DIGGING INTO CANADIAN SOILS: AN INTRODUCTION TO SOIL SCIENCE



Canadian Society of Soil Science
University of Saskatchewan

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Digging into Canadian Soils: An Introduction to Soil Science

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Licensing

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About the Authors

Digging Into Canadian Soils: An Introduction to Soil Science is comprised of 17 chapters written by 41 authors and co-authors. This textbook represents the ideas, inspiration and hard work of many members of the ***Canadian Society of Soil Science/Société Canadienne de Science du Sol***, and is dedicated to the soil scientists who came before us, and to those students who will continue the soil science journey after us.

Canadian Society of Soil Science
Société Canadienne de la Science du Sol

The Canadian Society of Soil Science is a non-governmental, non-profit organization that brings together researchers and scientists, engineers, technologists, administrators and students involved in the profession of soil science. The primary objectives of the Society are to:

- promote the wise use of soil for the benefit of society,
- facilitate information and technology exchange among people in the professional soil science community,
- promote research and the practical application of findings in soil science, and
- foster the integration of students into the professional soil science community.

The overarching mission of the Society is to "*nurture the discipline of soil science in Canada and ensure its relevance in the future*". What better way to nurture our discipline than to have a Canadian soil science textbook for the next generation of soil scientists?

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Written, reviewed and edited by members of the Canadian Society of Soil Science, *Digging into Canadian Soils: An Introduction to Soil Science* introduces students to the core disciplines of soil science (pedology, soil biology and microbiology, physics, chemistry, and fertility and nutrient cycling)—presenting the concepts and vocabulary needed by students just beginning their soil science journey.

The planning and development of this textbook were overseen by an editorial board consisting of: Maja Krzic, University of British Columbia; Fran Walley, University of Saskatchewan; Amanda Diochon, Lakehead University; Maxime C. Paré, University of Québec in Chicoutimi | UQAC; and Rich Farrell, University of Saskatchewan; and includes contributions by 41 authors and co-authors, and a small but dedicated team of resource people whose technical assistance was integral to the production of this textbook. Will Engle and Lucas Wright (UBC Centre for Teaching, Learning and Technology) and Heather Ross (Gwenna Moss Centre for Teaching and Learning at the University of Saskatchewan) provided invaluable advice and help throughout the various stages of eBook development. We would also like to acknowledge the anonymous reviewers whose expertise, dedication, and insights helped shape the content of this textbook.

A special thanks is extended to the *Canadian Society of Soil Science (CSSS)* and *BCcampus* for their financial support and to the CSSS membership at large for their enthusiastic encouragement and support. The French translation of this textbook (*Introduction à la science du sol : de la théorie à la pratique en sols canadiens*) would not have been possible without the financial support of l'*Association québécoise de spécialistes en sciences du sol (AQSS)* and the *University of Québec in Chicoutimi (UQAC)*. This was a true community effort and we hope you find that this textbook meets the needs of both the teachers and learners of soil science in Canada, and that it is both an informative and enjoyable read.

Maja Krzic, Fran Walley, Amanda Diochon, Maxime C. Paré, & Rich Farrell

Preface

Soil science has long been a core discipline in many universities and colleges across Canada. Traditionally, soil science as an academic discipline was linked to agriculture programmes, with many of these being among the first to be introduced at universities and colleges across Canada. Over time, however, soil science has spread far beyond its traditional roots and is now foundational in many programs including those in environmental science, geography, and environmental engineering to name a few. A recent Canadian study examining soil science courses in Canada identified 80 post-secondary academic units offering soil science courses (Diochon et al. 2017, Can. J. Soil Sci. 97: 122–132), including 63 offering introductory soil science courses (Krzic et al. 2018, Can. J. Soil Sci. 98: 343–356) with an estimated yearly enrolment of more than 3,000 students.

Several introductory soil science textbooks are available and commonly used as resources; however, until now, none of these textbooks specifically address the soils of Canada or use the Canadian System of Soil Classification when describing soils. The need for a Canadian textbook has long been discussed by members of the Soil Education Committee of the Canadian Society of Soil Science, and in 2018, a proposal to forge ahead with the development of an introductory soil science textbook was enthusiastically supported by the membership of the Canadian Society of Soil Science at its annual general meeting. This textbook is a result of those discussions. ***Digging Into Canadian Soils: An Introduction to Soil Science***, available in both English and French, provides an introduction to the core sub-disciplines of soil science, and introduces the concepts and vocabulary needed by students just beginning their soil science journey. Additionally, the textbook provides supplementary materials that are regionally specific, or may be of specific interest beyond what might be considered core soil science disciplinary material. Importantly, the textbook also is intended to introduce students to the Canadian System of Soil Classification using examples from across Canada, the world's second largest country by area, and to the Canadian Society of Soil Science, whose members share a common passion for soil science and are keen to share and instill this passion with students across the country and beyond.

Digging Into Canadian Soils: An Introduction to Soil Science is comprised of 17 chapters written by 41 authors and co-authors that cover the core sub-disciplines of soil science (eight chapters), five chapters that describe the distribution of soils across Canada, four special topics chapters, and a Glossary of common terminology used in soil science. Each chapter was overseen by an editor and reviewed by external reviewers. The book chapters are written at a level that assumes an introductory university level understanding of chemistry and biology.

This textbook represents the ideas, inspiration and hard work of many members of the ***Canadian Society of Soil Science/Société Canadienne de Science du Sol***, and is dedicated to the soil scientists who came before us, and to those students who will continue the soil science journey after us.

The Editors: Maja Krzic, Fran Walley, Amanda Diochon, Maxime C. Paré, and Rich Farrell

CHAPTER OVERVIEW

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1.1: Introduction

Fran Walley, Maja Krzic, Amanda Diochon, Maxime C. Paré, and Richard Farrell

Learning Objectives

On completion of this chapter, you will be able to:

1. Define soil as a physical entity
2. List and describe the four major categories of ecosystem services
3. Explain how soil and soil processes contribute to ecosystem services
4. Describe soil as natural capital and describe how natural capital may increase or decrease

THE SOIL BENEATH OUR FEET

Embarking on a study of soil starts with the contrasts that soils present. Soil is as common as the dirt beneath our feet. It sticks to our boots and tracks into our homes. We curse it when we get a vehicle stuck in the mud, or when it ruins a pair of shoes. Parents may be distraught when their child tastes a scoop of it. We are horrified when disaster strikes and a mudslide destroys everything in its path. Yet, soil also is the foundation to terrestrial life as we know it, and is amazingly complex and surprisingly rich with life. Soil impacts virtually every aspect of our lives. It provides a medium for production of food, fibre and bioproducts from energy biomass to fancy cosmetic products. It provides the physical foundation for buildings and roads. Since the very beginning of human civilization, soil has provided materials for our artistic expressions through paintings, sculptures and pottery. During antiquity, the Greek philosophers Empedocles and Aristotle recognized soil among the four main elements (i.e., “earth”, air, fire and water) used to explain the nature and complexity of all matter. Intrinsically linked to the global ecosystem, soil impacts and regulates climate, both emitting and sequestering greenhouse gases. Hydrological cycles are regulated by soil at local to global scales. Soil is a vast reservoir of biological diversity, replete with complex habitats. From this biodiversity, humankind has discovered and harnessed invaluable products including lifesaving antibiotics, microbial agents for the biocontrol of pests and plant diseases, and organisms capable of remediating polluted and contaminated areas. Soil has been recognized as the final frontier for human exploration (see *Soils – the Final Frontier*, 2004, Science, Vol. 304, Issue 5677), and holds the promise of untapped biological resources yet to be discovered. Did you know that when NASA lands on another planet or an asteroid they collect surface materials to search for evidence of life? We are connected to soil, physically, psychologically and spiritually. We know the soil of our homelands by sight and feel and smell, and we long to reconnect and to stay connected. Soil – humble and mighty, and right there, beneath our feet.

So how does one start the journey of discovering soil in all its many known facets? In this textbook, we introduce soils through the various sub-disciplines in which soil scientists have traditionally organized themselves. You will be introduced to soil genesis, soil organic matter, soil physics, soil chemistry, and you also will learn the inherent linkages between these areas when you delve into soil biodiversity and ecology, soil nutrient cycling and soil classification. You will travel virtually across Canada and learn about the dominant soil types and processes, and begin to further explore advances in digital soil mapping, soil micromorphology, and the importance of soil health and soil reclamation and remediation. We hope that as you study, you also will take the time to read about the Canadian soil scientists who have authored chapters in this textbook and hear about their journey as soil scientists – perhaps one of them is your instructor.

WHAT IS SOIL?

Defining soil is a bit tricky, largely because the definition may depend on your particular perspective. A farmer may view soil as the medium that supports plant growth, whereas an engineer may view it as the medium on which construction takes place—a foundation for roads or buildings, for example. These and other definitions of soil are further complicated by the complexity of soil itself. Soils are highly varied, at local, regional and global scales, so arriving at a single definition that applies equally to all soils is challenging indeed. The *Canadian Society of Soil Science* provides two definitions of soil (Gregorich et al., 2002):

1. The unconsolidated mineral or organic material on the immediate surface of the Earth that serves as a natural medium for the growth of land plants.
2. The natural, unconsolidated mineral or organic matter on the surface of the Earth that has been influenced by parent material, climate, macro- and micro-organisms, and relief, all acting over a period of time to produce a material different from which it

was derived in many physical, chemical, biological, morphological properties.

These definitions tell us a lot about soil . . . it is a natural substance and a living medium that may contain both mineral and organic materials and, uniquely, it is found at the uppermost surface of planet Earth. Martian soil, therefore, is not considered in this definition! The word “natural” also figures prominently, and the reference to macro- and micro-organisms tells us that soil is alive, and the living soil supports the growth of living plants and organisms. Indeed, without living organisms, soil ceases to function as a biological entity, and ultimately, could not support life on our planet. These definitions also reveal that soils have a “parent” and the parent materials ultimately contribute to the unconsolidated mineral or organic material that transforms into soil over time. Perhaps then, we need to know more about the parent material. We know our own parents have influenced our genetic make-up. Just as you may have blue eyes and curly hair like your mother and are tall like your father, soils also have characteristics akin to their parent material. These definitions also reveal that soil is a dynamic entity, ever changing and ever changed by the external environment. Indeed, humans have had, and continue to have, an enormous impact on soil, but it is important to note that soil continues to change with or without human action, and even the passage of time alone can bring about changes in a soil.

SOIL AND ECOSYSTEM SERVICES

Beyond defining soil as a physical entity as described above, soil also can be defined based on the *ecosystem services* it provides and supports. Ecosystem services are the benefits that humans derive directly and indirectly from ecosystems. Although the notion of ecosystem services has long been recognized by scientists, particularly those working in the area of sustainability science, the concept was formalized in a report of the *Millennium Ecosystem Assessment* (MEA) (MEA, 2005), an initiative conducted under the auspices of the United Nations, and called for by then Secretary-General Kofi Annan in 2000. The MEA report defined an ecosystem as “a dynamic complex of plant, animal and microorganism communities and the nonliving environment interacting as a functional unit” (2005, p. v). The report grouped ecosystem services into four main categories: (1) **provisioning services**; (2) **regulating services**; (3) **supporting services**; and (4) **cultural services**. Soils provide and contribute to many functions within ecosystems, and humans derive many benefits from soil, both directly and indirectly (Figure 1.1), all of which fall within one or more of these four main categories.

Provisioning services of soil are perhaps the easiest to recognize. Our relationship with soil is very direct: we grow more than 95% of our food, animal feed, fibre and, in some instances, our fuel in soil. Soils provide these goods and play a crucial role in ecosystem functioning (Adhikari and Hartemink, 2016). The sustained provision, or flow, of these goods or products from soil is dependent on the health of the soil and its ability to support plant growth now and for future generations. It follows that provisioning services are dependent both on the intrinsic chemical, physical, and biological properties of the soil as well as those properties that can be managed to improve or sustain the productivity of the soil (Dominati et al., 2010). Equally, soil degradation processes, such as soil erosion or soil salinization, can reduce soil quality and hence it’s capacity for provisioning services.

The **regulating services** that soils provide are many and easily overlooked. For example, we know that the levels of atmospheric carbon dioxide, an important greenhouse gas, are increasing. Soils can sequester, or retain, carbon from the atmosphere. Specifically, through the process of photosynthesis atmospheric carbon dioxide is transformed, or “fixed”, by plants into simple carbon molecules that are incorporated into structures within the plant. A similar transformation occurs within photosynthetically active soil microorganisms. These organic components, when returned to the soil, contribute to the soil organic matter thereby “sequestering” carbon that originated in the atmosphere as carbon dioxide. When this occurs the soil is effectively a sink for carbon but when this same organic matter decomposes soils also can be a source of carbon either as carbon dioxide or methane. The degree to which soils act as either a sink or source of these greenhouse gases can be controlled, to some degree, by appropriate management strategies such as minimum tillage practices. Similarly, the degree to which soils emit nitrous oxide, another very powerful greenhouse gas, is directly dependent on soil mineral nitrogen in the soil, which is a a manageable soil property in agroecosystems. Other regulating services provided by soils include water management and flood mitigation, biological control of pests and diseases, and detoxification of unwanted substances, to name a few. These examples of the regulating services that soils provide are not exhaustive, and other soil functions occurring at both the micro and macro scales—some of which have yet to be explicitly identified—are likely to fall within the category of regulating services.

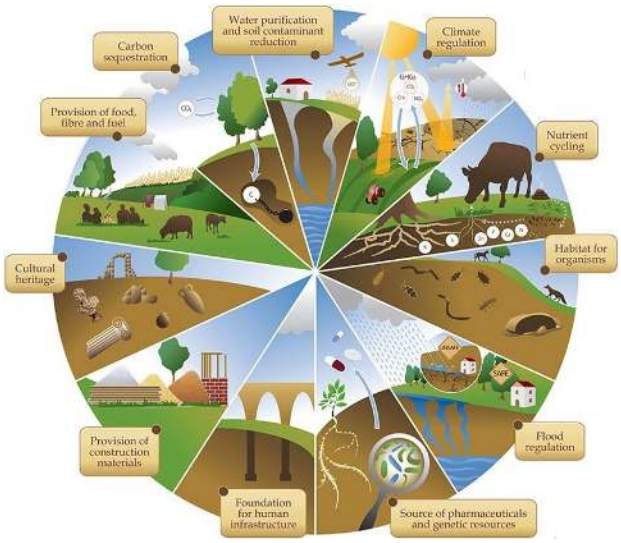


Figure 1.1 Schematic diagram representing some of the many functions that soils deliver. These functions contribute to ecosystems services, categorized as provisioning, regulating, supporting and cultural services. To which of these ecosystem service categories would you assign each of the above functions? © Food and Agriculture Organization of the United Nations. Reproduced with permission. Licensed under a CC BY-NC (Attribution NonCommercial) license.

Cultural services include the many non-material benefits derived from soils, such as spiritual enrichment, knowledge, sense of place, and connection to the land that soils provide. Many people and communities feel a deep connection to the land and the soil of their homeland. Clearly soil and the notion of soil matters at many levels. For many, the soil of their homeland represents them and who they are as a people—its colour or its smell after a rain, for example, reminds them of home and provides a sense of place.

Finally, **supporting services** provide support for the production of all other ecosystem services. Soil formation is an example of a supporting service that ultimately must take place to support all other ecosystems services. Other examples of supporting services include nutrient cycling, water cycling and soil biological activity, without which the other ecosystem services would not be achievable.

SOIL AS NATURAL CAPITAL – STOCKS AND FLOWS

The degree to which any individual soil contributes to the four categories of ecosystem services is dependent on inherent properties of the soil, and the degree to which these properties are expressed in a particular environment. For example, it is easy to recognize the provisioning services of soil—our food, feed, fibre, biomass and wood all grow in soil—but do all soils contribute equally to provisioning services? The answer, of course, is no. Some soils are inherently more fertile and thereby capable of providing greater crop yields, or are more capable of provisioning growth of forests for lumber production, or growth of biomass for use in energy production. Indeed, there are measurable chemical, physical and biological properties of soil that contribute to the ability of a soil to provide the different types of ecosystem services. In this sense, we can view soils through an economic lens and link the provision of ecosystem services to the perceived worth of a particular soil.

Linking the ecosystem services to economic concepts has led to the suggestion that soil can be viewed as natural capital or stock from which ecosystem services and useful goods flow (Dominati et al., 2010) (Figure 1.2). It follows that if the natural capital can grow, it also can be depleted. Moreover, certain soil properties contribute directly to the natural capital and either are inherently present and largely unchangeable or are changeable only very slowly (e.g., soil texture, presence or absence of stones, orientation and slope, etc.), or can be managed directly to enhance the natural capital. For example, organic matter is known to contribute significantly to the productivity potential of soil (i.e., its natural capital), and farmers can use minimum tillage and other management practices to enhance soil organic matter levels, thereby increasing the natural capital of the soil. Conversely, excessive tillage can lead to soil degradation and loss of soil organic matter, thereby reducing the natural capital of the soil.

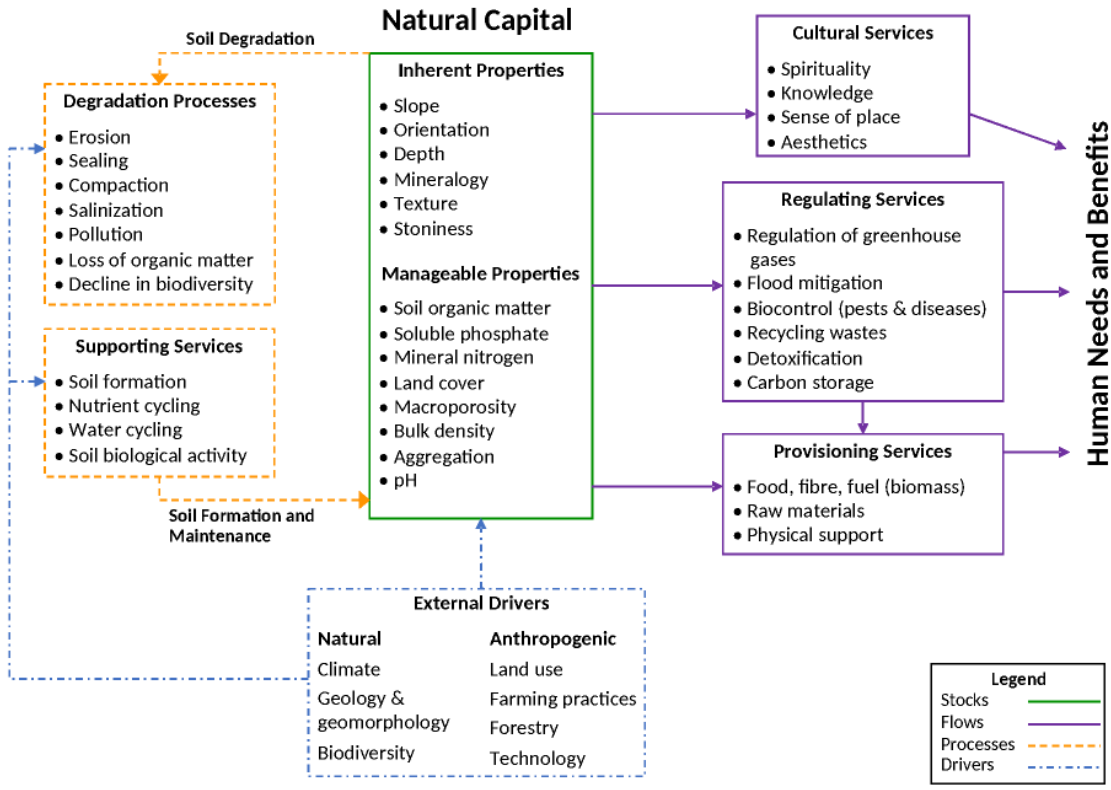


Figure 1.2. Framework demonstrating soil natural capital and flows of ecosystem services for the benefit of humans, as influenced by external drivers. The list of inherent and manageable properties contributing to the natural capital of soil represent examples of soil characteristics contributing to soil natural capital; the list is not exhaustive, nor are the lists of the ecosystem services that flow from the natural capital of soil. Adapted from Dominati et al. (2010); adapted by R. Farrell and licensed under a CC BY (Attribution) license.

Identifying and understanding inherent and manageable soil properties allows us to understand what ecosystem services a given soil can provide, and ultimately the magnitude of the flows of these services. External drivers such as climate change that can affect soil development and stability or anthropogenic drivers such as land use including soil “sealing” (i.e., urban sprawl or highways that seal the soil from other potential uses), farming practices, forestry, and other human activities that directly impact soils, all influence the soil’s natural capital; thus it is essential that we understand how these drivers can affect soil sustainability. The natural capital of soil is well worth preserving to ensure the sustainable flow of ecosystem services, and to meet human needs now and in the future.

In the following chapters, you will learn about the various physical, chemical and biological properties of soil, and the many interactions that occur between these properties that ultimately influence soil processes within ecosystems. As you learn more about the various soil properties in the following chapters, you will recognize that specific soil properties contribute to specific soil functions. Ultimately, identifying and understanding this connection can help us determine how best to manage soils to protect and sustain the desired functions from which humans derive enormous benefits. Our lives—and our livelihoods—depend on it!

BEYOND THE SCIENCE

As you read this textbook, and learn more about soils, you also will meet the soil scientists who have contributed to the creation of this book. You will learn about the many professions in which they work, as well as the vast fields of research to which they contribute. You also will learn that soil science is a passion for many, and that this passion goes far beyond the science and represents a lifelong journey of discovery and wonder for the most humble, and yet enormously complex ecosystem—soil.

A passion for soil has led to many descriptions of soil in the literature, including this from L.H. Bailey’s introduction to a very early but surprisingly relevant introductory soil science textbook published in 1909 “The Principles of Soil Management” (The MacMillam Company, Pennsylvania) (p. xxix-xxx):

"This marvelously thin layer of a few inches or a very few feet that the farmer knows as "the soil," supports all plants and all [humankind], and makes it possible for the globe to sustain a highly developed life. Beyond all calculation and all comprehension are the powers and the mysteries of this soft outer covering of the earth. We do not know that any vital forces pulsate from the great interior bulk of the earth. For all we know, the stupendous mass of materials of which the planet is composed is wholly dead; and only on the veriest surface does any nerve of life quicken it into a living sphere. And yet, from this attenuated layer have come numberless generations of giants of forests and of beasts, perhaps greater in their combined bulk than all the soils from which they have come; and back into this soil they go, until the great life principle catches up their disorganized units and builds them again into beings as complex as themselves."

Can You Dig It!



Dr. Dan Pennock at the international headquarters of the Food and Agriculture Organization (FAO), in Rome, Italy (licensed under a CC-BY-NC-ND license).

Did you know that the United Nations supports and promotes healthy soils?

The Food and Agriculture Organization (FAO) of the United Nations has a long history of promoting healthy soils as a basic prerequisite for the provision of essential ecosystem services, particularly as it relates to the global provision of food, fiber, fodder and other essential products. Indeed, the first World Soil Charter was introduced in 1981 at an FAO conference, and was heralded as a policy instrument to provide guidance to governments around the world for the management of soils—with a particular goal of addressing land degradation, which was identified as a major constraint to expanding agricultural production. More recently, the United Nations declared 2015 the International Year of Soils and endorsed a Revised World Soil Charter, which is widely available on-line (search World Soil Charter). The Revised World Soil Charter has an expanded goal of addressing new and emerging issues, such as soil pollution, climate change and urban sprawl. So what is the connection to the Canadian Society of Soil Science? Importantly, Canada is a member country of the United Nations, and has a Canadian representative on its scientific advisory committee, the Intergovernmental Technical Panel on Soils. In 2015, the Canadian representative was Dr. Dan Pennock, a Fellow of the Canadian Society of Soil of Science and a primary author of the 2015 Revised World Soil Charter. The charter outlines principles and guidelines for action by a range of international stakeholders to fight against soil degradation and to encourage sustainable soil management. Clearly, Canadian soil scientists are having a global impact!

THOUGHT EXERCISES

1. From your perspective (university student, farmer, future agronomist, global citizen, environmentalist, engineer, etc.), and in the context of ecosystem services, what is soil?
2. In addition to the soil functions listed in Figure 1.1, what are some other examples of functions that soils provide?

REFERENCES

Adhikari, K. and A.E. Hartemink. 2016. Linking soils to ecosystem services – A global review. Geoderma 262:101-111.

Bailey, L.H. 1909. Introduction. In L.H. Bailey (Ed.), The principles of soil management, The rural textbook series (pp. xxix-xxx). The MacMillan Company of Canada, Ltd., Toronto. 531 pp.

Dominati, E., M. Patterson and A. Mackay. 2010. A framework for classifying and quantifying the natural capital and ecosystem services of soils. Ecological Economics 69:1858-1868.

Gregorich, E.G., L.W. Turchenek, M.R. Carter, and D.A. Angers. 2002. Soil and Environmental Science Dictionary. CRC Press, Boca Raton, FL. 577 pp.

Millennium Ecosystem Assessment. 2005. Ecosystems and human well-being: Synthesis. Washington, DC: Island Press.

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Fran Walley (licensed under a CC-BY-NC-ND license)

After taking an introductory Soil Science course during my early university years, I worked as a summer assistant in the Department of Soil Science, University of Manitoba. I knew then that I had found my passion. It seemed to me then, as it does now, that soil science brought together everything that I loved. I love being outdoors, and soils research is all about being outside and in nature. I also love biology and understanding how the natural environment works. To me, soil science is applied biology, and I love to apply my knowledge of biology to soils. Over the years, soil science also has allowed me to grow as an environmentalist and contribute to sustaining our precious environment. What's not to love?

*

Maja Krzic, Associate Professor, Faculty of Land and Food Systems / Faculty of Forestry, University of British Columbia, Vancouver



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Growing up in Belgrade, former Yugoslavia (now Serbia), the only plants that I could encounter were those in parks and in pots at balconies of apartment buildings where we lived. Nevertheless, I was always intrigued with plants and decided to study agronomy to learn about them and their role in food production. I did not even know that there is such a discipline as soil science until I took a pedology course in the 2nd year of university and I was hooked! Being a soil scientist allowed me to work in a variety of ecosystems ranging from agricultural crop settings to grasslands, forests, urban brownfields and even golf courses in several countries around the world. However, the most rewarding part of my career are interactions with students and helping them learn about soil and its importance.

*

Amanda Diochon, Associate Professor, Department of Geology, Lakehead University, Thunder Bay



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When I was little you could often find me getting my hands dirty making mud-balls for a neighborhood mud battle, mud-stew for our imaginary village, and digging holes to reveal buried treasures. My very first independent science experiment was in middle

school when I tested the effects of soil pH and nutrients on plant growth for the science fair. It's a classic and I even placed! I still get to do all those things but with a different purpose and perspective. Now I get to make "soil balls" when discussing soil texture, "soil shakes" to measure changes in dissolved organic carbon in response to disturbances like forest harvesting and fire, and dig soil pits to expose the beautiful horizons beneath our feet and reveal the stories that the soil can tell us about its past, present and potential future. I may be older but I approach the soil with the same passion, wonder and excitement as that little girl. Perhaps you will too!

Maxime C. Paré, Professor, Université du Québec à Chicoutimi



Maxime Paré (licensed under a CC-BY-NC-ND license)

I became interested in soil science at the same time that I became involved in agriculture: as a child. I grew up in a small French-Irish town in Quebec (Saint-Ambroise-de-Kildare), and many of my friends' dads were dairy farmers. While helping them out during the haymaking season, I remember asking myself questions about basic yet important observations, such as, Why do some fields yield more and produce better quality hay than others? Soil pedogenesis and differences in soil quality quickly came to mind! Later, soil science became even more important during my undergraduate years at Université Laval (Quebec City). My first job as an undergraduate research assistant was working for Dr Léon-Étienne Parent, who undoubtedly propelled my interest for the discipline into the stratosphere! At that time, Professor Parent had numerous fertilization trials spread throughout the province, involving a multitude of crops. What I suspected as a child suddenly became real science; namely, these things were measurable, quantifiable, and replicable! Very exciting, no? Still now, every time I read a book or I review a paper, I learn something new about this discipline. This constant learning makes me revisit those feelings and questions I had as a young boy and keep the soil science fire burning inside me. Knowledge is my fuel and this book is a great opportunity to pass on the flame to the next generation!

*

Rich Farrell, Associate Professor, Department of Soil Science, University of Saskatchewan



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I am a native Rhode Islander and though I grew up in the suburbs, I have been interested in soil science since I was 13 when a friend of mine jokingly suggested doing a science project on soils because "there's nothing alive in soils" . . . . oh, was he wrong! So, I did my project on "Soils and Soil Conservation", and managed to win 2nd prize in the local science fair—though one of the judges said that the study of soils wasn't real science . . . . oh, was he wrong! So, here I am many years later still studying what I love. My research focus is on environmental agronomy—including the effects of agricultural practices on soil quality and greenhouse gas production/mitigation in agroecosystems. I am also co-director of the Prairie Environmental Agronomy Laboratory (PEARL), which supports research aimed at developing strategies that enhance or maintain agro-ecosystem capacity and integrity and establish the low GHG footprint of prairie cropping systems.

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1.2: Soil Genesis

Dan Pennock

Learning Objectives

After completing this chapter, the student will be able to:

- Explain why Canadian soils are young by global standards and the effect of their age on soil formation
- Describe the two main classes of soil parent materials at the national level
- Relate the two main genetic pathways for forest soils in Canada to the two broad classes of parent materials
- Identify the two main pathways for soil formation under water-saturated conditions in wetlands
- Explain the effect of frost-action in permafrost soils on soil formation
- Distinguish between grassland soils, forest soils, and wetland soils in terms of the effect of organic additions and the action of soil animals on soil formation
- Identify two distinct genetic pathways that occur in clay-rich parent materials in grasslands
- Explain the dominant pathway of soil formation in older, non-glaciated regions outside of Canada

INTRODUCTION TO SOIL GENESIS (PEDOLOGY)

A key goal of sustainable soil management is to match the practices we adopt with the specific properties of the soil. This can be difficult, however, when soils show major differences in their properties over very short distances. These differences in soil properties result from both properties they have inherited and the action of soil-forming processes during soil development. Both of these factors are studied in the sub-discipline of soil genesis or pedology, which is the focus of this chapter.

Soil genesis is underlain by two major concepts:

1. soil properties result from the action and interaction of soil-forming (or pedogenic) processes; and
2. the type and intensity of these pedogenic processes are governed by environmental factors – the conditions of the natural world in which the soils occur and the impact of human activity on these environmental factors.

Soil formation starts when fresh earth materials begin to be transformed by physical, chemical, and/or organic pedogenic processes. For the great majority of the Canadian land mass this moment of transformation began soon after the retreat of the great ice sheets that once covered most of Canada and the draining of the vast lake and river systems that followed the retreating ice front. The fresh glacial sediments were soon home to an assemblage of microbial, plant, and animal life that began to transform the glacial sediments into soil. In some places the landscape was then further modified by processes such as flowing water, wind, and gravity throughout the post-glacial period. The final wide-spread transformation of soils began with the European settlement of Canada and the imposition of their land uses on the soil.

This brief overview of soil formation in Canada establishes two key points needed to understand Canadian soils. First, Canadian soils are very young compared to most global soils (Sanborn, 2016). Post-glacial soil formation in Canada begins at about 17,000 years before present (yrs BP) at about the 49th Parallel (which is now our southern border with the United States of America). In comparison, in the center of the continents of Australia and Africa soil formation has been occurring since the start of the Mesozoic era 250 million years ago (Paton et al., 1995). Even compared to the southern areas of the USA or Europe, Canadian soils are very young.

In the ancient land surfaces of Australia or Africa the original rocks and sediments the soils formed in have been fundamentally altered by soil formation. In most Canadian soils, however, there has been little alteration of the rocks and sediments the soil formed in and hence many soil properties are directly inherited, as is the shape of the land surface itself. In soil science, the rocks and sediments the soil forms in are referred to as the *parent material* of the soil, and for Canadian soils inheritance of physical and chemical properties from the parent material is the major determinant of soil formation.

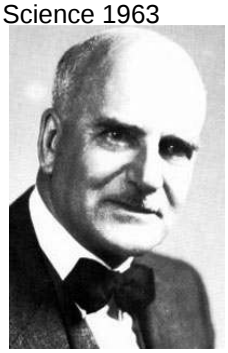
Second, in Canada the characteristic assemblages (or *biomes*) of climate, vegetation, and animals– for example, the tundra, Boreal forest, or grasslands – have largely been in place since soon after the ice sheets retreated. Most Canadian soils have developed in a stable biome and the soil properties reflect the distinctive set of soil processes associated with that biome. Hence in Canada we

have characteristic zones of soils that are distinct and different from one another, resulting from these distinct and different sets of processes. In more ancient land surfaces the soils may reflect many episodes of environmental change, and the soil properties often reflect long-vanished environmental conditions.

Hence, at the national level, the *soil formation factors* that control the pedogenic processes are **time** (the duration of soil formation), **parent material** (the size and chemical composition of mineral material), **climate** (temperature and moisture within the soil), and **organisms** (vegetation, animals, and microbes that transform the mineral material). As we will see, at more local levels two other factors also control specific pedogenic processes: **topography** (which controls flows of water and sediments) and the presence of **groundwater**. Finally all soil properties can be profoundly influenced by **human activity**. These seven soil formation factors are examined in greater detail in Chapter 8 of this textbook.

Can You Dig It!

J.H. Ellis, Fellow of the Canadian Society of Soil Science 1963



Joseph Henry Ellis. Photograph courtesy of the Canadian Society of Soil Science and licensed under a CC BY-NC-ND (Attribution NonCommercial NoDerivatives) license

Canadians on the whole generally tend to be rather humble people and are not known for great pronouncements of patriotism or promoting of our own. It is for this reason that very few know of the "groundbreaking" work of a Canadian soil scientist named Joseph Henry Ellis (1890-1973) whose work and musings greatly advanced our understanding of soils of Western Canada—and particularly of Manitoba. In his book *The Soils of Manitoba* (published in 1938) he described his understanding of why the soils in Manitoba were so diverse and distributed as they are:

"Soils are natural objects which have developed at the surface of the earth as the result of the action of climate on the geological deposits that are under the influence of organic life. Soils differ from the geological material over which they lie. These differences are shown by the physical and chemical characteristics of the soil which may be enumerated under the headings of colour, texture, structure, consistency, intrusions and concretions or products of soil weathering, reaction, etc. The sum of these characteristics determines the soil type (soil group)."

Professor Ellis went on to describe the factors that control the nature and distribution of soils in Manitoba (p 14). He listed these factors as:

1. *"The climate, or the temperature and moisture within the soil";*
2. *"The vegetation, which determines the type of organic matter added to the soil";*
3. *"The parent material, or the geological deposits which determine the minerals on which the soil is formed, and in turn affect the texture, the water retention capacity, and the mineral reserve";*
4. *"The position in which the soil is found in relationship to the topography";*
5. *"The presence or absence of ground water within the soil profile";*
6. *"The age or length of time the soil has been under the influence of its environment"; and*
7. *"In the case of cultivated soils – the modifying effects of culture or the work of man."*

A contemporary of Professor Ellis in the United States, Professor Hans Jenny (1899-1992) of the University of California at Berkeley, published a work in 1941 called *Factors of Soil Formation: A quantitative system of pedology* (Jenny, 1941) in which he identified what he originally called Factors of Soil Formation—later referred to as State Factors (Jenny, 1980). He identified climate, organisms, relief (topography), parent material, and time as the five factors of soil formation that explained the distribution

and nature of soils. Both Hans Jenny in the USA and Joe Ellis in Canada had identified similar factors to explain the distribution and nature of soils.

The above excerpt was adapted from an article originally published by Les Fuller in 2010 (http://www.prairiesoilsandcrops.ca): Chernozemic soils of the prairie region of western Canada. Prairie Soils and Crops 3: 37−45.

THE GEOLOGICAL LEGACY

The great majority of Canadian soils have formed in glacial or post-glacial sediments and often the bedrock occurs well below the zone of soil formation. The glacial and post-glacial sediments were, however, derived from bedrock pulverized by the glaciers and hence the nature of the pulverized bedrock is a very important control on the genetic pathway the soil follows. The minerals present in the parent material initially determine the chemical environment of the soil, especially its acidity and base cation concentration (see Chapter 5). As well, the particle size, or texture (see Chapter 4) of the mineral material is also originally inherited from the pulverized bedrock and then is further sorted by other geomorphic agents such as flowing water or wind. The Virtual Soil Science Learning Resources website has an excellent overview of the importance of parent material.

At the broadest level there are three major classes of rocks in Canada (Figure 2.1): igneous, sedimentary, and metamorphic. Igneous rocks form within the crust (referred to as intrusive or plutonic rocks on Figure 2.1) or through volcanic activity (volcanic rocks). Sedimentary rocks form from sediments or chemical compounds that were eroded and deposited in ancient oceans or land surfaces and then through time subsequently made into rock (or lithified). Metamorphic rocks are formed through transformation of existing igneous or sedimentary rocks by temperature or pressure or by chemical factors and their properties reflect the parent rock (i.e., igneous or sedimentary) from which they are formed.

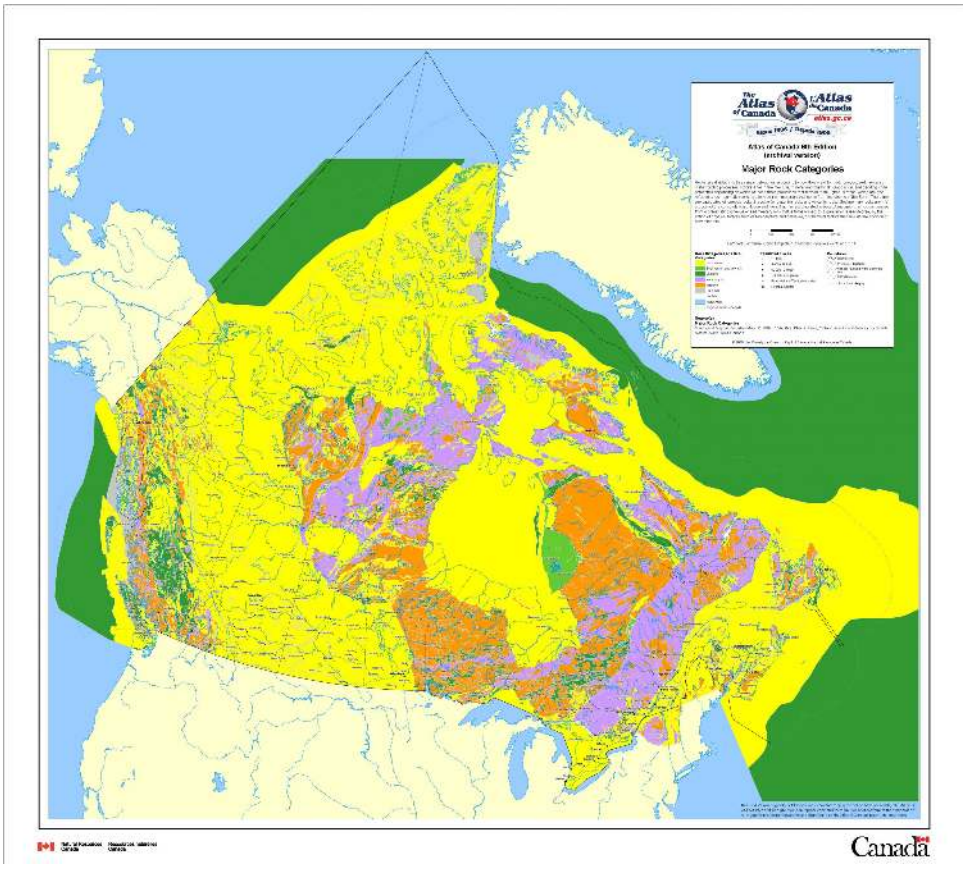


Figure 2.1. Major rock types in Canada (adapted from the National Atlas of Canada). Sedimentary rocks are shown in yellow. Orange (igneous) and purple (metamorphized igneous rocks) areas in Central and eastern Canada are the Canadian Shield. Volcanic (dark green), igneous and sedimentary rocks form a complex pattern in the western Cordillera. © Natural Resources Canada; used with permission and licensed under the Open Government Licence – Canada. https://open.canada.ca/en/open-government-licence-canada

Igneous rocks are dominated by silicate (i.e., oxygen-silicon dominated) minerals such as micas, feldspars, quartz, olivine, pyroxene, and amphiboles, all of which are formed through the cooling and solidification of magma. The cooling process in the earth's crust typically produce minerals of large, sand-sized particles and the resulting rock mass is very resistant to crushing and pulverization by glaciers. The greatest extent of igneous rocks is associated with the Canadian Shield, which is one of the most ancient rock bodies in the Earth. There are also extensive areas of igneous rocks in the interior and west coast of British Columbia.

Sedimentary rocks form through deposition of material (particles and chemical ions) eroded from ancient landscapes. Sedimentary rocks are dominant in the Prairies, eastern Rocky Mountains, and through to the Arctic Ocean and Arctic islands. They are also dominant in southern Ontario and Quebec and much of the Maritime provinces.

Sedimentary rocks can be broadly divided into three main classes: clastic rocks (formed from consolidation of preexisting particles such as sand, silt or clay); biochemical (or biogenic, meaning formed through the actions of living organisms) rocks such as limestone or coal; and chemical rocks formed by precipitation of ions to form salts such as halite (rock salt) The sediments that form sedimentary rocks were commonly deposited in ocean (or marine) environments. The sediments that form sedimentary rocks were originally deposited in layers and hence the sedimentary rocks are bedded. Their bedded nature and the structure of the minerals present in them make them more susceptible than igneous rocks to breakup and pulverization by glaciers.

The sediments that form clastic rocks are sorted into sand- (sandstone), silt- (siltstone), or clay-sized (shale, mudstone) particles. Clay minerals in shales are silicates with a distinctive layer or sheet structure (called phyllosilicates) that form in the near-surface zone of the earth at much lower pressures and temperatures than the minerals in igneous rocks. Clay minerals in Canadian soils are almost entirely inherited from these parent rocks. For pedogenetic purposes there is an important distinction between clay minerals such as smectite, which allows water between its structural layers and hence swells and shrinks as soil moisture conditions change, and other major minerals such as mica, vermiculite and chlorite that do not shrink and swell. Smectite clay minerals are in large part inherited from the Cretaceous marine sedimentary rocks that underlie the Prairie ecozone (see Chapter 14 for more information on minerals in Canadian soils).

Limestone is the most common biogenic rock and the dominant mineral in limestone is calcium carbonate. Chemical rocks often contain a wide variety of salts formed between cations (positively charged ions) such as sodium and magnesium and anions (negatively charged ions) such as carbonate and chloride.

The differences in mineral composition between the igneous and sedimentary rocks strongly influences the chemistry of the soil and, at a very broad level, determines the genetic pathway the soil will follow. First, the presence of carbonates in parent materials formed from sedimentary rocks maintains the soil pH in a range between 5 to 8.5, whereas soils formed in parent materials derived from igneous rocks are typically more acidic (pH levels from 3.5 to 5). As we will see in Chapter 8, this threshold for pH is an important classification criterion in the Canadian System of Soil Classification (Soil Classification Working Group, 1998). Second, base cations such as calcium, magnesium, potassium, and sodium are plant nutrients (see Chapter 7), and the higher content of these cations in sedimentary rock-derived parent materials makes them inherently more fertile than the more acidic, igneous-rock derived soils.

THE GLACIAL LEGACY

The Canadian land surface was greatly changed by the ice sheets of the Pleistocene Epoch, which lasted from 2.5 million years ago to approximately 11,700 years ago (Subcommission on Quaternary Stratigraphy). The Pleistocene Epoch was followed by the Holocene Epoch, which extends from the end of the Pleistocene to the present day. During the Pleistocene Epoch there were many episodes of glacial ice formation, advance, and retreat but for soil science in Canada we are most concerned with the final glaciation. In North America this final episode is informally called the Wisconsinan glaciation.

Glacial sediments

The great ice sheets are responsible for the majority of sediments that Canadian soils have formed in and for the shape of the land surface itself (Figure 2.2). The texture and mineralogy of glacial sediments result from the pulverization of the underlying bedrock surface by the ice sheets; the ice sheets break up the bedrock and then grind it into smaller and smaller pieces during transport. The two main types of rocks differ greatly in the resistance they offer to the ice sheets. Igneous rock surfaces are generally much more resistant to disturbance by glaciers, and yield sediments high in gravel and sand-sized particles dominantly of quartz. Sedimentary rocks are much more readily pulverized, and yield a mixture of particles sizes from gravel to clay.

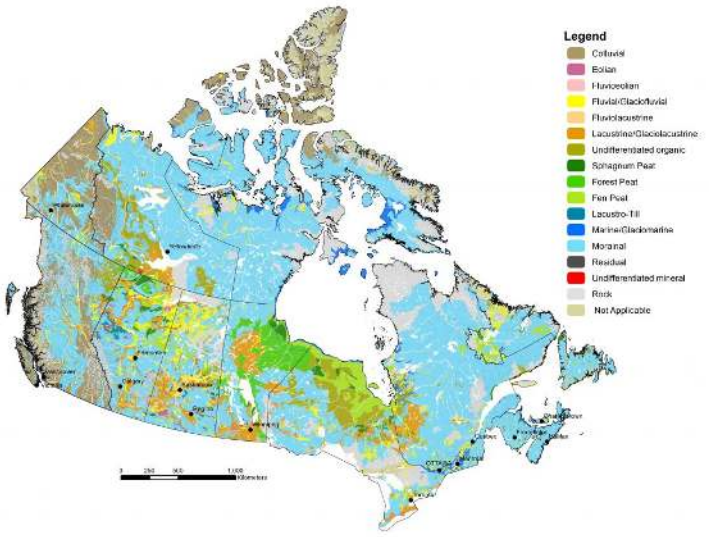


Figure 2.2. Map of parent materials for Canadian soils. Created from the 1:1,000,000 Soil Landscapes of Canada. © Darrel Cerkowniak, Agriculture and Agri-Food Canada; licensed under a CC BY (Attribution) license.

Initially the pulverized rock was transported by the great ice sheets and considerable mixing of crushed rock from different sources occurred. This mixed sediment is called till or moraine, which are synonymous terms. Morainal parent materials are by far the most common in Canada (Figure 2.2). In areas underlain by igneous rock, till is dominated by gravel and sand-sized material (Sims and Baldwin, 1991); in sedimentary rock areas, tills are more commonly a mixture of gravel, sand, silt, and clay sized particles. Till can be deposited directly as a basal layer under the ice, as material which melts out from a stagnant ice mass, or as material which flows off the surface of a stagnant ice mass.

At the end of the Wisconsinan glaciation the ice mass began to melt faster than new ice formed, and the front of the glacier began to recede (Dyke, 2004). The period of glacial retreat produced vast amounts of melt water, which carved channels into the underlying till and bedrock surfaces. The melt-water from the glaciers carried a high sediment load, and this material was deposited in huge floodplains and deltas. River or stream systems are generally called fluvial systems and deposits from glacial rivers are called glacio-fluvial sediments. Glacio-fluvial sediments are most common in the Prairie provinces but can be found throughout Canada (Figure 2.2).

In some locations, the glacial melt water was impounded in glacial lakes in front of or even on top of the ice sheets. Deltas of coarse sediments formed at the point where rivers entered the glacial lakes, and the reduced flow of the water could no longer suspend these larger materials. Fine sand, silt, and clay sediments were sequentially deposited in the relatively still waters further out in the lakes. Lake systems are called lacustrine systems and deposits from glacial lakes are called glacio-lacustrine sediments. The most extensive areas of glacio-lacustrine deposits occur in Manitoba, the Clay Belt of Ontario and Quebec, and around Great Slave Lake in the North-West Territories (Figure 2.2). Smaller glacio-lacustrine surfaces occur in southern Saskatchewan and Alberta and the high silt and clay contents make for excellent agricultural soils.

At the continental margins the land surface rebounded from the compression caused by the great ice sheets and near-shore areas of marine sediments were exposed to the atmosphere and soil formation; these are called marine or glacio-marine sediments. Flooding of the St. Lawrence valley by the ocean left deposits of glacio-marine clays in eastern parts of the Mixedwood Plains ecozone around Ottawa and Montreal. The Hudson Plain ecozone south of Hudson Bay has glacio-marine sediments overlain by deposits of peat formed in the post-glacial period.

In some locations glacial erosion scoured away any sediments, and left behind exposed rock surfaces (Figure 2.2). These exposed rock surfaces are most common in the Canadian Shield. Rock is also the surface material in many parts of the mountains of western Canada. Due to the short time for soil development in Canada little or no alteration of the exposed rock surface has occurred.

Glacial landscapes

While the largest terrain features such as the western Cordillera, Appalachian Mountains, and the High Arctic mountains are the result of mountain-building episodes triggered by plate tectonics, the shape of the landscape elsewhere in Canada is primarily the

result of glacial erosion and deposition.

At the broadest scale there are differences in glacially-induced surface forms between areas underlain by igneous rocks and those of sedimentary rock. The resistant, igneous rocks yield limited amounts of tills high in sand and gravel, and tills are generally thin and the rolling underlying bedrock surface is evident. There are also many landforms associated with flowing water at the front or within the ice sheet such as outwash plains and eskers (i.e., sediments from rivers flowing within the ice mass).

In the regions underlain by sedimentary rocks, the pulverization of the rocks yielded much greater quantities of glacial sediment and often there is little expression of the underlying bedrock surface. For example, in the Prairies there is a band of thick (100 to 300 m) of glacial sediments that runs parallel to the Canadian Shield and generally gets thinner as we move to the south-west of the plains (Fenton et al., 1994). In Southern Ontario, the glacial sediments are thickest (200 – 260 m) at the contact with the Canadian Shield and then again in Lake Erie Lowland Ecoregion. In areas with thick glacial deposits a range of surface forms associated with specific types of glacial deposition occur (Table 2.1).

Table 2.1. Materials and surface appearance of major glacial sediments

Sediment Type	Material Description	Surface Appearance
Till	Till is deposited directly from glacial ice, and is composed of an unsorted mixture of sand, silt, clay, and gravel. The gravel is distributed throughout the material. The dominant soil textures are loams.	Till landscapes are often hummocky (knolls and sloughs) or undulating. They commonly have gravel on the soil surface.
Glacio-fluvial gravels	The channels of the swiftly flowing glacial streams were composed of well-sorted sand and gravel layers.	Relatively flat landscapes; former bars in the rivers may form low, flat topped hills.
Glacio-fluvial sands	The floodplains of the former glacial streams and the deltas where the streams entered water bodies are often composed of well-sorted sand deposits (sand, loamy sand, or sandy loam). The size of the sand ranges from coarse sand to very fine sand.	The surface expression is often undulating; few stones are apparent.
Glacio-lacustrine silts and clays	The still waters of the glacial lakes allowed the fine silt and clay deposits to settle out of the water and be deposited. These layers may be well-sorted silt (silt or silt loam), clay (heavy clay, clay, or clay loam), or alternating thin layers of silt and clay (called varves).	Generally these are flat landscapes with few hills. Stones are very uncommon. In a few places, the lakes were formed on top of the ice sheets; these glacial deposits have a hummocky appearance.

The Holocene Legacy

Many areas of the Canadian land surface have been further modified by geomorphic processes since glaciation, through the actions of wind, water, gravity, or saturation (Table 2.2) during the Holocene.

Table 2.2. Materials and surface appearance of major post-glacial sediments in Canada

Sediment Type	Material Description	Surface Appearance
Loess	A layer which was transported and deposited by the wind, termed ‘eolian’. The loess forms a silty mantle over top of the underlying glacial sediments	Generally very smooth landscapes with long, gentle slopes
Dune sand or cover sand	Often the glacio-fluvial sands originally deposited in deltas or shorelines that have been reworked by the wind during the Holocene.	Complex series of slopes characteristic of sand dune areas.
Fluvial or alluvial	Flooding of the current fluvial channels leaves behind deposits of sand, silt, and clay in a floodplain.	Confined to the floodplains of the fluvial systems; generally level landscapes or gently undulating.
Colluvium	Transport of sediment from steep slopes of river channels, escarpments etc.; deposition commonly occurs at the base of the slope.	Associated with steep slopes.
Peat	Organic layer formed by incomplete decomposition of plant inputs due to saturated conditions	Generally level wetland surface occupied by trees, shrubs, and/or mosses
Marine	Shoreline and coastal sediments exposed due to land surface rebound following deglaciation	Generally level surfaces with raised beach ridges

Wind can be an important geomorphic agent, especially for lacustrine and fluvial sediments dominated by silt and sand. Geomorphologists refer to wind action as eolian processes. Immediately following glaciation wind action on floodplains and lakes detached silt from the surface and deposited silt layers (or loess) downwind of the source area. Loess areas are found throughout the globe and are often very important agricultural areas, but only limited areas of post-glacial loess occur in Canada – the only large area mapped on the parent material map of Canada is in southwestern Saskatchewan. Thin (10 to 30 cm) loess caps are very common in alpine and many Arctic landscapes as well but are too discontinuous to appear on the national map.

Glacial deltas and floodplains were also sources of sand-sized particles that were eroded by wind and then deposited in characteristic sand dune areas. Some areas like the Great Sand Hills of Saskatchewan remain active to this day. Sand dune areas are also commonly found in the shorelines of modern lakes and oceans throughout Canada.

Water is also a major geomorphic agent and is responsible for landform creation in modern floodplains throughout Canada. Like wind, water transports and separates sediments into different size fractions, and floodplains often have discrete areas of gravel, sand, silt, and clay deposition. Often floodplains are built up over time by different flood events and distinct layers of different sediments occur; these are referred to as cumulic layers by soil scientists (Figure 2.3).



Figure 2.3. Cumulic sand and gravel deposits in a river floodplain. © Dan Pennock, Univ. of Saskatchewan (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

Wherever relatively steep slopes occur sediments are transported downslope by gravity in a process known as colluviation. The texture of the colluvial deposits reflect the properties of the source sediment. In mountain areas like the Western Cordillera colluvial transport of till mantles is very common, and these colluvial layers often consist of loams with a high stone content.

Finally, one of the largest areas of Holocene-aged soils is the Hudson Plains Ecozone, which was exposed to the atmosphere and soil formation as the ocean receded due to the rebound of the land surface following melting of the great ice sheets. The recession of the ocean is marked by a series of raised beach ridges that mark the former shoreline of Hudson Bay.

Paleosols and climate change

Paleosols (ancient soils) are soils that have formed under climate and vegetation regimes very different from those that currently occur in an area. The properties of paleosols can be used to reconstruct past climate and are an important tool in the paleoclimatic research.

Only a few areas of Canada were not glaciated and the soils of these unglaciated regions reflect soil formation over much longer time periods than most Canadian soils (Sanborn, 2016). The west-central area of the Yukon was not glaciated, and the paleosols found in this area are more fully developed than the soils found adjacent to them in glaciated regions. The paleosols often have a mantle of loess over the older soil, and ample evidence of mixing caused by frost processes in the soil (Figure 2.4) (Sanborn, 2016). The other area with unglaciated soils is the Cypress Hills of southwestern Saskatchewan, which were not overridden by glaciers. Again the paleosols of the Cypress Hills are much more highly developed than adjacent soils and also have a layer of loess capping the older soil layers (Figure 2.5) (Sanborn, 2016).

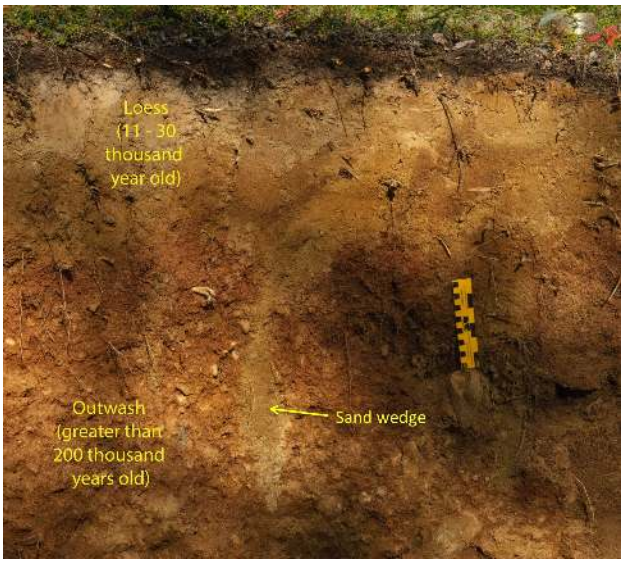


Figure 2.4. Yukon paleosol. The lower glacial outwash deposits have a clay-rich, reddish soil horizon indicating a long period of soil formation and soil weathering. The upper wind-blown loess deposit is much younger and is less weathered. The sand wedge into the outwash deposit is evidence of a period of intensive frost action that affected the lower soil layer. Photo courtesy of Paul Sanborn. © Paul Sanborn, UNBC; licensed under a CC BY (Attribution) license.



Figure 2.5. Cypress Hills (southern Saskatchewan) paleosol. The lowest layer is a weathered conglomerate rock dating from the pre-glacial Tertiary period and has an ancient clay-rich layer formed at its surface . The layer overlying the conglomerate has been subjected to intensive frost action (cryoturbation) during the glacial periods and then capped with a wind-blown loess layer during the Holocene period. The surface soil is a typical grassland soil formed in the past 11,000 years. Photo courtesy of Rene Barendregt and Paul Sanborn. © Rene Barendregt (Univ. of Lethbridge) and Paul Sanborn (UNBC); licensed under a CC BY (Attribution) license.

Paleosols are also found where land surface instability during the Holocene has lead to burial of land surfaces and the soils that formed on them. Buried paleosols can be found throughout Canada in fluvial and eolian deposits . They are also common in colluvial sediments (Figure 2.6), especially in mountain areas and in colluvial layers in sloping landscapes in the grasslands of the Prairie Ecoregion.



Figure 2.6. Paleosols in colluvial deposits, Highwood River valley, Alberta. The sediments on this river terrace include both river-laid sediments and sediments from the surrounding slopes. The slope (or colluvial) sediments have several organically enriched layers within them. © Dan Pennock, Univ. of Saskatchewan (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

Can You Dig It!

Where's the oldest soil in Canada?

The overwhelming majority of soils in Canada are (by global standards) very young since they have formed since the retreat of the great ice sheets began about 17,000 years ago. A few locations, however, emerged from the ice sheets much earlier than others and some points escaped glaciation altogether. The Yukon paleosol shown in Figure 2.4 formed on surfaces that were ice-free from about 200,000 years ago to the present day. The Cypress Hills paleosol in Figure 2.5 may be even older – these soils are found on a 300 km2 plateau in southwestern Saskatchewan and southeastern Alberta that was not overridden by glaciers at any point during the great ice ages. Some high-elevation buttes in southwestern Alberta also have very old soils – the soils and paleosols at Mokowan Butte in south-western Alberta have been extensively studied and the youngest soil here developed between 500,000 and 65,000 years old. The oldest paleosol at Mokowan Butte dates back to before 720,000 years ago. In his 2016 paper on the imprint of time on Canadian soils, Paul Sanborn notes that a large area in western and northern Yukon was never glaciated and presumably holds an array of very old soils but that most of the area remains to be explored by pedologists.

OVERVIEW OF SOIL-FORMING PROCESSES

The study of soil formation and classification is called **Soil Genesis or Pedology**.

Specific groups or classes of soils result from the action and interaction of the environmental conditions, and that these soil horizons will differ from place to place within a field and within a region or country. Hence, the distribution of soil horizons and soil properties is not random – they are responding to a set of environmental controls. The classification of soils in Canada is covered in Chapter 8.

Soil formation begins when freshly exposed sediments or rocks at the surface of the earth begin to be altered by various organic, chemical, and physical processes. Initially the alterations caused by these soil-forming agents are very slight, but through time they significantly alter the nature of the original materials. In some cases the minerals themselves are altered to new forms but in almost every case the physical organization (or **structure**) of the materials, both mineral and organic, is changed by soil formation. Soil formation creates new soil structural units ranging from small granules to large clods of soil, and these structural units are critical for air and water movement in the soil (soil physics reference).

For many (but not all) soils in Canada horizontal layers that parallel the land surface begin to develop in the soil (Figure 2.7). These layers have distinct sets of soil forming processes associated with them, and are called horizons by soil scientists. The set of horizons exposed when a soil pit is dug is called the soil profile. The most readily recognized horizons are those with layers that have contrasting colours due to the action of soil-forming processes, but in other soils the differences between horizons are more

subtle. In some Canadian soils the processes act to mix the soil material and hence distinct horizons do not form but these soils are far less common than those with distinct horizons. As we will see in Chapter 8 different horizons are assigned letters in classification systems such as the Canadian System of Soil Classification. In this chapter we will refer to them more generically according to how they are formed rather than the formal conventions of the classification system.



Figure 2.7. Example of soil horizons along a hillslope in southern Saskatchewan. The soil horizon thicken and the number of horizons increases downslope due to higher soil moisture levels in lower slope positions. © Don Acton (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

The **soil-forming processes** that create soils are bundled into four major groups (Figures 2.8 and 2.9): additions, losses, transfers, and transformations. The Virtual Soil Processes Project (Krzic et al. 2008) has an excellent overview video and website for the four soil processes.

Early microbial colonizers add carbon and nitrogen to a developing soil, largely at or very near the soil surface. As plants begin to colonize, addition of organic matter (OM) occurs both at the soil surface (when above-ground OM such as leaves and branches fall to the soil surface) and within the upper part of the soil (from below-ground OM such as roots and soil organisms of all sizes). Additions of water (and the ions and chemical compounds within the water, termed **solutes**) occur from precipitation falling on the surface, within the soil as laterally flowing soil water, and from groundwater moving from depth to the soil surface (Figure 2.8). Finally soil from elsewhere that has been eroded by water, wind, or tillage erosion can be deposited on the soil surface.

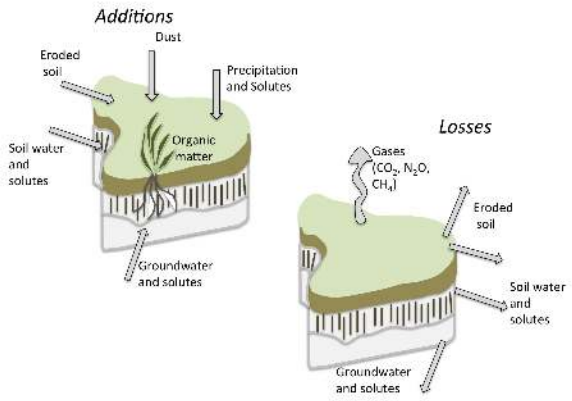


Figure 2.8. Additions and Losses. © Dan Pennock, Univ. of Saskatchewan; licensed under a CC BY (Attribution) license.

A significant fraction of the OM added to soils is lost as gases (especially carbon dioxide [CO_2]) back to the atmosphere after decomposition of the organic material by soil organisms (Figure 2.8). Other losses include water and solutes lost through water flow within the soil and from the soil to the groundwater. The complete loss of solutes from the profile is termed **leaching**. As well as dissolved mineral ions, very fine OM can be leached from the soil into the underlying materials. Finally, in some situations soil is lost from the soil surface by water, wind, and tillage erosion.

Transfers are processes that occur within the soil rather than between parts of the landscape (Figure 2.9). In soil genesis we distinguish between transfer processes that remove material from a horizon (termed eluviation processes) and those that lead to

gain of material in a horizon via deposition (termed illuviation processes).

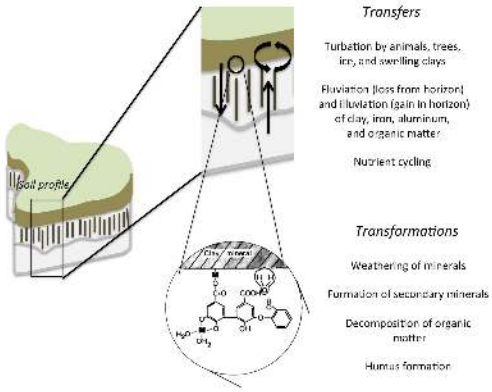


Figure 2.9. Transfers and Transformations. © Dan Pennock, Univ. of Saskatchewan; licensed under a CC BY (Attribution) license.

Transfers of solutes and solid particles occur as soil water flows between different layers of the soil. Very fine clay particles can be removed from upper soil layers and deposited deeper in the soil. In some soils, complexes between iron and aluminum and OM form and are also transported vertically within the soil and deposited at depth. Other transfers follow an initial stage of transformation, where (for example) ions are released from a mineral in an upper soil layer. In some cases, these ions are completely lost from the soil (in the process of leaching) but in others reformation of new minerals (called secondary soil minerals) occurs deeper in the soil.

Mixing (also called turbation) between layers of the soil is another type of transfer and occurs due to both organic and physical agents. Many soil organisms such as earthworms and mammals that reside in the soil physically displace soil from one location to another, as do processes such as the toppling of trees in the forest (Figure 2.10). Mixing also occurs through physical processes such as the shrinking and swelling of certain clay minerals and the freezing and thawing that occurs in almost all Canadian soils.

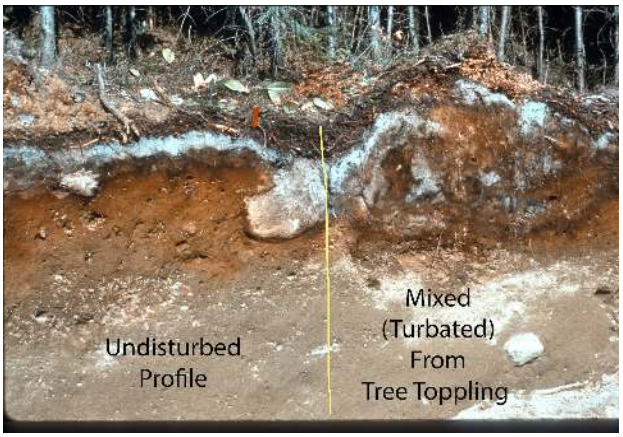


Figure 2.10. Example of turbation by trees. © Saskatchewan Center of Soil Research; licensed under a CC BY (Attribution) license.

Transformations of mineral and organic materials within the soil are very important for both soil genesis and for plant nutrition (Figure 2.9). The release of ions from minerals or the complete transformation of mineral particles from their initial form to new, secondary minerals is generally called weathering. Weathering can occur from physical processes such as freezing and thawing of soils or from chemical processes as discussed below.

There is a very complex set of pathways by which fresh OM added to the soil is transformed by soil micro- and macro-organisms into a range of new products, which is generally termed decomposition (see section SOM). In well-drained soils (i.e., where the soil pore space is normally filled with air), decomposition creates highly transformed, reactive organic complexes called humus. In poorly drained conditions (i.e., where the pore space is often or completely water filled) decomposition is retarded or absent, and organic materials instead accumulate and create layers of peat.

WEATHERING

Weathering is an umbrella term that refers to both the chemical alteration of soil minerals and physical disruption of minerals, which leads to smaller sized soil particles. Physical weathering in Canadian soils is limited due to their relative youth; hence, the size of soil mineral particles is inherited from the parent materials

The dominant control of chemical weathering is the hydrogen-ion concentration of the soil water in contact with minerals, leading to the rupture of chemical bounds via a process called hydrolysis. The small size of the hydrogen ion allows it to penetrate the mineral structure and to displace much larger cations from their positions within the mineral structure. Generally, the smaller the ionic charge (valence) of the cation, the more readily it is displaced during hydrolysis: sodium (Na^{+}) and potassium (K^{+}) are the most readily displaced, followed by divalent ions (Ca^{2+}, Mg^{2+}, Fe^{2+}), and trivalent ions (Fe^{3+}, Al^{3+}) and Si^{4+} (Table 2.3). Heavily weathered tropical soils are dominated by aluminum and iron, and the iron gives them a deep red colour.

Table 2.3. Composition of average igneous rocks and three surface soils of increased weathering: temperate forest soils are the least weathered and tropical forested soils the most weathered. All values reported as percentages. Adapted from H.L. Bohn et al. (1979). Soil Chemistry. John Wiley and Sons, New York.

Element	Average Igneous Rock	Temperate	Sub-Tropical	Tropical
		Forested soil	Forested soil	Forested soil
SiO_2	60	77	80	26
Al_2O_3	16	13	13	49
Fe_2O_3	7	4	5	20
TiO_2	1	0.6	1	3
MnO	0.1	0.2	0.3	0.4
CaO	5	2	0.2	0.3
MgO	4	1	0.1	0.7
K_2O	3	2	0.6	0.1
Na_2O	4	1	0.2	0.3
P_2O_5	0.3	0.2	0.2	0.4
SO_3	0.1	0.1	0.1	0.3

The main source of hydrogen is the reaction between gaseous carbon dioxide and water to produce carbonic acid (H_2CO_3), which ionizes to form bicarbonate (HCO_3^{-}) and hydrogen ions (discussed in more detail in Pennock et al., 2010). Hydrogen ions are also produced by various organic acids that form during the decomposition of organic materials. See Chapter 5 for further information on soil reaction and pH.

The hydrogen ions act in distinct ways upon different classes of minerals. Silicate minerals dominate the mineral fraction of sediments formed from igneous rocks. The basic building block of silicate minerals is formed by a silica cation surrounded by four oxygen anions (reference to section on clay minerals). The various classes of silicates are caused by the inclusion of other ions along with the basic silica-oxygen units. An important class of silicates are those that have a layer structure (i.e., phyllosilicates) that form in the near-surface zone of the earth's crust at much lower pressures and temperatures than many other silicate minerals. This class of minerals is generally called the clay minerals, which is somewhat unfortunate since clay also refers to a size range of mineral particles as well.

Silicate minerals high in easily displaced cations (e.g., Na^{+} and K^{+}) experience collapse of their structure after cation displacement and the mineral disintegrates. At the other extreme, quartz is composed solely of silica and oxygen and is the most resistant silicate mineral to weathering; indeed many of the most highly weathered landscapes in the world are dominated by sand-sized quartz minerals (Paton et al., 1996). Hence, under most conditions in Canada, quartz is little affected by weathering processes. Weathering

of clay minerals in Canadian soils is largely restricted to acidic Podzolic soils where transformations and losses of clay minerals through weathering in the bleached surface horizon occurs (Kodama, 1979).

Sediments formed from sedimentary rocks contain high contents of mineral carbonates and more soluble salts along with silicate minerals. The most important in the Canadian context are carbonates and more soluble salts formed through precipitation of ions in ancient seas and lakes. Compared to silicate minerals carbonates and salts are readily dissolved in water (i.e., they are soluble); hence, they are readily weathered in the soil.

The final class of non-silicate minerals are oxides of aluminum and iron. Most metal oxides are weathering products from primary igneous silicates and are dominant in heavily weathered soils such as those in sub-tropical and tropical areas. Because almost all Canadian soils are relatively young (in geological terms), iron and aluminum oxides are less common in Canadian soils but are important in colour formation in acidic forest soils and water saturated soils.

Cations freed by mineral weathering are available to be taken up through plant uptake, moved by mobile soil water laterally or vertically, or react with other ions to form new, secondary minerals. Low valence cations such as sodium (Na^{+}) are most readily released by weathering and are the most likely to remain in solution and hence be removed from soil in the process of leaching. Cations such as calcium (Ca^{2+}) and magnesium (Mg^{2+}) are released by weathering in the upper part of the soil but typically reform as secondary carbonate minerals lower in the soil. The higher valence ions (Fe^{3+} and Al^{3+}) are the most likely to remain in place and form new secondary soil minerals. These secondary minerals dominate older, more heavily weathered soils outside of Canada.

Role of water and organisms in chemical transformation of soil materials

Carbonic acid (an important driver of hydrolysis in soils) results from the initial reaction of water and carbon dioxide. Generally, the higher the availability of both water and carbon dioxide, the greater the amount of hydrolysis and hence weathering of soil minerals that will occur. Plants and organisms that live in the soil contribute to hydrolysis by raising the CO_2 concentration in the topsoil by 10 to 100 times atmospheric levels, thereby increasing H^{+} concentration. Generally, mineral weathering will be greatest where the availability of water and carbon dioxide is greatest; for example, in the coastal rainforest of British Columbia. The least weathering will occur where both water and CO_2 are the most limited, such as very dry, cold soils of the high Arctic.

Organisms also accelerate weathering through the process of chelation where certain organic molecules caused by decomposition form complexes with mineral cations (especially relatively immobile ions such as iron and aluminum) and extract them from mineral structures. These mineral-organic complexes can then be moved through the upper soil and deposited deeper in the soil profile. Lichens are very active in the promotion of chelation, and they are typically early colonizers of soil surfaces.

MAJOR SOIL FORMATION PATHWAYS IN CANADA

Thus far, we have seen that shape of the landscape and the parent materials that soils form in were largely inherited from glacial and post-glacial geomorphic processes. These parent materials are then transformed through weathering, which is closely linked with organisms and the availability of water.

Although soil water availability and organisms are the main controls on national-level soil formation, the temperature plays an important secondary role. Most all Canadian soils experience periods of freezing and thawing every winter, and frost action is an important physical control on soil formation. The role of freezing and thawing is best expressed in the soils of Taiga and Arctic ecozones typical of northern Canada, where frost action is indeed the dominant control on soil formation. In more temperate regions, temperature also controls, in part, the rate at which microorganisms decompose organic materials and the type of decomposition product produced, leading to distinctive soil horizons in forested soils.

At the national level, the regional patterns of moisture availability and temperature give rise to characteristic assemblages of vegetation and animals and these have been classified into distinct ecozones (Figure 2.11). These ecozones have characteristic bundles of soil-forming processes associated with them, and soils in these ecozones follow distinct pathways of soil formation. Forests, grasslands, and tundra ecozones all have distinct surface soil layers associated with them due to the interplay between their characteristic plants and the extraordinary range of organisms that dwell in the soil.

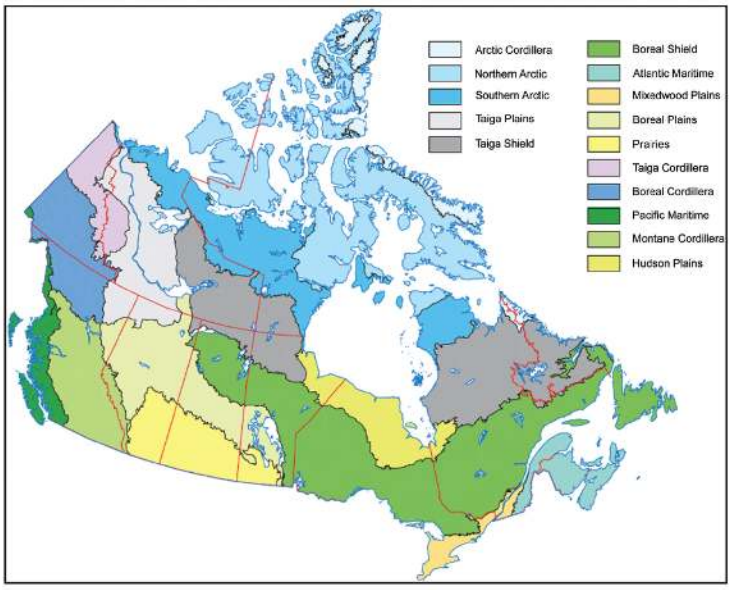


Figure 2.11. Ecozones of Canada. © Darrel Cerkowniak; adapted from Smith et al. (2011) and licensed under a CC BY (Attribution) license.

Forest Soils

In forested biomes the great majority of fresh OM is added to the soil surface in the form of fallen leaves, branches, and dead trees themselves. The freshly added OM undergoes decomposition by soil animals and microorganisms and typically the organic mat on the soil surface is layered, with fresh OM on the top and highly decomposed material directly overlying the mineral soil surface. The degree of decomposition depends on the water available, the temperature, and the nature of the OM added to the surface.

For most forested biomes in Canada the OM is not incorporated into the mineral layer underneath and there is a sharp contact between the two. Mixing of the two layers does not occur because there are no soil-dwelling organisms (such as earthworms or small mammals) that cause mixing to occur. Earthworms were eliminated from soils in northern North America by the great ice sheets and they are only slowly recolonizing the southern parts of Canada. Small mammals that live part of their lives within the soil are also largely absent from forested sites (Zaitlin and Hayashi, 2012). In areas of Canada where earthworms have invaded forested sites (such as Southern Ontario and Quebec), the OM becomes incorporated into the surface mineral layer and a mixed organic-mineral surface layer occurs (informally called **mull** by soil scientists) (Figure 2.12) . Where earthworms are absent, the organic mat instead forms a sharp and distinct contact with the mineral layer; litter layers of this type are informally called mor (Figure 2.13). Moder is a form of organic layer intermediate between mor and mull. An excellent on-line resource to distinguish among these three forms is at the Virtual Soil Science Learning Resources forest floor website.



Figure 2.12. Mull horizon in a sandy forest soil. The surface mineral soil layer consists of organic matter intimately mixed with the sand mineral matter. © Darwin Anderson (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

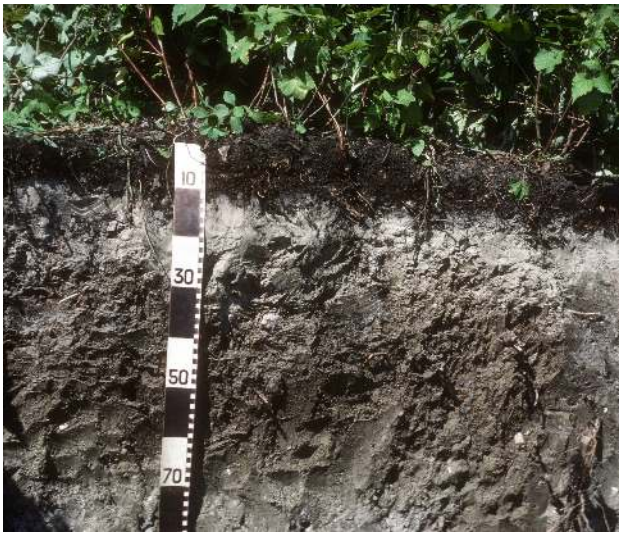


Figure 2.13. Mor horizon in loamy forest soil. The surface mineral layer lies over top of the grayish surface mineral layer and is not mixed into the mineral layer. © Darwin Anderson (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

The nature of the parent materials (i.e., sediments derived from igneous rocks versus those from sedimentary rocks) strongly controls the soil genetic pathways of forested biomes. Glacial sediments derived from igneous rocks are generally dominated by quartz sand and sand-sized silicate minerals and salts and carbonates are absent. Hence, these soils have little inherent ability to neutralize hydrolysis by hydrogen ions in the soil water. The high hydrogen-ion levels (or acidity) is further augmented by the decomposition of organic materials, which releases organic acids from the OM.

The reaction between the acids in soil water and the sandy surface material produces a characteristic genetic pathway (Figure 2.14). Initially, the acids react with very fine iron particles, which form thin, reddish or red-brown skins on the resistant quartz grains

(Paton et al., 1996. p. 125). This causes an overall reddening of the upper part of the soil. In the second stage, the organic molecules produced by decomposition form complexes with the iron coatings and strip them from the quartz grains, causing bleaching of the uppermost layer of the soil. This bleached layer characteristically has grayish or grayish-white colours.



Figure 2.14. Genetic pathway for soils in acidic, coarser parent materials. The soil on the left is minimally weathered with an organic mat overlying parent material. The soil in the middle has (from the top) an organic mat, thin greyish bleached layer, reddened layer, transitional yellow-red layer, and unaltered parent material. The soil on the right has an organic mat, thick white-gray bleached layer overlying a dark, iron and OM enriched layer that grades to parent material. © "Image A: Roly St. Arnaud (Saskatchewan Center of Soil Research); Image B: Saskatchewan Center of Soil Research; Image C: Ken Van Rees (Saskatchewan Center of Soil Research)"; all images licensed under a CC BY (Attribution) license.

In the third stage, the iron-organic complexes are transferred deeper into the soil where they are deposited, creating a layer of concentrated organic compounds and iron (and, to a lesser extent, aluminum). These depositional layers have black or reddish colours, and the colour contrast with the overlying bleached layer is very striking (Figure 2.14). The process of the formation of organic-mineral complexes in the upper soil and their transfer and deposition in the underlying soil is called podzolization (Sanborn et al., 2010, p. 857-866). The upper layer, where iron is lost from the quartz grains, is an example of an eluvial horizon; the underlying layer, where the organic-mineral horizons are added, in an illuvial horizon.

The genetic pathway is very different for soils formed in sediments derived from sedimentary rocks. These sediments are typically rich in carbonates such as calcium carbonate ($CaCO_3$) and have loamy textures (i.e., a mixture of sand, silt, and clay). Initially the carbonates are found throughout the soil profile and act to neutralize (or buffer) the acids produced by organic decomposition. This neutralization process gradually destroys the carbonate minerals and a layer of carbonate-free material forms at the soil surface and thickens with time. The calcium ions released by weathering are transported to depth in the soil profile where they re-form into new, secondary calcium carbonate minerals (Figure 2.15). The loss of calcium from the upper soil is termed decalcification and the gain of newly formed calcium carbonate at depth is called calcification.



Figure 2.15. Calcium carbonate deposited along root channels in a forested soil developed on till. The original, inherited carbonate minerals were dissolved in the overlying soil and reformed in the root channels deeper into the soil. © Dan Pennock, Univ. of Saskatchewan (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

The loss of calcium carbonate from the upper soil creates a more acidic soil chemical environment in the upper soil that allows very fine clay-sized particles to be released into the soil water and transported to the underlying soil layer, where they are deposited. This process of clay transfer (called lessivage) creates layers with contrasting textures – lower clay content in the upper eluvial layer and higher clay content in the illuvial layer (Figure 2.16) (Lavkulich and Arocena, 2011). These texture-contrast layers can also occur due to layering of sediments during the glacial or Holocene periods – for example, when a wind-borne silt is deposited over an underlying loamy till. Soil scientists call these layered horizons lithological discontinuities. There is no field-based method to distinguish between texture-contrast horizons caused by lessivage and those due to lithological discontinuities and indeed lessivage may well occur in a profile with a lithological discontinuity.

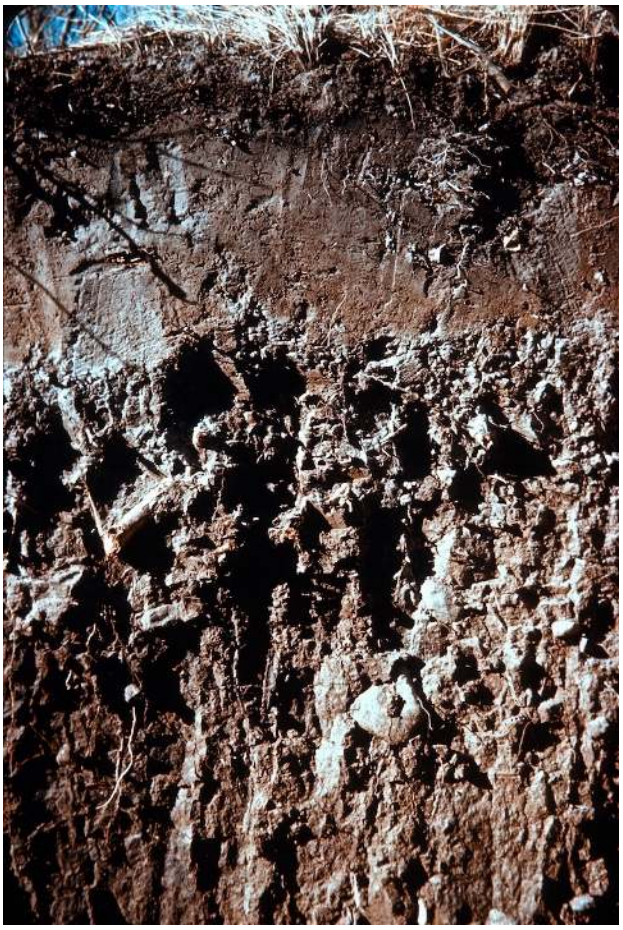


Figure 2.16. Forested soil with texture-contrast horizons. The uppermost mineral horizon has a lower clay content than the underlying layer. The structure of the two layers also differs greatly, with a rougher, more strongly developed structure in the higher clay layer. © Roly St. Arnaud (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

Forested biomes also contain many sites where water ponds and wetlands form. The genetic pathways of these wet soils differs greatly from the better drained forest locations discussed above and are discussed separately below.

Wetland Soils

Wetlands are a common feature of the Canadian landscape (Figure 2.17). Globally one of the largest extents of wetlands is found in the Hudson Plain ecozone south of Hudson Bay. Wetlands are also very common at the contact between the plains of western Canada and the Canadian Shield – many northward flowing rivers from the plains are unable to carve channels through the resistant igneous rocks of the Shield and water ponds at the contact. There are also countless small wetlands distributed throughout almost every landscape in Canada due to local-scale topographical unevenness.

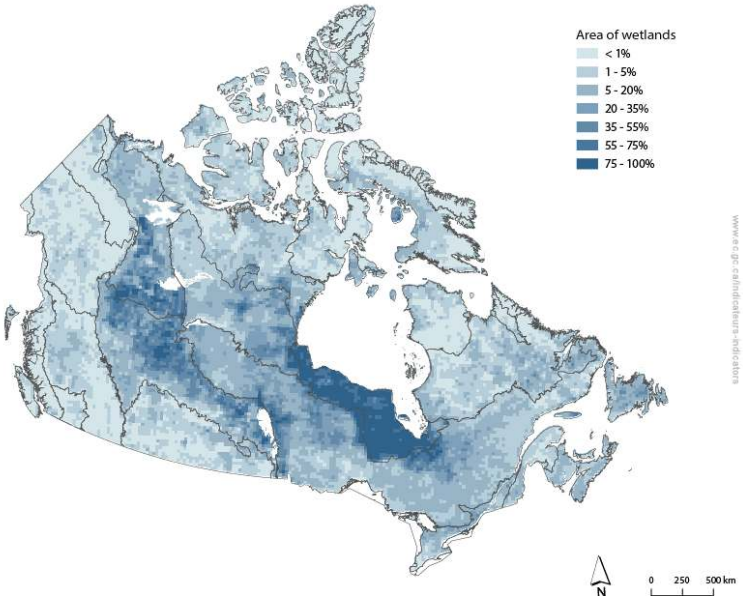


Figure 2.17. Extent of Canada's wetlands (circa 2000). © Environment and Climate Change Canada; used with permission and licensed under the Open Government Licence – Canada. https://open.canada.ca/en/open-government-licence-canada

The presence of prolonged periods of water saturation in the soil profile fundamentally alters the pathway of both OM decomposition and mineral weathering (Bedard-Haughn, 2011, p. 768-773). Microorganisms decompose organic material to feed off the nutrients and energy contained within the OM. The energy (in the form of electrons) that is liberated during decomposition must be taken up by other ions or organic compounds or else decomposition is unable to proceed. In soil with oxygen in the soil pore space (called aerobic conditions), oxygen itself takes up the electrons and decomposition continues until only highly resistant organic compounds are left. Where water displaces oxygen in the pore space (called anaerobic conditions) alternative electron acceptors must be used instead. The most common mineral electron acceptor is iron (Fe). In aerobic soils Fe occurs as $Fe^{3+}$and is responsible for the reddish colours found in soils. In anaerobic conditions the Fe^{3+} ion accepts an electron and its charge is reduced to Fe^{2+}. This form of iron is mobile and can be transferred within the soil, and the loss of iron creates dull soil colours, often with a bluish-grey hue. If a soil becomes partially or wholly aerobic (when water drains out of the soil column) some of the reduced iron may re-oxidize to a reddish state, giving the soil overall a mottled appearance (Figure 2.18). The complex processes associated with these chemical state transformations are called redox (reduction-oxidation) processes, and in soil science are referred to as gleying. See the Oxidation-Reduction (Redox) Processes section in Chapter 5 for more information on these processes.



Figure 2.18. Gleying. The middle part of the soil profile exhibits alternating gray and reddish soil colours (or mottles) overlying a primarily reddish (oxidized) iron zone. The water table is visible at the base of the photo at about 1 m depth. © Darwin Anderson (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

In soils with prolonged water saturation all mineral electron acceptors may ultimately be used up and organic decomposition essentially ceases. A largely undecomposed organic mat (called peat) begins to build up from the mineral soil surface and can reach up to several meters in thickness (Tarnocai, 1990; Kroetsch et al., 2011, p. 814-816) (Figure 2.19). The very slow decomposition processes produce methane (CH_4), which is an important greenhouse gas. Often the organic mat will contain layers with different degrees of decomposition, ranging from almost entirely undecomposed material (called fibric material), partially unrecognizable material (mesic), through to largely unrecognizable material (humic material). In some cases the organic material is mixed with mineral particles, creating muck soils.



Figure 2.19. Peat deposit. The peat at the bottom of the profile is more decomposed (and hence less fibrous) than the peat at the top of the profile. © Saskatchewan Center of Soil Research; licensed under a CC BY (Attribution) license.

In northern Canada, accumulation of organic material is very common due to the low evapotranspiration rates and the natural conversion of forests into wetlands is widespread. Ultimately, *Sphagnum* mosses become the dominant vegetation on the landscape and thick organic mats can occur. This process of organic soil development is termed paludification.

In more southerly locations (such as the Prairie and Mixedwood Plains ecozones), the warmer temperatures and hence higher decomposition rates prevent the formation of thick organic mats. Wetland soils in these locations are instead typically mineral soils with soil colours (dull colours, mottling) reflecting the gley conditions.

A final type of thick organic deposit is associated with wet conditions (typically 3000-4000 mm of precipitation per year) in the coastal rainforest of the Pacific Maritime ecozone of British Columbia. The lush coastal rainforest produces large amounts of organic material and the high rainfall levels retard decomposition, leading to a thick buildup of forest-derived organic materials (called **folic** materials) (Fox and Tarnocai, 2011, p. 828-234).

Tundra and Arctic Cordillera Soils

In the cold climate of the Canadian Arctic, the soil-forming processes discussed above (for example, organic matter decomposition, transformations of minerals) are still present but they occur at very slow rates due to the low annual temperatures and short frost-free period. Soils in these regions have permanently frozen layers (called permafrost) within them and any water in these layers is frozen. The presence of ice leads to unique soil-forming processes, which are generally termed cryogenic processes.

The main cryogenic process affecting soil properties is mixing (or cryoturbation) caused by the growth of ice lenses in the soil (Tarnocai and Bockheim, 2011, p. 750-757). When ice forms in the soil it draws unfrozen soil water along a thermal gradient to the

ice front. This causes the lens of ice to grow and displace mineral and organic soil material, mixing together the soil at the ice-water interface (Figure 2.20). This mixing is more pronounced in the surface layer of the soil, which undergoes seasonal thawing and freezing and is called the active layer. The pressure exerted by the ice also causes sorting of coarse rock fragments in the soil and the occurrence of unique features such as stone circles and polygons, earth hummocks, and ice wedge polygons. Turbation is most pronounced in loamy and fine-textured soils, and although permafrost occurs in coarse-textured sand and gravel soils the soils often show little or no evidence of cryoturbation.

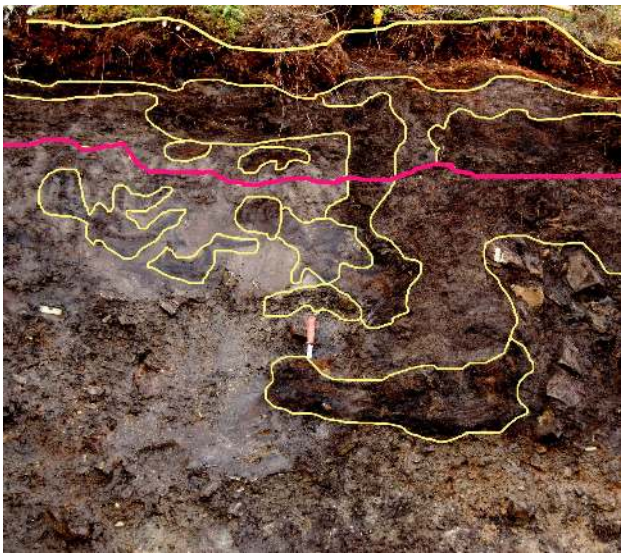


Figure 2.20. Effects of cryoturbation on soil horizons. The soil above the pink line is the active layer and the part below the pink line is permafrost. The yellow lines delineate patches of horizons that have been disrupted and mixed by frost action (cryoturbation). © Paul Sanborn and Scott Smith; licensed under a CC BY (Attribution) license.

At the southern boundary of the tundra in the Taiga ecological region, permafrost becomes discontinuous in the landscape, with patches of permafrost in lower-lying landscape positions and non-permafrost soils in uplands. Because of the dependence of permafrost on prolonged freezing conditions these transitional areas are very likely to be affected by increasing temperatures due to human-induced global warming.

Similar conditions occur at high elevations in the western Cordillera and permafrost soils also occur in the Alpine tundra. The Alpine tundra can be home to burrowing mammals and a turbated surface layer high in organic matter often occurs.

Grassland Soils

The greatest extent of grasslands (or at least former grasslands, as much of this area has been cultivated) in Canada is the Prairies ecozone. The Prairies ecozone is underlain by sedimentary rocks dominated by clay-rich marine shales and limestone and hence the glacial and Holocene sediments found in the region are normally high in carbonates and (in places) other more soluble salts. The landforms of the Prairies are dominantly formed by glacial deposition and include extensive areas of moraine, glacio-fluvial sands, and glacio-lacustrine silts and clays (Figure 2.2). Eighty-one percent of Canada's agricultural land is located in the three Prairie provinces (160 Mha in total; 62 Mha in Saskatchewan, 51 Mha in Alberta, and 18 Mha in Manitoba; Census 2011).

Two groups of soil-forming processes dominate this region: (1) below-ground OM additions from grasses (i.e., fibrous roots) and subsequent mixing by animals and (2) weathering of carbonates and salts and their transport through the soil and landscape (Pennock et al., 2011, p. 725-732). There are also two additional genetic pathways that are found in specific areas, one associated with physical mixing due to high clay contents, and the other due to the effect of high soil sodium (Na^+) levels on soil formation.

The majority of biomass in grasses is situated below-ground in their extensive root systems, which reflects the need for the grasses to fully exploit the soil for water in this region of significant water deficit. In the example shown in Figure 2.21 (Slobodian et al., 2002) the aboveground parts of the grasses had a biomass of 1,060 g m^{-2}, compared with 2,532 g m^{-2} belowground in the root systems. As roots age and die they are transformed by microorganisms through a complex set of processes into resistant humus, which can be stored for hundreds of years in the soil. The storage of soil organic carbon (SOC) is several times that of the living root material in the soil (Figure 2.21).

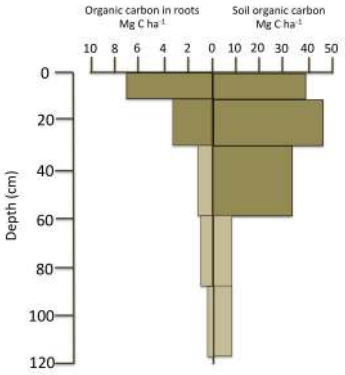


Figure 2.21. Depth distribution of organic carbon in roots and in soil organic carbon from mid-slope soils near Saskatoon, Saskatchewan. The units express the mass of carbon (Mega (106) grams or tonnes) per hectare in the specified depth increment. Data extracted from Slobodian et al. (2002) and figure created by Dan Pennock. .

The belowground OM inputs into the upper soil are then turbated by the action of soil-dwelling mammals (such as ants, ground squirrels, badgers, moles) and normally the organically-enriched upper soil is more mixed and extends more deeply than the root layers themselves (Figure 2.21). Often the evidence of burrowing animals in the form of infilled burrows is evident in these grassland soils (Figure 2.22).



Figure 2.22. The organically enriched surface horizon formed by high organic carbon inputs from roots and mixing of soil by burrowing animals and other macroorganisms. © Roly St. Arnaud (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

Decomposition of the root biomass also leads to high levels of CO_2 gas in the upper soil, also contributing to the destruction of salts and carbonates by chemical hydrolysis. As discussed above, the salts such as halite are readily weathered and release cations such as Na^+ into the soil solution, where it can be leached from the soil into the groundwater. The main carbonate mineral $CaCO_3$ is more slowly dissolved in the upper part of the soil column and the Ca^{2+} ion is transferred deeper into the soil, where it typically precipitates out of solution and reforms secondary calcium carbonate minerals. Therefore the grassland soils normally have an area of carbonate depletion in the upper soil and carbonate concentration lower in the soil profile.

The more mobile sodium ions are transported through the groundwater system for distances ranging from a few hundred meters to many kilometers. Ultimately the groundwater flow system can lead to them being brought back to the soil surface (termed groundwater discharge) (Chapter 4). Continued evapotranspiration of the discharging groundwater can lead to increasing levels of soil salts through time (termed salinization) and these high salts levels can be injurious to plant growth (Figure 2.23).



Figure 2.23. Soil salinity. The white film on the surface is formed by salt crystals that have precipitated on the surface. Grass growth is suppressed on the saline soil and instead salt-tolerant *Salicornia rubra* dominates the vegetation cover. © Darwin Anderson (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

Sodium is also responsible for a distinct set of soil-forming processes typical of areas where sodium-rich bedrock occurs close to the soil surface or where sodium-rich groundwater reaches the soil surface (Acton, 1990; Miller and Brierley, 2011). When sodium binds to reactive clay particles it causes the clay to be released from the mass of soil into the soil pore water (Chapter 5) and the sodium-saturated clay is subject to transport from the surface layer to lower layers in the soil. This process (called **solonization**) leads to formation of a clay-enriched, high-sodium layer deeper in the soil. Once this layer has formed it can lead to pronounced mineral weathering and bleaching of the overlying layer and creation of a distinct eluvial, gray-coloured horizon; this process is called **solodization** (Figure 2.24). See the section on Salinity and Sodicity in Chapter 5 for more information on these processes.



Figure 2.24. Solodization. This prairie soil has an organically enriched upper layer underlain by two layers that are typical of a soil that has undergone solodization. The lowest layer has high clay and sodium content, and the perching of water on this dense layer leads to weathering and bleaching of the upper part of the clay-rich layer. © Darwin Anderson (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

A final distinctive soil-forming process associated with grasslands is physical mixing of soil caused by the shrinking and swelling of some clay minerals (Brierley et al., 2011). Many locations in the Prairies have glacial-lacustrine sediments with high smectite clay contents. The smectite clays originate in the marine shale bedrock found throughout the Prairie region. These clay minerals can absorb water (and swell) when the soil is saturated and release water (and shrink) when the soil is dry. This physical process causes cracking of the soil (Figure 2.25) when the soil is dry and turbation (sometimes referred to as argillipedoturbation or self-mulching) of the soil when the soil is wet, leading to inversion of soil material. The turbation mixes together the surface soil material and prevents the formation of distinct soil horizons with clear boundaries with adjacent horizons. This set of processes are called **vertic** processes.



Figure 2.25. Cracks formed due to shrinking of expansive clays at dry soil condictions – an example of vertic processes. © Darwin Anderson (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

The grasslands also have countless wetlands associated with them and these wetlands are important habitats for migratory birds. The wetlands in the Prairie ecozone periodically dry out; hence, OM deposits do not form; instead they are dominated by mineral soils showing gleyed features (dull colours, mottling).

Formation of Soil Structure

As discussed in Chapter 4, the structure of the soil plays a very important role in formation of pores that consequently impact air, gas, and water movement in the soil as well as the ability of roots to penetrate into the soil to extract water and nutrients. The initial structure of a soil is inherited from the parent sediments but many of the soil-forming processes discussed above lead to distinctive arrangements (called aggregates or peds) of mineral and organic materials and soil pores (Figure 2.26).

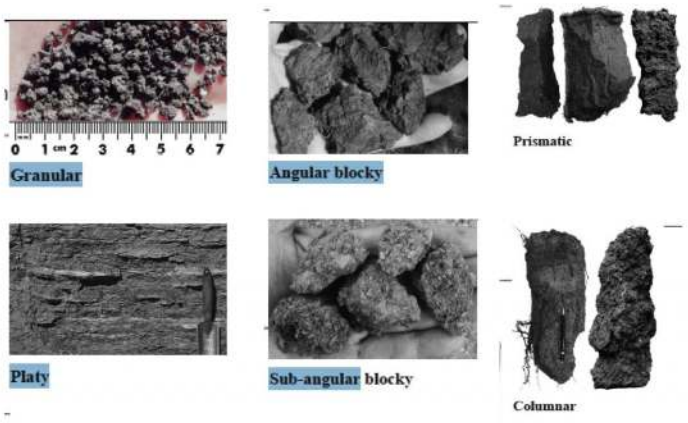


Figure 2.26. Main types of soil structural units or aggregates. From Watson and Pennock (2016). Reproduced with permission; licensed under a CC BY-NC (Attribution NonCommercial) license.

The key property of the parent sediment that influences the types of aggregates formed is the clay content. Clay particles are reactive (due to their small size and presence of charges on their surfaces) and can bond together into large and stable structural units. Soils dominated by sand-sized particles are largely non-reactive; hence, sandy soils have weakly developed and readily disrupted aggregates (if they are present at all).

Two sets of soil-forming processes modify the initial, sediment-derived structure. The first is the addition, transformation, and mixing of OM that occurs in grassland soils, tundra soils (due to small mammals), and in forest soils where earthworms have invaded the forests (such as in southern Ontario and Quebec). The dense root network of grasses acts as netting to bind solid soil material together and the larger soil organisms such as earthworms create soil aggregation by ingesting and excreting the soil material. The result of these agents is formation of a relatively loose granular aggregates (Figure 2.26), and such soil structire is well suited to root penetration and growth.

The second bundle of processes are the physical processes, especially those associated with freezing and thawing of soil water and swelling and shrinking of clay minerals. In some cases freezing and thawing in the bleached eluvial horizon in forest soils creates a horizontally layered platy structure (Figure 2.26). In loamy or clay-dominated soils the physical processes create a range of sizes and shapes of aggregates (Figure 2.26). In some cases (such as the high Na+ soils of grasslands), the structural units become so dense that they prevent root penetration and limit plant growth (clay-rich layer in Figure 2.25). Another example of growth-limiting structural units occurs in forested landscapes where deposition of iron-organic compounds and clay can clog soil pores and solidify into a near-solid mass. Depending on the type of cementing agent these features are called **duric horizon or placic horizons.**

Apparent structural units caused by human activity, such as cultivation, are referred to as clods, which do not cohere in the same manner as structural units that are naturally formed.

Human Activity as a Soil Forming Factor

So far in this chapter we have only considered soil formation under natural conditions but in many parts of Canada soils have also been transformed by human use. The impact of humans on soil formation has been most pronounced since settlement by Europeans began. Initially this impact was largely through conversion of land to agriculture and exploitation of forests but in the 19th and 20th Centuries the impact of mining and petroleum exploitation also became important. In the 20th Century the great expansion of urban areas has also profoundly affected the soil landscape.

The conversion of grassland and forested landscapes to agricultural production affects almost all of the processes that occur in soil. In terms of soil genesis the almost universal impact of agricultural conversion is a loss of SOM due to the combined effects of the breakup of aggregates and exposure of SOM to microbial decomposition and accelerated microbial decomposition and soil erosion (discussed below). Globally studies have shown that an average about 30% of the original soil organic carbon is lost when conversion from forest or grassland to cropland occurs (FAO, 2019). In forested sites with leaf litter lying on the surface, the clearance of trees and ploughing incorporates part of the leaf litter layer into the soil, resulting in an organically enriched mineral surface horizon.

Clearing of the land surface also leads to increases in soil erosion by water, wind, and tillage. Soil erosion is the detachment and transport of soil from one position and its deposition elsewhere (FAO, 2019). Water erosion involves soil detachment and transport by water. Soil can be detached by both raindrop impact and by flowing water, and transport is by water flowing on the surface. Where water concentrates on the soil surface, channels can be incised into the surface, creating shallow rills (Figure 2.27A) or deeper gullies (Figure 2.27B). The eroded soil may be deposited within the field or at the field edge but in some cases it is carried directly to stream systems, where it causes many environmental problems.

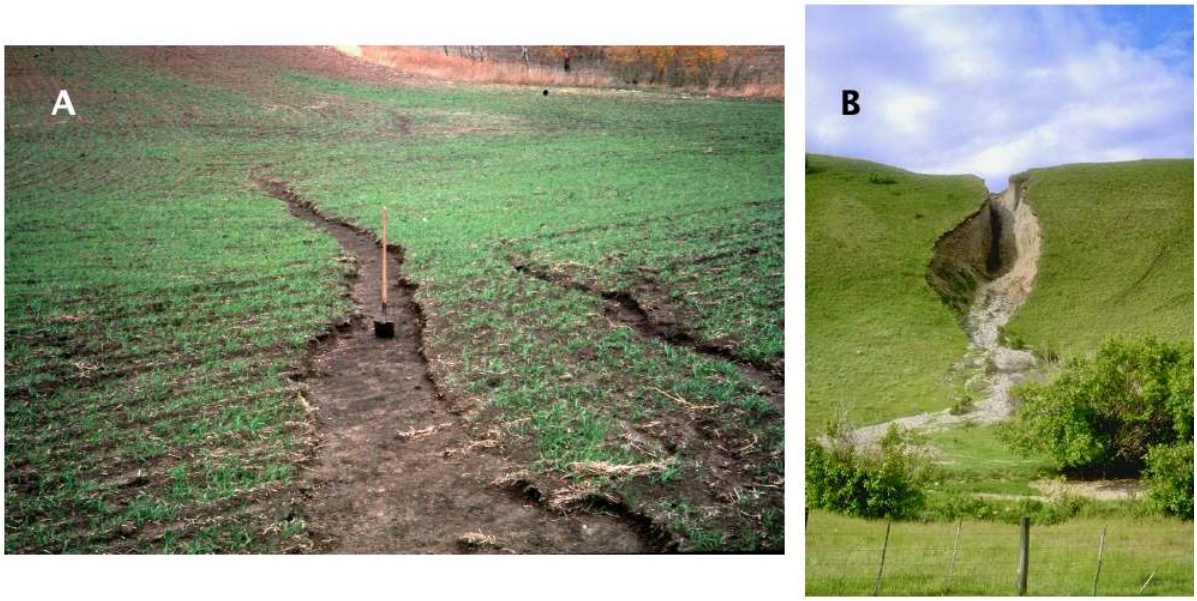


Figure 2.27. Water erosion in Saskatchewan (left) and British Columbia (right). **(A)** Rill erosion in which flowing water has incised through the loose, organically enriched topsoil and then widened its channel (or rill) when it encountered a denser soil layer. These rills will be filled in with soil the next time it is tilled. Image courtesy of Dan Pennock, Univ. of Saskatchewan (Saskatchewan Center of Soil Research). **(B)** Gully erosion in which water channelled across unprotected land washes away the soil along the drainage lines. Image courtesy of Maja Krzic, Univ. of British Columbia. Both images licensed under a CC BY (Attribution) license.

In wind erosion the wind itself detaches and transports soil particles. Wind erosion can occur anywhere where the soil is dry and bare and the land surface is flat. The worst episodes of wind erosion in Canada occurred in the southern part of the Prairie ecozone in the 1930s (hence, the Dirty Thirties) and lead to widespread land abandonment and depopulation.

Tillage erosion is a more insidious type of erosion. Tillage operations by farm implements inexorably remove soil from upper slopes and deposits it in lower slope positions during farm operations, yet the effects of any one operation are almost undetectable (FAO, 2019).

In terms of soil formation the main effect of soil erosion is to remove part of the organically enriched surface soil. In tillage erosion the amount of soil removed in a year may be very small (0.5 to 1 mm) but the cumulative effect year-after-year can be substantial. In water erosion rills and gullies can remove part or all of the organically enriched layer, as can wind erosion during very severe events. This loss (or truncation) of the surface layer leads to a loss of the most fertile part of the soil for plant production. It may also bring a growth-limiting lower soil layer (for example, with high sodium levels) closer to the surface, further reducing plant growth. Where deposition of eroded soil occurs in lower slope positions, the upper soil layer becomes unnaturally over-thickened. These are termed **cumulic** soils (Figure 2.28). See Chapter 14 (Soil MIneralogy) for further information on the transformations of minerals by weathering.



Figure 2.28. Cumulic soil formed by deposition of water-eroded soil at the base of a slope. The original topsoil from upslope has been buried under less fertile subsoil. The aluminum stick is 1 m long. © Dan Pennock, Univ. of Saskatchewan (Saskatchewan Center of Soil Research); licensed under a CC BY (Attribution) license.

Resource extraction and transport activities such as mining for minerals or for bitumen or pipeline development also fundamentally alter the soil, in some cases by completely excavating the soil and underlying sediment. Currently, regulations are placed to require some level of reconstruction of a soil but inevitably the constructed soil will differ from the original soil profile. Soils resulting with major disruption from human activities are termed **anthropogenic** (i.e., formed by humans).

A final (in every sense of the word) impact of humans on soil genesis is the complete stripping and/or burial of soils that occurs where infrastructure such as houses, roads, and shopping centers is constructed. This irrevocable loss of soil is termed **soil sealing** and globally is one of the major threats to soil. In Europe, for example, soil sealing was judged in 2015 (Montanarella et al., 2016) to be the major threat to the soil in that region. In Canada, the largest areas of urban development also correspond to our most productive soils and hence this human-induced destruction of soil is a significant concern here as well (Montanarella et al., 2016).

Soil Genesis in Non-glaciated Terrain

The genetic pathways found in Canada are also common in other glaciated landscapes of the Northern Hemisphere although the duration of soil formation is typically longer in northern portions of Europe and Asia than in Canada due to differences in the timing of the glaciations. Soil formation in regions beyond the margin of the great Northern ice sheets can, however, differ greatly from the glaciated regions, and the genesis of these non-glaciated soils is key to understanding the particular management issues that occur in these regions.

The dominant (in terms of area) pathway of soil formation in non-glaciated regions is associated with subtropical and tropical regions, which have moist to wet rainfall patterns, consistent warm temperatures, and old land surfaces where soil formation has been occurring for hundreds of thousands to millions of years. An additional difference is that the soils in these regions often develop directly into the rocks present at the surface whereas many Canadian soils have developed on thick glacial sediments that are absent from unglaciated landscapes. Hence, major differences in soil development may occur in close proximity due to the presence of different rock bodies at the surface. These landscapes also typically have many active turbation agents such as termites and ants and the effects of turbation can be very pronounced.

In relatively young and less weathered landscapes, clay transfer in the soil is very well developed and a thick layer of clay accumulation occurs in the middle horizons of the soil. However, due to the abundant moisture and warm temperatures in subtropical and tropical regions, processes such as clay translocation and chemical weathering of the soil minerals inherited from the parent materials is very advanced. Chemical weathering of silicate minerals destroys many of the original minerals such as feldspars or olivines, and neo-formation of iron and aluminum oxides occurs, which gives these soils a strong red hue (Figure 2.29). Along with the destruction of the silicate minerals many of the base cations (Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}) have been leached out of the soil profile, and hence these soils are generally less inherently fertile than Canadian soils. Soils with these features are common

in the south-eastern United States and much of South-east Asia. See Chapter 14 (Soil Mineralogy) for further information on the transformations of minerals by weathering.



Figure 2.29. Moderately weathered subtropical soil (termed an Ultisol in the classification system used in the USA) from Bibbs County, Alabama. Scale is in inches. Photo courtesy of USDA Natural Resources Conservation Service. © USDA Natural Resources Conservation Service is licensed under a Public Domain license.

The ultimate stage of this genetic pathway is associated with areas of South America and Africa around the equator. Chemical weathering of the clay minerals themselves occur and the silica is removed from the clay minerals themselves in the process of **desilication**. Some of the ions released reform simple clay minerals such as kaolinite but these soils are dominated by clay-sized iron and aluminum oxides, which often gives them a deep red colour. This end-point of chemical weathering is termed **laterization**. Again, the extensive weathering and leaching of bases makes these soils inherently infertile compared to many Canadian soils. Because of the very long duration of soil formation, the soils are also extensively turbated by soil-dwelling organisms, and the soils often show limited development of horizons because of this mixing (Figure 2.30) (Paton et al., 1996). As well, lines of stones within the soil are often observed, which are also associated with turbation. Soils in these landscapes are often very complex due to the imprint of many periods of soil formation under a range of climatic and vegetation conditions and the simple zonation associated with Canadian soils is absent.



Figure 2.30. A profile of a red, highly weathered soil in a tropical area. The upper part of the soil shows little sign of layering due to long-term mixing by organisms. The horizontal sheets in the subsoil are layers of ironstone that are being slowly transformed into soil by weathering. Photo courtesy of USDA Natural Resources Conservation Service. © USDA Natural Resources Conservation Service is licensed under a Public Domain license.

There are also very extensive areas of arid and semi-arid climates in the non-glaciated parts of the globe. The greatest extent of these is in a band running from North Africa through the Middle East and central Asia. Extensive areas also occur in the south-west United States and in much of Australia as well. The driest parts of these regions experience high levels of wind erosion from both natural and human-induced causes (FAO, 2019), and this soil loss coupled with very low water availability for plant growth and soil formation often greatly limits soil formation. In regions with more stable soil surfaces (such as the south-west USA), soils with thick layers of calcium carbonate occur and in certain conditions these may become cemented into a layer informally called **caliche**.

A final distinctive soil-formation pathway is associated with deposits of ash from volcanic activity. These soils are most common in the islands of the “Ring of Fire” such as Japan and parts of Indonesia. In Canada, there are thin volcanic ash additions to many soils in the Western Cordillera and in some soils in the Yukon a distinct, thin volcanic ash can be found (Figure 2.31). Weathering of these volcanic minerals is often very rapid and distinct and highly reactive minerals result from the weathering of the volcanic minerals. The new minerals readily bind with organic matter and phosphorus, and often thick organically enriched horizons occur. These properties are generally called **andic** properties and give these soils a very high natural fertility compared to other soils of surrounding regions.



Figure 2.31. Orthic Dystric Brunisolic soil formed in till deposited by the last glaciation in the Carmacks, Yukon, area. The prominent whitish layer immediately below the organic horizons is a relatively unweathered deposit of the White River volcanic ash which is approximately 1100 years old. Further information on similar soils can be found in Dampier et al. (2011). Photo courtesy of Paul Sanborn; licensed under a CC BY (Attribution) license.

SUMMARY

1. Canadian soils are young by global standards – almost all Canadian soils are post-glacial (i.e., less than 17,000 to 6,000 years old).
2. Because of their youth, many properties (for example, their mineralogy and texture) of Canadian soils are inherited from the parent material in which they form.
3. At the national level there are two general classes of parent materials: (i) acidic, coarse-textured glacial drift derived from igneous rocks and (ii) neutral to alkaline, loamy glacial drift derived from pulverization of sedimentary rocks. Parent materials derived from sedimentary rocks have varying contents of carbonates and more soluble salts; parent materials derived from igneous rocks do not.
4. Most Canadian soils have formed under a stable assemblage of climate, vegetation, and animals since deglaciation, which results in a characteristic zonation of soils.
5. The two main genetic pathways for forest soils in Canada are determined by the two broad classes of parent materials. Soils forming in acidic parent materials experience higher rates of chemical weathering and transfers of iron, aluminum and OM within the profile. Soil forming in weakly acidic to alkaline parent materials initially experience chemical weathering of carbonates and salts followed (in some cases) by transport of fine clay particles within the profile.
6. In wetlands, prolonged water saturation limits decomposition of OM and causes a distinctive set of soil-forming processes – the gley processes – to occur. In forested and northern environments, OM builds up to form thick layers of peat. In drier areas of Canada, mineral soils with characteristic dull colours and reddish iron mottling occur in wetlands.

7. In the Arctic and tundra regions of Canada, layers of permanently frozen soil (or permafrost) occur. Mixing of soil material caused by frost action often disrupts the formation of distinct soil layers in these soils.
8. Grassland soils in Canada, are overwhelmingly associated with alkaline parent materials. High OM inputs from the roots of grasses and mixing by soil organisms creates an organically enriched surface soil layer.
9. Two distinct genetic pathways occur in clay-rich parent materials in the grasslands. One pathway involves mixing of soil by shrinking and swelling of clays. The second occurs in parent materials high in sodium and involves the development of a clay-rich middle layer.
10. The dominant pathway of soil formation in older, non-glaciated regions involves significant transformations of minerals into secondary clay minerals and iron and aluminum oxides.

STUDY QUESTIONS

1. Soils are not rocks, yet the underlying rock has a major influence on soil properties. Why is this?
2. The parent material for most Canadian soils has a glacial origin. What are the four main classes of glacial sediments? How would you rank them in terms of suitability for crop growth? Briefly explain your ranking.
3. Where would we find the oldest soils in Canada?
4. What are the four soil-forming processes? For figures 2.12, 2.14B, 2.16, 2.18, 2.19, 2.20, 2.22, 2.24, and 2.29 identify (in your opinion) what the most important of the four processes is and briefly explain your answer.
5. Why does podzolization occur in some forest soils and clay translocation (or lessivage) in others? Based on your understanding of the two processes, which type of soil would be better for tree growth?
6. Wetland soils are a major reservoir for carbon. What process is responsible for this?
7. How do soil mammals contribute to topsoil development in grassland soils?
8. What are the three causes of soil erosion? Which one is most likely to be affected by climate change in your region?
9. Why are soils in Canada generally more fertile than soils in non-glaciated region?

REFERENCES

Bedard-Haughn, A. 2011. Gleysolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 763-779.

Brierly, J.A., Stonehouse, H.B., and Mermut, A.R., 2011. Vertisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 903-916.

Dampier, L., Sanborn, P., Smith S., Bond J., and Clague, J. 2011. Genesis of upland soils, Lewes Plateau, central Yukon. Part1: Soils formed on Pleistocene glacial deposits. Can. J. Soil Sci. 91: 563-578.

Dyke, A.S. 2004. An outline of North American deglaciation with emphasis on central and northern Canada. Developments in Quaternary Science 2 Part B: 373-424.

Ellis, J.H. 1938. The Soils of Manitoba. Soils Department, University of Manitoba. Project No.14, Manitoba Economic Survey Board. Published by The Department of Agriculture and Immigration, Province of Manitoba, 112 pp.

Environment and Climate Change Canada. 2016. Canadian Environmental Sustainability Indicators: Extent of Canada's Wetlands. Available at: www.ec.gc.ca/indicateurs-indicators/default.asp?lang=en&n=69E2D25B-1.

FAO. 2019. Soil erosion: the greatest challenge to sustainable soil management. Food and Agriculture Organization of the United Nations. Rome. 100 pp. Available at: http://www.fao.org/3/ca4395en/ca4395en.pdf

Fenton, M.M., Schreiner, B.T., Nielsen, E., and Pawlowicz, J.G. 1994. Chapter 26 – Quaternary Geology of the Western Plains; Pages 413-421 *in* G.D. Mossop and I. Shetsen (comp.) Geological Atlas of the Western Canada Sedimentary Basin, Canadian Society of Petroleum Geologists and Alberta Research Council, URL: https://ags.aer.ca/publications/chapter-26-quaternary-geology-of-the-western-plains (Accessed August 22, 2019).

Fox, C.A. and Tarnocai, C. 2011. Organic soils of Canada: Part 2. Upland Organic soils. Can. J. Soil Sci. 91: 823-842.

Fuller, L. 2010. Chernozemic soils of the prairie region of western Canada. Prairie Soils and Crops 3: 37−45. http://www.prairiesoilsandcrops.ca

Jenny, H. 1941. Factors of soil formation: A system of quantitative pedology. McGraw–Hill Inc. New York. (**Note:** A pdf copy of this book is in the public domain at the Soil & Health Library.

Jenny, H. 1980. The soil resource. Origin and behavior, Springer-Verlag, New York. Pp. 377.

Kodama, H. 1979. Clay minerals in Canadian soils: Their origin, distribution, and alteration. Can. J. Soil Sci. 59: 37-58.

Kroetsch, D.J., Geng, X., Chang, S.X., and Saurette, D.D. 2011. Organic soils of Canada: Part 1. Wetland Organic soils. Can. J. Soil Sci. 91: 807-822.

Krzic M., Sanborn, P. Watson, K., Bomke, A.A., Crowley, C., Doree, A., and Dyanatkar, S. 2008. Soil formation and soil processes. The University of British Columbia, Vancouver, University of Northern British Columbia, and Thompson Rivers University, Kamloops. https://processes.soilweb.ca/

Lavkulich, L.M. and Arocena, J.M. 2011. Luvisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 781-806.

Montanarella, L., Pennock, D.J., McKenzie, N., and 26 others. 2016. World's soils are under threat. Soil 2: 79-82.

Miller, J.J. and Brierly, J.A. 2011. Solonetzic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 889-902.

Paton, T.R., Humphreys, G.S. and Mitchell, P.B. 1996. Soils A New Global View. Yale University press, New Haven and London. Pp. 234.

Pennock, D.J., Bedard-Haughn, A., and Viaud, V. 2011. Chernozemic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 719-747.

Sanborn, P. 2016. The imprint of time on Canadian soil landscapes. Quaternary International 418: 165-179.

Sanborn, P., Lamontagne, L., and Hendershot, P. 2011. Podzolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 843-880.

Sims R.A. and Baldwin. K.A. 1991. Landform features in northwestern Ontario. For. Can., Ontario Region. Sault Ste. Marie, Ont. C0FRDA Rep. 3312. Ont. Min. Nat. Resour., Thunder Bay. On. NWOFTDU. Rep. 60. 63 p. Available at: http://cfs.nrcan.gc.ca/pubwarehouse/pdfs/9148.pdf

Smith, C.A.S., Webb, K.T., Kenney, E., Anderson, A., and Kroetsch, D. 2011. Brunisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 695-717.

Tarnocai, C. and Bockheim, J.G. 2011. Cryosolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 749-762.

University of Idaho. College of Agricultural and Life Sciences. The Twelve Soil Orders. https://www.uidaho.edu/cals/soil-orders

VandenBygaart, A.J. 2011. Regosolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 881-887.

Watson, K. and Pennock, D. 2016. Section 3. Soil Profile Description. From: D. Pennock, K. Watson, and P. Sanborn. 2016. Field Handbook for the Soils of Western Canada. Pedology Subcommittee, Canadian Society of Soil Science. Available at https://soilsofcanada.ca/documents/S...escription.pdf

Zaitlin, B. and Hayashi, M. 2012. Interactions between soil biota and the effects on geomorphological features. Geomorphology 157-158:142-152.

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Dan Pennock is a Professor Emeritus (or, more simply, a retired professor) of the Department of Soil Science at the University of Saskatchewan. His research career focused on how soils are shaped by their place in the landscape and by the processes (like erosion or water movement) that occur in landscapes. He taught courses in many areas of soil science and geography, and his love of teaching was recognized by the university with a Master Teacher award in 2006. He was made a Fellow of the Canadian Society of Soil Science in 2010. Since his retirement he has collaborated on a number of projects on sustainable management of soils with the Global Soil Partnership of the Food and Agriculture Organization of the United Nations.

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1.3: Soil Organic Matter

Sylvie Quideau, Myrna Simpson, and Adam Gillespie

Learning Objectives

On completion of this chapter, students will be able to:

1. Define soil organic matter and list its main components
2. Evaluate the importance of organic matter on soil physical, chemical and biological properties and processes
3. Explain the sources and turnover processes of soil organic matter
4. Discuss the spatial distribution and land management of soil organic matter
5. Critically evaluate how soil organic matter is measured and quantified
6. Recognize the role of soil organic matter in global issues

INTRODUCTION

Soil organic matter has been increasingly in the news lately, in part because the vast majority of terrestrial carbon is contained in soils. As such, soils play an important role in the global carbon cycle. In addition to serving as a key carbon store, soil organic matter is one of the central attributes of soils. Organic matter influences virtually all of soil properties and overall health, including its physical structure, nutrient status, and biodiversity. Soil organic matter is composed of plant and animal residues at various stages of decomposition, microbial biomass, and products of microbial synthesis. Organic matter accumulation in soils reflects the balance between plant litter inputs and losses through microbial respiration, erosion, and leaching. Carbon stocks may vary widely among soils and across the landscape as a function of parent geological material, climate, vegetation, and topography.

WHAT IS SOIL ORGANIC MATTER?

Soil organic matter is comprised of both living and non-living components. The living component includes soil macro- and micro-fauna, and soil microbial communities, which may be active or dormant. The non-living portion of soil organic matter is derived from dead plant and faunal inputs into the soil (Figure 3.1). These inputs undergo various biogeochemical processes and are either lost, preserved or transformed in soil. The non-living component of soil organic matter is currently hypothesized to be a complex mixture of different biologically-derived molecules that are at various stages of oxidation and preservation (Schmidt et al., 2011). The chemical structure of soil organic matter has been under debate for many decades and it was previously believed that the non-living component of soil organic matter was made of recognizable plant- or animal-derived biomolecules and humic substances, large by-products that resulted from both biological and chemical decay of organic inputs to soil (Schmidt et al., 2011; Lehmann and Kleber, 2015; Can You Dig It Box 1). A more current paradigm recognizes the key contribution of microbial communities in the formation of soil organic matter (Simpson et al., 2007). As organic inputs are decomposed and processed by microorganisms, the majority of the carbon is respired as CO_2, but some is incorporated into microbial biomass, can interact with clay minerals or is partially degraded and remains in the soil. Consequently, soil organic matter consists of a complex mixture of plant, animal or microbial-derived residues at various stages of decomposition, and of new molecules synthesized by microbes. Soil organic matter can also interact with soil minerals to form organic matter – mineral complexes that contribute to soil aggregation and may increase the stability of organic matter in soil by hindering microbes from physically accessing organic matter substrates that are protected through these organic matter – mineral interactions (Oades, 1988; Baldock and Skjemstad, 2000).

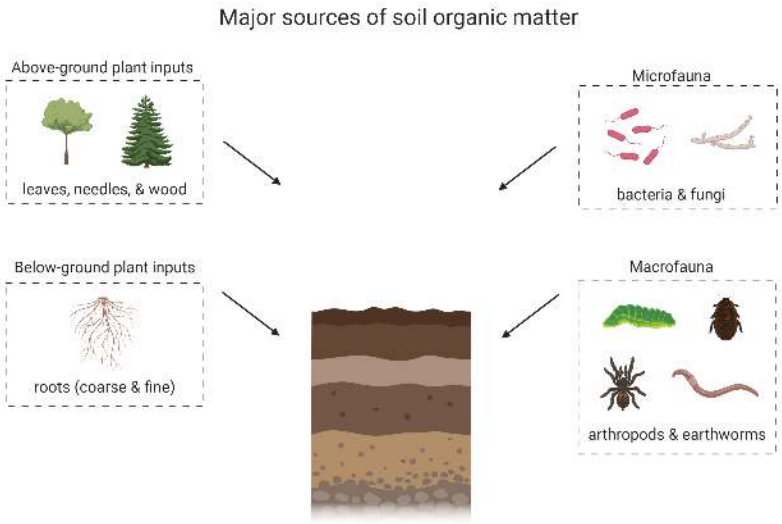


Figure 3.1. Major sources of inputs that form soil organic matter. The types and quantities of inputs vary with different ecosystems and land management practices. © Myrna Simpson. Figure created with BioRender (www.biorender.com). © Myrna Simpson; licensed under a CC BY (Attribution) license.

Organic matter plays several key roles in soil properties and soil processes. The presence of organic matter improves soil stability by promoting aggregation which reduces the potential for soil erosion. Soil organic matter also improves water retention in soil and has a high cation exchange capacity (from 100 to 500 cmol kg^{-1}) which contributes to the total cation exchange capacity of the soil, depending on the amount of soil organic matter present. This improves the soil's ability to retain important cations (Ca^{2+}, Mg^{2+}, K^{+}, and Na^{+}), that can enhance the buffering capacity of the soil. Soil organic matter is also the most biogeochemically active and dynamic portion of the soil and it is important for regulating many processes related to the global cycling of elements such as carbon. Soil organic matter acts as a storehouse and a slow release fertilizer for many plant nutrients, including nitrogen, phosphorus and sulfur. Finally, organic matter supports a large and varied faunal and microbial community.

WHY IS SOIL ORGANIC MATTER IMPORTANT?

Soil organic matter is intimately linked to numerous key physical, chemical, and biological soil properties. It is so important for soil function that it is almost impossible to find a soil property that is not influenced by soil organic matter in some way. Throughout this textbook, every chapter refers to how soil organic matter is influential on the overall expression of soil physical, chemical and biological properties. Table 3.1 shows many of the important components that soil provides to crop production, and how soil organic matter influences these components. In many of these instances, we can see that the properties of soil organic matter allow it to mitigate plant growth limiting conditions and accentuate important plant growth-promoting process in soils.

***Soil Water*:** Soil organic matter has the ability to buffer water supply in soil. High organic matter increases water infiltration during intense rain events, limiting ponding, soil saturation and anoxic conditions which limit plant growth. Soil organic matter also has the ability to retain water in times of drought and water deficit. Together, these two phenomena act to stabilize the hydrodynamics of plant water supply. Sandy soils benefit hydrologically from increased organic matter because they are intrinsically droughty as a result of large inter-particle pore spaces that do not hold water tightly.

***Soil Structure and Compaction*:** Soil aggregation is a strong indicator of healthy soil structure because it reduces stress on root growth and allows for better water infiltration. Indeed, granular structure tends to be encouraged in soils with higher organic matter, whereas soils exhibiting management-induced blocky structure also tend to have shown soil organic matter depletion. Compaction, which is always a risk in heavily managed soils, can be mitigated in soils with higher organic matter levels, particularly in the particulate, or 'light' organic matter fraction (see *Soil Organic Matter Composition* section). Finally, clay soils can be more easily managed when organic matter is high because it reduces the inter-particle cohesion and plasticity.

***Soil nutrient status*:** Organic matter contains a wide range of functional groups that can increase cation exchange capacity. Indeed, soil organic matter may contain between 50 to 90% of the cation-adsorbing ability of a soil. Also macronutrients of nitrogen,

phosphorus and sulphur have very important storage pools in soil organic matter, becoming available to plants through microbial mineralization.

***Soil cation exchange capacity*:** Soil organic matter contributes to the total cation exchange capacity of soil. Soil organic matter cation exchange capacity ranges from 100 to 500 cmol kg^{-1} and is considerably higher than most minerals found in soil. Soil organic matter has a pH-dependent cation exchange capacity due to the dissociation of its functional groups with increasing pH of the soil solution. Cation exchange capacity is important for the ability of a soil to retain important nutrients, which include the major cations Ca^{2+}, Mg^{2+}, K^{+}, and Na^{+} that also contribute to the buffering capacity of the soil.

***Soil biology and energy storage*:** Chemical bonds in soil organic matter contain energy, and this energy is consumed by microbes for growth and metabolism. It is this stored energy that drives microbial nutrient cycling and all the other plant-beneficial associations that soil biology maintains with plants.

Table 3.1. Contributions of soil organic matter to fundamental soil properties

Function component	Function characteristics/processes	Effect of Soil Organic Matter
Soil Structure	Aggregate stability	Promotes formation of stable soil aggregates
	Compaction	Particulate organic matter helps resist compaction
	Erosion	Aggregate stability helps reduce erosion
	Crusting	Helps reduce preconditions to crust formation (i.e., erosion and aggregate destruction)
Nutrient cycling	Accept, hold, and release cations	High cation exchange capacity of soil organic matter improves the ability to retain base cations
	Oxy-anion supply (N, P, S)	Organic N, P and S pools are source of plant available macronutrients via microbial mineralization
Regulate water	Infiltration and retention	Encourages stable and divers pore structures to receive, store, and release moisture for plant use
	Water supply	Adequate water retention to buffer and reduce effects of drought
	Buffers extreme rainfall events	Encourages infiltration to reduce erosion and pooling in extreme rain events
Microbial function	Microbial diversity	Wide range of substrates supports many metabolic pathways, encouraging microbial diversity
	Store and release (recycle) energy	Primary repository of chemical energy used in microbial respiration, drives all major nutrient cycles
	Plant growth promotion	Supports higher levels of microbes which may benefit plant growth through disease suppression.
Medium of plant growth	Seed germination and root growth	Improves seed imbibement and access to nutrients, reduces physical force on growing roots
	Buffer acidity, sodicity	High CEC of soil can buffer and reduce adverse chemistry
Waste management	Sequester biotoxic elements	High CEC can reduce metal mobility
	Degrade organic substances	Supports diverse microbial population that can degrade xenobiotics

WHERE DOES SOIL ORGANIC MATTER COME FROM AND WHY DOES IT PERSIST?

Carbon Inputs to Soils

Soil organic matter can have a diverse range of inputs, which are preserved or transformed through various biological and chemical reactions to form soil organic matter (Schmidt et al., 2011; Kögel-Knabner and Amelung, 2013). Some inputs can also be stabilized in the soil (in the form of soil organic matter or through interactions with clay minerals; Oades, 1988; Kleber et al., 2007) or destabilized and leave the soil through erosion or export with water (Bailey et al. 2019). Soil inputs vary with ecosystem properties and land management practices but predominantly include necromass from soil animals (macro- to micro-scale) and above- and below-ground plant litter (Schmidt et al., 2011; see Figure 3.1). Some soils may also receive inputs such as pesticides, herbicides, fertilizers, and biochar/charcoal from processes related to land management and other anthropogenic activities. Plant inputs are believed to be a major contributor to soil organic matter (Kögel-Knabner, 2002), but some studies suggest that microbial sources are underestimated (Simpson et al., 2007). The nature and quantity of soil organic matter inputs depends on the soil environment as well as ecosystem management. Another consideration is that soil organic matter composition is closely related to soil development (see Chapter 2 for more information) as younger soils may not have as much soil organic matter as older soils which have developed a thick O or A horizon over time, depending on the soil and environmental properties. For example, Chernozemic Ah horizons are more enriched in soil organic matter, especially Black Chernozems which are found in cooler climates. These ecosystem conditions encourage soil organic matter accumulation. In contrast, Ap horizons contain less soil carbon than their native counterpart (Ah) because of agricultural land use. As well, organic matter accumulates in peatlands due to high water saturation and anaerobic conditions that limit microbial activity.

The key factors that control soil formation (abbreviated as *clorpt*) are: climate, organisms (biota), relief (topography), parent material, and time (Jenny, 1941). These also play a role in soil organic matter formation and biogeochemistry. The composition (quality) and amount (quantity) of inputs will vary with different ecosystems and this is controlled by soil forming factors such as climate. Climate also determines the types of biota and also, their activity. Topography can determine the availability of oxygen which is required for the aerobic degradation of inputs as well as stored soil organic matter. Topography can also mediate soil water content, which is required for plant growth. Moisture content may also limit microbial decay of plant inputs and enhance organic matter accumulation, such as in wetlands. The parent geologic material and formation of secondary minerals is also important for soil organic matter because clay minerals can bind soil organic matter which increases soil organic matter stabilization over time.

Climatic conditions such as mean annual temperature and mean annual precipitation are important for determining soil inputs and soil organic matter processes over both short and long-time scales. For instance, temperature and moisture will control the types of plants that inhabit an ecosystem and cooler climates favour coniferous-dominated versus deciduous-dominated trees in forests. The types and quantity of shrubs and grasses that are present are also determined by climate and other soil properties. Figure 3.2 shows the ranges of mean annual temperature and precipitation for different biomes. These parameters also contribute to the quantity and quality of inputs to soil and the extent to which they are transformed as warmer climates favour soil organic matter degradation (through microbial respiration) but also encourage plant growth. Cooler climates may limit plant growth but can also restrict microbial degradation and bioturbation by macrofauna.

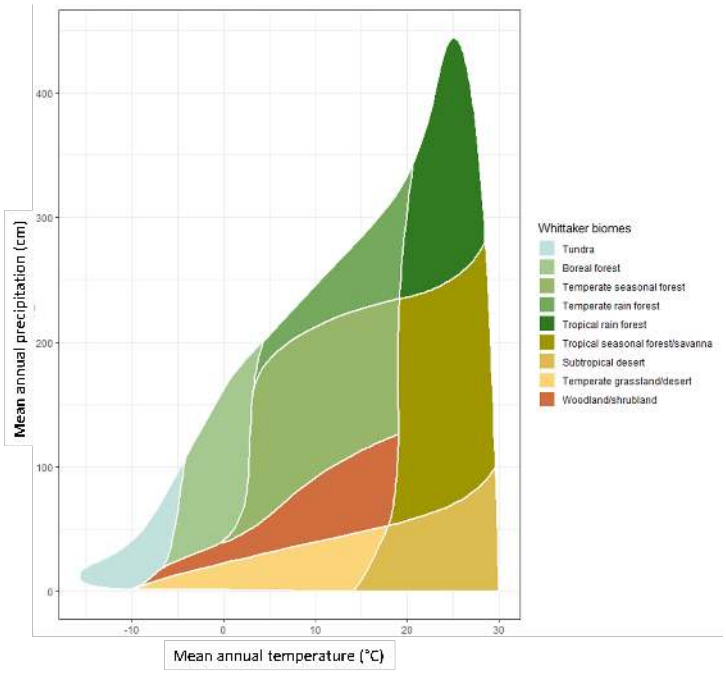


Figure 3.2. Biomes in relation to mean annual temperature and precipitation (also referred to the "Whittaker biome plot"). These ecological properties determine soil organic matter preservation and degradation processes. The plot was created by M. Simpson using R Studio with ggplot2 and the plot biomes package; biomes are based on Ricklefs (2008). © Ricklefs, R. E. 2008; adapted by Myrna Simpson is licensed under a CC BY (Attribution) license.

As described previously (see Figure 3.1), there are varying types of inputs that enter soil and some are preserved via sorptive interactions with clay minerals and some are degraded by native soil biota. The chemistry of soil organic matter inputs has been studied for several decades to help understand soil organic matter formation processes as well as predict soil organic matter stability. The major types of inputs and their sources are summarized in Table 3.2.

Table 3.2. Major soil inputs and their properties that contribute to soil organic matter formation and persistence (Kögel-Knabner 2002; Schmidt et al. 2011; Kögel-Knabner and Amelung 2014)

Input	Source(s)	Chemistry	Stability
Lipids[a]	Plants and microbes	Small molecules	Varies (low-high) with carbon chain length and functional group
Proteins/peptides[b]	Plants and microbes	Macromolecules	Medium
Cellulose (hemicellulose)	Plants and microbes	Macromolecule	Low
Lignin	Plants	Macromolecule	Medium-high
Chitin	Fungi & Arthropods	Macromolecule	Medium-high
Cutin	Plants (waxy coating on leaves)	Macromolecule	High
Suberin	Plant roots	Macromolecule	Low-High
Char/biochar	Pyrolyzed plant material	Macromolecule	High

[a]Lipids are a wide range of compounds that have ranges of carbon lengths as well as functionalities. Some compounds in this group have alcohol or acid functionalities.

[b]This encompasses a large group of molecules that have varying chemical properties and stability

Degradation of Carbon Inputs

Soil organic matter inputs (Table 3.2) have varying chemistry, which can contribute to their susceptibility for microbial degradation after being introduced into the soil environment. Some inputs are degraded more rapidly than others. Many of the inputs are large molecules (macromolecules) that decompose in stages. For example, cellulose is a complex carbohydrate but once its macromolecular structure is broken down, it produces simple sugars that can be readily degraded. Lignin, which is another macromolecule composed of a network of aromatic (phenolic) components, is more difficult to break down and is primarily degraded by fungi. Therefore, decomposition of plant inputs is a function of the chemistry of the inputs, the availability of microbes and enzymes, as well as nutrient content, and other environmental factors (such as oxygen and water).

Under aerobic conditions, organic matter inputs are degraded via respiration (Equation 1.3). Respiration, also referred to as mineralization, is an oxidation reaction where oxygen acts as the primary electron acceptor (Bohn et al. 1985). The net products of this reaction (Equation 1.3) are carbon dioxide and water. As shown, a basic form of organic matter inputs (CH_2O) yields 4 electrons (Equation 1.1), which then require the equivalent amount of oxygen (Equation 1.2) to balance the net reaction (Equation 1.3):

$$CH_2O + H_2O \rightarrow CO_2 + 4e^- + 4H^+ \quad (1.1)$$

$$O_2 + 4e^- + 4H^+ \rightarrow 2H_2O \quad (1.2)$$

$$CH_2O + O_2 \rightarrow CO_2 + H_2O \quad (1.3)$$

More complex forms of organic matter require more chemical energy for degradation (Bohn et al. 1985). For example, if empirical formulae are considered for both lignin ($C_{2.9}H_{2.9}O$) and cellulose ($C_{1.2}H_2O$), it is evident that the degradation of plant material (estimated as two-thirds cellulose and one-third lignin for the purposes of this example; Equation 2.1) will require less chemical energy and less oxygen (which acts as a terminal electron acceptor) than soil organic matter (empirically represented as $C_{2.2}H_{2.2}O$; Equation 2.2). This requirement is one reason why more complex forms of organic matter are stabilized over other forms and persist longer in soil (Table 3.1).

$$\text{Plant matter decomposition: } C_{1.7}H_{2.2}O \rightarrow 1.7C^{4+} + H_2O + 7e^- + 0.2H^+ \quad (2.1)$$

$$\text{Soil organic matter: } C_{2.2}H_{2.2}O \rightarrow 2.2C^{4+} + H_2O + 9e^- + 0.2H^+ \quad (2.2)$$

Inputs from plants and microbes are respired to CO_2 based on oxidation as shown in Equation 1.3. But the rate at which respiration occurs is dependent on ecosystem properties, microbial biomass and diversity, and nutrient availability. Figure 3.3 shows the fate of fresh plant material in the soil carbon cycle. The addition of fresh plant residues often fuels microbial activity (referred to as soil priming; Kuzyakov et al. 2000), which can then result in continued degradation of stored soil carbon based on energetics described in Equations 2.1 and 2.2. This may then in turn destabilize stored soil carbon and reduce soil carbon stocks. In some cases, fresh plant material decomposition may also increase soil carbon stored (recall, soil carbon includes both the living and non-living components). Therefore, how fresh plant additions alter soil carbon can vary but this is dependent on the quality of the inputs, presence of microbial degraders and enzymes, oxygen and water content, and time. In the example highlighted in Figure 3.3, fresh plant carbon is partly respired to CO_2 and also stored in the form of soil carbon. The total amount of carbon doesn't change (carbon balance exists) but the form of the different carbon pools (atmospheric CO_2, soil organic matter, soil microbes, etc.) varies with time, space and ecosystem properties. For example, this process may occur more quickly in aerobic soils in warmer regions of the world. In other ecosystems, the time required for plant carbon to be converted into other carbon forms can be much longer (several decades at least). Under anaerobic conditions, the process of respiration (conversion of plant carbon to CO_2) is much slower due to the limited availability of O_2 (see Equation 2.1). Consequently, the time scale for plant carbon conversion into other forms varies with climate and other ecosystem properties, as well as the quality of plant carbon.

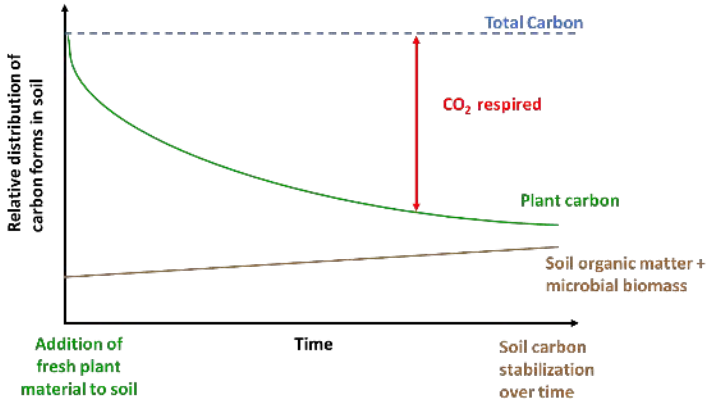
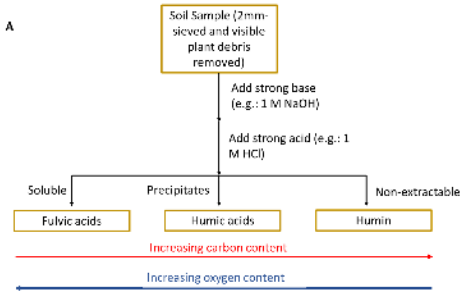


Figure 3.3. Illustration of what happens to freshly added plant carbon when it enters the soil. Over time, plant carbon is converted to soil carbon (which includes both non-living soil organic matter and carbon found in living soil organisms) and respired carbon (CO_2). The time it takes to convert plant carbon into other forms is dependent on the ecosystem properties and availability of oxygen (see Equation 1). © Myrna Simpson; licensed under a CC BY (Attribution) license.

Can You Dig It!

Humic substances



Traditional method of isolating humic substances. © Myrna Simpson; licensed under a CC BY (Attribution) license.

It was historically believed that the long-term stability of soil organic matter was due to the conversion of plant and animal inputs into humic substances through different biological and chemical reactions. The traditional fractionation method used to isolate different components of soil organic matter (Schmidt et al. 2011; Lehmann and Kleber 2015) uses a series of strong bases and strong acids to separate out soil organic matter based on its solubility (see side Panel A).

Part of the soil organic matter was not extractable in either strong base or strong acid. The resulting fractions were then characterized to learn more about soil organic matter composition (see side Panel B). These humic substance fractions (fulvic acid, humic acid and humin) were historically analysed using elemental analysis and other low-resolution chemical analysis methods. This resulted in soil organic matter scientists interpreting that these fractions were of high molecular weight (large macromolecules). It was then believed that these large macromolecules were stable in soil in the long-term and the main process by which soil inputs are stabilized in soil environments.

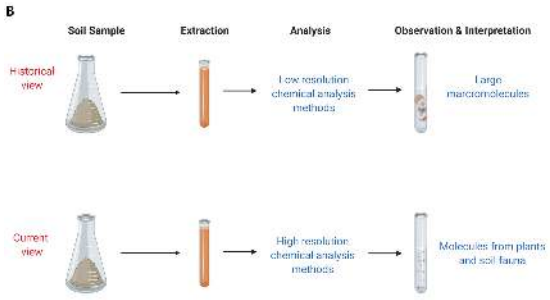


Illustration of the classical and current techniques to isolate different humic substances. Prepared using BioRender (www.biorender.com) and based on Schmidt et al. (2011). © Myrna Simpson; licensed under a CC BY (Attribution) license.

Starting in the late 1990s, the advancement of chemical analysis methods allowed scientists to obtain high resolution data about the chemistry of soil organic matter (Kelleher and Simpson 2006; Schmidt et al. 2011). These methods showed that soil organic matter was actually made of recognizable biomolecules from plants and animals at varying stages of degradation (see Panel B). These methods also showed that these biomolecules can associate with each other and behave like larger molecules (consistent with the historical view). Scientists are further exploring the current view and how this new information can be used to predict and explain the stabilization of soil organic matter in different environments and with climate change.

Persistence of Soil Organic Matter

Many inputs enter soil and are transformed into soil organic matter or mineralized to CO_2. But some soil organic matter can reside in soil for decades to hundreds of years. Part of this long-term persistence is due to the strong interaction between soil organic matter and clay minerals found in soil (Oades 1988; Schmidt et al. 2011; Simpson and Simpson 2012; see Figure 3.4). Clay minerals can stabilize soil organic matter through the formation of organic matter-mineral complexes in soil (Figure 3.4b). Both amorphous and crystalline clay minerals can bind soil organic matter but to varying extents and via different chemical interactions such as covalent, ionic or weak van der Waals interactions (Feng et al. 2005; Kleber et al. 2007). Although the manner in which soil organic matter interacts with clay minerals differs in each soil environment, it is clear that the binding of organic matter to clay minerals increases its persistence (Baldock and Skjemstad 2000). Once bound to clay minerals, soil organic matter is less available for biological degradation because microbes cannot gain easy access to organic matter sequestered on clay surfaces. As such, soil organic matter bound to clay minerals is more persistent in the long-term as it is not as readily used as a microbial substrate. Other mineral surfaces that facilitate organic matter persistence are iron and aluminum oxides in acidic soils. In neutral and alkaline soils, the presence of polyvalent cations such as calcium and magnesium stabilizes organic matter by forming bridges between negatively charged organic molecules and clay surfaces. The type of clay minerals is also important in controlling how much carbon can be ultimately stored in a soil. Clay minerals with higher surface areas (2:1 clays) have a higher ability to adsorb carbon than 1:1 clays. In Canada where soils are still relatively young, clay mineralogy mostly reflects that of the geological parent material.

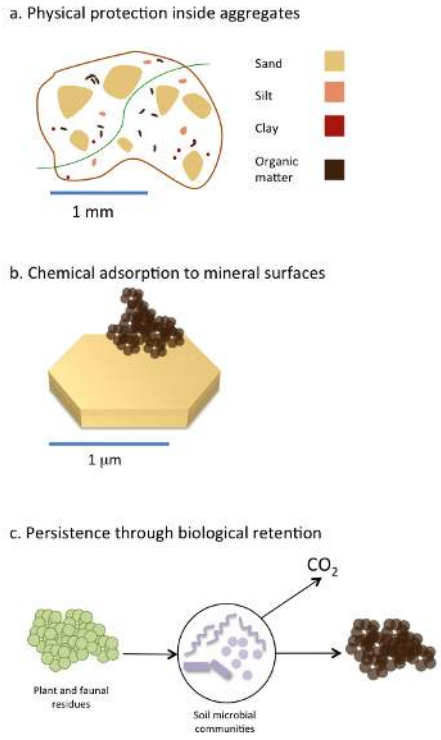


Figure 3.4. Mechanisms of organic matter stabilization. © Sylvie Quideau; licensed under a CC BY (Attribution) license.

In addition to the interaction of organic matter with mineral surfaces, physical protection inside aggregates (Figure 3.4a) is considered to be an important mechanism for organic matter persistence (Six et al. 2002). In that case, organic matter mostly persists in soils because it is spatially separated from the decomposing microbial communities. Furthermore, the lower availability of oxygen within aggregates may become a limiting factor to microbial activity. In boreal ecosystems, which cover the majority of

Canada's land area, fire is a recurrent disturbance that produces pyrogenic carbon, often referred to as black carbon or char. Some pyrogenic carbon, containing highly condensed aromatic structures, may persist in soils and constitute a long-term carbon sink (Ohlson et al. 2009). However, the persistence of pyrogenic carbon in soils greatly varies depending on its formation conditions (Soucemarianadin et al. 2015)

Traditional views highlighted the importance of organic matter chemical structure in explaining its persistence in soils. A newer paradigm questions the validity of this assumption and suggests instead that molecular structure alone does not explain organic matter accumulation and stability in soils. In addition, microbial metabolism, and more specifically carbon utilization efficiency plays a role in determining organic matter persistence (Cotrufo et al. 2013). Under this new model (Figure 3.4c), it is proposed that more labile molecules, because they are preferentially taken up by microbes hence retained with the soil, constitute the main portion of stable soil organic matter.

HOW CAN WE MANAGE SOIL ORGANIC MATTER?

Environmental Factors

Soil carbon storage results from the balance between carbon inputs and outputs. Carbon sequestration occurs when aboveground and belowground (*i.e.*, leaf and root-derived) inputs exceed outputs, which are usually dominated by CO_2 fluxes from the soil surface, but may also include in some instances important contributions from methane (CH_4) and leaching of dissolved organic carbon (DOC). Plant residues are the major substrate for soil organic matter formation. When microbial communities decompose plant litter, the majority of carbon contained in these residues is eventually evolved as CO_2. However, a small portion remains in the soil as soil organic matter, sometimes for centuries or even millennia. Figure 3.5 shows the distribution of soil carbon (related to soil organic matter) of the soils of Canada. The intricate balance of these inputs and outputs, as well as organic matter stabilization to clay minerals, is what determines soil organic matter persistence.

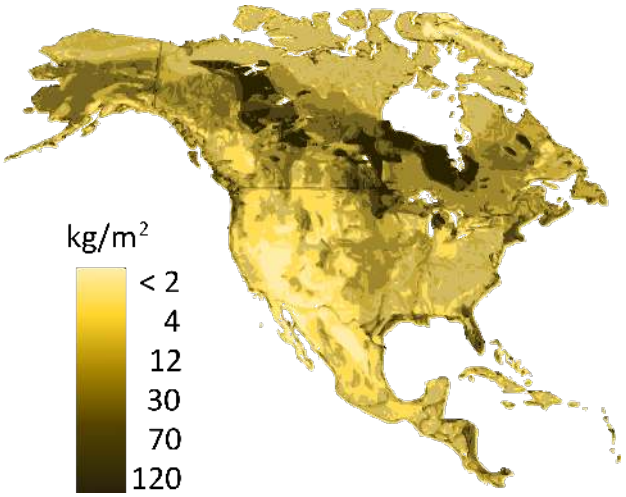


Figure 3.5. Soil carbon distribution from soils in North America. Adapted and generalized from USDA sources © Adam Gillespie. CC-BY. © USDA adapted by Adam Gillespie is licensed under a CC BY (Attribution) license.

The rate of carbon accumulation varies greatly among soils, which reflects the influence of environmental factors; *i.e.*, soil forming factors (climate, organisms, parent geological material, relief and time) on pedogenic processes. Climate directly affects soil carbon stocks by controlling the balance between litter inputs and decomposition outputs. Increasing precipitation increases plant growth and litter production, and decreasing temperature decreases decomposition losses. At the global scale, carbon accumulation in soils is more influenced by temperature than precipitation and tends to increase in high-latitude soils where low temperatures limit decomposition. The largest stocks of carbon are found in the northern permafrost region, where both Organic soils and Cryosols are present. In arid environments, water availability may limit plant growth and plant residue inputs to soils, and consequently lead to lower soil carbon stocks.

Across Canada, soil carbon stocks also vary as a function of precipitation, where high local precipitation along both the Atlantic and Pacific coasts allows for the establishment of dense forest vegetation and consequently, large surficial litter inputs. In addition, placement of the vegetation inputs directly affects the distribution of carbon within the soil profile as well as total soil carbon

stocks. In grassland soils where belowground (root) inputs can be plentiful, carbon stocks can be greater and not decrease as sharply with depth as they do under forest vegetation. Lastly, while vegetation may be the main factor controlling surface organic carbon, the soil parent geological may also be a key factor for carbon sequestration in the mineral soil profile. Soil carbon and clay content are positively correlated, both because of the formation of organic matter-mineral complexes resistant to decomposition losses, and because of the indirect effects of clay that promotes greater fertility and plant growth, hence increases carbon returns to the soil.

At the landscape level, organic carbon accumulation is favoured in lower slope positions where soil conditions are wetter and tend to inhibit decomposition losses. Similarly, in the northern hemisphere, soil carbon stocks tend to be greater on north-facing slopes compared to south-facing slopes, where lower temperature and lower evaporation increase soil moisture and decrease decomposition.

Average carbon stocks in the different mineral soil orders found in Canada vary from around 11-12 g m^{-2} for Regosols, Brunisols, Luvisols, Chernozems and Solonetzs to 16-20 g m^{-2} for Podzols and Gleysols. In forest soils, carbon stocks increase from well-drained Luvisolic and Brunisolic soils to poorly drained Gleysolic and Organic soils (Figure 3.6). Carbon stocks in Organic soils may be an order of magnitude higher than in other soil types and average 134 g m^{-2}. Within mineral soil orders, Cryosols contain the highest stocks averaged at 41 g m^{-2}. Cryosols also occupy the largest area in Canada, as they constitute more than one third of Canadian soils. Cryosols and Organic soils contain roughly the same amount of carbon, which is by far the largest amount compared to other soil orders. Taken together, these two soil orders hold about 80% of the total carbon stocks contained in Canadian soils. Podzols and Brunisols contain 5-10%, and the remaining orders < 5%.

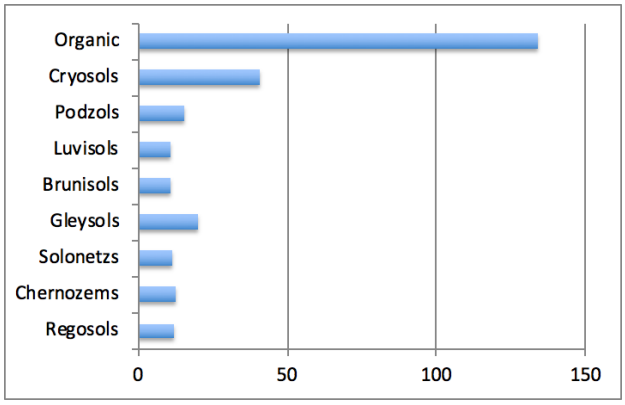


Figure 3.6. Carbon stocks (kg m^{-2}) in different Canadian soil orders (adapted from Shaw et al. 2008; Tarnocai 2008). Adapted by Sylvie Quideau and licensed under a CC BY (Attribution) license.

Pedogenic processes (i.e., the processes of soil formation) can directly affect carbon stocks and carbon stability in soils. In turn, soil carbon content and soil classification are directly related for many soil types. The most obvious example may be the case of Organic soils, where organic matter accumulation is at the core of their formation and classification. Paludization, the accumulation of thick organic layers on the mineral soil surface, creates the fibric (Of), mesic (Om), and humic (Oh) horizons representative of Organic soils. In upland forests, littering, the thinner, surficial accumulation of fresh plant residues, is responsible for forest floor accumulation, which can be separated into three distinct horizons (LFH) based on the degree of decomposition (see Chapter 8 for further details).

Humification, the formation of humic substances from plant residues, and melanization, the darkening of a soil color due to carbon addition, are both responsible for the accumulation of carbon within the top mineral soil layers, in particular in Chernozemic horizons. The formation of these horizons is favoured by the extensive network of grass roots that contributes large exudate and residue inputs. Bioturbation by earthworms and stabilization of organic matter by calcium are two common processes contributing to carbon accumulation in Canadian Chernozems. The extent of darkening is used to differentiate between great groups of Chernozems and Solonetzs, with in increasing amount of carbon: Brown, Dark Drown and Black great groups. The absence or presence of a carbon-rich Ah horizon is also used to differentiate great groups of Brunisols, Luvisols, Gleysols, and Regosols; i.e., Melanic and Sombric Brunisols, Gray Brown Luvisols, Humic Gleysols and Humic Regosols. Lastly, carbon content is used to separate subgroups of several additional mineral soil orders. For instance as opposed to the processes just mentioned, podzolization results in carbon accumulation at depth. Mostly, during podzolization, carbon is added from soluble organics leached from the soil

surface. The other carbon input at depth would result from root exudates or root turnover. Podzolic great groups are partially differentiated based on their carbon content, with the Ferro-Humic Podzols containing more carbon in their subsoils than the Humo-Ferric Podzols.

Lastly, processes that are not directly linked to humification may nonetheless increase soil carbon. Gleization, which results in anaerobic conditions, may lead to the accumulation of carbon in Gleysols. Extreme cold, which also inhibits decomposition, can result in thick surficial carbon accumulation characteristic of Organic Cryosols, and, in the case, of Turbic Cryosols, in carbon accumulation in the mineral horizons, which can also be found at depth due to cryoturbation, the process of soil mixing due to freezing and thawing.

Land Management

In Canada, conversion of native land to arable agriculture resulted in a loss of about 25% of the carbon present in the soil surface layers (0-30 cm) prior to the onset of cultivation. In other words, carbon lost over the 19th century from Canadian cultivated grassland soils would have resulted in an emission of about 1 Pg (Janzen et al. 2018).

Carbon losses may be attributed to removal of crop residues, as well as to increased erosion and decomposition. In some cases, fire was used to clear land, which resulted in increased losses. Increased decomposition, rather than erosion is typically he main culprit following cultivation, and results from the physical disruption of soil aggregates by tillage, which exposes previously inaccessible organic matter to the degradative actions of soil microorganisms. The majority of carbon lost upon cultivation consists of the light fraction, which is composed of still “young”, only partially decomposed plant residues. Most of the loss occurs rapidly, *i.e.*, during the first twenty years of soil conversion to arable agriculture.

Because of past losses of carbon from agricultural soils, these have significant capacity to store more carbon. Several management practices can promote soil carbon gains. Specifically, some can decrease soil carbon losses including: reduction in tillage intensity and reduction in summer fallow. Some measures contribute by increasing carbon inputs to soils, either by adding carbon from other sources (e.g. manure, biochar), or by promoting greater plant inputs to soils (e.g., by growing cover crops, improving residue retention, increasing use of perennials, and optimizing fertilization). Chemical fertilization (i.e., addition of inorganic fertilizers), by promoting plant growth, typically results in carbon accumulation in soils. The annual carbon storage rate for soils receiving inorganic fertilizers has been estimated at 23 ± 13 g C m^{-2} yr^{-1} (VandenBygaart et al., 2004). Application of organic amendments such as animal manure or biosolids (human waste) also leads to an increase in soil carbon, especially when applied regularly over many years or even decades. However, this may also increase the generation of greenhouse gases, in particular nitrous oxide. An upper limit in soil carbon sequestration may be reached, which corresponds to the fact that the soil matrix may become saturated with carbon as it has reached its full storage capacity. Fine-textured soils have a higher ability to store carbon than coarser-textured soils.

Conservation tillage, including reduced and no tillage, can promote soil carbon accumulation, although recent studies have shown that the increase is likely confined to the upper layer of soil (0-10 cm), and that tillage has no effect on carbon stocks integrated over the full 0-130 cm depth interval (Mary et al., 2020).

Soil organic matter can also be impacted by other anthropogenic activities. It is well documented that conversion of native lands for use in farming can lower soil organic matter content, especially with tillage. The incorporation of crop residues, as well as conservation tillage practices, is being used to preserve soil organic matter in agricultural lands. Crop residues are applied to the soil surface after harvest and are reincorporated into the soil organic matter pool. The application of charred residues (biochar or through slash and burn agriculture) improves both the organic matter content and cation exchange capacity of the soil and alters the soil pH, all of which benefits both plants and soil microbes. Char can be long-lived in soil, and though the benefits of its application vary with time as the composition of the char can change, the addition of char does contribute to the overall soil organic matter continuum found in managed soils.

While deforestation typically results in a loss of soil carbon, afforestation may, at least in part, restore the original soil carbon levels. Reforestation in the boreal zone results in smaller rates of soil carbon increase compared to other climates, likely because tree growth rates are slower there (Laganière et al., 2010). Previous land use and the type of disturbance is also an important factor controlling the rate of soil carbon increases following afforestation. Afforestation of agricultural lands yields to greater increases in soil carbon stocks compared to pastures or grasslands (Mayer et al., 2020). Afforestation of soils disturbed by industrial activities, such as surface mining, can also lead to a significant increase in soil carbon levels (Prescott et al., 2019). If suited to the site

conditions, deciduous trees help to increase soil carbon stocks faster than conifers do; in the long run, deciduous trees also contribute to higher carbon stocks in the mineral soils compared to the forest floors.

Table 3.3. Effects of land use changes on soil organic carbon stocks. Compiled from VandenBygaart et al. (2004); Horwath and Kuzyakov (2018); and Janzen et al. (2018)

Management practice	Location	Carbon storage rate	Δ soil carbon
		(g C m^{-2} yr^{-1})	(%)
Grassland conversion to agriculture	Eastern Canada		-30 to -36
Grassland conversion to agriculture	Western Canada		-11 to -59
Forest conversion to agriculture	Eastern Canada		-4 to –49
Forest conversion to agriculture	Western Canada		-11 to –36
Overall conversion to agriculture	Canada		-24 ± 6
Organic amendments	World	+ 40 to 62	
Chemical fertilizers	Canada	+ 23 ± 13	
Chemical fertilizers	World	+ 16 to 32	
Chemical fertilizers + organic amendments	World	+ 33 to 52	
Conventional to no tillage	Canada	+ 5 to 16	
Cover crops	World	+ 15 to 23	
Afforestation of agricultural soils	World	+ 9 to 51	

HOW DO WE MEASURE SOIL ORGANIC MATTER?

There are two major categories of soil organic matter analysis: total elemental analysis; and chemical speciation. Both groups of analyses are important for answering a range of environmental and agronomic questions. Importantly, soil organic matter is related to soil organic carbon but these are not the same measurement. This is an important distinction because the chemical character of soil organic matter is important for its environmental persistence, behavior, and turnover dynamics. Soil organic matter content is almost universally reported in basic soil tests, and new soil health analytical packages are now reporting results which target rapid-cycling soil organic matter fractions to track management improvements. There are, however, several different approaches and techniques used to operationally measure soil organic matter, and they are not equivalent. This section of the chapter discusses and compares the common methods of measuring soil organic carbon and soil organic matter and presents considerations to keep in mind when choosing methods and interpreting results.

Table 3.4. Common methods of measuring soil carbon

Method	Measures	Details	Strengths	Disadvantages
Dry combustion	Carbon	Sample combusted in O_2 atmosphere	Direct measure of soil C; can be automated; can measure N as well	Carbonates in calcareous soils will confound this measurement
		CO_2 evolved is measured directly, usually >1000ºC		
Loss on ignition (LOI)	Organic matter	Sample combusted in O_2 atmosphere	Simple equipment	No discrimination of C from other components of organic matter
		Measures mass loss	Bulk analysis	Structural water contributes to mass loss
		Usually ~500ºC		
Wet oxidation	Organic matter	Chemically oxidized with chromate ($Na_2Cr_2O_7$)	Targets organic matter only	Chromate is carcinogenic
		Back-titrated with Fe(II) or measured spectroscopically	Carbonate does not react	Reaction is incomplete
				Chromate reacts with other soil components

Dry Combustion

Dry combustion is a widespread technique used to determine total carbon and nitrogen simultaneously. Air dried and pulverized soil samples (usually <250 µm diameter) are introduced into a combustion furnace at about 1250°C. The samples are combusted in a stream of O_2, producing CO_2 and N oxide gases (*i.e.*, NO_x). The pulse of O_2 serves to ensure that all the C and N are oxidized. Following combustion, the gases are swept out of the combustion tube in a flow of helium into a reduction column filled with chopped copper wire. This column is held at 600°C and serves to scavenge O_2 from any NO_x in the gas stream, thus converting all N to N_2. Following the reduction column, the gas stream is dried by passing through a water trap containing magnesium perchlorate. The gas stream is then passed into a packed column gas chromatograph, which separates the CO_2 from N_2, with the N_2 exiting the column first. CO_2 and N_2 are detected using thermal conductivity detectors or using IR detectors.

Modifications of these methods can include a separate pretreatment to measure any C present in carbonates by neutralization of carbonate with acid and back titration of the excess acid, which is then subtracted from total C to give organic C. Alternatively, carbonates may be destroyed by treatment with dilute acid before C determination. Other laboratories use combustion at 860°C to selectively decompose only organic carbon. Instrumental methods featuring automated versions of DC methods are now routinely used to determine total C and organic C in many laboratories.

Loss on Ignition

The Loss on Ignition (LOI) is a measure of the change in weight as a result of heating to drive off volatiles and/or remove components by thermal decomposition and/or burning in an oxygen-rich atmosphere. At any temperature, the LOI will be the sum of several processes, including:

1. Loss of moisture adsorbed on to mineral grains (mostly removed by heating at 105°C).
2. Loss of structurally-bound water from within mineral structures (mostly removed by heating to between 600 and 1000°C).
3. Oxidation of organic matter, generally at ≤ 500°C, releasing CO_2, CO and H_2O.
4. Thermal decomposition of carbonate minerals (e.g. magnesite at 750 to 950°C or calcite at 850 to 950°C) releasing CO_2.

5. Oxidation of metals (in particular Fe^{2+}) in an oxygen-rich atmosphere adding iron oxides.

Wet Oxidation

Also known as the 'Walkley Black' method, this uses chromate to chemically oxidize organic C in the sample. Oxidation is facilitated by heat generated from the addition of sulfuric acid. The excess chromate is determined by titration, and the quantity of carbon that was oxidized is calculated from the amount of chromate reduced. This analytical procedure used to be the standard method for measuring soil organic carbon but suffers from several limitations. There is incomplete oxidation of the carbon present in the soil samples, and a correction factor is needed. However, this correction factor may vary among soil types, which leads to bias in the results. For instance, carbon present as charcoal may not be included, and would lead to underestimates of total soil carbon.

Soil Organic Matter *vs* Soil Organic Carbon and the Conversion Issue

Soil organic carbon (SOC) and soil organic matter (SOM) are not the same quantity. It has, however, become an automatic practice to convert between SOC and SOM using Equation 3:

(3) $\%SOM = 1.72 \times \%SOC$

This assumes that carbon molecules comprise 58% of the total mass of soil organic matter. This is based on results reported by Waksman and Stevens (1930), and even they reported that this was a conversion factor that should be regarded with caution. Soil organic matter also contains hydrogen- and oxygen-containing functional groups, which contribute to the overall soil organic matter mass. While it also is assumed that hydrogen contributes *ca*. 4% and oxygen contributes *ca*. 30% to the overall composition, we know that this depends on the decomposition state of soil organic matter. Indeed, hydrogen and oxygen indices that were originally applied to petroleum characterization (Lafargue et al., 1998) can also be used to infer decomposition status in soils (Gregorich et al., 2015; Sebag et al., 2006). Soil organic matter is also an important store of macro-elements crucial in plant nutrition (*i.e.*, N, P, and S). N, P, and S are usually present in relatively low concentrations in soil organic matter (<5% for N, <2% for P and S). A critical review of the commonly-used conversion factor (Pribryl, 2010) and showed that the majority of studies through the years report that this conversion ought to be higher. Figure 3.7 shows that actual relationships between soil organic matter and soil organic carbon vary significantly from the standardized correction, and typically favour higher conversion factors (*i.e.*, lower proportion of C) in soil organic matter.

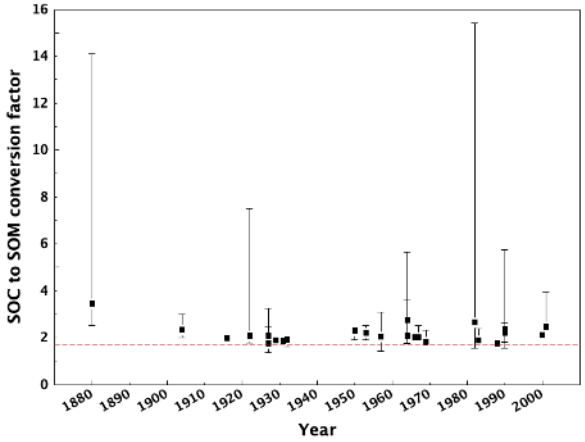


Figure 3.7. Historical comparison of factors used to convert % soil organic carbon to % soil organic matter. Year refers to year in which papers were published; the red horizontal line marks the commonly-used conversion factor (1.72) as a reference. Adapted from Pribyl (2010); adapted by R. Farrell and licensed under a CC BY (Attribution) license.

The most effective approach to converting between soil organic matter and soil organic carbon would be to conduct separate direct analyses. Otherwise, be sure to clarify which values were analyzed, and how conversions were carried out.

The Carbonate Issue

Paleozoic carbonate-based (limestone) bedrock is common in many agricultural regions in Canada, mostly in the Prairie provinces and along the Great Lakes and St Lawrence river lowlands. In some areas, particularly those experiencing tillage erosion, carbonates are present in surface soils as well as in parent material horizons. Problems arise when quantifying soil organic carbon

in soils because combustion-based analytical methods will decompose carbonates to CO_2, thus inflating the overall estimate of organic carbon. Three main approaches are used to differentiate carbonate-C from organic C: heat treatment; acid treatment and; wet oxidation (Table 3.5).

Table 3.5. Methods for sample pretreatment to account for inorganic C (i.e., carbonates) when analyzing soils for total or organic C

Method	Specifics	Benefits	Disadvantages	
Heat Treatment	Indirect organic carbon measurement	Simple, bulk processing possible	Some carbonates decompose at lower temperatures	
	Muffle furnace at 500ºC		Some soil organic carbon resistant at 500ºC	
	Calculate soil organic carbon by difference			
Acid treatment	Acidify with HCl	Simple, bulk processing	Reduced lifetime of C analyzer	
	Acidify with H_2SO_3 (sulfurous acid)	Safer for analysis	Requires special evaporative apparatus	
Wet oxidation	Redox with chromate to target only organic carbon	Targets organic carbon only	Chromate is carcinogenic	
			Redox reaction with other soil components	

Heat treatment is used in many commercial labs. In this scenario, the soil sample is split into two subsamples. The first half is measured directly by combustion (see *Dry Combustion* section) giving a measure of total carbon for that soil. The second half is placed in a muffle furnace at *ca*. 500°C, and this removes organic carbon only, leaving behind the inorganic C. This sample is then measured by combustion, and the organic C is calculated as the difference between the two values. This approach has merit because the common carbonate minerals in Canadian soils have decomposition temperatures above 500°C (*i.e.*, calcite > 700°C, dolomite > 650°C). There are some soils, however, which contain siderite with a lower decomposition temperature around 440°C.

Acid treatment is frequently used because it is inexpensive and does not require specialized apparatus. Acid decomposes carbonates to water and CO_2. Two acids are used in laboratory settings: hydrochloric and sulfurous. Hydrochloric acid is relatively easy to work with, but it will reduce the lifetime of carbon analyzers. Sulfurous acid offers the benefit of being volatile, and so evaporates after use. It does, however, require proper ventilation and care in use (Baldock and Skjemstad, 2008).

Wet oxidation, as mentioned above, is a method that targets organic carbon only. The carbonates which may be present in the sample do not participate in the reaction because they already are fully oxidized. This method, however, is based on the use of chromic acids, which are carcinogenic, and may also participate in other redox reactions, thus biasing estimates of organic C.

Soil Organic Matter Composition

In addition to measure soil carbon forms for quantifying soil carbon, soil organic matter composition can also be measured. In many cases, the composition of the soil organic matter is important for better understanding the cycling of carbon, nitrogen and oxygen as well as soil health. Therefore, it is becoming more important to supplement soil carbon measurements with more advanced measures of soil organic matter composition. However, measuring soil organic matter composition is challenging because of the nature of soil organic matter. As reviewed previously in this chapter, soil organic matter is a complex mixture of compounds that are of various sizes, polarity, and stage of oxidation (biodegradation). This continuum of organic compounds makes it challenging to isolate and characterize the composition of soil organic matter which is why in many cases, scientists use multiple techniques that provide atomic or molecular level information (Kögel-Knabner, 2000; Simpson and Simpson, 2012; Derenne and

Nguyen Tu, 2014). For this reason, in many cases the soil organic matter is extracted into different fractions. One common and more traditional approach is the fractionation based on solubility in strong base and strong acid, which yields three fractions (fulvic acid, humic acid and humin) that can then be characterized using other techniques. The criticism of this method is that these fractions are not representative of any biogeochemically relevant processes that occur in soil environments (Lehmann and Kleber, 2015). Consequently, another method of fractionating soil organic matter is using density-based methods which separate particulate organic matter (which is not bound to clay minerals) and organic matter associated with different size fractions (for example: sand, silt and clay sized fractions; (Gregorich et al., 1996). Other studies have emphasized the physical fractionation method because it is believed to be more representative of soil organic matter biogeochemical cycling (Mikutta et al., 2006; Sollins et al., 2009; Hatton et al., 2012). For example, the particulate organic matter fraction (also called the light fraction) is fast cycling and accessible to microbes whereas the clay fraction holds soil organic matter that may not be biogeochemically active and is the portion of soil organic matter that is more persistent. Different sized soil aggregates can also be isolated and studied, depending on the research questions or objectives of the analysis.

There are many different advanced and sophisticated techniques that have been developed over the last several decades and applied to study the composition of soil organic matter in whole soils and soil fractions. These methods have improved our understanding of soil organic matter chemistry and persistence in the environment. Methods that use mass spectrometry and nuclear magnetic resonance spectroscopy, two commonly used tools in chemistry, have enabled the understanding of the molecular level composition of soil organic matter (Kögel-Knabner 2000; Simpson and Simpson 2012; Derenne and Nguyen Tu 2014; Simpson et al. 2018). Mass spectrometry techniques are coupled with chromatography methods which separate components before they are analyzed in the mass spectrometer. This has allowed researchers to unravel some of the molecular complexity of soil organic matter and also identify specific compounds within the mixture. Mass spectrometry is also used to measure different stable and radiogenic isotopes of important elements such as carbon, nitrogen, and oxygen. Stable isotopes can provide information about the different plant sources and degree of microbial processing of organic matter (through measuring isotope fractionation (Whalen et al. 2014; Kohl et al. 2015)) whereas radiogenic isotope measurements of carbon (radiocarbon) provides information about the age of organic matter and can be used to better understand the mechanisms of persistence (Trumbore and Druffel 1995; Quideau et al. 2000).

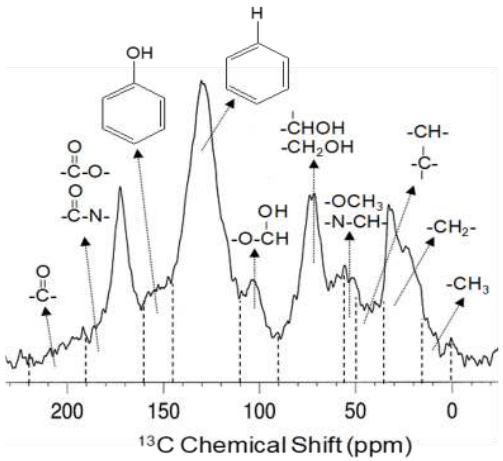


Figure 3.8. SOM characterization by Nuclear Magnetic Resonance Spectroscopy. From Simpson and Simpson (2012). Reproduced with permission and licensed under a CC BY-NC (Attribution NonCommercial) license.

Nuclear magnetic resonance spectroscopy is another method that has provided unparalleled insight into the composition of soil organic matter. Nuclear magnetic resonance spectroscopy can analyze samples in the solid-state or solution-state which is ideal for the analysis of whole soils and various soil fractions (insoluble or soluble) because this technique provides molecular information about all the structures present (Figure 3.8). The Canadian Light Source, at the University of Saskatchewan, is a national research facility where advanced synchrotron light methods can also be used to study soil organic matter composition. These synchrotron light methods can be thought of as powerful x-rays that provide information about the composition of different elements in matter, such as soils. Synchrotron methods can also provide information about interactions between soil organic matter, clays and other materials found in soil (Gillespie et al. 2015). As such, these techniques, as well as others, are rapidly changing our understanding of soil organic matter composition and the underlying mechanisms that control its stability in the environment. Future research

endeavours will continue to develop new methods and apply these to better understand the long-term sustainability of soil organic matter in a changing world.

Nuclear magnetic resonance (NMR) spectroscopy has been used for several decades to study the composition of soil organic matter. There are many different types of NMR experiments that can be used (see Simpson et al. 2018 for more detail) but one of the most commonly applied methods in soil science is solid-state ^{13}C NMR (Preston et al. 1997). This technique allows for the analysis of all the carbon forms in soil organic matter in a whole soil sample or even different soil physical fractions. An example ^{13}C NMR spectrum (from Simpson and Simpson (2012) for an Ah horizon from an Orthic Black Chernozem (Malmo Soil Series) is shown in Figure 3.8, along with all of the different structural components that can be identified. The table links these different carbon forms to soil organic matter sources. These structures relate to different plant- and microbial-derived compounds that allows researchers to study soil organic matter sources, turnover, and preservation.

Table 3.6. ^{13}C nuclear magnetic resonance (NMR) chemical shift assignments and relation to soil organic matter structural components shown in the NMR spectrum of the Orthic Black Chernozem (Ah horizon). Chemical shift assignments compiled from: (Baldock et al. 1992; Preston et al. 1997; Salloum et al. 2002; Simpson and Simpson 2012)

Chemical shift range (ppm)	Molecular component	Soil organic matter source
0-45 ppm	Unsubstituted alkyl carbon: includes straight chain methylene carbon (30-34 ppm) and terminal methyl groups (15 ppm). Branched methylene carbon is found more downfield (35-45 ppm)	Lipids, cutin, suberin, lignin side chains, amino acids/peptides
45-65 ppm	Substituted alkyl carbon such as that found in amines (45-46 ppm) and methoxyl groups (56 ppm)	Lipids, lignin
65-95 ppm	Oxygen substituted carbon, ring carbons in carbohydrates, and carbons in ethers	Cellulose and simple sugars, amino acids/peptides, chitin
95-110 ppm	Di-oxygen substituted aliphatic carbon and anomeric carbon in carbohydrates (105 ppm)	Cellulose and simple sugars
110-145 ppm	Aromatic carbon	Black carbon/biochar, lignin and aromatic amino acids
145-160 ppm	Phenolic carbon	Lignin, suberin
160-220 ppm	Carboxylic, amide and ester carbon, carbonyl carbon	Lipids (including fatty acids), cutin, suberin, and amino acids/peptides

SOIL ORGANIC MATTER AND GLOBAL ISSUES

Global Carbon Cycle

Rising atmospheric CO_2 levels have brought attention to the important role that soils play in the global carbon cycle. Soils are the largest terrestrial store of carbon, and scientists and policy makers alike have advocated increasing soil carbon sequestration as a way to remove some CO_2 from the atmosphere.

Soil carbon stocks are most commonly expressed as kilograms (kg or 10^3 g), petagrams (Pg or 10^{15} g). Sometimes, gigatonne (Gt) is used as a unit; note that one Gt = one Pg. Stocks refer to the total amount of carbon contained within a given depth (*e.g.* the top 30 cm or 50 cm of soil). There are still major uncertainties around the estimates of global soil carbon stocks. Uncertainties are linked to limited data on deep carbon stocks (below 30 cm, to either 1 m or 3 meter depth), as well as to limited data in some geographical areas (such as in northern high-altitude soils). In addition, soil carbon stocks remain hard to model accurately as they depend on the interactive effects of several environmental factors, namely climate, vegetation, and the soil geological parent material.

In 2017, the Food and Agriculture Organization (FAO) of the United Nations launched the Global Soil Organic Carbon Map (GSOCmap) initiative. More than 100 countries contributed to produce the first global soil map ever produced from harmonized national soil maps. This map can be assessed on the GSOCmap portal. Based on these data, the global soil organic carbon stocks for topsoil (0-30cm) represent 680 Petagrams (Pg). Over 70% of these stocks are held by 14 countries, with the largest contributor being the Russian Federation (147.9 Pg – 21.9%), followed by Canada (80.2 Pg – 11.9%), and the United States of America (54.4 Pg – 8.0%).

The majority of global soil carbon stocks lie below 30 cm. Estimates to 1 m average 1,505 Pg, and up to 3,444 Pg of carbon that may be stored between 0 and 3 m. This is more than what the atmosphere and vegetation combined store.

Agricultural activities and land use change account for about 25% of the CO_2 anthropogenic emissions. Sequestering additional carbon in soils could help offset these emissions. The "4 per mille initiative- Soils for Food Security and Climate" (also known as the 4 PT- or 4 per thousand) was launched in 2015 during COP21, which stands for the "21st Conference of the Parties" to the United Nations Framework Convention on Climate Change (UNFCCC). This initiative aspires to compensate for the current increase in atmospheric CO_2 by increasing global soil carbon stocks by 4 per 1000 (or 0.4 %) per year (see 4per1000). Agriculture and associated land-use change contributes a little under 15% of total anthropogenic CO_2 emissions to the atmosphere (Figure. 3.9).

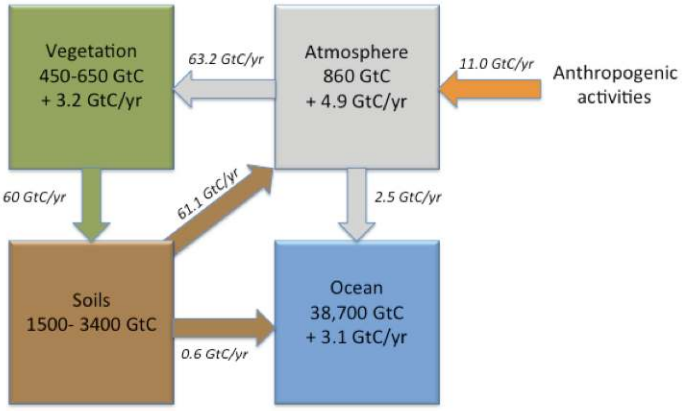
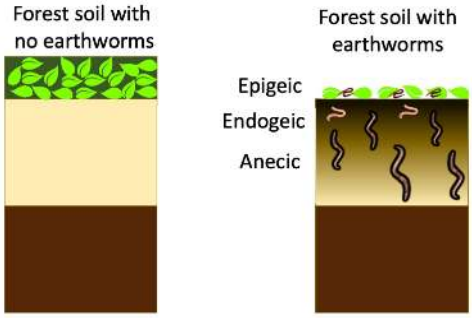


Figure 3.9. The global carbon cycle, showing stock estimates, rates of changes, and net fluxes between global reservoirs. Anthropogenic activities include emissions from fossil fuel reserves and land use changes. Redrawn from Lal (2018) with amounts from Friedlingstein et al. (2019). Amounts are averages for 2009-2018. Adapted by Sylvie Quideau and licensed under a CC BY (Attribution) license.

Climate Change and Soil Carbon and Climate Change

Climate change is altering many biogeochemical processes on the planet. Changes in both temperature and moisture have changed the quantity and quality of plant inputs to soil, and have shifted microbial activity, diversity and their subsequent substrate usage (Melillo et al. 2017; Pold et al. 2017; Lajtha et al. 2018). As described earlier, soils are a major reservoir of carbon. These carbon reservoirs are major carbon sinks in the global biogeochemical cycle. However, climate change is shifting some sinks to sources due to the fundamental alteration of processes that regulate soil carbon storage. For example, soil warming experiments have shown that warming can promote microbial activity which then increases soil organic matter degradation and decreases the total amount of soil carbon stored (Melillo et al. 2017). Sensitive ecosystems found in the Arctic are particularly vulnerable because of thawing permafrost which is contributing previously stored carbon into biogeochemically active zones (Schuur et al. 2015). Increases in fire frequency and intensity in the boreal forest are increasing soil carbon losses and threatening the ability of this biome to act as a carbon sink (Walker et al. 2019). Current predictions show a global loss of soil carbon via enhanced respiration across all ecosystems (Bond-Lamberty et al. 2018). Scientists are actively studying how sensitive soil carbon is to climate change impacts in the hopes to determine how to preserve soil carbon stocks. The future of soil carbon remains uncertain and continued soil carbon and soil organic matter losses may result in widespread declines in soil fertility and health. As such, understanding the nature and persistence of soil organic matter and how it is impacted by climate change will result in the development of informed climate change mitigation strategies that preserve and protect soil carbon in the future.

Earthworms- friends or foes?



Earthworm effects on forest soils. Small epigeic earthworms live in the forest floor and feed on plant litter, while endogeic earthworms inhabit the mineral soil; larger anecic species also feed on plant litter but create vertical burrows in the mineral soil. © Sylvie Quideau; licensed under a CC BY (Attribution) license.

Beneficial effects of earthworms on soil fertility are numerous. In agricultural soils, they have been reported to increase organic matter, decrease bulk density and increase drainage, and boost crop yield.

In the boreal forest soils of Canada, however, where they are not part of the native biological communities, they are rapidly changing the distribution of soil carbon stocks.

Estimated global carbon stocks in boreal soils range between 367 and 1716 Pg, with a median value at 1095 Pg. About one third of all boreal forests is in Canada. Earthworm invasion results in a reduction or even disappearance of the forest floor, with some of the carbon being transferred to the mineral soil. In the short term, earthworm invasion may result in a net release of carbon to atmosphere. In the longer term, the verdict is still out.

Can You Dig It!

Important Canadians in Soil Organic matter Research

***Dr. Caroline Preston**, Pacific Forestry Centre, Natural Resources Canada.*

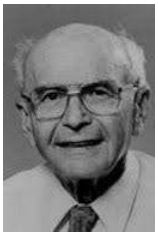


Caroline Preston (licensed under a CC-BY-NC-ND license)

Dr. Preston is an expert in soil chemistry and is considered one of the pioneers in applying the ^{13}C-NMR technique in the study of the complex structure of soil organic matter in both agricultural and forest soils. Her research program embodied the following aspects: improving the understanding of the cycling of C, N and P in the soil and use this understanding to enhance the storage of C (climate change implications) and to manage soil N and P and most significantly, pioneering the application of spectroscopic techniques (13C-NMR in particular) in soil science. Among her many important contributions is her work on the chemical structures in forest fire-derived char. Her persistence and understanding of both the methodology of NMR and of soil systems developed an innovative and highly influential body of research which has impacted our understanding of soil organic matter around the world. Dr. Preston received her undergraduate and graduate training in chemistry from McMaster University (1970)

and the University of British Columbia (1975), respectively. In 1991, she was elected a fellow of the Chemical Institute of Canada. In 1992, Dr. Preston received the Barringer Award for Applied Spectroscopy from the Spectroscopy Society of Canada and is a fellow of the Canadian Society of Soil Science.

***Dr. Morris Schnitzer**, Agriculture and Agri-Food Canada*



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Dr. Schnitzer is a soil chemist who was a pioneer in the study of soil organic matter. His contributions to this field are unparalleled, with his main contributions centering on the analytical methods of extracting, characterizing and modelling humic substances. Many feel he is the grandfather of humic substance study; his publication record includes over 350 refereed articles and many books. He was born to a Jewish family in Germany and fled the Nazis throughout the war. This odyssey has been detailed in his autobiography. He settled in Montreal where he studied soil chemistry and pioneered the early use of gas chromatography, mass spectrometry and NMR to study soil organic matter. For example, in the 1960s, he started a long-term investigation on the oxidative degradation of humic substances and whole soils using very early forms of gas chromatography and mass spectrometry. Never afraid of controversy, he also pioneered the use of pyrolysis mass spectrometry which led to his assertion of the prevalence of N-containing structures in soil organic matter. Dr Schnitzer obtained his B.Sc. with first class honors in 1951, M.Sc in 1952, and Ph.D. in 1955, all from McGill University in Soil Chemistry. He is a fellow of the Canadian Society of Soil Science, Soil Science Society of America, American Society of Agronomy, and Royal Society of Canada.

***Dr. Con (Constantine) Campbell**, Agriculture and Agri-Food Canada*



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Dr. Con Campbell, is recognized worldwide as one of the foremost specialists in research on soil organic matter and all aspects of nitrogen in soils and crops. Through his research he demonstrated that soil organic matter is a key environmental indicator on health status of prairie soils; his research has received attention nationally and internationally and is used in the development and testing of plant growth models. Con Campbell's incredible contributions have been recognized; he has been awarded the Order of Canada as well as the Saskatchewan Order of Merit. In his research, he was an early leader in the use of carbon dating to measure the dynamics of soil organic matter turnover. He conducted studies that refined the concept of potentially mineralizable nitrogen, thus providing research basis to optimize fertilizer nitrogen use efficiency on the Canadian prairies. Dr Campbell was born in Montego Bay, Jamaica where he graduated from Cornwall College before obtaining his BSA and MSA in soil science from Ontario Agricultural College of the University of Toronto in 1960 and 1961. He earned his Ph.D. at the University of Saskatchewan in 1965, and immediately joined the Swift Current Research Station in as a research scientist. He is a Fellow of the Canadian Society of Soil Science, the Agricultural Institute of Canada, the Soil Science Society of America, and the American Society of Agronomy.

SUMMARY

- Organic matter is one of the most important attributes of soils. It decreases soil bulk density, increases its water retention, cation exchange capacity, and buffering capacity. Organic matter is central to the biogeochemical cycling of many plant nutrients and helps support the huge biodiversity present in soils.

- Soils contain both organic and inorganic carbon. Several methods exist to measure soil total carbon and soil organic carbon. Organic carbon is the main constituent of soil organic matter, but caution must be applied when converting between soil carbon and organic matter.
- Plant residues are the major substrate for soil organic matter formation. Soil organic matter is composed of these residues at various stages of decomposition, and also contains, in some cases in large amounts, dead faunal and microbial biomass, and products of microbial synthesis.
- Soils store vast amounts of carbon, more than twice as much as the amount of carbon in the atmosphere and terrestrial vegetation combined.
  Soil carbon is not evenly distributed in the landscape, but instead will vary as a function of soil forming factors.
- The majority of soil carbon in Canada is contained in Cryosols and Organic soils where cold temperature inhibits carbon decomposition losses.
- Initial cultivation of Canadian prairies caused the loss of about a quarter of the carbon contained in these soils. Some of this soil carbon can be restored using conservation agriculture practices, and planting to perennials.

STUDY QUESTIONS

1. Where does soil organic matter come from?
2. What is the relationship between soil organic matter and major elements (carbon, oxygen and nitrogen)?
3. How does soil organic matter contribute to the total soil cation exchange capacity?
4. Which environmental factors determine the persistence of soil organic matter in different parts of Canada?
5. What happens to plant material (fresh residues) when added to soil? Where does the carbon go?
6. Based on some basic redox principles, why are some forms of soil organic matter preferred microbial substrates over others?
7. How do we measure soil organic matter content in soil samples?
8. If information about the soil organic matter composition is needed, what types of methods could be used? Are these methods easy to perform or do they require advanced equipment or facilities to carry out these measurements?

REFERENCES

Bailey, V. L., Pries, C. H. and Lajtha, K. 2019. What do we know about soil carbon destabilization?. Environ. Res. Lett. 14: 083004. doi: 10.1088/1748-9326/ab2c11.

Baldock, J.A., Oades, J.M., Waters, A.G., Peng, X., Vassallo, A.M., and Wilson, M.A. 1992. Aspects of the chemical structure of soil organic materials as revealed by solid-state 13C NMR spectroscopy. Biogeochemistry 16 : 1–42.

Baldock, J.A. and Skjemstad, J.O. 2000. Role of the soil matrix and minerals in protecting natural organic materials against biological attack. Org. Geochem. 31: 697–710. doi: 10.1016/S0146-6380(00)00049-8.

Bohn, H.L., McNeal, B.L. and O'Connor, G.A. 1985. Soil Chemistry. 2nd edn. New York, Wiley.

Bond-Lamberty, B., Bailey, V.L., Chen, M., Gough, C.M. and Vargas, R. 2018. Globally rising soil heterotrophic respiration over recent decades. Nature 560: 80–83. doi: 10.1038/s41586-018-0358-x.

Cotrufo, M. F., Wallenstein, M.D., Boot, C.M., Denef, K. and Paul, E. 2013. The Microbial Efficiency-Matrix Stabilization (MEMS) framework integrates plant litter decomposition with soil organic matter stabilization: Do labile plant inputs form stable soil organic matter? Glob. Change Biol. 19: 988-995. doi: 10.1111/gcb.12113.

Derenne, S. and Nguyen Tu, T. T. 2014. Characterizing the molecular structure of organic matter from natural environments: An analytical challenge. C. R. Geosci. 346: 53–63. doi: 10.1016/j.crte.2014.02.005.

Feng, X., Simpson, A.J. and Simpson, M.J. 2005. Chemical and mineralogical controls on humic acid sorption to clay mineral surfaces. Org. Geochem. 36: 1553–1566. doi: 10.1016/j.orggeochem.2005.06.008.

Friedlingstein, P, Jones, M W, O'Sullivan, M et al. 2019. Global carbon budget 2019. Earth System Science Data 11: 1783-1838.

Gillespie, A.W., Philips, C.L., Dynes, J.J., Chevrier, D., Regier, T.Z. and Peak, D. 2015. Advances in using soft x-ray spectroscopy for measurement of soil biogeochemical processes. Adv. Agron. 133:1–32. doi: 10.1016/bs.agron.2015.05.003.

Gregorich, E.G., Gillespie, A.W., Beare, M.H., Curtin, D., Sanei, H., Yanni, S.F., 2015. Evaluating biodegradability of soil organic matter by its thermal stability and chemical composition. Soil Biology and Biochemistry 91, 182–191. https://doi.org/10.1016/j.soilbio.2015.08.032

Gregorich, E. G., Monreal, C.M., Schnitzer, M. and Schulten, H.R. 1996. Transformation of plant residues into soil organic matter: Chemical characterization of plant tissue, isolated soil fractions, and whole soils. Soil Sci 161: 680–693. doi: 10.1097/00010694-199610000-00005.

Hatton, P.J., Kleber, M., Zeller, B., Moni, C., Plante, A.f., Townsend, K., Gelhaye, L., Lajtha, K, and Derrien, D. 2012. Transfer of litter-derived N to soil mineral-organic associations: evidence from decadal 15N tracer experiments. Org. Geochem 42: 1489–1501. doi: 10.1016/j.orggeochem.2011.05.002.

Horwath, W.R. and Kuzyakov, Y. 2018. The potential for soils to mitigate climate change through carbon sequestration. Dev. Soil Sci. 35: 61-92. doi: 10.1016/b978-0-444-63865-6.00003-x.

Janzen, H.H., Campbell, C.A., Gregorich, E.G. and Ellert, B.H. 2018. Soil carbon dynamics in canadian agroecosystems. Pages 57-80 in R., Lal, J.M. Kimble, R.F. Follett and B.A. Stewart, eds. Soil Processes and the Carbon Cycle. CRC Press, Boca Raton, Fla. doi: 10.1201/9780203739273.

Jenny, H. 1941. Factors of Soil Formation. McGraw-Hill, New York. doi: 10.1097/00010694-194111000-00009.

Kelleher, B. P. and Simpson, A. J. 2006. Humic substances in soils: Are hey really chemically distinct? Environ. Sci. Technol. 40: 4605–4611. doi: 10.1021/es0608085.

Kleber, M., Sollins, P. and Sutton, R. 2007. A conceptual model of organo-mineral interactions in soils: self-assembly of organic molecular fragments into zonal structures on mineral surfaces. Biogeochemistry 85: 9–24. doi: 10.1007/s10533-007-9103-5.

Kögel-Knabner, I. 2000. Analytical approaches for characterizing soil organic matter. Org. Geochem. 31: 609–625. doi: 10.1016/S0146-6380(00)00042-5.

Kögel-Knabner, I. 2002. The macromolecular organic composition of plant and microbial residues as inputs to soil organic matter. Soil Biol. Biochem. 34: 139–162. doi: 10.1016/S0038-0717(01)00158-4.

Kögel-Knabner, I. and Amelung, W. 2014. Dynamics, chemistry, and preservation of organic matter in soils. pp. 157-215 In H.D. Holland and K.K. Turekian (eds), Treatise on Geochemistry, Second Edition, vol. 12, Elsevier, Oxford. . doi: 10.1016/B978-0-08-095975-7.01012-3.

Kohl, L., Laganiere, J., Edwards, K.A., Billings, S.A., Morrill, P.L. Van Biesen, G. and Ziegler, S.E. 2015. Distinct fungal and bacterial δ13C signatures as potential drivers of increasing δ13C of soil organic matter with depth. Biogeochemistry 124: 13–26. doi: 10.1007/s10533-015-0107-2.

Kuzyakov, Y., Friedel, J. , and Stahr, K. 2000. Review of mechanisms and quantification of priming effects. Soil Biol. Biochem. 32: 1485–1498. doi: 10.1016/S0038-0717(00)00084-5.

Lafargue, E., Marquis, F., Pillot, D., 1998. Rock-Eval 6 applications in hydrocarbon exploration, production, and soil contamination studies. Revue de l'Institut Francais du Petrole 53, 421–437.

Laganière, J., Angers, D. A. and Paré, D. 2010. Carbon accumulation in agricultural soils after afforestation: A meta-analysis. Glob. Change Biol. 16: 439–453. doi: 10.1111/j.1365-2486.2009.01930.x.

Lajtha, K., Bowden, r.D., Crow, S., Fekete, I., Kotroczo, Z., Plante, A., Simpson, M.J. and Nadelhoffer, K.J. 2018. The detrital input and removal treatment (DIRT) network: Insights into soil carbon stabilization. Sci. Total Environ. 640–641: 1112–1120. doi: 10.1016/j.scitotenv.2018.05.388.

Lal, R. 2018. Digging deeper: A holistic perspective of factors affecting soil organic carbon sequestration in agroecosystems. Glob. Change Biol. 24: 3285-3301. doi: 10.1111/gcb.14054.

Lehmann, J. and Kleber, M. 2015. The contentious nature of soil organic matter. Nature 528: 60–68. doi: 10.1038/nature16069.

Mary, B., Clivot, H., Blaszczyk, N., Labreuche, J. and Ferchaud, F. 2020. Soil carbon storage and mineralization rates are affected by carbon inputs rather than physical disturbance: Evidence from a 47-year tillage experiment. Agr. Ecosys. Environ. 299: 106972.

doi: 10.1016/j.agee.2020.106972.

Mayer, M. Prescott, C., Abaker, W.E.A., Augusto, L., Cécillon, L., Ferreira, G.W.D., James, J., Jandl, R., Katzensteiner, K., Laclau, J.-P., Laganière, J., Nouvellon, Y.; Paré, D., Stanturf, J.A., ,Vanguelova, E.I. and Vesterdal, L.2020. Influence of forest management activities on soil organic carbon stocks: A knowledge synthesis. Forest Ecol. Manag. 466: 118127. doi: 10.1016/j.foreco.2020.118127.

Melillo, J. M., Frey, s.D., DeAngelis, K.M., Werner, W.J., Bernard, M.J., Bowles, F.P., Pold, G., Knorr, M.A. and Grandy, A.S. 2017. Long-term pattern and magnitude of soil carbon feedback to the climate system in a warming world. Science 358: 101–105. doi: 10.1126/science.aan2874.

Mikutta, R., Kleber, M., Torn, M.S. and Jahn, R. 2006. Stabilization of soil organic matter: Association with minerals or chemical recalcitrance?. Biogeochemistry 77: 25–56. doi: 10.1007/s10533-005-0712-6.

Oades, J.M. 1988. The retention of organic matter in soils. Biogeochemistry 5: 35–70. doi: 10.1007/BF02180317.

Ohlson, M., Dahlberg, B., Okland, T., Brown, K.J. and Havorsen, R. 2009. The charcoal carbon pool in boreal forest soils. Nature Geoscience 2: 692-695. doi: 10.1038/ngeo617.

Pold, G., Grandy, A.S., Melillo, J.M. and DeAngelis, K.M 2017. Changes in substrate availability drive carbon cycle response to chronic warming. Soil Biol. Biochem. 110: 68–78. doi: 10.1016/j.soilbio.2017.03.002.

Prescott, C.E., Frouz, J., Grayston, S.J., Quideau, S.A. and Straker, J. 2019. Rehabilitating forest soils after disturbance. Developments in Soil Science 36: 309-343.doi: 10.1016/b978-0-444-63998-1.00013-6.

Preston, C.M., Trofymow, J.A., Niu, J. and Sayer, B.G. 1997. 13C nuclear magnetic resonance spectroscopy with cross-polarization and magic-angle spinning investigation of the proximate-analysis fractions used to assess litter quality in decomposition studies. Can. J. Bot. 75: 1601–1613. doi: 10.1139/b97-872.

Pribyl, D.W. 2010 A critical review of the conventional SOC to SOM conversion factor. Geoderma. 156:75-83, doi:10.1016/j.geoderma.2010.02.003

Quideau, S.A., Anderson, M.A., Graham R.C., Chadwick, O.A. and Trumbore, S.E. 2000. Soil organic matter processes: characterization by ^{13}C NMR and ^{14}C measurements.Forest Ecol. Manag. 138: 19–27. doi: 10.1016/S0378-1127(00)00409-6.

Ricklefs, R.E. 2008. The Economy Of Nature. W. H. Freeman and Company, New York.

Salloum, M.J., Chefetz, B. and Hatcher, P. G. 2002. Phenanthrene sorption by aliphatic-rich natural organic matter. Environ. Sci. Technol. 36:. doi: 10.1021/es015796w.

Schmidt, M.W., Torn, M.S., Abiven, S., Dittmar, T., Guggenberger, G., Janssens, I.A., Kleber, M., Kögel-Knabner, I., Lehmann, J., Manning, D.A.C.,Nannipieri, P., Rasse, D.P., Weiner, S. and Trumbore, S.E. 2011. Persistence of soil organic matter as an ecosystem property. Nature 478: 49–56. doi: 10.1038/nature10386.

Schuur, E.A.G. McGuire, A.D., Schädel, C., Grosse, G., Harden, J.W., Hayes, D.J., Hugelius, G., Koven, C.D. , Kuhry, P., Lawrence, D.M., Natali, S.M., Olefeldt, D., Romanovsky, V.E., Schaefer, K., Turetsky, M.R. , Treat, C.C and Vonk, J.E. 2015. Climate change and the permafrost carbon feedback. Nature 520: 171–179. doi: 10.1038/nature14338.

Sebag, D., Disnar, J.R., Guillet, B., Di Giovanni, C., Verrecchia, E.P., Durand, A., 2006. Monitoring organic matter dynamics in soil profiles by “Rock-Eval pyrolysis”: Bulk characterization and quantification of degradation. European Journal of Soil Science 57, 344–355.

Shaw, C.H., Banfield, E. and Kurz, W.A. 2008. Stratifying soils into pedogenically similar categories for modeling forest soil carbon. Can. J. Soil Sci. doi: 10.4141/CJSS07099.

Simpson, A.J., Simpson, M.J., Smith, E. and Kelleher, B.P. 2007. Microbially derived inputs to soil organic matter: Are current estimates too low?. Environ. Sci. Technol. 41: 8070-8076. doi: 10.1021/es071217x.

Simpson, A.J., Simpson, M. J. and Soong, R. 2018. Environmental nuclear magnetic resonance spectroscopy: An overview and a primer. Anal. Chem. 90. doi: 10.1021/acs.analchem.7b03241.

Simpson, M. J. and Simpson, A. J. 2012. The chemical ecology of soil organic matter molecular constituents. J. Chem. Ecol. 38:768–784. doi: 10.1007/s10886-012-0122-x.

Six, J., Conant, R.T., Paul, E.A. and Paustain, K. 2002. Stabilization mechanisms of soil organic matter: Implications for C-saturation of soils. Plant Soil 241: 155-176. doi: 10.1023/A:1016125726789.

Sollins, P., Kramer, M.G., Swanson, C., Lajtha, K., Filley, T., Aufdenkampe, A.K., Wagai, R and Bowden, R.D. 2009. Sequential density fractionation across soils of contrasting mineralogy: evidence for both microbial- and mineral-controlled soil organic matter stabilization. Biogeochemistry 96: 209–231. doi: 10.1007/s10533-009-9359-z.

Soucemarianadin, L. N., Quideau. S.A., Wasylishen, R.E. and Munsen, A.D. 2015. Early-season fires in boreal black spruce forests produce pyrogenic carbon with low intrinsic recalcitrance. Ecology 96: 1575–1585.

Tarnocai, C. 2008. The amount of organic carbon in various soil orders and ecological Provinces in Canada. Pages 81-92 in R., Lal, J.M. Kimble, R.F. Follett and B.A. Stewart, eds. Soil Processes and the Carbon Cycle. CRC Press, Boca Raton, Fla.. doi: 10.1201/9780203739273.

Trumbore, S. E. and Druffel, E. R. M. 1995. Carbon isotopes for characterizing sources and turnover of nonliving organic matter. Pages 7-22 in R.G. Zepp and C. Sonntag, eds. Role of Nonliving Organic Matter in the Earth's Carbon Cycle. John Wiley& Sons Ltd., Chichester..

VandenBygaart, A. J., Gregorich, e.G., Angers, d.A. and Stoklas, U.F. 2004. Uncertainty analysis of soil organic carbon stock change in Canadian cropland from 1991 to 2001. Glob. Change. Biol. 10: 983-994 doi: 10.1111/j.1365-2486.2004.00780.x.

Walker, X. J., Baltzer, J.L., Cumming, s.g., Day, N.J., Ebert, C., Goetz, S., Johnstone, J.F., Potter, S., Rogers, B.M., Schuur, E.A.G, Turetsky, M.R. and Mack, M.C. 2019. Increasing wildfires threaten historic carbon sink of boreal forest soils. Nature 572: 520-531. doi: 10.1038/s41586-019-1474-y.

Whalen, J.K., Gul, S., Poirier, V., Yanni, S.F., Simpson, M.J., Clemente, J.S., Feng, X., Grayston, S.J., Barker, J., Gregorich, E.G., Angers, D.A., Rochette, P., and Janzen, H.H. 2014. Transforming plant carbon into soil carbon: Process-level controls on carbon sequestration. Can. J. Plant Sci. 94. doi: 10.4141/CJPS2013-145.

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I am a Professor in Soil Biogeochemistry with a special interest in carbon fluxes and organic matter processes. The focus of my research is in three areas: 1) quantifying the underlying environmental factors controlling organic matter accumulation and distribution in soils; 2) relating measurable organic matter quality indices to soil functioning (nutrient fluxes and microbial biodiversity); and 3) predicting the response of soil organic matter to natural and anthropogenic disturbance.

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I am an Assistant Professor of Soil Biochemistry in the School of Environmental Sciences. I have a strong interest in using innovative research and instrumentation to link land management with soil health and sustainable land use. In particular, my research focus is on organic matter characterization using spectroscopic and thermal methods, carbon cycling and modelling.

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1.4: Soil Physics

Sandra Brown, Asim Biswas, Jean Caron, Miles Dyck, and Bing Si

Learning Objectives

On completion of this chapter, students will be able to:

1. Define relationships between texture, bulk density, porosity and compaction
2. Describe how soils retain water and the forces driving water movement in soils
3. Relate water retention to soil properties, and water availability for plants
4. Characterize how soil water movement varies with tension and pore-size distribution
5. Describe the major mechanism of gas exchange in soils
6. List the main mechanisms of solute transport in soils and their relevance for plants
7. Outline soil thermal properties and factors influencing soil temperature and heat flow
8. Explain the effect of soil water on soil consistency

INTRODUCTION

Soil physics is the study of the solid, liquid and gaseous phases of soils, and their interactions. Soil texture, structure and bulk density reflect how soil mineral and organic particles combine to form the soil matrix and pore spaces. Pore spaces hold gases and water. Understanding soil water retention and soil water movement is crucial in determining water availability for plants and soil organisms, infiltration and drainage, runoff and erosion. Many soil nutrients are transported in the soil as solutes in soil solution. Soil aeration and gas exchange govern CO_2 emissions from the soil and O_2 availability for plants. Soil thermal properties regulate soil temperature with depth and determine how quickly (or slowly) a soil warms up in the spring. Soil strength is influenced by soil texture and water content, and determines the susceptibility of a soil to slope failure and compaction.

Understanding the impacts of land management on soil physical properties can help us develop alternative practices for managing soils in a variety of ecosystems (forest, agriculture, urban). For example, in forest operations, soils may be compacted on landings, where logs are stored and loaded onto trucks by heavy machinery. But not all soils are equally susceptible to compaction, nor will all soil types require mechanical loosening of the soil prior to tree replanting. Forest practitioners need to consider soil physical properties, such as texture and water content, as part of their assessment of the need for costly rehabilitation measures. Agriculture accounts for 85% of ammonia (NH_3) emissions in Canada (Bittman et al., 2017), in large part associated with the application of animal manure to fields. In regions with a large number of dairy and poultry farms, such as the Lower Fraser Valley in British Columbia, farmers are faced with excessive amounts of manure that are often applied to cropped fields. By understanding gas exchange in the soil-plant-atmosphere system, researchers from Agriculture and Agri-Food Canada in Agassiz were able to develop a manure spreader that injects the dairy manure into the soil between corn rows. This type of machinery reduced both air pollution (through reduction of nitrous oxide N_2O emissions) and groundwater pollution (through reduction of leaching losses of nitrates) from agricultural fields. In urban environments, the construction of houses commonly involves the removal of vegetation and topsoil from the site, and purposefully compacting the soil to reduce settling. Once construction is finished, a thin layer of topsoil is spread and seeded with grass. But the water holding capacity of this thin layer of soil is insufficient to support live grass in the summer. In the driest parts of Canada, cities such as West Kelowna in British Columbia have introduced bylaws, which require a minimum 30 cm topsoil thickness, aimed at reducing outdoor water consumption by increasing the volume of plant available water held in the soil. These are just a few examples of how soil physics can help us solve a range of soil-related environmental challenges.

SOIL IS A THREE-PHASE SYSTEM

Soil is a complex three-phase system, comprised of: solids (soil mineral particles and organic matter), liquids (the soil solution: water, dissolved nutrients, chemicals, and gases), and gases (e.g., N_2, O_2, CO_2) (Figure 4.1). The magnitude and interaction among the three phases determine the behaviour and functionality of soil.

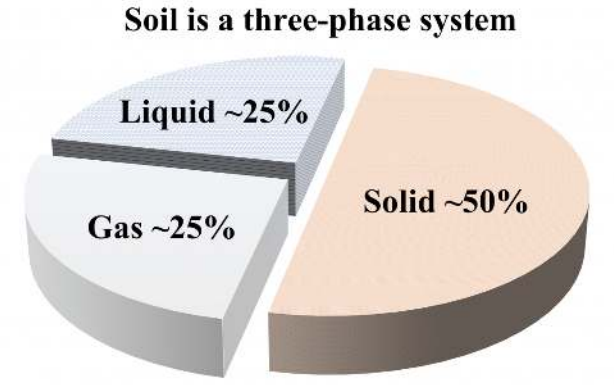


Figure 4.1. Soil is a three-phase system; solid, liquid and gas. The proportion of each phase may vary, but ideally, it consists of 50% solid, 25% liquid, and 25% gas. © Asim Biswas, Univ. of Guelph is licensed under a CC BY (Attribution) license.

Solid phase

The solid phase of a soil system can be comprised of soil particles that are either mineral (i.e., rocks, stones, cobbles, gravel, sand, silt, clay) or organic (i.e., soil organic matter) in nature. Ideally, the solid phase occupies approximately 50% of the soil by volume (Figure 4.1). However, this may vary between 30 and 60% of the total soil volume based on physical conditions as well as management impacts. For example, in a sandy soil, larger mineral particles may occupy a greater volume of soil. Similarly, compaction may increase the volume of solid particles within a specified volume. While the solid mineral content is generally stable over time, the soil organic matter content may change relatively quickly. The soil mineral content consists of primary particles such as crystalline minerals (e.g., quartz, aluminosilicates) and amorphous gels (e.g., oxides, hydroxides and hydrous oxides of iron, aluminum, and manganese). These primary soil particles interact with each other. For example, amorphous gels may coat crystalline particles and form secondary particles or aggregates (as described in Chapter 2: Formation of Soil Structure). Thus, the mineral content is comprised of particles of different shapes, sizes, and chemical composition. Organic materials also contribute to the total solid phase of a soil system, but generally occupy a smaller proportion (3-5%) in mineral soils. Soil organic matter is generally comprised of live organisms, and plant and animal residues in various stages of decomposition. These organic materials can bind other mineral and organic materials together, and form secondary units (called aggregates), which also contribute to the solid phase of a soil system. The composition and surface characteristics of solids dictate the behaviour of soil and determine the interaction of solid phase with liquid (e.g., soil water, plant nutrients) and gaseous (e.g., soil air) phases.

Liquid phase

The liquid phase of a soil system consists of soil water and various nutrients, chemicals, and gases dissolved in soil water (sometimes referred to as soil solutes), forming a soil solution. Both the amount of water in soil (water quantity) and the chemical composition (water quality) contribute to the liquid phase of a soil system. Ideally, the liquid phase consists of 25% of the total volume of soil. However, it is more dynamic than the solid phase as it may vary between <1% (completely dry conditions) to approximately 50% (all spaces between the solid particles being filled with liquids). The spaces between the solid particles within a soil system are known as soil pores, and when all the pores are filled with water, the soil is considered saturated. The pores that are not filled with water within a soil system, will be filled by the gaseous phase (i.e., soil air) and thus the liquid and gaseous phases share the pore spaces within a soil system.

Gaseous phase

The gaseous phase of a soil system consists of soil air, which is a mixture of gases commonly including nitrogen (N_2), oxygen (O_2), water vapour and carbon dioxide (CO_2). Ideally, soil gases comprise about 25% of the total soil volume, but it is highly dynamic in nature. Generally, the pore spaces that are not occupied by the liquid phase (i.e., soil water), will be occupied by the gaseous phase (i.e., soil air). The amount, composition and mobility of gaseous vary with time and position in the soil profile. Soil air also varies in composition from atmospheric air. For example, soil air contains higher amount of carbon dioxide and lower amount of oxygen than atmospheric air.

Soil separates and soil texture

Soil separates

The solid phase of a soil system is comprised of various primary and secondary particles that are mineral or organic in nature, and occur in various amounts, shapes, sizes, and chemical and mineralogical compositions. For example, some of the particles are coarse enough to be seen with the naked eye, while others are small enough that they can only be seen with a microscope and exhibit colloidal properties.

The mineral particles in soils are divided into size classes. **Coarse fragments**, >2 mm in size are separated from the **fine earth fraction** (<2 mm). Within coarse fragments, soil particles between 2 mm to 8 cm in diameter are named as gravel, 8 to 25 cm in diameter are named as cobbles and >25 cm in diameter is are named as stones or boulders. These coarse fragments can affect the selection of land management practices (e.g., selection of tillage implements), but they contribute little to basic soil functions such as water retention and the capacity to store and release plant nutrients.

The fine earth fraction (<2 mm) is grouped into three major categories: sand, silt and clay. The scheme for separating soil particle sizes following the Canadian system of soil classification is shown in Table 4.1.

Table 4.1. The soil separate classification scheme of Canadian system of soil classification (CSSC) and the diameter ranges for each size fraction

Soil separate	**Particle diameter**	
	mm	**µm**
Very coarse sand	2.0-1.0	2000-1000
Coarse sand	1.0-0.5	1000-500
Medium sand	0.5-0.25	500-250
Fine sand	0.25-0.10	250-100
Very fine sand	0.10-0.05	100-50
Silt	0.05-0.002	50-2
Clay	≤0.002	≤2
Fine Clay	≤0.0002	≤0.2

The term soil separates is used to describe the mineral particles (sand, silt, clay) that make up the fine earth fraction.

Sand – the largest group of soil separates, defined by the diameter from 2 mm to 0.05 mm according to CSSC. The sand fraction is subdivided into five sub-fractions (very coarse, coarse, medium, fine, and very fine). The sand particles are predominantly comprised of the mineral quartz but may also have fragments of other primary minerals such as feldspar, mica and others. Sand particles are generally spherical in nature, with jagged edges, hard and abrasive, and feel gritty and loose. Sand contributes to excessive drainage and low plant available water.

Silt – consists of particles with intermediate size between sand and clay. The size range for silt is 0.05 mm to 0.002 mm (CSSC). Silt particles are comprised of the same minerals as sand. When rubbed between your fingers, silt particles exhibit a smooth feeling. Due to their smaller size relative to sand, silt has larger specific surface area per unit mass (described later).

Clay – the smallest size fraction of soil mineral particles, and it is the colloid fraction. The diameter of clay particles is less than 0.002 mm and the diameter of fine clay particles is less than 0.0002 mm (CSSC). Clay particles are typically plate-like or needle-like in shape and generally are comprised of secondary minerals (e.g., aluminosilicates). Among the fine earth particles, clay size particles have the largest specific surface area per unit mass and exhibit a unique physicochemical activity. Clay particles typically carry a negative charge and are often highly reactive. Clays are generally sticky and typically exhibit plastic behaviour. Clay particles can absorb water, causing soil to swell and shrink/crack when dry. While relatively inert sand and silt fractions can

constitute the soil skeleton, clays are considered the flesh to that skeleton. Together, all three size fractions, in various configurations, constitute the matrix of the soil.

Soil texture

The term soil texture is used to express the relative proportions of the various soil separates in a soil. Qualitatively soil texture represents the feel of the soil materials, while quantitatively soil texture is the relative proportions of the various sizes of particles (i.e., sand, silt, clay) in any soil. Soils are grouped into texture classes based on the relative proportions of sand, silt and clay, and there are 13 such classes in the CSSC (Figure 4.2). The textural class – loam is used to represent soils with about equal proportions of sand, silt, and clay. Generally, a textural class dominated by a single particle size is less suitable for plant growth than a loam. Information on soil textural class is very important as it can guide the soil's suitability for plant growth and dictate the land use capability, as well as soil management. The movement and retention of water and solutes (e.g., nutrients), as well as heat transfer and aeration, are all affected by soil texture.

In determining the soil textural class using the Canadian soil textural triangle (Figure 4.2), mark the value of sand content on the bottom axis reading from left to right and then take the clay content and mark the value on the vertical axis reading from bottom to the top. Draw a line from the bottom axis identified point parallel to the vertical axis and draw another line from the vertical axis identified point in parallel to the bottom axis. Mark the point where two lines cross and identify the soil textural class. If the point falls on the line of two soil textural classes, the class with finer particles or high clay content will get the designation (customary). Though the third axis presents the silt content, often it is enough to identify textural class with sand and clay data.

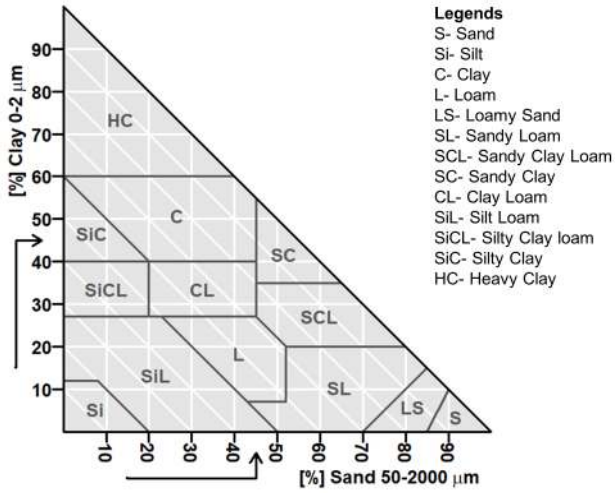


Figure. 4.2. Soil textural triangle following Canadian system of soil classification. The figure is created in R using package 'soiltexture'. © Asim Biswas is licensed under a CC BY (Attribution) license

Specific surface area of a soil is the total surface area of soil particles per unit mass of soil or per unit volume of soil particles. It is commonly expressed as square meters per gram of soil (mass) or per cubic centimetre (volume) of soil particles. The specific surface area depends on the size as well as shape of the particles. Smaller sized particles contribute to large surface area per unit mass or volume. Similarly, soil particles with flattened or elongated shapes can expose greater surface area per mass or volume than the soil particles of cubic or spherical shape. Clay particles, in addition to their small size, are generally of plate like shape, and they contribute a large surface area per unit mass or volume of soil. While sand particles can have specific surface area of about 1 m^2 g^{-1}, clay particles can have specific surface area of as high as several hundred square meters per gram of soil. Specific surface area of any soil material is a fundamental property and is correlated with other important properties such as cation exchange capacity, water retention, nutrient availability, and mechanical properties including plasticity, cohesion and strength.

Soil Structure

Mineral and organic particles in soils are arranged by various forces and at different scales to form distinct structural units called peds or aggregates. Soil structure refers to the arrangement of sand, silt, clay and organic particles into aggregates or peds. These aggregates affect the nature of the system of pores in a soil. Consequently, soil structure has a major influence on water and air movement as well as root growth and the movement of macro- and meso-fauna.

How do Aggregates Form?

Aggregation is the result of several biological, physical and chemical processes across different scales – the atomic or molecular level, the microscopic level, and the macroscopic level or visible scale. Figure 4.3 illustrates aggregate formation across different scales.

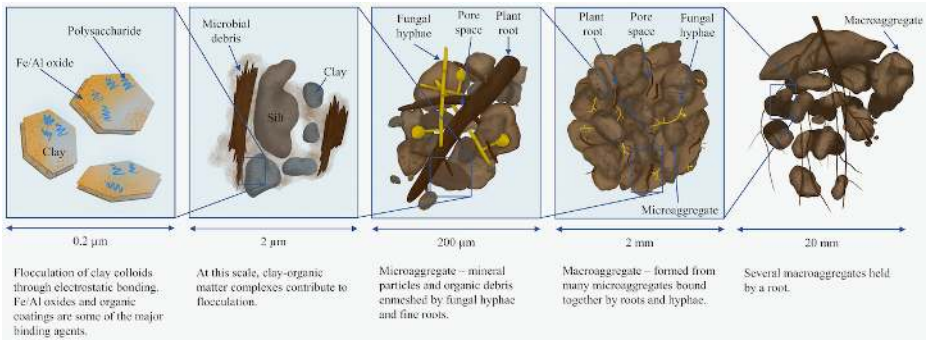


Figure 4.3. Aggregate formation across different scales. Modified from Tisdale and Oades (1982) and adapted from Krzic et al., SoilWeb200 CC-BY. Adapted and modified by Lewis Fausak and licensed under a CC BY (Attribution) license.

At the molecular level, clay particles (with negative charges on their surfaces) attract cations and polar water molecules. Polyvalent cations and water molecules are able to link clay particles by acting as bridges between them. When the soil dries out, the number of water molecules making up these bridges decreases and the clay particles are brought close together. This enables flocculation, which is the first step in aggregate formation.

Organic compounds are also important in forming bridges between clay particles. Long chain molecules (e.g., polysaccharides) are able to come in close contact with the clay, linking them together. Aluminum and iron oxides and hydroxides can perform similar role, creating complex linkages between clay particles, thus, contributing to aggregate formation.

Sand and silt particles carry almost no charge on their surface, but can be linked into micro-aggregates at the microscopic level (≤250 micron) if they have a coating of clay or organic compounds on their surfaces.

At the macroscopic level (≥250 micron), the soil micro-aggregates are bound together into macro-aggregates by fungal hyphae, fine plant roots and other stabilizing agents. It is generally the shape of these macro-aggregates that is described by soil scientists, since they are important for providing the soil with a system of large pores.

A number of factors are recognized as being important for aggregate formation, that all play a role in bringing individual charged clay particles close to each other so that flocculation can occur. These processes include:

- wetting and drying cycles
- swelling and shrinking of clay particles
- freezing and thawing

All of these physical and chemical processes are associated largely with soils that have a high clay content (i.e., in finer textured soils). In sandy soils, aggregate formation is entirely dependent on biological processes, which include:

- burrowing activities of earthworms,
- production of organic gels by bacteria and fungi that serve as cementing agents and stabilize floccules; for example, hyphae of soil fungi which produce a sticky substance called glomalin that acts as a cementing agent, and
- enmeshment of mineral particles by a network of fine roots and fungal hyphae.

The activity of microbes and roots, and the presence of organic matter produces larger aggregates (i.e., macro-aggregates).

Common types of cementing agents in soils are:

- organic compounds
- clay coatings
- Fe/Al oxides
- carbonates

Remember that Aggregation = Flocculation + Cementation, where:

Flocculation: is the interaction among soil (clay) particles, when they are brought close together. These interactions include electrostatic and van der Waals forces, and/or hydrogen bonding; and lead to formation of microscopic floccules (or clumps); and

Cementation: is the stabilization of floccules by a cementing agent, such as organic compounds, carbonates, Fe and Al oxides, or clays.

Dispersion of clays

The net negative charge of phyllosilicate particles tend to make them repel each other. When these repulsive forces are strong, particles cannot approach each other and they are said to be dispersed. Mutual repulsion is enhanced by monovalent cations at low concentrations, because under those conditions there is little neutralization of repelling negative charges. Consequently, soil dominated by Na^+ ions will have clay particles in a dispersed state, more so than a soil where Ca^{2+} or other polyvalent cations are dominant. Dispersion of soil particles is undesirable, since dispersed particles lead to small pores. Water infiltration is thus greatly reduced, leading to standing water at the soil surface.

Types of Aggregates

Soils can either be structureless or aggregated. Single-grained soils are one example of a structureless soil. These soils are dominated by sand particles and there is no aggregate formation. The lack of structure in a sandy soil is due to the fact that sand particles, comprised of primary minerals such as quartz, carry a limited charge. This results in very limited bonding and consequently a very limited flocculation in coarse textured soils.

Massive soils are another example of a structureless soil. These soils have no visible aggregates and are typical of massive clays. When dry, these soils are hard to break apart. The lack of structure in a clay soil can prevent plant growth, as a massive clay lacks sufficient medium and large pores for root development. This type of soil has a very slow rate of water infiltration.

In aggregated soils, soil particles are associated in distinct aggregates or peds. Soil aggregates can be characterized in terms of their: shape (or type), size (fine, medium, or coarse), and distinctness (or strength, such as strong, moderate, or weak). Many types of soil aggregates occur in soils with four principal shapes being spheroidal, platy, block-like (angular blocky with sharp edges and sub-angular blocky with rounded edges), and prism-like (vertically oriented columnar aggregates with rounded tops and prismatic aggregates with flat tops). These structural types (and some sub-types) are illustrated below (Figure 4.4).

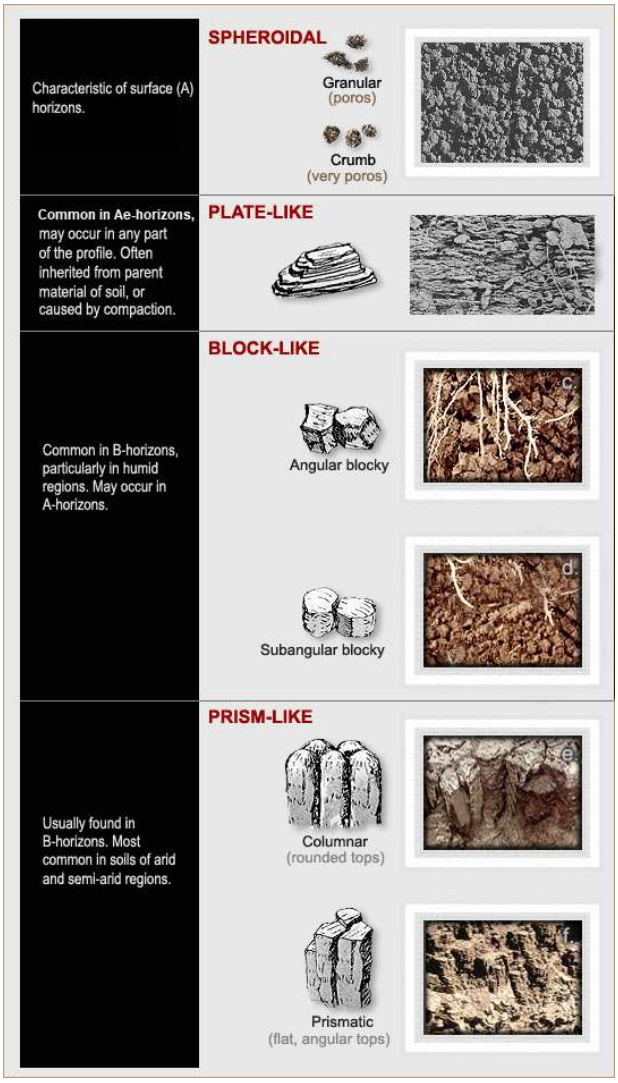


Figure 4.4. Types or shapes of soil aggregates found in mineral soils. Source: © Krzic et al. 2013. SoilWeb200 is licensed under a CC BY-NC-SA (Attribution NonCommercial ShareAlike) license.

Aggregate shape is affected partly by composition and partly by differences in soil development processes. For example, clayey soils often have blocky structure; the columnar structure is frequently associated with sodium clays in semi-arid climates; platy structure often occurs near the soil surface as a result of the freeze-thaw activity.

Soil Structure and Pore Space

Soil structure directly influences the type of pore spaces that exist in a soil. Pores can be classified as: macro-pores (diameter > 0.08 mm) that occur between aggregates or individual grains in coarse-textured soils. Macro-pores readily allow the movement of air, the drainage of water, and provide space for roots and organisms in the soil. Micro-pores (diameter < 0.08 mm) occur within aggregates or between individual clay-size particles. Micro-pores are usually filled with water and are too small to allow much movement of air. Water movement in micro-pores is slow and some of the water tightly held by soil colloids is unavailable to plants.

A well-aggregated clay soil tends to have a balance of large “aeration and drainage” pores, and small “water-retention” pores (Figure 4.5). Good soil structure is associated with small or medium size aggregates with abundant pores both within and between aggregates.

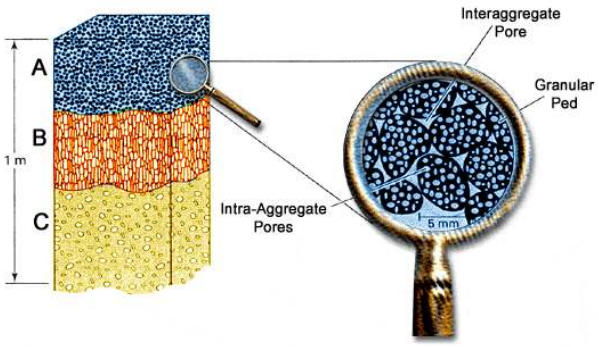


Figure 4.5. Schematic of an A horizon with a porous granular structure, pores between peds or inter-aggregate pores (shown as triangular spaces in the magnifying glass) and within peds or intra-aggregate pores (shown as blue rounded dots). Source: © Krzic et al. 2013. SoilWeb200 is licensed under a CC BY-NC-SA (Attribution NonCommercial ShareAlike) license.

Soil Structure and Water Flow

Soil structure directly influences the size of pores, and through that it impacts other important soil properties such as the rate of water infiltration, water retention, aeration, and drainage (Figure 4.6). Soil with massive structure or platy aggregates can inhibit water flow, while water moves rapidly through pores in soils with granular aggregates and in single grain soils.

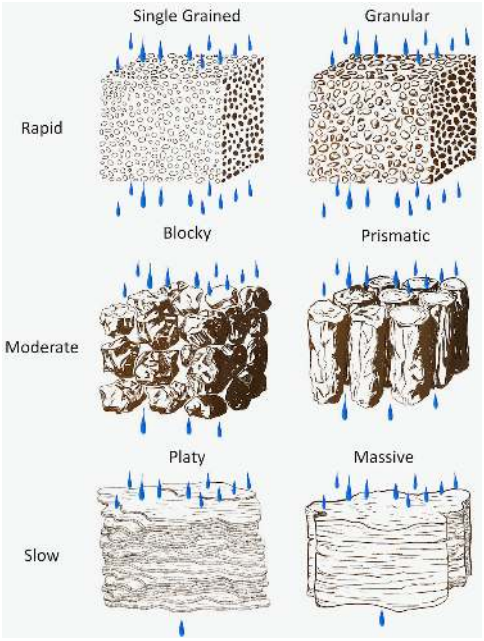


Figure 4.6 Water movement as affected by type (shape) of aggregates. Source: USDA 1991. Modified by Lewis Fausak and licensed under a CC BY (Attribution) license.

Impact of Soil Management on Soil Structure

Soil structure is a very important property for water flow and proper aeration in soils. Yet, it is easily subject to deterioration when stresses are applied onto the soil. The stress may be imposed by cultivation, trampling by animals, heavy traffic from logging, dirt bikes and ATVs, etc. Good structure (i.e., balanced macro- and micro-pores) is important for plant growth because it enhances drainage and aeration, while maintaining plant available water within aggregates. If the soil is subject to management practices involving machinery use, it can be vulnerable to compaction (reduction of total pore space by crushing aggregates and infilling of the pore space). Structure can easily deteriorate under the stresses imposed by tractors and farm implements, and even by raindrop impact or foot traffic. Compaction is especially serious when soils are repeatedly subjected to heavy machinery. For this reason, people who manage soils for plant growth respect several soil management principles to maintain good soil structure:

1. Not running machinery on soil when it is very wet. When water content is very high, the soil cannot offer much resistance to an applied stress. Heavy equipment can easily get stuck in wet soils.

2. Maintaining an adequate level of calcium or Ca^{2+} content by applying some ground limestone, $CaCO_3$ to keep the clays flocculated. Calcium is a divalent cation which creates bridges among soil particles, supporting the formation of micro- and macro-aggregates.
3. Maintaining or adding soil organic matter which can contribute to aggregate strength by binding soil particles together and by encouraging soil microbes which secrete substances that help bind soil particles together.

Repeated tillage, especially under wet conditions, causes the reorientation of particles and compaction of soil just below the depth ploughing. The resulting dense layer of soil, called a “plow pan”, can impede drainage and rooting. Occasional deep cultivation with a “subsoiler” implement, when the soil is dry, may be appropriate to break up such a pan.

Mass-Volume relationships

Mass-volume relationships can be used to define many soil physical properties, including porosity, soil bulk density, and the relative proportions of water and air occupying pore space in a soil.

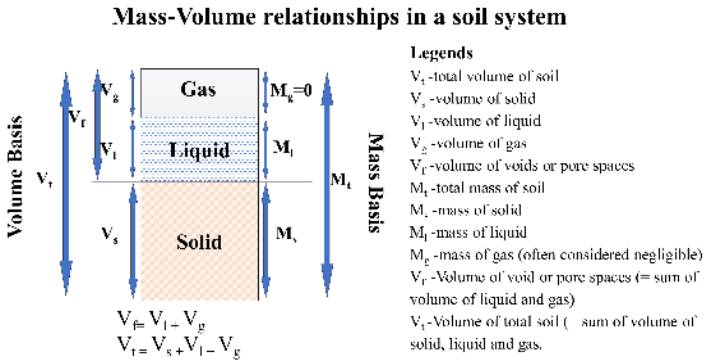


Figure 4.7. Mass-volume relationships between three phases of a soil system. © Asim Biswas, Univ. of Guelph is licensed under a CC BY (Attribution) license.

The total mass of soil (Figure 4.7) is comprised of the mass of solid (M_s), the mass of liquid (M_l) and the mass of gas (M_g). Note however, that the mass of the gas is negligible, and can be assumed equal to zero.

(1) $M_t = M_s + M_l + M_g$

Similarly, the total volume of soil is V_t is comprised of the volume of solid (V_s), the volume of liquid (V_l), and the volume of gas (V_g). As the volumes of liquid and gas make up the volume of **voids** or pores,

(2) $V_f = V_l + V_g$

and the total volume of soil,

(3) $V_t = V_s + V_l + V_g$

In the SI system, the unit of mass is kilogram (kg) and the unit of volume is cubic meters (m^3). Similarly, in the centimeter-gram-second (CGS) system the unit of mass is gram (g) and the unit of volume is cubic centimeter (cm^3).

Densities (ρ):

The relationship between mass and volume for any material in a phase is related to its density. The relationship can be written as

(4) $Mass = Volume \times Density$

(5) $Density = \frac{Mass}{Volume}$

The unit of density is mass unit over volume unit or kg m^{-3} (SI unit) or g cm^{-3} (CGS unit).

Particle density (ρ_p): The density of solid (or soil) particles can be calculated as

(6) $\rho_p = \frac{M_s}{V_s} = \frac{\text{Mass of solid}}{\text{Volume of solid}}$

The unit of ρ_p is kg m^{-3} (SI unit) or g cm^{-3} (CGS unit). In calculating ρ_p, it is difficult to measure the volume of solids in a soil system; unlike the total volume of soil which includes the volume of air and water. Based on the most common minerals found in

soil (i.e., quartz and feldspar), the ρ_p of a mineral soil ranges between 2,600 kg m^{-3} and 2,700 kg m^{-3} (or between 2.6 g cm^{-3} and 2.7 g cm^{-3}). Typically, the average value, 2,650 kg m^{-3} (or 2.65 g cm^{-3}) is used for calculating other soil physical properties. Similarly, ρ_p of organic materials is about 1,300 kg m^{-3} (or 1.30 g cm^{-3}).

Liquid or water density (ρ_l): The density of liquid (or water) can be calculated as

(7) $$\rho_l = \frac{M_s}{V_l} = \frac{\text{Mass of liquid}}{\text{Volume of liquid}}$$

The unit of ρ_l is kg m^{-3} (SI unit) or g cm^{-3} (CGS unit). For most calculations, the density of water is assumed 1,000 kg m^{-3} (or 1.0 g cm^{-3}).

Dry ***bulk density*** (ρ_b): The dry bulk density can be calculated as

(8) $$\rho_b = \frac{M_s}{V_t} = \frac{\text{Mass of solid}}{\text{Volume of total soil}}$$

The unit of ρ_b is kg m^{-3} (SI unit) or g cm^{-3} (CGS unit). Note that ρb accounts for only the mass of solid. As the name implies, ρ_b includes the mass of dry soil. The ρ_b of most mineral soils varies between 1,000 to 1,800 kg m^{-3}. However, this can be impacted by natural processes (e.g., shrink/swell, aggregation, freeze/thaw) and management practices such as tillage, logging, or grazing. Thus, bulk density is often considered a dynamic property at or near the soil surface, though it is reasonably constant lower in the soil profile. Unlike mineral soils, organic soils (soils with organic matter >30% by weight) generally have ρb ranging between 800 and 1,000 kg m^{-3} with values as low as 500 kg m^{-3} in soils with very high levels of undecomposed organic matter (e.g., peat soils).

Water content (θ)

Water content (θ_w) is the amount of water (or liquid) present in a soil system. It can be calculated on a mass basis (known as gravimetric water content, θ_w) or volume basis (known as volumetric water content, θ_v), as the ratio of M_w/M_t or V_w/V_t, respectively. Being a ratio, it is dimensionless. However, the unit of numerator and denominator are often maintained to indicate the type of water content or the calculation procedure. For example, the unit of θ_w is kg kg^{-1} or g g^{-1} and the unit of θ_v is m^3 m^{-3} or cm^3 cm^{-3}; thus, the unit indicates the type of water content and the calculation procedure. For example, a value of 0.22 kg kg^{-1} of soil indicates the amount of gravimetric water content or θ_w of 0.22. Without an indication of units or specification about the type of water content (gravimetric or volumetric), the value of 0.22 alone can be confusing and misleading. This ratio can further be multiplied by 100 to express the water content in percentage, i.e., 22% water content on a mass or gravimetric basis in this example.

Mass-based or ***gravimetric water content*** (θ_w): The mass-based water content also known as gravimetric water content can be calculated as,

(9) $$\theta_w = \frac{M_l}{M_s} = \frac{M_t - M_s}{M_s} = \frac{\text{Mass of liquid (water)}}{\text{Mass of solid}}$$

This is expressed as either dimensionless with specification of gravimetric water content or kg kg^{-1} or g g^{-1}.

Volume-based or ***volumetric water content*** (θ_v): The volume-based water content also known as volumetric water content can be calculated as,

(10) $$\theta_v = \frac{V_l}{V_t} = \frac{\text{Volume of liquid (water)}}{\text{Volume of soil}}$$

This is expressed as either dimensionless with specification of volumetric water content or m^3 m^{-3} or cm^3 cm^{-3}. However, measurement of the volume of liquid in a soil system is very difficult. Thus, often the volumetric water content is estimated from a simple relationship between gravimetric water content and bulk density (see Eq. 14).

Porosity

Porosity is the amount of pore space (or the space occupied by liquid and gas phases) in a soil system. It is an index for the relative volume of pores in the soil. The total volume of pores including that occupied by liquid and gas phases is known as total porosity (f). The pores occupied by gas (or air) is known as air-filled (or aeration) porosity and the pores occupied by liquid (or water) is known as water filled porosity. It is the ratio of two volumes (volume of pores to volume of soils) and the units are m^3 m^{-3} (SI Unit) or cm^3 cm^{-3} (CGS unit). This ratio can be converted to percentage by multiplying with 100.

Total porosity (f): indicates the total pores or voids in a soil system and is calculated as

(11) $$f = \frac{V_f}{V_t} = \frac{V_l + V_g}{V_t} = \frac{\text{Volume of voids (liquid + gas)}}{\text{Volume of soil}} = \frac{V_t - V_s}{V_t}$$

The unit of f is $m^3\ m^{-3}$ (SI unit) or $cm^3\ cm^{-3}$ (CGS unit). The ratio can also be converted into percentage by multiplying with 100. In general, for a mineral soil, f varies between 0.3 and 0.6 $m^3\ m^{-3}$. For sandy soils or soils with an abundance of large particle sizes, f tends to be smaller, while for clay dominated soils or soils with finer particles, f tends to be larger. Remember, large size particles can create larger pores than the smaller sized particles. However, the total number of pores and pore volume are higher in soils with finer particles. Note that total porosity does not provide any information about the pore-size distribution. Surface soils tend to have high porosity and the volume of pores generally decrease with depth in the soil profile as the soil bulk density tends to increase.

Water-filled porosity (f_{w}): the pores filled with water in a soil system and is calculated as,

(12) $$f_w = \frac{V_l}{V_t} = \frac{\text{Volume of liquid (water)}}{\text{Volume of soil}} = \theta_v$$

The unit of f_{w} is $m^3\ m^{-3}$ (SI unit) or $cm^3\ cm^{-3}$ (CGS unit).

Air-filled porosity *or* ***aeration porosity*** (f_{a}): the pores filled with gas (air) in a soil system and is calculated as,

(13) $$f_a = \frac{V_g}{V_t} = \frac{\text{Volume of gas (air)}}{\text{Volume of soil}}$$

The unit of f_{a} is $m^3\ m^{-3}$ (SI unit) or $cm^3\ cm^{-3}$ (CGS unit). Air-filled porosity is the proportion of the bulk soil occupied by air at any point in time. It is related to the total porosity, and volumetric water content, as both air and water share the same pore volume. Air-filled porosity can then be determined by subtracting the volumetric water content (cm^3 of water per cm^3 of soil) from the total porosity of the soil. As soil water content varies in time and with soil depth, so does air-filled porosity. Its value will be minimal following significant rainfall or irrigation events. Typically, values should be higher than 0.10 to 0.15 $cm^3\ cm^{-3}$ at field capacity, following drainage.

Relationships among soil physical parameters

While many physical quantities can be calculated from the mass-volume relationships of the three phases, measurement of mass and volumes of different phases are sometimes difficult. Alternatively, easy to measure parameters may be used to estimate parameters that are more difficult to measure. Some useful formulas are provided below.

Relationship between volumetric water content (θ_v), gravimetric water content (θ_w) and bulk density (ρb) can be written as,

(14) $$\theta_v = \theta_w \rho_b$$

Relationship between total porosity (f), bulk density (ρ_b), and particle density (ρ_p) can be written as,

(15) $$f = 1 - \frac{\rho_b}{\rho_p}$$

Relationship between total porosity (f), air-filled porosity (f_{a}), and volumetric water content (θ_v) can be written as,

(16) $$f_a = f - \theta_v$$

Compaction

When soil is subjected to pressure, it tends to compress and increase in density. The principal mechanism of soil compression is the reduction of porosity through the partial expulsion of air and/or water from the compressing soil body. For example, static pressure or vibration can cause particles to reorient within the same volume and create a closer packing arrangement. During this reorientation, total porosity, in particular macro-porosity, reduces. In the case of saturated soils, the decrease in porosity takes place as the water is pushed out of macro-pores, unlike in a dry soil where air is pushed out. The expulsion of air is faster than the expulsion of water, which is a slow process as the largest size pores empty first and then the smaller and smaller sized pores. In soils with intermediate soil water content, the air will be expulsed first and then the water. The term compaction is generally applied to the compression of an unsaturated soil body, resulting in a reduction in air volume. While the term consolidation, is generally used for the compression of saturated soil by squeezing out water.

In an agricultural context, a soil is considered to be compacted when the total porosity, in particular the air-filled porosity, is low and restricts aeration, as well as when the soil is so dense and its pores are so small that root penetration and drainage are impeded. Compaction also creates a major issue in field management, particularly tillage. Compaction can occur at the surface (e.g., surface crust) or at the subsurface (e.g., hardpan). A major cause of man-made compaction is related to the use of heavy machinery for agricultural operations. For example, as high as 90% of the field surface can be traversed by tractor wheels during traditional farming operations. Compaction of forest soils is also associated with heavy machinery use, particularly on landings (areas where harvested tress are stored, processed and loaded onto trucks); where it may restrict or prevent the growth of planted tree seedlings or natural regeneration, and where some rehabilitation of compacted areas may be required. Examples of compaction in urban settings include the use of heavy machinery during construction, and excessive trampling by people or bicycles.

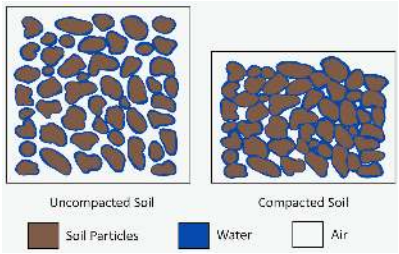
Can You Dig It!

What does camping have to do with soils?



Campsite. © Benjamin Hollis is licensed under a CC BY (Attribution) license.

Hiking, mountain biking and tenting can all have significant impacts on the soil. Even by walking, our body weight compresses the soil. Soil compaction reduces the soil porosity and increases soil bulk density. In particular, macro-pores are collapsed, which reduces infiltration and the movement of water through the soil, and can result in overland flow and erosion during rainfall events. The reduction in pore space also reduces aeration, which can negatively affect soil organisms and vegetative growth. Extensive trampling may result in a loss of vegetation, which is commonly seen along hiking trails and at campsites. Campsites typically have a about a 10 m^2 high-impact area.



Reduction in pore space as a result of compaction. © Lewis Fausak is licensed under a CC BY (Attribution) license.

In high-intensity camping, the impact on soils occurs very quickly, and restoration may take decades. But park rangers and recreation managers are working to create low-impact camping opportunities. In areas where dispersed camping is permitted, tents can be moved daily to reduce the potential for long term impact. Studies suggest that low impact campsites did not compact surface soils to the extent that the root growth of the surrounding vegetation was hindered. But selecting the site to pitch your tent matters; sandy soils are not as susceptible to compaction as fine textured soils. So, avoid sensitive soils and vegetation, stay on the trails, and remember that leave-no-trace camping is more than just packing out your garbage!

References:

Brevik, E.C. and Tibor, M.A. 2014. Impact of camping on soil properties of Strawberry Lake, North Dakota, USA. EGU General Assembly Apr 27-May 2, 2014. Vienna, Austria.

Eagleston, H., and Marion, J.L. 2017. Sustainable campsite management in protected areas: A study of long-term ecological changes on campsites in the boundary waters canoe area wilderness, Minnesota, USA. Journal for Nature Conservation 37: 73-82. doi:10.1016/j.jnc.2017.03.004

Leave No Trace Canada. 2009. Leave no trace principles. https://www.leavenotrace.ca/principle-travel-camp-durable-surfaces

Marion, J.L. and Cole, D.N. 1996. Spatial and temporal variation in soil and vegetation impacts on campsites. Ecological Applications 6(2): 520-530.

WATER RETENTION IN SOILS

Soil can retain water for a substantial amount of time, supporting the growth of plants and soil organisms. Despite the continuous pull from gravity, the water, which enters soil after precipitation or irrigation, stays in the soil for enough time so that plant roots can extract the water to survive. The water is held in the soil in such a way that gases can move through the soil pore spaces and allow oxygen to reach to the roots. Thus, it is critical to understand how soil retains water and acts as reservoir to support life on land. There are two important characteristics of soil water: the amount of water in a given amount of soil (soil water content); and the forces holding water onto the soil matrix (soil water potential). Soil water influences many processes including gas exchange with the atmosphere, the movement of nutrients and chemicals to plant roots, soil temperature, and swelling and shrinking. The forces exerted by the solid matrix on water influences how water is absorbed by plant roots, the drainage that may occur due to gravity, and the extent of upward movement of water and solutes against gravity. These processes are possible because water in soil behaves quite differently than the water in a drinking glass or a swimming pool. Water in soil is strongly associated with the solid particles, particularly soil colloids; and the interactions of water molecules with soil particles changes the behaviour of both.

Properties of water

Soil water influences many processes in soil mainly due to the unique structure of the water molecule. Water is a simple molecular compound, which contains two hydrogen atoms that are covalently bonded to an oxygen atom by sharing electrons in a V-shape. This V-shaped and non-symmetric arrangement of the atoms in the water molecule produces an electric field. The hydrogen atoms, on one side, tend to exhibit electropositive behaviour, while the oxygen atom, on the other side, tends to exhibit electronegative behaviour creating a dipole characteristic. Because of the uneven distribution of electron density, water molecules exhibit polarity and contribute to many properties that allow water to play a unique role in the soil. Water molecules link together as positively charged hydrogen atoms attracts negatively charged oxygen atoms and form a chain-like grouping. Further, due to the polarity of water molecules, they are attracted to charged ions (e.g., Na^+, K^+, Cl^-) and colloidal surfaces (e.g., clay and soil organic matter).

Hydrogen bonding

Because of the polarity of a water molecule, the electronegatively charged oxygen atom can develop an attraction force with the electropositively charged hydrogen as two water molecules come close together. This attraction force creates an intermolecular bond known as hydrogen bond between the protons of the hydrogen atom of one water molecule and the oxygen atom of the other. This force of attraction effectively bonds water molecules together.

Adhesion and cohesion forces

Hydrogen bonds create two forces that are responsible for water retention and movement in soils (Figure 4.8a). The attraction force between water molecules and solid particles is known as **adhesion** force (force between two different materials). The attraction force between water molecules is known as **cohesion** force (force between similar materials). Generally, adhesion forces are much stronger than the cohesion forces as water molecules are tightly held onto the soil-solids (also known as adsorption). These tightly bound water molecules further hold onto other water molecules through cohesion forces. Together, the adhesion and cohesion forces make it possible for soil solids to hold water and to allow water movement through the soil. Increasing distance between the water molecule and the solid surface decreases the binding attraction between the water molecule and the solid surface, thus soil water farther away from soil colloids may not be retained onto the surface of a soil particle (e.g., clay).

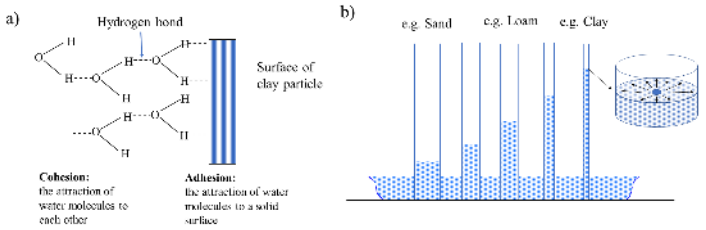


Figure 4.8. Cohesion and adhesion forces present between water and soil particles (a), and capillary rise of water in different sized (diameter) pores present in different soil textures (b). © Asim Biswas, Univ. of Guelph is licensed under a CC BY (Attribution) license

Capillary rise

If a cylindrical glass tube is kept in contact with a liquid media such as water in a reservoir open to the atmosphere, water will spread over the inside wall of the glass tube. Due to the stronger adhesion force, the attraction between water molecules and the glass wall will be stronger than the cohesion force between the water molecules. Due to this attraction, the glass wall will "pull" water and there will be a rise of water within the glass tube. This is known as capillary rise (Figure 4.8b). The height of capillary rise of water will be determined by the closeness of the glass walls which creates the attraction force and holds the water against the force of gravity. If the walls are close (or the diameter of the tube is small), the adhesion force will be stronger than the cohesion force, and the water will rise higher. On the other hand, if the diameter of the tube is larger, the cohesion force may surpass the adhesion force and the height of rise of water may be small. Consider the situation in soil, where soil pores created by soil particles may be considered as a buddle of tubes. Now, small pores created by small sized particles (e.g., clay) will have a greater height of rise of water compared to the large pores created by the large sized particles (e.g., sand).

Soil water content

Soil water content is the amount of water present in soil. The maximum amount of water a soil can hold is equal to the total volume of pores. When all the pores are filled with water, the soil becomes saturated with water and the water content is known as **saturated water content** (Figure 4.9). As that saturated soil starts to dry, the amount of water that the soil can hold is determined by various factors including the attraction force between the water and soil particles or the adhesion force. In a saturated condition, all pores are filled with water and the water molecules form layers of water on the soil particles. However, not all the water molecules are equally attracted to the soil particles. Water molecules closest to the soil particle are attracted more strongly than the water molecules far away from the soil particles. Thus, as the distance between soil particles and water molecules increases, the adhesion force between them decreases. Therefore, the water molecules in large sized pores that are farther away from a soil particle surface are weakly held by the soil particles (i.e., the adhesion force is weak). Now, if the saturated soil is allowed to drain freely, the gravitational force surpasses the adhesion force between the soil particles and water molecules. The soil drains until the force created by gravity and the force between soil particles and water molecules equalizes. The amount of water that can be drained by the gravitational force is known as **gravitational water** (Figure 4.5). The gravitational water leaves a soil system quickly and is not usable by plants or microbes.

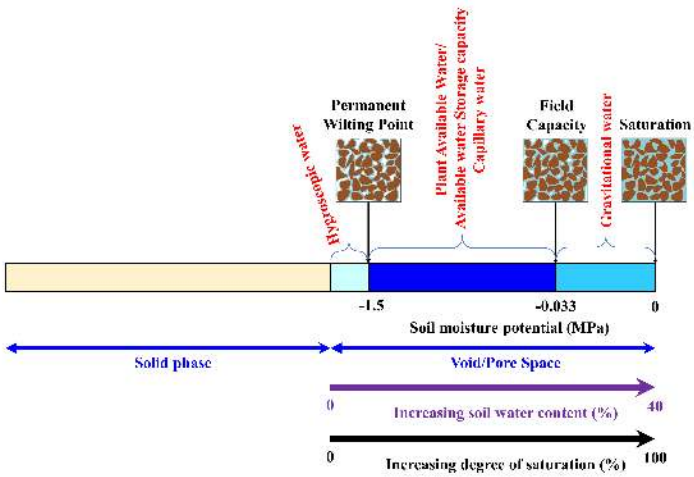


Figure 4.9. Common terms used to describe soil water content. Adapted from Brady & Weil (2002) and McCauley et al. (2005). Modified by Asim Biswas. CC BY. © Devine and O'Geen (2019) adapted by Asim Biswas, Univ. of Guelph is licensed under a CC BY (Attribution) license.

Once the gravitational water has drained, water is held in soil within the smaller sized (micro- or capillary) pores. Due to their smaller size, the adhesion force between soil particles and water molecules in **micro-pores** and the cohesion force between the water molecules can hold the water within these pores. External forces, such as the force created by plant roots, can take the water out of these pores. However, as the water molecules closer to the soil solid are attracted by increasing amount of forces, the external force must be stronger than the attraction force to take water out of those pores. Plant roots can only exert pressures up to a certain extent and can uptake a limited amount of water present in soil. The amount of water plants can extract from soil is known as **plant available water** or **available water storage capacity (AWSC)**. As this water is retained in the capillary pores and held by cohesion forces, it is also known as **capillary water** (Figure 4.9). Beyond this limit of attraction, the water molecules are attached to the soil particles mainly through adhesion forces. This is known as **hygroscopic water**. Hygroscopic water generally represents a small amount of the water in soil. This water is not available to plants as plants cannot exert enough force to take hygroscopic water out of the soil. The forces acting between the soil and water molecules can be quantified as soil matric potential, which is described later in this chapter.

As soil starts to dry from a saturated condition, the gravitational water drains out first, followed by plant available water, which decreases as the soil continues to dry out, and eventually may become unavailable as hygroscopic water. The water content when the gravitational water stops, and the plant available water starts is known as **field capacity** (Figure 4.9). Qualitatively, field capacity is described as the amount of water left in soil two to three days after being saturated (from irrigation or precipitation) and allowed to freely drain by gravitational force. Similarly, the water content at which plants cannot extract any more water from soil is known as **permanent wilting point**. It is the amount of soil water at which plants cannot take up any more water from soil, start to wilt, and cannot recover (even if water is added subsequently).

Soil water potential

In the previous section, soil water content was defined as the amount (mass or volume) of water in a given mass or volume of bulk soil: gravimetric and volumetric water content, respectively. The measurement of soil wetness is required for understanding the soil water balance, irrigation management, drought assessment and flood forecasting. Soil water also has energy and measurement of the energy of soil water is required to predict the movement of water in soil. Like all mass, soil water moves from locations of relatively high potential to locations of relatively low potential.

What is energy?

Indeed, all mass has energy. The units of energy are Joules (J). Unlike mass, however, energy is difficult to define and conceptualize. An example of the definition of energy from an introductory physics textbook is as follows:

"[energy is a] quantity that is associated with a *state* (or condition) of one or more objects"

"[energy is] a number that we associate with a system of one or more objects. If a force changes one of the objects by, say, making it move, then the number changes"

These definitions seem a bit "wishy washy", certainly not as precise as mass (i.e., the amount of matter in an object). The concept of energy becomes clearer when qualified as energy associated with: (1) motion: kinetic energy; (2) separation between an object and an object with known energy: potential energy (gravitational, electrical); and (3) temperature: heat energy.

Soil water potential – units

Because the movement of water in soil is slow (very low velocity), the kinetic energy of soil water is less important than the potential energy of soil water. Hence, soil water energy is simply called **soil water potential**. Similar to the mass of water in soil that is quantified per unit mass of soil (gravimetric water content) or per unit volume of soil (volumetric water content), soil water potential is also quantified per unit of soil water:

1. potential energy per unit mass: J kg^{-1}
2. potential energy per unit volume: J m^{-3} or Pa
3. potential energy per unit weight: J N^{-1} or m (equivalent height of water)

Which units are used to express soil water potential are a matter of choice and it is possible to convert between all three soil water potential units.

Potential energy is always quantified relative to a reference reservoir

Soil water potential is quantified as the difference between the potential energy of soil water and water in a reference reservoir. The reference reservoir used to quantify soil water potential is *pure water at atmospheric pressure at a specified elevation*.

An example of the reference reservoir would be the surface of water in a beaker sitting on a lab bench. By convention, we set the potential of the reference reservoir equal to zero in order for the potential of the reference reservoir to be the same regardless of its units (i.e., J kg^{-1}, Pa, or m). Further, this convention allows quick assessment of whether soil water potential is greater than, equal to, or less than the reference reservoir, making it a convenient and intuitive way to quantify and express soil water potential.

Work and formal definition of soil water potential

According to the Soil Science Society of America, the formal definition of soil water potential is: "The amount of work that must be done in order to transport an infinitesimal quantity [mass, volume or weight] of pure water, at a specified elevation and at atmospheric pressure [i.e., the water in the reference reservoir], to the soil water at the point under consideration [without changing its temperature]" (Soil Science Society of America, 1997).

In other words, soil water potential is equal to the amount of work that we must do on water in the reference reservoir to move it to a location in the soil without changing its temperature.

To better understand this definition, the concept of work must be defined. Work has a formal definition in physics, which can be stated as *an action on an object that changes the potential energy of the object*. For example, by lifting a box from the floor to the top of a table the gravitational potential of the box has increased in proportion to its change in elevation (from floor to table).

So, how much work must be done (i.e., how much effort is required) to transport the water in the reference reservoir to a location in the soils? Analogies to familiar phenomena will help us answer this question. Like soil, sponges have pores and can hold water. Imagine using a dry sponge to clean up some water that has been spilled on a table (or better yet, actually do it). The water on the table is analogous to the water in the reference reservoir and the sponge is analogous to the soil. Without any effort on your part, the water spontaneously moves into the sponge. How much work have you done on the water to transport it into the sponge? Did you break a sweat? No, it did not take any effort at all. In fact, it took less than zero effort, meaning that you did "negative work" on the water (i.e., the sponge did all the work). Now imagine how much effort is required on your part to transport the water from the sponge back to the table surface. You squeeze the sponge to the release the water. Does the sponge release all of the water? No, regardless of how hard you squeeze, the sponge still holds on to a small amount of water. In this case, you did "positive work" on the water in the sponge to transport it back to the table and you still need to do a lot more work to remove the remaining water in the sponge.

Drawing from the analogy of the sponge being the soil and water on the table being the reference reservoir, it can be concluded that the potential of water in the soil is less than the reference reservoir because negative work is required for the soil to absorb water and positive work is required to extract water from the soil. Further, it can be concluded that it is the forces of attraction between soil particles and water molecules that affects the soil water potential. In order to quantify the total potential of soil water, we need to consider all of the forces on soil water affecting its potential and they are summarized in Table 4.2.

Table 4.2. Components of total soil water potential

Forces effecting potential energy	Name of soil water potential component	Reference reservoir	Magnitude of component potential in the reference reservoir
Adsorptive forces between soil particles and water, capillary forces (above the water table; unsaturated soil)	Matric potential, h_m	Water at atmospheric pressure	0
Osmotic forces from dissolved solids	Osmotic potential, h_o	Pure water[1]	0
Gravitational forces, elevation	Gravitational potential, h_g	Water at a reference elevation	0
Pressure from overlying water (below the water table; saturated soil)	Pressure potential, h_p	Water at atmospheric pressure	0
All of the above	Total potential, $H = h_m + h_o + h_g + h_p$	Pure water at atmospheric pressure at a reference elevation	0

[1]In pure water the number of dissolved solute particles is zero, and therefore the osmotic potential is zero.

Matric potential is a result of the attraction of water molecules to the soil solids due to **adsorption** and **capillarity** (similar to the sponge). As the matric potential reduces the degree of freedom of movement of water relative to the reference reservoir, it is a negative potential. A dry soil has a very low matric potential, while a wet soil has a higher matric potential (i.e., closer to zero).

Osmotic potential is a result of the attraction of water to solutes, which reduce the energy level of water in soil solution. Osmotic potential can impact the uptake of water by plant roots in saline soils, but the solute concentration in most soils is low and does not affect water movement.

Gravitational potential is the result of the pull of gravity, and is defined relative to a specific elevation (e.g., top of the soil profile).

If the soil is completely saturated with water, under a **water table** for example, then there is no matric potential but a pressure potential resulting from the weight of the overlying water (such as the pressure on your eardrums as you dive at the bottom of a swimming pool).

At any given location in the soil, the total potential of the water is the sum of the soil water potential components, and soil water will flow from locations of high total potential to locations of low total potential. Not every soil water potential component contributes to the total soil water potential in all circumstances. For example, above the water table, the pressure potential is zero because the forces that increase the pressure potential are water pressure forces below the water table. Below the water table, matric potential is zero because the forces that change matric potential, adsorption and capillarity, are very weak in saturated soil.

Soil water retention curve

Soil water has both mass and energy. The amount of soil water (mass or volume) is quantified through measurement of gravimetric or volumetric water content, and soil water potential is quantified relative to reference reservoirs of known potential. Perhaps the next obvious question is whether the amount of soil water (mass or volume) has any relationship to the potential of soil water?

As it turns out, Edgar Buckingham asked the same question back in the late 1800s. He discovered that there was a relationship between the amount of soil water and the soil water potential. He showed that there was a positive relationship between soil water matric potential and soil water content.

Edgar Buckingham was an American soil physicist that understood that mass moves from locations of high total potential to locations of low total potential. It follows that if the total potential of soil water is the same at all locations, there will be zero water flow. Buckingham created a simple soil water system where he was confident that total soil water potential was the same at every location. He placed soil columns of 1.2 m height in a shallow pool of water. Initially the soil in the columns were dry and this caused water from the reservoir to move into the soil, but after a long time, the movement of water from the reservoir into the soil ceased, and the total potential of the soil water was equal at every location in the soil.

Buckingham reasoned that it was the capillary and adsorptive forces that caused water to move from the reservoir into the soil, and that water was held in the soil against the force of gravity – therefore, gravitational and matric potential were the two key components contributing to the total soil water potential. In fact, under the constraint of equal total potential at all locations, gravitational and matric potential must have equal magnitude, but opposite signs. The height to which the water rises in the soil depends on the size of the pores (Figure 4.10a). In soils with smaller pores, the water rise will be higher because of greater capillary forces in small pores. By measuring the water content of the soil at different locations along the soil columns, Buckingham was able to show the relationship between matric potential and soil water content. It is this relationship, between moisture content and the matric potential, which defines the soil water retention curve (Figure 4.10b).

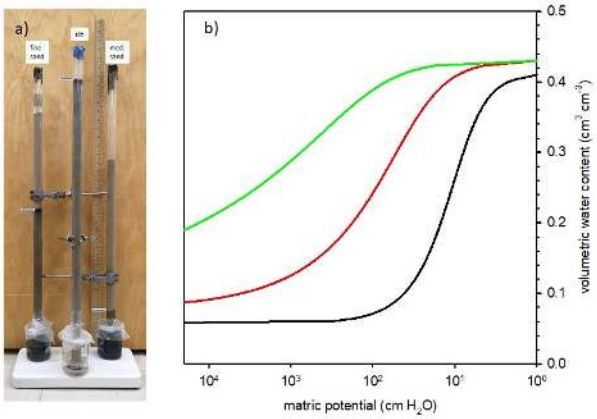


Figure 4.10. Soil water retention simulation of Buckingham's experiment with columns of fine sand, silt and medium sand (blue lines shown the height to which the water from the reservoir has risen) (a) and soil water retention curves for three soils of contrasting texture (b). © Miles Dyck is licensed under a CC BY (Attribution) license.

Each soil has a unique water retention curve, which can be expressed mathematically as a function: h_{m}(\theta) or $\theta(h_m)$ (i.e., matric potential is a function of water content and *vice versa*). To better understand why each soil has a unique water retention curve, soil physicists adopt what is called the *capillary tube model of soil pores.* This model acknowledges that soil pores behave in a similar way to capillary tubes even though soil pores do have the same shape as capillary tubes. As observed by Buckingham, when a column of dry soil is placed in a water reservoir, water spontaneously rises into the soil just like water rises in a capillary tube placed in a water reservoir. It is the adsorptive and capillary forces (the forces associated with matric potential), that cause the water to rise up into the soil column. Therefore, the height of capillary rise equation can be modified to relate matric potential to the effective radius of water filled pores as follows:

(17) $h_m \leq \frac{-0.15}{R*}$

where, h_m is the matric potential (cm H_2O), and $R*$ (cm) is the effective radius of water filled pores.

Therefore, the soil water characteristic curve is a representation of the soil's cumulative pore-size distribution.

Water availability to plants

Information on how different soils hold and release water is important in understanding water availability for plants and microbes, and forms the basis of water management practices in agriculture, agro-forestry, land reclamation and urban ecosystems. The **available water storage capacity** (AWSC), provides one approach to assess the proportion of soil water available for plant uptake. Recall that AWSC is defined as the difference in volumetric water content between an upper limit, referred to as field capacity, and a lower limit, referred to as permanent wilting point (Figure 4.9). The available water is calculated for the rooting zone (i.e., m of water per rooting depth). This provides an indication of the amount of irrigation (mm) needed to refill the water reservoir between permanent wilting and field capacity, and answers the two basic questions on when to irrigate and how much water to apply. However, this approach assumes that water is equally available to plants from field capacity down to the permanent wilting point.

Irrigation consists of artificially applying water to maintain its availability to the roots so that plant stress is reduced. Only about two thirds of AWSC is fully available without causing water stress for the plant as at some point, the rate of extraction of water by the plant is faster than the capacity of the soil to supply it (Reicosky and Ritchie, 1976; Rekika et al., 2014). Therefore, not only is maintaining the water supply important, but also the timing at which this water is transferred to the plant though the soil matrix. Better timing can be achieved with irrigation management tools, such as tensiometers and soil-moisture probes, which evaluate water availability within the soil matrix. In the past, overhead and furrow irrigation were extensively used to replenish the soil water reservoir. Currently, drip irrigation and subsurface irrigation are increasingly used, resulting in managing water availability to the plant within a limited soil volume. By adapting irrigation to the specific environmental growing conditions, we can ensure optimal crop productivity and optimal water use (Caron et al., 2015; 2017).

SOIL WATER MOVEMENT

Saturated and unsaturated soil water flow

Water flows in soil from locations of high total potential to locations of low total potential. The magnitude of vertical soil water flow under saturated and unsaturated soil conditions are described by two very similar equations. Flow of water in saturated soils (i.e., when all soil pores are water-filled) is described by Darcy's Law:

(18) $q_w = -K_s \frac{\Delta H}{\Delta z}$

where q_w is the soil water flux density with units of $\frac{m^3 H_2O}{m^2 soil \cdot s}$; K_s is called the saturated hydraulic conductivity having units of $\frac{m^3 H_2O}{m^2 soil \cdot s}$ per unit of hydraulic gradient; and $\frac{\Delta H}{\Delta z}$ is the hydraulic gradient (i.e., the change in sum of pressure and gravitational potentials over some depth interval, Δz); is dimensionless when h_p is expressed in m H_2O.

The negative sign in Darcy's law is used to specify flow direction. Upward flow (i.e., flow in the increasing *z* direction, is positive and downward flow, in the decreasing *z* direction, is negative). Downward flows are associated with positive gradients ($\frac{\Delta H}{\Delta z} > 0$), meaning the total potential decreases with decreasing depth.

Edgar Buckingham acknowledged that soils were often unsaturated (i.e., only a fraction of the pores were water-filled), and under these conditions, the hydraulic gradient depends on changes in matric potential within the soil, and that the hydraulic conductivity under unsaturated conditions changes with soil wetness. The result is the Buckingham-Darcy Flux Law:

(19) $q_w = -K(\theta) \frac{\Delta H}{\Delta z}$

where, $K(\theta)$ is the hydraulic conductivity as a function of water content, simply referred to as the unsaturated hydraulic conductivity, and $\frac{\Delta H}{\Delta z}$ is the hydraulic gradient (i.e., the change in sum of pressure and gravitational potentials over some depth interval, Δz)

Hydraulic Conductivity

The **hydraulic conductivity** (K) is a soil hydraulic property that represents the soil's ability to facilitate water flow. Each soil has a unique hydraulic conductivity that is related to the pore-size distribution of the soil. Figure 4.11 shows K as a function of matric potential for soils with three different textures. When saturated, soils with large, continuous pores will have a much greater hydraulic conductivity compared to soils with small, discontinuous pores. As soil water content decreases, the hydraulic conductivity decreases because: (1) the total volume of water-filled pores decreases; and (2) the size of water-filled pores decreases (Figure 4.12). Therefore, soil hydraulic conductivity is a function of the volumetric water content.

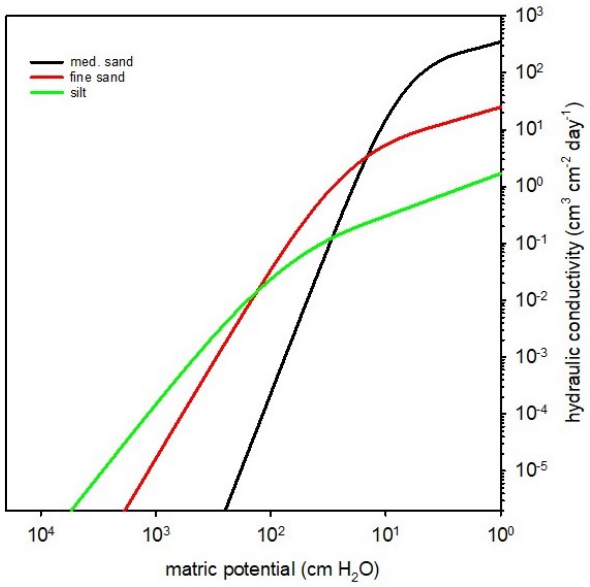


Figure 4.11. Soil hydraulic conductivity curves for three contrasting soil textures. © Miles Dyck is licensed under a CC BY (Attribution) license.

At saturation, the hydraulic conductivity of each soil reaches the maximum value (K_s), the saturated hydraulic conductivity, but decreases exponentially as the soil becomes drier (y-axis for hydraulic conductivity is on a logarithmic scale). The high Ks of the loamy sand relative to the loam and silty clay loam soils is a reflection of the larger pores associated with coarse-textured soils.

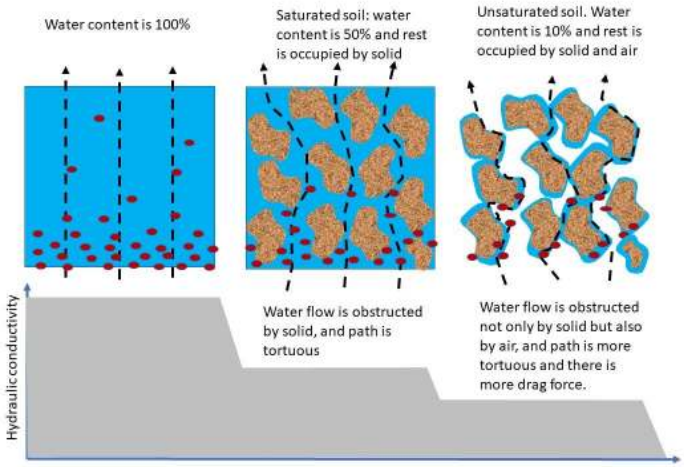


Figure 4.12. Reduction in flow pathways with decreasing soil water content. © Bing Si is licensed under a CC BY (Attribution) license.

Hydrological processes and the soil water balance

Infiltration and runoff

Infiltration describes the entry of water into the soil through the soil surface, and it determines the surface partitioning of water coming from rainfall or irrigation into either infiltration or runoff (Figure 4.13). The infiltration rate quantifies the volume of water entering a cross-sectional area of soil per unit time (e.g., mm hr^{-1}). The infiltration rate will generally drop over time due to the gradual saturation of the soil matrix. The infiltration capacity of the soil will depend on both the pore-size distribution and the continuity of the pores, in addition to the antecedent water content. When the rate of supply exceeds the infiltration capacity of the soil at a given time, runoff will occur. Runoff is an important process in water erosion as runoff waters carry and redistribute sediments along slopes, and create rills or gullies at the surface soil.

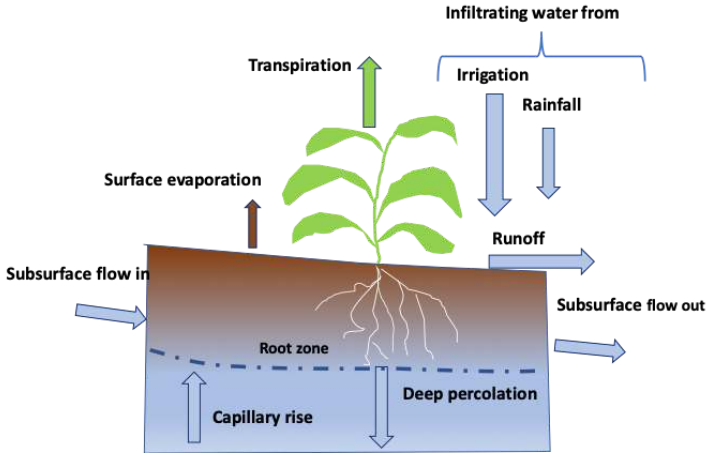


Figure 4.13. Partitioning of water in the soil. Source: FAO Irrigation and Drainage Paper No. 56 (CC BY-NC-SA). Modified by Jean Caron. © FAO Irrigation and Drainage Paper No. 56 adapted by Jean Caron is licensed under a CC BY-NC-SA (Attribution NonCommercial ShareAlike) license.

Drainage, capillary rise and deep percolation

At the bottom of the soil profile, three components interact to affect upward or downward water movement. **Percolation** refers to the downward movement of water through soil. Drainage is partitioned into two components: surface drainage, which refers to management practices at the land surface to carry excess water towards gullies and ditches; and subsurface drainage which refers to the fraction of water moving downward that is intercepted by a drain tile network, commonly made up of corrugated plastic pipes. Finally, the groundwater component can supply water to the soil through water table fluctuations under saturated conditions or through capillary rise. All three components (percolation, drainage and capillary rise) are time-dependent and therefore have units of volume of water per unit surface area per unit time.

The soil water balance

The soil water budget represents the balance between inputs and outputs from a given soil unit (Eqn. 20). Soil water storage (ΔS) is one component in a water balance. It refers to the change in water storage volume within a given soil unit over a specified time period, and is usually expressed in equivalent height of water per unit time (e.g., mm day^{-1}).

(20) $P + I - ET + Q_i - Q_o - \Delta S - DP = 0$

where, P = precipitation; I = irrigation; ET = evapotranspiration; DP = drainage and deep percolation; Q_i = other inflow (e.g., capillary rise, and water table fluctuations); and Q_o = other outflow, (e.g., runoff and lateral flow).

Water balance measurements (e.g., using measuring devices called lysimeters) are time and resource consuming, and thus estimates may be made from basic site measurements of soil water content and potential, soil properties and weather data over a given time interval.

SOLUTE TRANSPORT

Fertilizers, pesticides and growth promoters are frequently used in agriculture, and these agrochemicals have greatly contributed to increased crop yields since the 1950s (Ritchie and Roser 2013) . Fertilizers and pesticides are also commonly used in urban settings (e.g., private lawns, city parks) as well as golf courses and playfields. However, excessive application of chemicals and improper management, reduces nutrient use efficiency, enhances greenhouse gas emission and increases the risk of environmental pollution. A prerequisite for optimum application and proper management of chemicals in soil is to understand how chemicals move in soil and toward plant roots.

As described previously, soil is a three-phase system. Chemicals in the form of dissolved constituents (i.e., solutes) can move through soil in the liquid phase. The transport of solutes in soil solution is important for the uptake of nutrients by plant roots. There are two main mechanisms of solute transport in soils: diffusion and mass flow.

Molecular diffusion

Molecules in a liquid have an internal kinetic energy. As a result, a molecule in a static liquid is in constant motion. The direction in which an individual molecule moves is constantly changing due to its collision with neighboring molecules, known as Brownian motion. While individually, each molecule moves randomly, without a preferred direction, collectively, more ions or molecules tend to move from areas of higher concentration to area of lower concentration (Figure 4.14). In fact, the larger the concentration difference over a certain distance, the faster the spread. To quantitatively measure the rate of the spread we define the flux, as the mass of solute moved perpendicularly across a unit area per unit time (kg m^{-2} s^{-1}), following Fick's law of **diffusion**.

(21) $$q_s = -D_s \frac{\Delta C}{\Delta x}$$

where: q_s is the chemical flux (kg m^{-2} s^{-1}); ΔC is the concentration difference between two locations ($\Delta C = C_2 - C_1$) separated by distance Δx ($\Delta x = x_2 - x_1$); and D_s is and a proportionality constant called the diffusion coefficient (m^2 s^{-1}).

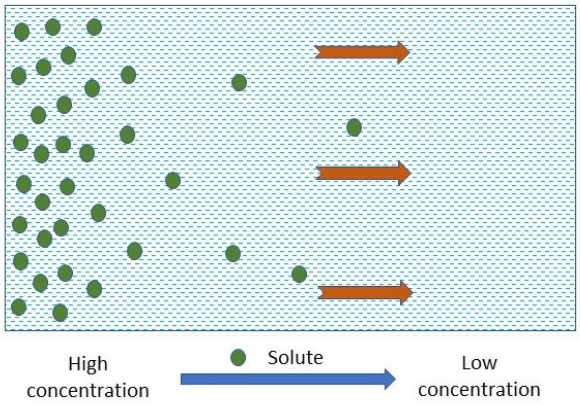


Figure 4.14. Solute (e.g., Cl^-, NO_3^-) flows from high concentration to low concentration in water. © Bing Si is licensed under a CC BY (Attribution) license.

Chemicals are gradually spread out by diffusion with time, and will eventually be uniformly distributed in the media. For a banded fertilizer, fertilizer moves, due to the concentration gradient, from area of high concentration (where the fertilizer band is) to area of low concentration in the surrounding soil. As illustrated by Figure 4.15, the chemicals spread out from the vicinity of the fertilizer band and are distributed horizontally in a bell shape, with the peak at the center of the fertilizer band. With time elapsing, the bell shaped distribution curve extends further away from the fertilizer band, resulting in reduced peak concentration.

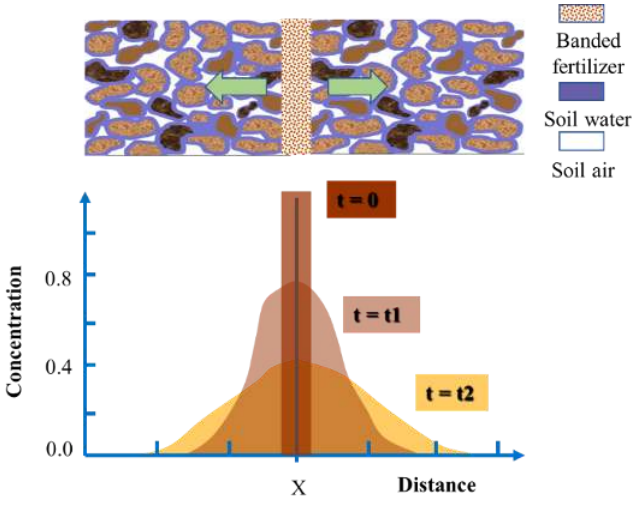


Figure 4.15. Spreading of banded fertilizer with time (t = 0, t_1 and t_2, where $t_2 > t_1 > 0$) due to diffusion. © Bing Si. CC BY.

In addition to concentration gradients, the diffusion coefficient also determines how fast diffusion occurs and how large the chemical flux is. The volume of water-filled pores for transport, the tortuosity of the transport pathway and drag force all effect the

diffusion coefficients in soil.

Mass flow

Solute transport by diffusion occurs within soil solution, but the water as a whole remains stationary. However, when water is under a pressure gradient in the soil, it will move from high potential to low potential, and any constituents in solution or suspension will move with the water. This movement of water and solutes together due to a pressure gradient, is called **mass flow**. The amount of a solute transported per unit area of soil by mass flow is given by the solute flux (q_s, kg m^{-2} s^{-1}), which is the product of the water flux and the solute concentration in the solution:

(22) $q_s = q_w C$

where, q_w is the water flux (m s^{-1}) and C is the solute concentration in water (kg m^{-3}).

Nutrient transport to roots

Plants develop a network of fine roots for water and nutrient uptake. Through direct contact with soil, roots can intercept nutrients. However, plant roots generally have only direct contact with about 1% of soil volume. Thus, roots need additional mechanisms to access nutrients and water from the remaining 99% of the soil volume.

Mass flow – As a plant transpires water, water is absorbed from soil through the root system. While water flows from the soil in the vicinity of the root hairs, mobile nutrients are carried along with the water, effectively fetching nutrients that are away from the root surface. Mass flow accounts for nutrient uptake of mobile nutrients, such as nitrogen and sulfur. Applying fertilizers in the root zone elevates the nutrient concentration in the solution, enhancing root uptake.

Diffusion – Roots first use up nutrients directly around the root hairs. As the concentration of nutrients around the root system drops, nutrient concentrations may be still high at locations that are slightly farther from the fine roots. This creates a concentration gradient, which moves nutrients from higher concentrated areas toward the surface of roots by diffusion. Note that the chemical flux is proportional to the concentration gradient, consequently the larger the concentration gradient between the root surface and the surrounding soil, the faster the flow of nutrient to the root surface. A large concentration gradient can be created by applying nutrients as close to the root zone and as frequently as possible.

When the transpiration rate is high, mass flow can be dominant mechanism for nutrient uptake. But at low transpiration, nutrient uptake mainly relies on diffusion. When root hairs are not densely distributed in the soil, diffusion from soil to roots can be a bottleneck for root nutrient uptake.

SOIL AERATION AND GAS EXCHANGE

Soil aeration is a critical parameter controlling plant and microbial growth. Accumulation, exchange and consumption of gases all take place in soils. While storage of gases takes place predominantly in large pores, exchange processes are related to gas diffusivity, a property dependent on pore shape, size and connectivity. Accumulation or depletion of a specific gas is controlled by concentration gradients and the presence of sinks and sources of microbial or plant origin. For gases with low solubility in water (e.g., O_2, CO_2), exchange processes take place mainly through the air-filled pores.

As densification of the soil matrix will affect both total porosity and pore-size distribution, poorly structured and compacted soils will experience an important decrease in air-filled porosity, associated with the loss of total porosity and an increase in bulk density. The use of air-filled porosity in characterizing aeration processes is in part related to the ease of measurement and interpretation of the data, but also due to the fact that gas exchange for most soil gases, largely occurs through air-filled pores. As a consequence, air-filled porosity is often used to characterize the potential productivity of the upper soil layer.

Soil air

The composition of the soil air is influenced by the relative ratio of gases (e.g., N_2, O_2, CO_2) in the atmosphere, and by sources and sinks within the soil. Near the soil surface, the soil readily exchanges gases with the atmosphere. However, roots and microorganisms uptake O_2 and produce CO_2 through respiration. As a result, soil air is typically lower in O_2 and higher in CO_2. If the gas exchange is restricted (e.g., by wet conditions), anoxic conditions which inhibit plant growth may occur.

Gas diffusion

The main mechanism by which gases move in the soil is through **diffusion**. Under steady state conditions, a specific gas molecule in the soil air (e.g., O_2 or CO_2) will move according to its concentration gradient following Fick's first law of diffusion.

Fick's first law:

(23) $J = -D(\frac{dC}{dx})$

where, J = diffusion flux (g m^{-2} s^{-1}); D = soil gas diffusivity (m^2 s^{-1}); and $\frac{dC}{dx}$ = gas concentration gradient (g m^{-2})

For example, the diffusive flux of O_2 will be driven by the concentration gradient ($\frac{dC}{dx}$), with the oxygen diffusing from a zone of higher partial pressure (or concentration) in the atmosphere above the soil to a zone of lower partial pressure (concentration) within the soil, as a result of root and microbial respiration (Figure 4.16). For a given concentration gradient, the rate of diffusion is determined the diffusivity (D). The diffusivity of oxygen in air is about 10,000 times larger than the diffusivity in water. Thus the oxygen diffusivity of a soil is strongly influenced by pore size, air-filled porosity and the connectivity of air-filled pores.

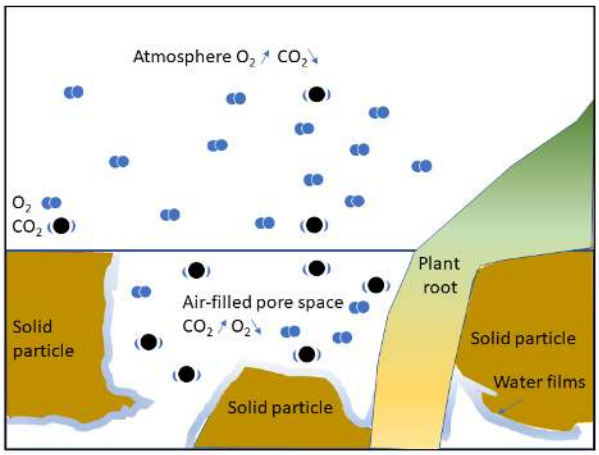


Figure 4.16. Diffusion of O_2 and CO_2 driven by concentration gradients of individual gasses near the soil surface. Microbial and root respiration will also lead to a change in local gas concentration as a result of consumption (of O_2) and production (of CO_2). © Jean Caron is licensed under a CC BY (Attribution) license.

The concentration of a gas (e.g., O_2 or CO_2) at depth within the soil depends on both diffusion and biological activity, since organisms both respire and consume gases in the soil. The change of concentration over time associated with oxygen uptake within the soil profile due to root or microbial activity is illustrated in Figures 4.16 and 4.17. The decrease in soil O_2 level at depth is associated with both higher biological activity (i.e., O_2 consumption) and lower diffusivity (i.e., restricted O_2 exchange with the atmosphere). Both biological activity and diffusivity may be affected by land management practices such as irrigation, the addition of organic mater, or compaction due to heavy machinery. As the soil water content increases, gas diffusivity will decrease and therefore the percentage of oxygen will be lower at depth (i.e., moving from blue to green lines in Figure 4.17). Adding compost will increase biological activity, thereby increase O_2 consumption and result in a lower percentage of oxygen at depth (i.e., moving from dashed to solids lines in Figure 4.17). Soil compaction will reduce air-filled porosity, decrease D and result in a decrease in the percentage of oxygen as a function of depth. Gases, such as methane (CH_4), nitrous oxide (N_2O), and ammonia (NH_3), are also exchanged within the soil-plant-atmosphere continuum through similar diffusion and biological transformation mechanisms, and these gases play an important role in climate change.

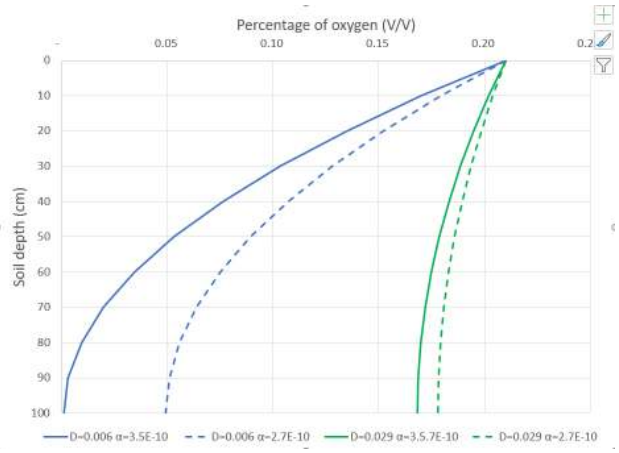


Figure 4.17. Decrease in soil O_2 levels with depth associated with biological activity (α) and gas diffusivity (D). © Jean Caron is licensed under a CC BY (Attribution) license.

SOIL TEMPERATURE AND HEAT FLOW

Soil temperature is an important property that is essential for many soil processes and reactions that include, but are not limited to, microbial activities, nutrient cycling, root growth and greenhouse gas emissions. Soil temperature also affects seed germination, and therefore is an important consideration for determining seeding date in cold regions such as the Canadian Prairies where winter temperatures drop well below 0°C. For example, the ideal temperature for most crops is around 10°C for seed germination. However, pulses, some peas, lentils and chickpeas, can germinate at around 5°C. Because earlier seeding means a longer growing season, earlier seeding usually means higher yield. Conversely, later seeding can result in later harvesting and greater risk of frost and cold temperatures in the fall, which could greatly reduce crop yield. Therefore, soil temperature is an essential soil property for understanding processes in soil and soil management.

Soil thermal properties are important in determining how quickly a soil heats up in spring, how heat is lost from the soil in winter, and how rapidly the temperature in surface layers change. The heat capacity and thermal conductivity of soils are two important properties, which vary depending on the mineral versus organic content of soils, and with soil water content.

Soil heat capacity: the ability of a substance to hold or store heat. The greater its heat capacity, the more heat it can gain (or lose) per unit rise (or fall) in temperature. Specific heat capacity is the amount of heat (energy) needed to cause a 1°C change in temperature of a unit volume of soil (i.e., the higher the heat capacity, the more energy is needed to cause a small change in temperature). Water has a high heat capacity, while the heat capacity of air is negligible. In soil, as the water content increases, the increase in soil heat capacity is linear (Figure 4.18b). Consequently, a dry soil warms up faster than a wet soil, because the energy required to raise the temperature of a dry soil is lower than that needed for a wet soil. The heat capacity of solid organic matter exceeds that of solid mineral particles, but organic layers are much more porous than mineral soil layers. Thus, the heat storage capacity is lower for organic than mineral soils. Soils with a high heat capacity are resistant to temperature change.

Thermal conductivity: the ability of a soil to conduct heat. Thermal conductivity (λ) describes heat flow in response to a temperature gradient. The thermal conductivity of mineral particles in a soil is about one order of magnitude greater than organic matter; and the thermal conductivity of water is an order of magnitude greater than air (Table 4.3). Organic layers, such as the forest floor, tend to have lower λ than for mineral soils, in large part due to their high porosity and greater air filled porosity. In a mineral soil, thermal conductivity is influenced by soil water content. Since water is a good conductor of heat relative to air, as the water content increases and water films bridge between adjacent solid particles, heat conduction increases quickly (Figure 4.18a). Thus, wet soils conduct heat better than dry soils, while dry soils are good insulators.

Table 4.3. Thermal conductivity (λ) of soil components (modified from de Vries, 1963)

Soil component	λ (J/msK⁰)
Quartz	8.368
Clay minerals	2.930
Water	0.594
Organic matter	0.251
Air	0.026

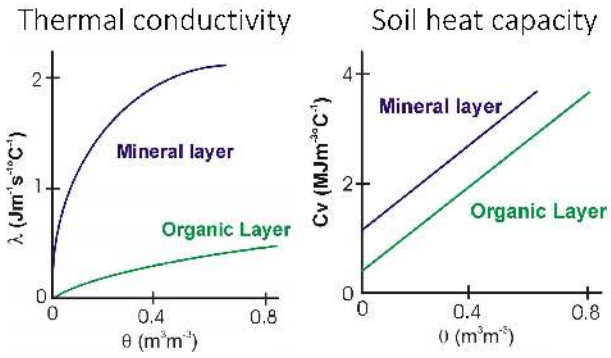


Figure 4.18. Thermal conductivity (a) and soil heat capacity (b) in soil mineral and organic layers. Source: © Krzic et al. 2013. SoilWeb200 is licensed under a CC BY-NC-SA (Attribution NonCommercial ShareAlike) license.

How is soil temperature affected by soil thermal conductivity and heat capacity?

Temperature is a measure of the kinetic energy of individual molecules. It is important for germination and plant growth as well as for activity of soil organisms, and how soil temperature varies over time depends on both climatic factors (e.g., radiation inputs) and soil thermal properties. To give an example, we can compare the surface temperature regimes of two mineral soils with identical texture, SOM, porosity and bulk density, but different soil water content (wet versus dry soil). Initially both soils have the same temperature. On a hot summer day, solar radiation reaching the soil surface is converted to heat energy, which tends to raise the soil surface temperature. The wet soil has a higher thermal conductivity, so it quickly conducts heat downward, away from the surface. Moreover, its high heat capacity means that the surface material would require a large increase in heat content to undergo a temperature increase of 1°C. In contrast, the dry soil cannot conduct heat downward very readily (because of low *thermal conductivity)*, so the surface heat content builds up, and even a small rise in heat content would cause a large rise in temperature (because of the low heat capacity). Thus, the dry soil surface is subjected to extreme temperature fluctuation, and the wet surface is not.

SOIL CONSISTENCY AND STRENGTH

Soil consistence refers to the resistance of soil material to deformation, and the tendency of the soil mass to cohere (stick together) and adhere (stick to other materials). Soil consistence is strongly affected by the clay content and the water content (Figure 4.19). Imagine that we start with dry powdery clay and begin adding water, while we work the sample between our fingers. The shrinkage limit indicates the soil water content where a change in volume occurs (specifically where a loss in soil water no longer results in a reduction in soil volume). At moderately low water content, the soil crumbles, although some pieces stick together. The soil in this condition is said to be **"friable"**. As we add water, the water content reaches a critical limit or boundary, which is called the **plastic limit.** The soil suddenly changes in behaviour; it no longer crumbles, but becomes "*plastic*,**"** that is, we can deform it and mold it under the stresses applied by our fingers. As we keep adding water to the soil, it remains plastic until another threshold is reached. Suddenly, it is no longer moldable into another shape which it will tend to retain. Instead, the sample has a tendency to flow when stress is suddenly applied. The critical water content at this boundary between plastic and liquid behaviour is called the **liquid limit.**

These threshold water contents (shrinkage limit, plastic limit and liquid limits) are called **Atterberg limits** (Figure 4.19). Unlike soil scientists who tend to want to document the resistance of a soil to deformation to preserve the state of a soil for growing trees or crops, engineers want to document how well a soil can resist a stress for building purposes.

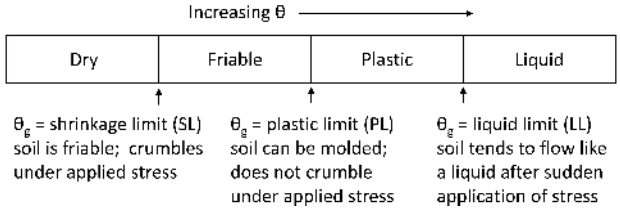


Figure 4.19. Atterberg's limits. Source: Krzic et al. (2013) SoilWeb200 CC BY-NC-SA. © Krzic et al. 2013. SoilWeb200 is licensed under a CC BY-NC-SA (Attribution NonCommercial ShareAlike) license.

SUMMARY

- The soil is a three-phase system: solid, liquid and gas. The magnitude and interaction between the phases determine the behavior and functionality of soil.
- Soil texture (i.e., % sand, silt and clay) influences many other soil properties, including water retention, aeration, drainage, and susceptibility to compaction.
- Adhesion and cohesion forces are responsible for the retention of water in soils.
- Field capacity, permanent wilting point and available water storage capacity (or plant available water) are terms used to describe the water content of a soil for management purposes.
- Water movement in soil is driven by differences in soil water potential.
- Hydraulic conductivity represents the soil's ability to facilitate water flow, and is related to the size and connectivity of water-filled pores.
- Solutes in soils are transported by diffusion and mass flow.
- The main mechanism of gas movement in the soil is through diffusion.
- Soil thermal conductivity and heat storage capacity determine how rapidly a soil heats up (or cools down).

STUDY QUESTIONS

1. Define each of the following terms:
    - specific surface area
    - fine earth fraction
    - soil water potential
2. Define each of the following terms and identify important distinctions between paired terms:
    - soil texture and soil structure
    - particle density and bulk density
    - field capacity and permanent wilting point
3. Draw the soil water characteristic curves for a sandy soil and a loam (include axis labels).
4. Define soil water potential, and explain how the direction of water flows in response to a gradient in total water potential.
5. List and define the individual forces, which make up total soil water potential.
6. What is the relationship between matric potential and the size of soil pores?
7. How might the soil water characteristic curve be used to obtain information about the soil pore-size distribution?
8. Briefly describe differences in solute transport by mass flow and by diffusion.
9. How do water-filled pores affect the diffusion of oxygen in soils relative to air-filled pores?
10. How does compaction by heavy machinery influence oxygen distribution in soils?
11. In temperature regions (like Vancouver B.C.) where winters are wet and mild, soils may take a long time to warm up in spring. Why do wet soils take a long time to warm up?

WORKED EXAMPLES

Example #1 – bulk density & gravimetric water content

Zineb, a student at the University of British Columbia, collected a soil core from the UBC Farm during summer. The sample is weighed, oven dried at 105ºC for 24 hours (or constant weight) and re-weighed.



Photo by Zineb Bazza, University of British Columbia, Vancouver @ CC BY

The following data were obtained:	
Volume of the core:	330 cm^3
Weight of the metal core:	56.0 g
Fresh weight of the core + soil	499.4 g
Oven dry weight of the soil:	386.1 g

Calculate the bulk density and gravimetric water content of this soil sample (show formulas used and units).

Step 1: Calculation of bulk density

We know $\rho_b = \frac{M_s}{V_t}$

where M_s = mass of solids and V_t = total soil volume

Thus, $\rho_b = \frac{M_s}{V_t} = \frac{386.1\,g}{330\,cm^3} = 1.17\,g\,cm^{-3}$

Step 2: Calculation of gravimetric water content

We know that $\theta_w = \frac{M_l}{M_s} = \frac{M_{fresh} - M_{oven\ dry}}{M_s}$

where

$M_{fresh} = (fresh\ mass\ of\ soil + core) - mass\ of\ core = 499.4\,g - 56.0\,g = 443.4\,g$

Thus, $\theta_w = \frac{443.4\,g - 386.1\,g}{386.1\,g} = \frac{57.3\,g}{386.1\,g} = 0.15\,g\,g^{-1}$

Example #2 – gravimetric & volumetric water content

A soil sample collected from the Chaudière-Appalaches region of Quebec has a gravimetric water content (θ_w) of 0.19 g g^{-1} and a bulk density (ρ_b) of 1.26 g cm^{-3}. What is the volumetric water content (θ_v) of this sample in cm^3 cm^{-3}? Show all formulas, units and full calculation.

We know that $\theta_v = \frac{V_l}{V_t}$ and $\theta_w = \frac{M_l}{M_s} = \frac{M_t - M_s}{M_s}$

Now, $\rho_l = \frac{M_l}{V_l}$ or $V_l = \frac{M_l}{\rho_l}$ and $\rho_b = \frac{M_s}{V_t}$ or $V_t = \frac{M_s}{\rho_b}$

Thus, $\theta_v = \frac{V_l}{V_t} = \frac{\frac{M_l}{\rho_l}}{\frac{M_s}{\rho_b}} = \frac{M_l \cdot \rho_b}{M_s \cdot \rho_l} = \theta_w \frac{\rho_b}{\rho_l}$

Now, if we consider the density of liquid (or water in this case) = 1 g cm^{-3} (in CGS unit), then we can simply write $\theta_v = \theta_w \rho_b$

For the soil sample from the Yukon (showing full units):

$$\theta_v = \theta_w \rho_b = 0.19 g g^{-1} \times \frac{1.26 \frac{g}{cm^3}}{1 \frac{g}{cm^3}} = 0.24 \frac{cm^3}{cm^3}$$

Example #3 – bulk density & porosity

A loam soil in Southern Alberta has a porosity of 48%. What is bulk density of this soil in kg m^{-3}? Show all formulas you use and the full calculation including units.

Assuming a particle density of $2.65\frac{kg}{m^3}$ and given a porosity of 48% or $0.48\frac{cm^3}{cm^3}$, the bulk density is $1,378\frac{kg}{m^3}$

This calculation is done by rearranging and solving the following equation:

Porosity $= 1 - \frac{\rho_b}{\rho_s}$ where ρ_b is soil bulk density and ρ_s is soil particle density.

Therefore: $\rho_b = 2.65\frac{g}{cm^3} - (2.65\frac{g}{cm^3} \times 0.48) = 1.378\frac{g}{cm^3} = 1,378\frac{kg}{m^3}$

Example #4 – Soil water movement

Miles, who teaches introductory soil science at the University of Alberta, took measurements of soil water tension (T) at 3 depths, before and after a large rain event. Unfortunately, some of the measurements were lost.

A) Fill in the two tables below using the following information:

- the soil solution contains no salts and the soil is well vented to atmosphere
- all three points (or depths) of observation are above the water table (or free water surface)
- the reference point is set at the soil surface

[*hint: remember that* $T = -\text{matric potential } (h_m)$]

Table I

Point	Depth in soil (m)	T (m)	h_g (m)	h_m (m)	h_o (m)	h_p (m)	h_t (m)	Direction of flow (↓↑)
A	0	2.3						
B	-0.1	2.45						
C	-0.25	2.9						

Table II

Point	Depth in soil (m)	T (m)	h_g (m)	h_m (m)	h_o (m)	h_p (m)	h_t (m)	Direction of flow (↓↑)
A	0	5.2						
B	-0.1	4.4						
C	-0.25	4.65						

B) Which table belongs to the measurements taken before the rain event, and which after? Briefly explain.

Solution:

- determinie h_g : since the reference elevation is at the surface, and the units are m of water, h_g is the same as the depth, i.e., increasingly negative as we move deeper in the soil profile and farther below the reference elevation
- recall that the matric potential (h_m) = $-T$

- as the soil solution contains no salts, the osmotic potential $h_o = 0$
- as the soil is well vented, and the soil profile is above the water table, $h_p = 0$
- recall $h_t = \Sigma(h_g + h_m + h_o + h_p)$
- recall that soil water moves from high to low potential, thus you can determine the direction of flow

Table I

Point	Depth in soil (m)	T (m)	h_g (m)	h_m (m)	h_o (m)	h_p (m)	h_t (m)	Direction of flow (↓↑)
A	0	2.3	0	-2.3	0	0	-2.3	
								↓
B	-0.1	2.45	-0.1	-2.55	0	0	-2.55	
								↓
C	-0.25	2.9	-0.25	-3.15	0	0	-3.15	

Table II

Point	Depth in soil (m)	T (m)	h_g (m)	h_m (m)	h_o (m)	h_p (m)	h_t (m)	Direction of flow (↓ ↑)
A	0	2.3	0	-5.2	0	0	-5.2	
								↑
B	-0.1	2.45	-0.1	-4.4	0	0	-4.5	
								↓
C	-0.25	2.9	-0.25	-4.65	0	0	-4.9	

B) Table I shows the data after the rain, since the direction of water flow in the surface part of the profile (i.e., from point A to B) is downward. Downward water movement implies that water is entering the soil.

Table II shows data before the rain. In Table II, the water is moving upward, from point B to point A, which is an indication that that part of soil profile is drier (probably due to combination of evapotranspiration and water uptake in that part of soil by roots and soil organisms).

Example #5 – Solute transport / nitrate leaching in soil

Fernanda, a student at the University of Saskatchewan conducted a column experiment. The diameter of the soil column was 20 cm. She applied a constant rate of NO_3^- solution on the top of soil column. When the amount of solution collected from the bottom of the column reached steady state (i.e., the flux was constant with time), 30 mL of solution was collected in 30 minutes.

A) Calculate the soil water flux

The water flux is $q_w = \frac{V}{At}$ where A is the cross-sectional area of the column (πr^2); thus,

$$q_w = \frac{30\,cm^3}{3.14 \cdot (10\,cm)^2 \cdot 30\,min} = 0.00318\,cm\,min^{-1}$$

Fernanda also took a sample of the collected solution and measured the NO_3^-concentration using ion Chromatography (IC). The measured concentration of NO_3^- was 10 µg L^{-1}.

B) Calculate the NO_3^- flux

The NO_3^- flux can be calculated using Equation # 22:

$$q_s = q_w C = 0.00318\,cm\,min^{-1} \cdot 10\,\mu g\,L^{-1} \cdot 0.001\,L\,cm^{-3}$$

$$q_s = 3.18 \cdot 10^{-5}\,\mu g\,cm^{-2}\,min^{-1}$$

Alternatively, we can calculate the flux based on the definition of chemical flux:

$$mass\,of\,chemicals = V \cdot C = 30\,cm^3 \cdot 10\,\mu g\,L^{-1} \cdot 0.001\,L\,cm^{-3} = 0.3\,\mu g$$

thus,

$$q_s = \frac{m}{At} = \frac{0.3\,\mu g}{3.14 \cdot (10\,cm)^2 \cdot 30\,min} = 3.18 \cdot 10^{-5}\,\mu g\,cm^{-2}\,min^{-1}$$

REFERENCES

Bittman, S., Sheppard, C., and Hunt, D. 2017. Potential for mitigating atmospheric ammonia in Canada. Soil Use and Management 33(2): 263-275. doi:10.1111/sum/12336

Caron, J., Pelletier, V., Kennedy, C. D., Gallichand, J., Gumiere, S., Bonin, S., Bland, B., and Pepin, S. 2017. Guidelines of irrigation and drainage management strategies to enhance cranberry production and optimize water use in North America. Canadian Journal of Soil Science 97(1): 82-91.

Caron, J., Price, J. S., and Rochefort, L. 2015. Physical properties of organic soil: adapting mineral soil concepts to horticultural growing media and Histosol characterization. Vadose Zone Journal 14:1-14.

De Vries, D.A. 1963. Thermal properties of soils. *In* W.R. van Wijk (ed.) Physics of Plant Environment, John Wiley & Sons, New York. pp. 210-235.

Gregorich, E.G., Turchenek, L.W., Carter M.R., and Angers D.A. 2001. Soil and environmental science dictionary. CRC Press, Boca Raton.

Krzic, M., Wiseman, K., Dampier, L., Grand, S., Wilson, J. and Gaumont-Guay, D. 2013. SoilWeb200: An Online Educational Tool for the APBI 200 course: Introduction to Soil Science. The University of British Columbia, Vancouver [http://soilweb200.landfood.ubc.ca]

Narasimham, T. N. 2007. Central ideas of Buckingham (1907): A century later. Vadose Zone Journal 6(4):687-693. doi:10.2136/vzj2007.0080

Reicosky, D. C., and Ritchie, J. T. 1976. Relative importance of soil resistance and plant resistance in root water absorption. Soil Science Society of America Journal 40(2):293-297.

Rekika, D., Caron, J., Rancourt, G. T., Lafond, J. A., Gumiere, S. J., Jenni, S., and Gosselin, A. 2014. Optimal irrigation for onion and celery production and spinach seed germination in Histosols. Agronomy Journal 106(3):981-994.

Ritchie, H., and Roser M. 2013. Crop Yields. Published online at OurWorldInData.org retrieved from https://ourwolrdindata.org/crop-yields.

Tisdale, J.M. and Oades, J.M. 1982. Organic matter and water-stable aggregates in soils. European Journal of Soil Science 32(2):141-163. doi.org/10.1111/j.1365-2389.1982.tb01755.x

USD-NRCS. 1991. Soil-Water-Plant Relationships. Section 15, Chapter 1, National Engineering Handbook, 2nd edition. https://directives.sc.egov.usda.gov/...tent=18350.wba

About the Authors

Sandra Brown, Assistant Professor Teaching, Faculty of Land and Food Systems, University of British Columbia, Vancouver, BC



Sandra Brown (licensed under a CC-BY-NC-ND license)

Shortly after the 1980 eruption of Mount Saint Helens, I was in the red zone – grey, bleak, ominous, yet beautiful. One year later, that landscape had already begun to transform. I was hooked, I wanted to learn more about this transformation from sediment to soil. Andisols (soils formed in volcanic ash) are my favorite, in part because of their ability to retain water, in part because they are great for growing Colombian coffee!

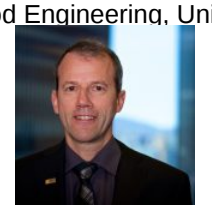
Asim Biswas, Associate Professor, School of Environmental Science, University of Guelph, Guelph, ON



Asim Biswas (licensed under a CC-BY-NC-ND license)

Growing up on a small subsistence farm in a remote village in India, I recognized soil as the most valuable resource any farm family can have. As I started to play with soils and grow crops with elders, one question always fascinated my young mind, ‘what is there in soils that a small seed can become a big plant or tree?’ My curiosity about soils lead me to study soil science at university. My current research focuses on increasing the productivity and resilience of our food production in an environmentally sustainable way.

Jean Caron, Professor, Soil Science and Agri-Food Engineering, Université Laval, Quebec City, QC



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At the age of 12, I read an article in Reader’s Digest about Norman Borlaug, a famous agronomist. I made my decision then to become an agronomist and that has never changed. I became interested in the power of math, physics and technology, which lead me to soil physics. As a Professor, I work on water movement in soils, drainage and irrigation management, and soil conservation. Being in the field investigating and solving problems with growers and agricultural engineers is, for me, the most exciting part of my job; having the opportunity to work with passionate farmers and professionals.

Miles Dyck, Associate Professor, Faculty of Agricultural, Life and Environmental Sciences, University of Alberta, Edmonton, AB



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I teach courses in soil physics and soil formation/classification. My research focuses on how soil physical processes and properties are affected by human activities. Growing up on a farm in southwestern Saskatchewan, I became curious about how the soils landscapes around my home had developed. In university I learned about soils of Canada and I was hooked! Soils are important for food production, and water and air quality. I am grateful to have the opportunity to make a living by learning about soil.

Bing Si, Professor, College of Agriculture, University of Saskatchewan, Saskatoon, SK



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When I was a teenager, I enjoyed helping out my mom on the farm. But after a long day of harrowing soil using a hoe, I felt the soil was still as hard as a rock. The soil on our farm was clayey and low in organic matter, and when dry, it formed a hard crust. With the question on how to improve the soil, I registered as an agriculture major in university. My current research tries to answer: Why sandy soil can sustain a more water-demanding forest, but not a crop? What is the contribution of deep soil (beyond the 1 m depth) to plant growth?

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1.5: Soil Chemistry

Darshani Kumaragamage, Jim Warren, and Graeme Spiers

Learning Objectives

On completion of this chapter, students will be able to:

1. Define and describe the importance of the major soil chemical properties in sustaining life on Earth
2. Describe and compare the structures, charge development, and properties of organic and inorganic soil colloids
3. Explain the influence of soil minerals on soil properties
4. Define cation exchange capacity, explain the importance of cation exchange and anion adsorption, and identify the factors influencing ion exchange in soils
5. Discuss the causes of soil acidity and alkalinity, characteristics of strongly acidic and alkaline soils, and their effect on plant growth
6. Describe the causes of soil salinity, major characteristics of saline, saline-sodic and sodic soils and reclamation of salt-affected soils
7. Describe the importance of redox reactions in nutrient dynamics and greenhouse gas emissions
8. Explain the basic principles in measuring cation exchange capacity, pH, electrical conductivity and redox potential in soils, and the interpretation of their values

INTRODUCTION

Soil chemistry is a branch of soil science that deals with the chemical composition, chemical reactions and chemical properties in soils. Abiotic phases of the soil include solids (organic matter and inorganic minerals), liquids (soil water), and gases (soil air), while the biotic phase consists of living organisms. Soil, therefore, is a dynamic living system in which ions and molecules constantly may move from one phase to the other, while interacting with each other.

Soil chemical properties are based on the concentrations, and/or proportions of dissolved species in soil water and/or on the ion exchange complex. Soil chemical properties such as cation exchange capacity (CEC), pH, redox potential (Eh or pe), and electrical conductivity (EC) are important as they influence nutrient availability, plant growth, fate of pollutants, biological activity, etc. For example, CEC is a measure of the amount of negative charge capable of adsorbing cations on mineral or organic surfaces, usually expressed in centimoles of charge [cmol(+)] per kg of soil. Soil pH is a measurement of the concentration of H^+ in the soil solution, while the Eh (or pe) is related to concentrations of oxidized/reduced species of redox-sensitive elements such as iron (Fe^{3+}/Fe^{2+}) . Electrical conductivity (EC) is another important soil property that provides an indication of the abundance of soluble salts in a soil.

Most of the important chemical properties of soils are controlled by reactions that occur between the soil solution and surfaces of colloidal (particles <0.002 mm diameter; Brady and Weil, 2010) soil mineral particles and soil organic matter. In the first part of this chapter we will focus on the chemistry of the soil solution and the colloidal fraction. The remaining sections of the chapter will focus on important soil chemical properties; namely, CEC, anion exchange, soil pH, EC, and Eh, with emphasis on their importance in sustaining life on Earth.

SOIL Solution

Soil water, including dissolved solutes, is referred to as the soil solution and is the lifeblood of soil. All chemical reactions, mineral precipitation/dissolution reactions, ion exchange reactions, redox reactions, and nutrient uptake by plants occur in, or are mediated by the soil solution. Without soil water, very few chemical or biological reactions would occur in soil; in fact, there would be no life on this planet.

The soil solution is very complex and contains a large variety of cations and anions (as both as free ions and complex ions) as well as dissolved organic molecules, usually in low concentrations. For soil nutrients to be taken up by plant roots, they must exist in a dissolved form in the soil solution. At least 17 elements are required to support the growth of most plants (see Chapter 7), with an additional four elements essential for the growth of some plants (Havlin et al., 2013). For example, some important cations and anions that plants absorb from soil solution are: nitrogen as ammonium (NH_4^+) and nitrate (NO_3^-), phosphorous as several

phosphate species (e.g., HPO_4^{2-} and $H_2PO_4^-$), potassium (K^+), calcium (Ca^{2+}), magnesium (Mg^{2+}), and sulphur as sulphate (SO_4^{2-}). All nutrients are absorbed from the soil solution by plant roots.

The soil solution also contains other non-nutritive ions and molecules, some of which are innocuous, while some are toxic, depending on concentration. These include some trace elements (e.g., lead, arsenic), pesticides (e.g., chlorpyrifos, glyphosate), antimicrobial ions and compounds (e.g., disinfectants, antibiotics), and complex organic molecules. These may also be taken up by plant roots. Nutrients and other chemicals in the soil solution may also be lost from the soil through surface runoff and/or downward leaching, particularly under excessive rainfall and irrigation events.

SOIL COLLOIDS

The soil colloidal fraction is defined as that fraction of the soil made up collectively of small (<0.002 mm) inorganic and organic particles (Brady and Weil, 2010). The high specific surface area and the presence of electric charges on most particles are responsible for important soil characteristics such as swelling capacity, the ability of the soil to retain water, cation and anion adsorption and ion exchange.

The importance of colloids in soil relates to the relationship of their size with the amount of surface area available to react with the soil solution. Table 5.1 shows the relationship between particle size and the magnitude of surface area associated with particle size range for a clay loam soil. Specific surface area is defined as the surface area per unit mass of the solid particle and expressed as $m^2 g^{-1}$ of soil. As can be seen in Table 5.1, even though this soil contains equal proportions of sand, silt, and clay by mass, by far the majority of the surface area is associated with the smallest size fraction.

Table 5.1. Relationship between particle size range and specific surface area of soil particles.

Particle Size Fraction	**Size Range**	**Median Diameter**	**Percent of Clay Loam Soil [a]**	**Specific Surface Area [b]**	**Percentage Surface Area**
	(µm)	**(µm)**	**(%)**	**($cm^2 g^{-1}$)**	**(%)**
Sand	2000-50	1000	33.3	22.6	0.2
Silt	50-2	25	33.3	90.9	0.8
Clay	<2	1	33.3	11,320	99

[a] Assumed a clay loam soil with equal proportions of sand, silt and clay. [b] Calculated based on uniform spherical particles with median particle diameter.

Mineral phases in the colloidal or clay-size fraction are primarily phyllosilicate and oxy-hydroxide minerals (Figure 5.1). Organic colloids are also in the clay-size range and are composed of small particles of humus that are relatively stable and resistant to further biological decay. These diverse colloidal particles may either exist independently in the soil environment or they may occur in association in organo-mineral complexes (Newman and Hayes, 1990).

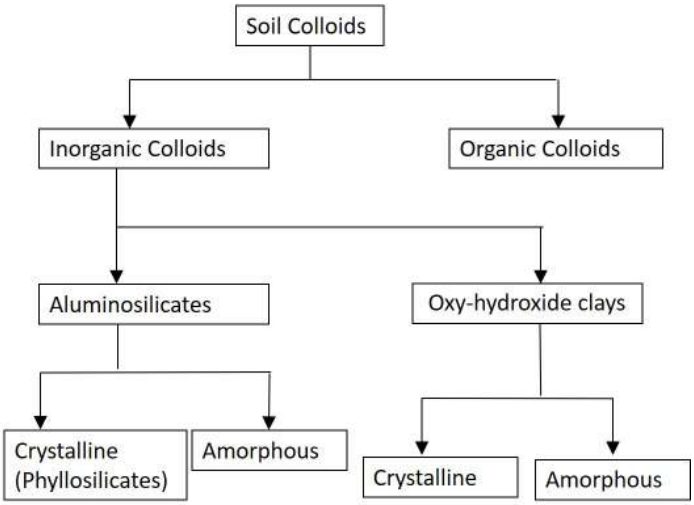


Figure 5.1. A classification of soil colloids. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Phyllosilicates

It is important to understand that there are differences among aluminosilicate clay minerals (also called phyllosilicates or layer silicates) that make up the vast majority of mineral colloids and that these impart substantial differences in important soil properties, such as ion exchange.

Phyllosilicates (*phyllon* meaning leaf or sheet in Greek) are crystalline minerals, with a structure consisting of layers of repeating atomic units. There are two fundamental units composed of layers of oxygen and hydroxyl groups bonded with silicon, aluminum, magnesium and other cations. The fundamental layers are composed of tetrahedral and octahedral sheets that occur in different combinations to form the different clay minerals.

Tetrahedral sheets are composed of repeating tetrahedron units, each typically containing a silicon atom surrounded by four oxygen atoms: one apical oxygen atom and three basal oxygen atoms forming a four-sided (tetrahedral) configuration (Figure 5.2A). Tetrahedral sheets extending in two dimensions are formed by sharing the basal oxygen atoms between silicon atoms (Figure 5.2B).

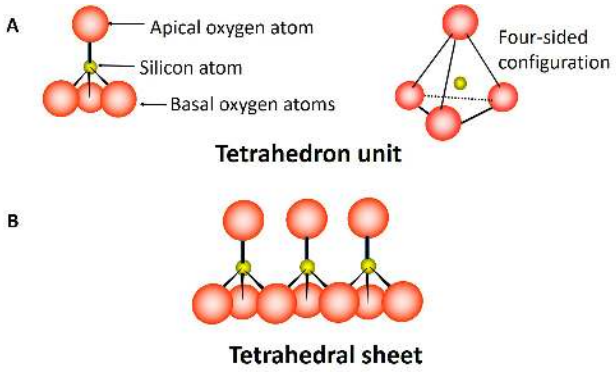


Figure 5.2. Arrangement of atoms in a silica tetrahedron (A) and linked in a tetrahedral sheet (B). © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

The octahedral sheet is composed of multiple octahedron units (Figure 5.3A), each unit containing aluminum (or magnesium) atoms surrounded by six oxygen atoms and/or hydroxyl groups. Three oxygen atoms (or hydroxyl groups) occur in the lower plane, the remaining three oxygen atoms (or hydroxyl groups) occur in the upper plane with the aluminum or magnesium atoms sandwiched in between, forming an eight-sided (octahedral) configuration. These units combine by sharing oxygen atoms to form an octahedral sheet extending in two dimensions (Figure 5.3B).

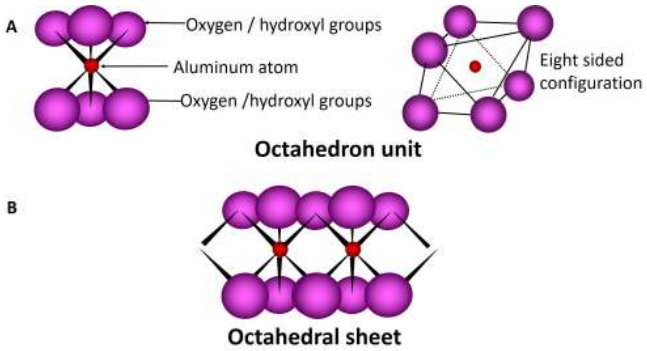


Figure 5.3. Arrangement of atoms in an octahedron (A) and in an octahedral sheet (B). © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Tetrahedral and octahedral sheets combine to form unit layers of different phyllosilicates through the sharing of the apical oxygen atom of the tetrahedral sheet with oxygen atoms in the octahedral sheet. Phyllosilicates composed of one tetrahedral sheet and one octahedral sheet are called 1:1 type phyllosilicates; those composed of one octahedral sheet sandwiched between two tetrahedral sheets are called 2:1 type phyllosilicates (Figure 5.4). The ideal unit cell formula for a 1:1 clay mineral is $Si_4Al_4O_{10}(OH)_8$, while that of a 2:1 clay mineral is $Si_8Al_4O_{20}(OH)_4$.

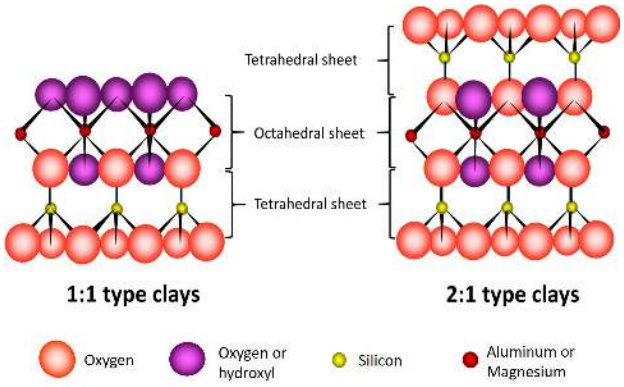


Figure 5.4. Layer arrangement of 1:1 and 2:1 silicate clay minerals. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Phyllosilicates consist of many layers (1:1 type or 2:1 type) stacked one on top of the other with different forms of bonding between the layers. The region located between two adjacent layers is called the interlayer region. The basic chemical behaviour and physical properties of the different phyllosilicates are dictated in large part by the characteristics associated with the interlayer region of the clays.

Isomorphous Substitution

In addition to their large surface area (see Table 5.1), the phyllosilicates are also electrically charged due to substitution of the central cations within the tetrahedral or octahedral sheets by other ions of similar size (see Table 5.2) but different valence. This imparts a net negative charge to the structure of the mineral.

Isomorphous substitution is defined as “replacement of one atom by another of similar size in a crystal structure without disrupting or changing the crystal structure of the mineral” (Canada Department of Agriculture, 1976). When a substituting cation is of a lesser valence than the cation it is replacing, there is an excess of negative charges associated with the oxygen atoms surrounding it and therefore a net negative charge is imparted to the structure. For example, trivalent aluminum (Al^{3+}) may substitute for some of the tetravalent silicon (Si^{4+}) in the tetrahedral sheet. The lesser charge of the aluminum compared to silicon creates a charge deficit and a negative charge within the tetrahedral sheet of the structure. Similarly, in the octahedral sheet, Mg^{2+}, Fe^{2+}, or other divalent cations, may substitute for some trivalent Al (Al^{3+}), resulting in an extra negative charge associated with the octahedral sheet of the structure. These negative charges attract positively charged cations to the surface of the minerals to neutralize the negative charge. These charge neutralizing cations are located mostly within the interlayer region of the phyllosilicates. The abundance of negative charge sites (i.e., the **charge density**), the location of the negative charges (tetrahedral

vs. octahedral sheet), and the type of cation in the interlayer region neutralizing the charge (e.g., Na^{+}, Ca^{2+}, Mg^{2+}, K^{+} or other) determines the degree to which the phyllosilicates expand or swell, and their capacity to retain cations.

Table 5.2. Radius of common cations and location in phyllosilicate structures

Cation	Radius (nm)	Location
Si^{4+}	0.041	tetrahedra
Al^{3+}	0.05	tetrahedra, octahedra, interlayer
Fe^{3+}	0.064	octahedra
Fe^{2+}	0.076	octahedra, interlayer
Mg^{2+}	0.065	octahedra, interlayer
Ca^{2+}	0.099	Interlayer
Na^{+}	0.095	Interlayer
K^{+}	0.133	Interlayer

Phyllosilicate Minerals

There are about 50 species of phyllosilicate minerals, but only five are common in Canadian soils (see Chapter 14). These are sketched in Figure 5.5 and described as follows:

Kaolinite (Figure 5.5A) is a 1:1 type clay mineral held together by hydrogen bonds in the interlayer region. Hydrogen bonding occurs between the hydroxyl in the octahedral sheet of one unit layer, and the (basal) oxygens in the tetrahedral sheet of the adjacent layer. The hydrogen bonds hold the layers closely together to form a very stable structure. This combined with very little isomorphous substitution inhibits the intrusion of cations and water into the interlayer region. This mineral is non-expandable since water molecules cannot enter the interlayer space. Kaolinite exhibits very little swelling when wet, or shrinkage when dry. The interlayer is not exposed and therefore the specific surface area is low. With almost no isomorphous substitution, kaolinite minerals have a low surface charge and low cation exchange capacity.

Can You Dig It!

Rocks that are rich in kaolinite are known as kaolin which is derived from the Chinese word "Gaoling" (literally meaning 'High Ridge'), which is the name of a village in southeastern China where kaolinite was mined historically. Kaolinite in its pure form is white and its minimal shrink/swell characteristic and ability to not crack when dried made it highly prized for making porcelain and "fine China" dishware.

Smectite (Figure 5.5B) is a group of expandable 2:1 type phyllosilicates. Montmorillonite and beidellite are the two most common species of the smectite group of minerals in Canadian soils. The amount of isomorphous substitution in both minerals is moderate compared to other 2:1 type phyllosilicates (e.g. hydrous mica), and the two species are differentiated based on the location of the layer charge in the structure. Most of the layer charge in montmorillonite is a result of isomorphous substitution in the octahedral sheet, whereas in beidellite the layer charge derives mostly from isomorphous substitution in the tetrahedral sheet. Hydrogen bonding does not occur in 2:1 clay minerals since there are no hydroxyl groups associated with the basal oxygens of the tetrahedral sheets of adjacent layers facing each other. Thus, the structural layers are only held together by electrostatic forces of whatever neutralizing cation happens to be in the interlayer region to neutralize the negative charges associated with isomorphous substation within the layer structures. Exchangeable cations in the interlayer are usually surrounded by six water molecules in octahedral coordination. Smectite minerals have a high cation exchange capacity. Since the layers are loosely bound, both water molecules and cations can freely enter the interlayer space. Therefore, these clays have a high specific area, and are expandable; swelling when wet and shrinking when dry.

Vermicullite (Figure 5.5B) is a 2:1 type phyllosilicate similar to smectite but with a layer charge intermediate between hydrous mica/illite (see below) and the smectites. The mineral has limited expandability, with moderate shrink/swell characteristics. Most isomorphous substitution occurs in the tetrahedral sheet imparting a high surface charge and a high cation exchange capacity. The

structure is capable of trapping K^+ ions in the interlayer region similar to hydrous mica/illite, but as the layer charge is slightly lower, the trapped K^+ can be released back into the soil solution unlike hydrous mica/illite.

Hydrous Mica also called **illite** (Figure 5.5C) is another 2:1 type clay mineral, Hydrous mica is derived from the partial weathering of mica, a primary mineral common in the granitic rocks of the Canadian Shield. The structure of hydrous mica/illite is similar to the smectites and vermicullite, except that the layer charge is higher with isomorphous substitution occurring almost exclusively in the tetrahedral sheet where Al^{3+} is substituted for Si^{4+}. Potassium almost exclusively occupies the interlayer region, neutralizing the layer charge and holding the structure together. The potassium ion (K^+) is almost an exact fit for the hexagonal shaped holes that occur in the tetrahedral sheets of the 2:1 layers, and thus gets even closer to the charged sites and strongly binds adjacent layers together. This prevents entry of water into the interlayer region and renders minerals of the mica group non-expandable. So even though the layer charge is higher than the smectites and vermiculite, the cation exchange capacity of hydrous mica/illite is low, since the charge is neutralized by K^+.

Chlorite (Figure 5.5D) is a 2:1:1 (or 2:2) type phyllosilicate. The mineral has a 2:1 type structure, but with an additional octahedral sheet—composed of an aluminum, magnesium or iron hydroxide layer in six-fold (octahedral) coordination—in the interlayer region between the 2:1 type layers. This interlayer hydroxyl layer does not share oxygens with the main 2:1 layer, instead it is hydrogen bonded to the 2:1 layer; consequently, the mineral is non-expandable. The amount of isomorphous substitution in the structure is variable, but the charges are neutralized by the positively charged octahedral sheet in the interlayer, thus the ion exchange capacity is low.

Amorphous aluminosilicates are yet other aluminosilicates found in soil but these are not phyllosilicates. These are described as amorphous or as minerals of short-range order, meaning they lack the defined structural units of the phyllosilicates. Allophane and imogolite are examples of amorphous aluminosilicate minerals. These minerals are formed in abundance in volcanic ash soils found in countries such as Japan and New Zealand, but they are also being found in very small amounts in some Podzolic and Solonetzic soils in Canada. Amorphous aluminosilicates are non-expandable. They are extremely small (<4 nm diameter) with a very high specific surface area and a very high ion exchange capacity.

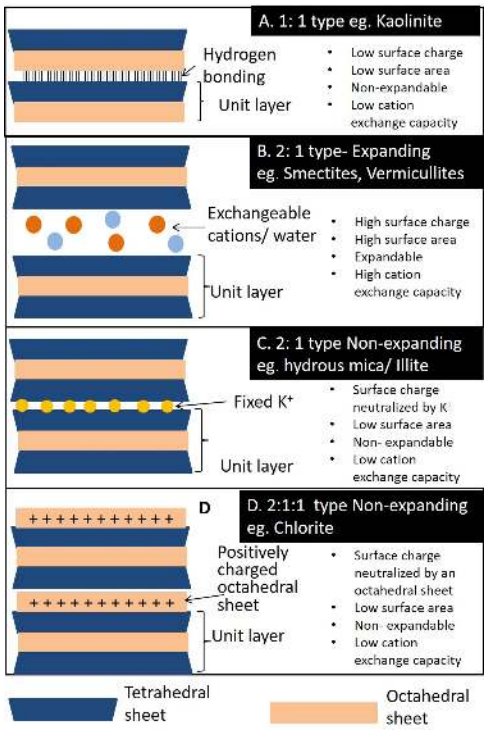


Figure 5.5. Arrangement of tetrahedral sheets and octahedral sheets in common phyllosilicates and some key properties. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Oxy-hydroxide Minerals

Oxy-hydroxide minerals (sesquioxides) are mainly oxides, hydroxides and hydrous oxides of iron, aluminum, and manganese, and are among the youngest mineral phases found in soil. Some are well crystallized (e.g., gibbsite, hematite, goethite), while others are amorphous (e.g., ferrihydrite). Crystalline oxide clays are composed of oxygen and/or hydroxyl groups, usually in an octahedral arrangement with iron, aluminum, or manganese. The octahedral structures of these minerals are held together by H-bonding. Non-crystalline (amorphous) oxy-hydroxides are complex iron, aluminum, and manganese minerals that are typically precipitated very quickly from the soil solution in response to wetting and drying cycles. Precipitation of these minerals occurs when the solubility product for a combination of cation and anion concentrations is exceeded, and usually other solutes including carbonate, phosphate and sulphate may be trapped in the structure during the rapid precipitation. Amorphous oxy-hydroxides are metastable and convert slowly to their more crystalline forms.

Both crystalline or amorphous oxy-hydroxide minerals do not carry permanent negative charges but do have a high surface charge associated with surface hydroxyl groups. As such, the exchange capacity of oxy-hydroxides is very pH dependent with the variation due to protonation of hydroxyl groups at low pH producing positive charges, and dissociation of hydroxyl groups at high pH producing negative charges. Oxy-hydroxide minerals are **amphoteric** due to their variable cation exchange capacity; meaning they typically have a net cation exchange capacity at high pH and a net anion exchange capacity at low pH.

Oxy-hydroxides are usually found in small quantities in young Canadian soils, but are more common in highly weathered soils found in warm, humid regions in other parts of the world. They are highly reactive and have a very high specific surface area. Presence of these minerals significantly influences soil properties, such as ion adsorption and exchange.

Organic Colloids (Humus)

Organic colloids, generally known as humus, consist of highly decomposed organic matter. These colloids have a complex structure, and are chemically composed mainly of carbon, hydrogen and oxygen. On the molecular scale, humus may be described as a complex polymeric mixture of random organic components and is probably one of the least understood aspects of soil chemistry. Organic colloids may be highly reactive chemically, with high water holding capacities allowing absorption of water. Compared to inorganic colloids, organic colloids have greater specific surface area and higher surface charge. Surface charges are predominantly negative and pH dependent arising from partial dissociation of hydroxyl (-OH), carboxyl (-COOH), and phenolic groups, which are associated with central carbon structures of varying size and complexity. On a per unit mass basis, organic colloids impart a greater influence on soil properties than inorganic colloids even though their abundance in soil is generally far less than that of the inorganic colloid fraction.

Charges on Soil Colloids

Charges on soil colloids arise from the following two mechanisms:

1. **Isomorphous substitution** within the crystalline structure of phyllosilicates – leading to formation of mostly negative, permanent charges. As described earlier in this chapter, isomorphous substitution is the replacement of one atom by another of similar size in a crystal lattice without disrupting or changing the crystal structure of the mineral. The replacement of a higher valent cation with a lower valent cation (e.g., substitution of Mg^{2+} for Al^{3+} in the octahedral sheet) results in an unbalanced negative charge on the mineral surfaces. These negative charges are permanent and thus are not affected by pH.
2. **Protonation/Deprotonation** of functional groups on the surfaces of oxy-hydroxide clays, the edges of phyllosilicate clay minerals and organic compounds, give rise to positive and negative charges, referred to as pH dependent (variable) charge. Protonation of functional groups occurs under acidic (low pH) conditions whereas dissociation of hydroxyl protons predominates under basic (high pH) conditions (Figure 5.6). Thus, the relative amounts of positive and negative surface charge vary depending on soil pH, with negative charges predominating under high pH conditions and positive charges predominating at low pH conditions. The pH at which the number of positive charges is equal to the number of negative charges is referred to as the point of net zero proton charge or zero point of charge. Humus particle surfaces are similar to inorganic colloids, containing functional groups (e.g., carboxyl [-COOH], phenolic [R-OH]) that can dissociate (deprotonate) depending on the pH, giving rise to a negative charge on organic colloids. As the pH increases, dissociation is greater, and the negative charges increase. In acidic soils, conditions favour the non-dissociated functional group, thus negative charges are low.

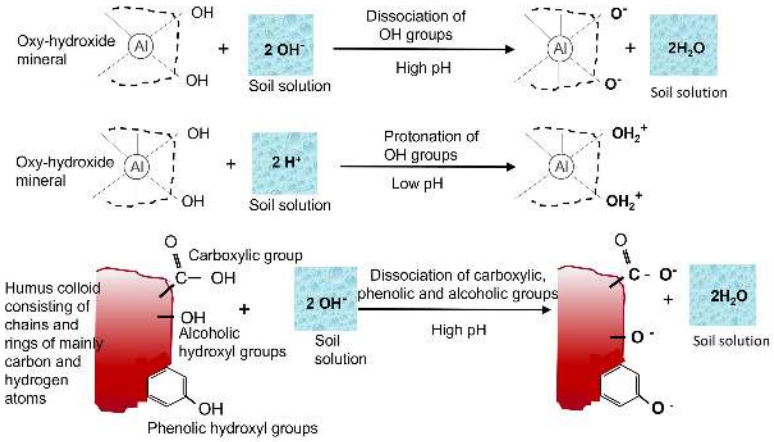


Figure 5.6. Variable pH dependent charges in oxy-hydroxide clays and soil humus particles. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Influence of Soil Colloids on Soil Properties

The amount and type of inorganic and organic colloids present in the soil has a profound influence on soil properties. For example, soils containing a high proportion of humus have high water holding capacities. Soils having a high proportion of freely expanding clay minerals, such as smectite, shrink when dry and swell when wet. Vertisols, fine textured soils containing >60% clay of which at least half is smectite, are typically very sticky when wet and form wide cracks when dry.

Phyllosilicates, oxy-hydroxides, and humus particles have electrical charges on their surfaces, which contribute to the total charge in the soils, and thus to the cation exchange capacity. The ranges of CEC associated with the different soil colloids are presented in Table 5.3. For example, a soil containing 40% smectite clay with CEC of 100 cmol(+) kg^{-1} may contribute up to 40 cmol(+) kg^{-1} to the total CEC of the soil. Given its high charge density, humus can be a significant contributor to the total CEC even in mineral soils—particularly under neutral to high pH conditions where pH dependent negative charges dominate. Depending on the characteristics of the humus, the pH, and interactions of humus with other soil particles, a 3% humus content (assuming a CEC of 400 cmol(+) kg^{-1}), may contribute up to 12 cmol(+) kg^{-1} to the CEC of the soil. The contribution of iron oxy-hydroxide clays to CEC is also pH dependent. However, compared to humus and phyllosilicates with high CEC, the contribution of iron oxy-hydroxides towards soil CEC is negligible. Conversely, oxy-hydroxides together with 1:1 type layer silicates may exhibit positive charges under acidic conditions and are thus a main source of **anion exchange capacity**.

Table 5.3. Ranges of CEC in different soil colloids. Adapted from Brady and Weil (2010) and Soil Survey Staff (2014)

Colloid	CEC Range [cmol(+) kg^{-1}]
Kaolinite	2 to 16
Chlorite	10 to 40
Hydrous Mica/Illite	20 to 40
Montmorillonite	60 to 100
Vermiculite	100 to 150
Organic matter (humus)	150 to 400
Amorphous aluminosilicates	5 to 350
Fe/Al oxy-hydroxides	~0 to 3

CATION EXCHANGE IN SOILS

Negatively charged surfaces on soil colloids attract cations by electrostatic forces. Cations retained by soil colloids may be eventually released to the soil solution through exchange with other cations in the soil solution. The cations most commonly held

by soil colloidal particles are Ca^{2+}, Mg^{2+}, K^+, Na^+, H^+ and Al^{3+}. Of these, Ca^{2+}, Mg^{2+}, K^+, and Na^+ are so-called base-forming cations, while H^+ and Al^{3+} are acid-forming cations. Exchangeable H^+ and Al^{3+} are found in increasing abundance in acidic soils.

Cation exchange is a process by which cations in the soil solution exchange with cations attached to the surfaces of clay minerals and humus. Cation exchange occurs on a charge-for-charge basis as illustrated in Figure 5.7.

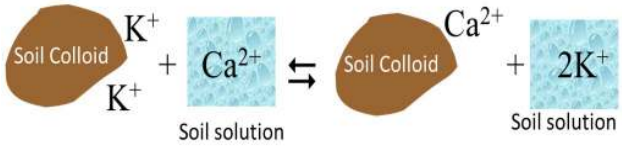


Figure 5.7. A schematic representation of a cation exchange reaction in soil. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Cation exchange reactions are **stoichiometric**. Thus, for divalent-monovalent exchanges (e.g., exchange between Ca^{2+} and K^+) one divalent cation exchanges with two monovalent cations. The reaction is rapid (almost instantaneous) and reaches equilibrium quickly. The cation exchange reaction is usually reversible, unless the exchanged cation is fixed in the interlayer space of a phyllosilicate (e.g., K^+ in hydrous mica/illite).

Cation exchange capacity (CEC) is defined as the sum of total exchangeable cations that a soil can adsorb (Brady & Weil, 2010). The CEC is usually expressed in centimoles of positive charge per kilogram of exchanger (cmol(+) kg^{-1}) or millimoles of positive charge per kilogram of exchanger (mmol(+) kg^{-1}). Thus a soil with a CEC of 10 cmol(+) kg^{-1} can adsorb 10 cmol of a monovalent cation (e.g., Na^+) or 5 cmol of a divalent cation (e.g., Ca^{2+}). In this example, if Na^+ exchanges with another monovalent cation (e.g., K^+), 10 cmol of K^+ would exchange with 10 cmol of Na^+. However, if the exchange is with a divalent cation (e.g., Ca^{2+}) only 5 cmol of Ca^{2+} will exchange with 10 cmol of Na^+.

Since it is the negative charges (including pH dependent negative charges) in the soil colloids that are responsible for the CEC of a soil, the amounts of inorganic and organic colloids, the type of clay minerals present, and the soil pH are the primary determining factors controlling the CEC of a soil.

Can You Dig It!

Ion exchange in soils is important to sustain life on Earth. Because of ion exchange, plant nutrients are retained in soil and released when needed by plants. Thus ion exchange occupies a central position unique to soils in their ability to supply plant nutrients and act as a filter to retain contaminants from reaching groundwater. Ion exchange retains toxic compounds, reducing their mobility, thus, protecting water resources from pollution.

Determination of Cation Exchange Capacity

The CEC of soil is generally estimated by saturating all cation exchange sites with an index cation and then replacing and measuring the amount of the index cation released to solution. The first step is to saturate a known mass of oven-dried soil with an index cation (e.g., Na^+ or NH_4^+). This is accomplished by equilibrating the sample with a solution containing the index cation buffered at pH 7.0 (usually sodium acetate or ammonium acetate). After ensuring that all cations are displaced by the index cation, the excess index cation is removed from the soil sample by leaching with a non-polar solution such as 95% ethanol. The amount of index cation held by the exchange sites in the sample must then be determined by leaching the soil with a solution containing a different cation (e.g., K^+) to replace the index cation on the exchange sites. Then the concentration of index cation in the leachate is measured and used to calculate the CEC of the soil (Hendershot et al., 1993).

Base Saturation

The proportion of CEC saturated by base-forming cations (typically Ca^{2+}, Mg^{2+}, K^+, and Na^+) is known as base saturation. **Base saturation percentage** is calculated using the equation:

(1) $$Base\ saturation\ \% = \frac{\sum\ Exchangeable\ base-forming\ cations}{CEC} \times 100$$

Both the CEC and the sum of exchangeable base-forming cations should be expressed in the same units (e.g., cmol(+) kg^{-1}). Since the dominant exchangeable base-forming cations in soils are Ca^{2+}, Mg^{2+}, K^+, and Na^+, the equation is usually written as:

(2) $$Base\ saturation\ \% = \frac{exch.\ Na^{+} + exch.\ K^{+} + exch.\ Ca^{2+} + exch.\ Mg^{2+}}{CEC} \times 100$$

Note that for divalent cations such as Ca^{2+} and Mg^{2+}, exchangeable cations in cmol (+) should be calculated considering the charge; i.e, 1 cmol Ca^{2+} = 2 cmol of charge (+). Base saturation is positively related to soil pH. If the pH is low (<7) then the base saturation is also low. And, as soil pH increases, base saturation also increases approximately linearly between pH 4 and 7.

ANION ADSORPTION AND EXCHANGE

Soils are able to retain anions in a similar fashion to retaining cations, particularly in the lower pH range when the surfaces of some colloids (e.g., oxy-hydroxides) have positive charges. Retention of anions through adsorption varies from one anion to another. Some anions are retained through non-specific adsorption mechanisms, while others are retained through specific adsorption mechanisms.

Non-specific adsorption refers to an anion that is retained by positively charged sites and the anions are weakly held through electrostatic forces. This type of adsorption is common for anions like chloride (Cl^{-}) and NO_3^{-}. These anions are weakly held on positively charged surfaces and they can freely exchange with other anions in soil solution; consequently, they may be easily lost through leaching. For non-specific adsorption to occur, the colloidal surfaces must be positively charged; thus, soil containing minerals with pH dependent charges have a high anion exchange capacity, particularly at low pH. Soils of the tropics contain high proportions of oxy-hydroxides and kaolinite, which exhibit positive charges under low pH conditions, thus in these soils, anion adsorption through non-specific mechanisms can be high.

Specific adsorption occurs when the ion becomes chemically bonded to the surface of the colloid. Mechanisms are specific to the anion being adsorbed. For example, phosphate anions (e.g., $H_2PO_4^{-}$) may be specifically adsorbed to certain colloidal surfaces through exchange of a surface hydroxyl on the colloid for an oxygen atom that is part of the anion (Figure 5.8). The strongly bonded anion is not mobile in soils and is less available to plants. Specific adsorption is only minimally reversible, resulting in permanent loss of the ion from the soil solution and from uptake by plants. In extreme cases, the original mineral surface may become completely coated with the adsorbed ion. The common anions that are retained in soil through specific adsorption are phosphate, sulphate, silicate, and carbonate.

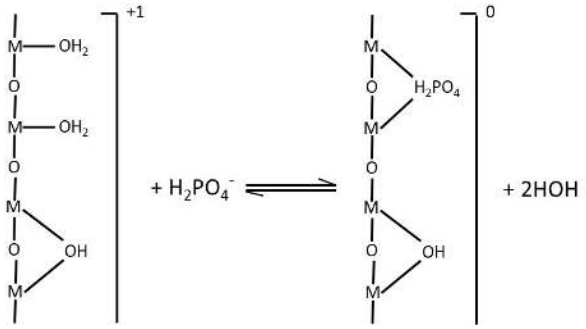


Figure 5.8. Mechanism of specific adsorption of phosphate to colloidal surfaces. Adapted from D. Shreeja (n.d.) by Darshani Kumaragamage. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

SOIL pH

Soil pH is a measure of the **activity** (cf. concentration) of hydrogen ions in the soil solution and describes the acidity or alkalinity in the soil. For our discussion herein we define activity as “effective” concentration, whereby activity is roughly equal to concentration in dilute solutions but generally deviating from concentration (typically becoming less than the concentration) as the total concentration of salts dissolved in the soil solution increases. For current purposes, we assume that activity and concentration are equal.

The standard pH scale spans from 0 to 14, with a pH value of 7 being neutral. Pure water has a pH value of 7.0. Values of pH less than 7 are acidic and become increasingly acidic as values approach zero. Values of pH greater than 7 are alkaline, or basic, and become increasingly alkaline as values approach 14. The pH scale is logarithmic, which means that for every one-unit change in pH there is a 10-fold change in the activity of hydrogen ions in the soil solution. For example, a soil with pH 6 has ten times more

hydrogen ions in solution than a soil at pH 7, and a soil with pH 8 has ten times less hydrogen ions than a soil at pH 7. A comparison of soil pH values for some soils with some common acids and bases are shown in the Figure 5.9.

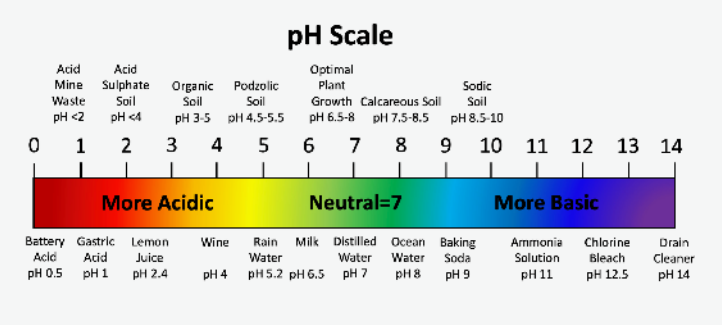


Figure 5.9. The pH scale with some common materials and soils. Adapted from Brady and Weil (2010). Adapted by Lewis Fausak and licensed under a CC BY (Attribution) license.

The primary factors influencing the natural soil pH are the geological parent materials of the soil and reaction with carbon dioxide (CO_2) dissolved in meteoric water. Organic acids derived from the decay of vegetation also influence soil pH. Generally, mineral soils in Canada are formed on one of two very broadly defined parent materials: those developed on granite-based parent materials, commonly referred to as acidic parent materials, and those developed on carbonate-based parent materials, commonly referred to as alkaline parent materials. Granitic-based parent materials are derived mostly from rocks of the **Precambrian Shield**, while carbonate-based parent materials, typically limestone and dolostone (dolomite), are derived from Paleozoic aged rocks or younger. Soils formed from granitic-based parent materials are generally in the acidic pH range. Conversely, soils formed on carbonate-based parent materials typically have soil pH values in the neutral to alkaline range.

Rainwater equilibrates with atmospheric gases, particularly CO_2, and has a pH of approximately 5.6. Other factors such as atmospheric pollutants (e.g. sulphur dioxide which hydrolyzes to sulphuric acid) typically drive the pH lower. The general trend for all soils is to be driven to the acidic end of the pH scale with increasing length of exposure to precipitation, which may cause leaching of basic cations. Granitic parent materials generally contain very little basic mineral material and therefore the pH of granitic-based soil is easily driven to the acid end of the pH scale. Podzolic soils, typically developed on granitic parent materials, have pH values in the range of 4.5 to 6.0. Although it may appear somewhat counter intuitive, the pH of soils containing carbonate minerals are buffered against the acidifying effects of dissolved CO_2 in the soil solution. Calcareous soil materials typically have pH values in the range of 7.5 to 8.5.

Soil pH may be buffered in some soils. That is, although pH reflects the concentration of H+ in solution at the time of measurement, there are additional sources of H^+ on exchange sites referred to as “exchangeable acidity” that buffer and maintain the pH in solution. Through exchange of H^+ between the soil solution and exchange sites, high CEC soils can buffer soil pH, which means they will resist a change in pH when an acidic or basic compound (e.g., a fertilizer) is added. Acidity, measured immediately in the soil solution, is referred to as “**active acidity**”, while acidity on the exchange sites is referred to as “**reserve acidity**”. Some soils are highly pH buffered with a large amount of reserve acidity (e.g., soils with high clay and/or organic matter content), while others are not (e.g., sandy soils with low organic matter content).

Soil pH affects plant nutrient availability. However, not all nutrients are available at a specific pH level. Soil pH values in the range of 6.5-8.0 are considered optimum for the availability of most nutrients to plants. Generally, extreme pH values are detrimental to nutrient uptake and plant growth. For example, acidic pH values increase the solubility of aluminum, manganese and iron, which are potentially toxic to plant roots when in high concentrations in the soil solution. These elements also may react with plant nutrients to render them unavailable to plants. Similarly, alkaline pH values result in excess amounts of calcium in the soil that also react with nutrients (e.g., phosphate) so they are unavailable for plant use. General ranges of nutrient availably with pH in mineral and organic soils are shown in Figure 5.10.

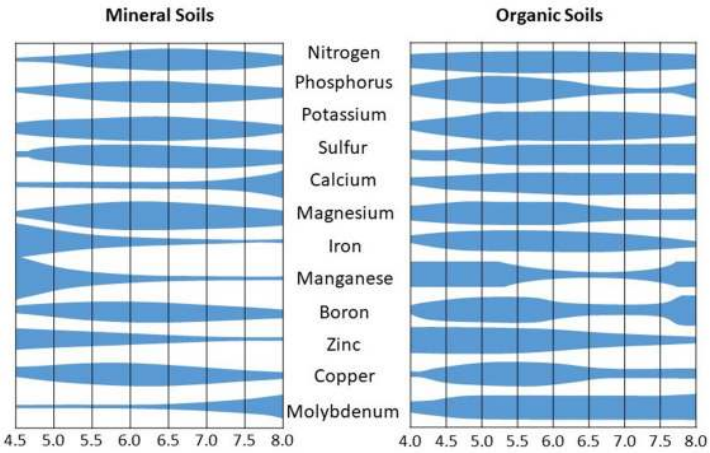


Figure 5.10. Nutrient availability in mineral and organic soils influenced by soil pH. Adapted from Munroe (2018) by Lewis Fausak and licensed under a CC BY (Attribution) license.

Acids and Bases

Soil pH values reflect the various reactions involving acid-base pairs in water, which itself dissociates:

$$H_2O \leftrightarrow H^+ + OH^-$$

An aqueous solution is said to be neutral (pH = 7) when the concentration of protons [H^+] exactly equals the concentration of hydroxyl ions ([OH^-]), with both equal to 10^{-7} moles per litre.

An acid is any chemical substance that can donate a proton when dissolved in water, while a base is any substance that can accept a proton. When an acid loses a proton it forms the conjugate base:

$$HL \leftrightarrow H^+ + L^-$$

In the above expression HL is an acid capable of donating a proton (H^+) to solution and L^- is the conjugate base, capable of accepting a proton. In this expression, the base (L^-) or conjugate base is also referred to as a ligand. The strength of an acid is the measure of its ability to dissociate in water and release protons to solution. A strong acid (e.g., sulphuric acid) is a substance that completely dissociates into protons and ligands.

A base is any chemical substance that can generate hydroxide ions (OH^-) or takes up H^+ ions in solution forming its conjugate acid.

$$B + H_2O \leftrightarrow OH^- + BH^+$$

Humified soil organic matter, although perhaps still poorly understood in terms of specific structural/chemical makeup, is the source of surface functional groups that influence the chemistry of the soil. Some of the most important organic functional groups include carboxylic (–COOH), phenolic-OH, alcoholic-OH, carbonyl (C=O) and quinone groups, as well as amine (-NH_2) and thiol (-SH) groups. The carboxylic acid, phenolic-OH and alcoholic–OH functional groups contribute to acidity and ion-exchange. These groups behave as weak acids in soil acting as a source of reserve acidity and serve as cation exchange sites. Other functional groups, such as aromatic acids, dissociate and function in a similar fashion.

Soil pH controls many of the chemical processes that take place in the soil, especially plant nutrient availability. Most soils have pH in the range of 5 to 8 with the optimal pH range for most agronomically important plants between 5.5 and 7.5, though many plant species have adapted to pH values outside of this range. For example, blueberries prefer soil with pH in the range of 4.3 to 5.5, growing very poorly in soils with pH in the alkaline range (pH>7).

If soil contains significantly more Na^+ in the soil solution compared to Ca^{2+} or Mg^{2+}, there may be a risk of sodium buildup. This ion buildup is of greater concern in arid or semi-arid regions (e.g., Canadian prairies) than in more humid regions (e.g., eastern Canada). Soils with naturally high sodium levels, or those that have received large quantities of sodium bicarbonate (Na_2CO_3) through irrigation, may have pH levels of 8.5 or higher. This high pH value is due to the presence of Na_2CO_3 precipitates in the soil which produce the elevated pH. This process is a characteristic of Solonetzic soils discussed later in this chapter.

Measurement of Soil pH

The pH of a soil is likely the most frequently measured soil chemical property. Soil pH measurements are typically made using a commercial pH meter equipped with a combination glass pH electrode. Combination electrodes are single electrodes that include both a glass electrode and a reference electrode. There are numerous methods for soil pH measurement, which depend on the application. Methods vary for the soil to solution ratios used; addition of an electrolyte and sample resting time prior to measurement. The methods include, but are not limited to measurements in: saturated paste; 1:1, 1:2, 1:2.5, 1:5, and 1:10 soil:water suspensions; 1:2 and 1:5 soil:0.01M $CaCl_2$ suspensions; 1M KCl; and 1.0M NaF (Soil Survey Staff, 2014; Hendershot et al., 2008). The soil pH measurement is that of the solution in contact with the soil solids and not of the soil solids themselves (Hendershot et al., 2008). Soil pH values measured in electrolyte (e.g., 1:2 soil to 0.01 M $CaCl_2$) suspensions are usually more stable and reproducible than those taken in water because the concentration of the electrolyte is constant. However, the addition of an electrolyte generally produces a decrease in soil pH compared to water; e.g., values measured in soil:water suspensions are typically 0.5–0.6 units higher than those in 0.01M $CaCl_2$, and 0.7 units higher than those in 1M KCl. This is because of the inclusion of exchangeable (reserve) acidity introduced to solution through ion exchange reactions involving Ca^{2+} or other introduced ion with Al^{3+} and/or H^+ ions on the exchange sites.

Although the pH of a soil containing a carbonate mineral should ideally be 8.2, soil pH measurements for these soils are commonly in the 7.6 to 8.5 range. This measured pH is a result of several factors. First, in addition to exchangeable ions such as Ca^{2+}, Mg^{2+}, Na^+ and K^+, the ion exchange sites on soil also contain small amounts of exchangeable aluminum and iron, which on entering the soil solution hydrolyze and react to reduce the measured soil pH. Second, although soils may contain carbonates, the mineral dolomite [$Ca,Mg(CO_3)_2$] reacts (dissolves) much slower compared to calcite ($CaCO_3$), and the pH may not stabilize within the timespan (e.g. within an hour) of the measurement. Third, typical procedures for pH measurements (e.g., Hendershot et al., 2008) recommend allowing prepared sample suspension to stand for one hour prior to pH measurement. This procedure provides a standard reaction time allowing easy adaptation for routine high sample throughput operations in commercial soil laboratories that follow standard ISO17025 accredited analytical methods. Although this time frame is sufficient to allow for ion exchange equilibration, it may or may not be sufficient to allow full equilibration for dissolution of carbonate and equilibration with CO_2 in the atmosphere. Other chemical reactions may also have an influence on the measured pH.

Soil pH Adjustment

The natural tendency is for soil to become more acidic through the natural acidity of rainwater, decay of organic matter and through application of fertilizers such as urea and anhydrous ammonia. However, soil pH may also be increased through additions of agricultural limestone. The timing and amount of agricultural "**lime**" applied depends on the crop to be grown, with lime application being commonly advised when soil pH values decrease below 6.0 (Munro, 2018). Finely ground limestone (primarily comprised of the mineral calcite; $CaCO_3$) or dolostone (primarily comprised of the mineral dolomite; $Ca,Mg(CO_3)_2$) are the most frequently used liming products, though many other products with acid neutralizing capacity may be used.



Images of the Martindale area of Sudbury in 1981 **(top)** and 2018 **(center)** after application of the "Regreening Recipe", which includes manual application of lime **(bottom)**. Photo credits: K. Winterhalder and P. Beckett.

The Sudbury Regreening Program

A century of sulphur dioxide fumigation and nickel particulate deposition by base-metal smelters, coupled with fire, soil erosion, and enhanced frost action left thousands of hectares of barren land with eroded soils and stunted, open birch-poplar woodlands in the Sudbury area.

The relict soils of the area are coarse textured, eroded Podzols that are acidic, metal-rich, nutrient poor and drought-prone—all of which have contributed to limit plant recolonization. The drought vulnerability of seedlings germinating in the soils was further enhanced by aluminium toxicity limiting root growth, and further exacerbated by enhanced frost action in the medium textured soil that lacked the protection of surface humus forms previously lost as a result of wind and water erosion.

A revegetation program that detoxified the soils enough to allow for successful colonization was developed based on field research in the 1970's. This research demonstrated that the surface application of ground dolomitic limestone [primarily $CaMg(CO_3)_2$] at 10 tonnes per hectare adjusted the soil pH from less than 3.0 to between 5.2 and 5.6. Application of the liming treatment, with or without an accompanying fertilizer and/or a grass-legume seed application, led to the natural colonization by woody species including birch, aspen, and willows. This colonization has further been aided by planting a mixture of native coniferous and deciduous species together in groups to form a "seed bank" for future colonization – with nearly 10 million trees being planted in the Sudbury region between 1976 and 2020.

Adjustment of soil pH towards the acid end of the scale may be required when growing certain crops such as potatoes, blueberries, azaleas or some other horticultural crops. Downward pH adjustment is typically accomplished with the addition of elemental sulphur, sulphur containing compounds or ammonia containing fertilizers; however, application of amendments to intentionally decrease soil pH is very uncommon.

SOIL SALINITY AND SODICITY

Soil **salinity** levels are controlled by the concentration of total dissolved mineral salts present in the soil solution. Salts occur naturally in both soil and water and may, under certain conditions, accumulate in soil. The accumulation of soluble salts in the soil is known as **salinization**. Dissolved salts that contribute to soil salinity include the cations Na^+, Ca^{2+}, Mg^{2+}, and K^+, and the anions Cl^-, SO_4^{2-}, HCO_3^-, CO_3^{2-}, and NO_3^-.

Sodicity is related to salinity, but refers specifically to the concentration of Na^+ in the soil solution relative to concentrations of Ca^{2+} and Mg^{2+}. The degree of sodicity, estimated by calculation of the **sodium adsorption ratio** (SAR), is a measure of the concentration of Na^+ in saturated soil extract relative to Ca^{2+} and Mg^{2+} concentrations. The SAR is calculated as the ratio of the Na^+ concentration divided by the square root of average [Ca^{2+} + Mg^{2+}] concentrations. The square root of [Ca^{2+} + Mg^{2+}] concentrations is used since these are divalent ions while Na is monovalent. Sodicity can be quantified as **exchangeable sodium percentage** (ESP); i.e., the exchangeable Na^+ expressed as a percentage of cation exchange capacity. If the SAR of the soil equals or exceeds 13, or if the ESP of the soil equals or exceeds 15%, the soil is termed sodic.

$$SAR = \frac{[Na^+]\ (mmol\ L^{-1})}{(0.5 \times [Ca^{2+} + Mg^{2+}])^{\frac{1}{2}}\ (mmol L^{-1})} \quad (3)$$

$$ESP = \frac{[Na^+]\ (cmol(+)\ kg^{-1})}{CEC\ (cmol(+) kg^{-1})} \times 100 \quad (4)$$

Development of Salinity in Soils

Salts in soils originate from natural and anthropogenic sources. Natural sources include weathering of primary minerals, atmospheric deposition, sea water intrusions, discharge of saline groundwater, and brines from natural salt deposits. Major anthropogenic sources of salts include irrigation with waters containing high salt concentrations (see Figure 5.11), fertilization with synthetic and organic fertilizers with high salt content including manure, biosolids, and compost. Road salts applied during winter in Canada to melt ice and improve road conditions, are also a localized source of salinity to soils.

Soluble salts accumulate in soils for various reasons:

- Dry climatic conditions – Soluble salts accumulate in arid and semiarid areas where rainfall is not sufficient to leach the salts through the soil profile. High evaporation losses from these soils result in the upward movement of water, which may bring salts from the deeper soil layers that then remain at the soil surface as visible crusts when water evaporates. Evaporative losses can range from 50–90% in arid regions resulting in 2- to 20-fold increase in levels of soluble salts.
- Drainage – Poor drainage (due to high water table or low subsoil permeability) can increase the salt content.
- Irrigation water quality – Salinity problems are common in irrigated lands as most irrigation waters contain varying concentrations of soluble salts. The water added to the soil is utilized by the crop or evaporates directly from the moist soil, leaving the dissolved ions behind to form saline crusts.
- Coastal regions – Soils in coastal areas may have high salt accumulation through sea water intrusions and saline groundwater.
- Other management practices – Salt accumulation may take place due to other management practices such as overuse of synthetic fertilizers, continuous application of manures, or yearly application of road salts that are washed into road-side ditches and/or blown onto nearby fields and vegetation by wind.

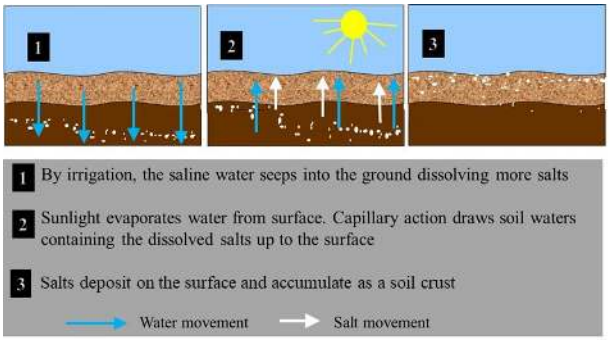


Figure 5.11. Salinization of soils by irrigation with saline water. © Graeme Spiers, Laurentian University is licensed under a CC BY (Attribution) license.

Measurement of Soil Salinity

The total dissolved salt concentration of a soil can be measured by extracting a soil sample with water and measuring the concentrations of individual salts in the extract. In practice however, measuring individual salts is time consuming, cumbersome and expensive; thus the electrical conductivity (EC) of a solution, which increases in direct proportion to the overall abundance of dissolved salts, is employed as an alternative index to assess and describe soil salinity. Soil extract EC is the preferred measurement to assess the overall salinity status of the soil. The measurement of EC—accomplished with a conductivity meter in soil testing

laboratories—is reliable, rapid and cost effective. The EC of a soil is determined in a saturated extract (i.e., in the soil solution extracted from a saturated soil paste) and expressed in units of decisiemens per meter (dS m^{-1}) or millisiemens per centimeter (mS cm^{-1}). Soils with an EC value of saturated soil extract >4 dS m^{-1} are classified as saline.

Properties of Salt-Affected Soils

Salt-affected soils can be classified as **saline**, **saline-sodic** or **sodic**, depending on the total concentration of Na^+ salts in the soil. Saline soils have high salt content but low concentrations of Na^+. Sodic soils have relatively low salt content but high concentrations of Na^+. Saline-sodic soils have high concentration of both salts and Na^+. Figure 5.12 illustrates the classification of salt-affected soils based on the EC, SAR and ESP.

Saline soils have high concentrations of total dissolved salts, with EC values greater than 4 dS m^{-1} (Figure 5.12). These soils have relatively low amounts of Na^+ in the soil solution or on the exchangeable complex, and thus SAR values are less than 13 and ESP values are less than 15%. The soil pH of saline soils is usually less than 8.5. The high salt concentrations in these soils promote flocculation of soil colloids (i.e., individual colloids are attracted together forming flocs or clusters).

Sodic soils have low total concentrations of dissolved salts, with EC values less than 4 dS m^{-1}. These soils, however, have relatively high concentrations of Na^+ in the soil solution and on the exchange complex relative to Ca^{2+} and Mg^{2+}. Therefore, SAR values are greater than 13 and ESP values are greater than 15% (Figure 5.12). Sodic soils typically have pH values greater than 8.5 due to the hydrolysis of Na_2CO_3. The colloids in these soils are dispersed due to the swelling in response to the high proportion of exchangeable sodium. Swelling and dispersion is greater when the exchange complex has a greater proportion of hydrated, monovalent ions particularly Na^+. When hydrated Na^+ ions come between clay particles, the electrostatic forces that bind the clay particles together are neutralized and additional layers of water molecules force the particles apart, thus dispersing the soil matrix materials. The high pH also disperses soil organic matter. Alkaline sodic soils classified as Black Solonetz have dispersed, dark coloured humic material coating the soil peds.

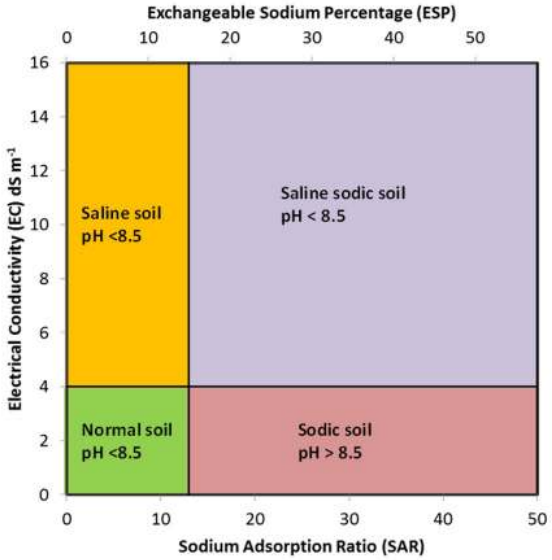


Figure 5.12. Classification of salt-affected soils. Adapted from McKenzie and Woods (2010) by Jim Warren and redrawn by Darshani Kumaragamage, Univ. of Winnipeg. © Jim Warren and Darshani Kumaragamage is licensed under a CC BY (Attribution) license.

Soils having high concentrations of dissolved salts as well as high concentrations of Na^+ are termed saline-sodic soils. Such soils have EC values greater than 4 dS m^{-1}, SAR values greater than 13 and ESP greater than 15%. Saline-sodic soils usually have pH values less than 8.5. These soils are flocculated because of the high concentration of dissolved salts (Sparks, 2003).

Effects of Soil Salinity and Sodicity on Ecosystem Eealth

Soil salinization, one of the most devastating global environmental concerns, is a major cause of land degradation and contributes to desertification of dry ecosystems. Worldwide, salinization of soils leading to salinity and sodicity have caused major reductions in cultivated areas and crop productivity.

Both salinity and sodicity affect soil properties, plant growth, and microbial activity. High salt concentrations will adversely affect plant growth as the dissolved salts lower the osmotic potential of the soil solution. Since plant roots take up water primarily by osmotic mechanisms, they are unable to absorb water from highly saline soils, with some plants being more sensitive to saline conditions than others. This resultant condition is known as “physiological drought” since plant roots are unable to take up water even when the water content in the soil may be adequate for normal growth. Similarly, microbes in the soil may be affected because of an inability to obtain water due to osmotic effect.

Table 5.4. Sensitivity of some plants to saline conditions

Sensitive	Moderately Sensitive	Moderately Tolerant	Tolerant
(EC < 4 dS m^{-1})	**(EC 4-7 dS m^{-1})**	**(EC 7-9 dS m^{-1})**	**(EC 9-12 dS m^{-1})**
Vegetables	Alfalfa	Fescue	Barley
Maple	Corn	Aspen	Wheat
Tomatoes	Grasses	Oats	Willow
Potatoes	Clover	Soybean	
Note: plants capable of surviving EC in the range of 12–15 dS m^{-1} are considered halophytes; however, very few, if any, plants can survive EC values >15 dS m^{-1}			

The high concentration of sodium in sodic soils favours swelling of expandable clay minerals and dispersion of clay particles, resulting in poor soil structure, reduced permeability and infiltration due to blockage of the soil pores, and formation of surface crusts. Low infiltration and permeability may also increase soil erosion and may lead to saturation of the soil surface during irrigation and rain/snowmelt events, in turn restricting air entry into the soils. This leads to anaerobic conditions in soils which also negatively affects plant growth.

Salt-Affected Soils: Distribution Globally and in Canada

Soil salinization is a global problem. Salt-affected soils occur on all continents, mostly in arid and semi-arid areas. Salinization is of lesser concern in humid areas as rainfall is sufficient to leach excess salts from the soil profile. Salinized areas in the world increase at a rate of about 10% annually because of natural and anthropogenic causes. Based on the 1970–1980 world soil map, the total area of salt affected soils worldwide is estimated as 397 million ha of saline soils and 434 million ha of sodic soils (FAO, 2020). Figure 5.13 illustrates the distribution of saline and sodic soils worldwide (Zheng, 2014). Globally land degradation due to salinization has been reported in the Aral Sea Basin (Amu-Darya and Syr-Darya River Basins), the Indo-Gangetic Basin in India, the Indus Basin in Pakistan, the Yellow River Basin in China, the Euphrates Basin in Syria and Iraq, the Murray-Darling Basin in Australia, and the San Joaquin Valley in the United States (Qadir et al., 2014).

There are an estimated 6 to 8 million ha of salt-affected soils in Canada, located primarily in the Canadian prairies (Miller and Brierley, 2011). In 1995, the risk of salinization was rated as moderate to very high affecting an estimated 34% of farmland area in the Prairie Provinces (Acton and Gregorich, 1995), with the estimate decreasing to 12% of farmland area by 2001 (Eilers et al., 2010). This decrease in the risk of salinization potential in agricultural lands in the Prairie Provinces is largely due to the reduction in summer fallow, and conservation practices such as conversion to permanent cover.

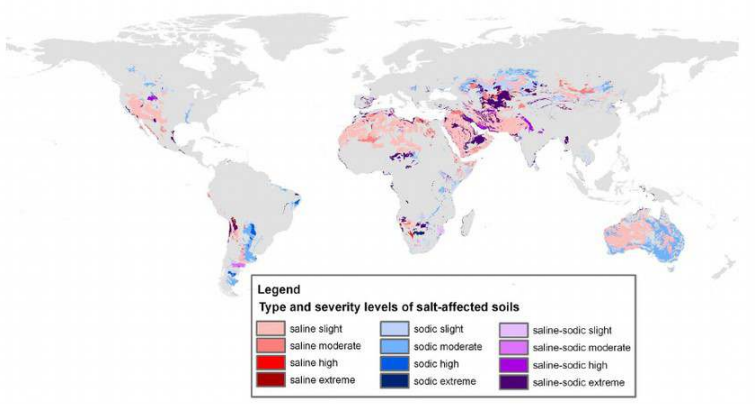


Figure 5.13. Global distribution of saline and sodic soils. Source: Zheng (2014), used with permission.

Can You Dig It!

Surface and subsurface salinity in the Canadian Prairies – A threat to soil quality and sustainable ecosystem productivity

Soil salinization is a typical process for the Canadian prairies and adjacent areas of the Boreal Plains. It is estimated that at least 1,431,00 ha in the region are affected by salinity. The two images below show white saline crust, which often occurs adjacent to shallow wetlands (sloughs) on the prairies. The salts are dominantly calcium and magnesium sulphates (gypsum and epsom salts) derived from evaporation of saline groundwater discharge from the bedrock below the regional glacial till veneer. The surface expression is the result of a gently rolling topography combined with a high water table in the wetland.



Surface expression of a severely salt affected soil from southwest Manitoba **(left**; Photo credit: Robert G. Eilers). Chernozemic to Solonetz catena near Drumheller, Alberta **(right**; Photo credit: G.A. Spiers). Note the saline crust in the lowland area.



Roots on the surface of a Brown Solodized Solonetz column impacted by salt efflorescence at base (Photo by G.A. Spiers)

The image to the right shows the effects of subsurface salinity. The most common expression of salt-affected soils is irregular growth of crop plants across a field. The soil surface may look normal, but salinity is often greater at depth in the root zone. Plants typically germinate and grow after seeding due to available moisture at the surface. As the soil dries and plant water consumption increases, the roots penetrate deeper searching for water. At some depth they can encounter higher levels of salt in the subsoil, which acts as a barrier to accessing moisture. The result is stunted or irregular plant growth, and possibly death due to lack of moisture.

Reclamation of Salt Affected Soils

Methods of removing salts from surface soils include scraping, surface flushing and leaching. Since the salts are highly soluble, leaching with good quality water is the most effective and commonly used method of reclaiming saline soils. While saline soils are aggregated and have high infiltration rates, sodic and saline-sodic soils are usually dispersed and have low infiltration rates. Thus, it is difficult to leach the salts and sodium with water from sodic and saline-sodic soils since they are typically dispersed. Thus, to reclaim sodic and saline-sodic soils, application of soil amendments such as gypsum ($CaSO_4 \cdot 2H_2O$) is usually needed. The added Ca^{2+} promotes soil aggregation and improves water infiltration when monovalent Na^+ is replaced by divalent Ca^{2+}. The displaced Na^+ and excess salts can then be leached by irrigation with good quality water. However, because leached salts may accumulate in the groundwater, saline drainage water may be diverted through subsurface drains. Other ways of removing salts from salt-affected soils include growing salt tolerant crops (e.g., barley, sunflower) and planting trees that will take up the dissolved cations and anions from soil solution. Preventing salinization by using best management practices (e.g., use of good quality irrigation water in agricultural lands, managing water below root zone through subsurface drainage) is a better approach than allowing salinity to increase in soils and then attempting to reclaim the salt-affected soils.

OXIDATION-REDUCTION (REDOX) REACTIONS

Oxidation-reduction reactions, or redox reactions, involve the transfer of electrons and the conversion of an element from one valence state to another. Understanding the behaviour of redox sensitive elements in soil is central to understanding the behaviour of Gleysolic and Organic soils, and gleyed soil subgroups that occur within all mineral soil orders in Canada (SCWG, 1998). Internationally, about 130 million ha of land, accounting for about 10% or all arable land on the planet, is temporarily flooded for rice production generating reduced soil conditions and emphasizing the importance of understanding redox reactions (IRRI, 1997).

Redox reactions are typically written as half reactions (i.e., the electron and change in oxidation state are explicitly stated in the equation). The reversible reaction is always to be written as a reduction reaction with the electron (e^-) on the left side of the equation:

$$\text{oxidants} + e^- \leftrightarrow \text{reductants}$$

where oxidants are chemical species capable of accepting electrons and reductants are chemical species capable of donating electrons. For example:

$$Fe^{3+} + e^- \leftrightarrow Fe^{2+}$$

Redox measurements are useful to provide an understanding of chemical behaviour in wet and poorly drained soils (e.g., Gleysolic or Organic soils and gleyed subgroups; see Figure 5.14) as a high redox potential (>300 mV) indicates good availability of O_2 (aerobic conditions), whereas a low redox potential (<300 mv) indicates limited availability of O_2 (anoxic conditions).



Figure 5.14. Poorly drained Gleysol from Delta, BC during wet fall/winter. © Maja Krzic, UBC is licensed under a CC BY (Attribution) license.

Key redox sensitive elements in soil include oxygen (O_2), nitrogen (NO_3^-, NO_2^-, NO_x, N_2O, N_2), manganese (Mn^{4+}, Mn^{2+}), iron (Fe^{3+}, Fe^{2+}), sulphur (SO_4^-, S^{2-}), and carbon (CO_2, CH_4). These are the main species commonly involved in redox reactions in soils and constitute a reaction sequence indicating the tendency of each element to become reduced:

Oxygen > Nitrogen > Manganese > Iron > Sulphur > Carbon

Typically, elements on the left of this sequence dominate the prevailing oxidation state of the other elements to the right of it. For example, the presence of oxygen dominates electron acceptance over all other elements in the sequence and thus dictates that each of the other species must donate their electrons and be converted to their oxidized form. Thus, in the presence of oxygen (i.e., oxidizing environments) the stable valences of the other elements are NO_3^-, Mn^{4+}, Fe^{3+}, SO_4^{2-}, and CO_2.

Table 5.5. Half reactions, standard electrode potentials (E°), typical Eh range in soils and mediating micro-organism group for dominant redox species in soil

Half Reaction	E° (millivolts)	Eh range (millivolts)	Micro-organism
$O_2(aq) + 4\ H^+ + 4\ e^- \leftrightarrow 2\ H_2O$	1230	600 to 350	Aerobes
$2\ NO_3^- + 12\ H^+ + 10\ e^- \leftrightarrow N_2(g) + 6\ H_2O$	1240	400 to 200	Denitrifiers
$Mn(VI)O_2(s) + 4\ H^+ + 10\ e^- \leftrightarrow Mn2+ + 2\ H_2O$	1220	300 to -50	Mn-reducers
$Fe(III)(OH)_3(s) + 3\ H^+ + e^- \leftrightarrow Fe^{2+} + 3\ H_2O$	930	50 to -50	Fe-reducers
$SO_4 + 9\ H^+ + 8\ e^- \leftrightarrow S2- + 4\ H_2O$	250	-150 to -200	Sulphate reducers
$CO_2(aq) + 8\ H^+ + 8\ e^- \leftrightarrow CH_4(aq) + 2\ H_2O$	170	-150 to -250	Methane fermenters

The notation used to identify redox species in in the aqueous phase (i.e., soil solution) is different than that use to identify elements occurring in the solid phase. For example, iron species in the aqueous phase are typically noted as Fe^{2+} and Fe^{3+}, whereas iron species present in soil minerals and other solids are expressed as Fe(II) and Fe(III).

Generally, oxidative decomposition (decay) of organic matter is mediated by various microorganisms to derive energy (see Table 5.5). Consequently, organic matter is the primary source of electrons in soil. Conversion/decay of organic matter varies considerably from complete conversion to CO_2 (gas) to partial and incomplete oxidation and conversion to intermediate products (i.e., conversion to humus), but all involve the reduction of redox species to their reduced form. Because the composition of soil organic matter is highly complex (see Chapter 3) there is no general chemical formula attributed to it (van der Park, 2006).

Figure 5.15 is an example of an Eh-pH diagram, or Pourbaix diagram, a graphical representation showing the approximate ranges for Eh and pH in soil. Soil pH values generally range from about 4 to 10, with Eh bounded on the top and bottom by the redox stability of water. The Eh varies with pH according to Nernst equation by -59.16 mV per pH unit at 25 ℃ (298.15K). Figure 5.15 also indicates the approximate ranges for oxic conditions (O_2 reduction reaction is dominant and mediated by aerobic organisms); sub-oxic conditions (nitrogen, iron and manganese reduction reactions are dominant and mediated primarily by facultative organisms); and anoxic conditions (sulphur and carbon reduction reactions are dominant and mediated by anaerobic organisms) in soil.

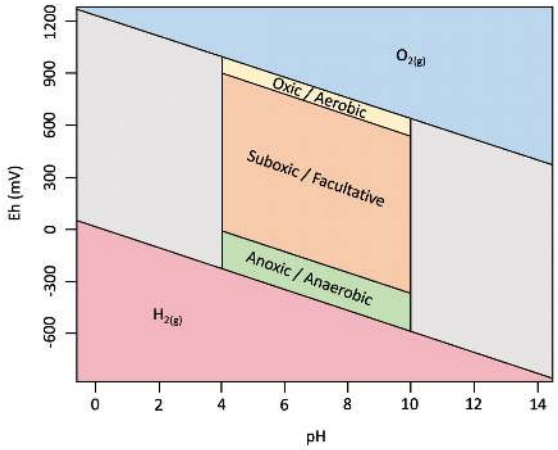


Figure 5.15. Common Eh and pH ranges for soils including approximate ranges for oxic, sub-oxic and anoxic environments. Figure created by James Warren, adapted by Lewis Fausak and licensed under a CC BY (Attribution) license.

Oxygen

In non-saturated (aerobic) soils, oxygen occurs in both the soil gas phase (O_2 g) and in the soil solution (aqueous) phase (O_2 aq). Although similar in composition to that of the atmosphere (about 20% O_2 by volume), the concentration of oxygen in soil air is typically somewhat lower as oxygen in the soil gas phase is constantly consumed by plant roots and microbial respiration. Under oxygenated (non-saturated) conditions, oxygen is effectively in infinite supply capable of accepting electrons and completely oxidizing soil organic carbon to CO_2 (g). Continued introduction of oxygen to topsoil due to cultivation practices is the primary reason for the reduction in soil organic matter levels as a result of oxidation reactions.

Although the concentration of oxygen in the soil gas phase is usually high, the concentration of dissolved oxygen in water is far less, being generally on the order of 10 mg L^{-1} (at 15 ℃) but varying with temperature. As dissolved oxygen becomes consumed, it is replenished by the normally abundant pool of oxygen in the soil gas phase. Aside from the large difference in the abundance of oxygen, the mobility of oxygen is roughly 10,000 times slower in water compared to oxygen in the gas phase. Consequently, as soils become increasingly saturated with water the oxygen supply from air becomes increasingly limiting until all, or almost all, of the soil pore space becomes filled with water, and oxygen transport is possible only via the aqueous phase. The redox potential decreases as the soil effectively becomes sub-oxic or even anoxic/anaerobic (see Figure 5.15).

Can You Dig It!

Seasonally anoxic soils during spring snowmelt flooding



Flooded farmland in southern Manitoba during the spring snowmelt period. Photo credit: Don Flaten.

Most farmlands in the Canadian prairies get flooded during the spring snowmelt period. Seasonal flooding can make the soil sub-oxic or anoxic as the redox potential decreases. This will have many environmental impacts. For example, phosphorus concentration in floodwater may increase when Fe(III) and Mn(IV) are reduced to divalent forms releasing the phosphate associated with them. Released phosphorus may end up in lakes causing algal blooms. Anoxic conditions may also enhance emissions of greenhouse gases such as nitrous oxide (N_2O) and methane (CH_4).

Nitrogen

The primary reason for the depletion of available oxygen from soil is saturation with water. As discussed above the amount of dissolved oxygen in soil water is very limiting and diffusion of oxygen through water is very slow. As oxygen becomes limiting/depleted, typically under temporary saturated conditions, nitrate-reducing microbes become active initiating the process of denitrification. The process of denitrification is a major pathway for the loss of fertilizer nitrogen applied as a nutrient source for crops. Nitrate (NO_3^-) is reduced to nitrogen gas (N_2) with nitrite (NO_2^-), nitric oxide (NO) and nitrous oxide (N_2O) as intermediate species. Soluble nitrite is a short-lived aqueous intermediate while nitric oxide (NO) and nitrous oxide (N_2O) intermediates are involved in the destruction of stratospheric ozone (O_3) layer. Nitrous oxide is also an important greenhouse gas with a global warming potential (GWP) 265–298 times that of carbon dioxide.

Manganese

As nitrate becomes depleted, manganese reducing bacteria take over. Although the total content of manganese in soil is typically only about 750 to 1000 mg kg^{-1} (Ure and Berrow, 1982; Bowen, 1979) it is still a major redox species in soil. Manganese exists in soil mainly as Mn^{2+}, or as Mn(VI). The soluble form prevalent in anoxic soil is Mn^{2+} (aq), which when oxygen is introduced oxidizes and precipitates manganese as Mn(IV), usually as non-crystalline (amorphous) oxy-hydroxides. Manganese oxides appear as black coatings on soil particles, or as black nodules, some as large as 2 cm in diameter. When present, manganese solids act as scavengers for nutrient anions (e.g., phosphate).

Iron

Iron is the fourth most abundant element in soil and is responsible for diverse mottles characterized by the drastic differences in colouration between reduced (Fe^{2+}) and oxidized (Fe^{3+}) valences in wet and water-logged soils. Strongly reduced soils with Eh values in the range of -200 mV typically have gleyed (dull grey or practically white) colour due to the presence of reduced iron. This reduced soil condition means that dissolved oxygen as well as nitrate and manganese are essentially absent and that most dissolved iron is present as the Fe^{2+} species, with most iron compounds (responsible for the grey colour) occurring as Fe(II) oxides and hydroxides (e.g., ferrous hydroxide [$Fe(OH)_2$]. When the water table lowers or oxygenated water infiltrates into a reduced soil (perhaps through a root channel, open pore or fracture), ferrous iron transfers electrons to the oxygen resulting in the formation of oxidized [Fe(III)] compounds. These are observed as mottles occurring as reddish blotches on an otherwise dull grey background. Mottles are a very common morphological feature in Gleyed sub-groups or Gleysolic soils that are periodically wet or have a fluctuating water table. Mottles are diagnostic features in the identification of moderately well, imperfectly, and poorly drained soils.

Sulphur

Although there are several oxidation states for sulphur, only two oxidation states are prevalent in soil. Under aerobic conditions sulphate (SO_4^{2-}) is the stable form with an oxidation state of +6, while sulphide (S^{2-}) with an oxidation state of -2 is the stable form under strongly reducing conditions (McBride, 1994). Sulphate is the primary form of sulphur taken up by plants, precipitating primarily as the mineral gypsum ($CaSO_4 \cdot 2H_2O$) when there is sufficient dissolved calcium (Ca^{2+}) present in the soil solution. Sulphate may be reduced to sulphide which has very low solubility in water under anoxic conditions forming either odourous hydrogen sulphide gas (H_2S with the aroma of rotten eggs), or insoluble metal sulphides such as FeS_2 if soluble metals are present.

The introduction of oxygen into sulphide containing soils results in the re-oxidation of metal sulphides:

$$FeS(s) + {}^{15}/_{4}\, O_2 + {}^{7}/_{2}\, H_2O \leftrightarrow Fe(OH)_2 + 4H^+ + 2\, SO_4^{2-}$$

This reaction is common in reclaimed coastal soils high in sulphides, in Northern Prairie soils derived from sulphide containing shales (referred to as acid sulfate soils) and in mine wastes (tailings and waste rock) containing pyrite and other metal sulphides.

Carbon

Ultimately if sulphate is limited, then methane producing bacteria continue the process of organic matter decomposition under anaerobic conditions as follows:

$$2CH_2O \leftrightarrow CO_2 + CH_4$$

Methane production typically occurs under stagnant, water saturated conditions in wetlands. Under these conditions the organic decay process is slow and incomplete, resulting in the accumulation of soil organic matter and the formation of peatlands and organic soils.

Measurement of Redox

The direct measurement of redox (oxidation-reduction) potential in soil is a quandary. Fundamentally, redox measurement is extremely simple, being similar to pH measurement in mechanics, but very different from pH with respect to fundamental chemistry and interpretation. Redox measurements or measuring the "concentration of electrons" is not analogous to pH because free electrons do not exist in soil solutions in the same way as hydrogen (H^+) ions. Redox is typically measured as a voltage (usually reported as volts or millivolts), which represents a measure of the potential for electrons to be transferred from one element to another. This potential measure provides an indication of which redox species are the most receptive to gaining electrons at the time of the measurement. Actual times for electron transfer are typically much slower than the time required to complete a redox measurement, with reactions being mediated by specific biological organisms as discussed above. The redox measurement thus indicates the tendency or likelihood that a redox reaction will occur, while the reaction itself may not truly occur to any meaningful extent as the measured redox condition may be only temporary.

Regardless of difficulties with measurement, redox measurements provide critical data for the description and analysis of wet or seasonally saturated soils. Measurements are actually recordings of voltage (Eh) between a reference electrode and a sensor electrode inserted into a soil. The most common sensor electrodes are usually constructed of platinum wire (Pt electrode). The voltage results from an exchange of electrons between a redox couple such as ferrous (Fe^{2+}) and ferric (Fe^{3+}) iron during the process of reduction and oxidation. In soils that have fluctuating wet and dry conditions, wide seasonal fluctuations in Eh values occur (Fiedler et al., 2007).

Although free electrons do not exist in soil, the concept of free electrons—analogous to describing free-proton activity and pH values—is useful as a first step in describing and understanding soil redox status. Similar to pH, "electron activity" may be express as pe (analogous to pH) which is the negative common logarithm of electron activity:

(5) $pe = -log(e^-)$

Values for pe are not measured directly, but derived from Eh through the Nernst equation:

(6) $Eh\ volts = E^o - \frac{RT}{nF}\ ln\frac{reductants}{oxidants} = 0.05916\ pe$

where E^o is the standard electron potential for the half-reaction, R is the molar gas constant (8.31 joules/mole/K), T is absolute temperature (298.15 K); F is the Faraday constant (96,490 coulombs/mole) and n is the number of electrons transferred in the half reaction.

Although the measurement of Eh using a platinum electrode is a simple procedure, it is also subject to measurement uncertainty. Principle issues (Sposito, 1989) include:

- voltage measurement in soil is usually due to more than one-half reaction;
- concentrations of redox species in solution may be too small to provide electron transfer detectable by the Pt electrode;
- several redox species are not readily reactive;
- Pt electrodes are easily contaminated or 'poisoned' by oxides or other coatings; and,
- the salt content of the soil solution (and its ability to conduct electrons) may differ greatly from the calibration buffer.

These uncertainties mean that Eh measurements are considered qualitative measures that provide a general indication of the soil redox status sufficient to differential oxic, suboxic, and anoxic environments (Figure 5.14). The Eh data should not be used to numerically quantify the concentrations of redox species but can be used to determine which redox couples are most active at the time of measurement. Quantitative measurements for redox species should involve direct measurement of the species of interest; for example, dissolved oxygen or ion specific electrode measurement, or multi-species techniques such as ion chromatography.

SUMMARY

- Soil chemistry involves the study of the chemical composition, chemical properties and chemical reactions involving soil solids (organic matter and inorganic minerals), liquids (soil water), gases (soil air), including interactions with plants and soil

organisms. Soil chemical properties have huge implications for global human and environmental health. Through the influence on fate of nutrients and contaminants, soil chemical properties play a role in crop production and food security, biodiversity, water and air pollution, global climate change through greenhouse gas emissions etc. Knowledge of soil chemistry is fundamental in understanding the fate of nutrients and contaminants in the soil system, predicting their environmental impact and finding solutions to mitigate impacts.

- Soil chemical properties are largely influenced by the type and quantity of inorganic minerals present in the soil and the organic matter (humus) content. Inorganic colloids mainly include phyllosilicates (layer silicate minerals), amorphous aluminosilicates and oxyhydroxide minerals.
- The presence of surface charges in both inorganic minerals and humus particles is responsible for ion adsorption and exchange in soils. Most charges are permanent (not dependent on soil pH) and negative, while others vary with pH and may be negative or positive.
- Negative charges on soil particles can attract and adsorb cations, which can be released back to the soil solution through ion exchange. Cation exchange capacity, defined as the sum of total exchangeable cations that a soil can adsorb, is an important soil chemical property that influences the retention of cationic plant nutrients supporting plant growth and retention of cationic pollutants minimizing groundwater pollution.
- Soil pH is a measure of acidity and alkalinity, which controls many of the chemical processes that take place in the soil including availability of plant nutrients. The primary factor influencing the natural soil pH are the geological parent material from which soils are derived. Liming is a practice commonly used to raise the pH of acidic soils.
- Soil salinity gives an indication of the salt content of the soil and is determined by measuring the electrical conductivity of a saturated soil paste extract. Depending on the electrical conductivity and the relative concentration of sodium in soil solution or exchangeable complex, salt-affected soils are classified as saline, saline-sodic and sodic soils. Irrigation with saline waters, particularly in arid climates, is one of the major anthropogenic causes of salinity. Excess salinity in soil can decrease plant available water and affect plant growth.
- Oxidation-reduction (redox) reactions in soil involve the conversion of an element from one valence state to another. Key redox sensitive elements in soil include oxygen (O_2), nitrogen (NO_3^-, NO_2^-, NO_x, N_2O, N_2), manganese (Mn^{4+}, Mn^{2+}), iron (Fe^{3+}, Fe^{2+}), sulphur (SO_4^{2-}, S^{2-}), and carbon (CO_2, CH_4). Redox reactions influence dynamics for certain plant nutrients (eg., nitrogen, sulfur, iron, manganese) and gaseous losses from soils, including greenhouse gas emissions (nitrous oxide and methane).

STUDY QUESTIONS

1. Explain why montmorillonite has a higher shrink-swell capacity than kaolinite.
2. Define isomorphous substitution. Briefly explain how isomorphous substitution causes negative charge development on clay mineral surfaces.
3. Explain why Vertisols in Canada are typically very sticky when wet and form wide cracks when dry.
4. Define cation exchange capacity. Why is soil cation exchange capacity important in protecting and enhancing biodiversity and ecosystem health?
5. Consider the structure and properties of phyllosilicates summarized in Figure 5.5 and answer the following True/False questions.
    - Presence of interlayer hydrogen bonds makes kaolinite a relatively non-expandable clay mineral.True or False
    - 2:1 type clay minerals are composed of one tetrahedral sheet sandwiched between two octahedral sheets.True or False
    - There are no interlayer hydrogen bonds in 2:1 type clay minerals since hydroxyl groups are not exposed.True or False
    - Montmorillonite and illite (hydrous mica) are both 2:1 type phyllosilicates. Montmorillonite freely expands while illite has a limited expandability.True or False
    - Illite (hydrous mica) has a high layer charge due to isomorphous substitution but its cation exchange capacity is relatively low.True or False
6. A soil with a cation exchange capacity of 42 cmol(+) kg^{-1} has the following cations on the exchange complex: Na^+ = 7 cmol(+) kg^{-1} soil, Ca^{2+}= 12 cmol(+) kg^{-1} soil, K^+= 6 cmol(+) kg^{-1} soil, and Mg^{2+}= 8 cmol (+) kg^{-1} soil). Calculate the base saturation of this soil.
7. Explain how a change in soil pH influences the cation and anion exchange capacities of soils.

8. Differentiate between active and reserve acidity.
9. “Soil pH is greatly influenced by the parent material from which soil is formed”. Explain the statement giving examples of parent materials in Canada to support your answer.
10. Briefly explain how soil colloids buffer soil pH changes caused by acidic or basic inputs, such as nitrogen containing fertilizers and liming amendments.
11. Briefly explain how soil pH influences nutrient availability to plants.
12. Define saline, sodic and saline-sodic soils. Outline their major characteristics and effects on ecosystem health.
13. Briefly explain how irrigation under arid climates can influence soil salinity development.
14. A soil with a cation exchange capacity of 20 cmol(+) kg^{-1} has exchangeable Na content of 8 cmol(+) kg^{-1}.
    - Calculate the exchangeable sodium percentage (ESP) of the soil.
    - If the electrical conductivity of the saturated extract of this soil is 5.6 dS m^{-1} (decisiemens per meter), how would you classify the soil? (Hint: See Figure 5.11)
    - How can such a soil be reclaimed?
15. Identify the key redox sensitive elements in the soil. Give one oxidized form and one reduced form for each of these elements.
16. Describe how the redox state of the soil influences nutrient availability and greenhouse gas (nitrous oxide and methane) emissions.

WORKED EXAMPLES

Describe the different mechanisms by which negative charges are developed on surfaces of soil colloids.

Two main mechanisms are responsible for the development of negative charges to surfaces of soil colloids. One is isomorphous substitution in phyllosilicate minerals where central cation in the octahedral or tetrahedral sheet in the mineral is substituted by other cations. The substitution occurs with ions of similar sizes and a negative charge is developed when a higher valent cation is replaced by a lower valent cation. For example, if the Si^{4+} in the tetrahedral sheet is replaced by Al^{3+} (as in illite or hydrous mica), it will give rise to a surface negative charge. Similarly, in the octahedral sheet, Al^{3+} may be substituted by Mg^{2+} which creates a negative charge on the mineral. The negative charges due to isomorphous substitution is permanent and not dependent on pH.

The second mechanism of development of negative charges on soil colloids is the deprotonation (dissociation) reactions depending on pH. Oxyhydroxide clays and the edges of phyllosilicate and amorphous silicate clays have hydroxyl groups which can dissociate at high pH developing a negative charge on surface. Organic colloids (humus) also can carry pH dependent negative charges due to dissociation of functional groups such as carboxyl and phenolic groups. These charges increase with increasing pH.

Explain why the clay mineral montmorillonite has a higher CEC than kaolinite (hint: see Figure 5.2)

Montmorillonite is a 2:1 phyllosilicate clay mineral with 2 silicon tetrahedral sheets enclosing an aluminum octahedral sheet. Isomorphic substitution, e.g. of magnesium (Mg^{2+}) for some of the aluminum (Al^{3+}) in the octahedral sheet results in a permanent negative charge. In addition, montmorillonite clays expand when wetted, as water enters the interlayer space between adjacent tetrahedral sheets. This increases the available surface area for cation adsorption, since in montmorillonite has both external & internal surfaces available for cation adsorption. The combination of net negative charge due to isomorphic substitution and high specific surface area gives montmorillonite a high CEC.

Kaolinite, on the other hand, is a 1:1 phyllosilicate clay mineral, which consists of single silicon tetrahedral sheets alternating with single aluminum octahedral sheets. There is very limited isomorphic substitution, and therefore the kaolinite has a limited number of negative charges. Additionally, being a 1:1 phyllosilicate clay mineral, which does not expand, kaolinite has lower effective surface area on which cations can be adsorbed. Thus kaolinite has a lower CEC.

Fill in the characteristics of the phyllosilicates in the table below.

Mineral	Type	Isomorphous substitution	CEC	Surface area	Expandability
	(1:1, 2:1, 2:1:1)	**(Low/ High)**	**(Low/ High)**	**(Low/ High)**	**(Low/ High)**
Kaolinite	1:1	Low	Low	Low	Low
Montmorillonitee	2:1	High	High	High	High
Illite (hydours mica)	2:1	High	Low	Low	Low
Chlorite	2:1:1	High	Low	Low	Low

Calculate the CEC of the soil

A soil has 3% organic matter, 20% montmorillonite, and 17% kaolinite with the remaining 60% being sand and silt sized primary mineral grains. Calculate the cation exchange capacity of the soil in cmol (+) kg^{-1} (Note: assume cation exchange capacity of organic matter = 200 cmol(+) $kg^{-1,}$ montmorillonite = 100 cmol(+) kg^{-1} and kaolinitie = 10 cmol(+) kg^{-1}).

CEC of the soil = [0.03 x 200 cmol(+) kg^{-1}] + [0.20 x 100 cmol(+) kg^{-1}] + [0.17 x 10 cmol (+) kg^{-1}] = 27.7 cmol(+) kg^{-1}

Identify four major factors that contribute to soil acidity and briefly explain their contribution.

Major factors contributing to soil acidity are acidic parent material, high rainfall, application of ammonium fertilizers, and decay of organic matter. Soils derived from granite-based parent materials, commonly referred to as acidic parent materials are generally acidic in reaction. The natural acidity of rainfall, because of the dissolved carbon dioxide, is a cause of soil acidity. In areas of high rainfall basic cations in soils are leached, increasing the acidity. Application of ammonium fertilizers may also contribute to acidity because the nitrification of ammonium release protons. The decay of organic matter release organic acids contributing to soil acidity.

Name two natural and two anthropogenic causes of soil salinity and briefly describe how they contribute to salinity development.

Two natural causes of soil salinity are the dry climatic conditions and poor drainage. Soluble salts accumulate under dry climatic conditions where rainfall is not sufficient to leach the salts through the soil profile. High evaporation losses from the soils may bring the salts from deeper soil layers with the upward movement of water, which remain at the soil surface.

Under poor drainage conditions (due to high water table or low subsoil permeability), salts are not removed through drainage and leaching, which will result in accumulation of salts on the surface when water evaporates from the soil.

Two anthropogenic causes of salinity are irrigation with saline waters, particularly in dry regions, and overuse of fertilizers. Irrigation water will move downward and dissolve more salts. When the evaporation is high under a dry climate the water moving up will bring dissolved salts. Water will evaporate leaving the salts on the surface. Fertilizers (which are salts containing different nutrients) can increase the salt content of soils if used inappropriately.

Consider the following data for four mineral soils that all have a clay loam texture:

Soil Property	Soil #1	Soil #2	Soil #3	Soil #4
pH	8.8	8.1	7.5	7.6
electrical conductivity (dS/m)	2.7	12.2	1	4.3
base saturation (%)	90	90	60	86
cation exchange capacity	12	11	19	9
(cmolc/kg)				
exchangeable sodium (%) or ESP	17.3	16.8	2.1	2.5
Salinity classification	Sodic	Saline-sodic	Non saline	Saline

A) Determine the salinity classification (i.e., saline, sodic, saline-sodic, or non saline) for each of these 4 soils. [Hint: see Figure 5.12]

B) Which of the above soil(s) would most likely be in a dispersed state? Briefly explain your answer based on the information provided.

Soil #1 is most likely be in dispersed state since it has a high ESP of 17.3% while the salt concentration is low (with EC of 2.7 dS m^{-1}).

C) What are the implications of a dispersed soil for plant growth?

Since individual soil particles are in dispersed state, there is no flocculation and consequently no aggregate formation. The consequence of this is that soil will have poorly developed soil aggregate structural units and be dominated by micropores, which prevent air & water movement and also impede root growth. Thus, plant growth will be impaired.

Identify the common oxidized and reduced forms of the following elements in soils: manganese, iron, sulfur, and nitrogen.

- manganese: Mn^{4+}, Mn^{2+} (reduced)
- iron: $Fe^{3}+$, Fe^{2+}
- sulphur: SO_4^{2-}, S^{2-}
- nitrogen: NO_3^-, NO_2^-, NO_x, N_2O, N_2

REFERENCES

Acton D.F., and Gregorich L.J. 1995. The health of our soils: toward sustainable agriculture in Canada. Agriculture Agri-Food Canada, CDR Unit, Ottawa.

Bowen H.J.M. 1979. Environmental chemistry of the elements. Academic Press, London, New York.

Brady N.C. and Weil, R.R. 2010. The Nature and Properties of Soils 13th Ed. Pearson Education.

Canada Department of Agriculture; Research Branch 1976. Glossary of Terms in Soil Science, Publication 1459.

Eilers W., MacKay R., Graham L. and Lefebvre A. (eds.). 2010. Environmental Sustainability of Canadian Agriculture: Agri-Environmental Indicator Report Series — Report #3. Agriculture and Agri-Food Canada, Ottawa, Ontario.

FAO 2020. Food and Agriculture Organization of the United Nations. FAO Soil Portal. http://www.fao.org/soils-portal/en/

Fiedler S., Vepraskas M.J., and Richardson J.L. 2007. Soil Redox Potential: Importance, Field Measurements and Observations. Advances in Agronomy 94:1-54.

Havlin J.L., Tisdale S.L, Nelson W.L., and Beaton J.D. 2013. Soil Fertility and Fertilizers: An Introduction to Nutrient Management, 8th ed. Upper Saddle River, NJ: Prentice Hall.

Hendershot W.H., Lalande H., and Duquette M. 2008. Soil Reaction and Exchangeable Acidity. Pages 173-178 IN Carter, M.R. and Gregorich, E.G. (eds.) Soil Sampling and Methods of Analysis. 2nd Edition. Canadian Society of Soil Science. CRC Press, Baton Rouge.

Hendershot W.H., Lalande H., and Duquette M. 1993. Soil Reaction and Exchangeable Acidity. Pp. 167-176 in Carter M.R. (ed.) Soil Sampling and Methods of Analysis. 1st Edition. Canadian Society of Soil Science. CRC Press, Baton Rouge.

International Rice Research Institute (IRRI). 1997. Rice Almanac, 2nd Ed. International Rice Research Institute, Manila.

McKenzie R.H. and Woods S.A. 2010. Management of sodic soils in Alberta, Agdex 518-20, Alberta Agriculture. https://www1.agric.gov.ab.ca/$depart...18-20.pdf?Open Element

McBride M.B. 1994. Environmental Chemistry of Soils. Oxford University Press. New York. 406pp.

Miller J.J. and Brierley J.A. 2011. Solonetzic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91(5):889-902.

Munroe J. (ed.) 2018. Soil Fertility Handbook. Publication 611 3rd edition, Ontario Ministry of Agriculture, Food and Rural Affairs. 236pp.

Newman A.C. and Hayes M.H. 1990. Some properties of clays and of other soil colloids and their influences on soils. In: De Boodt M.F, M.H. Hayes, A. Herbillon (Eds). Soil colloids and their associations in aggregates. Springer Science & Business Media; Boston. MA, USA. pp. 39-55.

Qadir M., Quillérou E., Nangia V., Murtaza G., Singh M., Thomas R.J., Drechsel P., and Noble A.D. 2014. Economics of salt-induced land degradation and restoration. In: Natural Resources Forum 38 (4):282-295.

Shreeja D. (n.d.) Phosphate Fixation in Soil: 3. Reactions of Anion Fixation. Available at https://www.soilmanagementindia.com/...-fixation/2252

Sparks D.L. 2003. Environmental Soil Chemistry. Academic Press, London, UK.

Sparks D.L. 2019. Fundamentals of Soil Chemistry. Encyclopedia of Water: Science, Technology, and Society, pp 1-11.

Sposito G. 1989. The Chemistry of Soils. Oxford University Press. New York. 277pp.

Soil Classification Working Group (SCWG), 1998. The Canadian System of Soil Classification. 3rd Edition, Research Branch Agriculture and Agri-Food Canada, Publication 1646.

Soil Survey Staff. 2014. Kellogg Soil Survey Laboratory Methods Manual. Soil Survey Investigations Report No. 42, Version 5.0. R. Burt and Soil Survey Staff (ed.). U.S. Department of Agriculture, Natural Resources Conservation Service. 1001pp.

Ure A.M., and Berrow M.L. 1982. The Elemental Constituents in Soils. Pages 94-204 IN Bowen, H.J.M. (ed.). Environmental Chemistry. Volume 2. Royal Soc. Chem. London.

Van der Perk M. 2006. Soil and Water Contamination from Molecular to Catchment Scale. Taylor and Francis, London 389pp.

Zheng T. 2014. Numerical Analysis of Modeling Concepts for Salt Precipitation and Porosity-Permeability Evolution during Brine Evaporation. Doctoral dissertation, Universität Stuttgart, Germany.

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I obtained a BSc and a MPhil in Agriculture from the University of Peradeniya, Sri Lanka, and a PhD in Soil Science from the University of Manitoba, Canada, specializing in soil physical chemistry. I have over 30 years of teaching experience at the University level and have taught a variety of courses in Soil Science including Introduction to Soil Science, Soil Chemistry, and Soil Fertility as well as courses in Environmental Science. My research interests include assessing, monitoring, and mitigating environmental impacts of fertilizers and animal manure, with special focus on phosphorus transformations and mobility in soils under different climatic and management scenarios. I have authored/coauthored 65 peer-reviewed journal publications.

My favorite memory of working with soil was when I conducted a field research in Manitoba on a very sticky, chernozemic, clay soil on one rainy day. I had to take soil samples from many plots, and as I stepped into the field, about one kilo of soil got stuck onto my boots. With the next step, it almost doubled, and with each step I took, my boots got heavier and heavier until I literally had to drag my feet to go from one plot to another.

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Jim Warren (licensed under a CC-BY-NC-ND license)

I am a P.Geo. and an environmental soil scientist with a diverse background in research, consulting and teaching. I am a graduate of both the University of Guelph (B.Sc. (Agr), Soil Science), and the University of Alberta (M.Sc. Soil Science; Ph.D. Soil Chemistry). I conducted research in agronomy, geochemistry, and mineralogy at the University of Guelph and University of Waterloo before becoming a private consultant in acid mine drainage. I am currently employed as a Land Resource Specialist conducting soil survey and pedology work with the Ontario Ministry of Agriculture Food and Rural Affairs, and am also an Adjunct Professor with the University of Guelph. I have delivered courses in soil science, soil mineralogy, soil chemistry, geochemistry, and pedology as a sessional lecturer at four universities (Alberta, Guelph, Waterloo, and Toronto). I have authored or co-authored over 100 publications in refereed journals; book chapters, conference proceedings and reports.

How did I get interested in soils? Well, I grew up on a farm where, as a teenager, I spent a lot of time on a tractor doing tillage operations and had a lot of time to think. One of the questions I asked myself during long days in the field was: “What makes plants grow?” With that I spent the next several decades trying to understand soils to answer that question. The quest continues . . .

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Graeme Spiers (licensed under a CC-BY-NC-ND license)

I teach in the School of the Environment, the Department of Earth Sciences and the Department of Biology, Laurentian University. I obtained a B.Sc. in Earth Science and Botany at the University of Waikato in New Zealand, and M.Sc. and Ph.D. degrees from the University of Alberta, specializing in Pedology, Clay Mineralogy and Chemistry. With commercial laboratory and consulting experience across Canada, I have been associated with research programs at the Universities of Alberta, Guelph, Waterloo, Laurentian and New Brunswick. My research has resulted in numerous technical and scientific publications and presentations in analytical chemistry, aquatic chemistry, environmental geology, ecology, plant biology and soil science. Prior to attending university on a full-time basis, I was a New Zealand dairy farmer for 10 years, earning my undergraduate degree as a long-term hobby between milkings. In Sudbury, I am involved in research initiatives examining the effects of both historical and current anthropogenic metal emissions on soils, rivers and lakes, as well as vegetation within the Sudbury Smelter Footprint. I am actively involved in a variety of research programs across Canada, and actively collaborates with researchers in Russia (Moscow State University and the Russian Academy of Sciences), Australia, New Zealand, China, Peru and South Africa.

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1.6: Soil Biodiversity and Ecology

Nathan Basiliko, Kari Dunfield, and Mario Tenuta

Learning Objectives

On completion of this chapter, students will be able to:

1. List and describe the major types and groups of biota that live in soils, including their size, distribution, and taxonomy
2. Define the metabolic strategies of soil biota in terms of how they harness energy and nutrients from the environment
3. Describe some of the major challenges and technological breakthroughs in how we study and learn about soil biota
4. Describe how biological communities are integral drivers of organic matter turnover, nutrient cycling, and soil fertility that are covered in detail in Chapters 3 and 7
5. Describe some of the important trophic interactions (i.e., who eats whom) among biotic groups in soils

INTRODUCTION

The mysterious community beneath your feet!

Since Schatz and Waksman's 1943 discovery of the lifesaving antibiotic streptomycin produced by soil bacteria, soil biology has grown to be one of the most exciting and cutting-edge disciplines. Microbiologists have gone on to discover numerous other vital classes of antibiotics produced by soil microbes. Have you ever had a bacterial or fungal infection and been prescribed an antibiotic?- thank soil microbes. But those discoveries only begin to scratch the surface of the value of the biotic communities that make soils their home. Soils are now recognized as arguably the most biodiverse habitat on Earth. The exceedingly complex physical and chemical makeup of soils you encountered in Chapters 3-5 has given rise to vast niche diversity and, thus, the evolution of millions, if not billions, of unique taxa of soil biota. These communities dynamically shape the soil environment, and in the process influence such vitally important ecosystem services ranging from plant and crop growth and health, to water quality, even the regulation of the global climate system through the storage of atmospheric carbon and exchange of atmospheric greenhouse gasses.

Can You Dig It!



Figure $PageIndex1$: Professor Waksman working on soil enrichment cultures with students.

As unlikely as it may sound, a soil scientist won the Nobel Prize in Physiology or Medicine for curing tuberculosis, once one of the world's most devastating respiratory diseases. Ukrainian born Selman Waksman spent most of his career as a professor at Rutgers University in New Jersey (U.S.A.), and was a pioneer in the early 20th century studying soil microbiology and organic matter.

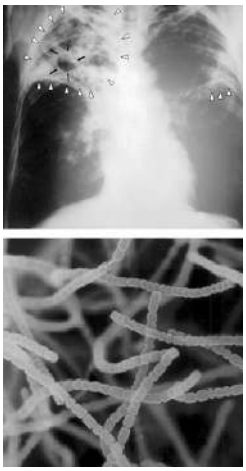


Figure $PageIndex2$: X-ray image of a human lung with severe *Mycobacterium tuberculosis* (top). Micrograph of filamentous *Streptomyces griseus* (bottom; the width of each filament is ~0.5 micrometers) the soil bacterium that produces streptomycin- the first antibiotic to effectively defeat *M. tuberculosis*.

His group was one of the first to tackle the complex chemical nature of soil humus (see Chapter 3), including how soil microorganisms play roles in its formation and decay. Waksman's research group also focused on the unique physiological properties of the fungi and bacteria they were culturing from soils, including that some produced powerful antibiotics.

Ph.D. student Albert Schatz was supervised by Waksman and characterized Waksman's large collection of Streptomyces cultures obtained from local soils and composts for the ability to kill the pathogen Mycobacterium tuberculosis, which led to eventual isolation of streptomycin, the first antibiotic that effectively killed the pathogen without also killing the patient! The story goes that, according to Schatz, Waksman was too afraid to work with the Mycobacterium cultures for which there was no cure at the time, and never ventured to the lab where Schatz was handling the TB-causing microbe. The pair published on the seminal discovery of streptomycin together- and streptomycin was later produced commercially saving countless lives. This discovery was not without controversy, however, as Schatz was initially shut out of any patent royalty payments, and the 1952 Nobel Prize in Physiology or Medicine was awarded solely to Waksman. The apparent rationale was that the difficult task had been obtaining the extensive soil bacteria cultures, while the selection procedures to test for antibiotics were more routine.

Key Terminology

Biological communities are made up of multiple populations interacting with each other. A population refers to all of the same species in a defined area (e.g., the population of humans in Canada; the population of redback salamanders living in the soil of a 1-hectare forest stand; etc.). An ecosystem refers to a biological community dynamically interacting with its abiotic environment; in soils this abiotic environment is the organic and mineral solid components and water and gasses in pore spaces. The concept of species is complicated for many groups of soil microorganisms and small invertebrate fauna. For these, DNA sequence similarity or uniqueness of taxonomic marker genes in relation to sequences in other organisms often defines unique or same taxa (or "species", if only loosely defined).

In this chapter we introduce the range of soil organisms, from sub-micrometer diameter (that is, less than 1/1,000,000 m) unicellular bacteria and archaea to the much-loved (but also occasionally loathed) night-crawler, to vertebrate fauna. What are their strategies for survival and growth, and how do they interact with each other? How do soil microbes and fauna dynamically interact with their physical environment (building on Chapter 4) and with plants? We will explore soil biota as key players driving the cycling of carbon and organic matter you read about in Chapter 3, and the cycling of essential nutrients like nitrogen that you will read more about next in Chapter 7.

An important theme will emerge in this chapter: we presently know very little about the breadth and depth of soil biodiversity, nor do we understand the exact functions of many populations and communities. This is especially true for the soil microorganisms and microarthropods, however, in the past decade, molecular biology tools for detecting and characterizing biodiversity have begun to give us a picture of the millions to billions of "known unknown" organisms dwelling in soils. This makes soil biology an exciting,

breakthrough field; however, we can almost certainly guarantee that if you are reading this chapter 10 years or so from now, at least some of what we now “know” will have changed!

SOIL BIODIVERSITY

Who lives in the soil?

One small handful of healthy surface soil contains more organisms than the whole population of humans on Earth, up to 10 billion individual cells in a single gram! Soil biota range from microscopic unicellular bacteria and archaea that lack nuclei and other organelles to warm-blooded borrowing small mammals like voles and woodchucks, as well as the hidden half- give or take- of plants. The range of sizes of organisms in soil ecosystems spans across nearly ten orders of magnitude! For example, the smallest soil bacterium is ~0.5 micrometers in diameter (0.000005 m), soil-dwelling amphibians and mammals might typically be ~10 cm long (0.1 m), plant roots are typically in the 0.1 to 10 m length range, while certain pathogenic soil forest soil fungi grow thread-like mycelia that extend up to 1 km (1000 m) and beyond! Traditionally soil scientists have lumped soil biota together roughly by some combinations of size and function (see Figure 6.1). Although it is useful to be aware of these groups because the terminology still is in use, here we introduce broad groups of soil biota a bit differently, beginning with the prokaryotes, that are the bacteria and archaea. The term prokaryote refers to single cellular organisms that lack internal membrane-bound organelles (e.g., nucleus, mitochondria, chloroplast), while eukaryote refers to organisms whose cell(s) contain(s) at least a nucleus, and often times other organelles.

Soil scientists and ecologists often use the following groups to describe soil biota:

- **Microflora:** bacteria and archaea (the two domains of life termed prokaryotes, that is, organisms lacking nuclei and other organelles), algae and fungi.
- **Macroflora:** plant roots.
- **Microfauna:** protozoa (single cellular eukaryotes that are motile), small nematodes, and small arthropods.
- **Mesofauna:** metazoans (referring to the animal phylum) with body sizes typically between 0.1 and 2 mm as well as the larger predatory and omnivorous nematodes.
- **Macrofauna:** all metazoan invertebrates and members of the chordate phylum (mammals, amphibians) larger than 2 mm.

However, this scheme can be confusing. For example, whereas the group name **microflora** implies “little plant”, only the algae and cyanobacteria are – like plants – photoautotrophs while many soil bacteria and all soil fungi are heterotrophs. In another example, although certain fungi (yeasts) are unicellular, others form large macroscopic (to massive) fruiting bodies and mycelia (see Figure $PageIndex4$).

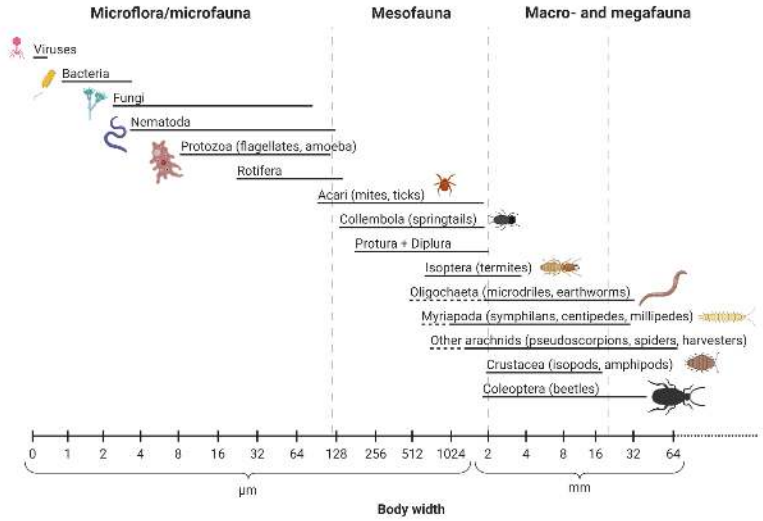


Figure $PageIndex3$: Representation of the main taxonomic groups of soil organisms based on body-size. © K. Dunfield; modified after Swift et al. (1979) and licensed under a CC BY (Attribution) license. Figure created using BioRender.com (www.biorender.com).



Figure $PageIndex4$: Despite fungi sometimes being referred to as “microflora”, branching threads of soil fungi called mycelia **(A)** and their fruiting bodies, including of the conifer tree pathogen *Armillaria ostoyae* **(B)**, are very macroscopic (i.e., visible to the human eye). Indeed a colony of *A. ostoyae* growing in the soil of a Douglas fir forest in Oregon (USA) covers multiple square kilometers and might be the largest organism (by area at least) on Earth! Photo credits: **(A)** Lex vB. (https://commons.wikimedia.org/wiki/F...C3%A9lium).jpg). **(B)** W.J. Pilsak. (https://commons.wikimedia.org/wiki/F...ia_ostoyae.jpg). Images licensed under a CC BY-SA (Attribution ShareAlike) license.

Bacteria and archaea

Being similar in size (typically just a one to a few micrometers in diameter), single cellular, and lacking organelles, it took until the early 1990s before bacteria and archaea were widely recognized as being clearly distinct groups from one another. Although they share some traits, bacteria and archaea diverged from each other evolutionarily a very long time ago. This divergence led to bacteria and archaea having fundamentally different cell envelopes (cell membranes, cell walls, and other outer layers), protein synthesis machinery, mechanisms for packaging their DNA, among many other major physiological differences. In fact, based on DNA sequence analyses, they are as dissimilar to each other as each of them is to higher plants or animals! Carl Woese from the University of Illinois (U.S.A) was the principal scientific force behind the first major redrawing of the tree of life (showing relatedness among all organisms) based on DNA sequences and proposed just three main branches representing the domains Bacteria, Archaea, and Eukarya. Despite their fundamental differences, both bacteria and archaea are small and can interact with micro-scale habitats in soil particles, and both have the ability to evolve quickly to new environments, conditions, and resources with common mechanisms of horizontal gene transfer. This means that they can have some niche overlap in soil environments, and a diverse range of metabolic strategies exists across both domains (Table 6.1), including:

- source of energy (ATP);

- source of molecules to provide electrons to the electron transport chain to produce metabolic energy as ATP; and
- source of carbon to build new organic molecules.

In surface soils, bacteria predominate over archaea in number and biomass. However, there are exceptions where lithotrophic archaea carry out very specific reactions of ammonia oxidation (see Chapter 7) in the nitrogen cycle, or producing methane in flooded, oxygen-free soils. The roles of many other soil archaea are not well known. The interesting strategy of obtaining energy (electrons) from the environment through lithotrophy literally means “eats mineral matter” (see Table 6.1).

Table $PageIndex1$. Common strategies (and related terminology, all ending with “-troph” roughly meaning food) for soil biota to acquire energy for ATP generation and carbon to build biomass. Note that only certain bacteria and archaea are lithotrophs, acquiring their electrons for redox and ATP generation from reduced mineral compounds

Source of energy/electrons	Source of carbon for building biomolecules	Termed:
Sunlight harnessed to split H_2O	CO_2 (inorganic carbon is chemically reduced or “fixed” to organic forms using the Calvin Cycle)	Photoautotroph (phototroph)
Reduced mineral (i.e., non-organic) compounds	CO_2 (inorganic carbon ‘fixed’ to organic forms using the Calvin Cycle or other assimilation pathways	Chemolithoautotroph
Reduced mineral compounds	Organic matter	Chemolithoheterotroph
Organic matter	Organic matter	Chemoorganoheterotroph (heterotrophs)

Many soil bacteria are heterotrophic (see Table 6.1). Most utilize organic carbon molecules that enter the soil ecosystem mainly as plant litters or from plant roots and are already in a chemically-reduced state (that is, carbon is in reduced form as organic molecules as opposed to carbon dioxide). Cyanobacteria that are commonly found in marine and aquatic habitats can also inhabit wet surface soil environments exposed to sunlight. Cyanobacteria are also known to live symbiotically and in mutualistic relationships with a number of plant species. Some lichens on the surface of soils also contain cyanobacteria as the photosynthetic symbiont. Unlike eukaryotic organisms, a good proportion of soil bacteria and some archaea have the ability to carry out cellular respiration in the absence of molecular oxygen (O_2), instead utilizing oxidized forms of other elements like nitrogen (N), sulfur (S), manganese (Mn) and iron (Fe) as the terminal electron acceptor in the production of ATP (the main energy currency in cells). These anaerobic respiration pathways often generate relatively more ATP than fermentation reactions, which certain eukaryotes like yeasts can carry out when O_2 is lacking, but also dramatically influence nutrient budgets in soils. Because of their small size, bacteria and archaea can reside in soil pores as small as 1 micrometer in diameter. Many are attached to the soil matrix, however, some remain **motile**, moving around and between particles via films of soil water.

Fungi

Soil fungi (generally divided into categories based on form as single cellular yeasts and multicellular, filamentous molds) are heterotrophs, more specifically, eukaryotic chemoorganotrophs (see Table 6.1). They play a very important role in the decomposition of woody and other cellulose and lignin-rich plant litters that are low in nutrients like nitrogen (see Chapter 3). Thus, they tend to dominate decomposition processes over bacteria in many forest soils, whereas bacteria may become dominant in more near-neutral pH and alkaline soils. Together, fungi and bacteria are often regarded as the main heterotrophic microbial decomposers in soils. Although fungal cell numbers are generally lower than bacteria in soils, their cell sizes are typically about ten times as large, and in many cases, the contribution to decomposition rates can be about equal between bacteria and fungi in slightly acidic soils. Fungi typically can withstand lower water availability (A_w; Chapter 4), while certain bacteria can outperform fungi in low- and no-oxygen (anoxic, also called anaerobic) conditions.

Certain fungi, called mycorrhizal fungi live symbiotically with most plant species (collectively the fungi and the plant hosts are called mycorrhizae). Specific mycorrhizal fungi are able to colonize the roots of specific plant species and extend their filamentous tufts of hyphae (the larger, often quite-macroscopic structures of hyphae are called mycelia), and at first glance might look like

plant roots themselves (Figure $PageIndex5$). Usually, the association is mutualistic in mycorrhizae, where the plant supplies sugars as sources of energy to the fungal partner, and in turn, the fungus aids the plant in acquiring nutrients from soil, warding off would-be pests or pathogens with antibiotic defense compounds, and helping improve soil structure (see Chapter 4) in the rhizosphere (i.e., the zone near plant roots) through the secretion of sticky protein and polysaccharide compounds. There are different types of mycorrhizal associations, and these are plant-specific, including ectomycorrhizal **fungi** that surround the outer layers of cells of plant roots, and arbuscular mycorrhizae that are generally obligate symbionts of plants (i.e., they can't grow without their plant host) and that extend through cell walls of the root cortex (Figure $PageIndex5$). Mycorrhizal mycelial networks in soils can be pervasive and connect roots from multiple plants- and occasionally between multiple plant species, and facilitate energy (sugars) and nutrient (nitrogen and phosphorus) flow between them! A soil internet of sorts.



Figure $PageIndex5$: Close up photo (**top left**) of ectomycorrhizal mycelium (white) associated with black spruce tree roots (brown). © A.D. Picard (https://commons.wikimedia.org/wiki/F...orhizes-01.jpg). Roots of Douglas Fir (**bottom left**) colonized by ectomycorrhizal fungi giving the appearance of 'stubby' roots © T. Trofymow. The roots feed the mycorrhizal fungi sugars and in turn, the fungi scavenge soil nutrients, provide protection from pests and pathogens, and improve the soil structure in the rhizosphere. The sexual fruiting structures of the fungus is evident on the soil surface as *Lactarius rubrilacteus* mushrooms (**top right**) © T. Trofymow. A root segment of flax is stained to show colonization of arbuscular mycorrhizal fungi (**bottom right**), with hyphae of the fungi evident as thread-like structures running between root cells. Arbuscules are the large deeply coloured blue appendages of the fungi. Arbuscules are in individual root cells, they don't kill the cell but provide phosphorus in exchange for energy from the plant © C. Welsh. Images by André Picard; T. Trofymow, Canadian Forest Service; and Cathy Welsh, University of Manitoba are licensed under a CC BY-SA (Attribution ShareAlike) license.

The life cycles of different soil fungi vary. Yeasts remain unicellular but can exist in either **(1n)** haploid or **(2n)** diploid forms under different conditions and divide through mitosis in either 1n or 2n state. The 1n body form is the main state of fungi, with the 2n state being very small and present only during sexual reproduction (genetic combination) when conditions are poor (e.g., moisture stress) or resources, like readily decomposable plant materials, are scarce. Introducing genetic variations might allow them to cope better with stress. The 2n state undergoes meiosis to return the fungus to the 1n haploid state. Stages of multicellular filamentous mould lifecycles can include fruiting body phases (mushrooms, puffballs) and single cellular spore phases. The majority of fungal growth, however, occurs by mitotic division of the 1n haploid state. Specific fungal species can be saprophytic (degrade soil organic matter and dead plant materials) or pathogenic on plants (cause plant diseases). Some of the most devastating plant diseases are caused by fungi such as rusts of cereals and grasses, and the honey fungus causing white rots of many deciduous and coniferous trees. The width of fungal hyphae means they can occupy soil pores 5 micrometers in diameter and greater.

Water moulds

Water moulds are also called oomycotes and are filamentous (hyphal forming) multicellular eukaryotes more related to the giant kelp of the oceans than fungi! They are called water moulds because they prefer living in water and in water-saturated soil. As they have hyphae, they are often confused with fungi. The normal body state is diploid (2n). They can reproduce asexually (without genetic combination) by mitosis to produce 2n motile zoospores that germinate into hyphae. Zoospores need water-filled pores to

use their flagella for motility. Sexual reproduction occurs by genetic combination of a male and female 2n hyphal strand followed by meiosis to produce a 2n non-motile oospore. Oospores are very thick-walled and can rest in the soil for years and withstand the stress of drying and freezing. Under wet and resource-abundant conditions (e.g., high density of root tips of crop seedlings), oospores germinate to produce many motile 2n zoospores. Species of oomycetes can be saprophytic or plant pathogenic. Plant pathogens can be nasty, such as the late blight pathogen of potato that helped cause the Irish Potato Famine of 1845-1849, Aphanomyces root rot of pulses, and Sudden Oak Death that is devastating oaks in California. Similar to fungi, the width of the hyphae of water moulds means they can occupy soil pores 5 micrometers in diameter and greater.

Green algae

Algae are non-plant, photosynthetic organisms (or "photoautotrophic"; see Table 6.1) whose cells contain organelles (i.e., they are eukaryotes). They are very important primary producers in aquatic ecosystems (e.g., lakes), and in aquatic and marine habitats span from single cellular to large seaweeds. Unicellular algae are found in wet areas on the surface of some soils where they receive sunlight, but typically are in very low abundance, and generally, algae don't represent an important input of newly fixed carbon through photosynthesis to terrestrial ecosystems relative to plants. Green algae reside on wet soil surfaces and surface water-filled pores greater than 50 micrometers in diameter.

An important exception to this is in lichens that can be found in soil biological crusts and colonizing rocks, bare soil surfaces, and standing and fallen woody plant parts. Lichens are interesting cases where two species join together to form an entirely new "species". Lichens represent an important mutualistic symbiosis (i.e., different organisms living together) and have both a "photobiont" and a "mycobiont". In about 90% of lichen "species", the photobiont is an alga, while the mycobiont is always a fungus. In this mutualistic relationship, the photobiont uses sunlight to fix CO_2 into organic matter (i.e., supplying food), while the fungal partner provides most of the shape and physical strength. The fungi in lichens (chemo-organo-heterotrophs; Table 6.1) also obtains water and essential nutrients from the environment (e.g., nitrogen, phosphorus) in the partnership, and provides the physical anchor the keep the lichen in place. In ~10% of lichens, the photobionts are photosynthetic bacteria called Cyanobacteria. Recently, some lichens have been identified as being a tripartite symbiosis between an alga, cyanobacterium, and fungus. Lichens have the ability to live in very harsh environments, and are important early colonizers of barren soils, and even rocks (Figure 6.4)! Lichens that contain a cyanobacterium also have the neat ability to biologicallly fix atmospheric nitrogen gas to ammonia, which is usable by the fungal partner and input of plant-available nitrogen to the soil. Thus, lichens serve many important ecological roles as colonizers providing cover in exposed and new soils such as Regosols (Chapter 8) preventing erosion, building soil through the input of organic matter and plant-available nitrogen to the soil, increasing chemical weathering of mineral surfaces by the production of organic acids, and initiating succession of plants on soils.



Figure $PageIndex6$: The lichen *Xanthoria elegans* growing in cracks of graywacke rock along the shore of Hudson Bay at Churchill, Manitoba. Lichens are often first colonizers of exposed land surfaces and important primary producers and biological nitrogen fixers in those environments. This lichen can survive extremely harsh conditions. In fact it survived being 18 months in space outside the International Space Station. That's one tough organism. © Mario Tenuta, University of Manitoba is licensed under a CC BY (Attribution) license.

Protozoa

Protozoa are single cellular eukaryotic (they have a nucleus and often other organelles) heterotrophic (they eat organic matter; see Table 6.1) microbes that can move about in water (they are motile). Although ranging in size, a typical protozoan in soil might be ten times larger than a soil bacterium, but still microscopic. Broadly, protozoa are grouped by how they move; amoeba shift their cytoskeleton and pull themselves with pseudopods, ciliates like paramecium are covered in short hair-like cilia that bend in unison to propel the cell, and flagellates use one or multiple long whip-like flagella for motility. Although much remains to be learned about the diversity and roles of protozoa in soils, many feed on bacteria, and in some cases fungi, and thus can play important roles in structuring the microbial community through this predation and contribute to the mineralization of carbon (Chapter 3) and nutrients such as nitrogen and phosphorus (Chapter 7) from the bodies of grazed microorganisms or small particles of soil organic matter. Protozoa are typically found in larger numbers in the rhizosphere where both organic matter and bacteria are abundant compared to root-free areas of soils. The populations of protozoa can oscillate with those of bacteria in soil, where populations increase in response to the growth in bacteria such as after rewetting of soil or addition of animal manures to the soil. As swimmers by definition, protozoa typically reside in the films of waters on soil particles and in water-filled pore spaces, and numbers are low in drier soils. Protozoa reside in water-filled pores greater than 50 micrometers in diameter.

Nematodes

Nematodes are multicellular eukaryotic, nonsegmented roundworms. They constitute as much as 90% of the number of multicellular animals in soil, and their numbers often exceed several million per m^2 of A horizon (Chapter 2) soil! They are not worms such as earthworms; those are segmented worms and a very different group of soil organisms. Nematodes vary with species from 0.5 mm to 5 mm in length. Nematodes are long and narrow with either tapered or rounded ends (head and tail). Compared to the organisms we have learned about so far, nematodes are very advanced in the degree of specialization of their body components, having a mouth and esophageal system for feeding, digestive tract, anus for the elimination of wastes, glands and openings for osmoregulation with soil water, male or female reproductive organs, openings and various chemo and tactile sensory structures, and nervous system and rudimentary brain. Nematodes move through water-filled or near water-filled soil pores greater than 50 micrometers in diameter.

Can You Dig It!

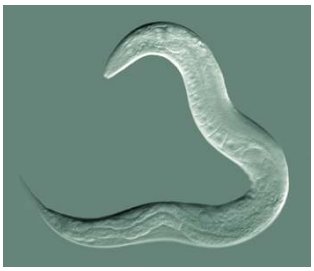


Figure $PageIndex7$: *Caenorhabditis elegans*, adult hermaphrodite © B. Goldstein; licensed under a CC BY-SA (Attribution ShareAlike) license.

Nematodes *are very hardy and fascinating animals. The most famous nematode is Caenorhabditis elegans, which is thought to have been first collected from a stockpile of animal manure in England. This nematode is a bacterial feeder and present in humid temperature soils with a plentiful supply of plant residues. Most C. elegans individuals are hermaphroditic worms, self-fertile females. It can be dried or frozen for decades and still be revived. It has been to space on many missions by NASA, was the first multicellular organism to have its genome sequenced, its cell development from egg to adult fully characterized, and first and only organism to have its nerve cells and nervous system completely mapped out. C. elegans, as scientists call the nematode, has been the subject of many breakthroughs in recent science history, including research leading to three Nobel Prizes, 2002 and 2006 in Physiology or Medicine, and 2008 in Chemistry.*

A nematode species can be assigned to a specific trophic group being plant, bacterial, fungal, algal, omnivorous, or predatory feeding (Figure $PageIndex8$). Plant feeders are herbivores, and some can cause very damaging diseases such as the soybean cyst nematode, root-knot of tomato and potato early dying, but most plant-feeding nematodes do limited damage to plants. Omnivorous nematodes can feed on fungal hyphae, fine roots and hairs, other nematodes and mesofauna. Predatory nematodes feed on nematodes and mesofauna.

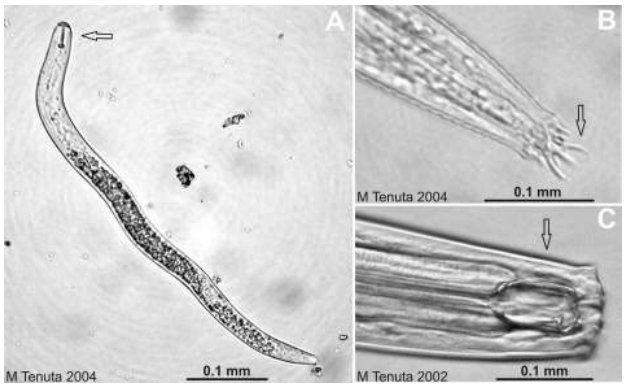


Figure $PageIndex8$: Micrographs of nematodes. It is fairly easy to determine what soil mesofauna feed on based on their feeding structures or 'mouths'. **(A)** a root lesion nematode (*Pratylenchus penetrans*) that is a plant-parasitic nematode with a 'pin-like' spear structure called a stylet (arrow) is protruded into root cells for feeding. **(B)** The head area of a bacterial-feeding nematode of the genus Acrobeles. Bacterial-feeders have small mouth cavities called a stoma where bacteria are ingested. The arrow points to cool appendages called probole of uncertain purpose. They may aid in scraping bacteria biofilms for feeding. **(C)** The head area of a predatory nematode that preys on other nematodes and mesofauna. The arrow points to a large mouth cavity called a buccal cavity. Within the cavity is a spur called a 'stationary tooth' that helps rip the surface of prey allowing their contents to be ingested. © Mario Tenuta, University of Manitoba is licensed under a CC BY (Attribution) license.

The ecological roles of nematodes involve introducing plant fixed carbon to the soil food web, increasing mineralization, particularly that for nitrogen from their food sources, regulating population and community diversity of food webs, and reducing levels of plant pathogenic and plant damaging nematodes through predation.

Nematodes are excellent bioindicators of the effect of management practices on soil health as the presence and population of individual species respond to populations of food sources in the soil food web with some species of nematodes being sensitive to soil management practices (e.g., tillage, fertilizers, pesticides, moisture, rotations, etc.) and stresses (e.g., contaminants such as metals). Many nematodes are parasites of plants. Some nematodes are r-strategists that respond quickly in population to levels of favourable food resources (e.g., bacteria) and soil conditions (e.g., moisture and temperature). Others are K-strategists that thrive in soils with little stress and fluctuating conditions; these are slow-growing, have long life spans for a nematode (one to many years). Then there are nematodes intermediate to r- and K-strategists that are usually most abundant in soil.

Microarthropods

Arthropods have exoskeletons, paired legs, and jointed appendages (e.g., think of insects and spiders). In soils, *microarthropods* are often dominated by two groups: springtails (of the sub-class Collembola, which is often used as a scientific name) and oribatid mites, or Oribadita (Figure $PageIndex9$). Collembola and Oribatida fall in the body-size category of soil "mesofauna" (see Figure $PageIndex3$), ranging from ~0.1 to 2mm in size, and similar to all of the groups of biota described above, are found in soils globally. Typically between 10 and 100 unique species of microarthropods exist throughout the profile of one square meter of soil and on decomposing vegetation aboveground, and microarthropod communities can contain up to 10,000 individuals throughout the profile of a one-m^2 of soil. Soil microarthropods perform important functions through their feeding activities as fungivores (grazing on fungi), herbivores (e.g., grazing on small plant roots), decomposers (grazing on soil organic matter and plant litters), and predators (eating other microarthropod). Similar to protozoa and nematodes, some species of soil microarthropods have highly specialized feeding behaviours (e.g., specialists grazing on very specific types of fungal hyphae), while many of the herbivores and decomposers are generalists. Soil microarthropods can selectively influence the structure of the biological community (e.g., by selectively feeding on certain fungi or other arthropods), distribute microbial cells through the soil profile, and their fecal pellets can promote the formation of aggregates, which together with burrowing activities, improves soil structure (see Chapter 4). Microarthropods can reproduce sexually or asexually, and Collembola are typically thought of as early colonizers of recently disturbed soils (r-strategist) while Oribatida tend to colonize and reproduce slower and maximize numbers in older-growth ecosystems (*K- strategist*).

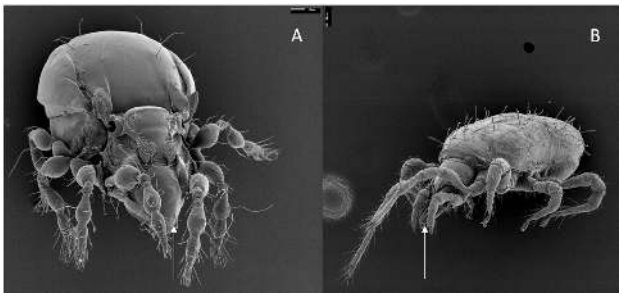


Figure $PageIndex9$: Scanning electron micrographs of a fungal-feeding mite of the genus Quadroppia **(A)** and a predatory mite of the genus Ololaelaps **(B)**. The piercing/chewing structure called a chelicera is shown with arrows for each. For the fungal-feeding mite, the chelicera is very small because fungal hyphae are small themselves. Thus the chelicera is not large enough to be seen here. In contrast, the chelicera of the predatory mite is needs to be larger to pierce the hard bodies of mites it preys on. © Heather Proctor, University of Alberta is licensed under a CC BY (Attribution) license.

Other invertebrates

Soils and associated plant litters are also home to a very large diversity of other larger arthropods (ants, beetles, and others) with many generalist and specialist detritivores, fungivores, herbivores, and even carnivores (e.g., centipedes; scientific name Myriapoda). Annelids (worms; scientific name Oligochaeta) are also important macroscopic invertebrates found in soils, including a variety of earthworms. Earthworms are often revered by farmers and gardeners as being both indicators of healthy soils and contributors to soil health, but these sorts of earthworm paradigms don't tell the full story (Figure $PageIndex10$). On the one hand, earthworms, notably the nightcrawler (*Lumbricus terrestris*) create deep borrows and casts that can improve soil structure, increase pore spaces in compacted soils, and speed up nutrient turnover from plant litters. In intensively managed ecosystems with repeated tillage, harvesting and organic residue management (e.g., crop stover, livestock manure), earthworms arguably do help. However, in unmanaged or less intensively managed ecosystems like long-harvest-rotation temperate and boreal forests, earthworms do more harm than good. There are very few native earthworm species in Canada; the exceptions are on Vancouver Island and in Yukon. The reason is that Canada was nearly entirely glaciated (97%) during the last glacial maximum of the Pleistocene Epoch (ending ~12k years ago), and earthworms have likely only become widespread following the proliferation of resource exploitation by European settlers that involved the movement of soil on tillage equipment and improper management of worms as fishing bait. Particularly in forest ecosystems, earthworm invasion- particularly by *L. terrestris* changes the structure of the soil profile, including largely eliminating organic horizons that are important habitat for plant seedlings, fungal networks, and fauna, including certain birds that nest in the forest floor. Additionally, nutrient retention in mature forest ecosystems is particularly high; long-lived trees producing mainly woody biomass need only small amounts of nutrients like nitrogen and phosphorus to grow each year, and correspondingly, the release of nutrients through decomposition in soils occurs slowly (see Chapter 7). Earthworm invasion leads to very rapid stimulation of microbial decomposer activity and the release of a finite pool of nutrients at a rate too fast for trees to use. It is no surprise that earthworms like *L. terrestris* have earned the title "ecosystem engineer", alongside the likes of beavers, given the severity of the changes they facilitate to soils. For example, Charles Darwin estimated earthworms in a meadowed grass soil in England transported 5 cm of soil from below the surface soil to the surface over ten years!



Figure $PageIndex10$:A sign at Canada's most visited zoo praising earthworms (the burrowing nightcrawler, *Lumbricus terrestris* middle photo) for turning "dirt" into something valuable… hey, should have consulted a soil scientist! Although earthworms can play beneficial roles in heavily-managed agricultural and garden soils, they are non-native in Canada, highly invasive, and have adverse effects in many forest ecosystems. On the right are photos from an experiment showing ingestion leaf and burial activity of a single worm over just a few weeks in an Ontario sugar maple forest floor. Canada's forests and soils are not adapted to such a rapid turnover of organic matter and nutrients. © Nathan Basiliko and Michael McTavish, Laurentian University is licensed under a CC BY (Attribution) license.

Vertebrates

The biomass of vertebrates in soils is quite low compared to smaller biota, however, they can have an important influence in ecosystem functioning by way of altering soil physical and chemical properties (via tunnelling, mixing soils, and their excrement) and by selectively or indiscriminately feeding on other soil biota. A range of mammals (gophers, prairie dogs, groundhogs, etc.), amphibians (e.g., salamanders), and reptiles (skinks, lizards, turtles, etc.) spend at least portions of their lifecycles in burrows or under litter in the surface of soils. Even bird species like the ovenbird spend a portion of their life nesting in the forest soils of Central Canada and the North Central USA. Some vertebrates like the redback salamander are useful indicators of physical (tillage, forestry-related activities) and chemical (e.g., pollution) disturbance because they are reliant on stable habitat and food supply in the form of a robust arthropod community, and because they breathe through their skin.

Viruses

Viruses are smaller than even bacteria (0.05 to 0.2 micrometers), are not organisms and not "alive" in the accepted meaning of the term. However, despite being non-cellular entities, they are composed of biomolecules and play immense roles in disease and genetics in most, if not all, organisms and ecosystems. The simplest viruses are made up solely of genetic material (DNA or RNA) and a protein shell called a capsid. Virus particles are entirely dependent on their hosts- they "hijack" the host's genome with their genetic code and instruct the host's metabolic machinery (i.e., ribosomes for protein synthesis and DNA or RNA synthases for nucleic acid synthesis) to build 10s to 100s of new virus particles. This inevitably leads to cell death, and in the case of single cellular organisms like bacteria or protozoa, kills the host. Viruses that infect microorganisms are also called phages and broadly grouped based on the type of organism they infect (e.g., bacteriophage for bacteria, mycophage for fungi). Some viruses are vectored (transmitted) by soil organisms. For example, some plant-feeding nematodes harbour plant viruses in their esophagus that infect plants with feeding by the nematode. Little is known about the ecological importance of viruses in soil. However, because viruses are often specific to a species of an organism, they may be more diverse than any group of organisms.

Soils can serve as a reservoir for plant and animal viruses that cause disease; however, the role of viruses in ecosystem functioning is much less clear. The number of virus particles per gram of surface horizon soil can be as high as one billion. This is like concentrations seen in surface waters of aquatic ecosystems- and in the past two decades, virus roles in aquatic ecosystem functioning have become well understood. By lysing phytoplankton (cyanobacteria and unicellular algae, the primary producers in aquatic ecosystems) and heterotrophic bacteria, viruses are responsible for shaping biotic communities, and for releasing as much as one-quarter of all carbon fixed into biomass from CO_2 during photosynthesis. Soil viruses may play similar roles, but with much higher biological *richness* (the number of unique species) and much more species *evenness* (i.e., without one or a few dominant taxa) in soils than waters, soil viruses will very likely have their own nuances in how they shape biological communities and ecosystem functioning.

Plant roots

Roots are soil biota, influencing soil development and properties as much as any group of organisms we have explored in this chapter. The biomass of roots is often higher than for all other soil organisms combined in soil that is vegetated. For example, a soil covered in perennial grasses can typically have 10 to 90 t ha^{-1}, compared to fungi (2-5) and bacteria (1 to 2), and up to 2.5, 0.5, 0.2 t ha^{-1} for earthworms, protozoa and nematodes, respectively. The size of roots varies, meaning they occupy a variety of soil pore sizes, handy for obtaining nutrients and water in the soil. Root hairs are about 10 micrometers in diameter, fine roots up to 1 mm, and coarse roots above 1 mm in diameter.

Plant roots are very important in the ecology of soils. They are a major source of organic carbon materials and hence energy for soil heterotrophs and saprotrophs, they lower soil moisture and nutrients levels, consume oxygen through respiration, and alter soil chemical (e.g., soil pH) and physical (e.g., promote aggregation and macropore formation) properties. For more information on soil chemical and physical properties, see Chapter 3, Chapter 4, and Chapter 5.

Distribution of soil biota

Where does the bulk of soil biota reside within soil profiles? The simplest strategy for answering is to “follow the carbon”. Because most of the non-plant soil community are heterotrophic (see Table $PageIndex1$), and reliant on organic matter for both carbon and a source of electrons for generating ATP/energy, numbers of most of the types of soil biota introduced above are highest near the surface of soils and near plant roots in the rhizosphere and Ah horizon of soil (Figure $PageIndex11$). Even lithoautotrophic organisms (Table \(PageIndex{11}\:) tend to cluster around the organic matter inputs and other community members, as they are still dependent on the soil for many of their non-carbon nutrients (e.g., reduced mineral nutrients such as ammonium [NH_4^+] or ammonia [NH_3] for nitrifying bacteria) for growth and metabolism. That said, even in lower abundance, microbes and other heterotrophs alike are still found throughout soil profiles if there are resources available. The importance of microbial biofilms in and around primary soil particles and peds is also worth mentioning. Biofilms form when even distantly related microbes signal to each other to produce sticky extracellular compounds that provide stable habitat (and trophic interactions within the biofilm), and thus allow microbial communities to build their own niches. Although motile soil microorganisms are important- including the protozoans that are defined as being motile, most soil bacteria are attached to soil particles or reside in biofilms.



Figure $PageIndex11$: Fence post showing rot where the wood was in contact with the Ah horizon of soil. The Ah horizon has the most presence of soil organisms and biological activity © Mario Tenuta, University of Manitoba is licensed under a CC BY (Attribution) license.

What about the larger-scale distribution of soil organisms? For example, are the same species and communities found in similar soils on different continents? This is an intriguing question that, in some regards, is not perfectly answered yet. For larger organisms, in the absence of humans facilitating biological invasions by intentionally or unintentionally introducing species, the answer is probably “no”. Think of the nightcrawler example above- there are still “invasion fronts” of these earthworms in forests of Canada and the USA today following their introduction to North America post-colonization by settlers. What about for the very smallest organisms that tend to dominate the non-plant biological activity in soils- the bacteria, archaea, protozoa, and unicellular phases of fungi? Pioneering Dutch environmental microbiologists Martinus Beijerinck in the early 1900s and later Lourens Baas-

Becking are credited with the hypothesis that “Everything is everywhere, but the environment selects”. In essence, there is a global ‘seeding’ of microorganisms given their small size (e.g., could be transported globally on small dust particles in the atmosphere), and they flourish when conditions and resources are right for them, regardless of where those conditions and resources are. Body-size-threshold paradigms have been proposed to explain (lack of) constraints on the natural distribution of biota (including soil biota). However, given that highly-mobile humans have touched nearly every soil environment on Earth, the role of introduced/exotic species is arguably equally as important.

How do we know who lives in the soil?

Studying what we cannot easily see- both because it is buried, but also because it is small- makes fields of soil biology/ecology both challenging and exciting. Analytical artifacts and biases are present in studies of any of the groups of soil biota we introduced above, but on most fronts, soil biologists are rapidly advancing the techniques used to study soil populations, communities, and dynamic interactions. It is surprisingly recently that we have discovered what we don’t know about the scope of soil biology. This perhaps began for soil bacteria with what is called “the great plate count anomaly”. Here, it was noted that the number of cells that could be extracted from soil and viewed/counted using a microscope was orders of magnitude greater than the number of colonies that could be grown on microbiological media in the lab from the same sample (Figure 6.9). The term anomaly was probably too harsh. Extracting and viewing cells from soil is relatively free from selective bias, however, trying to simulate soil conditions and resources in a synthetic medium in the lab is extremely challenging- it is no surprise that only a small fraction of the soil bacterial cells that could be viewed could be grown in the lab. Think back to what you have already learned in Chapter 3, Chapter 4, and Chapter 5 in this book. The chemical, physical, and ecological diversity and complexity in soils are immense at very small scales (i.e., humus molecules with molecular weights in the tens of kg per mole range; a single gram of clay with more surface area than a soccer pitch!). This small-scale complexity has given rise to virtually limitless biological niches that soil microbes and other organisms have evolved to exploit. Just within the bacterial realm a gram of surface soil in a productive ecosystem contains 10s of billions of individual cells from 10s to 100s of thousands of unique taxa. Likewise, the amount of biomass of soil biota is also staggering. A cultivated field can typically have 20 t ha^{-1} of roots, which is about the weight of 25 cows and 15 t ha^{-1} of other soil organisms being the weight of 20 cows! That alone is an overwhelming prospect, and this single section in a single chapter of an introduction to soil science textbook could comprise multiple textbooks on its own. Below we touch on just a few examples of both the challenges and excitement of methods for answering questions on soil biodiversity and ecology.

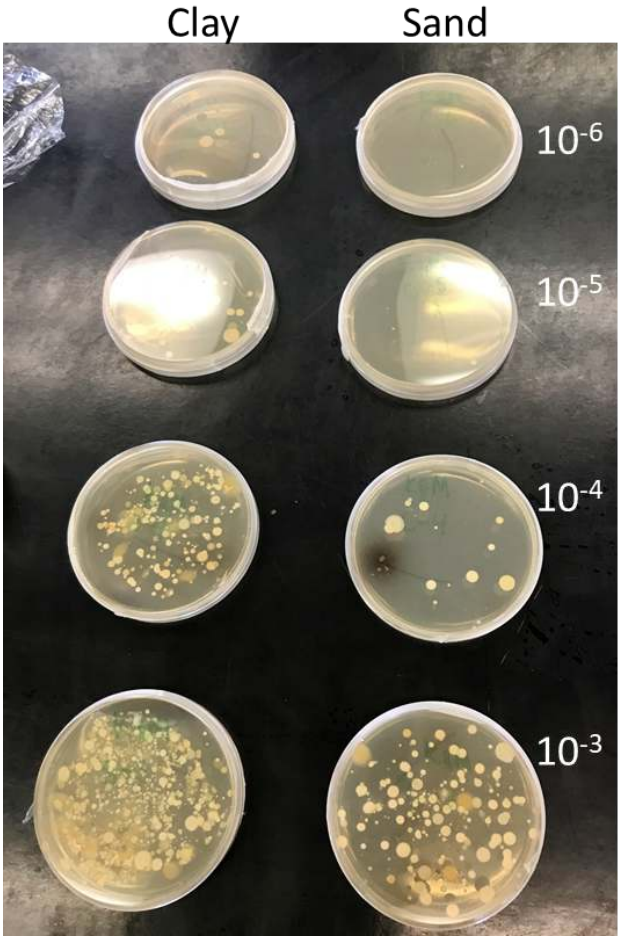


Figure $PageIndex12$: Dilution spread plating of soil suspensions onto agar medium to growth and count population levels of fast-growing aerobic heterotrophic bacteria. Agar plates after six days produced by senior undergraduates for a course are shown. 0.1 mL of 10^{-3} to 10^{-6} dilution suspensions were transferred to plates. Colonies produced are where one bacterial cell from the suspension came to rest and grew. Notice the number of colonies decreases as the dilution increases of the soil suspension. The students were able to estimate the number of bacteria in the soil from the 10^{-4} plate of sand and 10^{-5} plate of clay. They estimated the populations to be 22,000,000 for a gram of the clay soil and 1,800,000 for a gram of the sand soil. This is typical that clay soils have higher microbial populations and activity. Despite that these numbers might appear large, this approach for counting bacteria still drastically underestimates the numbers of bacteria present in soil, because it selects only for the ones that can grow on media in the laboratory. © Mario Tenuta, University of Manitoba is licensed under a CC BY (Attribution) license.

In-situ observations

Depending on the target population or community, different methods exist for studying soil biota in-situ. The soil “contact slide” approach involves inserting a glass slide into soil for a number of days to weeks and then observing attached microorganisms with microscopy. Invertebrates can be observed with pitfall traps, which are buried smooth-surface containers that are open on the top and allow soil fauna to fall in, but not climb out. Often times pitfall traps are partially filled with a preservative like ethanol to kill and prevent decomposition of the trapped fauna. (Find out how to make your own pitfall trap in the You Can Dig It! Activity #1). A particularly interesting and simple technique is commonly used to measure earthworms in soils: mustard powder (the same you would find in the spice section of your local grocery store) is mixed in a weak solution and poured over a defined area of soil. The mustard irritates the worms and they flee their borrows to the surface temporarily, allowing them to be counted, weighed, and identified to species, and there appears to be no long-term negative physiological effects from this contact (see You Can Dig It! Activity #2). Plant roots can be observed and tracked in-situ using clear tubes inserted in soils and a “rhizotron” camera that is inserted periodically in each tube. Image analysis software helps in identifying the same roots over time and quantifying their growth and death. Finally, for groups of soil biota that perform a specific task (often called “guilds”), rates of activities can be

measured by quantifying changes in either reactants or products over time. For example, in upland soils, aerobic methane-oxidizing bacteria consume the trace atmospheric gas methane, an important global greenhouse gas even though its concentration in air is only approximately 0.00002%. Placing a chamber over a defined area of soil and measuring the decrease in methane trapped in the chamber over time gives a reasonable estimate of how active the methane-oxidizing bacteria are in the soil under field conditions.

Isolation approaches

A common approach in soil biology has been to isolate species from their soil environment and community so they can be studied directly (Figure 6.10). For microorganisms, the goal is often to obtain a pure culture (i.e., a population growing on an artificial medium in the lab) where physiological traits can be measured, including what resources are used, what products produced, and how changing conditions like temperature influence growth rates and survival. Much has been learned by cultivating microbes, but the approach is highly selective, and selecting one species from a community of thousands to millions is challenging. In the past few decades, biomarker fingerprinting approaches described below for characterizing soil microbes- particularly those using DNA sequences to identify specific taxa- have helped guide isolation and cultivation approaches towards environmentally relevant organisms and ignore “weeds” that inherently grow well under simple laboratory conditions.



Figure $PageIndex13$: Isolation approaches for studying soil biota. **(Left)** A streak plate approach where microbial cells extracted from soil have been dragged across a 10 cm diameter petri dish with a solid medium in decreasing concentration. Note the isolated colonies that represent single populations. Source: US Centers for Disease Control and Prevention. © J. Gathany. (https://upload.wikimedia.org/wikiped...a_Plate_01.png). Like many mesofauna, nematodes are recovered from soil water suspensions **(Right)** using Baermann funnels filled with water and with soil submerged and wrapped in tissue. Nematodes move out of soil into water and fall by gravity to the end of the rubber tube. © M. Tenuta. For an excellent set of videos with Professor Terry Tollefson from the University of Saskatchewan carrying out soil faunal extractions and visualization, visit https://www.youtube.com/watch?v=TCenRcKbf7U and https://www.youtube.com/watch?v=VuHznslr8aI © US Centers for Disease Control and Prevention (CDC)/ James Gathany; Mario Tenuta, University of Manitoba is licensed under a CC BY (Attribution) license.

Isolation of soil invertebrates differs from the isolation of microorganisms in that individuals are large enough that they can be observed and characterized on their own with microscopy (i.e., growing populations on artificial media is not needed to study them). Nematodes can be extracted with a Baermann Funnel system, where the soil is placed on a cloth mesh in the funnel and saturated. Nematodes are washed downward and through the cloth, while soil particles remain. A common approach for isolating collembolans and mites (groups of soil “mesofauna” see Figure $PageIndex3$:) is the use of a Tulgren (sometimes also called Berlesse) Funnel extraction system. Here a light above a funnel filled with soil is a repellent for small soil fauna and they move downward until they drop through the stem of the funnel and into a container with a liquid preservative like ethanol below. Although individual specimens can be obtained with these approaches, taxonomic identification is still a major challenge and requires a great deal of time and expertise.

Biomarker approaches

Soil biota contains different unique biomolecules that can be extracted and characterized chemically to help study community composition, the relative size of a particular population or functional guild of organisms (e.g., nitrogen-fixing bacteria, see Chapter 7), or metabolic functioning of a soil community. These different biomolecules include cell membrane lipids that vary across broad groups of biota (for example, Archaea have a unique ether-lipid structure of their cytoplasmic membranes), sugar polymers that makeup cell walls (e.g., cellulose in plants and algae versus a compound called chitin in fungi and some invertebrates or a compound called peptidoglycan in bacteria), and very importantly nucleic acids (DNA and RNA). DNA can be extracted relatively easily from soil biota by pulverizing a soil matrix and binding the negatively charged nucleic acid mixture to a silica powder with

high cation exchange capacity where it can be washed and then re-dissolved in pure water. Often taxonomic “marker genes” are targeted for sequencing and quantification (Figure $PageIndex14$: and Figure $PageIndex15$). These are genes or other regions of an organism’s genome that are present in many or all organisms. For example, genes involved in producing a portion of the ribosome (protein synthesis factory present in every known organism) are commonly used to show similarity or dissimilarity in bacteria and archaea (16S rRNA gene), and fungi (18S rRNA gene or Internal Transcribed Spacer (ITS)), while a gene encoding for an oxidase enzyme involved in cellular respiration (Cytochrome Oxidase 1 (CO1) gene) is often used to show the relatedness of animal taxa. It is assumed that a more similar gene sequence between organisms means that they are evolutionarily more closely related. Techniques including the polymerase chain reaction (PCR) and high throughput DNA sequencing have become very accessible in research laboratories, and community fingerprints based on the DNA sequences of millions of individuals in a soil sample (i.e., how many of which individual taxa are in the soil) can be generated relatively cheaply and quickly (Figure $PageIndex14$: and Figure $PageIndex15$). There are other nucleic acid based approaches, for example using quantitative PCR (qPCR) for estimating the number of copies of a particular gene in an original soil DNA extraction- and here genes could be associated with very specific functional guilds, for example like encoding for a portion of the nitrogenase enzyme complex involved in nitrogen fixation. Gene transcripts (i.e., messenger RNA or mRNA) can also be quantified with a variation of qPCR to characterize gene expression related to particular biochemical function. “Shotgun” sequencing of DNA and RNA extracted from soils is also a very powerful tool for characterizing the community and biochemical potential and functioning of soil communities. Referred to as metagenomic or environmental genomic sequencing, random fragments of nucleic acids are sequenced and analyzed, giving a picture of the overall community with less selective bias than PCR-based sequencing approaches.

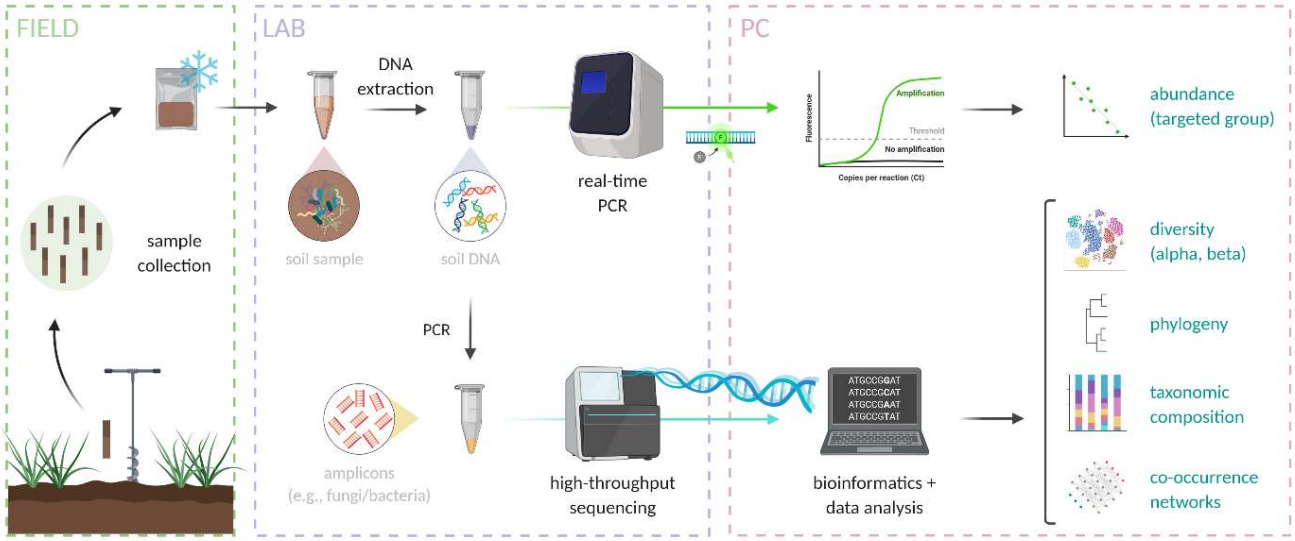


Figure $PageIndex14$: Schematic representation of steps involved in studying microbial communities in soil using DNA approaches. Representative soil samples are taken from the field and immediately stored on ice. Once taken back to the lab, DNA is extracted and stored at -80°C for further analysis. Quantitative PCR can determine the abundance of a target gene, providing an estimate of gene copies per gram of soil. High-thoughput sequencing of PCR amplicons of taxonomic biomarker genes (16S, ITS, CO1) can be combined with bioinformatics approaches to provide a picture of microbial diversity, phylogeny, taxonomic composition and even co-occurrence. Up to date protocols for soil DNA extraction, 16S and ITS sequencing on the Illumina platform, and initial bioinformatics can be found on the Earth microbiome project (EMP) https://earthmicrobiome.org/. © Kari Dunfield, Univ. of Guelph (created using Biorender) is licensed under a CC BY (Attribution) license.



(Figure $PageIndex{15}$: DNA and RNA sequence fingerprinting is common for characterizing soil community diversity and for identifying individual soil specimens or isolates. Gel electrophoresis micrographs **(Left)** showing polymerase chain reaction (PCR) to identify plant-parasitic nematodes. A. Amplification by PCR of nematode DNA shows a positive identification for the nematode, *Ditylenchus weischeri*, in lanes 1, 2, and 3. DNA of nematodes in lanes 4 through 8 do not have the gel band and are thus other species. B. DNA from nematodes in lane 5, 6, and 7 amplified in a different PCR reaction indicating those to be *Ditylenchus dipsaci*. © M. Madani. CC BY. Benchtop high-throughput DNA sequencing system **(Right)** capable of sequencing 1M bp of DNA at the cost of ca. $100 in the year 2020. For comparison, the sequencing cost of the first human genome less than 20 years earlier was over $100,000,000. © Konrad Förstner. (https://upload.wikimedia.org/wikiped..._sequencer.jpg) CCO 1.0 Universal. Images by Mehrdad Madani (University of Manitoba) and Konrad Förstner are licensed under a CC0 (Creative Commons Zero) license.

Although technologies for less selective nucleic acid extraction and quantification and inexpensive sequencing are ever-improving, a major challenge in the use of nucleic-acid sequence-based analyses of soil organisms and communities lies in the fact that very few taxa of soil biota globally have been described in detail (e.g., through isolation and cultivation approaches). For example, there are only approximately 10k identified unique strains of bacteria in culture collections globally, while up to potentially billions of unique bacteria across global soils. Fingerprinting approaches have taught us much about the diversity and ecology of soil biota over the past few decades, but at least at the time this chapter is being written, they have also shown us that there are many "known unknowns" waiting to be discovered!

SOIL ECOLOGY

What do soil biota do?

Ecology describes the interactions within a biological community and the dynamic interplay with the abiotic environment. Soil biota influence the soil environment in many ways that are covered throughout this book. For example, in Chapter 3 you encountered the still somewhat mysterious production and breakdown of soil organic matter that intimately involves soil communities, in Chapter 4 the role of biota in the formation of soil structure was highlighted, and next in Chapter 7 you will learn how key nutrient elements are transformed in soils, in large part a result of biological processes. Here in the present chapter, we take the approach of illustrating soil biological activities from the perspective of the organisms and communities. For example, the soil processes of nitrification and denitrification—where nitrogen is converted from one mineral form to another—are exceptionally important in terms of nutrient availability and loss from soils, and have regional and global importance in terms of how they affect water quality and the composition of the atmosphere. But if we step back and evaluate the directive of the microorganisms involved in these processes, they are- much more earnestly- just harnessing redox gradients in the environment to generate energy, grow, reproduce and successfully stay dormant in between periods of resource availability and proper environmental conditions. Next, we briefly present a number of important soil and related ecosystem functions carried out by soil communities in terms of the biochemical goals of the underlying guilds of organisms driving the processes that have broader ecosystem consequences. This complements the ecosystem-scale perspectives covered in other chapters.

Decomposition, cycling carbon, and methane

With the exception of plant roots, soil biota are predominantly heterotrophic and rely on fixed organic matter as the source of C for building biomass (anabolism) and as a source of energy (see Table 6.1). The initial input of food (i.e., organic matter) in soils is predominantly from plant litters and root turnover, however, organic amendments like manure can also be important large sources in crop and pasture systems. As you learned in Chapter 3, the quality and quantity of different plant litter influence rates of decomposition and formation of soil humus, and an important fraction of soil organic matter appears to be microbial by-products made from processing organic matter, rather than partially broken-down plant structures. In well-aerated soils, simple, and nutrient-rich compounds like sugars, starches, nucleic acids, and polypeptides can easily be mineralized to CO_2 for energy generation and

incorporated in biomass. More complex polymers like cellulose and chitin and aromatic polymers like lignin take longer to be converted back to mineral forms. Carbon dioxide is an important atmospheric greenhouse gas, and the role of soils in the global climate system and as mediators of global climate change is determined in large part by the rate of net carbon inputs to plant biomass minus the rate of release of CO_2 by heterotrophic respiration as plants subsequently die and decompose.

What happens to heterotrophs when soils become saturated and no longer contain O_2 rich air in pore spaces? In many soils, this happens periodically after a rain or snowmelt event; however, wetland soils are defined by prolonged saturation. Decomposition rates are much slower without O_2, and this is one of the factors that contribute to the formation of organic soils (see Organic Soil order in Chapter 8); essentially, rates of plant growth exceed rates of heterotrophic mineralization over long periods of time. In bogs and fens (i.e., peatlands, the predominant ecosystem types with organic soils), soil profiles consisting of partially decomposed wetland plants can extend many meters deep (Figure 6.13)! Without oxygen, heterotrophic soil biota must rely on other biochemical pathways besides aerobic respiration to generate energy. Obligate aerobes only carry out metabolism when O_2 is available, while obligate anaerobes cannot function in the presence of O_2. The term facultative anaerobe (or facultative aerobe) refers to an organism that can generate energy and grow under O_2 and O_2-free conditions. Fermentative organisms carry out glycolysis- generating 1 or 2 ATP molecules per glucose molecule consumed- and then fermentation reactions to regenerate the molecules needed for glycolysis to continue. Fermentations generate reduced compounds like alcohols, organic acids, and H_2 gas as waste products, and typically yield 18 to 36 times less ATP than glucose can provide an aerobe. Anaerobic respiration pathways are used by certain anaerobic soil bacteria and archaea to process plant litters and soil organic matter. In these reactions, compounds other than O_2 are used in cellular respiration as the terminal electron acceptors. Denitrification and sulfate reduction are important examples of anaerobic respiration; denitrification in soils is typically carried out by facultative anaerobes and can yield up to 1/2 of the ATP per unit substrate as aerobic respiration, while sulfate reduction is carried out by strict anaerobes and yields much less energy. Methanogenesis is an important terminal anaerobic decomposition step in wetland soils (See Figure $PageIndex16$:). A group of strictly anaerobic archaea utilize the waste products from fermentative soil bacteria (notably acetic acid and H_2) and produce methane (CH_4). Although energy yields from methanogenesis are relatively small, methane is a globally important greenhouse gas that is much more potent than CO_2 per molecule, and changing methane emissions from wetland soils represent an important soil role in biosphere feedbacks to global climate change. Groups of bacteria called methane-oxidizing bacteria or methanotrophs can consume methane and use it both as a source of carbon and energy. Indeed, methane is a good source of electrons for ATP generation (similarly, it is a good source of heat in natural gas!), but it has high activation energy, and most methane oxidation in soils occurs under aerobic conditions, with O_2 used as an electron acceptor. In wetland soils, methanotrophs reside in surface soil horizons where they can obtain methane produced under anaerobic conditions from below and O_2 from the atmosphere above; they can be an important “filter” that keeps a portion of methane from reaching the atmosphere and contributing to the global greenhouse effect. Interestingly in upland soils, particularly those not disturbed by agriculture, methanotrophs consume atmospheric methane, representing an important global sink for this greenhouse gas, consuming about 5-10% of all methane emitted annually.



Figure $PageIndex16$: A poor fen wetland ecosystem northwest of Sudbury, Ontario. The inset photo on the left shows bubbles of methane-rich gas (noted with a white arrow) escaping from the surface around a metal corer that was produced by anaerobic soil microorganisms. The inset photo on the right is core sample of the upper 50 cm of organic 'peat' soil with clearly intact plant parent material. The stainless steel "Russian" corer that was used to extract the core is also in the photo. The depth of the organic soil/peat deposit in this site is 6 m. © Nathan Basiliko, Laurentian University is licensed under a CC BY (Attribution) license.

Nutrient cycling

The same types of activities by heterotrophic soil biota that break down organic matter and mineralize carbon as CO_2 are also responsible for the release of important mineral forms of nutrients like nitrogen (as ammonium) and phosphorus (as phosphate). Uptake of nutrients for anabolism by non-plant soil biota is often termed immobilization, as the resources are at least temporarily unavailable to support plant growth, whereas plant root uptake of nutrients is termed assimilation. Certain groups of prokaryotes are also responsible for transforming mineral forms of nutrients as well. This includes the very specialized, energy-consuming process of nitrogen fixation where N_2 in the atmosphere is chemically reduced and converted into available N build into amino acids (see below and Chapter 7) and the energy-producing processes of oxidizing reduced mineral forms or reducing oxidized mineral forms of nutrients. In the case of nitrogen, this involves the chemolithotrophic (Table 6.1) process of nitrification where ammonium or ammonia is used as a source of electrons that are typically transferred to O_2 for ATP generation (producing nitrite and nitrate), and the anaerobic heterotrophic process of denitrification where nitrate is used as an electron acceptor during organic matter decomposition. These processes influence the availability of different forms of mineral N explored more in Chapter 7, and ammonium oxidation (the first step in nitrification) releases H^+ ions and lowers soil pH, a consideration for farmers when spreading ammonium-based fertilizer salts. Both processes of nitrification and denitrification are multi-step and mediated by multiple enzymes. The potent greenhouse gas nitrous oxide (N_2O) is an intermediate product and reactant in both. Although this means that soils can be either sinks or sources for atmospheric nitrous oxide, denitrification-related losses of nitrous oxide from fertilized agricultural soils represents one of the most important sources of this gas that contributes to both climate change and degradation of the stratospheric (upper atmosphere) ozone layer. Similar to nitrogen, sulphur and iron in reduced or oxidized forms can be both sources of electrons for chemolithotrophic microorganisms or electron acceptors for anaerobic heterotrophic respiration.

Phosphorus is essential for growth, because it is found in many critical biomolecules including nucleic acids, phospholipids, and ATP, and is often considered the second most limiting nutrient for productivity after N. Low soil P can decrease plant yields, while excess soil P that enters freshwater bodies due to surface runoff can cause a bloom in algal populations called eutrophication. The P cycle is unique because phosphorus does not exist in a gaseous form. Inorganic P present in the soil solution as orthophosphate ions ($H_2PO_4^-$ and HPO_4^{2-}) is the only available form for uptake, but is usually found in very low concentrations (around 1-10 µM), because it is very reactive, and binds to cations and clays. P is primarily recalcitrant in soils in inorganic (sorbed P, primary and secondary minerals) and organic (organic matter, microbial biomass, living biomass) forms. Microorganisms can aid in plant P uptake through several mechanisms:

1. iincreasing root interactions with orthophosphates by extending existing root systems through symbiotic associations with mycorrhizal fungi or stimulating root growth through the release of hormones (phytostimulation) by plant growth promoting

rhizobacteria (PGPR);
2. competing with plants for available P in the soil solution, immobilizing it into microbial biomass, and then regulating supply of P through microbial biomass turnover, and;
3. solubilizing inorganic P through a range of metabolic processes conducted by a wide range of bacteria and fungi. These mechanisms include proton pumps, or organic acids, which solubilize precipitated (solid) forms of P (e.g., Ca-phosphates), siderophores that chelate metal ions associated with complexed forms of soil P, and microbial enzymes such as phosphatases that can hydrolyse organic P or cellulolytic enzymes that mineralize organic matter (see Chapter 7 for an explanation of "P fixation" v available forms of soil P).

Microorganisms with these capabilities can be used to increase the availability of soil P to plants. A fungus, *Penicillum bilaiae* was first isolated by researchers at an Agriculture Canada Research Station in Lethbridge Alberta (Kucey, 1983) and was commercialized as the first phosphate inoculant in Canada, JumpStart™, registered under the Canadian Fertilizers Act, for use on wheat in 1990, and is now registered for most crops grown in Canada.

Pollutant transformation

Soil biota, particularly some bacteria and archaea, play important role in the cycling of metal and organic pollutants. Similarly to iron, certain metals like arsenic can serve as both electron donors and electron acceptors for certain prokaryotic microorganisms depending on the conditions and resources in soils (e.g., concentration of oxygen and available organic matter), and the bioavailability and toxicity of arsenic is altered as its oxidation state is changed through microbial redox processes. Unlike with iron, however, arsenic has no structural role in biomolecules and indeed causes oxidative stress to cells and is severely toxic to certain organ systems in complex organisms. Cadmium is an element that is often present in trace amounts in mined agricultural fertilizers and can accumulate over repeated fertilization to levels of concern for food and feed supply. Unlike with iron and arsenic, prokaryotic microorganisms in soils don't commonly harness reduced or oxidized forms for redox and ATP generation. The redox states of other metals like copper and nickel can be altered by certain soil prokaryotes for both energy generation and for anabolic use or detoxification. Copper, for example, is needed in small amounts for important microbial enzymes related to hydrolysis of organic matter and oxidation of methane by methanotrophic bacteria, while a trace amount of nickel is required for methanogenic archaea to produce methane. Both nickel and copper can be quite toxic in even moderate concentrations though-think that copper sulfate is the key additive in antimicrobial "pressure-treated" lumber, for example. For metals that are trace nutrients, much remains to be learned about how biota walk the line between utilization of and protection from the same elements.

In wetland soils, certain anaerobic bacteria and some archaea influence to the toxicity of mercury (Hg). In its elemental form, mercury is semi-volatile and can be transported globally in warm air masses. Primarily from coal burning, inputs of mercury deposition have more than doubled in many soils. Luckily elemental mercury (like you would find in an old thermometer) is not very bioavailable to organisms in soils, however in wetland soils where there is abundant organic matter and sulfate (e.g., like the peatland in Figure 6.13), anaerobic sulfate-reducing bacteria transform inorganic mercury into a bioavailable organic form called methylmercury. Methylmercury travels from soil water to surface waters where it bioaccumulates in organisms and biomagnifies in aquatic food chains.

Metal pollutant elements like arsenic, cadmium, and mercury cannot be fundamentally created or destroyed by soil biota, however, their toxicity and bioavailability can be changed by soil bacteria by altering redox states, or incorporation into a simple organic molecule like methylmercury. Organic pollutants that are typically synthesized for industrial processes or formed as inadvertent by-products during high-temperature industrial processes, can be fundamentally destroyed. Human-synthesized organic compounds like classes of triazine herbicides introduced in the 1960s and widely used in agriculture today initially were not biologically altered by soil bacteria. However, by the mid-1990s heterotrophic bacteria were isolated that had the ability to grow on the herbicides and break them down to a non-toxic form! Rapid evolution, including through horizontal gene transfer in bacteria, plays an important role in the ability to catalyze the breakdown of human-synthesized pollutant compounds. Soil bacteria certainly can't process all of our organic pollution. In the case of crude oil and other hydrocarbon spills in soil, bacterial enzymes exist to catalyze breakdown; however, they tend not to be very active because hydrocarbon fuel pollutants tend to have very little nitrogen and phosphorus in them that is also needed for bacterial growth.

Plant-microbe interactions

There are many important interactions between plants and microorganisms in soils. The term symbiosis refers to different organisms living together, and although it is sometimes used to imply a mutualistic relationship, symbioses can also involve commensal interactions where only one of the organisms benefits and the other is generally unharmed, and **parasitic** relationships where one organism is harmed and one benefits (e.g., see on pathogens and parasites below). The rhizosphere is an important zone were plants and microbes interact. Heterotrophic decomposers gain energy and carbon from plant root turnover and root exudates, and in turn, mineralize organic forms of nitrogen and phosphorus into inorganic plant-available forms. This is a broad case of generally mutualistic symbiosis, however, when decomposing low-nitrogen-content plant litters like woody tissues, the relationship can shift and microbes can actually immobilize soil nutrients for a period of time to grow on the high carbon, low nitrogen material, leading to plant nutrient deficiencies.

Other plant-microbe relationships are much more direct. Symbiotic nitrogen fixation involves plants "farming" nitrogen-fixing bacteria in root nodules- the bacteria are fed carbon compounds pulled from the plant's Krebs cycle, and in turn, the bacteria use a portion of the energy to convert atmospheric N_2 into a chemically-reduced bioavailable form for the plant that is pushed back into the Krebs cycle and subsequently amino acid synthesis, allowing the plant to produce high-N containing compounds like chlorophyll and rubisco (a key enzyme in photosynthesis). Only certain plants and microbes can form these mutualistic symbioses. Legumes pair with a group of bacteria called rhizobia, while members of some shrubs like alder and sweet gale pair with members of the bacterial phylum Actinobacteria. Nitrogen fixation is exceptionally costly- with up to 16 ATP consumed per single molecule of atmospheric N_2 reduced, and thus symbiotic N fixation is a good adaptation- as plants can readily produce energy-rich compounds through photosynthesis, whereas free-living heterotrophic soil N fixing bacteria cannot. The rates of N fixation between free-living and symbiotic root-associated bacteria are astoundingly different, with input rates of available N to soil jumping from 1 kg N ha^{-1} $year^{-1}$ up to 200 kg!

Another important plant root-microbe symbiosis that was introduced above is mycorrhizae (see Figure 6.3). Many plant species allow colonization of roots by mycorrhizal fungi and supply the fungus with sugars, while the fungus helps the plant acquire nutrients like phosphorus, wards off plant parasites, and improves the soil structure in the rhizosphere. The symbiotic relationship in mycorrhizae is usually mutualistic, and recent research has shown that both plants and fungi have important mechanisms to choose partners who are going to facilitate the other in the relationship. In this regard, a fungus cannot utilize carbon-rich food provided by the plant without also doing its part and helping the plant acquire important limiting nutrients from soil like phosphorus.

Trophic interactions

In soil, the transfer of organic energy and nutrients in the form of molecules in the body of organisms determines the kinds and population levels of organisms in those communities. The concept of the soil food web is used to describe the community of organisms in soil for all or part of their lives. Food webs can depict the transfer of energy or nutrients between organisms within a community, but usually, it focuses on the flow of organic energy. These flows can be extremely complex, governed by the feeding needs of organisms. The steps in energy transfer (feeding of one organism on another) are referred to as trophic interactions.

The basis of most soil food webs is the production of organic energy by autotrophic organisms (largely photosynthesizing plants and referred to as primary producers). This energy is introduced to the soil through the dropping of dead above-ground parts (leaves, stems, trunks, crop residues) and roots. Organic energy and nutrients in dead plant parts and roots are used or released by decay organisms that are heterotrophic and called decomposers. Many species of bacteria and fungi are decomposers. From here on, the trophic interactions are done by heterotrophic organisms collectively called consumers.

Have you ever wondered how large dead plant parts disappear in time in soil? The feeding action of microarthropods that chew and shred those parts into smaller pieces is key. They are called detritivores and their main feeding interest is consuming the decomposers on dead plant parts. Their activity produces smaller dead plant parts, which in turn contains a greater surface area for continued colonization and decay by decomposers.

Living roots are also fed upon directly by a wide range of organisms including many bacteria and fungi, root-feeding nematodes, and microarthropods collectively referred to as herbivores and sometimes primary consumers. Herbivores can be detrimental or beneficial to the health of plants. Some bacteria and fungi, called plant-pathogens, target roots to feed by either killing root parts

before feeding or feed on living cells, slowly draining roots of energy and leading to the eventual death of roots (Figure $PageIndex17$:).



Figure $PageIndex17$: Top: morphology of the soil fungal pathogen of potato, Verticillium dahliae, isolated from Manitoba fields. **(A)** Colony of the fungus after placement of a single spore onto potato dextrose agar 18 days earlier. **(B)** and **(C)** Conidiophore and whorl phialide bearing dispersal cells called conidia. **(D)** and **(E)** Clusters of fungal cells with darkened pigmentation for survival in harsh conditions of soil called Microsclerotia. **(F)** Close up of individual Conidia. © Oscar Molina; licensed under a CC BY (Attribution) license. **(Bottom)** A survival structure also called conidia, of the fungus that cause Early Blight of Tomato and Potato, *Alternaria solani* extracted from a potato field soil in Ontario. Conidia of this fungus has 6-10 cells. They are also pigmented to help survive in soil. © M. Tenuta; licensed under a CC BY (Attribution) license.

Soil fauna that kill parts of roots or compromise the health of plants are referred to as plant-pests (Figure $PageIndex18$). Many root-feeding nematodes and microarthropods are plant-pests. Beneficial herbivores obtain energy from roots and promote the health of those plants and thus called plant growth promoting rhizo-organisms (PGPR). They live in the rhizosphere (in the soil around or on and in roots) to obtain energy from plants. Examples of beneficial herbivores include biological nitrogen-fixing bacteria in the roots of legumes (e.g., bacteria of the order Rhizobiales in which the genera *Bradyrhizobium* forms nodules on soybean roots) and plants of the alder family (ex., bacteria in the genus *Frankia*: Figure 6.16). Examples of beneficial fungal herbivores are ecto- and arbuscular mycorrhizal fungi (Figure $PageIndex5$:).



Figure $PageIndex18$: Many nematodes are of great economic importance. Here roots of potato growing in the Andes Mountains in Peru are parasitized by females of the potato cyst nematode. The females are evident as small tiny white spots on roots. Because of the devastation this nematode causes in reduced potato yield and marketability (tuber surface is warted, blemished and spotted), soil and plants infested with it are legally prevented from entering many countries such as Canada. © Mario Tenuta (University of Manitoba) is licensed under a CC BY (Attribution) license.



Figure $PageIndex19$: Nodule of the nitrogen-fixing bacterium *Frankia alni* attached to the roots of alder (Alnus sp.). Each nodule can contain billions of symbiotic bacteria that benefit from a stable environment and organic matter to consume from the plant; in turn the bacteria fix large amounts of otherwise unavailable atmospheric N_2 for the host plant. © W. Cranshaw. (https://commons.wikimedia.org/wiki/File:Frankia_alni_at_Alnus_sp._(02).jpg); licensed under a CC BY (Attribution) license.

Numerous organisms feed on decomposers, plant-pathogens, and beneficial bacteria and fungi and are called microbial grazers or sometimes secondary consumers. They are considered carnivores as they do not feed directly on living or dead plants. Grazers often specialize to prey on bacteria or fungi, but rarely both. Feeding on bacterial grazers involves taking up the whole organism by the predator. Examples of grazers include amoeba, ciliate and flagellate protozoa, rotifers, bacterial feeding nematodes, and Collembola. Examples of fungal grazers include tardigrades (Figure $PageIndex20$:) and fungal-feeding nematodes, mites, and Collembola.

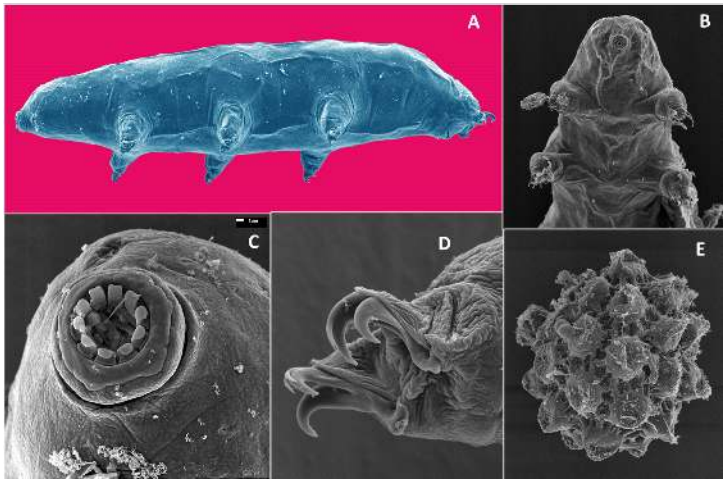


Figure $PageIndex20$: Scanning electron micrographs of tardigrades or water bears. A. Lateral (side) view. B. Ventral (tummy) view with head at top. C. Anterior (head) close-up showing the mouth where the stylet at centre would protrude from to feed. D. Tardigrades have claws at the end of their legs that are helpful to move in soil and perhaps latch onto prey. E. Tardigrade egg that has a heavy surface armour to survive many years in soil. Tardigrades are renown as 'extreme organisms', capable of surviving harsh conditions. Tardigrades that were dried and encased in amber crash landed on the moon in an ill-fated moon mission in 2019. Scientists believe the tardigrades will revive if placed in suitable conditions. © Matthew Boeckner, University of Alberta is licensed under a CC BY (Attribution) license.

Tertiary consumers feed upon microbial grazers and plant-pests. They can be fully carnivorous or omnivorous (feeding on combinations of fungi, soil animals and living roots and other plant parts. Examples of tertiary consumers include omnivorous nematodes, entomopathogenic nematodes (nematodes that kill and feed on microarthropods), nematode-trapping fungi (fungi that immobilize and feed on nematodes), and predatory nematodes and mites.

The trophic interactions of soil food webs can be described as diagrams showing trophic organisms or groups and the direction of energy flow. Soil food webs are analyzed by carefully sampling and processing soil and using methods for recovering and enumerating organisms present. The feeding preference of recovered organisms is either well documented as for those of economic importance such as pathogens and pests, or apparent—based on feeding structures of the organism. For example, predatory mites have a more massive piercing structure (chelicera) than fungal-feeding mites because of differences in prey size.

An example of a soil food web from the prairies of northeastern Colorado is shown in Figure $PageIndex21$: Organisms of similar trophic levels (preferred feed source) are given in boxes with the direction of food web energy flow in trophic interactions (food source/prey to consumer/predator) shown with arrows. The general trophic levels of primary producers, herbivores, decomposers, microbial grazers, and tertiary consumers are given. Note that the apex predator (top carnivore) in this food web are predaceous mites.

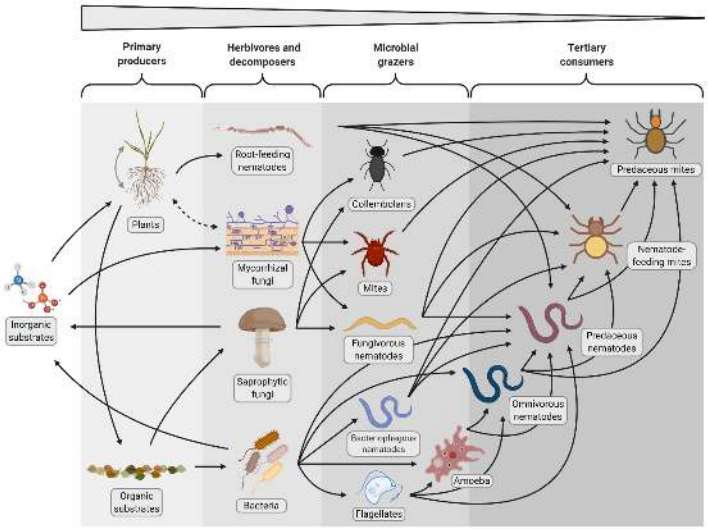


Figure $PageIndex21$: Representation of detrital food web in shortgrass prairie. Organic substrates consist of both labile and resistant substrates. Flows omitted from the figure for the sake of clarity include transfers from every organism to the substrate pools (death) and transfers from every animal to the substrate pools (defecation) and to inorganic N (ammonification). Triangle represents the ecological pyramid principle, demonstrating energy transfer up trophic levels. Note – this site was too dry for earthworms, therefore they are not represented. Adapted from Hunt et al. (1986). Adapted by Kari Dunfield, Univ. of Guelph, using Biorender (Biorender.com), and licensed under a CC BY (Attribution) license.

Primary producers and consumers in a food web have important roles in the soil, such as to balance populations of beneficial and detrimental organisms to plants. Food webs that are out of balance are prone not to promote the growth of plants. Balance in soil food web structure is controlled by bottom-up and top-down processes. Bottom-up processes involve the level of a resource determining the population of consumers in the soil food web. For example, where the levels of microbial and tertiary consumers are related to the density of roots in the soil, bottom-up control is evident (i.e., level of energy input via roots determines populations in higher trophic interactions). Top-down processes are where the levels of the predator determine that of the prey. In agriculture and forestry, top-down control of pest organisms is applied in biological control practices. Here the health of a plant affected by an herbivore is improved by the addition of predator of the later. Examples of biological control agents include entomopathogenic nematodes that kill lawn grubs or predatory mites to control populations of gnats and thrips.

Soil food webs conform to the ecological pyramid principle. Soil food webs usually have much higher amounts of energy in the biomass of primary producers (i.e. the plants), and the amount of energy/biomass decreases sequentially in decomposers, microbial grazers and finally tertiary consumers, whose overall biomass is smallest in a given area of soil. This observation is strong evidence that bottom-up control of soil food webs is important. Figure 6.18 shows that for a typical grassland plant, roots have the largest biomass of all soil organisms, followed by fungi and bacteria (presumably largely decomposers) then microbial grazers. When present in soils, earthworms, feeding on anything in ingested soil, likely derive most of their energy from detritus, bacteria and fungi.

The same reasons shaping the ecological pyramid above-ground is in effect in soil. Each trophic interaction (feeding activity) results in loss of organic energy as heat. This means there is less energy transferred up the food web. Typically, we don't think of organisms in soil as producing heat as mammals do. But if you have ever composted, you know temperatures can get as high as 65°C at the centre of large piles. Compost is an excellent insulator for heat, thus retaining heat produced by decomposers (Figure $PageIndex22$:). Other reasons for an ecological pyramid are predators expend energy seeking prey and predators can live longer than prey.



Figure $PageIndex22$: Decomposers during the active phase of composting (early on in composting when degradation is most rapid) generate great amounts of heat. Here the heat generated is evident in a compost pile of cattle manure where the air temperature on this day was -25°C but the internal temperature of the compost was 44°C (112°F). © Megan Westphal, University of Manitoba is licensed under a CC BY (Attribution) license.

SUMMARY

1. The very large diversity of chemical, physical, and ecological niches in soils have selected for diverse biota spanning from micrometer-sized bacteria to vertebrate fauna; soil biota dynamically responds to and shapes their soil environments.
2. Small size, high numbers, and lacking databases and expertise in taxonomy represent important challenges to our understanding of the scope of diversity in soils- there are many new discoveries to come!
3. Different metabolic strategies are used by soil organisms in terms of how they harness energy and nutrients from the environment; however, heterotrophic activity predominates.
4. These metabolic processes represent biotic directives of generating ATP, surviving, and reproducing, but from the soil ecosystem perspective, the associated resource exploitation essential drives processes such as organic matter and nutrient cycling (influencing soil fertility), among many others.
5. Soil food webs depict trophic interactions between organisms and the flow of organic energy through biota; trophic levels in soil are primary producers, herbivores, decomposers, microbial grazers, and tertiary consumers; and bottom-up control where the amount of organic energy input to soil from plants generally controls the structure (biomass) of soil trophic levels.

SUGGESTED READING

Coleman, D., Callaham, M. and Crossley, D. 2017. Fundamentals of Soil Ecology. 3rd Ed. Academic Press, London, UK.

De Wit, R. and Bouvier, T. 2006. 'Everything is everywhere, but, the environment selects'; what did Baas Becking and Beijerinck really say? Environ. Microbiol. 8: 755-758.

Hunt, H.W., Coleman, D.C., Ingham, E.R., Ingham, R.E., Elliott, E.T., Moore, J.C., Rose, S.L., Reid, C.P.P., and Morley, C.R. 1987. The detrital food web in a shortgrass prairie. Biol. Fertil. Soils 3:57–68.

Kingston, W. 2004. Streptomycin, Schatz v. Waksman, and the Balance of Credit for Discovery. Journal of the History of Medicine and Allied Sciences 59:441-462.

Kucey, R. M. N. 1983. Phosphate-solubilizing bacteria and fungi in various cultivated and virgin Alberta soils. Can. J. Soil Sci. 63:671_678

Madsen, E.L. 2015. Environmental Microbiology: From Genomes to Biogeochemistry. 2nd Edition. Wiley-Blackwell, Hobokon, NJ.

Nair, P. 2012. Woese and Fox: Life, rearranged. P.N.A.S 109: 1019–1021.

Paul, E.A. 2015. Soil Microbiology, Ecology and Biochemistry, 4th Edition. Academic Press, Boston, MA.

Richardson, AE., and Simpson, R.,J. 2011. Soil microorganisms mediating phosphorus availability update on microbial phosphorus. Plant Physiol. 156: 989-996. DOI: 10.1104/pp.111.175448

Schmitt, C.L., and Tatum, M.L. 2008. The Malheur National Forest:Location of the World's Largest Living Organism [The Humongous Fungus]. United States Department of Agriculture- Forest Service. Available from http://www.fs.usda.gov/Internet/FSE_...ev3_033146.pdf [accessed 26 October 2020].

Simard, S., Perry, D., Jones, M., Myrold, D.D., Durall, D.M. and Molina, R. 1997. Net transfer of carbon between ectomycorrhizal tree species in the field. Nature 388:579–582

Swift, M.J., Health, O.W., Anderson, J.M. 1979. Decomposition in Terrestrial Ecosystems. University of California Press, Berkley, CA.

STUDY QUESTIONS

1) Why is the diversity of soil bacteria so high? What roles do soil properties you learned about earlier in the text and trophic interactions among soil biota play in this high level of diversity?

2) What are biomarker and isolation based approaches to studying soil biological communities? Why are biomarker approaches particularly useful for characterizing the diversity of small-sized biota in soils?

3) Give one example each and describe the nature of the relationship in a mutualistic and a parasitic symbiotic relationship between soil organisms.

4) Where is biomass typically largest in the soil profile and why? How does this relate to the dominant metabolic strategy (e.g., heterotrophy, lithotrophy, autotrophy) in belowground ecosystems?

5) Imagine a sterile soil entirely devoid of biota. Could this soil support plant group? Why or why not?

6) Describe differences in soil biota in a flooded (anoxic) wetland soil relative to a well-drained upland soil.

7) How does soil biota influence the cycling of carbon in both soils and the atmosphere?

8) Draw a hypothetical soil food web.

Can You Dig It!

Activities

Activity 1: Make a Pitfall trap to catch Soil Invertebrates and Insects

Adapted from: https://www.nhm.ac.uk/discover/how-to-make-pitfall-trap-to-catch-insects.html

You will need:

- a hand trowel
- a small plastic container or plastic cup
- a tray
- ID guides (optional)

1. Choose a location for your trap on flat ground near vegetation.
2. Use a trowel to dig a small hole.
3. Place a clean plastic container in the hole. Fill in any empty space around the container with soil. Make sure that the top of the pot is level with the ground, or you won't catch anything.
4. Leave your trap overnight.
5. Empty the trap into a tray to see what creatures wandered in. Use ID guides or apps such as iNaturalist supported by the Canadian Wildlife Federation and the Royal Ontario Museum Biodiversity Program https://inaturalist.ca/) to help you identify what kind of invertebrates they are.
6. Record your findings: take pictures, make a note of what you caught, the date and location.
7. Carefully release the creatures, returning them to a safe, sheltered place.
8. Return the area back to how you found it.

Adapted from: Christina Curell, Michigan State University Extension – June 29, 2016

https://www.canr.msu.edu/news/earthworms_can_be_an_indicator_of_soil_health

Materials that you will need to measure worms in the field:

- Tape measure
- 2 liters of tap water
- Hand trowel or shovel
- Container to collect worms
- Solution of 2 tablespoons of mustard powder dissolved in 2 liters of water

Once you have your materials gathered you are ready to count earthworms.

Step 1: Measure a square foot in the test area and dig down 12-inches.

Step 2: Collect and count the number of worms found. If possible, differentiate worms by type. For example, label as earthworms, red worms, etc.

Step 3: (Optional) Level out the bottom of the hole, and pour the mustard solution slowly. Deep burrowing worms should come to the surface within 5 minutes. Collect and count the worms that come to the surface.

Step 4: Count and record the total number of earthworms that are collected.

About the authors:

Nathan Basiliko, Professor and Canada Research Chair in Environmental Microbiology, Department of Biology, Laurentian University



Nathan Basiliko (licensed under a CC-BY-NC-ND license)

My research group focuses nearly entirely on soil biology and the roles in biogeochemical cycles. However, because my academic “home” for the past 7 years at Laurentian University has been in our Biology department, I don’t teach soil science courses anymore (I used to). I miss teaching introductory soil science in particular, but my current position does give me a new opportunity to squarely build soil science in to other courses- including introductory and advanced microbiology and ecosystem ecology courses. I hope that this has some broader impact by engaging students with soil science who otherwise would not take soil science courses.

Kari Dunfield, Professor in Applied Soil Ecology and Canada Research Chair in Environmental Microbiology of Agro-Ecosystems, School of Environmental Sciences, University of Guelph



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Kari is a Professor in Applied Soil Ecology and Canada Research Chair in Environmental Microbiology of Agro-Ecosystems in the School of Environmental Sciences at the University of Guelph (Guelph, Ontario). Her research group works at the intersection of microbiology, ecology, and soil science, and takes an inter-disciplinary approach to find solutions to complex global environmental challenges, including, recycling nutrients and minimizing greenhouse gas emissions in agriculture systems; breaking down pollutants in brownfield sites; and understanding pathogen transport from agricultural systems into waterways. I love studying soil microbiology because it is a combination of field and lab based research. In my group you need to own a pair of steel toed boots and a lab coat, because in the morning you could be going out to the field, using soil probes and shovels to get soil and root samples, and in the afternoon you could be back in the lab doing delicate and sterile molecular biology techniques.

Mario Tenuta, Professor of Applied Soil Ecology and Senior Industrial Research Chair in 4R Nutrient Stewardship, Department of Soil Science, University of Manitoba



Mario Tenuta (licensed under a CC-BY-NC-ND license)

As a Soil Ecologist, the work is very diverse and full of new challenges and learning opportunities. My research involves tackling the issue of lowering the amount of greenhouse gases emitted from agricultural soils. For this we do field research to understand how microbes are involved in nutrient cycling, how soil conditions affect microbial activities and functions and how farmer practices of fertilizer and manure use affect microbes. There is lots of opportunity for laboratory analysis of microbes, nutrients in soil, nutrients in plants, and of crop productivity. We take that understanding to recommend practices farmers can use to lower emissions of important greenhouse gases such as nitrous oxide (N_2O). Soil is everywhere and the work also provides lots of opportunity to travel locally and the world doing research, teaching and development

The goal of inviting this team of authors was to ensure that a range of ecosystems (including forest, prairie, wetland, agricultural, permafrost-affected) and groups of soil biota was covered within their combined expertise. All three of the co-authors are active researchers in the field of soil biology and have each held or currently hold Canada Research Chair positions at their universities.

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1.7: Soil Nutrient Cycling

Joann Whalen, Noura Ziadi, Jeff Schoenau, Maxime C. Paré, David Burton, and Tom Bruulsema

Learning Objectives

On completion of this chapter, students will be able to:

1. Know the forms of soil nutrients that can be acquired by plants
2. Know the factors controlling nutrient transformations into plant-available forms
3. Understand how to evaluate soil fertility with soil testing and plant analysis
4. Understand that inorganic and organic fertilizers can be applied to replenish the soil nutrient supply
5. Be familiar with the 4R Nutrient Stewardship approach to soil nutrient management
6. Be aware of the environmental implications of soil nutrient management
7. Be acquainted with some examples of soil fertility improvement in Canada

INTRODUCTION

Soils support plant growth. Seeds germinate in the warm, moist soil environment. The seedling emerges at the soil surface, which needs to be porous and uncompacted, with no surface crust. Elongation of the radical and development of the root system occurs easily in soil with a crumbly, loose structure. The root system has two vital functions. The first is anchorage. Many plants grow a thick, strong primary root that is approximately as tall as the stem. A plant that is firmly rooted in the soil is unlikely to be dislodged by normal rainfall events and windstorms. The second vital function is resource acquisition, since most of the water and many essential elements for plant growth are absorbed through the root hairs (Table 7.1).

Table 7.1. Essential plant elements, and the forms of nutritive cations, anions and molecules absorbed by plant roots (adapted from Havlin et al., 2014)

Essential element	Concentration	Nutritive form(s) absorbed		
	(μg g^{-1} dry mass)	Cation	Anion	Molecule
C	450000			CO_2 [1]
H	450000			H_2O [2]
O	60000			CO_2, H_2O
N	15000	NH_4^+	NO_3^-	N_2 [3]
K	10000	K^+		
Ca	5000	Ca^{2+}		
P	2000		HPO_4^{2-}, $H_2PO_4^-$	
S	2000		SO_4^{2-}	SO_2 [4]
Mg	1000	Mg^{2+}		
Fe	100	Fe^{2+}, Fe^{3+}		Soluble chelate [5]
Cl	100		Cl^-	
Mn	50	Mn^{2+}		Soluble chelate
B	20		$H_2BO_3^-$	H_3BO_3, soluble chelate
Zn	20	Zn^{2+}		Soluble chelate
Cu	6	Cu^{2+}		Soluble chelate
Mo	0.1		MoO_4^-	Soluble chelate
Ni	0.01	Ni^{2+}		Soluble chelate

[1] CO_2 gas is absorbed through leaves and converted to carbohydrates $(CH_2O)_n$ via photosynthesis.
[2] H_2O, as water vapour or liquid water, is absorbed through foliage and roots.
[3] Symbiotic and associative N_2-fixing prokaryotes that grow with the plant transform N_2 gas into NH_4^+ via biological nitrogen fixation, and then transported into plant cells.
[4] Less than 10% of the plant's sulphur requirement comes from SO_2 gas absorbed through leaves.
[5] Soluble chelates are readily absorbed by the root.

Soil fertility refers to the ability of soil to supply nutrient ions for plant growth. Soils have inherent fertility, which depends upon four factors:

1. parent material, which is weathered to release nutrients;
2. vegetation, which produces organic residues that are decomposed to release nutrients;
3. organisms that participate in weathering and are responsible for decomposition and nutrient cycling; and
4. climate, which determines the rates of biological, chemical and physical processes that contribute to or deplete the soil nutrient supply.

Inherent soil fertility is a good predictor of what plant species will grow in a given area under natural conditions, without human intervention. When we harvest agricultural crops to feed people and livestock, we remove some nutrients from the soil-plant system. Consequently, agroecosystems are usually fertilized to replenish or increase the soil nutrient supply. Selecting the right fertilizer and other soil amendments requires careful decision, since it is expensive to purchase and apply these materials to

agricultural land, and because overuse can have negative environmental consequences. Farms across Canada can follow nutrient management plans to guide the application of fertilizing materials to agricultural soils, for the best agronomic, economic and environmental outcomes.

NUTRIENTS IN THE SOIL-PLANT SYSTEM

Fine roots and root hairs grow in soil pores, which contain air and water. Soil pore water is a dilute electrolyte solution containing water-soluble nutrient ions. To satisfy their water requirements, plants absorb water from soil via osmosis and rely on aquaporins to selectively conduct water molecules in and out of cells, which regulates the cell volume and internal osmotic pressure. Water-soluble ions in the soil pore water enter the root hairs and diffuse through the epidermis. However, the root surface must maintain electrical neutrality. Any ion that enters the root causes the release of a counter-ion such as H^+, OH^- or HCO_3^- (Figure 7.1). For example, when a positively charged ion such as potassium (K^+, a cation) passes through the root epidermis, the root releases one H^+ ion into the surrounding soil pore water, and absorption of a calcium ion (Ca^{2+}) results in the release of 2 H^+ into the soil pore water. Anion absorption is similarly balanced, such that uptake of nitrate (NO_3^-) causes the root to release an OH^- ion or a HCO_3^- ion into the soil pore water.

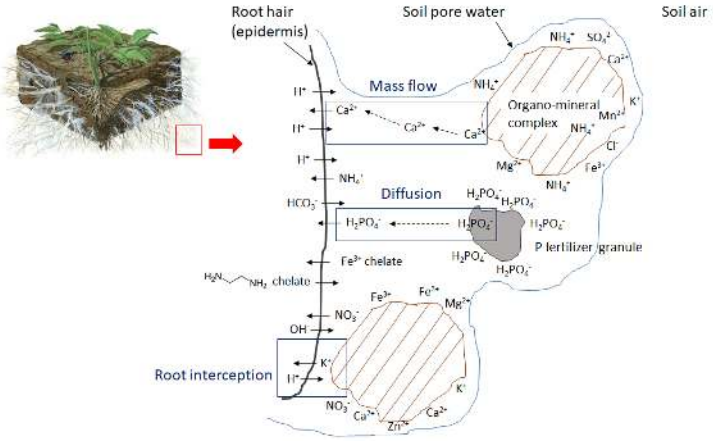


Figure 7.1. Nutrient ions move through the soil pore water and are absorbed by root hairs resulting in the release of counter-ions (H^+, OH^-, HCO_3^-) to maintain electrical neutrality at the root surface. Ions that travel along with water are absorbed by mass flow. Diffusion involves ion movement across a diffusion gradient, from high concentration to low concentration, near the root. Roots can intercept ions retained in organo-mineral complexes. Roots release chelates that bind to metals ions, increasing metal mobility and uptake by the root. © Joann Whalen is licensed under a CC BY (Attribution) license.

Once inside the root, water-soluble ions can move through the apoplast of the cortex, a transport pathway that transports water and dissolved substances along cell walls and extracellular space outside of plasma membranes. There are two other pathways for water-soluble ion movement: the symplastic and transmembrane pathways. In the symplastic pathway, water and solutes move along the cytosol and cross the plasma membranes of adjacent cells via the plasmodesmata. Alternatively, in the transmembrane route, the dissolved minerals and water move from cell to cell by crossing the cell wall, exiting one cell before they enter the next cell. The suberized layer surrounding the endodermis acts as a physical barrier, and ions must be transported through the plasma membrane via protein channels by binding with transporter proteins. Some of the transporter proteins responsible for nutrient ion flux in plants are listed in Table 7.2. Next, nutrient ions enters the xylem and are transported along with water to the stem, leaves, flowers, seeds and other components. Nutrient ions may be incorporated into structural compounds within these plant organs, or support metabolic processes in the cytoplasm. Furthermore, nutrient ions may be remobilized and translocated to other tissues or organs if they are needed in another part of the plant.

Table 7.2. Transporter proteins mediate the flux of nutrient ions across the plasma membranes of vascular plants. Adapted from Reid and Hayes (2003)

Nutrient ion	Transporter protein	Comments
NH_4^+	High affinity AMT transporter proteins	
NO_3^-	Nitrate transporter protein (NRT2) family	
$H_2PO_4^-$ and HPO_4^{2-}	PHT1 transporter protein	PHT1;1 through PHT1;9 were identified in *Arabidopsis thaliana*
K^+	K^+ transporter protein (Trk) family	
SO_4^{2-}	SO_4^{2-} transporter proteins	Plants also possess the cysteine (Cys) transporter, methionine (Met) transporter, and glucosinolate (GSL) transporter
Ca^{2+}	Ca^{2+} transporters, Ca^{2+}-ATPases and H^+/Ca^{2+} antiporters	Transporter proteins that move Ca^{2+} across the membrane, against the electrochemical gradient, are specifically referred to as an ion pump.

Plant uptake depletes the nutrient ion concentration in soil. Ions may move together with water through mass flow or be intercepted by growing roots that make contact with exchangeable ions on organo-mineral surfaces, a process referred to as root interception (Figure 7.1). Ions may also move across diffusion gradients from a zone of high concentration to a zone of low concentration in pore water close to the root (Figure 7.1). The concentration of nutrient ions in soil pore water is also affected by multiple, concurrent biological, chemical and physical processes, illustrated in Figure 7.2.

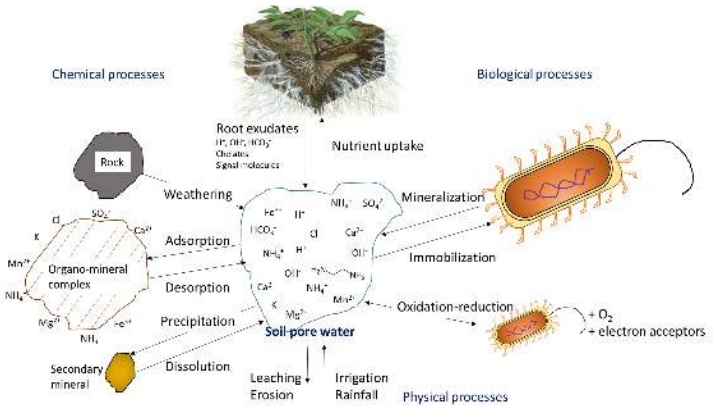


Figure 7.2. Biological, chemical and physical processes that affect the concentration of nutrient ions in the soil pore water. © Joann Whalen is licensed under a CC BY (Attribution) license.

Biological activity is responsible for the mineralization of organic compounds, which releases water-soluble nutrient ions. Mineralization requires a specific hydrolytic enzyme, produced by plant roots or soil microorganisms, to cleave the covalent bonds in an organic compound. Soil enzymes function in the presence of water, often outside of biological cells. Mineralization of soil organic matter and organic residues releases ammonium (NH_4^+), phosphate (HPO_4^{2-}, $H_2PO_4^-$) and sulfate (SO_4^{2-}) into the soil pore water. Immobilization refers to absorption of nutrient ions by organisms other than plants. Ions are immobilized in soil microbial biomass for a limited period of time (days to weeks) while the microorganism grows and reproduces, and are released at the end of their lifespan. The mineralization-immobilization-turnover concept applies to any nutrient that is covalently bound in organic matter, including carbon, nitrogen, phosphorus and sulphur. An illustration of the concept (Figure 7.3) shows how microorganisms produce enzymes to hydrolyze proteins, releasing NH_4^+ in the soil pore water, which is then immobilized inside growing microbial cells, and finally re-released into the soil pore water upon lysis of those cells.

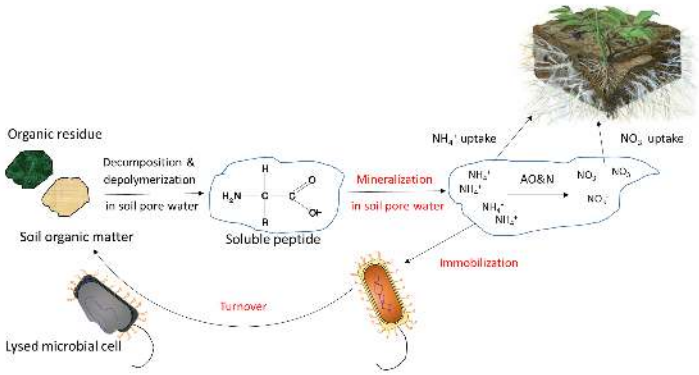


Figure 7.3. The mineralization-immobilization-turnover process. Organic residues, soil organic matter and lysed microbial cells undergo decomposition, which involves physical fragmentation and solubilisation before organic compounds are depolymerized by extracellular enzymes of microbial origin. Monomers such as the illustrated peptide then undergo further enzymatic degradation in the soil pore water, a process referred to as mineralization. The ammonium (NH_4^+) released into soil pore water may be taken up by plants, immobilized in microbial cells, or undergo microbial-mediated transformations called ammonia oxidation and nitrification (AO&N) that produce nitrate (NO_3^-). Similarly, NO_3^- is available for uptake by plants and microorganisms. The NH_4^+ and NO_3^- used by microorganisms for protein synthesis and other metabolic functions will be released when microbial cells are lysed, at the end of their lifespan. © Joann Whalen is licensed under a CC BY (Attribution) license.

The oxygen concentration in soil pore water affects the transformation of nutrient ions. Oxygen moves approximately 1000 times faster through air-filled pores than through soil water, so it is not uncommon for oxygen limitation to occur in soil pore water. Limited oxygen availability can stimulate denitrification, a biological reaction in which the nitrogen electron acceptors gain electrons and decrease their oxidation state, e.g.,

$$NO_3^-{}_{(aq)} \rightarrow NO_2^-{}_{(aq)} \rightarrow NO_{x(g)} \rightarrow N_2O_{(g)} \rightarrow N_{2(g)}$$

which also changes the chemical form of nitrogen from water-soluble NO_3^- to dinitrogen (N_2) gas. Chemical reduction of elements such as Mn, Fe, and sulfate in anoxic (low-oxygen) soil is another process that alters the chemical speciation and quantities of these nutrient ions in soil pore water.

Chemical reactions that add nutrient ions to soil pore water include weathering, which occurs when the broken, exposed surfaces of mineral surfaces are exposed to acidic compounds, such as carbonic acid. Dissolution refers to the solubilisation of nutrient ions from clays and colloidal materials in the presence of water. In dry soils, dissolution is less common because a low volume of soil pore water will increase the proximity of nutrient ions with reactive ions and molecules that can form precipitates. Surface complexation reactions on organo-mineral surfaces can either increase the concentration of nutrient ions, through desorption reactions, or remove ions from the soil solution through adsorption reactions.

The concentration of nutrient ions is highly dependent on the water content in soil pores. Irrigation and rainfall add water to the soil pores, thereby diluting the water-soluble nutrient ions. In humid regions that have more precipitation than evapotranspiration, soil pores may periodically reach 100% water-filled pore space. This will limit oxygen diffusion, which affects oxidation-reduction reactions, and impedes the growth of most plants, except those with morphological adaptations that allow them to tolerate inundation. Agricultural activities generally occur on land that has good natural drainage or where drainage systems are installed to encourage water movement below the cultivable topsoil. Drainage removes water, but it also leaches water-soluble nutrient ions through the profile, along with the water. Removal of excess water from topsoil via drainage or through evaporation will increase the relative concentration of water-soluble ions remaining in soil pore water.

SOIL FERTILITY AND PLANT NUTRITION

Plants acquire most of their essential nutrients ions from the soil pore water, a dynamic environment. In agriculture, much effort is made to understand the soil nutrient supply. Does a particular soil have enough nutrients to support the plant needs, or do we need to add fertilizer containing nutrients to avoid a plant nutrient deficiency?

Soil fertility testing is an accurate way to assess the nutrient supply in agricultural soil. The main purpose of soil fertility testing is to determine whether a soil contains sufficient nutrients to meet the needs of an annual crop during its lifespan (one growing season). In addition, soil fertility testing can determine whether there is an adequate nutrient supply for perennial crops during their

establishment and production phases, which span multiple growing seasons. Soil fertility testing relies on chemical and biological analyses by certified laboratories. The results of these tests are a good indicator of the plant-available nutrient supply. This concept relies on an understanding of two facts. First, the plant will acquire nutrient ions from the soil pore water throughout the growing season according to its nutrient demands. Second, the soil pore water will be replenished with water-soluble nutrient ions at a rate that reflects the combined biological, chemical and physical processes that contribute and remove ions from the soil pore water, during the growing season. For this reason, soil testing methods such as the Mehlich 3 extraction remove nutrient ions from soil pore water and other sources (Figure 7.4), a good approximation of the nutrient supply that will become available to plants during the growing season.

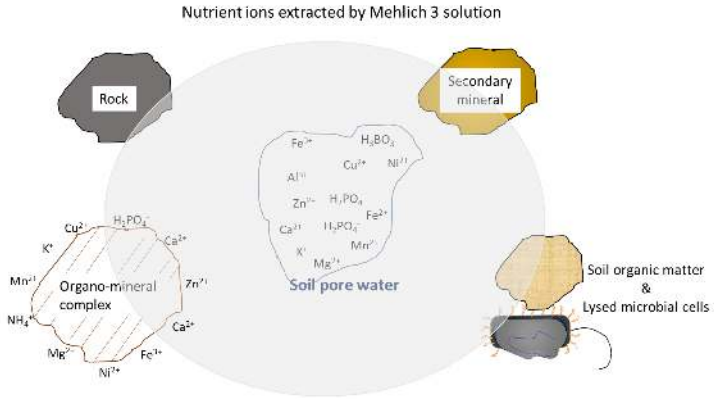


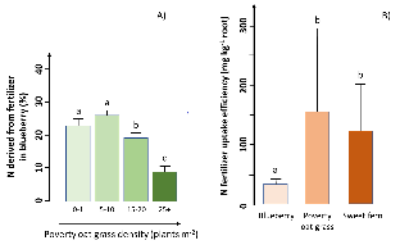
Figure 7.4. The Mehlich 3 extractant was developed by Dr. Adolph Mehlich to estimate the plant availability of many macronutrients and micronutrients in soil. In Eastern Canada, the fertilizer recommendations are based on the nutrient concentrations in Mehlich 3 extracts. The chemical solution extracts nutrients from the soil pore water, as well as some of those that are contained in rocks, in precipitated minerals such as iron phosphates and aluminum phosphates, adsorbed on cation and anion exchange sites, and contained in microbial biomass and organic matter. The chemical composition of the Mehlich 3 extractant is 0.2 N CH_3COOH + 0.25 N NH_4NO_3 + 0.013N HNO_3 + 0.015N NH_4F + 0.001 M EDTA (Mehlich, 1984). © Joann Whalen is licensed under a CC BY (Attribution) license.

Another purpose of soil fertility testing is to assess the probability of a profitable response to fertilizers and soil amendments. This evaluation is part of an integrated plan to improve the growing conditions for crops. Farms set yield targets for their crops, according to the production goals and economic returns they hope to achieve. The maximum yield that can be expected for a particular crop depends on its genetic and physiological characteristics, which govern the plant's photosynthetic capacity, its ability to acquire water and nutrients from soil, and its ability to convert these resources into grain, oilseed, starchy tubers, fruit or other marketable products. Furthermore, the maximum yield is influenced by the plant's response to environmental conditions during the growing season. Unfavorable weather can provide too much or too little rainfall for the crop. Heat stress resulting from too hot temperatures during summer months, or frost damage in the spring and fall months, are stressful for crops. Farms also follow good agronomic practices to control growth limiting factors. For instance, the number of plants occupying a given area is optimized to intercept all available solar radiation, but too-close spacing could cause competition among plants and should be avoided. Weeds, pests and disease-causing organisms must be controlled to prevent yield losses. Finally, soils need to provide adequate nutrients, at the right time during crop development, to achieve the expected yield and quality outcomes.

Can You Dig It!

When Blueberry Gets Wild with Nitrogen Fertilizer

The availability of nitrogen (N) in the soil solution limits plant growth and crop yields in most agroecosystems. That is why farmers often fertilize their fields with one or several forms of N. Nevertheless, the efficacy of N fertilization depends on several factors including the presence of weeds, which can compete with the crop for the acquisition of fertilizer-derived N. Inappropriate N fertilization can indeed be counterproductive as it can stimulate weed growth rather than that of the targeted crop.



N fertilizer uptake and uptake efficiency. © Marty et al. 2019 is licensed under a CC BY (Attribution) license.

Researchers conducted a study in a lowbush blueberry (Vaccinium angustifolium Aiton) commercial field of the Lac Saint-Jean region, Québec, Canada, to assess the efficacy of N fertilization in the presence of two common weed species (poverty oat grass and sweet fern). The researchers found a much higher aboveground biomass production and a larger amount of N in blueberry than in both weed species, indicating that blueberry captured most of the added N fertilizer. However, blueberry growth and fertilizer uptake decreased with increasing weed density (Figure A). When poverty oat grass density was more than 15 plants m^{-2}, this weed captured about the same amount of fertilizer as blueberry despite a much smaller belowground biomass, resulting in a significant negative impact on blueberry production. In addition, the researchers found that fertilizer uptake efficiency (the fertilizer uptake per gram of root) was about four times lower in blueberry than in both weed species (Figure B). ***In other words, compared to the weeds, blueberry requires four times more belowground biomass to acquire the same amount of N from fertilizer.***

The complete story is provided in Marty et al. (2019)

Since soil fertility testing quantifies the plant-available nutrient concentrations, it can be used to interpret the crop response to nutrient inputs. Soil fertility testing should be accompanied by measurements of the crop response to nutrient inputs in controlled environments such as growth chambers and greenhouses, and in the field. In soils with low to medium nutrient supply, crop growth will probably be improved by adding fertilizer. On the other hand, crops planted in soils with high to excessively high soil fertility rarely benefit from fertilizer application (Figure 7.5). Since fertilizers cost money to purchase, transport and apply to fields, it is not profitable to apply fertilizer when no yield improvement is expected, since this will not give an economic return on the investment.

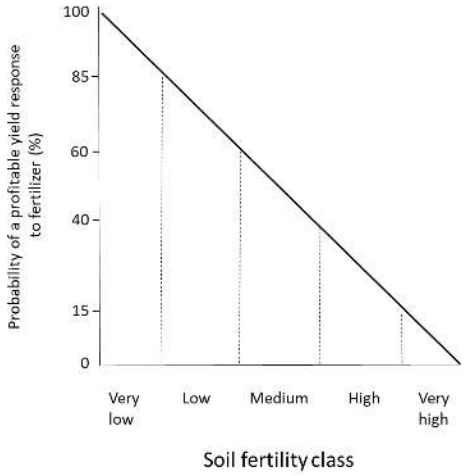


Figure 7.5. The probability of a profitable yield response from fertilizer declines as the soil fertility level goes from very low to very high. © Joann Whalen is licensed under a CC BY (Attribution) license.

Plant nutrition is determined by measuring the nutrient content of plants, either in a specific tissue or in the whole plant. When plants receive ample nutrients from the soil and fertilizer, if needed, their nutrient concentration will be in the sufficient range. When the soil nutrient supply is insufficient and no fertilizer is applied, the plant will be deficient in that nutrient. Plants that accumulate more nutrients than they need for their metabolic processes could have an excessively high or toxic level of nutrients (Figure 6). Nutrient toxicity is uncommon for macronutrients such as N, P, K, Ca, Mg and S because the active transport processes that transfers water-soluble nutrient ions through the endodermis and into the xylem are tightly regulated at a genetic level. Toxicity is sometimes observed when micronutrients such as B, Fe, Mn, Zn and Cu are applied at too-high rates to the foliage because these elements can damage the sensitive, photosynthetic cells in leaves. However, these micronutrients can be absorbed safely through

the root system because plants have defense mechanisms that control the uptake, accumulation and translocation of micronutrients and other metals. This involves retaining the ion in root cells, where they are detoxified through complexation with amino acids, organic acids or metal-binding peptides and/or sequestered into vacuoles. These mechanisms will prevent micronutrient translocation to the sensitive leaf tissues.

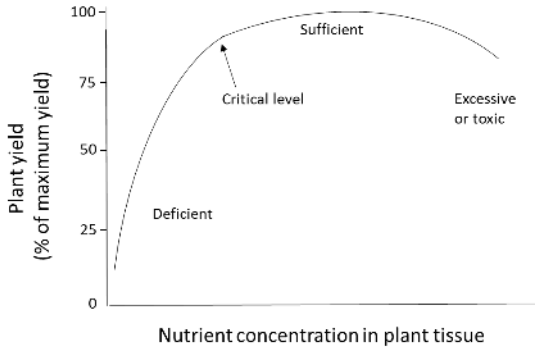


Figure 7.6. Plant yield is related to the concentration of essential macronutrients and micronutrients in plant tissue. As the nutrient concentration in plant tissue increases towards the critical level, plant yield increases. Above the critical level, the plant contains sufficient nutrients for normal growth and can continue to absorb nutrients without a yield increase. An excessive nutrient concentration in plant tissue can be toxic to the plant, resulting in yield loss or plant death. © Joann Whalen is licensed under a CC BY (Attribution) license.

Nutrient requirements are known for all agricultural crops. Crops that have less than the critical nutrient level for development will not grow normally, will be stunted, and they may display deficiency symptoms such as chlorosis (yellowing of the leaves). This may be a sign that the plant lacked nutrients needed to produce chlorophyll, since a well-nourished plant will produce dark green leaves for photosynthesis. Even if there are no visible symptoms of a nutrient deficiency, inadequate plant nutrition can produce other undesirable outcomes:

- weak stems that are susceptible to lodging during windstorms, hailstorms and intense rainstorms;
- few flowers or inflorescences;
- seed abortion;
- poor-quality seeds (low protein content, lacking essential amino acids, low lipid content); and
- loss of marketable yield due to little seed/grain production, small fruits and tubers.

The best way to determine if crop growth is limited by lack of nutrients is to sample and analyze plant tissue. This involves taking a sample of the plant at an appropriate growth stage, generally during vegetative growth or in the early reproductive growth stage. The plant sample is then sent to a certified laboratory for analysis, and the results can be interpreted by checking reference tables that present the sufficient level of the macro- and micro-nutrients in a particular crop (Table 7.3). Since every crop has unique nutrient requirements at each of its growth stages, the reference table has to be specific for the crop species of interest. Furthermore, plant tissue analysis is done early in the growing season because the results may indicate that fertilizers should be applied to correct or prevent a nutrient deficiency. Timely analysis of plant tissue can guide the farmer to make an in-season nutrient application that will prevent yield losses. However, the diagnosis from a plant tissue test might arrive too late to take corrective action. For example, the crop may be too large for fertilizer applicators to enter the field. Another possibility is that the deficient nutrient is not mobile in the plant, meaning it is too late in the plant's development to overcome the growth limitation associated with this nutrient deficiency. Therefore, soil fertility testing is the most important diagnostic tool to avoid a negative economic outcome that occurs when crops have inadequate nutrition during the growing season.

Table 7.3. Sufficiency ranges for macronutrients and micronutrients in corn plants, in relationship to corn growth stages (based on Jones, 2003)

Element	V4-V6 stage (30 cm tall)	V10 stage (prior to tasselling)	R1 stage (initial silk)
Nitrogen (%)	3.50-5.00	3.00-5.00	2.70-4.00
Phosphorus (%)	0.30-0.50	0.25-0.45	0.25-0.50
Potassium (%)	2.50-4.00	2.00-2.50	1.70-3.00
Calcium (%)	0.30-0.70	0.25-0.50	0.21-1.00
Magnesium (%)	0.15-0.45	0.13-0.30	0.20-1.00
Sulphur (%)	0.15-0.50	0.15-0.50	0.21-0.50
Boron (mg kg^{-1})	5-25	4-25	5-25
Copper (mg kg^{-1})	5-20	3-15	6-20
Iron (mg kg^{-1})	50-250	10-100	20-250
Manganese (mg kg^{-1})	20-300	15-300	20-200
Molybdenum (mg kg^{-1})	0.10-0.30	0.10-0.30	0.10-0.20
Zinc (mg kg^{-1})	20-60	15-60	25-100

FERTILIZERS AND SOIL AMENDMENTS

Fertilizer is any material that supplies water-soluble nutrient ions for crop production. Plants absorb nutrient ions from the soil pore water, without preference or discrimination among fertilizer sources. Therefore, any fertilizer that releases water-soluble nutrient ions could be beneficial to plants. Characteristics of some fertilizers that can increase the NH_4^+ and NO_3^- supply in soil are listed in Table 7.4.

Table 7.4. Characteristics of some nitrogen fertilizers that can supply NH_4^+ and NO_3^- for crops, based on IPNI (2012) and Munroe (2016)

Fertilizer	Guaranteed analysis[1]	Comments
Anhydrous ammonia	82-0-0	Applied as a compressed gas; requires special equipment and training
Calcium ammonium nitrate	27.5-0-0	Granular; water-soluble
Ammonium sulfate	21-0-0 + 24% sulphur	Granular; water-soluble
Urea	46-0-0	Granular; water-soluble Hydrolyzed by urease enzymes to release NH_4^+
Urea ammonium nitrate	28-0-0	Solution containing 50% urea and 50% ammonium nitrate
	Typical analysis[2]	
Liquid pig slurry	2.2-2.7-1.9	Solution (>90% water) with C:N ratio <5
Dairy slurry	1.2-1.8-2.9	Solution (>90% water) with C:N ratio <10
Chicken manure	2.4-1.1-1.0	Decomposed solid, C:N ratio <20
Beef cattle manure	0.8-1.8-2.6	Partially-decomposed solid, C:N ratio <25

[1]Guaranteed analysis is based on the %N-%P_2O_5-%K_2O in the fertilizer, on a dry weight basis.
[2]Typical analysis of manure is based on the %N-%P_2O_5-%K_2O, on a wet weight basis

Fertilizer sources can be classified based on how they are manufactured, but from the perspective of plant nutrition, it is more interesting to consider how quickly the water-soluble nutrient ions are released into the soil pore water. For example, granular monoammonium phosphate purchased from a fertilizer company dissolves in soil water to release water-soluble NH_4^+ and $H_2PO_4^-$ ions. Another solid fertilizer is chicken manure, which contains about 4% total nitrogen and 0.5% total phosphorus (equivalent to 1.1% P_2O_5). The total nitrogen is composed of uric acid, protein and NH_4^+, and the total phosphorus contains orthophosphate compounds (50-75% of the total P) and organic phosphorus compounds such as myo-inositol (25-50% of the total P). Because chickens do not produce phosphatase enzymes in their digestive tract, only 10 to 30 per cent of the organic phosphorus in their food is absorbed by the animal and the undigested phosphorus is excreted. When chicken manure is applied to soil, the organic phosphorus compounds dissolve in soil water and are degraded by phosphatase enzymes that were produced by plant roots and soil microorganisms. Phosphatase enzyme hydrolyze organic phosphorus to release phosphate (HPO_4^{2-}) ions into the soil pore water (Figure 7.7). The dissolution of granular monoammonium phosphate generally occurs faster than the hydrolysis of organic phosphorus compounds in chicken manure, which is why salt-based fertilizers are a ‘fast-release’ fertilizer and manure is considered a ‘slow-release’ fertilizer.

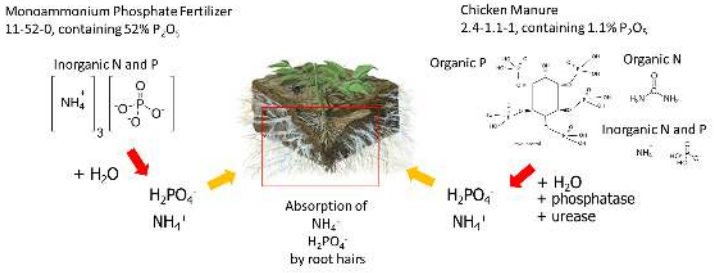


Figure 7.7. Commercial fertilizer such as monoammonium phosphate dissolves in water. Organic fertilizers such as chicken manure also contain NH_4^+ and $H_2PO_4^-$ ions, since inorganic forms can be >50% of the total nitrogen and total phosphorus in manure. Urease and phosphatase enzymes, produced by plant roots and soil microorganisms, are responsible for hydrolyzing organic nitrogen and organic phosphorus compounds in chicken manure. Soluble NH_4^+ and $H_2PO_4^-$ ions are absorbed through root hairs and used for plant metabolic processes. Water-soluble fertilizers are considered to be "fast-release" fertilizer and organic fertilizer is a relatively "slow-release" fertilizer. The numbers under each fertilizer are the guaranteed analysis based on the %N-%P_2O_5-%K_2O in the fertilizer. Monoammonium phosphate is a compound fertilizer that contains 11% N, 52% P_2O_5 and no potassium. Chicken manure is a compound fertilizer that, on average, contains 2.4% N, 1.1% P_2O_5 and 1% K_2O. © Joann Whalen is licensed under a CC BY (Attribution) license.

A soil amendment is a material that is added to improve the soil condition, but it does not necessarily provide nutrients for plant growth. Ground lime is often applied to acidic soils to increase the soil pH and reduce the concentration of Al^{3+} and Mn^{2+} in the soil pore water, since these elements can be damaging to crop roots. Neutralizing soil pH to around 6.0–7.0 provides more water-soluble NH_4^+, $H_2PO_4^-$ and K^+, and increases the activity of soil bacteria, which are beneficial for plant growth and maintain soil structure. Other soil amendments include mulch and biochar. These carbon-rich materials are not expected to be a substantial source of plant-available nutrients, but they contain organic matter that can retain soil moisture and exchangeable ions, build soil structure and provide substrate for diverse heterotropic microorganisms. Since carbon-rich amendments stimulate microbial growth, they must be used sparingly to avoid the possibility that fast-growing microorganisms will immobilize nutrient ions from the soil pore water, thereby depriving plants of plant-available nutrients.

SOIL NUTRIENT CYCLES

Agricultural crops absorb nutrient ions from the soil pore water to produce edible biomass (e.g., grain, oilseed, forage, fruit and tubers) and non-edible biomass, some of which is harvested for animal bedding and for other purposes (i.e., second-generation biofuels are made from non-edible stems and leaves). Fertilizer addition will replenish the plant-available nutrients, but the nutrient ions from fertilizer will not necessarily remain in the soil pore water because they are subject to the biological, chemical and physical reactions of soil nutrient cycles, described below.

Nitrogen Cycling

Reactions that contribute plant-available nitrogen

The soil nitrogen cycle (Figure 7.8) is well studied, due to the central importance of nitrogen for chlorophyll production, Calvin cycle reactions that take place in chloroplasts during photosynthesis, and protein synthesis in plants. One way that agricultural crops can obtain nitrogen is through a symbiotic or associative relationships with N_2-fixing prokaryotes. The best known of these are soil bacteria in the Rhizobium genera, but there are many others (Table 7.5). The N_2-fixing prokaryotes possess the nitrogenase enzyme, which can transform atmospheric N_2 into ammonia (NH_3), the precursor for protein synthesis, through the reaction:

$$N_2 + 8H^+ + 16ATP + 8e^- \rightarrow 2NH_3 + H_2 + 16ATP + 16P_i$$

The nitrogenase complex reduces N_2 to NH_3, and it also reduces H^+ to H_2. The energy (ATP) for this reaction comes from the breakdown, via glycolysis, of photosynthates that are transferred from the host plant to the symbiont. The NH_3 (g) product diffuses out of the bacteriod's protoplasm into the root cortex, where it is protonated with 2 H^+ ions, without additional energy, to yield:

$$NH_3 + 2H^+ \rightarrow 2NH_4^+$$

The NH_4^+ then binds to glutamate to produce glutamine, or is transformed into ureides such as allantoin, allantoic acid or citrulline, before it is translocated from the root to other parts of the plant.

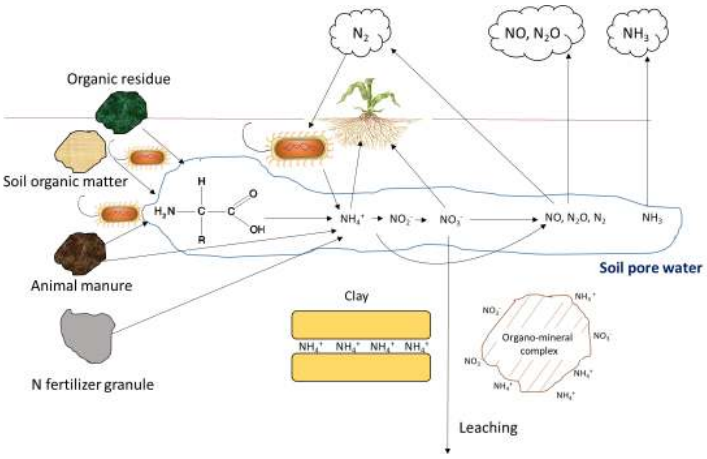


Figure 7.8. Simplified version of the soil nitrogen cycle. © Joann Whalen is licensed under a CC BY (Attribution) license.

Legumes in the Fabaceae family form a symbiosis with N_2-fixing bacteria, which inhabit nodules in their roots. The symbiosis is initiated when roots secrete flavonoids, which are chemo-attractants for the nodulating bacteria (rhizobia) to attach to the root hairs. The flavonoid signal activates the nodulation (nod) genes in the rhizobia, which synthesizes Nod factors that bind to surface protein receptors at the subapical root tip. This causes the root hair to curl and entrap the rhizobia, and the plasma membrane invaginates to form the infection thread, a tubular structure that extends from the root hair tip to lower cells of the root cortex. Rhizobia move through the infection thread and multiply, and root cortex cells proliferate into the nodule primordia, which will eventually become a fully-formed nodule enveloped in a membrane that prevents oxygen diffusion. This protective membrane prevents nitrogenase from being inactivated by oxygen. At this stage, the rhizobial bacteria become bacteroids capable of biological N_2 fixation.

Although N_2 is abundant (78% of the Earth's atmosphere), the biological N_2 fixation reaction is energetically demanding. In theory, 16 ATP are required to reduce one molecule of N_2 to $NH_{3\ (g)}$, but the actual energetic requirement is generally higher due to inefficiencies in the process. Therefore, biological N_2 fixation is a strictly controlled process. All other available forms of nitrogen, such as NH_4^+, NO_3^- and amino acids, will inactivate the nitrogenase complex and inhibit the expression of the nif genes that encode for the nitrogenase components. In addition, the reaction products $NH_{3\ (g)}$ and H_2 also inhibit nitrogenase activity. Consequently, most legumes acquire 40-75% of their nitrogen requirement from biological N_2 fixation, and the rest is absorbed as water-soluble NH_4^+ and NO_3^- from the soil pore water.

Associative N_2-fixing bacteria live in the rhizosphere (in the soil pores and on soil particles close to plant roots) or as endophytes (within the plant tissues) of grasses such as sugarcane, rice, wheat, sorghum, maize and others. Some common genera are Azospirillum and Herbaspirillum, and there are many others (Table 7.5). These bacteria do not form nodules. Instead, they live on the root surface or within the intercellular space and xylem vessels of stems and leaves. They obtain energy from plant photosynthates and transfer a portion of the NH_4^+ generated from N_2 fixation to the plant. There is considerable variability in their contribution to crop nitrogen requirements. In sugarcane, *Gluconacetobacter diazotrophicus* can supply 0-60% of the nitrogen requirement under field conditions. Cereals and oilseeds may obtain 1-25 kg N ha^{-1} y^{-1} from their associative N_2-fixing bacteria. The rest of the nitrogen needed by these crops must be absorbed from the soil pore water as NH_4^+ and NO_3^- ions.

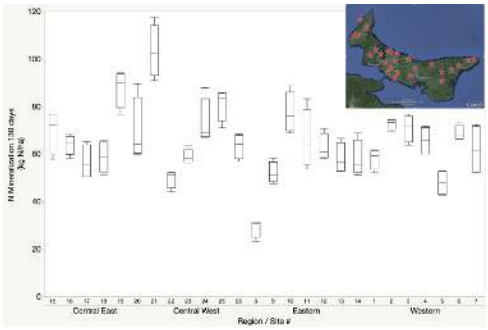
Table 7.5. Nitrogen-fixing prokatyotes that can supply nitrogen to plants. Adapted from Rascio and La Rocca (2013)

Class	N_2-fixing prokaryote	Host plant
Symbiotic	Allorhizobium spp. (fast-growing) Azorhizobium spp. (fast-growing) Bradyrhizobium spp. (slow-growing) Mesorhizobium spp. (intermediate-growing) Rhizobium spp. (fast-growing) Sinorhizobium spp. (fast-growing)	Legumes (Fabaceae family)
	Anabaena azollae (cyanobacteria)	Water fern (Azolla)
	Frankia spp. (actinobacteria)	Woody shrubs, e.g., in the families Betulaceae, Casuarinaceae and Myricaceae
Associative	Azospirillum spp. Bacillus spp. Beijerinckia fluminense Derxia spp. Enterobacter spp. Erwinia spp. Gluconacetobacter diazotrophicus Herbaspirillum spp.	Cereals (Poaceae family) Oilseeds (Brassicaceae family)

Can You Dig It!

The importance of measuring soil N supply: A Prince Edward Island example

Nitrogen is a very dynamic nutrient—both critical to crop production and with the potential for significant environmental impact. Thus it is essential that we use nitrogen as effectively and efficiently as possible. In determining the need for nitrogen additions, we cannot forget to account for the nitrogen that will be supplied by the soil. In semi-arid and subhumid agroecosystems, such as those in the Canadian Prairies, residual nitrate-nitrogen (NO_3-N) is an important source of plant N for the subsequent crop and should be measured in soil fertility tests and accounted for in fertilizer recommendations. In cool, humid agroecosystems such as those in Eastern Canada, overwinter loss of nitrogen is such that there is often limited carry over of nitrate (Zebarth et al., 2003). As a result, in-season nitrogen mineralization is an important crop input in both types of agroecosystems.

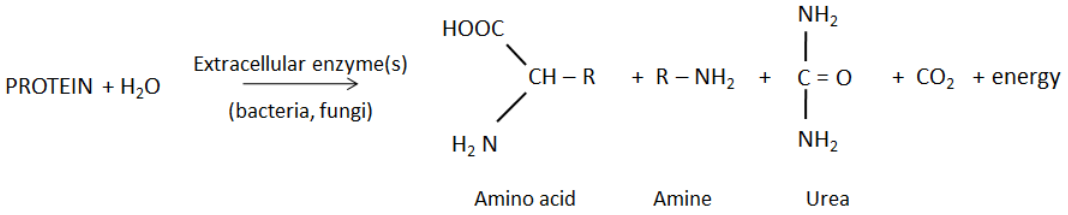


Nitrogen mineralization potential of PEI soils

There are a variety of approaches to estimate the nitrogen mineralization potential that range from measures of the quality of the organic matter to the activity of the microbial community (Dessureault -Rompré et al., 2011). The variation in nitrogen mineralization potential can vary substantially as a result of soil moisture, temperature, soil type, cropping system or soil management. A survey of nitrogen mineralization potential of fields in three-year potato rotations in Prince Edward Island demonstrated that potential N mineralization over a 130-day growing period could range from 30–110 kg N ha^{-1}. While there was on average greater N mineralization potential associated with central regions of the province, the greatest variation was from field to field, emphasizing the value of site-specific soil testing to quantify N mineralization potential.

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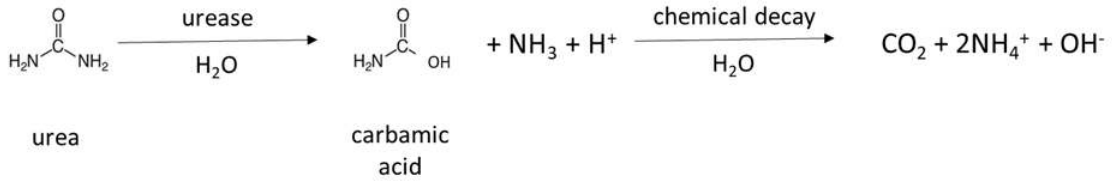
Biological mineralization of proteins, also referred to as nitrogen mineralization, is the main source of NH_4^+ in unfertilized soil. This naturally-occurring process releases water-soluble NH_4^+ from decaying plant, animal and microbial residues, which were recently added to soil or are part of the soil organic matter. Soil bacteria and fungi produce protease and peptidase enzymes, as well as amino acid hydrolase enzymes that cleave the amine groups from amino acids and urease, which degrades the urea molecule. The general mineralization reaction is:



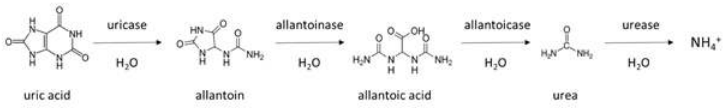
Nitrogen mineralization occurs more quickly with increasing soil temperature and is optimal when soil moisture reaches 60% water-filled pore space. Soil conditions that favour bacterial activity (pH from 6 to 7.5, low salinity) will stimulate mineralization, since bacteria have a faster generation time and greater capacity to produce protein-degrading enzymes than fungi.

Nitrogen mineralization is also affected by the C:N ratio in organic residues. The C:N ratio is the mass of the total organic carbon divided by the mass of total nitrogen in an organic substance. Organic residues with a larger proportion of organic carbon provide substrates for heterotrophic microorganisms, which use carbon for metabolic processes and biomass production, and respire CO_2. If their growth becomes limited by a lack of nitrogen in the organic residue, heterotrophic microorganisms will immobilize NH_4^+ and NO_3^- from the soil pore water. Organic residues with a C:N ratio > 20 such as woodchips, straw and grass-based residues tend to be rich in organic carbon and relatively impoverished in total nitrogen, which causes immobilization. Organic residues with a C:N ratio < 20 such as food-based compost and decomposed animal manure are relatively enriched in total nitrogen, so they will mineralize and increase the NH_4^+ concentration in soil pore water.

Many fertilizers are a source of NH_4^+ in soil. Granular fertilizer such as ammonium sulfate, $(NH_4)_2SO_4$, is water-soluble and dissolves to release NH_4^+ (Table 7.4). Urea fertilizer is hydrolyzed by the urease enzyme, which is produced by naturally-occurring soil microorganisms and plants, to release NH_4^+:

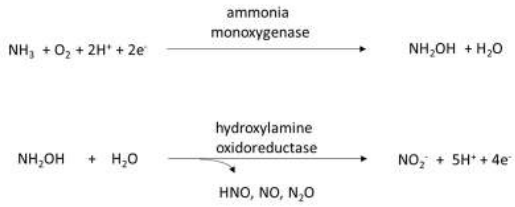


A similar reaction occurs in animal manure, which contains urea from animal urine. Chicken manure collected from laying facilities without bedding is enriched with urea-like molecules, due to the fact that 80% of the nitrogen excreted is in the form of uric acid. It takes several days to a week to transform uric acid to NH_4^+ in a warm, moist soil according to the general reaction:

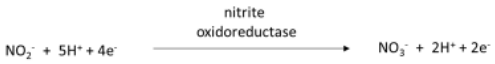


Pig manure that is collected by spray-washing the barns and stored in a liquid form also has a high proportion of NH_4^+, due to urea hydrolysis and protein degradation that occurs in the manure storage. For example, liquid pig slurry contains >90% water, has a C:N ratio of 3, and the NH_4^+ form represents 75% of the total N content. The high proportion of readily-degradable uric acid in chicken manure, and the large amount of water-soluble NH_4^+ in liquid pig slurry, makes these manures “fast-release” nitrogen fertilizer, compared to animal manure mixed with bedding, which has a lower water content.

Ammonia oxidizers and nitrifiers are responsible for transforming the NH_4^+ to NO_3^-. The ammonia oxidizers are a diverse group of prokaryotes, including bacteria and archaea, which possess the amo and hao genes that encode for the enzymes that catalyze the following 2-step reaction:



Nitrifiers are prokaryotes and eukaryotes that possess the nxr gene and are capable of converting nitrite (NO_2^-) into NO_3^-, as:



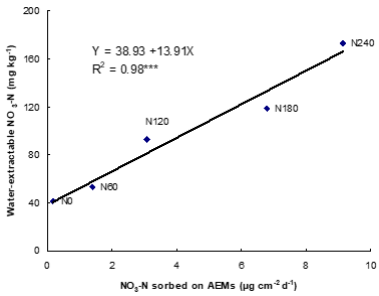
Ammonia oxidation and nitrification reactions occur quickly in field soils during the growing season. Consequently, NH_4^+ released from nitrogen mineralization or added in water-soluble fertilizer or animal manure will be transformed into NO_3^- within days to weeks. This understanding becomes important when deciding when to apply fertilizer, because we would like crops to absorb most of the water-soluble NH_4^- and NO_3^- from fertilizer as quickly as possible after the fertilizer application. Otherwise, reactions in the nitrogen cycle will remove these nutrient ions from the soil pore water, thereby reducing the amount of plant-available nitrogen for crops and increasing the risk of environmental pollution.

Can You Dig It!

Soil mineral nitrogen as predicted by ion exchange membranes

Ionic exchange membranes (IEMs) and plant root simulators were developed to determine N availability for agricultural and environmental studies. Indeed, IEMs contain positively or negatively charged surface functional groups that adsorb ions, such as NH_4^+ and NO_3^-, by electrostatic attraction. To maintain the equilibrium between the soil solid phase and soil pore water, IEMs continually remove NH_4^+ and NO_3^- that is mineralized from organic N or desorbed from organo-mineral surfaces.

One way to evaluate the NH_4^+ and NO_3^- concentration is to bury IEMs in the field for a period of time that depends on the objectives. Alternatively, the NH_4^+ and NO_3^- concentration can be assessed in a standardized laboratory environment. Once removed from the soil, IEMs are collected, washed with distilled water in the field to remove adhering soil particles and placed in individual tubes containing either NaCl (anionic exchange membranes) or HCl (cationic exchange membranes), and analyzed in laboratory for NO_3^- and NH_4^+, respectively. Soil nitrate and ammonium fluxes (supply rates) measured by IEMs are reported in µg cm^{-2} d^{-1} (Ziadi et al., 1999).



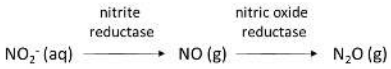
Relationship between soil NO_3-N following a water extraction and NO_3-N adsorbed on ionic exchange membranes in forages fertilized with various N rates in an experiment conducted at four site-years in eastern Canada (Ziadi et al., 1999). © Noura Ziadi is licensed under a CC BY (Attribution) license.

In general, there are significant correlations between IEMs and chemical extractions (Ziadi et al., 1999), and between crop N uptake and nitrates adsorbed on IEMs. Plant N uptake was more closely related to soil N as evaluated by IEMs than estimated by chemical extractions, suggesting that IEMs provide a better index of soil N availability than chemical extractions. Over all sites and treatments (mineral and organic N), NO_3^- and NH_4^+ adsorbed on IEMs increased with increasing N fertilizer rates. SoilNO_3^- and NH_4^+ adsorbed on IEMs varied largely with growing seasons and development stage of crop indicating the variation of soil N availability to plants. The in situ use of ionic exchange membranes as a soil testing procedure provide a more precise evaluation of N availability to different crops (forages, corn and potato) than standard methods. Based on its simplicity, rapidity, and low cost, IEMs have many practical advantages over chemical extractions for assessing soil N availability in eastern Canada.

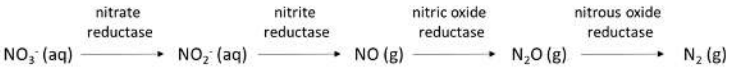
Processes that deplete plant-available nitrogen

The NH_4^+ and NO_3^- ions in the soil pore water are removed through biological, chemical and physical processes. Nitrogen uptake by plants and immobilization by microorganisms is driven by nitrogen requirements for protein synthesis and formation of genetic material. In plants, nitrogen is a major component of proteins of the Calvin cycle (e.g., about 20-30% of the leaf nitrogen is found in the RuBisCO enzyme that initiates carbohydrate production), the thylakoids and the chlorophyll molecule and thus essential for CO_2 fixation associated with photosynthesis.

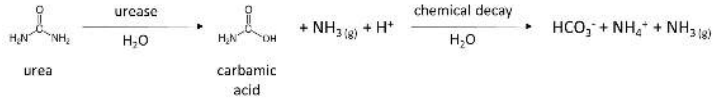
Soil biological reactions that use NO_2^- and NO_3^- as electron acceptors are common. These reactions permit soil bacteria and fungi to acquire oxygen when the soil environment is anoxic. High soil moisture content can create anoxic conditions, as can intense biological activity that consumes oxygen and produce carbon dioxide, such as in the rhizosphere or in the detritusphere (i.e., around decomposing organic matter). Nitrifier-denitrification is one such reaction, and is expected when the soil moisture content is from 50 to 80% of water-filled pore space.



Denitrification is another biological reduction reaction that consumes NO_3^-, and it is expected when the soil moisture content exceeds 80% water-filled pore space.



A chemical reaction that depletes the NH_4^+ concentration is $NH_{3\ (g)}$ volatilization, which occurs when NH_4^+ is deprotonated. During urea hydrolysis, this can be the result of carbonic acid formation:



Animal manure can lose up to 50% of the total nitrogen excreted by the animal as $NH_{3\ (g)}$ during manure storage and land application. Urea fertilizer is also susceptible to $NH_{3\ (g)}$ loss, especially in alkaline soils that have an excess of OH^- and HCO_3^- in the soil pore water, since H^+ associates with these counter-anions rather than with NH_3. Volatilization of $NH_{3\ (g)}$ is undesirable because the transfer of gaseous nitrogen from agricultural fields to the atmosphere reduces plant-available nitrogen for the crop. Eventually, the $NH_{3\ (g)}$ will return to the ground through wet and dry deposition, which results in the unintentional fertilization of non-agricultural land and aquatic systems.

Ammonium fixation is a chemical reaction that removes NH_4^+ from the soil pore water. This occurs when NH_4^+ adsorbs to exchange sites on organo-mineral surfaces. Ammonium fixation is greater in clayey soils, which have greater reactive surface area and more adsorption sites in the interlattices (2:1 clays only), than sandy soils. Due to its 1+ charge and small ionic radius, NH_4^+ is likely to bind to the interlattice, which reduces the amount of NH_4^+ in the soil pore water. This could account for a reserve of as much as 90–460 mg NH_4^+ kg^{-1} soil in fine-textured clays. In contrast, the NO_3^- molecule is loosely retained in the soil matrix due to its size and negative charge, and is more likely to leach through the soil profile and below the plant root zone. Nitrate leaching is affected by soil porosity and the water infiltration rate after snowmelt, rainfall and irrigation events. Finally, both NH_4^+ and NO_3^-

are susceptible to loss from soil through erosion and surface runoff processes. Wind and water erode soil particles that contain adsorbed NH_4^+, while runoff water transports NO_3^- from saturated soils through overland flow.

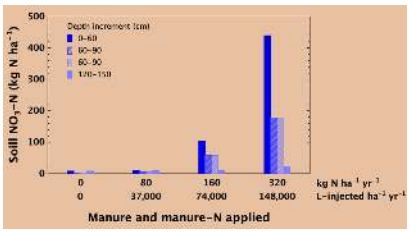
Can You Dig It!

Optimizing the nitrogen application rate of animal manure



Low disturbance injection of liquid swine manure at research plots in east-central Saskatchewan. Photo credit: Dr. Tom King. © Tom King is licensed under a CC BY (Attribution) license

Balancing the rate of nitrogen addition with removal in the harvested crop over time prevents excessive accumulation of nitrate in the soil. Unused nitrate can slowly migrate down below the rooting zone over time with deep percolating water. In a trial in east-central Saskatchewan, liquid swine manure was added every year for eight years at a recommended agronomic rate of 37,000 L ha^{-1} yr^{-1}, adding about 80 kg N ha^{-1} yr^{-1} which was about equal to the amount of N removed in the grain harvest each year. This rate resulted in little or no accumulation of nitrate in the soil profile, nor was there evidence of any deep leaching compared to an unfertilized control (lower right).



Fall nitrate concentrations in the soil profile (0-150 cm) after eight annual applications of liquid hog manure at Dixon, Saskatchewan (adapted from Stumborg et al. 2007). © Stumborg et al. adapted byTom King is licensed under a CC BY (Attribution) license.

In the unfertilized control, crop production over the years was greatly reduced due to lack of available nitrogen while the agronomic rate produced good yields. At annual application rates that were double (74,000 L ha^{-1} yr^{-1}) and especially quadruple (148,000 L ha^{-1} yr^{-1}) the recommended rate, soil nitrate contents in the top 60cm as well as the depths beneath were elevated compared to the agronomic rate and the unfertilized control. Large amounts of unused nitrate that accumulate in the soil profile over time can slowly migrate to greater depths. If leached deep enough, it can become unreachable by plant roots and be a potential source of ground water contamination. With appropriate application rate that matches crop need and removal, this can be prevented.

Phosphorus Cycling

Reactions that contribute plant-available phosphorus

Plant-available phosphorus is derived from biological and geochemical processes (Figure 7.9). The biological source is organic phosphorus, which is hydrolyzed by phosphatase enzymes from plant roots and soil microorganisms. Organic phosphorus compounds are abundant in plant residues and in undigested plant materials. Inositols are major phosphate storage compounds that contribute to cell wall biosynthesis, cell-to-cell communication, storage and transport of plant hormones and other stress-related molecules, and osmoregulation. Degradation of myo-inositol releases water-soluble $H_2PO_4^-$, as follows:

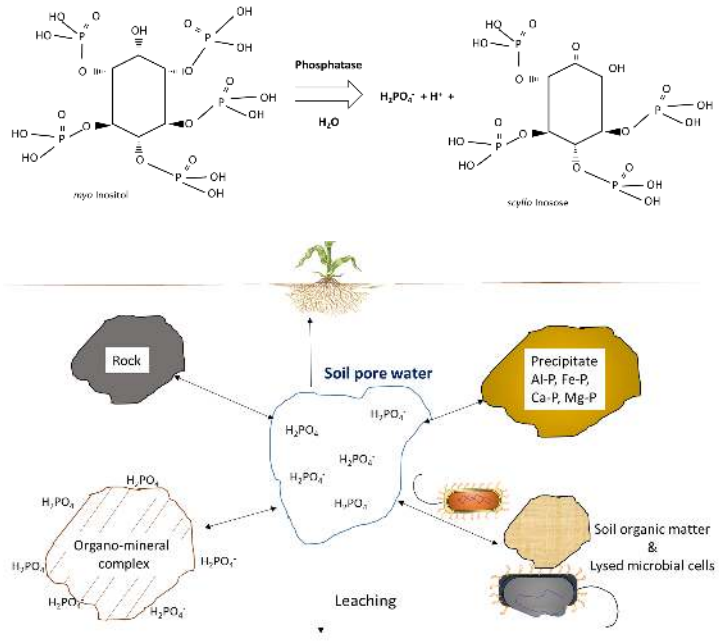


Figure 7.9. Simplified version of the soil phosphorus cycle. © Joann Whalen is licensed under a CC BY (Attribution) license.

Primary minerals such as apatite undergo weathering to release $H_2PO_4^-$ into soil pore water. Other geochemical sources include $H_2PO_4^-$ that is desorbed from clay surfaces or dissolved from secondary minerals, the iron, aluminum, calcium and magnesium phosphates. Weathering and dissolution processes require water. The dissolution of calcium phosphates and magnesium phosphates are accelerated by the presence of carbonic acid ($H_2CO_3 \leftrightarrow HCO_3^- + H^+$) and other acidic substances, however, iron phosphates and aluminum phosphates become less soluble in acidic soils. Inorganic acids that contribute to the solubilisation of $H_2PO_4^-$ from calcium phosphates and magnesium phosphates include H_2SO_4 and HCl. Organic acids produced by plant roots and soil microorganisms lower the pH in their surroundings, resulting in localized $H_2PO_4^-$ solubilization; in addition, organic acids can displace $H_2PO_4^-$ that is adsorbed onto clay and organo-mineral surfaces. Organic acids that are effective in this regard include acetic, lactic, malonic, malic, tartaric, oxalic, citric, ρ-hydrobenzoic and salicylic acids. Soil bacteria that produce organic acids, such as those from the genera Pseudomonas, Bacillus, Rhizobium and Burkholderia, can be isolated, cultivated and re-inoculated into soil to increase $H_2PO_4^-$ released via dissolution, but have not proven to be reliable at solubilizing $H_2PO_4^-$ under field conditions. Arbuscular mycorrhizal fungi also produce organic acids that increase water-soluble $H_2PO_4^-$, which they then absorb through their extra-radial hyphae. Since arbuscular mycorrhizal fungi are plant symbionts, some of the $H_2PO_4^-$ acquired by the fungus can be transferred to the plant host through low and high affinity phosphorus transporters.

Long-term impact of tillage practices and phosphorus fertilization on soil phosphorus forms as determined by ^{31}P nuclear magnetic resonance spectroscopy

Conservation tillage practices (minimum or no-till [NT]) are characterized by minimal soil disturbance and mixing and have been used more frequently in recent years to reduce off-site losses of nutrients associated with eroded particles, including phosphorus (P). However, by maintaining crop residues and fertilizers on the soil surface, the relatively immobile nutrients that do not readily move down the soil profile remain at or near the soil surface. Therefore, NT management systems often result in high concentrations of nutrients at the soil surface (0–5 cm) but sharply decreasing concentrations below this depth. The impact of tillage practices and P fertilization on soil P forms as determined using ^{31}P nuclear magnetic resonance (NMR) spectroscopy was investigated under a long-term corn-soybean rotation experiment established in 1992 in Québec that compared NT with conventional (mouldboard plow; MP) tillage. The results showed that:

- *total phosphorus varied with soil depth under the NT treatment— accumulating in the top unfertilized soil layer (0–5 cm) and in the 0–10 cm layer where P fertilizer was applied;*
- *pH, available P, and total carbon were significantly higher in the topsoil (0–5 cm) of the NT treatment than in the deeper layers (5–20 cm); total N also accumulated in the surface 10 cm;*

- *conversely, nutrient distributions were homogeneous throughout the soil profile under MP management—suggesting that stratification in NT results from the retention of crop residues at the soil surface;*
- *^{31}P-NMR spectra showed concomitant stratification of orthophosphate in NT plots that received 35 kg P ha^{-1};*
- *accumulation of scyllo-IP6 and nucleotides in the deeper layers was possibly due to their preferential movement through the soil column, whereas the pyrophosphate and DNA concentrations were greater at the surface (0–5 cm) soil layer than the deeper layers; and*
- *total C and N were similarly affected, suggesting that DNA was synthesized in greater amounts under NT owing to the higher organic matter accumulated at the soil surface in comparison to conventional tillage.*

Overall, it appears that the accumulation of labile inorganic P at the surface of the NT soil may increase the potential for soluble inorganic P loss in surface runoff and lead to the loss of organic monoesters draining through different hydrological pathways. ***More information available at Abdi et al. (2014).***

Reactions that deplete the plant-available phosphorus

The $H_2PO_4^-$ ions in the soil pore water are removed through biological, chemical and physical processes. Phosphorus uptake by plants and immobilization in microbial biomass are necessary for energy storage and transfer reactions, as well as the formation of nucleic acids and phospholipids.

Phosphorus retention reactions reduce the $H_2PO_4^-$ concentration in soil pore water because $H_2PO_4^-$ adsorbs to organo-mineral surfaces and to the surface of iron and aluminum sesquioxides, namely iron oxide (Fe_2O3), iron hydroxide ($Fe(OH)_3$) and aluminum hydroxide ($Al(OH)_3$). These reactions are common in acidic soils, but occur in all soils to some extent. In general, soils with a higher clay content will adsorb more $H_2PO_4^-$, particularly when they contain 2:1 expanding lattice clays such as smectite. The hydrated sesquioxides of iron and aluminum bind $H_2PO_4^-$ tightly and render it unavailable for plant uptake. In addition, $H_2PO_4^-$ precipitates with Fe, Al and Ca present in the soil pore water. This produces precipitates such as $Fe(OH)_2H_2PO_4$ and the variscite-like compound, $Al(OH)_2H_2PO_4$ in acidic soils. In neutral to alkaline soils, the $H_2PO_4^-$ precipitates with Ca to form slightly soluble dicalcium phosphate dihydrate, $CaHPO_4$. Further reaction with $Ca(OH)_2$ and $CaCO_3$ in alkaline soils leads to the formation of insoluble hydroxylapatite (e.g., $[Ca_3(PO_4)_2]_3{\bullet}CaOH_2$) or insoluble carbonated calcium apatitie (e.g., carbonato-apatite; $[Ca_3(PO_4)_2]_3{\bullet}CaCO_3$).

The phosphorus saturation concept is another way to describe the capacity of soil to retain chemically-bound $H_2PO_4^-$. Soil is shaken with an extracting solution such as Mehlich 3 solution, the amount of extractable P and Al is measured, and the ratio is calculated as:

(1) $$P\ saturation = \frac{P\ in\ Mehlich\ 3\ solution}{Al\ in\ Mehlich\ 3\ solution}$$

Soil with a high degree of phosphorus saturation has elevated $H_2PO_4^-$, relative to the amount of extractable Al that could bind with $H_2PO_4^-$. Therefore, more water-soluble $H_2PO_4^-$ remains in the soil pore water and could leach after snowmelt, rainfall and irrigation events. In addition, erosion of soil particles that contain adsorbed $H_2PO_4^-$ will transport particulate P, while surface runoff will transport the dissolved phosphorus compounds at the soil surface through overland flow. The magnitude of phosphorus loss through these pathways depends upon the concentration of water soluble $H_2PO_4^-$and particulate P in the surface soil, as well as the amount of leaching and severity of surface erosion processes.

Sulphur Cycling

Reactions that contribute plant-available sulphur

Plant-available sulphur can originate from weathering of parent material, atmospheric deposition, dissolution of gypsum ($CaSO_4$) and mineralization of organic matter (Figure 7.10). Plutonic rocks such as pyrite contain elemental sulphur (So) that is oxidized to SO_4^{2-} during the weathering process. Most oxidation reactions are mediated by autotrophic fungi and bacteria, such as Thiobacillus, which is abundant in the rhizosphere.

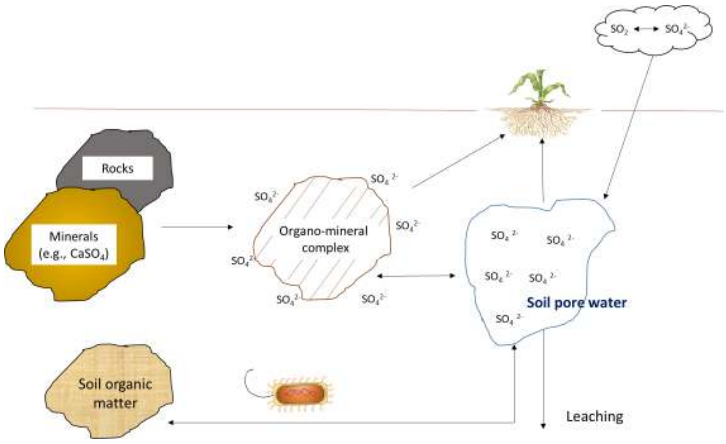
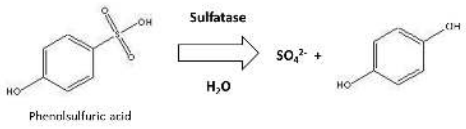


Figure 7.10. Simplified version of the soil sulphur cycle. © Joann Whalen is licensed under a CC BY (Attribution) license.

The conversion of So to SO_4^{2-} is extremely variable in soils, due to natural variation in the distribution of sulphur oxidizing microorganisms. In general, sulphur oxidation reactions occur rapidly in warmer soils with moisture content with around 60% water-filled pore space and soil pH from 6-7.5. In addition, SO_4^{2-} is released from the dissolution of hydrated salts in soil, such as $CaSO_4 \cdot 2H_2O$ and $MgSO_4 \cdot 7\ H_2O$.

Sulphur enters soil from atmospheric sources through dry and wet deposition. Several natural phenomena emit sulphur to the atmosphere, including particulate sulphur compounds that come from weathering of continental crusts (dust) and sea spray. Small particles of sulphur are transported by wind and deposited onto the surface of soil or vegetation. Gaseous emissions of sulphur dioxide, hydrogen sulfide, dimethyl sulfide and other gases from volcanic eruptions, swamps and coastal wetlands are another source. In the atmosphere, these gases tend to be oxidized to sulphuric acid and aerosol sulfate, and then return to the ground in rainfall and snow. These biogenic sources of sulphur account for <10 kg S ha^{-1} y^{-1} in terrestrial ecosystems. Since the Industrial Revolution, the anthropogenic sources of sulphur make a significant contribution to the soil sulphur cycle, and could exceed >50 kg S ha^{-1} y^{-1} in agricultural and forested lands in the proximity of industrial activities. Combustion of fossil fuels, biomass burning, waste incineration and the use of explosives all emit sulphur dioxide and therefore contribute through wet deposition to the sulphur content in soil. However, a reduction in air pollution has lowered the amount of sulphur released from these sources.

Organic sulphur is mineralized by soil enzymes – sulfatases – of plant and microbial origin to release SO_4^{2-}, as illustrated below:



Like other mineralization reactions, sulphur mineralization increases with higher soil temperatures and is optimal when soil moisture content is around 60% water-filled pore space and soil pH is near neutral. Sulphur mineralization is higher around plant roots, which produce sulfatase. Also, plants absorb some of the SO_4^{2-} in soil solution, which could limit the amount of SO_4^{2-} available for microorganisms and therefore induce them to synthesize sulfatase. When fresh plant or animal residues are added to soil, the organic carbon to organic sulphur (C:S) ratio of the material will determine its susceptibility to mineralization. Organic residues with a C:S ratio <200 are mineralized readily and release SO_4^{2-} into soil solution. However, organic residues with a C:S ratio >400 that have a low sulphur content could lead to SO_4^{2-} immobilization by microorganisms, to meet their metabolic requirements.

Reactions that deplete the plant-available sulphur

The SO_4^{2-} ions in soil pore water are removed through biological, chemical and physical processes. Biological uptake of SO_4^{2-} by plants and microorganisms is necessary for the synthesis of essential amino acids (methionine, cysteine, homocysteine, and taurine), for production of glucosinolates, vitamins, as an enzyme cofactor and for chlorophyll production. In anoxic soils, sulfur electron acceptors gain electrons and decrease their oxidation state, so SO_4^{2-} will be reduced to elemental sulphur So and then to S^{2-} by sulphur-reducing bacteria in the phylum Firmicutes, including the genera *Desulfotomaculum*, *Desulfosporomusa*, and *Desulfosporosinus*.

Several chemical reactions reduce the SO_4^{2-} concentration in soil pore water. In calcareous soils, the SO_4^{2-} can be co-precipitated with $CaCO_3$ to form a $CaCO_3 \cdot CaSO_4$ complex. This accounts for about 10% of the total sulphur in calcareous soils. Second, SO_4^{2-} binds to anion exchange sites on organo-mineral surfaces. However, SO_4^{2-} is not as tightly bound to anion exchange sites as HPO_4^{2-} and $H_2PO_4^-$. Consequently, SO_4^{2-} is susceptible to leaching and may be lost through erosion and runoff processes.

Potassium, Calcium, and Magnesium

The mineral elements—potassium, calcium, and magnesium—cycle through geochemical processes (Figure 7.11). Although plants and other organisms require substantial amounts of potassium, calcium and magnesium, these elements function in their ionic forms within biological tissues. For instance, potassium exists predominately as K^+ in the vascular solution of plant tissues and in the cytoplasm of plant cells. It is translocated along with water in living plants. It may exist between layers of organic compounds such as polysaccharides, proteins and lipids, but it does not bind covalently to those substances. Leaching of K^+ and other mineral elements such as Ca^{2+} and Mg^{2+} occurs when rainfall comes into contact with plant and animal residues, before biological decomposition begins.

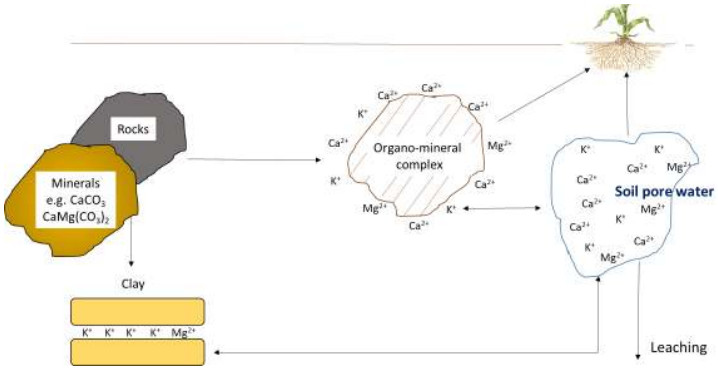


Figure 7.11. Simplified version of the soil potassium, calcium and magnesium cycles. © Joann Whalen is licensed under a CC BY (Attribution) license.

Reactions that contribute plant-available potassium, calcium and magnesium

The amount of K^+, Ca^{2+} and Mg^{2+} in the soil pore water and on the cation exchange sites depends on the nature and weathering of the parent material. Organic residues contribute minor amounts of water-soluble K^+, Ca^{2+} and Mg^{2+}, compared to parent material. Potassium originates from primary minerals, such as feldspar, microcline, mica and biotite, and can also be added as fertilizer.

Soil Fertility Response to Intensive Crop Nutrition



Grower's field in south-central Ontario monitored to assess impact of intensive nutrient management on soil fertility. © Tom Bruulsema is licensed under a CC BY (Attribution) license.

From 2002 to 2006, a corn-soybean rotation in a grower's field in south-central Ontario was monitored to assess changes in response to intensive management. The trial consisted of seven management combinations; three varying K rates at the grower's level of inputs, and four at a more intensive input level—varying both K rate and zone placement.

The soil is a London loam with good drainage. Initial soil fertility levels were 8 ppm Olsen P (low), and 107 ppm ammonium-acetate extractable K (medium). Five treatments used the conservation tillage practice of the grower: fall chisel plow with spring secondary tillage. The other two treatments used fall zone tillage followed by spring zone tillage. Each treatment plot consisted of a strip eight rows wide that ran the full length (300 to 450 m) of the field (see photograph at right).

Yields averaged over the five years increased modestly in response to both increased K rates and management intensity. No combination of input intensity and K rate was more economically viable than the grower's current level of management. There was no evidence of a greater requirement for K with higher input intensity.

Treatment	Fertilizer N-P_2O_5-K_2O (kg ha^{-1})	Seeding rate	Mean Corn Yield (Mg ha^{-1})	2006 Soil Test (ppm)	
				Olsen-P	AA-K
Control, zero K	147-71-3	normal	9.8	10	76
Control, grower K	147-71-33		10.2	9	77
Control, high K	147-71-200		10.6	7	140
Intensive, grower K	278-200-33	high	10.7	26	91
Intensive, high K	278-200-200		11.8	22	145
Intensive, zone-till, grower K	278 200 33		10.7	21	80
Intensive, zone-till, high K	278-200-200		11.1	17	119

Reference: Dean et al. 2007. Corn response to intensive crop nutrition. Better Crops 91:20-21.

Soil test levels changed little where P and K inputs were at the grower's rates but increased significantly where higher rates were applied. Grower rates resulted in net surpluses of P and net deficits of K, though the changes were not large enough to influence soil test levels. At the high rate of each nutrient, however, surpluses of P and K were large enough to substantially increase soil test levels for both nutrients. This work demonstrates that economically optimum crop yields require sufficient nutrient inputs to maintain soil fertility.

Calcium is released from weathering of primary minerals such as anorthite ($CaAl_2Si_2O_3$), followed by pyroxenes and amphiboles, with smaller amounts of Ca coming from the weathering of biotite and apatite. Substantial amounts of Ca^{2+} are released from the dissolution of secondary minerals such as limestone ($CaCO_3$), dolomitic limestone ($CaMg(CO_3)_2$), and gypsum ($CaSO_4 \cdot 2H_2O$), all of which are major sources of Ca^{2+} in Canadian soils.

Magnesium in soils originates from rocks such as biotite, dolomite, hornblende, olivene and serpentine. Magnesium is often a substituted cation in clay minerals, such as chlorite, illite, montomorillonite and vermiculite. Weathering of rocks and clays releases Mg^{2+} into soil pore water, as does dissolution of secondary Mg minerals such as dolomitic limestone ($CaMg(CO_3)_2$), epsomite ($MgSO_4 \cdot 7H_2O$) and bloedite ($Na_2Mg(SO_4)_3 \cdot 4H_2O$).

There is a chemical equilibrium between the amount of K^+, Ca^{2+} and Mg^{2+} in soil pore water and that on cation exchange sites. Smaller ions, particularly H^+, can replace these mineral elements on exchange sites, leading to desorption of K^+, Ca^{2+} and Mg^{2+} into soil pore water. The balance between exchangeable and soluble mineral ions is the key determinant of plant-available K^+, Ca^{2+} and Mg^{2+} concentrations. Since pH affects the charge on organo-mineral surfaces (i.e., neutral to alkaline soils have more negatively charged surfaces than acidic soils), the exchange capacity is related to soil pH, clay and organic matter content.

Reactions that deplete plant-available potassium, calcium and magnesium

The K^+, Ca^{2+} and Mg^{2+} ions in soil pore water are removed through biological, chemical and physical processes. Biological uptake of K^+ is essential for enzyme activation (more than 80 enzymes require K^+ as a cofactor), transport processes and osmotic regulation. The major functions of Ca^{2+} in plants are for maintenance of cell membranes, as well as cell division and elongation. An essential component in chlorophyll, Mg^{2+} is also a structural component in ribosomes and responsible for transfer reactions involving phosphate-reactive groups.

Adsorption reactions are the dominant chemical process that removes K^+, Ca^{2+} and Mg^{2+} from the soil pore water. Adsorption can occur on the outer surfaces ("o" position), inner surfaces ("i" position) and edge ("e" positions) of clay minerals. Furthermore, these ions can be adsorbed in the interlattice of 2:1 clays in mica, illite and vermiculite. Due to its 1+ charge and small ionic radius, K^+ tends to bind in the interlattice whereas Ca^{2+} and Mg^{2+} remain on outer surfaces of clay minerals. This reduces the amount of K^+ in the soil pore water due to potassium fixation, which can account for up to 20% of the total K contained in soils with 2:1 clay minerals. The K^+, Ca^{2+} and Mg^{2+} ions in soil pore water move out of the topsoil by leaching. This is associated with anion leaching, since anions that move through the soil pore water are accompanied by cations to maintain electrical neutrality. Erosion of clay-rich particles is another way that exchangeable K^+, Ca^{2+} and Mg^{2+} ions are depleted from the topsoil.

Micronutrient Elements

Weathering is the primary source of most micronutrients, including iron, copper, manganese, zinc, nickel, boron and molybdenum. Plant-available chlorine may be dissolved from salts such as NaCl and KCl, or it may enter the soil in rainfall and irrigation water. Micronutrients are generally concentrated in the vascular solution and cytoplasm of plants, so they are present in organic residues

and released during the initial physical fragmentation of decomposing organic residues. Consequently, plant and animal residues are a naturally-occurring source of essential micronutrients. Each micronutrient has specific functions in plants, listed in Table 7.6.

Table 7.6. Functions of micronutrients in plants, according to Jones (2003)

Element	Functions
Iron	Fe-containing proteins, including heme proteins and Fe-S proteins, are important for photosynthesis, chlorophyll production and ribonucleic acid synthesis. Cytochromes are Fe-containing heme proteins, and ferredoxin is the most well-known Fe-S protein.
Manganese	Important in redox processes, including electron transport in photosynthesis and detoxification of oxygen-free radicals. An enzyme cofactor for biosynthesis of secondary metabolites such as phenolics, lignins and flavonoids.
Copper	Cu-containing enzymes react with molecular O_2 and are important in photosynthesis, respiration, detoxification of superoxide radicals and lignification reactions.
Zinc	An enzyme activator that improves binding of enzymes and substrates involved in carbohydrate metabolism, protein synthesis and detoxification of superoxide radicals (O_2^-) to protect membrane lipids and proteins against oxidation.
Nickel	Required for the proper functioning of the urease enzyme that catalyses the reaction $CO(NH_2)_2 + H_2O \rightarrow 2\ NH_3 + CO_2$. Any crop that produces ureides requires Ni.
Boron	Forms stable complexes with organic compounds that contributes to cell wall formation and stabilization, lignification and xylem differentiation.
Molybdate	Involved in electron transfer reactions. Mo-containing enzymes catalyse biological N_2 fixation, and are responsible for pollen formation in plants.
Chloride	Mobile anion essential for osmoregulation (e.g., cell elongation, stomatal opening) and charge compensation (e.g., balance charges between cations and anions in the cytoplasm). Essential co-factor for photosynthesis.

Plant uptake of micronutrients is improved at higher temperatures and in plants with prolific root systems, since greater evapotranspiration will move the water-soluble forms of micronutrients to the root surface, where they can be absorbed. Soil moisture has to be sufficient to maintain micronutrient ions in the soil pore water, but not so high as to create anoxic conditions that affect ion absorption by roots. For instance, anoxic conditions prevent uptake of Fe^{2+} by roots, but may increase absorption of toxic Mn^{2+} ions. Soil pH is often critical. The amount of plant-available iron, copper, manganese, zinc and nickel declines as the pH increases to the neutral and alkaline range. For example, iron deficiency can occur in plants growing in calcareous soils with pH >7.5. Plants that are adapted to this environment will respond by altering root morphology to increase the growth of root hair, release protons to acidify the rhizosphere and make Fe^{3+} more soluble, reduce the Fe^{3+} to Fe^{2+} with a plasma membrane-associated Fe^{3+} chelate reductase enzyme, and transport Fe^{2+} across the plasma membrane with an iron regulated transporter. Adapted plants in the Poaceae family have a different strategy, which involves chelation of Fe^{3+} by the phytosiderophores before the Fe^{3+} soluble complex is absorbed by the root.

Plants need far less of the micronutrient elements than macronutrients (Table 7.1). Soil reserves are often sufficient to meet plant requirements, particularly when in soil with pH 6–7 and when adapted cultivars are planted. We can improve micronutrient nutrition by adding fertilizer containing the deficient micronutrient(s) to the soil or directly to the plant foliage. With foliar

application, a small amount of micronutrient fertilizer is sprayed onto the foliage in a solution with an adjuvant (to improve the adhesion of fertilizer solution to the leaf surface) during the plant's vegetative growth or early reproductive growth stages. Water-soluble micronutrients such as H_3BO_3, Cu^{2+}, Mn^{2+}, MoO_4^- and Zn^{2+} and soluble chelates are absorbed through pores in the leaf and move through the plant's vascular system to organs and tissues where they are required. Foliar fertilization is advantageous because it delivers a small amount of micronutrients and avoids adsorption and precipitation reactions that occur in soil. Due to the sensitivity of photosynthetic cells in leaves, the application rate and timing of micronutrients to foliage must be selected with care.

Non-essential Elements

Non-essential elements fall into two classes: (1) elements that are beneficial for some, but not all plants, and (2) elements that are not needed for plant growth. Elements that stimulate the growth of some plants include silicon, sodium and selenium. Silicon is beneficial for rice, sugarcane and other grass species. It accumulates preferentially in the stem tissues and appears to have a role in maintaining the structural integrity of the vascular system, which is required to transport water, nutrients, carbohydrates and other substances throughout the plant. Higher silicon concentrations are associated with greater drought tolerance. Sodium fertilization can improve the growth of some C4 plants, celery, spinach, sugar beet and turnips. Selenium appears to have general benefits related to oxidative stress tolerance and greater resistance to tissue damage caused by pathogens and herbivores.

Some elements can absorbed in quantities that are toxic to plants. In acidic soils with water-soluble Al^{3+} levels of 10 to 20 mg kg^{-1} or more, plants can be experience aluminum toxicity. The decline in plant root functions is related to oxidative damage of cell membranes, and less cell division and elongation at the growing tips due to Al^{3+} replacement of Ca^{2+}. In sensitive plants, the presence of Al^{3+} in leaf tissues may reduce stomatal conductance and chlorophyll content, which reduces the photosynthesis rate. Plants grown in acidic soils are also prone to manganese toxicity when they absorb more Mn^{2+} than required for metabolic functions, resulting in oxidative stress and destruction of indole-3-acetic acid (auxin), an important plant growth hormone. Manganese toxicity symptoms are expected to be more severe in anoxic soils because oxygen limitation will reduce several low-solubility compounds, such as MnO_2, Mn_2O_3 and Mn_3O_4, into Mn^{2+}. Liming acidic soils with $CaCO_3$ or another neutralizing substance that raises the soil pH will increase quantity of oxides and hydroxides, which form precipitates with Al^{3+} and Mn^{2+}. Since precipitated aluminum oxides/hydroxides and manganese oxides are not water-soluble, the risk of plant toxicity diminishes with increasing soil pH. Installing drainage systems that increase the soil oxygen level should reduce Mn^{2+} toxicity in plants.

In salt-affected soils, an abundance of sodium (Na^+), SO_4^{2-} or Cl^- creates salt stress for plants. Due to osmosis, root cells tend to lose water when the Na^+, SO_4^{2-} and Cl^- levels are greater outside the root than inside the root. This can lead to physiological disorders within roots and other plant tissues, such as ion toxicity, altered metabolic processes, oxidative stress, membrane disruption, and less cell division and expansion. Consequently, plants require more water to grow in salt-affected soils. Transpiration is responsible for the movement of water and the accumulation of water-soluble salts near the root. Evaporation also causes salt movement into topsoil, particularly in semi-arid and arid climates where annual water loss is greater than annual precipitation. Soil management can be effective in removing excess Na^+ and Cl^- from the root zone. Sprinkler irrigation combined with improved drainage could leach these salts through the soil profile and away from plant roots. Salt-tolerant crops should be planted in salt-affected soils.

STRATEGIES FOR MANAGING SOIL NUTRIENTS

Soil fertility on farms cannot be managed without knowledge of the actual fertility status, based on observations of plant growth, yield measurements, plant tissue analysis and soil fertility testing in each field. A diagnostic evaluation of soil fertility also requires information about the amount of crop residues left in the field after the harvest, how much and how often agricultural lime and other soil amendments are spread on fields, and how much fertilizer is applied each year. We should also know the crop rotation, tillage, irrigation and drainage conditions in each field. This information can explain why soil organic matter and nutrient levels are increasing, stable or declining across the farm.

Although it is tempting to believe that we should always try to increase the soil organic matter and nutrient levels, this is not always achievable or desirable. There are physical limits to the amount of organic matter that can be retained in soil, related to climatic conditions (primarily temperature and rainfall) that determine the crop residues remaining after harvest, soil microbial activity and soil texture. Increasing the concentration of water-soluble nutrients in soils with high or excessively high fertility could increase the risk of nutrient loss to the environment through biological reactions (e.g., nitrogen loss in gaseous forms such as NH_3, N_2O and N_2)

and via leaching and erosion, which transport water-soluble NO_3^- and $H_2PO_4^-$ and particulate-associated nutrients to downstream environments. Consequently, we must use fertilizers and other amendments prudently and judiciously.

The value of measuring residual soil nitrate to assess nitrogen use *efficiency*

Residual soil nitrate at the end of the growing season is N that was not used by the crop. Thus, measuring how much nitrate remains in the soil profile following harvest provides a means of assessing how effectively we have managed N sources. In semi-arid and subhumid climates, this test is used to adjust the N fertilizer requirements for the next growing season. However, in cool humid agroecosystems where there is a high potential for non-growing season loss of nitrate, the residual soil nitrate test is an indicator of the risk of nitrate loss into aquatic systems. In Agriculture and Agri-Food Canada's Agri-environmental Indicator series (Clearwater et al., 2016) the difference between N inputs and N outputs is referred to as Residual Soil Nitrogen (RSN) and has been estimated for the agricultural regions of Canada (see Figure below).

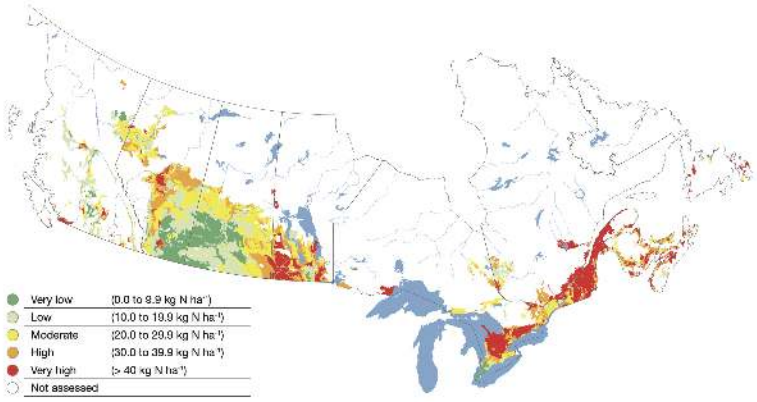


Figure 11-2). © Agriculture and Agri-Food Canada; used with permission and licensed under the Open Government Licence – Canada. https://open.canada.ca/en/open-government-licence-canada

This indicator points to potential for a large amount of nitrogen remaining in the soil in the fall after harvest in Eastern Canada with typical values exceeding 40 kg N ha^{-1}. Recent surveys of soil nitrate—measured to a depth of 45 cm—in Prince Edward Island suggest this estimate may indeed be conservative for potato production in that province with values averaging over 60 kg N ha^{-1}. The implementation of 4R practices reduced RSN in potato production in PEI by as much as 30%.

Since crops are the intended beneficiary of nutrient in agricultural lands, fertilizer should be applied in a way that optimizes nutrient acquisition by the crop. The 4R nutrient stewardship approach is a decision-making framework that guides farms in selecting the '**right**' actions related to fertilizer use, and is summarized below:

1. **Choose the right source of fertilizer** that will release water-soluble nutrients for the crop. The right source of fertilizer also depends on its availability and price. Fertilizers produced on farms, such as manure or compost, could be a readily-available, inexpensive fertilizer. Apply these fertilizers to meet crop nutrient requirements first, before purchasing commercial fertilizers that will supply the rest of the required nutrients. Is a fast-release fertilizer needed to overcome a crop nutrient deficiency? Consider a fertilizer source that dissolves in water. A controlled-release fertilizer, which is biologically or geochemically stabilized and releases water-soluble nutrients slowly, could be suitable for annual crops with a long growth period or perennial cropping systems.
2. **Choose the right rate of fertilizer** that permits your crop to reach its yield potential. When selecting the right rate, account for residual nutrients from decomposing crop residues and animal manure that were applied in previous growing seasons. Applying more nutrients than the crop can use will not improve the yield, although it could make weeds grow better. From an economic perspective, it is better to apply a little less fertilizer than you need, particularly when the growing conditions are sub-optimal due to a cold spring, flooding or drought. Applying more fertilizer during a poor growing season will not make the crop grow better and is a waste of money.
3. **Choose the right time to apply fertilizer** so that you can synchronize the supply of water-soluble nutrients with periods of high nutrient demand by the crop. Annual crops have an exponential growth phase for several weeks during their vegetative growth period, which slows down after floral initiation. In humid and irrigated soils, the best strategy is to apply fertilizer just before the exponential growth phase, so the plants will absorb nutrients efficiently. In some crops, fertilizer should be applied at

the beginning of the reproductive stage, to promote the development of high-quality seed. However, applying required fertilizers at the beginning of the growing season is a better strategy in semi-arid and subhumid regions because nutrient release is slow when precipitation occurs unevenly during the vegetative growth and initial reproductive stages. Perennial crops can be fertilized before they renew their vegetative growth each year. Perennial forages (i.e., hayfields) that are harvested two, three or four times a year can be fertilized after each successive cut, before the vegetative regrowth begins.

4. **Choose the right place to apply fertilizer** so that you can maximize uptake of water-soluble nutrients by the plant root. Precise placement is required to avoid direct contact of fertilizer salts with seeds and seedlings, but not so far that water-soluble nutrients leach from the root zone before they can be absorbed by root hairs. The most efficient methods for annual crops involve placing fertilizer in the planted row (i.e., between seeds or tubers) or beside the planted row so that transpiration aids the movement of nutrient ions through mass flow (e.g., NO_3^-, SO_4^{2-}) and diffusion (e.g., $H_2PO_4^-$, K^+) processes. Furthermore, as fertilizers dissolve or mineralize, they will release soluble nutrients that are accessible to the proliferating root system. Seeds can also be treated with fertilizer, as long as the fertilizer has a low salt content so there is no osmotic stress during germination and early seedling development. Foliar application is a rapid way to deliver small quantities of some micronutrient fertilizers, but there is a fine line between a micronutrient deficiency and a micronutrient toxicity in leaf tissues.

The 4R nutrient stewardship principles are the fundamentals of nutrient management on farms, but must be combined with soil and water conservation practices to be effective. The 4R Plus framework that is used on modern farms connects 4R nutrient management and soil and water conservation practices. The goal of the 4R Plus approach is to provide nutrients when the crop needs them, to enhance soil fertility and to improve water quality. Soil and water conservation practices that can enhance soil fertility and nutrient management include:

1. Reduce tillage to minimize soil disturbance and improve soil structure.
2. Leave crop residue on the soil surface to protect it from the impact of rainfall.
3. Grow cover crops to protect the soil and support the soil microbial community between cropping seasons.
4. Apply adequate fertilizer and manure using the 4R principles of right source, right rate, right time and right place.
5. Value manure applications, since manure can enhance soil biology and soil fertility.
6. Control field traffic to reduce soil compaction.
7. Extend crop rotations and include perennial crops to build soil organic matter.
8. Establish contour buffer strips, field buffers and stream buffers to protect sensitive environments within the field and in downstream environments.
9. Highly erodible areas can be planted with perennial crops or with trees, rather than annual crops.
10. Use prescribed or rotational grazing to minimize soil erosion.

Impact of long-term tillage and synthetic fertilization on soil parameters and crop yields under a corn–soybean rotation in eastern Canada

In eastern Canada, the corn (Zea mays L.)–soybean [Glycine max (L.) Merr.] rotation represents a dominant cereal–legume association. Although the economic benefits of this rotation are recognized, research is required to evaluate its long-term impact on soil quality and crop productivity under varying management alternatives such as reduced tillage or varying rates of synthetic fertilization. Indeed, the fundamental role of long-term field experiments in agronomic research and development has been recognized for decades by the international scientific community as essential for assessing the effects of management practices on soil fertility, soil quality, and crop productivity.

In 1992, a long-term research experiment was established in southern Quebec on a clay loam to evaluate the effect of tillage [mouldboard plow (MP) vs. no-till (NT)], synthetic N (0, 80, and 160 kg N ha^{-1}) and P fertilization (0, 17.5, and 35 kg P ha^{-1}) on soil functioning and crop yields of a grain corn–soybean crop sequence. Synthetic fertilizers were applied only to the corn phase of the rotation, while tillage was performed each year.

Results obtained 12 to 20 years after initiation of the study indicated that NT treatment enhanced organic C accumulation, NO_3-N, P and K availability, and microbial activity in the upper soil layer likely due to accumulation of crop residues at or near the soil surface. This stratification can benefit soil structure but might impact surface water quality over the long term. Conversely, the MP treatment was characterized by a more uniform distribution of organic C and nutrients to tillage depth and higher microbial activity near the bottom of the plow layer. Strong links were established between soil tests P (e.g. Mehlich-3 and Olsen P) and P balance, explaining how P can accumulate in time with continuous agricultural practices. Soil N_2O emissions were increased by N

fertilization in corn but not affected by tillage. Grain corn yield linearly responded to fertilizer N addition in all years but soybean yield was little affected by residual N. Except for soil P status, fertilizer P had limited impact on both crops on this site but lack of replenishment would eventually lead to soil P depletion. Overall, MP gave greater yields than CT in several years (see above Figure)—by an average of 10% in corn and 13% in soybean—and provided generally a better profitability. The lower performance of CT in this site compared with MP is probably because cooler and wetter soil conditions in spring due to surface crop residues.

***Note:** this site is still operating and other studies are under investigation. For more details see Ziadi et al. (2014).*

In the future, soil fertility managers will rely more on precision agriculture technologies to achieve the 4R Plus goals. The greatest advantage of precision agriculture, from the perspective of managing nutrients, is that it allows for the development of spatially-explicit maps of the soil and hydrological conditions in agricultural fields. The multiple layers of information in these maps permit us to evaluate individual parameters, such as soil pH or the soil potassium concentration across a field, or we can evaluate the interactions between multiple parameters (e.g., are changes in soil pH associated with fluctuations in the water table level across the field?). Rather than relying on an average soil fertility value for a large field, we can observe gradients in soil fertility that exist across the field. Knowing the zones of low, medium and high fertility allows us to adjust fertilizer application rates and meet the needs of the crop growing in a particular zone, rather than applying a uniform rate of fertilizer across the field. Soil and hydrological maps can also identify the vulnerable areas in fields where leaching or erosion are more likely to occur. Hence, the farm that adopts precision agriculture technologies can save money by applying fertilizers precisely, where they are needed, and establish buffer zones in the sensitive areas of the field and at the interface between fields and non-agricultural land. Precision agriculture technologies are well aligned with the 4R Plus goals and should be of great benefit to manage soil fertility in the future.

SUMMARY

1. Water-soluble nutrient ions in the soil pore water are absorbed through root hairs of plants.
2. Plant-available nutrients in the soil pore water can be transformed into unavailable forms through biological reactions, such as plant uptake, immobilization and reduction. Chemical precipitation and adsorption reactions also remove plant-available nutrients from the soil pore water. Physical loss of plant-available nutrients occurs when soils are leached and eroded.
3. Soil pore water is replenished with plant-available nutrients through mineralization, oxidation, weathering, dissolution and desorption reactions.
4. Soil fertility testing describes the soil nutrient supply and is required to make fertilizer recommendations. Crop nutrition is explained by plant tissue analysis, which determines whether the crop has sufficient nutrients or is deficient in nutrients required to achieve the economically optimal yield.
5. Inorganic and organic fertilizers can be applied to replenish the soil nutrient supply and improve crop nutrition. Dissolution of inorganic fertilizers releases water-soluble nutrient ions quickly, whereas organic fertilizers contain some organic compounds that must be mineralized to release water-soluble nutrient ions, making them a relatively slow-release fertilizer.
6. The 4R Nutrient Stewardship approach to soil nutrient management focuses on selecting the right source of fertilizer, to be applied at the right rate, at the right time and in the right place to get the greatest percentage of fertilizer nutrients into crop biomass. Achieving a high efficiency of fertilizer nutrient use in a cost-effective way is the goal.
7. Soil nutrients are susceptible to loss from the soil-plant system, so farms should be aware of the environmental implications of soil nutrient management.
8. The “Can you dig it?” boxes provide examples of soil fertility testing and soil nutrient improvement in Canada.

SUGGESTED READING

International Plant Nutrition Institute. 2012. 4R plant nutrition: a manual for improving the management of plant nutrition. North American version. IPNI, Norcross, GA, USA.

Havlin, J.L., J.D. Beaton, S.L. Tisdale, and W.L. Nelson. 2014. Soil fertility and fertilizers: an introduction to nutrient management. 8th ed. Pearson Prentice Hall, Upper Saddle River, NJ.

Munroe, J. (ed). 2016. Soil fertility handbook. 3rd edn. Ontario Ministry of Agriculture, Food and Rural Affairs Publication 611. Queen’s Printer for Ontario, Toronto. http://www.omafra.gov.on.ca/english/crops/pub611/pub611.pdf

Reid, K. 2014. Improving your soil: A practical guide to soil management for the serious home gardener. Firefly Books Ltd., Richmond Hill, Ontario, 272 p.

STUDY QUESTIONS

1. How do soil nutrients get absorbed by plants?
2. Complete the following table.

Nutrient	Macronutrient or Micronutrient?	Source in soil?	Role in plant nutrition
N			
P			
Ca			
S			
Cu			
Fe			

3. Give an example of a biological reaction that can deplete the amount of plant-available nitrogen in the soil pore water.
4. What is a chemical reaction that reduce the amount of plant-available potassium in the soil pore water?
5. Why does the application of a fertilizer containing NH_3 tend to acidify the soil?
6. Nitrogen is often the most limiting nutrient for agricultural crops to reach their yield potential, so is it a good idea to add excessively high rates of nitrogen fertilizer to agricultural land? Why or why not?
7. What is the best diagnostic tool to determine the fertilizer requirements of crops: soil fertility testing or plant tissue analysis?
8. Micronutrient fertilizers can be applied to crop foliage. What is the advantage of this fertilization method?
9. What are some considerations to help you choose the right fertilizer source for your crop?
10. What are suitable soil and water conservation practices to prevent nutrient loss through erosion?

REFERENCES

Abdi, D., Cade-Menun, B.J., Ziadi, N., and Parent, L.-É. 2014. Long-term impact of tillage practices and phosphorus fertilization on soil phosphorus forms as determined by 31P nuclear magnetic resonance spectroscopy. J. Environ. Qual. 43(4): 1431–1441.

Bender, R.R., J.W. Haegele, M.L. Ruffo, and F.E. Below. 2010. Nutrient uptake, partitioning, and remobilization in modern, transgenic insect-protected maize hybrids. Agron. J. 105(1): 161–170.

Carter, M.R. and Gregorich, E.G. 2008. Soil sampling and methods of analysis. 2nd ed. Canadian Society of Soil Science. CRC Press, Boca Raton, FL.

Centre de référence en agriculture et agroalimentaire du Québec. 2010. Guide de référence en fertilisation. 2e édn. Centre de référence en agriculture et agroalimentaire du Québec, Ste Foy, QC.

Chantigny, M.H., Angers, D.A., Morvan, T., and Pomar, C. 2004. Dynamics of pig slurry nitrogen in soil and plant as determined with N-15. Soil Sci. Soc. Am. J. 68(2): 637–643.

Deen, B., Lauzon, J., and Bruulsema, T. 2007. Corn response to intensive crop nutrition. Better Crops 91: 20–21. http://www.ipni.net/publication/bett...48874B7A852579 800081D5CE/$FILE/Better%20Crops%202007-2%20p20.pdf

Dessureault-Rompré, J., Zebarth, B.J., Chow, T.L., Burton, D.L., Sharifi, M., Georgallas, A., Porter, G., Moreau, G., Leclerc, Y., Arsenault, W.J., and Grant, C.A. 2011. Prediction of soil nitrogen supply in potato fields in a cool humid climate. Soil Sci. Soc. Am. J. 75(2): 626–637.

Drury, C.F., Yang, J., De Jong, R., Huffman, T., Reid, K., Yang, X., Bittman, S., and Desjardins, R. 2016. Residual Soil Nitrogen. Pages 114–120 In R.L. Clearwater, T. Martin, and T. Hoppe (Eds.) Environmental Sustainability of Canadian Agriculture: Agri-Environmental Indicators Report Series. Report #4. Ottawa, ON: Agriculture and Agri-Food Canada

Elbon A., and Whalen, J.K. 2015. Phosphorus supply to vegetable crops from vesicular arbuscular mycorrhizae: a review. Biol. Agric. Hort. 31(2): 73–90.

Eskin, N., Vessey, K., and Tian, L. 2014. Research progress and perspectives of nitrogen fixing bacterium, Gluconacetobacter diazotrophicus, in monocot plants. Int. J. Agron. 2014, 13 p. Article 208383. https://doi.org/10.1155/2014/208383

Evans, J.R. 1989. Photosynthesis and nitrogen relationships in leaves of C_3 plants. Oecologia 78(1): 9–19.

Fleury, D. 2020. 4R nitrogen management for wheat and canola. Top Crop Manager 19 August 2020 https://www.topcropmanager.com/4r-ni...at-and-canola/

Follett, R.H., Murphy, L.S., and Donahue, R.L. 1981. Fertilizers and soil amendments. Prentice-Hall, Inc., Englewood Cliffs, NJ.

Havlin, J.L., Beaton, J.D., Tisdale, S.L., and Nelson, W.L. 2014. Soil fertility and fertilizers: an introduction to nutrient management. 8th ed. Pearson Prentice Hall, Upper Saddle River, NJ.

He, Z.Q., Honeycutt, C.W., Griffin, T.S., Cade-Menun, B.J., Pellechia, P.J., and Dou, Z.X. 2009. Phosphorus forms in conventional and organic dairy manure identified by solution and solid state P-31 NMR spectroscopy. J. Environ. Qual. 38(5): 1909–1918.

International Plant Nutrition Institute. 2012. 4R plant nutrition: a manual for improving the management of plant nutrition. North American version. IPNI, Norcross, GA, USA.

Jones, J.B., Jr. 2003. Agronomic handbook: management of crops, soils and their fertility. CRC Press, Boca Raton, FL.

Karamanos, R.E. 2008. Nitrogen and phosphorus recommendations for Manitoba for today's production costs. Presented at the 2008 Manitoba Agronomists Conference, Winnipeg, MB
http://umanitoba.ca/faculties/afs/MA...is%20Karamanos %20paper.pdf

Kour, D. Rana, K.L., Kaur, T., Yadav, N., Yadav, A.N., Kumar, M., Kumar, V., Dhaliwal, H.S., and Saxena, A.K. 2021. Biodiversity, current developments and potential biotechnological applications of phosphorus-solubilizing and -mobilizing microbes: A review. Pedosphere 31(1): 43–75.

Kumar, V., Vogelsang, L., Schmidt, R.R., Sharma, S.S., Seidel, T., and Dietz, K.-J. 2020. Remodeling of root growth under combined arsenic and hypoxia stress is linked to nutrient deprivation. Frontiers in Plant Science. 23 October 2020. https://doi.org/10.3389/fpls.2020.569687

Leytem, A.B., Smith, D.R., Applegate, T.J., and Thacker, P.A. 2006. The influence of manure phytic acid on phosphorus solubility in calcareous soils. Soil Sci. Soc. Am. J. 70(5): 1629–1638.

Marty, C., Lévesque, J.-A., Bradley, R.L., Lafond J., and Paré, M.C. 2019. Contrasting impacts of two weed species on lowbush blueberry fertilizer nitrogen uptake in a commercial field. PLoS ONE 14: e0215253.

Mehlich, A. 1984. Mehlich-3 soil test extractant: a modification of Mehlich-2 extractant. Commun. Soil Sci. Plant Anal. 15(12): 1409–1416.

Mortvedt, J.J., Cox, F.R., Shuman, L.M., and Welch, R.M. 1991. Micronutrients in agriculture. SSSA Book Series 4, American Society of Agronomy, Madison, WI.

Munroe, J. (ed). 2016. Soil fertility handbook. 3rd edn. Ontario Ministry of Agriculture, Food and Rural Affairs Publication 611. Queen's Printer for Ontario, Toronto. http://www.omafra.gov.on.ca/english/...611/pub611.pdf

Ontario Ministry of Agriculture, Food and Rural Affairs (OMAFRA). 2006. Nutrient management planning. Best Management Practices Series, Guelph, ON.

Rascio, N., and La Rocca, N. 2013. Biological nitrogen fixation. Reference Module in Earth Systems and Environmental Sciences. Elsevier, Amsterdam. doi:10.1016/B978-0-12-409548-9.00685-0

Reid, R., and Hayes, J. 2003. Mechanisms and control of nutrient uptake in plants. Int. Rev. Cytol. 29: 73–114. doi: 10.1016/s0074-7696(03)29003-3.

Ruiz Herrera, L.F., Shane, M.W., and López-Bucio, J. 2015. Nutritional regulation of root development. WIREs Dev. Biol. 4: 431–443. doi: 10.1002/wdev.183

Sims, J.T., Maguire, R.O., Leytem, A.B., Gartley, K.L., and Pautler, M.C.. 2002. Evaluation of Mehlich 3 as an agri-environmental soil phosphorus test for the Mid-Atlantic United States of America. Soil Sci. Soc. Am. J. 66(6): 2016–2032.

Stumborg, C., Schoenau, J.J., and Malhi, S.S. 2007. Nitrogen balance and accumulation pattern in three contrasting prairie soils receiving repeated applications of liquid swine and solid cattle manure. Nutr. Cycl. Agroecosyst. 78: 15–25.

Westerman, R.L. 1990. Soil testing and plant analysis. 3rd edn. SSSA Book Series 3, American Society of Agronomy, Madison, WI.

Whalen, J.K. 2014. Managing soil biota-mediated decomposition and nutrient mineralization in sustainable agroecosystems. Advances in Agriculture 2014: 13 p. Article 384604. doi: 10.1155/2014/384604

Whalen, J.K., Thomas, B.W. and Sharifi, M. 2019. Novel practices and smart technologies to maximize the nitrogen fertilizer value of manure for crop production in cold humid temperate regions. Adv. Agron. 153: 1–85.

Zebarth, B.J., Leclerc, Y., Moreau, G., Gareau, R., and Milburn, P.H. 2003. Soil inorganic nitrogen content in commercial potato fields in New Brunswick. Can. J. Soil Sci. 83(4): 425–429.

Ziadi, N., Angers, D.A., Gagnon, B., Lalande, R., Morel, C., Rochette, P., and Chantigny, M.H. 2014. Long-term tillage and synthetic fertilization affect soil functioning and crop yields in a corn–soybean rotation in eastern Canada. Can. J. Soil Sci. 94 (3): 365–376.

Ziadi, N., Simard, R.R., Allard, G., and Lafond, J. 1999. Field evaluation of anion exchange membranes as a N soil testing method for grasslands. Can. J. Soil Sci. 79(2): 281–294.

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Joann is a Chief Editor for Soil Biology and Biochemistry, an academic journal that publishes original, scientifically challenging research articles of international significance that describe and explain biological processes occurring in soil. She is a Fellow of the Soil Science Society of America and of the Canadian Society of Soil Science. Joann is the first international scientist to be elected for a 3-year term as President of the American Society of Agronomy (2022-2024). She grew up in a farming community near Sussex, New Brunswick and has always been interested in the practical aspects of growing food. In her opinion, soil fertility specialists are the most broadly-trained agricultural scientists because they have to muster knowledge from so many fields – physics, hydrology, chemistry, biochemistry, plant biology, microbiology and genetics – to understand the complex and dynamic soil processes responsible for crop growth. To make sense of soil responses, we have to dig deeper, which comes quite naturally to a soil scientist!

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Noura has a distinguished career in soil science especially soil fertility and plant nutrition and she's conducting her research at the national and international scales (France, China, Switzerland, Finland, Tunisia, Iran, Saudi Arabia) within multi-disciplinary teams including government and university researchers and industry partners. Noura published more than 155 papers, five reviews and 11 book chapters including 37 papers with the Canadian Journal of Soil Science (CJSS). Noura joined the CSSS in 1994 and has served as Eastern Councillor (2008-2010), President-Elect (2017), President (2018) and Past-President (2019). Noura is an associate editor (AE) for the CJSS since 2011, has handled 107 manuscripts for this journal and was been recognized as a best AE in 2017. Noura has been recognized for her contributions to soil science and to the CSSS with a Fellow Award in 2015.

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I teach soil science to undergraduate students, most of whom are enrolled in the biology program at UQAC. I teach soil science as an introductory class, and we look at soil chemistry, biology, nutrient cycling, and pedology. It is always interesting how initially most students view soil science negatively—I usually do a small survey during the first class of the semester—and how their views change over the semester as the students realize the importance of soil for biology, life, and the general well-being of the planet. I

also include soil science in most of my field teaching activities. I like to show the links between spheres, such as how what we see when we look up, e.g., the traits of plants and trees, strongly influences what we find when we look down, i.e., in the soil, and vice versa. Finally, soil science allows me to do research and support many graduate students across a variety of research disciplines, from various agricultural domains (wild blueberry, haskap, potatoes, cash crops, etc.) to forestry and the environmental sciences. How ever could I be more stimulated?

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Since July 2019, Tom Bruulsema is Chief Scientist with Plant Nutrition Canada, providing support for the nutrient stewardship programs of Fertilizer Canada, The Fertilizer Institute, and the International Fertilizer Association. Prior to the dissolution of the larger IPNI, he was vice-president for the Institute's programs in the Americas, and research globally. From 2015 to 2017 he led the Institute's Phosphorus Program. Dr. Bruulsema has been recognized as a Fellow of the American Society of Agronomy, the Soil Science Society of America, and the Canadian Society of Agronomy. He was previously a research associate at the University of Minnesota (1994), and an agronomist with the Mennonite Central Committee in Bangladesh (1986-1990).

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1.8: Soil Classification and Distribution

Dan Pennock

Learning Objectives

On completion of this chapter, students will be able to:

1. Explain the role that diagnostic horizons play in classification of soils under the Canadian System of Soil Classification (CSSC)
2. List the seven soil-forming factors that control the distribution of soils at national and local levels
3. Explain how the CSSC uses soil pH to distinguish between acidic parent materials derived from igneous rocks and neutral-alkaline parent materials derived from sedimentary rocks
4. Identify the three main soil orders (Brunisolic, Podzolic, and Luvisolic) typically associated with forested landscapes in Canada
5. Understand the effect of permafrost layers on Arctic and tundra soils of the Cryosolic order
6. Identify the three main soil orders (Chernozemic, Solonetzic, Vertisolic) typically associated with grassland landscapes
7. Explain how the two orders of wetland soils (Gleysolic, Organic) differ
8. Understand how the usefulness of a given soil survey for land use planning is determined by the scale of the survey
9. Identify the major classes used in the Canada Land Inventory

INTRODUCTION

Classification and mapping of soils in Canada was a major activity in soil science throughout the latter half of the 20th century and the maps that resulted from this program are still widely used in many sectors. As well, mapping of soils at intensive scales for activities such as pipeline development, mine reclamation, and soil contamination remediation are major activities by private sector consultants in Canada. Hence, an understanding of the classification system used in Canada and how soil maps can be used is essential for many soil scientists.

The placement of soils into the correct class is based on the horizons that we can see in the soil profile. Distinctive sets of soil horizons result from the response of the soil-forming processes (discussed in Chapter 2) to soil forming factors. These sets of soil horizons differ from point to point over short distances – within a field, for example – and even more so over large distances such as a county, rural municipality, or province. The sets of horizons are the basis for classifying individual soils into the taxonomic system used in Canada, the Canadian System of Soil Classification (CSSC) (Soil Classification Working Group, 1998). Moreover, the distribution of soil horizons and soil properties is not random – they are responding to a set of environmental controls and once we understand the controls we can predict the distribution of soils and hence map their distribution.

Soil maps are a key source of information in many types of land use planning. The usefulness of a map for any particular purpose depends on the scale of the map – the more detailed the scale, the more certain we can be that map accurately reflects the actual distribution of soils at the site. Evaluating the reliability of soil maps (or surveys) requires a good understanding of the interplay between map scale and map accuracy. Digital Soil Mapping, which draws upon sophisticated computer and remote sensing tools, is increasingly being used to produce soil maps at all scales. Digital map products allow the accuracy of the map to be reliably estimated (see Chapter 17 for more information on Digital Soil Mapping).

SOIL HORIZONS AND THE PEDON

An initial stage in the evolution of any natural science is the naming (or classification) of the objects under study. Often there is an informal name that will differ regionally or in different language groups (e.g., the Blue Jay or Jack Pine) and a formal, scientific name that is used internationally (at the species level, C*yanocitta cristata* and *Pinus banksiana*). The classification system used to identify and name the objects is formally called a taxonomic system.

Soils are unlike birds or trees insofar as there are no naturally occurring soil individuals (such as an individual tree or bird). The soil forms a layer on the Earth's surface whose properties vary continuously through space – the soil at any given location will be different (although often in very subtle ways) from the soil even a footstep away. Moreover, only the uppermost millimeter of the

soil is even visible to us – the vast majority of the soil is underneath the ground surfaces where it is invisible to even sophisticated remote sensing devices.

Given this continuous variation is soil properties, the definition of a basic soil individual is therefore arbitrary. For the purposes of classification soil scientists developed the concept of a pedon – formally defined in Canada as:

"It [the **pedon**] is the smallest, three-dimensional unit at the surface of the earth that is considered as a soil. Its lateral dimensions are 1 m if ordered variation in genetic horizons can be sampled within that distance or if these horizons are few and faintly expressed." (Soil Classification Working group, 1998; p.9)

Typically, the way we observe the soil is to dig a hole in the ground, which is referred to as a soil pit. The pit is the means for describing the soil profile and for taking samples from the soil which we analyze in the laboratory.

Mineral and Organic Soil Horizons

As we learned in Chapter 2, soils are composed of one or more distinct layers, which are formally called soil horizons. These layers may differ from each other in terms of their colour, structure, texture (the particles which compose it) or many other properties. In every soil classification system, these layers are assigned distinctive alphabetic symbols as a form of shorthand for their characteristics.

The horizon description system begins by splitting soil horizons into two distinct groups: mineral and organic horizons. **Organic horizons** are those which contain 17% or more organic carbon by weight; **mineral horizons** have less than 17% organic carbon by weight.

The three master mineral horizons are the A, B, and C horizons.

A horizon: The A horizon occurs at or near the soil surface. In the CSSC two very distinct bundles of processes can form A horizons. The first is the addition, mixing, and transformation of organic matter (OM) to form an organically enriched mineral horizon (Ah). The second bundle involves the bleaching of the horizon by acids and (in many cases) the loss of mobile soil components such as clay, OM, iron, and aluminum. The loss of soil material is called eluviation, and these are classified as Ae horizons. The CSSC is unusual for having two very different types of A horizons – in the major international soil classification systems, the eluvial horizon is assigned an uppercase E, not an Ae, label as is used in the CSSC.

B horizon: The B horizon is a layer that has undergone substantial change from the parent material. In aerobic soils this may be due to the formation of red surface coatings on mineral grains, deposition of materials (clay, OM, iron, aluminum) from the Ae, loss of highly weatherable carbonate or salts, or the formation of distinct soil structural units. In anaerobic soils, water saturation can lead to processes that form the gley soil properties, such as dull coloration or the presence of mottles.

C horizon: In many cases (especially in parent materials high in sand) the C horizon has experienced little or no soil formation and closely resembles the parent material the soil formed in (assuming that no layering of parent materials occurs in the profile). In soils formed in carbonate or salt-rich parent materials, dissolved ions from these minerals that originate from the A and B horizons may re-form new, secondary minerals in the C horizon.

The main organic soil horizons are the **O horizons** and the **LFH horizons**. The O horizons are organic layers formed in bogs, fens, swamps, and marshes and are informally called peat layers. They are further subdivided based on their degree of decomposition into highly fibrous layers of recognizable plant origin (Of for fibric), layers of intermediate decomposition (Om for mesic), and highly decomposed material (Oh for humic).

The LFH horizons occur from the accumulation of leaves, twigs and woody materials on the mineral soil surface in forests or shrublands and which overlies a mineral soil. The L layer is from readily recognizable leaf litter, the F from moderately decomposed organic material (known as folic material), and the H from highly altered humic material.

THE CANADIAN SYSTEM OF SOIL CLASSIFICATION

Taxonomy is the science of classification, and the CSSC is a rigorous taxonomic system. The System is a comprehensive method for assigning pedons or profiles to the soil classes; if the system is correctly used, a pedon can only be assigned to one specific class. The System is also a hierarchical system: each group occupies a distinct position within the overall system. The major levels

used in the classification are shown in Table 8.1. The first three levels (order, great group and subgroup) are used in soil classification in the field whereas the final three (family, series, and phase) are primarily used in soil mapping.

Table 8.1. Taxonomic levels in the Canadian System of Soil Classification ordered from highest (the most general) to lowest (the most specific)

Class	Description
Order	Orders are based on properties of the profile that reflect the nature of the overall soil environment and the effects of the dominant soil-forming process.
Great Group	Great groups are subdivisions of each order. The great groups reflect differences in the strengths of the major processes or a major contribution of a process in addition to the dominant one.
Subgroup	Subgroups are subdivisions of great groups. Profiles are differentiated on (1) how closely they correspond to the central concept of the great group (termed Orthic); (2) transitional stages grading towards soils of another order; and (3) additional special features within the profile.
Family	Families are used to group series (see below) of the same subgroup that have similar response to management, engineering and related uses. The family level is seldom used in Canada.

The assignment of a given soil profile to its taxonomic class is done by comparing the properties of the horizons in the field to a list of distinctive characteristics, called **diagnostic properties**. These diagnostic properties are the basis for classification of any pedon into the formal classes of the CSSC (Table 8.2).

Table 8.2. Diagnostic horizons and main characteristics of the ten orders in the Canadian System of Soil Classification

Order	Diagnostic	Nature of the soil environment and dominant soil-forming process
	Horizon	
Chernozemic	Ah (humus),	Grassland soil whose diagnostic horizon is formed by high levels of organic matter additions from the roots of grasses and turbation by burrowing animals.
	Ap (ploughing)	
	Ahe (humus + eluviation)	
Solonetzic	Bn (sodium) or Bnt (sodium + clay)	Grassland soil with high sodium levels in the B horizon; usually associated with a clay-rich B horizon and often with saline C horizon material.
Podzolic	Bf (iron) or	Forest soil normally associated with coniferous vegetation on igneous-rock derived parent materials. High acidity in the A horizon results in formation of a bleached Ae horizon and formation of organic-mineral complexes that are transferred to the B horizon.
	Bh (humus)	
Luvisolic	Bt (clay)	Forest soil found in areas with parent materials derived from sedimentary rocks. Dominant property is higher clay content in the B than the Ae horizon caused (in part) by loss of clay from the Ae horizon and its deposition in the Bt horizon.
Brunisolic	Bm (minimal transformation)	Forest soil whose properties are not strongly enough developed to meet the criteria for the Luvisolic or Podzolic Orders.
Gleysolic	Bg, Cg (gleying)	Wetland soil found throughout Canada wherever temporary or permanent water saturation cause formation of gleyed features (mottles, blue-grey colours) in the profile.
Regosolic	No B horizon or B horizon less than 5 cm thick	A minimally developed soil found throughout Canada wherever conditions prevent the formation of B horizons (unstable slopes, sand dunes, floodplains etc.).
Vertisolic	Bss or Css	Grassland soil associated with high clay glacio-lacustrine landscapes; characterized by shrinking and swelling of clays.

Order	Diagnostic	Nature of the soil environment and dominant soil-forming process
	(slickensides)	
	and Bv (vertic mixing)	
Cryosolic	By, Cy, (cryoturbation)	Soil of arctic and tundra regions with permanently frozen layers in the soil profile.
	Cz (permafrost)	
Organic	O (Organic)	Wetland soils associated with the accumulation of organic materials (peat) in water-saturated conditions.They are most commonly associated with the Boreal forest and tundra regions.

ENVIRONMENTAL CONTROLS ON SOIL DISTRIBUTION

In the middle of the 20th century several pedologists independently developed functional relationships between the properties of soils and the environmental controls (or factors) on the processes that form the properties. The best known of these relationships is by the American soil scientist Hans Jenny in his famous book the *Factors of Soil Formation* (1941). At about the same time, the great Canadian pedologist Joe Ellis identified seven factors in his book the *Soils of Manitoba* (1939). The work of the two authors overlapped for several of the factors but each statement had unique elements and combining them gives us a comprehensive set of **soil-forming factors**:

1. **Climate:** the temperature and moisture within the soil.
2. **Organisms:** the vegetation, animals, and microbes that affect the soil.
3. **Parent material:** the mineral matter in which the soil is formed, and which in turn affect the texture, water retention capacity, and the mineral reserve.
4. **Topography (or relief):** the shape of the land surface.
5. **Time:** the duration of soil formation.
6. **Groundwater:** the presence or absence of groundwater within the soil profile.
7. **Human activity:** the soil-modifying effects of human use of the soil.

Note that groundwater and human activity are the two soil-forming factors that make a difference between Ellis' and Jenny's groupings of these factors. In Chapter 2 we saw how five of these factors (climate, organisms, parent material, time, and human activity) have controlled the main bundles of soil-forming processes responsible for Canadian soils at the national level. The result of these bundles of soil-forming factors is to create a geographic zonation of soil orders in the different regions of Canada (Figure 8.1).

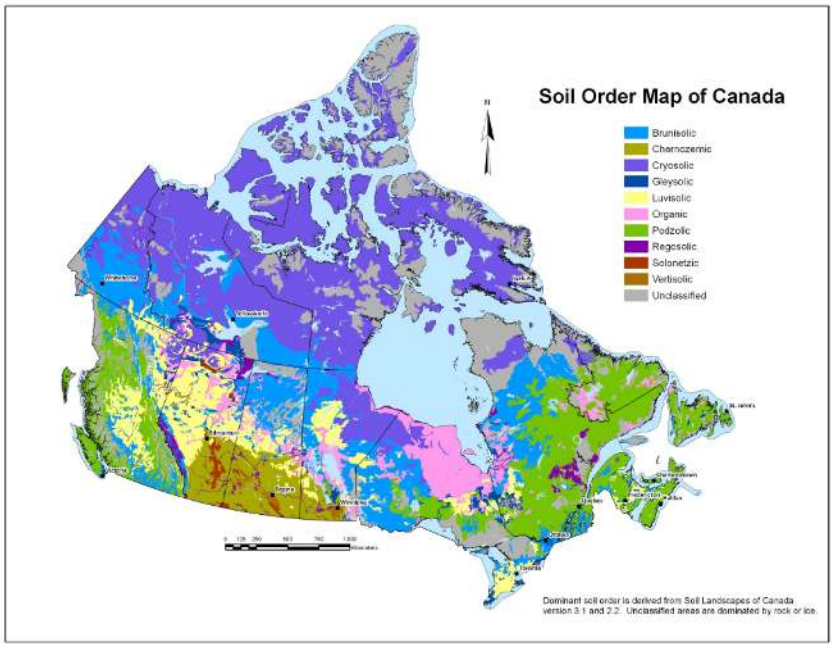


Figure 8.1. Map of soil orders in the Canadian System of Soil Classification. © Agriculture and Agri-Food Canada; used with permission and licensed under the Open Government Licence – Canada.https://open.canada.ca/en/open-government-licence-canada.

Local controls on soil distribution: the soil catena

At the more local level – for example in an individual hillslope or hillslope catchment – topography and groundwater have a major influence on the distribution of individual soil classes. As we learned in Chapter 2 much of the local topography in Canada is the result of glacial or post-glacial geomorphic processes and the basic shape of landforms does not change on human-time scales. The shape of the land surface has a very strong influence on the pattern of water movement both on the surface of the soil and laterally within the soil and this water movement and concentration is the dominant control on the local-scale pattern of soil formation (Figure 8.2).

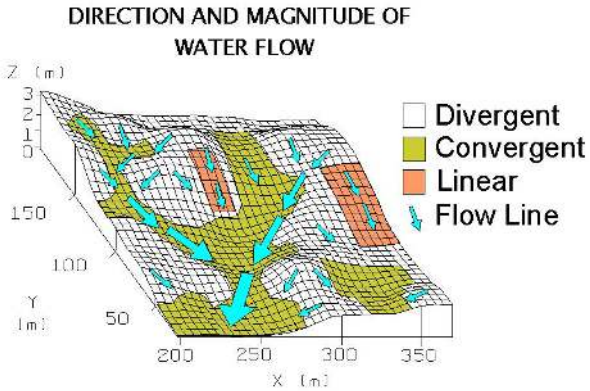


Figure 8.2. Schematic diagram of water flow in a hillslope. Slope segments where water flow diverges will be drier than the hillslope average, and segments where flow concentrates will by moister than the average. © Dan Pennock, Univ. of Saskatchewan, is licensed under a CC BY (Attribution) license.

For dominantly aerobic soils (i.e., those with air-filled pore space) the greater the amount of water present in the soil profile, the greater the action of the soil-forming processes discussed in Chapter 2. For example, hydrolysis (literally the splitting of water molecule) releases the hydrogen ions that drive much of chemical weathering; hence, the greater the water available in the profile, the greater the chemical weathering that occurs. Water also transports mobile ions and dissolved OM out of the profile through the process of leaching; again, the greater the water flowing through the soil, the greater the leaching of solutes from the soil. Finally,

microbial decomposition of OM (and the formation of humus) is greatest where optimum soil moisture and soil temperature conditions occur; drier soil moisture levels reduce the activity of the microbial community.

If, however, the pore space in soil profile is water-filled for extended periods of time, the soil becomes dominantly anaerobic and the chemical regime becomes dominated by redox processes (Chapter 5). Under those conditions, soil formation follows the gleyed pathway – reduction and mobilization of iron, formation of blue-grey soil colours, and slower (or no) decomposition of OM added to the soil.

Water movement in hillslopes can cause a major range in soil moisture conditions (and hence different soil classes) to occur over short distances (i.e., 10 m to a few hundred meters) (Figure 8.3). Generally upper slope positions shed moisture, and soil development will be minimally expressed in these positions. As soil moisture increases downslope, the degree of soil formation (expressed through soil thickness and development of soil horizons) will increase. Finally, a threshold may be passed where the soil becomes dominantly anaerobic and the gleying processes become dominant. This pattern of soil development along hillslopes repeats itself over many landforms, and the pattern is called a soil **catena** (*catena* is Latin for chain) – literally a chain of soils along the slope, genetically linked to one another.

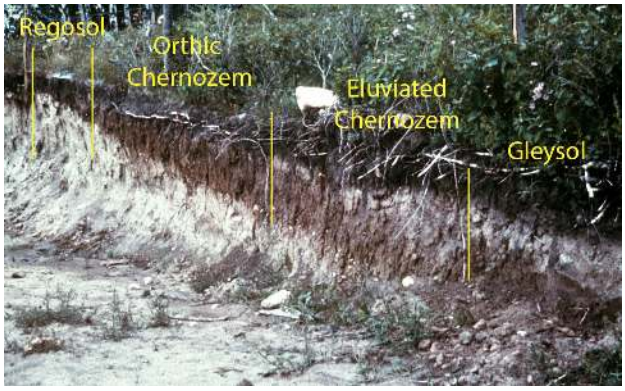


Figure 8.3. Soil catena exposed along a hillslope near Saskatoon, Saskatchewan. At the top of the slope thin soils (with only Ah/Cca/Ck horizons) of the Regosolic soil occur. The B horizon thickens downslope, and Ah/Bm/Cca/Ck horizons of the true (or Orthic) Chernozems occur. Towards the base of the slope a grayish, Ae horizon begins to develop (the Eluviated Chernozems) and finally Gleysolic soils (with Bg horizons) occur at the bottom of the slope. © Don Acton (Saskatchewan Centre for Soil Research) is licensed under a CC BY (Attribution) license.

As well as water transport within the soil there are also flows of water through sediments and even bedrock deep in the Earth's surface. These groundwater flow systems can transport water (and solutes) over distances of hundreds of meters to many kilometers. In some topographical settings the pressures within the groundwater system cause the groundwater to move vertically upwards (or discharge) towards the soil surface. The groundwater can then be lost to the atmosphere due to evapotranspiration, and often any solutes in the groundwater precipitate out to form new mineral salts and carbonates. This is a major source of salinity in Canadian Prairiers. In areas of Canada with well developed river drainage systems, the groundwater more typically discharges into the stream system and soils affected by discharge are less common.

FOREST SOILS

Luvisols, Podzols, and Brunisols

The key taxonomic distinction in forested areas is between soils formed in generally coarse-textured glacial parent materials derived from acidic, base-poor (e.g., lacking magnesium, calcium, potassium and sodium) igneous rocks and those formed on glacial and post-glacial parent materials derived from high base-status sedimentary rocks. In the CSSC, the criterion used to distinguish between soils of the two types of parent materials is the soil pH (as measured using 0.01 M calcium chloride; see Chapter 5 for more information about soil pH and acidity). At the national level, soils developed on the igneous-rock derived parent materials have an acid pH < 5.5; soils on sedimentary rock parent materials have pH values in the neutral-alkaline range (i.e., ≥ 5.5). For the purposes of this chapter the soils in each group will be referred to by their pH range (i.e., acid vs. neutral-alkaline) rather than the full parent material description.

At the onset of soil formation four soil-forming processes begin to modify the upper portions of both types of parent materials: additions of OM to the soil surface, transformations of added by OM by the microbial community, chemical weathering of minerals

in the parent material, and development of soil structure. In the earliest phase of soil formation, the alteration of the parent material is limited to a thin layer at the surface of the soil. In the CSSC, these soils are classified into the **Regosolic Order** (VandenBygaart, 2011). In most forested regions Regosolic soils only occur in unstable landscapes such as sand dune areas or active floodplains and are not readily mapped on a national-level map (Figure 8.4).

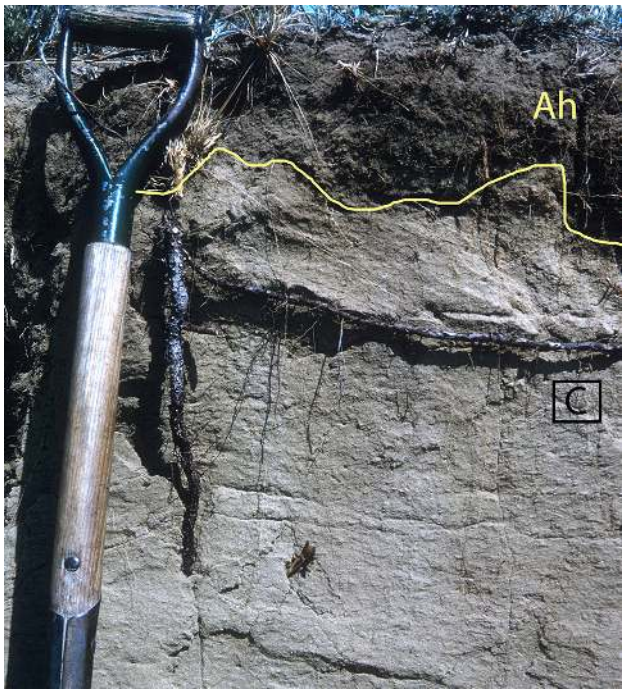


Figure 8.4. Regosolic soil formed in a sand dune. Regosolic soils lack a B horizon completely (as in this case) or have a B horizon less than 5 cm thick. © Roly St. Arnaud (Saskatchewan Centre for Soil Research) is licensed under a CC BY (Attribution) license.

In acid parent materials, chemical weathering in the surface horizon begins to form a bleached horizon (Ae), which is underlain by a horizon with red coatings on the sand grains (Bm). Soils with acid Bm horizons over 5 cm thick are classified into the **Brunisolic Order** (Smith et al., 2011) (Figure 8.5). In some regions organic-iron or organic-aluminum complexes form in the upper soil and are deposited in the B horizon (i.e., podzolization). These horizons may be high in iron (Bf), aluminum, or organic materials (Bh) or all three materials (Bfh or Bhf), and are diagnostic of the **Podzolic Order** (Sanborn et al., 2011) (Figure 8.6).

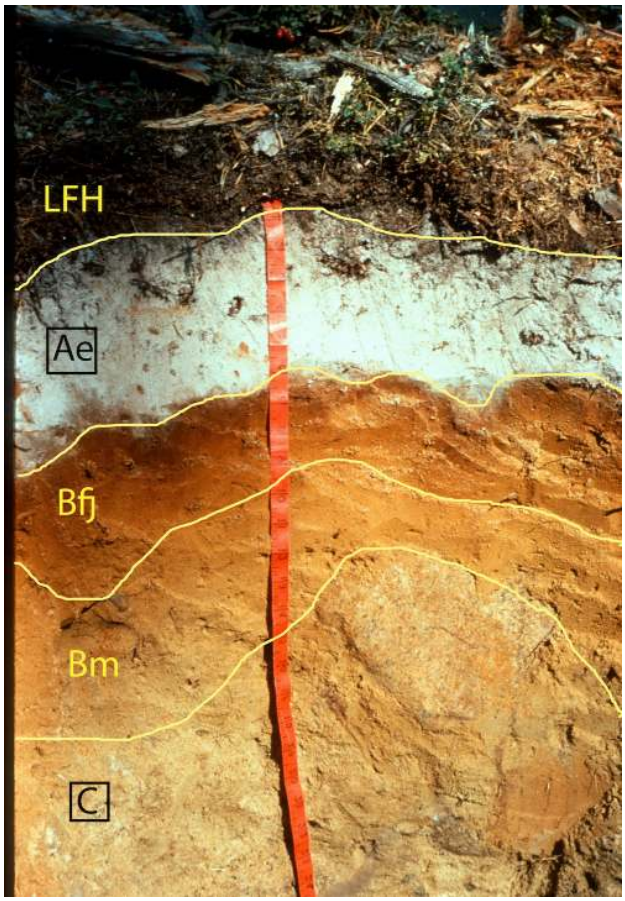


Figure 8.5. Eluviated Dystric Brunisol. The leaf litter directly overlies the mineral Ae horizon. The thick bleached Ae is underlain by a horizon that has some of the characteristics of a Bf (iron-rich) horizons but fails to meet the criteria set in the CSSC and is therefore assigned a j (juvenile) suffix as well as the f suffix. It is underlain by the Bm horizon. Either the Bfj or the Bm is diagnostic of the Brunisolic order. © Ken Van Rees, Univ. of Saskatchewan (Saskatchewan Centre for Soil Research) is licensed under a CC BY (Attribution) license.

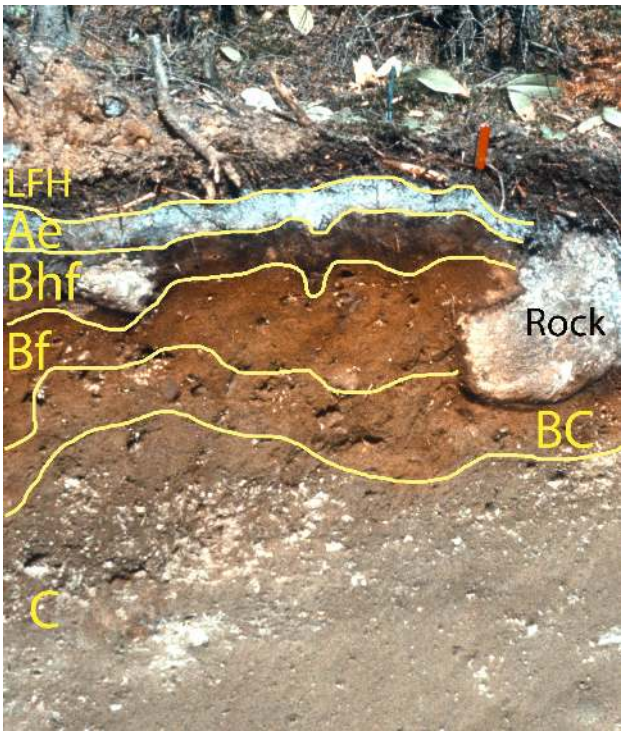


Figure 8.6. Orthic Humo-Ferric Podzol. The Bhf (humus-ferric iron) and Bf (ferric iron) horizons are diagnostic of the Podzolic order. © Saskatchewan Centre for Soil Research is licensed under a CC BY (Attribution) license.

The final major geographic distinction within the acid Brunisolic and Podzolic orders depends on whether earthworms or other burrowing animals are active in the surface horizon of the soils. Where earthworms are present the LFH horizons are mixed with the surface mineral layer and an Ah horizon occurs; where they are absent a sharp contact between the LFH and Ae horizon occurs. These differences are recognized in the great group level of the CSSC – acid Brunisols and the Podzols that have an Ah horizon at least 10 cm thick present are placed in the Sombric great group (Figure 8.7).

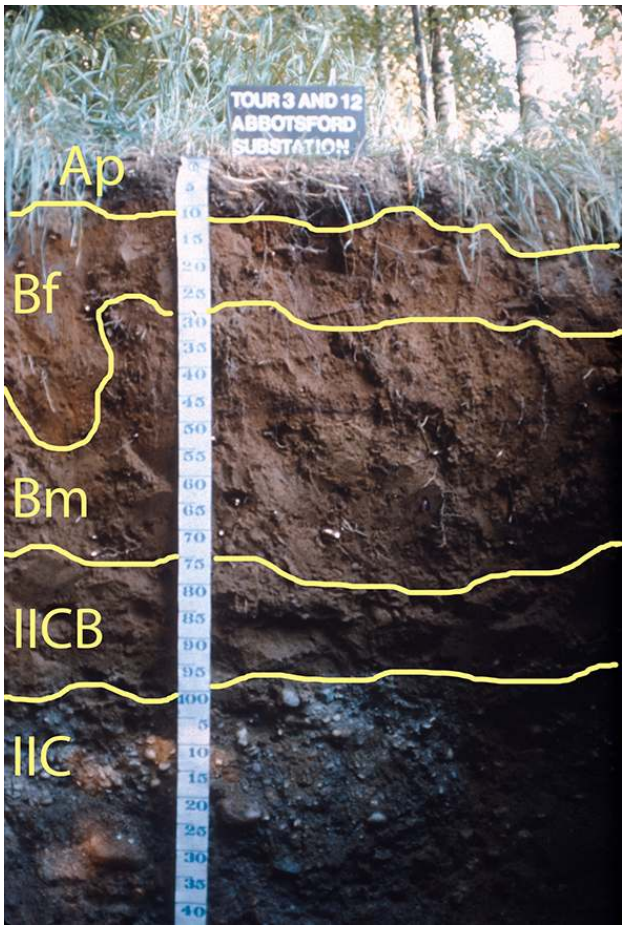


Figure 8.7 Sombric Podzol. Soils of the Sombric great groups of the Podzols and Brunisols have an Ah or Ap horizon at least 10 cm thick that overlies the diagnostic iron-enriched (Bf) or minimally transformed (Bm) horizon. Slide from Abbotsford Research Station (Slide IUSS E-4). © Saskatchewan Centre for Soil Research is licensed under a CC BY (Attribution) license.

The regional climate is the major control on the distribution of acid Brunisolic and Podzolic soils (Figure 8.1). Generally Podzolic soils occur in wetter and warmer forested regions with acidic, igneous-rock derived parent materials, while Brunisolic soils can be found in drier and cooler regions of the continental interior (Smith et al. 2011, Sanborn et al. 2011). Podzolic soils are dominant in the Atlantic Maritime and the eastern portion of the Boreal Shield ecozone as well as throughout higher elevation portions of the Montane Cordillera and southern areas of the Boreal Cordillera. The base-poor, acidic classes of the Brunisolic order occur throughout the Taiga Shield ecozone and the Boreal Shield west from NW Ontario through the northern Prairie Provinces. They also dominate the Boreal Cordillera ecozone.

There are also large areas of the Boreal and Taiga Shield Ecozones where the dominant surface material (in terms of area covered) is rock (mapped as Unclassified in Figure 8.1). These are the iconic landscapes painted by Canadian artists such as Tom Thompson and the members of the Group of Seven in the first half of the 20th Century. There are pockets of sand and gravel (with Brunisolic or Podzolic soils) and Organic soils in depressions scattered throughout these landscapes, but the igneous or metamorphic rocks are the dominant type of surface material.

The genetic pathways for soils developed on neutral-alkaline parent materials derived from sedimentary rocks are very distinct from those discussed above but (somewhat confusingly) some of these soils are also classified into the **Brunisolic Order**. The first stage in the formation of these sedimentary-rock derived soils would involve the same four soil-forming processes (OM additions and transformations, chemical weathering, and structure formation). Because these parent materials typically have carbonates and more soluble salts present, the effect of weathering is to destroy carbonates and salts in the upper soil and transfer the ions released deeper into the soil (in the case of carbonates) or leach them completely from the soil (for the more highly soluble salts). This creates a carbonate- and salt-depleted B horizon, which is also designated a Bm horizon. There may also be reddening of the Bm horizon after removal of carbonates. Soils with a neutral-alkaline Bm horizon greater than 5 cm thick are again classified into the

Brunisolic Order (Figure 8.8) but into different great groups (the Eutric and Melanic Great Groups) than the more acidic versions (the Dystric and Sombric Great Groups). More developed versions of these Brunisolic soils also begin to have bleached Ae horizons present.

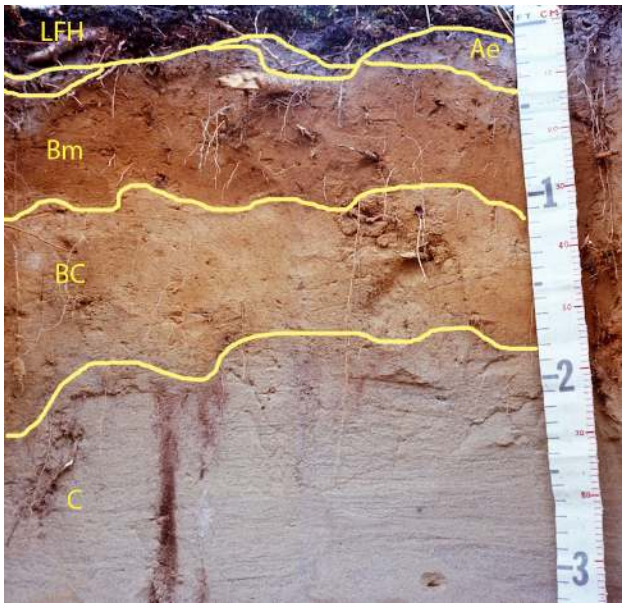


Figure 8.8. Orthic Eutric Brunisol. A thin, patchy Ae is underlain by a reddish Bm horizon, which grades into a transitional horizon (BC) with characteristics of both the B and the C horizons. © Saskatchewan Centre for Soil Research is licensed under a CC BY (Attribution) license.

The removal of carbonates allows clay particles in the upper soil to be detached from the soil mass, moved vertically in the soil, and deposited in the B horizon in the process of lessivage. This creates a textural contrast between the bleached Ae and the B; and the layer with higher clay content is denoted with a Bt horizon label. As discussed in Chapter 2, clay-enrichment from lessivage is very difficult to distinguish from textural differences caused by layering in parent materials and in practise all B horizons with higher clay contents than the Ae are given the Bt label. Soils with a Bt horizon are classified into the **Luvisolic Order** (Lavkulich and Arocena, 2011) (Figure 8.9).

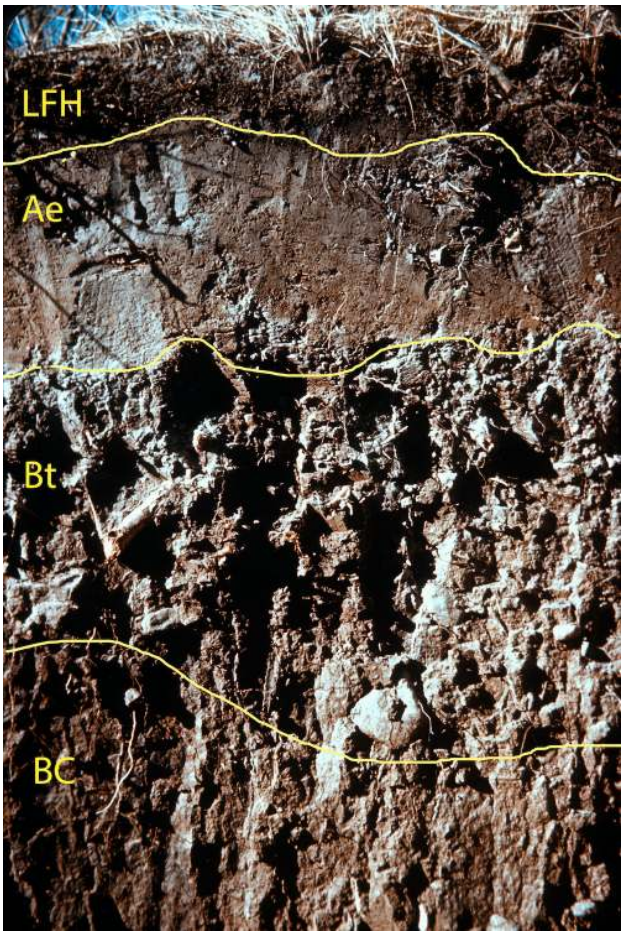


Figure 8.9. Orthic Gray Luvisol. The Ae has a lower clay content than the Bt horizon. Note how the clay loam Bt has a much more strongly developed structure relative to the Ae. © Roly St. Arnaud (Saskatchewan Centre for Soil Research) is licensed under a CC BY (Attribution) license.

As with the soils developed on acid parent materials, a further distinction is made between neutral-alkaline Brunisolic and Luvisolic soils with an organically enriched surface mineral horizon (Ah) caused by mixing (usually by soil-dwelling organisms) and those with a sharp contact between the Ae and LFH horizons. Again these differences are recognized in the CSSC at the great group level (Figure 8.10). The Gray Brown Luvisol and Melanic Brunisols have an Ah horizon greater than 10 cm thick present and the Gray Luvisols and Eutric Brunisols do not.

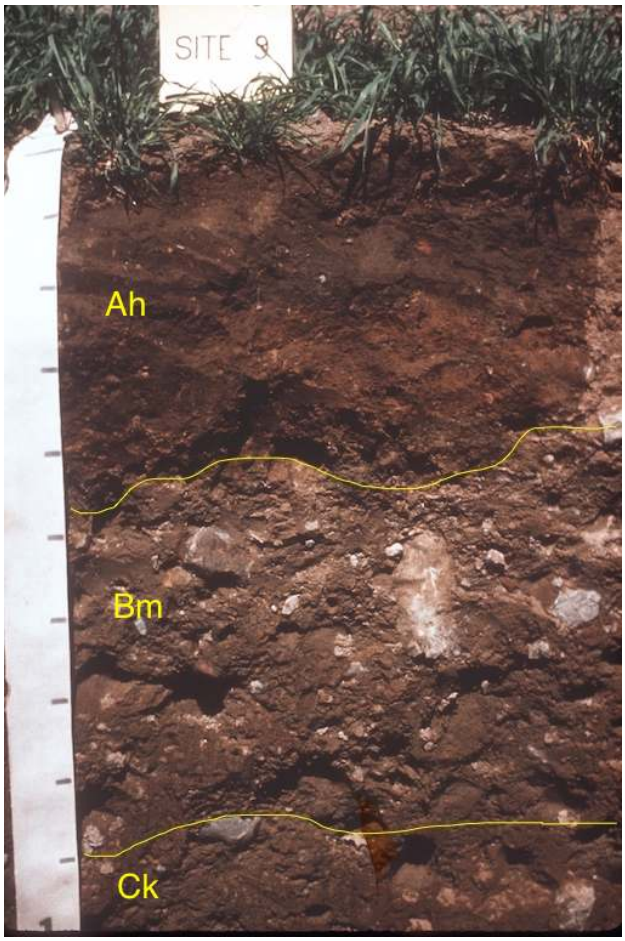


Figure 8.10. Melanic Brunisol. This sandy Ontario soil has a thick Ah horizon overlying the diagnostic Bm (minimally weathered horizon). Compare to Figure 8.8 where no Ah is present. © Saskatchewan Centre for Soil Research is licensed under a CC BY (Attribution) license.

The greatest extent of Gray Luvisolic soils is in the Boreal Plains ecozone of the Prairie Provinces and southern portion of the Taiga Plains ecozone (Figure 8.1). This ecozone is dominated by mixed deciduous and coniferous forests. A similar mix of vegetation and soils occurs in the interior of British Columbia in the Montane Cordillera ecozone. The Gray Brown Luvisolic and Melanic Brunisolic soils are primarily found in the Mixedwood Plains ecozone in southern Ontario and in parts of the Atlantic Maritime ecozone in New Brunswick and Nova Scotia. A final area of Gray Luvisolic soils occurs on the areas of clayey glacio-lacustrine deposits in north-central Manitoba, eastern Ontario and western Quebec in the Boreal Shield ecozone.

WETLAND SOILS

Gleysols and Organic Soils

The two orders for wetland soils in the CSSC are differentiated based on the presence or absence of O horizons. Where the O horizon is at least 40 cm thick, wetland soils are classified into the **Organic order**; where the O horizon is less than 40 cm or absent altogether, wetland soils are classified into the **Gleysolic order** (Figure 8.11). As we saw in Chapter 2, the distribution of wetland soils is controlled by the duration of water-saturated or anaerobic conditions and the effect of anaerobic conditions on the decomposition of organic materials and soil chemical environment.

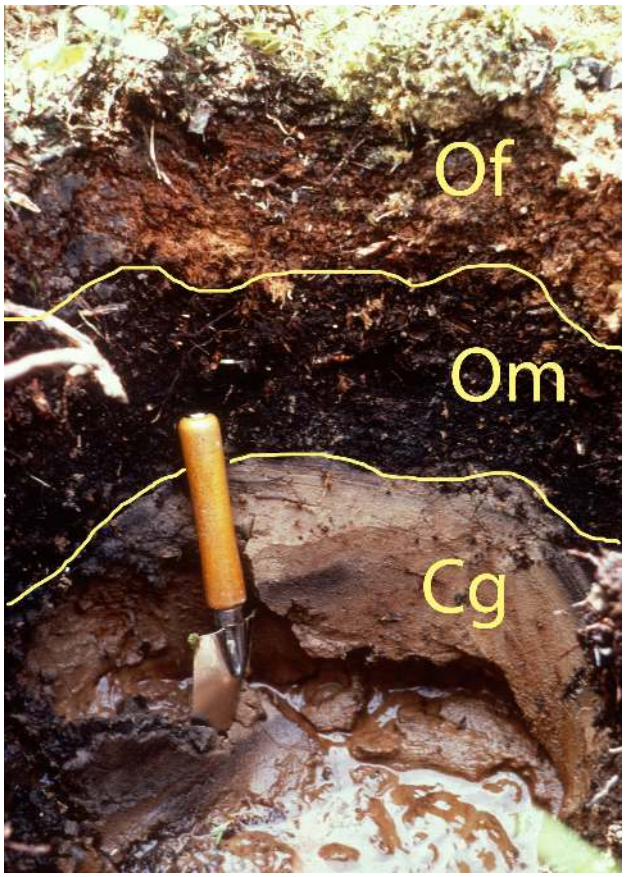


Figure 8.11. Rego Gleysol, peaty phase. The O horizons are less than 40 cm in thickness, and directly overlie the diagnostic Cg horizon. © Ken Van Rees, Univ. of Saskatchewan (Saskatchewan Centre for Soil Research) is licensed under a CC BY (Attribution) license.

Soils of both orders occur in every province of Canada but they rarely are dominant in the landscape (Figure 8.1). More commonly they are found in depressional positions in a landscape where water accumulates and saturated soil conditions occur. Hence, these soils are common in soil catenas throughout Canada but are rarely the dominant soil order present.

Soils of the Organic order dominate the Hudson Plains ecozone, which is one of the largest wetland areas in the world (Kroetsch et al., 2011). They also occur in large areas in the Boreal Plains ecozone at the contact with the Boreal Shield ecozone (Chapter 2; Figure 2.11). They also occur throughout the forested landscapes of Canada in depressional sites or in areas of sluggishly moving water flow.

Soils with organic layers greater than 40 cm occur in two other settings in Canada. The first is in the wet, temperate rainforests of coastal British Columbia, where the very high precipitation levels (typically greater than 3000 mm per year) cause water saturation and retards the decomposition of leaf litter and other aboveground forest biomass (Fox and Tarnocai, 2011). This leads to the buildup of this material (generally termed folic material) and soils in this region are classified into the Folic great group of the Organic Order (Figure 8.12). There are also substantial areas of soils with permafrost in the southern Yukon and south-west NWT that have an organic layer greater than 40 cm thick. These soils are discussed in the following section on Arctic and Tundra soils.

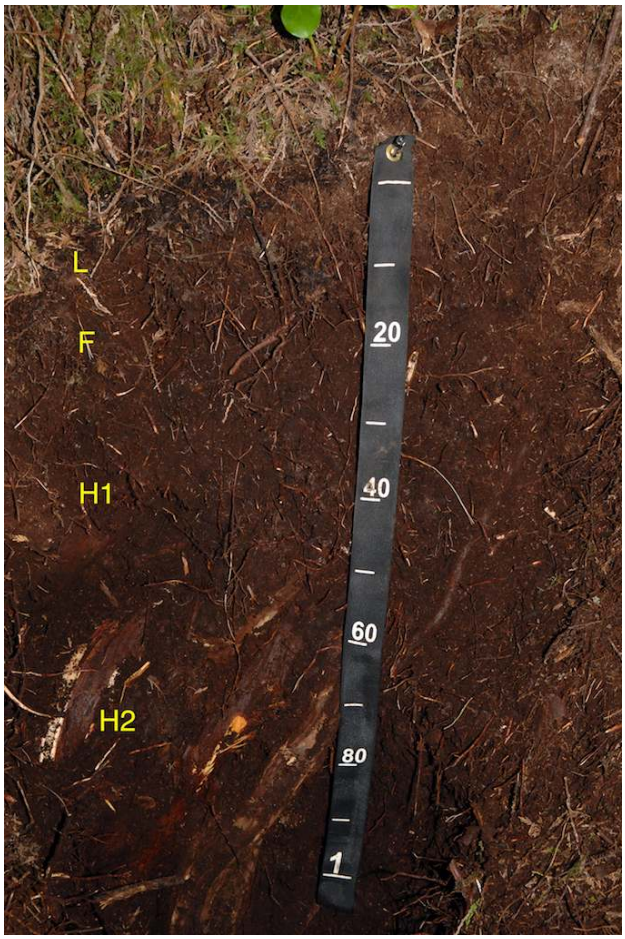


Figure 8.12. Soil of the Folisol Great Group, Central coast of B.C. These soils have thick F (fibric) and H (humic) horizons formed by accumulation of forest floor organic material in wet conditions. Thick tree roots are evident from 60 to 80 cm. © Paul Sanborn is licensed under a CC BY (Attribution) license.

Gleysolic soils are mineral soils that exhibit evidence of gleying conditions: blue-grey colours, reddish mottles and low soil chromas indicative of prolonged water saturation (Bedard-Haughn, 2011) (Figure 8.13). Horizons with these features are assigned a lowercase g label. This label can be assigned to A horizons (Aeg) or to a range of B and C horizons. The Ah horizon is typically too dark to allow visible expression of the gley properties.

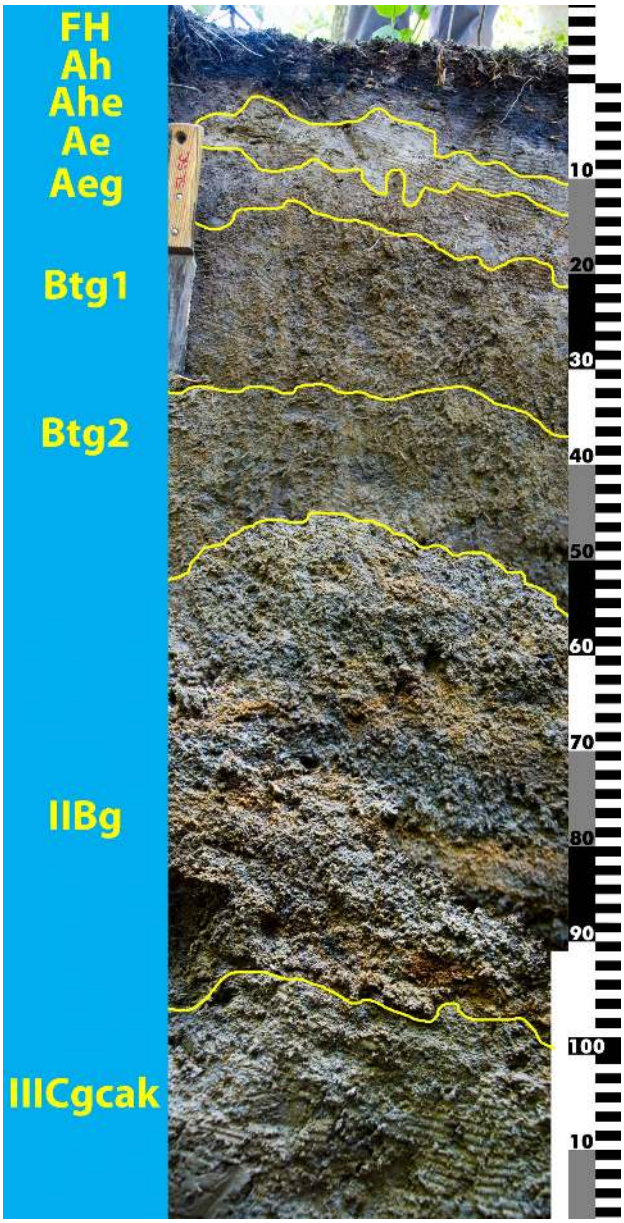


Figure 8.13. Humic Luvic Gleysol. The soil has an Ah horizon greater than 10 cm thick (Humic) and there are two B horizons with both higher clay contents (Luvic, t suffix) than the Aeg and mottling (g). The roman numeral II indicates that the material below 50 cm is formed in a different parent material. The zones of reddish (oxidized iron) material interspersed with the grey material (reduced iron) in the IIBg horizon are good expressions of mottling. Photo courtesy of Kent Watson. © Kent Watson is licensed under a CC BY (Attribution) license.

Soils of the Gleysolic order are only dominant in small areas of Canada such as in parts of the Mixedwood Plains ecozone in southernmost Ontario and south-west Quebec (Figure 8.1). In both of these areas, the Gleysolic soils occur on high clay glacio-lacustrine or glacio-marine parent materials through which the precipitation drains very slowly, creating anaerobic conditions and Gleysolic soils (Bedard-Haughn, 2011). An example of the catenary position of Gleysolic soils is shown for the Huron catena in Southern Ontario (Figure 8.14). The lower slopes in this landscape have silty clay loam and clay loam parent materials, while the upper slopes have loams or sandy loams. Water concentrates in the lower slopes and water movement (or **drainage**) through the finely textured soil is very slow. This leads to prolonged saturation and the creation of gley properties. The dominant soil (in terms of area) is Luvisolic, hence the area would appear on the national map as Luvisolic. Digital soil mapping (as discussed in Chapter 14) allows for the creation of digital maps where the individual soils can be more readily identified and mapped.

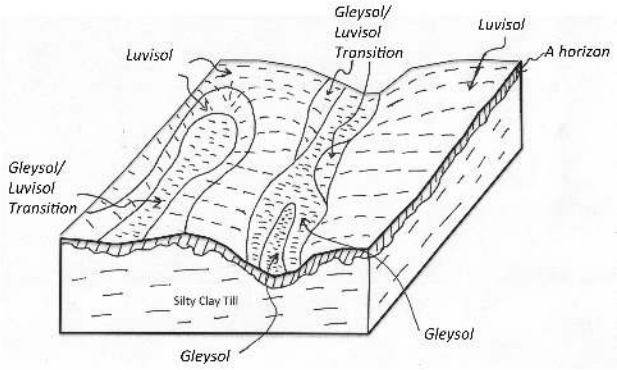


Figure 8.14. Distribution of soils and parent materials in the Huron catena, Waterloo County Ontario. Both the concentration of water flow and the more clay-rich parent materials in lower slope positions cause prolonged periods of water-saturation and the formation of Gleysolic soils. Adapted from Presant and Wicklund (1971; Figure 9) by Dan Pennock, Univ. of Saskatchewan. © Dan Pennock is licensed under a CC BY (Attribution) license.

ARCTIC AND TUNDRA SOILS

Cryosols

In the CSSC soils with a permafrost layer within 1 to 2 m of the soil surface are classified into the **Cryosolic order**. Cryosolic soils are the dominant soil in Canada, covering about 2.5 million km^2 or about 35% of the soil area of Canada (Tarnocai and Bockheim, 2011) (Figure 8.1). Cryosolic soils completely dominate the Arctic Cordillera and Northern Arctic ecozones and the northern portions of the Taiga Cordillera, Plains and Shield ecozones. In the southern portions of the Taiga ecozones, the permafrost becomes discontinuous and the Cryosolic soils intermingle in the landscape with soils of the Brunisolic order.

The Cryosolic soils follow three broad pathways, which are recognized at the great group level in the CSSC. In the first pathway, the growth of ice lenses in the soil causes considerable mixing of soil material, which disrupts horizons and creates profiles with a complex pattern of discontinuous horizons. These soils are classified into the Turbic great group of the Cryosolic order, which is the dominant class in terms of area (Figure 8.15). Describing horizons in the Turbic Cryosols can be complex due to the discontinuous nature of the horizons (Figure 8.16). The diagnostic suffixes for the Cryosolic order are the y suffix (for horizons affected by cryoturbation as shown by disrupted and broken horizons) and the z suffix (for a frozen layer). Both of these suffixes can be used with any of the master horizons (i.e., A, B, C, or O). Pockets of Ah horizon and Bm horizons also often occur in Cryosolic landscapes.

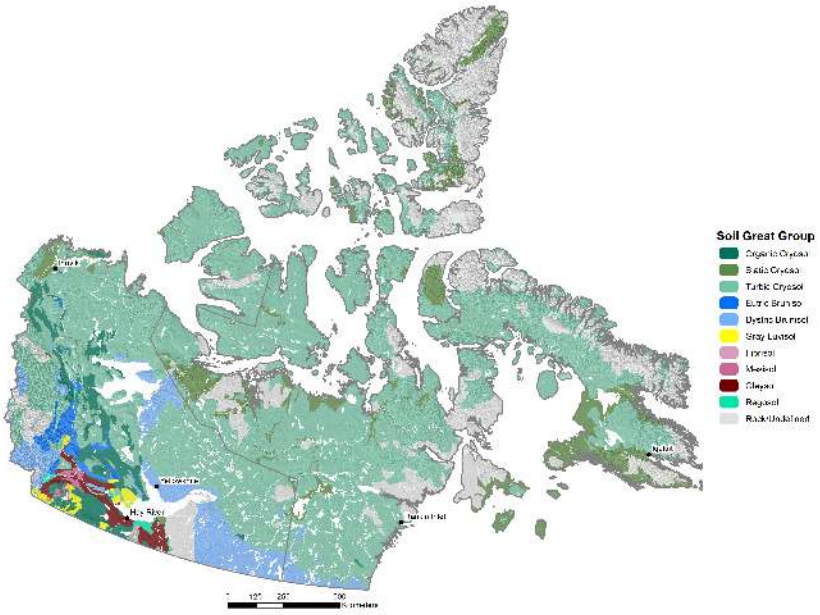


Figure 8.15. Map of soil great groups in the Northwest Territories and Nunavut. The Turbic great group of the Cryosols is the dominant soil class in the region, and grades to Dystric (i.e. acidic) Brunsols at the southern border of the Cryosols. This corresponds to the border between continuous permafrost and discontinuous of patchy permafrost. Map is a reproduction of an official work published by the Government of Canada and was provided courtesy of Darrel Cerkowniak, Agriculture and Agri-Food Canada. © Darrel Cerkowniak, Agriculture and Agri-Food Canada; licensed under a CC BY (Attribution) license.

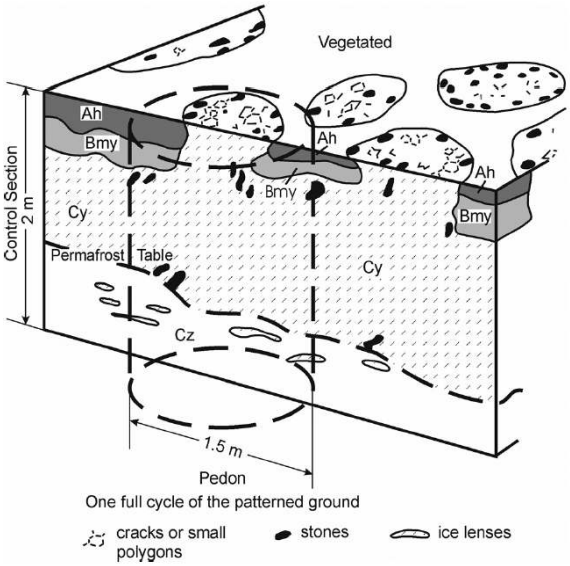


Figure 8.16. Turbic Cryosol. Cryoturbation has mixed the surface organic layer into the lower soil and highly disturbed the mineral B and C horizons. The fine-texture of the parent materials facilitates water retention and frost churning. © Tarnocai and Bockheim (2011); licensed under a CC BY (Attribution) license.

In some smaller areas along the coastlines of the regions where the soil parent materials are too coarse to permit water transfer to form ice lenses, mixing due to ice action does not occur. These soils are classified into the Static great group (Figure 8.17). Finally, in the (relatively!) more temperate areas of the southern Yukon and south-west NWT, we can find Cryosols with an organic layer thicker than 40 cm. These are classified into the Organic great group of the Cryosolic order (Figure 8.15).



Figure 8.17. Brunisolic Static Cryosol. The thick fibric organic layer (Of) overlies a Bm horizon. The diagnostic Cz horizon occurs within 1 m of the surface. The coarse parent materials hold little water, which prevents frost-induced cryoturbation from occurring.
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GRASSLAND SOILS

Chernozems, Solonetz, and Vertisols

Three soil orders of the CSSC are found only in grassland or forest-grassland transition regions: the Chernozemic, Solonetzic, and Vertisolic orders (Anderson and Cerkowniak, 2010). The Chernozemic order is the dominant order, and the Solonetzic and Vertisolic orders are primarily associated with particular parent materials (Figure 8.1). Overall, the grassland regions are dominated by loam and clay loam till parent materials: 35.1 million hectares (Mha) of loam to clay loam till out of a total of 53.1 Mha of farmland in the region (Pennock et al., 2011). All areas of the grasslands have a significant soil moisture deficit ranging from 200 mm in the Brown soil zone in southern Saskatchewan to 80 mm in the transition grassland-Mixedwood forest zone of the Boreal Plains ecozone (Pennock et al., 2011).

Grassland soils in Canada are overwhelmingly found only on neutral-alkaline parent materials. They are characterized by high **base saturation** and the presence of carbonates and more soluble salts in the parent materials. In the first stage of soil formation the growth of grasses (with their lush root mass) and mixing by soil-dwelling mammals transforms the surface mineral soil into an Ah horizon, with the suffix h indicating humus formation. The combined action of hydrolysis and the organic acids produced in this layer begin to dissolve the salts and carbonates below the Ah horizon and a leached Bm horizon forms along with the start of a layer of re-formed calcium carbonate (Cca) deeper in the soil. When the Ah and Bm are very thin these soils would be classified into the **Regosolic order** (Figure 8.18), but within a few hundred years of soil formation they would become classified into the Chernozemic order.

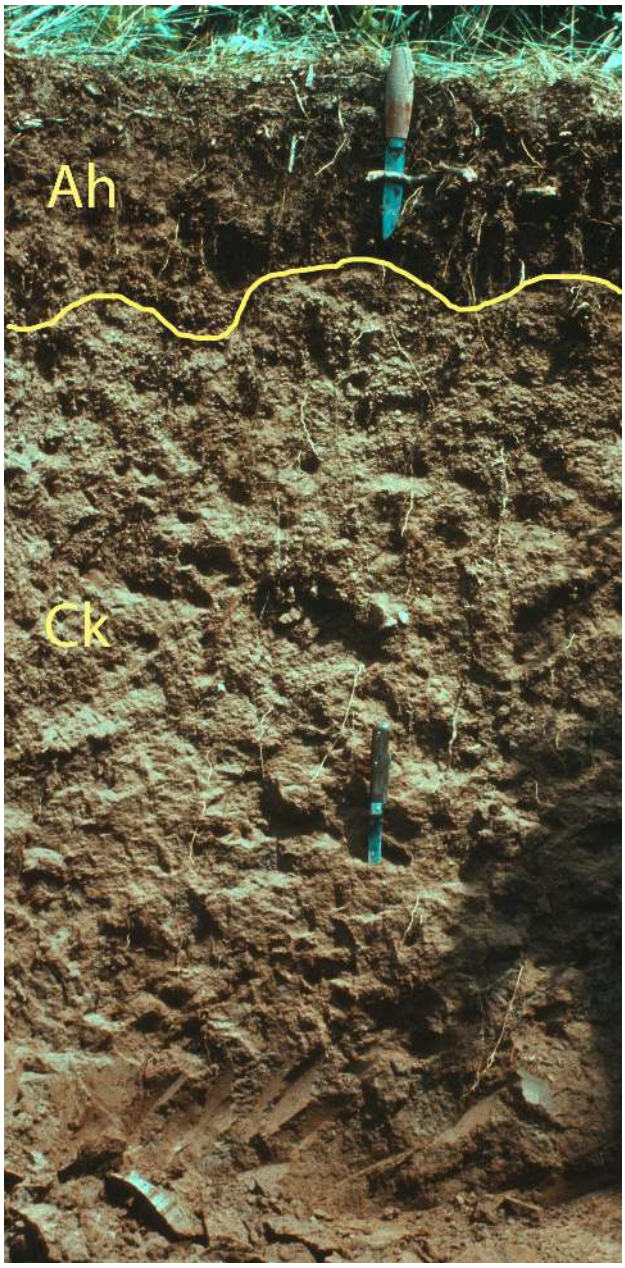


Figure 8.18. Humic Regosol in loamy till. Additions of grass roots and their transformation by microbes create an Ah horizon within a few hundred years but weathering of the inherited calcium carbonate (Ck) to create a B horizon is a slower process. © Saskatchewan Center of Soil Research is licensed under a CC BY (Attribution) license.

The diagnostic horizon of the **Chernozemic order** is an Ah horizon that has dark soil colour (due to OM enrichment), high levels of bases (especially calcium), and is at least 10 cm thick (Figure 8.19). The driest part of the Prairies is the south-eastern area of Saskatchewan and the south-western portion of Alberta; soil moisture increases as we move north from this area, and the colour of the Ah horizon grows darker as the amount of OM stored increases in response to the moister conditions. Often the Ah and the Bm horizon show evidence of the activity of burrowing animals, which is recognized with the u suffix (e.g., Ahu, Bmu). Chernozemic soils also typically have a well-developed C horizon with secondary calcium carbonate (Cca horizon), which overlies a C horizon with the original calcium carbonate inherited from the parent material (Ck horizon).

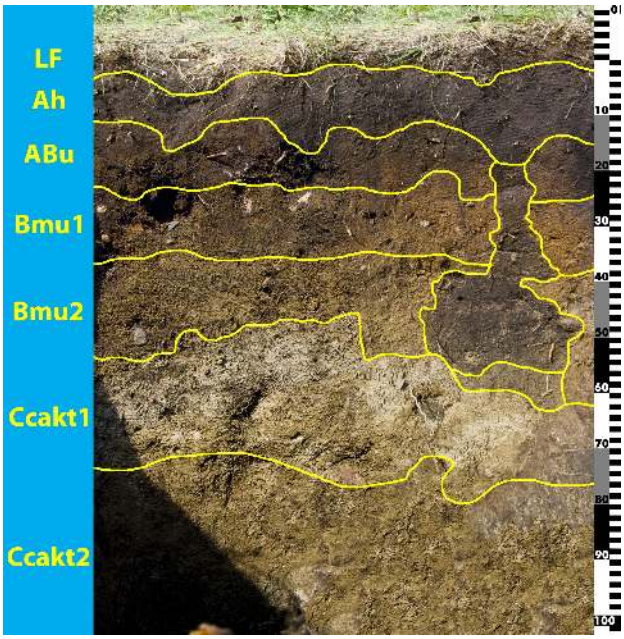


Figure 8.19. Orthic Black Chernozem. Evidence of burrowing animal activity is evident in the in-filled burrow filled with Ah material that extends into the Bmu2 horizon. Secondary enrichment of calcium carbonate is evident in the uppermost C horizon. Photo courtesy of Kent Watson. © Kent Watson is licensed under a CC BY (Attribution) license.

The climatically controlled increases in OM storage (and surface soil colour) leads to a clear zonation of soils that is reflected in the great groups of the Chernozemic order (Brown, Dark Brown, Black, and Dark Gray) (Figure 8.20). The northernmost extent of the Chernozemic order is in the Peace River Lowlands of Alberta and the boundary between grassland and forest trends southeast through Alberta, Saskatchewan, and Manitoba. The geographic position of the transition is controlled by the dominant movement of the air masses in the region, which break through the Cordilleran barrier in the low elevations west of the Peace River area and then trend southeast as they pass through the region (Pennock et al., 2011).

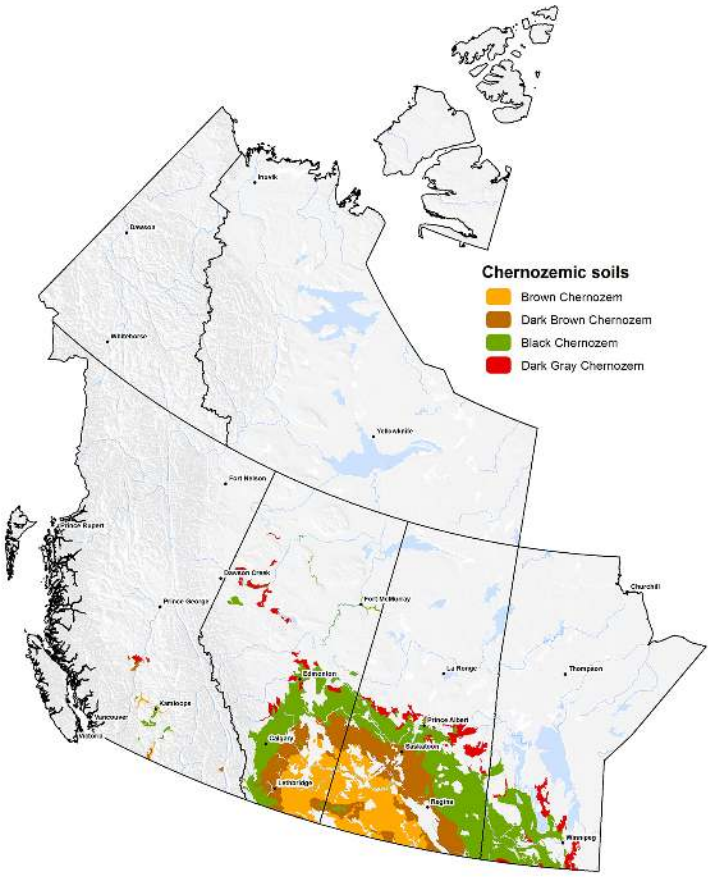


Figure 8.20. Regional distribution of the great groups of the Chernozemic order in western Canada. The area of Dark Brown and Black Chernozems in SW Saskatchewan and SE Alberta is associated with the Cypress Hills Upland. Map is a reproduction of an official work published by the Government of Canada and was provided courtesy of Darrel Cerkowniak, Agriculture and Agri-Food Canada. © Darrel Cerkowniak, Agriculture and Agri-Food Canada; licensed under a CC BY (Attribution) license.

Throughout much of the Prairie region in the transitional zone the Chernozemic soils grade into the Luvisolic soils of the Boreal Plain ecozone. The soils associated with this transitional zone often have A horizons with evidence of both organic additions and bleaching (the Ahe horizon) overlying a weakly developed (or juvenile) texture contrast horizon, designated as a Btj horizon. The soils of the transition are recognized at the great group level as Dark Gray Chernozems or as Dark Gray Luvisols.

The land surface in many areas of the Prairies was formed by melting of immobile ice masses resulting in a hummocky surface form with innumerable small wetlands. A characteristic catena of soils occurs in these small catchments, with Regosolic or thin Chernozemic soils in upper slopes, Chernozemic soils with increasingly thick Ah horizons in mid- and lower-slopes, and Gleysolic soils in the depressions (Figure 8.3).

The other two grassland orders are associated with particular types of parent materials. **Vertisolic soils** form in clayey glacio-lacustrine parent materials with a high content of smectite clay minerals (Brierley et al., 2011). This causes the soil to shrink in dry periods and form deep cracks that can extend a meter or more into the soil (Anderson, 2010). In wetter periods the clays swell and close the cracks and generally cause mixing or turbation of the soil material. The mixing prevents well-expressed horizons from forming. The turbated horizons are denoted with a v suffix (e.g., Bv, Cv). The displacement of soil material along the cracks also forms polished surfaces (called **slickensides**) when one soil mass slips by another during turbation. These features are assigned a ss label (e.g. Bss). The presence of horizons with both the v and ss suffixes are required for classification into the Vertisolic order (Figure 8.21).

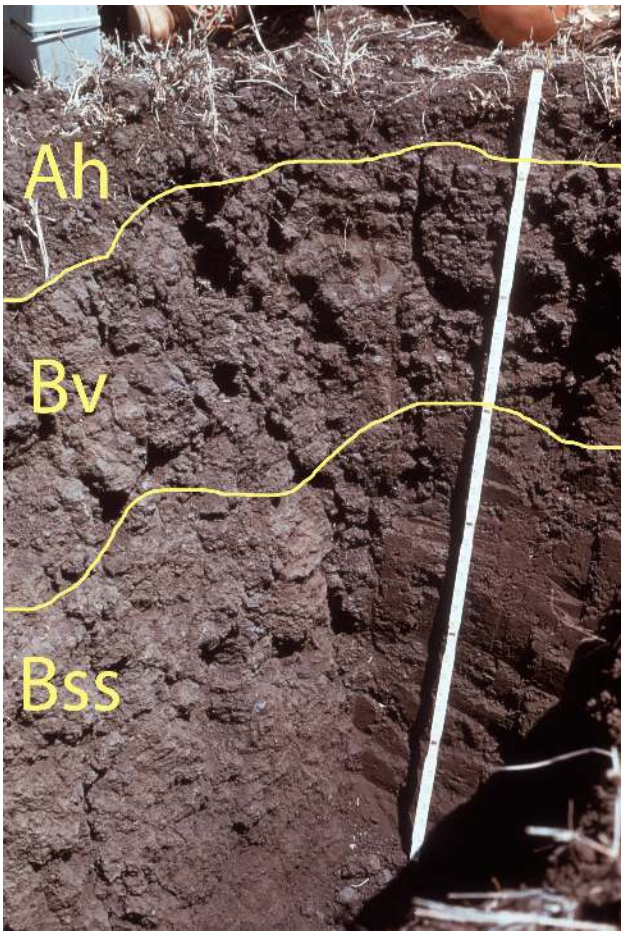


Figure 8.21. Orthic Vertisol. Mixing of the horizons by shrinking and swelling of clays prevents the formation of distinct soil horizons in soils of the Vertisolic order. © Roly St. Arnaud (Saskatchewan Centre for Soil Research) is licensed under a CC BY (Attribution) license.

High concentrations of the base cation sodium (Na^+) are responsible for the distinctive soil horizons of the Solonetzic order (Miller and Brierley, 2011) (Figure 8.22). Typically for Chernozemic soils the ratio of exchangeable calcium:exchangeable sodium is 30 or greater, but in Solonetzic soils the ratio is below 10. The high sodium levels result from complex interactions between parent material, topography, the groundwater regime, and time (Anderson, 2010b). Horizons with exchangeable Ca:Na ratios less than 10 are assigned an “n” suffix (e.g., Bn). In the field, they often have high clay content and dense, impermeable structure (often informally called hardpan) and are additionally assigned a “t” suffix (Bnt). The ponding of water on the surface of the Bnt can lead to destruction of the upper B horizon and formation of an Ae/AB layer.

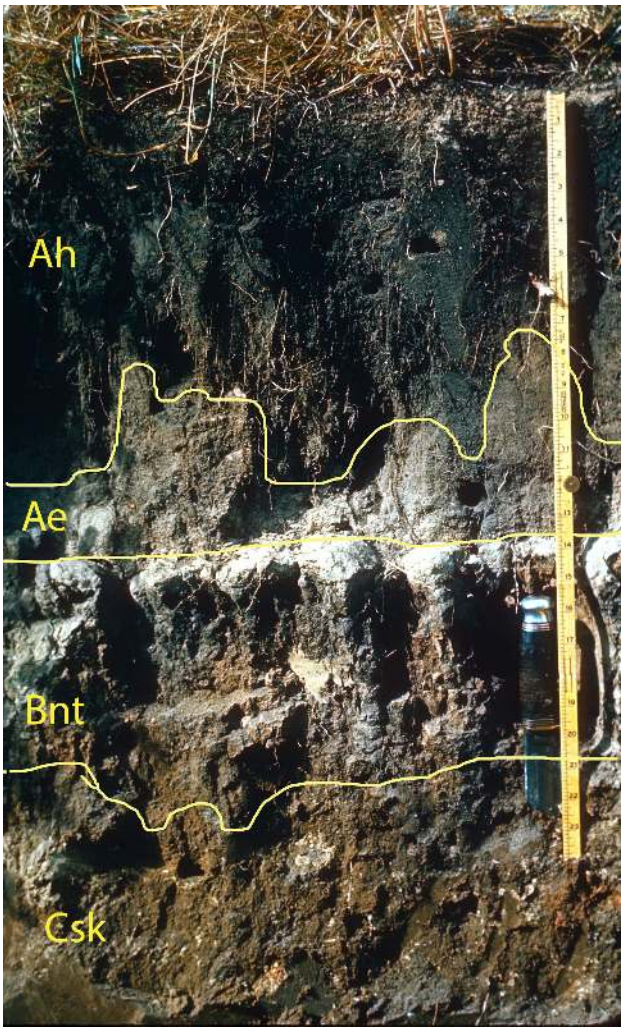


Figure 8.22. Black Solodized Solonetz. The thick Ah horizon overlies a bleached Ae. The characteristic (but not always visible) round-tops of the columnar structure of the Bnt horizon are undergoing bleaching. White flecks of salts are visible in the Csk horizon. © Roly St. Arnaud (Saskatchewan Centre for Soil Research) is licensed under a CC BY (Attribution) license.

Probably the soil property of greatest interest to producers is **salinity** but the CSSC does a poor job of recognizing salinity in its taxonomic hierarchy. Horizons with salts inherited from the parent material are assigned an s suffix (e.g., Csk) and those with secondary salts formed through precipitation of ions from groundwater are assigned an sa suffix (e.g., Asa, Bsa). Soils with these horizons are not, however, recognized at the order, great group, or subgroup level of the CSSC but are designated as a phase – for example, Orthic Brown Chernozem, saline phase.

HUMAN IMPACT ON SOIL DISTRIBUTION

Regosols and Anthroposols

The most widespread impact of human activity on soil horizonation (and hence classification) is due to human-induced increases in soil erosion, especially erosion caused by tillage in agricultural landscapes (FAO, 2019). Tillage causes mixing of the surface layer to the depth of tillage (typically between 10 to 25 cm). The tilled surface layer is denoted with a p (ploughed) suffix (Ap). As erosion removes surface soil, ploughing incorporates B and C horizons into the Ap horizon, which can limit plant growth where a growth-limiting B or C horizon exists. Often tillage erosion and mixing eliminates the B horizon totally, and the Ap/C or Ap/Cca sequence is classified into the **Regosolic soil order** (Figure 8.23). Soil catenas in agricultural landscapes commonly include Regosolic soils in the most eroded positions. Deposition of soil at the base of the slope causes overthickening of the A horizon in the depositional positions but typically this has little effect on the classification of the soil.

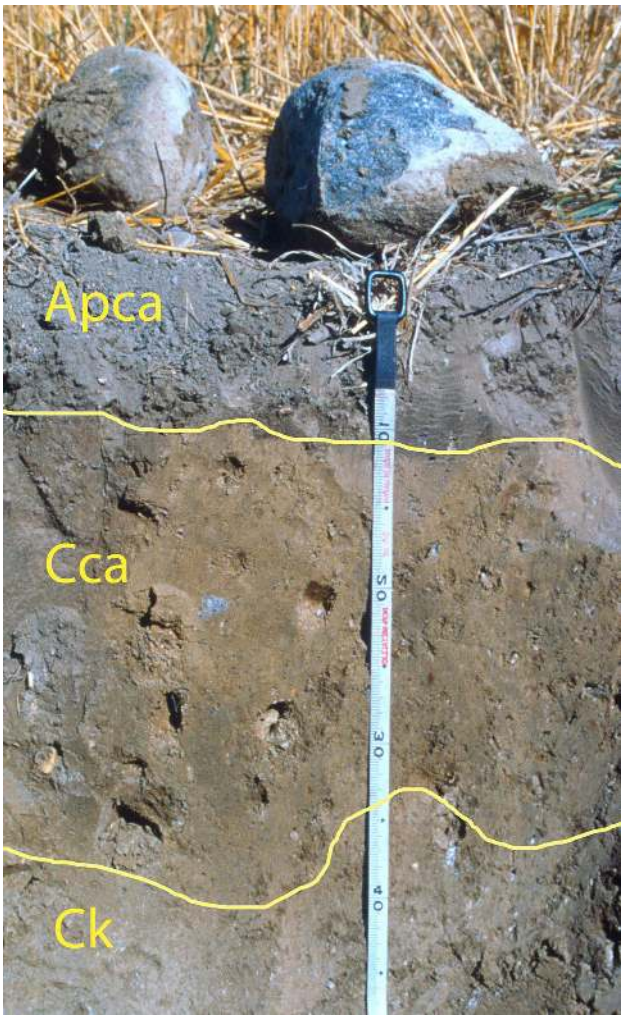


Figure 8.23. Orthic Regosol. In the native state this soil would have been a thin Chernozemic soil. Tillage erosion has removed the B horizon and incorporated the Cca horizon into the surface Apca layer. This process causes formation of Regosols on the knolls of eroded cultivated fields throughout Canada. © Dan Pennock, Univ. of Saskatchewan (Saskatchewan Centre for Soil Research), is licensed under a CC BY (Attribution) license.

A second type of widespread human impact on soils is caused by removal, storage and replacement of soil material associated with various forms of resource exploitation – for example, trenching for pipelines or removal of surface material to allow mining of bitumen in the oilsands or mineral resources elsewhere. The highly disturbed or reconstructed soils can not be adequately described using the terminology of the CSSC, and Naeth et al. (2011) suggested that a new order called the **Anthroposols** should be adopted to describe them. The diagnostic horizon of the Anthroposols would be the D (for disturbed) horizon. Many other national and international systems of soil classification have already adopted orders to account for these soils. The Anthroposol order has not been officially adopted into the CSSC at the time of writing (summer of 2019).

GLOBAL SOILS

Because of the relative youthfulness of Canadian soils and the limited range of climates found here there are many classes of global soils that do not occur in Canada. The CSSC is designed only for soils found in Canada and we have to use other classification systems for global soils that do not have Canadian equivalents.

Unfortunately, there is no single agreed-upon international soil classification system in place. The two main international systems are Soil Taxonomy (developed by the United States Department of Agriculture) and the World Reference Base for Soil Resources, which was originally developed by the Food and Agriculture Organization of the UN and is now edited by a working group of the International Union of Soil Sciences. The situation is further complicated because the highest taxonomic levels in each system do not correspond very closely to orders in the other system, but despite this overlap in classes a general table of concordance between

the two systems (and with the CSSC where appropriate) can be developed (Table 8.3). For our purposes (at the introductory level) we will only use the USDA soil classes as many sources used by Canadian students use the USDA classification. Overall, the lack of a single taxonomic system for soils is a very unsatisfactory state of affairs, leading to significant confusion among generations of students and among scientists and policymakers who wish to use soil information in their work.

Table 8.3. Correlation between classes of Soil Taxonomy, the World Reference Base for Soil Resources, and the Canadian System of Soil Classification

	Canadian System of Soil Classification	World Reference Base	Soil Taxonomy[1]
	Order	**Reference Soil Group**	**Order[2]**
Soils with strong human influence			
Altered by long human use		Anthrosol	
Contains significant human artifacts		Technosol	
Permafrost Soils			
Permafrost-affected	Cryosol	Cryosol	Gelisol
Grassland soils			
Clay-enriched subsoil, high sodium concentration	Solonetzic	Solonetz	Natric great groups
High shrink-swell clays, slickensides	Vertisol	Vertisol	Vertisol
High soluble salt concentration		Solonchak	Salids
Dark, well-structured Ah, secondary $CaCO_3$	Chernozemic (Black GG)	Chernozemic	Mollisol
			(Udolls)
Dark Ah, secondary $CaCO_3$	Chernozemic	Kastanozem	Mollisol
	(Dark Brown, Brown GG)		(Ustolls, Xerolls)
Dark Ah, no $CaCO_3$		Phaeozem	Mollisol
			(Udolls, Albolls)
Dark Ah, low base status	Brunisol	Umbrisol	
	(Sombric)		
Arid Zone Soils			
Accumulation and cementation by silica		Durisol	Aridisol
			Durids
Accumulation of secondary gypsum		Gypsisol	Aridisol
			Gypsids
Accumulation of secondary carbonates		Calcisol	Aridisol
			Calcids
Wetland Soils			

	Canadian System of Soil Classification	World Reference Base	Soil Taxonomy[1]
Thick organic Layer	Organic	Histosol	Histosols
Mineral soil, gleyed properties	Gleysolic	Gleysol	Aquic suborders
Stagnant water, abrupt textural difference		Planosol	Albaqualfs,
			Albaquults,
			Argiabolls
Stagnant water, no textural contrast		Stagnosol	
Forest soils, younger soil surfaces			
Illuviation of iron and aluminum	Podzolic	Podzol	Spodosol
Higher-clay B horizon	Luvisolic	Luvisol	Alfisols (High-activity clays)
Reddening of B horizon	Brunisolic	Cambisol	Inceptisol
Soils of older soil surfaces			
Illuviation and accumulation of Fe, irreversible hardening when dry		Plinthosol	Plinth-
			descriptor
Low-activity clays, high Fe oxides		Nitisol	Alfisols, Ultisols (Kandic GG)
Dominance of kaolinite and oxides		Ferrasol	Oxisol
Interfingering of coarse material into redder layer		Retisol	Gloss-
			descriptor
Clay-enriched subsoil, low-activity clays, low base status		Acrisol	Ultisols (with low-activity clays
Clay-enriched subsoil, low-activity clays, high base status		Lixisol	Alfisols (with Low-activity clays)
Clay-enriched subsoil, high-activity clays, low base status		Alisol	Ultisols (with high activity clays)
Soils with Little or No Profile Differentiation			
Very sandy soils		Arenosol	Psamments
Stratified fluvial,lacustrine, or marine	Regosol	Fluvisol	Fluvents

	Canadian System of Soil Classification	World Reference Base	Soil Taxonomy[1]
sediments	(Cumulic)		
No significant profile development	Regosol	Regosol	Entisols
Thin soil or with many coarse fragments		Leptosol	Entisols
Soils in volcanic sediments, complex aluminum minerals		Andosols	Andisols

[1]Correlation between USDA and WRB from IUSS Working Group WRB. 2015. World Reference Base for Soil Resources 2014, update 2015. International soil classification system for naming soils and creating legends for soil maps. World Soil Resources Reports No. 106. FAO, Rome. (http://www.fao.org/3/i3794en/I3794en.pdf)
[2]Soils that correlate at lower taxonomic levels in Soil Taxonomy are shown in italics.

The additional classes in the USDA/WRB systems (compared to the CSSC) are required (in part) because of the higher degree of chemical weathering in many global soils compared to the relatively youthful, post-glacial soils in Canada. Both the WRB and USDA systems use two criteria as indicators of the degree of weathering. The first, is the base status of the soil – high base status (i.e., the soil exchange complex is dominated by calcium, magnesium, potassium and sodium) indicates limited weathering and leaching, and low base status (i.e., replacement of the four bases by aluminum and hydrogen) indicates greater weathering and leaching. The second, is the exchange capacity or activity of the clay minerals present – complex, inherited clay minerals such as illite or smectite are high activity clays whereas simpler, secondary clay minerals such as kaolinite are low activity. Formation of low activity clays and oxides is associated with older, highly weathered landscapes.

In Canada, the major forest soil on neutral-alkaline parent materials is the Luvisol. On older, non-glaciated surfaces the clay minerals in the Luvisols undergo further weathering and the complex, inherited clays are gradually transformed into low-activity clay variants of the Alfisol order (Figure 8.24). In the next phase of soil formation, chemical weathering further transforms the primary, inherited minerals into secondary iron and aluminum oxides and soils with very-well developed texture contrast horizons and high oxide levels occur. These are classified into the **Ultisol order** in Soil Taxonomy (Figure 8.24). Finally, the inherited clay minerals are weathered completely and the soil becomes dominated by the simple clay mineral kaolinite and clay-sized iron and aluminum oxides. These soils also often experience very high levels of turbation by soil-dwelling organisms such as ants and termites. Overall the soils are often very thick and have relatively limited horizon development. These soils are classified as **Oxisols** in the USDA (Figure 8.24). In some situations a horizon with very high oxide levels forms in the soil, which is called plinthite; when exposed to the atmosphere this material can irreversibly dry, creating rock-like petroplinthite. These soils are informally referred to as laterites.

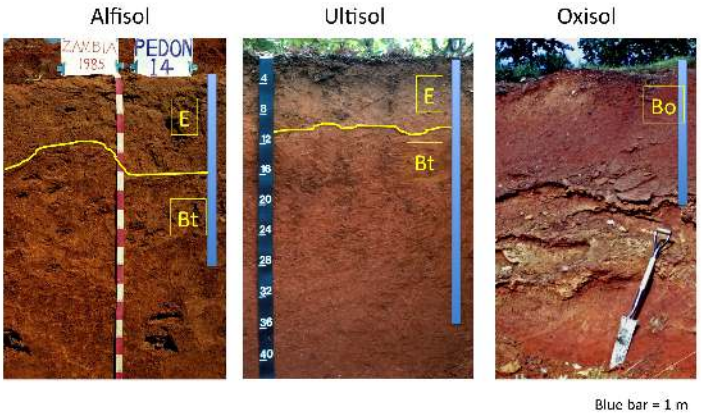


Figure 8.24. Major classes of soils in highly weathered, older land surfaces. The Alfisols and Ultisols of these regions have Bt horizons with low-activity clays. In the Oxisols, the clay minerals themselves are largely destroyed through desilication and the soils are dominated by iron and aluminum oxides. Photos courtesy of the U.S. Department of Agriculture, Natural Resources Conservation Service; licensed under a Public Domain license. Reproduced with permission.

Ultisols are the dominant soil order in the Southwestern USA and throughout much of south-east Asia (Figure 8.25). They also occur fringing the soils of the Oxisol order in the older land surfaces in South America and equatorial Africa. Many of the Alfisols found in South America and the Sahel region of Africa would be of the low-activity clay variants. Paton et al. (1996) argue that soil classification systems such as the USDA Soil Taxonomy are inadequate for countries like Australia, which generally has old land surfaces with complex environmental histories.

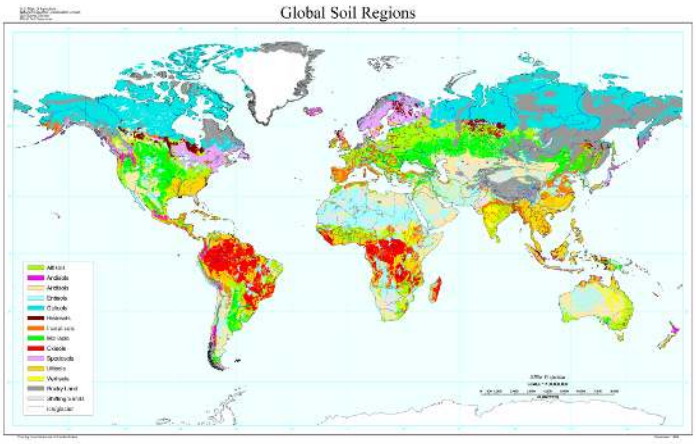


Figure 8.25. Global soil map of soil orders in Soil Taxonomy. Map courtesy of the Oak Ridge National Laboratory. Photo courtesy of the U.S. Department of Agriculture, Natural Resources Conservation Service; licensed under a Public Domain license. Reproduced with permission.

This brief summary necessarily simplifies the great complexity of soil formation on older, highly weathered land surfaces, many of which have undergone substantial changes in environmental conditions over time (Paton et al., 1996). It does, however, highlight a major difference between Canadian soils and these more heavily weathered soils: many Canadian soils have high levels of primary (i.e., inherited from the parent material) minerals present which release a steady supply of mineral nutrients (such as calcium, magnesium and potassium) for plant growth through time. For agricultural soils in the Canadian Prairies (as an example), it is relatively rare to add potassium from processed potash rock, but many more heavily weathered soils require potassium fertilizer additions to achieve optimum crop yields.

True hot desert soils do not occur in Canada. Desert soils in Soil Taxonomy are grouped into the **Aridisols** order. Desert soils with cemented layers of silica, gypsum, and calcium carbonate are recognized at lower taxonomic levels in the USDA system (Table 8.3). Many desert areas also have thin, weakly developed soils of the Entisol order, comparable to Regosolic soils in the CSSC.

Both the USDA and WRB systems also have classes for soils formed in volcanic sediments, primarily volcanic ash. These soils have a distinct suite of properties and are classified in the **Andisol** order. They occur in small areas associated with active volcanic

areas in plate boundary areas (e.g., Japan, southern Chile).

SOIL SURVEYS

The primary reason for developing soil classification systems is to allow soils to be consistently mapped and the histories of soil classification and soil mapping are inextricably intertwined. Both began in the early 20th century as countries began to systematically inventory their natural resources. Soil mapping has always been referred to as soil surveying, and the maps themselves are referred to as soil surveys.

In Canada, soil surveys began in Ontario in 1914 followed by British Columbia and the Prairie provinces in the 1920s, Quebec and the Atlantic Provinces in the 1930s, and finally the Yukon, Northwest territory and Nunavit in the 1940s and 1950s (Anderson and Smith, 2011; SCWG, 1998, Chapter 1). Federal government involvement in soil surveys essentially ceased across Canada in the early 1990s, but some provinces have continued surveys since that time.

Soil survey therefore began before a national classification was in place—the initial version of the CSSC was published in 1970. Subsequent editions of the CSSC were published in 1987 (2nd Ed.) and in 1998 (3rd Ed.). This can be a source of confusion to users who consult paper surveys published through the main era of surveys (from 1945 to the early 1990s) as the classification used may be archaic and not readily translated to the 1998 edition of the CSSC. Updating of the classification systems has occurred in the on-line versions of the soil surveys (as discussed below).

There have been a wide variety of scales used for the soil surveys in different provinces. The scale at which a survey is carried out largely determines its usefulness for land use planning and resource assessment – generally the smaller the scale, the more limited the map is for planning purposes. Generally, soil surveys in British Columbia, Manitoba, Ontario, Quebec, Nova Scotia, New Brunswick, P.E.I., and Newfoundland were produced at scales between 1:20,000 and 1:63,360 (i.e., Survey Intensity Levels 2 and 3, Table 8.4). Surveys in Alberta were a mix of scales between 1:50,000 and 1:126,730 and surveys in Saskatchewan, the Yukon and Northwest Territories, and Nunavut are at scales of 1:100,000 or greater. Information on all surveys is available on-line at http://sis.agr.gc.ca/cansis/publications/surveys/index.html.

The basic process used in field work for soil surveys is similar across Canada (although each province has its own variant of the basic process). First, a Survey Intensity Level (SIL) is chosen based on the purpose of the survey (Table 8.4). The SIL determines the number of soil profiles (inspections) that will be described per unit area. Normally, the soil surveyor chooses an inspection point that they believe is most representative of the area being described. At each profile (using soil pits or soil exposures at roadsides, trenches etc.) the soil profile is divided into horizons and a standard suite of soil properties are described. Each horizon is assigned labels according to the CSSC and these labelled horizons are used to classify into the correct order, great group, and subgroup level of the CSSC. The geographic position, landform context of the point is recorded, and the nature of the parent material is also recorded.

The soil surveyor must then delineate the area of land that they believe the soil they have described is representative of. As we have seen in Chapter 2, in Canada the parent material is of primary importance in determining the characteristics of the soil and hence the described soil pit is (hopefully) representative of the parent material it occurs in. The extent of the parent material was normally mapped using aerial photographs – different types of parent material often have characteristic surface forms associated with them (Tables 2.1 and 2.2).

The combination of a given soil class and its parent material is called the soil series, which is the basic unit for the construction of soil maps across Canada (although it is confusingly referred to by different names in different provinces). By convention across Canada series are assigned geographic names based on the closest locality to the point at which they are first described. This means that the names for series are specific to each province, although by referencing the soil taxonomic class and parent material, correlations across provinces are readily done. In some provinces, the soil phase is used to further subdivide soil series – soil drainage class is a commonly used phase criteria in much of Central and Atlantic Canada for example.

For soil maps beyond the most intensive SIL 1 level it is not possible to show individual soil series on the traditional paper map. Instead map units that contain multiple series are used. Often these map units can be related to soil catenas as described earlier in this chapter (although this differs by province). In a simple map unit one series dominates, but in complex units two or more series may occur. In this case the label for each mapped area (called map polygons) will often give an estimate of the percentage of each series in the polygon.

As the scale of survey increases, the number of series in each map polygon will normally also increase and hence the variability in soils in each polygon becomes greater. This variation limits the usefulness of the map for specific planning purposes and hence the broadest surveys at reconnaissance through exploratory levels (Table 8.4) are only of use for very generalized planning purposes.

Table 8.4. Usual scale, inspection intensity, and general objective of the survey intensity levels used in soil surveys in Canada. From Agriculture and Agri-Food Canada (1987).

Survey Intensity Level (SIL) and Name	Usual scale	Inspection (soil pit) Intensity	Objective
SIL1	1:5,000	One per 1 to 5 ha	Very intensive planning e.g. for high value crops (nurseries), septic fields, or baseline and post-disturbance pipeline surveys
Very detailed	(1 cm on map equals		
	50 m on ground surface)		
SIL 2	1:20,000	One per 2 to 20 ha	Local planning for groups of farms, catchments, small parks, irrigation management
Detailed	(1 cm on map equals		
	200 m on ground surface)		
SIL 3	1:50,000	One per 20-200 ha	Limited number of uses for county planning, large parks, major stream catchments, irrigation districts
Reconnaissance	(1 cm on map equals		
	500 m on ground surface)		
SIL 4	1:100,000	One per 100-1000 ha	Limited number of uses: provincial level suitability assessments, regional plans
Broad	(1 cm on map equals		
Reconnaissance	1,000 m (1 km) on ground surface)		
SIL 5	1:250,000	Greater than 1 per 1000 ha; often widely separated	Very limited general uses: broad regional or provincial plans
Exploratory	(1 cm on map equals		
	2,500 m (2.5 km) on ground surface)		

The province-specific approaches to mapping make it very difficult to develop a Canada-wide soil map based on the provincial surveys. To address this problem Agriculture and Agri-Food Canada produced the *Soil Landscapes of Canada* map at a scale of 1:1,000,000 (http://sis.agr.gc.ca/cansis/nsdb/slc/index.html). This map is the only comprehensive soil map of Canada. Each polygon on the SLC map describes one or more particular type of soil and the landscape associated with it. Version 3.2 of the SLC map was released in March 2011 and is the basis for the maps (Figure 8.1) used in this textbook.

CANADA LAND INVENTORY

The Canada Land Inventory (CLI) was a major effort by a number of federal government agencies to map and classify natural resources in Canada. The CLI was established by the Agricultural Rehabilitation and Development Act of 1961 and mapping began in 1963 and was essentially complete in 1975-76. The CLI covered approximately 2.6 million km^2 (or about 25% of the Canadian land mass) and included assessment for agriculture, forestry, outdoor recreation, and wildlife (Pierce and Ward, 2013). The capability of mapped areas for each activity was placed into 7 classes, with class 1 being the highest capability and class 7 the lowest. Classes 1 to 3 were deemed to be the prime lands for each activity.

The determination of CLI class is made by first assessing the limitations present in each area and severity of the limitation. The area of land being assessed was then assigned the value of the most severe limitation present to arrive at its final class (i.e., from 1 to 7).

Information on soils was of greatest relevance for the agricultural and forestry CLI classes. The agricultural CLI classes required information on ten main criteria: undesirable soil structure and/or low permeability, erosion damage, low natural fertility, inundation risk, low moisture holding ability, surface stoniness, adverse climate (heat units available for growth and/or moisture), salinity, shallow bedrock, and adverse topography. Information on the human context of agriculture (e.g., distance to markets, road infrastructure) was not considered. Note that the soil taxonomic class was not part of the CLI system – this lack of use of the soil taxonomic system in assessments like the CLI was one factor that limited the use of soil surveys by the broader community and contributed to the withdrawal of federal funding for surveys in the 1990s.

The forestry CLI classification included the soil factors used in the agricultural system. It also included factors relating to soil moisture (soil moisture deficiency, soil moisture excess), permeability and rooting depth (physical restriction to rooting caused by dense soil horizons or shallow bedrock) as well as various climate factors.

CLI classes are still used in some jurisdictions in Canada for land use planning purposes. In Ontario, for example, the 2014 Provincial Policy Statement under the Planning Act requires that prime agricultural areas (i.e., those in CLI classes 1 to 3) must be protected and designated for long-term agricultural use (OMAFRA, 2019b).

Can You Dig It!

"Oh, give me land, lots of land under starry skies above…."

How can we compare different provinces and countries in terms of their natural endowment of agricultural land? The most used measure is the area of arable land per person (measured in hectares per person). In Canada we can use the Canada Land Inventory to readily compare across provinces. The highest is Saskatchewan, with an amazing 14.67 ha per person; the lowest is Quebec with only 0.28 ha per person, followed closely by BC and Ontario. However even Quebec has higher values than the global average – according to the World Bank, in 2016 the global average was only 0.19 ha per person. This average has declined from an average of 0.37 ha per person in 1961. Only a few countries have a greater natural endowment than Canada (which has an average of 1.21 ha per person according to the World Bank). For major countries only Kazakhstan (1.65 ha/person) and Australia (a whopping 1.90 ha/person) outstrip Canada.

Soil resources in Canada (Canada Land Inventory Capability Classes 1, 2 and 3*‡) compared with population (estimated for 2019). Table compiled by Tim Moore, McGill University.

Region	CLI 1	CLI 2	CLI 3	ΣCLI 1-3	Population	ΣCLI 1-3 /population
	(thousands of km^2)				(million)	(ha per person)
Atlantic	0	5.89	22.78	28.67	2.4	1.19
QC	0.6	9.54	13.63	23.77	8.4	0.28
ON	22.49	23.6	32.79	78.87	14.4	0.55
MB	1.83	25.56	25.61	53.01	1.4	3.79
SK	10.72	64.46	100.82	176.00	1.2	14.67
AB	7.08	38.97	62.88	108.93	4.3	2.53
BC	0.7	3.98	10	14.68	5.0	0.29

*From the Canada Land Inventory https://sis.agr.gc.ca/cansis/publica...agr/index.html
‡For descriptions of the Land Capability Classes, see https://sis.agr.gc.ca/cansis/nsdb/cli/classdesc.html

SOIL SURVEY

In many areas of the world including parts of Canada, traditional soil survey has been replaced with computer-assisted approaches that are generally termed Digital Soil Mapping (DSM). DSM draws upon all the resource information (including existing soil maps and soil profile data) available for a given location to predict the likely range of soil properties and soil taxonomic classes present at that point. The predictions are especially accurate where high-resolution topographic information is available such as that produced by laser light surveying (i.e., light detection and ranging, or LIDAR). The algorithms that produce the predictions and maps can also estimate the likely error associated with the predictions and hence allow users to assess the uncertainty associated with the maps. The techniques associated with DSM are discussed in more detail in **Chapter 17**.

SUMMARY

1. Soils in Canada are classified according to the Canadian System of Soil Classification (CSSC), a formal taxonomic system with a hierarchy of six levels.
2. The highest level in the CSSC, the order, is based on diagnostic horizons that reflect the effects of the dominant soil-forming process. There are 10 soil orders. The next level, the great groups, are based on differences in the strength of the dominant process or contribution of an additional soil-forming process.
3. The distribution of soils at different scales (e.g., hillslope, catchment, region, province) is controlled by seven soil-forming factors: climate, organisms, parent material, topography, time, groundwater, and human activity.
4. Different classes of forest soils occur on the two main parent materials types in Canada. The CSSC distinguishes between forest soils formed in acidic parent materials (pH levels less than 5.5) and those formed in neutral-alkaline parent materials (pH levels of 5.5 or greater).
5. Forest soils formed in acidic parent materials are classed into the Podzolic order and the Brunisolic order depending on the amount of iron, aluminum, and OM deposited in the middle or B horizon of the soil.
6. Forest soils formed in neutral-alkaline parent materials are classified into the Brunisolic soils and into the Luvisolic order depending on the differences in clay content between the B horizon and the mineral horizon that overlies it.
7. The presence of permafrost layers is diagnostic for soils of the Cryosolic order and the presence and absence of frost mixing is recognized at the great group level.
8. The organically enriched surface soil horizon of grassland soils is diagnostic of the Chernozemic order. Grassland soils on clay-rich parent materials with mixing of horizons due to expanding clays are classified into the Vertisolic order, and those with high sodium contents into the Solonetic order.
9. Wetland soils with thick (> 40 cm) mats of peat are classified into the Organic order, and mineral wetland soils that have gley features such as dull colors and mottling are classified into the Gleysolic order.

10. The scope of the CSSC is limited to those soils found in Canada. For soils not found in Canada either the American USDA system (Soil Taxonomy) or the United Nations system (World Reference Base) is used. The existence of multiple classification systems is a weakness of soil science as a discipline.
11. The soils of Canada have been surveyed (or mapped) at various scales since the beginning of the 20th century. The usefulness of a given soil survey for land use planning is determined by the scale of the survey.
12. Soils data were extensively used in the Canada Land Inventory, which was a major effort in the 1960s and 1970s to map the capability of activities such as agriculture and forestry. The results of the inventory are still used for land use planning in some jurisdictions in Canada.

SUGGESTED READING

Soil Classification and Description for Canada:

Images for Canadian soils and additional soil features required for classification can be found in the Imagery Collection of the Soils of Canada website

Information on soil orders and soil classification keys are in the *Field Handbook for the Soils of Western Canada*. Available at https://soilsofcanada.ca/links.php

USDA Soil taxonomy: Good summary of the properties of each order in soil taxonomy is available from the University of Idaho at https://www.uidaho.edu/cals/soil-orders

In addition to the Field Handbook, there are two books that are very useful in field description of soils:

Heck, R., Kroestch, D.J., Lee, H.T., Leadbetter, D.A., Wilson, E.A. and Winstone, B.C. 2017. Characterizing Sites, Soils & Substrates in Ontario – Volume 1 Field Description Manual. School of Environmental Sciences, University of Guelph. Information about this book is available at:

https://www.uoguelph.ca/ses/sites/default/files/CSSSO%20Vol1%20Announcement.pdf

Watson, Kent. 2007 (Revised 2009, 2014). Soils Illustrated. International Remote Sensing Surveys Ltd., Kamloops, B.C. Information on this book is available at:

https://www.edwardkentwatson.com/soils-illustrated.html

STUDY QUESTIONS

1. What are the seven factors of soil formation? Which factor do you think is most responsible for the sequence of soils seen in Figure 8.3?
2. Which three diagnostic horizons result from the addition of organic material? What setting (e.g., forest, tundra, wetland etc.) would each one likely to be found in?
3. Which two diagnostic horizons result from turbation of soil material? What factor of soil formation do you think is most responsible for each horizon?
4. In the 1930s parts of the Prairie Provinces experienced very severe wind erosion. What soil would have been associated with the wind-eroded surfaces? What factor of soil formation do you think was most responsible for these soils?
5. Salts are associated with very dry soils yet their distribution in the landscape is largely determined by water. Briefly resolve this conundrum.
6. What is key threshold used by the Canadian System of Soil Classification (CSSC) to distinguish between the two main types of parent materials for forest soils?
7. The great Canadian pedologist Paul Sanborn has argued that Folisols are the hardest soils to dig a soil pit in. Based on Figure 8.12 why do you think that this is the case?
8. What two soil forming factors are most responsible for the soil zones (Figure 8.20) of the Canadian Prairies?
9. What two criteria of weathering are used by Soil Taxonomy and the World Reference Base systems of soil classification?

10. You have been asked to find a route for a pipeline using a soil survey completed at Survey Intensity Level 4. Is this feasible? Briefly explain your answer.

REFERENCES

Agriculture Canada. Research Branch. 1987. Soil survey handbook. Volume 1. Technical Bulletin 1987-9E. Available at: http://sis.agr.gc.ca/cansis/publications/manuals/1987-9/1987-9-soil-survey-handbook.pdf

Anderson, D.W. 2010. Vertisolic soils of the Prairie Region. Prairie Soils and Crops Journal 3: 29-36. Available at: https://prairiesoilsandcrops.ca/volume3.php

Anderson, D.W. and Cerkowniak, D. 2010. Soil formation in the Canadian Prairie region. Prairie Soils and Crops Journal 3: 57-64. Available at: https://prairiesoilsandcrops.ca/volume3.php

Anderson, D.W. and Smith, C.A.S. 2011. A history of soil classification and soil survey in Canada: Personal perspectives. Can. J. Soil Sci. 91: 675-694.

Bedard-Haughn, A. 2011. Gleysolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 763-779.

Brierly, J.A., Stonehouse, H.B., and Mermut, A.R., 2011. Vertisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 903-916.

DREE. 1970. The Canada Land Inventory. Objectives, scope and organization. Second Ed., Report No. 1. Ottawa 1970. http://publications.gc.ca/collections/collection_2019/eccc/RE63-1-1970-eng.pdf

Ellis, J.H. 1938. The Soils of Manitoba. Economic Survey Board. Province of Manitoba. September , 1938. 112 pp. Available at: http://sis.agr.gc.ca/cansis/publications/surveys/mb/mbs006/mbs006_report.pdf

FAO. 2019. Soil erosion: the greatest challenge to sustainable soil management. Rome. 100 pp. Available at: http://www.fao.org/3/ca4395en/ca4395en.pdf.

Fox, C.A. and Tarnocai, C. 2011. Organic soils of Canada: Part 2. Upland Organic soils. Can. J. Soil Sci. 91: 823-842.

Jenny, H. 1994. Factors of Soil Formation. A System of Quantitative Pedology. New York: Dover Press (Reprint, with Foreword by R. Amundson of the 1941 McGraw-Hill publication). Download available from NRCS website.

Kroetsch, D.J., Geng, X., Chang, S.X., and Saurette, D.D. 2011. Organic soils of Canada: Part 1. Wetland Organic soils. Can. J. Soil Sci. 91: 807-822.

Lavkulich, L.M. and Arocena, J.M. 2011. Luvisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 781-806.

Miller, J.J. and Brierly, J.A. 2011. Solonetzic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 889-902.

Naeth, A., Archibald, H.A., Nemirsky, C.L., Leskiw, L.A., Brierley, A.J., Bock, M.D., VandenBygaart, A.J., and Chansyk, D.S. 2012. Proposed classification for human modified soil in Canada: Anthroposolic order. Can. J. Soil Sci. 92: 7-18.

OMAFRA 2019a. Classifying Prime and Marginal Agricultural Soils and Landscapes: Guidelines for Application of the Canada Land Inventory in Ontario. http://www.omafra.gov.on.ca/english/landuse/classify.htm#groupings

OMAFRA 2019b. FAQ Prime Agricultural Areas. http://www.omafra.gov.on.ca/english/landuse/prime-ag-areas.htm

Paton, T.R., Humphreys, G.S. and Mitchell, P.B. 1996. Soils A New Global View. Yale University Press, New Haven and London. 234 pp..

Pennock, D.J., Bedard-Haughn, A., and Viaud, V. 2011. Chernozemic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 719-747.

Pierce, T.W and N. E. Ward. The Canada Land Inventory. The Canadian Encyclopedia. https://www.thecanadianencyclopedia.ca/en/article/canada-land-inventory.

Sanborn, P., Lamontagne, L., and Hendershot, P. 2011. Podzolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 843-880.

Smith, C.A.S., Webb, K.T., Kenney, E., Anderson, A., and Kroetsch, D. 2011. Brunisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 695-717.

Soil Classification Working Group. 1998. The Canadian System of Soil Classification. Agric. And Agri-Food Canada. Publ. 1646 (Revised). 187 pp. Available at: http://sis.agr.gc.ca/cansis/publications/manuals/1998-cssc-ed3/cssc3_manual.pdf

Tarnocai, C. and Bockheim, J.G. 2011. Cryosolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 749-762.

VandenBygaart, A.J. 2011. Regosolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 881-887.

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Dan Pennock is a Professor Emeritus (or, more simply, a retired professor) of the Department of Soil Science at the University of Saskatchewan. His research career focused on how soils are shaped by their place in the landscape and by the processes (like erosion or water movement) that occur in landscapes. He taught courses in many areas of soil science and geography, and his love of teaching was recognized by the university with a Master Teacher award in 2006. He was made a Fellow of the Canadian Society of Soil Science in 2010. Since his retirement he has collaborated on a number of projects on sustainable management of soils with the Global Soil Partnership of the Food and Agriculture Organization of the United Nations.

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CHAPTER OVERVIEW

2: Digging Across Canada

2.1: Soils of British Columbia and Yukon- The Western Cordillera

2.2: Soils of the Prairie Provinces

2.3: Soils of Ontario

2.4: Soils of Quebec

2.5: Soils of the Atlantic Provinces

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2.1: Soils of British Columbia and Yukon- The Western Cordillera

Les Lavkulich

Learning Objectives

- On completion of this chapter, students will be able to:
- Describe the geologic history of British Columbia and Yukon
- Identify factors responsible for soil distribution in the region
- Relate the major land uses of the region to soil properties

INTRODUCTION

British Columbia and Yukon are the most diverse areas of Canada in terms of geologic history, physiography, climate, biologic species and land use (Valentine et al., 1978). The diversity in environmental conditions is the result of the geologic history of the region and the complex pattern of mountain chains and valleys that resulted from the Earth-building forces (Figure 9.1). This diversity in environmental conditions in turn results in the greatest diversity of soils in Canada.

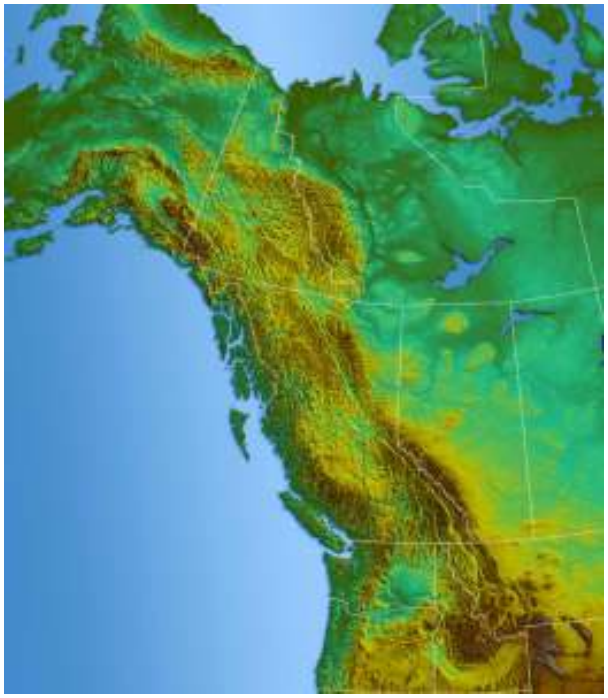


Figure 9.1. Map of northwest Canada showing the relief or British Columbia and Yukon. Source: Mapsofnet.com, http://mapsof.net/british-columbia/northwest-relief (CC-SA). © Mapsofnet.com is licensed under a CC BY-SA (Attribution ShareAlike) license.

Most of British Columbia and Yukon are located in the Pacific Cordillera, a mountain chain (or cordillera) extending for a distance of over 6,000 km along the western side of North America, from Mexico to Alaska. Also known as the North American Cordillera or Western Cordillera, the Pacific Cordillera includes parts of Alberta and the Northwest Territories, including the Rocky and Coast Mountains. It is characterized by landscapes of mountains, isolated valleys, broad plateaus, and coastal islands. This complex topography contributes to topographically induced climate zones and a complex assemblage of vegetation types (Scudder, 1997), resulting in four major terrestrial ecozones (Taiga Cordillera, Boreal Cordillera, Montane Cordillera, and Pacific Maritime).

GEOLOGIC HISTORY

Located on the Pacific Ring of Fire, a zone characterized by active volcanoes and earthquakes, the western Cordillera is the result of the contact of the Cascadia subduction zone, that extends from Vancouver Island to northern California, and the Juan de Fuca, and North American tectonic plates (Monger and Price, 2002).

The Pacific Cordillera is the youngest of the three geologic regions in western Canada and is known for frequent earthquakes and sporadic volcanic eruptions (Figure 9.2). A series of crustal collisions, volcanic episodes, and periods of metamorphism formed the complex landscapes of the Cordillera (Monger and Price, 2002). The geologic history has resulted in a wide range of bedrock formations that include igneous, sedimentary, and metamorphic rocks, and resulted in a variety of mineral and other exploitable resources, notably copper, asbestos and abundant coal and fossil fuel deposits. Although there are hundreds of volcanic fields and centres in British Columbia, there are only 18 true volcanoes. The largest group is located in the north near the coast, while others are scattered throughout the province (Figure 9.2). The largest and most persistent volcanoes are the Mount Edziza complex and Level Mountain in northwestern British Columbia, which have had volcanic activity for millions of years. The Mount Edziza volcanic complex had about 20 eruptions over the past 10,000 years, with the most recent eruption in Canada in 1904 (Souther, 1992). The Mount Edziza volcanic complex was designated as a provincial park in 1972 to protect the volcanic landscape. Another relatively recent eruption was in the 18th century, when the Tseax Cone erupted killing over 2,000 people due to the toxic volcanic gases released during the eruption (Williams-Jones et al., 2019).

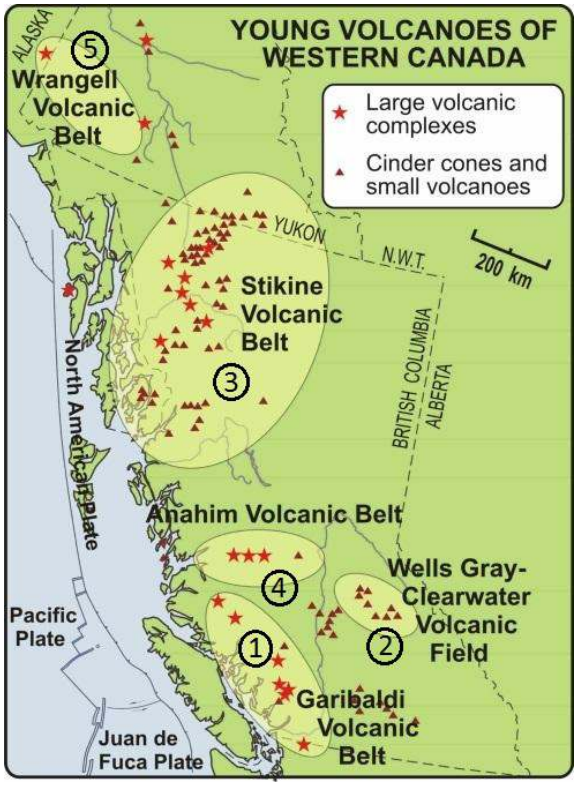


Figure 9.2. Canada has five potentially active volcanic areas, all of which are located in British Columbia and the Yukon Territory. Source: *Volcanoes of Canada. Where are Canada's volcanoes?* (2020). © Natural Resources Canada; licensed under the Open Government Licence – Canada.

In the Yukon, there are more than 100 volcanoes that have been active during the Quaternary period (covers 2.58 million years ago to present). Around 800 AD, Mount Churchill near the Alaska-Yukon border erupted and the ejected ash, covering a large part of the Yukon Territory. The ejected ash, or tephra, is known as the 'White River Ash' (Reuter et al., 2020). The layers of volcanic ash deposits may still be seen in the soils throughout the central and southern Yukon (Dampier et al., 2011; Preece, et al., 2014).

The Cordillera region was almost entirely covered by ice during the Pleistocene glaciation that extended from the coastal mountains in Alaska, south to the Yukon and British Columbia (Figure 9.3). Glaciation resulted in a lowered global sea level, including the formation of the Bering land bridge, which linked Asia with North America creating the ice-free Beringia region. The St. Elias Mountains interrupted and prevented ice sheets from forming in Beringia. Thus, the west-central and northern Yukon remained ice-free. This cold and arid grassland was the home to woolly mammoth, horses and lions (MacDonald et al., 2012).

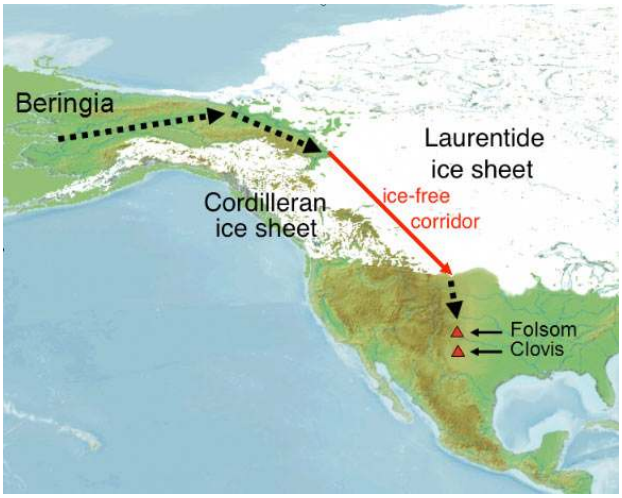


Figure 9.3. Maximum glaciation during Pleistocene – green areas within the glaciers were ice free. Image credit: Roblespepe. http://www.sci-news.com/paleontology...ins-03930.html (CC BY-SA 3.0). © Roblespepe is licensed under a CC BY-SA (Attribution ShareAlike) license.

SOIL DISTRIBUTION

The soils of British Columbia and Yukon include all ten of the soil orders recognized by the System of Soil Classification for Canada (Figure 9.4). This remarkable diversity reflects the wide range of soil forming factors (time, parent material, climate, biota, topography and human activity) contributing to soil formation.

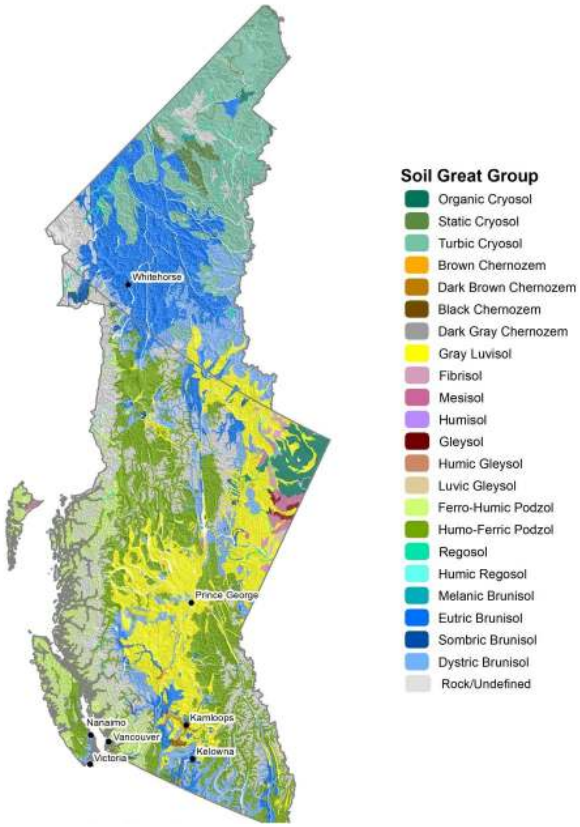


Figure 9.4. Soil great groups of British Columbia and Yukon. Map courtesy of Darrel Cerkowniak, Agriculture and Agri-Food Canada. This figure is a reproduction of an official work published by the Government of Canada. © Darrel Cerkowniak, Agriculture and Agri-Food Canada, is licensed under a CC BY (Attribution) license.

The complex geological and glacial history contributed to a complex pattern of soil parent materials. **Parent materials** range from present-day alluvium, colluvium and marine sediments, to organic deposits, and a complete suite of morainal (glacial drift)

deposits, residual, bedrock, and mine wastes of tailings and rock dumps (Figure 9.5).

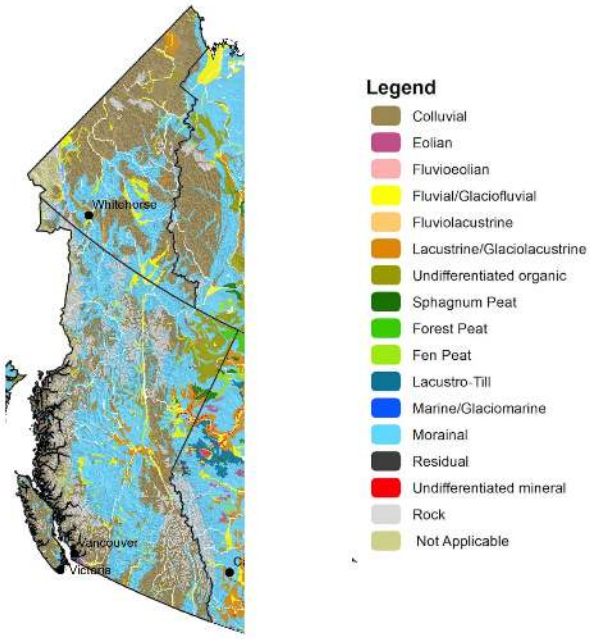


Figure 9.5. Parent Materials of soils in British Columbia and Yukon. Map courtesy of Darrel Cerkowniak, Agriculture and Agri-Food Canada. This figure is a reproduction of an official work published by the Government of Canada. © Darrel Cerkowniak, Agriculture and Agri-Food Canada, is licensed under a CC BY (Attribution) license.

The particle size and chemistry of colluvial and morainal parent materials are largely inherited from the bedrock from which the parent material originated. The bedrock formations of British Columbia and Yukon have examples of all three general classifications of rock types – igneous, sedimentary and metamorphic (see Chapter 2). The Coast Mountains are predominately interlocking igneous and metamorphic rocks ranging from granites and granodiorites to basalts, schists and gneisses, while the eastern perimeter is primarily folded and faulted sedimentary shales, limestones and sandstones. The interior mountains and plateaus are dominated by a complex of sedimentary, metamorphic, volcanic and scattered igneous intrusions.

Climate also influences the pattern of soil distribution, and is most strongly controlled by the mountain chains within the Cordilleran region. The mountain chains control the pattern of precipitation, with generally wetter conditions on the western slopes and drier conditions on the eastern slopes and in intermontane valleys. A dominant north-south temperature pattern is typical along with altitudinal temperature gradients in the mountain chains. These climatic patterns strongly influence the pattern of vegetation distribution; hence, the Western Cordillera contains both the most desert-like vegetation assemblages and the wettest rainforest in Canada, within four major terrestrial ecozones (Taiga Cordillera, Boreal Cordillera, Montane Cordillera, and Pacific Maritime)

Biota is a soil forming factor that exerts a strong influence on soil distribution. The pattern of forest, tundra, grassland, and rockland distribution in British Columbia and Yukon varies greatly and is controlled by the impact of the mountain chains on weather patterns and climate. Interestingly, there is far greater diversity of plant species within the forests of this region than in comparable areas of the boreal forest or in the Prairie region (Andrews et al., 2019).

Human activities within the region reflect the diversity of soil types in the Western Cordillera. An interesting example of the impact of human activity of soil formation can be found at the Gulf Islands in British Columbia. Prolonged use by the Indigenous Peoples over thousands of years led to formation of unique anthropogenic soils from the Neptune soil series (Figure 9.6). These soils consist of shallow gravelly loamy sand to gravelly sand marine deposits mixed with clam and oyster shells, organic debris, and sometimes human artifacts. They are characterized by high calcium carbonate content from the shells, which prevents any significant profile development as indicated by the absence of a B horizon. Other examples of the impacts of human activity on soil formation in the Western Cordillera date to the period of European settlement and introduction of mining activities, beginning with the famous Klondike (1847) and Cariboo (1858) gold rushes, followed by extensive mining of asbestos and metals—notably copper, iron, and molybdenum—and fossil fuels (Levenson, 1992). One legacy of these activities is the presence of the large areas of mine spoil landscapes with fresh, human-derived “parent material” that provide challenges for ecological restoration.



Figure 9.6. A soil profile of Neptune soil at the Galiano Island, British Columbia showing a layer of shell deposits. © Maja Krzic, Univ. of British Columbia is licensed under a CC BY (Attribution) license.

The interactions among the various factors affecting soil formation (see Chapter 2) are reflected in Canada's terrestrial ecozones (Figure 9.7). The ranges within the soil forming factors for the ecozones found in BC and Yukon are presented in Table 9.1.

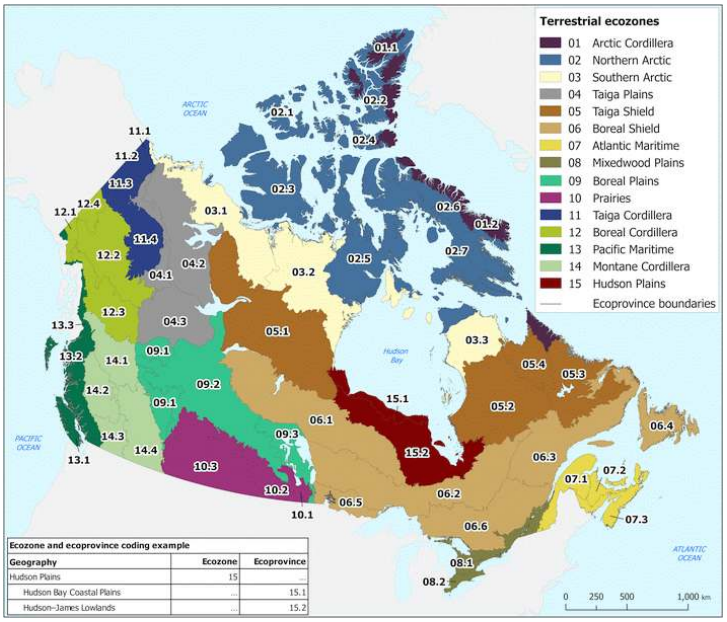


Figure 9.7. The terrestrial ecozones and ecoprovinces of Canada. © Statistics Canada; used with permission and licensed under the Open Government Licence – Canada. https://open.canada.ca/en/open-government-licence-canada

Table 9.1. Major climate, vegetation, parent material, topography, and soil attributes of the Taiga Cordillera, Boreal Cordillera, Pacific Maritime, and Montane Cordillera ecozones. Text is abstracted from Ecological Stratification Working Group (1996).

Taiga Cordillera Ecozone	
Climate	Annual precipitation ranges from less than 300 mm in the north to more than 700 mm in the southeast. Mean annual temperatures range from -10°C in the north to -4.5°C in the south.
Vegetation	Natural vegetation ranges from arctic tundra in the north, to alpine tundra at higher elevations, and taiga or open woodland in the south (white spruce and white birch), mixed with medium to low shrubs, mosses, and lichens.
Topography	Steep, mountainous topography, consisting of repetitive, sharply etched ridges and narrow valleys, predominates with foothills and basins also present.
Bedrock and Parent Materials	The bedrock is largely sedimentary in origin with minor igneous bodies. Much of the area is mantled with colluvial debris with frequent bedrock exposures and minor glacial deposits.
Soils	Dominantly Turbic Cryosols
Boreal Cordillera Ecozone	
Climate	The climate is marked by long, cold winters and short, warm summers and modified by vertical zonation and aspect. Mean annual temperatures range from 1.0°C to 5.5°C. Mean annual precipitation is lowest in valleys within the rain shadow of the coastal ranges (<300 mm) and increases in the interior ranges farther east, where up to 1500 mm of precipitation is received at higher elevations. Precipitation in the intermontane plateau areas ranges 300-600 mm annually.
Vegetation	The vegetative cover ranges from closed to open forest canopies over much of the plateaus and valleys. Tree species include white and black spruce, alpine fir, lodgepole pine, trembling aspen, balsam poplar, and white birch. At higher elevations, there are extensive areas of rolling alpine tundra.
Topography	This ecozone is characterized by mountain ranges that contain numerous high peaks and extensive plateaus, and are separated by wide valleys and lowlands.
Bedrock and Parent Materials	Bedrock is a mixture of sedimentary, igneous, and metamorphic rocks. Glacial drift, colluvium, and outcrops constitute the main surface materials.
Soils	Eutric Brunisols and smaller areas of Turbic Cryosols dominate the Yukon portion of this ecozone, and Humo-Ferric Podzols and rock outcrops dominate the BC portion.
Pacific Maritime Ecozone	
Climate	The ecozone has some of the warmest and the wettest climatic conditions in Canada. Mean annual temperatures range from 4.5°C in the north to 9°C in the Georgia-Puget Basin - Lower Mainland regions. Relative to the rest of Canada, there is little variation

Taiga Cordillera Ecozone	
	between the mean monthly temperatures through the year. The annual precipitation ranges from as little as 600 mm in the Gulf Islands of lower Strait of Georgia to over 4000 mm in the Coastal Gap region to the north. Overall, the zone typically receives 1500-3000 mm of precipitation per year.
Vegetation	Variations in altitude account for the presence of widely contrasting ecosystems within the ecozone, ranging from mild, humid coastal rainforest to cool boreal and alpine conditions at higher elevations.
Topography	Mountainous topography dominates, cut through by numerous fjords and glacial valleys and bordered by coastal plains along the ocean margin.
Bedrock and Parent Materials	Igneous and sedimentary rocks underlie most of the area. Colluvium and glacial deposits are the main surface materials.
Soils	Podzolic soils along the coastal areas with smaller areas of Folisols. Rock outcrops at higher elevations.
Montane Cordillera Ecozone	
Climate	The climate of the region ranges from subarid to arid and mild in southern lower valleys to humid and cold at higher elevations in the northern reaches. Moist Pacific air and the effect of orographic rainfall control the precipitation pattern such that both rain shadows and wet belts are generated within the ecozone. The rain shadow cast by the massive coast mountains results in some of the driest climates in Canada in the valley bottoms of the south-central part of the ecozone. Mean annual temperatures range between 0.5°C in the northwest (Skeena Mountains) to 7.5°C in the Okanagan area along the Canada-United States border. Annual precipitation drops from 1200-1500 mm in the mountains and ranges to the west, to 500-800 mm in the north and interior and rises again to 1200 mm in the mountains and ranges along the British Columbia-Alberta border. Precipitation falls below 300 mm in the arid valleys and plateaus to the south.
Vegetation	Vegetative cover is extremely diverse; alpine environments contain various herb, lichen, and shrub associations, whereas the subalpine environment has tree species such as lodgepole pine, alpine fir, and Engelmann spruce. With decreasing elevation, the vegetation separates into three general forest groups. Shrubs and some remnant grasslands are found in the dry southern interior.
Topography	It is a rugged, mountainous ecozone that incorporates several major interior plains. The plains are more extensive in the north and extend out as intermontane valleys towards the southern half of the ecozone.
Bedrock and Parent Materials	Most of these plains and valleys are covered by glacial moraine and, to some degree, fluvial and lacustrine deposits, whereas the mountains consist largely of colluvium and rock outcrops.
Soils	Dominated by Gray Luvisols in the central portion with Humo-

Taiga Cordillera Ecozone	
	Ferric Podzols to the north and east and Eutric Brunisols in southern portions. Small areas of Chernozems in the semiarid southern portion.

UNIQUE FEATURES OF MAJOR SOIL TYPES

The soils of British Columbia and Yukon reflect the natural diversity of the region and each have representatives of all the soil orders recognized by the Canadian System of Soil Classification (Soil Classification Working Group, 1998). For example, some landscapes are prone to landslides and avalanches, on which weakly developed soils, the Regosols, can be found. Permafrost regions are dominated by Cryosols. Gleysols are most common in coastal regions. Chernozems and Vertisols are found on the glacial drift deposits in the grasslands of the British Columbia interior, although the occurrence of Vertisols, in general, is rare. Solonetzic and Luvisolic soils are predominant on the glacio-lacustrine parent materials common in the northeastern portion of British Columbia, namely the Great Plains region. Brunisolic soils are widespread from the mountainous alpine to central Yukon, to the lowlands of British Columbia and the eastern perimeter, the Rocky Mountains. Organic soils are widespread with larger areas found in the northern Yukon, and northern and west coastal British Columbia. The most productive forested soils in Canada, the Podzolic soils, dominate the west coast and the Pacific coastal islands. It is estimated that about 0.05 percent of British Columbia, or 45,000 hectares, has been affected by mining activity. Thus, the region has “soils” that may be classified as Anthroposols.

Regosols

Regosols are defined as soils that exhibit only initial weathering and formation of soil horizons, and thus often lack a recognizable B horizon. Landslides, avalanches and accelerated erosion common in the rugged topography (colluvial parent materials) and the broad floodplains (alluvial deposits) of the Fraser and Yukon Rivers of British Columbia and Yukon shorten the effects of time factor of soil formation to produce well-developed B horizons. In addition, Regosolic soils are common along the broad Fraser River Delta and on the mountainous landscapes. These steep areas with sparse vegetative cover are popular for outdoor recreational activities, such as skiing and the use of snowmobiles and cross-country recreational vehicles. Conversely, Regosols along flood plains and the Fraser Delta support highly productive and diverse agricultural production, once they are protected from periodic floods.

Brunisols

Brunisolic soils are widely distributed on a range of parent materials, both acidic and basic, and mainly form in glacial drift parent materials. These soils exhibit leaching of soluble salts and emerging chemical weathering resulting in a B horizon characterized by a slightly different colour compared to the C horizon or the parent material. Brunisols are associated with both temperate and boreal forests, and are also are common in alpine ecosystems. Brunisols may be found in association with Cryosols in regions of discontinuous permafrost in northern Yukon, and with Podzolic soils in more humid forested ecosystems of British Columbia as well as with Luvisols in the forest-grassland ecotones. They are the major soil order in eastern British Columbia and the Central Interior and the alpine zones of the Cordillera. The soils support extensive areas of commercial forests, dominated by lodgepole pine (*Pinus contorta* Douglas ex. Loudon), rangelands in the intermontane eastern British Columbia, as well as agriculture in the Interior plateaus and the Lower Mainland of British Columbia. In alpine regions, because of the cooler climatic regimes, the A horizon tends to be rather shallow and vulnerable to summer activities associated with grazing animals and motorized recreational vehicles. Brunisols provide the foundation for much of the scenic vistas of the Canadian Cordillera.

Podzols

The most common soil (i.e., largest geographical extent) in British Columbia is the Podzol. Podzols are predominant in regions with cool soil temperatures and humid climates. Podzolic soils are the true forested soils, found on acidic parent materials of Canada. These soils are not as common in the Yukon, as they are formed most commonly in temperate climates, although they do extend to the Boreal Forest and the Canadian Shield. Podzolic soils are acidic soils usually found on relatively coarse parent materials such as sandy till, glacial fluvial deposits and colluvium. The relatively coarse texture and the humid climate provide an ideal environment for coniferous forested ecosystems. This combination of humid climate, acidic parent material and presence of coniferous forest with its acidic litter results in relatively intense chemical weathering and the leaching of base-forming cations. The rapid decomposition of the acid forest litter and associated release of organic acids with strong chelating properties, enhances

movement of ions within the soil profile. The result is a soil with a recognizable forest floor (or LFH horizon sequence), a leached and acidic surface horizon (commonly an Ae) and an enriched B horizon with the accumulated weathering products that are dominated by the non-crystalline hydrated aluminum and iron oxides. The resulting morphological feature is the reddish coloured B horizon caused by the strongly pigmented iron oxides (Bf) as well as organic matter (Bhf).

Luvisols

Luvisolic soils are one of the two dominant soil orders in British Columbia and are less common in the Yukon. The soils are found in the Interior plateaus and are common in the Boreal forest. Luvisolic soils are dominant in forested landscapes commonly on loamy glacial tills or glacio-lacustrine deposits. Luvisols have a textural difference between the A and the B horizon, where the lower clay content Ae horizon is the result of a decrease in inherited clay-sized particles by the physical movement by percolating water and subsequent increase by deposition in the Bt horizon. This may be observed as oriented deposits of clay along the walls of the soil pores; and the features are called clay skins or cutans. In British Columbia, less so in the Yukon, these soils have been extensively converted from forests to agriculture, notably in the Central Interior and the northeastern Peace River region of British Columbia. The soils are highly prized for forest harvests, notably lodgepole pine and both coastal Douglas-fir (*Pseudotsuga menziesii var. menziesii*) and interior Douglas-fir (*Pseudotsuga menziesii var. glauca*). Once cultivated, the Luvisol loses its characteristic greyish coloured Ae horizon, while mixing of the LFH and the relatively thin Ah, leads to formation of an Ap.

Cryosols

Cryosolic soils are geographically limited to the northern Yukon in the areas of permafrost and in the Boreal and few locations on the higher mountain peaks in the Alpine regions. These soils are found on both mineral deposits (Turbic and Static Cryosols) and organic deposits (Organic Cryosols). Associated ecosystems include the Arctic and Alpine tundra, and Subarctic and Boreal forests. The cold temperatures have physically protected organic matter stores and preserved thousands of years of soil carbon. With disturbance, Cryosols rapidly lose the stored carbon that was physically protected and become carbon sources.

Chernozems

Soils of the Chernozemic order are limited in British Columbia and the Yukon, and they only occur in the Central Interior regions. They are found in the relatively more arid regions of the Cordillera. In British Columbia, beginning at the Canada – USA border in the south, which is almost a true desert, and moving north, the climate becomes more humid, which results in a sequence of Brown Chernozems that gradually transitions to Dark Brown and Black Chernozems. The Chernozems serve as the resource base for vineyards and the agriculturally important fruit growing soils of the Okanagan. In the northern section, the Chernozemic soils are associated with the ranching by providing both grazing lands and forage for the industry. These rangelands suffered serious overgrazing in the past. Fortunately, by the adoption of conservation and best management practices, these lands have been restored. Northwest of the Fraser River is where some of the largest cattle ranches in North America are found. Except for animal grazing, agricultural activities require irrigation to be efficient and economically feasible. Limited, and rather minor, areas of Chernozems are found in the Yukon, dominantly on south-facing aspects (Pennock et al., 2015).

Solonetz

Soils of the Solonetzic order are developed from the saline lacustrine parent materials in the northeastern portion of British Columbia and are located on the Great Plains physiographic region. Solonetzic soils are found where there is thin layer of glacial material overlying a shallow bedrock composed of saline and alkaline marine shales. Solonetzic soils generally occur on topographic relief that is level to undulating, especially on lowlands with restricted drainage. These soils present problems in agriculture, namely the acidic Ah horizon and the dense compacted Bnt horizon. It has been suggested that deep sub-soiling would aid in mitigating the concern as the deep sub-soiling has been successful in incorporating bases, notably calcium from depth which helped neutralize the acidic surface A horizon and increased the porosity of the dense B horizon.

Vertisols

Vertisolic soils are rare, or possibly do not occur, in British Columbia and Yukon. Vertisolic soils are associated with a level, or near-level, lacustrine surfaces that have high clay contents (>60%) dominated by the swelling clays. Soils with vertic properties are marked by swelling and shrinking during alternating dry and wet cycles sufficient to mix the upper solum of the soil matrix and are found in the Okanagan region of British Columbia. There is debate about whether these soils are true Vertisols or intergrades to the

surrounding Chernozems. The main concern with the management of these soils for agriculture is high water holding capacity, which restricts cultivation in the spring and the limited water infiltration following extreme rainfall events.

Gleysols

In low lying areas throughout the region and notably in the Lower Fraser Valley, Gleysols are common. Most occur on lacustrine, glacial marine and marine parent materials. The artificially drained Gleysols are some of the most productive agricultural soils in British Columbia. Those in the Fraser Valley experience an ideal climate to support a diverse range of agricultural crops and to support dairy and poultry industry. Gleysolic soils, and the associated Organic soils, in the Fraser Valley are of great importance for the cranberry production in Western Canada (Bertrand et al., 1991).

Organic Soils

Organic soils (soils that contain greater than 30% organic matter by mass, equivalent to 17% organic carbon) are widely spread throughout Yukon and British Columbia. Generally, Organic soils occur when organic matter deposition exceeds the decomposition rate. Formation of Organic soils is favoured by cold climates and/or prolonged saturation. Prolonged water saturation result in anoxic conditions that reduce the rate of organic matter decomposition. However, Folisols, which are one of four great groups of the Organic soil order, form under cool, humid and well aerated conditions. In these environments, large amounts of organic matter accumulate directly on the bedrock and the accumulation exceeds the decomposition rate even under aerated conditions. Folisols occur frequently in cool, moist, and humid forest ecosystems, particularly on the west coast of Canada in the Coastal Western Hemlock biogeoclimatic Subzone of British Columbia. Dominant tree species include western hemlock (*Tsuga heterophylla*), western red cedar (*Thuja plicata*), and some amabilis fir (*Abies amabilis*). Folisols occur very rarely in other upland regions of Canada.

In the northern Yukon, adjacent to the regions of continuous permafrost and throughout areas located at high elevation mountain positions, the cold environments are conducive to the formation of Organic soils. In the more temperate regions, low lying and relatively flat landscapes, especially on the finer textured lacustrine and marine parent materials, have inherently poor water drainage and over time the soil water becomes anoxic (absence or near absence of oxygen), which results in the formation of Organic soils. Organic soils are important sinks for carbon, as are Crysosols, and are equally vulnerable to the release of greenhouse gases following drainage or physical disturbance which cause a warming of soil climate or increased aeration. Organic soils vary from relatively undecomposed organic materials (fibric matter) commonly called peat to highly decomposed muck (humic matter). As indicated, Organic soils are associated with all other soil orders and thus there is a continuum or transition from these other soil orders to the Organic order. Historically, some areas of Organic soils have been subjected to “peat” harvesting, especially in British Columbia. In the Lower Fraser Valley, these harvested Organic soils areas have been cultivated, incorporating the residual organic materials with the underlying mineral parent materials and the fields converted to cranberry production. Also, in the Lower Fraser Valley, near Vancouver, Burns Bog is noted for its natural history. It was formed over 10,000 years ago during the Pleistocene and is the largest raised peat bog on the West Coast of the Americas. Burns Bog is habitat to more than 300 plant and animal species, and 175 bird species (Hebda, 2000).

Anthroposols

The Anthroposols, soils which have been profoundly modified or constructed by humans, typically have one or more of their natural soil horizons modified, removed or replaced (Naeth et al., 2012). Anthroposols occur throughout British Columbia and Yukon, as the entire region is known for metal, coal, oil and gas extraction and aggregate mining. These activities have resulted in local areas of mine “spoil.” In areas of metal mining, the primary concern is acid rock (mine) drainage. Most metal mines contain sulphide minerals, such as pyrite and pyrrhotite, which rapidly weather to form sulfuric acid when exposed to water and the oxidizing atmosphere. The leaching of this strong oxidizing acid causes considerable environmental concern with water quality and aquatic trophic levels. Waste rock dumps resemble colluvial deposits and are usually coarse textured with low available water holding capacity and poor nutrient retention. There are also environmental concerns regarding reclamation of the large open pits, formed by the mining extraction process, as these scars affect the hydrologic cycle. Dark-coloured, or black coal wastes have the additional problem of high absorption of solar energy, reaching temperatures that prevent plant survival. Oil and gas extraction and fracking sites leave behind non-commercial organic products that may contain toxic components that could negatively impact the environment.

APPLICATIONS OF SOIL INFORMATION

Canada, including British Columbia and Yukon, is rich in natural resources that support a healthy economy and contribute to global equity, notably through its export commodities and recreational attractions. The economies of British Columbia and Yukon, mainly based on agriculture, forestry, and mining and outdoor recreation depend directly and indirectly on these soil resources. Cumulatively these activities placed increased demands on the soil resource, modifying the diverse landscapes of British Columbia and Yukon by drainage projects, transportation networks, extractive industries, urban centers and organized recreational facilities.

One result of the historical pressures on soil resources was the establishment, in 1961, of the Canada Land Inventory (CLI) (CLI, 2002). The CLI provided a nationwide assessment of Canada's land base for renewable resources. The CLI assessed about 2.6 million km^2 or 25% of Canada's land base. The assessments were based on land capability for agriculture, forestry, outdoor recreation and wildlife (waterfowl, ungulates and sportfish). In the case of agriculture and forestry, the capability was based on soil limitations for each activity within the local climatic regimes (Tables 9.2 and 9.3). The classification for land capability for agriculture indicates the degree of limitation imposed by the soil in its use for mechanized agriculture. Class 1 has no limitations for the range of climatically suited crops, while Class 7 has no capability. The limitations include excess stoniness, shallow sola, excessive steepness and soil drainage.

Table 9.2. Land capability classes for agriculture according to the Canada Land Inventory (http://sis.agr.gc.ca/cansis/nsdb/cli/index.html, modified 2016-04-18).

Class	Land capability
1	No significant limitations to use for crops
2	Moderate limitations the restrict the range of crops or require moderate conservation practices
3	Moderately severe limitations that restrict the range of crops or require special conservation practices
4	Severe limitations that restrict the range of crops or require special conservation practices, or both
5	Very severe limitations that restrict their capability to producing perennial forage crops, but improvement practices are feasible
6	Capable of producing perennial crops only, and improvement practices are not feasible
7	No capability for crop use or permanent pasture
0	Organic soils (not placed in capability classes)

Forest Land Capability differs from the classification for agriculture by adding an estimate of forest productivity given as annual tree growth, expressed as the mean annual increment (MAI). The best lands for commercial forest production are Class 1, while Class 7 lands are not suitable for commercial forests. The Canada Land Inventory was developed to provide a national standard; however, each province was mapped independently, and as a consequence, regional variations are known to occur. For example, in British Columbia, forest growth exceeds the limits that were set as standards across Canada (McCormack, 1970). Forest productivity on the Podzolic soils of the west coast, including Vancouver Island, have greater growth rates than those defined for in the CLI for Class 1 land. Thus, Class I was expanded by increments of two cubic meters per ha per year into Class,1a, 8-10; Class, 1b, 10-12; Class 1c, 12-14; and Class 1d, 14-16.

Table 9.3. Land capability classes for forestry according to the Canada Land Inventory (http://sis.agr.gc.ca/cansis/publicat...aps/index.html, modified 2013-06-25).

Class	Land capability for growth of commercial forests	Productivity (MAI*)
		$m^3\ ha^{-1}\ yr^{-1}$
1	Land with no important limitations	Usually greater than 7.8
2	Land with slight limitations	6.4-7.7
3	Land with moderate limitations	5.0-6.3
4	Land with moderately severe limitations	3.6 – 4.9
5	Land with severe limitations	2.2-3.5
6	Lands with very severe limitations	0.8-2.1
7	Land unsuitable for the growth of commercial forests	Usually less than 0.8
*MAI – mean annual increment		

These two examples of capability classifications illustrate the importance of soil resources in developing land use guidelines and informing government policy. Although the CLI uses the word “land” the main criterion for assigning the capability classes are the properties of the soil.

Another example that has gained considerable international attention to conserve soils for agriculture is the action taken in British Columbia to maintain food security through agricultural land use planning. British Columbia is land poor in relation to the amount of productive soils for agriculture. Roussin et al. (2015) referred to agriculture in British Columbia as “pocket agriculture” to illustrate the nature of scattered pockets of agricultural lands within the diverse landscapes. In 1973, the government of British Columbia passed legislation to protect the most productive lands for agriculture (ALC, 2014). The basis of the legislation was soil capability for agriculture, notably Classes 1-4. Thus, all soil units that have that range of capability are included in the provincial Agricultural Land Reserve. Although the designation is agricultural land, the designation is not a dictation of land use, the designation serves as a mechanism, according to government regulation, to protect prime soils from development and to maintain the land for soil-based agriculture.

The importance of the soil resource also has been recognized by urban planners. In British Columbia, soil information is required to identify Best Management Practices required by the province for community site plan development. The soil information helps to identify sensitive environmental features for development planning. Examples for which soil information is required include: identifying sensitive areas and avoiding development in these areas; designing development to retain important ecosystem features and important ecosystem functions; designing sites to maintain hydrology and water quality; planning trails carefully; maintaining habitats in urban areas; restoring groundwater for storm water management, preserving or improving water quality and retaining water through rate control or detention; and developing an erosion and sediment control plan during construction (Ministry of Water, Land and Air Protection, 2004).

As demand for ecological goods and services increases, so does the need to understand how different soils and their unique properties impact food security, environmental quality, and public safety. The soils of British Columbia and Yukon, and the complex environment of the Western Canadian Cordillera, reveal the dependence of the region on the soil resource, from the need of Indigenous Peoples for food and shelter, to the needs of more recent economic development that encompasses mining, agriculture, forestry and urban development. Natural soil resources have been modified by extractive activities, conversion of forests to agriculture, and by installation of drainage systems for agricultural production and urban expansion. Changes to the landscape as a result of human activity have been recognized by the introduction of the concept of Anthroposols, as a new class of “parent materials” and incipient soils. The province of British Columbia has recognized the need for soil conservation, by legislation, to protect the most productive soils for agriculture. The soil is the “environmental indicator” and the understanding of soil processes and properties is essential for responsible management of this critical natural resource.

REFERENCES

ALC. Provincial Agricultural Land Commission. 2014. www.alc.gov.bc.ca

Andrews, M., R. Gonzalez, K. Abram, J. Mitchem, and E. Hernanadez. 2019. Biomes – Canada https://canada-exhibit.weebly.com/biomes.html

Bertrand, R.A., G.A. Hughes-Games, and D.C. Nikkel. 1991. Soil management handbook for the Fraser Valley. V. Soils and Engineering Branch Ministry of Agriculture, Fisheries and Food. Abbotsford, British Columbia. http://www.env.gov.bc.ca/esd/distdata/ecosystems/Soils_Reports/Docs/soil_mgmt_hbk_fraservalley.pdf

Canada Land Inventory (CLI). 2013. Land Capability for Agriculture [computer file]. Ottawa, ON: Agriculture and Agri-Food Canada.

Canada Land Inventory (CLI). 2002. Land Capability for Forestry [computer file]. Ottawa, ON: Natural Resources Canada.

Chilton, R.H. 1981. A summary of climate regimes of British Columbia. B.C. Min. Environ., Victoria, B.C.

Dampier L., P. Sanborn, S. Smith, J. Bond, and J. Clague. 2011. Genesis of upland soils, Lewes Plateau, central Yukon. Part 1: soils formed on Pleistocene glacial deposits. Can. J. Soil Sci. 91(4): 563-578, https://doi.org/10.4141/cjss2010-001

Ecological Stratification Working Group. 1996. A National Ecological Framework for Canada. Centre for Land and Biological Resources Research, Research Branch, Agriculture and Agri-Food Canada and State of the Environment Directorate, Environment Conservation Service Environment Canada. Ottawa Ontario.

Hebda, R. 2000. Burns Bog Ecosystem Review Synthesis Report. Environmental Assessment Office – Province of British Columbia.

Levenson, V.M. 1992. Pioneering Geology in the Canadian Cordillera. British Columbia. Geological Survey. BCMEMPR Open File 1992-19. pp. 1-55.

McCormack, R.J. 1970. The Canada Land Inventory. Land Capability Classification for Forestry. Report No. 4. Dep. Reg. Econ. Expan. Ottawa, Ontario. 72 p.

MacDonald, G.M., D.W. Beilman, Y.V. Kuzmin, L.A. Orlova, K.V. Kremenetski, B. Shapiro, R.K. Wayne, and B. Van Valkenburgh. 2012. Pattern of extinction of the woolly mammoth in Beringia. Nat. Commun. 3:893 https://doi.org/10.1038/ncomms1881.

Marshall, I.B., Schut, P.H., and Ballard, M. 1999. *A National Ecological Framework for Canada: Attribute Data*. Agriculture and Agri-Food Canada, Research Branch, Centre for Land and Biological Resources Research and Environment Canada, State of the Environment Directorate, Ecozone Analysis Branch. Ottawa/Hull. http://sis.agr.gc.ca/cansis/nsdb/eco...ort/index.html (accessed July 4, 2021).

Ministry of Water, Land and Air Protection. 2004. Environmental Best Management Practices for Urban and Rural Land Development. Ecosystem Standards and Planning Biodiversity Branch, Government British Columbia, Victoria, Canada.

Monger J., and R. Price. 2002. The Canadian Cordillera: Geology and Tectonic Evolution. CSEG Recorder, Feb. pp.17-36. http://74.3.176.63/publications/recorder/2002/02feb/feb02-canadian-cordillera.pdf

Naeth, M.A., H.A. Archibald, C.L. Nemirsky, L.A. Leskiw, J.A. Brierley, M.D. Bock, A.J. Vandenbygaart, and D.S. Chanasyk. 2012. Proposed classification for human-modified soils in Canada: Anthroposolic order. Can. J. Soil Sci. 92: 7-18.

Pennock, D.J. and P. Sanborn. 2015. Section 1: Soil Genesis and Geographical Distribution. From: D. Pennock, K. Watson, and P. Sanborn. Field Guide to the Soils of Western Canada. Canadian Society of Soil Science.

Preece, S.J., R.G. McGimsey, J.A. Westgate, N.J.G. Pearce, W.K. Hart, and W.T. Perkins. 2014. Chemical complexity and source of the White River Ash, Alaska and Yukon. Geosphere 10(5): 1020–1042. doi: https://doi.org/10.1130/GES00953.1

Reuter, J., B. Potter, S. Coffman, H. Smith, and N. Bigelow. 2020. Revisiting the timing of the northern lobe of the White River volcanic event in Eastern Alaska and Western Yukon. Radiocarbon 62: 168- 188.

Roussin R., J.E. Wilson, G. Utzig, and L.M. Lavkulich. 2015. Assessing the Potential for Pocket Agriculture in Mountainous Regions: A Case Study in West Kootenay, British Columbia, Canada. J. Agri. Food Syst. and Com. Develop. 6(1): 175-188.

http://dx.doi.org/10.5304/jafscd.2015.061.016

Scudder, G.G.E. 1997. Environment of the Yukon. In Insects of the Yukon. H.V. Danks, and J.A. Downes (Eds.). Biological Survey of Canada. Monograph Series No.2, Entomological Society of Canada, Ottawa, ON. pp. 13-57.

Smith C.A.S., K.T. Webb, E. Kenney, A. Anderson, and D. Kroetsch. 2011. Brunisolic soils of Canada: Genesis, distribution, and classification. Can. J. Soil Sci. 91: 695-717. doi:10.4141/CJSS10058

Soil Classification Working Group. 1998. The Canadian System of Soil Classification, 3rd ed. Agriculture and Agri-Food Canada Publication 1646.

Souther, J.G. 1992. The late Cenozoic Mount Edziza volcanic complex. In Geology, Edziza Volcanic Complex, British Columbia. Geological Survey of Canada Memoir, 420: 1-320.

Valentine K.W.G., P.N. Sprout, T.E. Baker, and L.M. Lavkulich. Editors. 1978. The Soil Landscapes of British Columbia. Ministry of the Environment, Victoria, BC.

Wiken, E.B.D., I Marshall, K. Larson, and H. Hirvonen.1996. A perspective on Canada's ecosystems: an overview of the terrestrial and marine ecozones. Canadian Council on Ecological Areas. Occasional Paper, no. 14, Ottawa.

Williams-Jones, G., R.W. Barendregt, J.K. Russell, Y. Le Moigne, R.J. Enkin, and R. Gallo. 2019. The age of the Tseax volcanic eruption, British Columbia, Canada. Can. J. Earth Sci. 57(10). https://doi.org/10.1139/cjes-2019-0240

About the Author

Dr. Leslie (Les) Lavkulich was raised on a farm in southern Alberta. He studied soil science at the University of Alberta and subsequently received a PhD at Cornell University.



Les Lavkulich (licensed under a CC-BY-NC-ND license)

Les joined the Department of Soil Science at the University of British Columbia where he became Professor and served as Department Head. His concern about the responsible management of resources, lead him to establish the interdisciplinary graduate program in Resource, Environment and Sustainability and more recently the Professional graduate program in Land and Water Systems.

I focused my energies on teaching, in a range of courses in Soil Science as well as Perspectives on Resources Management, Soil-Water Systems, and Communications in Resources Science. I have been awarded the Killam Teaching Award and the Margret Fulton Award for *Student Experience and Learning*. My research has focused on pedology, soil and environmental chemistry, mine/land reclamation, and soil-water systems. I served as a Canadian representative on the NATO Committee *Concerns of the Modern Society-Water Resources*, as well as a number of Canadian organizations, including the Association of Faculties of Agriculture in Canada, the Canadian Agricultural Research Council, and the Expert Committee on Soil Classification; and within British Columbia, the Chair of the BC Deputy Ministry Land Resources Steering Committee. I served as President of the Canadian Society of Soil Science and was give the honour of election as a Fellow of the Society.

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2.2: Soils of the Prairie Provinces

Dan Pennock

Learning Objectives

Upon completion of this chapter, students will be able to:

1. Describe the factors responsible for the distribution of soils in the Prairie Provinces
2. Identify the diagnostic soil horizons associated with soil orders and great groups (according to the Canadian System of Soil Classification) of the Prairie Provinces
3. Relate the major land uses of the region with the properties of the soils

INTRODUCTION

The soils of the three Prairie Provinces have experienced the greatest conversion to agricultural land uses of any region of Canada. In most of the southern portion of this region more than 80% of the land is agricultural (Prairie Farm Rehabilitation Administration, 2000). Throughout this region the specific type of agriculture is closely linked to the capability of the soil, and hence the pattern of agricultural land use is largely determined by the pattern of the soils themselves. Although we typically refer to Alberta, Saskatchewan and Manitoba as the Prairie Provinces, it is important recognize that the true prairies generally occupy only the southern portion of these provinces.

FACTORS CONTROLLING THE DISTRIBUTION OF SOILS IN THE PRAIRIE PROVINCES

The regional pattern of soil distribution (Figure 10.1) is determined by the bedrock geology and the effect of the Western Cordillera on the regional climate. These two factors largely determine the patterns of soil and vegetation in the region.

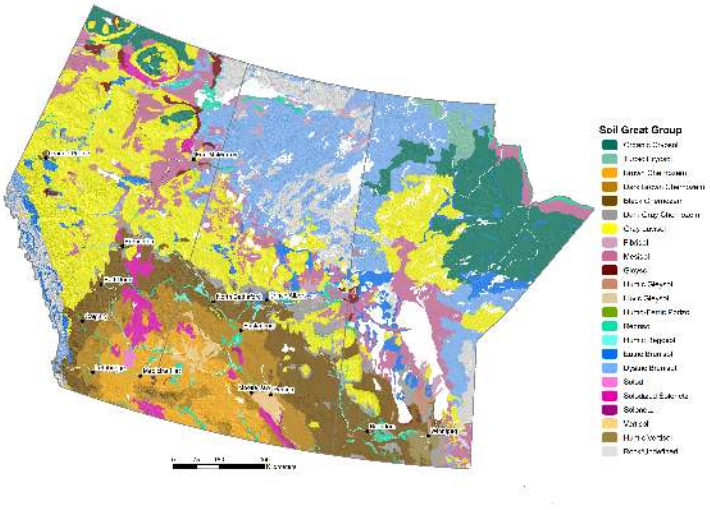


Figure 10.1. Map of soil great groups (according to the Canadian System of Soil Classification) for the Prairie Provinces. Map is a reproduction of an official work published by the Government of Canada and is based on the 1:1,000,000 Soil Landscapes of Canada map. Map courtesy of Darrel Cerkowniak, Agriculture and Agri-Food Canada. © Darrel Cerkowniak, Agriculture and Agri-Food Canada, is licensed under a CC BY (Attribution) license.

The Prairie Provinces are underlain by approximately ⅓ igneous and metamorphic rocks of the Canadian Shield and ⅔ sedimentary rocks (Figure 10.2). The Canadian Shield underlies the eastern and north-eastern portion of the region, and sedimentary rocks the remainder. As discussed in Chapters 2 and 8, the glacial sediments that form the parent materials of soils closely reflect the nature of the bedrock that underlies them; igneous rocks yield acidic, sandy and gravelly parent materials and sedimentary rocks produce neutral, alkaline, loamy parent materials. The parent materials produced by sedimentary rocks are a mixture of gravel and sand, silt, and clay.

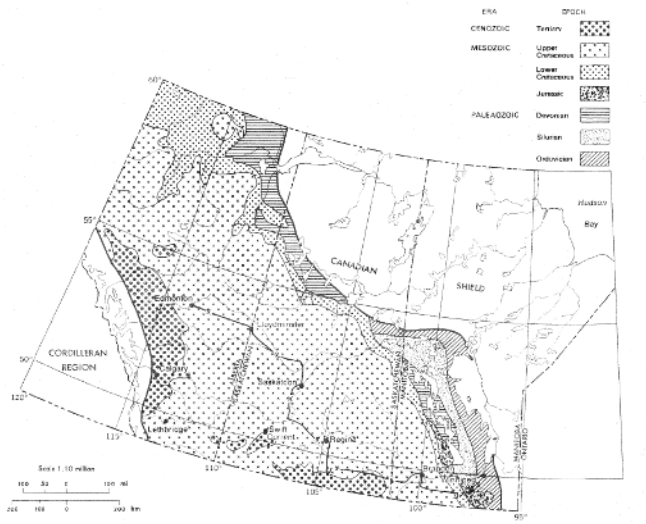


Figure 10.2. Map of bedrock geology of the Prairie Provinces. The rocks of the Canadian Shield are igneous and metamorphic. The sedimentary rocks are arranged from the oldest (Ordovician) to youngest (Tertiary). Reproduced from Michalyna et al. (1978) and licensed under the Open Government Licence – Canada.

Research on tills in this region has shown that rock fragments greater than 1 mm in diameter are derived from the Precambrian rocks of the Canadian Shield to the NE of the sedimentary rock area, whereas the sand, silt, and clay fractions normally reflect nearby geological formations (St. Arnaud, 1976). The glacial tills also contain carbonates and soluble magnesium and sodium salts. The carbonate content of the tills is highest (>40% calcium carbonate equivalent) in the Interlake area of Manitoba where surface exposure of Paleozoic limestones occurs. The carbonate content of tills decreases across the region to about 6 to 10% in much of southern Alberta and 1 to 6% in central Alberta (St. Arnaud, 1976). The carbonate content rises to 10 to 16% in the Cordilleran regions of Alberta in the region of deposition of glacial sediments from the Cordillera itself.

The northern boundary of the continuous Chernozemic region is found north of Edmonton in Alberta and then trends to the south-east until it ends just north of Winnipeg (Figure 10.1). This NW-SE trending boundary reflects in part the influence of the Western Cordillera on the passage of moist, temperate air masses from coastal British Columbia into the western interior (discussed in Pennock et al., 2011). The air masses are deflected to the north by the high mountain ranges of southern Alberta and pass through the Cordillera via the lower elevations of the Peace River area of British Columbia and Alberta around Grand Prairie. The air masses then trend to the SE through the rest of the Prairie region. The more temperate climates of the Peace River region, which occur because of this movement of air masses, allow extensive agricultural development throughout the region and even the development of a pocket of Chernozemic soils around Grand Prairie (Figure 10.1).

MAJOR SOILS OF THE PRAIRIE PROVINCES

At the broadest scale, the sedimentary rock region has multiple zones of soils that radiate out from SW Saskatchewan and SE Alberta (Figure 10.1). The dominant soil order in the southern part of this region is the Chernozemic order, which formed in the grasslands that were found in this area prior to European settlement. Throughout most of the region, the Chernozemic soils are bounded by soils of the Luvisolic order, which are associated with the mixed deciduous-coniferous forest characteristic of the northern part of this region.

Brunisolic soils are the dominant soil order on the Canadian Shield and consists of acidic forest soils develop on the sand and gravelly parent materials characteristic of this area. In north-central Manitoba a large area of glacio-lacustrine silts and clays were deposited (Figure 10.3) and Luvisolic soils have formed in these parent materials.

Closer to Hudson Bay, the rebound of the land surface after glaciation has created a poorly drained region of near-level beach ridges and former shorelines with saturated soils. The cold temperatures and permanent water saturation has led to the development of thick layers of peat and formation of permanently frozen layers within the peats. These soils are classified as Organic Cryosols in the CSSC (Figure 10.1).

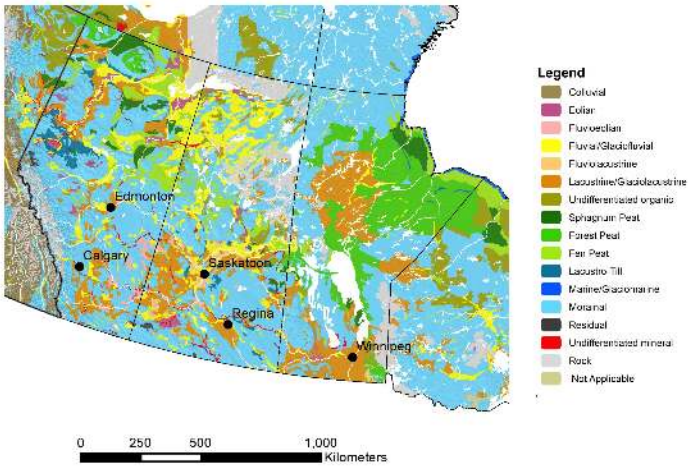


Figure 10.3. Major soil parent materials found in the Prairie Provinces. Map is a reproduction of an official work published by the Government of Canada and is based on the 1:1,000,000 Soil Landscapes of Canada map. © Darrel Cerkowniak, Agriculture and Agri-Food Canada, is licensed under a CC BY (Attribution) license.

The complex pattern of soils in northern and north-eastern Alberta (Figure 10.1) departs from the overall zonation of soils. In this area there are several major uplands (such as the Caribou Mountains) that are elevated plateaus occupied by Gray Luvisols and Organic Cryosols. The uplands are separated by wide valleys occupied by the Athabasca, Hay and Peace Rivers. The relatively level valleys have a complex mixture of sandy Brunisolic, loamy Luvisolic and Fibrisol Organic soils. The extensive area of oil sand extraction near Fort McMurray devastates these soil landscapes, and reclamation of the landscapes (and remediation of the spoil from oil sands operations) is a major activity for soil scientists in this region.

Finally, the Cordilleran region at the Alberta-British Columbia border has a complex mixture of forested soils and exposed rock faces. The regional map (Figure 10.1) shows areas of exposed rock and both Dystric and Eutric Brunisolic soils, but areas of Gray Luvisolic soils also are found as well as Cryosolic alpine tundra soils at higher elevations.

GRASSLAND AND PARKLAND SOILS

The zonal pattern of great groups of the Chernozemic order reflects the different types of grasslands that once occupied the area (and the regional climates responsible for the zonation of grasses) (Fuller, 2010; Pennock et al., 2011). These grasses were responsible for organic carbon inputs into the soil, and the colour of the diagnostic Ah horizons reflects differences in the amount of organic carbon additions from the various grass communities. The great groups of the Chernozemic order are based on the colour differences of the A horizons.

The major climate attribute that varies across the region is the mean annual water deficit (Pennock et al., 2011, Table 10.2). The driest portions of the region have annual water deficits of approximately 200 mm and the native grasses of this region were low grasses with limited above- and below-ground biomass. The lower organic carbon inputs lead to the development of soils of the Brown Chernozemic great group (Figure 10.4). The water deficit decreases west and north of this dry region, and greater organic carbon inputs lead to formation of Dark Brown Chernozemic soils and Black Chernozemic soils. The highest organic carbon levels occur in southern Manitoba where the annual water deficit is only approximately 70 mm.

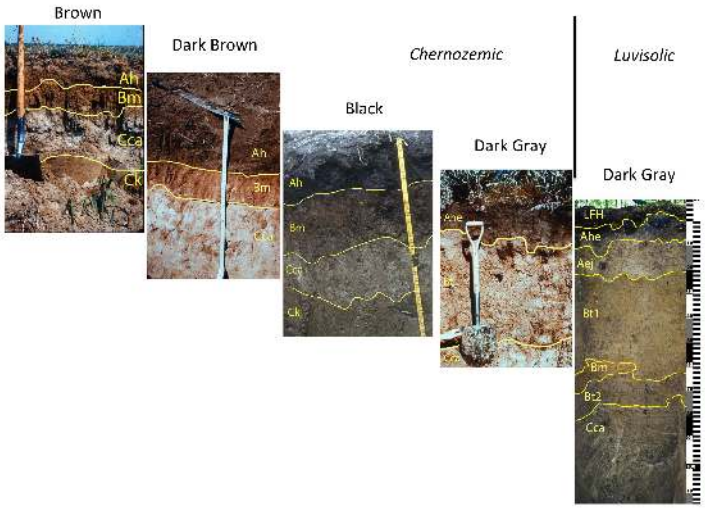


Figure 10.4. Examples of major soils of the Chernozemic and Luvisolic great groups found in the Prairie Provinces. Soils of the Chernozemic order have organically enriched A horizons (Ah) overlying decalcified and slightly reddened Bm horizons. The upper C horizon has concentrated secondary calcium carbonate (Cca) overlying largely unaltered parent material (Ck) with lower calcium carbonate contents. The Dark Gray Chernozem is transitional to the Gray Luvisolic soils and has an A horizon with both organic accumulation and bleaching (Ahe horizon). The forested Gray Luvisols have a leaf litter layer (LFH) on top of the mineral soil horizons. The uppermost bleached Ae horizon overlies a B horizon with higher clay levels than the Ae (Bt horizon). © "Brown, Dark Brown, Dark Gray Chernozem: Saskatchewan Center of Soil Research; Black Chernozem: Roly St. Arnaud (Saskatchewan Center of Soil Research); Dark Gray Luvisol: Kent Watson (Saskatchewan Center of Soil Research)" is licensed under a CC BY (Attribution) license.

Within these broad soil zones there are many pockets of other soil classes in the former grassland areas (Figure 10.1). During melting of the ice sheets glacio-lacustrine silts and clays were deposited in vast glacial lakes throughout the region, with large areas of deposition in west-central Saskatchewan, southern Manitoba, and north-central Manitoba (Figure 10.3). In the former, grassland region the high-clay soils of the Vertisolic order form in these glacio-lacustrine deposits (Anderson, 2010a; Brierly et al., 2011). The great groups of the Vertisolic order (Figure 10.5) reflect the differences in organic carbon inputs discussed for the Chernozems above: Vertisols in southern Saskatchewan are classified into the Vertisol great group, whereas the higher organic inputs that occur in the region south of Winnipeg result in Humic Vertisols (Figure 10.1).

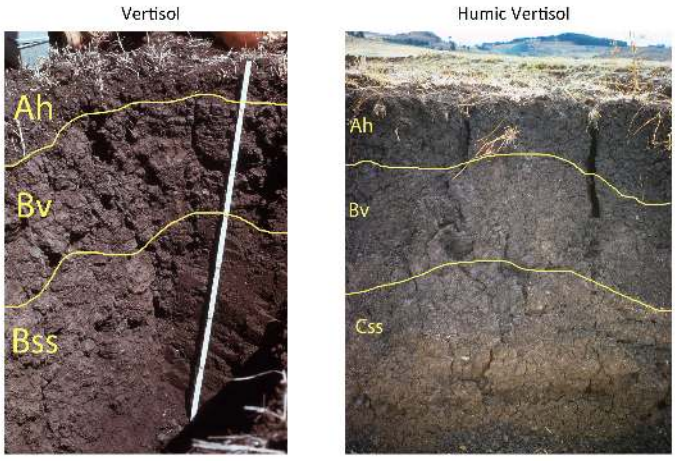


Figure 10.5. Great groups of the Vertisolic order. Mixing of soil by shrinking and swelling of clays prevents visually distinct horizons from forming. The mixed horizons are assigned a 'v' suffix, and in lower horizons polished surfaces (slickensides, assigned a 'ss' suffix) form on the face of fractures in the soil. The Humic great groups have higher organic matter inputs and a more distinct Ah horizon that the Vertisol great group. © "Vertisol: Roly St. Arnaud (Saskatchewan Center of Soil Research); Humic Vertsol: Darwin Anderson (Saskatchewan Center of Soil Research)" is licensed under a CC BY (Attribution) license.

The distribution of Solonetzic soils is largely controlled by the presence of sodium-rich bedrock within a few meters of the soil surface rather than a particular type of glacial sediment (Anderson, 2010b; Miller and Brierley, 2011). In both Alberta and

Saskatchewan, the marine shales of the Cretaceous-aged Bearpaw formation are the main bedrock type associated with Solonetzic soils. Movement of soil water and groundwater circulates sodium through the soil profile, which can lead to vertical movement of clay from the surficial Ah horizon soil to the B horizon and formation of a bleached, eluvial Ae horizon overlying the B horizon. The B overlies a saline and carbonate-rich C horizon (Csk).

The great groups of the Solonetzic order reflect the degree of development of the Ae horizon and the AB horizon, which forms through weathering of the upper B horizon. In the Alkaline Solonetz great group the Ae is very thin or absent, and the B horizon is high in sodium (Bn) but often without higher clay levels than the Ah (Figure 10.6). The Solodized Solonetz great group has an Ae greater or equal to 2 cm thick and an intact Bnt or Bn horizon with well-developed columnar structure. In the Solod great group the top of the Bnt has disintegrated through weathering and this weathered layer forms a transitional AB horizon between the Ae and the Bnt horizons. The Solodized Solonetz great group is the most common in the landscape (Figure 10.1).

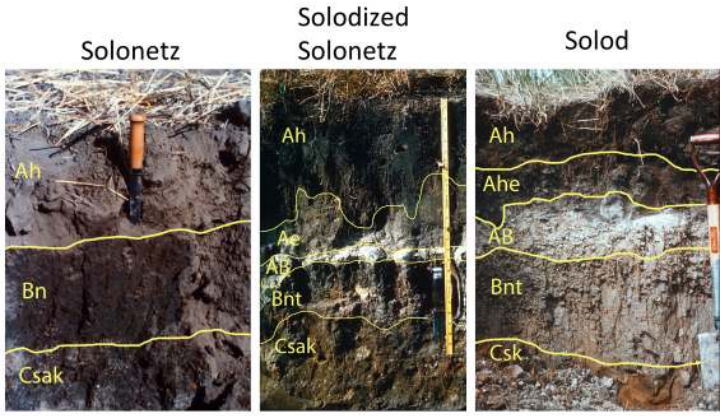


Figure 10.6. Great groups of the Solonetzic order. The great groups of the Solonetzic order reflect the progressive disintegration of the upper part of the Bnt horizon (a horizon enriched in both sodium and clay) to form a transitional AB horizon. The AB horizon begins to form in the Solodized Solonetz great group and reaches maximum thickness in the Solod great group. © "Solonetz: Darwin Anderson (Saskatchewan Center of Soil Research); Solodized Solonetz: Roly St. Arnaud (Saskatchewan Center of Soil Research); Solod: Saskatchewan Center of Soil Research" is licensed under a CC BY (Attribution) license.

Small areas dominated by Regosolic soils (VandenBygaart, 2011) are also found throughout the region. Regosolic soils are associated with unstable land surfaces (such as river floodplain or sand dunes) where soil development is limited to development of an Ah horizon either directly overlying the C horizon or with a thin (<5 cm) Bm between the A and C horizons (Figure 10.7). The Regosol-dominated areas in the Prairie Provinces are associated with sandy fluvial or glaciofluvial deposits along river valleys or areas of active or recently stabilized sand dunes such as the Great Sand Hills of Saskatchewan. As well, the knolls in many cultivated landscapes in the Prairie Provinces have experienced considerable soil loss due to tillage erosion, and thin Ap (p for ploughed) horizons directly overlay Cca horizons on these hill crests.



Figure 10.7. Regosolic soils. Regosolic soils occur in very young sediments (such as on active floodplains or in active sand dune areas) or due to erosion of surface soils in cultivated fields. In all cases, B horizon formation is either absent or less than 5 cm thick. © "River sediments: Darwin Anderson (Saskatchewan Center of Soil Research); Truncated soil: Dan Pennock (Saskatchewan Center of Soil Research); Dune Sand: Roly St. Arnaud (Saskatchewan Center of Soil Research)" is licensed under a CC BY (Attribution) license.

Many of the glacial landforms in the area are associated with melting in-place of stagnant ice masses, and the resulting land surface is a very complex assemblage of small knolls and depressions. This land surface is labelled as hummocky or knob and kettle on soil maps. Wetlands on these surfaces are often occupied by Gleysolic soils, whose Bg and Cg horizons have dull gray colours and, in some cases, patches of reddish oxidized iron within the dominantly gray material (Figure 10.8) (Bedard-Haughn, 2010, 2011). Texture-contrast horizons are also common in these soils, with a Btg horizons underlying a mottled Aeg horizon. Soils with a Btg horizon are classified into the Luvic Gleysol great group.

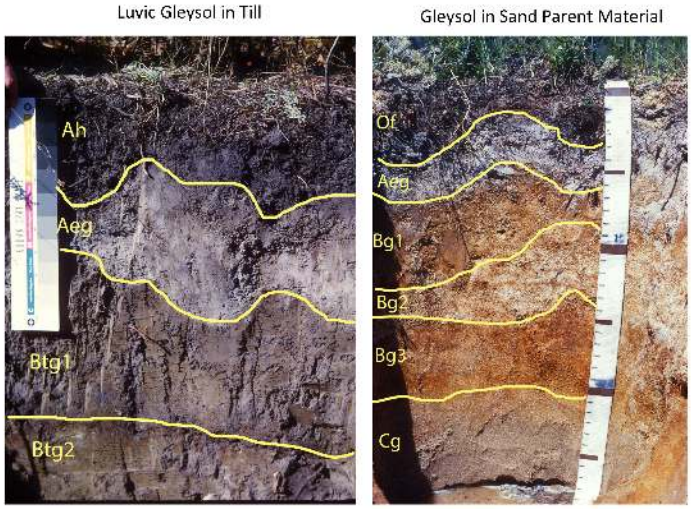


Figure 10.8. Gleysolic soils. The Luvic Gleysol great group soil developed in clay loam till has dull, grayish colours throughout and small reddish mottles in the Aeg horizon. The Gleysol great groups soil developed in sandy glaciofluvial parent materials has much brighter coloured mottles throughout with maximum expression of oxidized iron in the Bg3 horizon. © Both slides: Darwin Anderson (Saskatchewan Center of Soil Research) is licensed under a CC BY (Attribution) license.

A final soil class that is largely unmapped are saline soils. As discussed in Chapter 8, the Canadian System of Soil Classification (CSSC) does a poor job of identifying saline soils. This is in part due to the very dynamic nature of soil salinity, which fluctuates considerably due changes to soil water and groundwater flow patterns. Salinity can only be recognized in profile descriptions through the use of the suffixes s and sa in horizon labels (e.g., Csa, Cs, Bsa, Ahsa, etc.). Agriculture and Agri-Food Canada has mapped soil salinization risk for several decades as part of its Agri-Environmental Indicators program and the main areas that are vulnerable to salinization (i.e., areas mapped as moderate, high, and very high risk) are widespread throughout the Prairie region. The largest areas are the region stretching NE from Calgary to almost Saskatoon, SW Alberta, in SE Saskatchewan near Estevan,

and in areas of Manitoba south of Brandon and of Winnipeg. The AAFC maps show that the risk of salinization has decreased since the 1980s due to adoption of no-till cultivation practices and the near-elimination of tillage summer fallow throughout this region.

FOREST AND WETLAND SOILS

At the northern boundary of the Chernozemic soils the grasslands become increasingly interspersed with clumps of trees and shrubs and ultimately the forest forms an almost continuous cover over the soil. The transitional grassland-forest zone is referred to as the Parkland zone, and the continuous forest cover as the Mixedwood forest, reflecting the forest cover of deciduous and coniferous species.

In the Parkland, the inputs of organic carbon from grasses decrease and leaf litter from trees builds up on the soil surface as a LFH layer. The colour of the Ah horizon grows lighter, and bleaching of the horizon also begins to occur, forming a surface A horizon with both humus inputs (given an 'h' suffix) and bleaching (an 'e' suffix) – the Ahe horizon. These soils are classified into the Dark Gray Chernozemic great group.

North of the forest-grassland transition zone, organic inputs form a discrete layer laying on top of the mineral soil (the LFH layer) and the soils have a well-developed bleached layer (Ae horizon) overlying a B horizon with higher clay levels (Bt horizon). The Bt horizon is underlain by a C horizon with secondary (Cca) or primary (Ck) carbonates. These soils are classified into the Gray Luvisolic great group (Figure 10.9), which dominates the forested landscapes underlain by neutral-alkaline tills (Pettapiece et al., 2010; Lavkulich and Arocena, 2011). Gray Luvisols also occur in the large area of glacio-lacustrine parent materials in north-central Manitoba.

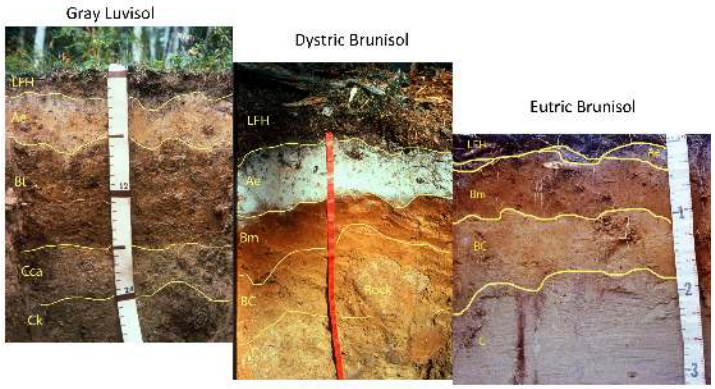


Figure 10.9. Luvisolic and Brunisolic soils. Soils of the Gray Luvisol great group have well developed texture contrast (Ae and Bt) horizons overlying the C horizons with primary (Ck) and secondary (Cca) calcium carbonate. Both of the great groups of the Brunisolic order develop primarily in sandy parent materials and have the diagnostic Bm horizon. The great groups differ on the acidity of the parent materials: Dystric Brunisols develop in acidic parent materials, and Eutric Brunsiols develop in neutral pH parent materials. © "Luvisol: Darwin Anderson (Saskatchewan Center of Soil Research); D. Brunisol and Eut. Brunisol: Saskatchewan Center of Soil Research" is licensed under a CC BY (Attribution) license.

At the boundary between the sedimentary rocks of the Interior Plains and the Canadian Shield a complex assemblage of forest and wetland soils occurs (Figure 10.1). The forested soils form in parent materials that include sandy sediments from the Canadian Shield and pulverized sedimentary rocks, which raises the pH of the parent materials above 5.5. These soils are classified into the Eutric Brunisol great group (Figure 10.9) (Smith et al., 2011). In low lying areas, water collects and forms large fens and bogs. In these wetlands peat builds up due to the permanent water saturation and soils of the Mesisol great group of the Organic order occur (Bedard-Haughn, 2010). These soils have intermediate levels of decomposition of organic materials (Figure 10.10).

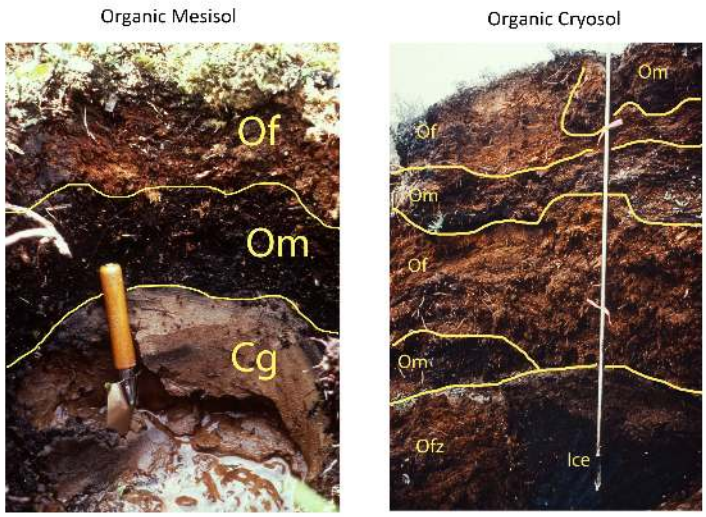


Figure 10.10. Organic and Cryosolic soils. Both soils have a thick buildup of peat in either largely undecomposed (Of) or moderately decomposed (Om) layers. The Organic Cryosol has permafrost within 1 m of the surface (note – total length of white tape is 1.2 m). © "Mesisol: Ken Van Rees (Saskatchewan Center of Soil Research); Cryosol: Saskatchewan Center of Soil Research" is licensed under a CC BY (Attribution) license.

The dominant soil of the Canadian Shield itself are Dystric Brunisols (Figure 10.9) that develop in the acidic (pH < 5.5) sand and gravel parent materials characteristic of this region. Like the Eutric Brunisols, the Dystric Brunisols soils are very sandy soils and soil formation is limited to reddening of the B horizon and, in some cases, formation of a bleached Ae horizon. In moister forested regions of Canada, these conditions lead to formation of soils of the Podzolic order (Sanborn et al., 2011). There are also significant areas where exposed rock is the dominant surface with only scattered pockets of soils interspersed in the landscape (Figure 10.1).

Finally, the near-level landscapes of the Hudson Plains ecozone are mainly occupied by organic soils that have a layer of permafrost (i.e., permanently frozen soil) within the soil profile. These soils have thick organic layers but, because of the permafrost, are classified into the Cryosolic order (Figure 10.10) (Tarnocai and Bockheim, 2010).

SOILS AND LAND USE

Soil properties have a major impact on agricultural capability throughout the Prairie Provinces. The agricultural capability of soils throughout southern Canada was evaluated by the Canada Land Inventory (CLI) (see Chapter 8). The criteria used in the CLI system were based on soil and climate and remain generally applicable although cropping systems used in the region have changed greatly since the period of the CLI inventory (roughly 1965-1975). At that time the dominant crop was spring wheat and summerfallow (i.e., a year of fallow with tillage of the soil to suppress weed growth following a year of wheat) was the dominant tillage practice. Currently, a much wider range of crops (e.g., canola, pulse crops, flax) is grown, and the great majority of producers practice continuous cultivation with high amounts of crop residue left on the soil surface (i.e., conservation tillage or no-till).

In the CLI rankings, classes 1 through 4 are considered to be capable of sustained use for cultivated field crops. In class 1 land, there are no significant constraints to crop growth, whereas in class 3 and 4 areas the range of crops is restricted and special management measures are needed to conserve the soil. Classes 5 and 6 are only suitable for forage production and are most commonly used for grazing of livestock.

The major soil property that determines agricultural capability in the CLI system is soil texture due primarily to its effect on water availability. Secondary properties of relevance are unfavorable soil structure associated with Solonetzic soils and the presence of salinity in the soil profile. Soils that have experienced erosion of the organic-rich A horizons through time are also downgraded – organic matter has many beneficial properties including a high water-holding capacity and high nutrient-supplying power.

All agricultural regions of the Prairie Provinces experience a water deficit (as discussed above) and hence the ability of the soil to hold water during dry periods is critical. This property is termed the available water capacity (see Chapter 4). Generally, fine-textured soils (clay, heavy clay, silty clay, clay loam) retain the highest amounts of water even in drought conditions and have high

available water capacity. Medium-textured soils (silty clay loam, sandy clay loam, loam, silt loam, fine and very fine sandy loam) have lower ability to hold water during droughts but have slightly higher available water capacity. Medium-coarse sandy loam, loamy sand, and sand soils have both very low ability to hold water during droughts and low available water holding capacities and hence are the lowest rated for agriculture.

The second major factor that determines the CLI class is the climate. In the Prairie Provinces, both moisture and heat can be limiting. Agriculture and Agri-Food Canada developed agro-climatic regions throughout Canada based on three climate factors: Mean annual precipitation, average number of frost-free days, and degree days above 5.5°C (Table 10.1 and Figure 10.11). The latter is a cumulative measure of the period when the temperature is high enough for significant plant growth. The class limits will have shifted somewhat due to climate change but the general pattern would be unchanged.

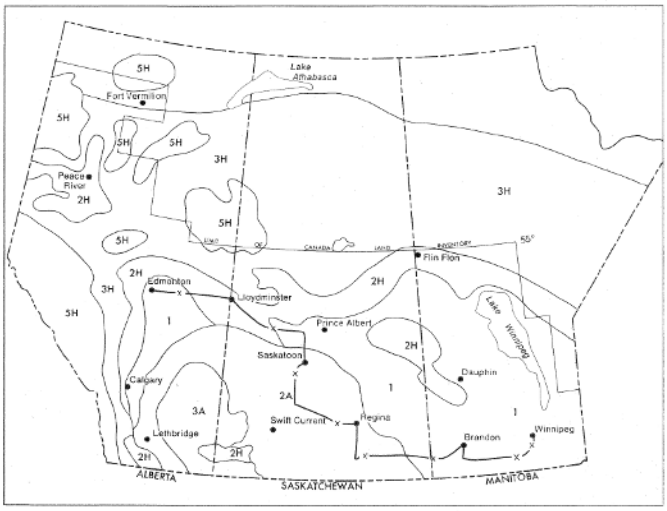


Figure 10.11. Agro-climatic regions of the agricultural areas of the Prairie Provinces. Reproduced from Michalyna et al. (1978) and licensed under the Open Government Licence – Canada.

Soil texture is directly inherited from the parent materials in the Prairie Provinces; hence, a clear link between agricultural capability and the soil classes described above. At the driest end of the spectrum are the Regosolic soils formed in sandy glacio-fluvial sediments or sand dune areas, which typically have CLI classes of 5 or 6. At the other end of the spectrum are Vertisol soils, which inherently have clay or heavy clay textures and typically have CLI class 1 or 2 ratings throughout the main agricultural region.

The capability ratings for Chernozemic soils depends primarily on texture and on climate zone. The general pattern is clear from Figure 10.12: fine-textured soils are the best agricultural soils in each climate region and coarse textured soils the worst. The range is greatest for medium-textured soils developed on glacial till, which are the most common in the region.

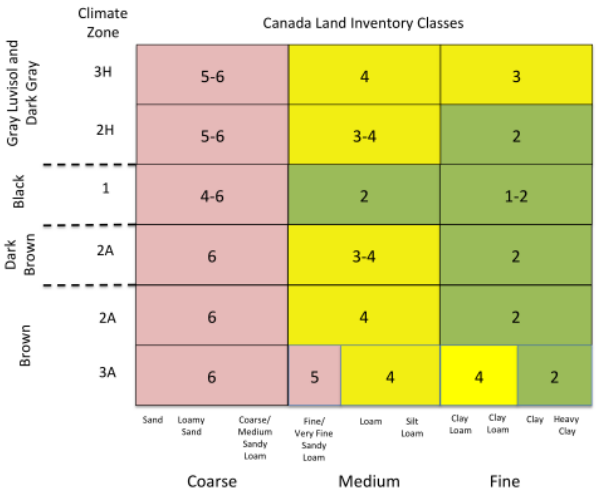


Figure 10.12. Matrix of Canada Land Inventory (CLI) classes associated with soil zones, soil textures, and agro-climatic zones in the Prairie Provinces. © Dan Pennock, Univ. of Saskatchewan is licensed under a CC BY (Attribution) license.

Gray Luvisolic soils have a number of limitations for annual crop production (Pettapiece et al., 2010). The Gray Luvisols occur in zones 2H and 3H (Figure 10.11) where lack of heat and frequent frosts limit growth. The Gray Luvisols lack an Ah horizon and typically have silty textured surface mineral horizons; both factors contribute to problems with seedbed preparation and crop growth generally. Compared to Chernozemic and Vertisols soils they are also deficient in nitrogen, phosphorus, and sulphur. The Gray Luvisols and related soils in the Aspen Parkland region are widely used for forage crop production and grazing – about two-thirds of western Canada's cattle and more than 80% of the forage crops are produced in this region (Thorpe and Anderson, 2010).

Secondary soil properties that strongly affect agricultural capability are factors such as the presence of high sodium soils or salinity, both of which are generally associated with class 5 or 6 soils. The Bn or Bnt horizon associated with the Solonetzic soils are difficult for plant roots to penetrate and hence only the horizons above these horizons can be fully exploited by roots for nutrients and water. As well, the Ah and Ae horizons above the dense B horizons have often experienced erosion, and the variable depth of surface soil over the B horizon often gives crops on Solonetzic surfaces a wavy appearance due to lower growth on thin soils and higher growth on thicker soils. Salinity has a severe effect on crop growth and on highly saline soils growth of any plant except saline-tolerant species is impossible.

An important landscape factor that affects the rating of medium-textured soils is topography – generally the rougher the topography, the lower the rating. Many medium-textured glacial till landscapes have many knolls and wetlands associated with them, and these cause difficulties in farm operations and uneven plant growth. Rough topography generally causes a given area to drop by at least one class in the CLI rating.

Table 10.1. Agro-climatic regions. (Adapted from Michalyna et al. 1978)

Climate subregion	Mean annual precipitation (cm)	Average frost-free period (days)	Degree-days above 5.5ºC	General comments
1	40-45	>90	1225-1670	Climate is suitable to permit growing of all dryland crops that are typical to the prairie region
2A	30-40	>90	1330-1670	Precipitation limits growth in approximately 50% of years
3A	27-35	90-120	1330-1670	Precipitation is a severely limiting factor for crop growth
2H	40-50	75-90	1060-1225	Precipitation is adequate but some frost damage to crops in approximately 30% of years
3H	27-35	<75	1000-1112	Precipitation adequate by high incidence of frost damage to crops
5H	27-35	<60	<950	Average frost-free period too short for annual crop production

Soil properties also play an important role in determining land capability for forest harvest operations. Forest harvest has increased in Alberta since early 2000s (from 82 thousand hectares (t ha^{-1}) in 2005 to 92 t ha^{-1} in 2016). However, it has declined in Saskatchewan (42 t ha^{-1} in 2005 to 22 t ha^{-1} in 2016) and Manitoba (14 t ha^{-1} in 2005 to 8 t ha^{-1} in 2016) (all statistics from Natural Resources Canada). In total, this represents about 13% (2005) to 15% (2016) of total forest area harvested in Canada.

The Canada Land Inventory also rated soils for forest capability in the portion of the forest where commercial forestry existed. The highest class of land in the Prairie Provinces was only class 3 due to climate constraints for tree growth throughout the region. Generally, forests on Gray Luvisol soils are rated as class 4 (Alberta and Manitoba) or Class 5 (Saskatchewan) with small areas rated one class higher in all provinces. Forests on both Eutric and Dystric Brunisolic soils are generally rated as class 6 due to significant growth constraints. Detailed ecosite guides have been prepared for each province (see Suggestions for Further Reading) that rate soils on soil texture and soil drainage – generally well-drained, fine- and medium-textured soils are the most highly rated in the forest landscape.

Wetland soils in the Prairie Provinces are key contributors to two transnational and global issues: the protection of waterfowl in North America and the removal of carbon from the atmosphere by increasing soil organic carbon levels (carbon sequestration). In the former grassland areas of the Prairies, the complex hummocky land surfaces have countless wetlands that are occupied by Gleysolic soils or, in some cases, saline soils. These wetlands are the seasonal home for vast numbers of migratory waterfowl, and protection of Prairie wetlands is essential for preservation of waterfowl habitat. The North American Waterfowl Management Plan (NAWMP) is a major agreement among Mexico, the USA and Canada and protection of Prairie wetlands is a key aspect of this transnational plan.

Gleysolic wetlands and especially Organic soils are repositories for large amounts of carbon. Drainage of these wetlands for agriculture causes mineralization of this carbon and its release (as carbon dioxide) to the atmosphere. The restoration of wetlands and of the wetland vegetation that surrounds them has the potential to draw carbon from the atmosphere and hence lessen the growth in carbon levels caused by human activity. The 2019 report by the Intergovernmental Panel on Climate Change on climate

change and land has highlighted the role that carbon sequestration by soils could play in greenhouse gas mitigation (Section 4.9.4 IPCC, 2019).

Finally, soils of all types have been extensively affected by resource extraction activities throughout western Canada. In Alberta alone, there are 422,000 km of oil and gas pipelines, 343,000 well sites (176,000 active wells, 90,000 inactive wells, and 77,000 abandoned wells) and 895 km^2 have been disturbed by surface mining for oil sand production (as of 2013); oil sand tailing ponds and associated structures totaled 220 km^2. (Sources: Agdex 878-4; Alberta Energy–Oil & Gas Liabilities Management; and Alberta Energy–Oil Sands, Coal and Mineral Operations). In all cases, legislation requires that the sites be reclaimed by replacing soil and re-establishing vegetation on the disturbed site so it can support activities similar to those it could have supported before it was disturbed. In situations where contamination has occurred the soils must be remediated (or cleaned up) to meet specific soil and groundwater standards. All of these steps require knowledge of soil properties and soil distribution, ensuring a continuing role for soil scientists form decades to come.

ADDITIONAL RESOURCES

Soil surveys in the Prairie Provinces can be queried on-line at the following sites:

Saskatchewan: sksis https://sksis.usask.ca/#/map
Alberta: AGRASID https://soil.agric.gov.ab.ca/agrasidviewer/
Manitoba: https://soils.gis4ag.com/manitoba-soil-series-interactive-online-map/

Soil information (surveys, CLI maps, historic publications, Canadian System of Soil Classification) is available on-line through the Canadian Soil Information Service (CanSIS) of Agriculture and Agri-Food Canada. An index to CANSIS can be found at: http://sis.agr.gc.ca/cansis/nsdb/index.html

Provincial-scale Canada Land Inventory maps are available at:
http://sis.agr.gc.ca/cansis/publications/maps/cli/1m/agr/index.html

Field Guides for Forest Ecosite Classification
Saskatchewan https://publications.saskatchewan.ca/#/products/31663
Manitoba https://cfs.nrcan.gc.ca/publications?id=12179
Alberta https://cfs.nrcan.gc.ca/publications?id=25327

STUDY QUESTIONS

1. Explain why the carbonate content of Prairie soils generally decreases from a high in central Manitoba to a low in central Alberta.
2. Why do Chernozemic soils occur as far north as Grand Prairie in Alberta (see Chapter 8, Figure 8.1)?
3. Looking at Figure 10.3 – on what parent materials south of the Canadian Shield would you expect to find soils of the Vertisolic order?
4. Based on the material presented in Chapter 10, rank the following four soils in terms of their ability to support crop growth from highest to lowest: Black Chernozem, Gray Luvisol, Humic Vertisol and Solodized Solonetz. Briefly explain your ranking.
5. What are the four main soil great groups found on the Canadian Shield in the Prairie Provinces (see Figure 10.1)? Which of the four great groups is “out of place” on the Shield and why is it found there?
6. What secondary soil properties reduce the CLI class for agriculture soils to class 5 or 6?
7. Wetlands in the agricultural zone are often drained for agricultural purposes. What soil order is most common in these wetlands and what are two important ecosystem services that they provide?

REFERENCES

Anderson, D.W. 2010a. Vertisolic soils of the Prairie Region. Prairie Soils and Crops Journal 3: 29-36. Available at: https://prairiesoilsandcrops.ca/volume3.php

Anderson, D. 2010b. Solonetzic soils of the Prairie region. 2010. Prairie Soils and Crops Journal 3: 65-72. Available at: https://prairiesoilsandcrops.ca/volume3.php

Anderson, D.W. and Cerkowniak, D. 2010. Soil formation in the Canadian Prairie region. Prairie Soils and Crops Journal 3: 57-64. Available at: https://prairiesoilsandcrops.ca/volume3.php

Bedard-Haughn, A. 2010. Prairie wetland soils; Gleysolic and Organic. Prairie Soils and Crops Journal 3: 9-15. Available at: https://prairiesoilsandcrops.ca/volume3.php.

Bedard-Haughn, A. 2011. Gleysolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 763-779.

Brierly, J.A., Stonehouse, H.B., and Mermut, A.R., 2011. Vertisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 903-916.

Fox, C.A. and Tarnocai, C. 2011. Organic soils of Canada: Part 2. Upland Organic soils. Can. J. Soil Sci. 91: 823-842.

Fuller, L. 2010. Chernozemic soils of the Prairie region of Canada. Prairie Soils and Crops Journal 3: 37-45. Available at: https://prairiesoilsandcrops.ca/volume3.php

Intergovernmental Panel on Climate Change. 2019. Chapter 4: Land Degradation. IPCC Special Report on Climate Change and Land. Available at: https://www.ipcc.ch/report/srccl/

Kroetsch, D.J., Geng, X., Chang, S.X., and Saurette, D.D. 2011. Organic soils of Canada: Part 1. Wetland Organic soils. Can. J. Soil Sci. 91: 807-822.

Lavkulich, L.M. and Arocena, J.M. 2011. Luvisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 781-806.

Macyk, T.M., Ellis, J.G., Michalyna, W., and Smith. R.E. 1978. Guidebook for a tour across the southern portion of the Interior Plains of Western Canada Tour 11. 11th Congress International Society of Soil Science. Edmonton Alberta June 1978. Available at: http://sis.agr.gc.ca/cansis/publications/manuals/ca_1978_ISSS_Tour/index.html.

Michalyna, W., Smith, R.E., Ellis, J.G., and Macyk, T.M. 1978. Guidebook for a tour across the southern portion of the interior plains of western Canada from Winnipeg manitoba to Edmonton Alberta. Tour 2. 11th Congress International Society of Soil Science. Edmonton Alberta June 1978. Available at: http://sis.agr.gc.ca/cansis/publications/manuals/ca_1978_ISSS_Tour/index.html.

Miller, J.J. and Brierly, J.A. 2011. Solonetzic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 889-902.

Pennock, D.J., Bedard-Haughn, A., and Viaud, V. 2011. Chernozemic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 719-747.

Pettapiece, W., Robertson, J., and Anderson, D. 2010. Cultivated gray Luvisol soils of the Prairie region. Prairie Soils and Crops Journal 3: 73-83. Available at: https://prairiesoilsandcrops.ca/volume3.php.

Prairie Farm Rehabilitation Administration. 2000. Prairie Agricultural Landscapes. A Land Resource Review . PFRA, Regina. pp. 180.

St. Arnaud, R. J. 1976. Pedological aspects of glacial tills. Pages 133-155 in R.F. Legget, ed. Glacial Till. Royal Society of Canada, Ottawa, ON.

Sanborn, P., Lamontagne, L., and Hendershot, P. 2011. Podzolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 843-880.

Smith, C.A.S., Webb, K.T., Kenney, E., Anderson, A., and Kroetsch, D. 2011. Brunisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 695-717.

Soil Classification Working Group. 1998. The Canadian System of Soil Classification. 3rd Edition. Agriculture and Agri-Food Canada. Publ. 1646 (Revised). 187 pp. Available at: http://sis.agr.gc.ca/cansis/publications/manuals/1998-cssc-ed3/cssc3_manual.pdf

Tarnocai, C. and Bockheim, J.G. 2011. Cryosolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 749-762.

Thorpe, J. and Anderson, D. 2010. Soils of the rangelands in the Prairie region. Prairie Soils and Crops Journal 3: 46-56. Available at: https://prairiesoilsandcrops.ca/volume3.php

VandenBygaart, A.J. 2011. Regosolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 881-887.

SOIL SURVEYS, LAND INVENTORY MAPS & FIELD GUIDES

Soil surveys in the Prairie Provinces can be accessed on-line at the following sites:

Saskatchewan: sksis https://sksis.usask.ca/#/map

Alberta: AGRASID https://soil.agric.gov.ab.ca/agrasidviewer/

Manitoba: https://soils.gis4ag.com/manitoba-soil-series-interactive-online-map/

Canadian Soil Information Service (CanSIS) of Agriculture and Agri-Food Canada:

Soil information (surveys, CLI maps, historic publications, Canadian System of Soil Classification) is available on-line through CanSIS. An index to CANSIS can be found at: http://sis.agr.gc.ca/cansis/nsdb/index.html

Provincial-scale Canada Land Inventory maps are available at:

http://sis.agr.gc.ca/cansis/publications/maps/cli/1m/agr/index.html

Field Guides for Forest Ecosite Classification:

Saskatchewan https://publications.saskatchewan.ca/#/products/31663

Manitoba https://cfs.nrcan.gc.ca/publications?id=12179

Alberta https://cfs.nrcan.gc.ca/publications?id=25327

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Dan Pennock is a Professor Emeritus (or, more simply, a retired professor) of the Department of Soil Science at the University of Saskatchewan. His research career focused on how soils are shaped by their place in the landscape and by the processes (like erosion or water movement) that occur in landscapes. He taught courses in many areas of soil science and geography, and his love of teaching was recognized by the university with a Master Teacher award in 2006. He was made a Fellow of the Canadian Society of Soil Science in 2010. Since his retirement he has collaborated on a number of projects on sustainable management of soils with the Global Soil Partnership of the Food and Agriculture Organization of the United Nations.

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2.3: Soils of Ontario

Daniel Saurette, Jim Warren, and Richard Heck

Learning Objectives

On completion of this chapter, students will be able to:

1. Identify the ecozones of Ontario
2. Understand the spatial distribution of the major soil forming factors in Ontario
3. List the Soil Orders and Great Groups present in Ontario
4. Describe the relationships between soil forming factors and soil great group distribution in Ontario
5. Describe land use in Ontario by ecozone
6. Understand the relationships between soil distribution and land use in Ontario

INTRODUCTION

All of Ontario was once covered by the Laurentide Ice sheet of the Wisconsinan glaciation, with ice thicknesses exceeding one kilometre; this glacial event is responsible for the deposition of almost all the parent materials from which Ontario soils have developed. There are some exceptions of course, such as more recent alluvial or fluvial deposits in modern flood plains, and lacustrine materials from lakes, eolian materials such as active sand dunes and the continued accumulation of organic materials. The ice advanced from the north, forming many ice lobes, scraping soil from much of the pre-glacial landscape and depositing it south to the central and northern United States. Upon retreat, the continental glacier left behind the varied glacial landscapes we see today. The ice sheet retreated towards the northeast over millennia, slowly exposing parts of southern Ontario as much as 13,000 years ago, and finally the northernmost portions of the province approximately 8,500 – 6,500 years ago (Crins et al. 2009; Chapman and Putnam 1984).

Broadly speaking, the distribution of Soil Orders in Ontario aligns with the terrestrial Ecozones of Canada as present in the province: the Hudson Plains Ecozone, the Boreal Shield Ecozone, and the Mixedwood Plains Ecozone. Bedrock geology, parent materials, climate and vegetation play key roles in governing this alignment and will be discussed in the following sections.

ECOZONES OF ONTARIO

Hudson Plains Ecozone

The **Hudson Plains Ecozone**, commonly referred to as the Hudson Bay Lowlands, occupies the northern portions of the province from the border with Manitoba in the west to the Quebec border in the east, and parallels the shores of Hudson Bay and James Bay (Figure 11.1). This ecozone comprises approximately 25% of the land area of Ontario, and has its southern limit marked by the boundary between mostly Paleozoic (541 to 252 million years ago [mya]) and Mesozoic (252 to 66 mya) limestones of the Phanerozoic Eon, which underlie the ecozone, and the Precambrian Shield (Figure 11.2, Crins et al. 2009). These rocks are softer than those of the Precambrian Shield and are neutral to alkaline in reaction (refer to Chapter 5: Soil pH for more information). The climate is described as relatively cold and semi-arid, with average annual temperature ranging from -6.2 to 7 °C (Figure 11.3, Fick and Hijmans 2017) and mean annual precipitation ranging from 240 to 525 mm (Figure 11.4, Stewart and Lockhart 2005). January and July mean daily air temperatures are between -20 to -27.5 °C and 12 to 16 °C, respectively. Generally speaking, the Hudson Plains Ecozone, as the name would suggest, is a relatively flat landscape characterized by poor drainage. After deglaciation, sea water flooded the landscape, depositing vast clay and silt deposits over the limestone bedrock. Isostatic rebound, the process of land surface uplift or recovery from the weight of the continental glacier, is visible by the sequence of beach ridges formed parallel to the modern shoreline of Hudson Bay. The rate of isostatic rebound is greatest along the shores of Hudson Bay, estimated at a rate of 13 mm per year (Henton et al. 2006). The vast flat landscape is dissected by major rivers that have carved into the landscape and empty into the marine waters of Hudson Bay and James Bay.

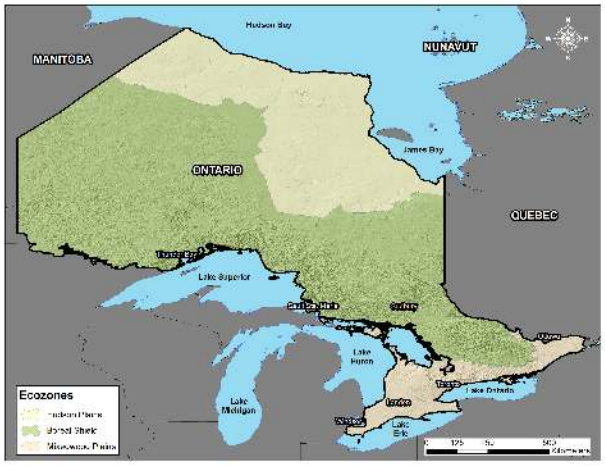


Figure 11.1. Ecozones of Ontario. © Daniel Saurette is licensed under a CC BY (Attribution) license.

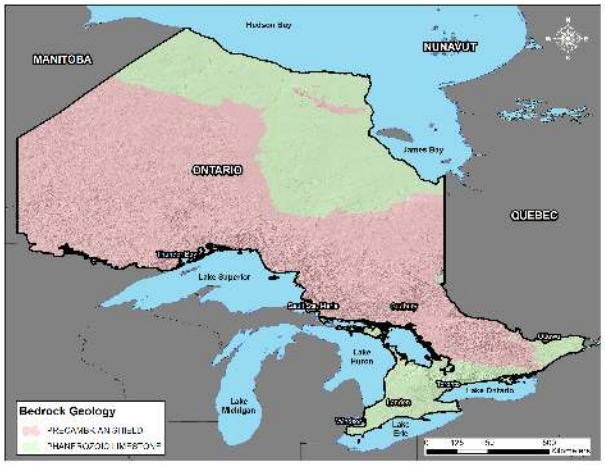


Figure 11.2. Bedrock geology of Ontario. © Daniel Saurette is licensed under a CC BY (Attribution) license.

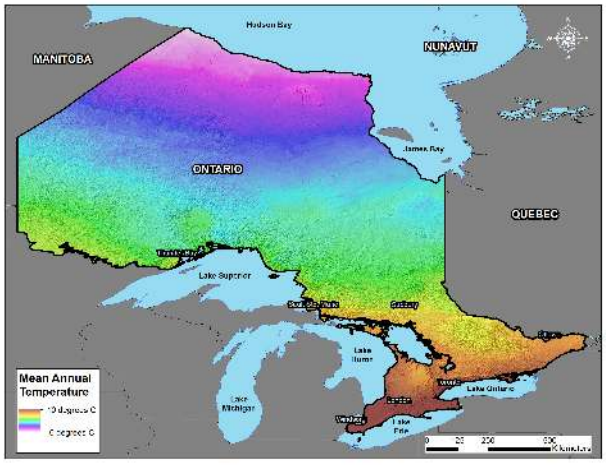


Figure 11.3. Mean annual temperatures in Ontario. © Daniel Saurette is licensed under a CC BY (Attribution) license.

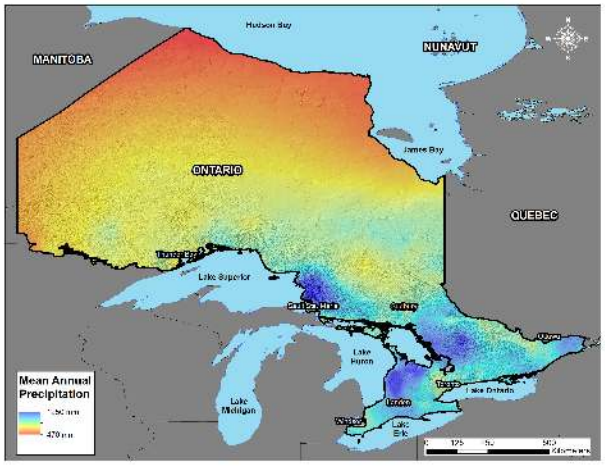


Figure 11.4. Mean annual precipitation in Ontario. © Daniel Saurette is licensed under a CC BY (Attribution) license.

Boreal Shield Ecozone

The Boreal Shield Ecozone occupies a central band of the province, extending from the Manitoba border in the west, to the Quebec border in the east, and occupies approximately 66% of the province (Figure 11.1). The Boreal Shield Ecozone is bordered by Paleozoic limestone bedrock in the north and south (Figure 11.2). Bedrock is Precambrian, more specifically from the Archean Eon (formed 4000 to 2,500 mya) and is dominated by gneisses and granites. These rocks form a major portion of earth's crust and are predominantly metamorphic rocks, formed when the Earth initially cooled enough to allow for molten rocks to solidify. These rocks are resistant to weathering, significantly more so than the neighboring limestones, and their reaction is typically acidic. The climate is described as relatively cold and moist, with average annual temperature increasing from -3.5 °C to 6.8 °C (Figure 11.3) and mean annual precipitation increasing from 517 mm to 1246 mm from west to the east across the ecozone. (Figure 11.4, Fick and Hijmans 2017). January and July mean daily air temperatures are -15 °C and 17 °C, respectively (Stewart and Lockhart 2005). The geographic extent of the ecozone is such that air temperatures can vary dramatically across the Boreal Shield, especially in proximity to the Great Lakes. Surficial deposits are complex and are dominantly the relics of the Wisconsinan Glaciation. Exposed bedrock, a result of the continental glacier scraping and relocating materials further south, is significant in extent and dominates large tracts of this landscape. Scattered throughout the bedrock exposures are morainal (till) deposits, glacio-fluvial deposits such as eskers and kames, glacio-lacustrine deposits from ancient lakes, and organic deposits. The thickness of these deposits varies considerably from thin veneers (10 to 100 cm) over bedrock to deep deposits many meters thick. This ecozone is known provincially as the Ontario Shield, and contains not only the conifer-dominated boreal forest, but also the important portions of Great Lakes-St. Lawrence forest region composed dominantly of deciduous hardwood forests.

Mixedwood Plains Ecozone

The Mixedwood Plains Ecozone occupies the most southern reaches of Ontario extending from the western tip of Lake Erie near Windsor, Ontario, to the Quebec border in the east, bounded by the Precambrian Shield to the north, Lake Huron to the west, and Lakes Erie and Ontario to the south (Figure 11.1). The bedrock that underlies the ecoregion is of the same origin as that of the Hudson Plains Ecozone: limestone and dolostone of Phanerozoic origin. It occupies the smallest area of the three ecoregions in Ontario; approximately 9% of the land surface. Despite the small footprint, this ecozone is the most extensively studied and managed area in Ontario, primarily because the major population centers are all located within this ecozone, and the most productive agricultural lands in Ontario (and Canada) are located here. For this reason, there is a disproportionate amount of data, and data of higher resolution, available in this part of the Province. The climate is described as cool to mild, having cool winters and warm summers, and is known to be one of the mildest climates in Canada (Crins et al. 2009), largely due to the moderating effects of the Great Lakes. January and July mean daily air temperatures range from -3 to -12 °C and 18 to 22 °C, respectively. Moisture is more abundant with the ecoregion receiving from 770 to 1,202 mm of precipitation annually, key to the development of soils which will be expanded upon in the subsequent sections. The surficial geology is extremely complex and has been the focus of numerous works (e.g., Chapman and Putnam 1984; Karrow and Calkin 1985). The Laurentide Ice Sheet began its retreat from Ontario along the southwestern peninsula which is currently surrounded by Lake Huron, Georgian Bay, Lake Erie and Lake Ontario. The ice split from southwest to northeast creating numerous ice lobes that retreated into today's modern lake basins, all the

while forming iterations of glacial lakes in the process at the foot of the glacier. This resulted in a series of terminal, interlobate and ground moraines, vast glacio-fluvial outwash plains and extensive glacio-lacustrine deposits. The retreat from southwestern Ontario also exposed the Niagara Escarpment, which arches from Niagara Falls to the tip of the Bruce Peninsula and the shorelines of Manitoulin Island. In the east, the Champlain Sea was formed and inundated portions of the Ottawa and St. Lawrence River valleys, depositing significant areas of glacio-marine sediment.

DISTRIBUTION OF SOILS IN ONTARIO

Soil distribution in the Hudson Plains Ecozone

The distribution of soils in Ontario is controlled by the relative importance or dominance of soil forming factors, and in many cases one soil-forming factor overrides many others. This is the case for the soils in the Hudson Plains Ecozone. The cold, semi-arid climate is the single most important factor. Despite the relatively small amount of annual precipitation, the cold annual temperature and short growing season do not allow for high rates of evapotranspiration, leaving much of the moisture in the soil. This is exacerbated by the dominantly clayey and silty textured glacio-marine parent materials which have high water holding capacities. This results in soils that are saturated most of the year and favor the accumulation of organic materials at the surface (Figure 11.5). In addition, in the cooler northwest portions of the ecozone, mean annual temperatures are so low that permafrost is a common occurrence. These conditions result in soils of the Organic Order and Cryosolic Order being dominant (Figure 11.6). At the Great Group level, Fibrisols of the Organic Order constitute the majority of the ecozone. Fibrisols are characterized by materials composed of relatively undecomposed organic materials, developed mostly from sphagnum mosses (SCWG 1998) and contain >17% organic carbon. In terms of taxonomy, Fibrisols have a dominantly fibric middle tier (40–120 cm), or middle and surface tiers if a terric, lithic, hydric or cryic contact occurs in the middle tier (Kroetsch et al. 2011). Fibrisols dominate because the cool temperatures slow the rate of decomposition of organic materials dramatically, resulting in the organic materials rated as classes 1 to 4 on the von Post scale of decomposition (SCWG 1998). In field diagnostics, when samples are squeezed in the hand, fibric materials yield clear to yellow-brown water and the plant remains are relatively unaltered (Kroetsch et al. 2011). More decomposed materials (classified as Mesisols) are found sporadically in small pockets typically at the southern edge of the ecozone and bordering the Canadian Shield. Mesisols are characterized as dominantly composed of mesic organic materials, or those materials in classes 5 and 6 on the von Post scale of decomposition (SCWG 1998). The extreme northwest of the ecozone, bordering northeastern Manitoba, is dominated by Organic Cryosols. Organic Cryosols are soils developed from organic deposits, similar to the Fibrisols and Mesisols, but differ in that they also have permafrost within one metre of the surface.

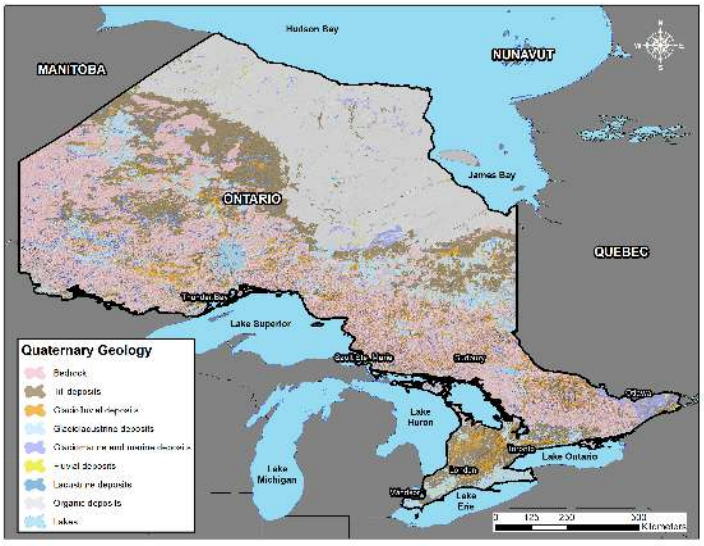


Figure 11.5. Quaternary geology of Ontario. © Daniel Saurette is licensed under a CC BY (Attribution) license.

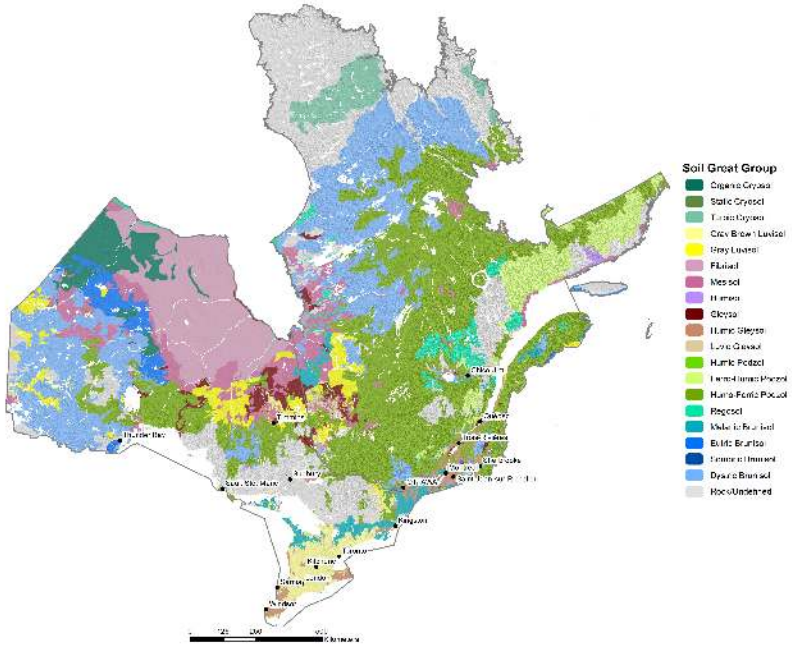


Figure 11.6. Distribution of Soil great Groups in Ontario and Quebec. Map is a reproduction of an official work published by the Government of Canada and was provided courtesy of Darrel Cerkowniak, Agriculture and Agri-Food Canada. © Darrel Cerkowniak, Agriculture and Agri-Food Canada; licensed under a CC BY (Attribution) license.

Soil distribution in the Boreal Shield Ecozone

Soils of the Boreal Shield Ecozone are more diverse than those of the Hudson Plain Ecozone. Whereas the Hudson Plains Ecozone is dominated by one soil forming factor, climate, the distribution of Soil Orders across the Boreal Shield Ecozone is governed by climate, parent material, topography and vegetation. Some of these soil forming factors exert their influence on pedogenesis over large regional scales (e.g., precipitation, parent materials), while others exert their influence over short distances (e.g., topography or relief).

In the west, generally from Thunder Bay and Lake Nipigon to the Manitoba border, soils of the Dystric Brunisol Great Group dominate (Figure 11.6). Dystric Brunisols are juvenile soils with weak profile development that have a Brunisolic Bm, Bfj, or thin Bf diagnostic horizon less than 5 cm thick and an acidic reaction. Dystric Brunisols either have no Ah or a very thin Ah horizon and a pH (0.01 M $CaCl_2$) in the upper 25 cm of the Bm horizon of <5.5 (Smith et al. 2011). The lack of an Ah horizon indicates conditions that favor mor humus forms, as opposed to mull or moder humus forms (see Chapter 2, section 2.6.1 for information on humus forms). These are forest soils generally formed on non-calcareous materials, typically transitional to Podzolic soils (Smith et al. 2011). In northwestern Ontario, these soils have developed on the sandy, acidic morainal deposits of the Canadian Shield. It should be noted that the Dystric Brunisols are interspersed with granitic bedrock outcrops, and in some areas the bedrock outcrops dominate as can be seen along the Manitoba border. Local depressions in the landscape are bedrock-controlled and therefore hold significant amounts of water, resulting in the development of Organic soils.

Northeast of Thunder Bay lies a band of soils dominated by the Humo-Ferric Podzol great group that parallel the north shore of Lake Superior, extending south of Timmins east to the Quebec border (Figure 11.6). The parent materials in this region are also sandy, non-calcareous, acidic morainal deposits of the Canadian Shield; but the Humo-Ferric Podzols are the succession soils of the Dystric Brunisols. Podzols in general are associated with parent materials that have low pH buffering capacity, vegetation that yields acidic litter materials (typically conifers), and climates where annual precipitation exceeds evapotranspiration, leading to a leaching environment (Sanborn et al. 2011). The difference from the western part of the Canadian Shield described above is the higher annual precipitation, most notably in proximity to and east of Lake Superior (Figure 11.4). Increased precipitation leads to increased leaching and movement of iron and aluminum into the B horizon.

A significant area north of Timmins, but south of the Hudson Plain, is dominated by Gray Luvisols, Gleysols and Mesisols; this area is known as the Great Clay Belt. The parent materials in this region were deposited by proglacial Lake Barlow-Ojibway, and consist dominantly of glacio-lacustrine clays and silts, and heavy-textured morainal (till) deposits (Figure 11.5). Numerous bedrock

exposures dissect the clay plain; associated with the bedrock ridges are the heavy-textured morainal deposits. These deposits are neutral to alkaline, carbonated, base-saturated parent materials, with low hydraulic conductivity and are relatively stone free. These characteristics favour the development of Luvisolic soils in upper to mid landscape positions. Gleysolic soils occupy the lower landscape positions and organic soils of the Mesisol great group, those in the middle stage of decomposition, occupy level and depressional landscape positions. Luvisolic soils are typically formed in forest environments, more commonly in mixedwood and deciduous forests, and in areas of sub-humid to humid moisture regimes and mild to very cold climates (Lavkulich and Arocena 2011). Diagnostics properties of Gray Luvisols include an eluvial Ae horizon and an illuvial Bt horizon, characteristic of the removal of clay from the Ae with subsequent deposition in the Bt horizon (Lavkulich and Arocena 2011). This translocation is evident by the increased clay content of the Bt horizon and visible clays skins on the peds of the Bt horizon.

Poorly-drained soils of the Gleysol subgroup occupy significant areas in this region. The great groups of the Gleysolic Order reflect the occurrence (or absence) of additional processes beyond that of the primary gleying process (Bedard-Haughn 2011). The Gleysol great group is a great group of the Gleysolic Order that has no secondary process; therefore, these soils are characterized simply by the occurrence of anaerobic conditions inferred by soil color and/or the presence of a water table (**see Chapter 8**). The clayey deposits of the Great Clay Belt favor water retention due to low hydraulic conductivities and infiltration rates, resulting in poor drainage conditions and the development of Gleysols. Another important factor is the dominantly level topography of the Great Clay Belt which leads to poor external drainage, which exacerbates the internal drainage conditions. Gleysols are widespread in the Great Clay Belt and occur in lower slope positions in morainal and glacio-lacustrine deposits, and in level areas of the clay plain. Mesisols account for the remainder of the dominant soils in the area, and, as previously described, are composed of organic deposits in the middle stages of decomposition on the von Post scale (classes 5 and 6). These soils occupy the vast low-lying and level portions of the landscape and typically support wetland communities.

The remaining portion of the Boreal Shield Ecozone, extending east from Sault Ste. Marie to the Quebec border, is dominated by exposed bedrock. This area of the shield contains exposed rock and thin drift (till) over bedrock of varying thickness. In the till deposits, materials are non-calcareous and acidic, and Dystric Brunisols and Humo-Ferric Podzols develop in complex associations with the bedrock. These types of landscapes are often described in soil maps and reports with land types or land units, which are groupings of soils of varying proportions, due to the complexity and high variability of the soils.

Soil distribution in the Mixedwood Plains Ecozone

Similar to the Boreal Shield Ecozone, the distribution of soils across the Mixedwood Plains Ecozone is governed by parent material, topography and vegetation; however, climate plays a less important role in controlling soil distribution within the ecozone. Unlike the precipitation gradient that results in separation of the Brunisolic from Podzolic soils in the Boreal Shield, soil development in the Mixedwood Plains is more reliant on the distribution of parent materials, or surficial deposits. The Mixedwood Plains Ecozone is primarily underlain by Paleozoic limestones, and as such, the parent materials are carbonate rich and typically neutral to alkaline in reaction. This provides the foundation for soil development.

Extensive glacio-lacustrine plains occur in southwestern Ontario north of Lake Erie (Figure 11.5). These are remnants of numerous iterations of glacial lakes that formed during the retreat of the Laurentide Ice Sheet. These areas correlate with areas dominated by Humic Gleysols (Figure 11.6). Soils of the Humic Gleysol great group differ from those of the Gleysol great group (previously described) due to the presence of a thick Ah (>10 cm) or Ap (>15 cm) horizon (Bedard-Haughn 2011) which indicates incorporation and decomposition of organic matter in the surface horizon (see Chapter 8: Wetland Soils for more information on Gleysols). Similar to the soils of the Great Clay Belt, these glacio-lacustrine plains are predominantly composed of clay-dominated materials, with areas of clay-rich tills deposited as a result of glacial re-advance over previously deposited glacio-lacustrine clays. The heavy-textured soils have low infiltration rates, low hydraulic conductivity, and the level landscape limits external drainage, resulting in development of Humic Gleysols. The remainder of the southwest peninsula of Ontario, and east along the north shore of Lake Ontario, is largely dominated by Gray Brown Luvisols. In terms of diagnostic properties, the Gray Brown Luvisols share the same as the Gray Luvisols (described above), however Gray Brown Luvisols have an Ah (mull) horizon at the surface (Lavkulich and Arocena 2011) which differentiates them from Gray Luvisols. Despite evidence that organic matter and decomposition differ between Gray and Gray Brown Luvisols, there is no general consensus on the influence on pedogenesis (Lavkulich and Arocena 2011); however, it is recognized that Gray Brown Luvisols form in deciduous-dominated hardwood forests of eastern Canada (maple, oak, ash) and that the Ah forms from the mixing of organic matter with the mineral surface due to faunal activity, primarily earthworms. A review of the Quaternary geology of Ontario map (Figure 11.5) reveals that the areas mapped as

Gray Brown Luvisols are predominantly composed of morainal and glacio-fluvial deposits. The till deposits are mainly medium textured, while the glacio-fluvial deposits vary from moderately coarse to very coarse textured. Both are calcareous parent materials, hence the trajectory to Luvisolic soils and not Podzolic soils.

North of the Gray Brown soils, in a narrow band running east-west, is an area of Melanic Brunisols. This area is at the intersection of the Canadian Shield metamorphic bedrock to the north and the Phanerozoic limestone bedrock to the south and is dominated by thin morainal deposits over bedrock. These morainal deposits are a mix of the granitic materials originating from the Shield and the medium-textured calcareous materials from the limestone, and as such the materials are moderately coarse in texture, can contain significant coarse fragments from gravel to stone in size, and are neutral to alkaline in reaction with calcareous parent material. This combination results in the formation of Melanic Brunisols. Brunisols are differentiated at the great group level based on the presence or absence of an Ah horizon ≥10cm in thickness and the pH of the underlying Bm horizon. For the Melanic Brunisols, the Ah horizon ≥10 cm thick is present, and the Bm has a pH ≥5.5 (0.01M $CaCl_2$) in the upper 25 cm (Smith et al. 2011). These soils occur in association with the Gray Brown Luvisols and have a similar mull surface horizon originating from the incorporation of forest litter. In this region, the parent materials are extremely calcareous, and more time is required to neutralize and remove the carbonates from the upper solum, which limits soil development and results in truncated or shallow soil profiles with carbonates occurring much closer to the surface than in the Gray Brown Luvisols to the south. East of Kingston, Ontario, and towards Ottawa is another region dominated by the Melanic Brunisols. These are developed mostly in the Grenville till, found east of the Frontenac Axis. The Frontenac Axis is an intrusion of Canadian Shield bedrock that dissects the Phanerozoic bedrock just east of Kingston. It should be noted that even though Brunisols dominate the Grenville till, lower lying areas and depressions contain Humic Gleysols.

Finally, the areas surrounding Ottawa and the confluence of the Ottawa and St Lawrence rivers are dominated by Humic Gleysols. These soils are developed in glacio-marine deposits from the Champlain Sea that inundated the region when it was depressed or lowered by the weight of the ice (isostatic depression) during glaciation and subsequent rebound after deglaciation. These materials are clay-rich, and for reasons previously explained, favor the development of Gleysolic soils. Champlain Sea clays are unique in that they were deposited in a brackish, or saline, marine environment. The salts deposited with the clays had positive charges binding the negatively charged clay particles together into a stable mass. When saline waters retreated and the materials were exposed to precipitation, the salts were dissolved and leached from the clays, in some cases being reduced from upwards of 30 g/l down to <1 g/l (Geertsema 2013). The increase in the repulsive forces between clay particles as a result of reduced salinity can render the clays unstable and susceptible to landslides and mass wasting. More than 250 landslides have been mapped in these clay deposits in Canada (see Chapter 12: Soils of Quebec for additional information on the Champlain Sea clay deposits). These clays are commonly referred to as quick clay or Leda Clay.

LAND USE

Land use in the Hudson Plains Ecozone

The Hudson Plains Ecozone is relatively untouched and pristine. It is dominated by sparse, slow-growing vegetation in bog and fen environments, dominated by black spruce, tamarack and birch. These low productivity forests are not suitable for commercial forest harvesting operations. The Ecozone is remote—there is little to no transportation infrastructure in place to access the region. Settlements are located along the coasts of Hudson Bay and James Bay, along major rivers, from which most derive their names. These include Fort Severn, Peawanuck (Severn), Attawapiskat, Fort Albany, Moosonee and Moose Factory. These settlements are only accessible by air or water, with the exception of Moosonee which is serviced by rail. Ontario's only diamond mine is located less than 100 kilometers west of Attawapiskat, but it is set to end production by the end of 2019. The Hudson Plains Ecozone is home to Polar Bear Provincial Park, Ontario's largest Provincial park, located along the shores of Hudson Bay and James Bay.

Land use in the Boreal Shield Ecozone

The Boreal Shield Ecozone is known for its abundance of natural resources, making it an ideal setting for the forestry and mining sectors. Ontario has 71 million hectares of forests, of which 43 million hectares are crown forests managed by the Ontario Ministry of Natural Resources (OMNR 2016). These managed forests are located mostly within the Boreal Shield Ecozone (Figure 11.7). The forest sector supports local communities, creating on average 51,660 direct jobs annually (OMNR 2016). Crown forests are managed in administrative zones (forest management units) and through forest management plans.

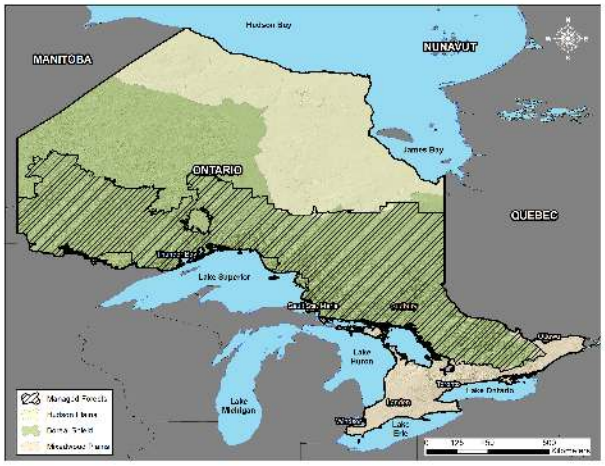


Figure 11.7. Extent of Crown managed forests in Ontario. © Daniel Saurette is licensed under a CC BY (Attribution) license.

The mining sector plays an important role in the Boreal Shield, mostly through extraction of various precious metals from the Precambrian Shield bedrock. In 2019, there were 39 active mines in Ontario, of which 33 were located in the Boreal Shield. Of these, 18 were gold mines, nine were base metal mines, one iron mine, one platinum mine and four industrial mineral operations (Ontario Mining Association-OMA 2019). The industry sustains 26,000 direct jobs and over 50,000 indirect jobs in the province and boasted a $10.1 billion value of mineral production in 2018 (OMA 2019). Mining operations span the entire Boreal Shield, but the areas surrounding Timmins, Sudbury and Wawa have the most intensive operations. A new major mining project, known as the Ring of Fire, located in northwestern Ontario, north of Thunder Bay, is currently under development.

Agricultural land use is not widespread in the Boreal Shield, but rather occupies smaller pockets of land scattered throughout the ecozone. These small pockets of agricultural land are typically associated with soils formed on deposits other than the coarse-textured morainal deposits of the Shield (e.g., glacio-lacustrine deposits from proglacial Lakes Agassiz, Barlow-Ojibway and Nipissing). Agricultural production is predominantly focused on livestock production, due to short growing seasons and less fertile soils; however, a warming climate has seen a transition from hay and livestock to cash crops such as corn, soybean, canola and small grains in Temiskaming Shores, Nippissing and Algoma.

Land use in the Mixedwood Plains Ecozone

A large portion of the Mixedwood Plains Ecozone has been converted from natural forests to agricultural production and urban settlements (Figure 11.8); it is the most densely populated ecozone in Canada (Crins et al. 2009). In 2018, agriculture accounted for 53.2% of the land use in the ecozone (Table 11.1), or approximately 4.5 million hectares of land (AAFC 2018). Forests accounted for over 23% of the ecozone (11.5% broadleaf, 8.4% mixedwood and 3.4% conifer), and urban or developed land covered over 11% (934,898 ha). Cash crops dominate the agricultural landscape, the largest three crops being soybeans, grain corn and cereals, primarily winter wheat (AAFC 2017). The clayey glacio-lacustrine plains of the far southwest, the Niagara Region above the Niagara Escarpment and the Ottawa Valley are dominated by corn, soybean and winter wheat crops. These areas are extensively drained by systematic tile drainage systems and/or networks of surface drains, which also serve as outlets for the tile drainage systems. Without drainage, the soils in this area would have limited agricultural production due to excess water. The area between the Niagara Escarpment and Lake Ontario is home to a thriving specialty crop industry that includes vineyards, orchards (apples, pears), tender fruit production (peaches, plums, cherries, apricots, etc.) and greenhouses (vegetables and floral industry). In winter, warm air over Lake Ontario rises, and the northerly winds push the air inland towards the Niagara Escarpment. This displacement of air results in cool air being drawn off the land surface and out towards the lake, creating a circular system which moderates the winter climate. The central part of southwestern Ontario, dominated by undulating morainal hills and glacio-fluvial deposits, is also dominated by corn, soybean and winter wheat cash crop operations. Moving northward toward the Bruce Peninsula, the rolling topography becomes steeper and cropping operations give way to cattle production with more land in pasture and forages.

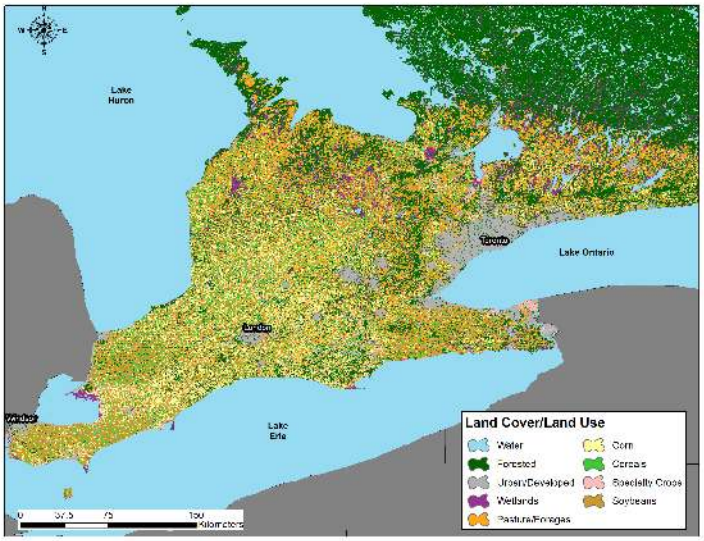


Figure 11.8. Land use in the Mixedwood Plains Ecozone. © Daniel Saurette is licensed under a CC BY (Attribution) license.

Table 11.1. Summary of land use in the Mixedwood Plains Ecozone based on Agriculture and Agri-Food Canada's 2018 Annual Crop Inventory (AAFC 2019)

Land Use	Area (hectares)	Area (acres)	% of total
Agricultural			
Pasture / Forages	1,501,517	3,708,747	17.9
Soybeans	1,296,229	3,201,685	15.4
Corn	1,064,740	2,629,907	12.7
Cereals	495,040	1,222,748	5.9
Vegetables	38,752	95,718	0.5
Specialty	25,385	62,701	0.3
Pulses	25,108	62,016	0.3
Fruits	8,498	20,990	0.1
Fallow	6,797	16,788	0.1
Oilseeds	6,656	16,441	0.1
Sod	3,514	8,680	0
Greenhouses	2,148	5,306	0
Other Crops	230	569	0
Subtotal	**4,474,614**	**11,052,295**	**53.2**
Native			
Broadleaf	963,805	2,380,597	11.5
Mixedwood	706,106	1,744,081	8.4
Wetlands	478,298	1,181,397	5.7
Shrubland	405,111	1,000,623	4.8
Conifers	283,120	699,306	3.4
Subtotal	**2,836,439**	**7,006,005**	**33.7**
Developed			
Urban/Developed	934,898	2,309,199	11.1
Subtotal	**934,898**	**2,309,199**	**11.1**
Water			
Water	158,144	390,615	1.9
Subtotal	**158,144**	**390,615**	**1.9**
Grand Total	**8,404,095**	**20,758,113**	**100**

Canada Land Inventory – Soil Capability Classification for Agriculture

Land use planning in Ontario is subject to the Provincial Policy Statement (PPS), issued under section 3 of the Planning Act (Planning Act 1990). The PPS provides a series of policies that "set out the government's land use vision for how we settle our landscape, create our built environment, and manage our land and resources over the long term to achieve livable and resilient communities" (Ontario 2020). Protection of agricultural land is afforded under Part V: Policies 2.0 Wise use and management of resources, Policy 2.3 Agriculture, which states:

"Prime agricultural areas shall be protected for long-term use for agriculture. Prime agricultural areas are areas where prime agricultural lands predominate. Specialty crop areas shall be given the highest priority for protection, followed by Canada Land Inventory Class 1, 2, and 3 lands, and any associated Class 4 through 7 lands within the prime agricultural area, in this order of priority."

The Canada Land Inventory (CLI) was a federal-provincial program initiated in 1961 and administered under the Agricultural Rehabilitation and Development Act (DREE 1969), commonly referred to as ARDA (see Chapter 8, section 8.12 for general information on the CLI). The Soil Capability Classification for Agriculture (henceforth referred to as Agricultural Capability) under the CLI program is a seven-class interpretive system used to assess the impacts of climate, soil and topography on the potential for production of field crops. Capability classes range from Class 1 to Class 7, with increasing limitations to crop production (Table 11.2). Limitations are noted as subclasses in the system, and soils are assigned to a class based on the type and severity of limitations (subclasses) for crop production.

Table 11.2. Description and summary of CLI agricultural capability classes and subclasses in Ontario

Capability Class	Description
Class 1	No significant limitations in use for crops
Class 2	Moderate limitations that reduce the choice of crops, or require moderate conservation practices
Class 3	Moderately severe limitations that reduce the choice of crops or require special conservation practices
Class 4	Severe limitations that restrict the choice of crops, or require special conservation practices and very careful management, or both
Class 5	Very severe limitations that restrict their capability to producing perennial forage crops, and improvement practices are feasible
Class 6	Unsuited for cultivation, but are capable of use for unimproved permanent pasture
Class 7	No capability for arable culture or permanent pasture
Capability Subclass (Limitations)	
Subclass C	Adverse climate: This subclass denotes a significant adverse climate for crop production as compared to the "median" climate which is defined as one with sufficiently high growing-season temperatures to bring common field crops to maturity, and with sufficient precipitation to permit crops to be grown each year on the same land without a serious risk of partial or total crop failures. In Ontario this subclass is applied to land averaging less than 2300 Crop Heat Units.
Subclass D	Undesirable soil structure and/or low permeability: This subclass is used for soils which are difficult to till, or which absorb or release water very slowly, or in which the depth of rooting zone is restricted by conditions other than a high water table or consolidated bedrock. In Ontario this subclass is based on the existence of critical clay contents in the upper soil profile.
Subclass E	Erosion: Loss of topsoil and subsoil by erosion has reduced productivity and may in some cases cause difficulties in farming the land e.g. land with gullies.
Subclass F	Low natural fertility: This subclass is made up of soils having low fertility that is either correctable with careful management in the use of fertilizers and soil amendments or is difficult to correct in a feasible way. The limitation may be due to a lack of available plant nutrients, high acidity, low exchange capacity, or presence of toxic compounds.
Subclass I	Inundation by streams or lakes: Flooding by streams and lakes causes crop damage or restricts agricultural use.
Subclass M	Moisture deficiency: Soils in this subclass have lower moisture holding capacities and are more prone to droughtiness.
Subclass P	Stoniness: This subclass indicates soils sufficiently stony to hinder tillage, planting, and harvesting operations.

Capability Class	Description
Subclass R	Consolidated bedrock: The occurrence of consolidated bedrock within 100 cm of the surface restricts rooting depth and limits moisture holding capacity. Conversely, in poorly drained soils the presence of the bedrock may, depending on depth, make artificial drainage impossible.
Subclass S	Adverse soil characteristics: This subclass denotes a combination of limitations of equal severity. In Ontario it has often been used to denote a combination of F and M when these are present with a third limitation such as T, E or P.
Subclass T	Topography: This subclass denotes limitations due to slope steepness and length. Such limitations may hinder machinery use, decrease the uniformity of crop growth and maturity, and increase water erosion potential.
Subclass W	Excess water: This subclass indicates the presence of excess soil moisture due to poor or very poor soil drainage. It is distinguished from Subclass I - water inundation which indicates risk of flooding from adjacent lakes or streams.

Since completion of the program in the mid 1970s provincial governments have gradually modified the system to meet provincial needs. While these modifications provide information that is more relevant at the provincial level, the differences across jurisdictions would now likely preclude direct comparison of soils across provincial boundaries without a thorough review of the differences between modern rating systems. The subtleties typically reside in how subclass limitations are applied. For example, in Saskatchewan, Subclass C (climatic) limitations are assigned based on heat and moisture deficiencies (Shields et al., 1968), while in Ontario, Subclass C limitations are assigned simply based on crop heat units (OMAFRA 2020). There are also situations where the criteria are quite provincially specific. Take for example Subclass D, in Ontario, which is determined by the presence and depth at which soil materials with > 35% or > 60% clay are found, rules that are not found or adhered to in other Provinces.

Eleven subclasses are recognized in Ontario. A detailed description of the CLI system and its application in Ontario is maintained by the Ontario Ministry of Agriculture, Food and Rural Affairs (OMAFRA 2020). All soils in the province have been assigned an Agricultural Capability rating, and these ratings form the basis for the protection of agricultural lands in Ontario. Soils in Classes 1, 2 and 3 are designated as prime agricultural lands; the soil mapping in Ontario forms the basis for delineating prime agricultural areas, which contain predominantly prime lands, but also potentially areas of Classes 4 through 7. OMAFRA's Agricultural System in Ontario's Greater Golden Horseshoe (GGH) is a prime example of the integration of CLI Agricultural Capability ratings into land use planning and preservation of prime agricultural areas for future generations. The data is available through the Agricultural System Portal and provides base data to assist municipalities in the GGH with update to their Official Plans.

Ontario's Provincial Soil

Did you know many provinces have designated a Provincial Soil? In Ontario, the provincial soil is the Guelph soil series, and it was officially named the provincial soil on November 30th, 2015. The Guelph soil is classified as a Brunisolic Gray-Brown Luvisol and is found on rolling hills and drumlins in Wellington County and surrounding areas, covering over 70,000 hectares. The Guelph series was the first soil described in the Preliminary Soil Survey of Southwestern Ontario, published in 1923, another reason for its selection as the provincial soil. In addition, the Guelph series is an agriculturally important soil, considered prime agricultural soil when not limited by slope, developed in loam to silt loam textured morainal materials, often with a loess cap that is free of stones.

SUMMARY

- Three Ecozones are found in Ontario:
  - Hudson Plains Ecozone
  - Boreal Shield Ecozone

- Mixedwood Plains Ecozone
- Major soil forming factors in Ontario are spatially distributed:
  - Hudson Plains: cold climate; low annual precipitation; limestone-dominated bedrock; suficial deposits comprised of marine silts and clays with extensive organic deposits.
  - Boreal Shield: annual precipitation and mean temperature gradients both increasing from west to east; igneous (Shield) dominated bedrock; surficial deposits comprised of morainal, glaciofluvial, glacio-lacustrine and organic material with extensive amounts exposed bedrock.
  - Mixedwood Plains: mild climate with sufficient precipitation across the region with a slight decrease from west to east; limestone-dominated bedrock; surficial deposits comprised of glacio-lacustrine plains, morainal deposits, glaciofluvial outwash and glaciomarine plains.
- The dominant Soil Orders present in Ontario include:
  - Cryosolic Order
  - Brunisolic Order
  - Gleysolic Order
  - Luvisolic Order
  - Organic order
  - Podzolic Order
- The soil forming factors and soil great group distribution in Ontario are related:
  - Cryosolic, Organic and Gleysolic soils dominate the Hudson Plains Ecozone due to the cool temperature and the clay and silt marine materials deposited after deglaciation.
  - Brunisolic and Podzolic soils dominate the Boreal Shield Ecozone, with Dystric Brunisols in the western half, and Podzols in the eastern half linked to a precipitation gradient from west to east.
  - The Mixedwood Plain Ecozone is dominated by Humic Gleysols which are associated with vast glacio-lacustrine and glacio-marine plains and depressional areas, and by Gray-Brown Luvisols which are associated with morainal deposits.
- Land use in Ontario:
  - The Hudson Plains Ecozone is mostly pristine, with very little infrastructure to access the region.
  - The Boreal Shield Ecozone is a resource-rich region in Ontario, and supports forestry, mining, and recreational opportunities.
  - The Mixedwood Plains Ecozone is the most densely populated ecozone in Canada and is an important agricultural region.
- Human interaction with the environment is intricately linked to soil distribution in Ontario, as it is around the world.
  - The cold, wet, and sometimes frozen soils of the Hudson Plain Ecozone do not provide opportunities for resource extraction or agricultural development, as thus the region remains relatively undisturbed by direct human activities.
  - The acidic, shallow, and low fertility soils of the Boreal Shield Ecozone are not suitable for agricultural development, and therefore support resource extraction and smaller urban settlements.
  - The neutral to alkaline, fertile soils and mild climate of the Mixedwood Plains Ecozone are ideal for a vast agricultural sector and large urban settlements it supports.

REFERENCES

Agriculture and Agri-Food Canada (AAFC). 2019. Annual Crop Inventory 2018. [Online]. Available: https://open.canada.ca/data/en/dataset/ba2645d5-4458-414d-b196-6303ac06c1c9 [11 December 2019].

Bedard-Haughn, A. 2011. Gleysolic soils of Canada: Genesis, distribution, and classification. Canadian Journal of Soil Science. 91: 763-770. doi:10.4141/CJSS10030.

Chapman, L.J. and Putnam, D.F. 1984. The physiography of Southern Ontario. Ontario Geological Survey. Special Volume 2, 270pp.

Crins, W.J., Gray, P.A., Uhlig, P.W.C., and Wester, M.C. 2009. The Ecosystems of Ontario, Part 1: Ecozones and Ecoregions. Ontario Ministry of Natural Resources, Peterborough, Ontario, Inventory, Monitoring and Assessment, SIB TER IMA TR-01. 71pp.

Department of Regional Economic Expansion. 1969. The Canada Land Inventory Soil Capability Classification for Agriculture. Report No. 2. 16pp.

Fick, S.E. and Hijmans, R.J. 2017. WorldClim 2: new 1-km spatial resolution climate surfaces for global land areas. Int. J. Climatol. 37: 4302-4315.

Geertsema, M. 2013.Quick clay. In Bobrowsky, P.T. (eds) Encyclopedia of Natural Hazards. Encyclopedia of Earth Sciences Series. Springer, Dordrecht. DOI: https://doi.org/10.1007/978-1-4020-4399-4_282

Henton, J.A., Craymer, M.R, Ferland, R., Dragert, H., Mazzotti, S. and Forbes, D.L. 2006. Crustal motion and deformation monitoring of the Canadian landmass. Geomatica. 60(2): 173-191.

Karrow, P.F. and Calkin, P.E., eds. 1985. Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30. 258pp.

Kroetsch, D.J., Geng, X., Chang, S.X. and Saurette, D.D. 2011. Organic soils of Canada: Part 1. Wetland organic soils. Canadian Journal of Soil Science 91: 807-822. doi:10.4141/CJSS10043.

Lavkulich, L.M. and Arocena, J.M. 2011. Luvisolic soils of Canada: Genesis, distribution, and classification. Canadian Journal of Soil Science 91: 781-806. doi:10.4141/CJSS2011-014.

Ontario Mining Association. 2019. Ontario Mining. [Online]. Available: https://www.oma.on.ca/en/ontariomining/OntarioMining.asp [11 December 2019]

Ontario Ministry of Agriculture, Food and Rural Affairs. 2020. Classifying prime and marginal agricultural soils and landscapes: guidelines for the application of the Canada Land Inventory in Ontario. [Online]. Available: http://www.omafra.gov.on.ca/english/landuse/classify.htm [7 February 2020].

Ontario Ministry of Natural Resources. 2016. State of Ontario's Natural Resources – Forests 2016. Queen's Printer for Ontario. Sault Ste. Marie, ON.

Ontario. Planning Act, RSO 1990. c.P.13. [Online]. Available: https://www.ontario.ca/laws/statute/90p13 [11 February 2020]

Sanborn, P., Lamontagne, L. and Hendershot, W. 2011. Podzolic soils of Canada: Genesis, distribution, and classification. Canadian Journal of Soil Science 91: 843:880. doi:10.4141/CJSS10024

Shields, J.A., Rostad, H.P.W., and Clayton, J.S. 1968. A guide to soil capability and land inventory maps in Saskatchewan (Second Edition). Publication No. M8. Saskatchewan Institute of Pedology. University of Saskatchewan. Saskatoon, Saskatchewan. 57pp.

Smith, C.A.S, Webb, K.T., Kenney, E., Anderson, A. and Kroetsch, D. 2011. Brunisolic soils of Canada: Genesis, distribution, and classification. Canadian Journal of Soil Science 91: 695-717. doi:10.4141/CJSS10058.

Soil Classification Working Group (SCWG). 1998. The Canadian System of Soil Classification. Agriculture and Agri-Food Canada. Publication 1646 (Revised). 187pp.

Statistics Canada. 2017. Cropland in Ontario grows despite fewer farms. Published by authority of the Minister responsible for Statistics Canada, Minister of Industry. Ottawa, ON. 9p.

Stewart, D.B. and Lockhart, W.L. 2005. An overview of the Hudson Bay marine ecosystem. Canadian Technical Report of Fisheries and Aquatic Sciences 2586, Fisheries and Oceans Canada, Winnipeg, Manitoba. 487pp.

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Dan's work focuses on soil survey, mapping and classification. Typically, I would be referred to either as a Pedologist, or Soil Surveyor. My work involves planning field campaigns to describe soil profiles, classify them, sample the soil to determine chemical and physical properties in the lab, and then interpret the data to develop soil maps. Traditionally this was done with the aid of air photos, but more recently the discipline has shifted to Predictive Digital Soil Mapping where we leverage the power of machine learning and geographic information systems (GIS) to create high-resolution digital soil maps.

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Dr. Jim Warren P.Geo. is an environmental soil scientist with a diverse background in research, consulting and teaching. He is a graduate of both the University of Guelph (B.Sc. (Agr), Soil Science), and the University of Alberta (M.Sc. Soil Science; Ph.D. Soil Chemistry). He conducted research in agronomy, geochemistry, and mineralogy at the University of Guelph and University of Waterloo before becoming a private consultant in acid mine drainage. He is currently employed as a Land Resource Specialist conducting soil survey and pedology work with the Ontario Ministry of Agriculture Food and Rural Affairs He is also an Adjunct Professor with the University of Guelph. He has delivered courses in soil science, soil mineralogy, soil chemistry, geochemistry, and pedology as a sessional lecturer at four universities (Alberta, Guelph, Waterloo, and Toronto). He has authored or co-authored over 100 publications in refereed journals; book chapters, conference proceedings and reports.

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Richard Heck (licensed under a CC-BY-NC-ND license)

Growing-up on a farm in south central Saskatchewan, I became aware of issues like soil salinity, wind erosion and variable crop production across the landscape. As I took my undergraduate courses, I came to understand the wonderful complexity, diversity and importance of soil. Ultimately, I chose to pursue a career in soils, and specifically in pedology. Over the years, I have had the opportunity to experience and study soils in various temperate, subtropical and tropical ecosystems. While my early work focussed on soil chemical and mineralogical characteristics, my more recent work is directed at quantifying soil structure and microstructure, within which the various soil processes function, as well as understanding the variability of soil magnetic susceptibility, as a means

of aiding in delineating soilscape moisture dynamics. The emphasis of my teaching is on helping the next generation also appreciate, by getting their hands ‘dirty’, the diversity of soils in our managed and natural landscapes.

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2.4: Soils of Quebec

Tim Moore

Learning Objectives

1. Describe the factors responsible for the distribution of soils in Quebec
2. Identify the diagnostic soil horizons associated with soil orders and great groups (according to the Canadian System of Soil Classification) of Quebec
3. Describe the impact of human activities on the soils of the region, and relate the major land uses to the properties of the soil

INTRODUCTION

Quebec contains a great deal of environmental diversity, not surprising because it is the province with the largest area of land, 1.36 million km^2, or 15 % of the Canadian total. A consequence of that diversity is the broad range of soils found in Quebec, spanning seven soil orders, ranging from fertile Brunisols and Gleysols in the south, to Podzols of the boreal forest, Brunisols in the subarctic and Cryosols in the north. This diversity reflects the range of soil forming factors and their control on soil forming processes (see Chapter 2).

SOIL FORMING FACTORS IN QUEBEC

The soil parent materials partially follow the underlying geology, varying from igneous and metamorphic rocks in the Canadian Shield to the metamorphic rocks in the Appalachian region to softer sedimentary rocks underlying the St. Lawrence valley (Bird 1980). However, the parent materials also reflect the emergence of Quebec from the last glaciation and the sediments laid down since then.

Quebec was covered by the Laurentide Ice Sheet until about 12,000 years ago when the southernmost part started to be freed by the ice sheet retreating northwards (Bird 1980; Richard and Occhietti 2005). With the land surface depressed by the mass of the ice, the sea rushed in to create the Champlain Sea in the St. Lawrence lowlands, depositing clay-rich marine sediments with a high initial salt content. As the land rose and emerged from the sea and glacial lakes, the clays were reworked and sandy deposits laid down, creating a rich mosaic of soil texture as rivers eroded downwards. The clay sediments laid down in the Champlain Sea were initially salt-rich and leaching of the salt and incision by rivers has created instability. This has resulted in over 700 landslides and earthflows in the St. Lawrence Lowlands, from Ottawa to Lac Saint-Jean, and some of which destroyed farmland and homes and killed people, such as at Nicolet in 1955 and at St. Jean Vianney in 1971 (Bird 1980; Carson 1978)

. Landslides continue to happen in Quebec and have had devastating consequences. Search the web for “Quebec landslides” for a number of images of the destruction associated with several of these events over the years. In northwestern Quebec, Lake Ojibway-Barlow formed, depositing fine sediments which now form the Clay Belt extending from Quebec into Ontario.

As the ice sheet continued to retreat northwards the land surface was covered with a thin layer of sandy/stony till. The last remnant of the ice sheet disappeared about 6,000 years ago in north central Quebec. Thus, Quebec soils have formed over the last 6,000 to 10,000 years, during which the climate has warmed and vegetation has moved northwards to occupy the glacial sediments.

The current climate varies from a mean annual temperature of 7°C around Montreal (45°N) to -5°C at Kuujjuaq in northern Quebec (58°N). Annual precipitation ranges from about 1000 mm in the south to about 550 mm in the north and the proportion of precipitation falling as rain varies from 80% in the south to 50% in the north. Runoff ranges from 400 mm in the south to 500 mm in the north, reflecting differences in evapotranspiration rates, and soils are not very moisture deficient, except for sandy, well-drained soils in the south.

Resulting from differences in climate, topography and parent material, the vegetation varies from deciduous forests in southwestern Quebec, mixed forests in the southern Appalachians and eastern Quebec through to dominantly coniferous forests in the Boreal Shield. In the north, the boreal forests give way to subarctic woodlands in the Taiga Shield, in which lichen forms an important ground cover and in northernmost Quebec into arctic tundra devoid of trees. Throughout Quebec, depressions collect water forming peatlands which contain plants adapted to wet conditions, such as sedges and mosses.

SOIL DISTRIBUTION IN QUEBEC

The soils of Quebec follow a classic ‘zonal’ distribution, from south to north (Figure 12.1), reflecting the dominating influence of climate and vegetation, but also the time for soil formation and the differences in parent material from south to north. Within these landscapes, topography becomes important, in producing shallow soils on steep slopes to poorly drained soils in depressions.

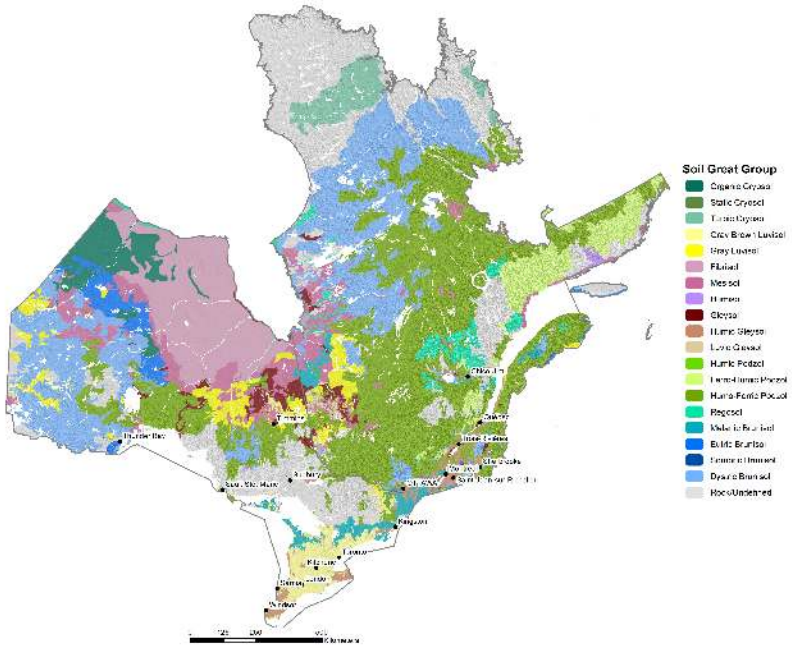


Figure 12.1. Map of Soil Great Groups in Quebec and eastern Ontario (according to the Canadian System of Soil Classification). Map is a reproduction of an official work published by the Government of Canada and is based on the 1:1,000,000 Soil Landscapes of Canada map. © Darrel Cerkowniak, Agriculture and Agri-Food Canada, is licensed under a CC BY (Attribution) license.

The dominant soil order in Quebec is Podzol, covering 39%, comprising mainly the Humo-Ferric great group and concentrated on the Boreal Shield and Taiga Shield terrestrial ecozones (Table 12.1). Podzols are characterized usually by a thick accumulation of organic matter at the surface, underlain by a leached, grey Ae horizon and then by a reddish brown B horizon showing the accumulation of iron and aluminum and/or organic matter, and defined by colour and chemical composition (see Chapter 8; Sanborn et al. 2011). Ferro-Humic and Humic great groups, with evidence of a larger translocation of organic matter within the soil profile than in the Ferric great group, cover 5% and occur primarily on the Canadian Shield along the North Shore in eastern Quebec, under a wetter climate.

Table 12.1. Coverage of soil Great Groups and Orders in Quebec

Great group	Order	Area (km^2)	Great Group (%)	Order (%)
Melanic Brunisol	Brunisol	9,714	0.71	17.4
Eutric Brunisol	Brunisol	3,279	0.24	
Sombric Brunisol	Brunisol	153	0.01	
Dystric Brunisol	Brunisol	223,940	16.44	
Turbic Cryosol	Cryosol	79,867	5.86	5.9
Organic Cryosol	Cryosol	65	0	
Static Cryosol	Cryosol	172	0.01	
Luvic Gleysol	Gleysol	4,360	0.32	2.7
Gleysol	Gleysol	14,210	1.04	
Humic Gleysol	Gleysol	17,863	1.31	
Gray Luvisol	Luvisol	22,553	1.66	1.7
Fibrisol	Organic	7,668	0.56	8.1
Humisol	Organic	5,327	0.39	
Mesisol	Organic	96,848	7.11	
Humic Podzol	Podzol	1,894	0.14	38.8
Humo-Ferric Podzol	Podzol	463,518	34.03	
Ferro-Humic Podzol	Podzol	62,825	4.61	
Regosol	Regosol	41,044	3.01	3
Undefined		303,190	22.26	22.3
Total		**1,362,237**		

Second in coverage is the Brunisol order, covering 17% dominated by the Dystric great group, mainly in the Taiga and Boreal Shield Ecozones, where profile development is not strong enough for soils to fall into the Podzol order. Brunisols are characterized by the absence of a single, dominant pedogenic process and the formation of a brown Bm horizon, indicative of some weathering (see Chapter 8; Smith et al. 2011). Both the Podzols and the Dystric Brunisolic great group form in the thin, acidic coarse textured materials left during glacial retreat. Dystric Brunisols can transform into Podzols over several thousand years of soil formation, as occurs in sandy deposits exposed by relative sea level fall along the Quebec North Shore, resulting in the Bm horizon being replaced by sesquioxide-cemented Bfc horizons (Moore 1976). Soils of the Melanic Brunisol great group, caused by the incorporation of organic matter into the upper mineral layer by soil-dwelling organisms, occur where the parent material is calcareous with deciduous forest cover, mainly in southwestern Quebec and the Clay Belt.

Organic soils, defined by an accumulation of organic matter over 40 cm thick, cover 8%, The Organic soils are dominantly Mesisols, in which most of the soil profile shows an intermediate degree of decomposition of the original plant tissues, whereas Fibrisols and Humisols show less and greater decomposition, respectively (see Chapter 8; Kroetsch et al. 2011). Where the topography and climate conspire to produce waterlogged conditions, decomposition of plant litter is slowed, producing organic soils with up to 5 m of peat. In some peatlands, such as bogs, the inherent decomposability of the plant litter is slow such as is found in *Sphagnum* mosses which are essentially the peat formers. The peatlands can be found in depressions underlain by clays in the St. Lawrence lowlands, as acid bogs and more nutrient rich swamps, and in waterlogged sections of the Boreal Shield and the

Clay Belt. The eastern coast of James Bay contains the extension of the Hudson Bay Lowlands, dominated by organic soils and covering 0.25 million km^2.

Cryosols, underlain by permafrost, cover 6% in the northernmost part of the province, in the Southern and Northern Arctic Ecozones, and extend down the Hudson Bay coast (see Chapter 8). In a colder climate and north of the tree line, sporadic permafrost terrain develops, including palsa, which are ice cored mounds in peatlands and wetlands (Tarnocai and Bockheim 2011). North of about 58°N continuous permafrost covers the landscape with Turbic Cryosols, in which freeze-thaw activities turnover, or cryoturbate, the soil profile.

Gleysols cover 3% and are formed under poor drainage conditions in depressions or on heavy textured parent materials such as the glacio-marine sediments in the St. Lawrence Lowlands and the Clay Belt. The high water table creates morphologic features such as grey coloured matrices that can include brown or reddish mottles indicative of oxidation-reduction processes or gleying (see Chapter 8; Bedard-Haughn 2011). This arises from the depletion of oxygen beneath the water table resulting in the reduction of iron from the insoluble ferric (Fe^{3+}) to soluble ferrous (Fe^{2+}) state, which then leaves the soil profile, producing a grey colour, or relocates to microsites which retain oxidizing conditions producing the distinctive mottles.

Luvisols, mainly of Gray Luvisol great group, cover 2%, and occur mainly in the Clay Belt. Luvisols form on clay-rich sediments containing calcium and magnesium and are characterized by the translocation of clay from the upper to middle part of the soil profile, producing the Bt horizon enriched in clay (Lavkulich et al. 2011).

The remaining mapped area comprises Regosols (3%) which are soils which are weakly developed and do not contain a recognizable B horizon at least 5 cm thick (see Chapter 8). They form on recently exposed sediments, on steep slopes leading to erosion and mass movement or where the parent material is resistant to weathering, such as on the Canadian Shield (VandenBygaart 2011). Finally, 22% of the area remains undefined, primarily in northernmost Quebec.

HUMAN-SOIL INTERACTIONS IN QUEBEC

Human activities have had a profound effect on soils in Quebec. The Algonquin and Iroquois Indigenous peoples of southern Quebec occupied the soils of the St. Lawrence lowlands and developed agriculture on the most fertile soils, which reached a peak in the 16th century. Riley (2013) argues that the arrival of Europeans disturbed the delicate balance between soils and the environment under indigenous management, causing an initial decline in indigenous population and then the widespread forestry operations in the 19th century, leading to erosion and degradation of the soils. With European arrivals, there was more intensive use of land for agriculture, concentrated along St. Lawrence River and also extensive forest harvesting starting in early part of 19th century. The varying soil texture provides a variety of viable activities, with arable cultivation on the clay-rich soils, tobacco on sandy soils near Joliette and apple orchards on the gravelly soils around the Monteregian Hills.

The imprint of human occupation is not restricted to the soils of southern Quebec. In northern Quebec, extremely nomadic people fished and hunted over 3000 years ago, though archaeological evidence is hard to find with poor preservation of bones in acid soils. Nevertheless, the soils contain evidence of occupation, such as hearths which changed the morphology of the soil and chemical changes associated with disposal of bones of fish and caribou. The best indicator of the latter is the increase in the phosphorus concentration of the soil, compared to undisturbed conditions, and changes in the fractionation of the phosphorus, in particular an increase in the calcium phosphate fraction (Moore 1986; Moore and Denton 1988).

The introduction of the Dominion Drainage Act from the mid-19th century allowed more extensive use of the potentially fertile but wet soils of St. Lawrence Lowlands, and led to the colonization of northern locations, such the Clay Belt in northwestern Quebec and around Lac Saint-Jean, which have heavy-textured soils.

The introduction of worms to Quebec forests, through fishing and farming and, has also affected soils. The forest leaf litter layer on the soil surface can be consumed by earthworms, and lost to the atmosphere as carbon dioxide, mixed with the underlying mineral horizon, or, in the case of deep-living earthworms, being incorporated into the mineral subsoil layers (Wironen and Moore 2006).

Although Quebec soils cover over 1 million km^2, soils of high agricultural potential cover a small area. The Canada Land Inventory identifies 600 km^2 of Class 1 soils (those with no significant limitation for crops), 10,000 km^2 of Class 2 soils (those that have moderate limitations) and 13,500 km^2 of Class 3 soils with moderately severe limitations, mainly wetness and fertility (Table 12.2). These soil classes comprise only 0.2, 2.9 and 4.2%, respectively, of the area surveyed by the Land Inventory (essentially south of 49° N), being located mainly in the land around Montreal, with pockets along the St. Lawrence River and in the Lac Saint-Jean

region, and small areas in the Clay Belt of northwestern Quebec. The Class 1 and 2 soils, especially when drained, have created the fertile 'Corn Belt' of agricultural land in the St. Lawrence Lowlands, now commonly in rotation with soya. Capability Class 1 to 3 soils are very restricted in area, and give an average of 0.28 ha (equivalent to a quarter of a football field) for each Quebecer, the smallest in any province in Canada and half the value in Ontario. Moreover, about 80% of the good quality soils (Classes 1 to 3) are covered by cropland, and this rises to 88% when pasture is included, a large percentage when compared to other parts of Canada. Thus, conservation of this limited area of good quality soils is important.

Table 12.2. The distribution of soils suitable for agriculture across Canada (Canada Land Inventory Capability Classes 1, 2 and 3*‡) by province or region and compared with population (estimated for 2019) and the area of cropland

Region	**CLI 1**	**CLI 2**	**CLI 3**	**ΣCLI 1-3**	**Population**	**ΣCLI 1-3 /population**	**Cropland†**	**Pasture†**	**Cropland / ΣCLI 1-3**
	(thousands of km^2)				**(million)**	**(ha per person)**	**(thousands of km^2)**		
Atlantic	0	5.89	22.78	28.67	2.4	1.19	3.1	0.9	0.11
QC	0.6	9.54	13.63	23.77	8.4	0.28	18.7	2.2	0.785
ON	22.49	23.6	32.79	78.87	14.4	0.55	36.5	5.3	0.463
MB	1.83	25.56	25.61	53.01	1.4	3.79	46.7	17.4	0.88
SK	10.72	64.46	100.82	176.00	1.2	14.67	163.9	65.1	0.931
AB	7.08	38.97	62.88	108.93	4.3	2.53	102.2	86.1	0.938
BC	0.7	3.98	10	14.68	5	0.29	5.8	16.4	0.395

*From the Canada Land Inventory http://sis.agr.gc.ca/cansis/publicat...agr/index.html
‡For descriptions of the Land Capability Classes, see https://sis.agr.gc.ca/cansis/nsdb/cli/classdesc.html
†Cropland defined as 'Land in crops' and pasture as 'tame or seeded pasture + natural land for pasture' in Statistics Canada Land Use (2016); https://www150.statcan.gc.ca/t1/tbl1...ers%5B0%5D=1.1

Legislation was introduced by the provincial government in 1978 to preserve agricultural land and agricultural activities, particularly the very limited good quality soils in the province. This was enacted in the region around Montreal, to prevent widespread 'suburban sprawl' outcompeting agricultural activities for the best soils. The agricultural zone, to which the legislation applied, covered 64,000 km^2 or only 12% of the land area of Quebec, further evidence of the paucity of good, agricultural soils. The legislation has been relaxed, so that suburban sprawl now occupies rich soils off the island of Montreal (see www.cptaq.gouv.qc.ca/fileadmin/en/publications/guides/Summary.pdf for an overview of the Act to preserve agricultural land and agricultural activities in Quebec). With climate change, the agricultural potential of soils in the Clay Belt and the Lac St. Jean region may improve.

Tree harvesting has occurred over the last three centuries in Quebec, initially on a small scale by indigenous communities, but more intensively with European settlement, so that few undisturbed forest stands remain. Tree harvesting occurred on a wide range of soils, mainly Podzols and Brunisols, but also Gleysols; unlike Finland, peatlands have not been drained to promote tree growth. Given the shallow nature of these soils and limited fertility, especially Podzols, tree harvesting may lead to exhaustion of nutrient supplies. Issue of exhaustion of nutrients through repeated harvesting operations poses a threat to the sustainability of forest operations on these soils. Thiffault et al. (2011) review the effects of whole-tree harvesting versus stem-only harvesting on soil properties in boreal and temperate forests. They show that, in general, whole-tree harvesting, as practiced in clearcutting, leads to larger reductions in soil carbon, nitrogen, phosphorus and base cations than stem-only harvesting, particularly on coarse-textured soils. However, there is great variability in the magnitude of the response arising from different forest types, climate and soil characteristics.

Extensive areas of bogs (Fibrisols and Mesisols, weakly and moderately decomposed peatlands, respectively) formed from slowly decomposing Sphagnum mosses and shrubs and trees, under acid, waterlogged conditions. They are of very limited value for agriculture, but substantial bog areas of Quebec, along with New Brunswick and Alberta, have been used to harvest horticultural peat moss. Flying along the St. Lawrence valley, one can see large bogs which have been drained, vegetation removed and the surface peat is dried and collected by giant vacuum cleaners. The peat is milled, dried and packaged for residential use (the bags of garden 'peat moss' bought in the spring), though the majority is used for horticultural purposes, such as a substrate for vegetables and flowers and, more recently, marijuana.

On the other hand, Mesisols and Humisols, peatlands formed from vegetation under higher nutrient conditions, and thus more decomposed, have been extensively drained in the St. Lawrence Lowlands for agriculture, particularly vegetables. Lowering the water table increases the rate at which the peat decomposes, releasing carbon dioxide, causes shrinkage of the peat as well as aeolian erosion, so that the peat surface is lowered by about 2 cm per year (Glenn et al. 1993; Parent et al. 1982). With soils originally between 1 and 1.5 m thick, these fertile soils have a limited lifespan.

Finally, 'climate change' is beginning to affect Quebec soil. This ranges from summer drought to mores spring flooding in the in the south to the thawing of Cryosol soils in the north, with resultant instability of transportation and housing sites.

REFERENCES

Bedard-Haughn, A. 2011. Gleysolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 763-779.

Bird, J.B. 1980. The Natural Landscapes of Canada. Second Edition, Wiley, Toronto.

VandenBygaart, A. J. 2011. Regosolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 881-887.

Carson, M.A. 1979. Le glissement de Rigaud (Québec) du 3 mai 1978 : une interprétation du mode de rupture d'après la morphologie de la cicatrice. Géographie physique et Quaternaire, 33: 63–92.

Glenn, S.M., Heyes, A., and Moore, T.R. 1993. Methane and carbon dioxide fluxes from drained peatland soils, southern Quebec. Global Biogeochemical Cycles 7: 247-257.

Kroetsch, D.J., Geng, X., Chang, S.X., and Saurette, D.D. 2011. Organic soils of Canada: Part 1. Wetland Organic soils. Can. J. Soil Sci. 91: 807-822.

Lavkulich, L.M. and Arocena, J.M. 2011. Luvisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 781-806.

Moore, T. R. 1976. Sesquioxide-cemented soil horizons in northern Quebec: their distribution, properties and genesis. Can. J. Soil Sci. 56: 333-344.

Moore, T.R., 1986. The spatial variability of soil properties and its application to archaeology. Canadian Geographer 30: 80-82.

Moore, T.R., and Denton, D. 1988. The use of soils in archaeological investigations in northern Quebec. pp. 25-37 in J. Bintliff, D.A. Davidson and E.G. Grant (eds.) Conceptual Issues in Environmental Archaeology. Edinburgh University Press.

Parent, L.E., Millette, J.A., and Mehuys, G.R. 1982. Subsidence and erosion of a Histosol. Soil Sci. Soc. Am. J., 46: 404-408.

Richard, P.J.H., and Occhietti, S. 2005. 14C chronology for ice retreat and inception of Champlain Sea in the St. Lawrence Lowlands, Canada. Quaternary Research 63: 353-358.

Riley, J. 2013. The Once and Future Great Lakes Country: an Ecological History. McGill-Queen's University Press.

Sanborn, P., Lamontagne, L., and Hendershot, P. 2011. Podzolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 843-880.

Smith, C.A.S., Webb, K.T., Kenney, E., Anderson, A., and Kroetsch, D. 2011. Brunisolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 695-717.

Soil Classification Working Group. 1998. The Canadian System of Soil Classification. Agric. And Agri-Food Canada. Publ. 1646 (Revised). 187 pp. Available at:

http://sis.agr.gc.ca/cansis/publications/manuals/1998-cssc-ed3/cssc3_manual.pdf

Tarnocai, C. and Bockheim, J.G. 2011. Cryosolic soils of Canada: Genesis, distribution and classification. Can. J. Soil Sci. 91: 749-762.

Wironen, M., and Moore, T.R. 2006. Exotic earthworm invasion of an old-growth forest in southern Quebec and effects on soil carbon and nitrogen. Can. J. For. Res. 36: 845-854.

Thiffault, E., Hannam, K.D., Paré, D., Titus, B.D., Hazlett, P.W., Maynard, D.G, and Brais, S. 2011. Effects of forest biomass harvesting on soil productivity in boreal and temperate forests – A review. Environ. Rev. 19: 278–309.

About the Author



Tim Moore (licensed under a CC-BY-NC-ND license)

My mother's family were market gardeners and she had a 'green thumb', so it is not surprising that I was exposed to soils early in life. I was professionally attracted to soils because they represent the interaction of many aspects of the environment and human activity. My research focuses on the relationships between soil and the environment, particularly the biogeochemical role of soils in the regulation of fluxes of gases, nutrients and elements between the soil and the atmosphere, the biosphere and the hydrosphere, and the impact of human activity on these fluxes. I am particularly interested in peatlands and wetlands and the controls on the cycling of carbon from these systems.

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2.5: Soils of the Atlantic Provinces

Brandon Heung, Kevin Keys, David Burton, and Derek Lynch

Learning Objectives

On completion of this chapter, students will be able to:

1. Understand the factors responsible for the distribution of soils in the Atlantic Provinces
2. Identify the diagnostic soil horizons associated with soil orders and great groups (according to the Canadian System of Soil Classification) of the Atlantic Provinces
3. Relate the major land uses of the region with the properties of the soils

INTRODUCTION

Seven of Canada's ten soil orders can be found in Atlantic Canada, but high annual precipitation rates and a preponderance of relatively coarse, acidic glacial till parent material has led to Podzols being the dominant soil in the region. In many cases, these soils are also very stony and/or shallow to bedrock, which further limits their agricultural potential. As a result, the Atlantic provinces (New Brunswick, Nova Scotia, Prince Edward Island, and Newfoundland and Labrador) account for only about 1.5% of Canada's agricultural land base (Statistics Canada, 2016), with forests covering the majority of land in the region. Although each province has locally important agricultural areas, only in Prince Edward Island does agricultural land dominate over forest land.

FACTORS CONTROLLING THE DISTRIBUTION OF SOILS IN THE ATLANTIC PROVINCES

Using the National Ecological Framework for Canada, the Atlantic provinces can be divided into four ecozones (Ecological Stratification Working Group, 1996; Figure 13.1), which describe regions that share similar geology, vegetation, soil, and climate. New Brunswick, Nova Scotia, and Prince Edward Island are all situated within the southern-most Atlantic Maritime ecozone, with the island of Newfoundland fully covered by the Boreal Shield ecozone and Labrador primarily in the Taiga Shield ecozone (with inclusions of the Boreal Shield ecozone in the southeast and the Arctic Cordillera ecozone in the north).

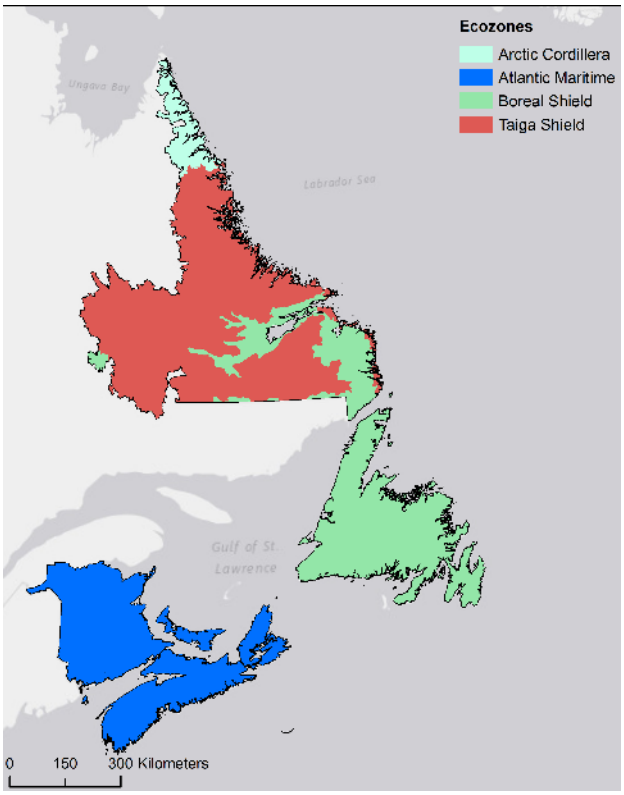


Figure 13.1. Ecozone map for the Atlantic provinces. The map is based on A National Ecological Framework for Canada (Agriculture and Agri-Food Canada, 1996). © Brandon Heung, Dalhousie University; licensed under a CC BY (Attribution) license.

Atlantic Maritime Ecozone

The climate of the Atlantic Maritime ecozone is largely influenced by its proximity to the Atlantic Ocean, which results in moist conditions and moderate temperatures (Figure 13.2). This ecozone receives an average 1250 mm of precipitation annually (range 970 to 1840 mm) and is the warmest in the region with an average annual temperature of 7°C (range 1 to 8°C) (Fick and Hijmans, 2017). Land cover is dominated by forests ranging from pure softwood, to mixedwood, to pure hardwood. Where site conditions do not limit growth, climax species include red spruce (*Picea rubens* Sarg.), eastern hemlock (*Tsuga canadensis* (L.) Carriere), eastern white pine (*Pinus strobus* L.), sugar maple (*Acer saccharum* Marsh), and yellow birch (*Betula alleghaniensis* Britton). In coastal areas and in higher elevations, balsam fir (*Abes balsamea* (L.) Mill.) and white spruce (*Picea glauca* (Moench) Voss) are more common. Bedrock geology is largely comprised of Paleozoic era formations (Figure 13.3) and are primarily sedimentary in nature; however, inclusions of volcanic and intrusive igneous rock are also common (Wheeler et al., 1997). Surficial geology is dominated by a variety of glacial till deposits, with glacio-fluvial deposits and rocky areas locally important (Figure 13.3).

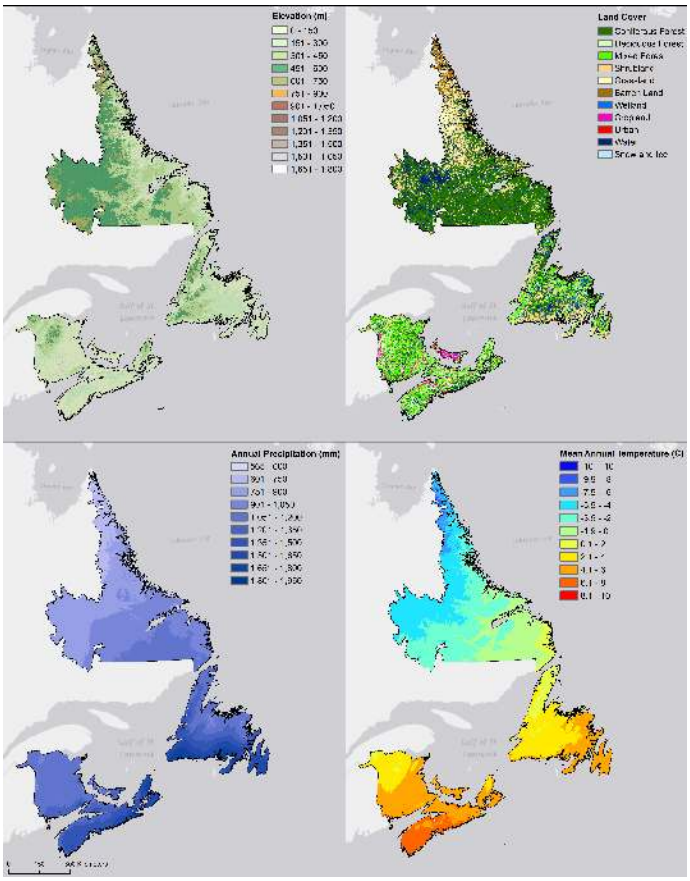


Figure 13.2. Maps of elevation, land cover, mean annual precipitation, and mean annual temperature for the Atlantic provinces. The elevation map is derived from the ALOS Global Digital Surface Model (Earth Observation Research Centre and Japan Aerospace Exploration Agency, 2019); the land cover map is derived from the North American Land Cover dataset (North American Land Change Monitoring System, 2019); and the mean annual precipitation and mean annual temperature maps are derived from the WorldClim version 2.0 dataset (Fick and Hijmans, 2017). © Brandon Heung, Dalhousie University; licensed under a CC BY (Attribution) license.

Figure 13.3. Maps of bedrock geology and surficial geology for the Atlantic provinces. The bedrock geology map is derived from the Geological Map of Canada (Wheeler et al., 1997) and the surficial geology map is derived from the Surficial Materials Map of Canada (Fulton, 1995). © Brandon Heung, Dalhousie University; licensed under a CC BY (Attribution) license.

In terms of soils, Humo-Ferric Podzols with diagnostic Bf horizons are the most common and are typically formed at lower elevations with milder climates (Figs. 13.4 and 13.5). However, Ferro-Humic Podzols with diagnostic Bhf horizons are also frequently found, especially where cooler temperatures and higher moisture levels have contributed to greater soil organic matter

accumulation. In addition, Gray Luvisols can be locally important and are somewhat unique in that they often exhibit a bisequal profile development where a Podzolic Bf forms over a clay-enriched Bt horizon (Figure 13.5). Dystric Brunisols with Bfj horizons can sometimes be found in less well-developed soils, while richer Sombric Brunisols with Ah and Bm horizons are locally important in parts of New Brunswick (Figure 13.4). Finally, with high precipitation and variable topography, poorly drained Gleysols (all great groups) and Organic soils (mainly Mesisols and Humisols) can be found scattered across the ecozone.

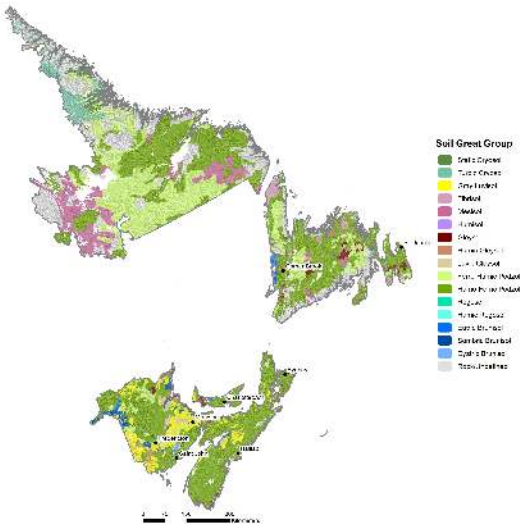


Figure 13.4. Soil great group map for the Atlantic provinces. Map is a reproduction of an official work published by the Government of Canada and is based on the 1:1,000,000 Soil Landscapes of Canada map. © Darrel Cerkowniak, Agriculture and Agri-Food Canada, is licensed under a CC BY (Attribution) license.

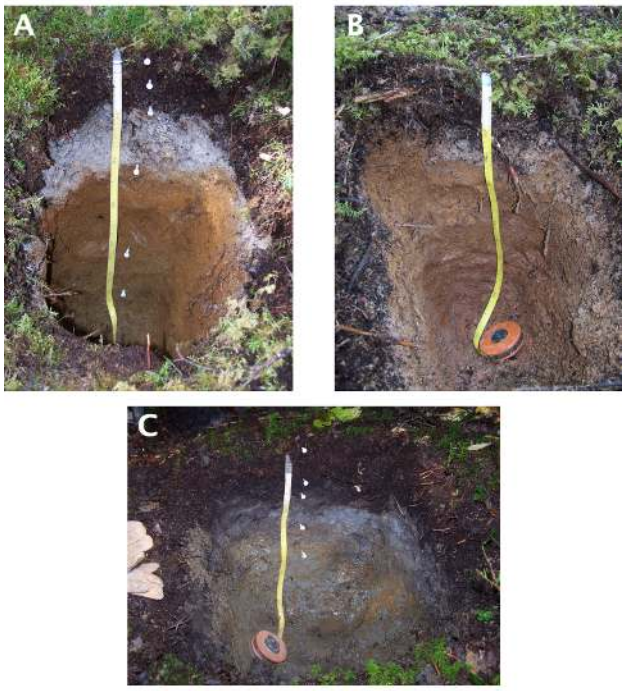


Figure 13.5. An Orthic Humo-Ferric Podzol (a), Gleyed Podzolic Gray Luvisol (b), and Orthic Gleysol (c) from Nova Scotia. © Kevin Keys; licensed under a CC BY (Attribution) license.

Boreal Shield Ecozone

The Boreal Shield Ecozone is by far the largest ecozone in Canada and extends from northern Saskatchewan to all of Newfoundland and southeastern Labrador (Figure 13.1). Although much of this ecozone is associated with a continental climate, proximity to the Atlantic Ocean results in generally warmer and moister conditions in Newfoundland and Labrador. This component of the ecozone receives on average 1230 mm of precipitation annually (range 800 to 1800 mm)—similar to the Atlantic

Maritime ecozone (Fick and Hijmans, 2017; Figure 13.2). Precipitation is highest along the southern coast of Newfoundland and decreases along a northwestern track towards Labrador. Mean annual temperature is about 2°C with a range of -5 to 6°C (Fick and Hijmans, 2017).

Unlike the Atlantic Maritime ecozone, the Boreal Shield has far greater variability in terms of the type and age of the bedrock (Figure 13.3). The eastern-most region of Newfoundland is dominated by old sedimentary and volcanic rock from the Precambrian era that transitions westward to younger Paleozoic rock. In Labrador and parts of northern Newfoundland, bedrock consists mainly of Precambrian metamorphic and intrusive rock. As with other ecozones, surficial geology is mainly comprised of glacial till deposits of variable thickness, but with a larger proportion of thin veneer tills and exposed rock (Figure 13.3). In addition, glacio-fluvial deposits can be significant in some areas, such as with terrace landforms adjacent to the Churchill River in eastern Labrador. Land cover is dominated by boreal softwood forests [black spruce (*Picea mariana* (Mill.) Britton, Sterns & Poggenb.), balsam fir, white spruce, eastern larch (*Larix laricina* (Du Roi) K. Koch)] with some trembling aspen (*Populus tremuloides* Michx.) and white birch (*Betula papyrifera* Marshall). Newfoundland and Labrador also has extensive coverage of bogs, fens, and heathlands.

Soils in the Boreal Shield ecozone are dominated by moderate to coarse-textured Humo-Ferric and Ferro-Humic Podzols (Figures 13.4 and 13.6). Humic Podzols (with diagnostic Bh horizons) are also found in some areas. Gleyed Subgroups of these soils are quite common, with cemented Subgroups (e.g., Orsteins with Bfc or Bhfc horizons) also frequently found along the coast and in coarser-textured soils (Woodrow and Heringa, 1987). Fibrisols and Mesisols are associated with the widespread bog and fen landforms. High moisture levels, variable topography, and near-surface bedrock also promote development of poorly drained Gleysols which can be found scattered across this ecozone, as well as in concentrated areas in eastern and central Newfoundland (Heringa, 1981; Figure 13.4).



Figure 13.6. An Orthic Humo-Ferric Podzol (a) and Orthic Ferro-Humic Podzol (b) from southern Labrador. © Kevin Keys; licensed under a CC BY (Attribution) license.

Taiga Shield and Arctic Cordillera Ecozones

The Taiga Shield ecozone is the main ecozone in Labrador and is the largest by area in the Atlantic region (Figure 13.1). At these latitudes, moderating effects of the Atlantic Ocean are offset by cold water currents from the north resulting in short, cool summers and long, cold winters (Ecological Stratification Working Group, 1996). The mean annual temperature is highest (1.5°C) along the eastern coast and decreases inland to -7°C. Average annual precipitation is 910 mm and ranges from 650 to 1250 mm. North of the Taiga Shield is the Arctic Cordillera ecozone (Figure 1). Here the average annual temperature is only -6°C (range -10°C to 0°C). In addition to being the coldest ecozone in the region, the Arctic Cordillera is the driest with an average annual precipitation of only 680 mm (range 560 to 910 mm). Both these ecozones are mainly underlain by Precambrian metamorphic and intrusive igneous rocks (gneiss and granite). Surficial geology is comprised of glacial till deposits of variable thickness, but with a large proportion of thin veneer tills and exposed rock in eastern and northern areas (Figure 13.3).

Land cover in the Taiga Shield ecozone is dominated by black spruce forests that can grade into stunted tuckamore in more exposed areas. At higher elevations and along the northern range of the ecozone, trees reach their habitat limits and the landscape transitions to alpine meadow and/or tundra vegetation (Ecological Stratification Working Group, 1996). Moving north, the Arctic Cordillera ecozone transitions to a landscape consisting of mainly tundra and exposed bedrock.

Humo-Ferric and Ferro-Humic Podzols are the dominant soils in the Taiga Shield ecozone, with significant inclusions of Organic Mesisols. In some areas, near-surface permafrost also results in the development of Organic Cryosols, while thin Organic Folisols can be found in association with bedrock exposures. Dystric Brunisols can also be found where soil development has been slowed by climate and/or age of parent material (e.g., younger colluvium deposits). Moving northward, near-surface permafrost becomes more extensive leading to development of both Organic and Turbic Cryosols. Turbic Cryosols show evidence of cryoturbation (y horizon suffix) where mineral horizons become disrupted, mixed, and/or broken through regular frost churning.

SOILS AND LANDUSE

Agriculture

Throughout Atlantic Canada, the capability of agricultural lands is evaluated using the Canada Land Inventory (CLI; see Chapter 8, Canada Land Inventory). The CLI is based on a rating system ranging from 1 to 7 whereby soils are categorized based on their potential and limitations for agricultural use. In the CLI rankings, Classes 1 through 4 are considered to be capable of sustained use for cultivated field crops. In Class 1 land there are no significant constraints to crop growth while Class 2 lands experience moderate limitations and therefore require moderate conservation practices. In Class 3 and 4 lands, the range of crops is restricted and special management measures are needed to conserve the soil. Classes 5 and 6 are only suitable for forage production and are most commonly used for grazing of livestock.

Within this region, there are no Class 1 lands (no significant limitations); furthermore, the distribution of Class 2 lands is limited to Prince Edward Island, northwestern New Brunswick, and the Annapolis Valley of Nova Scotia (Figure 13.7). Class 2 lands experience limitations due to climate in Prince Edward Island and Nova Scotia and low fertility in New Brunswick (Figure 13.8). Class 2 lands occupy approximately 588,000 ha of Atlantic Canada while Class 3 and Class 4 lands occupy approximately 2,275,000 ha and 2,507,000 ha, respectively. The majority of land in Nova Scotia and Newfoundland are Class 7 and are therefore not capable for arable culture or permanent pasture.

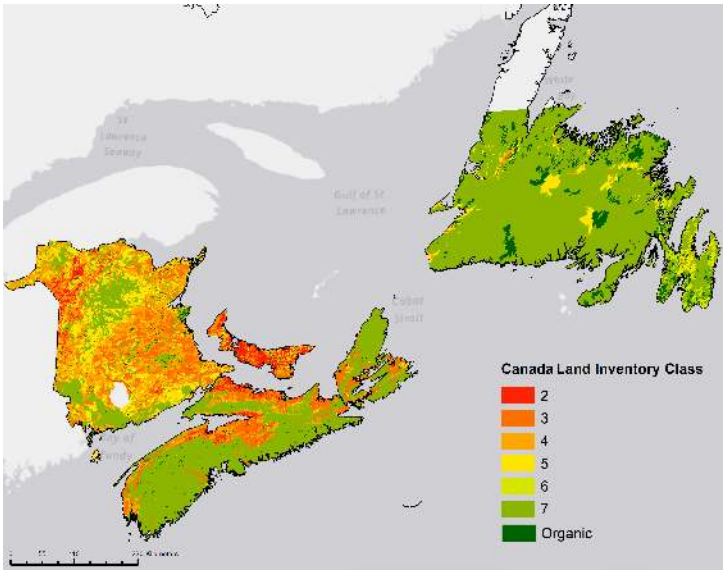


Figure 13.7. Canada Land Inventory Classes for agriculture in Atlantic Canada (Agriculture and Agri-Food Canada, 2013). Classes range from 1 to 7, where Class 2 lands represent moderate limitations whereby moderate conservation practices are required and Class 7 lands represent no capability for arable culture or permanent pasture. © Agriculture and Agri-Food Canada; used with permission and licensed under the Open Government Licence – Canada.https://open.canada.ca/en/open-government-licence-canada

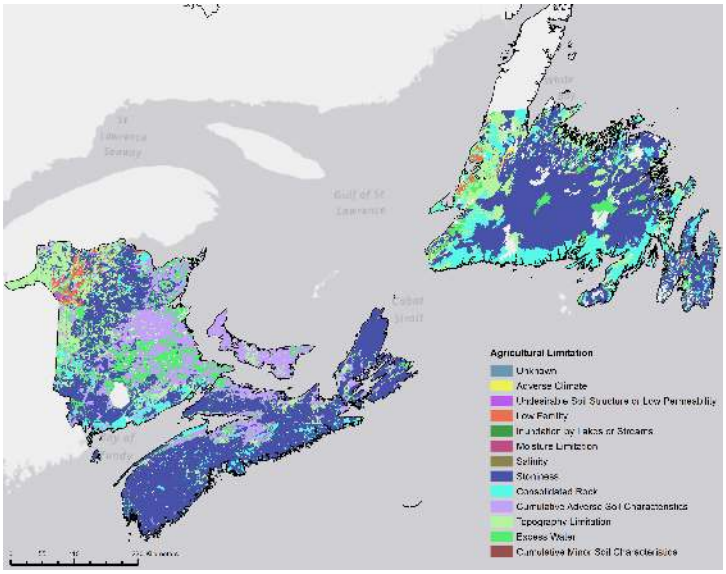


Figure 13.8. Agricultural limitations for Atlantic Canada based on the Canada Land Inventory (Agriculture and Agri-Food Canada, 2013). © Agriculture and Agri-Food Canada; used with permission and licensed under the Open Government Licence – Canada.https://open.canada.ca/en/open-government-licence-canada

In 2011, Statistics Canada reported 1,063,747 ha of agricultural land in Atlantic Canada—a decrease of 12.8% from 30 years earlier (Clearwater et al., 2016). As of 2011, approximately 40% of the land was cropland; 10% was pasture; and the remaining 50% of the land included woodlots, wetlands and farmyards. From 1981 to 2011 there was a 56% decline in area for pasture and a 7% decline in other land uses with a corresponding 8% increase in cropland. Lower disturbance land management has declined (pasture -56%, forage -4%, and cereal grains -37%) and are being replaced by higher intensity cropping systems characterized by higher disturbance and/or low crop residue return to soil, (corn +145%, potatoes, +18%, soybeans increased from no reported area in 1981 to over 25,000 hectares in 2011). This transition from low disturbance, land use (pasture, woodland, wetland) and the reduced frequency of forages to more intensive cropland has resulted in significant impacts on soil health and soil microbial composition (Mann et al., 2019).

Soil organic matter is of critical importance in the function and resilience of the predominantly coarse textured soils of Atlantic Canada and is a primary indicator of soil health. The percentage of cropland in Atlantic Canada experiencing moderate (-25 to -90 kg ha^{-1} y^{-1}) to large (> – 90 kg ha^{-1} y^{-1}) decreases in soil organic carbon increased from 61% in 1981 to 83% in 2011. The decline in soil organic matter has been highlighted as a major concern in Atlantic Canada, particularly Prince Edward Island (Nyiraneza et al., 2017), where a province-wide network for soil quality monitoring as affected by landuse has been in place since 1998. Nitrogen additions have also increased to support the more intensive cropping practices. From 1981 to 2011 the percentage of crop land having very low to low residual soil nitrogen had decreased by 85% while land rated with a high and very high levels of residual soil nitrogen increased by 138%.

In 2016, an assessment of soil health was initiated by the Atlantic Soil Health Lab at Dalhousie University. The lab examined the role of land management, as reflected in cropping system, and pedogenic origin, as reflected in soil order, in influencing measures of soil health using the Cornell Soil Health Assessment framework. This protocol for assessing soil health was developed in New York state and combines the measurement of soil physical, chemical, and biological properties selected based on their relevance, sensitivity, consistency, and cost-effectiveness (Moebius-Clune et al., 2016). Land management explained a greater percentage of the variability in soil health parameters (average R^2 = 23%) than did soil order (average R^2 = 8%). In particular, parameters relating to organic carbon quantity or quality were more highly correlated to land management (R^2 = 9 to 55%) than soil order (R^2 = 2 to 12%). These results highlight the importance of soil management in impacting soil health and soil function.

Forestry

Forests in the Atlantic region have evolved in relation to local climate and soil/site influences. However, more temperate conditions in the Atlantic Maritime ecozone means that soils there have more influence on tree species distribution and potential productivity than is the case in the Boreal Shield and Taiga Shield ecozones where climate is the main driver. For example, sandy and/or very

gravelly soils in the Atlantic Maritime ecozone tend to be associated with black spruce and pine species that are better adapted to dry, poor soils. These soils are typically Humo-Ferric Podzols and can sometimes be cemented. More loamy soils with higher fertility support a greater mix of tree species and tend to have a higher proportion of Ferro-Humic Podzols. Gray Luvisols have the potential to be very productive forest soils, but poor aeration and/or drainage problems often restrict potential growth and species mix. Shallow or stony soils in all ecozones can limit potential productivity and increase windthrow hazard, while surface boulders can reduce machine trafficability (Figure 13.9). Periodic windthrow disturbance in deeper soils can lead to mixing of soil horizons and create buried horizons which influence soil development (Figure 13.10).



Figure 13.9. Very stony soils found under a windthrown tree in Nova Scotia (a). Surface boulders in a burned black spruce forest in central Labrador (b). © Kevin Keys; licensed under a CC BY (Attribution) license.



Figure 13.10. Buried Ae and Bf horizons in a Humo-Ferric Podzol caused by tree windthrow disturbance (arborturbation). © Kevin Keys; licensed under a CC BY (Attribution) license.



Field following the soybean harvest on dykeland soils near Truro, Nova Scotia (photo D. Lynch; licensed under a CC BY (Attribution) license)

In the 1700's Acadian French settlers, through the construction of dykes, land forming and a drainage network combined with hand-made aboiteaux (or sluice gates) to evacuate the drainage water, generated extensive (over 33,000 ha) farmland landscapes from reclaimed tidal salt marshes in Atlantic Canada. Dominated by the 'Acadia' soil series (Regosol order), these new soils show little horizon development and are characteristically flat, silt loam to silty clay loams, imperfectly drained, and subject to periodic flooding and saturated conditions. As relatively fertile soils, Acadia soils have been used for centuries for forage production, and also more recently for silage corn and soybean production. In 2012, the over 1300 ha of reclaimed marshland landscape at Grand Pre in the Minas Basin in Nova Scotia, the best example of such a historically transformed marshland or 'polder' landscape in North America, was recognized as a UNESCO World Heritage site (https://whc.unesco.org/en/list/1404).

SUGGESTED READING

Detailed information on the major soil orders and great groups for the Atlantic Region may be found in the following references:

Bedard-Haughn, A. 2011. Gleysolic soils of Canada: Genesis, distribution, and classification. Can. J. Soil Sci. 91: 763-779.

Kroetsch, D.J., Geng, X., Chang, S.X. and Saurette, D.D. 2011. Organic soils of Canada: Part 1. Wetland Organic soils. Can. J. Soil Sci. 91: 807-822.

Lavkulich, L.M. and Arocena, J.M. 2011. Luvisolic soils of Canada: Genesis, distribution, and classification. Can. J. Soil Sci. 91: 781-806.

Sanborn, P., Lamontagne, L., Hendershot, W. Podzolic soils of Canada: Genesis, distribution, and classification. Can. J. Soil Sci. 91: 843-880.

Tarnocai, C. and Bockheim, J. 2011. Cryosolic soils of Canada: Genesis, distribution, and classification. Can. J. Soil Sci. 91: 749-762.

REFERENCES

Agriculture and Agri-Food Canada. 2013. ISO 19131 Canada Land Inventory (CLI) – Data Product Specification. Science and Technology Branch, Agriculture and Agri-Food Canada. pp.28.

Canada Centre for Remote Sensing (CCRS), Earth Sciences Sector, Natural Resources Canada, Comisión Nacional para el Conocimiento y Uso de la Biodiversidad (CONABIO), Comisión Nacional Forestal (CONAFOR), Insituto Nacional de Estadística y Geografía (INEGI), and U.S. Geological Survey (USGS). 2015. North American Land Change Monitoring System (NALCMS) Collection, Commission for Environmental Cooperation.

Clayton, J.S., Ehrlich, W.A., Cann, D.B., Day, J.H., and Marshall, I.B. 1977. Soils of Canada Volume 1. Research Branch, Canada Department of Agriculture, Ottawa. pp. 243.

Clearwater, R.L., Martin, T., and Hoppe, T. 2016. Environmental Sustainability of Canadian Agriculture (Report #4). Agri-Environmental Indicators Report Series. Agriculture and Agri-Food Canada, Ottawa, ON.

Colpitts, M.C., Fahmy, S.H., MacDougall, J.E., Ng, T.T.M., McInnis, B.G. and Zelazny, V.F. 1995. Forest Soils of New Brunswick. Agriculture and Agri-Food Canada & Natural Resources Canada. pp. 51.

Earth Observation Research Center and Japan Aerospace Exploration Agency. 2019. ALOS Global Digital Surface Model (DSM), ALOS World 3D-30m Version 2.2 Product Description. pp. 12.

Ecological Stratification Working Group. 1996. A National Ecological Framework for Canada. Research Branch, Agriculture and Agri-Food Canada, Ottawa. pp. 125.

Fick, S.E. and Hijmans, R.J. 2017. WorldClim 2: new 1-km spatial resolution climate surfaces for global land areas. Int. J. Climatol. 37: 4302-4315.

Fulton, R.J. 1995. Surficial Materials of Canada. Geological Survey of Canada, Natural Resources Canada, “A” Series Map 1880A.

Heringa, P.K. 1981. Soils of the Avalon Peninsula, Newfoundland. Report No. 3, Agriculture Canada. pp. 117.

Mann, C., Lynch, D.H., Fillmore, S., and Mills, A. 2019. Relationships between field management, soil health and microbial community composition. Appl. Soil Ecol. 144: 12-21.

Moebius-Clune, B.N., Moebius-Clune, D.J., Gugino, B.K., Idowu, O.J., Schindelbeck, R.R., Ristow, A.J., van Es, H.M., Thies, J.E., Shayler, H.A., McBride, M.B., Kurtz, K.S.M., Wolfe, D.W., and Abawi, G.S. 2017. Comprehensive Assessment of Soil Health: The Cornell Framework, 3rd ed. Cornell University, Ithaca, NY.

Nyiraneza, J., Thompson, B., Geng, X., He, J.X., Jiang, Y.F., Fillmore, S., and Stiles, K. 2017. Changes in soil organic matter over 18 yr in Prince Edward Island, Canada. Can. J. Soil Sci. 97: 745-756.

Rees, H.W., Fahmy, S.H., Wang, C. and Wells, R.E. 2005. Soils of Central and Northern New Brunswick. New Brunswick Soil Survey Report No. 12, Agriculture and Agri-Food Canada. pp. 137.

Soil Classification Working Group. 1998. The Canadian System of Soil Classification Version 3. Agriculture and Agri-Food Canada. pp. 187.

Soil Landscapes of Canada Working Group. 2010. Soil Landscapes of Canada Version 3.2. Agriculture and Agri-Food Canada.

Statistics Canada. 2016. Table 32-10-0153-01: Total area of farms and use of farm land, historical data. https://doi.org/10.25318/3210015301-eng.

Wheeler, J.O., Hoffman, P.F., Card, K.D., Davidson, A., Sanfrd, B.V., Okulitch, A.V. and Roest, W.R. 1996. Geological Map of Canada. Geological Survey of Canada, Natural Resources Canada, “A” Series Map 1860A.

Woodrow E.F. and Heringa, P.K. 1987. Pedoclimatic Zones of the Island of Newfoundland. Report No. 32, Canada Soil Survey. pp. 12.

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Dr. Brandon Heung is an assistant professor in soil science and geographical information systems (GIS) in the Department of Plant, Food, and Environmental Sciences. His research is in the area of pedometrics—a subdiscipline of soil science that integrates soil science, GIS, remote sensing, spatial statistics, and machine-learning to better understand that spatial and temporal patterns of soil. He has been developing digital soil maps for applications in agriculture and forested systems and he is also interested in understanding how predictive digital soil models may be used to address climate change and food security issues.

Kevin Keys, Nova Scotia Department of Lands and Forestry



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I have spent many years assessing, classifying, and interpreting soils across Canada from British Columbia to Labrador. However, most of my experience has been in Nova Scotia where I helped develop a forest ecosystem classification system focusing on soils and ecosite components. I am currently a soils specialist with the Nova Scotia Department of Lands and Forestry and I lead the Site Productivity Group within the Research and Planning Section. It's my job to conduct research and develop management tools and guidelines that help promote forest soil health and sustainable forest management in the province.

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CHAPTER OVERVIEW

3: Digging Deeper

3.1: Soil Mineralogy

3.2: Soil Health and Management

3.3: Soil Reclamation and Remediation of Disturbed Lands

3.4: Digital Soil Mapping

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3.1: Soil Mineralogy

Jim Warren and Graeme Spiers

Learning Objectives

On completion of this chapter, students should be able to:

1. Identify common mineral phases found in soil
2. Explain the difference between primary and secondary minerals
3. Describe the influence of continental glaciation on mineral distributions in Canadian soils
4. Describe the relationship of between particle size and soil minerals
5. Define isomorphous substitution and the basis for cation exchange capacity
6. Explain differences in the phyllosilicate composition of soils between eastern and western Canada

INTRODUCTION

Soil mineralogy is "concerned with the inorganic minerals found in the pedosphere and to the depth of weathering" (Finkl, 1983). Although soil mineralogy draws heavily on the disciplines of mineralogy, geology, inorganic chemistry and crystallography—which are scientific disciplines unto themselves—the reader is not required to have much prior knowledge in these areas to understand the information presented herein. The general approach taken here is based on the textural fractions of soil (sand, silt and clay) and the mineral components contained therein.

Importance of the Soil Mineral Fraction

Most soils are composed of inorganic mineral materials that comprise the overwhelming bulk of a soil. Well-structured soils are composed of about 50% solids of which most is mineral materials (Figure 14.1), with organic soils (>30% organic matter) being the obvious exception.

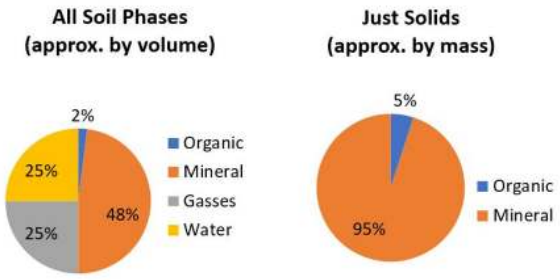


Figure 14.1. The general bulk composition of well-structured soil and soil solids. © Jim Warren is licensed under a CC BY (Attribution) license.

There are about 4,500 to 5,000 named minerals identified on this planet (MSA, 1997–2020; MSA, 2004–2020; Mindat, 2020), with most being either rare or found in isolated pockets (e.g., economic mineral deposits) deep in the Earth's crust. Minerals are naturally occurring inorganic crystals with a definable range of chemical composition. For example, by far the most common terrestrial mineral found in Canadian soils is α-quartz (α-SiO_2); pronounced "alpha-quartz" where alpha refers to the crystal form. Rock fragments are also found in soil but these differ from minerals in that rocks are "mineral composites" made up of physical mixtures of several individual minerals. For example, granite is an igneous rock making up the vast majority of the Canadian Shield. Granite is commonly composed of a mixture of four types of minerals: α-quartz, feldspars, amphiboles and micas which, due to the grinding and mixing actions in the glaciers, are also common minerals found in our soils. Rock fragments are only found among the larger soil particles (i.e., sand-size and larger) because as rocks are broken down into increasingly smaller fragments, their physical breakdown typically occurs along boundaries between adjacent minerals, which are the weakest points within rocks, leaving the individual discrete minerals as the smaller fragments.

Most minerals found in soil are of the common variety, meaning they are ubiquitous and present in large quantities usually with little or no economic value. The types of minerals in soil represent only a handful (<100) compared to all the known minerals;

however, their behaviour in soil as they interact with the soil solution (soil water), the soil atmosphere and soil organic matter have a huge bearing on the chemical characteristics and nature of our soils.

Soil minerals are the natural source of plant nutrients that are released slowly with time during chemical weathering. All plants require a minimum of 17 nutrient elements to complete their life cycle (see Chapter 7). With the exception of C, H, and O, which plants obtain from air and water, plants derive the remaining 14 elements (N, P, K, Ca, S, Mg, Fe, B, Cl, Mn, Zn, Cu, Mo, and Ni) mainly from the soil itself and through the addition of fertilizers, manures, and other amendments (Parikh and James, 2012; Singh and Schulze, 2015). An additional four elements (Na, Si, V, and Se) are essential to the growth of some plants (Havlin et al., 2005). Soil minerals also contribute to a soil's ability to retain plant nutrients through the processes of cation and anion exchanges.

SOIL PARENT MATERIALS

Canadian soils are very young compared to those in the tropics, which have never been reworked by glaciers (see Chapter 2). Canada contains more glaciated terrain than any other country in the world (Rutter, 2015) and almost all Canadian soils are developed on Late Wisconsin glacial sediments that were laid down between about 5,000 to 18,000 years ago. The only exceptions are some small areas such as the Cypress Hills region of southern Alberta and Saskatchewan; non-glaciated areas of the northern Yukon and small parts of the western Northwest Territories which were not covered with ice during Late Wisconsin glaciation (see Chapter 2). By comparison soils such as those in the southern United States, Australia, Africa, and South America are several million years old. Therefore, Canadian soils are composed of inherited minerals derived mostly from the underlying bedrock and through transport and deposition by glacial ice during ice advance or by glacial melt waters and wind during deglaciation.

PRIMARY *vs.* SECONDARY MINERALS

The Earth's crust may be described in the very broad context as being composed of two main lithologies: the oceanic crust composed mostly of mafic and ultramafic (Fe and Mg rich) rocks; and the continental crust which contains more granitic, metamorphic and sedimentary rocks. Since soils are located in terrestrial areas they are developed primarily from the continental crust rocks, which in the Canadian context is about 85% igneous and metamorphic rocks and 15% sedimentary rocks.

Minerals derived directly from geological and geochemical processes in the Earth's crust are referred to as primary minerals. Examination of mineral surfaces with a microscope may reveal some changes due to biogeochemical weathering (Spiers, et al., 1989), but most primary minerals in Canadian soils have been only minimally altered by chemical actions in the last 5,000–18,000 years.

Once brought to the Earth's surface most primary minerals are not chemically stable. Although they persist in our soils, given enough geological time they will eventually succumb to the biogeochemical processes of the environment and gradually release their elemental components to the environment. Primary minerals weather and dissolve at different rates. Generally, minerals containing Na, and K dissolve faster than those containing Ca and Mg, which in turn dissolve faster than minerals composed of Si, Al, and Fe. Iron is a slightly special case as it is the only major element with two valences (Fe^{2+} and Fe^{3+}) with solubility differing greatly depending on the chemical environment. Under reducing (absence of oxygen) conditions, Fe tends to be soluble and present as the Fe^{2+} ion. Under oxidizing conditions, Fe^{2+} reacts with oxygen (oxidized) to form Fe^{3+} ions which in turn form insoluble oxyhydroxide minerals (effectively the composition of rust). Other minor elements (e.g., Mn, As, U, etc.) are transformed similarly. The result is that as weathering proceeds soils become generally depleted in Na, K, Ca and Mg which are leached from the soils while becoming enriched in Si, Al, and Fe^{3+} (see Figure 14.2). This also explains why the oceans contain elevated concentrations of soluble salts (Na, K, Cl and SO_4). Carbonate minerals (principally $CaCO_3$) are formed or dissolved depending on the pH of their bearing solution. They precipitate under alkaline conditions and are dissolved under acid conditions. Atmospheric CO_2 reacts with the water to produce a weak acid (H_2CO_3) that dissolves carbonate (and other rock types). Deeper groundwater tends to be alkaline thus favouring precipitation of carbonate typically found deeper in soil profiles (C horizons).

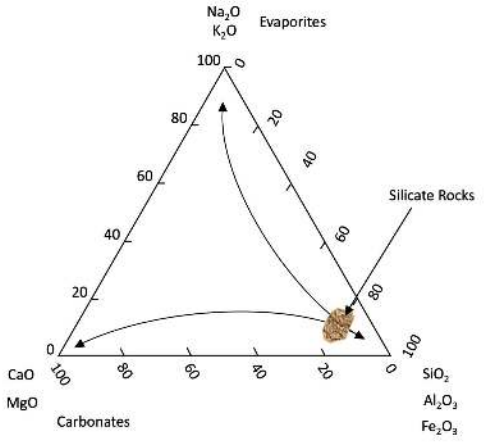


Figure 14.2. The general trend of chemical weathering in soil. Adapted from Warren (1994); adapted by Lewis Fausak. © Jim Warren adapted by Lewis Fausak is licensed under a CC BY (Attribution) license.

While primary minerals form by geological processes, secondary minerals are a by-product of chemical weathering formed either through the partial chemical alteration of primary minerals or through chemical precipitation from the reaction of components dissolved from other minerals. Most secondary minerals are hydrated aluminosilicates or Fe and Al oxyhydroxides that are stable in soils. Most hydrated aluminosilicates are classified as phyllosilicates or clay minerals which are the most abundant secondary minerals in Canadian soils. These are very fine-grained sheet-like particles with a very high surface area to mass ratio referred to as specific surface area (SSA). Since surface area is inversely related to particle diameter, the clay-size fraction accounts by far for the largest mineral portion of the available surface area associated with soil minerals (Chapter 5, Table 5.1).

Clay minerals are also the most chemically reactive inorganic phases in soils and both the amount and type of clay mineral influence many soil properties including:

- porosity and pore size;
- swelling behaviour;
- compressibility and compactability;
- sorption capacity;
- exchange capacity; and,
- hydraulic conductivity.

Can You Dig It!

Although secondary minerals are formed and are stable in soils at the surface of the Earth, the vast majority secondary minerals found in Canadian soils were formed in the distant past, prior to Late Wisconsin glaciation. Most soil minerals re-deposited during the last glaciation were derived from surface sediments formed and deposited, likely repeatedly, during the millennia prior to the last glaciation. Although secondary minerals continue to form in our soils in response to chemical weathering, the total amounts produced since the last glacial event are very small compared to the amounts re-deposited during many repeated glaciations in the past.

MINERALS vs PARTICLE SIZE

The average person with 20:20 vision can resolve individual particles as small as about 30 µm (0.03 mm) with the unaided eye. This is the approximate division (0.05 mm) between sand and silt-size soil particles (Figure 14.3). By comparison, the thickness of human hair is in the range of 0.1 to 0.04 mm. This means that the average individual with reasonably good eye-sight can easily distinguish individual sand-size particles with the naked eye (without magnification) but cannot differentiate between most individual silt particles and definitely cannot resolve clay-size particles. Examination of silt-size particles requires magnification through use of a light microscope while clay-size particles can only be resolved using electron microscopy (Smart and Tovey, 1981) or atomic force microscopy (Liu, 2003).

The distribution of soil minerals varies with particle size (Figure 14.3). Based on grain size fractions as defined in the Canadian System of Soil Classification (SCWG 1998), the sand-size fraction (particles with diameter in the range of 2-0.050 mm) are dominated by primary minerals. The primary minerals are mainly quartz, feldspars and other primary silicate minerals while the clay-size fraction (<0.002 mm diameter) is dominated by secondary phyllosilicates and iron and aluminum oxyhydroxides. The silt-size fraction (0.05-0.002 mm diameter) is intermediate in size between sand and clay and typically contains a mixture of both primary and secondary minerals.

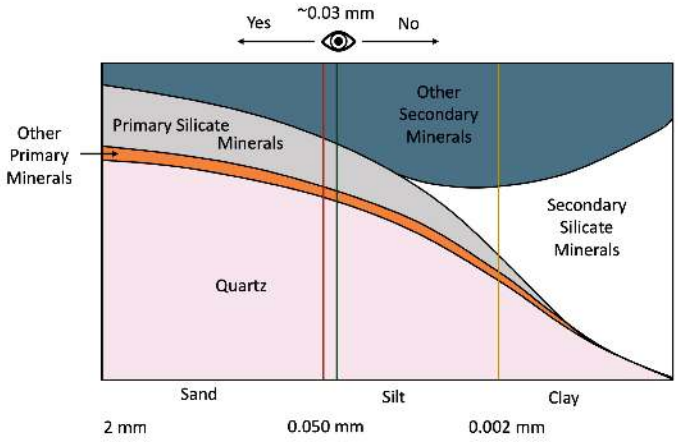


Figure 14.3. Size distribution of minerals in soil. Area within the figure denotes the relative abundance of minerals. Particle size fractions based on the Canadian System of Soil Classification (SCWG, 1998). © Loganathan (1987) adapted by Lewis Fausak is licensed under a CC BY (Attribution) license.

Elemental Abundance and Mineral Structures

There are 91 naturally occurring elements, all of which are found at some concentration in soil. Table 14.1 provides a list of 70 elements in order of decreasing abundance in crustal rocks and in soils. Most elements substitute for one another within the crystal structure of minerals depending on their size and valence. This will be discussed in more detail later in this chapter. Depending on solubility and resistance to chemical weathering, the content of some elements contained in soil minerals increase relative to crustal abundance (e.g., Si and O) while the concentration of most elements decreases as they are released to the environment (compare crustal elemental abundance with soil abundance; Table 14.1).

Table 14.1. Total content (mg kg^{-1}) of selected elements in order of abundance in continental crust rocks and in soils (Adapted from Bowen, 1979)

Element	Crust	Soil		Element	Crust	Soil
O	464000	490000		B	10	38
Si	281500	330000		Th	9.6	14
Al	81650	71000		Sm	6	6
Fe	53000	32000		Gd	5.4	3.5
Ca	38650	15000		Cs	3	3
Na	26150	10900		Dy	3	5.7
K	23950	18300		Yb	3	3.9
Mg	21950	8300		Hf	3	7.7
Ti	5050	5100		Be	2.8	1.5
P	1050	800		Er	2.8	3
Mn	950	760		U	2.7	2.2
F	625	270		Br	2.5	43
Ba	425	568		Sn	2	5.8
Sr	375	278		Ta	2	1.2
S	260	433		As	1.8	11
Zr	165	345		Ge	1.5	3
V	135	108		Mo	1.5	1.9
Cl	130	485		W	1.5	1.1
Cr	100	84		Eu	1.2	1.3
Rb	90	120		Ho	1.2	0.8
Ni	75	34		Tb	0.9	0.85
Zn	70	60		I	0.5	7.1
Ce	60	84		Lu	0.5	0.46
Cu	55	26		Tm	0.48	0.62
Y	33	28		Tl	0.45	0.25
La	30	41		Cd	0.2	0.6
Nd	28	44		Sb	0.2	1.7
Co	25	12		Bi	0.17	0.5
Sc	22	10		In	0.1	1
Li	20	31		Hg	0.08	0.1
Nb	20	14		Ag	0.07	0.05
Ga	15	21		Se	0.05	0.4
Pb	12	29		Pt	0.001	0.001

Element	Crust	Soil		Element	Crust	Soil
Pr	11	6.5		Au	0.001	0.001

Note: Does not include hydrogen, carbon, nitrogen, noble gasses, or elements found in ultra-trace (<0.001 mg kg^{-1}) amounts. See https://earthref.org/GERMRD/10/ for the most updated version of this table. See Spiers et al. (1989a) for selected elements in Alberta soils.

Closer examination of elemental abundances indicates that the eight most abundant elements (O, Si, Al, Fe, Ca, Na, K and Mg) account for almost the entire mass (99%) of the Earth's crust (Table 14.2). Considering the atomic radii further indicates that almost 85% of the Earth's crust by volume is composed of oxygen as oxides. Specifically oxide based minerals (O^{2-}) form the fundamental framework of most minerals in the Earth's crust with the major cations (Si, Al, Fe, Ca, Na, K, and Mg) fitting between the oxygen atoms on the atomic level.

Table 14.2. Radius of O^{2-} and major cations, crustal abundance, radius ratios and predicted coordination numbers

Element	Atomic Radius (nm)	Abundance by Mass (%)	Abundance by Volume (%)	Radius Ratio	Predicted coordination number (CN)
O^{2-}	0.14	0.464	0.845	na*	na*
Si^{4+}	0.04	0.282	0.012	0.29	4
Al^{3+}	0.054	0.0817	0.0083	0.38	4
Fe^{3+}	0.065	0.053	0.0094	0.46	6
Fe^{2+}	0.078	Incl.**	Incl.**	0.56	6
Ca^{2+}	0.112	0.0387	0.0361	0.80	8
Na^{+}	0.097	0.0262	0.0285	0.69	6
K^{+}	0.151	0.022	0.0548	1.08	12
Mg^{2+}	0.072	0.024	0.0054	0.51	6
Total		0.9909	1		

Note: *na = not applicable. **: abundance of Fe^{2+} included with Fe^{3+}

The crystal structure of minerals at the atomic level may be easily visualized if we assume that oxygen and the other metal ions act as rigid spheres with fixed radii. Although not strictly true, this is a useful conceptual to visualize crystal structures. The geometry of these rigid spheres is governed by Pauling's Rules (Pauling 1929) where the first rule states that the distance between a cation and anion is the sum of their radii. Ions in a crystal structure tend to gather as many oppositely charged ions around itself as size permits. The number of O^{2-} ions that may surround a central cation is called the coordination number (CN). Large cations (e.g. K^{+}) have high coordination numbers because of their large volumes while small cations (e.g., Si^{4+}) have low coordination numbers (Figure 14.4). Also note that those cations with large atomic radii and low valence (K^{+} and Na^{+}) are the most mobile while those with smaller radii and high charge (Si^{4+}, Al^{3+} and Fe^{3+}) are far less mobile.

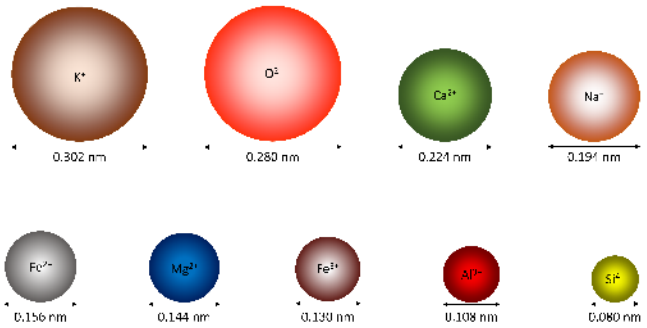


Figure 14.4: Relative sizes (diameters in nm) of the major soil ions. © Maxime Paré is licensed under a CC BY (Attribution) license.

Since ions tend to surround themselves with as many oppositely charged ions as possible, predicting coordination numbers requires using only the relative sizes of the ions expressed as the radius ratio:

(1) $$Radius\ Ratio\ RR = \frac{Radius\ of\ Cation}{Radius\ of\ Anion\ (O^{2-})}$$

Table 14.3 provides ranges of radius ratios correspond to common coordination numbers and the corresponding structural units. Some overlap and uncertainty arise because the assumption of treating ions as rigid spheres with constant radii is not strictly true. With this in mind, ions are typically interchangeable if their size does not differ by more than about 15%, and their valences do not differ by more than one unit. In reality, ions are not spherical and their radii change slightly with different coordination numbers. As a result, some elements have more than one common coordination geometry. For example, aluminum is commonly found in both tetrahedral (CN=4) and octahedral (CN=6) units. Substitution of cations within mineral structures is referred to as isomorphous substitution and will be discussed in more detail later in this chapter.

The geometry of ions about a central ion can be visualized as a coordinating polyhedron. In mineralogy, these polyhedra are usually constructed from the arrangement of anions; around the cations but sometimes the cations may be absent altogether. By far the most common polyhedral structures that make up most soil minerals are the tetrahedron (four-sided structure with four oxygen atoms coordinating around Si) and the octahedron (eight-sided structure with six oxygen atoms coordinating around aluminum; Figure 14.5). Since Si and O are by far the most abundant elements in soil, minerals with tetrahedral structures form the basis for many of the common primary minerals. Sheet-like phyllosilicates composed of combination of silica tetrahedrons and aluminum octahedrons are secondary minerals dominating the clay-size (<0.002 mm) size fraction. Phyllosilicates will be discussed in further detail later in this chapter.

Table 14.3. Radius ratios coordination numbers and associated structural units for most cations relative to structural oxygen (O^{2-}) in minerals

$R_{cation}/R_{(O^{2-})}$ Range	Coordination Number (CN)	Polyhedron Unit
0.155-0.225	3	Triangular
0.225-0.414	4	Tetrahedral
0.414-0.732	6	Octahedral
0.732-1	8	Cubic
>1	12	Cuboctahedron

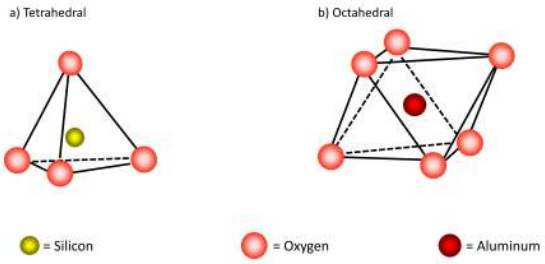


Figure 14.5. a) Sketches representing a single tetrahedral unit. b) A single octahedral unit. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

The fundamental difference among mineral structures is the number of shared oxygen atoms that occurs between adjacent tetrahedra. Although there are several types of mineral structures and classifications; generally speaking, most common soil minerals are composed of repeating silicate tetrahedron (SiO_4) units that fall into one of four simplified general types of structures:

- isolated silica tetrahedra (neosilicates);
- chains of silica tetrahedra (inosilicates);
- framework tetrahedra (tectosilicates); and,
- sheets of tetrahedra (combined with octahedral units; phyllosilicates).

Some common soil minerals listed according to the amount of shared oxygen atoms between adjacent tetrahedra are given in Table 14.4.

Table 14.4. Some of the common soil minerals

Oxygen Sharing in Tetrahedra	Common Mineral Groups	Example Mineral Name	Generalized Ideal Composition
Neosilicates	Olivines	Fosterite	Mg_2SiO_4
No sharing			
Isolated tetrahedra		Fayalite	$Fe(II)_2SiO_4$
Inosilicates	Pyroxenes	Enstatite	$MgSiO_3$
One oxygen shared		Diopside	$(Ca,Mg)SiO_3$
		Augite	$(Ca,Mg,Fe(II),Al)SiO_3$
Inosilicates	Amphiboles	Tremolite	$Ca_2Mg_5[Si_8O_{22}(OH)_2]$
Two oxygen shared		Actinolite	$Ca_2(Mg,Fe(II))_5[Si_8O_{22}(OH)_2]$
		Hornblende	$(Ca,Na)(Mg,Fe(II),Al)_5[(Al,Si)_8O_{22}(OH)_2]$
Phyllosilicates		Kaolinite	$Si_2Al_2O_5(OH)_4$
Three oxygens shared		Muscovite	$KAl_3(AlSi_3)O_{10}(OH)_2$
Tectosilicates all four oxygens shared	Framework Silicates	Quartz	α-SiO_2
		Feldspars	$(Ca,Na,K)(Al_xSi_{(3-x)})O_8$

MINERALS OF THE SAND-SIZE (2 mm – 0.05 mm) FRACTION

Quartz

Quartz (SiO_2) or more specifically, α-quartz (α-SiO_2), is the most common mineral found in Canadian soils, typically accounting for the majority of the sand fraction (Figure 14.3). Quartz is a framework silicate (tectosilicate) with all four oxygen atoms in its silica tetrahedrons shared among adjacent Al,Si_3 tetrahedra (Figure 14.6). It belongs to the hexagonal crystal system.

When tetrahedra share all oxygen atoms with adjacent tetrahedra, they form a very strong 3-dimensional framework of Si-O bonds. Quartz is essentially pure SiO_2. The negative charge of the O atoms is exactly balanced by Si requiring no other bonding ions. Under a microscope most specimens are usually irregular shaped lustrous fragments appearing either white, milky and sometimes clear. The 3D framework chemical structure of quartz is typical of tectosilicates. The repeating units are SiO_2 in a tetrahedral configuration. The Si-O bonds are very strong covalent bonds, highly resistant to weathering but overall not chemically stable in soil. The solubility of quartz is very low and the rate of weathering (dissolution) is extremely slow. Quartz is, therefore, highly persistent in soil with a slow rate of dissolution compared to most other minerals leading to a far longer persistence, with relative quantities building up over time (e.g., Ae horizons of Podzolic and Luvisolic soils). Visually quartz appears usually as clear, white, or gray glassy fragments but sometimes exists as yellow, brown, purple, pink or red fragments that can easily be seen with the naked eye or a hand lens. Quartz usually constitutes about 40-60% of the sand (2.0–0.05 mm) fraction, a major fraction of the silt-size range (0.05-0.002 mm) and is present in small amounts in the clay-size (<0.002 mm) fraction of Canadian soils as well (see Figure 14.3).

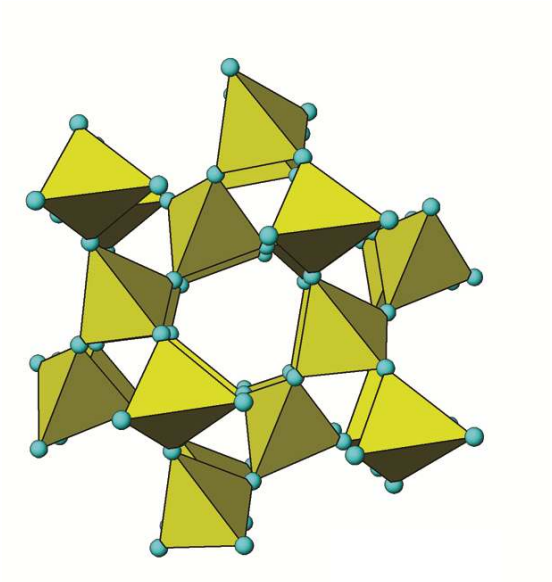


Figure 14.6. A schematic representation of the three-dimensional atomic structure of quartz. Illustration from Open-geology teaching website (https://opengeology.org/Mineralogy/1...al-structures/). © Dexter Perkins, University of North Dakota (open e-texbook Mineralogy 2020) is licensed under a CC BY-NC-SA (Attribution NonCommercial ShareAlike) license.

Feldspars

The feldspars are actually a group of minerals making up the second most common tectosilicate group found in soil (after quartz), typically accounting for 15-35% of the sand fraction (Figure 14.3). Feldspars make up a large portion of granitic and metamorphic rocks. Therefore, feldspars constitute a significant portion of our soil. There are many mineral species making up the feldspar group, with all having a general formula of $X(Al,Si)_4O_8$ where X is most commonly K, Na or Ca with Al and Si ions making up the tetrahedral units. Other ions such as Ba, Zn, Rb, Sr, and Fe^{2+} also substitute for X in feldspar structures (King 2020). Feldspars typically appear off-white, with various shades of red, orange, brown and occasionally green existing. The 3D framework of the feldspars is made up of units ranging from $AlSi_3O_8$ to $Al_2Si_2O_8$ in a “kinked ribbon” configuration (Figure 14.7a). The K, Na or Ca found in the voids of the kinks with sufficient Al substituting for Si within the kinked ribbon structure to produce a negative structural charge to neutralize the positive charge on the cations.

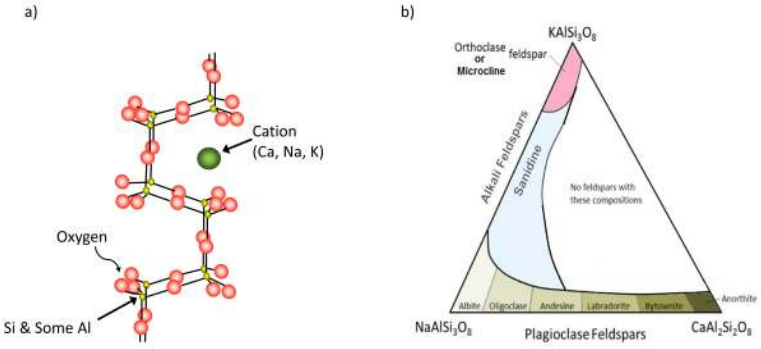


Figure 14.7. **(a)** Structural atomic arrangement (kinked ribbon) of feldspars. © Darshani Kumaragamage. CC BY. **(b)** A compositional ternary diagram for feldspar minerals. Adapted by Jim Warren from King (2020). © King 2020; Darshani Kumaragamage, Univ. of Winnipeg adapted by Jim Warren.

The most common feldspar minerals and their ranges in composition are illustrated in the ternary diagram (Figure 14.7b). Orthoclase and microcline are the mineral names given to feldspars with the same chemical composition dominated by potassium ($KAlSi_3O_8$) with minor amounts (up to about 25%) of Na substituting for K in the structure (Figure 14.7B). Orthoclase and microcline are polymorphs, which means they have the same chemical composition but different crystal forms. Orthoclase has a monoclinic crystal form while microcline is triclinic. Feldspars containing more than about 25% Na but less than 75% K constitute the mineral sanidine [$(K,Na)AlSi_3O_8$]. Feldspars within the full range of K and Na compositions (orthoclase/microcline, sanidine to albite) are generally referred to as the alkali feldspars (Figure 14.7B). Sodium and Ca feldspars occur in a continuous series called the plagioclase series with compositions ranging from $NaAlSi_3O_8$ to $CaAl_2Si_2O_8$ where the charge on the amount of Na and Ca in the structure is neutralized by substitution of Al for Si in the tetrahedral kinks. Plagioclase feldspars with specific compositions are given specific mineral names listed in Table 14.5. The plagioclase series of minerals is an example of a solid solution series where composition varies between two end-member minerals (i.e., albite; $NaAlSi_3O_8$ and anorthite; $CaAl_2Si_2O_8$) that share the same basic chemical formula but experience substitution of elements in one or more atomic sites.

Quartz and the feldspars have similar specific gravity and hardness, thus they tend to be transported and deposited together. Unlike quartz, feldspars have several cleavage planes along which the minerals fracture— making them brittle and more prone to mechanical breakdown upon impact during physical transport. These cleavage planes are weakness planes along which water and microbial exudates may infiltrate and accelerate their chemical alteration. Organisms such as bacteria and fungal hyphae also infiltrate these weakness planes to enhance feldspar degradation. Hence, feldspars tend to weather and dissolve at a faster rate compared to quartz (Cousineau, 2020).

Table 14.5. Range in composition for plagioclase feldspars

Mineral Name	% $NaAlSi_3O_8$	% $CaAl_2Si_2O_8$	Range of Composition
Albite	90-100	0-10	$Na(Al,Si_3O_8)$ – $Na_{0.9}Ca_{0.1}(Al_{1.1}Si_{2.9}O_8)$
Oligoclase	70-90	10-30	$Na_{0.9}Ca_{0.1}(Al_{1.1}Si_{2.9}O_8)$ – $Na_{0.7}Ca_{0.3}(Al_{1.3}Si_{2.7}O_8)$
Andesine	50-70	30-50	$Na_{0.7}Ca_{0.3}(Al_{1.3}Si_{2.7}O_8)$ - $Na_{0.5}Ca_{0.5}(Al_{1.5}Si_{2.5}O_8)$
Labradorite	30-50	50-70	$Na_{0.5}Ca_{0.5}(Al_{1.5}Si_{2.5}O_8)$ - $Na_{0.3}Ca_{0.7}(Al_{1.7}Si_{2.3}O_8)$
Bytownite	10-30	70-90	$Na_{0.3}Ca_{0.7}(Al_{1.7}Si_{2.3}O_8)$ - $Na_{0.1}Ca_{0.9}(Al_{1.9}Si_{2.9}O_8)$
Anorthite	0-10	90-100	$Na_{0.1}Ca_{0.9}(Al_{1.9}Si_{2.9}O_8)$ - $Ca(Al_2Si_2O_8)$

"Heavy" Minerals

"Heavy" minerals represent a generalized group of "other" primary minerals usually found only in minor amounts (2-5%) in the sand fraction of Canadian soils. Several of these minerals are dark coloured due their high content of Fe and Mg. Others such as garnet and hematite are dark red and zircon is yellow to colorless. Heavy minerals may be isolated from the sand and silt-size fractions based on density separation (Andò, 2020). Heavy minerals are not typically found in the clay-size fraction. Heavy minerals for the most part are generally defined as having particle density (specific gravity) greater than 2.90 g/cm^3 (Andò, 2020). Some common minerals found in the sand fraction of soils are provided in Table 14.6. Although the heavy mineral component is only a minor mineral component of the sand fraction, it may supply many critical plant nutrients through progressive weathering.

Once isolated, the heavy mineral fraction may be examined by several methods including light and electron microscopy coupled with computer machine learning techniques (Maitre et al., 2019), powder X-ray diffraction, and other physical or chemical methods. Heavy minerals generally include mafic minerals typically with high concentrations of Fe and Mg. Macroscopically pyroxenes and amphiboles can be difficult to distinguish from one another, as they are both dark-colored. Typically, pyroxene crystals are shorter crystals while amphiboles have longer needle-like crystals. Other minerals in the heavy mineral fraction may include garnets ($X_3Al_2Si_3O_{12}$ were X may be Mg, Fe(II), Mn, or Ca), magnetite (Fe_3O_4), chromite ($Fe,Cr(III)_2O_4$), hematite (Fe_2O_3), zircon ($ZrSiO_4$), rutile (TiO_2).

Table 14.6. Some common minerals in the sand-size fraction with associated densities

Density Fraction	Mineral	Density (g cm^{-3})
"Light" Fraction (Density <2.9 g cm^{-3})	quartz	2.65
	feldspars	2.55-2.76
"Heavy" Fraction (Density >2.9 g cm^{-3})	olivine	3.22-4.39
	augite (pyroxene)	2.96-3.52
	hornblende (amphibole)	3.02-3.45
	hematite (Fe_2O_3)	5.26
	magnetite (Fe_3O_4)	5.2

MINERALS OF THE SILT-SIZE (0.05 – 0.002 mm) FRACTION

The silt-size fraction of soils is intermediate in size and mineralogy between the sand and clay-size fractions. Hence, the silt fraction effectively contains a physical mixture of minerals found in both the sand-size fraction and the clay-size fraction. Readers are therefore directed to those sections of this chapter for further discussion.

Can You Dig It!

The terms "clay" and "clay mineral" are often used interchangeably and see commonplace use in the scientific literature; however, their use can be ambiguous and sometimes misleading. The term "clay" is typically used to refer to a clay soil or the clay-size fraction (particles <0.002 mm diameter). "Clay minerals" refer specifically to phyllosilicates which account for the majority of "clay" in soil but does not include oxyhydroxides and amorphous aluminosilicates. Although not phyllosilicates, numerous sand-size mineral described above may also be present in the clay-size fraction.

MINERALS OF THE CLAY-SIZE (less than 0.002 mm) FRACTION

As discussed previously, the clay-size fraction is comprised mostly of secondary hydrated aluminosilicates. These minerals are very fine-grained and thus have a very high SSA and account for the largest portion of the available and reactive mineral surface area in soil (Chapter 5, Table 5.1). In addition, minerals in the clay fraction have both permanent and pH variable charge capable of attracting and retaining plant nutrients with direct and indirect impacts on the soil fertility. In short, they are the most chemically active inorganic phases in soil.

SHEET SILICATES: PHYLLOSILICATES

By far the most common type of minerals in the clay-size fraction are the sheet or layer silicates called phyllosilicates (Greek; phyllon meaning leaf or sheet), which appear as flat sheet structures in photomicrographs (e.g., Beutelspacher and Van Der Marel, 1968). Phyllosilicates are crystalline minerals, with a structure consisting of layers of repeating atomic units; thus given the name "layer" or "leaf" aluminosilicates. The fundamental layers are composed of tetrahedral and octahedral sheets that occur in different combinations to form different phyllosilicate mineral species.

Tetrahedral sheets are composed of repeating tetrahedron (four-sided) units each containing a Si atom surrounded by four O atoms. Three of the four O atoms are shared with adjacent tetrahedral units, referred to as the basal O atoms. The one unshared O is referred to as the apical O (Figure 14.8). The shared basal O atoms combine to form sheets that continue in two dimensions *ad infinitum*.

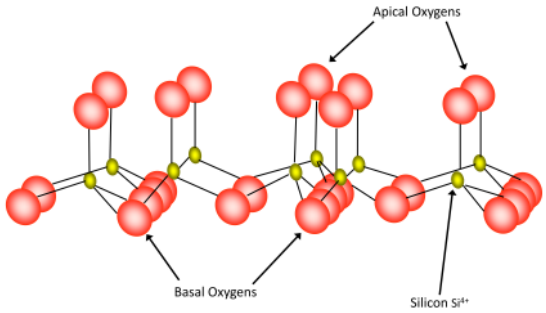


Figure. 14.8. Arrangement of atoms in the tetrahedral sheet. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

The octahedral sheets found in phyllosilicates are composed of multiple octahedron units, each unit containing Al and/or Mg atoms surrounded by six O atoms or OH^- groups. Three O atoms (or OH^- groups) occur in the lower plane, with three oxygen atoms (or hydroxyl groups) in the upper plane sandwiching aluminum or magnesium atoms between, forming an eight-sided (octahedral) configuration and a pseudohexagonal form. These units combine by sharing O atoms to form an octahedral sheet extending in two dimensions *ad infinitum* (Figure 14.9). When the central cation of the octahedron is trivalent, such as aluminum (Al^{3+}), only two of three (2/3) cation sites in an octahedral sheet require filling in order to neutralize the charge by the surrounding O atoms or OH^- ions. In this case, the octahedral sheet is referred to as dioctahedral (Figure 14.9). When the central cation of the octahedron is divalent, such as Mg^{2+}, then all three of three (3/3) cation sites in an octahedral sheet require filling in order to neutralize the negative charge of the surrounding O or OH^-. In this case, the octahedral sheet is referred to as trioctahedral (Figure 14.9).

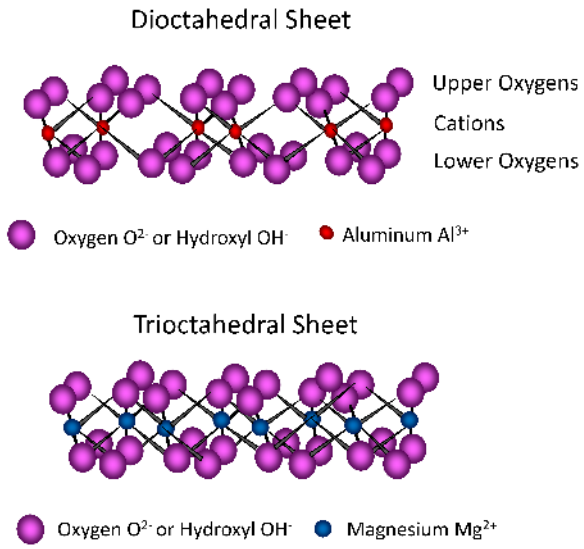


Figure 14.9. Arrangement of atoms in octahedral sheets. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Phyllosilicate structures have tetrahedral and octahedral sheets combined to form layers through sharing of the apical O atom of the tetrahedral sheet with O atoms in the octahedral sheet. Phyllosilicates composed of one tetrahedral sheet and one octahedral sheet are called 1:1 type phyllosilicates. Those composed of two tetrahedral sheets sandwiching one octahedral sheet between are called 2:1 type phyllosilicates. Phyllosilicates consist of many layers (1:1 type or 2:1 type) stacked one on top of the other with some form of bonding between the layers. The region located between two adjacent 1:1 or 2:1 layers is referred to as the interlayer region. Some 2:1 type phyllosilicates may also contain an additional octahedral sheet composed completely of hydroxides in the interlayer region. These are referred to as 2:1:1 or 2:2 type phyllosilicates. The basic behaviour and properties of the different phyllosilicates are dictated in large part by the characteristics associated with the interlayer region of the clays found in a given soil.

Isomorphous Substitution

In addition to their extremely large SSA, phyllosilicates are also electrically charged due to the substitution of the central cations within the tetrahedral or octahedral sheets for others of similar size but different valence (Table 14.2). Table 14.3 provides ranges of radius ratios corresponding to common coordination numbers and their corresponding structural units. Ions are typically interchangeable within a structure if their size does not differ by more than about 15% and valences do not differ by more than one unit. Substitution occurs at the time the phyllosilicates form. The substitution of a cation of lower valence for another of similar size imparts a net negative charge to the structure of the mineral because the charge on the surrounding O atoms is not balanced completely. For example, trivalent aluminum (Al^{3+}) may substitute for some of the tetravalent silicon (Si^{4+}) atoms in the tetrahedral sheet. The lesser charge of the Al compared to Si creates a charge deficit with a resulting net negative charge within the tetrahedral sheet of the structure. Similarly, in dioctahedral sheets, Mg^{2+}, Fe^{2+}, or other divalent cations, may substitute for some Al^{3+} resulting in an extra negative charge associated with the octahedral sheet of the structure. Similarly, Li^{+}, although occurring in low abundance (Table 14.1), also sometimes substitutes for Mg^{2+} in trioctahedral phyllosilicates (lepidolite).

To neutralize the negative charges produced by isomorphous substitution within the structure of phyllosilicates, positively charged cations are attracted from the soil solution to the surface of the minerals. These charged neutralizing cations that migrate into the interlayer region of the phyllosilicates are called exchangeable cations. These cations are capable of exchanging with other cations in the soil solution. Collectively, the overall abundance of negative charge on the clay structure is referred to as the cation exchange capacity (CEC). The abundance of negative charge sites (the charge density); the location of the negative charges (tetrahedral and/or octahedral sheet), and the type of cation in the interlayer region neutralizing the charge (e.g., Na^{+}, Ca^{2+}, Mg^{2+}, K^{+} or other) all determine the ability of the phyllosilicate to be expandable (swelling clay), and control the capacity to retain soil and plant nutrients.

The Interlayer Region

In addition to attracting and retaining cations, the nature of the interlayer region of some phyllosilicates may also be controlled by hydrogen bonding. In 1:1 type clay minerals, hydroxyl groups associated with the octahedral sheet in the structure face a layer of O atoms formed by the apical Os in the adjacent 1:1 layer (Figure 14.10). The result is that hydrogen bonding holds the 1:1 layers closely together forming a very stable structure. Consequently, all 1:1 type phyllosilicates do not expand and are thus non-swelling.

In contrast, hydrogen bonding does not occur in 2:1 clay minerals as basal O atoms in two adjacent layers face each other with no OH^- groups (Figure 14.11). Thus, the unit layers are held together by the weaker electrostatic forces associated with whatever neutralizing cation occupies the interlayer region to neutralize the negative charges associated with isomorphous substitution within the layer structures. The interlayer cations may be any combination of Ca^{2+}, Mg^{2+}, Na^+, K^+, NH^{4+}, plant nutrients (Zn^{2+} Cu^{2+}, Mn^{2+}, Ni^{2+}), iron (Fe^{2+}, Fe^{3+}), aluminum (Al^{3+}) or other cations. These cations are typically exchangeable and have some amount of water associated with the cation.

The degree of expandability of a 2:1 clay mineral depends on the total layer charge; the location of the layer charge (tetrahedral *vs.* octahedral sites) and the nature of the exchangeable cation. The 2:1 phyllosilicates with a low layer charge (<0.6 per formula unit)—usually associated with the octahedral sheet—typically exhibit the capability of swelling. Swelling properties are usually associated with the smectites (see Table 14.7). The 2:1 phyllosilicates with a higher layer charge (>0.6 per formula unit)—usually associated with the tetrahedral sheets—are far less likely to exhibit swelling properties. The non-swelling 2:1 phyllosilicatess are the vermiculites and micas (Table 14.7).

Water is a polar molecule, allowing development of hydration spheres with the exchangeable cation. Divalent cations such as Ca^{2+} and Mg^{2+} typically have two layers of water associated with it in the interlayer region of 2:1 clay minerals with lower total layer charge. Large monovalent cations such as Na^+ and K^+ obviously, having a lower charge, are not as capable as divalent cations at holding the 2:1 structures together and therefore may allow more water molecules to be incorporated into the interlayer region. Under these conditions the 2:1 structures separate to many times their original ("dry") configuration and the clay is said to be a swelling clay. This is typical of the smectite group of 2:1 phyllosilicates (see below). Note that in addition to water, polar organic molecules can also migrate into the interlayer region.

In some instances, the interlayer cation in 2:1 type phyllosilicates may become hydroxylated in six-fold (octahedral coordination). Such is the case with Al, Mg and Fe. In these cases, the interlayer becomes filled with an additional octahedral sheet which does not share O atoms with the main 2:1 layer but is hydrogen-bonded to the adjacent oxide surfaces forming a phyllosilicate which has a 2:1:1 (or 2:2) type structure. These types of phyllosilicates are referred to as the chlorite group or, more generally, chlorite.

Common Phyllosilicates

There are about 50 different species of phyllosilicates but most are rare or occur in soils found in association with specific geological deposits (e.g., halloysite in volcanic ash soils in humid regions). A general classification scheme for some of the more common phyllosilicates is shown in Table 14.7. The five species of phyllosilicates (kaolinite, smectite (montmorillonite and beidellite), micas (including hydrous mica or illite), vermiculite and chlorite) commonly identified in Canadian soils are discussed in detail below.

Table 14.7. A general classification scheme for some phyllosilicates

Type	Group x = charge per formula unit	Subgroup	Example Species	Idealized formula unit*
1:1	Kaolinite x~0	Kaolinite	Kaolinite	$Si_2,Al_2O_5(OH)_4$
			Halloysite	$Si_2,Al_2O_5(OH)_4\ 2H_2O$
		Serpentines	Serpentine	$Si_2,Mg_3O_5(OH)_4$
2:1	Pyrophyllite-talc x~0	Pyrophyllite	Pyrophyllite	$Si_4,Al_2O_{10}(OH)_2$
		Talc	Talc	$Si_4,Mg_3O_{10}(OH)_2$
	Smectite x ≈ 0.25–0.6	Dioctahedral	Montmorillonite	$0.33M+Si_4(Al_{1.67},Mg_{0.33})O_{10}(OH)_2$
			Beidellite	$0.33M+(Si_{3.67}Al_{0.33})(Al_{2.00})O_{10}(OH)_2$
			Nontronite	$0.33M+(Si_{3.67}Al_{0.33})(Fe(III)_{2.00})O_{10}(OH)_2$
		Trioctahedral	Saponite	$0.33M+(Si_{3.67}Al_{0.33})(Mg_{3.00})O_{10}(OH)_2$
			Hectorite	
	Vermiculite x ≈ 0.6–0.9	Dioctahedral	Dioctahedral vermiculite	$0.86M+(Si_{3.47}Al_{0.53})(Al_{1.67}Mg_{0.33})O_{10}(OH)_2$
		Trioctahedral	Trioctahedral vermiculite	$0.86M+(Si_{3.47}Al_{0.53})(Al_{1.67}Mg_{0.33})O_{10}(OH)_2$
	Micas x ≈ 1	Dioctahedral	Muscovite	$K_{1.00}(Si_{3.00}Al_{1.00})(Al_{2.00})O_{10}(OH)_2$
			Paragonite	
		Trioctahedral	Biotite	$K_{1.00}(Si_{3.00}Al_{1.00})(Fe(II)_{3.00})O_{10}(OH)_2$
			Phlogopite	
2:1:1	Chlorite x = variable	Di-Dioctahedral	Donbassite	$(Al_2(OH)_6)(Si,Al)_4Al_2O_{10}(OH)_2$
		Tri-Dioctahedral	Sudoite	$(Mg_3(OH)_6)(Si,Al)_4Al_2O_{10}(OH)_2$
		Di-Trioctahedral	Ferri-sudoite	$(Al_2(OH)_6)(Si,Al)_4(Mg,Fe)_3O_{10}(OH)_2$
		Tri-Trioctahedral	Clinochlore	$(Mg_3(OH)_6)(Si,Al)_4(Mg,Fe)_3O_{10}(OH)_2$

*The sequence of elements within the idealized formula unit follows the order: exchangeable cation (if any) or interlayer octahedral sheet; tetrahedral cations; octahedral cations; oxygens and hydroxyls. Within each sheet the major cation (e.g. silica) is followed by the isomorphically substituting cation(s).

Kaolinite is a 1:1 type clay mineral constructed of one tetrahedral sheet and one octahedral sheet (Figure 14.10). The 1:1 layers are held together by hydrogen bonds between the layers so that structure is non-expandable as water molecules cannot enter the interlayer region. Kaolinite exhibits very little swelling when wet, or shrinkage on desiccation. The interlayer is not exposed and therefore the specific surface area is low. The mineral has almost no isomorphous substitution; therefore, these kaolinite minerals have a low surface charge and low CEC, although they have minor variable charge sites at the edges. Kaolinite is the most common 1:1 type phyllosilicate and is ubiquitous in Canadian soils.

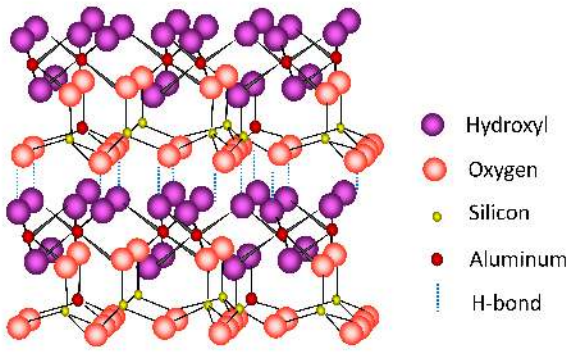


Figure 14.10. Idealized structure of kaolinite. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Smectite is a freely expandable 2:1 type phyllosilicate (Figure 14.11). There are two common species of smectite: montmorillonite and beidellite. The amount of isomorphous substitution in both minerals is moderate compared to other 2:1 type phyllosilicates. The layer charge for smectites is in the range of about 0.25 to 0.6 sites per formula unit $(M^{x+}(Si,Al)_4(Al,Mg)_2O_{10}(OH)_2)$ where M^{x+} generally refers to exchangeable cations (e.g., Na^+, Mg^{2+}, K^+, Ca^{2+}, etc.). Montmorillonite and beidellite fall within the same range of total layer charge but are differentiated based on the location of the layer charge in the structure (Table 14.7). Most of the layer charge in montmorillonite is seated in the octahedral sheet while most is associated with the tetrahedral sheet of beidellite. Specific surface area is high (about 800 $m^2\ g^{-1}$) and both water molecules and cations can enter the interlayer space.

Smectite-rich soils are sticky when wet and form large cracks under dry conditions. Heavy clay textured soils containing >60% clay of which at least half is smectite are defined as Vertisols in Canada (SCGW, 1998). These soils are described as “self mulching” as repeated wetting and drying cycles form slickenslides, which are polished surfaces on opposing ped surfaces along soil fissures formed where soil materials move back and forth past each other as large continuous cracks reopen and close with wetting and desiccation. The soil forming process causing the formation of slickenslides is called argilli-pedoturbation, being especially common in Vertisolic soils formed in the smectite-rich glacio-lacustrine parent materials found throughout the Prairie Provinces.

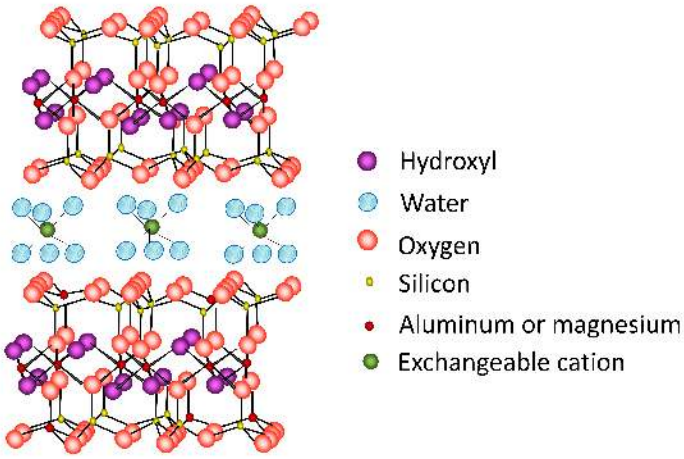


Figure 14.11. Idealized structure of smectite and vermiculite. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Micas, including hydrous mica/illite are a 2:1 type clay minerals with two tetrahedral sheets sandwiching an octahedral sheet. The 2:1 layers are held tightly together by large K^+ in the interlayer region (Figure 14.12). The structure of the micas is similar to that of the smectites except that the layer charge is much higher (about 1 per formula unit) and isomorphous substitution occurs almost exclusively in the tetrahedral sheets. Potassium almost exclusively occupies the interlayer region. The very high layer charge, combined with the location of the charge in the tetrahedral sheets very close to the interlayer region, prevents expansion of the mineral. Micas are primary minerals derived from granite and metamorphic rocks of the Canadian Shield. Hydrous mica/illites form in soil from the partial weathering of the micas. Micas and hydrous mica/illite are found in almost all soils in Canada. There are two common species of mica differentiated by the nature of the octahedral sheet, namely dioctahedral muscovite and trioctahedral biotite (Table 14.7).

Hydrous mica/illite is a partial weathering product of the micas, with some (up to 20-30%) of the interlayer K^+ being removed from the interlayer region and replaced with other exchangeable cations. There is also loss of tetrahedral Al^{3+} from the structure. Hydrous mica/illite is usually produced through the weathering of dioctahedral muscovite whereas the trioctahedral biotite structure typically breaks down totally due to relative ease of chemical attack on iron (Fe(II)) in the structure.

Although the basic structure of the micas and hydrous mica/illite are similar to smectite, the properties are very different primarily due to the higher amount of isomorphous substitution in the tetrahedral sheet, where Al^{3+} is substituted for Si^{4+} (Figure 14.12). The strong negative charge located in the tetrahedral sheets close to the interlayer surface is balanced by K^+, locking the adjacent layers together. Thus, the mineral has limited expansion as water molecules cannot enter the interlayer space. Since K^+ are trapped in the lattice preventing expansion, the CEC is low.

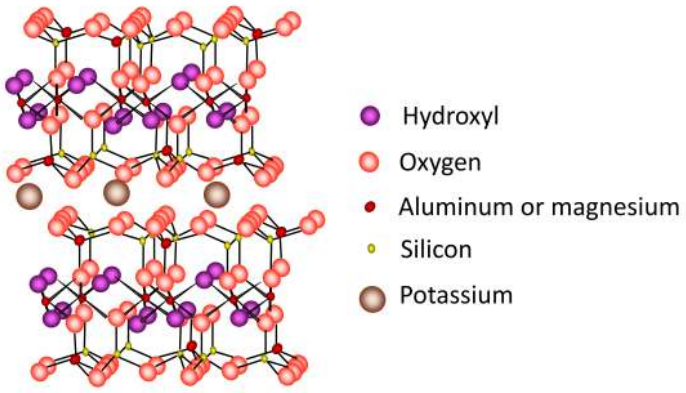


Figure 14.12. Idealized structure of muscovite mica. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Vermicullite is a 2:1 type phyllosilicate that is similar structurally to smectite and the micas. The layer charge is in the range of 0.6 to 0.9 sites per formula unit ($M^{x+}(Si,Al)_4Al_2O_{10}(OH)_2$), which is intermediate between the micas and the smectites (Figure 14.13). As in the micas, the layer charge is located almost exclusively within the tetrahedral sheet. The expandability of vermiculite is limited and it has moderate swelling and shrinkage characteristics. With a combination of a layer charge lower than the micas but higher than smectite and with most isomorphous substitution is in the tetrahedral sheet vermiculite has a high surface charge and a high CEC. As the structure is similar to the micas, vermiculite is capable of trapping K^+ ions in the interlayer region reverting the structure to partly resemble hydrous mica/illite.

Chlorite is the name for a mineral, as well as a group of 2:1:1 (or 2:2) type phyllosilicates constructed of a 2:1 layer structure with a hydroxide layer in the interlayer region (Kohut and Warren 2002). The mineral chlorite is the most common mineral species within the chlorite group. The structure is composed of an Al, Mg or Fe hydroxide layer in six-fold (octahedral) coordination in the interlayer region between 2:1 type layers (Figure 14.13). This interlayer hydroxyl layer does not share O atoms with the main 2:1 layer. The hydroxide layer is hydrogen bonded to the adjacent 2:1 layers and is non-expandable (Figure 14.13). The amount of isomorphous substitution is variable, but as the structure is non-expandable so the overall CEC is low.



Figure 14.13. Idealized structure of a tri-dioctahedral chlorite. © Darshani Kumaragamage, Univ. of Winnipeg is licensed under a CC BY (Attribution) license.

Exchange Capacity and Specific Surface Area

The clay-size fraction of most Canadian soils typically contain a mixture of the five common phyllosilicates along with small amounts of amorphous aluminosilicates and Fe and/or Al oxy-hydroxide. Proportions are mostly a function of the origins of the soil parent materials. Variations in the overall ion exchange and adsorptive properties of soils depends on the mix of phyllosilicates present. The two main differences among the phyllosilicates is their CEC and the amount of reactive surface area are summarized

in Table 14.8. Kaolinite has very little isomorphous substitution and is non-expanding. Consequently, it has a very low CEC and all surface area is associated with the external surfaces of the structure. Hydrous mica/illite and the chlorites have a significant amount of isomorphous substitution; however as their interlayer regions contain either tightly bound K^+ or an additional hydroxide layer, respectively, which inhibits layer expansion. Thus, CEC is limited and the structures are non-expandable. Smectites have an intermediate amount of isomorphous substitution coupled with the ability to admit exchangeable cations into the interlayer region. Consequently, both CEC and total surface area is high. Soils containing smectites are noted for their high plasticity, cohesion and marked ability to swell when wet and shrink when dried. Vermiculites have more isomorphous substitution compared to smectites and thus have a higher CEC; however, most of the charge in vermiculites is concentrated within the tetrahedral sheets closer to the interlayer region and therefore has a greater electrostatic attraction for exchangeable cations in the interlayer region. As a result, although the total surface area of vermiculite is similar to the smectites, layer expansion of vermiculite structures is limited due to restricted entry of water into the interlayer region.

Table 14.8. Ranges of cation exchange capacity (CEC) and internal and external surface area in common phyllosilicates and other colloids. (Adapted from Weil and Brady, 2017)

Phyllosilicate/Colloid	**CEC (cmol(+) kg^{-1})**	**External Surface Area (m^2 g^{-1})**	**Internal Surface Area (m^2 g^{-1})**
Kaolinite	2-16	5-15	~0
Chlorite	10 - 40	5-20	~0
Hydrous Mica/Illite	20 - 40	100 - 125	~0
Smectite	60 - 120	80 - 140	570 - 660
Vermiculite	100 - 150	70 - 120	600 - 700
Amorphous aluminosilicates	5 - 350	500 - 1450	~0
Fe/Al	~0 - 3	200 - 500	~0
oxy-hydroxides			

Can You Dig It!

Although not confirmed, some data suggest that phyllosilicates could be a template to synthesize RNA from simple components and was the catalyst for the abiotic origin of life itself on this planet (Berman, 2019). Clay minerals act as very efficient catalysts in the polymerization of amino acids and nucleotides. RNA adsorbed to clay minerals can be encapsulated within vesicles. Once formed, such vesicles could grow and divide by incorporating fatty acids thus mediating vesicle replication through cycles of growth and division (Brack, 2013)

DISTRIBUTION OF PHYLLOSILICATES IN CANADIAN SOILS

The distribution of phyllosilicates in Canadian soil varies somewhat by physiographic region (Kodama, 1979); most phyllosilicates are inherited from the parent materials deposited by the retreat of the continental glaciers during the late Pleistocene. For example, some clay minerals such as kaolinite, mica, (i.e., biotite and muscovite) and chlorite are inherited directly from the Canadian Shield. Others in the Cordilleran region and Interior Plains Regions are derived from glacial degradation of sediments from the Cretaceous period (65 million years ago). Comparison of data for clay-size separates of soils, for example, from the Interior Plains of western Canada with those found in the eastern Canada (St. Lawrence Lowlands) show a distinct difference between the suite of clay minerals (Table 14.9). The Interior Plains region of western Canada is commonly reported to contain smectite, clay-size mica, chlorite and some kaolinite, (Kodama, 1979; Dudas and Pawluk, 1982; Spiers et al. 1989a; Spiers et al. 1989b; Warren and Dudas, 1992). In contrast, the clay-size fraction of soils from the Western St. Lawrence Lowlands (Ontario) derived from much older sediments formed during the Devonian (359 million) and Silurian (443 million) periods and are dominated by clay-size mica, chlorite, vermiculite and kaolinite with some reports of traces of smectite (Kodama, 1979). Parent materials in both regions had contributions from the Canadian Shield. There is only weak evidence of alteration in the diverse clay mineral assemblages from the various physiographic regions and correlation is futile between generic types of soil. An exception is the Podzolic soils and to a

lesser extent Brunisolic and Gleysolic soils where inherited chlorite minerals appear to weather to expandable phyllosilicates (Kodama, 1979).

Table 14.9. Range of abundances for phyllosilicates typically found in the clay-size fractions of soils of the Interior Plains and from the Western St. Lawrence Lowlands

Interior Plains Region			**St. Lawrence Lowlands**	
Phyllosilicate	**g kg^{-1}**		**Phyllosilicate**	**g kg^{-1}**
Kaolinite	70-110		Kaolinite	120-150
Discrete Muscovite	250-340		Discrete Muscovite	350-450
Montmorillonite	470-490		Vermiculite	180-220
Chlorite	120-160		Chlorite	240-300
Vermiculite	trace		Smectite	trace

Amorphous Aluminosilicates

Other hydrous aluminosilicates are also found in soil but are different structurally from the phyllosilicates. These are described as amorphous (lacking form or shape) or nanocrystalline and do not have a definite ordered crystalline structure like the phyllosilicates. Allophane and imogolite are examples of amorphous aluminosilicate minerals, commonly formed during rapid weathering in volcanic ash soils found in humid climates of countries such as New Zealand, Chile, and Japan. Volcanic ash weathers rapidly and typically dissolves releasing large amounts of soluble aluminum and silica, which can precipitate as allophane or imogolite. Amorphous aluminosilicates are non-expandable, but are extremely small (<4 nm diameter) with an extremely high specific surface area (700-1500 m^2 g^{-1}) and very high ion exchange capacities. Although they occur in abundance in volcanic ash soils, they may occur in small amounts in some soils along with phyllosilicates. Imogolite, for example, is documented in some Humo-ferric Podzolic soils in British Columbia (Grand and Lavkulich, 2013; 2015); elsewhere in Canada (McKeague and Kodama, 1981; Wang et al., 1991) and Solodized Solonetzic soils in Alberta (Spiers et al., 1984).

IRON AND ALUMINUM OXYHYDROXIDES

Oxyhydroxide clays (sesquioxides) are secondary oxides, hydroxides and hydrous oxide minerals of Fe, Al, and Mn. Some are well crystallized (e.g., gibbsite, hematite, goethite), while others are amorphous. These minerals are the usual source of red, yellow, brown, and blueish hues observed in soil. Crystalline oxide clays are composed of sheets of oxygen and/or hydroxyl groups in an octahedral arrangement with Fe, Al or Mn. They do not have tetrahedral sheets of Si in their structures. The octahedral structures of these minerals are held together by hydrogen bonding, therefore they are non-expanding. Iron oxides and hydroxides once called “limonite” were later found to actually be a mixture of Fe oxyhydroxide minerals.

These minerals do not carry identifiable permanent negative charges, but do have a high surface charge associated with surface hydroxyl groups. As such, the exchange capacity of oxyhydroxides is highly pH dependent with the variation in surface charges due to protonation of hydroxyl groups at low pH producing positive charges and dissociation of hydroxyl groups at high pH producing negative charges. Oxyhydroxide minerals are referred to as being amphoteric due to their variable CEC. These minerals typically have a net CEC at high pH and a net anion exchange capacity at low pH ranges.

Oxyhydroxides are usually found in small quantities in Canadian soils (Grand and Lavkulich, 2013; 2015; Spiers et al., 1984). They are highly reactive with a very high specific surface area. Presence of these minerals significantly influences soil properties such as ion adsorption and exchange. For example, Fe oxyhydroxides are capable of adsorbing and retaining large amounts of phosphate, molybdate and borate ions from the soil solution as well as scavenging metals such as Cu, Pb, Zn, Co, Cr, and Ni. Some common Fe oxyhydroxides are listed in Table 14.10.

Table 14.10. Some common secondary iron oxyhydroxides found in soil

Mineral	Formula	Notes
Goethite	α-FeOOH	Most common and abundant iron oxyhydroxides in soils. Yellow to brownish yellow. Very fine particle size. Precipitates quickly from Fe^{3+} in solution. Very small (<0.5 um) needle-like crystals.
Lepidocrocite	γ-FeOOH	Found in poorly drained soils as bright orange mottles but NOT common. Formed through oxidation of precipitated Fe(II) $(OH)_2$ in the absence of CO_2 at pH 5-7. Extremely small (0.1-0.5 um) lath shaped crystals.
Hematite	α-Fe_2O_3	Second in abundance to goethite. Prevalent in tropical soils. Forms through dehydration of goethite and ferrihydrite. Very small amounts impart a red colour to soil.
Ferrihydrite (aka hydrous ferric hydroxide)	$(Fe(III))_2O_3 \cdot 0.5H_2O$ but also $5Fe_2O_3 \cdot 9H_2O$	Highly variable composition. Precipitates directly from Fe rich oxygenated water or bacteria through biological activity. Reddy-orange. Forms nanocrystals. Metastable.

CARBONATES

Some common carbonate minerals found in soil are presented in Table 14.11. The majority of soil carbonates originate from primary sources, being inherited from sedimentary rocks, and some are secondary precipitates usually found in subsurface horizons. The crystal structure of many carbonate minerals reflects the trigonal symmetry of the carbonate ion which usually occurs in combination with Ca, Na, U, Fe, Al, Mn, Ba, Zn, Cu, Pb, and other elements. There are about 80 known carbonate minerals; however, by far most common carbonate minerals found in Canadian soils are calcite ($CaCO_3$) and dolomite ($(Ca,Mg)CO_3$) derived primarily from Paleozoic bedrock. Most carbonate minerals typically contain trace amounts of other carbonate related elements (see above) substituted into their structures.

Calcite is sparingly soluble, but solubility increases at lower soil pH values and is not usually found in soils with pH values less than 6.0. Calcite typically dissolves from the solum and is leached to become re-deposited in the lower soil profile, sometimes observed as secondary calcite nodules (Miller et al., 1987; Wang and Anderson, 2000). Aragonite and other calcium carbonate minerals are neither common nor stable in soils.

Table 14.11. Some common carbonate minerals

Mineral	Ideal Formula	Notes
Calcite*	$CaCO_3$	Main mineral component in limestone. Typically unstable and dissolves in the upper solumn but can precipitate in the C horizon usually under subhumid or dryer climatic regimes.
Aragonite*	$CaCO_3$	Derived from fossil shells of geological origin. Unstable and dissolves in soil.
Dolomite	$(Ca,Mg)CO_3$	Derived from sedimentary rocks (dolostone) of geological origin. Unstable and dissolves in soil.
Siderite	$Fe(II)CO_3$	Derived from sedimentary or metamorphic rocks. Unstable in soil but very slow to dissolve.
Magnesite	$MgCO_3$	Highly soluble in soil.
Trona	Na_2CO_3	Highly soluble in soil. Found mostly in Solonetzic soils at high pH (~9.5)
*Calcite and aragonite are polymorphs with the same chemical formula but different crystalline structures.		

Dolomite ($Ca,Mg(CO_3)_2$) is a common calcium-magnesium carbonate mineral found in soil derived from dolostone bedrock. Like calcite, dolomite dissolves from the upper profile, contributing calcium to the precipitation of secondary calcite in the lower profile while the magnesium component is mostly leached or absorbed by plant roots. Some Mg^{2+} may substitute for Ca^{2+} within secondary carbonates but the amount is minor (<10%), primarily occurring if leaching is restricted (St. Arnaud, 1979).

Carbonates impart a number of properties to soil that include:

- maintaining high base saturation;
- maintaining high exchangeable Ca and Mg;
- maintainng pH >7;
- adsorbing/precipitating PO_4^{3-} ions;
- adsorbing/precipitating metals (e.g., Al, Zn, Pb, Fe, Ni, Mn); and,
- stabilizing clays and soil structure.

PHOSPHATE MINERALS

Primary Phosphate Minerals

The average crustal abundance of P is reported as 1050 mg kg^{-1}, while the total content of P in soils is typically in the range of 400–800 mg kg^{-1}. Phosphate can easily combine with about 30 different elements forming over 300 minerals that contain P (Bleam, 2017). Some are primary phosphate-containing minerals derived from sedimentary and igneous rocks, but most are minerals containing phosphate that have PO_4 substituted onto their structures as a minor “impurity”.

The most common “pure” primary phosphate minerals are referred to by the group name “apatite” and given the general chemical formula: $Ca_5(PO_4)_3(F,Cl,OH)$. Apatite usually occurs usually as very small hexagonal crystals with a variety of colours including green, bluish green, brown, white, yellow, red, gray and others depending on impurities. Names for specific mineral for the three common end members of apatite are as follows:

- fluoroapatite $Ca_5(PO_4)_3F$
- chloroapatite $Ca_5(PO_4)_3Cl$
- hydroxyapatite $Ca_5(PO_4)_3OH$

In nature, these minerals occur as admixtures, usually dominated by one of the three components (F, Cl and OH) depending on the source. Rock phosphate or phosphorite generally contains the highest concentration of these minerals and is mined mostly for the production of phosphate fertilizers. In soils, apatite can originate from almost any igneous or metamorphic rock occurring mostly as accessory phases (Nezat et al., 2007) but rarely as a major mineral phase in rock. The mineral structure may also contain substitutions with a huge variety of ions including carbonate (CO_3^{2-}), arsenate (AsO_4^{2-}), V, Th, Pb, Sr, Ba, U, and rare earth elements.

Secondary Phosphate Minerals

Generally phosphate in soil reacts readily with Ca^{2+}, Al^{3+}, and Fe^{2+} and Fe^{3+} forming a variety of secondary precipitates (Table 14.12). This includes phosphate derived from primary mineral sources and phosphate fertilizers. Iron phosphates (vivianite and strengite) are more common in sub-oxic soils where dissolved iron is more plentiful. Aluminum phosphates (varisite) are most often associated with acidic soils where phosphate first precipitates as an amorphous intermediate that eventually transforms into varisite (Lindsay et al., 1959). Secondary calcium phosphates generally occur in calcareous soils containing limestone (mineral calcite) as a calcium source (Kar et al., 2012). Secondary brushite is commonly precipitated in calcareous soils as a result of placing high concentrations of fertilizer in bands for crop production (McLaughlin et al., 2011).

Table 14.12. Some common secondary phosphate minerals that occur in soil

Group	Mineral Name	Ideal Formula	Occurrence
Iron	Vivianite	$Fe(II)_3(PO_4)\cdot 2.8H_2O$	Flooded/wet soils
Phosphates	Strengite	$Fe(III)(OH)_2(H_2PO_4)$ or $FePO_2\cdot 2H_2O$	Flooded/wet soils
Aluminum Phosphates	Variscite	$Al(OH)_2(H_2PO_4)$ or $AlPO_4\cdot 2H_2O$	Acidic soils
Calcium Phosphates	Brushite	$CaHPO_4\cdot 2H_2O$	Calcareous soils

SULPHATES AND SULFIDES

Sulphide minerals are not present in well to imperfectly drained soils but may found in some poorly drained Gleysolic or Organic soils. The sulphide mineral pyrite (FeS_2) is very common primary mineral in mine wastes but can occur in shales (a sedimentary rock contributing to some soils) and as a precipitate in some tidal soils in response to highly reducing conditions. Pyrite once exposed to the atmosphere is easily oxidized releasing iron and sulphuric acid:

$$2FeS_2 + 7O_2 + 2H_2O \rightarrow 2Fe^{2+} + 4H^+ + 4SO_4^{2-}$$

As reduced iron (Fe(II)) in the sulphide structure is released, it oxidizes and precipitates as an oxyhydroxide. Oxidation of reduced S combines with water to produce a strong acid (H_2SO_4). Anthropogenic exposition of sulfide minerals during mining operations or other excavations is primarily responsible for acid rock drainage (ARD). Soil pH values less than 2.0 may result from the oxidation of sulphides. Sulphate derived from the oxidation of sulphides or other sources form gypsum (Table 14.13) usually in response to high pH and the presence of soluble Ca.

Table 14.13. Some sulphur containing minerals

Type	Mineral	Ideal Formula
Sulphides	Pyrite	FeS_2
	Pyrrhotite	$Fe_{(1-x)}S$ (x = 0 to 0.2)
	Sphalerite	ZnS
	Galena	PbS
	Chalcopyrite	$CuFeS_2$
	Arsenopyrite	FeAsS
	Pentlandite	$(Fe,Ni)_9S_8$
	Cinnabar	HgS
	Marcasite	FeS_2
Sulphates	Gypsum	$CaSO_4 \cdot 2H_2O$
	Barite	$BaSO_4$

SUMMARY

In summary, the parent material of most minerals soils of Canada was deposited by the physical action of glaciers. Most minerals in these soils are derived either from the Canadian Shield or reworked and redeposited from other sources during glaciations in the millennia prior to the late Wisconsin glaciation. Primary minerals composing the majority of the sand-size fraction of Canadian soils are usually dominated by quartz and feldspars, with a minor quantity of heavy minerals. Chemical weathering of the sand size fraction contributes components to the formation of secondary minerals and to nutrients for plant uptake. Phyllosilicates dominate the clay-size fraction and constitute the most chemically active inorganic phase in soil. Differences between the phyllosilicate composition of the clay-size fraction found in regions of western and eastern Canada regions are due primarily to the source of the parent materials. Silt, intermediate in size range between the sand-size and clay-size fractions, contains a physical mixture of minerals from these fractions.

REFERENCES

Ando, S. 2020. Gravimetric separation of heavy minerals in sediments and rocks. Minerals. 10(3): 273

Berman, J.J. 2019. Evolutions clinical guidebook. Translating ancient genes into precision medicine. Academic Press. 357pp.

Beutelspacher, H., and H. Van Der Marel, 1968. Atlas of electron microscopy of clay minerals and their admixtures. A picture atlas. Elsevier. 333pp.

Bleam, W.F. 2017. Soil and environmental chemistry, 2nd Edition. Academic Press. 573pp.

Brack, A. 2013. Clay minerals and the origin of life. Chapter 10.4. Developments in Clay Science. Elsevier. 507-521pp.

Bowen, H J M, 1979. Environmental chemistry of the elements. Academic Press, London, New York.

Cousineau, P.A. 2020. Personal Communication.

Dudas, M.J. and Pawluk, S. 1982. Re-evaluation of the occurrence of interstratified and other phyllosilicates in southern Alberta soils. Canadian Journal of Soil Science. 62: 61-69.

Fanning, D., Rabenhorst, M. and R. Fitzpatrick. 2017. Historical developments in the understanding of acid sulfate soils. Geoderma 308: 191-206. http://dx.doi.org/10.1016/j.geoderma.2017.07.006

Finkl C.W. 1981. Soil mineralogy. In: Mineralogy. Encyclopedia of Earth Science. Springer, Boston, MA. https://doi.org/10.1007/0-387-30720-6_133

Grand, S and L.M. Lavkulich. 2013. Potential influence of poorly crystalline minerals on soil chemistry in Podzols of southwestern Canada. European Journal of Soil Science 64(5):651-660.

Grand, S and L.M. Lavkulich. 2015. Short-range order mineral phases control the distribution of important macronutrients in coarse-textured forest soils of coastal British Columbia, Canada. Plant and Soil. DOI 10.1007/s11104-014-2372-6

Havlin, J. L, Tisdale S.L, Nelson W.L., Beaton, J.D. 2005. Soil fertility and fertilizers: An introduction to nutrient management, 7th edition. Pearson. 528pp.

Kar, G., Peak, D. and J. Schoenau. 2012. Spatial Distribution and Chemical Speciation of Soil Phosphorus in a Band Application. Soil Science Society of America Journal 76(6):2297-2306. DOI: 10.2136/sssaj2012.0146

King, H.M., 2020. Feldspar. Geology.com. https://geology.com/minerals/feldspar.shtml

Kodama, H. 1979. Clay minerals in Canadian soils; their origin, distribution and alteration. Canadian Journal of Soil Science. 59: 37-58.

Kohut, C.K. and Warren, C.J., 2002. Chlorites. In Amonette, J.E. Bleam, W.F. Shultz, D.G. and J.B Dixon (eds.). Soil Mineralogy with Environmental Applications. Soil Science Society of America Book Series. No. 7, Madison, WI. 531-553pp.

Lindsay, W.L. Peech, M. Clark, J.S. 1959. Solubility criteria for the existence of variscite in soils. Soil Science Society of America Journal. 23(5): 357-360.

Loganathan, P. 1987. Soil Quality, Physical properties of the soil. http://www.fao.org/3/AC172E/AC172E04.htm#ch4

Liu, C. Li, X. Xu, F. Huang, P.M. 2003. Atomic force microscopy of soil inorganic colloids. Soil Science and Plant Nutrition. 49(1): 17-23.

Maitre J., Bouchard K., Bédard L.P. 2019. Mineral grains recognition using computer vision and machine learning. Computers & Geosciences 130: 84-93.

McKeague, J.A. and H. Kodama. 1981. Imogolite in cemented horizons of some British Columbia soils. Geoderma 25(3-4): 189-197.

McLaughlin, M.J., McBeath, T.M., Smernik, R., Stacey, S.P., Ajiboye, B. and G. Guppy. 2011. The chemical nature of P accumulation in agricultural soils—implications for fertiliser management and design: an Australian perspective. Plant Soil 349:69–87.

Miller, J. J., Dudas, M. J. and Longstaffe, F. J. 1987. Identification of pedogenic carbonate minerals using stable carbon and oxygen isotopes, X-ray diffraction and SEM analyses. Canadian Journal of Soil Science. 67: 953-958.

Mindat. 2020. https://www.mindat.org/

Mineralogical Society of America (MSA) 1997-2020. http://www.minsocam.org/msa/collectors_corner/faq/faqmingen.htm

Mineralogical Society of America (MSA) 2004-2020. http://www.handbookofmineralogy.org/search.html?p=all

Nezat, C.A. Blum, J.D. Yanai, R.D. and Hamburg, S.P. 2007. A sequential extraction to determine the distribution of apatite in granitoid soil mineral pools with application to weathering at the Hubbard Brook Experimental Forest, NH, USA. Applied Geochemistry. 22(11): 2406-2421.

Parikh, S. J. & James, B. R. 2012. Soil: The foundation of agriculture. Nature Education Knowledge. 3(10): 2

Pauling, L. 1929. The principles determining the structure of complex ionic crystals. Journal of the American Chemical Society. 51(4): 1010-1026.

Rutter, N.W. and White, M. 2015. Glaciation. The Canadian Encyclopedia. https://www.thecanadianencyclopedia.ca/en/article/glaciation

Smart, P. and Tovey, N.K. 1981. Electron microscopy of soils and sediments: Examples. Oxford University Press. 400pp.

Soil Classification Working Group (SCWG), 1998. The Canadian system of soil classification. 3rd Edition. Agriculture and Agri-Food Canada. Publication 1646. 187pp.

Spiers, G.A., M.J. Dudas and L.W. Turchenek. 1989a. The chemical and mineralogical composition of soil parent materials in northeast Alberta. Canadian Journal of Soil Science. 69: 721-738.

Spiers, G.A., M.J. Dudas, K Muehlenbachs and L.W. Turchenek. 1989b. Mineralogy and oxygen isotope geochemistry of clays from surficial deposits in the Athabasca Tar Sands area. Canadian Journal of Earth Sciences. 21: 53-60.

Spiers, G.A., Pawluk, S., M.J. Dudas. 1984. Authigenic mineral formation by Solodization. Canadian Journal of Soil Science. 64: 515 532.

St. Arnaud, R.J. 1979. Nature and distribution of secondary soil carbonates within landscapes in relation to soluble Mg^{++}/Ca^{++} ratios. Canadian Journal of Soil Science. 59: 87-98.

Wang, D. and Anderson, D.W. 2000. Pedogenic carbonate in Chernozemic soils and landscapes of southeastern Saskatchewan. Canadian Journal of Soil Science. 80: 251-261.

Wang, C., Ross G.J., Torrance, J.K., Kodama, H. 1991. The formation of Podzolic B horizons and pedogenic imogolite as influenced by microrelief within a pedon. Geoderma. 50: 63-77.

Warren, C.J. and Dudas, M.J. 1992. Acidification adjacent to an elemental sulfur stockpile: I Mineral weathering. Canadian Journal of Soil Science. 72: 113-126.

Weil, R.R. and Brady, N.C. 2017. The nature and properties of Soils. 15th Edition. Pearson Education. 1086pp.

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How did I get interested in soils? Well, I grew up on a farm where, as a teenager, I spent a lot of time on a tractor doing tillage operations and had a lot of time to think. One of the questions I asked myself during long days in the field was: “What makes plants grow?” With that I spent the next several decades trying to understand soils to answer that question. The quest continues . . .

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3.2: Soil Health and Management

Laura Van Eerd, Kate Congreves, Melissa Arcand, Yvonne Lawley, and Caroline Halde

Learning Objectives

On completion of this chapter, students will be able to:

1. Define the term “soil health” and understand its components
2. Compare methods for quantifying soil health
3. Explain the connection between improving soil health and the enhancement of ecosystem services
4. Identify threats to soil health that causes soil degradation, and how certain practices armour the soil against threats, support diversification or ‘perennialize’ annual cropping systems, and supply increased carbon inputs
5. Name the 6C practices of soil health management and explain how each contributes to improved soil health

INTRODUCTION

The potential for soil degradation remains a constant threat to the resiliency and sustainability of agricultural and natural ecosystems. When soil is degraded so too is its health. Efforts to improve or maintain soil health will not only benefit growers or land managers, but will also benefit the environment—a win-win scenario made possible because healthy soils go hand-in-hand with ecosystem services (Table 15.1). In reading this Chapter, you will learn about the history of soil health, its definition, how it relates to soil ecosystem services, and how it is measured and interpreted. In terms of improving soil health, we propose the 6C practices of soil health management as positive actions to improve soil health (Figure 15.1). They are dubbed 6Cs because they start with the letter C, emphasizing the importance of soil carbon and its influence on soil health. Here, you will learn about each of the 6Cs and how they influence soil health. You will quickly see that the concept of soil health pulls together much of what you learned in the other chapters, and the 6Cs provide you with management options that you can apply in your field, garden, or yard to improve soil health and plant productivity.

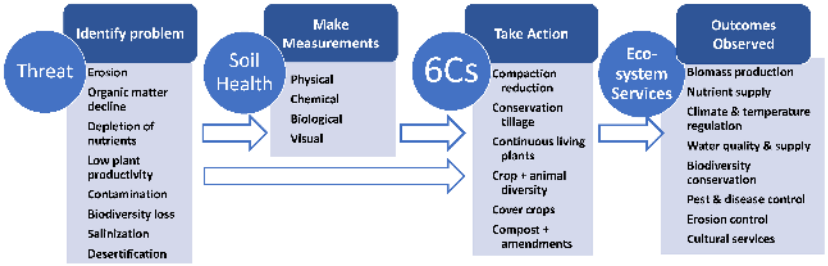


Figure 15.1. Schematic path of identifying a soil threat, measuring various soil health indicators, and implementing the 6Cs of soil health management that restore soil health and simultaneously enhance various ecosystem services. Adapted from the Soil Health Institute by Laura Van Eerd. © Soil Health Institute is licensed under a CC BY (Attribution) license.

THE IMPORTANCE OF SOIL HEALTH IS LONG RECOGNIZED

The soils in Canada are young, having developed over the 10,000 years or so since the retreat of the last glaciers (see Chapter 2). Canadian society has benefited from the fertility inherent to these young soils, still relatively rich in nutrients and organic matter, that are spread across our present-day landscapes. Still, Canadian farmers and land managers have had to keep a keen eye on how anthropogenic activities have degraded soils over the decades and centuries since the large-scale clearing of forests and native grasslands for resource use and extraction. We have had to respond by adapting new technologies and management strategies that are best suited to the specific climate and soil conditions present within any geographic location, so that the **soil functions** and capacity to provide ecosystem services are maintained—that the soil is healthy.

Concepts of soil health and the awareness that human actions affect the soil’s functions can be traced back in the written record for millennia. In Canada, those written records are not so old, but the Indigenous people and societies that flourished here had an awareness of the central role soils played in supporting their livelihoods. Prior to European settlement, Indigenous people managed

the landscape in myriad ways that directly and indirectly influenced soil properties and functions to serve their needs in promoting food provisioning.

For example, agriculture was present in the Great Lakes region prior to European settlement, where the Haudenosaunee people practiced Canada's first **intercropping** systems by planting squash, corn, and beans together. By blending these complimentary plant growth forms together, this practice helps support plant productivity by way of tending to soil nutrients and emphasizes the reciprocity between people and the soil: not only taking from the soil, but also giving back. Centuries later, intercropping is viewed as a reemerging soil health building strategy—and is an example of how plant diversification supports soil health. The linkage among people, soils and plants also extends to animals. On the Canadian Prairies, bison played a central role in shaping grassland ecosystems. These landscapes included Indigenous people that practiced prescribed burning to control woody species while maintaining palatable vegetation for bison grazing (Savage, 2011). Without the human influence on these animal-plant-soil interactions, would the geographic range of the organic matter-rich grassland soils, the Chernozems (see Chapter 10), that support Canadian Prairie agriculture exist as we know them today?

Can You Dig It!



Kirsten Kurtz. © Laura Van Eerd is licensed under a CC BY (Attribution) license.

Kirsten Kurtz, manager of the Cornell Soil Health lab, displays a poster of the '*Three Sisters in Soil*' painting (Left) made by Kurtz and students using only soils. This painting won Food and Agriculture Organization (FAO) of the United Nations' first global soil painting competition to honour World Soil Day (December 5th). Notice the baskets of corn, beans and squash (the three main agricultural crops of various Indigenous groups in North America) and the multitude of colours all derived from soil. Who knew soil could be so colourful?

In the *Three Sisters* planting method, three crops benefit from each other. The corn provides a structure for the beans to climb, eliminating the need for poles. The beans provide the nitrogen to the soil, and the squash spreads along the ground, blocking the sunlight and helping prevent the establishment of weeds.

For more info see: https://blogs.cornell.edu/hort/2018/01/12/three-sisters-in-soil-wins-global-soil-painting-competition/

DEFINING SOIL HEALTH

Today, the terms soil quality and soil health are commonly used as synonyms — distinguished mostly by preference rather than meaning. The most widely cited definition is that by Doran et al. (1996) – *soil health/quality is the capacity of a soil to function within ecosystems and land-use boundaries to sustain: (i) biological productivity, (ii) environmental quality, and (iii) plant and animal (including human) health*. As such, it invokes the linkage between soil functions and soil ecosystem services (Bünemann et al., 2018). **Soil functions** represent "what the soil does" or "how the soil behaves", rather than describing what the soil is. Soil functions are bundles of soil attributes and processes, which tie together to serve a role. The soil's capacity to function powers the extent to which ecosystem services are provided. Ecosystem services, many of which are directly and indirectly provided by soils, support the very survival and quality of all life (Table 15.1, Figure 15.2). Given the complex relationship soils have to ecosystem services, it is no surprise that pinpointing exactly what represents a healthy soil has been challenging. No single soil measurement quantifies soil health, rather it must be inferred from a multitude of soil attributes, processes, and contexts.

Table 15.1. Simplified linkages between a healthy and functioning soil and ecosystem services with example indicators

Soil functions/roles	Ecosystem service	Goods provided by soil ecosystems	Example soil health indicators
Habitat provisioning, regulation of biological populations; element and organic matter cycling; water cycling	Biomass production	Food, feed, fuel, fibre, and other materials	Primary productivity
			Organic matter and carbon
			Depth of A horizon
Element and organic matter cycling; water cycling	Nutrient supply	Supply and flow of nutrients and energy	Organic matter and carbon
			Active carbon
			Potentially mineralizable nitrogen
			Nutrient concentrations
Element and organic matter cycling; water cycling	Climate regulation	Capacity to store carbon and regulate greenhouse gas emissions, and mitigate climate change	Organic carbon
			Respiration
			Total nitrogen
Habitat provisioning, and regulation of biological populations	Temperature regulation	Mitigates against extreme temperatures fluctuations	Water holding capacity
			Organic matter
Water cycling; element and organic matter cycling; soil structure maintenance	Water quality & supply	Vegetation and edaphic factors regulate moisture infiltration, percolation, storage, flow, filtration and the quality of water	Available water capacity
			Infiltration rate
			Penetration
Habitat provisioning, and regulation of biological populations; element and organic matter cycling	Biodiversity conservation	Support flora and fauna, and maintain the viability of gene pools	Phospholipid fatty acid analysis
			qPCR
			Genomics
Habitat provisioning, and regulation of biological populations; water cycling	Pest & disease control	Regulate the balance and structure of communities and	Genomics

Soil functions/roles	Ecosystem service	Goods provided by soil ecosystems	Example soil health indicators
		populations, and the activity of predators and prey	
			Root assessment
Soil structure maintenance	Erosion control	Mitigate damage such as flooding or landslides	Aggregate stability
			Texture
			Organic matter
Habitat provisioning	Cultural services	Preserve benefits people derive from ecosystems in addition to those mentioned above, such as spiritual enrichment, development, recreation and aesthetic values	Primary productivity

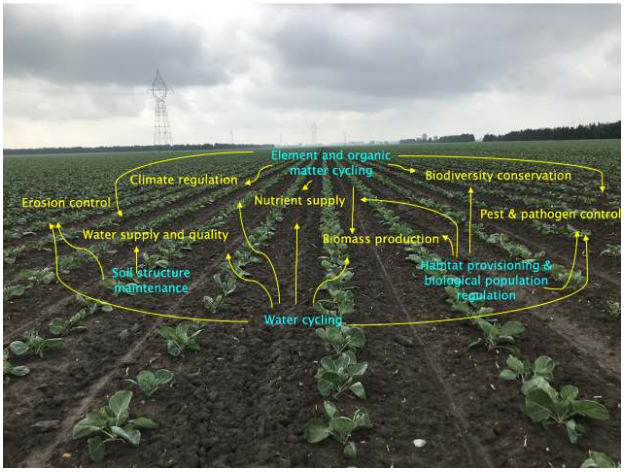


Figure 15.2. Linkages between soil functions (in blue) and ecosystem services (in yellow) © Kate Congreves, Univ. of Saskatchewan is licensed under a CC BY (Attribution) license.

SOIL HEALTH MEASUREMENT AND INTERPRETATION

Soil health research has dramatically expanded in the past three decades (Figure 15.3), and with knowledge comes use. This body of knowledge has largely fueled the uptake of soil health interest and testing by growers and land managers, worldwide. People are now using soil health tests to compare and track soil health measurements as they implement new practices aimed at improving agricultural, ecosystem, and soil resiliency. However, soil health measurements vary greatly and there is no global consensus. Multitudes of approaches exist for monitoring and deciphering a soil's health status, ranging from narrow and simple to multifaceted and complex (Figure 15.4). Approaches are chosen based on sampling ease, measurement criteria, analytical laboratory facilities, reliability and cost. Regardless of the differences, a single rule of thumb applies: *for a soil health test to be meaningful, it must target a specific and relevant soil function or ecosystem service.*

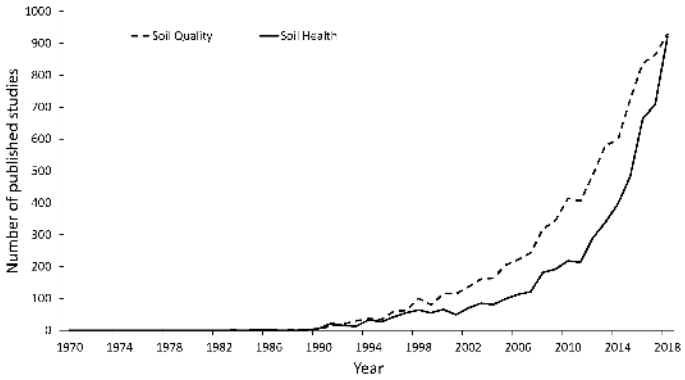


Figure 15.3. Since 1970, number of the studies focusing on the evaluation of soil quality and soil health in agroecosystems. © Inderjot Chahal is licensed under a CC BY (Attribution) license.

Visual Assessments

There is great value in soil education, and visual assessments of soil health indicators offer an opportunity to learn. Qualitative indicators that can be assessed in the field can provide experiential learning experiences and promote communication between growers or land stewards and specialists. By frequently paying close attention to the soil, visual assessments are a key part of land management planning and refining.

In fields, visual soil assessments generally require little time (less than an hour), minimal tools (a spade or shovel, and water), and offer immediate results. It can be said that visual soil health assessments focus primarily on surface soil (unless pits are dug for a profile assessment) and on physical attributes, compared to biological and chemical attributes. For example, by visually examining a soil sample, one can coarsely evaluate soil texture, aggregation, structure, compaction, and water infiltration—all soil physical components. With a bit more effort and access to simple tools, visual interpretations about chemistry (soil pH) and biology (respiration, earthworm counts) can be done. Some practitioners use reference cards to examine color, combine visual assessments with smell, or look at their soil samples under a microscope. The photo montage here shows visual assessments that rely on color change (Figure 15.4A and 15.4B), and expert knowledge (Figure 15.4C).



Figure 15.4. Various types of visual assessment of soil health: **(A)** Solvita soil respiration ‘paddle’ changes colour indicating greater CO_2 concentrations, **(B)** a soil protein test works by measuring the colour change with a spectrophotometer after the chemical reaction test in the lab, **(C)** Dr. Catherine Fox, Research Scientist at Agriculture and Agri-Food Canada, Harrow, ON conducts a structural assessment. Photo credits: © Kate Congreves, Univ. of Saskatchewan; Laura Van Eerd, Univ. of Guelph is licensed under a CC BY (Attribution) license.

To help with visual soil health assessments, score cards have been developed as a simple tool to guide soil, plant, water, and animal observations on quality. Some score cards also provide a visual key to help minimize misinterpretations by non-experts of soil science.

Not all visual-based soil assessments are low-tech—soil scientist are using the synchrotron or computerized axial tomography (CAT) scans as visual assessment techniques. The same CAT scans used in medicine have been used in soil to measure porosity, connectivity and crusting.

Despite the benefits of visually examining the soil, most visual assessments alone cannot evaluate the degree to which soil ecosystem services are performing—other than possibly erosion control, as inferred from soil physical attributes. As such, visual soil assessments may be more useful in inferring soil health status when combined with other analytical techniques.

Soil Health Tests

Analytical soil tests offer more precise information on soil physical, chemical, and biological pillars of soil health than visual examinations can offer. A list of common soil attributes that are measured via lab-based analytical approaches for each of the three pillars are listed in Figure 15.5, and most of which have already been discussed in earlier chapters of this book (see Chapter 4, Chapter 5, and Chapter 6). Having soil tests can be helpful for management planning; for example, many certified labs offer recommended fertilizer quantities based on soil nutrient levels and growers' yield goals.

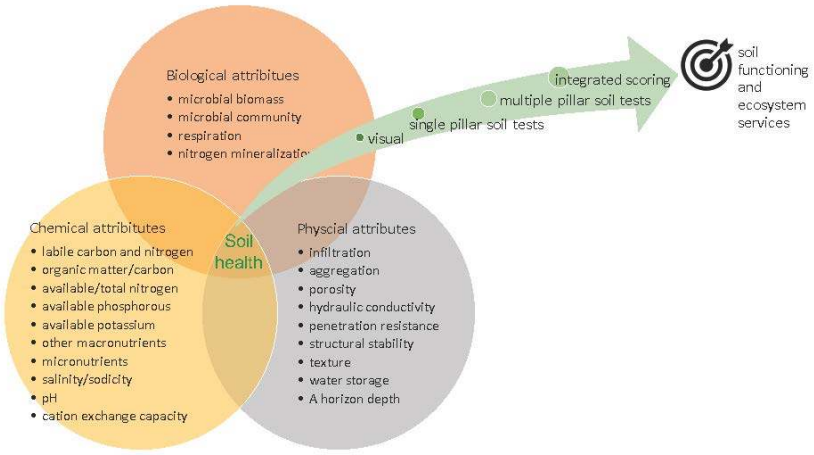


Figure 15.5. Holistic measurements and interpretations of soil health relies on three pillars of soil science: biology, chemistry, and physics. Soil attribute data should be linked with soil function and ecosystem service performance to interpret soil health in a meaningful way © Kate Congreves, Univ. of Saskatchewan is licensed under a CC BY (Attribution) license.

Currently available commercial soil health tests offer insight into one or more soil attributes (Table 15.2). These commercial tests are often biology- or chemistry-focused tests that quantify soil attributes and may be useful for monitoring trends over time and from one location to another. However, all three pillars of soil health should be considered for a holistic interpretation of a soil's health status. With this in mind, it must be acknowledged that increasing the number of attributes can increase the complexity of relationships between indicators, which can complicate interpretation and therefore management recommendations. Ideally, soil attribute data should be linked with soil function and ecosystem service performance to understand soil health in a more meaningful way.

Table 15.2. Soil health indicators that are included in various soil health tests. Which indicators would your soil health test include?

Soil Health Indicator	Comprehensive Assessment of Soil Health (CASH)	Soil Health Management Assessment Framework (SMAF)	NRCS Soil Test Kit (NRCS)	Haney Test (Haney)	Soil Health Institute Evaluation (SHI)	Soil Fingerprint Framework (AAFC)
Physical						
Texture	X	-	X	-	X	X
Aggregate stability	X	X	X	-	X	proposed
Penetration	X	-	X	-	X	-
Available water capacity	X	X	X	-	X	-
Saturated hydraulic conductivity	-	-	-	-	X	proposed
Bulk density	-	X	X	-	X	X
Infiltration	-	X	X	-	-	-
Erosion rating (RUSLE)	-	-	X	-	X	-
Slaking	-	X	X	-	-	-
Crustings	-	X	X	-	-	X
Structure + macropores	-	X	X	-	-	X
A horizon depth	-	-	X	-	-	X
Biological						
Plant growth/yield	-	-	X	-	X	proposed
Root assessment	X	-	X	-	-	development phase
Potentially mineralizable N	X	-	X	-	X	-
Protein	X	-	-	-	X	-
Enzymes	-	X	X	-	X	-
Microbial Biomass	-	-	-	-	-	-
Earthworms	-	X	X	-	-	development phase
NH3 released	-	-	-	X	-	-

Soil Health Indicator	Comprehensive Assessment of Soil Health (CASH)	Soil Health Management Assessment Framework (SMAF)	NRCS Soil Test Kit (NRCS)	Haney Test (Haney)	Soil Health Institute Evaluation (SHI)	Soil Fingerprint Framework (AAFC)
(SLAN)						
CO2 evolved (respiration)	X	X	X	X	X	-
Phospholipid Fatty Acid (PLFA)	-	-	-	-	X	-
Ester-Linked Fatty Acid Methyl Ester (EL-FAME)	-	-	-	-	X	-
Genomics	-	-	-	-	X	proposed
Reflectance	-	-	-	-	X	-
Chemical						
pH	X	X	X	-	X	X
Electrical Conductivity (EC)	-	X	X	-	X	X
Cation exchange capacity	-	-	-	-		-
Organic matter	X	X	X	-	X	X
Organic carbon	-	X	X	-	X	X
Active carbon	X	X	X	-	X	-
Water extractable organic C+N (WEOC, WEON)	-	-	-	X,X	-	-
Total Nitrogen	-	-	-		X	-
Nitrate-Nitrogen	-	X *Leaching	X	X	-	proposed
Ammonium-N	-	-	-	X	-	proposed
Extractable P, K	X,X	-	-	X,X	X,X	-
Extractable Ca, Mg	X,X	-	-	X,?	X,X	-
Extractable Mn, Zn	X,X	-	-	-	X,X	-
Extractable Na	-	-	-	-	X	-

Soil Health Indicator	Comprehensive Assessment of Soil Health (CASH)	Soil Health Management Assessment Framework (SMAF)	NRCS Soil Test Kit (NRCS)	Haney Test (Haney)	Soil Health Institute Evaluation (SHI)	Soil Fingerprint Framework (AAFC)
Extractable Fe, Cu	X,?	-	-	X *H3A	X,X	-
Salinity and Sodicity	X	-	-	-	X	-
Soil forming factors						
Environment and genetic soil formation processes	-	-	-	-	X	X
Slope (kind, position, gradient)	-	-	X	-	X	X
Surface conditions (incl. soil management)	-	-	X	-	X	X
Land use (incl. crop type)	-	-	X	-	X	X
Reference	Idowu et al. (2009)	Andrews et al. (2004)	Sarrantonio et al. (1996)	Haney et al. (2010)	Norris et al. (2020)	Fox et al. (2014)

Integrated Scores of Soil Health Tests

To be useful, soil health indicators must be clearly interpreted and ideally expressed relative to reference values. Reference values for soil attributes can be derived from an undisturbed soil (not impacted by anthropogenic activity—*which may be difficult to find, or otherwise not reflective of optimum soil functioning*) or a soil with superior primary productivity and environmental performance. As a framework for soil health scoring, soil health indicators at any given site can be compared to either a reference soil, or to the statistical distribution frequency of the given indicator (Figure 15.6) from a range of samples under comparable land-uses or regions.

A multifaceted approach to soil scoring involves the establishment of standard non-linear scoring functions, which have three general shapes (Figure 15.6) and represent indicators where: (i) more is better (e.g., soil organic matter), (ii) optimum is best (e.g., nutrients, pH), and (iii) less is better (e.g., salinity). The shape of the curve for each soil indicator is informed by expert opinion and/or literature. Each soil attribute value is transformed into a relative score based on where it falls on the scoring curve—as derived from the distribution frequency for a particular region or land-use. As such, when scoring curves are based on regional datasets – like the Cornell University's Comprehensive Assessment Soil Health (CASH), the resulting scores (0 to 100) indicate the rank relative to that particular region. When integrating the scores of many soil attributes, all the scores are averaged. The higher the score, the better the soil health.

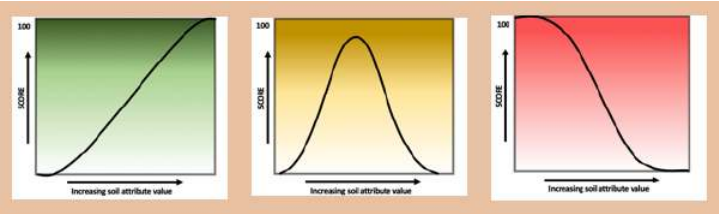
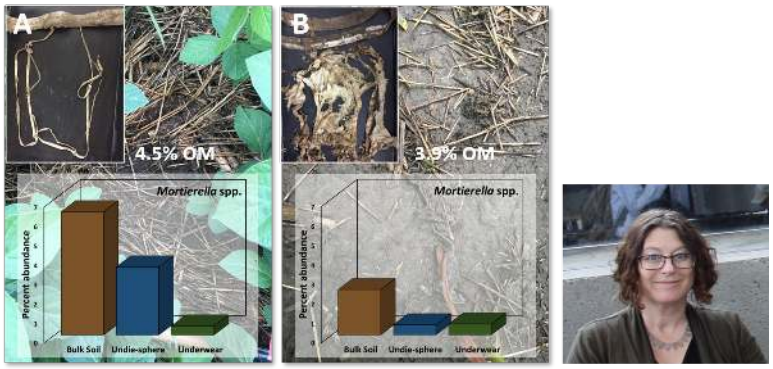


Figure 15.6. Soil health scoring curve scenarios: more is better (left), optimum is best (middle), less is better (right) © Adapted from Moebius-Clune et al. 2016 adapted by Athena Wu is licensed under a CC BY (Attribution) license.

Can You Dig It!

Soil Your Undies

Well that gets your attention. *Soil Your Undies* started by the Innovative Farmers Association of Ontario in 2017 to encourage people to learn more about their soil (http://www.omafra.gov.on.ca/english/premier_award/2017/events/h/ifao.htm). By burying cotton underwear in soil for up to 2 months, you can see how biologically active your soil is. Here's how it works. Cotton is plant-based and thus carbon-based. Soil fauna and microbes feed on carbon, so the more degraded your underwear the more active your soil. Collaborating with Ontario Ministry of Agriculture, Food and Rural Affairs (OMAFRA) Soil Management Specialist Anne Verhallen, soil microbiologist Dr. Lori Phillips (lower Right) quantified the microbial diversity associated with buried underwear. She extracted DNA from soil, the soil attached to the buried undies, the "undie-sphere", and from the buried undies. She used a combination of sequencing and quantitative PCR to examine microbial life in these soils. The figure below shows two soils from one farm in southwest Ontario with similar management (conservation tillage, 3-crop rotation and cover crops). Underwear decomposed more rapidly in biologically diverse soils, which had up to 15% more taxa than soils where underwear remained 'wearer-friendly'. These soils also had a higher abundance of fungi that specialize in breaking down carbon compounds (saprotrophs) as well as a high abundance of bacteria that free up the nitrogen these fungi require. For example, she observed that carbon decomposers (e.g., saprotrophic Mortierella fungi) and bacterial nitrifiers (such as Nitrosomonas spp.) were twice as abundant in soils where the carbon-rich undies decomposed completely in 2 months (Panel A) compared to where decomposition was incomplete (Panel B).



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Pretty nifty what your underwear can tell you. To give it a try, see the exercise at the end of this chapter.

SOIL HEALTH MANAGEMENT

We know that the activities of humans impact the soil; the goal is to minimize threats to soil health that result in soil degradation (or poor soil health). The main degradation threats to soil health include: erosion, compaction, soil organic matter decline, depletion or excess of nutrients, biodiversity loss, salinization, desertification and contamination or pollution (Figure 15.1)—many of which have been discussed in previous chapters. In this chapter we focus on land management practices that exacerbate or ameliorate these soil problems. Practices that maintain or improve soil health will (i) protect the soil against threats, (ii) support diversification or 'perennialize' annual cropping systems, and/or (iii) build soil carbon (Table 15.3). Here, we introduce six practices that land managers can use to build soil health and thereby protect against soil degradation (Figure 15.7). Purposefully, these six practices start with the letter C to emphasize the role of carbon in soil health. An explanation of what each practice is and how it enhances soil health follows in the sections below. These 6C practices can be implemented in the field, garden, yard or on any managed soil

—or can be combined together, for a positive influence on soil health (Figure 15.8). As you learn about each, it is very important to know that these practices in combination or partial combination will interact to influence aspects of soil health in various ways.

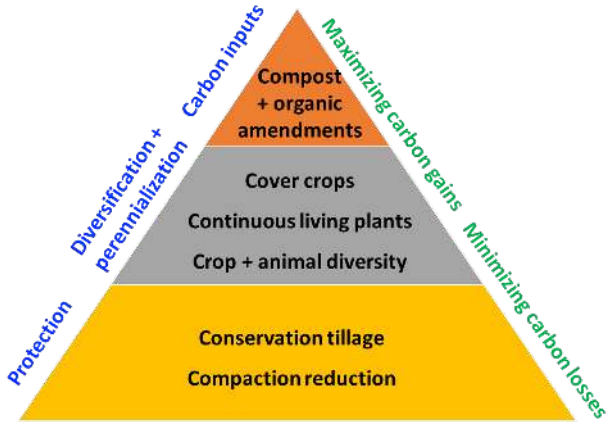


Figure 15.7. Schematic of the 6C practices of soil health management explaining how they enhance soil health (in blue on left) and by which mechanism (in green on right). © Kate Congreves (Univ. of Saskatchewan), Laura Van Eerd (Univ. of Guelph) is licensed under a CC BY (Attribution) license.

Table 15.3. The 6C management practices and principles that can be used to help achieve soil health

6C management practices	**Soil Health Principle**				
	Protection		**Diversity + Perennialization**	**Building Carbon**	
	Maximize soil cover	***Minimize soil disturbance***		***Carbon inputs + gains***	***Minimizing carbon + soil losses***
Compaction reduction		***		*	*
Conservation tillage	**	***		*	**
Continuous living plants	**	**	***	**	**
Cover crops	***		**	*	**
Crop +animal diversity			***	*	*
Composts and amendments	*			***	*

***principle goal and benefit; **expected benefit; *likely benefit

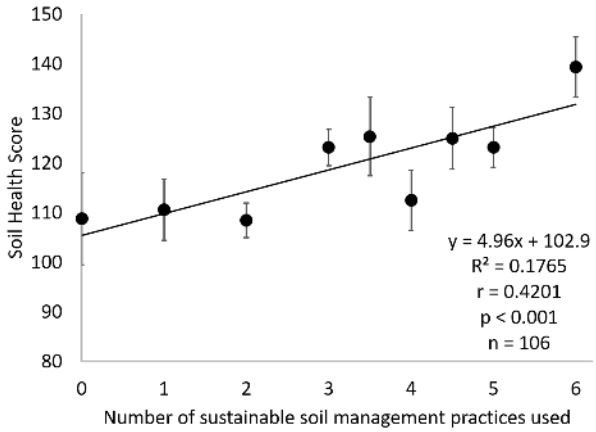


Figure 15.8. Positive relationship between the adoption of the 6C management practices and soil heath. Notice as more of the sustainable soil management practices are used soil health scores improve. Soil health was measured as commercially available Solvita labile amino-nitrogen (SLAN) on surface soil collected after 36 years of management at Elora, Ontario. © Laura Van Eerd is licensed under a CC BY (Attribution) license.

6C Strategies

6C–1: Compaction Reduction



What is soil compaction? As discussed in Chapter 4, soil structure and porosity are critical components of soil. Compaction occurs when a force is applied that presses soil particles together causing soil pores to collapse (i.e., pores are reduced in size and number), and bulk density to increase. As pores are reduced, the volume of water and air is reduced as is the space available for root growth. Soil compaction limits the rooting depth of plants and makes living conditions for the plant and microbial life in soil more challenging. Soil compaction can occur at the soil surface, at the depth of **tillage** (typically 20 cm), or even deeper in the soil depending on the type of equipment or management practices involved. As we will outline below, reducing soil compaction is an important but challenging step in improving soil health.

Why is reducing soil compaction important for soil health? As you recall from Chapter 6 on Soil Biodiversity, all of soil life happens in soil pores. The water and gases that plants and soil microorganisms depend on move through the soil pores. Compacted soil limits voids and pore space, movement of water and diffusion of gases. This can lead to saturated soils and if water cannot drain, then anaerobic conditions may result. All these conditions limit root and microbial growth. Adequate soil pore space is very important for soil health because it supports life.

Plant roots have a harder time pushing through compacted soil, more force is required (Figure 15.9). Think about a path through a green space on your campus where students are walking every day (Figure 15.10). When the soil is wet, the force of your footsteps (and thousands of others) compress the soil and cause compaction. Next time you walk by, take a look at the appearance of the soil and plants in the path and compare conditions a few feet away from the path. Perhaps the grass has stopped growing in the path and only "weedy" plants that have adapted to growing in compacted soil conditions are present. The next question is what will campus gardeners do to solve this soil health problem?

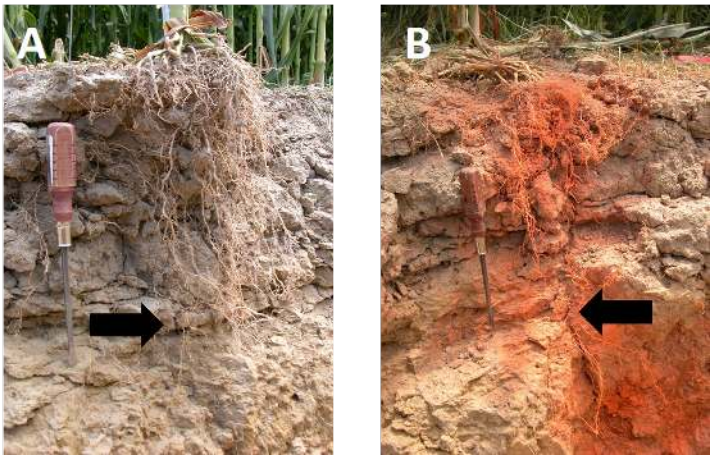


Figure 15.9. **(A)** Corn roots are restricted from growing beyond the compacted layer (indicated by the arrow) preventing roots from accessing water and nutrients deeper in the soil profile. **(B)** Corn roots (painted orange) are able to grow through the layer of soil compaction because deep tillage with a chisel plow created a crack through the compaction layer (indicated by the arrow) enabling roots to grow deeper into the soil profile. © Yvonne Lawley, University of Manitoba is licensed under a CC BY (Attribution) license.

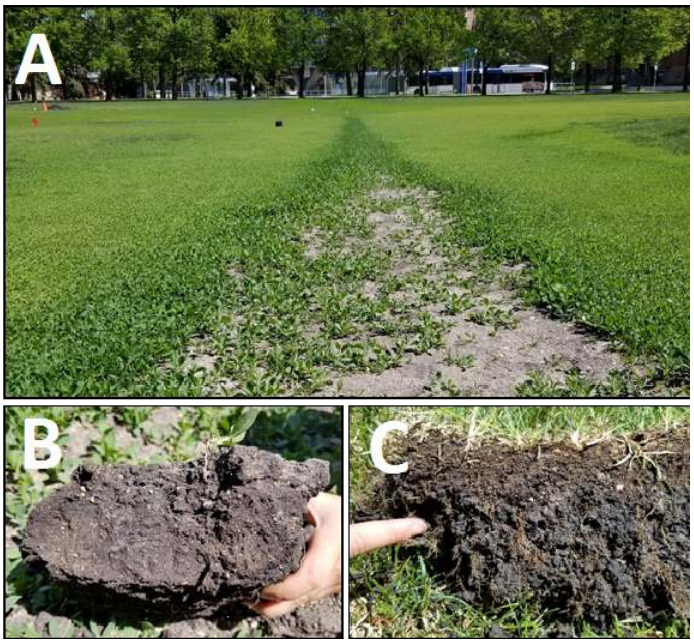


Figure 15.10. **(A)** Soil compaction from students walking across a green space on campus. **(B)** Soil in the compacted walkway had poor massive structure where few plants could grow. **(C)** Soil a few feet away from the path had soil aggregates, granular structure, and roots from healthy grass plants growing. © Yvonne Lawley, University of Manitoba is licensed under a CC BY (Attribution) license.

How to reduce and alleviate soil compaction?

Most would say that avoiding soil compaction in the first place is the best solution. That means avoiding traffic in fields or on your campus green space when soils are wet. After soil compaction occurs, something needs to happen to restore pores (voids) that were lost. In many parts of Canada, the freezing and thawing acts to **loosen** the soil and help alleviate soil compaction. Where this does not occur or if soil compaction occurs deeper in the soil, then a “deep ripper” tillage implement, such as a **chisel plow**, is often used to fracture or break up compacted soil layers (Figure 15.9B, 15.11A). Deep tillage, called **subsoiling**, with a chisel plow is expensive because it takes a lot of fuel and time. Tillage also needs to happen at a time of the year when the soil is dry enough to fracture and break up the plow pan (compaction layer), rather than smear and create a new deeper layer of soil compaction if soils are too wet. If the management practices that caused soil compaction in the first place do not change after the subsoiling, then action to alleviate compaction will likely be needed again after a few years.

An alternative approach to alleviating soil compaction is using plant roots to create channels (bio-pores) through layers of soil compaction (Figure 15.11B). This is called biodrilling or biological tillage (Cresswell and Kirkegaard, 1995; Chen and Weil, 2010). After the plants die and decompose, if these bio-pores are not disturbed through tillage, plants that grow the following year will grow through them to reach the soil underneath the compaction layer (Figure 15.12).



Figure 15.11. **(A)** A chisel plow is a tillage implement that is used to break through compacted soil layers and manage crop residues. **(B)** Cover crops like forage/diakon radish are being used for biodrilling to create root channels through compacted soil layers. Notice at the compacted layer how roots grow horizontally in the path of least resistance © Yvonne Lawley, University of Manitoba is licensed under a CC BY (Attribution) license.

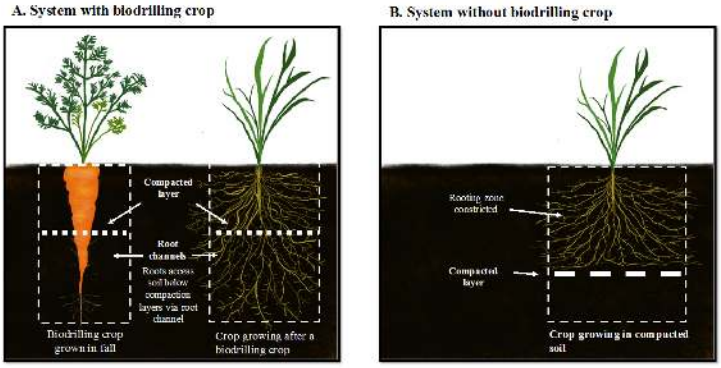


Figure 15.12. **(A)** By growing cover crops when there is adequate soil moisture, roots can biodrill channels through soil compaction layers that are used by the roots of subsequent crops to access the soil below. **(B)** In a system with a biodrilling crop, roots are restricted by the compacted layer © Figure created by Athena Wu (conceptualized by Yvonne Lawley and Kate Congreves) is licensed under a CC BY (Attribution) license.

You may be asking yourself the question, if soil compaction is bad for root growth then why can we use plant roots to solve this problem? There are three key factors that make biodrilling possible. First, plant roots must be growing at a time of the year when soils are moist and it is easier for plant roots to push through the tillage pan. If you have ever tried to push a stick into soil, you would know that more force is needed when the soil is dry compared to when the soil is wet. Second, not all plant roots have the same ability to push through compacted layers (Chen and Weil, 2010). Some plants have more root pressure than others to push through root restricting layers. For example, tap-rooted plants such as forage radish (Photo 11.11B), commonly used as cover crops had about twice as many roots as fibrous-rooted cereal rye cover crops at the 15–50 cm depth under compacted soil conditions (Chen and Weil, 2010). Third, leave these macropores intact by avoiding tillage in order to accumulate more and more root channels over time.

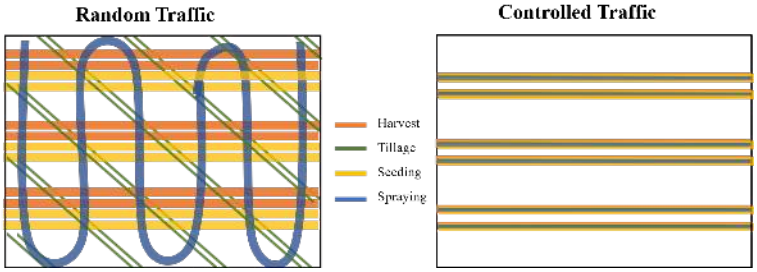
Why is it hard to reduce soil compaction? Think about your walk across that green space on campus. You should be walking on the side walk, especially when the soil is wet after a rain, but you take a short cut because you are late and you need to get to class. Or maybe it's just habit and you don't even think about it. It's the same with farmers or gardeners or landowners. Sometimes soil conditions can be too wet when something important needs to be done in the field. An increase in farm size plays a role, as having more ground to cover often means that in order to complete field operations in a timely manner, field activities like seeding, spraying, or harvesting may occur when conditions are too wet. As the weight of farm equipment has been increasing over recent decades so too has the force exerted on the soil. Monitoring axel loads and tire inflation on field equipment as well as avoiding field activities when soils are wet can help avoid soil compaction.



Adam Gurr (Brandon, Manitoba). © Adam Gurr is licensed under a CC BY-NC-ND license.

Controlled traffic farming is a new approach to addressing soil compaction. Controlled traffic farming limits where all field equipment is driven and intentionally concentrates soil compaction in permanent tramlines (think roadways but without gravel or pavement). This eliminates compaction because there is no wheel traffic in the rest of the field. Adam Gurr, who farms near Brandon, Manitoba, was concerned about the impact that soil compaction was having on soil health and plant performance on his farm. Adam now uses controlled traffic farming with global positioning system (GPS) to direct all of his farm equipment to travel on permanent tramlines in his fields.

Adam decided to adopt controlled traffic farming because it reduced soil compaction, reduced fuel consumption, and increased his ability to access fields (especially in wet years). He has measured the positive soil health impacts including quicker water infiltration, lower bulk density, and greater crop yields on his farm.



Comparison of field traffic patterns for random traffic and controlled traffic farming systems. © Athena Wu is licensed under a CC BY (Attribution) license.



(A) Seeding and **(B)** harvesting equipment travels in the same tramline tracks on Adam Gurr's farm near Brandon, Manitoba where he practices controlled traffic farming to improve soil health through compaction reduction. © Adam Gurr is licensed under a CC BY (Attribution) license.

Converting to controlled traffic farming may not work for every situation or farmer. The majority of farmers adopting it are already practicing reduced or no-till farming. It can also take several years to purchase or adapt all of your field equipment (tractors, sprayers, harvesters, grain carts) to have the same wheel width and be able to line up on tramlines.

6C–2: Conservation tillage



Soil tillage is the mechanical mixing of soil to various degrees. It is primarily used to prepare a good seedbed (Figure 15.13) ahead of seeding or transplanting, but also serves other purposes, such as: controlling weeds, breaking up soil compaction or crusting, facilitating faster decomposition of crop residues or organic amendments, and to incorporate amendments and soil-applied pesticides into the soil. Tillage can increase soil aeration, decrease soil moisture, and promote faster warming in the spring, leading to earlier or more uniform plant germination. Tillage also stimulates soil microbial activity including the breakdown of soil organic matter important in maintaining soil structure.



Figure 15.13. Compare these two fields that have been prepared for planting. **(A)** Thomas Dewavrin, observing his soil after ridge-tillage in late-April on his organic grain farm, located in Les Cèdres, QC. **(B)** Seeding into wheat residue in a no-till field. © Matthew Dewavrin; Yvonne Lawley, University of Manitoba is licensed under a CC BY (Attribution) license.

The method and timing of tillage will depend on soil type, cropping system, and grower goals. Although tillage is often used to solve problems (such as described in the compaction section), it can also create problems for soil health and plant productivity. By bringing crop residues deeper into the soil, tillage 'removes' soil protection that would otherwise protect the soil from water or wind erosion, thereby leaving the soil vulnerable to loss. Tillage also disrupts soil aggregates and arbuscular mycorrhizal fungi networks, both of which are important for maintaining soil structure in the long run. Tillage when soil conditions are not ideal (too wet or too dry) can lead to compaction, resulting in negative impacts on plant growth and soil health. To reduce these problems, farmers are increasingly reducing the number and intensity of tillage operations and adopting **conservation tillage** practices to conserve soil and water mainly by retaining at least 30% of the soil surface covered by plants or crop residues (Figure 15.14). In addition to the mechanism of improving soil health, we will outline the multitude of tillage practices and the reasons for regional and cropping system differences.

Types of conservation tillage: Conservation tillage is an umbrella term for many different forms of tillage that reduce the frequency or intensity of soil disruption, relative to conventional tillage or **plowing** (Figure 15.15). By reducing the number of passes over the field, tillage frequency can be reduced; by reducing the depth of tillage, tillage intensity can be reduced. Conventional tillage or plowing typically inverts the soil to approximately 20 cm depth. For more information on various tillage practices please see glossary and resources at the end of this chapter.



Figure 15.14. Tillage implements ranging from full inversion tillage **(A)** to various conservation tillage **(B –E)** to no-till **(F-G)**. Note the soil conditions and the amount of residue remaining on the soil surface. The goal of conservation tillage is at least 30% coverage. **(A)** Plowing. **(B)** Cultivator. **(C)** Conservation tillage implement with (from right to left; the direction of travel) discs/coulters, finger tines, and roller packer with **(C1)** resulting soil surface showing wheat residue. **(D)** Strip tillage implement designed to create tillage zones where crop rows will be planted and apply fertilizer, **(D1)** resulting soil conditions after stripping. **(E)** Disking corn residue. Note the minimal soil disturbance **(F)** No-till planter showing coulters to move residue away from the row. **(G)** roller crimper -an implement to terminate cereal rye without tillage. Photo credits: © "(A, B, E, F) – Yvonne Lawley; (C, C1, D, D1) – Laura Van Eerd; (G) – Caroline Halde (Université Laval)" is licensed under a CC BY (Attribution) license.

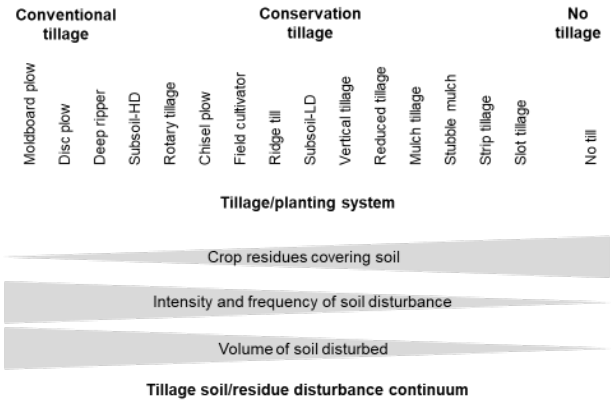


Figure 15.15. The frequency and intensity of soil disturbance under conventional or conservation tillage. Note that no-till is a type of conservation tillage. There can be varying degrees of disturbance from high (HD) to low (LD). © Adapted from Reicosky 2015 adapted by Julie Forest-Drolet is licensed under a CC BY (Attribution) license.

Context matters: Within Canada, the adoption of conservation tillage practices varies widely by region, and depends on climate, soil types, crop rotation, and cropping system. For example, **no-till** has been widely adopted by growers in drier regions (i.e., western Canada) because water is a limiting resource for crop growth, and growers use conservation tillage as means to conserve soil moisture for plant growth. But, in humid regions, such as British Columbia and eastern Canada, water is typically in excess in the spring, so many growers use tillage, mainly in the fall, to hasten soil drying so they can plant sooner. Consequently, the adoption of conservation tillage practices is relatively lower in eastern Canada than the western counterparts (Statistics Canada, 2016 Table 32). Having said that, farmers in humid climates are using conservation tillage. With time, soil health has improved with no-till, most noticeably reduced crusting and greater infiltration rates which allow soils to dry whereby planting is not delayed.

Reasons for regional differences not only reflect the differences in climate, but also differences in the types of crops that are predominantly grown. For example, the small grains and oilseeds produced in western Canada do not leave as much crop residue on the soil surface after harvest, compared to grain corn that is grown in eastern Canada. As such, managing large quantities of crop residues via tillage becomes an important part of preparing a seedbed.

Certain cropping systems are better suited to conservation tillage than others. For example, conservation tillage is well suited to grain and oilseed crop rotations where weeds can be chemically controlled. Likewise, in organic farming systems or systems where chemical weed control options are limited (e.g., specialty or vegetable crops), weeds are managed mechanically via tillage. Tillage practices are still common in vegetable cropping systems where small-seeded crops need a fine seedbed and good soil contact for germination, or where transplant equipment needs to deeply nestle a seedling into the soil. In these aforementioned systems, tillage is also used to incorporate crop residues to mitigate insect and disease pressure. These factors as well as time and cost of converting to conservation tillage will be front of mind when making the decision to switch practices.

How does conservation tillage enhance soil health? Conservation tillage maintains crop residues on the soil surface, providing protection (Table 15.3; Figure 15.7). The layer of crop residues, organic material or plants on the soil surface improves water infiltration by slowing surface run-off and wind speeds, and it retains moisture by reducing soil water evaporation and salinization in arid conditions (see continuous living plant section). The crop residues that remain on the soil surface provide a food source (carbon) and habitat for macrofauna and microorganisms, which is beneficial during the decomposition of crop residues as biological activity helps to form soil aggregates—a key to soil structure. Reducing soil disturbance also decreases the breakdown of organic matter already present in the soil and helps to maintain soil structure and aggregate stability. Adopting conservation tillage practices can improve soil health through a range of mechanisms, such as by influencing water cycling, soil structure, organic matter and nutrient cycling, habitat provisioning and biological population regulation (these are all soil functions; see Figure 15.2, highlighted functions in blue). Because conservation tillage influences many soil functions, it is not surprising that conservation tillage improves soil health—and that studies have generally found higher soil health scores under no-till than plow tillage (Figure 15.16)

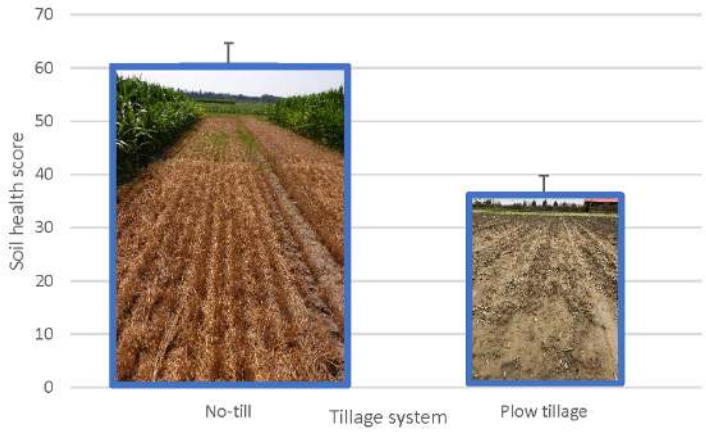


Figure 15.16. Lower soil health scores with fall plow tillage compared to no-till systems. Average (SE) scores from four long-term trials in Ontario. © Adapted from Congreves et al. (2015) by Laura Van Eerd, Univ. of Guelph. is licensed under a CC BY (Attribution) license.

It is universally accepted that conservation tillage minimizes erosion, which minimizes soil carbon losses. It is not difficult to imagine that if long-term research trials were conducted on sloped land subjected to erosion that great differences would be observed in soil health, organic carbon and primary productivity between conservation tillage and plow tillage. Given the regional climatic, soil, and production system differences outlined above, it should be expected that regional differences associated with the impact of no-till on soil organic carbon accumulation and soil health also occur. It is not possible to separate if this effect is climate-driven or simply due to the different crop rotations employed in various regions resulting in different quantities of carbon inputs. Regardless, surface soil health indicators and scores point to the value of adopting conservation tillage practices.

Sampling depth is an important consideration when comparing tillage systems. Many soil health tests only require sampling to the 0–15 cm depth, but plowing mixes the soil to 20 cm—and some tillage practices like subsoiling will mix soil to the 40 cm or 60 cm depth. The impact of tillage on soil health indicators will depend on the level of soil mixing brought by specific tillage practice. Nonetheless, in cases where conservation and conventional tillage practices have comparable levels of soil carbon throughout the whole profile, it is the surface levels of organic carbon under conservation tillage that help support many soil functions. Moreover, plant roots mainly occupy surface soil, where exudates and other carbon compounds are deposited. So, the question is: are the higher soil health scores that are often observed under conservation tillage vs. conventional tillage overestimated when soil sampling focuses only on the surface soil layers? The answer is still under debate in the scientific literature.

Unlike the other 6C soil health management strategies, there is greater uncertainty as to the magnitude of effect of no-till on organic carbon (Ogle et al., 2019) and soil health. Some of the uncertainty is due to differences in management among research trials. As previously stated, residue management can be challenging in no-till systems, which often lowers plant stands and primary productivity. With lower primary productivity, so too are the soil carbon inputs. Over time, researchers and farmers have largely overcome the challenges of residue management to establish a good seedbed by modifying planter equipment (Photo 15.14F). Ultimately, the benefits that conservation tillage imparts on soil ecosystem services (including plant productivity) drive the decisions to adopt this practice.

Challenges of conservation tillage: Transitioning from conventional to conservation tillage requires the purchase or modification of farm equipment to manage crop residues. Transitioning also requires a period of learning through trial and error. Moreover, there is a transition for soil life too, where microbial populations change. An obvious example is the change from annual to perennial weeds when tillage ceases. Practitioners of no-till often refer to two phases: a 'transition period' and 'mature' no-till. Management and challenges are greatest during the transition period. The greatest soil health gains are observed in the mature no-till phase, but it takes time and investment to get there.

As eluded to in the *Context Matters* section above, challenges with adopting conservation tillage practices will also vary among soil types and management systems. The transition period of converting to no-till systems is particularly challenging for poorly drained clay textured soils. Another challenge in no-till systems is the acidification of the soil surface, caused by the application of ammonium-based fertilizers and the stratification of nutrients such as P and K, which results in greater concentrations at the soil surface. Adopting no-till in organic farming systems still remains a challenge across Canada (Halde et al., 2015, 2017).

Conservation tillage is often described as a system rather than its individual practices or equipment. Any change in tillage impacts many other aspects of the cropping system that will also need to be adapted (e.g., cultivar selection, application of mineral fertilizers, weed control). Note that this concept also applies to the adopting other 6C strategies of soil health management. In addition to being prepared and dedicated, any change to an operation requires evaluation of the entire system, in order to be successful.

6C–3: Continuous Living Plants



In nature, soil is usually kept covered by living plants and plant residues. Think of a forest with its trees, understory species, and blanket of leaf litter or think of a native prairie grassland with its extensive root systems. Soils in these ecosystems have a continuous cover of mostly **perennial plants** and a layer of plant litter that remains at the soil surface. Continuous living plants protect the soil against disturbance, while tightly cycling nutrients and water and provide a steady supply of organic inputs from roots and litterfall to build soil organic matter (Table 15.3).

In contrast, agricultural systems where **annual plants** are typically grown, there are periods during the year in the early spring and late fall when living plants are not growing. Moreover, some farmland, largely on the Prairies, is intentionally left as fallow land to conserve soil moisture for subsequent grain crops. As far back as the early 20th century, however, soil scientists could measure the negative impact of frequent fallow on soil organic matter levels in prairie crop rotations (Janzen, 2001). The concerns and warning from these early scientists are linked to our modern concern and interest in soil health. The absence of living plants and their root systems means that soils are more susceptible to erosion. At the same time, there are no plants taking up nutrients or contributing organic matter—putting soil health at risk. Continuous living plants, primarily by perennializing cropping systems, is the only 6C soil health management strategy that contributes to all three soil health principles (Table 15.3) protect soil, diversity + perennialization and building carbon. We will outline how keeping continuous living plants supports soil health and what management practices can be employed to better approximate perennialization even in annual cropping systems. This soil health strategy lays the foundation for the other 6C centred around plants (Cover crops and Crop + animal diversity).

The root of the solution: Healthy soils support plant growth and healthy plant roots. This concept can also be flipped around. Plant roots support soil development and soil health (King et al., 2020). Increasing the duration, number, and type of plant roots in soil through the presence of continuous living plants is an important ingredient for building healthier soil. Living roots connect the soil

and the soil ecosystem to a plant's infrastructure, which collects solar energy through photosynthesis. Plants allocate roughly 40% of their net photosynthetic carbon towards roots (Jones et al., 2009)—this is a large amount of carbon that can contribute to increasing soil carbon and soil health. In general, perennial crops provide more carbon inputs than annual crops because their root systems are not only larger, but for a much longer time the plant is able to photosynthesize and release carbon to the soil from decomposition of aboveground residues, dead roots, and rhizodeposits from live root systems. The carbon released serves as fuel for the heterotrophic soil microbial communities (recall Chapter 6). In turn, the soil microbial communities can enhance soil health and function by cycling nutrients, building soil organic matter, and anchoring soil aggregates to build soil structure (see Chapter 4). The longer a crop is present in the soil, the more carbon it adds to build soil organic matter and to fuel important microbial-driven processes that underpin soil health.

Roots anchor the plant to the soil, in turn stabilizing the soil and reducing exposure to wind and rain erosion. Perennial crops can outperform no-till annual cropping in improving soil physical properties such as aggregate stability (Arshad et al., 2004; King et al., 2019). Perennial roots can also affect soil nutrient distribution and help mitigate nutrient losses, as roots are able to access nutrient sources at lower soil depths. The presence of continuous living plants results in improvements in soil structure and aggregation conferred by roots, rhizodeposits, and rhizosphere microbial communities. In turn, these improvements result in greater hydraulic conductivity and therefore greater infiltration and drainage, which is important in humid regions. In regions with limitted rainfall, continuous living plants can improve soil hydrological properties such as soil porosity and the water retained at field capacity (Basche and DeLonge, 2017). Increasing the duration of time that living plants grow and transpire water from deep in the soil profile can help reduce salinity (Chapter 5). Soil cover and living roots growing in the soil limits evaporation and reduces the upward movement of salts carried in water towards the soil surface. Thus, the presence of a continuous living plant can positively affect biological, chemical, and physical properties that support healthy soils—especially when a root system is present in the soil for a long time.

Aboveground plant components such as leaves and stems also play important roles in improving soil health. By having continuously living plants in the soil, the aboveground components contribute to protecting surface soil, reducing soil evaporation and runoff, minimizing displacement of soil during heavy rain, moderating soil temperatures, and providing carbon inputs. Whether the benefits of continuously living plants are due to the aboveground or belowground parts of the plant, the key point is that the longer plants are allowed to grow, the greater the benefits. So, in agricultural systems that are dominated by annual cropping, what strategies can farmers employ to increase the duration of living plants in their systems?

Strategies for increasing continuous living plants: Growing perennials is an obvious choice for increasing continuous living plants. Perennial systems, such as orchards (Figure 15.17), pastures, and forages dramatically increase the duration and depth of living roots in a crop rotation and their effects on soil health. For example, based on data from a long-term crop rotation trial in Saskatchewan, modeled organic carbon stocks were predicted to be highest in a diversified six year crop rotation that included three years of annual crops plus three years of alfalfa (a perennial plant), compared to a low diversity system that only included annual crops or even a diversified system with only annual crops (Lychuk et al., 2019). Overall, perennial forages can enhance soil health by affecting multiple soil properties. In Ontario, soil health scores were improved with the presence of a red clover cover crop that extends the duration of living plant cover in an annual cropping system (see section on cover crops below), but soil health improvements were even greater with alfalfa (Figure 15.18).



Figure 15.17. Ontario growers discuss their continuous living plant systems. Notice the soil is completely protected in both systems. **(Left)** Haskap production on Amy Strom's agro-tourism farm. Note the high plant diversity. **(Right)** Ellen Jennen discusses strawberry production under high tunnels where cover crops are grown to minimize crop pests and provide habitat for insect predators © Laura Van Eerd, Univ. of Guelph is licensed under a CC BY (Attribution) license.

Perennial crops are typically grown as forage or forage seed, which depends on livestock for consumption and clearly links with other 6C strategies of Compost + organic amendments and Crop + animal diversity. Perennial grains (e.g., wheat) and oilseeds (e.g., sunflower) are increasingly being considered as options to increase living cover in cropping systems. Perennial grains may be grown to varying degrees, ranging from permanent cover to being included as a phase in an annual crop rotation (Martens et al., 2015). In Canada, perennial grains are not widely grown, primarily because varieties have not yet been developed for Canadian growing conditions (Cattani and Asselin, 2017). Until breeding programs develop perennial grains and oilseeds suitable to Canadian growing conditions, perennial forage will likely remain as the dominant option for including perennials in Canadian cropping systems.

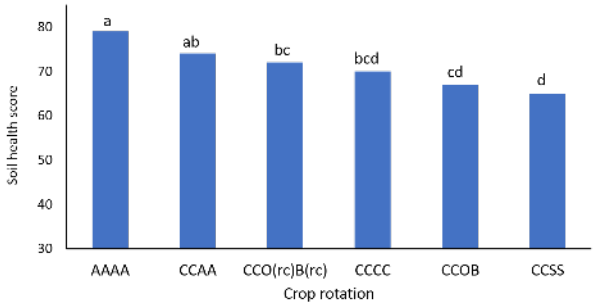


Fig. 15.18. Greater soil health scores with continuous cropping and cover crops. Soil health was measured as commercially available Solvita labile amino-nitrogen (SLAN) on surface soil collected after 36 years of management at Elora, Ontario. Based on a 4- year sequence of A = alfalfa, C = corn, O = oats, B= spring barley, S = soybeans, rc = red clover cover crop © Adapted from Chahal et al. (Submitted) adapted by Laura Van Eerd is licensed under a CC BY (Attribution) license.

There are some options to incorporate the concept of continuous living plants into annual systems. In Canada, annual crops may have living roots that grow for 3 to 5 months of the year. Land managers do have a few options to extend that duration, as well as to increase the number and type of living plants growing in soil (Figure 15.19). Growing winter annuals, such as winter wheat or fall rye in a crop rotation allows for living roots to be growing in the fall and early spring. Cover crops grown in the fall and early spring are another option; their benefits and limitations are outlined in more detail in the next section. Increasing crop diversity (described in Crop and Animal Diversity section) by inter-seeding two plant species together through relay cropping or intercropping can also add duration and increase the number and type of living roots in annual crop rotations. Agroforestry is the integration of woody plants such as trees and shrubs into agricultural systems and is another option for perennializing the system. Woody plants may be grown as shelterbelts or in alley cropping where annual crops are grown between rows of woody plants. The inclusion of woody plants to agricultural systems can add additional ecosystem services by increasing plant diversity, habitats for pollinators and other fauna, soil carbon storage, suppression of crop diseases and pests, and regulation of water and nutrient cycles.

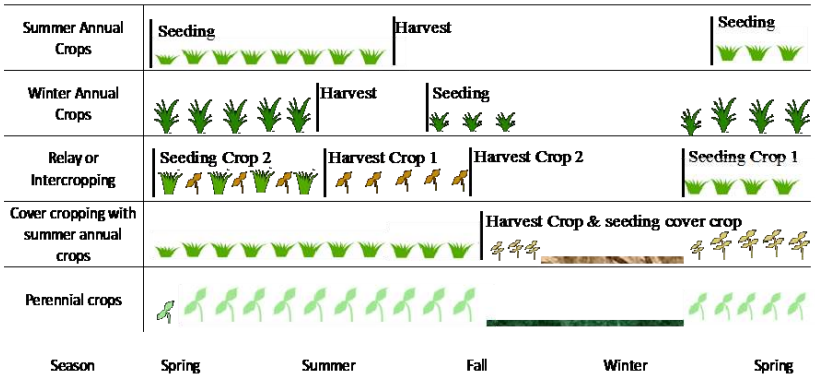


Figure 15.b19. Schematic Gantt chart of various cropping strategies to increase living plant cover. Note the green and brown lines indicate soil coverage as opposed to bare soil. © Athena Wu (conceptualized by Yvonne Lawley) is licensed under a CC BY (Attribution) license.

6C–4: Cover crops



The concept of cover cropping is not new. The use of cover crops dates back to ancient civilizations such as the Roman and Mayan empires, and likely even further back to the dawn of agriculture. Cover crops are plants grown to protect the soil and therefore serve an alternate purpose than growing a crop for its harvestable commercial product (i.e., grain, fruit, fibre, fuel). For this reason, cover crops are sometimes called service crops because of their role in influencing ecosystem services such as erosion control, nutrient supply, climate regulation, water quality and supply, and biodiversity conservation. Cover crops act as a 'catch crop', capturing nutrients that would otherwise be lost if not taken up by the main crop. On the other hand, cover crops may supply nutrients to the following main crop if used as **green manure** (similar to compost or livestock **manure**). It is important to note that different cover crops do not influence each ecosystem services in the same way. For instance, certain over-wintering **legumes** alter nitrogen dynamics, but are poor competitors with weeds during establishment. It is therefore important to match the cover crop species to the ecosystem goods and services desired. Here we will build on the concept of soil health developed in the Continuous living plants section, while introducing multiple approaches and challenges associated with cover cropping.

Types of cover crops: Almost any plant can be grown as a cover crop, but certain characteristics—such as rapid emergence and growth as well as relatively inexpensive seed costs—are more desirable. Cover crops are grouped into five general categories based on **ecological niche**, plant type, and growth form (Table 15.4). Legume cover crops have a stand-alone category due to their ability to fix nitrogen, whereas the other cover crops are categorized based on their warm- or cool-season classification, largely driven by their ability to survive the winter. Whether a warm- or cool-season plant is appropriate to grow depends on the time of year and the climate. To increase diversity, cover crops are often planted as mixtures called bi- or poly-cultures. When making a polyculture or a 'cover crop cocktail', a good approach is to mix from these five categories (Table 15.4; Figure 15.20).

Table 15.4. Common cover crop types with key benefits

Cover crop type	Example species	Key benefits
Legumes	Red clover	Nitrogen fixation
Warm-season (C4) grasses	Sorghum	Abundant biomass and carbon inputs
Warm-season broadleaves	Sunflower	Added diversity
Cool-season (C3) grasses	Cereal rye	Rapid growth and overwinter survival
Cool-season broadleaves	Radish	Rapid growth and deep roots for nutrient scavenging

Timing matters: In addition to time of year, the decision on what to plant depends on the desired ecosystem service, and the fit into the land management or cropping system. Cover crops can be sown before or after the main crop or during the growing season. Regardless of when the cover crop is sown, there must be sufficient time for growth. A general rule is that cover crops should accumulate approximately 1 Mg ha^{-1} of aboveground dry biomass to provide ecosystem services, such as those outlined in Table 15.1. A minimum of six weeks is needed for cover crops that rapidly grow (e.g., buckwheat and mustards), but most cover crops will need over two months of growth—this can limit options to growers in cooler and northern climates. Although some practitioners (such as organic growers, orchardists, or small acreage landowners with high value crops) take their fields out of production to plant cover crops, the timing and length of the main crop growing season dictates whether cover crops can be grown before or after the main crop (Figure 15.19). An alternative method to increase the duration of plant growth is by **underseeding** a cover crop directly under the main crop. For instance, red clover undersown into small grain cereals (Figure 15.18). The clover grows slowly under the grain canopy, and by grain harvest in August, the red clover cover is well established and has the rest of the year to grow (Figure 15.20C). Yet another approach is to take a section of land out of production for a season to gain full benefits of growing cover crops (see Continuous living plants section). For more information on how to incorporate specific cover crops into your system and region, consult resources provided at the end of the chapter.



Figure 15.20. Various cover crop approaches from full season **(A)** to planting after main crop harvest **(B)** to intercropping into growing main crops **(C-D1)**. **(A)** Sébastien Angers, an organic grain grower from Quebec, in one of his multi-species cover crop mixtures. This full-season polyculture includes sunflower, sorghum-sudangrass, pea, crimson clover, hairy vetch, and faba bean. **(B)** Undergraduate students collecting biomass approximately two months after planting oat cover crop in Ontario. **(C)** In Saskatchewan, red clover underseeded into wheat in early spring and by the time the crop starts to mature, the ground is covered by the clover. **(D)** Specifically, designed planter to plant three rows of cover crops into corn at V6 (6-leaf growth stage) to V8 (8-leaf growth stage) by canopy closure **(D1)** the annual ryegrass and red clover cover crops have established. Photo credits:© "(A) – Denis S Sébastien Angers; (B, D, D1) – Laura Van Eerd; (C) – Yvonne Lawley" is licensed under a CC BY (Attribution) license.

How do cover crops enhance soil health? The importance of plants for soil organic matter (Chapter 3), soil microbial communities (Chapter 6) and soil genesis (Chapter 2) have been outlined. Cover crops are intentionally grown to harness and maximize ecosystem services that plants can provide—and most directly influence soil health. Cover crops enhance soil health by increasing carbon inputs and decreasing carbon exports. Let's start with the carbon inputs. Cover crops increase primary productivity and thus carbon assimilation and carbon inputs mainly by extending the length of time plants are growing (Figure 15.19). Cover crops are not 'harvested'; no biomass is removed from the field and therefore no carbon is removed from the field by direct management decisions (except in some systems through livestock grazing; see section on Compost + Organic Amendments and Crop Diversity). The plant material, and carbon contained within, added to the soil enhances microbial activity, and through decomposition a portion is converted into soil organic matter. For example, in Eastern Canada using annual cover crops six times over eight years in a processing vegetable and grain system, aboveground carbon inputs from cover crops were approximately 9 Mg C ha^{-1}, given that 40% of net photosynthetic carbon is allocated belowground, surface soil stored up to 28% more organic carbon than the soils that did not have cover crops growing (Chahal et al., 2020) In this same medium-term cover crop experiment, primary crop productivity was equal to or better with cover crops and soil health tests scores improved by up to 17% compared to the no cover crop control (Figure 15.21; Chahal and Van Eerd, 2018, 2019).

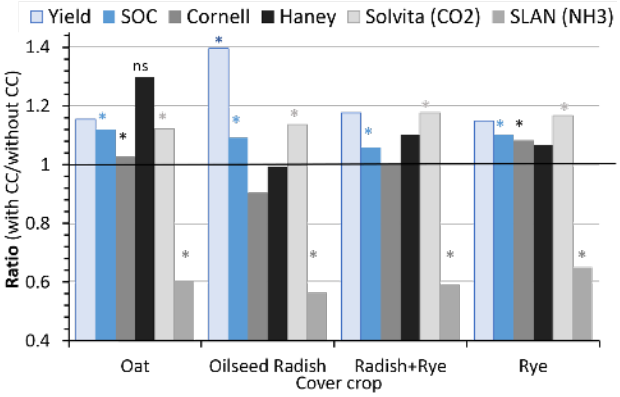


Figure 15.21. Better soil health scores and crop yield with cover crops (CC) compared to the no cover crop control in a medium-term cover crop (CC) experiment established in 2008 and sampled in 2016 in Ridgetown, ON. Data expressed as a response ratio, where values greater than 1 indicate cover crops had higher scores than the no cover. Cover crops were planted after main crop harvest; rye= winter cereal rye. SOC = soil organic carbon, Cornell = Comprehensive Assessment of Soil Health, Haney = Haney soil health test, Solvita = evolved CO_2-C, SLAN = Solvita labile amino-N as evolved NH_3. * indicates the cover crop was significantly different than the control and ns indicates not significantly different. © Adapted from Chahal and Van Eerd (2018, 2019) adapted by Laura Van Eerd is licensed under a CC BY (Attribution) license.

Cover crops not only add carbon through photosynthesis and partial decomposition, but also minimize carbon exports by mitigating erosion and run-off and promoting aggregation. Cover crops act as a physical barrier to the erosive forces of water and wind, which slows water movement and wind speed at the soil surface, thus reducing the force applied and amount of soil and carbon leaving the field. Cover crops can improve soil infiltration, which lowers runoff and water erosion. Cover crops improve soil water storage and supply by acting as a physical barrier to trap snow, which may lead to more soil moisture retention depending on the climate and soil temperatures. Through transpiration, cover crops lower soil moisture thereby increasing the capacity of soil to receive rainfall. Less soil runoff and erosion results in less carbon losses from the field. Carbon losses are also minimized through increased soil aggregation. Because cover crops reduce the amount of time that soils are bare, the plant residues, root exudates, and microbial associations enhances soil aggregation, further protecting carbon inputs against microbial decomposition and carbon loss. An increased amount of stable soil aggregates improves resistance to the physical force of wind and water erosion, and therefore helps maintain soil structure and mitigates surface crusting.

Cover crops enhance other soil health properties, living up to their other name as “service crops”. Cover crops provide additional food and habitats for soil microorganisms and macro-fauna through root exudates and the rhizosphere. As described above, some cover crop roots can penetrate hard plow pan layers, thus ameliorating the negative effects of compaction. While protecting surface soils, cover crops enhance nutrient retention capacity. Deep-rooted cover crops, such as Brassicas (e.g., mustards, radish, turnips), play a key role in moving macro- and micro- nutrients from deeper soil depths to surface and near-surface soil. Specific varieties of mustard species contain high concentrations of glucosinolates that can be managed as a bio-**fumigant** to suppress soil-borne pests and diseases, particularly in horticultural crops. Translocation of nutrients to near-surface soil along with nitrogen fixation of legumes enhances nutrient availability to the next crop. Through transpiration cover crops lower soil water content, which mitigates leaching losses of nutrients, such as nitrogen to groundwater. Moreover, cover crops increase soil organic matter, which is important for water holding capacity and nutrient cycling. While the aforementioned soil health attributes may be realized by plants in general, the fact that cover crops are in place when fields would otherwise be fallow, greatly extends primary productivity and the capacity to which these benefits can be attained.

The trouble with cover crops: With all the benefits cover crops can provide, why are only 13.7% of Canadian farmers reporting cover crop use (see Figure 15.22)? An obvious barrier is money. Cover crops are grown as ‘service crops’ that are not harvested; therefore, they do not produce a product to sell nor an income to farmers. Likewise, it is very difficult to put a dollar value on the ecosystem services that cover crops provide. Cover cropping has immediate costs (seed, planting, terminating), but expected benefits accrue over the long-term and many of these benefits are realized off-field (e.g., less soil erosion, less nitrogen and phosphorus runoff to surface water) while costs are borne by the farmer.

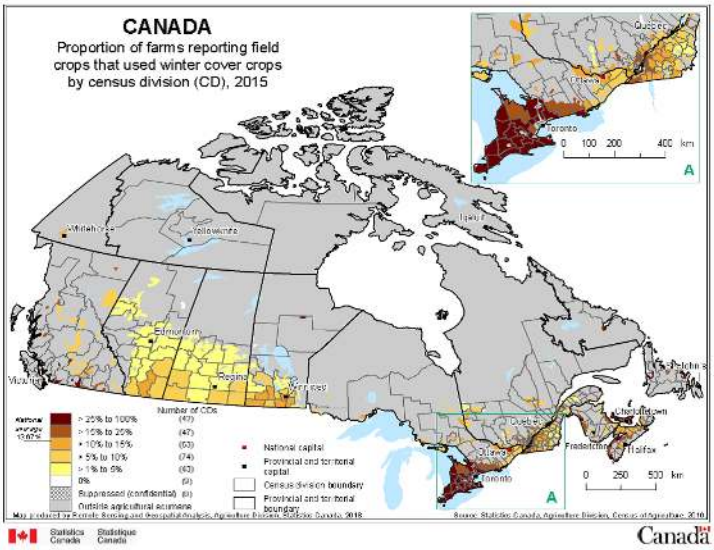


Figure 15.22. Map of Canada indicating cover crop usage based on the 2015 survey. Source: Statistics Canada – Land use, land tenure and management practices. © Statistics Canada; licensed under the Open Government Licence – Canada.

In many climates and cropping systems, there is the very short time period available for cover crops to grow. For many main crops like grain corn or in regions with a shorter growing season, particularly in the Prairies and northern Canada, there just is not a lot of time to grow a cover crop. Growing cover crops are even further challenged in the semi-arid climate of the Prairies because they can reduce soil water content, which limits main crop productivity. Other adverse effects on the main crop as a result of cover crop use include a potential increased incidence of pests in certain crops and hampered nutrient supply. Non-legume cover crops with high biomass and C/N ratios immobilize N, which makes it difficult to estimate adjustments to synthetic fertilizers. Although legume cover crops, such as clovers and alfalfa, provide mineral N to the following crop, there is no clear method to estimate fertilizer reductions.

Adoption of cover crops places greater demands on the grower's decision making; there is a learning curve and an extra level of management. Cover cropping also demands time and when farm sizes increase, so too do time constraints. For example, in the spring, cover crop residues must be managed and in cold, wet springs, soils with plant residues on the soil surface will not warm up nor dry out quickly, which causes planting delays. Further, it is not unusual to have uneven cover crop stands, which complicates management in the following crop. Despite these limitations, practitioners and researchers are finding innovative methods to overcome these limitations. Indeed, adaptation is on the increase, with national cover crop use increasing 5.5% from 2011 to 2018. Further efforts are needed to mitigate these challenges by developing agronomic solutions that work in Canada, while quantifying the resulting soil ecosystem and services of cover crops.

6C-5: Crop and animal diversity



While cover crops increase plant diversity and improve soil health, they are not a substitute for good crop rotation (i.e., main crop diversity). This is because cover crops are present in the soil for a shorter period of time than a main crop. A good crop rotation considers plant diversity temporally and spatially. Canadian organic standards (CAN/CGSB-32.310-2020) extend this concept further, where plant families rather than species must be rotated (i.e., tomatoes and potatoes belong to the same family – *Solanaceae*). All crops benefit from rotation, with estimates of 10% yield increases over continuous cropping and up to 25% in drought years. This phenomenon is well known in agriculture, such that contracts between growers and the companies buying their commodities typically include a clause stating that the crop should be grown in any particular field only once every four to five years. Similarly, in southwestern Ontario—and although farmers have the land base—processing tomato growers will rent land where *Solanaceae* plants have never been grown before. The expected yield gain of 12 to 15% (pers. comm. Cory Cowan CCA), derived from less soil-borne pathogens and compaction, more than compensates for the rental and added transportation costs.

Although the benefits of avoiding **monoculture** are well known, the mechanisms that explain why we see increased primary productivity with crop diversity are largely unknown. One obvious benefit is reduced pest pressure in diverse systems, particularly soil-borne pests. Another—but less obvious—contributing factor is enhanced soil health in diverse systems. Diverse systems tend to have higher soil organic matter. Because the same source of carbon inputs is provided in monoculture, the lack of diversity in carbon inputs selects for microbial communities that can decompose the specific source of carbon supplied by that crop (recall Darwin's natural selection). In more diverse systems there are many different types of carbon inputs, which influences the ability for microbes to use (decompose) carbon sources. Plant diversity likely promotes microbial diversity and functional redundancy. By mimicking nature and including a diversity of plants in any given field (over time and space), there will be a more diverse soil community. The diverse communities lead to more stable populations and confers resiliency to abiotic and biotic stressors. It is important to consider both temporal and spatial diversity—and combination of planting intercrops, cover crops, and perennials helps to address both. We will expand on the previous described 6C strategies and provide examples of how crop diversity enhances soil health.

It's not just how many you grow, it's what you grow: To build soil health, it is not enough to have high plant diversity; rather, it is key to grow plants with high carbon inputs and, if possible, avoid **soil depleting crops**. For example, soybean is considered a soil depleting crop. In the Prairies and elsewhere, many grass species are known for their high carbon inputs, particularly belowground carbon inputs. A diversity of plant species means that there will be diversity in root architecture, rooting depths, carbon composition and root exudates. Maximizing both the aboveground and belowground biomass can help build soil health. Deep-rooted crops include canola, sunflower, safflower, and alfalfa. Extensive root systems increase the rhizosphere volume—providing a 'bigger buffet table' for the microbes. The roots and resulting carbon inputs also strengthens the soil structure and aggregation, which improves soil resilience to threats such as erosion, compaction, and crusting.

Winter wheat: the soil builder in Eastern Canada: Based on the long-term tillage systems and crop rotation trials in Southwestern Ontario, rotations that include small grains, such as winter wheat, have 10% greater soil health values based on various soil health tests (Congreves et al., 2015) as well as greater organic carbon (Congreves et al., 2017) and 20% greater water stable aggregates (Figure 15.23; (Van Eerd et al., 2018). Soil health was equivalent to or greater than rotations with winter wheat and a red clover cover crop, as well as with the continuous living cover of alfalfa (Figure 15.18). The increases in organic carbon and soil health conferred by winter wheat resulted in at least 10% yield increases in grain corn and soybeans when winter wheat was in the rotation. Moreover, crop yields were more stable in extreme weather years and required 17% to 22% less fertilizer N with winter wheat in the rotation (Gaudin et al., 2015) and attributed to increased soil health. Thus, winter wheat increases agroecosystem stability and resiliency mediated through improved soil health.

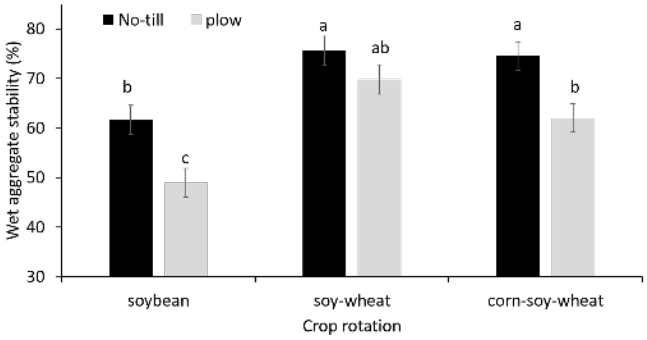


Figure 15.23. Crop diversity and a greater frequency of winter wheat in the crop rotation improves the stability of soil aggregates in a clay loam soil from the long-term experiment at Ridgetown Ontario © Adapted from Van Eerd et al. (2018) adapted by Laura Van Eerd is licensed under a CC BY (Attribution) license.

The mechanism by which winter wheat improves soil health is undoubtedly due to greater and diverse carbon inputs as winter wheat is a grass species with an extensive root system. Greater yields for corn and soybean when winter wheat is in the rotation goes hand-in-hand with greater amounts of crop residue returned to the soil (higher yields correspond to larger plants and more carbon). By including winter wheat in the rotation, different carbon compounds are added to the soil at different times of year, as previously mentioned. Winter wheat is planted in September, which provides carbon inputs during a time of year when other rotations would be fallow. Finally, winter wheat provides soil protection during the time of year when erosive forces of wind and

water are particularly strong (fall to early spring). Notice the similarities to cover crops and continuous living plants (refer to the previous two sections).

Even more ways to increase crop diversity: In addition to a good crop rotation, growing a mixture of crops can be achieved in a variety of ways. Polyculture and intercropping are techniques where two or more crops are grown simultaneously on the same land, whereas, relay and sequential cropping involves growing two or more crops in the same season (Figure 15.19). An example of this would be intercropping squash into winter wheat. While wheat crop yield is reduced to allow for the squash crop to be planted in early June, there are advantages for squash production. By wheat harvest in July, squash vines are elongating with the added advantage of fruit developing on crop residues as opposed to soil. This system reduces squash soil-borne disease pressure and minimizes labour during harvest as there is less soil to clean off the fruits. On the prairies, chickpeas and flax have shown promise as an intercrop system by using special equipment to separate the grains at harvest. Polycultures are mixtures of plants, planted primarily for ecological benefits. Some farmers use polycultures as a cover crop mix or a forage crop, while others use polycrop strips along field edges to support biological diversity. While critics would cite lost acreage as a negative factor of this planting approach, the beneficial insects that some plants and mixtures encourage helps with pest management.

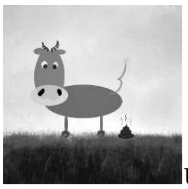
Not just plants, animals are important too: A diverse system also includes animals. There are positive effects of grazing on primary productivity and increased plant belowground carbon inputs. Moreover, with animals in the system carbon is returned to the soil with feces and manure. This is explored more in the section about Continuous living plants, which are often grazed or used as forage in livestock systems (e.g., alfalfa grown for cattle feed) and in the section on Composts and amendments, which are often plant material that has been partially decomposed or processed through animals.

Challenges: Similar to the other 6C soil sustainable management strategies, the decision to diversify crop rotations is based on many factors. Growing different crops temporally distributes finances and labour because crops are planted and harvested at different times, which may be viewed as an advantage (less of a time crunch at any one time point) or disadvantage (labor needed over a longer time period). Many crops require specialized planting or harvesting equipment, which requires capital investment and routine maintenance. Another important consideration is market access and fluctuating prices. Regardless, in the long run, soil health benefits are expected with diverse systems.



Figure 15.24. Tabitha Hastings shows the spatial crop diversity within beds (many species of greens grown together) and among beds at her Urban Sunset Farm in Highgate, ON. Temporal diversity of rotating crops with different families becomes challenging as there are many different vegetables are in the same family, for example cabbage, broccoli, cauliflower, and collard greens are all in the *Brassicaceae* family. Including grass cover crops such as oat or cereal rye can provide plant diversity and build soil health (not shown). Notice the various types of soil coverings; black fabric in the foreground used for weed control, white fabric row covers in the mid-ground used for frost protection and pest control and plastic hoop house in the background used for heat to extend the growing season © Laura Van Eerd, Univ. of Guelph is licensed under a CC BY (Attribution) license.

6C–6: Compost and amendments



Using the simplest definition, an organic soil amendment is any carbon-based material added to the soil. Composts are a type of organic amendment that has been partially decomposed before application. Thus, it is easy to see how applying composts and organic amendments directly applies to the soil health principles of building carbon (Table 15.3). For the most part, the sources of carbon material are largely regarded as 'waste products' from various sources such as humans, food, plants, animals (Table 15.5) where soil is viewed as 'the recipient' of this waste. It is unfortunate that these valuable carbon sources are viewed as wastes instead of as a critical resource in building soil health. This view really needs to change if real progress in building soil health and soil reclamation is to be made (see Chapter 16) for more information on reclamation). In this section, you will learn about different amendment types, the various mechanisms by which compost and amendments enhance soil health, and the practical challenges or barriers to adopting this practice.

Table 15.5. Common types of amendments available in Canada

Amendment	Sources	Phase[a]	Principle constraint	Reference
Generated On-Farm				
Animal manure	Livestock excrement	S, L, C	Pathogens	Maillard and Angers (2014)
Crop residues	Trimmings, culls and wash water	S, L, C	High N	
Non-Agricultural Source Materials (NASM)				
Biosolids	Sewage waste	L, S, P, C	Pathogens	Levesque et al. (2020)
Food waste	Processors, Municipalities	All	Salts	
Wood waste, yard trimming	Municipalities, Pulp and paper, Forestry products	S, C, U	N and P immobilization	Camberato et al. (2006)
Biochar	Pyrolysis	S	High energy input	Mechler et al. (2018)

[a] Dominant but not the only phase of available amendments (not in order). S=solid, L= slurry liquid, C=composted, U=unprocessed, P=processed (e.g., de-watered, pelletized).

Why bother composting? Organic amendments can be applied directly (e.g., culls after sorting and packaging vegetables on farm) or they can be processed, generally for ease of transport and application. Composting is one common form of processing. Composting is an aerobic process of decomposition mediated by microorganisms that results in stable organic material. It requires management of oxygen concentrations, moisture, temperature and carbon to nitrogen ratios (Figure 15.25); it is not enough to just leave a pile of manure in place over time. As materials decompose and carbon dioxide evolves, odor is minimized, pathogen loads are reduced (requires high temperatures), water is evaporated or leached, and the total volume of materials is reduced by approximately 50% which concentrates nutrients. For most practitioners, pathogen and volume reductions are the main reasons for composting. Because the most readily available materials are decomposed first, the remaining material (i.e., compost) is relatively stable. Nutrients are mainly in their organic form as opposed to mineral form, and thus are not as readily available (recall Chapter 7 Nutrient Cycling). From a crop management perspective, nutrient availability from composted manure is typically greater in years 2 and 3 than the year of application; hence, timing of application relative to crop nutrient uptake is important.

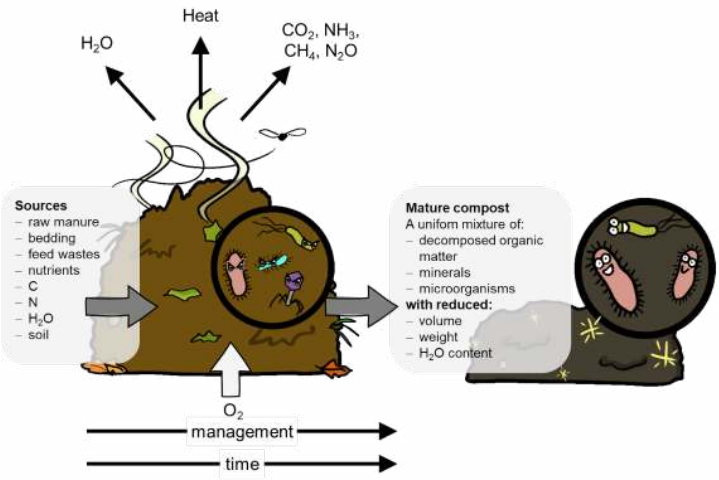


Figure 15.25. The composting process © Julie Forest-Drolet (conceptualized by Julie Forest-Drolet and Laura Van Eerd) is licensed under a CC BY (Attribution) license.

How do organic amendments enhance soil health? Similar to the other 'Cs' of soil health, organic amendments improve soil health through a number of mechanisms. Likely the most obvious and influential is increases in building soil carbon, which enhances soil infiltration, water holding capacity, microbial activity, CEC, nutrient availability. Physical improvements to soil health are expected due to the complex nature of the organic materials in the amendment. For instance, biosolids used in a semiarid grassland in British Columbia, increased soil aggregate size and stability, due to higher concentrations of carbon and nitrogen in aggregates (Wallace et al., 2009). The surface application of organic amendments also improves the availability of water and nutrients, and promotes plant growth, which in turn increases soil carbon (see continuous living plants). Meta-analyses (i.e., big data) showed that the quantity of carbon applied over time was positively related to soil organic carbon content (Maillard and Angers, 2014) and that amendments had a improved soil health of various biological, chemical and physical soil attributes (Norris and Congreves, 2018) (Figure 15.26). In fact, positive effects of organic amendments on soil health and crop productivity can be long-lasting (decades) and mitigates the negative effects of erosion (Larney and Olson, 2018).

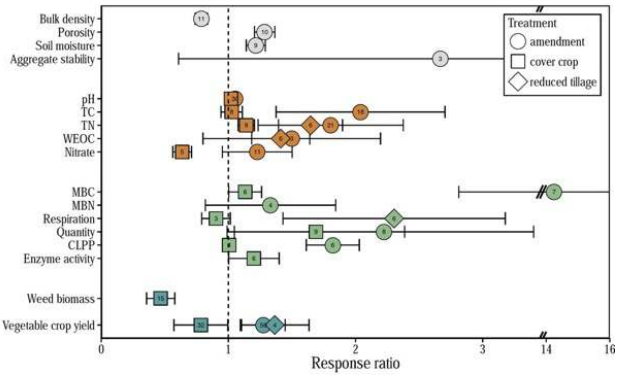


Figure 15.26. Meta-analysis of vegetable cropping systems showing how 3 of the 6C soil health management strategies (compost/amendments, cover cropping, and conservation tillage), as compared to conventional management, influence soil physical, chemical, and biological attributes and crop yields (Reprinted from Norris and Congreves (2018)). Symbols and confidence intervals that fall to the right or left of the dashed line represent cases where the soil health management practice increased or decreased the soil attribute value relative to the conventional practice, respectively. Where symbols and confidence intervals cross the dashed line, then the soil health management practice had no impact on the soil attribute value, relative to the conventional practice. Abbreviations: total carbon, TC; total nitrogen, TN; water extractable organic carbon, WEOC; microbial biomass carbon, MBC; microorganism quantity (colony forming units or counts), Quantity; community level physiological profiles, CLPP. © Norris and Congreves (2018) is licensed under a CC BY (Attribution) license.

Organic amendments alter soil microbial communities and depending on the degree of decomposition may stimulate microbial activity in soils. For example, biochar has been found to favor the establishment of plant beneficial bacteria and increased bacterial richness in a greenhouse study in Quebec, Canada (Lévesque et al., 2020). Likely, this occurred through increases in ecological micro-niches and nutrient availability in the peat-based growing medium. In New Brunswick, applications of organic amendment effectively suppressed soilborne potato diseases in potato fields. In these fields, the level of disease suppression was linked to the

quantity of carbon in compost, rather than its quality (e.g., C/N ratio, degree of decomposition) (Wilson et al., 2018). Similarly, in Ontario processing tomato fields with a soil-borne disease complex, crop yields improved with organic amendment application, which was attributed to changes in the rhizosphere and root colonized microorganisms (Van Eerd et al., 2021 – unpublished data). Thus, while it is generally accepted that organic amendments influence soil microbial communities, there is evidence that it might also be beneficial in disease suppression.

In addition to amount of carbon applied, many other factors also impact an amendment's effect on soil health, including climate, vegetation type, soil pH, soil microbial community, etc. Site-specific applications of amendments may be beneficial. In their review of amendments, Larney and Angers (2012) suggest that degraded soils have the greatest potential to benefit from composts and organic amendments in terms of soil health and primary productivity. Thus, there is an opportunity for site-specific application to degraded areas within a field (such as sandy knolls or compacted areas) (Figure 15.27A).



Figure 15.27. Ontario processing tomato growers use organic amendments to improve soil health and primary productivity. **(A)** Grower Kees Poppe targets applications of organic amendments to the sandy knolls on this farm. He has seen organic matter increase and yields improve in those areas, which he attributes to improved water holding capacity. Note how uniform the field looks in the first two photos, whereas before applying amendments the tomatoes on the sandy knolls would be less green with reduced growth. **(B)** Ontario Ministry of Agriculture, Food and Rural Affairs (OMAFRA) Soil Management Specialist Anne Verhallen discusses cover crop and crop diversity approaches to improve soil health. **(C)** Farmer Kurtis Allaer and family have a diversified operation including poultry manure that they apply to their tomato fields. © Laura Van Eerd, Univ. of Guelph is licensed under a CC BY (Attribution) license.

Choices, choices, choices: With so many amendments sources available (Table 15.5) and with so many different materials within each type of amendment (e.g., liquid, slurry, solid), the options are seemingly endless. The decision on what to apply depends on both soil conditions and the amendment composition, which largely centre around nutrient, salt and heavy metal concentration and pH. Application rates can be based on carbon or organic matter content, but are often made based on nutrient levels (particularly N, P). The decision on which amendment to apply depends on over-coming the challenges listed below as well as availability and, of course, costs.

Challenges: As with any management decision, there are trade-offs to consider when using organic amendments. Amendments, particularly of animal or human origin, contain pathogens. Although composting minimizes pathogen load, management must minimize pathogens in the crop and in the environment. Provincial rules and regulations regarding method and timing of manure and biosolid applications are designed to minimize the risk of pathogen and nutrient movement off field and to reduce odor. For instance, the timing of application to soil is often restricted to times of year when crops are not being grown (i.e., spring or fall). In many regions, soils are wet during the spring and late fall, so the application of amendments may lead to soil compaction. Soil compaction is exacerbated as organic amendments are bulky and requires specialized, heavy application equipment. Moreover, in order to minimize odor and surface nutrient losses, tillage may be required to incorporate the amendment. (recall Compaction reduction and Conservation tillage effects).

Although a significant quantity of organic amendment is necessary to observe enhancements in soil health, certain nutrients (e.g., P, K) and heavy metals (particularly Zn) accumulate in soil. Excessive levels can pose a risk to human, animal and environmental health. Likewise, some amendments are high in salts, which can burn plants and may create salinity problems in semi-arid climates. Finally, while the costs and nutrient benefits are immediate, the soil health benefits from amendments typically accumulate over time and following repeated applications. This is a challenge for cash-strapped landowners, however, long-term investments in soil health are expected to result in increased primary productivity and system resiliency, which is not unlike the other 6C soil health management strategies.

Other sustainable soil management approaches

In addition to the 6C strategies, there are other approaches that can be done to improve soil health (Table 15.6). These practices can occur in the field or around the field perimeter. Many are targeted to problem areas of a field and not all are applicable. As with the

6Cs, holistic considerations should be made when deciding to adopt any of these approaches.

Table 15.6. Additional approaches that can be used to help achieve soil health

Approach	Example	Soil Health Principle				
		Protect Soil		Diversity + Perennialization	Build Carbon	
		Maximize soil cover	Minimize soil disturbance		Enhance C inputs & gains	Minimize C & soil loss
Windbreaks	Tree barrier around field	♥		∗		⊕
Water erosion control structures	Vegetative strip where water flows	♥	⊕			∗
Land retirement	Convert to natural ecosystem	∗		♥	⊕	
Riparian zones	Vegetative strip before bodies of water	∗		⊕		∗
Physical remediation	Move soil from depositional sites (e.g. low areas or ditches)to eroded positions in landscape				⊕	∗
Contouring	Shaping land and growing crops perpendicular to slope	⊕				∗
Subsurface drainage	Move water off field quicker; compaction less likely		⊕		⊕	

♥ principle goal and benefit; ∗ expected benefit; ⊕ likely benefit

SUMMARY

Soil health is an important concept because it requires us to consider soil as a living system integrating biological, chemical, and physical properties. We can measure soil health and compare using visual soil assessments, soil health tests, or using integrated scores calculated from multiple soil health indicators. Through the ability of soil to function and provide ecological services, soil health connects people and their management decisions with soil, plants and animals.

Individual management practices influencing soil health can be categorized using the 6C framework: Compaction reduction, Conservation tillage, Continuous living plants, Cover crops, Crop and animal diversity, and Compost & amendments. Each of these management strategies has benefits and challenges. The 6C practices can be implemented individually or, ideally, in combination. As you probably noticed, there are a lot of connection between the 6C's. For example, using cover crops will increase the duration of continuous living plants and adds crop diversity. The timeline for measuring responses to new soil health management practices

range from years to decades depending on the management practice, soil characteristics, cropping system, and local environment. But adopting any of the 6Cs will have a positive impact.

It is an exciting time to be learning about soil health. When talking with farmers, agronomists, conservationist, and gardeners about soil health we often hear from them “I wish that I had learned this when I was in school/university.” We are confident that soil health will be on your mind when your life next intersects with soil. Soil science is continually breaking new ground. Our knowledge about soil health and the tools we have to characterize it are poised to increase exponentially during your lifetime. We invite you to continue learning about soil and join us to make these new discoveries.

SUGGESTED READING

Best Management Practices SOIL HEALTH http://www.omafra.gov.on.ca/english/environment/bmp/AF151.pdf

Magdoff and Van Es. 2009. Building Soils For Better Crops. Sustainable Soil Management. Third Edition. Sustainable Agriculture Research and Education (SARE). ISBN 978-1-888626-13-1 https://www.sare.org/Learning-Center/Books/Building-Soils-for-Better-Crops-3rd-Edition/Text-Version

ADDITIONAL RESOURCES

Infographic What is Soil Health by SARE outlines sustainable management practices. https://www.sare.org/Learning-Center/What-is-Soil-Health

Eastern Canada cover crop decision tool or MCCC cover crop selector tool

http://www.omafra.gov.on.ca/english/environment/bmp/soil-health.htm

https://www150.statcan.gc.ca/n1/pub/95-634-x/2017001/article/54903-eng.htm

Managing Cover Crops Profitably. Third Edition. Ed. A Clark. Sustainable Agriculture Research and Education (SARE)

https://www.sare.org/Learning-Center/Books/Managing-Cover-Crops-Profitably-3rd-Edition

NRCS Soil quality test kit. http://www.nrcs.usda.gov/wps/portal/nrcs/detailfull/soils/health/assessment/?cid=nrcs142p2_053873

STUDY QUESTIONS

1. Soil health tests currently available are weighted (or less commonly unweighted) averages of numerous soil indicators; where greater values indicate better soil health. This approach works well, but one criticism is that a soil could have a high test score but not support plant growth (i.e., doesn’t support primary productivity and thereby by definition is not a healthy soil). For instance, a soil with pH < 5.2 limits plant growth but all other soil indicators may be ok. Your task is to find a solution. How could this be a handled in a soil health index?
2. Using Table 2 and what you have learned from the other chapters (particularly Chapters 3-7), identify what is missing from the list. Why do you think it isn’t included?
3. Compare and contrast approaches to enhance soil health with approaches outlined in Soil Reclamation chaper of this textbook.
4. Pick two of the 6C strategies of soil health management. Compare and contrast how these approaches enhance soil health.
5. In science, there are always exceptions to the rule. There are situations where crop diversity would not be expected to enhance soil health. Provide two examples with different mechanisms and explain.
6. What are the different methods of incorporating cover crops into crop rotations? Pick one system and state advantages and disadvantages of incorporating cover crops.
7. What are three universal limitations to adopting the 6C practices of soil health management?

ONE MINUTE PAPERS

1. Choose one of the 6Cs. Describe how there are gains in soil organic carbon or how losses are minimized with the practice you selected.
2. You just purchased a brand new house (congratulations!) in a brand new neighbourhood (yikes- soil was degraded during building). You want to do the landscaping and know soil is important for successful plant growth. Describe how would you assess your soil and what you would look for to protect your investment in plants ensure they will thrive? Which 6C practices of soil health management would you use to improve the expected poor soil condition left by the builders?

OUTDOOR ACTIVITIES

1. **SOIL YOUR UNDIES:** Bury all-cotton underwear so that the waistband is at or near the soil surface. You want to maximize soil to cotton contact, so spread it out. Be sure to mark the area with a flag so that you can dig up the underwear 2 to 3 months later. Keep in mind that soil temperature and moisture influence soil activity, so you may want to adjust the length of time it is buried. Compare two different soils. Make a hypothesis: which soil do you think is the healthiest, most active? What does the amount of cotton remaining tell you about the soil. Was your prediction correct? Why or why not? For more information visit https://www.ifao.com/ and watch these videos https://www.youtube.com/watch?v=Ym0naU1oQmY [https://www.youtube.com/watch?v=akMT1ZZQ8PA]
2. **MEASURING SOIL COVERAGE:** This activity can be done almost anywhere – a field, a lawn, a disturbed forest. Stretch a measuring tape 25 meters. Every 50 cm, count when there is exposed soil as opposed to plant residue or debris. To be counted, the plant residue should be at least the size of a match (4 mm x 45 mm). Times the number by 2 and you have the percent of residue cover. Repeat 3 to 5 times in representative areas of the field to get an average value. To mitigate erosion, 30% coverage is recommended. A rope marked with electrical tape every 50 cm works well too.
3. **SOIL COMPACTION CAMPUS WALK:** Finding signs of soil compaction is not difficult, just take a walk around campus and look for green spaces where students are cutting through to get to class on time. Take a look at the appearance of the soil in the area where students walk compared to the rest of the green space. Think about what has happened to the number and size of pores in these two soils (Photo 17.3). Observe the number of grass plants growing in the area where students walk compared to the rest of the green space. Do you see other "weedy" plants growing where the grass is not? How have these plants adapted to grow where the grass is not? What management would you recommend to improve soil health in this area?
4. **DESIGNER SOIL HEALTH TEST:** Create your own soil health score card Using Table 17.2 and this web guide, create your own soil health test. Identify which variables you will include and why. Will all indicators be weighed the same? If not, why did you weight some indicators more than others? https://www.nrcs.usda.gov/Internet/FSE_DOCUMENTS/nrcs142p2_051879.pdf Answer: with justification, test must include chemical, physical and biological indicators and must include a measure of carbon quantity or cycling. LO#1-3:
5. **COMPOSTING AT HOME:** Now that you've learnt about the process and benefits of composting, why not start composting at home! In your backyard (or community garden), start composting your food and yard wastes. You can easily make your own compost bin! Add compost materials in layers, maintaining a balance between "green" materials (e.g., almost anything from your kitchen, fresh-cut lawn) and "brown" waste (e.g., dead leaves, old newspapers, etc.). Make sure to provide oxygen to your compost by mixing it well regularly (weekly or biweekly). Keep compost moist (but not too moist!). Take a handful of your compost and look at it. If you press it, is water running from it (i.e., too wet)? Any odor?

DIGITAL ACTIVITIES

1. **SOIL HEALTH from a FARMER's PERSPECTIVE:** Watch Farmer Fred Kirschenmann's TEDx Talk on Soil –from dirt to lifeline. Discuss the current constraints on farmers and farm systems. Although he doesn't explicitly say 6Cs, discuss how adopting the 6C practices of soil health management can provide various benefits and identify the benefactors.
2. **DECISION TOOL:** Use either the Eastern Canada or the Midwest cover crop decision tool to answer the questions for this case study. You own a CSA (Community Supported Agriculture) where you grow over 15 different vegetables. You purchased new land in August. This ground was tilled aggressively, often when the soil was too wet, and appears to have resistant weeds. You are planning on planting a cover crop in August with the goal of improving soil and crop productivity.
    - Based on the farm details, select three goals you would want your cover crops to achieve and explain why. Answer: goals compaction reduction, weed fighter, organic matter builder, perhaps another goal if well explained.
    - Based on the high ratings of these covers for the goals you selected, choose two cover crops you might want to grow. Based on the two cover crops you picked, look at the advantages and disadvantages in the info-sheets and choose the best cover crop option. State 2 reasons why you picked this cover crop. Give 2 reasons why you didn't pick the other cover crop. Answer: The cover crops should be rated 3 or 4 (out of 4) for the goals selected with reasonable explanations for selecting one cover over the other.

REFERENCES

Andrews, S.S., D.L. Karlen, and J.P. Mitchell. 2001. A comparison of soil quality indexing methods for vegetable production systems in Northern California. Agriculture, Ecosystems and Environment 1760: 1-21.

Bünemann, E.K., Bongiorno, G., Bai, Z., Creamer, R.E., De Deyn, G., de Goede, R., Fleskens, L., Geissen, V., Kuyper, T.W., Mäder, P., Pulleman, M., Sukkel, W., van Groenigen, J.W., and Brussaard, L. 2018. Soil quality – A critical review. Soil Biol. Biochem. 120: 105–125. Pergamon. doi:10.1016/J.SOILBIO.2018.01.030.

Camberato, J.J., Gagnon, B., Angers, D.A., Chantigny, M.H., and Pan, W.L. 2006. Pulp and paper mill by-products as soil amendments and plant nutrient sources. Can. J. Soil Sci. 86: 641–653. NRC Research Press Ottawa, Canada . doi:10.4141/S05-120.

Cattani, D.J., and Asselin, S.R., 2017. Extending the growing season: Forage seed production and perennial grains. Can. J. Plant Sci. 98: 235–246. doi.org/10.1139/cjps-2017-0212

Chahal, I., and Van Eerd, L.L. 2018. Evaluation of commercial soil health tests using a medium-term cover crop experiment in a humid, temperate climate. Plant Soil 427: 351–367. Springer International Publishing. doi:10.1007/s11104-018-3653-2.

Chahal, I., and Van Eerd, L.L. 2019. Quantifying soil quality in a horticultural-cover cropping system. Geoderma 352: 38–48. Elsevier. doi:10.1016/J.GEODERMA.2019.05.039.

Chahal, I., Hooker, D.C., Deen, B., Janovicek, K., and Van Eerd, L.L. 2021. Long-term effects of crop rotation, tillage, and fertilizer nitrogen on soil health indicators and crop productivity in a temperate climate. Soil Tillage Res. 213, 105121. https://doi.org/10.1016/j.still.2021.105121

Chahal, I., Vyn, R.J., Mayers, D., and Van Eerd, L.L. 2020. Cumulative impact of cover crops on soil carbon sequestration and profitability in a temperate humid climate. Sci. Rep. 10:13381| https://doi.org/10.1038/s41598-020-70224-6.

Chen, G., and Weil, R.R. 2010. Penetration of cover crop roots through compacted soils. Plant Soil 331: 31–43. Springer. doi:10.1007/s11104-009-0223-7.

Congreves, K.A., Hayes, A., Verhallen, E.A., and Van Eerd, L.L. 2015. Long-term impact of tillage and crop rotation on soil health at four temperate agroecosystems. Soil Tillage Res. 152. doi:10.1016/j.still.2015.03.012.

Congreves, K.A., Hooker, D.C., Hayes, A., Verhallen, E.A., and Van Eerd, L.L. 2017. Interaction of long-term nitrogen fertilizer application, crop rotation, and tillage system on soil carbon and nitrogen dynamics. Plant Soil 410: 113–127. doi:10.1007/s11104-016-2986-y.

Cresswell, H., and Kirkegaard, J. 1995. Subsoil amelioration by plant-roots-the process and the evidence. Soil Res. 33: 221–239. [Online] Available: http://www.publish.csiro.au/sr/SR9950221.

Doran, J.W., Sarrantonio, M., and Liebig, M.A. 1996. Soil Health and Sustainability. Adv. Agron. 56: 1–54. doi:10.1016/S0065-2113(08)60178-9.

Fox, C.A., Tarnocai, C., Broll, G., Joschko, M., Kroetsch, D., and Kenney, E. 2014. Enhanced A Horizon Framework and Field Form for detailed field scale monitoring of dynamic soil properties. Canadian Journal of Soil Science 94: 189-208.

Gaudin, Tolhurst, T.., Ker, A.P., Janovicek, K., Tortora, C., Martin, R.C., and Deen, W. 2015. Increasing crop diversity mitigates weather variations and improves yield stability. PLoS One 10: e0113261. doi:10.1371/journal.pone.0113261.

Halde, C., Bamford, K.C., and Entz, M.H. 2015. Crop agronomic performance under a six-year continuous organic no-till system and other tilled and conventionally-managed systems in the northern Great Plains of Canada. Agric. Ecosyst. Environ. 213: 121–130. Elsevier. doi:10.1016/J.AGEE.2015.07.029.

Halde, C., Gagné, S., Charles, A., and Lawley, Y. 2017. Organic No-Till Systems in Eastern Canada: A Review. Agriculture 7: 36. Multidisciplinary Digital Publishing Institute. doi:10.3390/agriculture7040036.

Haney, R. L. , Haney, E. B. , Hossner, L. R. and Arnold, J. G. 2010. 'Modifications to the New Soil Extractant H3A-1: A Multinutrient Extractant', Communications in Soil Science and Plant Analysis, 41: 12, 1513-1523.

Idowu, O.J., Van Es, H.M., Abawi, G.S., Wolfe, D.W., Schindelbeck, R.R., Moebius-Clune, B.N., Gugino, B.K. 2009. Use of an integrative soil health test for evaluation of soil management impacts. Renew. Agri. Food Syst. 24, 214-224.

Janzen, H.H. 2001. Soil science on the Canadian prairies – Peering into the future from a century ago. Can. J. Soil Sci. 81: 489–503. NRC Research Press Ottawa, Canada . doi:10.4141/S00-054.

King, A.E., Congreves, K.A., Deen, B., Dunfield, K.E., Voroney, R.P., and Wagner-Riddle, C. 2019. Quantifying the relationships between soil fraction mass, fraction carbon, and total soil carbon to assess mechanisms of physical protection. Soil Biol. Biochem. 135: 95–107. Pergamon. doi:10.1016/J.SOILBIO.2019.04.019.

Larney, F.J., and Angers, D.A. 2012. The role of organic amendments in soil reclamation: A review. Can. J. Soil Sci. 92: 19–38. Agricultural Institute of Canada . doi:10.4141/cjss2010-064.

Larney, F.J., and Olson, A.F. 2018. Wheat yield and soil properties reveal legacy effects of artificial erosion and amendments on a dryland Dark Brown Chernozem. Can. J. Soil Sci. 98: 663–677. Canadian Science Publishing65 Auriga Drive, Suite 203, Ottawa, ON K2E 7W6. doi:10.1139/cjss-2018-0025.

Lévesque, V., Jeanne, T., Dorais, M., Ziadi, N., Hogue, R., and Antoun, H. 2020. Biochars improve tomato and sweet pepper performance and shift bacterial composition in a peat-based growing medium. Appl. Soil Ecol. 153: 103579. Elsevier. doi:10.1016/J.APSOIL.2020.103579.

Maillard, É., and Angers, D.A. 2014. Animal manure application and soil organic carbon stocks: a meta-analysis. Glob. Chang. Biol. 20: 666–679. John Wiley & Sons, Ltd. doi:10.1111/gcb.12438.

Mechler, M.A.A., Jiang, R.W., Silverthorn, T.K., and Oelbermann, M. 2018. Impact of biochar on soil characteristics and temporal greenhouse gas emissions: A field study from southern Canada. Biomass and Bioenergy 118: 154–162. Pergamon. doi:10.1016/J.BIOMBIOE.2018.08.019.

Moebius-Clune, B., Moebius-Clune, D.J., Gugino, B., Idowu, O., Schindelbeck, R., Ristow, A., van Es, H., Thies, J., Shayler, H.A., McBride, M.B., Wolfe, D., and Abawi, G. 2016. Comprehensive Assessment of Soil Health The Cornell Framework Manual. 3rd edition. Cornell University, Ithica, New York. [Online] Available: http://www.scs.cals.cornell.edu [2016 Nov. 7].

Norris, CE, Bean, GM, Cappellazzi, SB, Cope, M, Greub, KLH, Liptzin, D, Rieke, EL, Tracy, PW, Morgan, CLS, Honeycutt, CW. 2020. Introducing the North American project to evaluate soil health measurements. Agronomy Journal. 112: 3195– 3215. https://doi.org/10.1002/agj2.20234.

Norris, C.E., and Congreves, K.A. 2018. Alternative Management Practices Improve Soil Health Indices in Intensive Vegetable Cropping Systems: A Review. Front. Environ. Sci. 6: XXX. Frontiers. doi:10.3389/fenvs.2018.00050.

Ogle, S.M., Alsaker, C., Baldock, J., Bernoux, M., Breidt, F.J., McConkey, B., Regina, K., and Vazquez-Amabile, G.G. 2019. Climate and Soil Characteristics Determine Where No-Till Management Can Store Carbon in Soils and Mitigate Greenhouse Gas Emissions. Sci. Rep. 9: 11665. Nature Publishing Group. doi:10.1038/s41598-019-47861-7.

Reicosky, D.C. 2015. Conservation tillage is not conservation agriculture. J. Soil Water Conserv. 70: 103A-108A. Soil and Water Conservation Society. doi:10.2489/jswc.70.5.103A.

Sarrantonio, M., J.W. Doran, M.A. Liebig, and J.J. Halvorson. 1996. On-farm assessment of soil quality and health. p.83-106. In: J.W. Doran and A.J. Jones (eds.) Methods for assessing soil quality. SSSA Spec. Publ. 49.

Savage, C.S. 2011. Prairie : A Natural History. Greystone Books.

Statistics Canada. Tillage practices used to prepare land for seeding (Table 32-10-0408-01) https://doi.org/10.25318/3210040801-eng https://www150.statcan.gc.ca/n1/pub/95-634-x/2017001/article/54903/catm-ctra-390-eng.htm

Thiessen Martens, J.R., Entz, M.H., and Wonneck, M.D., 2015. Review: Redesigning Canadian prairie cropping systems for profitability, sustainability, and resilience. Can. J. Plant Sci. 95, 1049–1072. doi.org/10.4141/CJPS-2014-173

Van Eerd, L.L., DeBruyn, A.H., Ouellette, L., Hooker, D.C., and Robinson, D.E. 2018. Quantitative and qualitative comparison of three wet aggregate stability methods using a long-term tillage system and crop rotation experiment. Can. J. Soil Sci. 98: 738–742. doi:10.1139/cjss-2018-0101.

Van Eerd, L.L., Y. Zhou, A.L. Turnbull, D.M. Johnston-Monje, G. Lazarovitz, S.A. Loewen. (n.d.). Managing tomato vine decline with soil amendments and transplant treatments: Elucidating impact on fruit yield, quality, and plant-associated microbial communities. Canadian Journal of Plant Science. (In revision: August 10, 2021)

Wallace, B.M., Krzic, M., Forge, T.A., Broersma, K., and Newman, R.F. 2009. Biosolids increase soil aggregation and protection of soil carbon five years after application on a crested wheatgrass pasture. J. Environ. Qual. 38: 291–8. J Environ Qual. doi:10.2134/jeq2007.0608.

Wilson, C., Zebarth, B.J., Goyer, C., and Burton, D.L. 2018. Effect of Diverse Compost Products on Soilborne Diseases of Potato. Compost Sci. Util. 26: 156–164. Taylor & Francis. doi:10.1080/1065657X.2018.1432430.

About the Authors

Laura L. Van Eerd, Professor, School of Environmental Sciences University of Guelph Ridgetown Campus



Laura Van Eerd (licensed under a CC BY-NC-ND license)

The goal of my research program is to enhance knowledge of nitrogen and carbon cycling in agricultural systems. I have published over 50 peer-reviewed papers (including *2018 Best Paper of the Year* in the Canadian Journal of Plant Science), supervised over 20 graduate students and received the *Pitblado Teaching Excellence Award*. My research has contributed to key policy documents on soil health and cover crops. As recognition of my impact, I was named one of six *Influential Women of Canadian Agriculture* in 2020.

What kind of work do you do? As a Professor of Sustainable Soil Management, my research program focuses on understanding how the various practices that farmers employ impacts their soil and crop. Thus, my research bridges agronomy and soil science. I am excited to be a co-executive director of Soils AT Guelph, an initiative to promote soil health knowledge by bringing researchers and farmers together. As Research Program Director for the Ontario Agri-Food Innovation Alliance, I lead the Soil and Water Stewardship theme by reviewing and awarding research projects valuing over $1M annually. I have a very rewarding career where everyday can be different; from teaching to research to extension service.

What did you like to do when you were a kid? As a farm kid, my favourite thing to do was explore outside. It's likely why I'm a scientist today. My younger sister and I would love playing in the fields, making mud pies, digging for earthworms, or breaking up soil crusts. Remember marveling at the odd patterns of curled up soil that resulted when a wet area finally dried. It fascinated me how the same soil could be so different; light and fluffy after tillage, spongy after a rain but later in the season have giant cracks when it was dry. Just thinking about it brings me back to the field behind the house on the home farm.

Kate A. Congreves, Assistant Professor, Department of Plant Sciences University of Saskatchewan



Kate Congreves (licensed under a CC BY-NC-ND license)

I am an Assistant Professor in the Department of Plant Sciences at the University of Saskatchewan. I grew up on a small farm in Ontario and went on to obtain a B.Sc. in Biology and a B.Ed. from Queen's University, then PhD in Land Resource Science from the University of Guelph. I specialize in *Environmental Agronomy*, and focus on nutrient cycling and crop nutrient use, soil health, and greenhouse gas emissions in agroecosystems. My research program is to understand the mechanisms that regulate nutrient and carbon transformations, and the implications for soil ecosystem services.

What kind of work do you do? As a Professor of Environmental Agronomy, I research farming practices that enhance food production in an environmentally friendly way. I design experiments and measure nutrient cycling, soil health, and greenhouse gas

emissions with the goal of understanding regenerative agriculture. I am grateful that my profession takes place in various settings: in the field, in the lab, in the classroom, and in the office—that come together to bring meaning in the form of student development, discovery, collaboration, and impact on the industry and environment.

How did you get interested in soil science? My interest in soil science traces back to my early years, when as a child I would garden with my mom. We would plant almost every vegetable I can think of—and tend to them all summer. I remember creating networks of pathways in the soil between the plant rows, where my brother and I would drive our dinky cars and Tonka trucks. Soil was a big part of our game! We would move it around, push it up against the plant roots, haul away the weeds, and hunt for earthworms which always showed up after the rain. Sometimes we would expand our tracks to the compost pile out back, where our pumpkins seemed to grow more voraciously than anywhere else. Immersed in that garden was where I started to think about the soil, consider different aspects of plants, and to appreciate the environment. I am grateful for that experience and to be able to carry it into my career today.

Melissa M. Arcand, Assistant Professor, Department of Soil Science University of Saskatchewan



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I am a soil biogeochemist with research interests focused on nutrient cycling, carbon storage, and soil health in agroecosystems. I grew up on a farm on the Muskeg Lake Cree Nation in central Saskatchewan, Treaty 6 Territory, and received my PhD in Soil Science from the University of Saskatchewan where I now do research and teach. I am the academic advisor for students in the *Kanawayihetaytan Askiy* program, designed to train students to work in resource management and land governance in Indigenous communities across Canada.

What kind of work do you do? I get to do exciting research and participate in teaching students soil science, but I also get to work closely with Indigenous students in the area of land and resource management in First Nations communities.

What is the most exciting thing about your job? The most exciting thing about my job is seeing students get excited about soil for the first time – usually when we're in the field together getting our hands dirty!

What is your favourite soil? My favourite soil is the Blaine Lake association, which is a Black Chernozem developed on medium textured lacustrine parent materials – and is the soil that supported much of my family's farm.

Yvonne Lawley, Assistant Professor, Department of Plant Science University of Manitoba



Yvonne Lawley (licensed under a CC BY-NC-ND license)

My area of research is agronomy and cropping systems with a focus on crops including soybeans, corn, and wheat, and a range of soil management practices from residue management, strip tillage, to cover crops. My research involves both small plot and on-farm field scale agronomy research. I enjoy communicating the results of my research to a wide range of audiences including farmers, agronomists, scientist in a range of disciplines, and especially in the classrooms where I teach at the University of Manitoba.

What kind of work do you do? As a professor, I teach about crop production and soil management every day. It may be in a large class, a small class, mentoring graduate students, or giving a presentation to agronomists and farmers. I love my job because I also get to ask new questions and learn every day. I work with graduate students to research new ways to grow food that are reliable for

farmers, good for our environment, and healthy for us to eat. My research looks at ways to improved crop and soil management on the Canadian prairies. Farmers have to make many decisions about how to grow their crops. Some of the decision we are examining in my research include: seeding rates, planting dates, fertilization, crop rotation, cover crops, and residue management. I am always excited to tell farmers about what we have learn from our experiments and to hear what new questions they have about growing crops or managing soil on their farms.

What did you like to do when you were a kid? When I was young, I loved to collect things from nature. I had a rock collection. I would pick up sticks everywhere I would find them. I spent hours finding acorns under the oak tree in front of my house. There were so many different types of flowers, leaves, and plants to find in the garden. When I was in school and university I loved learning how to identify, compare, and organize all these things I found in nature.

Caroline Halde, Professor, Département de phytologie Université Laval



Caroline Halde (licensed under a CC BY-NC-ND license)

I am an assistant professor in agroecology at Université Laval in Quebec City, QC, Canada. I grew up on an organic dairy farm near Montreal, QC and received a B.Sc. in Agronomy from Université Laval, a M.Sc. in Agriculture from Dalhousie University, and a Ph.D. in Plant Science from the University of Manitoba. I completed postdoctoral studies in plant ecology at Cornell University, working on weed control in organic farming systems. My research interests are interdisciplinary and include agronomy, soil health, weed ecology, and cover crops.

What kind of work do you do? As a university professor, I mentor students in their learning process about agriculture. I lead the Agroecology Lab at Université Laval, a research team investigating the impact of farming practices on soil health and crop production. Along with my team, I design and conduct field and lab experiments to test innovative farming practices, in collaboration with farmers. I also teach ecological farming to undergraduate and graduate students. Finally, I serve on a number of university committees, performing a variety of administrative duties.

What is your favourite soil? My favourite soil is the Sainte-Rosalie, a Orthic Humic Gleysol found in the St. Lawrence Valley in Quebec, Canada. This Gleysol is derived from marine deposits accumulated at the time of the Champlain Sea. It is a well-structured soil, with very fine clay, poorly drained, and rich in organic matter in the surface horizon. When drained, soils from the Sainte-Rosalie series are very fertile. It is the 3rd series of soils in importance among the soils mapped in Quebec. The Sainte-Rosalie soil, present on my family farm in southern QC, has been farmed by my ancestors for more than 200 years.

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3.3: Soil Reclamation and Remediation of Disturbed Lands

M. Anne Naeth, P. Michael Rutherford, and Allen M. Jobson

Learning Objectives

Upon completion of this chapter, students will be able to:

1. Define and differentiate common terms such as reclamation, remediation, and restoration
2. Describe the goals of reclamation and remediation
3. Understand how fundamental concepts in soils apply to reclamation and remediation
4. Understand the basics of common chemical, physical, and biological remediation strategies
5. Understand the challenges of reclamation and remediation and how to address them
6. Understand the beneficial use of waste materials and other amendments in reclamation and remediation

INTRODUCTION

Soil reclamation and soil remediation are components of the land reclamation process that take place on the landscape after disturbance of ecosystems and sites. These processes are critical to overall land reclamation success and are built around the premise that we can manipulate and accelerate soil development processes by altering specific soil properties.

Land reclamation is the process of converting disturbed or damaged land to its former or other productive uses. Reclamation encompasses all disturbed components of an ecosystem, including but not limited to, soils, hydrology, flora, and fauna.

Ecological restoration is the process of assisting recovery and management of ecological integrity; ecological integrity includes a critical range of variability in biodiversity, ecological processes and structures, regional and historical context, and sustainable cultural practices (Gann et al., 2019).

Remediation is the process of improving a contaminated site to prevent, minimize, or mitigate damage to human health and/or the environment. Contamination may be present in soils, surface water, and/or ground water. **Soil remediation** refers to a wide range of strategies that remove, destroy, contain, transform, or reduce availability of soil contaminants to humans and other receptors in the environment. These strategies may be conducted *in situ* or *ex situ*.

Revegetation is the process of providing barren or denuded land with a vegetation cover which replaces or tries to replicate pre-disturbance ground cover. Revegetation may be part of the reclamation strategy to reclaim or remediate soils.

Land reclamation is the umbrella term, with ecological restoration being a specific type of reclamation. Remediation and revegetation are specific components of the overall process.

Reclamation and remediation are usually associated with criteria developed within various jurisdictions. In many jurisdictions, reclamation is required by regulatory bodies. Normally these requirements are to return the land to equivalent capability, land performance that focuses on the degree and nature of limitation imposed by the physical characteristics of a land unit on a certain use, assuming a specific management system.

DISTURBANCES REQUIRING LAND RECLAMATION

Disturbance refers to variation in some ecosystem factor beyond its normal range of variation resulting in ecosystem change. A change in a system that interferes with its capacity to maintain a maximum range of tolerances for life, is known as degradation. Destruction refers to the complete degradation of a system to the point that it would have to be rebuilt to be useful, including changing of one ecosystem for development of another.

Disturbances can occur on the landscape as a result of human activities, such as mining, agriculture, and building of cities, towns, roads, pipelines, and power transmission lines. Surface mining generally causes more extensive landscape disturbance than underground mining. Disturbances can also occur from natural events such as floods, fires, hurricanes, and landslides. The intensity and properties of the disturbance will determine whether the land is degraded and whether human intervention is required to ameliorate the disturbance.

Soil disturbances change physical, chemical, and/or biological properties. For example, soil pH may be high or low relative to that tolerated by plants, soil fauna, or microorganisms. Salt contaminants may increase soil electrical conductivity. Physical mixing of horizons may occur when trenches or mines are dug, changing water and nutrient holding capacity. Use of heavy equipment can compact soils, increasing soil bulk density and penetration resistance, and reducing infiltration and percolation rates. Many of these changes can have serious impacts on plant rooting and microbial community function.

As land reclamation has become a requirement of companies seeking to alter the landscape for natural resource operations and/or land use changes, disturbances have become minimized and more controlled. In this way, land reclamation is easier to accomplish and the environmental impacts of the disturbance are reduced. Soil handling is one of the most important considerations in disturbances for natural resource extraction, and one that can be managed.

Disturbances associated with oil and gas operations, such as pipelines and well sites, now have strict construction guidelines in most jurisdictions. For example, soil salvage is mandatory and requires horizons of soil (layers) to be removed and stored separately. Upon pipeline installation or well site decommissioning, the soil horizons are required to be replaced in proper order.

Surface mining is considered one of the most intensive human disturbances on a landscape due to its aerial extent and size of the excavation. Although soils are stripped and require replacement to mimic the pre-disturbance condition, there is often a shortage of soil due to the mined material being removed from the system. Since the excavations are so large, there are major disruptions of the hydrologic regime. Excavations through overburden to desired ore bearing rock, minerals, coal, or bitumen create vast volumes of waste rock, unconsolidated soil materials, and/or tailings materials. Bitumen mining (oil sands) results in overburden and tailings ponds. These intense disturbances often require soil building and creation of end pit lakes to accommodate soil loss.

SOIL RECLAMATION

Soil reclamation is part of the overall land reclamation process and is usually addressed after removal of infrastructure, recontouring, and contaminant remediation, and before revegetation (Figure 16.1). An important concept in soil reclamation is end land use. Different criteria and reclamation processes may be followed depending on how the land will be used after reclamation. For example, regulatory requirements are more stringent for agricultural land and land that people will occupy than land used for industrial purposes. These potential end land uses can be similar to what was there prior to the disturbance or can require a complete change of ecosystem.





Figure 16.1. A typical coal mine reclamation process includes filling in the pit after coal removal **(top)**, recontouring to fit the adjacent landscape **(middle)**, and then seeding with desired plant species **(bottom)**. Photographs courtesy of M. Anne Naeth, licensed under a CC BY (Attribution) license.

Soil Replacement and Amendments

When soil has been stripped in horizons and stockpiled during disturbance it is normally replaced in its natural horizon sequence. Although admixing of horizons can be an issue, careful placement techniques can mitigate it. The most likely issue is compaction when the soil is replaced and this can be mitigated through tillage operations. Much damage to soils can be reduced by having construction activities occur under frozen, or dry conditions. Soil quality can be retained if soils are stockpiled only for short periods of time.

If soil chemical properties do not meet end land use requirements, amendments can be used to bring soils to an appropriate criteria. A common issue in soil reclamation is lack of organic matter and impacts on its associated properties. This can be addressed by adding an organic amendment such as animal manure or straw from a local agricultural operation, or forest floor material (LFH) from an area that has been deforested. Another common issue is lack of soil nutrients necessary for revegetation. These deficiencies can best be addressed by adding commercial fertilizers composed of the necessary nutrients such as nitrogen, phosphorus, and potassium. Fertilizers can be formulated in bulk to meet specific needs of a reclamation site.

A common and environmentally important process is to use waste materials in soil reclamation. This has the added benefit of removing many materials from the landfills where they would otherwise be placed. Sewage is available from waste treatment facilities from most mine camps and is a good source of nutrients and organic matter (Figure 16.2). However, there are concerns with raw sewage use due to possible presence of salts, metals, and disease causing microorganisms. Biosolids are widely used as soil amendments. They are sewage sludges which have been treated to lower pathogen risk. Other amendments can be used in reclamation, including calcium sources, paper mill waste, industrial sludges, biochar, wood ash, and commercial products. Each have specific properties than can be aligned with the soil properties that need improvement, and each often come with associated negative properties that require balancing. Many jurisdictions have regulations or guidelines for use of wastes and industrial by-products as soil amendments.



Figure 16.2. Sewage from a diamond mining camp to be used as an organic amendment in soil building. Photograph courtesy of M. Anne Naeth, licensed under a CC BY (Attribution) license.

Topographic variability of features on a landscape, can play an important role in determining plant communities, as can the scale of these features. Thus, micro topographic variability is needed on the constructed, reclaimed landscape to provide micro sites for plants and microorganisms, to reduce erosion, and to be aesthetically pleasing. This can be provided through features such as mounds, furrows, depressions, rocks, boulders, coarse textured substrates, mulches, and cracks.

Soil Building

When there is insufficient soil to be replaced or amended after a large scale disturbance, soil must be constructed. These human constructed soils are called Anthroposols (Naeth et al., 2012). Anthroposols can include soils that have had layers (of at least 10 cm depth) removed or modified by human activity, or where manufactured materials have been added to soil; overall human activity has altered the soil forming factors and a new pedogenic trajectory may have been introduced. Anthroposols are classified at the great group level based on material composition of the layers, specifically organic carbon content and presence of anthropogenic artefacts (materials made by humans such as plastic, glass, garbage, concrete).

Soil building requires several steps. The end land use for the disturbed area is assessed and the appropriate soil for that purpose delineated. To build a soil requires a mineral substrate, which can generally be found or produced in large amounts, and it forms the main mass of an anthroposol. Amendments can be added in smaller amounts to improve physical, chemical, and/or biological properties such as nutrient content, infiltration capacity, or pH. Selection of appropriate amendments and substrates or combinations is based on physical and chemical properties, cost, and availability of materials. The materials and combinations of materials are layered similar to natural soil horizons or may be incorporated into the soil.

Use of waste materials to build anthroposols is increasingly common, especially when a local source is lacking. Various waste materials are produced from different industries and human activities, including crushed ore and waste rock (mining), pulp and paper sludge (paper milling), and sawdust and wood chips (lumber milling) (Figure 16.3). Sewage, biosolids, and compost can be sourced from urban or industrial sites, and manure from agricultural sources.



Figure 16.3. Spoil from the mining process is stockpiled for use in reclamation. Photograph courtesy of M. Anne Naeth, licensed under a CC BY (Attribution) license.

Mineral materials are highly variable due to differences in industrial practices (extraction, processing) and source (parent material, location). They often have low organic matter and nutrient content and poor water holding capacity. They can have elevated metals and cause acid mine drainage (sulphur rich rock).

Many chemical, physical, and biological soil properties must be considered when building anthroposols as they are commonly limiting factors, especially when using waste materials. For example, waste rock containing pyrites can cause acidity problems, and coarse textured materials usually have low nutrient and water holding capacity, and tend to be prone to erosion. Thus the soil amendments must be closely matched with the property changes that are needed in the reclamation soil, being careful not to add any other constraints.

Erosion Control

Erosion is a problem at most disturbed sites prior to and during reclamation as the soil is bare and devoid of vegetation, exposing the surface to wind and water erosion. Erosion can be reduced on reclamation sites with contouring, amendments, and various materials such as erosion control blankets, and mulch. Many techniques, called bioengineering, include use of plant materials with techniques such as live staking (i.e., planting individual live wood cuttings into the soil). Many erosion control products are very effective, although some can add considerable cost to the reclamation project.

SOIL REMEDIATION

Soils may become contaminated through a variety of means, broadly categorized as point sources or non-point sources. Point sources include leaking fuel storage tanks, spills at industrial sites, waste disposal areas, and landfill areas. Non-point sources include atmospheric deposition. Although soils may be contaminated with non-chemicals, such as harmful bacteria or viruses, chemical contaminants are often the main issue.

The Canadian Council of Ministers of the Environment defines contaminants as chemical substances whose concentration exceeds background concentrations or that do not naturally occur in the environment (CCME 1997). Contaminated soils contain chemical substances at concentrations considered unsafe by regulators. The term pollutant is sometimes used, and is defined as a chemical or substance out of place, or present at higher than normal concentrations, that has adverse effects on non-target organisms (Pierzinski et al., 2005).

Soil remediation strategies entail removal or destruction of contaminants, reducing mobility of contaminants, or transforming them to less toxic forms. The main goal is to minimize risk to humans and other receptors.

Common Soil Contaminants

Soil contaminants can be broadly grouped as organic and inorganic. Organic contaminants include pesticides, organic solvents, and other industrial products (Table 16.1). The most common organic contaminants (mass or volume) at contaminated sites in Canada are petroleum hydrocarbons (PHCs), including crude oil and refined products such as fuels and lubricating oils. Dioxins, furans, and polycyclic aromatic hydrocarbons (PAHs) are organic compounds.

Table 16.1. Common soil contaminants

Common contaminants	Common sources
Organic contaminants	Refined and unrefined products, fuels, lubricants, oil
Petroleum hydrocarbons (PHCs)	Creosote, coal tar, asphalt, incomplete combustion, oil, wood
Polycyclic aromatic hydrocarbons (PAHs)	Degreasers, dry cleaning facilities, plastics industry
Halogenated hydrocarbons, aromatics	Electrical equipment, hydraulic oil
Chlorinated solvents	Incomplete combustion of waste containing chlorine
Polychlorinated biphenyls (PCBs)	Agriculture, pest control
Dioxins, furans	Chemical industries, military explosives
Pesticides	
Alcohols, ethers, nitrogen based	
Inorganic contaminants	Chemical, industrial, mine wastes, firing ranges, fly and bottom ash, metal recycling, battery disposal
Heavy and trace metals, metalloids	Acid rock drainage, chemical spills
Acids, bases	Biosolids, manures, salt storage, oil and gas industry
Excess nutrients and salts	Oil and gas, phosphate industry, uranium mining
Radionuclides	
(Adapted from Knox et al. 1999, CCME 2016)	

Inorganic contaminants include metals or salts, often as anions or cations. They can be beneficial to soil ecosystems at background concentrations, although at high concentrations may pose a risk to human or non-human receptors. For example, copper, molybdenum, cobalt, and zinc are important plant micronutrients, but are contaminants at high concentrations. Non-metal anions such as chloride can build up in soils from de-icing operations or chemical spills. At high concentrations they can increase soil electrical conductivity or have direct impacts on plants.

High concentrations of nutrients such as nitrogen, phosphorus, and sulphur, can be considered contaminants. They originate from waste materials that are inappropriately stored or disposed of on land, such as livestock waste applied on agricultural land. Application of these waste materials can be part of remediation and reclamation processes of contaminated sites.

Properties of Soils and Contaminants Relevant to Remediation

Selecting an appropriate remediation strategy is dependent on contaminant and site properties, including soil properties. For example, petroleum hydrocarbons vary in molecular weight, water solubility, volatility, and density; gasoline tends to have lower molecular weight and higher volatility than lubricating oils. Larger hydrocarbon compounds tend to be less volatile and less water soluble than smaller ones. Large, non-soluble hydrocarbons can exhibit strong sorption (are sorbed) to native soil organic matter (high sorption coefficient) making them difficult to degrade or remove, particularly in fine textured soils due to greater sorption and potential for entrapment within soil aggregates, and reduced movement of air and water.

Inorganic contaminants, such as metals, can be present in different forms in soils. Distribution among the various components will be affected by soil texture, mineralogy, pH, cation and anion exchange capacity, organic matter content, and aeration status. Remediation strategies need to take the various forms of chemical species into consideration.

Remediation Strategies

Remediation strategies for contaminated soil can be broadly classified as biological, chemical, or physical (Table 16.2). Many are a combination of more than one type, and can be applied either *ex situ* or *in situ*. *Ex situ* methods involve excavating the soil from the impacted zone and treating the soil either on or off site. *In situ* methods treat the contaminated soil in place. Contaminated sites often have contaminated surface and ground waters that may require different or complementary remediation strategies.

Table 16.2. Common remediation strategies for contaminated soils

Type	Category	Common contaminants	Location
Biological	Bioremediation, land farming, engineered biopiles or windrows	Petroleum hydrocarbons	Either
	Bioventing	Petroleum hydrocarbons	*In situ*
	Phytoremediation (rhizodegradation, phytoextraction, phytodegradation, phytostabilization, phytoextraction)	Petroleum hydrocarbons, other organic contaminants, metals	Generally *in situ*
	Slurry phase reactors or lagoons	Various organics	*Ex situ*
Chemical	Chemical oxidation / reduction	Organics, inorganics	Either
	Neutralization	Acids, bases	Either
Physical	Soil vapour extraction (SVE)	Volatile organics (gasoline)	*In situ*
	Soil flushing, soil washing, dual phase extraction in ground water	Organics, inorganics	Either
	Solidification, stabilization	Organics, inorganics	Either
	Electrokinetic separation	Organics, inorganics	Either
	Thermal desorption	Organics	Either
	Incineration	Organics	*Ex situ*
	Vitrification	Organics, inorganics	Either

Bioremediation is a *biological treatment* that uses soil microorganisms to treat contaminants. It can use natural biodegradation processes, or enhance them by adding commercial or other microbial preparations (bioaugmentation), or adding nutrients such as nitrogen or phosphorus and/or electron acceptors such as oxygen (biostimulation). Bioremediation strategies focus on microbial degradation of organic contaminants, although microbial induced chemical changes of inorganic chemicals can be considered bioremediation, such as converting more toxic chromium to the less toxic chromium.

Bioremediation can be applied in various technologies, such as landfarms, biopiles, or engineered reactors. Bioventing is an *in situ* treatment that enhances microbial degradation of contaminants through enhanced aeration in the contaminated zone. Use of plants to remove (phytoextraction), stabilize (phytostabilization), or destroy (phytodegradation) contaminants is a biological treatment

called phytoremediation. Rhizosphere bacteria can play an important role in biodegradation of organic contaminants during phytoremediation (rhizodegradation).

Chemical treatments can be used to convert contaminants to non-hazardous or less toxic chemicals, or to forms which are less mobile, more stable, or inert. Chemical oxidation using hydrogen peroxide, ozone, or permanganate can destroy compounds such PAHs, and is best for liquids such as ground water or soil slurries. Chemical reduction can convert toxic chromium to less toxic chromium and degrade some chlorinated organic solvents such as trichlorothene. Neutralization involves adding materials to adjust the pH of highly acidic or caustic soil.

Physical treatments use properties of the contaminant or contaminated medium to separate or immobilize the contaminant. Chemical additives can improve removal efficiencies. *In situ*, soil vapour extraction creates a vacuum in the vadose zone, draws vapours into an extraction well, then treats or destroys them above ground. Soil flushing occurs *in situ*, where the contaminated zone is treated with a solution, and mobilized contaminants are brought to the surface for disposal, treatment, or re-circulation. Treatment solutions can be water, basic, acidic, chelating or complexing, reducing, co-solvents, or surfactants. Dual phase extraction involves installation of piping to ground water and vacuum extraction of water and organic contaminants such as petroleum hydrocarbons. Soil washing is done *ex situ* and may involve separating soil fines (silt, clay) from coarse material thereby reducing the total mass and volume that needs to be treated.

Solidification and stabilization prevent or slow release of contaminants from soil by treatment with binding agents such as cement or asphalt. The soil may be treated with a material that will reduce contaminant solubility and mobility. For example, phosphate based materials often reduce mobility of lead, by converting lead into insoluble lead phosphate compounds. Hydrogen sulphide can convert many metals into insoluble respective pyrites.

Electrokinetic separation sends a low intensity direct current through the soil to separate metals, radionuclides, and organic contaminants. Positively charged species migrate to the cathode, and negatively charged species, such as inorganic and organic anions, migrate to the anode.

Thermal treatments are mainly used to remediate organic contaminants. Thermal desorption is typically *ex situ*, with heat applied to soil to evaporate or vapourize contaminants which are collected, treated, or destroyed. Temperatures of 90 to 320°C are required to treat volatile organic contaminants such as fuel components, 320 to 540°C to treat semi-volatile organics such as PAHs, PCBs, and lubricants. Incineration involves heating soil *ex situ* to 870 to 1370°C to destroy organic contaminants. Vitrification heats soil to 1400 to 2000°C to melt silica and convert it to stable glass and crystalline solids. Heat may be generated by electrodes and electrical resistance or plasma arc technology. Organic contaminants are volatilized, trapped, and treated or destroyed and inorganic contaminants, including radionuclides, are encased.

Manipulating Soil Processes to Enhance Remediation

Understanding soil properties and processes facilitates remediation of contaminated soils. For example, in bioremediation of petroleum hydrocarbon contaminated soils, a parent hydrocarbon compound is biodegraded to a simpler one or degraded completely to carbon dioxide. Three main factors influencing organic contaminant biodegradation can be manipulated. They include compound or chemical properties such as solubility, molecular weight, and sorption potential; microbial capabilities, such as whether microorganisms have appropriate enzyme systems to biodegrade the compound; and environmental conditions such as temperature, and soil pH, electrical conductivity, aeration status, water content, nutrient content, organic matter content, texture, and structure.

Crude oil contains hundreds of individual compounds of various molecular sizes, branching, and solubility. Larger, more complex compounds are generally less soluble and more strongly sorbed (or partitioned) to soil. Most surface soil horizons contain a consortium of heterotrophic microorganisms capable of degrading petroleum hydrocarbons. Soil organisms can be inoculated into the soil (bioaugmentation) in the bioreactor or landfarm, with soil known to have petroleum degrading microorganisms. Petroleum hydrocarbon bioremediation works best if petroleum concentrations are not so high that free product is present; concentrations less than 10% by mass are ideal.

The soil environment may be manipulated to enhance biodegradation. Petroleum degradation is generally most rapid under warm, moist, aerobic conditions near neutral pH. Soil pH can be modified with lime, wood ash, or elemental sulphur. Most soil microorganisms are mesophiles and biodegradation can be optimized at 25 to 35°C. The reactor can be aerated to enhance oxygen

levels, or heated in cold weather. Oxygen levels are between 1 and 10% in soil air and aerobic microbial activity is most rapid just below field capacity. Assuming no hydrophobicity, soil being land farmed or treated can be kept moist, near field capacity. Heterotrophic bacteria and fungi utilize petroleum hydrocarbons as a source of energy and carbon for growth. To create new cells, microorganisms need carbon, nitrogen, phosphorus, sulphur, and other nutrients. Nitrogen and sometimes phosphorus tend to be needed in greatest amounts; however, petroleum hydrocarbons are typically lacking in these elements. Crude oil typically has approximately 85% carbon. Assuming a carbon utilization efficiency of 50% and a mean cell C:N ratio of 10, microorganisms need approximately 1 mass unit of nitrogen per 20 mass units of petroleum carbon. Levels of 1 mass unit nitrogen per 40 to 50 mass units petroleum carbon may be sufficient as not all petroleum carbon will be degraded at the same rate. Nitrogen can be added as fertilizer such as urea or ammonium sulphate, or nitrogen rich organic sources such as manures and biosolids. Phosphorus should be available in a soil hydrocarbon mixture at an appropriate hydrocarbon carbon to phosphorus ratio of at least 50:1.

Not all petroleum hydrocarbons in soil are readily bioavailable. Compounds must come into contact with microbial enzymes for degradation to occur. Contaminants may be strongly sorbed to the soil, trapped within pores, or in a separate phase away from microorganisms and/or their enzymes. This partitioning or sorption effect is most prevalent for larger, hydrophobic petroleum compounds, and aged or weathered contaminated soils. For *in situ* hydrocarbon bioremediation as much free product hydrocarbon as possible should be removed through vacuum extraction, then the remaining adsorbed hydrocarbon can be addressed by bioremediation.

The soil can be manipulated to enhance bioavailability. Techniques include adding surfactants to decrease strong soil sorption and increase apparent solubility of petroleum hydrocarbons. Soil aggregates can be disrupted to improve contact between microbial enzymes and contaminants. Some systems, such as in slurry phase reactors, are ideal for this, but can greatly increase the treatment costs.

SUMMARY

- Soil reclamation and soil remediation are components of the land reclamation process whereby a disturbed landscape is returned to former or other productive uses; the goals of soil reclamation are dictated by the nature of the disturbance and the desired end land use.
- Soil reclamation of highly disturbed landscapes makes use of our understanding of soil formation and genesis; humans can accelerate soil forming processes in part through addition of various amendments, which includes utilization of various waste residuals and plants in the revegetation process.
- Anthroposols are soils which are built through human intervention.
- Soil remediation is a subcomponent of soil reclamation and is focused on removal, reduction, or containment of organic and/or inorganic contaminants with the goal of minimizing risk to humans and/or environmental receptors.
- A wide range of biological, chemical, and physical strategies are available to remediate contaminated soils; selection of the most appropriate technology is in part dictated by the nature of the contaminants and of the impacted soil and site.
- Bioremediation is a common approach to remediate petroleum hydrocarbon contaminated soils; the process can be optimized by manipulating soil and environmental conditions to enhance microbial degradation of the target compounds.

SUGGESTED READING

Alexander, M. 1999. Biodegradation and bioremediation. 2nd edition. Academic Press.

Adriano, D.C., J.M. Bollag, W.T. Frankerberger, and R.C. Sims. (co-editors). 1999. Bioremediation of contaminated soils. Agronomy Series 37. American Society of Agronomy, Crop Science Society of America, Soil Science Society of America, Madison, Wisconsin, USA.

Federal Remediation Technology Roundtable (FRTR). (ND). Remediation technologies screening matrix and reference guide 4.0. At https://frtr.gov/matrix2/section1/toc.html.

Naeth, M.A., H.A. Archibald, C.L. Nemirsky, L.A. Leskiw, J.A. Brierley, M.D. Bock, A.J. VandenBygaart, and D.S. Chanasyk. 2012. Proposed classification for human modified soils in Canada: anthroposolic order. Canadian Journal of Soil Science 92:7-18.

Pichtel, J. 2017. Fundamentals of site remediation. 2017. 2nd Edition. Government Institutes, an imprint of The Scarecrow Press Inc., Lanham, Maryland.

Reynolds, C.M. and H.D. Skipper. 2005. Bioremediation of contaminated soils. Pp 536-561 in Sylvia, D.M, P.G. Hartel Fuhrmann, and D.A. Zuberer. Principles and applications of soil microbiology. 2nd edition. Pearson Prentice Hall, Upper Saddle Riever, New Jersey.

Riser-Roberts, E. 1998. Remediation of petroleum contaminated soils – biological, physical and chemical processes. Lewis Publishers.

Suthersan, S.S., J. Horst, M. Schnobrich, N. Welty, and J. McDonaugh. 2017. Remediation engineering – design concepts. 2nd Edition. CRC Press, Taylor and Francis Group, Boca Raton, FL.

STUDY QUESTIONS

1. A 1 ha well site lease with little topsoil requires straw to increase organic matter in the upper 0.15 m. The topsoil has 1% organic matter, bulk density 1.50 Mg m^{-3}. Adjacent soils have 5% organic matter and this is required for the reclamation site. Determine the number of 1,200 lb round bales to get the upper 15 cm to 5% organic matter. To do this, calculate the mass of soil to be amended. Assume the organic carbon content of wheat straw is 45.5%. Account for mineralization given that bacterial respiration releases approximately 60% organic matter as CO_2 and 40% into soil.
2. Determine how much beef manure to apply to a fescue grassland, if incorporating into the soil 1 day after spreading. Assume recommended plant nutrient requirements for fescue grassland are 30 kg ha^{-1} nitrogen and 15 kg ha^{-1} P_2O_5.
3. A study was done to determine how much fertilizer should be added to a field to maximize the bioremediation of crude oil that was spilled at an industrial site. A laboratory incubation showed that the maximum rate of petroleum hydrocarbon (PHC) biodegradation occurred when 25.0 mass units of contaminant carbon is treated with 1 mass unit of nitrogen. Based on this relationship, calculate how many tonnes of urea fertilizer (45-0-0) needs to be applied per hectare (ha) to treat the top 15.0 cm depth of a crude oil contaminated field (crude oil concentration is 1.25 g oil 100-g^{-1} dry soil = 1.25% by weight). The fertilizer will be mixed into the top 15.0 cm of the field using a disker or rototiller after surface application. The bulk density of the soil is 1.15 Mg m^{-3}; and, the crude oil contains 85.0% carbon. Assume the N content of the urea is 45.0%
4. A greenhouse experiment was conducted to determine the feasibility of using the plant Indian mustard (*Brassica juncea*) to phytoremediate a lead (Pb) contaminated soil (1640 mg lead kg^{-1} soil). Indian mustard was seeded into the contaminated soil (in 4 L pots) and pots were placed in the greenhouse for 12 weeks (were watered and fertilized as needed). At the end of the experiment, the plants were harvested and the average concentration of Pb was determined in the above ground biomass (roots left in soil). The Pb concentration in above ground biomass was 1215 mg kg^{-1} dry biomass. Assume that typical above ground yield for the region of interest is 10.0 tonnes dry biomass per ha per growing season; soil bulk density is 1.22 g cm^{-3} (1.22 Mg m^{-3}), and Pb concentration in the field grown plants is equal to that in the greenhouse study. Calculate how many growing seasons (years) it would take the Indian mustard to reduce the soil Pb concentrations in a hectare furrow slice (HFS = 10,000 m^2 x 0.15 m = 1500 m^3 of soil) from 1625 to 300 mg Pb per kg of soil. In your opinion, is this remediation strategy practical for this level of Pb contamination?

REFERENCES

Canadian Council of Ministers of Environment (CCME). 1997. Guidance document on the management of contaminated sites in Canada. Canadian Council of Ministers of Environment PN 1279 C97-980244-X.

Canadian Council of Ministers of Environment (CCME). 2016. Guidance manual for environmental site characterization in support of environmental and human health risk assessment. Volume 1 guidance manual. Canadian Council of Ministers of Environment PN 1551 ISBN 978-1-77202-026-7 PDF.

Gann G.D., T. McDonald, B.Walder, J. Aronson, C.R. Nelson, J. Jonson, J.G. Hallett, C. Eisenberg, M.R. Guariguata, J. Liu, F. Hua, C. Echeverria, E.K. Gonzales, N. Shaw, K. Decleer, and K.W. Dixon. 2019. International principles and standards for the practice of ecological restoration. Second edition. *Restoration Ecology* S1-S46.

Knox, A.S., A.P. Gamerdinger, D.C. Adriano, R.K. Kolka, and D.I. Kaplan. 1999. Sources and practices contributing to soil contamination. Pp 53-87 in Adriano et al. (co-editors) Bioremediation of contaminated soils. Agronomy Series 37. American Society of Agronomy, Crop Science Society of America, Soil Science Society of America, Madison, Wisconsin, USA.

Pierzynski, G.M., J.T. Sims, and G.F. Vance. 2005. Soils and environmental quality. Taylor and Francis Group.

United States Environmental Protection Agency. 2017. Superfund remedy report. 15th edition. Office of Land Emergency Management, EPA-542-R-17-001. At: https://www.epa.gov/remedytech/superfund-remedy-report.

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Anne is a professional agrologist (PAg) and professional biologist (PBiol). She served as Associate Dean (Academic) and Associate Dean (Research and Graduate Studies). She worked in government and industry and served on the Alberta Environmental Appeals Board, on executives of numerous professional organizations and boards. Her land reclamation research program has led to extensive publications, policy changes, and development of a new soil order (Anthroposols). She has supervised over 75 graduate students and mentored dozens of postdoctoral fellows, research associates and undergraduates. She is a Vargo Distinguished Teaching Chair, Fellow of the Canadian Society of Soil Science, Fellow of the Society for Teaching and Learning in Higher Education, and recipient of numerous awards recognizing scholarship, teaching and community service, including the CLRA Noranda Land Reclamation Award, Mentors of the Millenium, Alumni Award of Excellence, Killam Professorship, Distinguished Agrologist, Rutherford Award of Excellence for Undergraduate Teaching, 18 Faculty Teacher of the Year Awards, the prestigious 3M Fellowship, and the University Cup (highest award from University of Alberta). She received the first University of Alberta NSERC CREATE grant to establish the *Land Reclamation International Graduate School* (LRIGS), the first school of its kind in the world for multi-disciplinary training and professional development of land reclamation students.

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Mike joined the Environmental Science program at UNBC in 2001. He teaches and conducts research into topics related to soil science, environmental quality and waste management. His research has focused on a wide range of topics including: biological interactions in soils, carbon and nitrogen cycling, radon and natural-occurring radioactivity, remediation and reclamation, and, the beneficial use of residuals (such as wood ash) as soil amendments. He is a P.Ag. that enjoys working on practical issues related to soil science.

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Al joined the Alberta Research Council in 1975 as a petroleum bioremediation specialist, moving to the Alberta Environmental Research Center (Vegreville) in 1978 as Head of Environmental Microbiology. In pursuit of other interests in agriculture Al moved to Saskatchewan in 1980 and farmed until 1987. He conducted reduced rate herbicide application trials with shrouded sprayers throughout Saskatchewan for Rogers Engineering Ltd. (Saskatoon). In 1991 he joined Beak Associates in Saskatoon which became SENTAR, then Stanley Environmental, then STANTEC (1994). Al consulted widely on soil environmental remediation projects for STANTEC in Alberta, Saskatchewan, Manitoba, Nebraska and Idaho. In 2008 he joined Ostrem Chemical Co. Ltd. (Edmonton) as a Product Development Specialist designing more environmentally acceptable products. From 1998 to 2018 Al was an adjunct professor in the Department of Renewable Resources, Faculty of Agricultural, Life and Environmental Sciences, University of Alberta.

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3.4: Digital Soil Mapping

Brandon Heung, Daniel Saurette, and Chuck Bulmer

Learning Objectives

Upon completion of this chapter, students will be able to:

1. Describe and rationalize a transition from conventional soil information to digital soil information
2. Link theories of pedogenesis to applications of digital soil mapping
3. Provide an overview of how digital soil information is used to generate soil maps

INTRODUCTION

Globally, Canada has the third and seventh largest asset bases in terms of forested and arable lands; hence, accurate and precise soil information is needed to maintain and improve our soil resources and to address significant environmental challenges, such as agricultural land loss or the decline in soil organic carbon and deterioration of soil health. Without information on the spatial patterns of soil, our ability to identify the most suitable locations to realize new agricultural and resource management opportunities, and to capitalize on our natural resources will be challenged in the future.

The demand for up-to-date soil information has been increasing to address emerging environmental issues such as sustainable food production; climate change regulation, adaptation, and mitigation; soil degradation; land resource management; and the provision of earth system services across all geographical extents (Sanchez et al., 2009; FAO and Global Soil Partnership, 2016). Additionally, better soil information is necessary for performing soil assessments; and the reduction and informing of risks for decision-making (Carré et al., 2007; Finke, 2012; Arrouays et al., 2014).

Canada has a long history of soil surveying—with the first soil survey being completed in Ontario in 1914 (McKeague and Stobbe, 1978)—and maps produced from the legacy soil surveys have been used to inform land management and planning for many years. Digitized versions of legacy maps are widely available online and are still being used to this day (e.g., British Columbia's Soil Information Finder Tool). Despite this, it has been well recognized by the soil science community that the approaches and techniques used in legacy soil survey may not be suitable for providing the precise and high-resolution information demanded by modern agricultural management and environmental assessment activities. With continued advances in computing, remote and proximal sensing technologies, and geographical information systems (GIS), soil surveying activities have transformed towards digitally-based techniques that can provide soil information that is more accurate and precise than what was previously available, and in an efficient manner.

This chapter will summarize the transition of conventional soil surveying approaches to digital soil mapping (DSM) approaches; present a theoretical framework of DSM; and provide an overview of how emerging technologies may be used to generate digital soil maps.

CONVENTIONAL SOIL MAPPING

Two major achievements contributed to the development of conventional soil survey methods in North America. The first achievement was the formalization of Hans Jenny's *Factors of Soil Formation* (1941) and the second was the formalization of national soil taxonomic systems in the US (Soil Taxonomy; Soil Survey Staff, 1975) and Canada (The Canadian System of Soil Classification, CSSC; Canada Soil Survey Committee, 1978; See Chapter 8). The classification systems described how soils were classified based on morphological properties that could easily be measured and quantified in the field. Jenny's *clorpt* model (Eqn. 1) characterizes the environmental conditions for which soils are found as a function of climate (*cl*), organisms (*o*), relief (*r*), parent material (*p*), time (*t*), and other local factors that influence soils (…):

(1) $S = f\,(cl,\ o,\ r,\ p,\ t,\ ...)$

The *clorpt* model was originally proposed as a method for studying how soils varied, quantitatively, as a function of various state factors. One of the key quotes that gets overlooked in Jenny's book is that he describes:

"The factors are not formers, or creators, or forces; they are ***variables*** *that define the state of a soil system."*

In other words, these factors are simply describing the environmental conditions from which a soil (and their properties) are found and therefore, these variables may be used as the basis for predicting soils using a model that relates the soil to the environment (i.e., soil-environmental model).

The adoption of Jenny's *clorpt* model led to dramatic improvements in soil survey compared to the early efforts. Soil descriptions and maps began to be organized around the factors of soil formation, especially by organizing soils into groups based on properties of their parent material, but also by linking fine-scale topographic variation to the boundaries between soils. This provided a rationale and system for linking soil unit boundaries to topographic features.

Soil classification systems have also led to the improvement of soil surveys because they focused on the most important soil properties for field surveyors to evaluate, and these were applied consistently across multiple survey projects, allowing for better correlation over large areas.

The conceptual improvements associated with the *clorpt* model of soil formation and the use of standardized soil profile information, coming from the application of a uniform soil classification system, were largely responsible for what we see today as the high information value of soil survey throughout the latter half of the 20th century. The digitization of these surveys has provided valuable information to land managers, and in some instances, it has also provided the raw material for the development of today's digital soil maps. This would not have been possible without these advancements.

Conventional Representations of Soil

Soils may be represented in several ways: a profile, pedon, polypedon, or map unit (Figure 17.1). The soil profile is a 2-dimensional representation and the pedon is a 3-dimensional representation that typically is 1–3 m laterally and 1–2 m vertically. In principal, the properties of the *pedon* should not vary horizontally (only vertically) and when several similar pedons are connected, it is referred to as a *polypedon*. The *polypedon* concept differs from a map unit because the map unit, a distinct polygon that is mapped by soil surveyors, is an areal representation of a polypedon or a polypedon complex. There are two types of mapping units: consociations and associations. Consociations are map units delineated based on a single taxonomic unit or series and may be referred to as a simple mapping unit whereas associations consist of two or more dissimilar soil taxa (multiple components) that occur in a pattern that is too complex to be resolved at the selected mapping scale (Hole and Campbell, 1985; Schaetzl and Anderson, 2005). In the case of complex map units, the proportion of each soil class within the map unit is specified in the map legend or the map symbol.

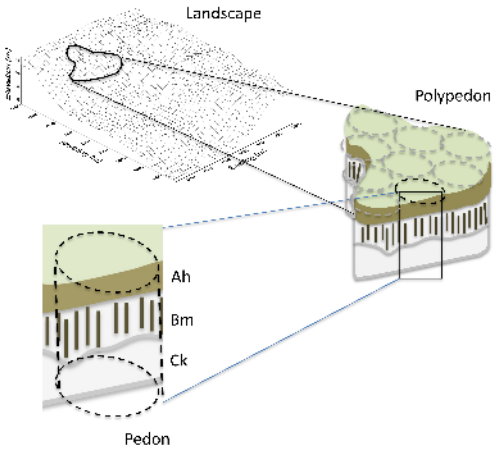


Figure 17.1. Illustration of the concept of the soil pedon and soil polypedon. © Dan Pennock, Univ. of Saskatchewan is licensed under a CC BY (Attribution) license.

Producing Conventional Soil Maps

MacMillan et al. (1992) describes the steps involved in making a conventional soil map, beginning with the preliminary steps of defining the objectives, compiling background information, and initial legend development. The preliminary (primarily office based) steps are followed by soil inspections, field mapping, and correlation. Finally, the map is produced, and a report written. The procedures used to produce soil maps were documented extensively during the period when these maps were being produced in

many countries throughout the world. In Canada, the procedures and considerations for meeting soil mapping objectives were documented in Mapping Systems Working Group (1981) and Coen (1987).

In a conventional survey, one key aspect of planning is to establish the survey intensity level, which expresses the amount of (field) work done, and the level of detail for the information collected. Guidelines for describing the survey intensity level are provided in Table 17.1 (Mapping Systems Working Group 1981; Coen, 1987).

Table 17.1. Soil survey intensity level guidelines adapted from Mapping Systems Working Group (1981)

Survey Intensity Level (SIL)	Common Name	Field Intensity	Method of Field Checking	Typical Publication Scale	Taxonomic Level
1	Very detailed	1 inspection per polygon	Foot traverse <0.5 km apart	1:5,000	Series
2	Detailed	1 inspection in >90% of polygons	Foot/vehicle traverse 2 km apart	1:20,000	Series or family
3	Reconnaissance	1 inspection in 60-80% of polygons	Foot/vehicle traverse 4 km apart	1:50,000	Series, family or subgroup
4	Broad Reconnaissance	1 inspection in 30-60% of polygons	Vehicle traverse 8 km apart/helicopter	1:100,000	Family or subgroup
5	Exploratory	1 inspection in <30% of polygons	Vehicle traverse 10 km apart/helicopter	1:250,000	Subgroup, great group or order

In Table 17.1, a conceptual diagram of the soil survey process is presented, where environmental data and field inspections are incorporated into soil landscape models and used to prepare a conventional soil map. The soil landscape model is a key feature of the conventional soil survey: these were mental models, which allowed surveyors to extend the information from the field inspections in one area to other areas with similar environmental characteristics, greatly improving map production at a time when access networks were more limited than they are today.

Aerial photographs, when they became widely available, were an important part of the process. In a conventional soil survey, the mapper first reviewed existing information and knowledge of soil-environmental relationships for the area. Aerial photographs were then reviewed in order to identify topographic and vegetation patterns, where the soil-environmental variables exhibited an external expression on the landscape in order to attempt the correlation of landscape characteristics to soil boundaries (soil-landscape relationships). Aerial photos became widely used during and after the 1960s, and air photo interpretation was carried out using a stereoscope, which allows the mapper to see the landscape in 3D—significantly increasing the quality of the soil maps being produced. Again, by referring to Jenny's *clorpt* model (1941) the underlying hypothesis was that areas with similar soil-environmental characteristics should have similar soils. A preliminary reconnaissance map was developed by drawing boundaries between soil units, and the preliminary map was then tested in the field where morphological classification was performed on the preliminary map units and linked to a soil taxonomic unit (e.g., series). When the series were identified, the mapper attempted to further delineate or adjust the boundaries of the map unit based on where the rate of change in soil properties was the greatest and to enclose relatively uniform areas within the map units.

Supplemented with field data and profile descriptions, map units with similar morphological characteristics are grouped into the same taxonomic unit (e.g., series) and the soil properties and the range of environmental conditions, from which the soils were found, were then described in the soil legend. Conventional soil surveys were produced at varying survey intensities, which reflects the amount of detail that is shown on a map and at its corresponding map scale (Table 17.1, Figure 17.2). Large parts of Canada were mapped this way, providing essential information for the development of agriculture and natural resources.

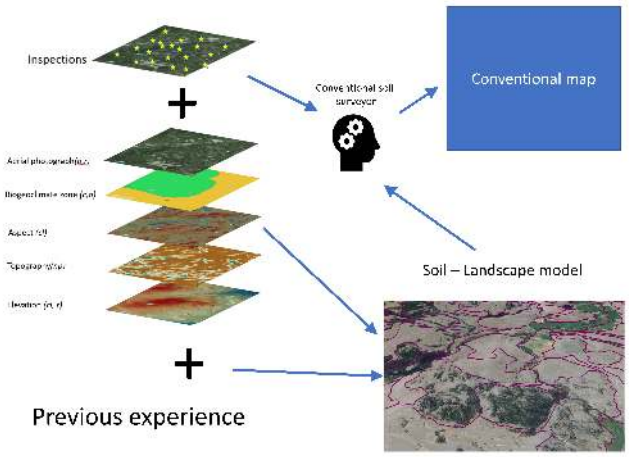


Figure 17.2. Conceptual framework for developing conventional soil maps using local information and expert knowledge by a soil surveyor to derive soil landscape models, with the use of aerial photographs to assign soils to landscape positions. © Chuck Bulmer, BC Ministry of Forests, Lands and Natural Resource Operations is licensed under a CC BY (Attribution) license.

The Evolution of Conventional Soil Maps

Because most of the conventional soil maps that were produced in Canada up to the 1990's were prepared for land inventory and regional planning purposes as 'semi detailed' maps, and because they were produced using analogue cartographic techniques (i.e., producing a hard copy map), they have certain limitations when compared to the current need for high resolution products. The chloropleth map representation used to present soil information on paper maps, and the lack of technology available at the time to acquire and process digital gridded soil and terrain data, are two key aspects of these limitations.

Firstly, the data in a chloropleth map is represented as discrete classes where the conditions are assumed to be homogenous within the map unit (Hole, 1978). Furthermore, it is also recognized that a significant amount of spatial generalization (i.e., simplification) within the map unit occurs due to inclusions of subdominant soils that are too small to be mapped at the map scale (Hole and Campbell, 1985). As a result, the purity of the mapping units is closely related to the complexity of the terrain, external expression of boundaries, survey effort, and mapping scale (Beckett, 1971). Although it would be ideal to have a soil map that consists of only simple mapping units, increasing the proportion of 'pure' map units within a map has been shown to result in an exponential increase in cost for developing the map (Bie et al., 1973).

Other issues may be related to the imprecision in map unit boundaries where the variability (or lack thereof) of the soil's surface does not necessarily coincide with the variability that may be occurring belowground (Hole, 1978). In addition, the changes in soil are not necessarily discrete (as suggested by the use of boundaries), but rather, they are fuzzy where the soil attributes between two neighbouring map units are an intergrade of the soil properties of the two units (Zhu and Band, 1994; Schaetzl and Anderson, 2005).

The final set of challenges stem from the delineation of map units based on the mental-models of soil-environmental relationships that were developed primarily to coincide with the objectives for a specific map product and area. These objectives varied from one map to another, and although they normally were described in the survey report, they are rarely suitable for incorporation into computer-based soil assessments spanning large areas with several smaller surveys. Consequently, inconsistencies manifested themselves in soil maps as mismatched boundaries of map units amongst different counties, states/provinces, and countries (Figure 17.3; Thompson et al., 2012; Dewitte et al., 2013). In addition, inconsistencies may also lead to problems such as having multiple soil series with the same soil properties, which results in redundancy, or even worse, where two soil series with the same name have completely different soil properties (Thompson et al., 2012).

Despite these limitations, in many parts of Canada and the world, legacy soil maps incorporate vast amounts of information derived from field inspections and soil surveyor knowledge. For these reasons, they represent an important source of training data for the development of digital soil maps to meet current needs.

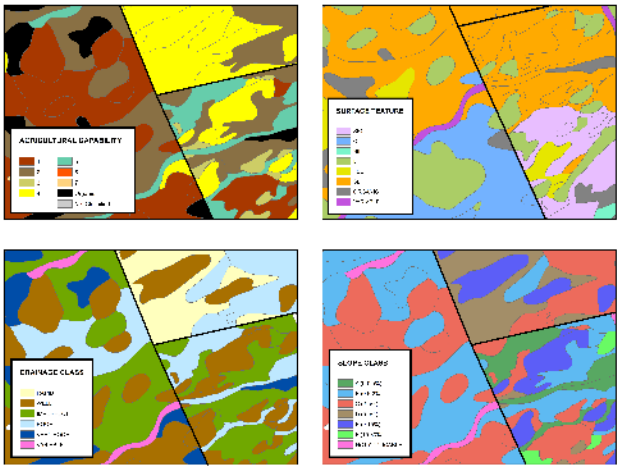


Figure 17.3. Mismatched soil survey boundaries at the intersection of Durham, Northumberland, and Peterborough counties in Ontario. © Daniel D. Saurette, Ontario Ministry of Agriculture, Food and Rural Affairs is licensed under a CC BY (Attribution) license.

DIGITAL SOIL MAPPING

Soil surveying, as a practice, has been evolving to take advantage of the advancements in computing technology (Minasny and McBratney, 2016; Rossiter, 2018), remote-sensing technologies (Mulder et al., 2011), proximal-sensing technologies (Viscarra Rossel et al., 2011), GIS, machine-learning techniques (Heung et al., 2016), and the increasing availability of geospatial datasets (McBratney et al., 2003; Minasny and McBratney, 2016; Scull et al., 2003). Hence, pedometrics—a branch of soil science that applies "mathematical and statistical methods for the study of the distribution and genesis of soils" (Webster, 1994)—has been an emerging field of research, globally. Although pedometric techniques for producing digital soil maps (DSM) have existed since as early as the 1970s (e.g., Webster and Burrough, 1972a, 1972b), technological advances have facilitated the production of DSM products since the 2000s (McBratney et al., 2003; Scull et al., 2003; Minasny and McBratney, 2016). Furthermore, these advances have also allowed for soil maps to be produced for progressively larger areas and in increasing levels of detail (Minasny and McBratney, 2016).

Some of the major achievements that have been made by pedometricians around the world have included, but are not limited to, the establishment of GlobalSoilMap.net, an international organization that aims to coordinate and produce global-scale digital soil maps; and the establishment of SoilGrids.org, an organization that was the first to develop a suite of global-scale map products. Within Canada, this is still an emerging and evolving area of research within the agricultural, forestry, and environmental sectors since 2010. With great success, the DSM community has been providing valuable information at multiple spatial scales to various stakeholders, including landowners, farmers, governments, and forest managers.

The steps for digital soil map production (after Kienast-Brown et al. 2017) are compared with those for conventional maps in Figure 17.4. The two methods of map development share several common steps, especially the need for detailed field inspections and an ability to correctly describe and interpret soil profile data. For digital soil maps, specifications for information content and level of detail are defined by the target spatial resolution, where the level of detail is inferred, and in the training data step, where the modeled attributes are defined. Some organizations have developed mapping specifications for digital soil map production (e.g., FAO and Intergovernmental Technical Panel on Soils, 2018).

Conventional and digital soil mapping differ primarily in the way that soil classes or attribute values are assigned to locations. In conventional soil mapping, a combination of field inspection, extrapolation and expert knowledge is used, while in digital soil mapping, field information and quantitative inference models are used to predict soil conditions at given locations. The conceptual framework for DSM, presented in Figure 17.5, highlights the role that machine learning algorithms play in the production of a digital soil map. Of critical importance is the need for field inspection data in both conventional and digital soil mapping approaches—DSM does not remove the need for trained pedologists and soil surveyors—the primary difference is in how the field data is used to construct the soil maps.

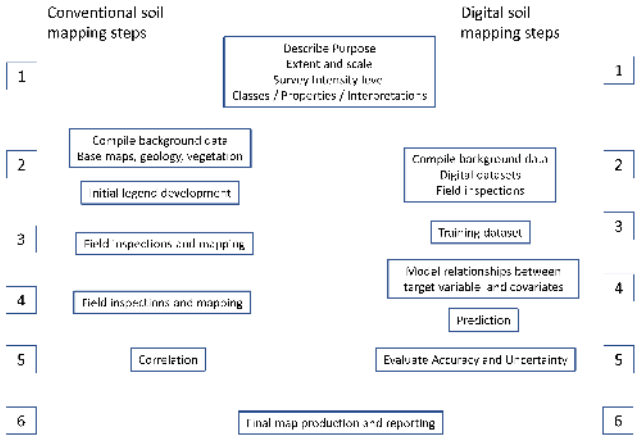


Figure 17.4. Steps in production of conventional and digital soil maps. Adapted from MacMillan et al. (1992) and Kienast-Brown et al. (2017). © Adapted from MacMillan et al. (1992) and Kienast-Brown et al. (2017). adapted by Chuck Bulmer, BC Ministry of Forests, Lands and Natural Resource Operations is licensed under a CC BY (Attribution) license.

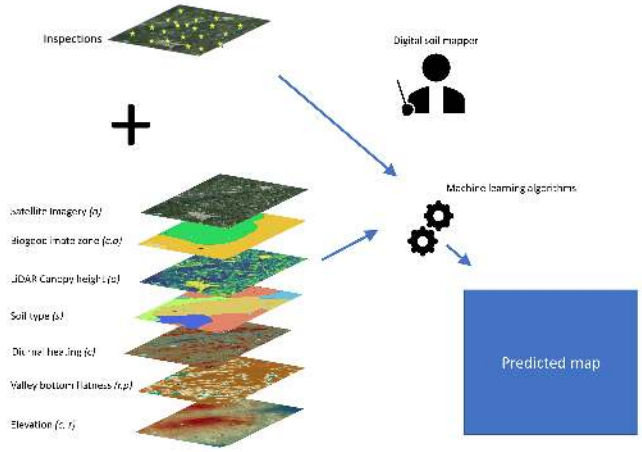


Figure 17.5. Conceptual process of creating a digital soil map, where the development and use of the soil landscape model are no longer held exclusively as mental models of the soil surveyor, and the digital soil mapper must assume the role of a conductor, guiding the process of supplying field inspection and environmental data (i.e., the training data) to algorithms that then assign soil classes and attributes to landscape positions. © Chuck Bulmer, BC Ministry of Forests, Lands and Natural Resource Operations is licensed under a CC BY (Attribution) license.

It is important to recognize that whereas conventional soil maps represent soil map units using polygons, DSM aims to produce maps using a raster representation. Raster data consist of a grid of two-dimensional cells (i.e. pixels), whereby each cell has a geographical location (e.g., longitude and latitude) and a corresponding value of a variable. Figure 17.6 shows a comparison of a conventional soil map that is represented in the polygon format, and a digital soil map that is represented in the raster format. Within DSM, the raster data may represent soil attributes (e.g., soil pH), a soil class (e.g., drainage class), or one of the environmental factors that are used to make spatial predictions of soils.

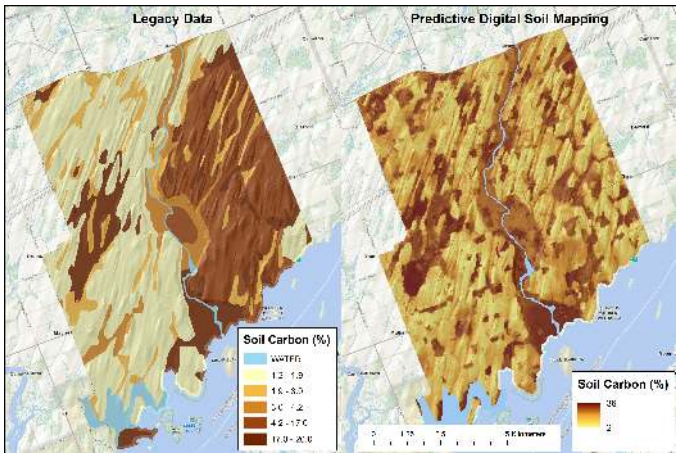


Figure 17.6. Comparison of a soil organic carbon (%) map digitized from a conventional soil survey map (left) and predictive digital soil map at a 20 m spatial resolution for the Keene area, Peterborough County, Ontario. © Daniel D. Saurette, Ontario Ministry of Agriculture, Food and Rural Affairs is licensed under a CC BY (Attribution) license.

Critical to the raster data representation is the concept of *spatial resolution*; i.e., the dimension of each cell with respect to the area that it is representing on the ground. For example, a spatial resolution of 10 m will indicate that each grid cell or pixel will represent a 10 × 10 m area on the ground. Ultimately, the spatial resolution will determine the level of detail or precision of a soil map and its uses. For example, a fine- or high-resolution map (e.g., 5–10 m spatial resolution) may be more useful for representing soil variability of individual agricultural fields for precision agricultural purposes whereas a coarse- or low-resolution map (e.g., 250–1,000 m spatial resolution) may be more practical for representing global-scale soil variability and the integration of that data into global climate models. Therefore, when selecting an appropriate spatial resolution, a soil mapper needs to consider what that map is used for and the spatial resolution of the input data that is required to generate those maps.

Of course, there is also a trade-off between spatial resolution and data size whereby higher resolution data are larger in size than lower resolution data. For example, when the spatial resolution is increased by a factor of two (e.g., decreasing the cell size from 10 × 10 m to 5 × 5 m), the size of the dataset could be increased by a factor of four because four pixels are now required to represent the same area as the original pixel. Figure 17.7 shows the relationship between spatial resolution and detail for topographic data.



Figure 17.7. Digital elevation model produced at 1 m, 15 m, and 25 m spatial resolutions. Original data was acquired from the Nova Scotia provincial LiDAR dataset. © Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.

The *scorpan* Model

Despite the wide use of Jenny's (1941) *clorpt* model, it is still largely a conceptual model; however, if we refer to his quote, we should recall that the *clorpt* factors are variables that describe the environment. To apply digital approaches, we need to recognize that these *clorpt* factors may also be represented using GIS—a computerized system that is able to acquire, store, analyse, and visualize spatial data. To fully make the theoretical transition from conventional to digital soil mapping approaches, McBratney et al. (2003) proposed the *scorpan* model as an extension of the *clorpt* model in the following:

(2) $S = f\,(s,\ c,\ o,\ r,\ p,\ a,\ n)$

The *scorpan* model shares the same variables as the *clorpt* model, which includes climate (*c*), organisms (*o*), relief (*r*), parent material (*p*), and time/age (*a*). The additional variables include *s*, which represents the intrinsic properties of the soil (e.g., the spectral properties of the soil) that may be captured by various remote and proximal sensors; and *n*, which represents the spatial

coordinates of a sample or the location relative to another geographical phenomenon (e.g., distance to river). Here, individual *scorpan* factors, or a combination of them, are used to predict the distribution of a soil property or soil class, *S*, of interest using a quantitative function, *f* (), that represents the soil-environmental relationship.

Representing the *scorpan* Factors

The effectiveness of the *scorpan* model is in its flexibility to integrate data from existing soil surveys and geospatial datasets acquired from multiple sources, such as remote sensing data, digital elevation data, climate data, land use data, and geological data (McKenzie and Ryan, 1999). These *scorpan* factors are represented in the raster format and therefore, each dataset also has a corresponding spatial resolution and projection system. Hence, each individual dataset may need to be re-projected into a common projection system and scaled to a uniform spatial resolution so that all the individual cells of the various datasets are spatially aligned (Figure 17.8).

The following provides a brief overview of the *scorpan* factors.

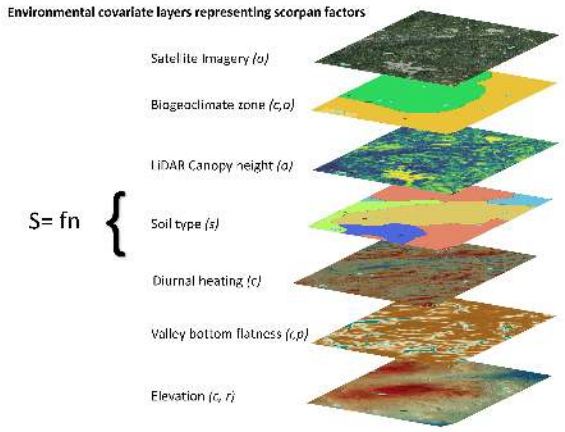


Figure 17.8. Representation of scorpan factors within a raster format. © Chuck Bulmer, BC Ministry of Forests, Lands and Natural Resource Operations is licensed under a CC BY (Attribution) license.

Soil (s)

The s factor is the second most frequently used factor in DSM studies (McBratney et al., 2003). Again, the s factor is founded on the concept that soils may be used to predict soils and therefore there are multiple sources from which this data may be acquired from: conventional soil survey data, proximal soil sensors, and remote sensors.

Conventional Soil Survey: Conventional soil maps represent a wealth of information derived from many years of field work by previous generations of soil scientists, and therefore they are extremely valuable sources of information that can be used for prediction purposes due to the relationship that different soil variables have with each other. For example, Paul et al. (2020a; 2020b) used the Soils of the Langley-Vancouver Map Area (Luttmerding, 1980) to generate digital layers of sand, silt, and clay percentages; and cation exchange capacity that were then used as predictors for mapping soil organic carbon and soil workability for the Lower Fraser Valley, BC. In both studies, predictors generated from the original conventional soil survey were identified as the most important layers. Although soil surveying approaches have transitioned to digital techniques, conventional soil maps should not be disregarded because they remain a valuable and complementary source of information within a DSM framework.

Remote Sensors: Remote sensors are designed to take measurements of the Earth's surface without any physical contact and are therefore mounted on an aircraft or a satellite. Here, it is important to recognize that data acquired from remote sensing could be used to represent multiple *scorpan* factors. Within the suite of remote sensors, spectral sensors (e.g., multispectral, hyperspectral) that detect reflected or emitted energy from the electromagnetic spectrum are most used in DSM (McBratney et al., 2003; Mulder et al., 2011). For example, Landsat 8 provides measurements for visible (e.g., red, green, blue wavelengths) and infrared (e.g., near infrared, short wavelength, and long wavelength) light. Alternatively, various spectral sensors may also be mounted on an aircraft or a remotely piloted aerial system (RPAS) if higher spatial resolution data is required (Figure 17.9).

There are several issues related to the use of spectral sensors for DSM purposes, especially given that soil measurements are often obscured by vegetation. Furthermore, atmospheric effects and topographic distortions may cause anomalous measurements. In

addition, when using spectral sensors, one needs to consider the day and time of which the imagery is taken. Lastly, remote sensors generally only acquire data that are representative of surface soils (5–6 cm; Adamchuk et al., 2017). Despite these issues however, under bare-soil conditions, measurements of the spectral properties of the soil have shown to correspond well with soil properties such as soil mineralogy, texture, soil organic matter, soil moisture, and other soil properties (Mulder et al., 2011).



Figure 17.9. Remotely piloted aerial system (Sense-Fly eBee) that is equipped with a multispectral sensor (Sequoia Red-Edge). Photo credit: Ian Manning, Nova Scotia Community College – Centre of Geographic Sciences Applied Research. © Ian Manning. CC BY.

Proximal Soil Sensors: Whereas remote sensors are aerial- or space-based, proximal sensors are a suite of ground-based sensors that are designed to measure properties of soils that may be correlated to other soil properties. Given that measurements are taken on the ground, proximal soil sensors can characterize soil variability at a higher spatial resolution than remotely sensed data; and as a result, these sensors are more conducive for mapping soils at the field-scale. Figure 17.10 shows an unmanned ground vehicle that has been equipped with a gamma radiometric sensor system.

Amongst proximal soil sensors, electro-magnetic induction (EMI) sensors have been the workhorses for DSM and precision agriculture research (Doolittle and Brevik, 2014). The EMI sensor measures the apparent electrical conductivity (EC_a) of the soil at multiple depth intervals. Combined with a global positioning system (GPS), EMI survey data are often used as a predictor of soil properties such as soil salinity, clay content, and water content; however, secondary soil attributes, such as soil bulk density and soil organic carbon may also be derived from EC_a measurements. The EMI sensors are all but one type of proximal sensor—others may include gamma radiometric sensors, ground penetrating radar, and electrical resistivity sensors (amongst many). Interested readers are encouraged to refer to Viscarra Rossel et al. (2011) and Adamchuk et al. (2017) for a detailed overview of these systems.



Figure 17.10. Unmanned ground vehicle (Korechi) that is equipped with a gamma radiometric sensor (Medusa M1000). The vehicle has autonomous capability for navigation and can carry a variety of sensors. The gamma radiometric sensor pictured here collects data that can be related to soil physical and chemical properties, which can be mapped using information from the vehicle's location. © Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.

Climate (c)

This factor is one of the least used *scorpan* factors (McBratney et al., 2003)—several possible explanations comes to mind. Firstly, the local climate is largely influenced by topography and as a result, topographic indices such as elevation and aspect may be used as a proxy for a climate variable due to the relationship between elevation and the environmental lapse-rate and the relationship between slope-face direction and temperature (Schaetzl and Anderson, 2005). However, as the extent of the study area increases to national- and global-scales, large-scale climatic patterns have been shown to be an important control on various soil properties—this was evident in the global *SoilGrids250m* product (Hengl et al., 2017). In terms of climate layers, common variables include mean annual temperature, mean annual precipitation, and evapotranspiration (McBratney et al., 2003). These datasets may be derived from sensors mounted on satellites; however, climate model data, often derived from weather stations, (e.g., WorldClim, ClimateNA) may also be used (Figure 17.11). In addition to using current climate conditions as inputs into a model, historical and projected future conditions could also be used to simulate the effects of changing climate conditions with respect to understanding the spatial and temporal patterns of soil properties.

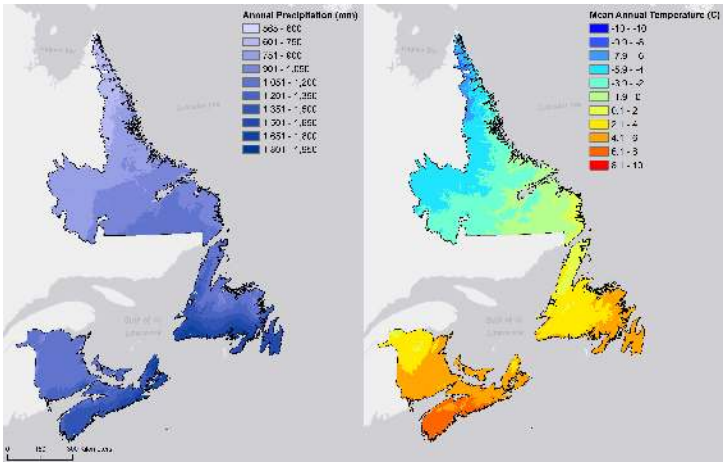


Figure 17.11. Maps of mean annual precipitation and mean annual temperature for the Atlantic Provinces. Data set from the WorldClim version 2.0 at a 1 km spatial resolution (https://www.worldclim.org/). © Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.

Organisms (o)

A major source of vegetation data used in DSM may be derived from satellite imagery where numerous vegetative indices have been developed based on satellite band-ratios (Mulder et al., 2011). Perhaps the most commonly used variable is the normalized

difference vegetation index (NDVI), which provides a measure of vegetative greenness as a function of the near-infrared and red wavelengths from multispectral imagery. The NDVI and other remotely sensed data such as thermal imagery have also been shown to be an effective predictor for soil moisture, soil colour, soil texture, and water holding capacity; and it is also effective in evaluating plant growth (Figure 17.12; Mulder et al., 2011). In addition to NDVI, similar indices may include the Soil Adjusted Vegetation Index (SAVI), Transformed SAVI (TSAVI), Modified SAVI (MSAVI) and the Global Environment Monitoring Index (GEMI). Studies such as Paul et al. (2020) have found utility of these variables for mapping soil organic carbon and clay in the Lower Fraser Valley, BC; while Heung et al., (2017) used similar variables for mapping soil great groups for the Okanagan-Kamloops Region, BC. Similar indices may also be calculated using imagery that are acquired from a RPAS that is equipped with a multispectral or hyperspectral sensor. Although these indices have been extensively used in DSM, the data will be affected by the season and year from which the image was taken. If a soil mapper is requiring information on crop cover, imagery taken throughout the growing season would be more useful than post-harvest imagery.

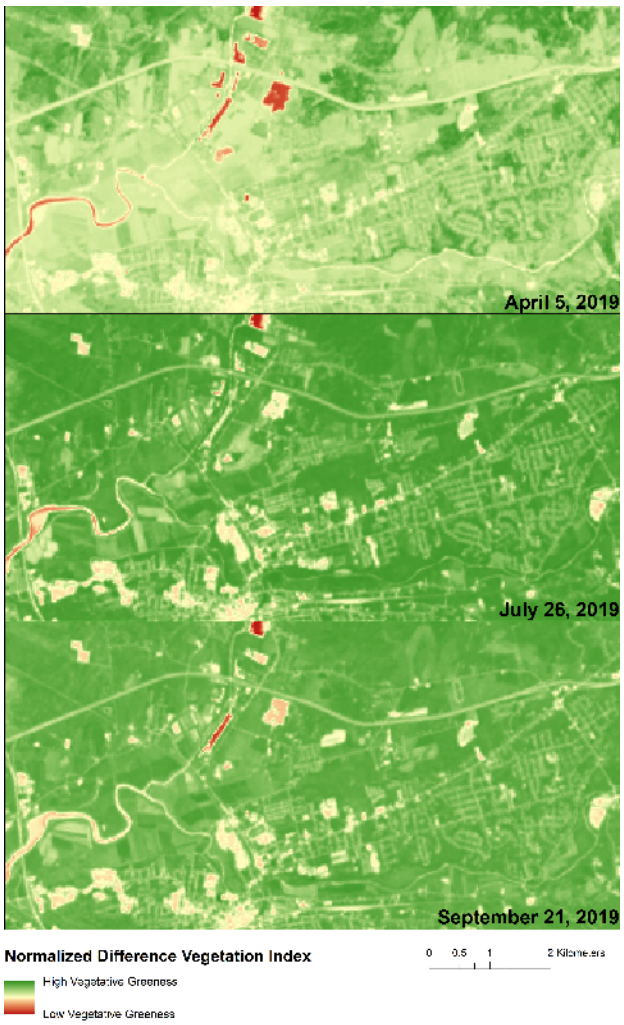


Figure 17.12. Normalized difference vegetation index maps produced from Landsat8 imagery for the Truro Region, Nova Scotia at a 30 m spatial resolution. © Siddhartho Paul, Post Doc with Heung, B. is licensed under a CC BY (Attribution) license.

In addition to using raw satellite imagery, researchers around the world have already processed vast quantities of imagery to generate landcover maps. For example, the North American Land Change Monitoring System provides landcover maps for 2005, 2010, and 2015 at a 30 m spatial resolution; and for a Canadian example, Agriculture and Agri-Food Canada have developed the Annual Crop Inventory data, which has been providing national-extent crop information since 2009 (Figure 17.13). However, it is necessary to recognize that these datasets are generated using predictive models; hence, it is important for a soil mapper to refer to the supporting documentation for these products and note their accuracy.

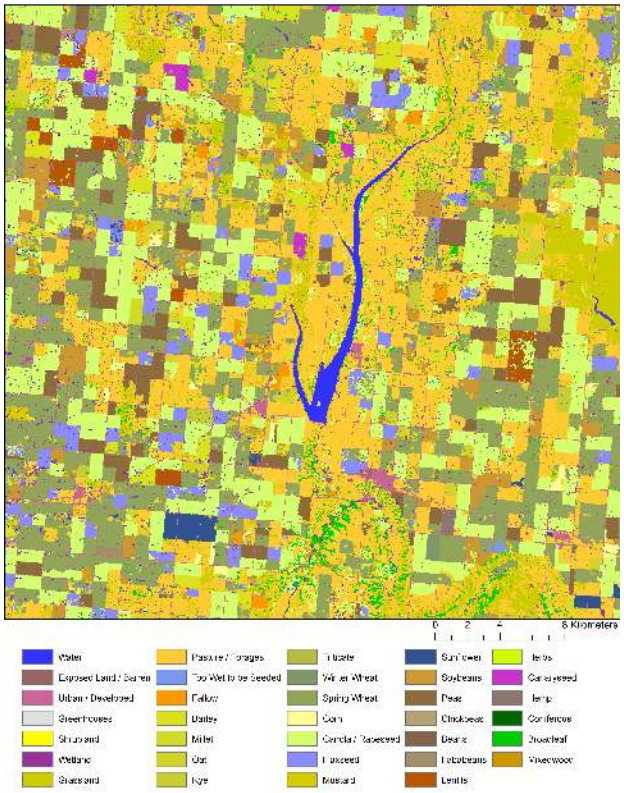


Figure 17.13. Annual Crop Inventory 2019 (Agriculture and Agri-Food Canada) for Moose Creek Region, Saskatchewan at a 30 m spatial resolution. © Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.

In other cases, crop data has also been used as a covariate for spatial prediction; for instance, crop yields are the result of the interaction between soils, plants, and atmosphere. Therefore, crop yield data may be used as an indicator for soil properties since plant growth is influenced by properties such as clay content, moisture content, and nutrient content for example (Shatar and McBratney, 1999; McBratney et al., 2000). In a forested setting, a possible opportunity may lie in the use of forest inventory data where forest variables such as basal area, gross total volume, stand density, stand height, and aboveground biomass might provide some insight into the soil properties. Especially with the developments in data acquired using light detection and ranging (LiDAR) technologies, forest inventory data might become more prevalent as a predictor (Woods et al., 2011; Treitz et al., 2012). Figure 17.14 shows a digital surface model that captures the variability in crop and tree height using LiDAR.

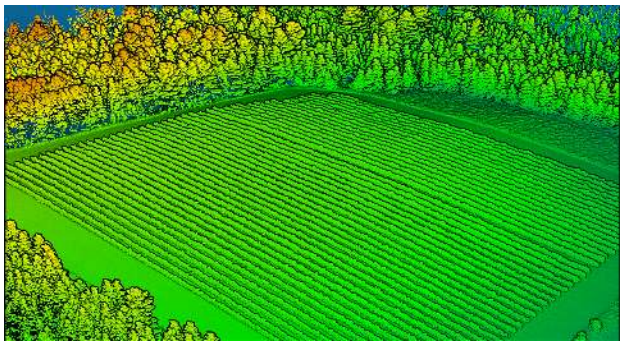


Figure 17.14. RPAS-acquired, LiDAR surface model for a vineyard in Middleton, NS. Image provided by Ian Manning, Nova Scotia Community College – Centre of Geographic Sciences Applied Research. © Ian Manning, Nova Scotia Community College – Centre of Geographic Sciences Applied Research is licensed under a CC BY (Attribution) license.

Relief (r)

It is well recognized that the r factor is the most used factor within DSM (McBratney et al., 2003). One of the staples in DSM research has been the use of a digital elevation model (DEM)—a raster that is made up of elevation values. DEMs may be

produced from multiple sources: digitized contour maps; interpolated from ground measurements; and acquired from remote sensing using satellite- or RPAS-mounted sensors.

DEMs are particularly useful because they are readily available and typically free to access; for example, there are various provincial and federal DEM product(s) that are accessible via online web portals, and in parts of the country where there is no DEM data available, global DEM products such as Shuttle Radar Topography Mission (SRTM) or Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) may also be accessed. As a general recommendation, prior to using any DEM, it is always important for a soil mapper to visually assess the DEM by generating a hillshade (3D-like) representation of the topographic surface (Figure 17.15). This will help the mapper evaluate the abundance and types of DEM anomalies, due to the quality of the raw data, and the methods used to generate the DEM; and also whether preprocessing techniques (e.g., smoothing, pit filling, or road removal) should be applied.

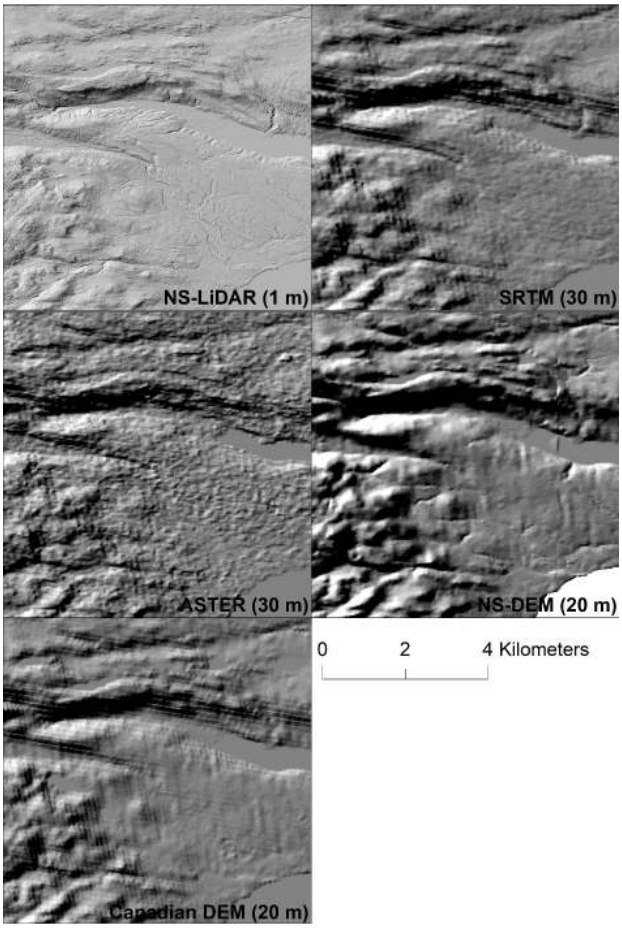


Figure 17.15. Comparison of hillshade models derived from Nova Scotia's LiDAR data; Shuttle Radar Topographic Mission (SRTM) data; Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) data; Nova Scotia's Enhanced DEM data; and the Canadian DEM data for a region west of Ingonish, Nova Scotia. © Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.

The value of a DEM comes from its flexibility in calculating a large suite of topographic variables. For example, a DEM can be used to characterize local-scale morphometry (e.g., slope, aspect, curvature), landscape-scale morphometry (e.g., relative slope position), and hydrological patterns (e.g., topographic wetness index) – all of which are influential on soil patterns (Figure 17.16). In fact, the analysis of land surface using various mathematical and statistical techniques, and the development of new topographic variables, are all part of the scientific discipline known as *Geomorphometry*.

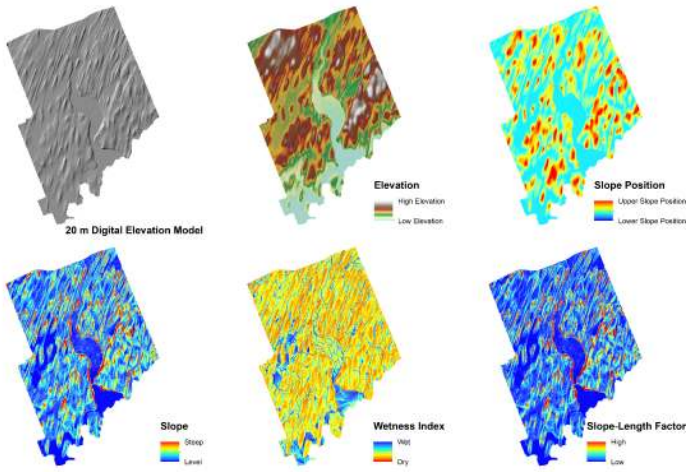


Figure 17.16. Topographic variables derived from a 20 m digital elevation model for the Keene area, Ontario, at a 20 m spatial resolution. © Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.

Within Canada, high quality DEMs have been used as the basis for mapping soil parent materials (Heung et al., 2014), soil types (Heung et al., 2016), and soil thickness (Scarpone et al., 2016) throughout BC; and mapping soil depth and textural classes throughout ON (Akumu et al., 2015; Akumu et al., 2016). In fact, those studies generally used only DEM-derived data.

Furthermore, the topographic variables may also be used to delineate landform features (Figure 17.17). In Pennock et al. (1987), combinations of planform (change in aspect) and profile curvature (change in slope) were classified into a series of seven landform elements (i.e., divergent/convergent shoulders; divergent/convergent backslopes; divergent/convergent footslopes; and level surfaces) that characterized the topographic controls on the flow of sediment and water. As a result, similar landscape classification schemes could be applied using a DEM (e.g., MacMillan et al., 2000, 2004) in a semi-automated approach.

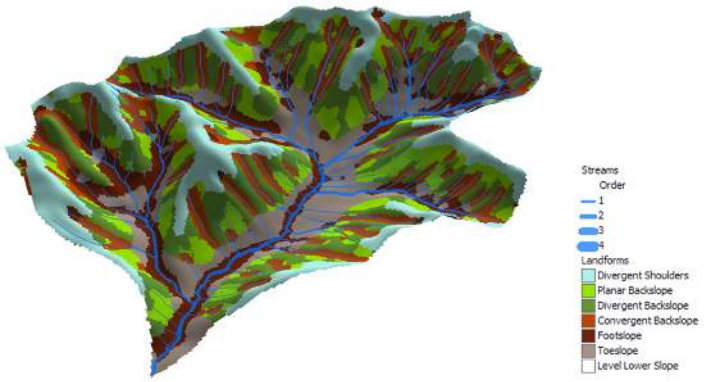


Figure 17.17. 3-dimensional representation of a landform classification map produced using a 25 m spatial resolution DEM for the Twin Sisters watershed, British Columbia. © Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.

Parent Material (p)

Information on soil parent materials have typically been acquired from digitized geological maps; hence, they share similar spatial problems to conventional soil survey maps whereby there may be a combination of complex map units; the amount of detail is controlled by the mapping scale; and detailed maps may be limited in coverage. As a result, soil parent material is an attribute of the soil that has previously been mapped using predictive approaches; for example, Heung et al. (2014) produced a parent material map for the Lower Fraser Valley, BC.

When using digitized geological maps, it is important to recognize the source of material from which the soil is derived (Figure 17.18). For example, a bedrock geology map would likely find greater use in non-glaciated environments where the soil is largely derived from weathered bedrock. However, Canadian landscapes are predominantly glaciated and therefore the overlaying sediment may have contrasting properties to the bedrock and hence, a surficial geology map would be more useful as it characterizes the material that has been transported and deposited on the bedrock. These mapping efforts have highlighted the close

associate between topography and parent material, particularly in glaciated environments, so combinations of topography derivatives are often used in DSM as a surrogate for parent material.

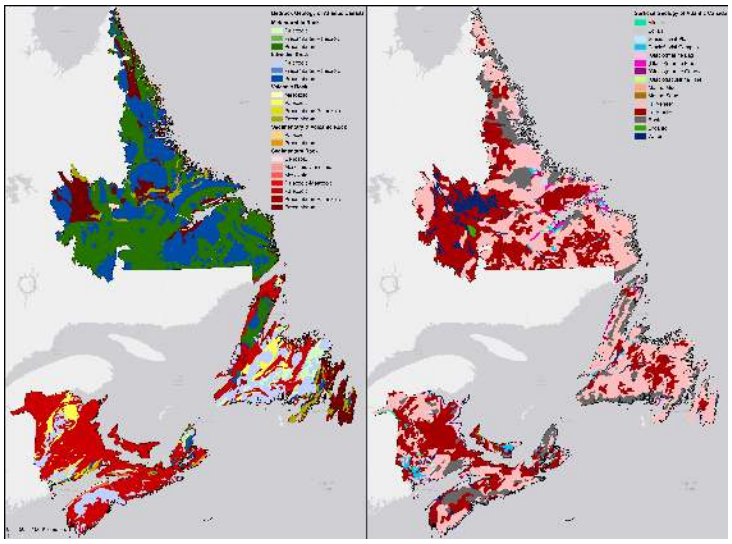


Figure 17.18. Maps of bedrock geology (left) and surficial geology (right) for the Atlantic provinces. The bedrock geology data are from the Geological Map of Canada (Wheeler et al., 1997) and the surficial geology map is derived from the Surficial Materials Map of Canada data set (Fulton, 1995). © Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.

Age (a)

The age factor may be used to describe how long pedogenesis has occurred and may be estimated as the age of the ground or material from which the soil has developed (McBratney et al., 2003). Within DSM, there are very few examples where this information has been used due to the difficulty in characterizing the age of soil within a format that would be conducive for use within a GIS. One potential option for incorporating the a factor into a DSM framework would be by incorporating information on how humans have modified the landscape and thereby influencing soil attributes and types.

Spatial Position (n)

The *n* factor may be incorporated in several ways: using the spatial coordinates of soil sample locations or using a raster layer that represents the distance to some geographical phenomenon.

Spatial Coordinates of Soil Samples: Predicting soil attributes may be carried out by using only the spatial coordinates of the soil samples themselves. Here, we should introduce Waldo Tobler's First Law of Geography (Tobler, 1970), which states:

"Everything is related to everything else, but near things are more related than distant things."

When applied to DSM, we can therefore say that two soil samples that are in close proximity are more likely to share similar soil properties than two soil samples that are located far from each other. As a result, the relationship between the sampling location, its corresponding soil value, and its distance to neighboring sampling locations can be used to predict (i.e., interpolate) soil values between those locations using geostatistical approaches. Although an overview of geostatistics is far beyond the scope of this textbook, students that have taken a spatial analysis class would be familiar with interpolators such as inverse distance weighting and kriging approaches.

Within Canada, geostatistical approaches have been used to characterize soil variability as early as the 1980's. In Raymond, Alberta, Chang et al. (1988) established a grid of 64 sampling points over a 20 × 25 m agricultural field to map sand content and soil salinity using kriging to facilitate irrigation practices on saline soils. Within a forested system located in southwestern BC, Chandler et al. (2008) applied a similar approach using gridded sampling and kriging to evaluate the localized, spatial patterns of forest floor nutrients as influenced by a bigleaf maple tree within a conifer-dominant stand.

Distance-Based Raster: Information on spatial position may also be incorporated by calculating the distance or proximity of each pixel within the study area to certain geographical phenomenon, or reference point, to capture contextual information about the landscape. For example, using a stream network, a distance-to-nearest-stream layer may be calculated; furthermore, similar

distance-based layers may be calculated to represent the proximity to the ocean, lakes, rivers, geomorphic features, and many other features. Studies such as Heung et al. (2014) have found that when mapping soil parent materials for the Lower Fraser Valley, BC, layers representing the distance to the nearest stream and to the Fraser River were important for predicting the distribution of fluvial materials.

Soil Sampling

The foundation of soil survey rests upon selection of representative sampling sites. With fewer resources available to complete large field programs and recent advances in computational tools, new techniques to optimize sample site selection are being used to improve accuracy of predictive models while gaining efficiencies in field sampling. Soil sampling in DSM seeks to gather information to understand the development and distribution of soils as expressed by soil forming factors (Jenny, 1941), or the *scorpan* model variables (McBratney et al., 2003). At its most basic, the sampling design seeks to sample representatively from the geographic area of interest and all the soil forming factors (i.e., covariates); the underlying assumption is that the spatial variability of soil properties being predicted can be explained by the environmental covariates.

There are many approaches to sampling design, including statistical and geometric approaches that seek to select samples randomly from all possible sampling locations (population) or distribute sample locations in geographic space based on spatial position or coordinates, and approaches that use ancillary data to select sample locations, commonly referred to as feature space approaches.

At the heart of every sampling program is the question of what attributes to measure at each field location. This question is closely related to a problem that was common for conventional soil mapping, namely the development of the map legend. Depending on the level of detail desired in the final product, and available survey resources, the need for agglomeration and subdivision of soil classes, the depth requirements for soil property evaluation, and the specific methods for field sampling and analysis all affect the final map, and the types of information in the training dataset. Ultimately, these considerations must support the original objectives set out for the DSM project.

The most commonly used geometric approaches include grid sampling (GS), simple random sampling (SRS), stratified random sampling (StRS), transect sampling (TS), cluster sampling (CS) and nested sampling (NS). Grid sampling divides the study area into an evenly-spaced regular grid with sample locations at the centre of each grid; the grid size is specified by the user based on some criteria, usually time, budget or prior knowledge of the required sampling density. Simple random sampling creates sampling sites that are selected completely at random (Brus et al., 2011) from the available study area with equal probability of being selected (Biswas and Zhang, 2018). Stratified random sampling is similar to SRS with the exception that the area is subdivided into smaller blocks, called strata, and SRS is applied to the strata (Brus and de Gruijter, 1997). The size of the strata can be used to weight the sampling, resulting in proportional sampling of the strata (Pennock and Yates, 2007). Transect sampling is a form of cluster sampling where samples are selected at equal distance intervals along a line (de Gruijter and Marsman, 1985). In cluster sampling, sample locations are grouped closely, a technique used to enable sampling in rough terrain with inaccessible areas without compromising accuracy (Biswas and Zhang, 2018). Nested sampling is typically applied in combination with other sampling designs and quantifies the variability of the data along different distances, a concept originating from geostatistics. Examples of sample designs are provided in Figure 17.19.

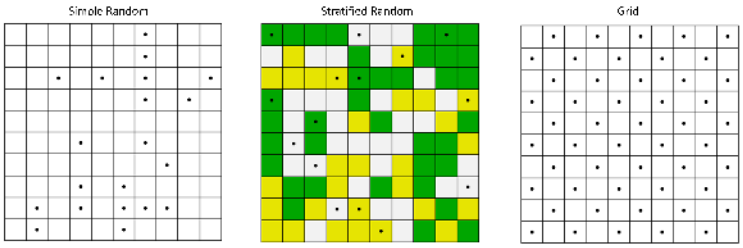


Figure 17.19. Examples of three sampling designs. Small black dots represent sampled raster cells. Shown are simple random sampling (left), stratified random sampling (middle) where colors (green, yellow and white) indicate a theoretical stratification in the data into three levels, and chessboard-style grid sampling (right). © Daniel D. Saurette, Ontario Ministry of Agriculture, Food and Rural Affairs is licensed under a CC BY (Attribution) license.

Feature space approaches consider the values of environmental covariates at the sampling locations. Fuzzy k-means sampling (FKMS) uses the k-means clustering algorithm to minimize the distance between sampling locations, but this distance is in covariate space, not geographic space (Brus, 2019). The conditioned Latin hypercube sampling (cLHS) was proposed by (Minasny

and McBratney, 2006) as a modification to Latin hypercube sampling (Mckay et al., 1979). With many covariates, a full factorial experiment is unfeasible—the Latin hypercube sampling allows sampling of all covariates, with one sample per strata (Brus, 2019).

An optimal sampling design should provide adequate coverage of both geographic and feature space. This can be assessed using various tools. For feature space coverage, typically a comparison of the feature space coverage of the sample sites and that of the entire study area can be completed to determine the adequacy of the sampling design. Tools exist to optimize sampling in terms of geographic space and in terms of feature space, however the fusion of these two sample optimization strategies remains an important area of research. One aspect lacking from existing sampling strategies is information regarding the optimal number of sites required to optimize the predictive ability of a model used for DSM. Interested readers are encouraged to refer to Brus et al. (2011) and Biswas and Zhang (2018) for a more detailed introduction to sampling design.

Sampling design and soil sampling present many technical and logistic challenges. During the sampling design phase, the soil mapper must consider features that are to be avoided during sample selection, especially given the use of computer algorithms, which cannot recognize these restrictions when selecting sample sites. For example, anthropogenic features such as roads, buildings, quarries, and natural features such as water bodies, are typically of little interest for soil mappers and should be removed from the layers of information used to design the sample plan. Similarly, in difficult terrain or remote areas, one might choose to restrict the sample selection process to a manageable buffer distance from access roads. Access to private lands for soil sampling can be challenging. Restricting the sampling design to public lands is in most cases impractical and not feasible, and as such permission from landowners must be acquired to complete soil sampling field work. A common issue that arises is the need for an alternate sampling site when access to private land is denied. Although beyond the scope of this introduction to DSM, researchers are developing strategies to modify and adapt sample design algorithms to account for these types of restrictions (Clifford et al., 2014; Malone et al., 2019).

Prediction of Soil Classes and Properties

Developing a soil map using DSM techniques largely requires the following three ingredients:

1. a suite of environmental layers (i.e., predictors or covariates) that represent the *scorpan* factors;
2. soil data that are geospatially referenced; and
3. a model that characterizes the relationship between the soil data and the environment in order to make predictions for unsampled locations.

In the previous sections, we have explored the different sources of environmental data and described the various spatial sampling approaches used to acquire the soil data. In this section we describe the process required to create a training dataset, which is then used to calibrate a predictive model to generate map outputs.

Typically, DSM techniques involve a form of *supervised learning* where a training dataset is first created by spatially intersecting the sample locations (with measured soil properties) with the underlying suite of environmental layers. Using a predictive model, the relationship between the soil response variable and the environmental predictor layers is established using a classification or regression function (i.e., model fitting). Once the predictive model is fitted with the training data, it is then applied to the set environmental layers to predict the soil properties or classes at unsampled locations.

Prior to making predictions, the soil mapper must recognize the type of soil data that is to be predicted. Soil data may take the form of *categorical data* or *continuous data*, whereby categorical data may be further subdivided into *nominal data* and *ordinal data*. Nominal data describe the qualitative aspects of soil and have no quantitative value and the most common example of nominal data would be the soil's taxonomic unit (e.g., soil order, great group, series) but may also include soil parent material class, soil textural class, and others. Ordinal data, another type of categorical data, represent values that have ordered units (i.e., values are relative to each other), where examples may include soil moisture and nutrient regime classes. With soil moisture regime, there are nine potential values ranging from 0 to 8, where a value of 0 represents 'very xeric' conditions while a value of 8 represents 'hydric' conditions; however, the exact difference between the values are not known. In contrast, continuous data represents measured soil properties such as soil depth, soil pH, clay content, soil organic carbon stock, and many more.

The distinction between categorical and continuous data types when representing soil information is critical because it largely determines which predictive model and methods for assessing model accuracy and uncertainty are appropriate. When predicting categorical soil data, we are limited to the use of predictive approaches that are suitable for classification purposes, whereas, the

prediction of continuous data requires the use of regression modelling approaches. An example of a digital soil map produced using a regression modelling approach is shown in Figure 17.6 (right) for soil organic carbon predictions and an example using a classification modelling approach is shown in Figure 17.20 for soil great group.

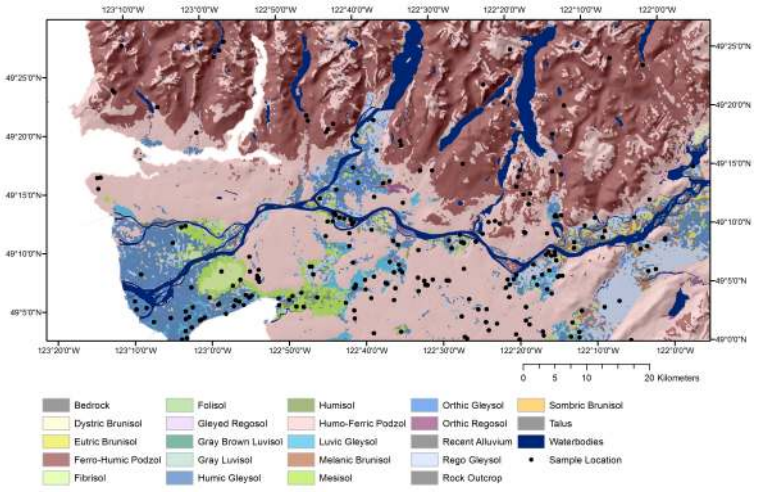


Figure 17.20. Soil great group map of the Lower Fraser Valley, British Columbia using support vector machine with radial basis function classifier at a 100 spatial resolution with a 72% overall accuracy. © Brandon Heung. CC BY.

A comprehensive overview of the modelling approaches is far beyond the scope of this textbook; however, interested readers should refer to McBratney et al. (2003), which provides an overview of various modelling approaches for regression and classification using machine-learning and geostatistical techniques. For a detailed overview of machine-learning techniques used for classification purposes in DSM, readers may refer to Heung et al. (2016).

It is important for readers to realize that there are many different types of predictive modelling techniques that are used in DSM, as well as in disciplines beyond DSM; however, there is not a single model that will consistently outperform another or is considered to be “best” in all situations. The choice in environmental predictor layers, the nature of the soil-environmental relationships (e.g., linear vs. non-linear relationships), the size of the training dataset, soil property, as well as a whole host of other factors, may affect a model’s performance. Furthermore, the intrinsic properties of the model, such as processing time, computational demand, and model complexity, will vary. As a result, it is encouraged that DSM practitioners compare a variety of modelling techniques and perform an accuracy assessment of all results; furthermore, a visual assessment of the mapping outputs should be carried out by a trained soil scientist to ensure that the soil maps are consistent with a pedological understanding of the landscape.

Given that much of the modelling process may be carried out in an automated way using statistical software such as *R*, the automation greatly facilitates model comparisons and the process of comparing models should be considered as ‘best practice’. For example, in Heung et al. (2016), identical input data were used to train 11 modelling techniques, where it was determined that despite the same inputs, the model outputs varied drastically when mapping soil great groups for the Lower Fraser Valley, BC (Figure 17.21).

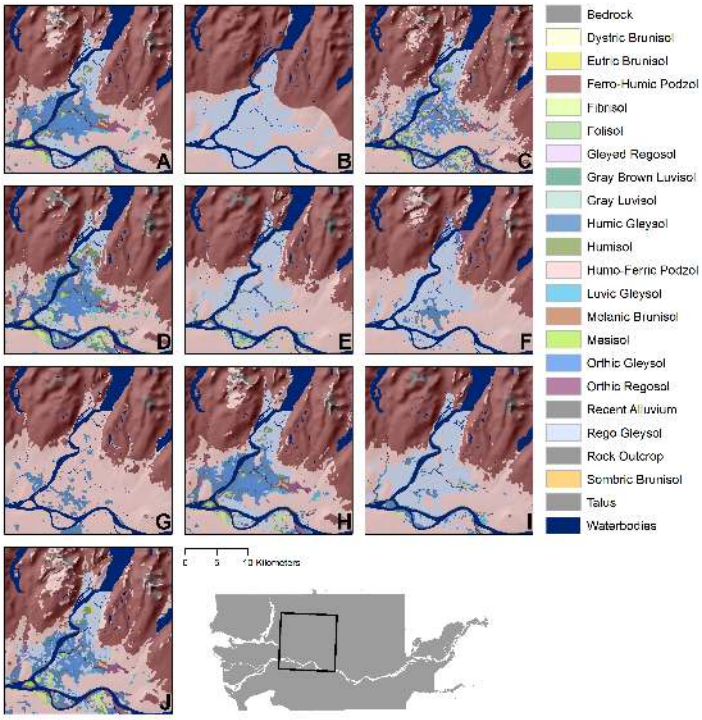


Figure 17.21. Close-up maps of soil great group predictions for the Lower Fraser Valley at a 100 m spatial resolution using a variety of classification algorithms. All predictions were generated using the same set of input data. Predictive models include (A) CART with bagging, (B) CART, (C) k-nearest neighbours, (D) logistic model tree, (E) multinomial logistic regression, (F) artificial neural network, (G) nearest shrunken centroid, (H) Random Forest, (I) linear support vector machine, and (J) support vector machine with radial basis function. © Brandon Heung. CC BY.

Accuracy Assessment

All maps approximate reality and all digital soil maps will deviate from the real world and whatever is generated from the model is only one out of an infinite number of realizations of a soil map. Again, this is demonstrated in Figure 17.21, which shows that different models could generate drastically different realizations of a soil map. Therefore, to quantify the quality of map predictions, an accuracy assessment needs to be carried out on all predictions. Here, we describe the term ‘accuracy’ as being the difference between the observed and predicted values at a location (Brus et al., 2011); whereby we can describe the accuracy as being ‘higher’ as the difference decreases.

Similar to selecting the appropriate model type to predict categorical and continuous soil variables, the choice of appropriate accuracy metrics also depends on this. However, in both cases, accuracy assessment must be carried out using an independent (i.e., validation or test) dataset that was not used to train the model.

When assessing the accuracy of categorical predictions (e.g., soil series, soil parent material type), we generally rely on *overall accuracy* and *Cohen’s kappa coefficient* as the main metrics. The overall accuracy is the proportion of observations that were correctly predicted for the independent dataset; whereas *kappa* accounts for the by-chance agreement between the observed and predicted classes. Both metrics range from 0 to 1, where a value of 1 represents high accuracy.

When assessing the accuracy of continuous predictions (e.g. bulk density, soil organic carbon), the mean square error (MSE), root mean square error (RMSE), coefficient of determination (R^2), and Lin’s concordance correlation coefficient (CCC) are often used. The MSE is effectively the average squared difference between the observed and predicted values; in comparison, the RMSE, is the square root of MSE and is expressed in the same units as the soil variable. In both cases, a lower MSE or RMSE represents a more accurate model. R^2 measures how closely the observed and predicted values follow along a best-fit regression line (i.e., how closely the values are correlated) and represents the proportion of soil variation that is explained by the model. It is critical to note that R^2 does not represent accuracy because it does not account for model bias whereby the model may consistently overpredict or underpredict a soil variable. Instead, the more appropriate metric is to use CCC, which represents the goodness of fit along a 45°

line on a scatterplot of observed and predicted values. Values of CCC range from 0 to 1, where values of 1 represent scenarios where the predicted values match the observed values and thus indicates high prediction accuracy and precision.

UNCERTAINTY IN DIGITAL SOIL MAPS

It is important to recognize that along with all environmental modelling activities, models are simply abstract representations of the real world and that there are uncertainties about the true properties of the soil. How we present soil patterns on a soil map is based on a predictive model, which is never truly free of errors; hence, quantitatively estimating the uncertainty of digital soil maps is just as important as validating it because it will allow users to assess the usefulness of the maps and how those maps may be used (Heuvelink, 1998). Uncertainties are accumulated and propagated from three sources: the measured soil property, the environmental predictors representing the *scorpan* factors, and the predictive model.

Uncertainty in Soil Properties: Whenever a soil sample is taken to a lab, there will always be differences in analytical techniques that may contribute errors when measuring various soil properties. In some situations, the soil response variables may be inferred from other soil variables that are easier to measure using pedotransfer functions whereby errors may be further propagated. Uncertainties could also arise when recording the spatial location of sample points when using a GPS, whereby some GPSs are more accurate than others.

Uncertainty in Environmental Predictors: There is inherent uncertainty with environmental predictors themselves. For example, vertical uncertainties may be present with DEMs as well as measurement uncertainty from remote and proximal soil sensors. When using conventional soil maps or geology maps in the polygon format, the composition of the polygon, polygon boundaries, and map scale may all contribute to uncertainty.

Uncertainty in Predictive Model: As previously discussed, the choice of predictive model may result in drastically different soil maps; and hence, the structure of the model may be another source of uncertainty. The relationships between the environmental predictors and soil properties may be linear, non-linear, or a combination of both, and certain types of models are more effective at capturing those relationships than others. For examples, a simple linear model would be effective at modeling linear relationships, whereas non-linear modelling approaches (e.g., classification and regression trees) are effective at capturing the non-linear, hierarchical relationships. Finally, some modelling approaches (e.g., model trees) are a hybrid of linear and non-linear models and are therefore able to capture both types of relationships.

Ideally, all digital soil maps should be accompanied with uncertainty maps, which estimate the uncertainty for each individual pixel—in practice, however, this is often not the case. Due to the importance of recognizing uncertainty in model outputs, international mapping standards (e.g., *GlobalSoilMap.net*) often require the inclusion of a 90% prediction interval map as well as the lower (5%) and upper (95%) prediction limits (Figure 17.22). Here, we assume that with a 90% confidence level, the real value of a soil property will be within the prediction interval. Therefore, when interpreting these uncertainty maps, the uncertainty will increase as the 90% prediction interval width increases. It should be noted that this representation of uncertainty only applies to continuous soil variables.

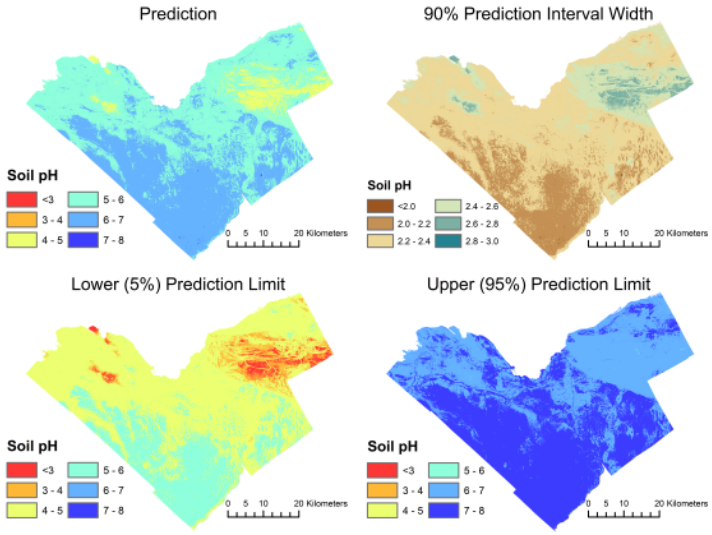


Figure 17.22. Series of maps used to represent uncertainty of digital soil mapping of soil pH in Ottawa, Ontario: prediction, prediction interval width (90%), and the lower (5%) and upper (95%) prediction limits. © Daniel D. Saurette, Ontario Ministry of Agriculture, Food and Rural Affairs is licensed under a CC BY (Attribution) license.

When representing uncertainty for predictions of categorical soil variables, class probability maps are used to calculate uncertainty for each pixel. Here, a set of probability rasters are produced for each class, whereby each pixel has a corresponding value that represents the probability of that class occurring (Figure 17.23). If an individual class has a high probability of occurrence, the uncertainty is low; however, if all the classes have an equal probability of occurrence, the uncertainty is high. Although there are many metrics that may be calculated from these class probability rasters (e.g., ignorance uncertainty, exaggeration uncertainty, confusion index), they are all effectively characterizing the spread in probability values amongst the various classes (Figure 17.24).

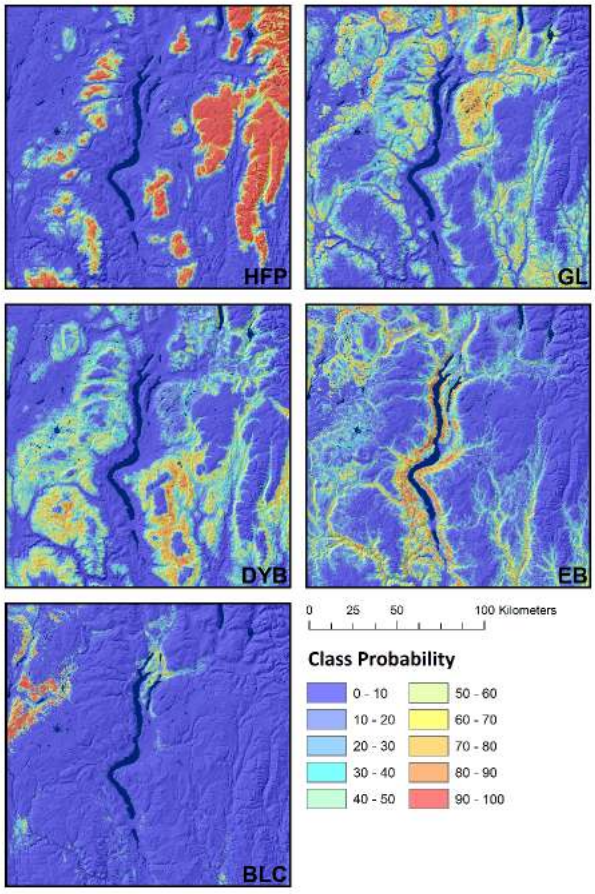


Figure 17.23. Class probability maps for soil great group prediction for the Okanagan-Kamloops region of British Columbia derived at a 100 m spatial resolution using the Random Forest classifier. Most frequently occurring soil great groups include Humo-Ferric Podzols (HFP), Gray Luvisols (GL), Dystric Brunisols (DYB), Eutric Brunisols (EB), and Black Chernozems (BLC). © Brandon Heung. CC BY.

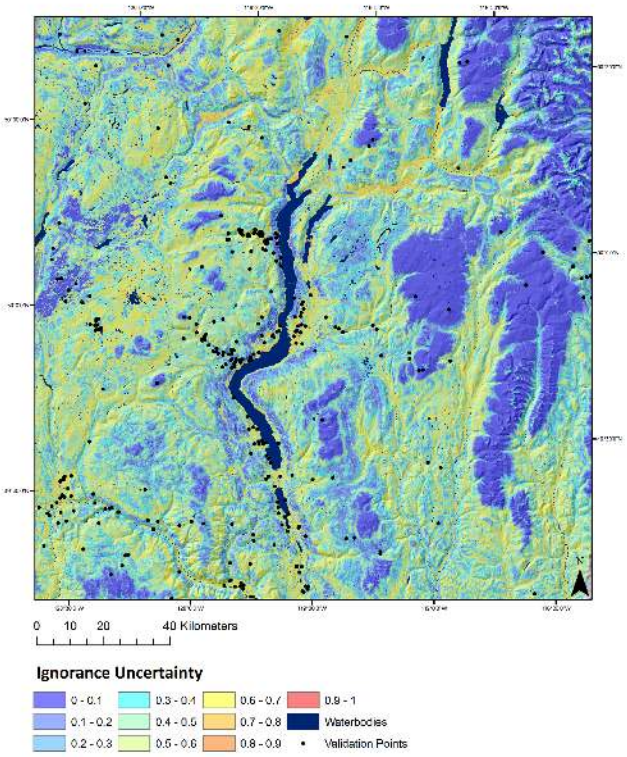


Figure 17.24. Ignorance uncertainty map produced from the soil great group probability maps at a 100 m spatial resolution for the Okanagan-Kamloops region of British Columbia. Higher values (red) represent higher uncertainty. © Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.

BEYOND DIGITAL SOIL MAPPING

Much effort has been made in improving the *accessibility, interpretability, understanding, and communication* of soil information to data users in some provinces. Online applications, such as the British Columbia Soil Information Finder Tool (SIFT), Agricultural Land Resource Atlas of Alberta, the Saskatchewan Soil Information System (SKSIS), Info-Sols (Québec) and the Ontario Agricultural Information Atlas, leverage legacy soil maps and provide a graphical user interface whereby a user could easily query those maps and acquire information related to soil properties as well as interpreted soil information including agricultural land capability and soil erosion risk.

Other platforms, such as SOILx, have explored the use of augmented reality in visualizing soil profiles on a mobile device. These platforms have the potential to improve decision-making processes, aid environmental impact assessments, and provide sustainable soil management information to users. However, these platforms are limited by the quality of the soil information that populates them and the user's ability to understand that information.

Traditionally, soil mappers have focused on the development of quantitative methods for predicting the soils over space. However, the other critical role of a soil mapper, which has received far less attention, is that of a "knowledge generator and provider" and understanding how soil information may be used to inform decision-making and resource-planning practices (Finke, 2012). Together with domain-experts (e.g., agronomists, foresters, environmental scientists, engineers), the role of the soil mapper is to collectively interpret the various digital soil maps and develop interpretive tools that assess soil threats and functions and facilitate the decision-making process. For example, a soil erosion risk map is probably more useful to a decision-maker than individual maps of soil properties (e.g., organic matter and soil texture maps) and other environmental data layers (e.g., slope and rainfall maps).

Within the pedometrics community, a large part of the discipline is devoted to using digital soil maps as ingredients for assessing soil functions and threats. This active area of research is referred to as *digital soil assessments* (DSA). While the details of DSA is beyond the scope of this chapter, interested readers should refer to Carré et al. (2007), who first introduces this field, and Finke

(2012), who subsequently discusses the scope of the field. Globally, there has been tremendous progress towards developing soil assessment tools using digital soil maps.

Applications of digital soil information

Applications may include, but are not limited to:

- Mapping management zones for variable rate fertilizer and irrigation applications (Fleming et al., 2000)
- Assessing regional-scale agricultural land suitability (Harms et al., 2015; Kidd et al., 2015)
- Estimating the potential gross margins for specific crops (Kidd et al., 2015)
- Evaluating national-scale soil erosion (Panagos et al., 2015)
- Assessing soil carbon sequestration potential (Angers et al., 2011; Akpa et al., 2016)
- Aiding the provision of site- and soil-specific insurance to manage soil threats (Cook et al., 2008)
- Assessing human exposure to soil contaminants (Caudeville et al., 2012)
- Characterizing wine regions (terrons and terroir) using climate and soil information (Carré et al., 2005; Coggins et al., 2019)
- Developing ecosite maps for stand-level forest management plans (Yang et al., 2017)
- Mapping landscape-scale soil health indices (Svoray et al., 2015)
- Farm-scale soil organic carbon audits and monitoring (de Gruijter et al., 2016; Malone et al., 2018)

Can You Dig It!

The Canadian Digital Soil Mapping Working Group



***(Top)** First meeting of the CDSMWG held in 2016, Quebec City; **(Middle)** First education workshop on digital soil mapping held at Simon Fraser University, 2017, and in Vancouver, BC; **(Bottom)** Members of the CDSMWG at the Canadian Society of Soil Science Annual Meeting, 2019 held at University of Saskatchewan, Saskatoon, SK. © Soils of Canada website (https://soilsofcanada.ca/digital-soil-mapping); Brandon Heung, Dalhousie University is licensed under a CC BY (Attribution) license.*

Over the last 20 years, the soil science community has evolved from the dominance of government soil surveyors and university scientists to a community where scientists with soil expertise are working in every public and private sector (e.g. agriculture, forestry, environment) and across Canada's diverse landscape. Unfortunately, this has led to the decentralization and dispersion of soil information and soil mapping expertise.

Out of that recognition, a major development in the national soil mapping landscape was the establishment of the Canadian Digital Soil Mapping Working Group (CDSMWG)—a national scientific network with researchers from multiple academic institutions, federal government departments, and provincial agencies. To provide a mechanism and platform for soil mappers to collaborate at the national-scale, the Pedology Committee of the Canadian Society of Soil Science (CSSS) established the CDSMWG in 2016.

This cross-country, cross-sector research network is tasked with coordinating national-scale DSM initiatives, disseminating digital soil information, and delivering educational workshops.

For example, the community developed the soil carbon maps of Canada for submission to the FAO's global carbon mapping project. All products developed by the CDSMWG are designed to meet international DSM standards of GlobalSoilMap.net and the Food and Agriculture Organization (FAO) of the United Nations. To date, the CDSMWG has been successful in its mandate; it was instrumental in developing and delivering a preliminary Canadian Soil Organic Carbon Map (CSOC map) as part of Canada's contribution to the Global Soil Organic Carbon Map (GSOC map) compiled by the Intergovernmental Technical Panel on Soils of the FAO in 2017 (FAO and Intergovernmental Technical Panel on Soils, 2018). Dedicated volunteers in the project clearly demonstrated the network's ability to effectively communicate and collaborate in the daunting task of developing the CSOC map.

In addition, members of the CDSMWG are also dedicated to training the next generation of digital soil mappers and have offered multiple workshops across the country thus far. Workshop attendees have included undergraduate and graduate students, university and government researchers, and industry participants.

THOUGHT EXERCISES

- Brainstorm some ways on how digital soil maps may be used for addressing environmental and resource management issues at local-scales (e.g., individual farm fields or forest stands), regional-scales (e.g., individual provinces), and national-scales (e.g., Canada).
- Digital soil data is becoming increasingly available in Canada. Please take some time to explore the following resources and brainstorm the benefits and challenges of digital soil mapping: British Columbia Soil Information Finder Tool; Saskatchewan Soil Information System; and Ontario AgMaps Geographic Information Portal. International resources available include: Australian Soil Resource Information System, and SoilGrids.

REFERENCES

Adamchuk, V., Allred, B., Doolittle, J., Grote, K., and Viscarra Rossel, R. 2017. Chapter 6: Tools for proximal soil sensing. In Soil Survey Manual, Soil Science Division Staff. Agricultural Handbook No. 18, United States Department of Agriculture, pp. 355-394.

Angers, D.A., Arrouays, D., Saby, N.P.A., and Walter C. 2011. Estimating and mapping the carbon saturation deficit of French agricultural topsoils. Soil Use Manage. **27**: 448-452.

Akpa, S.I.C., Odeh, I.O.A., Bishop, T.F.A., Hartemink, A.E., and Amapu, I.Y. 2016. Total soil organic carbon and carbon sequestration potential in Nigeria. Geoderma **271**: 202-215.

Akumu, C.E., Johnson, J.A., Ethreidge, D., Uhlig, P., Woods, M., Pitt, D.G., and McMurray, S. 2015. GIS-fuzzy logic based approach in modeling soil texture: Using parts of the Clay Belt and Hornepayne region in Ontario Canada as a case study. Geoderma **239-240**: 13-24.

Akumu, C.E., Woods, M., Johnson, J.A., Pitt, D.G., Uhlig, P., and McMurray, S. 2016. GIS-fuzzy logic technique in modeling soil depth classes: Using parts of the Clay Belt and Hornepayne region in Ontario, Canada as a case study. Geoderma **283**: 78-87.

Arrouays, D., Grundy, M.G., Hartemink, A.E., Hempel, J.W., Heuvelink, G.B.M., Hong, S.Y., Lagacherie, P., Lelyk, G., McBratney, A.B., McKenzie, N.J., Mendonça -Santos, M.L., Minasny, B., Montanarella, L., Odeh, I.O.A., Sanchez, P.A., Thompson, J.A., and Zhang, G-L., 2014. GlobalSoilMap: Toward a fine-resolution global grid of soil properties. Adv. Agron. **125**: 93-134.

Beckett, P.H.T. 1971. The cost-effectiveness of soil survey. Outlook Agr. **6**: 191-198.

Bie, S.W., Ulph, A., and Beckett, P.H.T. 1973. Calculating the economic benefits of soil survey. J. Soil Sci. **24**: 429-435.

Biswas, A., and Zhang, Y. 2018. Sampling Designs for Validating Digital Soil Maps: A Review. Pedosphere **28**: 1–15.

Brus, D.J., and de Gruijter, J.J. 1997. Random sampling or geostatistical modelling? Choosing between design-based and model-based sampling strategies for soil (with discussion). Geoderma **80**: 1–44.

Brus, D.J., Kempen, B., and Heuvelink, G.B.M. 2011. Sampling for validation of digital soil maps. Eur. J. Soil Sci. **62**: 394-407.

Brus, D.J. 2019. Sampling for digital soil mapping: A tutorial supported by R scripts. Geoderma **338**: 464–480.

Canada Soil Survey Committee, 1978. The Canadian System of Soil Classification. 1st ed. Research Branch, Canadian Department of Agriculture.

Carré, F., and McBratney, A.B. 2005. Digital terron mapping. Geoderma **128**: 340-353.

Carré, F., McBratney, A.B., Mayr, T., and Montanarella, L. 2007. Digital soil assessments: Beyond DSM. Geoderma **142**: 69-79.

Caudeville, J., Goovaerts, P., Carré, F., Bonnard, R., Arrouays, D., Cicolella, A., Jolivet, C.C., Saby, N.P.A., and Boudet, C. 2012. Spatial modeling of human exposure to soil contamination – an example of digital soil assessment. *In* Digital Soil Assessments and Beyond, CRC Press.

Chandler, J.R., Schmidt, M.G., and Dragicevic, S. 2008. Spatial patterns of forest floor properties and litterfall amounts associated with bigleaf maple in conifer forest of southwestern British Columbia. Can. J. Soil Sci **88**: 295-313.

Chang, C., Sommerfeldt, T.G., and Entz, T. 1988. Soil salinity and sand content variability determined by two statistical methods in an irrigated saline soil. Can. J. Soil Sci. **68**: 209-221.

Clifford, D., Payne, J.E., Pringle, M.J., Searle, R., and Butler, N. 2014. Pragmatic soil survey design using flexible Latin hypercube sampling. Computers and Geosciences. **67**: 62-68.

Coen, G.M. (ed.) 1987. Soil survey handbook. Volume 1. Technical Bulletin 1987-9E. Land Resource Research Center. Research Branch. Agriculture Canada. Edmonton, Alberta. 135pp.

Coggins, S., Malone, B.P., Stockmann, U., Possell, M., and McBratney, A.B. 2019. Towards meaningful geographical indication: Validating terroirs on 200 km2 scale in Australia's lower Hunter Valley. Geoderma Reg. **15**: e00209.

Cook, S.E., Jarvis, A., and Gonzalez, J.P. 2008. A new global demand for digital soil information. In Digital Soil Mapping with Limited Data. Springer, 31-41.

de Gruijter, J.J., and Marsman, B.A. 1985. Transect sampling for reliable information on mapping units. Pages 150–165 in D.R. Nielsen and J. Bouma, eds. Soil spatial variability: proceedings of a workshop of the ISSS and the SSSA, Las Vegas, USA, 30 Nov. – 1 Dec. 1984.

de Gruijter, J.J., McBratney, A.B., Minasny, B., Wheeler, I., Malone, B.P., and Stockmann, U. 2016. Farm-scale soil carbon auditing. Geoderma **265**: 120-130.

Dewitte, O., Jones, A., Spaargaren, O., Breuning-Madsen, H., Brossard, M., Dampha, A., Deckers, J., Gallali, T., Hallett, S., Jones, R., Kilasara, M., le Roux, P., Michéli, E., Montanarella, L., Thiombiano, L., van Ranst, E., Yemefack, M., and Zougmore, R. 2013. Harmonisation of the soil map of Africa at the continental scale. Geoderma **211-212**: 138-153.

Doolittle, J.A., and Brevik, E.C. 2014. The use of electromagnetic induction techniques in soils studies. Geoderma **223-224**: 33-45.

FAO and Global Soil Partnership, 2016. Global Soil Partnership Pillar 4 Implementation Plan: Towards a Global Soil Information System. Pillar 4 Working Group, 38 pp.

FAO and Intergovernmental Technical Panel on Soils. 2018. Global Soil Organic Carbon Map (GSOCmap) Technical Report. Rome, 162 pp.

Finke, P.A. 2012. On digital soil assessments with models and the Pedometrics agenda. Geoderma **171-172**: 3-35.

Fleming, K.L., Westfall, D.G., Wiens, D.W., and Brodahl, M.C. 2000. Evaluating farmer defined management zone maps for variable rate fertilizer application. Precis. Agric. **2**: 201-215.

Fulton, R.J. 1995. Surficial Materials of Canada. Geological Survey of Canada, Natural Resources Canada, "A" Series Map 1880A.

Harms, B., Brough, D., Philip, S., Bartley, R., Clifford, D., Thomas, M., Willis, R., and Gregory, L. 2015. Digital soil assessment for regional agricultural land evaluation. Glob. Food Sec. **5**: 25-36.

Hengl, T., de Jesus, J.M., Heuvelink, G.B.M., Gonzalex, M.R., Kilibarda, M., Blagotić, Shangguan, W., Wright, M.N., Geng, X., Bauer-Marschallinger, B., Guevara, M.A., Vargas, R., MacMillan, R.A., Batjes, N.H., Leenaars, J.G.B., Ribeiro, E., Wheeler, I., Mantel, S., and Kempen, B. 2017. SoilGrids250m: Global gridded soil information based on machine learning. PLoS ONE **12**: e0167948.

Heung, B., Bulmer, C.E., and Schmidt, M.G. 2014. Predictive soil parent material mapping at a regional-scale: A Random Forest approach. **Geoderma 214-215**: 141-154.

Heung, B., Hodúl, M., and Schmidt, M.G. 2017. Comparing the use of legacy soil pits and soil survey polygons as training data for mapping soil classes. Geoderma **290**: 51-68.

Heung, B., Zhang, J., Ho, H.C., Knudby, A., Bulmer, C.E., and Schmidt, M.G. 2016. An overview and comparison of machine-learning techniques for classification purposes in digital soil mapping. Geoderma **265**: 62-77.

Heuvelink, G.B.M., 1998. Error Propagation in Environmental Modelling with GIS. CRC Press.

Hole, F.D. 1978. An approach to landscape analysis with emphasis on soils. Geoderma **21**: 1-23.

Hole, F.D., and Campbell, J.B. 1985. Soil Landscape Analysis. Rowman and Allanheld, Totowa, NJ.

Jenny, H. 1941. Factors of Soil Formation: A System of Quantitative Pedology. McGraw-Hill, NY.

Kidd, D., Webb, M., Malone, B., Minasny, B., McBratney, A. 2015. Digital soil assessment of agricultural suitability, versatility and capital in Tasmania, Australia. Geoderma Reg. **6**: 7-21.

Kienast-Brown, S., Libohova, Z., Boettinger, J. 2017. Digital soil mapping. In C. Ditzler, K. Scheffe, and H.C. Monger (eds.) Soil survey manual, USDA Handbook 18, Government Printing Office, Washington, D.C., pp. 295–354. https://www.nrcs.usda.gov/wps/portal/nrcs/detail/soils/ref/?cid=nrcs142p2_054261

Luttmerding, H.A. 1980. Soils of the Langley-Vancouver Map Area. Report No. 15, British Columbia Soil Survey. BC Ministry of Environment, Kelowna, BC, Canada.

MacMillan, R.A., Nikiforuk, L., Fawcett, M.D., Whitson, I.R., McNeil, R.L. and Turchenek, L.W. 1992. Soil mapping systems; Alberta Research Council, ARC/AGS Open File Report 1992-22, 238 p. https://ags.aer.ca/publications/OFR_1992_22.html

MacMillan, R.A., Pettapiece, W.W., Nolan, S.C., and Goddard, T.W. 2000. A generic procedure for automatically segmenting landforms into landform elements using DEMs, heuristic rules and fuzzy logic. Fuzzy Sets and Systems 113: 81–109,

MacMillan, R.A., Jones, K., and McNabb, D.H., T.W. 2004.Defining a hierarchy of spatial entities for environmental analysis and modeling using digital elevation models (DEMs). Computers, Environment and Urban Systems 28(3): 175-200.

Malone, B., Hedley, C., Roudier, O., Minasny, B., Jones, E., and McBratney, A.B. 2018. Auditing on-farm soil carbon stocks using downscaled national mapping products: Examples from Australia and New Zealand. Geoderma Reg. **13**: 1-14.

Malone, B.P., Minansy, B., and Brungard, C. 2019. Some methods to improve the utility of conditioned Latin hypercube sampling. PeerJ 7, e6451.

Mapping Systems Working Group. 1981. A soil mapping system for Canada: revised. Land Resource Research Institute, Contribution No. 142. Agriculture Canada, Ottawa, 94 pp. http://sis.agr.gc.ca/cansis/publications/manuals/1981-smsc/81-142-soil-mapping.pdf

McBratney, A.B., Mendonça-Santos, M.L., and Minasny, B. 2003. On digital soil mapping. Geoderma **117**: 3-52.

McBratney, A.B., Odeh, I.O.A., Bishop, T.F.A., Dunbar, M.S., and Shatar, T.M. 2000. An overview of pedometric techniques for use in soil survey. Geoderma **97**: 293-327.

Mckay, M.D., Beckman, R.J., and Conover, W.J. 1979. A Comparison of Three Methods for Selecting Values of Input Variables in the Analysis of Output From a Computer Code. Technometrics **42**: 55–61.

McKeague, J. A. and Stobbe, P.C. 1978. History of soil survey in Canada 1914-1975. Research Branch, Canada Department of Agriculture, Ottawa, ON. Historical Series No. 11.

McKenzie, N., and Ryan, P. 1999. Spatial prediction of soil properties using environmental correlation. Geoderma **89**: 67-94.

Minasny, B., and McBratney, A.B. 2006. A conditioned Latin hypercube method for sampling in the presence of ancillary information. Comput. Geosci. **32**: 1378–1388.

Minasny, B., and McBratney, A.B. 2016. Digital soil mapping: A brief history and some lessons. Geoderma, **264**: 301-31.

Mulder, V.L., De Bruin, S., Shaepman, M.E., and Mayr, T.R. 2011. The use of remote sensing in soil and terrain mapping – a review. Geoderma **162**: 1-19.

Panagos, P., Borrelli, P., Poesen, J., Ballabio, C., Lugato, E., Meusburger, K., Montanarella, L., and Alewell, C. 2015. The new assessment of soil loss by water erosion in Europe. Environ. Sci. Policy **54**: 438-447.

Paul, S.S., Coops, N.C., Johnson, M.S., Krzic, M., Chandna, A., and Smukler, S.M. 2020a. Mapping soil organic carbon and clay using remote sensing to predict soil workability for enhanced climate change adaptation. Geoderma **363**: 114177.

Paul, S.S., Dowell, L., Coops, N.C., Johnson, M.S., Krzic, M., Geesing, D., Smukler, S.M. 2020b. Tracking changes in soil organic carbon across the heterogeneous agricultural landscape of the Lower Fraser Valley of British Columbia. Sci. Total Environ. **732**: 138994.

Pennock, D.J., Zebarth, B.J., and DeJong, E. (1987) Landform classification and soil distribution in hummocky terrain, Saskatchewan, Canada. Geoderma, 40: 297-315.

Pennock, D., and Yates, T. 2007. Soil sampling designs. Pages 1–14 in M.R. Carter and E.G. Gregorich, eds. Soil sampling and methods of analysis, Second Edition.

Rossiter, D.G. 2018. Past, present & future of information technology in pedometrics. Geoderma **324**: 131-137.

Sanchez, P.A., Ahamed, S., Carré, F., Hartemink, A.E., Hempel, J., Huising, J., Lagacherie, P., McBratney, A.B., McKenzie, N.J., Mendoça-Santos, M.L., Minasny, B., Montanarella, L., Okoth, P., Palm, C.A., Sachs, J.D., Shepherd, K.D., Vâgen, T-G., Vanlauwe, B., Walsh, M.G., Winowiecki, L.A., and Zhang, G-L. 2009. Digital soil map of the world. Science **325**: 680-681.

Scarpone, C., Schmidt, M.G., Bulmer, C.E., and Knudby, A. 2016. Modelling soil thickness in the critical zone for Southern British Columbia. Geoderma **282**: 59-69.

Schaetzl, R.J., Anderson, S., 2005. Soils: Genesis and Geomorphology. Cambridge University Press.

Scull, P., Franklin, J., Chadwick, O.A., and McArthur, D. 2003. Predictive soil mapping: A review. Prog. Phys. Geog. **27**: 171-197.

Shatar, T.M., and McBratney, A.B. 1999. Empirical modeling of relationships between sorghum yield and soil properties. Precis. Agric. **1**: 249-276.

Soil Survey Staff, 1975. Soil Taxonomy. A Basic System of Soil Classification for Making and Interpreting Soil Surveys. 2nd ed. U.S. Government Print Office, Washington, DC.

Svoray, T., Hassid, I., Atkinson, P.M., Moebius-Clune, B.N., and van Es, H.M. 2015. Mapping soil health over large agriculturally important areas. Soil Sci. Soc. Am. J. **79**: 1420-1434.

Thompson, J.A., Nauman, T.W., Odgers, N.P., Libohova, Z., and Hempel, J.W. 2012. Harmonization of legacy soil maps in North America: Status, trends, and implications for digital soil mapping efforts. In Digital Soil Assessments and Beyond. CRC Press, Leiden, Netherlands, pp. 97-102.

Tobler, W. 1970. A computer movie simulating urban growth in the Detroit region. Economic Geography **46**:230-240.

Treitz, P., Lim, K., Woods, M., Pitt, D., Nesbitt, D., and Etheridge, D. 2012. LiDAR sampling density for forest resource inventories in Ontario, Canada. Remote Sens. **4**: 830-848.

Viscarra Rossel, R.A., Adamchuk, V.I., Sudduth, K.A., McKenzie, N.J., and Lobsey, C. 2011. Proximal soil sensing: An effective approach for soil measurements in space and time. Adv. Agron. **113**: 243-291.

Webster, R. 1994. The development of pedometrics. Geoderma **62**: 1-15.

Webster, R., and Burrough, P.A. 1972a. Computer-based soil mapping of small areas from sample data – I: Multivariate classification and ordination. J. Soil Sci. **23**: 210-221.

Webster, R., and Burrough, P.A. 1972b. Computer-based soil mapping of small areas from sample data – II: Classification smoothing. J. Soil Sci. **23**: 222-234.

Wheeler, J.O., Hoffman, P.F., Card, K.D., Davidson, A., Sanfrd, B.V., Okulitch, A.V. and Roest, W.R. 1996. Geological Map of Canada. Geological Survey of Canada, Natural Resources Canada, “A” Series Map 1860A.

Woods, M., Pitt, D., Penner, M., Lim, K., Nesbitt, D., Etheridge, D., and Treitz, P. 2011. Operational implementation of LiDAR inventory in Boreal Ontario. Forest Chron. **87**: 512-528.

Yang, Q., Meng, F.-R., Bourque, C.P.-A., and Zhao, Z. 2017. Production of high-resolution forest-ecosite maps based on model predictions of soil moisture and nutrient regimes over a large forested area. Sci. Rep. **7**: 10998.

Zhu, A.X., and Band, L.E. 1994. A knowledge-based approach to data integration for soil mapping. Can. J. Remote Sens. **20**: 408-418.

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Brandon Heung (licensed under a CC BY-NC-ND license)

Dr. Brandon Heung is an assistant professor in soil science and geographical information systems (GIS) in the Department of Plant, Food, and Environmental Sciences. His research is in the area of pedometrics—a subdiscipline of soil science that integrates soil science, GIS, remote sensing, spatial statistics, and machine-learning to better understand that spatial and temporal patterns of soil. He has been developing digital soil maps for applications in agriculture and forested systems and he is also interested in understanding how predictive digital soil models may be used to address climate change and food security issues.

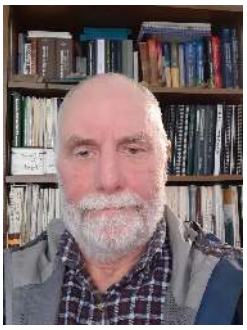
Daniel D. Saurette, Ontario Ministry of Agriculture, Food, and Rural Affairs; School of Environmental Sciences, Ontario Agricultural College, University of Guelph



Dan Saurette (licensed under a CC BY-NC-ND license)

My work focuses on soil survey, mapping and classification. Typically, I would be referred to either as a Pedologist, or Soil Surveyor. My work involves planning field campaigns to describe soil profiles, classify them, sample the soil to determine chemical and physical properties in the lab, and then interpret the data to develop soil maps. Traditionally this was done with the aid of air photos, but more recently the discipline has shifted to Predictive Digital Soil Mapping where we leverage the power of machine learning and geographic information systems (GIS) to create high-resolution digital soil maps.

Chuck E. Bulmer, British Columbia Ministry of Forests, Lands, Natural Resource Operations and Rural Development



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My work involves carrying out research on soil inventory and soil conservation and to provide advice to policy makers in BC regarding resource management questions that relate to soil. Some of the topics of interest to managers include reforestation, impacts of forest management on soil productivity, and soil processes affecting our response to carbon cycling and climate change. Scientific information is needed by forest managers to make good decisions in these areas and my role as a research soil scientist is to carry out research aimed at developing that information for BC, and conveying it to managers.

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