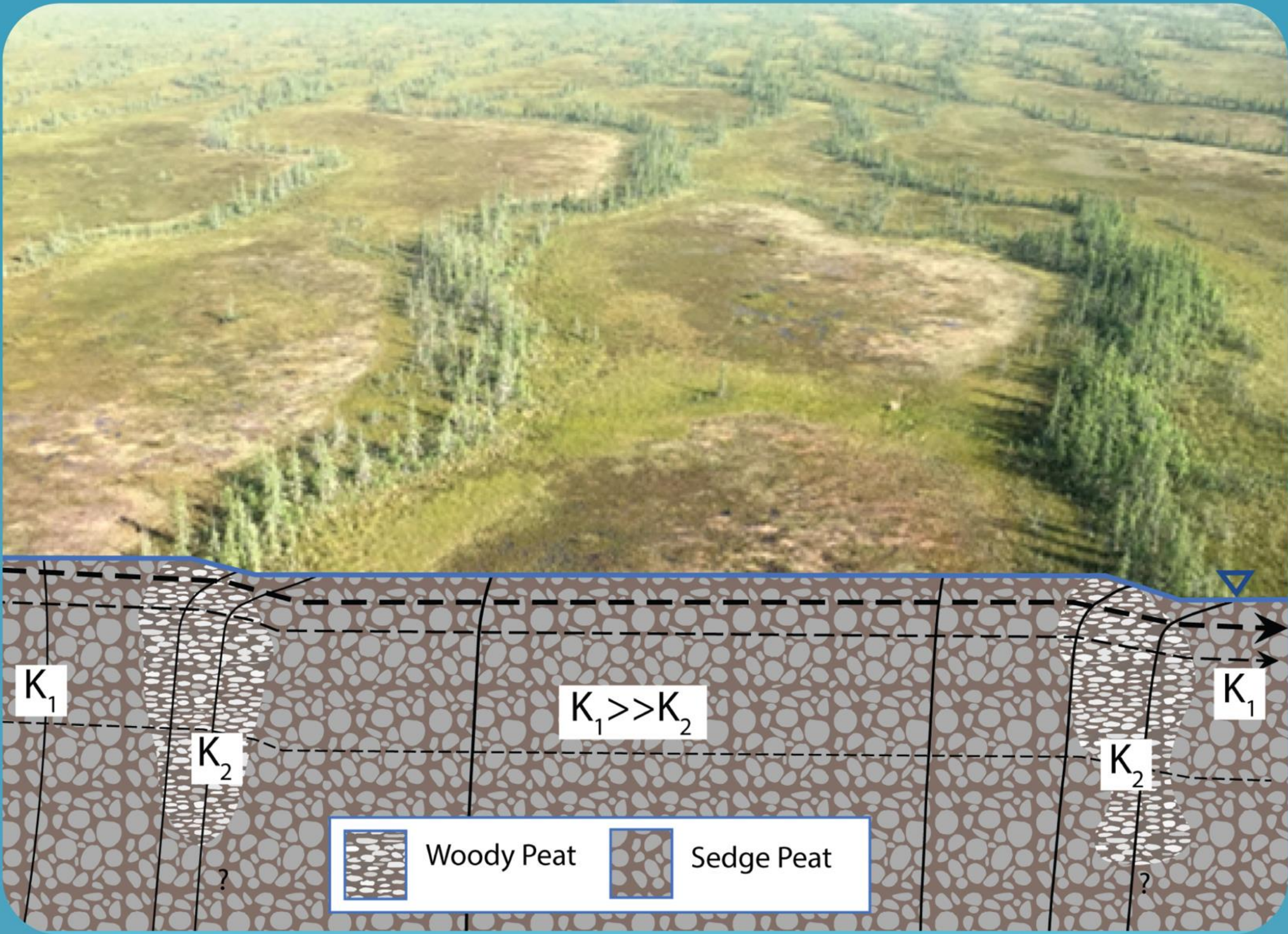

Groundwater in Peat and Peatlands

Jonathan S.Price, Colin P.R. McCarter and Willian L. Quinton

THE GROUNDWATER PROJECT

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THE GROUNDWATER PROJECT

Cover Image: Jonathan Price, 2018.

Dedication

This book is dedicated to the graduate students and post-doctoral fellows, whose tireless enthusiasm, insight, commitment, and hard work is the backbone of the advances in research that we seek.

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The Groundwater Project Foreword

At the United Nations (UN) Water Summit held on December 2022, delegates agreed that statements from all major groundwater-related events will be unified in 2023 into one comprehensive groundwater message. This message will be released at the UN 2023 Water Conference, a landmark event that will bring attention at the highest international level to the importance of groundwater for the future of humanity and ecosystems. This message will bring clarity to groundwater issues to advance understanding globally of the challenges faced and actions needed to resolve the world's groundwater problems. Groundwater education is key.

The 2023 World Water Day theme *Accelerating Change* is in sync with the goal of the Groundwater Project (GW-Project). The GW-Project is a registered Canadian charity founded in 2018 and committed to the advancement of groundwater education as a means to accelerate action related to our essential groundwater resources. To this end, we create and disseminate knowledge through a unique approach: the democratization of groundwater knowledge. We act on this principle through our website gw-project.org/↗, a global platform, based on the principle that

"Knowledge should be free, and the best knowledge should be free knowledge." Anonymous

The mission of the GW-Project is to promote groundwater learning across the globe. This is accomplished by providing accessible, engaging, and high-quality educational materials—free-of-charge online and in many languages—to all who want to learn about groundwater. In short, the GW-Project provides essential knowledge and tools needed to develop groundwater sustainably for the future of humanity and ecosystems. This is a new type of global educational endeavor is made possible through the contributions of a dedicated international group of volunteer professionals from diverse disciplines. Academics, consultants, and retirees contribute by writing and/or reviewing the books aimed at diverse levels of readers from children to high school, undergraduate, and graduate students or professionals in the groundwater field. More than 1,000 dedicated volunteers from 127 countries and six continents are involved—and participation is growing.

Hundreds of books will be published online over the coming years, first in English and then in other languages. An important tenet of GW-Project books is a strong emphasis on visualization; with clear illustrations to stimulate spatial and critical thinking. In future, the publications will also include videos and other dynamic learning tools. Revised editions of the books are published from time to time. Users are invited to propose revisions.

We thank you for being part of the GW-Project Community. We hope to hear from you about your experience with the project materials, and welcome ideas and volunteers!

The GW-Project Steering Committee

January 2023

Foreword

This book, Groundwater in Peat and Peatlands, is unique within the peat literature in that it describes peatlands specifically within the hydrogeologic context according to their characteristics and properties, indicates how this knowledge is acquired and explains the role of peatlands in the context of climate change. Peat is partially decomposed organic matter that accumulated under conditions of water logging and oxygen deficiency. Peat is permeable and is a form of an unconfined aquifer with water at or near the surface in areas of slow groundwater flow.

Peatlands occur predominantly in the northern part of the Northern Hemisphere, mostly in Canada, Fennoscandia, and Russia. These northern peatlands are either seasonally frozen or contain permafrost. There is also tropical peat, occurring mostly in Indonesia, Malaysia, Papua New Guinea, Democratic Republic of Congo, and Brazil.

Although peatlands occupy only 3% of the Earth's land area and few people live on or near peatlands, the state of peatlands is important to the well-being of the planet. Peat stores 30% of the earth's soil carbon. Sustaining this carbon storage requires a shallow water table. If the water table declines then oxygen will enter and oxidize the peat, releasing carbon dioxide into the atmosphere. This is part of a greenhouse gas feedback loop when a warmer climate dries out peat releasing carbon dioxide into the atmosphere, thus contributing to atmospheric warming. When the water table declines in peat, the degree of fire hazard rises markedly. Land use changes in Indonesia and Malaysia are severely impairing peatland function to support the production of palm oil, paper pulp, lumber, and agriculture. Burning peat can result in wildfires with global scale smoke plumes. Hence, keeping peatlands saturated is essential to the well-being of humans and the planet.

The authors have published extensively over decades with coverage of most aspects of peat related to groundwater and much more. Dr. Jonathan Price is an emeritus professor at the Department of Geography and Environmental Management, University of Waterloo, Canada; Dr. William Quinton is a professor at the Department of Geography and Environmental Studies; Wilfred Laurier University, Canada; and Dr. Colin McCarter is a Canada Research Chair in Climate and Environmental Change and an assistant professor in the Department of Geography and the Department of Biology and Chemistry, Nipissing University, Canada.

John Cherry, The Groundwater Project Leader

Guelph, Ontario, Canada, January 2023

Preface

Peatlands are wetlands whose soil comprises the partially decomposed remains of plants that accumulate in a way that both responds to—and controls—the flux and storage of surface water and groundwater within peatlands and runoff to downstream ecosystems. Water tables are generally at or near the surface, at least for much of the year. As the largest global terrestrial store of carbon, peatlands strongly influence global climate.

Depending on the setting, peatlands develop into distinct forms that peatland scientists generally categorize as bogs, fens, and swamps with peat soils. Bogs have accumulated sufficient peat depth, mostly due to the abundance of *Sphagnum* mosses, that they become topographically isolated and receive water only via precipitation, thus shed surface water and groundwater. Boreal and temperate bogs often have a sparse cover of stunted trees, whereas tropical systems can have a dense cover of large trees.

In contrast to bogs, fens and peat swamps generally receive water and constituent dissolved ions from adjacent mineral terrains, although flow directions may reverse. As with bogs, the quantity and quality of water received controls their plant community function and structure and, therefore, their soil properties and ability to store and transmit groundwater. Fens generally have a steadier supply of water than swamps, and thus a more stable water table that favors sedges and brown mosses and sometimes trees; swamps typically have a more episodic water exchange, thus variable water table, that is more favorable for woody vegetation.

In peatlands, more recently formed—thus less decomposed—soils occur in the upper layers; these newer soils can have extremely high porosity (≤ 95 percent) and are typically more permeable (saturated hydraulic conductivity up to 10 to 1000 $m\ d^{-1}$), whereas soils deeper than 0.3 to 0.5 m generally have low hydraulic conductivity (≤ 0.5 $m\ d^{-1}$). Consequently, the transmissivity feedback in peatlands exacerbates surface and groundwater flows when the water table is particularly high.

Understanding the hydrology and water quality of peatlands is key to effective land management where peatlands are common on the landscape.

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We deeply appreciate the thorough and useful reviews of and contributions to this book by the following individuals:

- ❖ Donald Siegel, Research Professor; Professor Emeritus, Earth and Environmental Sciences, Syracuse University, Syracuse, New York, USA;
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- ❖ Connie Bryson, Science Editor, Alberta, Canada;
- ❖ Everton de Oliveira, President of Hidroplan, Director-President of the Instituto Água Sustentável (Sustainable Water Institute), Brazil;
- ❖ John Cherry, Adjunct Professor, University of Guelph; Professor Emeritus, University of Waterloo, Canada;
- ❖ Matthys Dippenaar, Associate Professor, Engineering Geology and Hydrogeology, University of Pretoria, South Africa;

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1 Introduction

Peat is the living and dead, partially decomposed organic matter (mostly plants), that forms the soil matrix of peatlands. Peatlands are wetlands that contain peat; the peat deposit can be considered a shallow unconfined aquifer. This book focuses on the nature of soil water and groundwater exchanges to, from, and within peatlands and the adjacent ecosystems or underlying sediment. It considers the ecohydrological feedback between the landscape that controls the form and function of the peatland and the peat matrix itself that modulates groundwater flows.

The morphology of peatlands is controlled by the interaction of groundwater flow with the ecological processes that create and destroy peat. Where the water table is high, peat can form; where the water table is low, peat decomposes. The shape of the peat surface, in turn, controls the position of the water table and the associated surface water and groundwater flows. In peatlands, groundwater flow and the shape of the ground surface are tightly coupled.

Peatlands can occur as peatland complexes or isolated peatlands based on their connection to other peatlands. Peatland complexes exchange groundwater with the regional aquifer as well as with other types of peatlands. In contrast, isolated peatlands are linked only to the local groundwater system. Peat is the matrix of which all peatlands are built, and its character both reflects and modulates local and regional groundwater systems. Given the distinct nature, character, methods, and applications of groundwater hydrology in peat and peatlands, this book provides a resource for students, researchers, and practitioners pursuing groundwater studies in this type of land cover.

This book begins with definitions of peat and peatlands and how their interaction with the landscape controls their form and function. Then it focuses on the specific properties of peat that control groundwater flow and solute transport and examines the nature and outcome of disturbances to peatlands. The book closes with a discussion of appropriate methods for assessing the hydrology and hydraulic characteristics of peat.

This book is not intended to be a literature review; rather, it is a compendium and synthesis of groundwater and soil-water processes in peat and peatlands and methods for assessing them based strongly on our own collective knowledge and experience. We provide citations to encourage and support further exploration of the processes and issues.

2 Groundwater Controls on Peatland Form and Function

Peat consists of organic soil that accumulates on terrains where the water table is sufficiently close to the ground surface for long enough time to inhibit oxidation and decomposition of plant material. Under water saturated conditions, oxygen diffusion into the soil profile is slow, thus bacterial activity that drives decomposition is hindered, resulting in the accumulation of partially decomposed plant remains (peat). Wetlands that have accumulated a sufficient thickness of organic soil (commonly ~40 cm) are peatlands, and how they form and function reflect their hydrogeomorphic setting and climate (Brinson, 1993). Climate favors peatlands in northern boreal and subarctic latitudes where precipitation exceeds or closely matches potential evaporation, and in tropical areas where particularly high rainfall occurs (Figure 1).

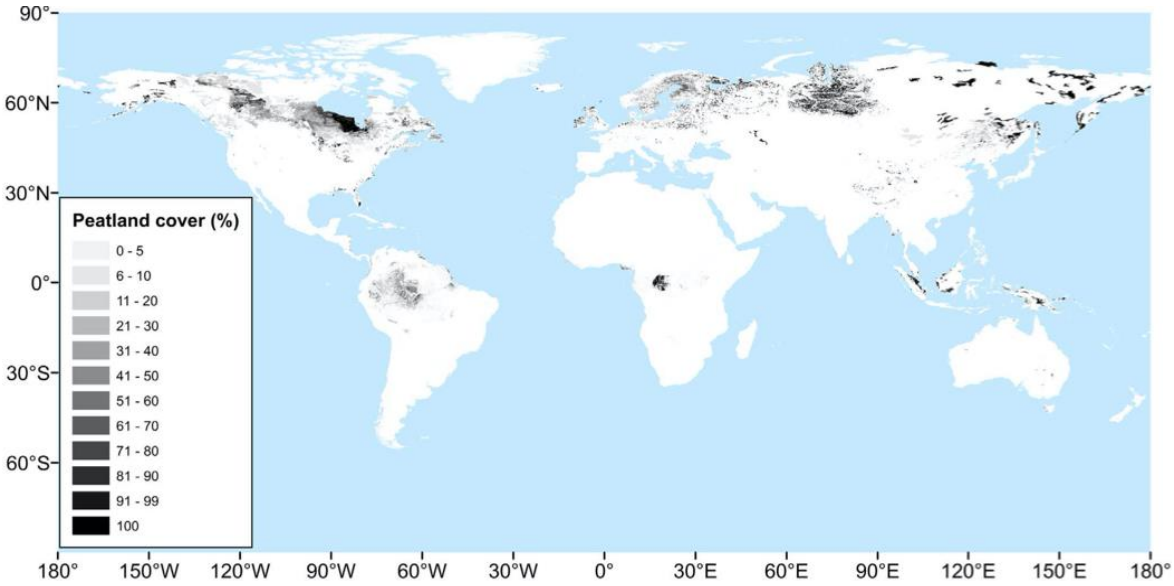


Figure 1 - Global peatland distribution (from Xu et al., 2018). The black shading classes are applicable to Canada and some smaller areas as shown. Elsewhere, black = peat and white = no peat. Data for this map can be accessed in more detail here↗.

Classifications of peatlands commonly reflect the origin, character, and degree of groundwater input. The porewater concentration of dissolved minerals and nutrients introduced by groundwater strongly affects their trophic status and geochemical and ecological function. Commonly, peatlands are categorized and named on the basis of the extent of groundwater interaction; and this convention is used herein. In the context of peatlands, groundwater consists of water below the water table that may be present at or near the ground surface. This book also discusses the variably saturated vadose zone that—from time to time—contains groundwater.

2.1 Hydrogeomorphic Setting and the Categorization of Peatland Systems

The combination of the hydrological and geomorphic settings of a peatland controls the rates and directions of flow, type and amount of dissolved minerals and nutrients from groundwater, and its plant and microbiological ecology. Peatlands in early stages of development commonly occur in valley bottoms and hollows/depressions in the landscape and remain saturated by groundwater discharge from local or regional aquifers in addition to precipitation and surface water.

For a given climate, the strength of groundwater interaction relies on the difference in hydraulic head (e.g., water table elevation) between the wetland and the adjacent upland or regional aquifer, as well as the permeability of the basal materials under the valley that limit the rates of groundwater exchange. Where inflow and precipitation rates are sufficient to cause persistent saturation of the ground surface so that hydrophytic (water-loving) vegetation grows, peat will form and accumulate. Groundwater interaction with peatland systems evolves with time because of strong ecohydrological feedbacks. For example, peat decomposition creates zones of lower hydraulic conductivity that reduce groundwater discharge rates and alters water -table fluctuations and hydraulic gradients. Because of this, we can broadly distinguish a range of peatland types based mostly on their groundwater relations: *swamps*, *fens,* and *bogs* as shown in Figure 2 (Zoltai and Vitt, 1995).

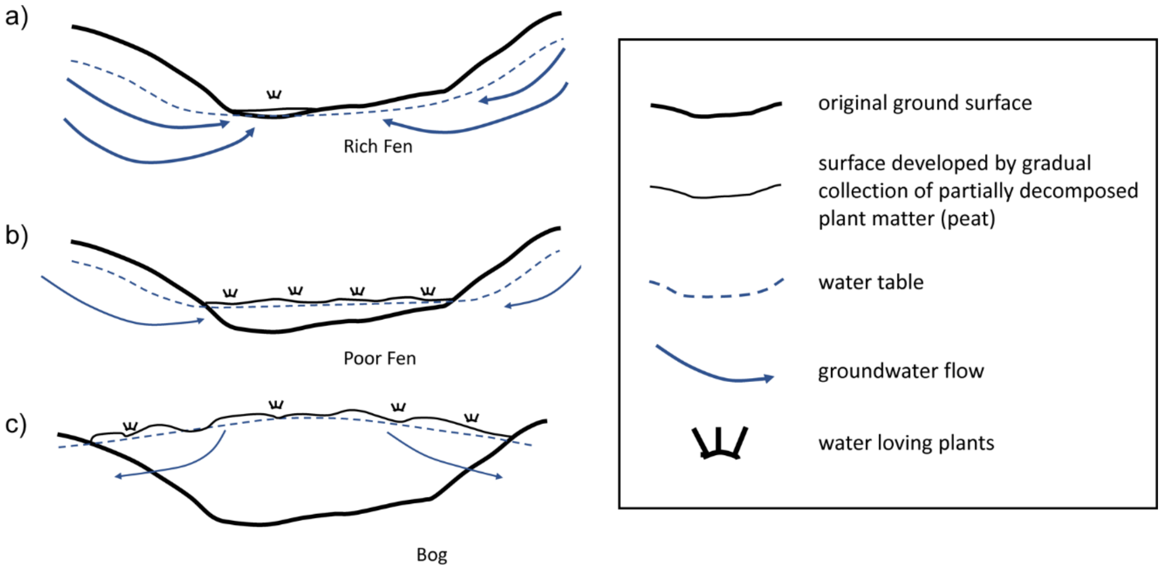


Figure 2 - Transition from a) an incipient rich fen peatland with relatively strong groundwater inflow, to b) poor fen where peat accumulation accompanied by a rise in water table diminishes groundwater inflow, to c) a raised bog where the groundwater flow direction is reversed. Swamps have a hydrogeological configuration similar to fens, although groundwater flow may be weaker compared to surface water inflow.

2.2 Swamps

Swamps are common in boreal, temperate, subtropical, and tropical landscapes but may or may not accumulate peat. Swamps commonly occur in valley bottom or riparian settings; they are dominated by woody plants, either large shrubs or trees (e.g., > 10 m tall), and their hydrological regime is highly episodic (Figure 3).

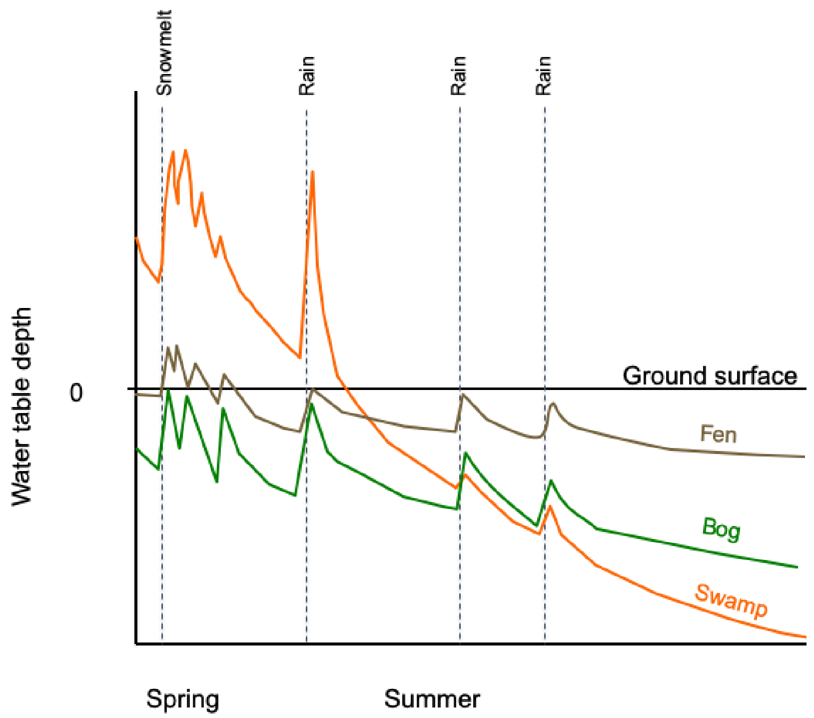


Figure 3 - Hypothetical seasonal hydrographs from temperate and boreal peatlands. Swamps display more extreme variability due to the strength of episodic water inputs; strong surface water inputs exaggerate water table response, while water table fluctuations below the surface may be exaggerated because of low drainable porosity, and strong tree canopy interception and transpiration. Fens have shallower water tables and less variability, due to sustained groundwater and/or surface water inputs. With their water tables in the upper layer of peat, the higher drainable porosity there reduces water table response to water exchanges. Bog water tables respond to episodic water inputs only (rainfall, snowmelt) and water loss is mostly due to evapotranspiration but also to small groundwater seepage losses; they are not modulated by groundwater inflow, and water tables are typically lower than in nearby fens. Values on the vertical axis scale were intentionally omitted but would be on the order of tens of centimeters.

Swamps may or may not receive substantial groundwater input but commonly flood in response to snowmelt or heavy rain or, in the case of riparian settings, overbank flows. Because of surface and groundwater outflows and/or high evapotranspiration, their water table declines steeply during dry periods. This decline is an essential feature since most large woody species cannot withstand persistent flooding. Peat swamps, therefore, must have a hydrological regime wherein drainage is sufficiently impaired to delay decomposition of plant materials, and/or the annual litter deposition is very high to accumulate sufficient organic matter to form peat.

In either case (peat or mineral swamp), water table drawdown must be sufficient to accommodate woody plant growth (Locky et al., 2005). We note, however, that some swamp trees such as bald cypress have root adaptations that permit them to tolerate extended periods of flooding (Harms et al., 1980), although these swamps are not typically peatlands.

In peat swamps, seasonally low water tables often result in highly decomposed peat. This may be enhanced by the labile (easily decomposed) nature of the litter and external mineral/nutrient-rich water inputs. The water in swamps is called *minerogenous*, referring to the input of mineral rich water from adjacent mineral soil uplands, and hence has *minerotrophic* ecologic composition and trophic status. The degree of mineral input depends on the rate and chemical composition of groundwater (and sometimes surface water) inflow that can buffer the organic acids released by decomposing plant matter; thus, swamps can be alkaline, intermediate in pH, or acidic.

2.3 Fens

In subarctic, boreal, and temperate landscapes, fens are typically groundwater- (and sometimes surface water) fed peatlands dominated by bryophytes and sedges, sometimes with tree or woody shrub cover; they are also considered to be minerogenous.

Fens occur in three broad categories that reflect the strength and character of the groundwater inputs, which strongly affect their ecological richness. *Rich fens* develop where strong mineral rich groundwater inputs occur and have pH > 7. *Moderate rich fens* have pH ~5.5 to 7. *Poor fens*, which have the lowest influence of inflowing groundwater at the surface, have pH ~4 to 4.5.

The water table in fens generally remains close to the ground surface (a key difference from swamps), commonly maintained by groundwater inflow from either adjacent upland (Figure 2) or regional aquifers. The water table depth typically occurs from the land surface to only 15 to 20 cm bgs (below ground surface) as shown in Figure 3. In fens and in swamps with peat accumulation, the water table rises with the accumulating soil layer, as drainage is limited by very small gradients. As the peat accumulates, the ground surface and underlying water table rise, a process that reduces the vertical hydraulic gradient and therefore the groundwater flux (including dissolved ions such as calcium) from adjacent uplands (Figure 2).

Where solute concentrations in groundwater are low or where peat decomposition above the underlying mineral soil is great, organic acids released by decomposing vegetation are not neutralized by the groundwater input, so the pH of pore water decreases. This results in a shift in vegetation to a surface dominated by *Sphagnum* mosses (Glaser et al., 2004), which are responsible for further acidification (Van Breemen, 1995). At this stage the peatland has transitioned from a moderate to a poor fen wherein the vegetation diversity is reduced. Therefore, as in swamps, the geochemical character of a fen can be alkaline, intermediate, or acidic.

In larger peatland complexes, the groundwater inputs to poor fens can be derived from more acidic bog *uplands* rather than mineral uplands. In such complexes, broad (10 to 100 m wide) *channel fens* or *ladder fens* form the basin drainage network, connecting aquatic systems that lie along their path. In rich and moderate fens, groundwater and surface flow maintain a relatively high nutrient status and neutral pH, while poor fens are usually found in areas of little or no horizontal flow, or where the primary groundwater source is an adjacent bog peatland.

In fens where relatively strong down-gradient water flows occur, a systematic microtopography can form in a repeating pattern of ridges and pools (sometimes called *flarks*) oriented perpendicular to the primary direction of flow (Figure 4). Ridges (sometimes called *ribs*) have lower hydraulic conductivity (Whittington and Price, 2006). Ribs and flarks result in a stepped water table in the down-gradient direction and sporadic periods of hydrological connectivity when the water table is within the high hydraulic conductivity near-surface peat, or as overland flow. The gradual differentiation between ridge and flark surface forms are associated with positive feedback favoring peat accumulation in better drained micro-sites and decay processes associated with flooding of low areas.

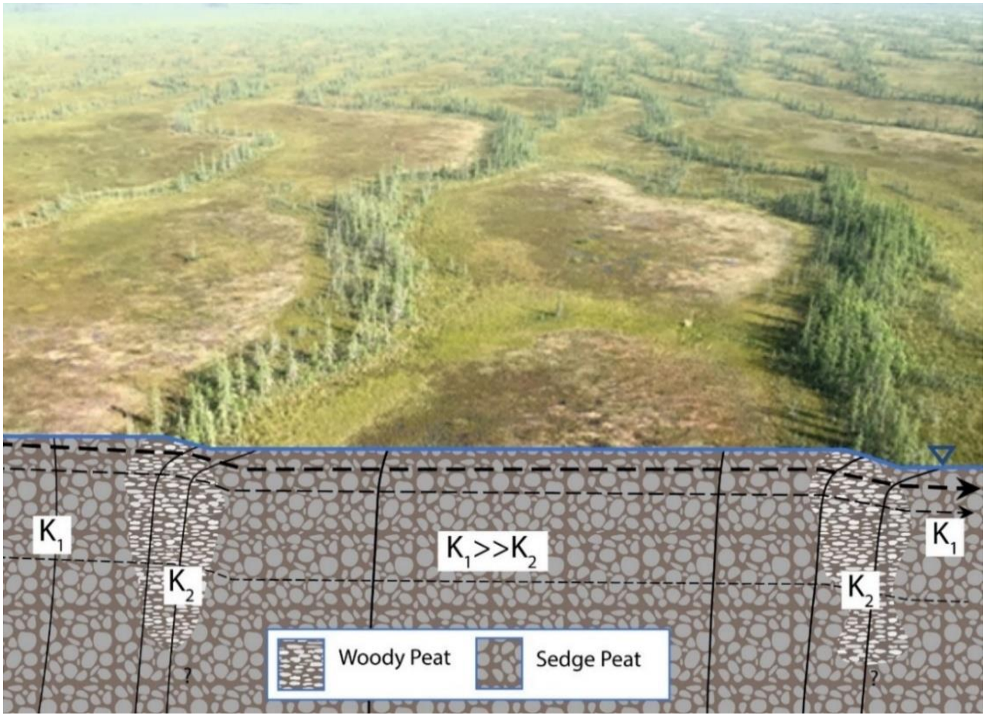


Figure 4 - Ridges (with trees) and flarks (open) in a moderate-rich fen. Peat stratigraphy and genesis (plant type) is distinct between ridges (woody, *Sphagnum*, and sedge) and flarks (sedge). Distinct peat in ridges may extend only part-way to the mineral substrate (Foster et al., 1988). Groundwater flow rates are controlled by ridges, which have lower hydraulic conductivity (K_2) than flarks (K_1), except perhaps in the upper ~10 cm as discussed in Section 4.2. Ridges impound flow, resulting in a stepped water table along the local hydraulic gradient. Flow is perpendicular to ridges, in this case left-to-right (dashed lines indicate flow, which is higher near the surface; vertical black lines are isopotential lines portraying contours of equal hydraulic head). In the foreground, ridge-spacing is approximately 40 m; in other fens, ridges can be much smaller and closer than shown here (Photograph by J. Price).

2.4 Bogs

Further accumulation of peat and the accompanying water table rise to an elevation at or above that in the adjacent mineral uplands (Figure 2), prohibits groundwater inflow from reaching the ground surface and plant rooting zone. In this case, the peatland transitions to a peatland exclusively fed by precipitation near its surface. Such raised peatlands are called bogs, at least in subarctic, boreal, and temperate zones.

Although succession to bog arising from peat accumulation is common, it is important to note that swamp and fen can be stable states and persist as such for millennia. Tropical peatlands can also become domed but given their predominantly woody vegetation they are referred to as swamps (Page et al., 2000). While local groundwater discharge from mineral sediments is absent in domed peatlands, regional groundwater discharge can, in some settings, periodically sustain the local water table (Siegel and Glaser, 1987). In northern regions, the formation of ground ice below peatlands or parts thereof can also raise the peatland surface above the surrounding terrain.

Bog peatlands are dominated by *Sphagnum* mosses, lichens, and herbs, along with small shrubs and stunted trees (Figure 5). Bogs are said to be *ombrogenous*, and thus *ombrotrophic* (“ombro” is derived from the Greek word for rain). The diminished minerotrophy favors a shift to acid-tolerant plants, especially *Sphagnum* mosses. *Sphagnum* mosses are more resistant to decay than most vascular plants and add to the vertical development of the emerging bog. Since *Sphagnum* releases organic acids, their presence further lowers the ambient pH. Van Breemen (1995) explains this in his wittily entitled article “How *Sphagnum* bogs down other plants.”

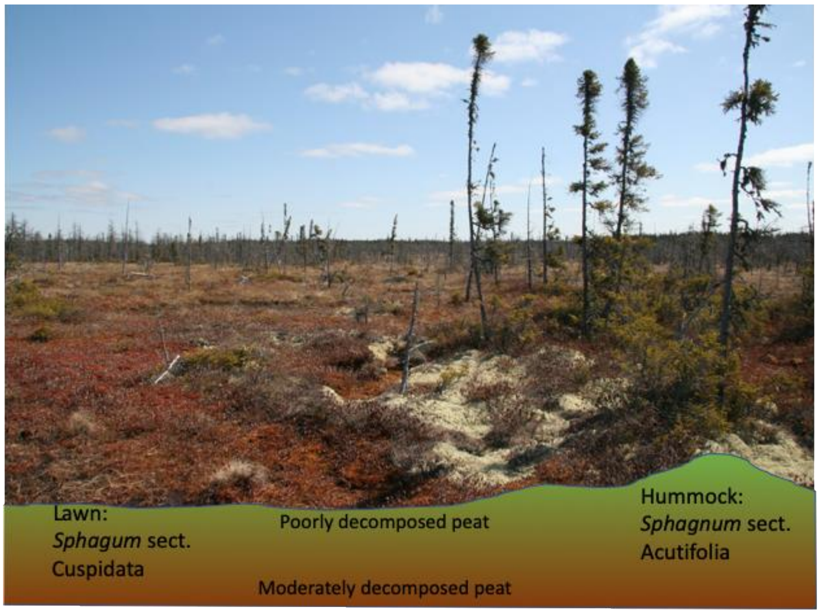


Figure 5 - Boreal bog dominated by *Sphagnum* mosses (left foreground) and stunted black spruce trees (right side and background), woody shrubs (right foreground), and sometimes lichen (bottom right). The dominant peat-forming plant is *Sphagnum*, which predominantly forms the groundwater matrix. *Sphagnum* mosses fall within different sections of the genus. Acutifolia species like S. *fuscum* and S. *rubellum* form relatively tight communities better adapted for being above the water table (better capillarity). Cuspidata species, like S. *papilosum* and S. *angustifolium* have looser growth forms and community structure and are situated closer to the water table. (Photograph by J. Price)

While bogs, by definition, are fed only by precipitation, they generally cannot be distinguished from fens with very weak groundwater inflow based on groundwater measurements. Rather, the distinction between bogs and poor fens is more commonly made on the basis of pH, calcium concentration, and presence/absence of obligate plant species such as *Juncus* spp. and *Equisetum* spp., which are characteristic of poor fens. Bogs generally have pH < 4 (Zoltai and Vitt, 1995).

In bogs, the rate of peat accumulation is greatest furthest from the peatland margins where *Sphagnum* is most likely to develop (Figure 2c). Near the peatland margins, mineral-rich water inhibits *Sphagnum* growth due to elevated concentrations of base cations, especially calcium. In the central peatland, the peat ground surface often rises to produce a peat dome, and such systems are often called *domed*, or *raised bogs*. The height of domed peatlands is discussed in Box 1⤵. In domed tropical peatlands the ultimate shape is first reached near boundaries (i.e., rivers) while carbon storage inland progresses at a rate proportional to the remaining interior area (Cobb et al., 2017). At the margin of a raised bog, a moat-like *lagg* can develop where groundwater discharge from both the bog and the adjacent mineral terrain converges.

Where groundwater discharge is relatively high, laggs are hydrologically, biogeochemically, and ecologically distinct (Figure 6). Where there is lower discharge, or where discharge from bogs continues landward (toward the mineral terrain), the laggs are less defined or even absent (Langlois et al., 2015, 2017; Howie and Tromp-van Meerveld, 2011). Laggs collect water and convey it laterally (i.e., into or out of the page in Figure 6), and thus can be an important feature that enhances peatland connectivity to surface water systems. The absence of the modulating effect of groundwater input to bogs (notwithstanding potential regional effects as noted earlier) typically results in a deeper and more variable water table than in fens; it is by nature more episodic, and commonly reaches 35 to 50 cm bgs (Figure 3). It is important to record the datum used for measurement of the water table when there is distinct microtopography that varies with time, such as in bogs. The water table depth is much greater beneath the hummocks than the hollows.

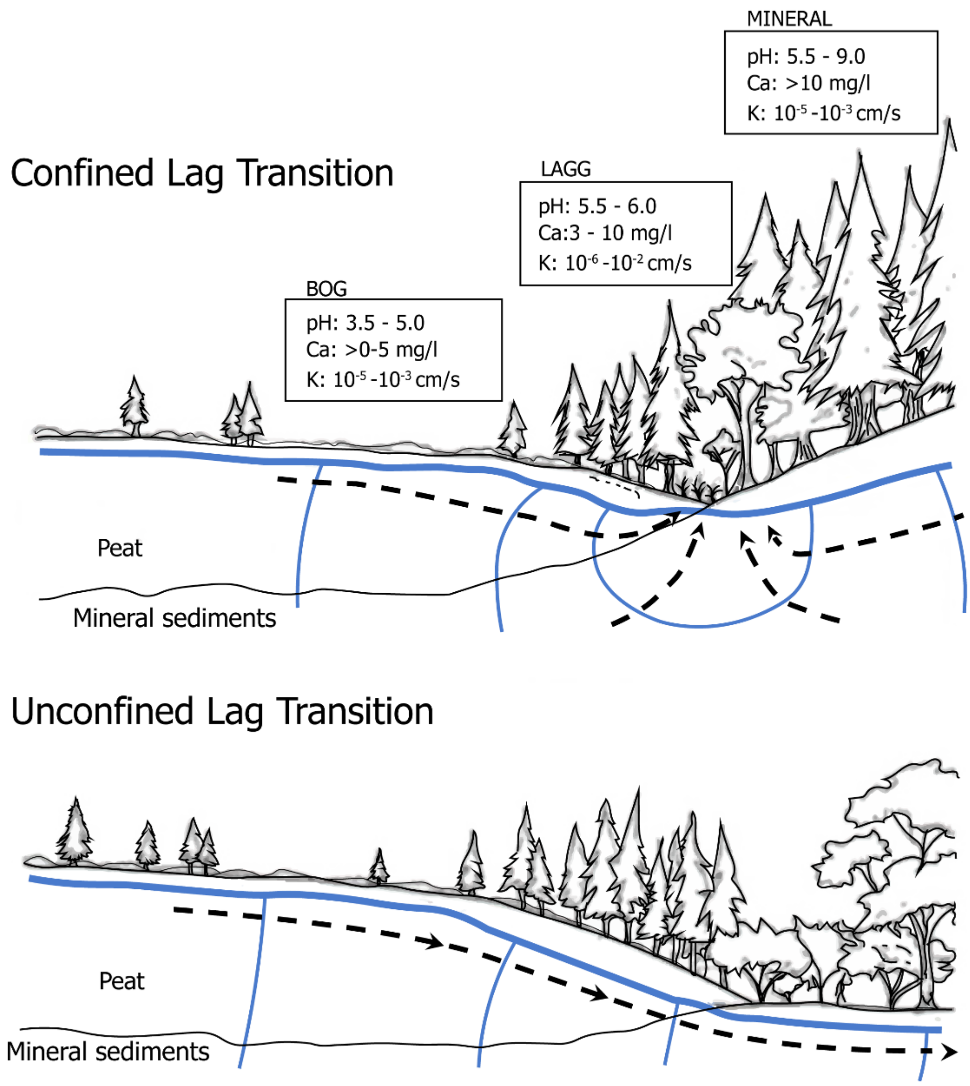


Figure 6 - Conceptual diagram of laggs, which occur at the margin of bogs. The thick blue line is the water table, while thin blue lines represent isopotentials. In the upper diagram, groundwater flow (black dashed lines) from the bog and adjacent mineral terrain converge to form a distinct lagg (confined lagg). In the lower diagram, where groundwater flow is unidirectional, a less distinct feature forms (unconfined lagg). In both cases, the groundwater chemistry is a mix of water from organic and mineral terrains (Modified from Howie and Tromp van Meerveld, 2011).

The hydrological function of bogs in the broader landscape is to store and periodically shed water. Peat can contribute groundwater to underlying mineral deposits, but this is generally minor due to the low vertical hydraulic conductivity associated with most basal peats and underlying mineral sediments. Bogs are unlikely to exist in a setting with high recharge or steep slopes, because excessive water loss would result in conditions too dry for a bog to form. However, in certain locations, including those with hyper-maritime climates with persistent rainfall and cool, foggy conditions, *blanket bogs* can form on undulating ground with slopes up to 15° (Chico et al., 2020). Blanket bogs are most common in northern latitudes, including northern parts of Europe and North America, but are also present in Patagonia, New Zealand, and the Falkland Islands.

2.5 Peatland Complexes

Over the long term, peat accumulation and the associated rise in water table elevation saturates adjacent areas so the peatlands can expand laterally, depending on local topographic gradients. Where climate is suitable, and Quaternary sediments are relatively flat and of sufficiently low permeability (Glaser et al., 2006), *peatland complexes* such as those in the Hudson Bay Lowland and Glacial Lake Agassiz area can develop. Peatland complexes like these host an assortment of bogs, fens, and swamps that dominate the landcover, coalescing as the peatlands develop. Over time, peat accumulation and system development alter the class of peatland and patterns of connectivity. Sometimes, flows from large bogs self-organize and drain through narrow fen systems (Figure 7), which often form into patterns of ridges and flarks (e.g., ladder fen, ribbed fen; NWWG, 1997). While bogs can also generate ribs and pools, the less distinct hydraulic gradients result in a disorganized pattern of pools/ponds on the top of the dome (Price, 1994). This is visible on the domed bog in Figure 7.

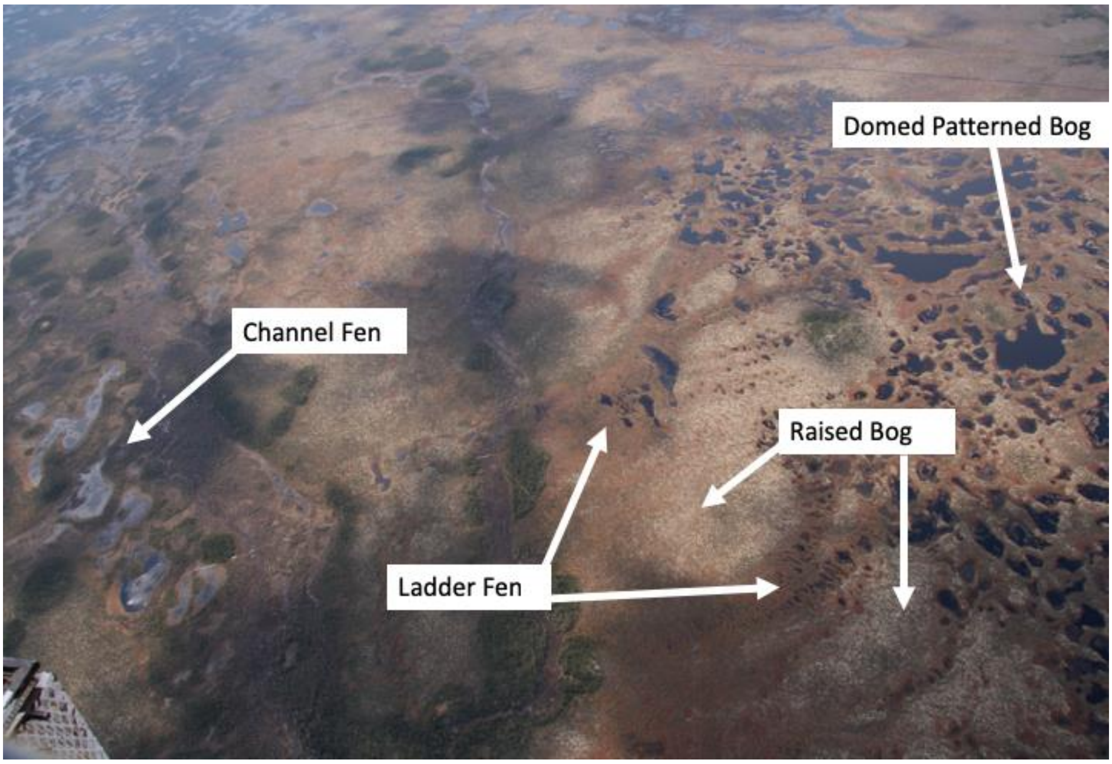


Figure 7 - Peatland complexes are assemblages of individual peatlands, where local hydrological and topographical gradients drive the exchange of water and nutrients among them. Here, a peatland complex in the James Bay Lowland has a large domed bog occupying the highest local elevation. It drains through ladder fens on its flanks and eventually to a large channel fen that comprises part of the regional flow system. Smaller raised bogs occur in interfluvial locations, including between adjacent ladder fens. (Photograph by J. Price)

2.6 Regional Processes

Peatland occurrence is related to latitudinal gradients of temperature and precipitation and zonal gradients of continentality (Rochefort et al., 2012), which control vegetation productivity and decay (thus peat formation) through energy and moisture availability. Bedrock geology and lithology also affect the regional distribution and character of peatlands and their groundwater relations. Regions like the Hudson Bay Lowlands have extensive peatlands (Figure 1) and the peatland complex (Figure 7) due to low permeability sediments in a favorable climate. Precambrian bedrock such as the Canadian Shield in eastern North America are also of low permeability, but irregular topography and thin soils result in isolated peatlands, more commonly bogs and poor fens, since they are poorly connected to groundwater.

Given that precipitation generally decreases poleward and with continentality, and potential evapotranspiration increases toward lower latitudes, bogs have a northern and southern limit in the Precambrian region of eastern North America (Damman, 1979). Regions with a higher seasonal water deficit, such as the more continental Western Boreal Plain, tend to have more fens than bogs (Vitt and Chee, 1990), given that groundwater exchanges in the deep sediments can augment the water budget. Oceanic climates are favorable for peatland formation, especially bogs, since they are cooler and wetter, whereas temperate climates are amenable to peatland formation, often swamps. However, peatlands in temperate climates are commonly more reliant on geomorphic controls that facilitate groundwater inputs, given their higher evapotranspiration losses.

Boreal and subarctic wetlands experience seasonal differences in their groundwater relations owing to the development of seasonal frost. Since the water table is always close to the ground surface, it freezes early in the fall season as air and shallow ground temperatures fall below 0° C. As the freezing front moves downward through the saturated peat profile, a relatively impermeable ice layer forms across the peatland. Such an ice layer can persist for over half the year during which time the peatland surface is hydrologically decoupled from the underlying groundwater. At higher latitudes, seasonal ice occurs over permafrost, ground that remains below 0 °C for at least two consecutive years. Permafrost presents a permanent (i.e., year-round) barrier separating peatlands from sub-permafrost groundwater systems and strongly influences peatland landscape development. Box 2↴ provides more detailed discussion of permafrost peatlands.

Tropical peatlands share many of the fundamental hydrological processes with peatlands of other regions, albeit without the cold-climate features described above. Given the high potential evapotranspiration in tropical latitudes, tropical peat swamps develop where rainfall is high. In subtropical areas, where rainfall is often much less but potential evapotranspiration is still high, swamps are less common, but can occur where groundwater input offsets the shortfall of precipitation. Box 3↴ provides additional information about tropical peatland hydrology.

3 Peat: A Porous Medium

In peatlands, peat is the aquifer material that governs the flow of water and thus its connectivity with local and regional groundwater aquifers. Water flow through peat is affected by the properties of the matrix, thus by its botanical source and the combined impact of decomposition and compaction, which vary with depth. Peat is a porous medium whose hydraulic character can be described with terms similar to those used for mineral sediments. However, the physical and chemical properties of peat that affect flow and transport are distinctly different from those in mineral soils (Table 1). Peat is a porous matrix with exceptionally high total porosity (up to 98 percent), only a portion of which conducts water flow.

Table 1 - Examples of properties of peat compared with mineral soils (Modified from Gharedaghloo and Price, 2017). Citations for values are available in Gharedaghloo and Price (2017).

Soil Property	**Peat**	**Mineral**
Porosity	86 – 98%	Sand 21 – 49% Clay 14 – 69%
Mobile porosity, ϕ_{mob} (total porosity, ϕ_t)	*Sphagnum* moss 53% (90%) Lightly decomposed 37% (87%) Humified 12% (75%)	Sand 23% (32%) Clay 8% (43%)
K_{sat} (m/d)	*Sphagnum* 10 – 10,000 Well decomposed sedge 0.001	Sand 0.004 – 40 Clay 5 x 10^{-12} – 0.5
Composition	Organics, lignin, cellulose, humic and fluvic acids, lipids, waxes, resin, bitumen	Minerals including quartz, feldspar, kaolinite, etc.
Wettability	Conditionally hydrophilic and hydrophobic	Generally hydrophilic
Pore structure	Dual porosity with interconnected macropores and immobile porosity	Single porosity (unless fractured)
Pore-size distribution	Bimodal	Unimodal

Peat consists of organic materials that begin to decompose at the time of their deposition. Hence, the organic materials deposited initially at the inception of the peatland eventually form the basal peat layer. Peat physical properties vary with time since deposition due to ongoing decomposition and compression typically results in more decomposed and consolidated peat with depth (Figure 8); however, less decomposed layers may occur if, for example, they were deposited during cooler and wetter periods that reduced decomposition.

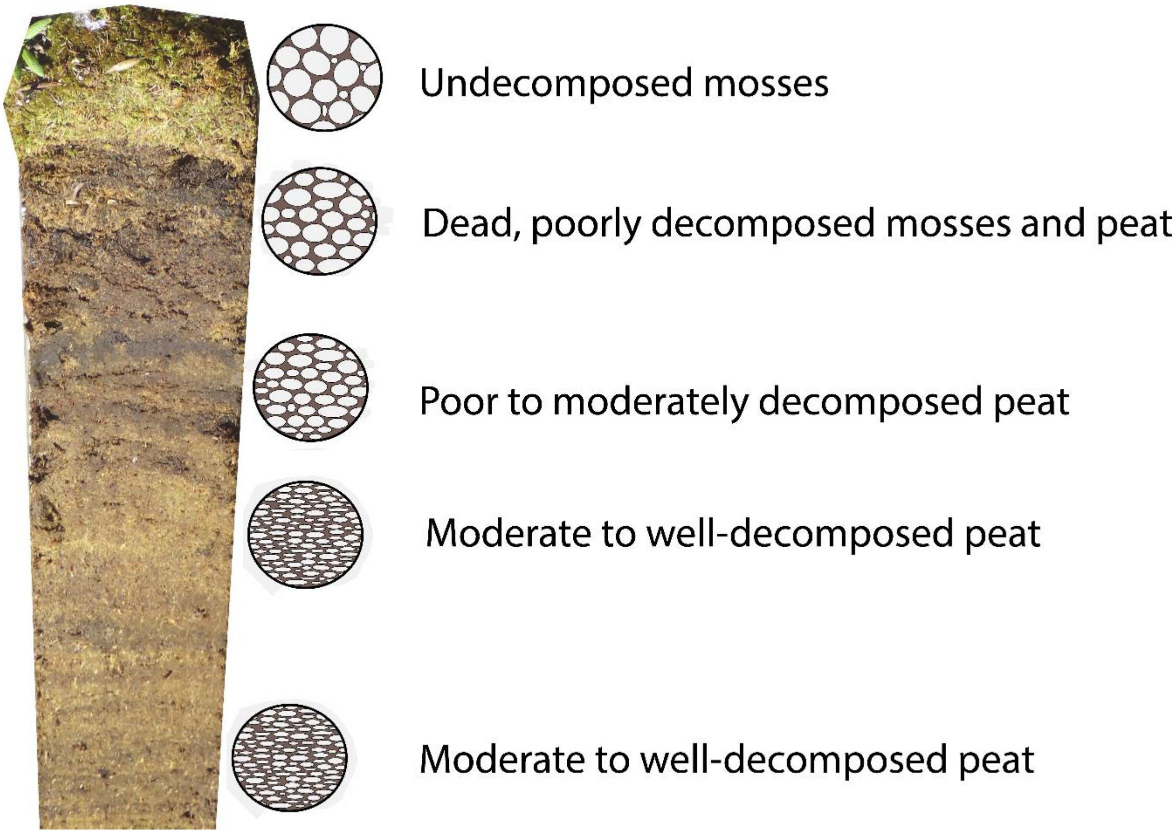


Figure 8 - Vertical profile of the upper ~60 cm of bog peat, dominated by *Sphagnum* mosses. In the accompanying graphic, organic matter is represented with black and gray, while pore space is white. The living moss occupies the topmost layer characterized by large open pores, resting on the remains of the dead, poorly decomposed mosses immediately below. At depth, the structure of the matrix is noticeably different, wherein the peat is more decomposed and partially compressed by the weight of the overlying material, reducing the pore-size and causing pore shapes to flatten. In this profile, indicated by "moderately decomposed peat," a distinct banding can be seen, possibly reflecting periods of higher and lower decomposition associated with warmer/drier and cooler/wetter periods of climate. In the lower section of the peat profile, moderate to well decomposed peat is indicated. In this particular profile, plant fragments are still visible at depth, so decomposition is moderate rather than well decomposed. Photograph of core is from Ahad and others, 2020, with modified perspective.

For peat to form, the average annual rate of plant matter accumulation must exceed the average annual rate of its decay. Vegetative matter is added at and near the ground surface, and decomposition occurs throughout the peat profile. In tropical forests, peat formation in tip-up pools formed by uprooted trees — which are often more than a meter deep — is an important zone of carbon accumulation (Domain et al., 2015). The upper layer is variably saturated and comprises relatively easily decomposed carbon compounds; decomposition by aerobic microbial processes in this zone is relatively rapid. Deeper peat is perpetually saturated (notwithstanding some biogenic gases that may be present), relatively resistant to decay, and thus decomposition by anaerobic bacteria occurs very slowly. At higher latitudes and/or altitudes, low ground temperatures also reduce the rate of peat decomposition.

The upper layer of the peat profile is sometimes called the *acrotelm*, whose depth is approximated by the mean annual maximum water table depth. Defining this operationally is difficult, but conceptually it is the zone of variable saturation. Below the acrotelm, where the peat is perpetually saturated, is the *catotelm* (Ingram, 1978). This relatively simplistic model of peatlands is useful for identifying distinctly different layers with contrasting hydrological, geochemical, and ecological processes (Rezanezhad et al., 2016). However, others have challenged it as overly simplistic, and recognize the importance of spatially distinct hotspots as being critical to peatland function (Morris et al., 2011), especially since many peatlands are characterized by various forms of microtopographic relief (e.g., hummock/hollows, ridges/depressions) that have distinctly different hydraulic and biogeochemical characteristics (Baird et al., 2016).

For the purpose of discussing groundwater processes, living mosses as well as dead but poorly decomposed and well decomposed plant material (peat) are all part of the matrix through which water and solute flow occurs. Since the physical structure of peat degrades as plant matter decomposes and pore spaces collapse, its hydraulic properties change accordingly. This includes a decrease of hydraulic conductivity, porosity, and drainable porosity with depth (Rezanezhad et al., 2016). For some peatlands— notably bogs and to a lesser extent, fens — the hydraulic conductivity can decrease by 4 to 5 orders of magnitude between the top and bottom of the acrotelm (approximately the upper 30 to 50 cm of peat), whereas within the catotelm, the hydraulic conductivity is generally lower than the minimum in the acrotelm and has less depth-dependent variability (McCarter and Price, 2017a). However, there are exceptions to this trend (Chason and Siegel, 1986). While the concept of acrotelm technically applies to swamps, the relatively amorphous degraded peat in many swamps does not commonly have distinct vertical patterns of variability of hydraulic conductivity, porosity, or drainable porosity.

Similarly, in disturbed peatlands where the acrotelm layer has been removed or highly degraded, the concept of a distinct acrotelm does not apply, or at least, is not useful. In most bogs and fens, however, the change in hydraulic character with depth both depends on, and controls, the hydrology of the system. This creates a critical ecohydrological feedback that regulates peatland form and function (Figure 9). Waddington and others (2015) provide a good discussion of hydrological feedbacks.

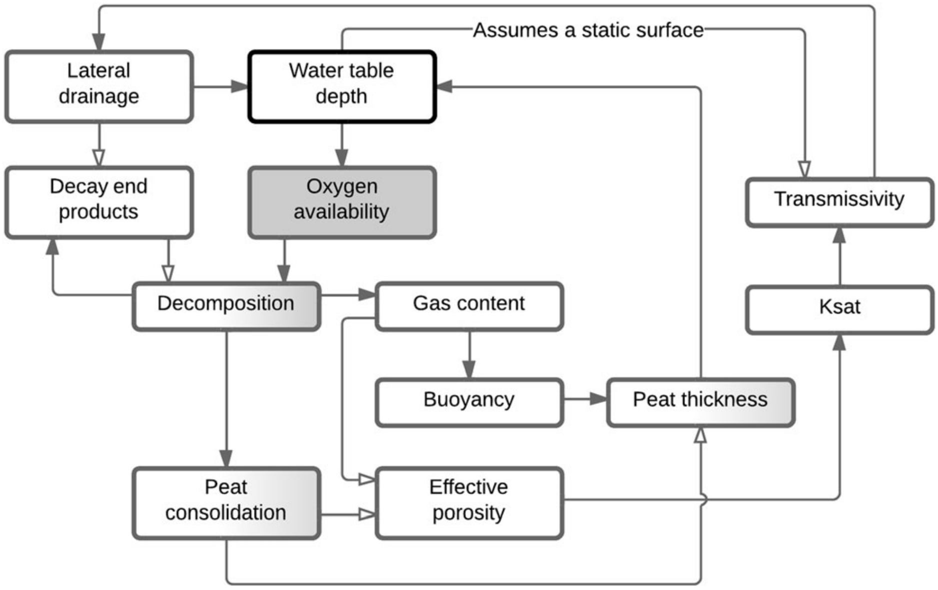


Figure 9 - The water table depth-peat decomposition feedback. Solid and outline arrows represent positive and negative relationships, respectively. Gray shading indicates a process in the unsaturated zone, white is the saturated zone and the white-to-gray transition implies both the saturated and unsaturated zone. Water table depth is a function of external water exchanges, for a given peat thickness. Peat thickness is controlled by organic matter input, subject to decomposition and consolidation, which affects peat structure including its porosity, hydraulic conductivity, and transmissivity. (From Waddington et al., 2015)

3.1 Saturated Zone Properties and Processes

The vertical profile of hydraulic properties can have a profound effect on the hydrology of peatlands. Bogs and fens shed water efficiently to adjacent ecosystems because of a water table depth—transmissivity feedback mechanism (Waddington et al., 2015). Transmissivity ($L^2 T^{-1}$), the product of the hydraulic conductivity and thickness of the flow layer considered, increases exponentially as the water table rises into the near surface layers where hydraulic conductivity is orders of magnitude higher (McCarter and Price, 2017a).

Conversely, when the water table is deep, the transmissivity is low; under such conditions, the saturated layer is thinner and occupies only the deeper, more decomposed peat layers where the hydraulic conductivity can be much lower. Exceptions may occur where macropores or more fibric layers allow preferential flow at depth. Extremely high transmissivities during periods of high water table enable many bogs to shed water by groundwater flow alone, without flow self-organizing into rivulets or streams. Although the effect of this mechanism can be dampened in the presence of preferential flow paths; this mechanism, governed by the pore-scale structures, can control water flow at the peatland scale (McCarter et al., 2020).

Peat is generally considered to be a dual porosity medium (Table 1). It includes relatively large pores between particles such as the leaves and branches of vegetative matter, as well as relatively small pores associated with more decomposed fragments and closed and dead-end spaces within the remains of plant cellular material (Figure 10).

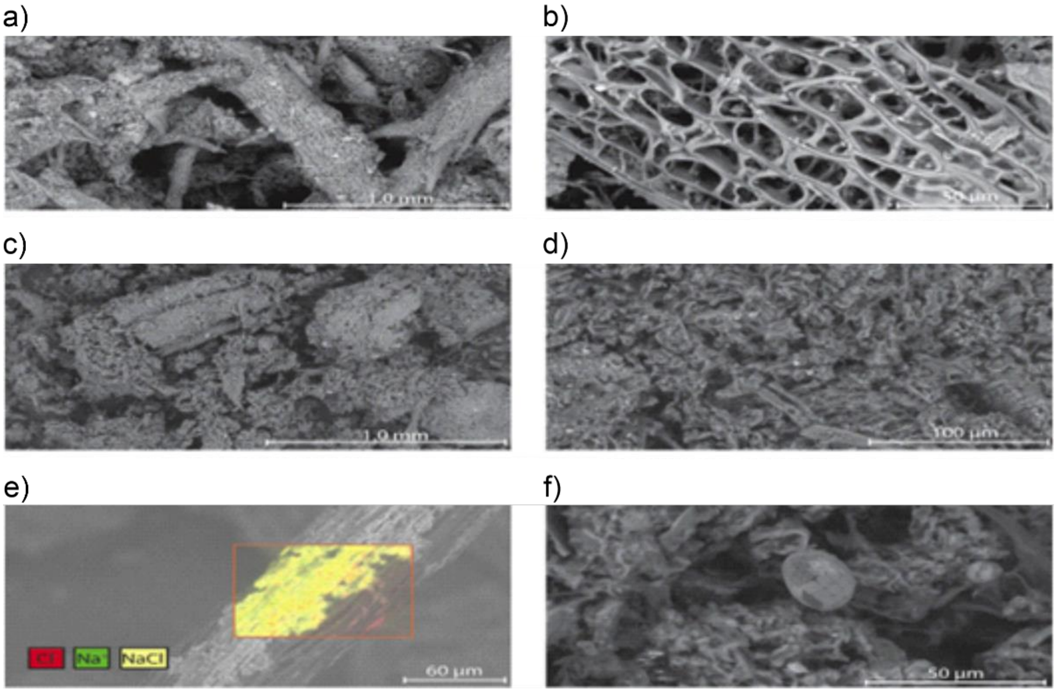


Figure 10 - Scanning Electron Microscope (SEM) images of different peat (Photograph modified from McCarter et al., 2020).

a) Undecomposed *Sphagnum* peat where the leaf and branch pore structure create an abundance of large pores.
b) Desiccated hyaline cells that comprise the majority of the immobile porosity of undecomposed *Sphagnum* — here, the cell walls are mostly absent, possibly degraded by sample desiccation for SEM analysis.
c) Decomposed *Sphagnum* peat (~30 cm bgs), with a greater proportion of smaller pores than undecomposed *Sphagnum* peat.
d) A magnification of the decomposed *Sphagnum* peat, highlighting the reduction in large pores.
e) Adsorbed Cl^-, Na^+, and precipitated NaCl on peat. Conditions in the SEM preclude any liquid water in the sample. The individual red and green spots suggest adsorption of the individual ions. The NaCl likely combines both adsorbed Na^+ and precipitated NaCl during imaging.
f) A testate amoeba within peat, highlighting size exclusion from the smaller pores. Samples were completely desiccated during the imaging process.

In some places, the peat has larger preferential flow paths caused by roots, expansion and contraction, and porosity caused by gas emission. These can produce localized areas of high permeability at depth where the permeabilities are otherwise very low. These and other larger pores can transmit water, while the liquid within smaller pores of saturated peat is mostly immobile. Thus, the total porosity (ϕ_t) of peat is the sum of the mobile porosity (ϕ_{mob}), and immobile porosity (ϕ_{im}) as shown in Figure 11. This dual porosity strongly affects the flow and storage of water and solutes. Porosity is a unitless quantity, being volume of water divided by volume of sample.

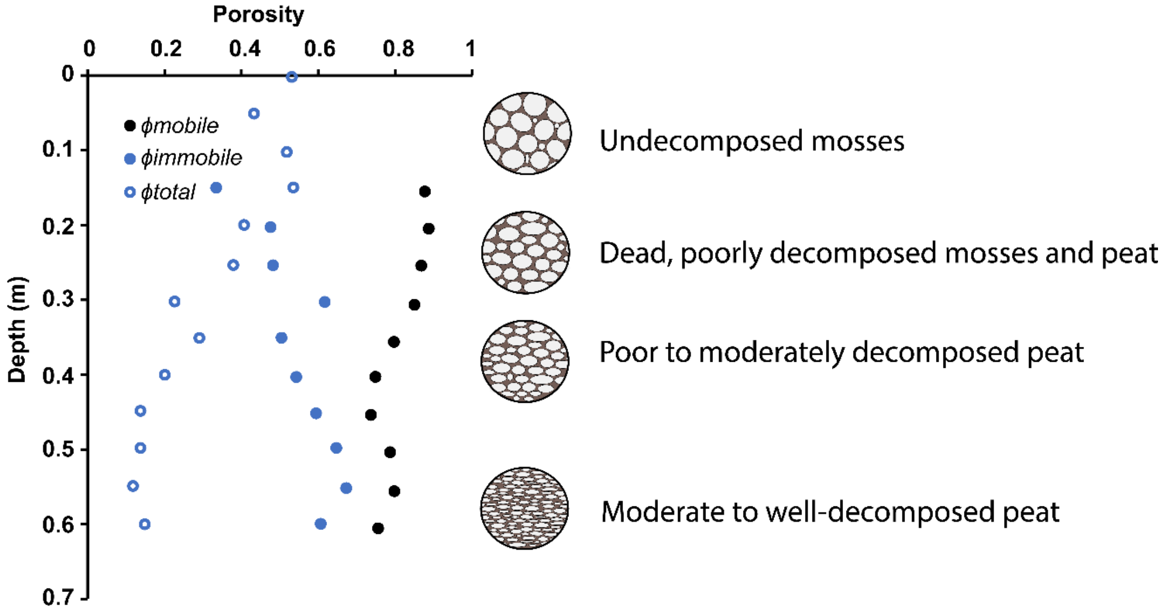


Figure 11 - Change in mobile water porosity (ϕ_{mob}), immobile water porosity (ϕ_{im}), and total porosity (ϕ_t) with depth below surface in a *Sphagnum*-dominated blanket bog peat (Modified from Hoag and Price, 1997). The graphics representing the state of peat decomposition are further described in Figure 8.

Poorly decomposed peat has a high proportion of mobile porosity (upper layer in Figure 11; Table 1), and flow occurs relatively easily (mobile water). Decomposition of peat reduces the fraction of larger pores, so in many peatlands ϕ_{mob} decreases with depth, while ϕ_{im} increases (Figure 11). The predominance of larger pores near the ground surface is responsible for the high saturated hydraulic conductivity in that part of the peat profile.

Deeper in the profile, the water table can rise rapidly in response to precipitation input (Figure 12) because much of the peat remains undrained, even at strong suctions. The proportion of porewater that can be drained gravitationally (ϕ_d) is < 0.05 at depth (Figure 13). As such, even during dry conditions, the highly conductive peat near the ground surface can be quickly re-activated into the runoff process. Water tables near the surface, where drainable porosity is high, exhibit a damped response (i.e., near-surface peat, with high drainable porosity, can gain or lose relatively large volumes of water for a given water table change).

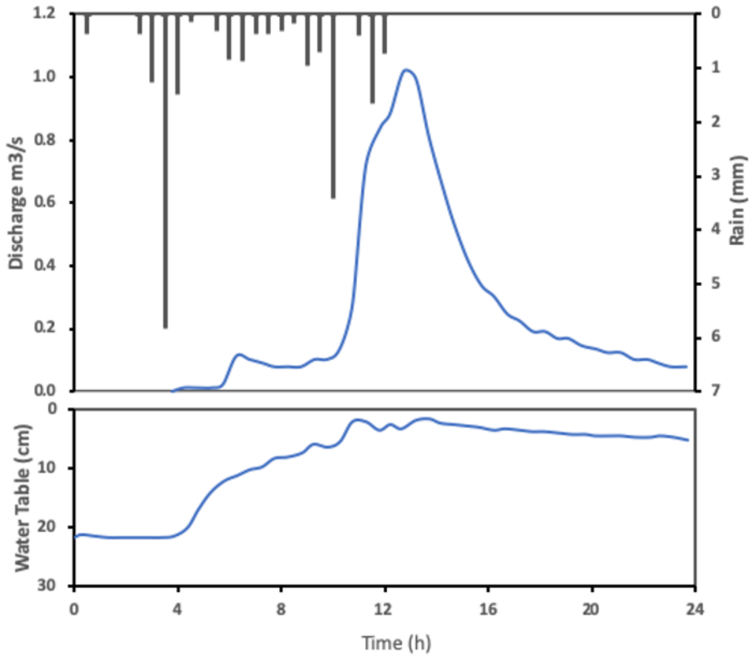


Figure 12 - Rainfall (bars) runoff response (upper diagram) and water table (lower diagram). Initially, there was no runoff while the water table was low. The largest rain event generated no runoff but caused the water table to rise — quickly at first (where drainable porosity is low), then less quickly nearer the surface (where drainable porosity is higher). At this location (Trout Beck, Cumbria, United Kingdom; 6 July, 1995), runoff was not initiated until the water table was within 10 cm of the surface, because of the large increase in peat transmissivity (Redrawn from Holden, 2006).

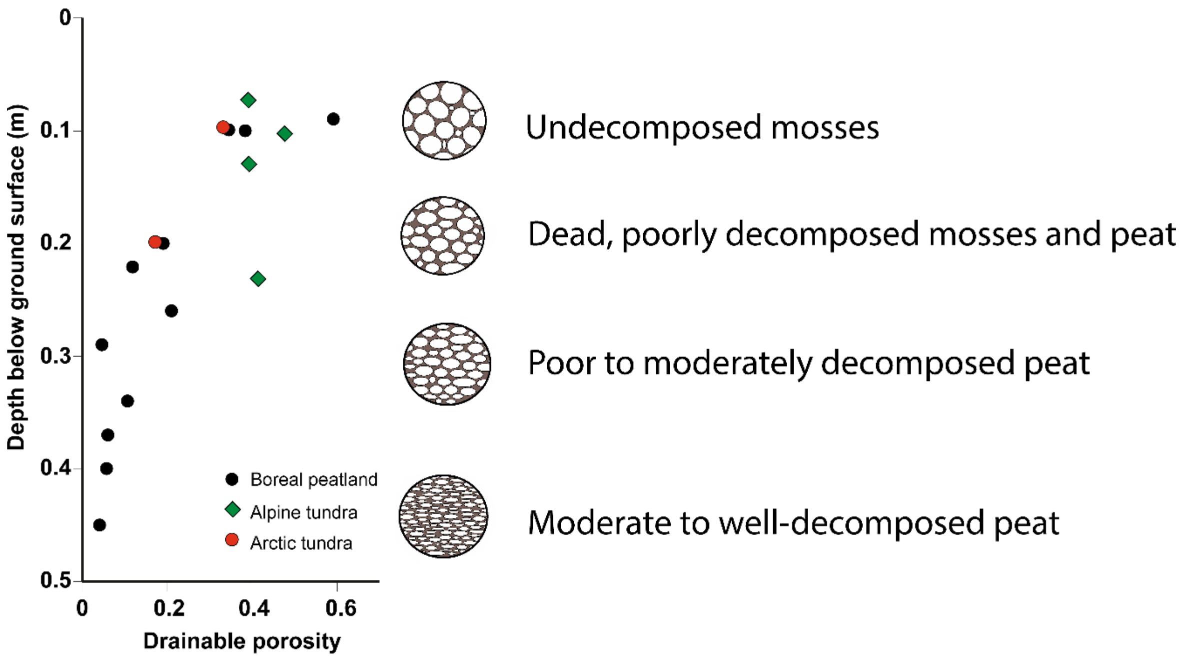


Figure 13 - Variation of drainable porosity (ϕ_d) with depth below the ground surface based on measurements made on discrete peat samples representing a range of depths (modified from Quinton et al., 2008). Peat accumulations are typically less than 30 cm at the two tundra sites presented here. For this reason, the number of observations presented for these sites are relatively few. By contrast, peat accumulations typically exceed 2 m at the boreal peatland site, although data are presented only for the upper 50 cm.

The water storage relations of some peatlands are confounded by the high compressibility of peat, which adjusts its volume depending on changes in water pressure based on water table position. This results in a process named mire breathing, in which the ground surface rises and falls with changes in water table depth. Sometimes this is referred to as peat-surface-oscillation (Fritz et al., 2008). This process is driven by changes in effective stress, σ_e (Equation 1), which is the load due to the weight of peat and water expressed as total stress, σ_T (Equation 2), above a given point in the peat profile offset by the buoyancy caused by the height of the column of water above that point, namely the water pressure (ψ).

$$\sigma_e = \sigma_T - \psi \tag{1}$$

$$\sigma_T = \rho_T g h \tag{2}$$

where:

σ_e = effective stress ($ML^{-1}T^{-2}$)

σ_T = total stress ($ML^{-1}T^{-2}$)

ψ = water pressure ($ML^{-1}T^{-2}$)

ρ_T = average density of the overlying peat and water (ML^{-3})

Thus, effective stress is highly sensitive to changes in water table elevation, such that the surface elevation can decline many centimeters in one day (Roulet, 1991; Whittington and Price, 2006). Given the high compressibility of peat compared to mineral matrixes, water table fluctuations result in volume changes. The implications of this can be important hydrologically, ecologically, and biogeochemically, since a drop in the water table elevation induces a drop in the ground surface elevation, so the depth to the water table below the ground is less than it would otherwise be. Hydrologically, this means saturated water flow can be sustained in the upper layer of the peat deposit, and that the nature of water storage relations is affected as discussed in the next paragraph. Having a relatively high and consistent water table affects the plant community type as well as the decomposition processes, thus the carbon exchanges.

Peatlands most susceptible to mire breathing tend to be those containing poorly decomposed peat with high water content. Mire breathing is exacerbated by large water table elevation changes. While the relationship of the water table elevation to water storage in unconfined aquifers is usually governed by specific yield (S_y) as discussed in Section 8.3, water storage changes associated with changes in aquifer volume can be important in peat. This is akin to the specific storage parameter (S_s) that is significant for confined aquifers and negligible for unconfined aquifers. For example, Price and Schlotzhauer (1999) showed that for a 1.8 m cutover peat deposit (cutover means that some of the peat mass has been harvested), an ~50 cm decline of the water table elevation caused the surface elevation to decline about 10 cm. The water storage change due to compression in the saturated portion of the peat deposit was greater than would be associated with water table lowering alone (i.e., pore water drainage volume calculated as the product of water level change and specific yield). In that system, the storage change (ΔS) needs to be calculated using Equation 3 which requires both S_y and S_s.

$$\Delta S = \Delta WT * (S_y + bS_s) \tag{3}$$

where:

ΔS = storage change (L^3/L^2)

ΔWT = change in water table elevation (L)

S_y = specific yield (-)

b = saturated thickness of the peat deposit prior to water table decline (L)

S_s = specific storage (L^{-1})

Estimating the water budget using only S_y (~0.05) without bS_s (~0.08) resulted in an error of more than 100%. At the same site, Price (2003) recorded a two-order of magnitude decrease in hydraulic conductivity 0.75 m below the surface as a consequence of seasonal subsidence. The hydraulic conductivity in a nearby undisturbed site, where water table fluctuations were much smaller, was reduced by about 50 percent.

A widely used parameter to quantify the hydraulic character of saturated peat is the hydraulic conductivity (K_{sat}). The hydraulic conductivity of saturated peat is related to the size and connectivity of pores, such that it increases proportionately with the square of the mean pore throat diameter; it is also affected by pore volume, distribution, and shape; all of which are dependent on the degree of decomposition and compaction. K_{sat} of a peat matrix is related to the hydraulic radius of pores (cross-sectional area/perimeter). Thus, with increased depth (hence, typically increased decomposition) the median hydraulic radius decreases as a result of the reduction in pore throat diameters, and perhaps also as a result of the flattening (due to compression) of pore shapes. Both of these changes increase hydraulic resistance, so tend to reduce hydraulic conductivity with depth.

While the distinct decrease in K_{sat} with depth has been reported by many researchers, others have reported higher than expected values at depth, possibly due to macropores associated with woody inclusions (Chason and Siegel, 1986) or layers of less decomposed peat that probably reflect periods of wetter/cooler climate during development.

A wide range of techniques have been used to measure K_{sat} in the field — including the use of tracers and water level recovery tests — and in the laboratory — including the use of permeameters and image analyses of samples as discussed in Section 8, *Methods and*

Approaches. The larger scale approaches to estimating K_{sat}, such as aquifer tests, can produce higher estimates because they are more likely to include macropores (Glaser et al., 2021).

Although absolute values of K_{sat} can vary widely for different peat types, hydrogeomorphic settings, and even within a single peatland, typical profiles of K_{sat} demonstrate common profile characteristics including the presence of a relatively thin (~0.1 m) top layer of living vegetation and lightly decomposed peat where average K_{sat} values are high (e.g., 10 – 1000 m d^{-1}) such as shown in Figure 14. Below this is a thicker peat layer in a more advanced state of decomposition where K_{sat} values and variability are typically lower (~0.5 – 5 m d^{-1}). Between these two layers can lie a transition zone where the rate of reduction in K_{sat} with increasing depth is large (Figure 14). While high K_{sat} of acrotelm peat is critical to the connectivity between adjacent peatlands or to the adjacent aquatic system, the typically lower K_{sat} of catotelm peat can modulate the water and solute exchanges between deeper groundwater and the peatland, although this is more likely to be controlled by low permeability underlying mineral sediments (Reeve et al., 2000).

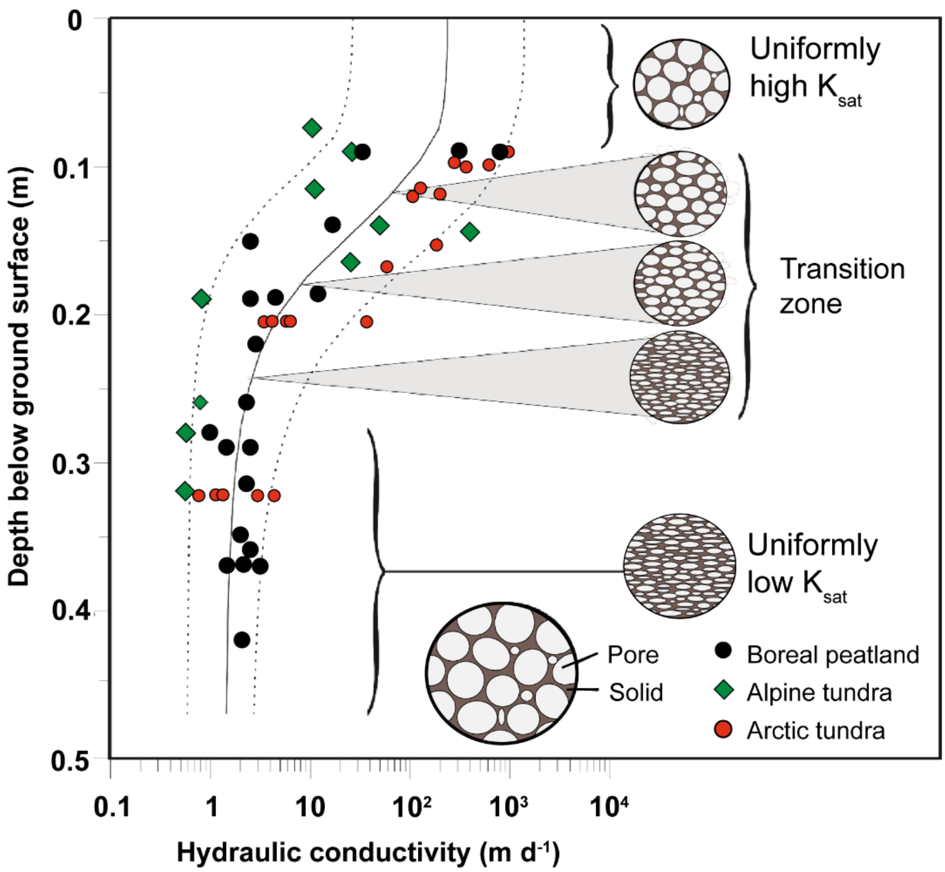


Figure 14 - Saturated hydraulic conductivity of peat overlying permafrost from boreal peatlands (Quinton et al., 2008), alpine tundra (Quinton et al., 2004), and arctic tundra (Quinton et al., 2000). The solid line indicates the best-fit of K_{sat} versus depth as determined by the least-squares method; the dashed lines indicate upper and lower envelopes of data points for the boreal peatland. Tundra values of K_{sat} occupy a similar range. The reduction in K_{sat} with increasing depth is largely accounted for by the decrease in the average pore diameter and pore hydraulic radius with depth (Modified from Quinton et al., 2008).

Peat hydraulic conductivity in a given setting can be temporarily reduced by peat consolidation caused by seasonal water table lowering (Price, 2003); it can also be underestimated due to a reduction in pore water pressure that causes peat structure to collapse during a bail test used to estimate hydraulic conductivity. Furthermore, Beckwith and Baird (2001) showed that biogenic gases, specifically methane, released from moderately to well decomposed peat can reduce the apparent hydraulic conductivity by blocking pores, although Glaser and others (2021) suggest it may cause pore dilation. This effect is less important in the upper layers of peat where the pore structure is larger (Kettridge et al., 2013). Rosenberry and others (2006) found gas volumes in peat from 2 to 20 percent, the pressure from which can influence hydraulic gradients (Kellner et al., 2004) such that increases in pressure at depth can cause flow reversals, or even an outward radial flow (Waddington and Roulet, 1997).

Finally, hydraulic conductivity is affected by water viscosity and there are strong temperature gradients in the near surface groundwater in peatlands that are large enough to cause substantial differences in viscosity, so care must be taken when comparing, or using, such values. More details on evaluating hydraulic conductivity are provided in Section 8.3, *Hydraulic Conductivity*.

3.2 Unsaturated Zone Properties and Processes

In the acrotelm, the rapid change in pore size and shape with depth enables important ecohydrological feedbacks that sustain critical biological processes and peat accumulation. Such feedback depends on the relationship between soil water retention and unsaturated hydraulic conductivity (K_{unsat}) within the peat, a relationship governed by the pore structures discussed above and statistically related to bulk density (Livett and Lennartz, 2019). As peat bulk density increases, so too does the ability of the peat to retain water. Higher water retention capacity increases the value of K_{unsat} at the same soil water pressure and, therefore, the ability of peat to conduct water. Rezanezhad and others (2010) found that for a soil water pressure of $\psi = -40$ cm, the large pores and much of the continuously connected water phase of the initially saturated *Sphagnum* peat samples had drained, and as a result could no longer conduct sufficient water to avoid ecohydrological stress.

Computed Tomography (CT), an imaging technique used in the medical field to generate a detailed picture of internal structures, provides insight into the water pathways in peat. It revealed that in addition to drainage of larger pores, smaller pores also drained, leaving a more irregular shape and smaller hydraulic radius of the remaining connected water pathways. This reduced the value of K_{unsat} and therefore constrained water movement. A visualization of air-filled pores and their diminishing size with depth is shown in Figure 15.

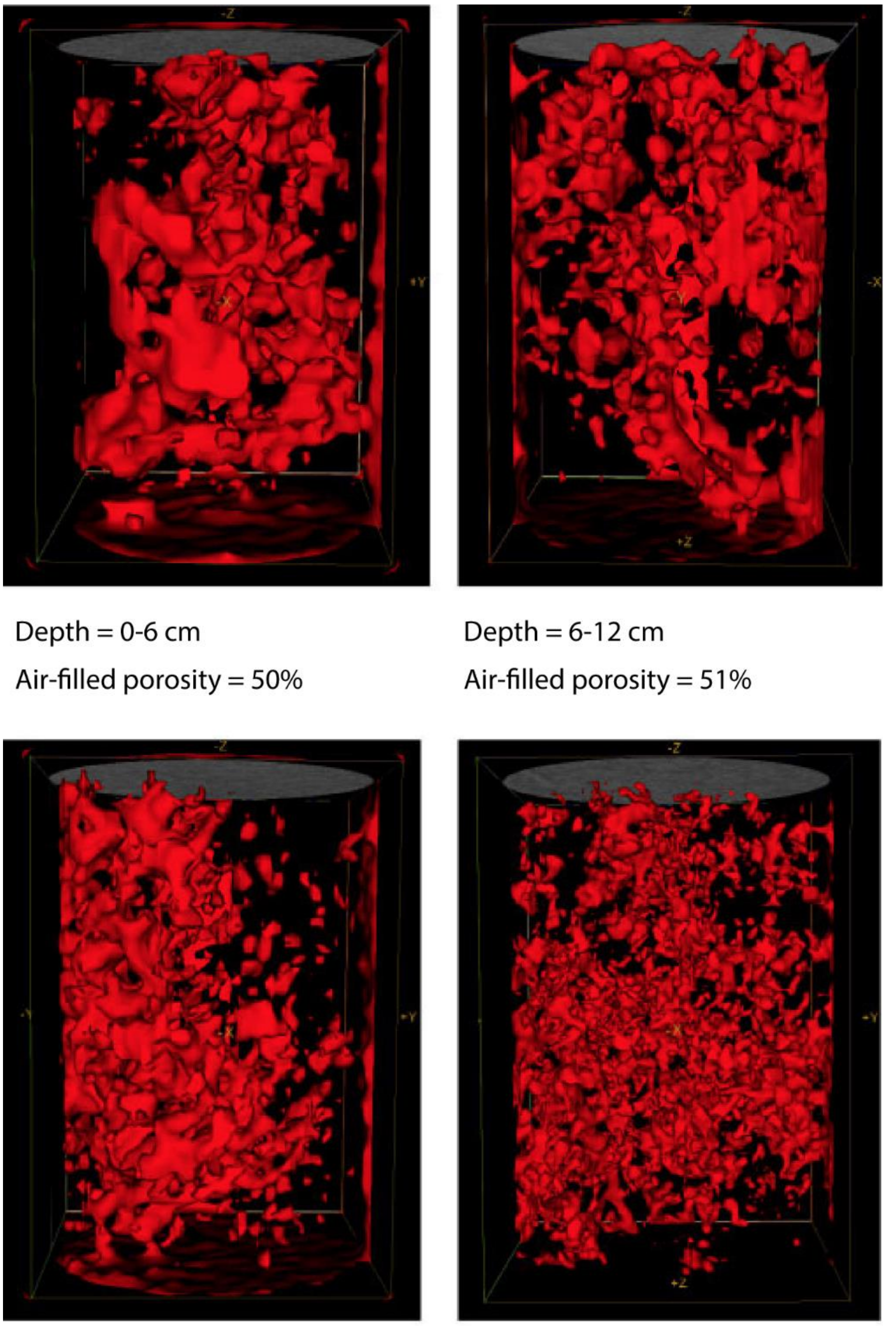


Figure 15 - Visualization of air-filled porosity (in red) of peat samples at increasing depth, all at ψ = -40 cm. In each 6-cm diameter core the total number of air-filled pores increased with depth from 7,956 to 13,267 to 15,041 and finally to 39,812 (Modified from Rezanezhad et al., 2010).

The three-dimensional images of Rezanezhad and others (2010) were used by Gharedaghloo and others (2018) to develop a pore network model that simulated the movement of water through individual pores and pore networks. These simulations showed that while hydraulic conductivity is isotropic (i.e., independent of the direction of measurement) at small scales, it becomes anisotropic (i.e., dependent on the direction of measurement) at larger scales due to the strong heterogeneity resulting from the layered structure of peat. This favors horizontal transport in the upper layers as discussed in Section 4.1, *Transport in Peat*.

At the near surface where *Sphagnum* peat is characterized by low bulk densities and high hydraulic conductivity, soil water content can decrease by 30 to 60 percent with a decrease of just 5 cm soil water pressure (Figure 16a). This dramatic decline in soil water content decreases K_{unsat} by at least an order of magnitude (Figure 16b), creating highly nonlinear soil water pressure—unsaturated hydraulic conductivity relationships. This creates conditions that decrease capillary rise of water to the surface of the moss, thus limiting evaporation. In peat with higher bulk densities, such as sedge fen peat, the decline in both soil water content and K_{unsat} is not as severe because water is preferentially retained in the smaller pores. Consequently, evaporation from peats of higher bulk density is rarely limited by capillary processes, and it is augmented by transpiration from vascular plants, which pull water upward in the soil profile. For a given negative pressure, peat from lower in the profile has higher water content than that from shallower layers because the smaller pores at depth remain continuously connected (Figure 16a). At saturation (i.e., greater than or equal to zero pressure) the shallow zone has the highest hydraulic conductivity (Figure 16b) because of its large pores. However, with a small decrease in pressure the largest pores drain, reducing pore connectivity thus hydraulic conductivity. Consequently, more decomposed peat at depth that retains water at decreasing pressures sustains a higher level of connectivity, thus hydraulic conductivity.

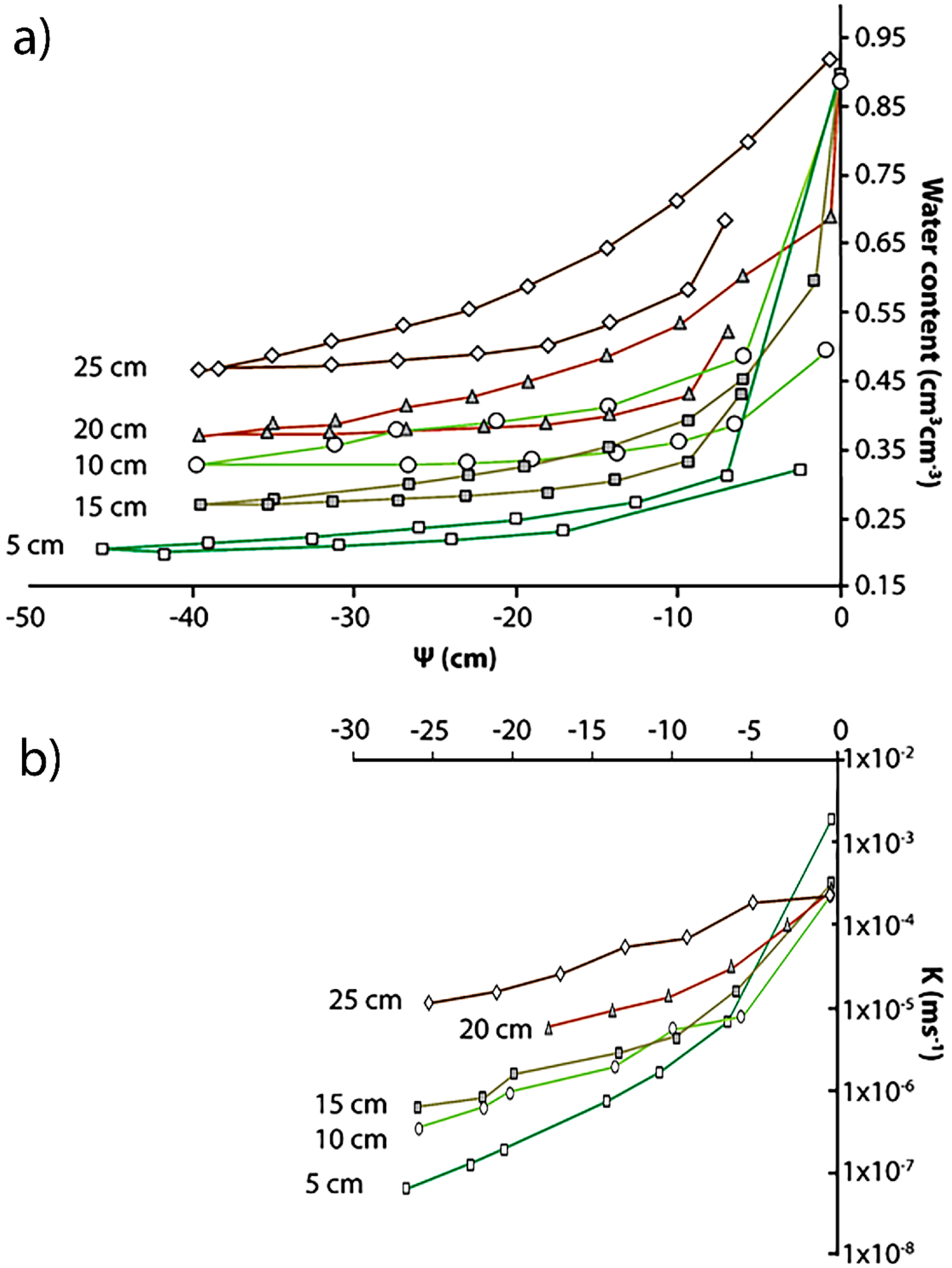


Figure 16 – a) Examples of soil water retention curves for drainage and rewetting of a bog peat. Hysteresis causes the rewetting curve to plot at lower water content for a given pressure. Soil water retention curves commonly have a shoulder in the pressure range near zero, representing sustained water content at slightly negative pressures; water content then declines substantially with further reduction in pressure as the matrix drains. In this example, the method was not able to test in this pressure range, so a shoulder is not visible. For a given negative pressure, deeper zones have higher water content than shallow zones because the pores are smaller thus maintain continuously connected zones of water. b) Hydraulic conductivity as a function of pressure for the same layers of peat. At saturation, i.e., zero pressure, the shallow zone has the highest hydraulic conductivity because of its large pores, but continuous zones of water decrease substantially in large pores following a small decrease in pressure. Consequently, more decomposed peat at depth that retains water at decreasing pressures, sustains a higher level of connectivity, thus hydraulic conductivity. Modified from Price and Whittington, 2010.

Hydraulic conductivity responds to soil freezing in a way similar to how it responds to soil drying since the water held in relatively large pores freezes readily, while that held in small pores (and therefore under greater tension) can remain unfrozen well below 0 °C (Kane and Stein, 1983). The hydraulic conductivity of both drying and freezing soils therefore decreases sharply as the water in large pores drains and/or freezes and the remaining liquid water is forced to flow through small pores and thin films (Watanabe and Flury, 2008). For this reason, the relationship between soil temperature and liquid water content, called the soil freezing characteristic, is analogous to the soil moisture characteristic of unfrozen soil (Miller, 1980).

The proportion of water in a variably saturated medium like peat can be described with the parameter *effective saturation*, S_e, as defined by Equation 4.

$$S_e = \frac{\theta - \theta_r}{\theta_s - \theta_r} \tag{4}$$

where:

S_e = effective saturation (-)

θ = volumetric soil water content (-)

θ_r = residual water content (-)

θ_s = volumetric water content at saturation (-)

The values of θ_s and θ_r can be approximated from soil-water retention curves such as those shown in Figure 16. When simulating flow in peat, gains and losses of water from the matrix must be calculated using the appropriate value of unsaturated hydraulic conductivity given the changing soil-water pressure. The most common method for describing the soil water retention and unsaturated hydraulic conductivity relationships in peat involves fitting the van Genuchten-Mualem (VGM) soil hydraulic property model (van Genuchten, 1980) to measured data. The VGM soil water retention model is defined by Equation 5.

$$S_e = \frac{1}{[1 + (\alpha h)^n]^m} \tag{5}$$

where:

h = soil water *tension* expressed as a positive value, equivalent to -ψ (L)

α = scaling parameter that is the reciprocal of the air entry pressure (L^{-1})

n = dimensionless shape parameter that is inversely related to the pore-size distribution (-)

m = defined as 1-1/n (-)

The parameter values of Equation 5 can be estimated by collecting a suite of values for S_e, α, and h experimentally in a laboratory. Then n can be determined by curve fitting.

The hydraulic conductivity in the VGM soil hydraulic property model (van Genuchten, 1980) follows the Burdine-Mualem conductance model (Mualem, 1976). Here, the VGM hydraulic conductivity model (van Genuchten, 1980) commonly follows Equation 6 in peat studies, where m is restricted as above.

$$K(h) = \frac{K_{sat}\,\{1 - (\alpha h)^{mn}[1 + (\alpha h)^{n}]^{-m}\}^{2}}{[1 + (\alpha h)^{n}]^{ml}} \tag{6}$$

where:

K	=	hydraulic conductivity
K_{sat}	=	hydraulic conductivity of saturated peat
l	=	scaling parameter related to the pore-size distribution

The other parameters are as defined for Equation 5 (Mualem, 1976).

Unlike mineral soils that typically fix l to 0.5, studies in peat have found that fitting l gives better agreement with the measured data. Although more complex models better describe the multi-modal soil water retention and hydraulic conductivity profiles (Weber et al., 2017), for most applications the van Genuchten-Mualem model provides a good representation of the underlying hydrological processes.

Details about the range of the soil retention/unsaturated hydraulic conductivity parameters for peat are provided in Box 4⤵.

4 Transport of Solutes

The structure of peat and peatlands that creates the inherent complexity in water flow significantly impacts the transport of nutrients, carbon, and contaminants. The complexity of transport is not only related to physical factors but also to biogeochemical conditions. Similar to transport in mineral soils, mass transport in peat and peatlands is controlled by the physical structure of the matrix, hydraulic gradients, the character of the pore network, as well as the properties of the solute.

The movement of solutes through the pore network along with flowing water is called the advective flux. If the solute does not react with the other solutes or the soil, it is considered a conservative solute. Biogeochemical conditions in peat are broadly governed by the reactivity of organic matter and presence of redox sensitive chemical species (i.e., NO^3-, SO_4^{2-}). As such, the hydrophysical structure of the peat aquifer both controls and is subsequently controlled by the mass transport processes (McCarter et al., 2020).

Rapid horizontal and vertical spatial changes in both physical and biogeochemical conditions in peatlands result in complicated solute transport and transformation processes at various spatial scales. Effectively, transport processes at two spatial scales govern the overall solute transport in peatlands:

- processes with the pores of the peat; and,
- processes at the scale of peatlands affected by the spatial distribution of pore properties and influenced by micro-topographic features.

These are discussed in subsections of this chapter.

4.1 Transport in Peat

Similar to water flow, the size, shape, and connectivity of the pore network governs the advective transport of solutes in peat. A critical property of the peat pore network that influences both water and solute flow is its dual porosity structure discussed in Section 3, *Peat: A Porous Medium*. In peatlands, the distribution of pores from a predominance of macropores in near-surface peat to more numerous but smaller diameter pores at depth results in rapid shifts of transport rates vertically and horizontally. The increase in the abundance of small pores with depth also coincides with an increase of immobile porosity, ϕ_{im}, that does not contribute to advective flow. Solutes transfer to the immobile porosity from the mobile porosity via diffusion, which is driven by chemical gradients. Solutes that migrate into immobile porosity are abstracted from the solute flowing in the mobile porosity (Figure 17) with the net effect of retarding migration of the solute plume (Hoag and Price, 1997).

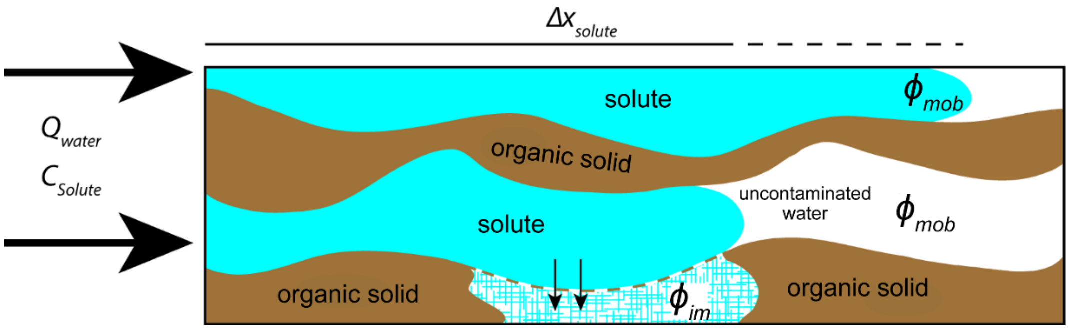


Figure 17 – Two pores within a peat without (top) and with (bottom) an immobile porosity. Blue represents pore water with dissolved solute, white represents pore water with zero concentration of solute, and brown represents organic matter. Both pores have the same input concentration (C_{solute}). The mobile porosity and its cross-sectional area are identical in both pores, and water flows through it at the same volumetric discharge rate (Q_{water}). The plume is shorter in the bottom pore as indicated by the solid portion of the black Δx_{solute} line above the figure because some of the solute mass that would otherwise be transported along with Q_{water} diffuses into the immobile porosity zone.

The parent material (e.g., *Sphagnum* moss, sedges, wood) has a large influence on the pore size distribution and the partitioning between the mobile and immobile porosity, thus plays an important role in solute transport. Estimation of mobile porosity and its influence on solute transport is explored in more depth in Box 5⤵. However, most of our understanding of solute transport in peat and peatlands has been determined in the upper 1 m of peat from *Sphagnum*-dominated peats from bogs or poor fens. McCarter and others (2020) and Rezanezhad and others (2016) provide good discussions on solute transport in peat.

The majority of solute transport in peatlands is through highly connected, near surface, large diameter pores (pore diameters > 70 μm), primarily consisting of macropores (pore diameters > 250 μm). Solute transport through these pores is predominantly via advective flow. The high degree of connectivity, coupled with the predominance of large pores, results in low dispersivities on the order of ~10^{-1} — 10^{0} cm (Hoag and Price 1997; McCarter et al., 2019). Dispersivity is a media and scale dependent property that represents both diffusion and variations in advection that are not accounted for in the flow portion of a solution in order to represent spread of a solute. As peat undergoes decomposition and compaction, such as with increasing depth below surface, dispersivity increases (Kleimeier et al., 2017).

Similar to water flow properties, *Sphagnum* and sedge peat differ in solute transport properties. In *Sphagnum* peat, the dispersivity and tortuosity is lower than that of sedge peat at an equivalent depth below the peat surface. This is due to the differences in pore structure, thus water flow, discussed previously. As such, the often systematic, small scale distribution of macropores and micropores, and the complexity of the pore network therein, is expressed at larger spatial scale, creating a clear region of enhanced solute transport. These regions of enhanced transport can be both vertically and/or laterally distributed as discussed in Section 3.2, *Unsaturated Zone Properties and Processes*.

The transfer of solutes within and between the mobile and immobile pore spaces is governed by the effective diffusion coefficient, which is both solute (the free water diffusion coefficient) and media specific (pore throat diameters). The free water diffusion coefficient is related to the ion size and the viscosity of the fluid—in the case of most peatlands, "fresh" water. The pore throat diameter is the size of the narrow portion of the opening between pores.

Although transfer of solute from mobile to immobile porosity zones is influenced by pore throat size, changes to the peat properties (increasing degree of decomposition) in the upper 50 cm of *Sphagnum* peat were not sufficient to decrease diffusion into the immobile porosity (McCarter et al., 2019). This is likely not the case in deeper, more decomposed peat, but there has yet to be measurements of this in deeper peats.

Although diffusion into the immobile porosity has been well established in peat, the mass transfer coefficients that are used to mathematically describe the process can be sufficiently high that solutes in the mobile porosity are essentially transferred to the immobile porosity instantaneously (McCarter et al., 2019). Under these circumstances, the peat can be simulated as a single porosity media with porosity equal to the mobile porosity (Simhayov et al., 2018). Currently, it is not known what specific discharge is low enough such that mass transfer rates between mobile and immobile pores are sufficiently high relative to the flow rate that dual porosity effects can be ignored (hence represented with a single porosity model). However, Kleimeier and others (2017) show that dual porosity effects were present at Darcy fluxes as low as 1.4×10^{-5} cm/s.

When the effective diffusion of solute in peat is low, the presence of an immobile porosity can lead to an elongated flushing of solutes, as the slower diffusive flux from the immobile porosity slowly transfers solutes to the mobile pore space (Hoag and Price 1997). This process can extend the period of contamination if a solute enters a peatland.

In peat, solutes are generally reactive due to the high organic matter content. Organic matter, whether considered peat or organic soil, removes cations from the pore water through adsorption. In peat, cation exchange capacities often exceed 100 centimoles per kilogram (cmol kg^{-1}) and can be much higher than in clays and clay loams which are typically 30—50 cmol kg^{-1} and in sands are typically 3—5 cmol kg^{-1} (Kyzoil, 2002; Rippy and Nelson, 2007). This results in many cations being essentially immobile within all but the near-surface and high hydraulic conductivity peats where soil water residence times are low.

The mechanisms that govern cation adsorption are complex in peat. In *Sphagnum* mosses, cation adsorption is thought to occur in two different regions: the pore space and inter-cellular spaces. In soils that are not composed of living and dead plant cells, adsorption primarily occurs on the interface between the pore water and solid phase. However, the presence of plant cells creates a secondary region for adsorption to occur. In the leaves of *Sphagnum*, ions are transferred across the *Sphagnum* leaf's cell membranes due to ionic gradients between the inter-cellular water and the pore water. Once within the inter-cellular space, cations adsorb to the cell wall. This allows for cations to affix throughout the leaf cell walls, not just the pore surface (Clymo, 1963; Richter and Dainty, 1989). Conversely, on the branches and stems of *Sphagnum*, direct ion exchange with the tissue surface is the dominant cation adsorption mechanism, which lowers the apparent adsorption capacity due to a decrease in available surface area.

The ability for cation adsorption depends on the physical location of the adsorption binding site and the highly variable chemical composition of the organic matter, thus the complexity of adsorption processes in peat. Most cations will undergo direct ion exchange with peat based on the peat's overall negative surface charge at relevant pH from 4 to ~7 (McCarter et al., 2020). However, the presence of carboxyl groups and/or reduced sulfur groups (among others) in peat creates a large heterogeneity in adsorption potential depending on the specific geochemistry of a given cation. These complexes do not readily desorb, leading to relatively stable long-term removal of metals—and most cations—from peatland pore waters (Pratte et al., 2018). The heterogeneity in adsorption processes leads to differential transport rates of cations based on their specific chemistry, organic matter composition of the peat, and prevailing geochemical conditions, thus predicting contaminant transport at large spatial scales in peatlands can be difficult as discussed in Section 4.2, *Transport in Peatlands*.

Transport processes in peat and peatlands are governed by the prevailing geochemical conditions of both pore water and the peat due to the strong control pore structure has on solute transport. The decomposition of organic matter, in this case peat, controls the specific pore structures and, in many cases, the layering of peat within a peatland. These decomposition processes are governed by the soil moisture content, delivery of microbiologically available nutrients (e.g., O_2, NO_3^-, SO_4^{2-}), labile carbon to act as an electron acceptor, and the timely removal of decomposition end-products (Bauer et al., 2007). It is the balance, or imbalance, of these processes that allows for the accumulation of organic matter in peatlands and gives peat its unique structure. McCarter and others (2020) provide a more detailed overview of these processes and their interaction with hydrological and solute transport processes in peat.

4.2 Transport in Peatlands

In peatlands with pronounced declines in hydraulic conductivity with depth and associated distribution of micro and macropores can induce large scale solute transport patterns. In near surface high hydraulic conductivity peat, solute transport follows that of water flow, with a large proportion (> 90 percent) of the total mass flux often transported within the upper few decimeters (McCarter and Price, 2017b). This is illustrated by the dissolved solutes. Due to the low dispersivity of these near surface pore networks, the resulting solute plumes are often elongated and narrow, both vertically and horizontally (Hoag and Price, 1995; McCarter and Price, 2017b).

Although local preferential flow paths, such as those created by layering of peat with different decomposition states, or roots, can induce solute fluxes well beyond that expected of the primary solute plume (Baird and Gaffney, 2000; McCarter and Price, 2017b), these are typically localized in their impact. As the hydraulic conductivity decreases to a more uniform value with depth, solute plumes are transported through a greater proportion of the peat aquifer. Thus, knowledge of the shape of the transmissivity profile is critical to understanding solute transport in peatlands.

In many peatlands, the surface topography is not uniform. The hummock—hollow topography that is a defining feature in many peatlands, notably bogs, leads to regions of preferential flow and transformation of solutes. Within the hollows, the near-surface hydraulic conductivity is higher than that at an equivalent elevation beneath adjacent hummocks. This lateral distribution of hydraulic conductivity results in preferential solute transport within and between the hollows, while limited lateral transport occurs through hummocks (Balliston et al., 2018). This inter-hollow transport occurs both as overland flow and in the near subsurface, where the hydraulic conductivity of near surface peat in hollows is typically high (Figure 18).

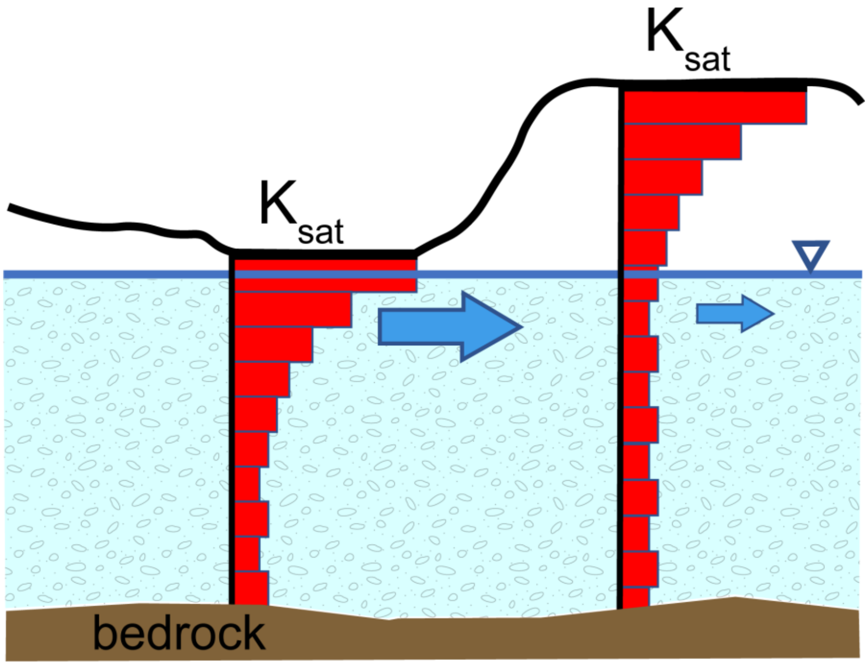


Figure 18 - Identical (arbitrary) hydraulic conductivity profiles with depth below the surface under a hollow (left) and hummock (right) result in much higher transmissivity beneath hollows when the water table is near the hollow surface. Similarly shaped hydraulic conductivity profiles are observed in fen ridge and flark systems as shown in Figure 4.

Although the permeability is typically lower in hummocks at the same plane as the peat in adjacent hollows, lateral solute transport still occurs. Within the hummocks, the available organic matter and longer residence times increase the potential for biogeochemical reactions and sequestration of contaminants. While the hummock—hollow topography is typically associated with *Sphagnum*-dominated peatlands, *tussocks* can form from dense groupings of sedges within sedge dominated peatlands, which are functionally comparable to the previously discussed hummocks. Hence, most solute transport is routed around tussocks. This complicated routing of solutes in peatlands makes accurately predicting solute plume development difficult.

Peatlands with distinct directional flow, such as ladder and northern ribbed fens (sometimes referred to as *aapa mires*, in Europe), often have a peat ridge—flark surface (e.g., Figure 4), or ridge—pool topography (e.g., Figure 7). Peat ridges are areas of elevated peat mounds, similar to hummocks, but they are linear and water flow is perpendicular to the peat ridges. Flarks are flat expanses of moss or sedges where the water table is near or above the surface. The spatial variation of the water table depth in relation to the near surface peat or the height of the water surface above the ground facilitates periods of extremely high solute transport rates during high water table periods. The majority of solutes are preferentially transported in these highly permeable groundwater layers or surface water (McCarter and Price, 2017b).

During high water table periods, solute transport is rapid, with rates on the order of hundreds of meters per day, which is extremely high for peatlands. However, the presence of open water pools that bisect these surface flow conducting peatlands can increase or decrease solute transit time, depending on mixing and storage of solutes, increased/decreased biogeochemical reactions, or preferential flow within the pools (McCarter and Price, 2017b). Similar to other peatlands, during low flow periods the water table resides within the lower hydraulic conductivity deeper peat and solute transport decreases exponentially, greatly reducing downgradient connectivity.

The peat—pool topography also alters the reactivity of different solutes (McCarter et al., 2017). Many redox sensitive or metal chemical species will be preferentially removed in the peat ridges, rather than in the pools, due to the abundance of organic matter (peat) in the ridges. While in the pool, nutrients such as nitrate or phosphate can be preferentially removed from the water column through algae uptake or other biochemical processes that are dependent on sunlight. Thus, the mobility of any given solute in these systems will not only depend on the hydrological transport rates but also on the partitioning between surface water and subsurface reaction rates at the peatland scale. It is this combination of water table-dependent transport and high spatial heterogeneity of reaction rates that makes predicting the pattern of reactive solute plumes in peatlands difficult.

5 Contamination in Peatlands

With increasing resource extraction activities such as: peat harvesting, mining, and associated transportation corridors in boreal areas; and with forest harvesting and agricultural development in tropical environs; there is increased risk of anthropogenic contaminant release in peatlands. In many cases, contaminants alter the ability of peatlands to provide important ecosystem services such as carbon sequestration. These alterations are driven by shifts in the cycling of nutrients and carbon or by direct disruption of biotic systems (i.e., by toxic metals) that are integral to peatland function. Sources of contamination range from direct aqueous inputs to long-range atmospheric deposition. The type of peatland, specific contaminant, and mode of input govern the contaminant mobility and changes of peatland functions. For example, in some peatlands, a chemical will be a mobile contaminant but in others it will be sequestered. Understanding the feedbacks between specific contaminants, delivery methods, and peatland function is critical to properly mitigating adverse human and environmental health impacts.

5.1 Atmospheric Pollutants as Contaminant Sources

Atmospheric deposition of contaminants is a pathway for the introduction of contaminants into peatland groundwater systems. Unlike other ecosystems, where toxic or trace metals can create local regions of decreased ecosystem functionality, the large mass of organic material (i.e., peat) limits both the toxicity and mobility of many toxic and trace metals. This is achieved, primarily, through adsorption onto organic matter. Interestingly, this makes peatlands excellent long-term records of metal deposition rates, with clear signals of the Industrial Revolution being captured in European peatlands (Livett et al., 1979; Shotyk et al., 1996).

Due to these sequestration processes, contaminants entering peatlands do not often result in groundwater contamination at larger spatial scales. Despite peatlands having an unparalleled ability to sequester metals over long (centuries) timeframes, changes to the hydrological, geochemical, or climatic conditions can upset the delicate balance of peatland functions that regulate metal sequestration, turning metal sinks into sources. These disruptions can release previously-sequestered metals into more susceptible systems such as drinking water aquifers, rivers, and lakes through direct groundwater inputs and/or surface erosion of peat. For example, Rothwell and others (2007) found Cu, Ni, Pb, Ti, V, and Zn were leached from peat into headwater streams, with Pb, Ti and V being mobilized by the release of dissolved organic carbon (DOC); Pb is also released by erosion (Rothwell et al., 2008).

Atmospheric pollution can be both relatively local sources such as NO^{3-} and SO_4^{2-}—important components in acid rain—or distal sources as in the case of mercury. More details on mercury in peatlands are provided in Box 6↴. Atmospheric deposition of these redox-sensitive pollutants can lead to enhanced mineralization (decomposition) of peat (Chapin et al., 2003). This can cause changes to water table variability and soil moisture retention, thus water and solute flows. The specific species of the contaminant will determine its reactivity and toxicity within peatlands, hence its persistence and mobility in peatland groundwater systems and its impact on peatland hydrology.

5.2 Direct Pollutants as Contaminant Sources

With increasing development pressures globally, particularly in the North, the risk of direct contamination to peatlands has increased. Direct contamination can be from overland exposure such as by wastewater treatment, vehicles traveling across peatland roads, spills associated with train derailment, or by subsurface pollution such as that from buried pipeline leaks. Depending on the vector of pollution, the processes that govern its movement and degradation will be different.

Dissolved contaminants can move quickly in the high near surface hydraulic conductivity layer in many peatlands. This can result in large areas of contamination, coupled with the potential for downgradient impacts. However, the upper layers of peatlands are commonly the most biogeochemically active zones due to the abundant and labile organic matter, high density of vascular roots, access to oxygen, warmer summertime temperatures, and exposure to sunlight. These properties allow for the rapid transformation or sequestration of most contaminants. Thus, many soluble contaminants do not remain long after the contaminant source has been stopped or fixed. The risk of these contaminant sources causing adverse environmental effects depends on the balance between enhanced near surface transport and reactivity of a given contaminant, as well as its toxicity.

In contrast to overland or near surface contaminants, deeper subsurface contaminants often remain in peatlands for longer periods of time because of the commonly low hydraulic conductivity matrix at depth, even after the source has been remedied. Subsurface contaminants introduced into the low hydraulic conductivity layers of the catotelm move slowly. The limited potential for transport must be balanced against the relatively recalcitrant organic matter and the general absence of electron acceptors (e.g., NO_{3^-}, SO_4^{2-}) at depth in most peatlands. These conditions are not conducive to the biodegradation of most contaminants. Geochemically, the buried peat in the catotelm strongly binds cationic contaminants, further limiting contaminant migration. In most cases, subsurface contaminants will remain in peatlands for long time periods, well beyond direct surface pollution.

5.3 Non-Aqueous Phase Liquids (NAPLs)

Up to this point, our descriptions of the transport processes and behavior of contaminants have been directed at water soluble substances. However, non-aqueous phase liquids (NAPLs) have dramatically different transport and reactivity properties, and behave differently in peatlands than in mineral soils. Like most contaminants in peatlands, the likelihood of transport or sequestration of these particular contaminants will depend on the specific peatland type, biological communities, method of introduction (e.g., surface or buried pipelines), and whether it is "lighter" than water (LNAPL, e.g., diesel fuel) or denser than water (DNAPL, e.g., bitumen or chlorinated solvents). While many of the processes governing the migration of NAPLs in groundwater are common to both mineral and peat systems, the chemical and mechanical properties of peat (Table 1) are so different than those of mineral sediments that the fate of NAPLs can be distinctly different.

The wettability of a soil matrix concerns the inclination of a pore fluid (including air, water, and NAPL) to spread over a pore surface in the company of the other fluids. This depends on the surface chemistry of the pore and the wetting history. Dry peat is used as a mopping agent in oil spills on water because the nonpolar hydrocarbon molecules preferentially wet the peat. If water wet, however, the affinity for NAPLs is reduced. The conditional wettability of peat therefore exercises a strong control on the mobility of NAPL.

In a peat soil profile, pores are generally already water wet due to capillary processes. However, the poorly decomposed near surface peat that comprises predominantly large pores drains readily, and the pores become increasingly air wet (more hydrophobic). Deeper, more decomposed peat has a smaller pore size distribution that retains more water; hence, pores may be predominantly water filled and remain water wet. When NAPL is introduced (e.g., a spill), it will preferentially enter the larger pores that are air wet and will be blocked from water filled (water wet) pores that are more prevalent at depth. The consequence of this is that spilled NAPL will preferentially spread in the upper layer and be precluded from entering more decomposed peat at depth. This is true for both LNAPL and DNAPL, although the latter will be more likely to exploit partially water wet pores and move downward through the peat. Gharedaghloo and Price (2017) provide more details of peat wettability and its implication for NAPL transport.

The wetting behavior of a soil controls the capillary—saturation relationship and the relative permeability—saturation relationship, which can be defined for both water and NAPL (Gharedaghloo and Price, 2019). In the presence of NAPL filled pores, water will recede into pore throats, and the low degree of water saturation will correspond with low (more negative) pressures. Simultaneously, the low degree of water saturation will reduce water permeability and water redistribution will be restricted. Conversely, NAPL saturation will be high and relatively more mobile. However, given a spill of finite volume, the spread of NAPL will decrease the NAPL saturation and, simultaneously, the NAPL permeability until the residual saturation level is reached. Thus, after initially spreading rapidly in the upper peat layer, the distribution of NAPL will stabilize, whereupon volatilization, microbial processes, and solubilization will further reduce the NAPL pool. Solutes released from the NAPL will tend to remain near the surface where their mobility and extent is constrained by microbial breakdown (Gupta et al., 2020) and adsorption (Gharedaghloo and Price, 2021), which is high in the organic rich peat matrix (Figure 19).

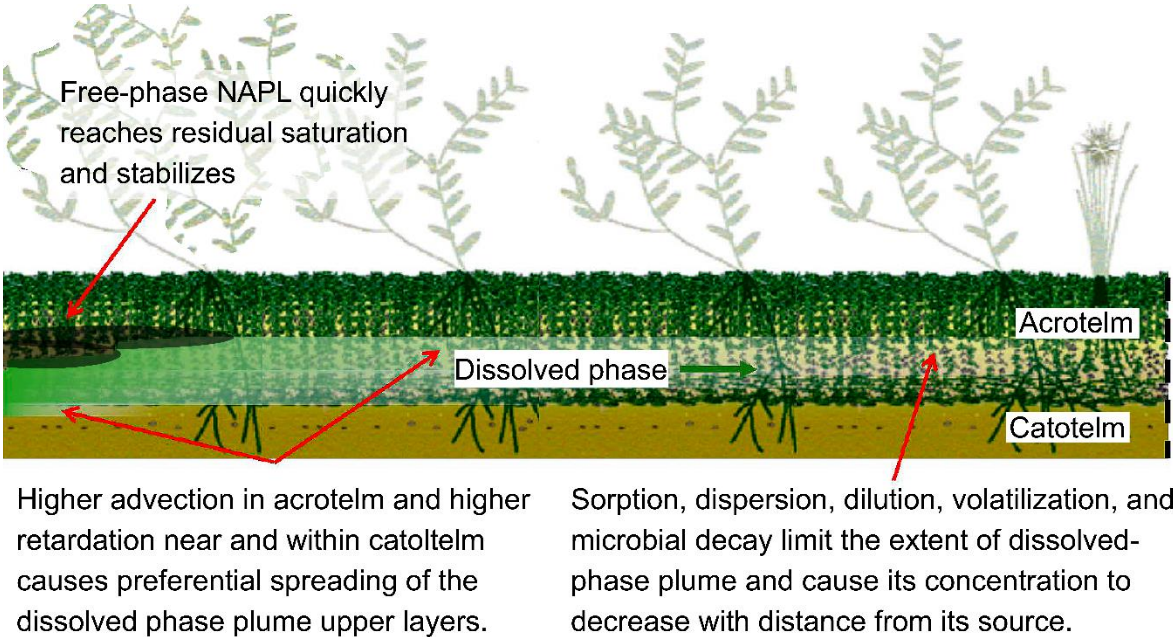


Figure 19 - Schematic of free-phase NAPL and dissolved phase redistribution through a peatland. There will be preferential spread of the NAPL in the shallow peat horizons until it reaches residual saturation. Soluble components will flow preferentially in the upper layer and their concentration will decrease with distance from the NAPL pool until the dissolved phase plume reaches its maximum extent (Modified from Gharedaghloo and Price, 2017).

6 Peatland Disturbance

Peatland disturbance can occur through intentional or unintentional anthropogenic activity. Intentional disturbance results from activities such as mining, agriculture, forestry, peat harvesting, urban expansion, and resource exploration; these disturbances tend to be localized. Unintentional disturbances occur from phenomena such as climate change, atmospheric pollution, and fire; these disturbances can be widespread. Given that peatlands are a major terrestrial store of carbon, loss of peat or loss of their peat-accumulation function is considered a major contributor to climate change. In most cases, disturbances, whether intentional or unintentional, can have large impacts on groundwater flow within peatlands and to surrounding systems.

6.1 Peatland Drainage

Most peatland disturbances are related to activities that drain or otherwise desiccate the peat. This can result in subsidence because of the increased effective stress associated with a lower water table; such subsidence is only partially reversible upon rewetting. Additional subsidence occurs with lowered water tables as a result of enhanced decomposition by aerobic microbes, which is non-reversible.

Both causes of subsidence alter the structure, thus hydraulic properties, of the peat; peat collapse reduces its porosity and permeability and increases its water retention capacity, thus decreases hydrological connectivity between the old cutover peat and the new moss layer in restored systems (Gauthier et al., 2018). In addition to peat structural changes, subsidence of the peat surface can result in a large-scale change in the groundwater relationship to and interactions with adjacent ecosystems because the lowered surface can induce increased surface and groundwater inflows from adjacent ecosystems.

In peatlands, the drainage efficacy increases with ditch depth and decreases with ditch spacing. Drainage efficacy is greatest adjacent to the ditch and diminishes with distance, such that common spacings range from 30 to 50 m, depending on the land management goals. To a certain extent, this is a self-regulating process since peat subsidence associated with drainage reduces the hydraulic conductivity of peat, thus its potential to drain. However, on steeply sloped peatlands, such as blanket peatlands common in the British Isles, drainage may be contoured to catch overland flow, increasing the efficiency of peatland drainage (Holden et al., 2006). Drainage directly alters the water balance by promoting seepage to the drains, but the lower water table can limit evapotranspiration losses. However, the connectivity of the surface to water stored in the vadose zone and below the water table is enhanced (Price, 1996) by the higher water retention capacity associated with more decomposed drained peat, and by surface subsidence that keeps the water table closer to the surface than it otherwise would be. The higher connectivity of the surface to stored water in this setting helps maintain soil evaporation.

Draining peatland for agriculture or forestry is common in Europe, but less so in North America. In addition to the impacts of drainage on peat physical structure noted above, the addition of fertilizers or nutrient-rich water for agriculture can preferentially enhance the decomposition of surface peats (Liu et al., 2017). This can create an inversion of the typical peat profile, where the densest peat is at the surface and the peat becomes less dense with depth. As decomposition occurs in these surface peats, a greater proportion of overland flow occurs, similar to those drained for horticultural peat extraction. Drainage of peatlands for forestry often results in even lower water tables due to enhanced water loss from transpiration. As the root systems of trees often penetrate much deeper into the phreatic zone than shallow—rooted shrubs common in peatlands, there is a greater loss of water directly from below the water table than with shrubs alone. The lowered water table may enhance local groundwater discharge or reduce groundwater recharge.

In tropical peatlands, especially in South East Asia, peatland degradation is driven by logging, drainage, large-scale plantations, and recurrent fires (Dohong et al., 2017). The extremely high hydraulic conductivity found in tropical peat domes exacerbates the impact of ditching, resulting in water tables >1 m below the surface, promoting rapid decay of organic material (Baird et al., 2017), thus subsidence-causing topographic irregularities. This makes uniform rewetting difficult following ditch blockage (Dohong et al., 2018). Wösten and others (2008) suggested water table levels more than 40 cm below ground surface result in degradation, thus these systems are very sensitive to change.

It is worth noting that peatland degradation in all settings can result inadvertently from inappropriately-situated roadways, having the same effect on peat properties as drainage on the down-gradient side. These effects may be partially reversible in the case of temporary roadways, depending on the extent and duration of disturbance (Elmes et al., 2021).

6.2 Peat Harvesting and Restoration

Peat harvesting typically occurs on *Sphagnum* dominated bogs, since *Sphagnum* moss is generally the preferred soil amendment for horticulture. At peat harvesting sites, drainage ditches are installed to decrease the water content of the peat and increase the bearing capacity for machines to access the site. Ditch spacing is commonly ~30 m, but narrower spacing is required for denser (low hydraulic conductivity) peat. The surface material (living and dead but poorly decomposed peat forming vegetation) is typically stripped to access the peat, and the underlying peat is removed in blocks or extruded—but nowadays is typically remove by vacuum harvesting.

For the latter, the surface is harrowed to break the capillary connection of the surface material from the underlying water source; this is allowed to dry in the sun and wind, then collected with industrial vacuums. At a peat extraction site in Canada (Bois-des-Bel, Quebec), median water table depth was ~65 cm, in contrast to an unharvested part of the same bog, where the median water table was ~25 cm (Figure 20).

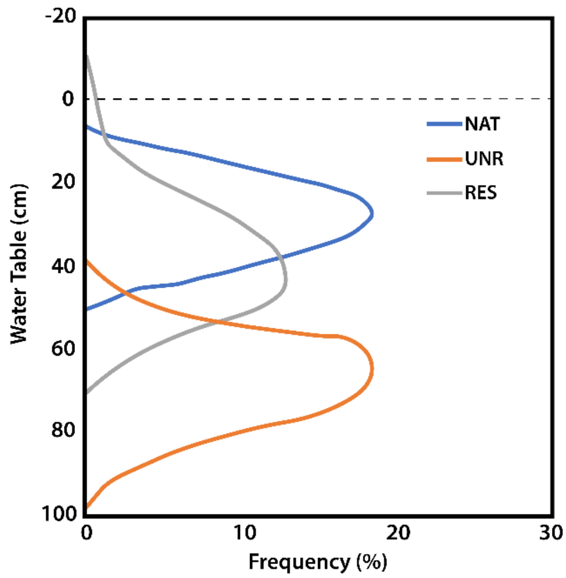


Figure 20 - Water table depth frequency at a natural bog site (NAT), a drained unrestored site (UNR), and a restored site (RES). For the natural and unrestored sites, the surface (dashed line) is at 0 cm. For the restored site, at which a ~15 cm layer of moss had regenerated, the interface between the cutover surface and regenerated moss is also shown at 0 cm; the regenerated moss occurred above that (Based on data from McCarter and Price, 2013).

Water table variation is greater at harvested sites, mostly because the drainable porosity of peat is reduced to ~0.05 from 0.35 to 0.55 in undisturbed peat (Price, 1996), due to peat consolidation caused by the lowered water table (decreased pore water pressure) and weight of machinery, as well as enhanced decomposition caused by aeration. Restoration of these degraded systems requires blockage of drainage ditches to reduce water loss. However, because of subsidence or peat cutting that alters the natural surface profile, as well as the altered hydraulic properties of the peat matrix, "normal" water table relations are not re-established until *Sphagnum* mosses regenerate sufficiently and become partially decomposed at their base, so that the hydraulic properties with depth transition in a similar manner to that in an undisturbed system (Taylor and Price, 2015).

While mosses are slow to regenerate on cutover peat, they can be introduced as part of the restoration process. Since *Sphagnum* mosses are non-vascular, connectivity between deeper peat and the surface relies on capillary rise of water through the matrix. However, in the early stages of regeneration, the compacted and more decomposed underlying cutover peat transitions abruptly to the mosses introduced for reestablishment. The result is a capillary barrier that restricts upward movement of moisture from the peat to the mosses (McCarter and Price, 2015), thus potentially restricting their growth and resilience.

6.3 Climate Change

The changing climate has an uncertain impact on peatland systems since peatlands have developed within a particular hydrogeomorphic setting as a consequence of the local climate that has prevailed since deglaciation. Water exchanges within peatlands, and between them and their surrounding landscape and/or fluvial systems, are a product of the climate that prevailed over their development. The rates of their carbon sequestration and decay, which dictate the rates of peat accumulation and their eventual form, are closely tied to climate (Figure 21). Climate change affects peatland—groundwater interactions indirectly through changes to floral and faunal communities, food webs, nutrient availability, the hydraulic structure of peat, its thermal state, or other ecosystem properties. Given the importance of groundwater to the water balance in many peatland settings, and the sensitivity of peatlands to water balance changes, even small changes to peatland—groundwater interaction could alter the greenhouse gas flux to the atmosphere (Figure 21), such as increasing the CO_2 emission for drying scenarios and increasing the CH_4 flux for wetting, potentially amplifying climate warming (Tarnocai, 2006). However, ecohydrological feedback processes that may dampen or amplify greenhouse gas fluxes are poorly understood. In large part, this is due to the lack of clarity on the rate, pattern, and trajectory of change of peatland ecosystems, their surrounding landscapes, and their interactions with groundwater systems.

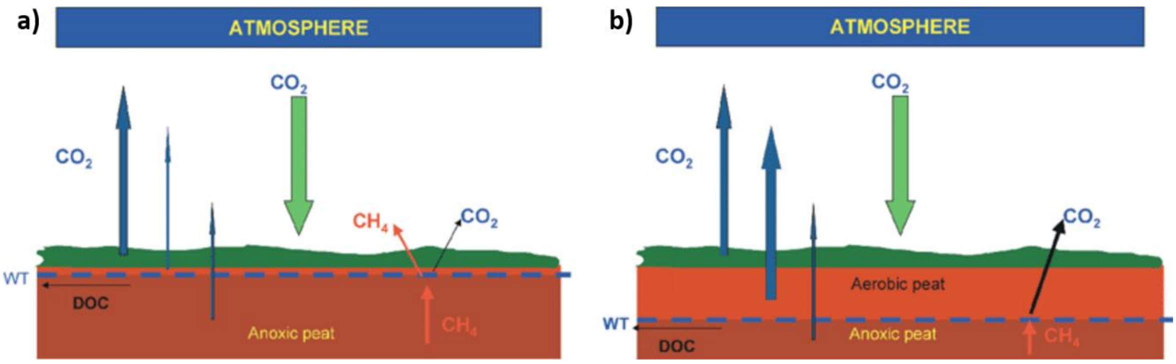


Figure 21 - A simplified representation of carbon exchanges at a) present and b) in a warmer climate. a) Carbon dioxide (CO_2) is sequestered from the atmosphere through plant photosynthesis. These plants form the peat deposit. CO_2 is released by the decaying plant litter and peat, especially in the aerobic zone above the water table (WT). Smaller quantities of methane (CH_4) are also released from the saturated, anoxic peat below the water table, although some of this is oxidized as it moves through the aerobic zone. b) A changing climate that results in a lower water table accelerates CO_2 loss to the atmosphere from the thicker aerobic zone, which is a positive feedback to climate warming. CH_4 is more likely to be oxidized under the low water table scenario, and since CH_4 is an important greenhouse gas this represents a negative feedback to climate warming. In most peatlands the larger CO_2 efflux will have a greater impact on global warming than reduced CH_4. Dissolved organic carbon (DOC) produced by decaying peat moves offsite via groundwater, and can be an important carbon source for downstream aquatic environments. A lower water table will reduce groundwater outflow, hence DOC export. Diagram from Renou-Wilson and others (2011).

Climate warming has the potential to transform peatlands because it increases the availability of energy needed to drive the hydrological, meteorological, and ecological processes that control their form and function (Carpino et al., 2021). In tropical regions, climate warming is expected to increase the frequency and severity of drought (and wildfire) and flooding events. Climate warming is generally thought to increase precipitation, since warmer air can hold more precipitable water. This could lead to the introduction and substitution of plant species, which would then lead to changes in local water and nutrient cycling and an alteration of the type of organic material for peat formation. For example, a lowered water table promotes the growth of woody vegetation, itself a form of carbon storage, although likely only important in its initial establishment. Ongoing effects of forest growth, however, include higher transpiration losses and precipitation interception that enhance drying of the peat. Over periods of decades to centuries, this could alter groundwater interactions through changes to peat hydraulic properties.

A climate warming-induced increase in peatland water temperature has the potential to increase the frequency and duration of hypoxic and anoxic conditions, which can reduce the growth rates of peat-forming species, although this is potentially offset where groundwater contributes significantly to peatland water balances. Sea-level rise driven by the melt of the Greenland and Antarctic ice caps will result in the loss of coastal wetlands in the tropics and elsewhere.

In the temperate region, climate warming is expected to increase the frequency and duration of droughts, putting more pressure on groundwater systems to maintain peatlands. Warmer winters would produce a higher frequency of mid-winter melt events, reducing the amount of snow on the ground at the end of winter. This would reduce the magnitude of the annual end-of-winter moisture recharge to peatlands. The absence of significant changes to total annual precipitation in some temperate regions can mask significant changes in the temporal distribution of precipitation. For example, Shook and Pomeroy (2012) reported an increase in the frequency of large, convective storms in the Canadian prairies and, as a result, greater flooding and hydrological connectivity of sloughs and other wetlands (Hayashi et al., 1998). Such changes have the potential to disrupt the recharge of local groundwater systems and thereby alter their role in sustaining peatland systems.

Perhaps the greatest impact of climate change on peatlands is expected in the boreal and subarctic regions, since it is these regions that contain most of the world's peatlands, have higher than global average projected temperature rise, and because the peatlands of these regions developed and function in the presence of seasonal ground ice and/or permafrost. The loss of ground ice/permafrost can profoundly alter the hydrological functioning of peatlands. For example, ground ice, whether seasonal or interannual, can impound water, thus limit drainage; however, as it thaws and the overlying ground surfaces subside, hydrological connections develop between previously impounded wetlands, allowing them to cascade shallow groundwater from one wetland to the next (Connon et al., 2015).

Seasonal ice and permafrost can impede flow between groundwater and surface water systems; the thaw of such impeding layers can increase groundwater interaction with wetlands, a process often referred to as *groundwater reactivation* (St. Jacques and Sauchyn, 2009). In permafrost regions, the thaw of ground ice can produce *taliks* (Connon et al., 2018; Devoie et al., 2019) where the depth of summer thaw exceeds the depth of winter re-freeze. Such layers provide a conduit for suprapermafrost groundwater exchange even during winter. The thaw of permafrost impoundments and the development of taliks can lead to the dewatering of wetlands (Haynes et al., 2018).Permafrost thaw and resulting land cover subsidence can also dramatically alter the local environment for peat formation and decay processes (Swindles, 2015), thus peatland hydrology (St. Jacques and Sauchyn, 2009). Additional information on groundwater in permafrost peatlands is provided in Box 2.

In short, it is difficult to predict the changes that will occur to any particular peatland as a result of climate change. Given the strong ecohydrological feedbacks in peatland systems, caution is advised in generalizing outcomes without a thorough understanding of ecosystem processes.

6.4 Peatland Wildfires

Global climatic and environmental conditions are testing the limits of peatlands to regulate their hydrological and biogeochemical functions; the cumulative impacts of disturbance on peatlands suggests these long-term ecosystem functions are at a tipping point. Of particular concern are the impacts of peatland wildfires—not only on the hydrological function of peatlands but also the negative impacts to environmental and human health. Under the warming climate, wildfires in peatland dominated landscapes are likely to increase, as peatlands are expected to become drier (Helbig et al., 2020). The susceptibility to fire is similar to that associated with drainage or dewatering associated with human activity (Figure 22).

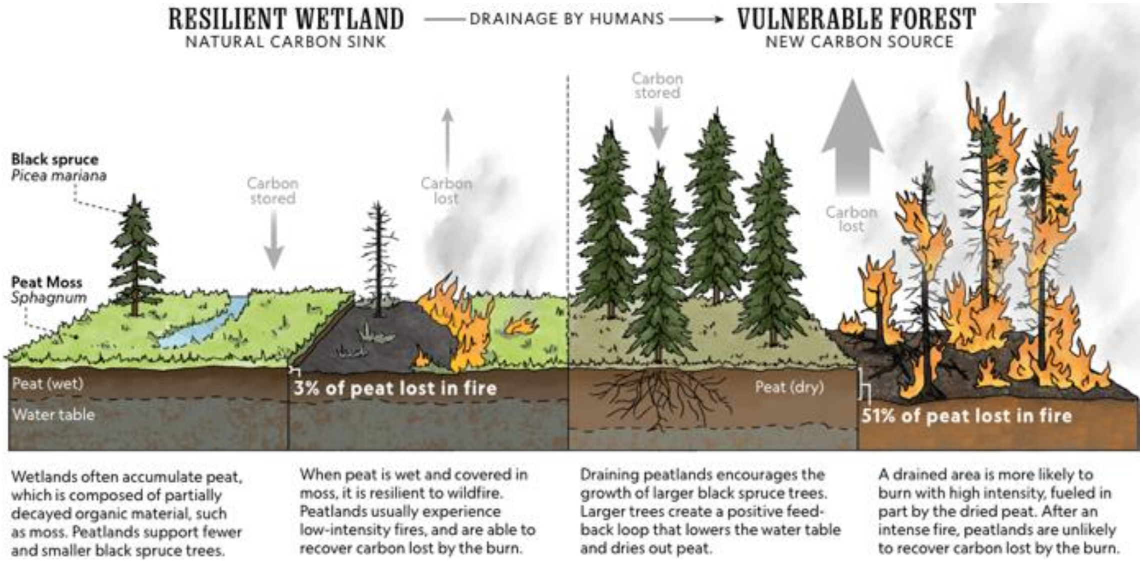


Figure 22 - The high water table in undisturbed peatlands provides a measure of protection against wildfire. While the surface may burn, the low intensity of the fire leaves sufficient nearby seeds and diaspores and a substrate more suitable for recolonization of typical peatland plants. Deeper water tables associated with drainage, dewatering, or climate warming intensify the severity of wildfire burns, consuming much of the peat deposit, and leaving the area devoid of viable genetic material to recolonize the surface—and with substrate properties entirely different than pre-burn (Graphic from Ebein, 2019 based on data from Wilkinson et al., 2018).

In addition to the large carbon loss associated with severe burns, the thickness of the remaining peat and the hydraulic character of the soil profile alters the groundwater regime and thus the function of any future recovered ecosystem. It is critical to understand the interactions between peatland wildfires and changes to peatland hydrology to better adapt to this growing disturbance.

Immediately after a peatland wildfire, several key changes to the physical and hydrological processes impact water flow. Changes to the physical and chemical structure of the peat from fires plays a critical role in what, and how much, percolates to the water table. As peat dries during combustion, the organic molecules become hydrophobic (Moore et al., 2017), reducing the volume of water percolating into the peatland through smaller diameter pore throats.

The degree of hydrophobicity is linked to not only the botanical origin of the peat but also the length and degree of heating, thus fire severity (Wilkinson et al., 2020). These changes result in a greater proportion of overland and near surface runoff being generated from burned peatlands (Sherwood et al., 2013) rather than by groundwater flow. However, the large proportion of macropores in the upper peat matrix can allow for rapid bypass flow from the surface to deeper, non-hydrophobic peats. Thus, the specific post-fire peat pore structure partly governs the total proportion of water that recharges the peatland aquifer. As some percolation occurs, the fine-grained ash particles are downwardly mobilized, increasing peat bulk density (Elmes et al., 2019), thus lowering saturated hydraulic conductivity (Ackley et al., 2021) and specific yield (Sherwood et al., 2013).

The loss of vascular vegetation following fire may reduce transpiration, leading to an increase in the water table height, increasing the transmissivity of the peat profile, and enhancing subsurface runoff from burned peatlands (Morison et al., 2020). However, increases in evaporation mediated by reduced shading have been found to partly offset reductions in transpiration and can lead to lower water tables (Thompson et al., 2014). In either case, reduced buffering of hydrological inputs and outputs caused by removing surface vegetation and ground cover may be exacerbated by the lower specific yield, driving greater water table variability. Concurrently, as the peatland vascular and bryophyte vegetation recovers from the fire, there are subsequent feedbacks to peatland hydrology driven by changes in atmospheric water exchange. As peatland vegetation returns after fire, the water table decreases due to increased evapotranspiration from both the vascular vegetation; the return of hydrophilic soils increases that connectivity of the peatland surface to the water table. It is not only the direct impacts of wildfires on peatland hydrology, but also the evolution of this particular disturbance over time and the compounding effects of climate change that makes peatland wildfires a particularly insidious hydrological disturbance.

7 Numerical Modeling in Peat and Peatlands

As with many branches of environmental science, numerical modeling has been used to simulate peatland processes, with varying degrees of success. Simulating water table position (and sometimes soil moisture) is a common objective because water table (and soil moisture) is so important to peatland function from hydrological, water quality and solute transport, climate, biogeochemical, and ecological points of view. Those focusing on ecology, biogeochemistry, and sometimes climate often use a simplified approach to hydrology to drive mechanistic processes relevant to their primary interest. Here, we focus only on modeling groundwater (including soil-water) hydrology and transport. For the most part, modeling of flow and transport in peat and peatlands have used numerical models developed for the porous matrix of mineral systems.

The fundamental processes represented are mostly similar, but some features of peat and peatlands provide additional challenges that increase uncertainty. These include the very low hydraulic gradients associated with the relatively flat landscape, highly and sometimes systematically irregular surface (i.e., microtopography), highly compressible matrix, heterogeneity of hydraulic properties, and the dual porosity character of peat. This section focuses on the use of models to simulate flow and transport in peat and peatlands, and the special challenges of using such models.

7.1 Numerical Flow and Transport Models

The sophistication of the model chosen for a project must reflect the nature of the domain being simulated, the degree of parameterization, availability of parameter data, and the objective. Modeling of peatland systems and their interaction with adjacent or underlying mineral groundwater systems requires a two- or three-dimensional approach.

Given the relatively thin peat layer (from a hydrogeological perspective) and the proximity of the water table to the surface, the dominant processes can usually be represented by a model that incorporates only saturated flow processes; with coupling to the atmosphere represented by an appropriate surface boundary condition. Freely available and well-documented models such as MODFLOW, which represent the partial differential equations for flow, provide a useful approach for assessing landscape scale influences on peatland function. For example, Reeve and others (2000) demonstrated the importance of the permeability of the underlying mineral deposit on the nature of vertical flow in bog peatlands. Where underlain by low permeability deposits, simulated flow in peatlands was primarily horizontal with laterally varying, isolated flow cells associated with meso-scale peatland features. In contrast, in areas with relatively high permeability underlying mineral deposits, simulated flow was primarily vertical. Quillet and others (2017) showed—using a saturated flow model—how a permeable esker deposit controlled the topographic slope of the adjacent peatland, and the importance of vertical heterogeneity in the peat deposit for maintaining appropriate water tables. Lower peat hydraulic conductivity associated with enhanced decomposition near bog peatland margins was shown to enhance water storage, which is favorable for peatland development (Lapen et al., 2005).

Sutton (2021) simulated three-dimensional saturated flow and transport in a constructed upland fen peatland system and showed the spatial pattern of salt contamination in the fen peatland reflected the design of the system, including features such as recharge basins in the upland. This, and most other studies, do not explicitly represent the effect of ground freezing in simulation studies. Given the proximity of the water table to the surface in peatlands, the hydraulic properties of the near-surface layers can change profoundly with freezing, and this requires explicit representation of thermal processes in order to simulate winter conditions (McKenzie et al., 2007).

A particular challenge associated with representing flow and transport at the landscape scale is the high degree of discretization required to represent the exponential decrease in hydraulic conductivity with depth, particularly in the upper layers. This feature of peatlands controls the transmissivity feedback mechanism that was described earlier, wherein horizontal flow is highly dependent on water table elevation because water table elevation dictates the extent to which high permeability layers are engaged in the flow process. Higher levels of model discretization require more detailed parameterization and thus, more computational resources. The latter argument diminishes with the availability of increasing computer processing power, but the need for accurate parameter values increases accordingly.

To represent vertical flow and transport, which is important to surface-vegetation—atmosphere transfers, a one-dimensional approach is often sufficient. Typically, these one-dimensional approaches explicitly represent variably saturated conditions in the profile. Given the availability, relative simplicity, and limited computational demand of freely available numerical models (e.g., Hydrus 1D; Simunek, 2005), considerable insight has been gained into what governs flow and transport in peat profiles. In general, this is done by solving for water flow with Richard's Equation in conjunction with a soil hydraulic property model, such as the van Genuchten—Mualem relationships (Mualem, 1976; van Genuchten, 1980), as discussed in Section 3.2 *Unsaturated Zone Properties and Processes* and represented by Equation 5 and 6. When simulating variably saturated water flow in peatlands, the upper boundary conditions (i.e., precipitation and evapotranspiration) are the key drivers of the hydrologic system.

7.2 Challenges of Numerical Modeling

The challenge to simulate water movement in the upper layers revolves around the exponential decline in saturated hydraulic conductivity. This is compounded by profound changes in saturation associated with macropores prevalent in the upper layers, especially in *Sphagnum* dominated systems. Evaporation from the drained moss surface generates extreme soil water pressures (< 10,000 cm H_2O) that cannot be confidently parameterized in pressure-moisture-conductivity relationships needed to simulate flow. Moreover, while precipitation is rapidly infiltrated, the role of macropores is key—but requires the use of dual permeability functions to simulate appropriately. While these functions are available in the Hydrus 1D model, for example, they have a greater number of parameters that need to be characterized either through measurement or statistical procedures. There is scant information about the values of these parameters in the peat literature, so the use of these functions modeling of water flow in peat and peatlands has been rare.

Like simulating water flow, numerically representing solute transport in peat and peatlands is difficult. The considerations outlined below apply to one-, two-, and three-dimensional approaches, although their application has been primarily in one-dimensional simulations due to the difficulties associated with parameterization. A primary consideration when modeling solute transport in peat is its dual porosity nature, which creates conditions where a given solute can enter the immobile porosity and be removed from the advective flux. Thus, solute transport is commonly characterized by the mobile—immobile solute transport model (van Genuchten and Wagenet, 1989) to account for the immobile porosity within peat (ϕ_{mob} and ϕ_{im} are described in Section 3.1, *Saturated Zone Properties and Processes*). However, there is some evidence that the transfer rate of solutes into immobile porosity can be sufficiently high in some peat that the mobile—immobile solute transport model simplifies to the advection—dispersion model (Simhayov et al., 2018; McCarter et al., 2019) or that peat can be represented by dual permeability models (Liu et al., 2017).

The advection dispersion solute transport model assumes that the mobile porosity is approximately the same as the total porosity, a key assumption that is not generally met when simulating solute transport in peat and peatlands. Conversely, dual permeability models are a group of models that simulate at least two different pore domains with dramatically different hydraulic and/or solute transport properties. For instance, in highly- degraded fen peat, macropores can accelerate the appearance of a solute pulse, but complete breakthrough of the solute plume occurs much later, often exhibiting a multi-modal flushing curve in breakthrough experiments (Liu et al., 2017).

Reactive solutes, such as cations or nutrients, require further parameterization beyond the hydrophysical parameters discussed above (i.e., characterization of biogeochemical parameters of the peat and the solute). Like most peat systems, these parameters are often not well characterized in the literature. Even assuming a conservative tracer such as chloride in peat can result in erroneous results. For instance, fitting of breakthrough curves using one-dimensional modeling has shown that anion adsorption (McCarter et al., 2018) and anion exclusion from narrow throat pores can operate in peat (McCarter et al., 2019), either retarding or accelerating solute breakthrough.

Furthermore, most reactive solutes are either simply modeled with first order decay/production coefficients or adsorption isotherms (often Langmuir, Freundlich, or Linear isotherms), depending on the specific processes in question. Simulating cation transport in peat requires representation of its cation exchange capacity, especially since this is strongly related to organic matter content (Gharedaghloo and Price, 2021). Similarly, first order decay/production coefficients are often used for most reactive solutes, regardless of the complexity of biogeochemical reactions being simulated. Numerical modeling of reactive solutes in peat is often limited by the inability to properly account for the complexity of organic matter composition (peat), mostly due to the lack of a detailed mechanistic understanding of the various biogeochemical processes and associated numerical expressions of such processes in peat.

In spite of the challenges, one-dimensional simulations of vertical flow and transport have been used to illustrate the nature of important hydrological processes in peat and peatlands. One-dimensional models have shown that the pore distribution in mosses controls evaporation from peatlands; certain *Sphagnum* hummock species better maintain moisture in spite of their elevated position (McCarter and Price, 2014); and, air entry pressure is positively correlated with carbon accumulation (Kettridge et al., 2016). However, simulations of capillary rise of water and solutes have shown that even adjacent and visually similar hummocks can have marked variability in hydraulic properties that strongly affect solute distribution (Balliston and Price, 2020).

While laboratory evaluation of hydraulic properties governing water retention and unsaturated hydraulic conductivity are useful for simulating flow conditions in a laboratory column, calibration of a one-dimensional model representing a field site can generate distinctly different parameter values (Elliott and Price, 2020). Simulation of these processes in multidimensional models is possible, but the difficulty in parameterization is compounded by the added spatial variability inherent in all peatland systems.

8 Methods and Approaches

The methods used to measure, monitor, and characterize groundwater processes in peat and peatlands are often derived from other media and geological settings. The unique structure of both peat and peatlands requires the adjustment of many well-established methods when applied to this complex media. Here we give broad descriptions of various methods used in peatland hydrology and, where needed, highlight fundamental methodological developments from the literature.

8.1 Well and Piezometer Installation and Use

Ideally, wells being used to estimate the capacity of an aquifer to deliver water should fully penetrate the aquifer. As many peat deposits are on the order of a few meters thick, this is generally not a problem. Wells in peatlands are normally used to determine the static water table (not aquifer yield), and so a fully penetrating well is not necessary. However, a well should extend past the acrotelm because, by definition, the water table can be below the bottom of the acrotelm. Practically, the well should be deeper to provide stability, ideally anchored in the mineral substrate. Many practitioners will use a 1.5 m length of PVC (polyvinyl chloride) pipe to allow at least 1 m penetration (and 0.5 m stick-up), which is sufficient in bogs and fens. In swamps, the water table can be deeper and so the length should be chosen accordingly.

Piezometers by their nature are designed to monitor hydraulic head at a given depth interval. The length of the slotted intake is up to the user; however, an intake length of 20 cm is often practical for depths within the catotelm (usually < 4 m thick). Even shorter slot lengths (10 cm or less) may be warranted, particularly in the upper layer where hydraulic conductivity can change profoundly with depth. We note that the chosen screen length is a function of the research goals, where smaller screen lengths increase the spatial accuracy of the measurement at the expense of increasing the number of measurement points required to mathematically obtain spatial averages that are otherwise generated by longer screen lengths. While very short screen lengths—or in some cases tubes open only at the bottom (for example to measure only hydraulic head)—can increase spatial discretization, time lag (time required for piezometers to equilibrate) may be excessive and instantaneous measurements may not reflect the current state of the system, particularly where hydraulic conductivity is low. With piezometers, the slotted interval should be centered with respect to the desired depth it is intended to represent (e.g., a 20 cm slotted intake placed 40 to 60 cm bgs represents average head or hydraulic conductivity at the 50 cm depth). Then, as long as the interval is below the water table, the hydraulic head and the head recovery during tests reflects the same part of the peat deposit, with the caveats discussed below.

Measuring the water level within a well or piezometer is often more problematic in peat than it is in mineral soils. This is because the water level will rise as the observer approaches the pipe (the peat is highly compressible) and electronic devices often do not work in bogs due to the very low ionic concentrations in the water. The high compressibility poses a related problem: the pipe elevation (and stick-up) can change between measurements. As the water table declines over the season, the peat surface can decline. For short pipes, their elevation may decrease accordingly. For longer installations, the pipes may be stable, but when the surface elevation decreases the stick-up increases. The seasonal surface fluctuations are generally reversible. It is good practice to place a ring or marker near the base of the pipe and manually measure stick-up each time a head measurement is made. Also, due to the uneven nature of the peatland surface, the stick-up always needs to be measured on the same side of the pipe. If the pipe length is sufficient (i.e., the base is stable), then the change in surface elevation can be deduced as discussed by Price and Schlotzhauer (1999).

Shallow pipes are also susceptible to frost heave. Anchoring the tube to a metal rod pounded into the mineral substrate can reduce their movement. Nevertheless, it is good practice to survey the pipe-tops each season if multiyear measurements are being made or if vertical or horizontal hydraulic gradients are being assessed. Use of a logging pressure transducer suspended from the well can reduce errors caused by the weight of an approaching observer. However, it is still subject to well/piezometer stability.

Surface level can also be measured with an ultrasonic sensor but is only necessary when surface elevation changes are large and part of the monitoring objective (Fritz et al., 2008). For water level measurement in a bog, where no electrical current passes through the weak ionic solution, a blowstick can be used. A blowstick is a length of flexible tube inserted into a slightly larger diameter graduated rigid tube; or a flexible graduated tube that allows the user to hear bubbling caused by blowing through the tube as it is inserted. The accuracy of these devices is about ± 5 mm, although better accuracy can be achieved by consistent use by the same operator.

Inaccuracy can be problematic in peatland studies if the intention is to measure vertical hydraulic gradients, which are commonly of a similar magnitude as the error of the measuring device, notwithstanding the errors caused by subsidence/compression and surveying. Calculating the vertical gradient between widely spaced piezometers (≥ 1 m) reduces the effect of the measurement error, proportionally. Vertical gradients can also be determined by measuring the hydraulic head in a piezometer relative to the local water table, measured in a well or shallow pit in the acrotelm. To reduce the errors associated with the compressible nature of peat, local boardwalks (e.g., 15 to 30 cm wide horizontal boards affixed to vertical wood pilings driven down to the mineral soil beneath the peat) are recommended so practitioners are not standing on the peatland surface.

8.2 Sampling Peat

Sampling soil for hydrophysical properties requires that the soil structure remains undisturbed. Consequently, peat soils require special considerations when sampling for hydrophysical properties because peat is a plastic, elastic, and compressible media that makes extracting undisturbed and intact samples difficult. This is particularly true for the easily disturbed delicate near surface mosses.

The easiest method for removing intact and undisturbed cores is with scissors or a serrated knife and guide housing, where the guide (e.g., a piece of PVC pipe, or stove pipe ideally ≥ 5 cm diameter) is gently placed on the peat surface, while cutting around the guide housing, which is then gently pushed into the ground, with care taken not to deform the delicate surface of the peat (Figure 23). Slight twisting of the guide housing can help sink the guide while reducing deformation. Once the desired depth has been reached, the peat outside and around the guide housing is excavated and the peat sample is carefully cut along the bottom of the guide housing to separate it from the peatland. This method works best for near surface (0 to 40 cm) peat and with shorter cores (< 20 cm).

Figure 23 - Sampling of the upper layer of bog peat using a 10 cm diameter, 5 cm tall PVC ring, using a knife and scissors. Lower layers can be sampled sequentially. Larger samples can be cut with serrated knives or a handsaw; storage in a rigid cooler or container is advised. (Photographs by J. Price, C. McCarter, and W. Quinton)

Sequential sampling with this method can be used down to the water table. A box-corer can also be used, such as a Wardenaar corer, which has sharp metal edges that alternately cut into the peat and is then squeezed to retain the sample as the sampler is withdrawn. If the climate allows, taking large samples in the winter (when the peat is frozen) can minimize compression, and subsamples can be cut with greater precision.

Sampling while the peat is frozen allows for much larger sample sizes (e.g., dimensions of a chest cooler) to be extracted without disturbing the peat structure. This method is best completed using a chainsaw to cut through the frozen surface peat. First, the desired outline dimensions are cut into the frozen peat, then a large area surrounding the desired sample is cut. These outer areas are removed to allow vertical access to the peat sample. The sample can then be cut horizontally from the bottom using a handsaw and the sample removed. The sample must be trimmed along the edges cut by the chainsaw to remove any disturbed peat. Although this method can facilitate large and undisturbed peat cores, it damages the surrounding peatland and should be done judiciously.

For sampling deeper peat, Russian and piston corers can be used. These are described by Pitkänen and others (2011) as well as Shotyk and Noernberg (2020), who compare volumetric sampling of peat using box, Russian, and piston corers. In all cases, near surface deformation of peat occurred, so one of the sampling techniques described above should be used in conjunction with a coring device.

8.3 Parameterization of Peat

Parameters that characterize the properties of peat are subject to considerable spatial variability, both between and within peatlands. Various parameters including hydraulic conductivity, porosity, water retention, and drainable porosity can be distinctly different in adjacent, visually similar hummocks. Awareness of the scales of variability of parameters governing the behavior of peat is important in predicting or interpreting its hydrology.

8.3.1 Hydraulic Conductivity

K_{sat} is commonly evaluated using the bail test method of Hvorslev (1951). To characterize the hydraulic conductivity in the acrotelm, a smaller slot length is generally required as discussed in Section 8.1, *Well and Piezometer Installation and Use*. This has the dual purpose of increasing the detail in the zone where the range of K_{sat} is highest and slowing the recovery rate, since this will likely be fast in the upper layers.

Hvorslev (1951) suggests slot length should be four times greater than pipe diameter to minimize error. Using this method with sufficient sampling depths generates a K_{sat} profile, that can be used to produce a transmissivity function (McCarter and Price, 2017c). A transmissivity function is important for calculating horizontal flow, given its sensitivity to water table elevation within the profile as discussed in Section 3.1, *Saturated Zone Properties and Processes*. Sometimes a well can be used to generate a transmissivity function (e.g., Price and Maloney, 1994). To achieve this, a bail test is required at the full range of water table elevations because the head recovery of a pumped well will depend strongly on the position of the water table, given the extreme vertical change in hydraulic conductivity in the acrotelm. In this case, the well will need to penetrate into the catotelm peat.

Because peat deposits are thin and near to the ground surface relative to most mineral aquifers, they commonly experience a large range in temperature, both seasonally and vertically within the peat profile. Head recovery can vary appreciably with temperature because temperature affects viscosity. *Dynamic viscosity* (μ; often expressed in units of Pascal seconds), which is a multiplier in the formula for *relative permeability* (k; often expressed in square meters), increases by 50 percent in water at 5 °C compared to viscosity at 20 °C, with a proportional reduction in permeability, thus hydraulic conductivity. While this is not relevant to the measured value of K_{sat} at one location in a peat profile, it becomes relevant when comparing peat properties down the profile or between peatlands, given their temperature differences. Comparing k, instead of K_{sat}, avoids this complication.

Laboratory determination of peat K_{sat} using a permeameter device follows the usual protocols used for mineral soils. However, core shrinkage or compression during sampling, transport, and preparation—including volume change on thawing—can lead to bypass flow down the inside wall of the permeameter. Paraffin wax (Hoag and Price, 1997) or Parafilm™ (McCarter et al., 2019) can be used to confine water within the peat sample during the tests.

Determining unsaturated hydraulic conductivity using fixed-plate pressure cells is problematic because peat shrinks away from the (upper) porous plate at lower (more negative) pressures (ψ). Using a peat sample between a floating upper pressure plate and fixed lower pressure plate (Price et al., 2008) can be used to control head differences, and thus calculate K_{unsat} at a range of pressure heads or moisture contents. Directing the water downward through the sample to generate a unit head gradient is the preferred method (McCarter et al., 2017c).

Alternatively, K_{unsat} can be determined using the evaporation method (Schindler et al., 2010). This requires the use of tensiometers to determine head (gradients). Using tensiometers in peat is common practice, but tensiometers do not work well in poorly decomposed peat (moss) because of poor contact between the peat (moss) and tensiometer cup. Tension infiltrometers provide an alternative approach that can be used in the laboratory or field. The tension infiltrometer releases water at a rate slower rate of seepage from ponded water by maintaining a small negative pressure on the water moving from a disk placed on a level peat surface.

8.3.2 Water Retention

Water retention experiments (including related tests such as for K_{unsat}) suffer from sample shrinkage at lower pressures. The common practice is to express the volumetric water content (θ_v) relative to the original (i.e., saturated) volume of the soil. However, the relation to the degree of saturation is non-linear and therefore not commonly established.

Estimating water retention at high pressure (high θ_v) requires a pressure plate or porous disk with high air entry pressure (e.g., Price et al., 2008).

For poorly decomposed (especially moss) samples, it is essentially impossible to measure ψ - θ_v points for $|\psi|$ less than the sample length, since water drains immediately to the base of the sample (Golubev et al., 2021). As a consequence, many retention curves for poorly decomposed peat do not exhibit an air entry pressure as shown by Figure 16. For lower pressures (low θ_v), a pressure chamber can be used, but extreme shrinkage can occur.

8.3.3 Bulk Density and Porosity

Bulk density (ρ_b) is generally based on the dry mass of solids and original (field) sample volume. To avoid combustion, peat samples should be dried at 95 °C or less for 24 hours or until a stable mass is achieved. Total porosity (ϕ_t) is equivalent to the saturated water content (θ_s). This can be evaluated directly in the laboratory based on the saturated mass (minus dry mass, accounting for sample volume and water density). However, sample swelling at saturation can confound the volume, and may result in apparent ϕ_t or $\theta_s > 1$, which is impossible. In this case, ϕ_t can be calculated on the basis of ρ_b such that $\phi_t = 1 - \rho_b/\rho_p$, where ρ_p is particle density. However, the range of particle density for peat can vary from ~0.9 to 1.5 g cm^{-3} (Gharedaghloo and Price, 2021; Redding and Devito, 2006), and < 0.9 to ~0.7 g cm^{-3} for undecomposed and lightly decomposed *Sphagnum* mosses (Whittington et al., 2021), so particle density should be assessed to acquire confidence in this method.

In peat, which comprises mobile (ϕ_{mob}), or immobile (ϕ_{im}) pores as shown in Figure 11, determining the distribution of porosity is methodologically challenging. McCarter and others (2019) found mobile porosity coincides with the drainable porosity at (ψ) = -100 cm, which is commonly measured during soil water retention experiments.

8.3.4 Specific Yield and Drainable Porosity

Specific yield (S_y) is the ratio of the volume of water that can drain by gravity from a saturated volume of material to the total volume of that saturated material. As such, S_y relates the change in water table elevation to change in storage and can be measured as the difference between the saturated moisture content and the moisture content following gravity drainage.

Specific yield is an important parameter in modeling saturated peat and estimating water budgets. Freeze and Cherry (1979, Section 2.10, subsection *Transmissivity and Specific Yield in Unconfined Aquifers*↗) characterize S_y as a profile property rather than a property of a distinct layer. While many researchers have reported S_y for specific layers of peat, such values actually refer to the drainable porosity (ϕ_d). The drainable porosity within the acrotelm can range from 0.45 near the surface to 0.048 at the base (Rezanezhad et al., 2016).

The specific yield of a peatland is thus the integrated value of a range of drainable porosities.

Determination of S_y can be done by comparing the amount of water added or lost from the peat profile (e.g., by measured rainfall or evapotranspiration) versus the measured water table elevation change. For example, the rain-to-rise ratio can be used, which is the amount of rain divided by the consequent water table rise (Dettmann and Bechtold, 2016). In the laboratory, the drainable porosity (ϕ_d) of a specific layer can be determined as the difference in the volume of water in a saturated, then drained, peat sample (typically drained for 24 hours), with respect to the total sample volume.

9 Wrap-up Section and Research Needs

Peatlands are shallow unconfined aquifers. Groundwater in peatlands embody the principles of an ecohydrological system including their development and function and, in turn, their groundwater relations. In a given hydrogeomorphic setting, subject to climate, groundwater exchanges are controlled by the character of adjacent and underlying mineral aquifers and aquitards, such that incipient peatlands form where persistent saturation hinders decay of plant material. This then forms the matrix, peat, that hosts the groundwater, in an ever-evolving ecohydrological cycle.

While the principles of groundwater flow and storage are not different from those in mineral matrixes, the physical and chemical properties of the peat matrix impart distinct characteristics and challenges to measuring and understanding groundwater relations. The distinct characteristics arise from the relative instability of the medium, which undergoes physical breakdown and consolidation on time scales orders of magnitude shorter than for mineral materials. The surface chemistry of peat particles facilitates oxidation-reduction reactions, imparts enormous cation-exchange capacity, and changes the wetting behavior that controls capillary relations. The interaction of peatlands with adjacent groundwater systems dictates their class, form and function (e.g., bog versus fen versus swamp) because groundwater inflows offset the tendency for acidification caused by organic acids released when peat and plant matter decay. The outcome of this chemical balance and wetness condition dictates the plant community composition (species) and form (e.g., mosses, sedges, or woody plants), which feeds back to the hydrology of the system by influencing the structure of the peat matrix, and water exchanges by evapotranspiration and runoff.

The challenges facing researchers and practitioners charged with evaluating peatland function span the range of scales from pore spaces to entire ecosystems. This book on groundwater in peat and peatlands has highlighted many of the important groundwater processes. However, there remain many uncertainties that present research challenges. Some of these uncertainties and associated challenges are discussed below.

The effect of global warming on peatland biogeochemistry and ecohydrology is much more complex and uncertain than described herein; changes in atmospheric temperature are predicted more confidently than changes in precipitation. The well-known Clausius-Clapeyron Equation indicates that vapor pressure rises non-linearly with air temperature and, as such, a warmer climate implies a potentially wetter atmosphere. However, it is not clear if the potential increase in precipitation is sufficient to offset potentially higher evapotranspiration losses, or whether systems will be required to adjust to a persistently drier state.

Nor are the hydrological impacts of potential changes to the seasonality of precipitation and temperature certain across the range of climates in which peatlands are found. For instance, higher latitude systems in subarctic Canada will likely have a shorter period of snowpack accumulation, thus less snow to melt and longer growing seasons. Vegetation shifts will be likely; drier systems will have more trees, thus more interception of precipitation and greater transpiration, contributing to a positive feedback loop that increases peat decomposition, alters hydraulic properties, and so on. Such shifts may produce compounding disturbances with increased frequency and intensity of peatland wildfires or other natural and human disturbances that can further alter the processes that sustain peatland functions.

The class of peatland could change, for example, if the weak groundwater inflow to poor fens ceases; then they will evolve to bogs. Increased decomposition rates could slow or even reverse the development of domed bogs, reconnecting such areas to groundwater systems. Protracted periods of summer drying that enhance woody vegetation may result in fens transitioning to swamps, or for peat swamps to lose organic soil and become mineral swamps. In short, our ability to predict these changes fundamentally relies on the mechanistic understanding of peat and peatland ecohydrology and its feedbacks with biogeochemical processes. Yet, the linkages between ecohydrology and biogeochemical processes remains a critical research question.

Peatland-scale processes that control flow direction, rates, and persistence produce feedback that results in distinct, patterned peatland forms. The cause of patterning in peatlands (ridge/flark, hummock/hollow) remains somewhat speculative—a feedback mechanism among biotic productivity, decay, water table elevation, and water flow. In the case of ridge-flark microtopography in fens, the orientation of ridges perpendicular to flow decreases the rate of drainage, increases surface (i.e., depressional) water storage, and enables a threshold (i.e., "fill and spill") runoff response once the water storage capacity of each flark is exceeded. However, ridge/flark systems such as ladder fens are a conduit for water loss from large domed bogs (as shown in Figure 7), capable of high water transmission rates following snowmelt or heavy rainfall.

While the ecohydrological and biogeochemical feedbacks that produce microtopography are uncertain, development of microtopography on measurable time scales is possible. Long-term monitoring of water levels over decades, combined with analysis of archived air-photograph or satellite images over the same period, indicate that microtopographic change accompanies persistent changes in peatland water storage. For example, *Sphagnum* lawns have been found to develop hummock-hollow topography in cases where peatlands lose water through sustained high rates of drainage. Such changes enable colonization of peatlands by trees on the relatively dry hummock surfaces, which presents the potential for feedback resulting in drying of the peatland through increased evapotranspiration. More research is required to evaluate whether such a process driven by increased drainage might also result from increased evapotranspiration driven by climate warming, and what those changes imply for downgradient ecosystems.

Like the flow of water in peat and peatlands, solute transport is complicated by ecohydrological feedbacks and linkages. However, studies of solute transport in peatlands have only recently begun to consider complex peat structures and peat surface chemistry. Solute transport in peat is not only subject to dual/multi-porosity processes; as an organic substrate, it is highly reactive with a wide range of chemicals, compounds, and elements. This makes understanding reactive transport in peat difficult but critical if we are to understand the feedbacks between ecohydrology and biogeochemistry that govern many key peatland processes.

At the peatland scale, it is thought that the linkages between hydrophysical peat properties, microtopography, and the movement of nutrients and carbon gives rise to regions of elevated nutrient/carbon cycling, resulting in both positive and negative feedbacks. However, the relative strength and importance of such processes and feedbacks have yet to be resolved. Understanding the movement of nutrients, carbon, and other elements/compounds from the pore to landscape scale of peatlands underpins much of our collective understanding of peatlands, but current limitations to our knowledge of processes in peat limits our understanding of peatlands.

The high compressibility of peat has been described. However, apart from the suggestion that decomposition decreases mean pore diameters and overburden increases pore compression—both of which reduce K—there is still much more to learn and know. For example, specific yield is almost exclusively used to relate water exchanges to water table position, but, given the compressibility of certain peats, coupled with extreme water table drawdown in some settings, the inclusion of specific storage may be essential to evaluating water storage changes. However, this approach has not been extensively adopted by the peatland hydrology community.

Peat's high compressibility results in *mire breathing* that causes the water table to be closer to the surface than it otherwise would be in a more rigid media, resulting in hydraulic properties that vary over short (hours to days) time scales. Characterization of hydraulic properties such as bulk density and porosity are mostly based on a fixed, sampled field volume, but peat cores shrink as soil water pressure decreases, so expressing porosity of a sample under tension, based on field volume of the sample, understates the proportion of saturation compared to what would be calculated if the reduced sample volume were used. Explicitly accounting for volume change in estimation and expression of hydraulic parameters needs more attention.

Advancements in hydrological modeling of peatlands are ongoing in several key areas. At the plot and point scale, modeling has focused on simulating the flux and storage of water, solutes, and energy, typically in one dimension. Modeling can incorporate mobile versus immobile porosity mechanistically, to demonstrate and evaluate the partition of water accordingly, in a peat matrix. However, some simulations, so presumably some peats, do not exhibit this behavior. A better understanding is needed about which peats (e.g., *Sphagnum* peat, woody peat) and their state of decomposition require complex porosity to be considered.

Studies undertaking 2- and 3-D simulations in peat and peatlands remain rare, yet the addition of extra dimensionality to modeling of water, solute, and energy fluxes will likely reveal new understandings of peatland hydrology, just as 1-D models substantially advanced understanding over the last few decades. At regional scales, groundwater modeling has focused on representing wetlands in land surface schemes to improve coupling of hydrological with atmospheric models. Incorporating feedbacks without unmanageable complexity is essential to realistically incorporating peatland functions into global climate models.

Modeling and quantification of water and solute fluxes in peat and peatlands requires a suite of parameters; ideally, these would be measured but commonly are estimated or taken from literature describing other sites. Unlike mineral soils with measurable components (percent clay, percent slit, percent sand, organic matter content, and bulk density) that facilitate the development and widespread adoption of pedotransfer functions, peat is predominantly organic in nature with very little mineral component. Several properties of peat lend themselves to the development of pedotransfer functions, however. Since the degree of decomposition increases with depth, there a systematic change in physical, hydraulic, and thermal properties of peat. As such, the degree of decomposition and bulk density as well as other key properties including the botanical origin of peat and its moisture content can be used to infer other properties that control the flux and storage of mass and energy. However, linking measurable chemical indices of the organic peat—i.e., carbon isotope ratios or C/N ratios—to hydrophysical properties has not been done. Thus, our lack of knowledge on the range of physical, chemical, and hydraulic properties across all peat types limits our ability to develop a universal pedotransfer function.

Our understanding of peatland hydrology is dominated by what we have learned from northern peatlands. This reflects not only the large area occupied by peatlands in the northern hemisphere but also the capacity of northern researchers to generate funding and regional interest in peat and peatlands from a scientific, industrial, and social establishment. Now, interest in the large peatlands of the subtropical and tropical regions is growing. Global interest has increased with recognition of their role in the global carbon budget because of air-quality impairment from extensive peat-fire smoke plumes, such as from Sumatra in 2015, and the role of land-clearance and agriculture in their demise.

Many processes governing the occurrence, growth, and degradation of tropical peatlands; mechanisms controlling water and solute flows; and even some basic approaches to restoration are shared with better-studied northern peatlands. However, the distinct climate, botanical origins of peat thus hydraulic structure, and scales of exploitation for resource development of tropical/subtropical peatlands are very different and require exclusive focus and extensive research. While an excellent cadre of scientists have reported on tropical peatland form, hydrology, and carbon biogeochemistry, the global importance of these peatlands warrants increased research effort.

10 Exercises

The exercises in this section provide an opportunity for readers to explore some common activities for estimating water, solute, and heat flow parameters and fluxes in peat. A summary of each exercise follows.

Exercise 1 invovles calculating hydraulic conductivity in peat usin the method of Hvorslev (1951) and data from a slug test. Two examples are provided. The first one requires a standard solution for a piezometer response common in porous media. The second example is for a response showing strong evidence of peat compressibility, which requires manual adjustment of the bail-test response curve to force the solution to fit the tail of the curve.

Exercise 2 estimates the flow of groundwater and a reactive solute from a peatland to a stream channel, illustrating that the decrease of K_{sat} with depth causes flux to be strongly dependent on the water table position. Depth-dependent values of K_{sat}, porosity (ϕ_{t}), mobile porosity (ϕ_{mob}), bulk density (ρ_b), distribution coefficient that controls solute retardation (R_f), are used to determine the solute flux.

Exercise 3 illustrates the importance of accounting for the character of the K_{sat} distribution, by posing a question similar to Exercise 2 with a more dramatic decrease in K_{sat} with depth.

Exercise 4 evaluates the change in thermal properties of a peat soil as it thaws, transitioning from saturated and frozen, to saturated and unfrozen, and then to unsaturated and unfrozen. The purpose is to illustrate how this affects heat flow in to and out of the ground.

Exercise 5 demonstrates the impact of changing soil water content on permafrost thaw which creates a feedback loop in peatland systems that promotes permafrost loss.

Exercise 1 - Determining Saturated Hydraulic Conductivity of Peat Using the Hvorslev Method

Determining hydraulic conductivity (K_{sat}) of peat can be done using the method of Hvorslev (1951), whose application was described by Freeze and Cherry (1979) for mineral aquifers. The method for determining K_{sat} in peat is identical and is based on timing the head recovery of a well or piezometer that has been bailed. The head recovery is non-linear, slowing exponentially as the water level in the well or piezometer equilibrates with the surrounding hydraulic head; thus, it should plot against time as a straight line on a semi-logarithmic plot. However, often the piezometer response in peat is not even log-linear, because a key assumption governing the response is often not valid. That is, peat is highly compressible compared to mineral sediments, so creating a large head gradient by bailing the piezometer causes the peat to compress and expel water into the piezometer rather than coming from the assumed "semi-infinite" expanse of the aquifer. This results in the pipe filling (head recovery) faster than it otherwise would, thus overestimating the "true" K_{sat} (Rycroft et al., 1975).

In this problem set, two data sets for bail tests in a peatland (ridge and flark of a patterned fen in northern Alberta) are provided. One has the expected log-linear response and another shows evidence of peat compression, in which there is not a log-linear response. This exercise demonstrates how to determine a reasonable K_{sat} value for both cases.

Bail tests involve removing water from a well or piezometer and monitoring the response. Here, we focus on using a piezometer, as the calculations are simpler because it has a fixed intake length. Comments on adapting the method to a well follow, the method being identical except for one aspect.

For a bail test, the more rapidly the water level in the pipe recovers, the higher the hydraulic conductivity. The measurements include H, which is the total hydraulic head at the start of the test (i.e., before water is removed) and H_0, which is the total hydraulic head value immediately after water is bailed from the well or piezometer (time = 0). The variable in this exercise is the time-dependent hydraulic head as the well or piezometer recovers.

Other key values include the dimensions of the intake, including its inner radius, r, and the external radius and length of the intake: R and L_e, respectively. These parameters are illustrated in the image associated with the equation for K_{sat}. The method of Hvorslev (1951) relies on an empirical shape factor for this type of piezometer to identify a time lag parameter, T_0, which is calculated as the time at which the dimensionless recovery $(H\text{-}h)/(H\text{-}H_0)$ reaches a value of 0.37. These values are used in the following equation developed by Hvorslev to determine K_{sat}.

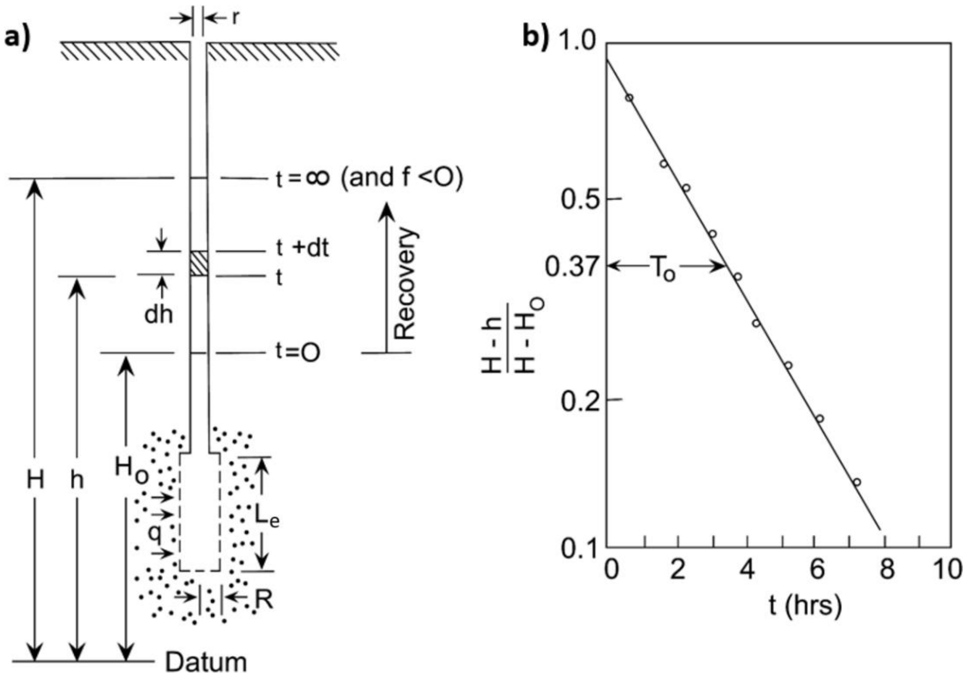
$$K_{sat} = \frac{r^2 \ln (L_e/R)}{2LT_o}$$

where:

K_{sat} = saturated hydraulic conductivity (LT^{-1})

R = external radius of the intake (L)

L_e = length of the intake (L)



a) Identification of values H, H_o, and h, at different times (t), as well as r, R, and L_e; b) K_{sat} is determined by the rate of recovery, which is portrayed here on a log-linear plot, from which T_o is estimated from a point on the line corresponding to $(H\text{-}h)/(H\text{-}H_o) = 0.37$. Diagram from Freeze and Cherry (1979).

As noted above, the response is commonly non-linear in particularly compressible peat. In this case, a straight line is drawn from the coordinates at the start of the test (1, 0), that is parallel to a line tangential to the asymptote of the recovery curve. This is described by Hvorslev (1951, page 41). Essentially, this approach uses the latter portion of the recovery data that better reflects the properties of the peat, rather than the initial portion that is an artifact of the test. To use this approach, plot the dimensionless recovery data $(H\text{-}h)/(H\text{-}H_o)$ versus time, estimate the slope of the lower (ideally straighter) section of the curve, and draw a line parallel to this lower part of the curve, with its origin at 1,0 on the semi-log plot (or 0.0 on a linear plot). Then determine the time lag parameter (T_o) from a point on the line corresponding to $(H\text{-}h)/(H\text{-}H_o) = 0.37$ on the semi-log plot. Inevitably, different users will estimate a slightly different slope of the tail portion of the curve, and thus calculate different values of K_{sat}. The result, however, will be closer to the true value than an uncorrected estimate.

Bail tests were done on a rib and flark of a ribbed fen in northern Alberta, Canada. The piezometer intakes were both 1.81-2.31 m below local ground surface. PVC pipes with a 5 cm-inside diameter were pushed into a pilot hole to the required depth; thus, their outside diameter (6.4 cm) represents the outside radius of the tube. In these examples, a logging pressure transducer was used to record the head recovery.

1) Access the data set #1 (Peat&Peatlands-Exercise-1-DataSets1&2.xlsx)↗ to calculate K_{sat} (m/d) for the piezometer located 1.81-2.31 m below ground surface in a rib of a patterned peatland.
2) Access the data set #2 (Peat&Peatlands-Exercise-1-DataSets1&2.xlsx)↗ to calculate K_{sat} (m/d) for the piezometer located 1.81-2.31 m below ground surface in a flark of a patterned peatland.

Click for solution to Exercise 1⤵

Exercise 2 - Calculating Water and Solute Flux from a Peat Deposit with Weak Depth Dependent Hydraulic Conductivity

The movement of contaminants is highly dependent on the hydrophysical processes operating within a peatland. Here we consider a peatland with a constant contaminant source some distance upgradient from an adjacent stream. The contaminant is weakly adsorbing and has reached geochemical equilibrium in the peatland.

The sorption of the contaminant can be represented by a linear adsorption isotherm with no limitation on adsorption sites, thus a retardation coefficient (R_f) can be determined as follows.

$$R_f = 1 + \frac{\rho_b K_d}{\phi_t}$$

where:

R_f = relative velocity of the solute with respect to the water (retardation coefficient)

ρ_b = bulk density (g cm^{-3})

ϕ_t = total porosity

K_d = adsorption coefficient (cm^{3}gm^{-1})

To simplify the evaluation, we assume the contaminant has come to equilibrium with the peat and that the amount of contaminant in the mobile water fraction (C_f) is proportional to the inverse of the retardation factor.

Under the high water table regime, the water table is at the peatland surface and the horizontal hydraulic gradient in all layers is (i) is 0.004. Under the low water table regime, the water table is 20 cm below the peat surface and the horizontal hydraulic gradient for all layers is 0.001. The relevant hydraulic and transport parameters are provided in the following table.

Soil and solute parameters

Peat Layer Depth	**0–10 cm**	**10–20 cm**	**20–30 cm**	**30–50 cm**	**50–150 cm**
Layer Thickness (cm)	10	10	10	20	100
ϕ_t	0.83	0.83	0.8	0.78	0.65
ϕ_{mob}	0.45	0.45	0.4	0.35	0.2
ρ_b (g cm^{-3})	0.12	0.12	0.15	0.18	0.25
K_{sat} (cm d^{-1})	2000	1500	1200	300	83
K_d (cm^{3}g^{-1})	15	15	17	23	35
R_f	3.2	3.2	4.2	6.3	14.5

Determine the proportion of contaminated water that will be exported to the adjacent stream under high and low water table regimes in several distinct peat layers.

Assume:

1) The flow within the peatland is parallel to the water table and that the bottom of the peatland is no-flow boundary.
2) The water table remains constant.
3) The width (w) of the flow face between the peatland and stream is 20 m.
4) The flow conforms to Darcy's Law.
5) There is a constant amount of contaminant at the interface but the concentration of contaminant in the pore water is 1 for a layer with a retardation factor of 1, while the concentration in the pore water is $1/R_f$) for layers with $R_f > 1$.

Contrast the two water table regimes and calculate the difference in proportion of contaminated water between the two scenarios. What proportion of the contaminant moves through the upper peat layer? How does this compare to the water flow?

Click for solution to Exercise 2↴

Exercise 3 - Calculating Water and Solute Flux with More Dramatic Decline in Hydraulic Conductivity

Peatlands can have a variety of hydraulic conductivity distributions ranging from relatively monotonic declines with depth—as calculated in Exercise 2—to exponential hydraulic conductivity profiles. In Exercise 2, a relatively monotonic decline in hydraulic conductivity drove slight differences in the proportion of both water and contaminant transport in the upper 30 cm of peat. Under low water table conditions, the proportion of water and contaminant in the saturated upper layer (20 – 30 cm) increased while the total mass of contaminant exported to the stream decreased by ~1 order of magnitude. Peatlands with a more pronounced exponential decline in hydraulic conductivity are common and can have extremely different transport behavior.

Using the same values for the following peatland hydrological parameters: K_{sat} (thickness weighted average = 409 cm d^{-1}), hydraulic gradients, flow-face dimensions, peat layer depths, and water table elevations; as well as the updated hydrophysical and reaction parameters provided in the following table; calculate the proportion of water flow and contaminant under both wet and dry conditions.

Soil and solute parameters.

Peat Layer (*d*)	**0–10 cm**	**10–20 cm**	**20–30 cm**	**30–50 cm**	**50–150 cm**
Layer Thickness (cm)	10	10	10	20	100
n_t	0.98	0.93	0.91	0.85	0.70
n_{mob}	0.40	0.35	0.25	0.20	0.10
ρ_b (g cm^{-3})	0.05	0.07	0.09	0.11	0.18
K_{sat} (cm d^{-1})	5000	1000	100	10	1
K_d (cm^3 g^{-1})	2	8	12	18	25
R	1.1	1.6	2.2	3.3	7.4

Compare and contrast the peatlands in Exercise 2 and Exercise 3. Think about how the shift to an exponential hydraulic conductivity distribution affects the release of contaminants to the adjacent stream. Which change in parameters, the hydraulic conductivity or partitioning coefficient/bulk density, has a greater impact on solute transport?

Click for solution to Exercise 3⤵

Exercise 4 - Changing Thermal Properties of Peat as it Thaws

As a peat profile thaws, it changes state from being:

- saturated and frozen; to
- saturated and unfrozen; and then to
- unsaturated and unfrozen (in other words, the soil thaws, then drains).

1. The volumetric heat capacity (J m^{-3} C^{-1}) is the quantity of heat needed to raise the temperature of a unit volume of soil. Compute the values of the volumetric heat capacity of a peat soil for each of these three stages using the appropriate choice from the following equations.

$$Cv_{WSA} = Cv_W\,(\theta) + Cv_S\,(1 - \phi_t) + Cv_A\,(\phi_t - \theta)$$

$$Cv_{IWS} = Cv_I\,(\phi_t - \theta) + Cv_W\,(\theta) + Cv_S\,(1 - \phi_t)$$

$$Cv_{WS} = Cv_W\,(\phi_t) + Cv_S\,(1 - \phi_t)$$

where:

C_v = volumetric heat capacity

θ = volumetric soil water content

ϕ_t = total porosity

W = subscript referring to water

S = subscript referring to soil

A = subscript referring to air

I = subscript referring to ice

Assume the soil profile is entirely peat. The results of Hayashi and others (2007) suggest the liquid soil moisture content for frozen soil is in the range of 0.15 to 0.2 (assume 0.2 for this case) and the drained water content is 0.5. For each constituent, the values of specific heat, cp (J kg^{-1} C^{-1}), mass density, ρ (kg m^{-3}), porosity, ϕ_t (-), and thermal conductivity, k_t (W m^{-1} °C^{-1}) are given in following table. Volumetric heat capacity is the product of specific heat capacity and density.

Thermophysical properties of constituents

	Specific heat	Density	Porosity	Conductivity
Air	1010	1.2	-	0.025
Ice (at 0º C)	2120	920	-	2.2
Water	4185	1000	-	0.57
Peat	1920	40	0.9	0.25

2. How would the transition from one condition to the next affect temperatures within the peat profile? What factors might effect the rate of transition from one condition to the next, and therefore the rate of temperature increases in the peat profile? For those who have access to commercial publications of the Wiley publishing house, Hayashi and others (2007) provides a thorough discussion of this topic.
3. Another important thermal property of porous media is its thermal conductivity (W m^{-1} $°C^{-1}$), defined as the amount of heat transferred through a unit area per unit time under a unit temperature gradient. Considering the very large difference in thermal conductivity values presented in the table above, describe how the transition from one compositional stage to the next effects the rate of ground thaw.

Click for solution to Exercise 4⤵

Exercise 5 - Effect of Moisture Content on Thaw

Permafrost underlies a large proportion of peatlands in the Boreal and Taiga ecozones. Permafrost thaw occurs when the thickness of ground thawing in summer exceeds the thickness of ground refreezing during winter. Explain how a change in soil moisture content can result in permafrost thaw. Reading the article by Connon and others (2018) is helpful when completing this exercise.

Click for solution to Exercise 5↴

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12 Boxes

Box 1 - Height of Domed Peatlands

Peat accumulates in a saturated environment because of the slow rate of decay of organic material in the catotelm. Away from the lagg (interface between bog and adjacent mineral terrain), toward the center of the peat massif, the horizontal hydraulic gradients are lower, thus drainage is slower, so more peat accumulates. This may be exacerbated by the central massif's isolation from minerotrophic water, unlike nearer the lagg where solute-rich water may accelerate decomposition and support vascular plants less resistant to decay than *Sphagnum*. These gradients and flows are illustrated in Figure 6.

The height of a bog dome can be greater in large systems—areas of higher precipitation excess over evapotranspiration—and where saturated hydraulic conductivity of the peat is low (slower drainage). Ingram (1982) provided a simplistic analytical model relating these parameters to bog dome height, although its simplicity is perhaps too much to be of practical importance (Belyea and Baird, 2006). In practice, the height to which a dome can reach is finite, because slow decomposition of a very thick peat deposit, even though saturated, eventually degrades an equivalent amount of organic matter as is added annually (Clymo, 1984).

Clymo (1987) estimates a range of heights between 0.5 to 10 m for temperate or boreal peatlands, although most are < 5 m. In tropical peatlands, where rainfall can be much higher, peat deposits of ~20 m can occur (Anderson, 1983), thus greater dome heights are possible.

Return to where text linked to Box 1⮥

Box 2 - Permafrost Peatlands

The boreal and subarctic regions contain approximately half of the earth's peatland area. In these regions, permafrost is typically discontinuous, with higher concentrations in low-lying terrains where peatlands predominate. In such terrains, permafrost is often restricted to below forested bogs known as *peat plateaus* that rise 1 to 2 m above permafrost free, often treeless, wetland terrain of collapse scar wetlands and channel fens (Figure Box 2-1). Unlike their surrounding wetlands, peat plateaus contain a well-developed vadose zone. Being a relatively dry layer, the vadose zone has a very low thermal conductivity, a property that enables it to thermally insulate and preserve the underlying permafrost, even in regions where the mean annual air temperature is close to or even above the freezing point.

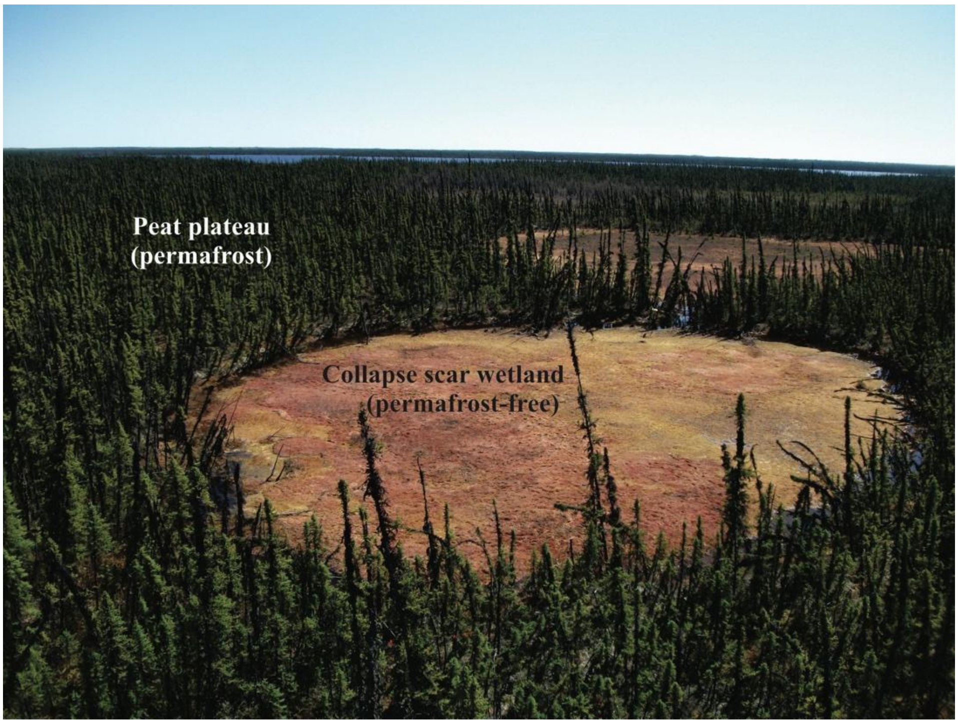

Figure Box 2-1 - Collapse scar wetlands (permafrost free) formed within a tree covered peat plateau. Scotty Creek, Northwest Territories, Canada. (Photograph by R. Connon)

Peat plateaus and collapse scar wetlands are typically arranged into distinct *plateau wetland complexes* separated by channel fens that collect the water draining from the complexes and convey it to the basin outlet. Because of their higher (1 to 2 m) topographic position, sloping surfaces, and the presence of relatively impermeable (ice-rich) permafrost near the ground surface, suprapermafrost (i.e., above the permafrost layer) groundwater drains either from the forested peat plateaus into the collapse scar wetlands or directly into the channel fens. The very high permeability of peat plateau surfaces precludes overland flow; as a result, drainage from plateaus is predominately through the saturated layer between the water table and underlying frost table.

The frost table is the base of this *suprapermafrost groundwater flow zone* because it is relatively impermeable, given that the underlying peat is frozen and saturated with ice and a small amount (< 15 percent of soil volume) of unfrozen water. The combined effect of the depth dependency of hydraulic conductivity and the nearly impermeable underlying frozen peat makes the degree of thaw of the seasonally frozen (active) layer a critical factor that controls the rate of subsurface runoff. Subsurface flow rates are high early in the thaw season since the frost table at that time is close to the ground surface; therefore, the overlying saturated layer occupies peat with high hydraulic conductivity values. Later in the thaw season, when the saturated layer is deeper in the peat, subsurface runoff rates are orders of magnitude lower.

Since the lateral flux rate of groundwater through the active layer in the case of permafrost peatlands—and through the thawed peat layer in the case of seasonally frozen peatlands—is closely coupled with the degree of ground thaw, groundwater flow from peatlands is closely coupled to the conduction of energy into the ground. The majority (> 85 percent) of the heat flux conducted vertically downward from the plateau ground surface into the peat profile is used to melt the ground ice (Hayashi et al., 2007). Owing to spatial variations of near surface soil moisture (i.e., wetter areas enable greater conduction of energy to the thawing frost table depth), the frost table depth can vary widely over short distances and therefore so too can the topography of the frost table.

Topographic variations of the frost table surface, including the slope of the frost table, control the rate and direction of groundwater flow. However, unlike the topography of other impermeable surfaces (e.g., bedrock), the frost table topography evolves when air temperatures are above 0 °C due to spatial and temporal variations of ground thaw. As a result, the rates and directions of subsurface flow from plateaus and from seasonally frozen peatlands often vary with time over the thaw season.

Peat plateaus are a type of bog that arises from the upward displacement of the ground surface resulting from the formation of permafrost. At the landscape scale over a period of decades to centuries, peat plateaus form within wetlands where interannual ice bulbs aggregate into permafrost bodies whose growth vertically displaces the overlying wetland terrain. This is followed by subsidence and collapse through a type of permafrost degradation called *thermokarst erosion*, a process that transforms tree-covered permafrost terrain (peat plateaus) back into treeless, permafrost free wetlands, known as collapse scars. Such erosion involves simultaneous permafrost thaw, ground surface subsidence, and inundation by adjacent wetlands. This process is an on-going and defining feature of the landscape type shown in Figure Box 2-2.

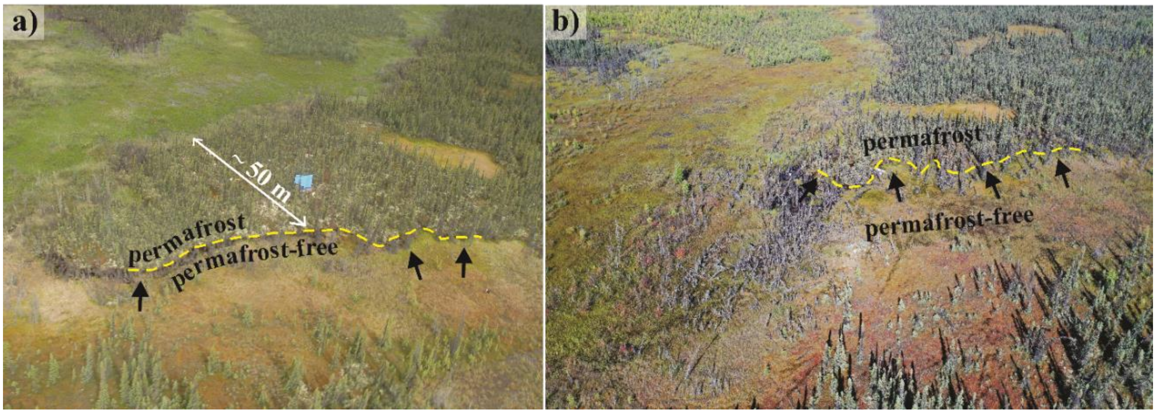

Figure Box 2-2 - Permafrost thaw-induced landcover change between a) 14 June, 2003, and b) 30 August, 2020, at Scotty Creek, Northwest Territories, Canada. The yellow, dashed lines indicate the boundary between permafrost (forested peat plateaus) and permafrost-free (treeless wetland) terrains. The black arrows indicate the direction of expansion of the permafrost-free terrain. (Photograph credits: (a) W. Quinton; (b) M. Dominico)

In a stable climate, thermokarst erosion is localized and does not result in a net loss of permafrost from a landscape since it is balanced by the re-establishment of permafrost below relatively dry areas of collapse scars, which over a period of decades to centuries can transition collapse scar wetlands back into peat plateaus. As such, the outcome of this process differs from that of climate-warming induced permafrost thaw since the latter results in a steady loss of permafrost terrain from a region.

The transformation of a forest underlain by permafrost to one that is permafrost free (Figure Box 2-3) is accomplished through a number of intermediate stages during which key processes (e.g., increased hydrological connectivity with the basin drainage network, partial wetland drainage, development of hummock topography, and tree seedling establishment) alter the ecohydrological environment and introduce new processes and feedbacks. The direction and relative importance of these and other processes and their associated feedbacks have not yet been quantified. The timescales of the transition are estimated to be less than half a century based on literature and observations of recent change captured by aerial/satellite imagery archives (Haynes et al., 2020). Ecohydrological modeling of the processes and associated feedbacks depicted in Figure Box 2-3 would greatly enhance the mechanistic understanding of this changing environment and its implication for peatland—groundwater interactions. This depiction identifies black spruce forest—either underlain by permafrost or permafrost free—as end members; that between these end-members is a wetland type defined as a *transition wetland* whose hydrological and ecological processes and properties evolve with time.

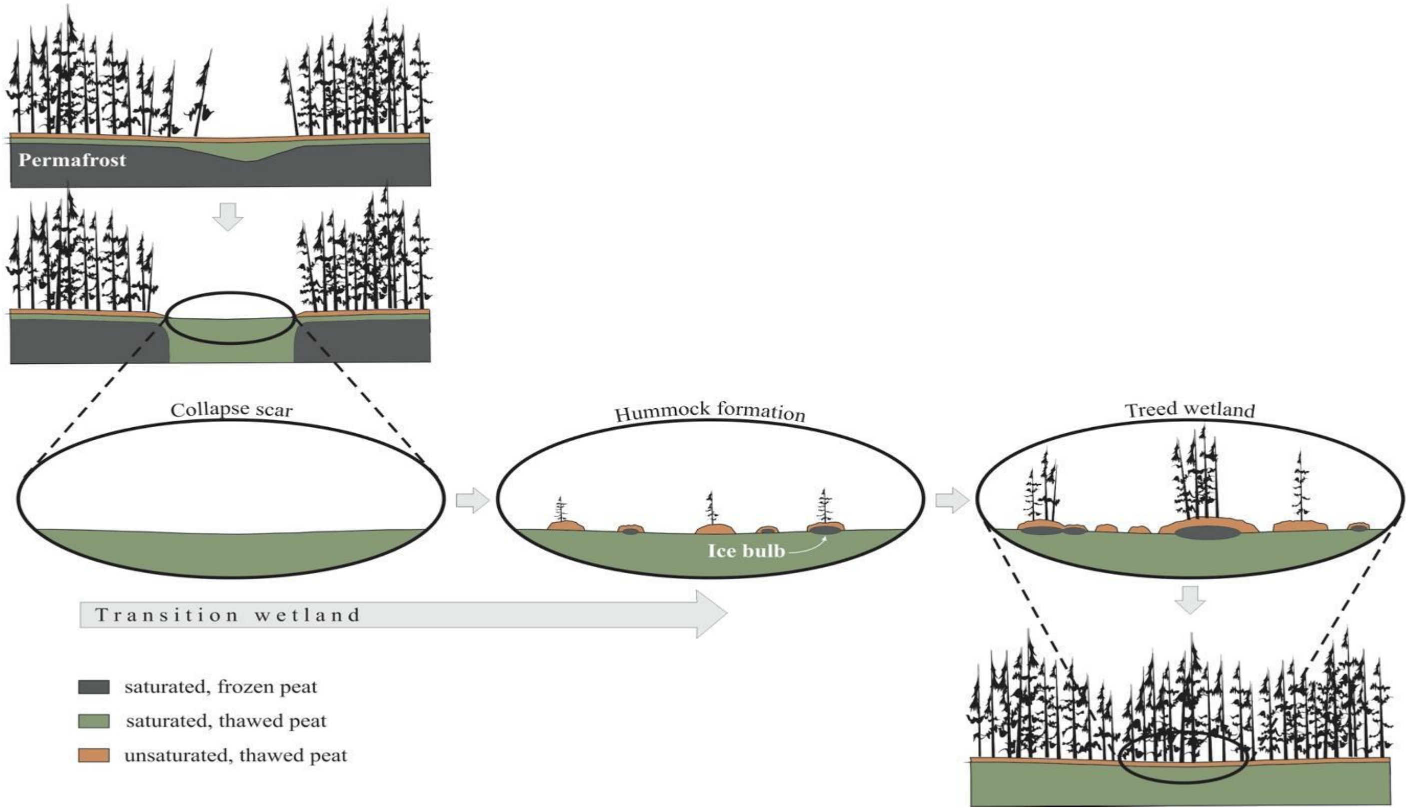

Figure Box 2-3 - Wetland transition following the permafrost thaw-induced development of a collapse scar wetland. Gray arrows indicate direct mechanisms and dashed lines represent potential feedbacks. Partial drainage of wetlands following the thaw of permafrost at their margins enables hummock development. Thermal insulation by the unsaturated surficial peat helps preserve seasonal ice bulbs that develop during winter due to greater exposure of hummocks to the atmosphere. The aerobic surfaces of hummocks promote re-establishment of black spruce, catalyzing the formation of treed wetlands, and ultimately the return of black spruce forests, although without permafrost (Modified from Haynes et al., 2020).

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Box 3 - Tropical Peatland Hydrology

Where tropical peatlands are common, rainfall can be extreme (~3800 mm/y; Wösten et al., 2007). Consequently, groundwater inflow is typically a minor contributor to the water balance, although in some peatlands it can be important geochemically (Grundling et al., 2015). Moreover, many tropical peatlands accumulate substantial layers of peat and become domed, similar to boreal bogs—thus ombrotrophic and moderately acidic (Baird et al., 2017). Water table fluctuations are primarily driven by rain events; by comparison, losses by groundwater outflow or evapotranspiration are much less variable. The water table commonly rises above the surface in wet seasons and can fall substantially in the dry season.

Where they have a domed profile, tropical swamps shed water. Groundwater outflow can be substantial, facilitated by extremely high hydraulic conductivity of the upper peat layers (470 m d^{-1}; Baird et al., 2017). These rates are similar to that in unconsolidated gravel. However, horizontal flow is tempered by the generally low hydraulic gradients typical of extensive tropical swamps. These gradients approximately follow the topographic slope created by the peatland dome. Such gradients range from 0.01 near the margins to 0.001 at the top (Page et al., 2006). Deeper peat hydraulic conductivity can be two orders of magnitude lower than that near the surface, so the transmissivity feedback mechanism described in the main text of this book also operates in tropical peatlands.

Ditching of vast areas of tropical peatland for Palm oil production results in water table lowering, commonly to ~1 m or more, compared to ~0 m at undrained areas outside the dry season. Drainage ditches increase local hydraulic gradients so that groundwater outflow, via discharge to the ditch network, can be substantial. Drainage also causes peat subsidence, in part from the loss of water pressure that buoys the peat but also from peat decomposition. Drainage of a South East Asian swamp resulted in nearly 1.5 m of subsidence in the first five years, mainly attributable to primary consolidation. Thereafter, surface subsidence of 0.05 m y^{-1} was attributed to peat decay (Hooijer et al., 2012).

Reclamation of these areas is problematic because primary consolidation is only partially reversible and subsidence caused by peat decay is non-reversible. Ditch blocking can be effective at raising water tables during the wet season, but flooding may ensue. During the dry seasons, the water table is often below the base of the ditch, so ditches have little effect (Putra et al., 2021).

Peat fires are a common outcome of drainage activities in tropical swamps and are exacerbated by seasonal dry periods and longer-term cycles such as El Niño. Page and others (2002) estimated that the 1997 El Niño peat fires in Indonesia released a mass of carbon equivalent to 13 to 40 percent of that generated by annual global fossil fuel use. Fire also lowers the peat surface, and as a consequence extensive flooding can follow in wetter periods. Burning reduces the hydraulic conductivity of near surface peat (Holden et al., 2014).

Return to where text linked to Box 3↥

Box 4 - Van Genuchten-Mualem (VGM) Variables and Parameters

In most cases, the volumetric water content at saturation, θ_s, is equivalent to the total porosity, $\phi_{mob} + \phi_{im}$, as discussed in Section 2.1. The residual water content, θ_r varies between 0.01 and 0.15 and is often estimated because measurements at extreme tensions ($h > 200$ cm, where $h = -\psi$) are less common in peat. The residual water content, θ_r, is a portion of the ϕ_{im} as measured by soil water retention curves. Both θ_s and θ_r are strongly related to bulk density and degree of decomposition, with θ_s decreasing and θ_r increasing with increase in bulk density.

Similar to both θ_s and θ_r, the VGM parameter n is related to bulk density, decreasing from ~0.2 to ~0.01 for bulk densities ranging from 0.02–0.8 g cm^3. However, this statistical relationship is weaker than those determined by porosity measures (Liu and Lennartz, 2019). The air entry pressure (characterized by $1/\alpha$) is strongly correlated with both macroporosity (discussed in Section 3.1) and K_{sat} (Liu and Lennartz, 2019). Broadly, α decreases with increasing bulk density because air entry into denser peat requires higher pressure. In sedge peat, α ranges from ~10^{-3} — ~10^{1} cm^{-1}, and in *Sphagnum* peat it ranges from 10^{-1}—10^{1} cm^{-1}. Unlike both sedge and *Sphagnum* peat that are constrained to relatively small ranges of α, woody peat spans a wide range of reported values, but the limited number of reported values may bias the true range for woody peat.

Values of K_{unsat} defined for a given tension, h—using Equation 6—employ a scaling parameter, l, related to the pore-size distribution. Typically, l is negative in peat and ranges from -6 to 1. To date, a statistical relationship between l and bulk density has not been defined; however, this may be due to the choice to fix l in some peat studies, while fitting l in other studies. Much less is known about how l varies with other physical peat properties as compared to what is known about soil water retention model parameters, due to the paucity of studies that encompass both soil water retention and unsaturated hydraulic conductivity.

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Box 5 - Mobile Water Porosity and Solute Transport

The water flux through a unit area of soil is a function of its hydraulic conductivity and the gradient across it, according to Darcy's law. The resulting water flux has the units of velocity but is more correctly called the *specific discharge, q* (cm/s).

Individual molecules of water and dissolved solutes travel faster than the rate indicated by the specific discharge because the solids of the matrix do not conduct water. The velocity of water and solute molecules is faster because the volumetric rate of discharge passes only through the cross sectional area of pores. This is called the *average linear pore-water velocity*, v_L (cm/s). Therefore, $v_L = q/\phi$, so knowing the porosity of a matrix is key to estimating the average linear velocity of a solute passing through it. It also affects the spread (dispersion) of that solute. If there is a dual-porosity matrix—as is the case for many peat soils—estimating the average linear velocity must be done using only the mobile water porosity (ϕ_{mob}).

Estimating ϕ_{mob} is not simple. One approach is to base it on the proportion of water that can be drained from a peat sample, as estimated by using water retention curves. These values vary considerably, depending on the peat's botanical origin and its state of decomposition (e.g., Figure 16). Since the rate of solute travel is sensitive to the value of ϕ_{mob}, this parameter can be estimated when fitting a model of solute breakthrough data. The dispersion parameter can also be included in the fitting process. Figure Box 5-1 shows a few attempts to simulate the measured solute breakthrough data by adjusting ϕ_{mob} values determined from a set of water retention curves.

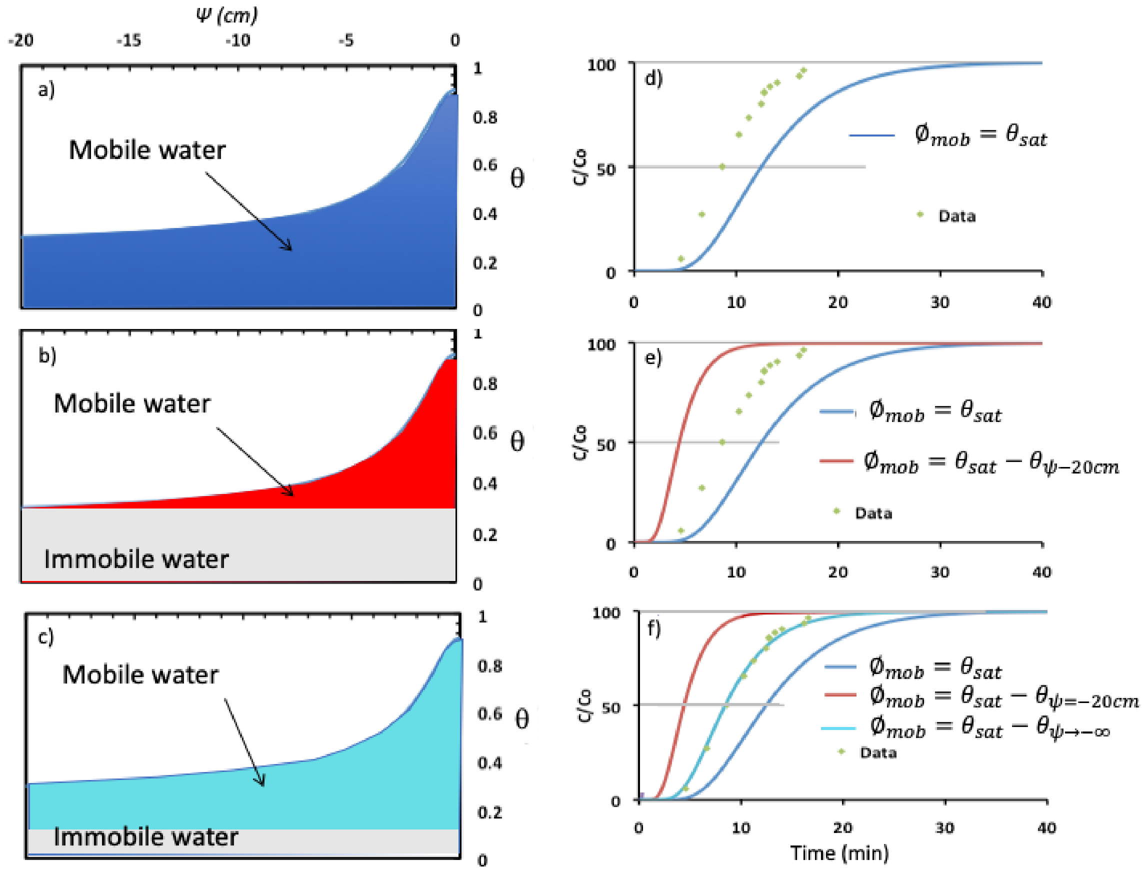


Figure Box 5-1 - A systematic illustration of the impact of using three different ϕ_{mob} estimates—a), b), c)—and associated water retention curves on the simulated solute breakthrough curves (solid lines) versus measured data (green points)—d), e), f)—in a laboratory column assuming no change in volume of the sample with changing pressure and drainage, and using a conservative solute. Breakthrough curves are a graph showing concentration of solute discharged from the column, expressed relative to the source concentration (C/C_o), as a function of time. The breakthrough curves were simulated using the Ogata-Banks solution (Ogata and Banks, 1961). In each case, the pore-water velocity was equal to $v = q/\phi_{mob}$, where q is the quotient of discharge volume and column cross-sectional area. The breakthrough curves shown in d), e), and f) correspond to ϕ_{mob} illustrated in a), b), and c). In the first case (a), all water is assumed to be mobile, thus the immobile porosity $\phi_{im} = 0$ and at full saturation (where $\psi = 0$) the mobile porosity ϕ_{mob} is 0.9. The resulting simulated breakthrough curve shown in (d) (solid blue line) is slower than the observed breakthrough (i.e., plots to the right of the measured data). In the second case (b), ϕ_{im} was determined by noting where the water content levels off with decreasing soil water pressure, suggesting that water can readily drain above this pressure, thus is mobile. This occurred at $\psi \sim$ -20 cm at which ϕ_{im} = 0.35, thus at $\psi = 0$ the mobile porosity ϕ_{mob} = 0.9-0.35 = 0.55. Using ϕ_{mob} = 0.35, the simulated breakthrough curve shown in (e), solid red line) is faster than the observed breakthrough (i.e., plots to the left of the measured data). Given that these two cases bracket the observed data, it must be that ϕ_{mob} lies between these values. For the third simulation, ϕ_{mob} was estimated by adjusting it until the simulated breakthrough curve matched the observed data as shown in (f). In this case ϕ_{mob} is the difference between θ_{sat} and a value of ϕ_{im} that may reflect the water content (0.11) that occurs as $\psi \to -\infty$ ($\phi_{mob} = \theta_{sat} - \theta_{\psi\to-\infty}$). Thus at full saturation (when $\psi = 0$) the mobile porosity ϕ_{mob} is 0.9 – 0.11 = 0.79. The resulting simulated breakthrough curve is shown in (f) by the solid turquoise line.

Return to where text linked to Box 5↑

Box 6 - Mercury in Peatlands

In many northern peatlands, mercury is considered a contaminant of concern where its bioavailability and ability to bioaccumulate in food webs is dependent on the source of groundwater and internal hydrology of a peatland (Branfireun et al., 2020). Gaseous elemental mercury has an atmospheric residence time of several months, which enables its global distribution and eventual deposition into ecosystems, often far from its original source.

The saturated, highly organic soils common in peatlands harbor a diversity of anaerobic microbes, particularly sulfate reducers and methanogens, which facilitate the transformation of deposited and/or stored inorganic mercury into methylmercury (Bishop et al., 2020). Methylmercury is a bioaccumulating neurotoxin that has been linked to adverse environmental and human health outcomes (Beckers and Rinklebe, 2017). Peatlands with weaker groundwater connectivity (i.e., bogs and poor fens) are often strong methylmercury sources in landscapes; while, more minerotrophic peatlands like rich fens or swamps can be net methylmercury sinks (Tjerngren et al., 2012). In general, the coverage of peatlands in many northern landscapes is often associated with increasing methylmercury concentrations in fish and other wildlife that local communities, such as indigenous communities in the James Bay Lowland, may rely on as a food source.

Return to where text linked to Box 6↥

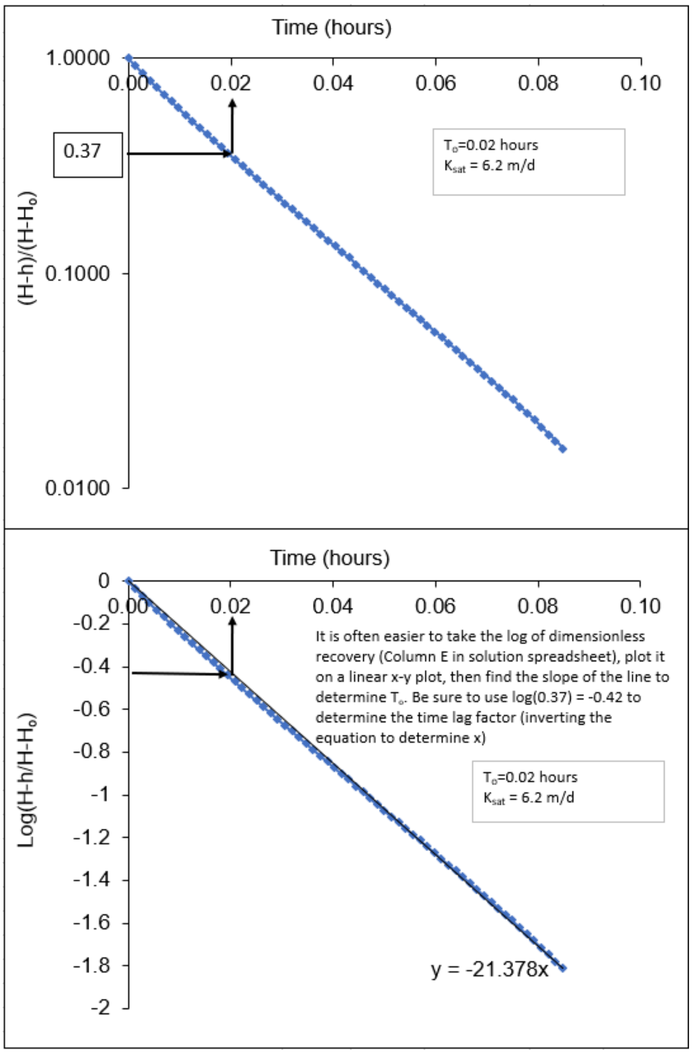
13 Exercise Solutions

Solution Exercise 1

1. An Excel spreadsheet with the solution for data set #1 is provided here (Peat&Peatlands-Exercise-1-SolutionsDataSets1&2.xlsx)↗. Time in seconds (column A) is expressed as hours (column B) and used to plot the data. The hourly rate is eventually multiplied by 24 to express the answer in m/d. The head measurements (h) are given in column C (m). Head recovery (H-h/H-H_o) is calculated in column D. Head recovery is plotted in log-linear space as shown in the top graph of the image below. Then a line is fit to the data. Although the data form a straight line in this case, the data often deviate from a straight line at later times; if the later points exhibit a distinctly smaller slope (see dataset #2), then follow the instructions for dataset #2. The time lag parameter (T_o) is interpreted from the data as the time when the line reaches (H-h/H-H_o) = 0.37. Sometimes, it is easier to work with linear-linear plots, especially when interpolating between numbers which is difficult on a log scale. The log of the head recovery values is calculated in column E. These values are plotted arithmetically as shown in the lower graph of the following image. The equation of the line can be calculated using the linear fit function of the Excel software. The value of T_o is estimated as x in the equation of the line by substituting 0.37 for y. The value of T_o is then used in the Hvorslev equation provided in exercise 1 along with the other shape parameters provided in exercise 1 to calculate $K_{sat,}$ which is ~6.2 m/d. The calculation is shown in the following equation presented in Exercise 1 and the graphs are shown in the following image.

$$K_{sat} = \frac{r^2 \ln{(L_e/R)}}{2 L_e T_o}$$

$$K_{sat} = \frac{(0.05\mathrm{m})^2 \ \ln{(0.5\mathrm{m}/0.064\mathrm{m})}}{2\ (0.5\mathrm{m}) \frac{0.02\mathrm{hr}}{24\frac{\mathrm{hr}}{\mathrm{day}}}} = 6.167\mathrm{m/d} = \sim 6.2\mathrm{m/d}$$

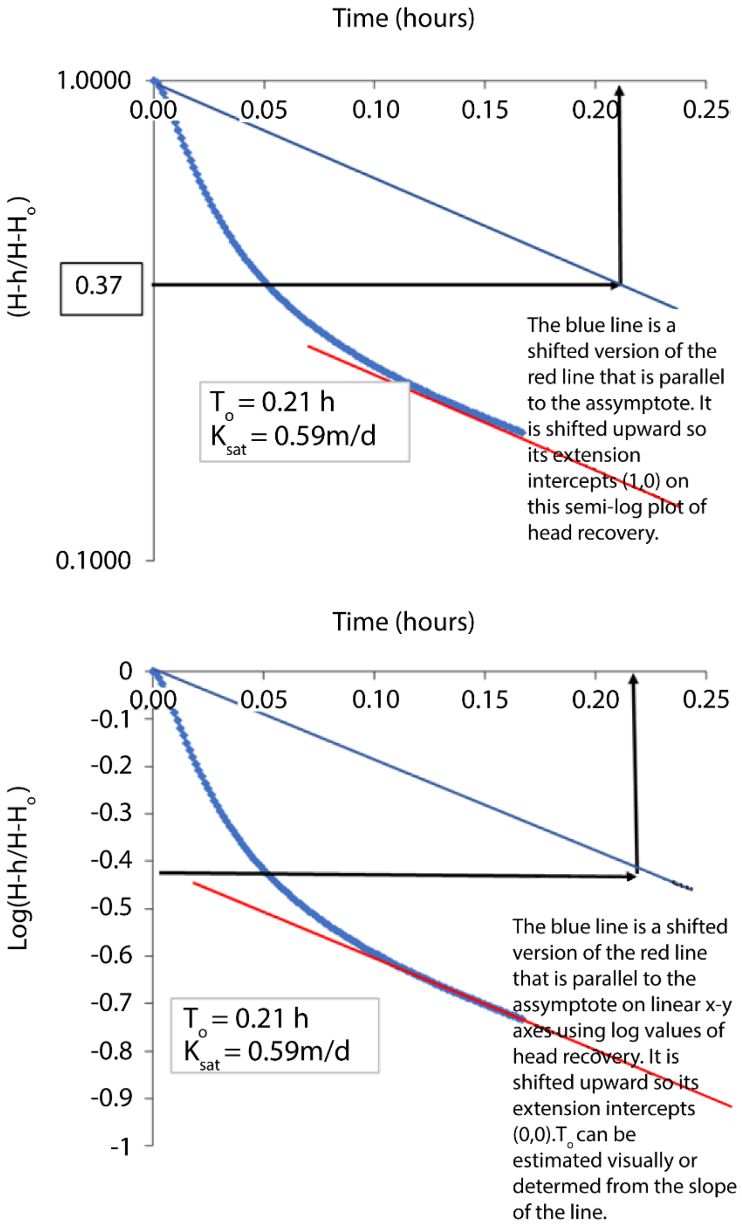


Head recovery plots for data set #1. The top graph is in log-linear space, from which T_0 is read ($T_0 = 0.02$ h). The bottom graph presents the same data plotted as log values (of head recovery) on a linear scale. T_0 can be read from the curve or, alternatively, estimated from the equation of the line provided on the graph. K_{sat} is then calculated using the Hvorslev equation.

2. An Excel spreadsheet with the solution for data set #2 is provided here (Peat&Peatlands-Exercise-1-SolutionsDataSets1&2.xlsx)↗. For dataset #2, the data do not plot as a straight line, likely as a consequence of peat compression caused by the large head-gradient imposed by pumping. If a straight line is fit to the entire data set, or to early-time data then the estimation of T_0 will result in too small of a time lag parameter, hence will overestimate K_{sat}. To obtain a more representative value, use only the late time data, thus draw a line parallel to the asymptote (i.e., late-time) head recovery curve (red line), then shift that line upwards so that it starts at the origin (1,0 in the semi-log plot) as shown in the image below (thin blue line). This rate (thin blue line) is more reflective of the actual rate of recovery after the initially rapid water inflow caused by water squeezed from collapsing peat around the intake. Read T_0 from

this line (0.21 h) and calculate hydraulic conductivity. $K_{sat} = 0.59$ m/d. T_0 could be calculated from a slope of this new line, but it would require using coordinates estimated from the line, so it is redundant and more work than simply estimating a value of T_0 from the line. The calculation is shown from the following equation.

$$K_{sat} = \frac{(0.05\text{m})^2 \ \ln(0.5\text{m}/0.064\text{m})}{2\ (0.5\text{m})\frac{0.21\text{hr}}{24\frac{\text{hr}}{\text{day}}}} = 0.59\text{m/d} \sim 0.6\text{m/d}$$



Head recovery plots for data set #2. The top graph is in log-linear space. The data do not plot as a straight line, suggesting enhanced initial recovery is not reflective of the true rate. An asymptote to the tail of the recovery curve is drawn and shifted up. The value of T_o is read from this line as $T_o = 0.21$ h. The bottom graph presents the same data plotted using log values of head recovery on a linear scale. T_o is read in the same way but using -0.43 on the y-axis which is the log of 0.37. K_{sat} is then calculated from Equation 5 as 0.6 m/d.

Return to Exercise 1↥

Solution Exercise 2

To complete these calculations, first calculate the average linear groundwater velocity (v_L) as shown in the following equation, at the stream interface for both the high and low water scenarios in each peat layer.

$$v_L = \frac{K_{sat}\ i}{\phi_{mob}}$$

where:

K_{sat}	=	saturated hydraulic conductivity for a given layer (cm d^{-1})
ϕ_{mob}	=	mobile porosity for a given layer

Next, calculate the proportion of water (p_w) flowing into the stream from each layer (i) as shown in this equation. In this case the sum of l_i is 150 cm.

$$p_{w\,i} = \frac{v_{L\,i} \frac{l_i}{\sum_1^n l_i}}{\sum_{i=1}^{i=n} v_{L\,i} \frac{l_i}{\sum_1^n l_i}}$$

Water velocity in cm per day and proportion of water by layer

		0–10 cm	10–20 cm	20–30 cm	30–50 cm	50–150 cm	Thickness weighted average velocity
Wet	v_{Li}	17.78	13.33	12.00	3.43	1.66	4.44
	p_{wi}	0.27	0.20	0.18	0.10	0.25	
Dry	v_{Li}			3.00	0.86	0.42	0.68
	p_{wi}			0.34	0.19	0.47	

Using a base concentration of 1 mg cm^{-3} for the maximum condition when $R_f = 1$, the amount of contaminant in a layer designated by i (M_{ci}) exported to the stream is the product of the volumetric flow rate of the groundwater and the contaminant concentration which is assumed to be proportional to $1/R_f$, as follows.

$$M_{ci} = v_{Li}\ l_i\ w\ C_i$$

$$M_{ci\ with\ R=1} = 17.78\ \frac{\text{cm}}{\text{d}}\ 10\ \text{cm}\ \ 2000\ \text{cm}\ \ 1\frac{\text{mg}}{\text{cm}^3}\ 1\frac{1\ \text{g}}{1000\ \text{mg}^3} = 356\ \frac{\text{g}}{\text{d}}$$

For an $R_f = 3.2$, the following expresses the mass of contaminant exported to the stream.

$$M_{ci} = v_{Li}\ l_i\ \ w\frac{C_i}{R_{fi}}$$

$$M_{c1} = 17.78\ \frac{\text{cm}}{\text{d}}\ 10\ \text{cm}\ \ 2000\ \text{cm}\ \frac{1\frac{\text{mg}}{\text{cm}^3}}{3.2}\ 1\frac{1\ \text{g}}{1000\ \text{mg}^3} = 111\ \frac{\text{g}}{\text{d}}$$

The proportion of contaminant (p_c) flowing into the stream from each layer (i) is the quotient of the layer M_c and the sum of M_c for all layers.

$$p_{C\,i} = \frac{M_{ci}}{\sum_{\mathrm{i=1}}^{\mathrm{1=n}} M_{ci}}$$

The results are shown in the table below.

Mass of contaminant in grams per day exported to stream and proportion of contaminant by layer

		0–10 cm	10–20 cm	20–30 cm	30–50 cm	50–150 cm	Total Mass (g/d)
Wet	M_{ci}	112	84	57	11	2	267
	p_c	0.42	0.32	0.21	0.04	0.01	
Dry	M_c			14	3	1	18
	p_c			0.81	0.15	0.03	

Under wet conditions, the decline in hydraulic conductivity and other hydraulic properties of peat within the upper 20 cm of the profile results in a smaller difference in proportion of water flux to the stream from the upper layers 1 and 2 (0.07) than contaminant flux (0.09). Even with a further decline in hydraulic conductivity and an increase in partitioning coefficient in the next peat layer (20 to 30 cm), the proportion of contaminant exiting to the stream from this third layer (0.19) is greater than the proportion of water flux from the layer (0.18). This suggests that hydraulic conductivity is a strong control on the export of this weakly adsorbing contaminant. If the partitioning coefficient was far larger, such as that of inorganic mercury (~$10^{3\text{-}5}$ mL g^{-1}), the relatively small changes in hydraulic conductivity would not impact the total mass of contaminant exported to the stream to such a degree.

The lower saturated thickness under dry conditions (low water table) resulted in a decrease in contaminant export by more than an order of magnitude (267 g/d versus 18 g/d).

Return to Exercise 2↥

Solution Exercise 3

Using the same procedure as in the solution for Exercise 2, the calculated water and contaminant velocities and proportions are listed in the following tables.

Water velocity and proportion

		0–10 cm	10–20 cm	20–30 cm	30–50 cm	50–150 cm
Wet	v_{Li}	50.00	11.43	1.60	0.20	0.04
	p_{wi}	0.78	0.18	0.03	0.01	0.01
Dry	v_{Li}			0.40	0.05	0.01
	p_{wi}			0.67	0.17	0.17

Mass of contaminant in grams per day exported to stream and proportion of contaminant by layer

		0–10 cm	10–20 cm	20–30 cm	30–50 cm	50–150 cm	Total Mass (g/d)
Wet	M_{ci}	907	143	15	1	0	1066
	p_c	0.85	0.13	0.01	0.00	0.00	
Dry	M_{ci}			10	1	0	11
	p_c			0.95	0.05	0.00	

The exponential decline in hydraulic conductivity with depth creates a preferential flux of water (0.78) and contaminant (0.85) through the upper 10 cm of the peatland under the wet condition. Even under dry conditions, it is the upper 10 cm of the saturated peat that transports the majority of both water (0.67) and contaminant (0.95). In the uppermost saturated peat layer, the proportion of contaminant flux is higher than the proportion of water flow; below this upper layer, the proportion of water flow is higher than the proportion of contaminant flux. This is driven by the increase in retardation coefficient with depth that is common in peatlands, highlighting the complex interaction between solute transport with the hydrophysical and geochemical properties of peatlands. In short, the exponential decline in hydraulic conductivity has a very strong control on the distribution of contaminants within a peat profile.

Relative to the peatland in Exercise 2 that has a more monotonic decline in hydraulic conductivity, there is more mass of contaminant exported to the in the stream (1066 g d^{-1} versus 267 g d^{-1}) in the exponential decline peatland under the wet conditions because of the substantially higher K_{sat} in the top layer and smaller R_f in the top two layers. Assuming the same Rf profile, the difference remains large between the exponential and monotonic K_{sat} profiles (1066 g d^{-1} versus 396 g d^{-1}). However, under dry conditions, the contaminant mass exported to the stream is reduced by ~40% between the exponential (11 g d^{-1}) and monotonic (18 g d^{-1}) peatlands due to the 2 times larger sorption in the monotonic case. These two examples and their comparison highlight the necessity of understanding both the hydrological and geochemical properties of peatlands to properly predict, remediate, and/or mitigate potential impacts from disturbances.

The mass of contaminant in grams per day exported to stream and proportion of contaminant by layer when R_f is the same as Exercise 4 but with the exponential hydraulic conductivity profile

		0–10 cm	**10–20 cm**	**20–30 cm**	**30–50 cm**	**50–150 cm**	**Total Mass (g/d)**
Wet	M_{ci}	316	72	8	1	0	396
	p_c	0.80	0.18	0.02	0.00	0.00	
Dry	M_{ci}			5.3	0.3	0.0	5.6
	p_c			0.95	0.05	0.00	

Return to Exercise 3↥

Solution Exercise 4

Part 1: The volumetric heat capacity for each constituent is the product of specific heat and mass density, calculated using the following equation.

$$Cv = cp\ \rho$$

The first phase in the thawing process is a saturated and frozen soil column. The volumetric heat capacity for this phase is derived from the following equation using a theta of zero because the total porosity is ice and the liquid water content is zero.

$$Cv_{IWS} = Cv_I(\phi_t - \theta) + Cv_W(\theta) + Cv_S(1 - \phi)$$

where:

Cv_{IWS} = volumetric heat capacity for a mixture of ice, water and soil [J m^{-3} C^{-1}]

Cv_I = volumetric heat capacity for ice = (2120)(920)=1,950,400 J m^{-3} C^{-1}

Cv_W = volumetric heat capacity for water = (4185)(1000)=4,185,000 J m^{-3} C^{-1}

Cv_S = volumetric heat capacity for soil = (1920)(40)=76,800 J m^{-3} C^{-1}

θ = volumetric soil moisture [-]

ϕ = porosity [-]

$$\boldsymbol{Cv_{IWS}} = 1{,}950{,}400(0.9 - 0.2) + 4{,}185{,}000(0.2) + 76{,}800(1 - 0.9) = \mathbf{2,209,960\ J\ m^{-3}\ C^{-1}}$$

The second phase is characterized by saturation of the soil and in the absence of ice and is estimated from the following equation.

$$Cv_{WS} = Cv_W(\phi_t) + Cv_S(1 - \phi_t)$$

$$\boldsymbol{Cv_{WS}} = 4{,}185{,}000(0.9) + 76{,}800(1 - 0.9) = \mathbf{3,7774,180\ J\ m^{-3}\ C^{-1}}$$

The final phase is characterized unfrozen, unsaturated soil with a drained water content θ_d of 0.5, thus volumetric heat capacity is estimated from the following equation.

$$Cv_A = (1010)(1.2) = 1{,}212\quad \mathrm{J\ m^{-3}\ C^{-1}}$$

$$Cv_{WSA} = Cv_A(\phi_t - \theta_d) + Cv_W(\theta_d) + Cv_S(1 - \phi_t)$$

$$\boldsymbol{Cv_{WSA}} = 1{,}212(0.9 - 0.5) + 4{,}185{,}000(0.5) + 76{,}800\ (1 - 0.9) = \mathbf{2,100,644\ J\ m^{-3}\ C^{-1}}$$

The following table presents the computed values of volumetric heat capacity for the three soil conditions using **kilojoules**.

Volumetric heat capacity for each phase in soil thaw.

Phase		kJ m^{-3} C^{-1}
(1) Saturated and Frozen	Cv_{IWS}	2210
(2) Saturated and Unfrozen	Cv_{WS}	3774
(3) Unsaturated and Unfrozen	Cv_{WSA}	2100

Part 2: First divide the peat profile into a number of computational layers, and conclude that the uppermost layer will be the layer that thaws first, thus would be the first to transition from an ice-water-soil mixture to one of water and soil only, and finally to an unsaturated mixture of water-soil-air. Assuming that each layer below also thaws and drains, they too would follow this same transition, although the initiation of the transition would be delayed with increasing depth below the ground surface. Peat layers composed of an ice-water-soil mixture warm relatively quickly. For layers near the ground surface, this condition is relatively short-lived since such layers are the first to thaw. Deeper layers can remain in this condition for extended periods. With the transition to the next composition stage of water and soil only, the rate of warming reduces considerably due to the increase in specific heat that accompanies the phase change of ice to water. Layers that transition to the final compositional stage of water-soil-air, can warm readily given the much lower specific heat of air than water. The rate and pattern of warming of the peat profile therefore depends on the rate that its individual layers thaw and drain. Precipitation inputs including the amount of snow water equivalent present at the end of winter, the slope of the ground surface, and the hydraulic properties of the peat in each computational layer each effect the rate of thaw and drainage and therefore indirectly affect the rate and pattern of peat profile warming.

Part 3: Given the high porosity of peat, this soil type is susceptible to large variations in thermal conductivity with variations in soil moisture content and phase changes. When frozen and saturated, peat is a highly effective thermal transmitter. During winter, this enables the peat profile to conduct energy toward the atmosphere at high rates in response to the upward directed thermal gradient. As the soil thaws, it becomes saturated with liquid water. In this condition it is still a highly effective thermal conductor and conducts energy from the ground surface into the peat profile at high rates in response to the downward directed thermal gradient. For this condition, ground thaw proceeds at a high rate. With continued thaw and drainage, layers in the upper part of the peat profile become unsaturated and as a result, their thermal conductivity decreases. For this condition, the near-surface layers become effective thermal insulators, and the rate of ground thaw therefore decreases.

Return to Exercise 4↑

Solution Exercise 5

Exercise 5 does not have a numerical solution, so this is an answer more than a solution. In response to this question, it should be recognized that the permafrost below Boreal and Taiga peatlands is typically composed of ice-saturated peat. Given the very high porosity of peat (as shown in the table presented with Exercise 4) and that water occupies a smaller volume than an equivalent mass of ice, the melt of ice occupying the pores of such permafrost results in a volume reduction of the layer of permafrost that thawed. Permafrost thaw therefore results in the subsidence of the overlying ground surface. This repositions the ground surface so that it is closer to the underlying water table which increases the moisture content and therefore the thermal conductivity of the near surface layers. This process accelerates the rate of permafrost thaw. The rate of ground thaw, whether it involves permafrost thaw or not, occurs preferentially due to local disturbances of the ground surface. Preferential thaw imposes local hydraulic gradients directing drainage toward the thaw depressions. This process also increases the local soil moisture content, and therefore the ability of the peat profile in areas of preferential thaw to transmit energy downward to the thawing front, increasing the depth of thaw and drawing horizontal drainage from greater distances. This sequence of positive feedbacks produces a 'talik' (i.e., unfrozen layer) between the bottom of the seasonally refrozen ground and the underlying permafrost table. Prior to talik formation, permafrost loses energy to the atmosphere during winter in response to the upward-directed thermal gradient (as indicated by the solution to Exercise 4). However, once a talik forms, the thermal gradient is directed toward the permafrost throughout the year, thus talik formation accelerates permafrost thaw. More information on this topic is included in Connon and others (2018) and Kurylyk and Hayashi (2015).

Return to Exercise 5↥

14 Notations

A	subscript referring to air
α	scaling parameter that is the reciprocal of the air entry pressure (L^{-1})
b	saturated thickness of the peat deposit (L)
C_v	volumetric heat capacity ($M^{-2}L^{2}T^{-2}K^{-1}$)
C_{v_i}	volumetric heat capacity for ice ($M^{-2}L^{2}T^{-2}K^{-1}$)
$C_{v_{IWS}}$	volumetric heat capacity for a mixture of ice, water, and soil ($M^{-2}L^{2}T^{-2}K^{-11}$)
C_{v_s}	volumetric heat capacity for soil ($M^{-2}L^{2}T^{-2}K^{-1}$)
C_{v_w}	volumetric heat capacity for water ($M^{-2}L^{2}T^{-2}K^{-1}$)
ΔS	storage change (L)
ΔWT	change in water table elevation (L)
h	soil water tension expressed as a positive value, equivalent to $-\psi$ (L)
I	subscript referring to ice
K	hydraulic conductivity (LT^{-1})
K_{sat}	saturated hydraulic conductivity (LT^{-1})
K_u	thermal conductivity of the unfrozen soil ($MLT^{-3}K^{-1}$)
L_e	effective length of bore intake (L)
L	mass-based latent heat of water (L^2T^{-2})
l	scaling parameter related to the pore-size distribution (-)
m	1-(1/n) (-)
n	dimensionless shape parameter inversely related to pore-size distribution (-)
p_c	proportion of contaminant (-)

ϕ porosity (-)

ϕ_{im} immobile porosity (-)

ϕ_{mob} mobile porosity (-)

ϕ_t total porosity (-)

ψ water pressure expressed as height of a column of water (L)

p_w proportion of water porosity (-)

R external radius of the bore intake (L)

R_f relative velocity of the solute with respect to the solvent (-)

ρ_b bulk density (ML^{-3})

ρ_T average density of the overlying peat and water

ρ_w density of water (ML^{-3})

S subscript referring to soil

S_e effective saturation (-)

S_s specific storage (L^{-1})

S_y specific yield (-)

σ_ε effective stress ($ML^{-1}T^{-2}$)

σ_T total stress ($ML^{-1}T^{-2}$)

θ volumetric soil water content (-)

θ_r residual water content (-)

θ_s volumetric water content at saturation (-)

T_s surface temperature of the soil (K)

ν_a apparent contaminant velocity (LT^{-1})

ν_L average linear groundwater velocity (LT^{-1})

W	subscript referring to water
$X(t)$	distance between the ground surface and the bottom of the thawed layer at a given time (L)

15 About the Authors

Dr. Jonathan Price is Professor Emeritus at the Department of Geography and Environmental Management at the University of Waterloo, examining the hydrology of peat dominated wetlands. His research focus is on restoration of peatland used for peat extraction and peatland creation in post-mined landscapes. He led the conceptualization, design, and evaluation of a groundwater-fed peatland and its watershed in the Athabasca oil sands area of Alberta, Canada, for the reclamation of a mined landscape. Dr. Price has also done pioneering work on contaminant transport in peatlands, including solutes and hydrocarbons. He has supervised over 60 graduate students and post-doctoral fellows and authored and coauthored over 200 peer-reviewed journal articles on topics including soil-water physics, micrometeorology, water quality, contaminant transport, ecology, and soil development as well as basin scale hydrology of wetlands.

Dr. Colin McCarter is an assistant professor and Canada Research Chair in Climate and Environmental Change at Nipissing University. He completed his PhD in 2016 from the University of Waterloo and held several post-doctoral fellowships, including a NSERC Post-Doctoral Fellowship. Dr. McCarter's research focuses on how the interactions and feedbacks between ecohydrological, biogeochemical, and soil physical processes control the transport of nutrients, carbon, and contaminants in northern landscapes under a changing climate. Dr. McCarter is particularly interested in how the structure of organic wetland soils, like peat, govern water and contaminant flow and the implications for restoring industrially contaminated peatlands.

Dr. William Quinton is a Professor in the Department of Geography and Environmental Studies at Wilfrid Laurier University. He has studied in the Canadian Arctic since 1987 and in the Mackenzie River valley region since 1991. In 1999, he established the Scotty Creek Research Station and since then has led several major research studies in the southern Northwest Territories that focus on the impacts of permafrost thaw on hydrological processes. Dr. Quinton was a Canada Research Chair in Cold Regions Hydrology (2005–2015) and played a leading role in developing the Laurier GNWT Partnership Agreement, the Laurier Institute for Water Science, and related initiatives. He also served as the Director of the Cold Regions Research Centre, President of the Canadian Geophysical Union Hydrology Section, National Representative to the International Association of Hydrological Sciences, and Chief Delegate to the Northern Research Basins Working Group. He has published over 100 journal articles and supervised over 50 graduate students. Close collaborations with Indigenous communities are a central theme of Dr. Quinton's work.

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Modifications to Original Release

Changes from the Original Version to Version 2

Original Version: March 24, 2023, Version 2: November 13, 2023

Page numbers refer to the original PDF.

page ii, added indication that this is version 2 of the book

page 65, elaborated on the names of the Excel files for Exercise 1, Data Sets #1 and #2 and corrected the links to the download page for the files.

pages 93 and 94, elaborated on the names of the Excel files for the solution to Exercise 1, Data Sets #1 and #2 and corrected the links to the download page for the files