Controls on dissolved organic matter concentrations in soils and streams from a forested wetland and sloping bog in southeast Alaska†

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ABSTRACT
Understanding the sources and removal mechanisms for dissolved organic matter (DOM) in soils is critical to determining the importance of peatlands in influencing streamwater DOM concentrations. Soil saturation and soil temperature are the primary factors that influence DOM cycling in soils, but the relationships among these factors vary by soil type and climate. Despite the extent of peatlands that occur within Pacific coastal temperate rainforest (CTR) watersheds, there is little information describing how DOM storage and export patterns are related to soil saturation and temperature in the region. We measured soil water tables, soil temperatures and redox potential and compared these measurements to fluctuations in dissolved organic carbon (DOC) and nitrogen (DON) concentrations in a forested wetland and sloping bog in southeast Alaska to address this key information gap. DOC concentrations ranged from 5 to 140 mg C l$^{-1}$ in wetland soils, 11 to 46 mg C l$^{-1}$ in streams, and varied greatly in response to changes in water table, redox potential and soil temperature. Similarly, DON concentrations ranged from 0-03 to 2-4 mg N l$^{-1}$ in wetlands, 0-2 to 0-6 mg N l$^{-1}$ in streams and concentrations also reflected seasonal changes in physical measures. Depth to water table and soil temperature were significant factors related to the concentration of DOC in forested wetland soils and streams, while soil temperature was a significant factor that influenced stream DOM and DON concentrations. Comparing soil solution and stream DOM concentrations indicated that N is retained in bogs, while both dilution and biotic/abiotic retention mechanisms control DOM export in forested wetlands. The relationships among soil saturation, temperature and concentrations of DOC and DON provide preliminary models for estimating DOM fluxes from wetland soils in the CTR. Furthermore, these results can help calibrate regional watershed carbon flux models to predict the potential impacts of climate shifts and management activities on future wetland soil and stream DOM concentrations in CTR watersheds. Copyright © 2009 John Wiley & Sons, Ltd.

KEY WORDS acrotelm; peatlands; peat hydrology; biogeochemistry

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INTRODUCTION
A fundamental question regarding the stabilization or loss of dissolved organic matter (DOM) in soils is the magnitude of DOM sources and sinks (McDowell, 2003). DOM is produced by the decomposition of plant litter and cycling of microbial biomass (Kalbitz et al., 2000; Michalzik et al., 2001). These biotic processes are influenced by soil temperature, soil aeration and organic matter quality (Berg et al., 1993; Schimel et al., 1994; Bridgham et al., 1995). Low temperatures and saturated soils increase peat accumulations via reduced decomposition and accumulation of recalcitrant organic material during decomposition (Berg and McLaugherty, 2008). In mineral soils, abiotic factors such as iron, aluminium and clay content influence DOM stabilization into soil organic matter (SOM) (McDowell and Wood, 1984; McDowell and Likens, 1988; Moore et al., 1992; Qualls and Haines, 1992; Kaiser and Zech, 1998). The retention and release of DOM in organic soils is also influenced by abiotic and biotic binding that can limit DOM solubility and lead to retention of DOM by the complex interaction among functional groups (Qualls, 2000; Lumsdon et al., 2005). The solubility of DOM and hydrologic flushing may be more critical than overall production in controlling the amount of DOM exported from organic soils (Worrall et al., 2008). Therefore, the integration of soil physical properties and chemical cycles with hydrologic export is needed to fully understand the mechanisms of SOM storage and export (Eckhardt and Moore, 1990; Dillon and Molot, 1997; Gorham et al., 1998; Aitkenhead et al., 2005). The delivery of terrestrial DOM from soils through groundwater to streams varies with the seasonal balance between production and loss from solution and the potential for hydrologic transport (Boyer et al., 1997; Yano et al., 2000; Pastor et al., 2003). If the production of DOM in the soil exceeds the capacity for storage or internal consumption, DOM is available for export from the soil.
Understanding the DOM cycle of northern peatlands is particularly important because nearly one-third of the estimated global soil carbon pool is stored within northern biomes (Gorham, 1991). In particular, organic matter has accumulated over the past 6000 years in the coastal temperate rainforest (CTR) of southeast Alaska, resulting in the formation of extensive peatlands (Heusser, 1952, 1960). These peatlands store most of the terrestrial carbon in the region (Alexander, 1989; Leighty, 2006). There is some evidence for increased DOM export in these soils and associated streams (Emili and Price, 2006). There is some evidence for increased DOM export from peatlands to aquatic ecosystems due to climate warming in Great Britain (Freeman et al., 2001; Worrall et al., 2004). While there is evidence that decreased sulphate deposition may also be a factor (Evans et al., 2006), the potential that stored SOM may be exported as DOM through increased microbial activity and decomposition under warmer conditions (Nadelhoffer et al., 1991; Evans et al., 1999) is of concern in the CTR. Increased DOM concentrations in the soils could be exported via shallow subsurface flowpaths, which could intensify with increased precipitation (Dalva and Moore, 1991; Qualls, 2000; Emili and Price, 2006). However, increased precipitation may also inhibit decomposition and subsequent DOM production. Therefore, understanding how DOM concentrations vary in response to seasonal changes in soil temperature and saturation is important for elucidating DOM cycling in peatlands of the CTR.

There are three primary fates for DOM produced in the soil matrix: (1) export from the soil, (2) consumption in situ by microbial respiration, and (3) retention within the soil matrix through adsorption. There is currently no integrated understanding of how these primary pathways are influenced by soil saturation and temperature in the CTR of southeast Alaska. We compared measurements of soil water table level, redox potential and soil temperature to soil solution and streamwater concentrations of dissolved organic carbon (DOC) and nitrogen (DON) to expand our understanding of the soil physical factors related to DOM concentrations and export in two dominant peatland soil types in southeast Alaska. We established empirical relationships between the physical fluctuations and the response of dissolved concentrations in soils and streams from 2 years of data in these two soil types.

**MATERIALS AND METHODS**

**Site description**

Research was conducted in the McGinnis Creek watershed near Juneau, AK, which is approximately 123 km from the open ocean and 23 km from the marine waters of the inside passage (58.5°N, 134.5°W; Figure 1). The McGinnis Creek watershed is 1720 ha of recently deglaciated slopes with sparse vegetation and thin soils in the upper watershed, and an older, uplifted marine terrace with old-growth spruce and hemlock forest interspersed with in wetlands at the lower part. The study area is located in the lower watershed near the mouth of McGinnis Creek, where peat has formed over uplifted marine sediments that were deposited along valley slopes smoothed by glacial ice. The slope bog site (MB) has approximately 4 m of peat over a slowly permeable glacio-marine deposit identified as the Gastineau formation (Miller, 1973). Vegetation consists of several sphagnum mosses (*Sphagnum* spp.), ericaceous shrubs, sedge (*Carex* spp.) in isolated spots, and shore pine (*Pinus contorta* var. *contorta*). This site is typical of the palustrine emergent (National Wetland Inventory, NWI, USA) or slope bog (National Wetlands Working Group, NWWG, Canada) wetland type. The slope bog has a subtle slope from a topographic high occurring near the sampling area downward toward a coalescing outlet channel. Main hydrologic inputs are precipitation near the topographic high that coalesces into shallow surface drainage zones that form depressions and small tributary channels that flow toward the west fork of McGinnis Creek (Figure 1).

The forested wetland (MF) is a shallower peatland than the slope bog with approximately 0.7 m of organic material over Gastineau formation (Figure 1). The forested wetland vegetation consists of sphagnum moss (*sphagnum* spp.), and forbs including both obligate wetland plants such as skunk cabbage, *Lysichiton americanum* and plant species common to upland (e.g. bunchberry, *cornus Canadensis*) soils. The overstory is dominated by western hemlock (*Tsuga heterophylla*) and Sitka spruce (*Picea sitchensis*), and Alaska blueberry (*Vaccinium* spp.) is common in the understory. This site is typical of the palustrine forested (NWI) category.

**Soil description and sampling**

Soil profile descriptions were developed for each site (Soil Survey Division Staff, 1999). The soil was characterized to 2-m depth in the slope bog using an auger from 132 to 200 cm, whereas the forested wetland was described down to the impermeable horizon at 162 cm (Table I). A modified syringe adjusted to 5-ml capacity and a 100-mesh sieve was used to calculate rubbed and unrubbed fibre content (Lynn et al., 1974; D’Amore and Lynn, 2002). Samples from each soil horizon were dried and ground to pass through a 177 μm (80 mesh) sieve for total C and N analysis by combustion on a CHN analyser. Soil pH was determined on moist 2.5 cm³ subsamples in 4 ml of 0.01 m CaCl₂. Bulk density was determined by measuring pore volume on 125 cm³ samples carved from soil pit faces with sharp knives. Organic matter particle density was assumed to be 1 g cm⁻³. Gravimetric water content was determined on three replicate 50 g subsamples of soil dried at 105 °C to a constant weight.

Water tables were measured using pressure transducers attached to dataloggers in three 50-cm deep wells constructed from 1.25 in PVC pipe and inserted at 10 m
Figure 1. Map of Alaska, location of McGinnis Creek watershed, and experimental design with sub-catchments of the forested wetland (MF) and bog (MB), soil solution sampling sites and outlet stream sample locations near Juneau, Alaska.

intervals across the site. Saturated hydraulic conductivity was determined using the Hvorslov calculation (Freeze and Cherry, 1979) on slug and withdrawal tests in the wells during the fall of 2004. Soil temperature was recorded with thermistor probes placed at 10 and 20 cm depths. Redox potential was measured with five replicate platinum electrodes constructed similar to Faulkner et al. (1989) at 10, 20 and 30 cm depths. We used a theoretical anaerobic threshold that was between 400 and 500 mV based on the relationship between $eH$ and pH 4–5 (Stumm and Morgan, 1981). We used a broad threshold because determining the specific threshold of aerobic conditions in a system of mixed potentials driven by many electroactive species in the soils is uncertain. Hourly measurements for soil temperature, water table levels and redox potential were obtained by field data recorders and aggregated into daily averages for analysis. The redox dataloggers failed in 2005 leaving a gap in redox data from August to October.

Grab samples were used to collect streamwater from the small tributaries draining the bog and forested wetland sites. Soil pore water samples were taken at both sites from four zero-tension lysimeters installed at the morphologic horizon breaks in the soils located at 10 and 20 cm. Soil pore water was composited from the two lysimeters at each depth, yielding one sample from each depth.
Soil solution and streamwater samples were field-filtered, Gelman A/E glass fibre filters and stored in the refrigerator until analysis within 1 week of collection. DOC and total dissolved nitrogen (TDN) concentrations were analysed using a high temperature combustion, Shimadzu TOC/TN-V analyser. Ion chromatography (Dionex, ICS-2500 and ICS-1500) was used to measure NH$_4$–N (minimum detection 5 μg l$^{-1}$) and NO$_3$–N (minimum detection 2 μg l$^{-1}$) on syringe-filtered (nominal pore size 0.45 μm) samples. DON concentrations were calculated by subtracting NH$_4$–N and NO$_3$–N from TDN concentrations. Because DON was the dominant fraction of TDN in all water samples (≥ 85% for all samples), we report DON concentrations only throughout the paper.

**Statistical analyses**

Pearson correlation coefficient (SAS Institute, V. 9.3) was used to test the correlation between DOC and DON concentrations in lysimeters and streams with soil temperature and depth to water table. Multiple linear regressions (SAS Institute, V. 9.3) were used to test the combination of soil temperature and water table depth as explanatory factors for soil solution and stream DOC and DON concentrations. DOC and DON concentration in lysimeters and streams were natural log transformed to meet regression assumptions where appropriate. Models of soil solution and streamwater relationships were determined to be significant at α ≤ 0.05 and analyses were conducted separately for the bog and forested wetland.

**RESULTS**

**Pedologic development and morphological features related to carbon and nitrogen storage**

The two peatland soils were distinguished from one another at the soil series level by the depth and decomposition of organic matter. The MB soil was classified as a Typic cryohemist in the US system of soil taxonomy (FAO designation: Histosol; Canadian Soil Taxonomy: Mesisol). This soil was similar to the Kina soil series mapped throughout the Tongass National Forest. The MF soil was classified as a Terric cryohemist (US Soil Taxonomy; FAO: Histosol; Canadian: Mesisol) that was closely associated with the locally mapped Maybeso soil series (Terric cryosaprist; FAO: Histosol; Canadian: Humisol), except for the degree of fibre decomposition. The wide range of variability in decomposition of these soils has been recognized and the Terric cryohemist and Terric cryosaprist soils are closely related, though not recognized as distinct soil series in the Tongass National Forest soil resource inventory (D’Amore and Lynn, 2002).

Both the MB and MF sites had soils that reflect the results of Holocene peatland development and contained deep (>50 cm) and concentrated (>35%) organic carbon stocks that were extensive reservoirs of organic matter (Table I).

There were different physical attributes and distribution of the organic material in the surface horizons of the soils. The MB soil had a greater retention of fibre after rubbing in the surface horizons than the MF soil, but both soils had a similar bulk density of approximately 0.20 g cm$^{-3}$ (Table I). The MF and MB soils had about the same carbon storage to 1 m depth (approximately 11 kg C m$^{-2}$), but size distribution of the material varied as indicated by the rubbed fibre analysis. There was a zone of carbon enrichment from surface to subsurface horizons in MB, but carbon depletion in the same horizons at MF. The second horizons (Oi2 and Oe2) were then followed by carbon enriched horizons (Oe and Oe3 at MB and MF, respectively) in both soils, along with a decrease in rubbed fibre at MB (Table I). Both MB and MF soils had a pattern of nitrogen depletion and then enrichment in the top three horizons (Table I).

**Rainfall, soil hydrology and redox potential**

Increases in soil water tables for MB and MF were less pronounced in the spring and early summer during both years, due to lower overall rainfall (Figure 2A, B). However, heavy, sustained rainfall in the late summer...
and autumn caused more pronounced changes in water table levels at both sites. Early spring soil saturation was promoted by snowmelt indicated by the higher soil water table measurements in May 2005 (Figure 2B). We did not capture the maximum spring saturation in 2004, but the measurements during both years showed the water table drawdown in late spring to mid-summer (Figure 2A, B). The water table level at MB reached a maximum depth of approximately 25 cm below the soil surface during both years, while the drawdown at MF reached a maximum of approximately 35 cm during both years. These decreases in the free water table left approximately 10–20 cm of surface soil available for air infiltration at MB and about 20–35 cm of soil at MF during the early part of the growing season in both years (Figure 2A, B). There was less precipitation during 2004 and the duration of unsaturated conditions above 20 cm in both soils was longer than during 2005 for both sites (Figure 2A, B).

The water table fluctuations at MF were responsive to late summer storms with increases in water tables of 10–15 cm during both years (Figure 2A, B). The MB site was less responsive to individual events in the dry year of 2004 (Figure 2A) than the wet year in 2005 (Figure 2B). There was a stairstep pattern indicative of thresholds in the pore pressure within the peat at MB in 2004 and a more uniform response to rainfall in 2005 (Figure 2A, B). The water tables had uniform responses to rainfall at MF in both 2004 and 2005, but reached a maximum threshold of saturation at approximately 20 cm below the soil surface during 2004, and 10 cm in 2005. The water tables at MB reached a maximum of approximately 10 cm below the surface in each year (Figure 2A, B).

The soils at the two sites had permeable surface horizons over impermeable subsurface horizons as indicated by saturated hydraulic conductivity (Ksat) measurements (Figure 3). The Ksat at 10-cm depth at both
sites was greater than $10^{-2}$ cm s$^{-1}$ (Figure 4). The Ksat decreased to less than $10^{-4}$ cm s$^{-1}$ at MF and less than $10^{-5}$ cm s$^{-1}$ at MB at 25-cm depth. Both sites had nearly impermeable soil matrix conductivities at 40-cm depth where Ksat was less than $10^{-6}$ cm s$^{-1}$. The pattern of slightly higher Ksat in the MF soils at each depth was consistent throughout the profiles.

The surface soil at both sites was unsaturated for the entire season during 2004. Redox potential measurements above 500 mV confirmed that the unsaturated soil was aerobic at 10-cm depth at MB and at 20-cm depth at MF (Figure 4A,C). For MB, reducing conditions were stronger deeper in the soil profile (Figure 4A). Redox potential at 30 cm remained consistently anaerobic, while fluctuating conditions at the 20-cm depth declined from aerobic to anaerobic during the 2004 season (Figure 4A). The 30-cm redox potential measurements at MF fluctuated early in the 2004 season, but declined into the anaerobic range once the soil was saturated at 30-cm depth (Figure 4C).

The surface soils were aerobic during the early part of the season at MB in 2005, but steadily declined into the anaerobic range by July in response to increased soil saturation (Figure 4B). The soil surface at MF was aerobic during the early part of 2005, but dropped into the anaerobic zone once the soil became saturated at 20-cm depth by June (Figure 4D). In contrast to the aerobic conditions observed throughout the measurement period in the surface soils at MB in 2004, heavier rainfall in 2005 caused soil saturation and the onset of anaerobic conditions by mid-summer (Figure 4B). The subsurface redox potentials at MB declined into the anaerobic range precipitously, and much earlier in 2005, than in 2004. The patterns of redox potential and soil saturation were similar between years with one exception. While the soil
at 20-cm depth remained aerobic during 2004 at MF, the redox potentials dropped into the anaerobic range during 2005 due to higher levels of soil saturation.

**Concentration of DOC and DON in soils**

In the drier year of 2004, the MB soils were usually unsaturated at 10-cm depth (Figure 2), and water samples were obtained only during rainfall events. Measurements at 10 cm during these sampling periods had concentrations of DOC ranging from 5 to 42 mg C l⁻¹ (Figure 5A) and DON concentrations ranging from 0-2 to 1.8 mg N l⁻¹ (Figure 6A). During the wetter year of 2005, water samples were consistently available with DOC concentrations from 7 to 44 mg C l⁻¹ (Figure 5B) and DON concentrations from 0-1 to 1.6 mg N l⁻¹ (Figure 6B). Water samples were always available at 20-cm depth at MB due to permanent saturation during both 2004 and 2005. Concentrations of DOC at 20-cm depth in 2004 ranged from 15 to 40 mg C l⁻¹ in 2004 and 18 to 34 mg C l⁻¹ in 2005 (Figure 5A, B), whereas concentrations of DON were 0-3 to 1.2 mg N l⁻¹ in 2004 and 0-4 to 0.8 mg N l⁻¹ in 2005 (Figure 6A, B).

![Figure 5. DOC concentrations and soil temperatures for the forested wetland (MF) and sloping bog (MB) in 2004 and 2005. No measurement indicates a lysimeter with zero concentration. (A) MB 2004; (B) MB 2005; (C) MF 2004; (D) MF 2005.](image-url)

**Table II. Pearson correlation table for temperature, depth to water table and concentrations of DOC and DON at the MB and MF sites.**

<table>
<thead>
<tr>
<th></th>
<th>Bog</th>
<th>Forested wetland</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>10 cm (°C)</td>
<td>20 cm (°C)</td>
</tr>
<tr>
<td>DOC 10 cm</td>
<td>0.41 (0.13)</td>
<td>0.41 (0.13)</td>
</tr>
<tr>
<td>DOC 20 cm</td>
<td>0.33 (0.10)</td>
<td>0.39 (0.04)</td>
</tr>
<tr>
<td>DON 10 cm</td>
<td>0.25 (0.37)</td>
<td>0.30 (0.28)</td>
</tr>
<tr>
<td>DON 20 cm</td>
<td>−0.04 (0.84)</td>
<td>0.01 (0.97)</td>
</tr>
<tr>
<td>DOC stream</td>
<td>0.72 (&lt;0.01)</td>
<td>0.74 (&lt;0.01)</td>
</tr>
<tr>
<td>DON stream</td>
<td>0.59 (&lt;0.01)</td>
<td>0.55 (&lt;0.01)</td>
</tr>
</tbody>
</table>

Correlation coefficient and strength of correlation as indicated by P value in parentheses. DWT, depth to water table.
The DOC concentrations in the bog soils were better correlated with soil temperatures than depth to water table (Table II), but there were no significant relationships between temperature or depth to water table found in the regression analysis (Table III). We did not detect any relationship between DON concentrations and soil temperature or depth to water table (Table II).

The MF soils were unsaturated during 2004 at 10-cm depth (Figure 2), but water was consistently available for sampling due to throughflow after rainfall events. The 10-cm DOC concentrations peaked early in the measurement period during both years. The concentration of DOC at 10 cm ranged from 53 to 90 mg C l\(^{-1}\) in 2004 (Figure 5C) and 24 to 140 mg C l\(^{-1}\) during 2005 (Figure 5D). Concentrations in the 20-cm lysimeter ranged from 28 to 86 mg C l\(^{-1}\) during 2004 and 26 to 132 mg C l\(^{-1}\) during 2005. Concentrations of DON at 10 cm were 1–2–2–3 mg N l\(^{-1}\) during 2004 (Figure 6C) and 0–2–2–4 mg N l\(^{-1}\) during 2005 (Figure 6D). The concentration of DON at 20 cm at MF was similar to that at 10 cm during 2004 with concentrations from 0–1 to 2–1 mg N l\(^{-1}\) and 0–6 to 2–4 mg N l\(^{-1}\) during 2005. There was a distinct seasonal pattern in DOC and DON concentrations at both depths at MF during 2005. DOC concentrations peaked before the mid-summer temperature maximum at MF in 2005 (Figure 5D). Soil solution DOC concentrations were related to both soil temperature and depth to water table (Table II) and the combination of these variables was significantly related to the soil solution DOC concentration in the 20-cm lysimeters (Table III).

**Export of DOC and DON from soils to outlet streams**

Concentrations of DOC in the outlet stream at MB were similar to soil solution concentrations at 20-cm depth during both years (Figure 5A, B). Mean stream DOC concentrations at MB ranged from 11.8 to 41.2 mg C l\(^{-1}\) (Figure 5A, B). The mean stream DOC concentrations at MB were 26% and 17% higher than the 10-cm lysimeters in 2004 and 2005, respectively. Mean stream Don concentrations were 4% and 5% lower than the 20-cm lysimeters at MB in 2004 and 2005, respectively. The streamwater DOC concentration was significantly related to soil temperature at both 10 and 20 cm depths, but did not have a significant relationship with depth to water table (Table II). The concentrations of DON in the outlet stream at MB were lower than the soil solution concentrations at 10 and 20 cm throughout most of the sampling period and ranged from 0.2 to 0.6 mg N l\(^{-1}\) (Figure 6A, B). The mean stream DON concentrations...
at MB were 28% and 4% lower compared to the 10-cm lysimeters in 2004 and 2005, respectively. Mean stream DON concentrations compared to the 20-cm lysimeters at MB were 20% and 23% lower in 2004 and 2005, respectively. Streamwater DON concentrations at MB were consistently about 0.5 mg N l⁻¹ throughout the measurement period (Figure 6A, B).

Concentrations of DOC exported in the outlet stream at MF were lower than soil solution concentrations during both years (Figure 5C, D). Mean stream DOC concentrations at MF were 39% and 17% lower than the 10-cm lysimeters in 2004 and 2005, respectively. Mean stream DOC concentrations compared to the 20-cm lysimeters at MB were 26% and 30% lower in 2004 and 2005, respectively. There was a consistent outlet streamwater DOC concentration of approximately 30 mg C l⁻¹ maintained throughout the measurement period at MF despite more variable and higher concentrations in the soil (Figure 5C, D). The most striking decrease from soil to streamwater DOC concentration occurred in 2005 between the peak soil DOC concentrations of 120–140 mg C l⁻¹ and the streamwater DOC concentrations of 30–40 mg C l⁻¹ (Figure 5D). DON concentrations in the outlet stream at MF are lower than the soil solution concentrations throughout most of the sampling period (Figure 6C, D). The mean stream DON concentrations at MB were 53% and 33% lower compared to the 10-cm lysimeters in 2004 and 2005, respectively. The mean stream DOC concentrations compared to the 20-cm lysimeters at MB were 28% and 43% lower in 2004 and 2005, respectively. The outlet stream DON concentration at MB was approximately 0.5 mg N l⁻¹ throughout the measurement period (Figure 6C, D).

The DOC : DON ratios in MB ranged from 3 to 65 in soil solution and 40 to 75 in the outlet stream across the measurement period. This led to an increase in the DOC : DON ratio from approximately 40 : 1 in soil solution to 60 : 1 in streamwater (Figure 7). For MF, DOC : DON ratios ranged from 11 to 99 in the soil solution and 40 to 76 in streamwater, which resulted in an average increase of approximately 40 : 1 in soil solution to 60 : 1 in streamwater (Figure 7). There is a much broader range of DOC and DON values in the MF lysimeters compared to MF streams than the MB site. While the stoichiometric ratios are similar, the patterns for MF and MB are quite different. The DOC : DON at MB shifts from soil to stream due to a shift in DON values, while the MF site shifts from soil to stream due to decreases in both DOC and DON concentrations.

**DISCUSSION**

The diplotelmic system proposed by Ingram (1978) provides a model for comparing DOM cycles in peatlands of the CTR with peatlands from other regions around the world. The duration of saturation, saturated hydraulic conductivity and redox potential measurements from this study confirmed the presence of an aerobic surface layer (acrotelm) and a permanently saturated subsurface layer (catotelm) consistent with the diplotelmic model. The complex mechanisms controlling retention and release of DOM in soils (Michalzik et al., 2001) and preferential flowpaths (e.g. Worrall et al., 2002, 2003) are not completely represented by the diplotelmic model. However, the diplotelmic system creates a horizontally stratified template to examine the thermodynamic constraints on metabolic potential exerted by other variables. Dramatic changes in redox potential driven by vertical movement of the water table created a mode shift in biochemical reactions that control mineralization rates, available metabolic pathways and terminal electron acceptors and determine the amount and quality of DOM available for transport and cycling. The distinct difference in behaviour between MB and MF indicate that these functionally distinct landscape units must be recognized in nutrient cycling and process oriented hydrologic models. Given the difference in topography, water source and hydraulic conductivity between the two functional units we would predict that they will respond at different rates and magnitudes to impending changes in mean temperature and seasonal rainfall distribution predicted by climate models (SNAP, 2009). Accurate models at anything less than regional scales should incorporate the differences in controlling dynamics between these two widespread landscape units.
The hydrologic fluctuations associated with the acrotelm/catotelm were previously described in southeast Alaska, but neither the implications for soluble DOM linkages between organic soils and streams nor an explicit description of the diplotelmic model were considered. Siegel (1988) found that peatlands in the Juneau area were not a significant component of regional groundwater flow (i.e. recharge) because of the strong vertical zonation in the peat. Water movement through the soil matrix is closely associated with the degree of organic matter decomposition (Boelter, 1969) and the degree of decomposition and associated water movement generally decreases with depth (Mathur and Levesque, 1985). The low permeability peat deposits in the catotelm, or aquihart, have a high water holding capacity due to abundant pore space created by organic material fibres in the soil matrix (Boelter and Verry, 1977; Siegel, 1988). Once this layer becomes saturated, water flows through the highly conductive surface horizons (Gorham and Hofstetter, 1971; Chason and Siegel, 1986) and creates sheet flow across the peat surface (Bishop, 1968).

The depth and duration of aerobic conditions in the CTR acrotelm is similar to other regions (Holden and Burt, 2003; Worrall et al., 2002, 2003), but the frequency and magnitude of water table fluctuations and resulting variability in the redox potential are much more dynamic in response to events and seasonal changes. The hydrodynamics and intensity of the biogeochemical exchange in the acrotelm also differs between the MB and MF sites and provides an explanation for the local strength of the association between temperature and redox potential. The water table drawdown combined with the expansion of aerobic conditions increases organic matter decomposition (Qualls et al., 1991; Palmer et al., 2001) creating a pool of soluble organic matter subject to hydrologic transport. Water moving through the conductive acrotelm perched on the less permeable catotelm flushes this soluble material to outlet streams. Therefore, the combination of SOM processing and hydrodynamic fluctuations in the MF acrotelm lead to both large soil solution DOM concentrations, and high stream DOM concentrations. The oxidation of SOM between storm flushing events is critical to the replenishment of the DOC reservoir in the acrotelm (Worrall et al., 2002). In addition, reduced hydrologic flushing influences the potential C storage or loss in peatlands, not the concentration of available DOC in the soil (Pastor et al., 2003). These findings highlight the need for increased DOC and DON at MF (Fröberg et al., 2003). Measurements made with platinum electrodes in our study appear to be sensitive to soil matrix conditions and indicate possible shifts in terminal electron acceptors (Dettling et al., 2006). The reduced SOM is the primary source of electrons for microbial communities in the soils and redox potential is an indicator of soil biogeochemical behaviour that can be used as a surrogate measure to predict subtle shifts in organic matter transformation (Fiedler and Kalbitz 2003; Mitchell and Branfireun 2005). Though we do not have direct measurements of decomposition, the observed fluctuations in redox potential provide a relative measure of the intensity of electron transfers and associated SOM transformation. The deeper and more sustained aerobic conditions and associated redox fluctuations at the water table interface in the acrotelm at MF indicate an environment favourable for mineralization of SOM along with fresh litter from conifers at the site. This is confirmed by the lower quantity of rubbed fibre in the soils at MF compared to MB as a surrogate for degree of decomposition. The persistence of the perennially aerated zone in MF is conducive to SOM mineralization and provides a continual source of DOM available for consumption within the soil or export to streams.

The variation of soil and stream DOM concentrations in MF is associated with the changes in temperature and the hydrodynamics in the acrotelm. The water table drawdown combined with the expansion of aerobic conditions increases organic matter decomposition (Qualls et al., 1991; Palmer et al., 2001) creating a pool of soluble organic matter subject to hydrologic transport. Water moving through the conductive acrotelm perched on the less permeable catotelm flushes this soluble material to outlet streams. Therefore, the combination of SOM processing and hydrodynamic fluctuations in the MF acrotelm lead to both large soil solution DOM concentrations, and high stream DOM concentrations. The oxidation of SOM between storm flushing events is critical to the replenishment of the DOC reservoir in the acrotelm (Worrall et al., 2002). In addition, reduced hydrologic flushing influences the potential C storage or loss in peatlands, not the concentration of available DOC in the soil (Pastor et al., 2003). These findings highlight the need for increased DOC and DON at MF (Fröberg et al., 2003). Measurements made with platinum electrodes in our study appear to be sensitive to soil matrix conditions and indicate possible shifts in terminal electron acceptors (Dettling et al., 2006). The reduced SOM is the primary source of electrons for microbial communities in the soils and redox potential is an indicator of soil biogeochemical behaviour that can be used as a surrogate measure to predict subtle shifts in organic matter transformation (Fiedler and Kalbitz 2003; Mitchell and Branfireun 2005). Though we do not have direct measurements of decomposition, the observed fluctuations in redox potential provide a relative measure of the intensity of electron transfers and associated SOM transformation. The deeper and more sustained aerobic conditions and associated redox fluctuations at the water table interface in the acrotelm at MF indicate an environment favourable for mineralization of SOM along with fresh litter from conifers at the site. This is confirmed by the lower quantity of rubbed fibre in the soils at MF compared to MB as a surrogate for degree of decomposition. The persistence of the perennially aerated zone in MF is conducive to SOM mineralization and provides a continual source of DOM available for consumption within the soil or export to streams.
to understand the coupled system of DOC concentrations and hydrologic fluxes in soils.

The MB site DOM concentrations are linked to changes in temperature, but not to depth to the water table. Although a soluble pool is created, the hydrodynamics limit movement through the soil and transfers to adjacent streams. This highlights the importance of differences in water table changes on present and potential DOM concentrations in CTR peatlands. The lower stream C and N concentrations at MB compared to MF are likely a result of the nutrient cycling capacity in the acrotelm combined with less efficient hydrologic delivery to the outlet stream at MB. The MB site is subject to tight internal recycling of N (Fellman and D’Amore, 2007) and appears to be subject to N retention (Moore et al., 2004). The MB site does not have a readily available supply of labile plant material as biomass is limited to lower stature plants such as sphagnum mosses creating less labile plant material. Overall, nutrients are retained and mineralization rates of organic material are low at MB. The transport capacity at MB is limited due to the thin acrotelm, the low hydraulic conductivity near the bog surface and the low topographic gradient.

The range of DOC and DON concentrations in the MF outlet stream are very narrow compared to the lysimeter concentrations at the site. Moreover, this narrow concentration range is not observed in the bog outlet stream. The substantially different DON, but not DOC, concentrations in MB soil solution compared to streamwater confirms that there may be high biotic retention of N in the bog. This finding is consistent with other studies of freshwater wetlands that have found N can limit net primary production (Aerts et al., 1999). The substantial decrease in DOC and DON concentrations as water moves through the soil matrix to the stream at MB may be caused by dilution with lower DOC groundwater, consumption of DOC through soil heterotrophic activity (Bengtson and Bengtsson, 2007), or adsorption to mineral soils (McDowell and Likens, 1988). Therefore, lower streamwater DOC concentrations at MB may be a product of both dilution and abiotic/biotic removal processes. The dilution by low DOC groundwater and overland flow are not as substantial factors at MB because the bog lysimeters were already sampling diluted water from the water table. DOC and DON are used as a substrate for soil heterotrophic metabolism in MB soils, while DON is more closely associated with soil temperature. In the MF site, both dilution and biotic/abiotic retention mechanisms control DOM export while DOC and DON concentrations in streams are closely associated with both soil temperature and depth to water table. These results provide the basis for modelling C and N export for these prominent peatland types in the CTR of southeast Alaska. These relationships must be tested in other CTR watersheds along with adjacent forested uplands to determine the aggregate influence of the wetland-upland landscape mosaic on stream DOM concentrations and chemical quality in forested watersheds. Relative proportions of bog and forested wetlands within CTR watersheds vary widely from 0% to 40% and 0% to 70% of watershed area, respectively. Therefore the significant difference in hydrologic and biogeochemical behaviour we have documented in this study emphasizes the need to understand process connections at the level of significant functional units in building larger scaled models of carbon and nitrogen flux across the varied terrain of the CTR biome. This understanding is critical in predicting future trends in carbon and nitrogen flux to the coastal ecosystem under changing climatic regimes.

CONCLUSIONS

This study confirms the substantial output of DOC and DON from wetlands to CTR streams and provides background information on the contrasting hydrologic and biogeochemical cycles in two important wetland types in the CTR of southeast Alaska. The presence of the diplotelmic system in CTR peatlands is confirmed by seasonal fluctuations in water tables that influence the production and export of dissolved C and N. The greater hydraulic conductivity within the acrotelm and the depth and duration of aerobic conditions coupled with water table fluctuations creates the potential for larger quantities of DOC and DON production and export to streams from forested wetlands compared to bogs. A comparison between soil solution and stream DOM concentrations indicate that N is retained in bogs and the export of DOC and DON is closely associated with soil temperature. In the MF site, both dilution and biotic/abiotic retention mechanisms control DOM export while DOC and DON concentrations in streams are closely associated with both soil temperature and depth to water table. These results provide the basis for modelling C and N export for these prominent peatland types in the CTR of southeast Alaska. These relationships must be tested in other CTR watersheds along with adjacent forested uplands to determine the aggregate influence of the wetland-upland landscape mosaic on stream DOM concentrations and chemical quality in forested watersheds. Relative proportions of bog and forested wetlands within CTR watersheds vary widely from 0% to 40% and 0% to 70% of watershed area, respectively. Therefore the significant difference in hydrologic and biogeochemical behaviour we have documented in this study emphasizes the need to understand process connections at the level of significant functional units in building larger scaled models of carbon and nitrogen flux across the varied terrain of the CTR biome. This understanding is critical in predicting future trends in carbon and nitrogen flux to the coastal ecosystem under changing climatic regimes.

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