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Chapter

Soil Genesis of Histosols and Gelisols with a Emphasis on Soil Processes Supporting Carbon Sequestration

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Abstract

Based on the U.S. Soil Taxonomy Histosols are soils that have a histic epipedon, which is a surface horizon that exhibits a sufficient abundance of soil organic matter to be distinctively different than other soil orders predominantly composed of clastic materials. Gelisols are soils that have permafrost, with histels being a suborder that is dominated by organic materials. Collectively, these soil orders are abundant in peatland ecosystems. The abundance of soil organic material is primarily a consequence of climate, topography, hydrology, vegetation. Peatland ecosystems have been a major research arena; however, added research attention is being directed to the potential release of carbon because of accelerated climate change. This review focuses of the structure and dynamics of organic soils and an understanding of their creation, evolution and ultimate fate. Attention is focused on degraded peatland net primary productivity because of potential forthcoming differences attributed to rainfall, temperature, vegetation, hydrology and permafrost disappearance.

Keywords: peatlands, net primary productivity, climate change, soil organic matter, soil respiration

1. Histosols

Organic soils are soils which have diagnostic horizons with more than 20% organic matter and essentially reside in marshes, bogs, and swamps where anaerobic soil conditions support a low rate of organic matter decomposition relative to the rate of organic matter production. Thus, organic soils are observed to have a carbon input rate that is initially greater than the carbon loss rate resulting in an annual carbon accumulation, then with continued soil genesis the rate of carbon input approximately equals the carbon loss rate and a carbon quasi-equilibrium is attained.

These organic soils are frequently associated with extremely wet landscapes, or extremely acidic soils, or soils lacking available nutrients or some combination of these influences. Organic soils (Histosols) as defined in the United States [1] are soils that have an abundance of organic soil materials with additional criteria specifying that they lack sufficient andic properties and lack permafrost plus these soils possess certain thickness, water saturation duration, and decomposition status associated with their fresh and rubbed fiber contents. According to the United States Keys of Soil Taxonomy [1], organic soil horizons have (i) 12% organic carbon (approximately 21% soil organic matter) if the clay content is 0% and (ii) 18% organic carbon if the clay content is 60% or greater. For horizons that have clay contents between 0 and 60% the organic carbon content is a linear relationship to clay content involving the 12% organic carbon if the clay content is 0% and 18% organic carbon if the clay content is 60%.

Histic epipedons are surface organic horizons that are water saturated for at least 30 days in most years (typically an aquic soil moisture regime) are generally 0.2 to 0.4 m thick and have sufficient organic carbon as a function of clay content. Folistic epipedons are surface horizons that are not water saturated for at least 30 days in most years (not artificially drained), typically are more than 0.20 m thick, and are largely composed of 75% or more sphagnum fibers or have a bulk density of less than 0.1 g cm⁻³. The Keys of Soil Taxonomy [1] partition histic epipedons into fibric, hemic and sapric materials. Fibric materials (Of) are minimally decomposed where three quarters or more of its volume is made up of fibers after rubbing the sample. Sapric materials (Oa) are highly decomposed; less than one-sixth of the volume of sapric material contains fibers after a sample is rubbed. Hemic materials (Oe) are intermediate with respect to decomposition. In general, fibric materials possess a very low bulk density (0.05 to 0.15 Mg m⁻³), a large total pore space (85%) with a high distribution of large pores spaces, a low bearing capacity, and a hydraulic conductivity ranging from 1.6 to 30 m day⁻¹.

Generally, the Histosol soil order is recognized if more than half of the upper 0.8 m of the soil profile is organic or if organic soil material rests on rock or fragmental material showing interstices filled with organic material. In colloquial terms the Histosol order contains soils formally described as bogs, moors, peatlands, muskegs, fens or are composed of peats and mucks. Histosols make up about 1% of the world's glacier-free land surface (325 to 375 million ha). Suborders of Histosol order are based on the degree of organic material decomposition and the length of water saturation. The Histosol suborders are: Fibrists, Hemists, Saprists and Folists. The World Reference Base for soil resources [2] states that Histosols are soils having a histic or folic horizon either 0.1 m or more thick from the soil surface to a lithic or paralithic contact or 0.4 m or more thick and starting within 0.3 m from the soil surface, and having no andic or vitric horizon starting within 0.3 m of the soil surface.

2. Histosol soil forming processes

Histosols occur in all latitudes; however, Histosols are particularly common in the boreal zone, a feature Histosols share with Spodisols. The dominant feature of Histosols is the accumulation of organic materials, which may be characterized as:

Organic material content=organic matter input – organic matter loss (1)

The rate of organic matter decomposition in Histosols is usually very slow, a feature attributed to specific conditions of climate, topography and hydrology. In boreal biomes, cool summer temperatures restrict microbial activity, with biologic zero being approximately 4 to 5°C. Low soil temperatures must be further associated with anoxic soil conditions to support Histosol genesis. In tropical climates, warmer temperatures support greater ecosystem productivities; however, the combined effects of precipitation, topography and hydrology may create anoxic

soil conditions for a sustained time interval to restrict soil organic matter decomposition. Topography influences Histosol formation by directing water flux within the landscape position. Lateral groundwater may create seepage on sideslopes, whereas peatlands may form in poorly-drained basins. Fens occur where surface water inflow or groundwater discharge concentrates nutrient rich water. Pocosins or bogs on coastal plains or interior flatlands are frequently located on slightly raised interfluvial positions.

The degree of soil organic matter decomposition has a significant influence on soil properties. Buol et al. [3] reviewed literature to describe the soil genesis and classification of Histosols. Key soil properties that are influenced based on the degree of soil organic matter decomposition include: organic carbon, total nitrogen, carbon to nitrogen ratio, cellulose content, pH, cation exchange capacity, bulk density, water contents at field capacity and permanent wilting point, hydraulic conductivity. Upon transition from fibric to sapric soil conditions the following properties typically increase in magnitude: total nitrogen, pH, cation exchange capacity, bulk density, and the water contents at field capacity and permanent wilting point. Most notably the vertical and horizontal hydraulic conductivities decrease on transition from fibric to sapric soil conditions. However, many Histosols exhibit greater soil organic matter decomposition with increasing soil profile depth, thus the corresponding reduced hydraulic conductivity and increased water content at greater soil profile depth support continuance of the sapric condition.

Buol et al. [3] alluded to two adjacent Histosols in Michigan that differ in nutrient sources. The Napoleon soil series (dysic, mesic Typic Haplohemists) receives nutrients only from precipitation and dry deposition, whereas the Houghton (euic, mesic Typic Haplosaprists) primarily receives nutrients from seepage water that transverses calcareous sandy glacial till. The Napoleon mucky peat has an Oa1-Oa2-Oe1-Oe2 horizon sequence, with all horizons having a pH near 4, whereas the Houghton muck has an Oa1-Oa2-Oa3-Oa4-Oa5-Oa6 horizon sequence with all horizons having a pH near 7. Vegetation associated with the Napoleon mucky peat comprised various maples, swamp white oak, and dogwood, whereas the Houghton muck is vegetated with marshy grasses. Thus, water chemistry dramatically influences the soil's pH and exchangeable cation expression and coupled with hydrology influences vegetation establishment.

Aide and Aide (two authors of this manuscript) have unpublished field data of several soil series in northeastern Wisconsin. The Lupton series (Euic, frigid Typic Haplosaprists) are very deep, very poorly-drained organic soils formed in depressions on lake and outwash plains. The horizon sequence is Oa1-Oa2-Oa3-Oa4-Oa5 and has little inorganic material, a very low bulk density, a pH in 0.01 M CaCl₂ of 5.7 to 6.0 and a cation exchange capacity ranging from 107 to 199 cmol kg⁻¹ across multiple pedons. The dominant surrounding soil consists of pedons of the Padus series (coarse-loamy, mixed, superactive, frigid Alfic Haplorthods). The tupical Padus horizon sequence is A-E-Bs1-Bs2-E/B-B/E-2C. The texture is sandy loam above the lithologic discontinuity and sandy textured at greater depths (2C). These very deep, well-drained and very strongly acidic pedons are moderately deep to stratified sandy outwash with an abundance of clay films in the B material of the E/B and B/E horizons. The organic carbon content of the A horizon is less than 2% and the cation exchange capacity is very low, reflecting the sandy loam texture and diminished quantity of soil organic matter. Water extracts from both soils show an abundance of calcium, reflecting that calcium is the dominant exchange cation. These two soils have very distinctive profiles, whose properties are directly related to the contrasting oxidation-reduction environments imposed by the local hydrology.

Parent materials for Histosols are mostly hydrophytic plants [1]. Sphagnum consists of both living and dead tissue from the genus Sphagnum, with approximately 380 species. Sphagnum leaf tissue consists of chlorophyllose and hyaline cells, with the former having photosynthetic activity and the latter consisting of larger, clear and non-living cells with a large capacity to hold and store water. The cell walls contain an abundance of phenolic compounds that are resistant to decomposition. Sphagnum also has a substantial uptake capacity for calcium, magnesium and other nutrients, predisposing the underlying mineral soil to an acidic reaction. Typically, Sphagnum is the dominant plant genus in mires, raised bogs and blanket bogs. Other plant species commonly associated with Sphagnum include sedges, various dwarf shrubs, *Betula nama* (Dwarf birch) and *Salix* spp. (Willows).

Paludification or the geologic accumulation of organic materials across a landscape is influenced by soil pH, soil temperature, microbial activity, nutrient availability, oxidation-reduction and vertebrates (example: beavers or *Castor canadensis*). One criterion for paludization is the maintenance of anaerobic soil conditions sufficient to inhibit plant material decomposition. In glacial lake settings or ox-bows in fluvial systems, sediment infusion may occur resulting in lacustrine sediment accumulation. When sediment accumulation is sufficient to permit acceptable light levels to penetrate to the submerged sediment surface and if the water oxygen levels are appropriately anaerobic then plant material preservation prevails. When Histosols evolve because of sediment deposition with subsequent soil organic matter accumulation then this process is termed terrestrialization.

3. Gelisols

In the United States the Keys of Soil Taxonomy support 12 soil orders at the highest level of soil taxonomy [1]. Gelisols (Cryosols in the World Reference Base of Soil Resources [2]) are soils that have permafrost within two meters from the soil surface. Permafrost is a soil climatic condition where soil material has continuous temperatures at or below 0°C. Because of the permafrost requirement, Gelisols occur extensively in boreal, subarctic and arctic environments and comprise approximately 18 km² (13.4%) of the ice-free land area [1]. Gelisols having a short period of seasonal thawing have an upper zone that thaws, creating an "active layer" approximately a few cm to 1.5 m thick. This active layer may experience soil forming processes, including sufficient biotic activity to form histic epipedons (suborder histels) [3].

The boundary between the active layer and permafrost is termed the "permafrost table". In moist soil and with the return of winter conditions, soil freezing begins at the permafrost table and also at the soil surface, which subsequently finalizes in the active layer. Thus, the active layer experiences freezing fronts from both the soil surface and from the permafrost table, giving rise to compaction and a loss of any soil structure. In the active layer of many Gelisols, dark streaks of organic matter that are distinguished from the soil matrix colors, suggesting soil material redistribution because of cryoturbation. The permafrost table is frequently impermeable to percolating water and therefore develops an accumulation of soil organic matter.

In very cold and low precipitation areas Gelisols are mostly shallow and relatively featureless soils; however, where temperatures are relatively mild and precipitation is more extensive, Gelisols are deeper and likely have an active layer that exhibits accumulation of soil organic matter. Gelisol vegetation includes lichens, moss, liverwort, sedge, grass and boreal forest species. Soil inhabiting organisms include prokaryotes (most notably N-fixing Azotobacter), fungi, actinomycetes, anthropoids, nematodes, protozoa and algae [1, 3].

Solifluction may occur on sloping landscapes. Cryopedogenic processes include cryoturbation causing a reduction in soil profile horizonation (Haploidization), soil structure formation, seasonal ice lens formation above the permafrost table, landscape collapse (thermokarst), and the formation of redoximorphic features. Additionally, soil carbon pool sizes, redistribution within the soil profile, and bioavailability are strongly affected by (1) cryoturbation, which is the soil-mixing action of freeze/thaw processes, and (2) by the presence of permafrost itself, which has strong controls over soil temperature and moisture and runoff. Overall, permafrost affected soils represent 16% of all soils on the globe, and contain up to 50% of the global belowground soil carbon pool [4]. Histels are Gelisols consisting of organic materials, with suborder groups listed as: (i) Folistels, (ii) Glacistels [have the upper boundary of a glacic layer (75% or more visible ice)], (iii) Fibristels, (iv) Hemistels, and (v) Sapristels.

Tarnocai et al. [4] performed an extensive review of carbon pools in the northern permafrost region, noting that approximately 3.56 x 10⁶ km² in this region at peatlands. These authors provided data illustrating that Histels (66.6 kg m⁻²) and Histosols (69.6 kg m⁻²) have the highest soil organic carbon contents. Histels alone are estimated to contain 184 Pg C, whereas histosols contribute 94.3 Pg C. Turbels show extensive soil organic carbon incorporation to deeper soil depths because of cryoturbation.

4. Organic carbon and peatlands

Peatland ecosystems are well represented in the majority of the world's biomes. In this manuscript we define a biome as a community of associated ecosystems characterized by their prevailing vegetation and by organism adaptation to that particular environment. Different sources define the types and number of biomes differently; herein, we specify six biomes: (i) tundra, (ii) taiga, (iii) grassland, (iv) deciduous forest, (v) desert, and (vi) tropical rainforest. Tundra, taiga and tropical rainforests are commonly accepted biomes having considerable expanses of peatlands; however, examples do exist in grassland and deciduous forest biomes.

Peatlands, as defined by the National Working Group (Canada), are wetlands containing more than 0.4 m thickness of peat [5]. Ombrotrophic peatlands or oligotrophic peatlands include soil and vegetation which receive water and nutrients primarily from precipitation, thus they are environments isolated hydrologically from the surrounding landscape. Given that rainfall is acidic because of equilibrium with the partial pressure of CO₂ and the rainfall nutrient composition is relatively low, ombrotrophic peatlands are typically considered nutrient deficient and exhibit reduced microbial activity. Frequently the vegetation is dominated by Sphagnum mosses. Minerotrophic peatlands are wetlands whose water availability comes mainly from nutrient-enriched surface waters that have neutral to alkaline pH reactions. Typically, minerotrophic wetlands have a high-water table, low internal drainage and exhibit moderately-well to well-decomposed sedges, brown mosses and related vegetation.

Carbon content is variably defined to represent the carbon concentrations on a surface area basis or a soil volume basis. Typically, carbon content defined as the mass of carbon per unit land area (kg carbon m⁻²) is presented to indicate landscape variability, whereas carbon content on a volume basis (kg carbon m⁻³) is presented to indicate intra-pedon or inter-pedon differences. Carbon content as expressed as the carbon concentration per volume is a soil or landscape property influenced by bulk density and horizon depth. Carbon accumulation is the net gain or loss of carbon content, typically at century or millennial scales. Peatlands reside on nearly

2.7% of the global land surface, yet peatlands possess a significant portion of the terrestrial soil carbon pool with deep soil organic matter accumulations created over millennia. Estimates suggest that boreal and subarctic peatlands contain 455 Pg C [6] and 462 Pg [7], repectively. Boreal peat deposits tend to be deeper than subarctic peatlands, a feature attributed to long carbon accumulation intervals [8].

Peat-forming systems have been partitioned into acrotelm and catotelm zones [9]. The acrotelm portion of a peat-forming soil system is defined as the relatively more oxygenated (oxic) upper portion of the peat forming soil system, where aerobic decomposition is comparatively greater, the hydraulic conductivity is more rapid and the bulk density typically ranges from 0.1 to 0.4 g cm⁻³. Conversely the catotelm is the suboxic to anoxic lower portion of the peat-forming soil system that is characterized by a comparatively slower hydraulic conductivity and a bulk density typically ranging from 0.8 to 1.2 g cm⁻³.

Soils being open thermodynamic systems receive water and particulate soil organic matter and energy at their boundaries, most notably at the soil-atmosphere interface. Matter and energy may also be transferred by lateral flow at the pedonpedon interface or vertical flow at the soil-sediment interface. Water infiltration and percolation within the acrotelm is rapid; however, percolation slows substantially in the catotelm, creating the upper oxic and deeper anoxic oxidation–reduction regimes within the soil profile. As soil organic matter decomposition progresses at the base of the acrotelm, the resulting loss of pore space, attributed to an increase in the bulk density, supports water retention and conversion of the lowermost portion of the acrotelm into that of the catotelm, thus elevating the acrotelm-catotelm boundary with progressive soil development.

The primary vegetation productivity (P [=] g cm⁻²) is the annual production of particulate organic matter and its subsequent incorporation in the soil's surface horizons. The transformation of particulate matter to humus is predicated on soil temperature, microbial acidity, the soil's oxidation–reduction status, pH and nutrient availability. The rate of organic matter accumulation per unit surface area (x) is the difference between the annual production of particulate organic matter per unit area and the rate of soil organic matter loss per unit area, expressed as a first-order linear ordinary differential equation:

$$dx / dt = P - \alpha x, \qquad (2)$$

(3)

where α is the decay coefficient, and t is time (years). Integration using an integration factor provides a solution:

From Clymo [9] typical decay constant values include $\alpha = 0.05$ and 0.15 year⁻¹. Also, from Clymo [9] typical annual production of particulate organic matter values includes: 150 and 450 g m⁻² yr.⁻¹. Using Eq. 3, The mass accumulation is presented for two scenarios: (i) P = 450 g m⁻² yr.⁻¹ and $\alpha = 0.15$ year⁻¹ (upper line in **Figure 1**) and (ii) P = 150 g m⁻² yr.⁻¹ and $\alpha = 0.05$ year⁻¹ (lower line in **Figure 1**). The scenario (i) P = 450 g m⁻² yr.⁻¹ and $\alpha = 0.15$ year⁻¹ provides a greater annual production of particulate organic matter and a faster rate of decay, such that the ratio P/ α is a limit point as t approaches infinity. The asymptotic approach to P/ α as a limit point implies that the net annual accumulation of organic matter ultimately becomes constant.

 $\mathbf{x} = (\mathbf{P} / \alpha) (1 - e^{-\alpha t}).$

Street et al. [10] in Svalbard considered the influence of phosphorus (P) on the decomposition potential of carbon stocks. Nitrogen additions supported carbon stock reductions because of enhanced soil organic matter decomposition; however, the combination of added nitrogen and phosphorus supported an increase in the



MASS ACCUMULATION OVER TIME

Figure 1.

Illustration of mass accumulation per year (0 to 3500 g m^{-2} yr.⁻¹) versus time (40 years) using Eq. (2). The primary vegetation productivity was 150 and 450 g m^{-2} yr.⁻¹ and the decay coefficients were 0.05 and 0.15 year⁻¹, respectively.

carbon stocks because of stimulated plant production. In Poland, Sienkiewicz et al. [11] investigated Histosol soil organic carbon and its relationship to total nitrogen, dissolved organic carbon and dissolved organic nitrogen. Carbon and nitrogen loss rates were independent, and soil organic carbon losses were dependent on the soil organic carbon content. The ratio of dissolved organic carbon to soil organic carbon increased with respect to the intensity of soil organic matter decomposition. Turunen et al. [12] investigated wet deposition of nitrogen (0.3 to 0.8 g nitrogen m⁻² yr.⁻¹) in ombrotrophic peatlands in eastern Canada, noting that nitrogen additions supported a greater diversity of vascular plants.

Qui et al. [13] modeled northern peatland areas and carbon changing aspects during the Holocene. They recognized that the net primary production (NPP) and heterotrophic respiration increased over the past century in response to climate change and increased atmospheric CO_2 activity. In their study net primary productivity was a greater influence than heterotrophic respiration, with 11.1 Pg C accumulated carbon storage since 1901, with the majority of the carbon storage increase occurring after 1950.

5. Research studies focusing on soil chemistry with emphasis on low molecular weight carbon species

The literature is replete with compelling research documenting biologically mediated geochemical pathways that are instrumental in creating vibrant biomes that have substantial accumulations of soil organic matter. Microbial populations secrete extracellular enzymes that are specific for degrading organic functional groups. The effectiveness of these extracellular enzymes is a complex function of (i) peat chemistry and litter quality, (ii) nutrient status, (iii) moisture content, (iv) plant community composition, (v) microbial community representation, and (vi) temperatures [14]. The absence of oxygen may also result in the accumulation of phenolic compounds that impost a negative feedback on microbial activity. Key enzyme activities important to mineralization include: (i) alpha-glucosidase, (ii) beta-glucosidase, (iii) cellobiohydrolase, (iv) N-acetylglucosaminidase, (v) acid phosphatase, and (vi) leucine aminopeptidase.

Fox [15] reviewed literature involving low-molecular-weight organic acids. Low-molecular weight organic acids are approximately 10% of a typical forest soil's dissolved organic carbon pool, but they may have a disproportionate influence on soil processes, including metal complexation. Common low molecular weight organic acids include: acetic, aconitic, benzoic, cinnamic, citric, formic, fumaric, gallic, lactic, malic, maleic, malonic, p-hydroxybenzoic, phthalic, protocatechuic, oxalic, salicylic, succinic, tartaric, and vanillic. Common functional groups include (i) acidic groups [carboxylic (R-COOH), enolic (R-CH=CH-OH), phenolic (Ar-OH) and quinones (Ar = O)], (ii) neutral groups [alcoholic OH (R-CH₂OH), ethers (R-CH₂-O-CH₂-R), ketones (R-C=O (-R)), aldehydes (R-C=O(-H)) and esters (R-C=O(-OR)) and (iii) neutral nitrogen-bearing amines $(R-CH_2-NH_2)$ and amides (R-C=O(NH-R)). When considering root extracts oxalic, citric and malic are quite abundant. Sources of low molecular weight organic acids are root respiration, leaching from the litter floor, decomposition of soil organic matter, and rainfall. Herbert and Bertsch [16] further detailed dissolved and colloidal organic matter in the soil solution. Based on their review of literature dissolved organic matter is primarily composed of hydrocarbons, chlorophyll, carotenoids, phospholipids and long-chain fatty acids, tannins, flavonoids and other polyphenols, fulvic and humic acids, aromatic and aliphatic acids, and proteins /amino acids. In most studies the dominant organic materials were humic substances.

Kane et al. [17] measured pore water chemistry associated with an artificiallyinduced warming of a nutrient poor fen. The dissolved organic carbon (DOC) concentration was greater in the warmed fen $(73.4 \pm 3.2 \text{ mg L}^{-1})$ compared to the untreated check (63.7 \pm 2.1 mg L⁻¹). The amount of dissolved organic nitrogen (DON) was greater in the warmed fen; however, the DON/DOC ratio was smaller. The reduced DON/DOC ratio was primarily attributed to a smaller capacity of the microbial community to yield labile nitrogen via the decomposition process and the greater utilization efficiency of the nitrogen by the microbial community. In Manitoba (Canada) Aide and Cwick [18] studied Eluviated Eutric Brunisols having an Of-Bm-C horizon sequence and Orthic Eutric Brunisols having an Oh or Of-Bm-C horizon sequence. Located in the glacial Lake Agassiz these soils formed in fine-graine lacustrine sediments interspersed with organic soils and fens. The surface horizons of the Eluviated Eutric Brunisols possessed organic carbon contents ranging from 19.8 to 29.4% with C/N ratios of 29.5 to 27.4, whereas the surface horizons of the Orthic Eutric Brunisols possessed organic carbon contents ranging from 27.3 to 41.7% with C/N ratios of 39.5 to 25.4. The C/N ratios and associated nitrate-N concentrations suggests that nitrogen limits the rates of soil mineralization. In a near companion manuscript Aide et al. [19] documented that the silty sediments were dominated by hydroxy Al-interlayered vermiculite, smectite, hydrous mica, and kaolinite in the clay separate. The potential for potassium fixation by vermiculite was reduced by Al-interlayering.

Van Cleve and Powers [20] isolated state factors involved in carbon storage in forest soils, noting the role of climate, parent material, topography, vegetation, and soil organisms. The chemistry of soil organic carbon, including root exudates and leachates, strongly influence the microbial processing of detritus, the materials synthesized in this process and the intensity of the roles that low and high molecular weight organic acids have in soil development. Observed effects show that synthesized products are more resistant to further decomposition and possessed smaller nitrogen contents, which over time supports soil organic matter accumulation.

6. Research studies having a focus on carbon loss as greenhouse gas emissions

Peatlands are an important terrestrial carbon sink and any increased microbial activity may result in soil organic matter oxidation, with subsequent CO_2 release. Northern peatlands historically have had the benefit of cool to frigid temperatures that limit microbial activity. Low oxygen activity attributed to water saturation further limits mineralization. Climate change may result in warmer soils, with the cavate that the effective length of the increasingly warmer summer interval is also increased. The encroachment of vascular plants will be expected to proceed, leading to a positive feedback on microbial activity. Thus, studies on peatland functioning in higher latitudes and their potential to accelerate climate change are becoming commonplace [14].

In Canada, Dieleman et al. [21] established mesocosms, where peat production of dissolved organic carbon was measured. The production of dissolved organic carbon from peat was estimated to be a function of temperature, CO_2 concentration and the influence of the water table, wherein increased temperatures increased the dissolved organic carbon contents, lowered water tables increased decomposition rates and reduced pore water dissolved organic carbon concentrations. In the Alaskan arctic Euskirchen et al. [22] established eddy covariance flux towers across various ecosystems for three years to document peak CO_2 uptake patterns. Peak CO_2 uptake centered from June to August at a mean of 51 to 95 g C m⁻² across the various ecosystems. Warmer spring seasons promoted greater CO_2 uptake patterns, whereas warmer late seasons supported greater soil respiration rates, reducing the Net Ecosystem Exchange (NEE).

In Canada, Frolking et al. [23] employed the Holocene Peat Model to simulate the vegetation community composition and the annual net primary productivity. Northern peatlands take up CO₂ at rates of 40 to 80 g carbon m⁻² yr.⁻¹, with carbon leaching as DOC at rates of 10–20 g DOC m⁻² yr.⁻¹. Decomposition was estimated to be 95% of the Net Primary Productivity. Similarly, Frolking et al. [23] observed undisturbed Canadian peatlands and determined that these peatlands were a weak sink for carbon and a moderate source of methane emission. McLoughlin and Webster [24] performed a review of peatland dynamics, primarily within the Hudson Bay Lowlands. Long term carbon accumulation, CO₂ sequestration, peat depth and land age were positively correlated. Carbon dioxide sequestration showed the greatest variability, with bogs (–1.7 to 1.5 g carbon m⁻² day⁻¹), fens (–4.3 to 1.6 g carbon m⁻² day⁻¹), and palsa peat (–0.8 to 1 g carbon m⁻² day⁻¹). Methane and evapotranspiration were greater in the wettest ecosystems, with methane emission for bogs (3.3 to 28 mg carbon m⁻² day⁻¹), fens (0.1 to 204 mg carbon m⁻² day⁻¹), and palsa peat (–1.6 to 24 mg carbon m⁻² day⁻¹).

On paludified soils Schneider et al. [25] measured methane (CH₄) flux for forest and peatland areas. Open peatlands exhibited a methane emission rate of 21.9 ± 1.6 g m⁻² yr.⁻¹ in contrast with forested peatland transition zones (7.9 ± 0.5 g m⁻² yr.⁻¹). The forested peatland transition zones demonstrated an inflow of less acidic surface water that supported a higher biological diversity and greater plant productivity. These authors noted that methane emission was more influenced by increased temperatures than the water table depths. In Sweden, Sagerfors et al. [26] established eddy covariance measurements across oligotrophic mires. Based on the vertical exchange of CO₂ their sites were a net sink for carbon (55 ± 7 g carbon m⁻² yr.⁻¹). The non-growing seasons exhibited a carbon loss; however, the growing season sequestration of carbon more than compensated for the non-growing season carbon loss. Wickland et al. [27] observed changes in CO₂ and methane exchanges on a black spruce (*Picea mariana*) lowland experiencing permafrost melting. Sites were partitioned as peat soils having permafrost, thermokarst wetlands, and thermokarst edges, with thermokarst edges having greater methane emissions. Ernakovich et al. [28] measured greenhouse gas emissions from thawed permafrost with simulated oxic and anoxic redox environments. Carbon dioxide emission was supported by an active microbial community and a labile dissolved organic carbon pool. Increased methane production was related to soils with a labile litter pool. Carbon dioxide emission was 30 to 450 times the methane production in an anoxic soil and carbon dioxide emission was 500 to 2500 times the methane production in an oxic soil.

In Canada, Webster et al. [29] investigated net ecosystem exchange and methane emissions for bogs, nutrient-poor fens, intermediate-rich fens across seven ecozones. During the growing season, the net ecosystem exchange, per season, was -108 ± 41.3 Mt. CO₂ and the methane emissions were 4.1 ± 1.5 Mt. CH₄. Converting methane to CO₂ global warming potential for a 25 to 100-year event, the total sink was -7 ± 77.6 Mt. CO₂e. The boreal plain peatlands exhibited the greatest net ecosystem exchange, whereas the boreal shield peatlands exhibited the highest methane emissions. In the discontinuous permafrost zone of western Siberia, Shirokova et al. [30] showed that permafrost thaw supported an increase in soil subsidence and the development of thermokarst lakes. Soil subsidence was related to soil carbon decomposition and mobilization to water resources.

Jackowicz-Korczynski et al. [31] observed methane emission from subarctic Swedish mires. A permafrost free mire having tall graminoid vegetation showed methane emission rates of $6.2 \pm 2.6 \text{ mg CH}_4 \text{ m}^{-2} \text{ hr.}^{-1}$. The annual emission was 24.5 to 29.5 g CH₄ m⁻² yr.⁻¹, with most of the emission during the summer months. In Wales (UK), Fenner et al. [32] investigated ombrotrophic peat or acid mires. Artificially enhanced CO₂ and warming produced increased concentrations of dissolved organic carbon. Higher concentrations of phenolic compounds were associated with the increase in dissolved organic carbon. The influence of increased temperature promoted microbial activity, whereas increased CO₂ content increased the supply of photosynthate to the soil because of greater root exudates. The effect of the temperature and elevated CO₂ were to synergistically decrease the C/N of the dissolved organic carbon. In Indonesian tropical peatlands, Uda et al. [33] noted that land drainage influenced CO₂ emissions from drained oil palm landscapes.

Aurangojeb et al. [34] contrasted a drained Histosol and an adjacent mineral soil in Sweden, noting that the Histosol N₂O emissions were 49.9 \pm 3.3 µg N₂O m⁻² hr.⁻¹, whereas the adjacent mineral soil N₂O emission was 8.0 \pm 3.3 µg N₂O m⁻² hr.⁻¹. The N₂O difference was attributed to the mineral soil having greater mycorrhizal N demand reducing the N availability. Leifeld et al. [35] investigated four temperate ombrotrophic peatlands across central Europe and determined that ash content is related to land drainage and land management, thus ash may be an indicator of historical decomposition but this protocol should be used only in pristine study areas.

7. Peatlands, net primary productivity and climate change

Net primary production is critical to developing large carbon contents in peatlands. Net primary production is a function of climate, vegetation, topography, the natural of the parent materials, and land use. Investigating Swedish peatlands, Chaudhary et al. [36] investigated drivers of biotic and abiotic peatland dynamics. For patterned ground they noted that plant species, hydrology, nutrient status, plant productivity and decomposition rates vary between hummock and hollow positions. Typically hollows possessed taller productive graminoid species that

showed faster decomposition rates than sphagnum. Hummock positions possessed more shrub species that preferentially lowered the water table. In interior Alaska, O'Donnell et al. [37] studied Gelisols having a 30-day enhanced temperature incubation period, noting that the dissolved organic carbon concentration and its associated aromaticity increased at higher incubation temperatures. At these higher temperatures the dissolved organic materials contained more hydrophobic organic acids, polyphenols, and condensed aromatics and smaller concentrations of lowmolecular weight hydrophilic and aliphatic compounds. Dissolved labile organic materials were preferentially mineralized, with the dominant kinetic controls being temperature and substrate lignin contents.

Wang et al. [38] correlated that increased mean annual air temperature was associated with increased active layer thickness. In a Siberian low arctic landscape, Frost et al. [39] documented seasonal and long-term changes to active layer temperatures and noted that vegetation and snow cover were important predictors of active layer thickness. Summer soil temperatures decreased with increasing shrub cover and soil organic matter thickness. Compared with open tundra, mature shrubs depressed summer soil temperatures; however, mature shrubs altered the insulative snowpack and fostered warmer winter soil temperatures.

In Canada Kroetsch et al. [5], working with the National Wetlands Working Group, noted that peatlands were routinely identified when peat depths exceeded 0.40 meters. Fibrisol, Mesisol and Humisol great groups were partitioned based on rubbed fiber content, von Post scale, pyrophosphate and depth of the surface, middle and bottom tiers. The key diagnostic genetic processes of organic soils included: (i) additions from litter, fine roots, soil organic matter deposition and low molecular weight organic acid exudation from sphagnum, feather mosses and related plant species, (ii) losses attributed to decomposition, (iii) transfers of dissolved organic carbon because of fluctuating water tables, leaching and burrowing organisms, (iv) transformations attributed to soil organic matter decomposition, O₂ status, nutrient availability, and toxins.

Glaser et al. [40] observed Hudson Bay Lowlands peatland development from a chronological perspective, relating the length of time for isostatic rebound to elevate the landscape and developing a transect of peatland sites ranging from comparatively younger to older sites. They observed that the resulting transects consisted of a sequence consisting of (i) basal tidal marshes in the youngest sites, (ii) Larix (Larches) dominated swamp forests, (iii) Picea (Spruce) forested bogs, and ending with (iv) non-forested bogs in the oldest sites. This sequence of peatlands was viewed as a predictable vegetation succession influenced by changes in hydrology and other factors derived from continuing isostatic rebound. Conversely, in western Siberia, peatlands demonstrated an increase in carbon accumulation upon transition from the northern region to the southern region [41]. The northern peatlands exhibited a carbon content of 7–35 kg carbon m⁻², whereas the southern peatlands exhibited a carbon content range of 43–88 kg carbon m⁻². The carbon content was estimated to be a complex function of soil organic matter quality (lignin content) and the predominant vegetation (vascular plants versus bryophytes).

8. Peatlands and their conservation

Karofeld et al. [42] noted Estonia's decline of pristine mires and investigated a method for mire reconstruction, involving the removal of oxidized peak layer followed by the spreading of plant fragments to increase the effective development of bryophyte and vascular plants. Along with maintaining the presence of a highwater table, the reconstruction effort was deemed successful. Miettinen et al. [43] employed satellite images to document the role of fire and logging on the loss of Sumatra's pristine peat swamps. In Indonesia, Swails et al. [44] investigated soil respiration as a climatic driver in undrained forest settings and adjacent oil palm plantations. They documented that oil palm plantations with a reduced water table exhibited a higher soil respiration rate $(0.71 \pm 0.04 \text{ g CO}_2 \text{ m}^{-2} \text{ h}^{-1})$ than forested sites $(0.58 \pm 0.04 \text{ g CO}_2 \text{ m}^{-2} \text{ h}^{-1})$.

Across Poland, Grzywna [45] documented drainage-induced Histosol subsidence ranges from 9 to 33 cm. Nicia et al. [46] demonstrated that restoration of peatlands in Poland has potential to increase the organic carbon content, the C/N ratio and increase the pH in acidic fens. Richardson [47] noted the development sequence of alkaline mires (fens) in the Everglades (Florida) and the role of changing hydrology during the Holocene. In Wisconsin, Adhikari et al. [48] used digital maps and soil profile data to spatially quantify carbon stocks and subsequently estimated the fate of carbon stocks with improved land use management. The average baseline soil organic carbon stock was 90 mg ha⁻¹ and with improved land management the soil across the state could increase the carbon stocks by 20 mg ha⁻¹. Mollisols were predicted to have the greatest potential for increasing carbon stocks, whereas Histosols and Spodisols were likely to lose carbon stock. Frazier and Lee [49] investigated Wisconsin Histosols partitioned as fibrists, hemists and saprists. Saprists possessed the highest carbon content, whereas the fibrists possessed the least carbon content, a feature related to chemical changes associated with the humification process.

9. Future possibilities

The fate of peatland ecosystems is integral to global sustainability. As scientists, we are acutely aware that carbon stored in peatland ecosystems may be released to the atmosphere, contributing to climate change acceleration. The precise drivers of peatland respiration, the role of the microorganism communities, organic acid leaching, soil mineralization, and other soil carbon pathways are reasonably well understood, but they are not sufficiently formalized into a coherent and interconnected model to provide detailed information concerning near-term peatland degradation [50–55]. Thus, a critical need exists to predict on a regional level specific changes to peatland dynamics because of the multi-faceted nature of accelerated climate change. With this process focus on peatland dynamics, best management practices are slow carbon de-sequestration.

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