Organic Matter Processing in Arctic Lake Sediments

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Abstract

Kenneth Fortino: Organic Matter Processing in Arctic Lake Sediments
(Under the direction of Stephen C. Whalen)

Lakes are a common landscape feature in the arctic and interact with carbon cycling through the sequestration of organic matter in their sediments. My research assessed the relative importance of landscape- and within-lake-scale factors on organic matter cycling in arctic lake sediments. Sediment organic matter mineralization (measured as sediment oxygen demand) varied between -8 and 40 mmol O_2 m^{-2} d^{-1} and was proportional to water temperature and oxygen concentration. There was greater variation in organic matter mineralization within lakes than between lakes, suggesting that variation in mineralization occurs primarily at the within-lake scale. At the both the landscape- and within-lake-scale, sediment slurries with greater percent organic matter had a higher rate of dissolved inorganic carbon (DIC) production and the effect of oxygen exposure on DIC production was greatest in those sediments with the lowest percent organic matter. Despite the variation in mineralization, organic matter content of the sediments (17 to 69%) varied primarily among lakes (i.e., at the landscape-scale) and was driven by organic matter inputs via benthic primary production. My results suggest that the attenuation of light by dissolved organic carbon in the water indirectly influences organic matter storage in arctic lake sediments. In addition to the direct limitations of organic matter input via benthic photosynthesis, the amount of light attenuation indirectly alters sediment organic matter cycling via changes to the distribution of temperature and oxygen within the lake. Light attenuation was inversely proportional to the depth of thermal stratification, which determines the distribution of temperature and oxygen in stratified lakes. I estimated

that a doubling of the light attenuation would result in a 30% decrease in the area of the sediments exposed to the relatively warm and oxygenated waters of the epilimnion. The interconnection between these factors provides a potential climate change feedback to arctic carbon cycling. Changes in terrestrial organic matter inputs to lakes due to climate change will alter transparency and the depth of the thermocline, changing the distribution of light, temperature and oxygen in the lake and thus the factors limiting the production and decomposition of organic matter in the lake sediments.

To Bailey, without whom none of this would have been possible.

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All of this research was performed in collaboration with the other members of the Geomorphic Trophic Hypothesis (GTH) project and the assistance and shared data has proven invaluable. In particular the lake maximum depths used in Chapters 3 and 4 are a result of GTH sampling efforts.

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1 Benthic organic matter processing in lakes

Globally lakes are estimated to cover > 3% of the area of the Earth's landmasses and small lakes (< 1 km^2) represent approximately 54% of the area covered by freshwater (Downing et al., 2006). Nearly all of the world's lakes are net heterotrophic systems and export CO_2 to the atmosphere (Cole et al., 1994). These emissions account for approximately 15% of the total CO_2 production from freshwater systems annually (Cole et al., 2007). However, lakes are simultaneously a sink for organic carbon via the burial of organic matter in their sediments (Cole et al., 2007; Sobek et al., 2009). Organic matter sequestration in lake sediments is estimated to represent 22% of total annual carbon burial in all freshwater systems (lakes, reservoirs, wetlands, and groundwater) (Cole et al., 2007).

1.1 Organic matter accumulation in lake sediments

The impact of sediment processes on whole-lake or landscape carbon cycling depends on both the rate of organic matter inputs to lake sediments and the fate of the material once deposited. The areal burial rate of organic matter in lake sediments can vary dramatically. Sobek et al. (2009) compile a range of 0.22 to 1140 g C m^{-2} y^{-1} for a global survey of lakes, with the median value much closer to the lower end of the range at 27 g C m^{-2} y^{-1}. Regionally the range of sediment organic matter accumulation rates is much more constrained. Molot and Dillon (1996) found organic carbon accumulation rates ranged 19 to 24 g C m^{-2} y^{-1} for 7 Ontario lakes and Kortelainen et al. (2004) estimate the average Holocene organic carbon accumulation rate to be 0.2 to 8.5 g C m^{-2} y^{-1} in a large survey of Finnish lakes.

Dean and Gorham (1998) found that lake size affected areal organic carbon accumulation rates and estimate the average accumulation rate of large lakes (> 5000 km^2) to be 6.7% the accumulation rate of small lakes. However, Einsele et al. (2001) showed that the

range of organic carbon accumulation rates in a sample of the world's largest lakes was 2 – 23 g C m^{-2} y^{-1}, which is not dissimilar to the ranges for smaller lakes described above.

1.2 Limitations on sediment organic matter mineralization

Once organic matter has been deposited in the sediments, limitations on heterotrophy drive both net CO_2 production and organic matter sequestration in the sediments. The net production of CO_2 from the whole lake derives from the mineralization of allochthonous organic matter and can be closely coupled to (Kortelainen et al., 2006) or independent of (Kling et al., 1991) sediment processes. Despite a variable influence of sediment metabolism on net CO_2 production from lakes, the sequestration of organic carbon is almost exclusively a sediment process. A net accumulation of sediment organic matter (i.e., sequestration) can only result from an imbalance between organic matter inputs and losses (Capone and Kiene, 1988; Canfield, 1994; Burdige, 2007). Sediment organic matter inputs may, in part, be derived from production in the water column or the watershed. However, on a whole--lake scale, organic matter losses depend largely on the factors limiting organic matter mineralization within the sediments.

Lindeman (1942) places bacteria and "ooze" at the center of his conceptualization of lake trophic dynamics, highlighting the connections between sediment heterotrophy and all other aspects of lentic material cycling. Short-term variation in the rate of benthic respiration is typically measured as a change in the concentration of dissolved oxygen or dissolved inorganic carbon (DIC) in the water overlying the sediments of cores removed from the lake or in chambers isolating a portion of the lake bottom (Table 1). Among the studies collected for this review, 18 used O_2 flux and 4 used DIC flux as a measure of sediment respiration. Sediment respiration rate ranged from 2 to 278 mmol m^{-2} d^{-1} of O_2 consumption or DIC production, resulting in a 139 fold difference between the highest and lowest measures (Table 1). This range is skewed, however, by one exceptionally high measurement. With the exception of Lake Mohegan, there are no measurements from the 82 lakes in Table 1 greater than 151 mmol m^{-2} d^{-1}. Thus, the majority of systems

reviewed had benthic respiration rates that ranged between 2 and 151 mmol m^{-2} d^{-1} of O_2 consumption or DIC production, or a 75.5 fold range. This is greater than, but not dramatically different from the 21-fold range of variation reported in a review of lake sediment respiration rates by Pace and Prairie (2005).

1.3 Factors limiting sediment organic matter mineralization

1.3.1 Temperature

Temperature, the availability of oxygen, and the supply of carbon have all been shown to alter the rate of sediment respiration (Table 2). Like all metabolic processes, respiration rate would be expected to increase with temperature (Gillooly et al., 2001). Baulch et al. (2005) experimentally warmed the littoral zone of a boreal lake 4.5° C and found that dark respiration by the epilithon increased 29 – 103% in the warmed treatments. Hargrave (1969) suggests that temperature is the principal factor regulating benthic respiration rates in lakes and found a very strong relationship between the log oxygen uptake rate by sediments in incubated cores and log temperature both across systems (r = 0.85) and within a system (r = 0.86). Subsequent research has shown that the relationship between temperature and benthic respiration rate changes with temperature. Granéli (1978) found that the Q_{10} (the change in respiration rate per change in 10° C) for oxygenated sediments from 7 Swedish lakes ranged from 2 to 3 for temperatures between 5 and 10° C but from 1.3 to 1.6 for temperatures between 15 and 20° C. This pattern is also found in anaerobic sediments. The Q_{10} for methane production from deep lake sediments declined with increasing temperature from 3.86 in the 5 to 15° C range to 1.8 in the 20 to 30° C range (Nozhevnikova et al., 1997).

The importance of temperature in controlling benthic respiration rates also appears to be related to the range of temperature variation within a lake. Baulch et al. (2005) measured the relationship between epilithic respiration and water temperature in a single lake during a 23-year period. The relationship was significant but showed a distinct wedge shape with much more variation in respiration rate at high temperatures, suggesting that

temperature may act as a limiting factor at low temperatures but other factors become more relevant as temperature moderates. Supporting this idea, Pace and Prairie (2005) found that although temperature has a significant effect on respiration rate, the range of the temperature-standardized and un-standardized respiration rates was similar. The authors suggested that some of the effect of temperature on benthic respiration may be due to covariance with other factors.

1.3.2 Oxygen availability

In marine systems, the time organic matter spends exposed to oxygen is thought to be a significant factor affecting the burial efficiency (Hartnett et al., 1998). Recently, oxygen exposure time has also been shown to be significantly correlated with organic matter burial efficiency in lake sediments (Sobek et al., 2009). Oxygen is used both as the final electron acceptor for aerobic respiration and in the enzymatic hydrolysis of organic matter with oxidase and peroxidaese enzymes (Burdige, 2007). As a result, there is an interaction between the impact of oxygen availability on mineralization rate and the recalcitrance of the organic matter (Kristensen, 2000). Canfield (1994) found that in marine sediments, oxygen reduced organic matter preservation only in sediments with low rates of labile organic matter deposition. Labile organic matter from eutrophic lake sediments decomposed at similar rates in aerobic and anaerobic incubations but the absence of oxygen limited the mineralization of refractory organic matter from the same system (Lehmann et al., 2002).

1.3.3 Carbon supply

Correlations between water column primary production (i.e., a source of organic C) and benthic respiration have been well established (Wetzel, 2001). Benthic respiration rates are strongly correlated with epilimnetic total phosphorus (a surrogate for volumetric pelagic primary productivity) (Simčič and Brancelji, 2002; Pace and Prairie, 2005) and an estimated 40 and 70% of sedimented within-lake production is mineralized in eutrophic and oligotrophic lakes respectively (Pace and Prairie, 2005). Furthermore, fertilization exper-

iments have shown that increases in pelagic primary productivity will increase sediment respiration rate (Jones and Simon, 1980; Sugai and Kipphut, 1992; O'Brien et al., 2005).

The importance of allochthonous organic matter inputs for benthic respiration rates is less clearly established. A number of studies (Prairie et al., 2002; Houser et al., 2003; Jonsson et al., 2003; Sobek et al., 2005) have found significant correlations between whole--lake respiration rates and the concentration of dissolved organic carbon (DOC) in the water column (Table 2), yet direct evidence that the input of allochthonous organic matter affects sediment respiration is more equivocal. Jonsson et al. (2003) found that sediment respiration was positively correlated with DOC concentration ($r = 0.75$) in 51 lakes in Sweden. Huttunen et al. (2002) showed that the DIC flux from 2 very shallow boreal ponds was strongly correlated with the input organic carbon from their watersheds. Given the depth of these ponds, this response likely represents the effect of sediment respiration. Sweerts et al. (1986) showed that the presence of a few millimeters of flocculent organic material was sufficient to double the respiration rate of sandy littoral sediments, but this flocculent material was only partially allochthonous. Finally, Lasenby (1975) found that areal hypolimnetic oxygen demand was strongly correlated with Secchi depth ($r = 0.85$) but uncorrelated with seston mass. Since the Secchi depth incorporates dissolved as well as particulate organic matter, this correlation with transparency but not seston suggests a decoupling of epilimnetic production and the hypolimnetic decomposition and may point to the importance of allochthonous material (Lasenby, 1975).

Other studies have found little to no relationship between carbon supply and benthic respiration rates. Granéli (1978) saw no effect of the addition of freshly sedimented material to cores or the redistribution of sediment within cores on oxygen uptake from the sediments of 7 Swedish lakes. Further, Hargrave (1969) found no relationship between the oxygen uptake rate in cores and the total organic matter, protein, carbohydrate, or calorific content of the sediments. Thus, despite strong evidence that pelagic primary productivity (i.e., autochthonous organic matter supply) is coupled to benthic respiration, the effect of allochthonous substrate supply on benthic metabolism is less clear.

1.4 Future research

The importance of lake sediment organic matter processing in the global carbon cycle is clear, yet our understanding of how lakes will respond to and influence future alterations to the carbon cycle remains incomplete. Global assessments of sediment metabolic processes show considerable variation among lakes over large geographic scales (Sobek et al., 2009) but regional studies show that variation within-lakes may exceed differences among lakes on a landscape (Hobbie et al., 1980; den Heyer and Kalff, 1998), suggesting that the scale of variation in organic matter processing may be primarily at the within-lake scale. The factors affecting sediment respiration (i.e, temperature, oxygen availability, and carbon supply) potentially vary at both within-lake and landscape scales so it is necessary to partition the impact of these factors between scales to understand how lake carbon cycling feeds back into the regional and global carbon cycle.

The arctic is expected to respond more dramatically to climate change than lower latitudes. Observed changes include a decrease in snow cover, the warmest temperatures in 400 years (Overpeck et al., 1997; Chapin et al., 2005) and the expansion of shrub vegetation cover (Sturm et al., 2005). Additionally, arctic lakes have been shown to be sources of carbon to the atmospheric via the outgassing of CO_2 and CH_4 derived from soil and lake respiration (Kling et al., 1991). This sensitivity of arctic systems to climate variability combined with the role of lakes in global carbon cycling (Cole et al., 2007) indicates the need to understand the factors affecting arctic lake respiratory processes.

1.5 Research Objectives

This dissertation was undertaken to address the central question: *What is the relative importance of landscape and patch-scale environmental factors in limiting the mineralization of organic matter in arctic lake sediments?*

Specific objectives were to:

1. Quantify the effect of temperature and oxygen availability on summertime benthic respiration rate at the within-lake scale;

2. Quantify differences in the lability of the sediment organic matter among lakes at a landscape-scale;

3. Quantify the landscape-scale variation in the distribution of temperature, oxygen, and organic matter in arctic lakes during the summer and assess their influences on sediment metabolism.

I address these objectives with a combination of laboratory experiments and field surveys conducted in the vicinity of the Toolik Lake Biological Field Station in northern Alaska (68°38' N, 149°38' W). The region surrounding the field station is typical of low latitude arctic regions. Plant cover is a mixture of tussock and shrub tundra (Walker, 2000) and the land surface is underlain by continuous permafrost (Ping et al., 1998). The region has an annual mean temperature between -10° and -8° C and annual precipitation of 140 to 270 mm of which 40% is snow (Ping et al., 1998). During the summer, the region experiences 24–h daylight and an average summer temperature of 11° C (Oechel et al., 2000). The region was discontinuously glaciated during past glacial advances which resulted in a diversity of land surface and lake ages (Hamilton, 2002). The numerous lakes in the region are typically oligotrophic with spatially variable nutrient limitation and tend to be relatively small and shallow (Levine and Whalen, 2001). Fish diversity is low within the lakes and is primarily controlled by landscape-level factors (Hershey et al., 2006).

Table 1: Range of values for sediment respiration rates (mmol m^{-2} d^{-1}) reported from the literature. Respiration was measured as either the flux of dissolved O_2 into or dissolved inorganic carbon (DIC) from the sediments.

Reference	Range §	Method	Location
Adams et al. (1982)	9.4	O_2 flux; benthic chambers	Lake Erie
Baulch et al. (2005)	2.1 – 8.3	DIC flux; benthic chambers	Lake 239, Ontario, Canada
Blanton and Winkhofer (1972) †	8.8 – 10.9	O_2 flux; benthic chambers	Lake Erie
Burns et al. (1996)	3.75 – 11.9	O_2 flux; benthic chambers	3 lakes, New Zealand
Cornwell and Kipphut (1992)	7.6	O_2 flux; benthic chambers	Toolik Lake, Alaska, USA
Edberg and Hofsten (1973) ‡	87.5	O_2 flux; not reported	Lake Norrviken, Sweden
Fillos (1977) ‡	53.1 – 278.1	O_2 flux; not reported	Lake Mohegan, New York, USA
Gardiner et al. (1984) ‡	104.4	O_2 flux; not reported	Green Bay, Lake Michigan, USA
Gelda et al. (1995)	25.6 – 90.3	O_2 flux; core incubations	Onondaga Lake, New York, USA
Granéli (1978)	11.5 – 28.8	O_2 flux; core incubations	5 lakes in Sweden
Hargrave (1969)	2 – 43	O_2 flux; core incubations	Marion Lake, British Columbia, Canada
Hayes and MacAulay (1959)	5.2 – 33.2	O_2 flux; core incubations	12 lakes in Canada
den Heyer and Kalff (1998)	17 – 60	DIC flux; core incubations	9 lakes in Quebec, Canada
Jonsson et al. (2003)	5.3 – 57.8	DIC flux; benthic chambers	16 lakes in Sweden
Lasenby (1975)	11.2	O_2 flux; core incubations	Dicky Lake, Ontario, Canada
Liikanen et al. (2002)	4.8 – 28	DIC and O_2 flux; core incubation	Lake Kevätön, Finland
Linsey and Lasenby (1985)	26 – 32	O_2 flux; core incubations	Sharpe Bay of Jack's Lake
Lucas and Thomas (1972) †	12.5 – 21.9	O_2 flux; benthic chambers	Lake Erie
Ramlal et al. (1994)	32 – 47	O_2 flux; benthic chambers	Lake 18, high arctic, Canada
Sehgal and Welch (1991)	6.3 – 12.8	O_2 flux; core incubations	Scriber Lake and North Lagoon, Washington, USA
Sweerts et al. (1986)	10.8 – 42.6	O_2 flux; benthic chambers	Lake 302S, Ontario, Canada
Range	2 – 278		

§ If no range is reported, the value is the mean.
† cited in Adams et al. (1982)
‡ cited in Gelda et al. (1995)

Table 2: Summary of studies reporting a correlation between sediment respiration rate and environmental variables. A dash (–) indicates that the results are from the above reference.

Reference	Correlated Factor (r)	Method	Location
Baulch et al. (2005)	Temperature (0.53)	O_2 uptake in benthic chambers	Lake 239, Ontario, Canada
Granéli (1978)	log Temperature (positive, r not reported)	O_2 uptake in cores	6 Swedish lakes
–	Oxygen concentration (positive, r not reported)	O_2 uptake in cores	
Hargrave (1969)	ln Temperature (0.86)	O_2 uptake in cores	Marion Lake, British Colombia, Canada
–	ln O_2 concentration (0.79)	O_2 uptake in cores	
den Heyer and Kalff (1998)	Sample Depth (-0.42)	Core DIC production	9 lakes in Quebec
–	Temperature (0.51)	Core DIC production	
Jonsson et al. (2003)	DOC concentration (0.75)	O_2 uptake in benthic chambers	51 lakes in Sweden
Muri and Simčič (2004)	Organic Carbon (0.49 – 0.63)	Slurry O_2 uptake	4 lakes in Slovenia
Pace and Prairie (2005)	log Total P (0.83)	O_2 uptake in cores	a review of numerous lakes
Simčič and Brancelji (2002)	Total P (0.87)	Slurry O_2 uptake	15 lakes in Slovenia

2 Factors Affecting Sediment Oxygen Demand in Arctic Lakes

2.1 Introduction

Lake sediments have recently been recognized as an important component of the global carbon cycle, and may annually store as much as 30 to 60% of the organic carbon stored in marine sediments (Cole et al., 2007). The fate of the organic matter deposited in lake sediments depends on the factors controlling its mineralization, which in turn depends on the the factors limiting the metabolic activity of sediment microorganisms (Canfield, 1994; Burdige, 2007).

The metabolic activity of sediment microorganisms is frequently quantified as the dark oxygen flux across the sediment-water interface, i.e., sediment oxygen demand (SOD) (Pace and Prairie, 2005). Oxygen is consumed by aerobic respiration or through the oxidation of the reduced products of anaerobic metabolism and, therefore, is tightly coupled to microbial organic matter mineralization (Sweerts et al., 1991; Kristensen, 2000; Torgersen and Branco, 2007). The magnitude of mineralization is reflected in SOD because, excluding the flux of non-oxidized metabolites from the sediments (e.g., NH_4, CH_4), the most significant sediment reactions involving organic matter mineralization directly or indirectly create demand for oxygen (Sweerts et al., 1991; Soetaert et al., 1996; Burdige, 2006).

Although most lake sediments efficiently sequester organic matter (Dean and Gorham, 1998; Einsele et al., 2001), variation in environmental conditions can alter SOD and thus organic matter mineralization (Pace and Prairie, 2005). Temperature has direct effects on both the metabolic rate of sediment microorganisms (Thamdrup et al., 1998) and the diffusion of oxygen within the sediments (Boudreau and Jørgensen, 2001) and therefore can have strong effects on SOD (Hargrave, 1969; Granéli, 1978).

SOD may also be affected by the oxygen concentration of the water overlying the sediments, since oxygen must diffuse across the sediment-water interface (Boudreau and Jørgensen, 2001). However, since the oxygen concentration gradient across the sediment--water interface is in part created through oxygen consumption within the sediments, observed changes in SOD due to fluctuations in water column oxygen concentrations (Hargrave, 1969; Granéli, 1978; Archer and Devol, 1992) should reflect real limitations to organic matter mineralization.

Finally, SOD can be affected by the quality of sediment organic matter (Sweerts et al., 1986; Kristensen, 2000), since regardless of the metabolic pathway, it is the mineralization of organic matter that directly or indirectly creates oxygen demand in the sediments (Torgersen and Branco, 2007). Significant increases in SOD have been reported as a result of elevated organic matter inputs (Sugai and Kipphut, 1992; Dedieu et al., 2007), and across systems SOD is correlated with whole system productivity (Pace and Prairie, 2005), suggesting widespread substrate limitation of SOD. Unlike the effects of temperature and oxygen concentration however, the impact of organic matter quality on SOD is not always observed (Hargrave, 1969; Granéli, 1978).

Variations in SOD have been observed at both a patch-scale (i.e., within a lake across space or time) and landscape-scale (i.e., among different lakes on the landscape) (Chapter 1). Within lakes, seasonal variation in SOD can range between 32 – 47 mmol O_2 m^{-2} d^{-1} for an oligotrophic, high-arctic lake (Ramlal et al., 1994) to 26 – 90 mmol O_2 m^{-2} d^{-1} for a hypereutrophic lake (Gelda et al., 1995). Presumably this seasonal variation is due to corresponding changes in environmental conditions. Hargrave (1969) found that seasonal changes in temperature alone could produce a range of SOD from approximately 2 – 43 mmol O_2 m^{-2} d^{-1} in Marion Lake, B. C.

Variation in sediment organic matter mineralization among lakes in a region is often similar to what is observed within, lakes suggesting only a minor influence of landscape--scale variables on sediment mineralization rates (den Heyer and Kalff, 1998). Pace and Prairie (2005) report a range of SOD values between 1.6 and 33 mmol O_2 m^{-2} d^{-1} in a review

of lakes that span productivity and size gradients. They attribute the observed variation primarily to differences in lake productivity (after standardizing SOD measurements to 10° C).

Fewer measurements of SOD have been performed on arctic lakes, yet overall the patterns appear similar to other regions. SOD measurements collected by Ramlal et al. (1994) for a high arctic lake are similar in overall magnitude (32 – 47 mmol O_2 m^{-2} d^{-1}) to the lakes reviewed by Pace and Prairie (2005). Hobbie et al. (1980) found dramatic variation in the sediment dissolved inorganic carbon (DIC) production of shallow tundra ponds. In these systems the summer DIC flux from the sediments varied from approximately 0 to 42 mmol DIC m^{-2} d^{-1}, with much greater variation within ponds than among ponds. The only published measurements of sediment respiration that I am aware of from the Toolik Lake region of the Alaskan arctic come from Toolik Lake and deviate from the above patterns. Cornwell and Kipphut (1992) found very little variation in net SOD (using clear benthic chambers) in Toolik lake (7.1 – 8.1 mmol O_2 m^{-2} d^{-1}) across a depth range from 3 to 7 m. Although the restricted range and magnitude of the measurements might have been due to the photosynthetic production of O_2, the lack of variation across depths which should have had much different light environments suggests that photosynthetic oxygen production does not completely explain the lack of variation.

In this chapter, I evaluate the variation in SOD due to differences in temperature and oxygen availability both within and among 3 shallow arctic lakes with the goal of understanding the factors affecting the range of SOD in shallow arctic lakes and the scale over which those factors vary. Specifically I test the hypothesis that variation in SOD in shallow arctic lakes is primarily the result of fluctuations in temperature and oxygen availability at a patch-scale.

2.2 Study Site

The three lakes in this study are located in the vicinity of the Toolik Lake Biological Field Station (68°38'N, 149°38'W) (Fig. 1). The study area is characteristic of the Arctic

Foothills region of Alaska and is underlain by continuous permafrost with predominantly tundra vegetation (Ping et al., 1998). The region has an annual mean temperature between -10° and -8° C and annual precipitation of 140 to 270 mm, of which 40% is snow (Ping et al., 1998). During the summer, the region experiences 24–h daylight and average summer temperatures of 11° C (Oechel et al., 2000). The three lakes used in the study are small, oligotrophic (Table 3) and lack permanent inlets. All three lakes are ice–covered for approximately 9 months.

2.3 Materials and Methods

2.3.1 Sample Collection and Incubation

Sediment cores were collected using a a K-B style gravity corer. The cores were standardized for incubation by extruding the upper 15 cm of sediment (and overlying water) from each core into a 25 cm long by 4.8 cm i.d. plastic incubation core. The incubation cores were sealed with acrylic or polycarbonate tops and bottoms. The core tops were beveled toward a center opening to exclude all air from the core when sealed. A magnetic stir bar was suspended approximately 1 cm above the sediment-water interface in each core. During incubation, the cores were arranged around a central array of magnets turning at 1 rpm, which slowly turned the stir bars within the cores and prevented stratification of the overlying water. The cores were incubated in a 750 L temperature controlled (± 1° C.) water bath. Sampling was performed via two septum–sealed ports fitted to the core top. One port was oriented vertically and allowed for the insertion of a cannula into the overlying water of the core. The other port was arranged perpendicular to the first and permitted the simultaneous replacement of water removed during sampling. The replacement water was collected from the lake at the same time and depth as the cores with a beta–style 2.2–L Van Dorn sampler. Between sampling events, the replacement water was stored in a 4–L plastic bottle in the same incubator as the cores.

2.3.2 Oxygen Measurements

The cores were allowed to acclimate approximately 1 h prior to the initial sampling. During sampling, 4 ml of overlying water was slowly removed from the core through the vertical port using a stainless steel cannula attached to a plastic syringe. The syringe was purged with 1 ml of the sample and the remaining volume was used to determine the oxygen concentration by Winkler titration (Carpenter, 1965) modified for the small volume. Plastic stops were designed to fit onto the syringe plunger to ensure that a repeatable volume was retained in the the syringe following each purge. The exact volume of sample retained in each syringe following purging was determined gravimetrically. Titrations were conducted immediately following sampling to minimize storage artifacts.

2.3.3 Experimental Details

Temperature Dependence Experiment I assessed the impact of temperature on the magnitude of SOD by incubating cores collected at 3 m in each lake at 4 different temperatures representing a realistic range of temperatures experienced by the sediments during the arctic summer. The cores were incubated in the dark. The oxygen concentration of the overlying water in the cores was sampled three times over approximately 24 h.

Oxygen Availability Experiment I measured the effect of oxygen availability on SOD by calculating the change in SOD that resulted from the reduction in the oxygen concentration of the overlying water during an approximately 48 h dark incubation. For the sediments from Lakes E–4 and S–3, this was accomplished by extending the incubation time of the cores used in the 12^{o} C treatment of the temperature dependence experiment. The response of the sediments from Lake GTH 91 were assessed in a separate experiment at 9^{o} C. Samples were taken at 5 time points over the 48 h period.

Light Availability Experiment I evaluated the impact of light availability on net sediment oxygen production by measuring the effect of different irradiance levels on oxygen flux from the sediments of all three lakes. In the experiment using sediments from GTH 91

the temperature was held at 10° C. In the experiments using sediments from E–4 and S–3, the temperature was approximately matched to the water temperature at the depth from which the cores were collected (15° C). Irradiance was provided by a 1000 W full spectrum grow lamp suspended above the incubation chambers. Different light levels were produced in each core by encasing the upper half of the core with sleeves made from neutral density screens. In the experiments conducted in E–4 and S–3, the lower portion of the core was covered in aluminum foil to ensure that only the sediment-water interface was exposed to the light. The irradiance in each core was measured with a Biospherical Instruments Quantum Light meter with a 4π sensor. The light probe was inserted through a notch in a specially constructed top so that irradiance could be measured with the top in place. The irradiance in each core was measured in the exact location that the core occupied during the incubation to account for spatial variation in the light environment of the 750 L incubation chamber. The concentration of oxygen was sampled at three time intervals over a 24–h period.

2.3.4 Lake Bathymetry, Thermal Stratification and Light

The bathymetry of each lake was mapped by combining a lake perimeter measurement determined with a Tremble Geo Explorer GPS with sonar transects collected with a Garmin GPSMAP 180 sonar. The lake bottom profile was extrapolated from the sonar measurements using a triangulated irregular network to convert the observed depths into 1 m contour lines in ARC–GIS (ESRI, 2006). Surface area of each depth interval was calculated using the 1 m contour intervals.

Temperature and dissolved oxygen profiles were collected in each of the three lakes throughout the ice–free periods of 2005–2008 using a YSI Model 85 meter. Profiles were taken at the deepest point in the lake in 0.5 m depth intervals. The oxygen meter was calibrated prior to sampling each lake. For each sampling event the thermocline depth was calculated as the depth with the greatest change in temperature from the preceding depth. If no two successive depths within the lake had a temperature change greater than

1° C, then the lake was considered unstratified. The boundaries of the epilimnion and hypolimnion were defined as the depth above or below which the slope of the change in temperature with depth was $< 1^{o}$ C m^{-1} (Wetzel, 2001). The area-weighted temperature of the epilimnion and hypolimnion was calculated as the weighted average of the temperature of each depth using the proportion of epilimnetic or hypolimnetic sediment area as the weights. Since sediment area was only avialable at 1 m resolution, the temperature data were collapsed to 1 m resolution as well. The oxygen deficit below the thermocline was estimated by dividing the area-weighted oxygen concentration of the hypolimion into the area-weighted oxygen concentration of the epilimnion and converting to a percent.

Measurements of the light environments of the lakes were collected variously during the ice–free period of 2005 through 2008. Measurements of photosynthetic photon flux density (PPFD) were taken in 0.5 m increments at the deepest point in the lake with a LiCor 192-SA 2π underwater quantum sensor and a LiCor 250 light meter. Profiles were stopped when the light level reached approximately 1% of the irradiance immediately below the air–water interface. The light attenuation coefficient (K_d) was calculated as the linear slope of the ln–transformed PPFD data (μE m^{-2} s^{-1}) versus depth.

To estimate the percent sediment surface area below the compensation point (i.e., the irradiance level at which photosynthetic oxygen production equals respiratory oxygen consumption) in each lake for the summers of 2005–2007, I used the mean K_d calculated in 2008 and the hourly mean irradiance data collected from just below the water's surface at the Toolik Lake meteorological station (Shaver, 2005, 2006, 2007) to estimate the amount of light in each lake at 1 m depth increments. The light estimate combined with the sediment area estimated from the bathymetric analysis was used to calculate the proportion of observations that were below the compensation point for each year. The mean 2008 K_d value was used because there was greater seasonal coverage in 2008 K_d data relative to the other years of the study.

The total hypolimnetic oxygen consumption attributable to SOD was estimated in lake GTH 91 as the ratio of total hypolimnetic sediment oxygen consumption to the rate of

hypolimnetic oxygen loss. I estimated sediment oxygen consumption (mmol O_2 d^{-1}) as the product of the sediment area (m^2) in a 1 m slice below 6, 7, and 8 m and SOD at 4.5° C (mmol O_2 m^{-2} d^{-1}). SOD at 4.5° C was estimated using the linear model from the temperature dependence experiment. The rate of hypolimnetic oxygen loss (mmol O_2 d^{-1}) at the same depths was calculated as the change in oxygen concentration between Julian days 207 and 213 in 2006 (mmol O_2 m^{-3} d^{-1}) multiplied by the volume of the lake in a 1 m slice below each depth (m^3).

2.3.5 Statistics and Calculations

In the temperature dependence and light availability experiment, SOD was calculated as the sum of the change in oxygen concentration of the overlying water between the successive time intervals. In the oxygen availability experiment, SOD was calculated for each time interval as the change in oxygen concentration of the overlying water from the preceding time point. All oxygen concentration measurements were corrected for oxygen added via the replacement water. Fluxes were normalized to the surface area of the core and expressed as an hourly rate.

Two points were removed from the analysis of the oxygen availability experiment due to obviously unrealistic values. The effect of temperature and the source of the sediments (i.e., lake) on SOD was analyzed with ANCOVA. The effect of the oxygen concentration of the overlying water and sediment source on SOD was evaluated using a repeated measures ANCOVA. All analysis were performed using JMP software (Ver. 4.0.4, SAS Institute Inc., Cary, NC, 1989–2007) or R (R Development Core Team, 2009).

2.4 Results

2.4.1 Temperature Dependence Experiment

The temperature dependence experiment tested SOD across the range of temperatures typically experienced by the lake sediments during the summer (Table 4). SOD ranged between -3.3 and 39.0 with a median of 15.1 mmol O_2 m^{-2} d^{-1} across all sediments and

temperatures. Temperature and sediment source (i.e., lake) explained 42% of the variation in SOD and there was significant and positive relationship between SOD and temperature across all three lake sediments (slope ($\pm$ $CI_{95\%}$) = 0.61 ($\pm$ 0.24) mmol O_2 m^{-2} d^{-1} per degree C) and no significant interaction between temperature and sediment source (Fig. 2; Table 5). After accounting for the variation due to temperature there remained a significant effect of sediment source on SOD. Median SOD was 12.6, 16.88, and 15.7 mmol O_2 m^{-2} d^{-1} in the sediments from lakes E–4, S–3 and GTH 91, respectively.

2.4.2 Oxygen Availability Experiment

In the oxygen availability experiment, I tested the effect of the oxygen concentration of the overlying water on SOD. The oxygen concentration of the overlying water ranged between 0.1 and 0.3 mmol O_2 L^{-1} (Table 6) and SOD ranged between -7.7 and 39.8 with a median of 11.4 mmol O_2 m^{-2} d^{-1} (Fig. 3). The highest SOD measurements were similar in magnitude to those from the temperature experiment but the reduction in oxygen in the overlying water reduced SOD at a rate ($\pm$ $CI_{95\%}$) of 49 $\pm$ 21 mmol O_2 m^{-2} d^{-1} per mmol O_2 L^{-1} to levels below those seen in the temperature dependence experiment (Fig. 3).

As with the temperature dependence experiment, there were significant differences in SOD among the sediments collected from the different lakes after accounting for the variation due to oxygen concentration and no interaction between oxygen concentration and sediment source (Table 7). The median SOD was 10.1, 14.8, and 9.8 mmol O_2 m^{-2} d^{-1} in the sediments from lake E–4, S–3 and GTH 91, respectively.

2.4.3 Light Availability Experiment

In all three lakes SOD was affected by irradiance and the sediments became net–autotrophic at irradiance levels greater than approximately 50 μE m^{-2} s^{-1} (Fig. 4). At light levels less than 50 μE m^{-2} s^{-1} the sediments of GTH 91, E–4 and S–3 show a net flux of O_2 into the sediments in 96, 90, and 100% of observations respectively. Although there are fewer observations at light levels greater than 50 μE m^{-2} d^{-1}, almost all show net efflux of O_2

(Fig. 4).

2.4.4 Stratification and Light

Although a complete time course for the development of thermal stratification does not exist for the duration of the study, there are clear differences in the thermal stratification regimes of the lakes. Lake E–4 was stratified in all of the profiles collected in 2006 and had a mean thermocline depth of 3.3 m (Table 8). In 2008, the mean thermocline depth was the same as 2006 but the lake mixed during the end of July. Lake S–3 was sampled in 2005, 2006 and 2008. In 2005 the lake was unstratified in early June and early August but stratified for the remainder of the summer. In 2008, S–3 was stratified by the first sampling date in late June but had mixed by the end of July. The mean thermocline depth was 3.8 m in 2005 and 3.2 m in 2008. There was only one profile collected in 2006 and the lake was stratified with a thermocline depth of 2 m. The deeper lake (GTH 91) was always stratified when sampled, with a mean thermocline depth of 4.2, 3.6, and 3.9 m for 2005, 2006 and 2008 respectively (Table 9).

Thermal stratification substantially affected the distribution of temperature and oxygen in the lakes (Tables 8 and 9). The polymixis of the shallow lakes resulted in more temperature variation, particularly in the deeper portions of the lake. The loss of oxygen from the deeper water of the lake was also related to thermal stratification and mixing. The bottom waters of the deep lake (GTH 91) had a median hypolimnetic O_2 deficit of 57% of epilimnetic O_2 concentration, which persisted throughout the sampling period (Table 9). The shallow lakes show less severe and more variable oxygen depletion in the bottom water relative to the deep lake (Table 8).

Comparison of K_d between lakes indicates that the two shallow lakes (E–4 and S–3) show a general increase in clarity as the summer progressed in years for which extensive data were collected (Table 8). In 2008, I sampled the lakes within 24 h of each other to assess between-lake differences in K_d independent of this seasonal pattern. In all of the 2008 observations, S-3 has lower light attenuation (i.e., lower K_d) than E-4. The deep

lake (GTH 91) showed a trend of increasing clarity during the summer in 2006 but less variation in clarity and no clear seasonal pattern in 2008 (Table 9).

The percentage of the lake sediments estimated to be below 50 μE m^{-2} s^{-1} (i.e., the compensation point) ranged between 42 and 60% in the shallower lakes (Lakes S–3 and E–4) and 70 to 79% in the deep lake (Table 10). In all of the lakes, the sediment area below the compensation point increased between 2005 and 2007. Although Lake S–3 consistently has the smallest percentage of its sediment area below 50μE m^{-2} s^{-1}, this percentage is only slightly less than what is estimated for Lake E-4. Sediment oxygen demand was estimated to contributed 67% of the total hypolimnetic oxygen consumption in lake GTH 91 in late summer 2006 (Table 11).

2.5 Discussion

SOD ranged between -7.7 and 39.8 mmol O_2 m^{-2} d^{-1} with a median of 13.8 mmol O_2 m^{-2} d^{-1} across all the lakes. This range of measurements is similar to the SOD range of 1.6 to 33 mmol O_2 m^{-2} d^{-1} reported by Pace and Prairie (2005) in a review of studies using similar methods. As far as I am aware there are no other published measurements of gross SOD from the Alaskan low arctic, yet the range of SOD observed in my study is similar to the range of DIC flux measurements (-0.4 to 42 mmol DIC m^{-2} d^{-1}, (Hobbie et al., 1980)), and SOD (32 to 47 mmol O_2 m^{-2} d^{-1}, (Ramlal et al., 1994)) collected in other shallow arctic systems at higher latitudes.

Cornwell and Kipphut (1992) measured a net SOD range of 7.1 to 8.1 mmol O_2 m^{-2} d^{-1} in Toolik Lake (which is on the same landscape as the lakes in my study). These measurements are lower and much less variable than I observed, but represent *in situ* net SOD collected over 2 to 7 day periods and do not likely reflect the range of environmental conditions found in my study or in the environment. My measurements of dark SOD across a greater range of environmental conditions and lakes is likely more representative of the actual variation in sediment SOD (and mineralization) in this region of the arctic.

The variation in SOD among sediments from within a lake in both the temperature de-

pendence and oxygen availability experiments greatly exceeded differences in SOD among the three lakes, suggesting that patch-scale variation in environmental conditions has a greater influence on SOD than landscape-scale differences among lakes (Hobbie et al., 1980; den Heyer and Kalff, 1998). Across all of the lakes, variation in SOD was significantly related to variation in temperature and oxygen concentration. SOD increased with temperature regardless of the sediment source and there were no significant differences in the relationship among the different lakes. Pace and Prairie (2005) calculate a mean slope of 0.65 $\pm$ 1 for the log–log SOD to temperature relationship in review of lake SOD experiments. Similar treatment of the present data gives a mean log–log SOD to temperature slope of 0.26 $\pm$ 0.08 across all three lakes, which although lower than what was observed by Pace and Prairie (2005), is within the large standard deviation of their collected measurements. This observation indicates that the sediments from the lakes in this study are at the lower end of temperature sensitivity relative to the lakes they report.

In addition to reduced sensitivity to temperature, I found that SOD increased linearly with temperature across the temperature range tested, whereas previous studies have observed that the response of SOD to temperature is greatest at lower temperatures ($<$ 10^{o} C) (Hargrave, 1969; Granéli, 1978; Pace and Prairie, 2005). This difference in response appears to be the result of greater SOD at low temperatures and lower SOD at high temperatures in my study. I measured a median SOD at 2^{o} C across lakes of 12.1 mmol O_2 m^{-2} d^{-1}, whereas the lakes surveyed in Hargrave (1969) do not achieve SOD rates this high until temperatures of approximately 9^{o} C. A similar comparison at higher temperatures shows that the median SOD at 17^{o} C in the sediments from the lakes in my study is 19.7 mmol O_2 m^{-2} d^{-1} while Hargrave (1969) predicts an SOD of 39 mmol O_2 m^{-2} d^{-1} at 17^{o} C. This reduced sensitivity of the SOD response to temperature may be due to selection for sediment microbial communities that perform more efficiently at lower temperatures (Madigan et al., 2000) in the persistently cooler temperatures found in arctic lakes.

Reduced oxygen concentrations in the water overlying the cores limited SOD in the sediments of all three lakes. Theoretical models (Bouldin, 1968) and measurements (Ras-

mussen and Jørgensen, 1992) show that if the rate of oxygen consumption within the sediments is constant with respect to oxygen concentration, then the flux of oxygen across the sediment-water interface will vary as a function of the square root of the oxygen concentration. Nonetheless, SOD appears to vary linearly with the oxygen concentration in the overlying water across the range of oxygen concentrations used in the oxygen availability experiment. It is likely that my experiment did not evaluate the relationship at oxygen concentrations low enough to include the point where the linear and square root models clearly diverge.

Hargrave (1969) found an approximate reduction of 42 mmol O_2 m^{-2} d^{-1} per mmol O_2 L^{-1} in Marion Lake, B.C. and Park and Jaffe (1999) calculate an approximate reduction of 63 mmol O_2 m^{-2} d^{-1} per mmol O_2 L^{-1} using a numerical model of sediment oxygen dynamics. Both of these estimates agree well with the estimated decline of 50 mmol O_2 m^{-2} d^{-1} per mmol O_2 L^{-1} across the three lakes observed in my study.

The correlation between SOD and the oxygen concentration in the overlying water is likely due to diffusion limitation resulting from a smaller oxygen concentration gradient (Granéli, 1978; Boudreau and Jørgensen, 2001). However, the reduction in SOD will still have implications for sediment organic matter mineralization. Assuming that sediment porosity remains constant, an increase in the oxygen concentration of the overlying water will increase the diffusion rate of oxygen into the sediments (Hartnett et al., 1998). If the sediment oxygen consumption pathways are saturated then the rate of oxygen loss with depth into the sediments will remain constant but the oxygen penetration depth will increase. Alternatively, if the oxygen consumption pathways are limited by oxygen availability then the loss of oxygen with depth will increase with no change in the penetration depth. The cases need not be mutually exclusive but both cases provide additional oxygen for sediment organic matter mineralization (Epping and Jörgensen, 1996).

The diffusion of oxygen into the sediments can also be limited by the thickness of the diffusive boundary layer (i.e., the thin layer of water above the sediment-water interface where diffusion is the dominant transport process)(Higashino et al., 2004; Glud et al.,

2007; Bryant et al., 2010). Variation in the thickness of the diffusive boundary layer over lake sediments is primarily determined by the turbulent energy of the benthic boundary layer, which can vary over hourly time-scales due to seiching (Lorke et al., 2003; Bryant et al., 2010). In lakes with relatively large seiches, SOD can vary as much as 85% over a single seiche (Bryant et al., 2010). Variation in the thickness of the diffusive boundary layer in neither the cores nor lakes from which the sediments were collected was measured so it is impossible to assess the degree to which the cores reproduce the conditions found in the lakes. All of the cores were incubated under identical conditions so differences in the thickness of the diffusive boundary layer among cores is likely minimal. Accordingly variability in my SOD values represent only a portion of the *in situ* variation in SOD.

After accounting for the variation in SOD due to temperature and oxygen concentration, I found significant differences in SOD associated with the source of the sediments. The differences among the sediments from the different lakes were small relative to the effects of temperature and oxygen concentration but suggest that other factors may be affecting SOD on a landscape scale. Due to the confounding effects of oxygen concentration and temperature not accounted for in the analysis of the temperature dependence and oxygen availability experiments, respectively, differences in SOD among sediments from the different lakes need to be interpreted cautiously. Nonetheless, among the shallow lakes, the sediments from lake S–3 consistently had higher SOD than those from lake E–4. One possibility is that sediment metabolism differs due to variation in sediment substrate quality along a gradient of lake productivity (Sugai and Kipphut, 1992; Pace and Prairie, 2005).

Nutrient and chlorophyll data (Table 3) suggest that all of the lakes in this study have similar and low water column productivity. Despite this, the organic content of the surface (0 – 3 cm) sediments of S–3 is greater than E–4 (Chapter 5) and the sediments from S–3 were more labile than those of E–4 when tested under controlled conditions (Chapter 3). These results indicate greater quantities and a higher quality of sediment organic matter in lake S–3, relative to lake E–4, which may result in greater SOD in the

former. Nonetheless, the differences in SOD due to sediment source were small relative to the effects of temperature and oxygen availability and are not likely to have significant impacts on the variation in SOD over short temporal scales in intact sediments.

The distribution of temperature and oxygen in the lakes appears to be determined primarily by thermal stratification. The epilimnetic water was between 1.2 to 8.9° C warmer than the bottom waters in the shallow lakes and 2.6 to 10.9° C warmer than the bottom waters in lake GTH 91. Stratification also had effects on oxygen availability. In the deeper lake (GTH 91) the continual summer stratification resulted in substantial oxygen deficits in the hypolimnion, and hypolimnion oxygen concentrations rarely exceeded 60% of those found at the epilimnion. The majority of the volume of both shallow lakes was above the thermocline and thus well oxygenated, however the small hypolimnion became rapidly depleted in oxygen during sustained stratification. Yet, even these oxygen deficits were eliminated with the typical loss of stratification at the end of July.

In addition to the effects of thermal stratification, the availability of oxygen is also controlled by benthic photosynthesis. Greater than 40% of the sediment surface of the shallow lakes and 20% of the sediment surface of lake GTH 91 were estimated to be above the photosynthetic compensation point during the summer. Photosynthetic sediments can have greater oxygen penetration depths (Epping and Jörgensen, 1996) and should overall have much lower oxygen limitation relative to sediments unable to support photosynthesis.

In lake GTH 91 sediment oxygen demand is estimated to contribute approximately 67% of the total hypolimnetic oxygen consumption. This estimate of the importance of SOD to hypolimnetic oxygen consumption is greater than has been generally reported for other systems. Algesten et al. (2005) found that sediment respiration was less than 10% of whole lake respiration in 15 unproductive subarctic lakes. Kling et al. (1991) estimated that sediment DIC production from lakes in the same region as my study ranged 13 to 50% of total lake DIC production. Although not directly comparable because they are estimated differences within a single deep lake, den Heyer and Kalff (1998) calculated that sediment respiration was 27% of total respiration at 3 m and 18% of the total respiration

at 10 m. The estimated relative contribution of SOD to hypolimnetic oxygen consumption lake GTH 91 is also greater than what was observed (Lasenby, 1975; Cornett and Rigler, 1984) in similar temperate lakes (27 – 40%).

2.5.1 Conclusions

These results extend the observations of SOD to a relatively understudied region of the Arctic. Consistent with the previous but limited measures of arctic lake SOD, the range of SOD that I measured was toward the low end of what has been observed at lower latitudes. Sediment oxygen demand (and therefore organic matter mineralization) was primarily controlled by factors varying at the within-lake scale (i.e., temperature and oxygen availability). Although temperature significantly affected SOD, these systems were less sensitive to temperature than lakes in temperate and sub-arctic regions.

Table 3: Morphometric characteristics and water column nutrient and chlorophyll a (Chl a) concentrations of study lakes. SRP refers to soluble reactive phosphorus. Concentration below the detection limit of the analysis are indicated with "bd".

Lake	Location	Surface Area (ha)	Maximum Depth (m)	NH_4 (μM)	NO_3 (μM)	SRP (μM)	Chl a (μg L^{-1})
E-4		4.0	4	bd	0.5	0.4	1.7
S-3		4.2	5	0.2	bd	bd	1.4
GTH 91		2.5	10	0.5	0.1	bd	1.5

Table 4: Experimental conditions of the temperature dependance experiment. The O_2 Conc. Range is the range of initial oxygen concentrations in the cores.

Lake	Date	Temperature (°C)	O_2 Conc. Range (mmol O_2 L^{-1})	Replicates
GTH 91	21 Jun 2006	2	0.25 – 0.32	10
	21 Jun 2006	9	0.25 – 0.30	10
	19 Jun 2006	12	0.25 – 0.31	10
	19 Jun 2006	17	0.24 – 0.28	10
S-3	11 Aug 2006	2	0.26 – 0.30	9
	11 Aug 2006	8	0.27 – 0.32	9
	8 Aug 2006	12	0.26 – 0.29	9
	8 Aug 2006	17	0.24 – 0.28	9
E-4	2 Aug 2006	2	0.23 – 0.30	9
	2 Aug 2006	7	0.24 – 0.33	8
	4 Aug 2006	12	0.24 – 0.27	9
	4 Aug 2006	17	0.20 – 0.24	8

Table 5: Results of the ANCOVA assessing the effects of temperature and sediment source on sediment oxygen demand from the temperature dependance experiment. Temp. refers to the temperature of the incubation and Lake is the source of the sediments.

Source	*df*	SS	F	p
Temp.	1	1056.2	62.4	< 0.0001
Lake	2	202.8	6.0	0.003
Temp. * Lake	2	17.6	0.5	0.596
Error	106	1792.7		

Table 6: Experimental conditions of the oxygen availability experiment. The O_2 Conc. Range is the range of initial oxygen concentrations in the cores.

Lake	Date	Temperature (° C)	O_2 Conc. Range (mmol O_2 L^{-1})	Replicates
GTH 91	26 Jul 2006	9	0.17 – 0.30	9
S-3	8 Aug 2006	12	0.11 – 0.29	9
E-4	4 Aug 2006	12	0.10 – 0.27	9

Table 7: Results of a repeated measures ANCOVA assessing the effect of the oxygen concentration in the overlying water on sediment oxygen demand from the oxygen availability experiment. Oxygen refers to the concentration of oxygen in the water overlying the sediment cores and Lake is the source of the sediments.

Source	*df*	df_{denom}	F	p
Intercept	1	123	243.1	< 0.0001
Oxygen	1	123	8.1	< 0.006
Lake	2	123	11.2	< 0.0001
Lake * Oxygen	2	123	0.8	0.44

Table 8: Summary of temperature and oxygen profiles collected from the shallow lakes (E–4 and S–3) during 2005–2008. Thermocline Z is the thermocline depth in m, while "ns" indicates that the lake was not stratified. K_d is the light extinction coefficient (m^{-1}). O_2 Deficit is the area-weighted mean oxygen concentration of the hypolimnion (mg O_2 L^{-1}) divided by the area-weighted mean oxygen concentration of the hypolimnion (mg O_2 L^{-1}) multiplied by 100. Epi. and Hypo. Temp. are the area-weighted mean temperature of the epilimnion and hypolimnion respectively. If there was no defined hypolimnion (i.e., the metalimnion contacted the bottom of the lake) the temperature of the deepest measurement is recorded and indicated with an "∗". Missing data is indicated with a dash (–), uncalculated results are indicated with "NA".

Lake	Year	Date	Thermocline Z	K_d	O_2 Deficit	Epi. Temp.	Hypo. Temp.
E–4	2006	20 Jun	2.5	1.19	90	10.1	5.4
		24 Jun	2.5	–	63	13.8	5.6
		29 Jun	3.0	1.09	98	13.5	5.9
		7 Jul	4.0	1.08	80	11.8	9.0∗
		15 Jul	4.0	0.91	–	13.0	11.1∗
		2 Aug	4.0	0.73	–	12.6	10.3∗
	2008	30 Jun	3.0	1.16	114	15.1	8.2∗
		9 Jul	3.5	0.92	56	15.3	9.4
		14 Jul	3.0	0.90	49	15.0	10.1∗
		21 Jul	ns	0.76	NA	12.6	11.4∗
		30 Jul	ns	0.81	NA	10.0	NA
S–3	2005	8 Jul	ns	–	NA	11.1	NA
		19 Jul	3.0	–	97	14.4	10.0
		25 Jul	4.5	–	104	15.2	12.7
		8 Aug	ns	0.78	NA	13.0	NA
	2006	22 Jun	2	0.97	73	14.2	5.8∗
	2008	30 Jun	2.5	0.98	63	15.4	7.4
		9 Jul	3.5	0.87	100	16.7	7.8∗
		14 Jul	3.5	0.78	–	16.3	8.8∗
		22 Jul	ns	0.69	–	11.4	NA
		31 Jul	ns	0.75	–	9.1	NA

Table 9: Summary of temperature and oxygen profiles from the deep lake (GTH 91) during 2005–2008. Thermocline Z is the thermocline depth in m, while "ns" indicates that the lake was not stratified. K_d is the light extinction coefficient (m^{-1}). O_2 Deficit is the area-weighted mean oxygen concentration of the hypolimnion (mg O_2 L^{-1}) divided by the area-weighted mean oxygen concentration of the hypolimnion (mg O_2 L^{-1}) multiplied by 100. $\bar{x}\ \alpha(z)$ is the area-weighted average of the ratio of the sediment area to water volume in the hypolimnion. Epi. and Hypo. Temp. are the area-weighted mean temperatures (°C) of the epilimnion and hypolimnion respectively. There were two distinct thermoclines on 26 Jul 2006. Missing data are indicated with a dash (–).

Year	Date	Thermocline Z	K_d	O_2 Deficit	Epi. Temp.	Hypo. Temp.
2005	22 Jul	3.5	–	53	12.6	5.6
	28 Jul	4.5	–	39	14.9	5.4
	3 Aug	4.5	–	26	12.8	5.3
2006	19 Jun	2.0	1.04	95	7.2	4.6
	23 Jun	2.0	–	78	11.6	4.6
	29 Jun	3.0	0.95	76	12.9	4.6
	3 Jul	4.0	–	–	11.9	4.5
	7 Jul	3.5	0.87	59	11.5	4.4
	19 Jul	5.0	0.87	60	11.9	4.5
	26 Jul	5.5	0.94	49	14.3 (13.2)	4.4
	1 Aug	4.0	0.56	32	13.4	4.4
2008	30 Jun	2.5	0.79	57	15.6	4.7
	9 Jul	3.5	0.83	59	15.3	4.9
	14 Jul	3.5	0.68	53	15.1	4.4
	21 Jul	4.5	0.84	–	13.1	5.0
	30 Jul	5.5	0.72	–	10.3	4.4

Table 10: The percentage of sediment area below the photosynthetic compensation point during the summers of 2005 – 2007. The sediments are estimated to be a sink for oxygen at irradiance levels less than 50μE m^{-2} s^{-1} based on the results of the light experiment.

	2005	2006	2007
Lake	12 Jun – 20 Aug	26 Jun – 24 Aug	24 Jun – 31 Aug
GTH 91	70	76	79
S–3	42	54	58
E–4	48	56	60

Table 11: Estimate of the proportion of total hypolimnetic oxygen consumption due to sediment oxygen consumption in lake GTH 91. Z is the depth in the lake in m. HOD is the estimated hypolimnetic oxygen demand (mmol O_2 m^{-3} d^{-1}). Vol is the volume of a 1 m thick slice of the hypolimnion below depth Z in m^3. HOC is the hypolimnetic oxygen consumption (mmol O_2 d^{-1}). SOD is the estimated sediment oxygen demand at 4.5° C (mmol O_2 m^{-2} d^{-1}). Area is the sediment area in a 1 m slice below depth Z in m^2. SOC is the sediment oxygen consumption at depth Z (mmol O_2 d^{-1}). Total is value of each variable for the whole hypolimnion (6 - 8 m).

Z	HOD	Vol	HOC	SOD	Area	SOC	SOC:HOC
6	3.7	7441	27532	10.5	1603	20518	0.75
7	7.0	5838	40866	10.5	1614	20659	0.51
8	5.5	4221	23215	10.5	1585	20288	0.87
Total			91613			61465	0.67

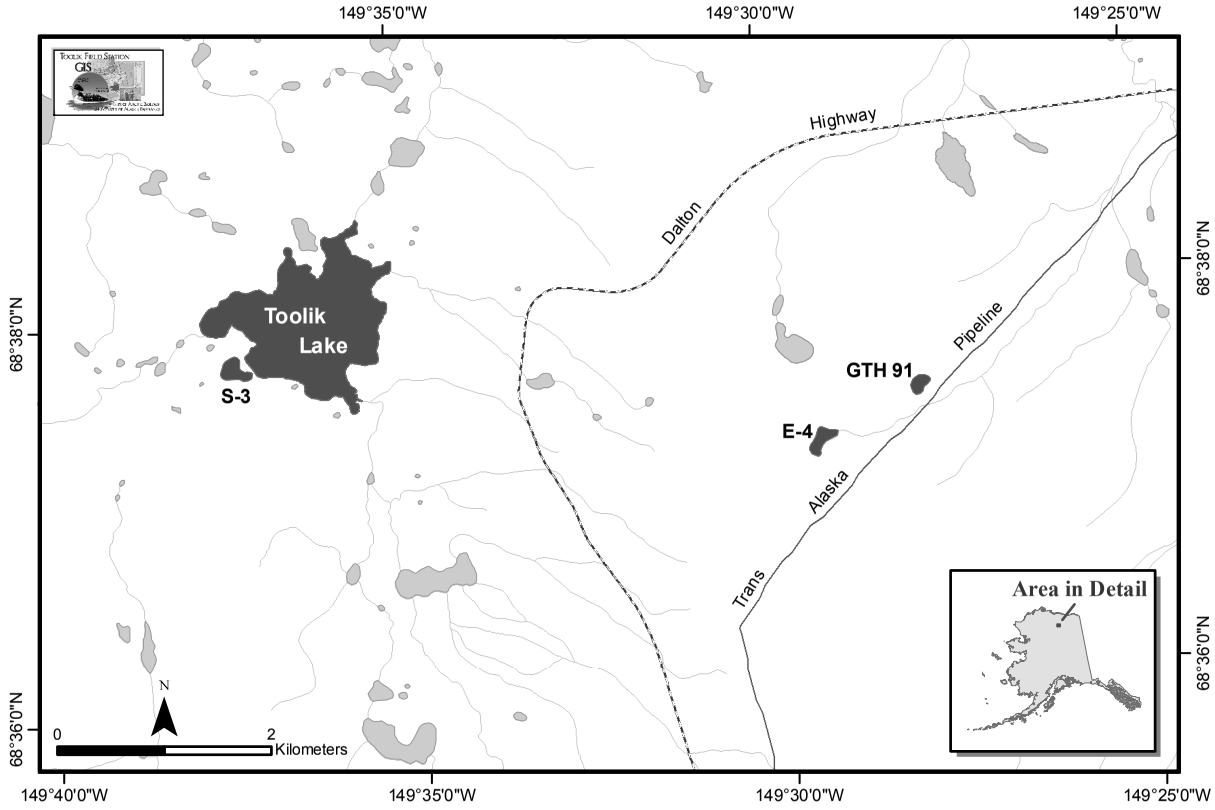

Figure 1: A map of the lakes used in the study of factors affecting sediment oxygen demand.

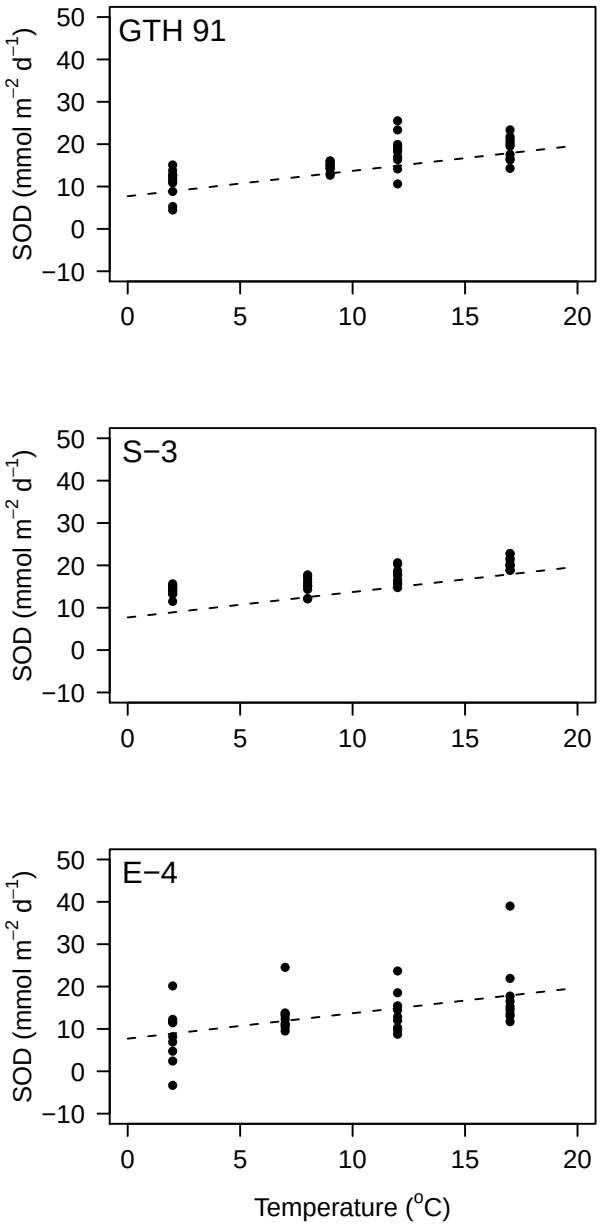

Figure 2: The relationship between temperature and sediment oxygen demand (SOD) in each lake. Each point represents the SOD from one core. The dashed line is the least squares regression based on the ANCOVA of temperature and sediment source.

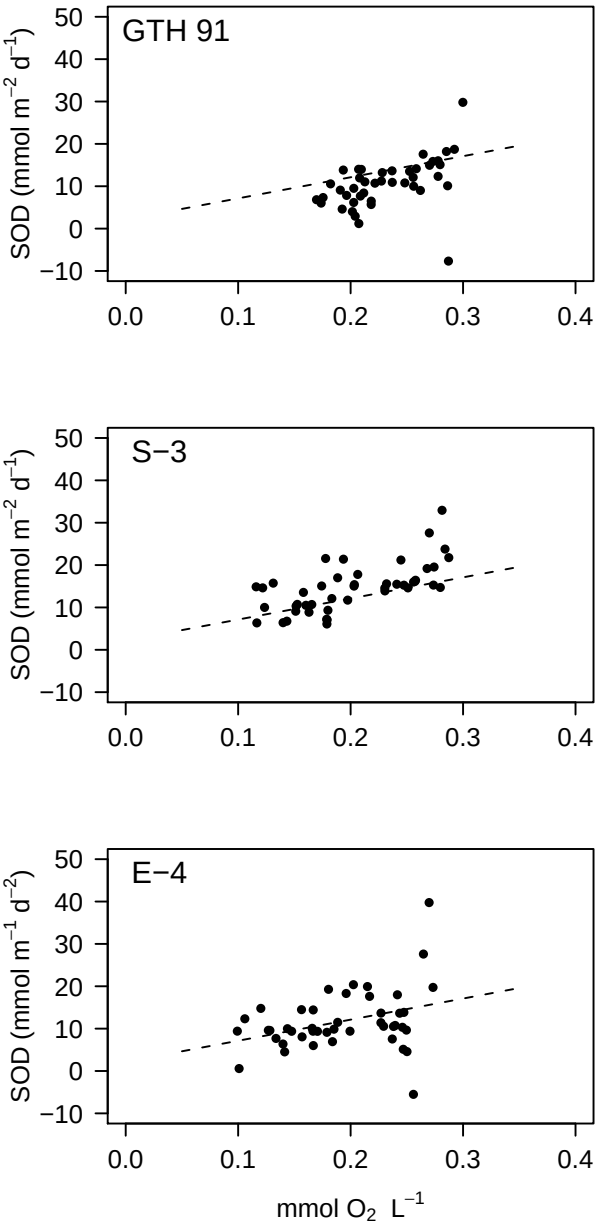

Figure 3: The effect of bottom water oxygen concentration on sediment oxygen demand (SOD) in sediments from lakes GTH 91, S-3, and E-4. Each point represents SOD from a single core and the dashed lines represent the least squares regression based on the repeated measures ANCOVA of all three lakes.

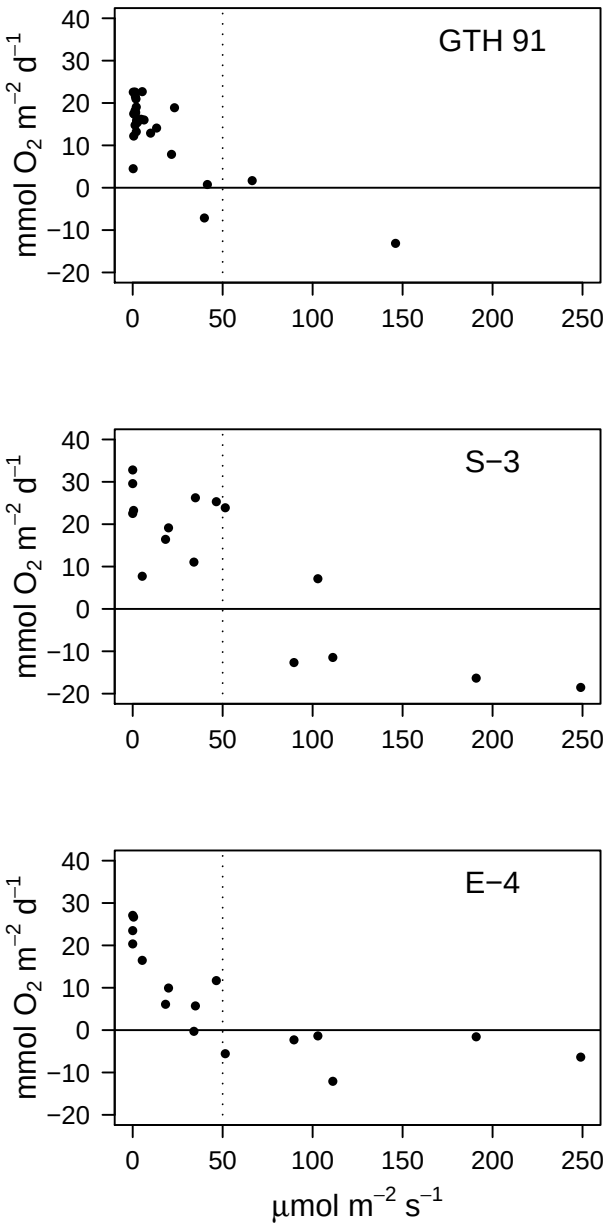

Figure 4: The effect of irradiance on net sediment oxygen demand (SOD) in sediments from lakes GTH 91, S-3, and E-4. Each point is the flux from a single core. The horizontal line indicates the sediment water interface. Points above the line represent fluxes out of the sediments and points below the line represent fluxes into the sediments. The vertical (dashed) line indicates the 50μmol m^{-2} s^{-1} irradiance level where the sediments switch from net oxygen consumption to net oxygen production.

3 The impact of sediment source and oxygen availability on dissolved inorganic carbon production (DIC) in lake sediment slurries

3.1 Introduction

The burial of organic matter in lake sediments is a major component of the global carbon cycle (Cole et al., 2007; Sobek et al., 2009) as lakes are estimated to bury between 5 and 14 g C m^{-2} y^{-1} in their sediments (Dean and Gorham, 1998; Stallard, 1998) representing 22% of the annual carbon burial in freshwater systems globally (Cole et al., 2007). Although the factors controlling lentic burial are less well documented than in marine systems, organic matter burial appears to be controlled both by the sedimentation rate and limitations on the mineralization of organic material once deposited (Sobek et al., 2009).

Sediment organic matter mineralization is limited by environmental factors (e.g., temperature, oxygen and electron acceptor availability) (Capone and Kiene, 1988; Andersen, 1996; Hartnett et al., 1998; Hulthe et al., 1998; Kristensen and Holmer, 2001; Lehmann et al., 2002) and/or qualities of the organic matter itself (e.g., molecular structure) (Hansen and Blackburn, 1991; Kristensen and Holmer, 2001). The paradigm of sediment organic matter preservation in marine systems predicts that the factors controlling organic matter preservation change with the time since sediment deposition. On sub–decadal and decadal time scales, the majority of sediment organic matter loss results from selective mineralization of labile material (Burdige, 2007). This selective loss of labile compounds increases the relative amount of refractory material in the organic matter pool, reducing the mineralization rate with time. On century to millennial scales, the fate of this refractory organic matter depends more on mechanisms that exclude organic material from mineralization

processes, such as anoxia or mineral sorption (Hulthe et al., 1998; Burdige, 2007). The range of studies in freshwater is more limited, but research suggests that organic matter mineralization in freshwater sediments follows a similar model (Capone and Kiene, 1988; Bastviken et al., 2003), although mineral sorption may be minor (Sobek et al., 2009).

The overall effect of environmental change on burial efficiency and the long-term fate of buried organic matter in a particular region will depend on the relative importance of structural and environmental limitations to sediment organic matter mineralization (Sobek et al., 2009). If the principal factors controlling organic matter mineralization are environmental, then changing conditions may result in the breakdown of previously stored organic matter. However, if the preservation of sediment organic matter is based on the structural lability of the organic matter, then environmental change will have little impact on stored organic matter.

Arctic lakes in the region surrounding Toolik Lake are similar to other lakes in the world in that they have organic sediments (Chapter 5) which sequester carbon for long time periods (Cornwell and Kipphut, 1992). In Chapter 2, I show that virtually all of the explainable variation in sediment oxygen demand (and by inference organic matter mineralization) in sediment cores from 3 shallow, low arctic lakes is due to variation in temperature and oxygen availability (i.e., environmental factors). This sensitivity of mineralization rate to these environmental conditions indicates that changing conditions could have substantial effects on organic matter preservation in these systems, irrespective of the sediment source. However these findings are based on short-term incubations of intact cores and it is unclear whether these findings reflect the limitations to mineralization under more persistent environmental change. Chapter 5 shows significant differences in organic matter burial efficiency between lakes in the same region, suggesting that some of the variation in organic matter burial on a landscape-scale is due to differences in the source (and quality) of sediment organic matter between lakes.

In this chapter, I quantify sediment organic matter lability (DIC production) from sediments of different diagenetic age under oxic and anoxic conditions to assess the rel-

ative role of environmental conditions and structural characteristics in limiting organic matter preservation in low arctic lakes. I test the hypothesis that sediment organic matter preservation in shallow, low arctic lakes is limited primarily by environmental conditions, specifically isolation from oxic environments. This information will aid predictions about the factors affecting burial efficiency under future environmental conditions.

3.2 Materials and Methods

Sediments were collected using a K–B style corer on 25 June 2008 (Julian day 177) from a depth of 3 m in lakes E–4 and S–3. A detailed description of these lakes can be found in Chapter 2. Due to differences in sediment bulk density, 11 and 22 cores were required to collect similar amounts of sediment from lakes E–4 and S–3, respectively. Sediment slurries from each lake were prepared separately but identically. The sediments from the cores were pooled into "surface" (1 – 2 cm) and "deep" (9 – 10 cm) fractions. The surface and deep sediments were then each divided into the oxic and anoxic treatments. The sediment batches were diluted to an estimated 22 mg dry sediment ml^{-1} using lake water collected from 3 m in conjunction with the core collection. The amount of dilution needed was calculated using the bulk density of the sediments determined in 2007. Following dilution, the slurries were passed through a 200 μm mesh to exclude macrofauna. The lake water used to dilute the anoxic treatments was bubbled with N_2 gas continuously to remove oxygen. Twenty ml of slurry was added to each of 3 glass serum bottles (160 ml) for each treatment combination. The oxic treatments were left open to the atmosphere and the anoxic treatments were immediately sealed with a rubber stopper and purged 10 times with N_2. Ten ml of the slurry from each treatment was added to a pre–weighed glass scintillation vial and dried at 50° C for at least 48 h to determine dry sediment mass. The dry sediment was then ashed at 550° C for 4 h to determine organic matter content by loss on ignition.

The incubation began on Julian day 177 when all of the bottles were randomly arranged on an orbital shaker turning at approximately 100 rpm. The bottles were incubated in

total darkness and sampled on Julian days 178, 180, 184, 187, 193, 203, and 211. At the beginning of each sampling event, the oxic treatments were sealed with a rubber stopper and the stoppers of the anoxic treatments were briefly pierced with a syringe to equalize the pressure in the bottle. Approximately 1 h after the bottles were sealed (or equilibrated), a 4 ml sample of the headspace gas was removed with a syringe (sample T_0). Two ml of the gas sample were injected into a Shimadzu GC–8A, thermal conductivity gas chromatograph (operating conditions: column = 2 m length x 1/8 od porapak N, carrier gas = ultra high purity He at 30 ml min^{-1}, column temperature = 50° C, injector and detector temperature = 90° C, current = 140mA, precision = CV < 2% at 10 replicates of 1000 ppm CO_2) to determine the concentration of CO_2. The remaining 2 ml of sample was injected into a Shimadzu GC–8A, flame ionization detector gas chromatograph (operating conditions: column = 1 m Mol Sieve 5A (60/80), carrier gas = ultra high purity N_2 at 33 ml min^{-1}, column temperature = 90° C, injector and detector temperature = 140° C, precision = CV < 1% at 10 replicates of 10 ppm CH_4) to determine the CH_4 concentration. On Julian day 178, a second set of samples (T_1 and T_2) were collected at approximately 4 and 7 h after the T_0 sample. On all other sampling dates the T_1 and T_2 samples were collected approximately 3 and 6 h from the T_0 sample. Methane was not measured on Julian days 178 and 187.

Following the T_2 sample collection, the pH of each slurry was measured. The bottles remained sealed until the pH measurement was taken to minimize changes in pH due to atmospheric exposure. Following the pH measurements, the anaerobic bottles were resealed and purged 10 times with N_2 to remove any oxygen that entered during the sampling. Evaporative losses in the oxic treatments were determined gravimetrically on Julian days 190 and 201. Lost volume was replaced with deionized water. The temperature of the incubation room was measured with a mercury thermometer inserted into a water bath adjacent to the incubation setup.

3.2.1 Calculations and Statistical Analysis

The total inorganic carbon or CH_4 production rate between the two sampling intervals (T_0 – T_1 and T_1 – T_2) was calculated as the change in mass of headspace CO_2 or CH_4 plus the masses of aqueous DIC or CH_4, normalized to the incubation time and the dry mass of organic matter in the slurry. The mass of aqueous DIC and CH_4 were calculated using Henry's Law and, in the case of DIC, the equilibrium constants for HCO_3^-, CO_3^{2-} (David, 1996-1997), and the pH of the slurry.

Differences in the total inorganic carbon flux were analyzed using a 3–way repeated measures ANOVA with lake, Julian day, and treatment as factors plus all interactions. The effects of oxygen (oxic vs. anoxic) and sediment depth (surface vs. deep) were assessed using orthogonal contrasts of the least squared means (LS–means) of the treatment factor. All analyses were performed using JMP software (Ver. 4.0.4, SAS Institute Inc., Cary, NC, 1989–2007). The methane flux was not analyzed statistically.

3.3 Results

During the first 16 d, the incubation room maintained a mean (± SD) temperature of 19.7° C (± 0.7). Thereafter, the temperature fluctuated more dramatically due to inadequate temperature control (Fig. 5). Since sediment metabolic rates are significantly affected by temperature, as well as length of incubation I cannot distinguish between temperature and time effects following Julian day 193 (Fig. 5). Because of the confounding effects of temperature and time, only data collected during the first 16 days of the incubation (i.e., Julian days 177–193) were used in the analyses.

The initial percent organic matter content of the sediments differed between the sediments collected from the different lakes and depths (Table 12). Percent organic matter content was positively related to DIC production ($F_{1,6} = 14.07$, $p = 0.01$) and predicted 70% of the variation in DIC flux across all sampling dates (Fig. 6). Variation in DIC flux was significantly affected by the source of the sediments, the incubation time, and the experimental treatment. All interactions were significant except the 3-way interaction

between the lake from which the sediments were collected, the incubation time, and the experimental treatment (Table 13).

The source of the sediments (i.e., lake) had the largest effect on DIC production (Fig. 7). Overall, the mean DIC flux from lake S–3 sediments was 291% greater than the flux from lake E–4 sediments. The next largest effect on DIC flux was depth from which sediments were taken. The orthogonal contrast between the surface and deep sediments showed a significant effect of depth on the DIC flux irrespective of the effect of lake or the presence of oxygen. The mean DIC flux from the surface sediments was 190% greater than from the deep sediments, overall. The presence of oxygen had the weakest effect on the DIC flux. The orthogonal contrast between the oxic and anoxic sediments was significant but the mean DIC flux from the oxic sediments was only 60% greater than the flux from the anoxic sediments. When the variation due to other variables was accounted for there was a slight but significant decline in mean DIC flux with incubation time (i.e., Julian day).

In addition to the main effects, there were significant interactions between the lake from which the sediment was collected, the sediment depth, and the oxygen availability (i.e., the lake by treatment interaction), which can be illustrated by looking at the effects of sediment depth and oxygen within each lake. Comparison of the DIC flux from the same depth zone in the presence and absence of oxygen illustrates the oxygen sensitivity of organic matter mineralization (Fig. 7). In the sediments from lake S–3, the deep sediments in anoxic conditions produced twice as much DIC as the deep sediments in oxic conditions. In the treatments containing surface sediments from lake S–3, the oxic treatments produced an average 40% more DIC than anoxic treatments. The deep sediments from lake E–4 had very low DIC production under anoxic conditions and the presence of oxygen raised the mean DIC production by 642%. The effect of oxygen was less dramatic in the surface sediments of lake E–4 and the mean DIC flux increased by only 120% in the oxic treatments relative to the anoxic treatments.

Comparison of the DIC flux from the surface and deep sediments measures any increase

in organic matter recalcitrance with burial time. Differences in this comparison between oxic and anoxic conditions suggests the stability of this recalcitrance in changing conditions (i.e., oxygen availability). There was only a slight difference in mean DIC production between the surface and deep sediments from lake S–3 in the absence of oxygen but 200% greater mean DIC flux in the surface sediments under oxic conditions (Fig. 7). Lake E–4 had 160% greater mean DIC production in the surface sediments than in the deep sediments under oxic conditions. In the absence of oxygen the mean DIC production from the shallow sediments of lake E–4 was 780% greater than the DIC production from the deep sediments.

Substantial methane production was only observed in the surface sediments in the absence of oxygen (Fig. 8). The mean flux from the surface anoxic sediments of S–3 was 1.7 times greater than the mean flux from the surface anoxic sediments from E–4. Overall methane production was 2 orders of magnitude lower than DIC production from the same sediments.

3.4 Discussion

Dissolved inorganic carbon production (i.e., organic matter mineralization) from the sediment slurries was significantly affected by both the source of the organic matter (i.e., sediments) and the presence of oxygen during the incubation. These results indicate that both structural lability and environmental conditions (i.e., anoxia) may affect sediment organic matter preservation in low arctic lakes.

The greatest differences in DIC production resulted from the source of the sediments. The mean DIC flux from the sediments of lake S–3 was overall 291% greater than the flux from lake E–4 and the mean DIC flux from the shallow sediments in both lakes was 190% greater than the flux from the deep sediments. In both comparisons the greater DIC flux occurred in sediments with higher percent organic matter content. Since the DIC flux is normalized to grams organic matter in the slurry, the elevated DIC production from sediments with higher percent organic matter reflects differences in sediment organic

matter quality and not quantity. The effect of sediment source on DIC flux in these lakes is similar to the response of DIC flux to organic matter lability in both marine and freshwater systems. Hansen and Blackburn (1991) found a 220% increase in DIC flux following an addition of labile organic matter (algal detritus) to marine sediment cores and Bastviken et al. (2003) found 108% to 267% greater DIC production in predominantly authochthonous versus allochthonous lake sediments.

The differences in organic matter lability between depths is likely related to the selective mineralization of reactive organic matter as the sediments age (Burdige, 2007). The source of the differences in organic matter lability between lakes is less clear but is likely due to differences in epipelic production. Both lakes are ultra–oligotrophic and have very low phytoplankton biomass (Table 3), suggesting that phytoplankton sedimentation is minor. Despite clear differences in the percent organic matter content and lability of the sediments, there are only minor differences between the lakes in depth and nutrient concentration (Table 3). Lake S–3 has slightly lower light attenuation (Table 8) and it is likely that the greater proportion of labile organic matter in lake S–3 sediments is due to greater benthic primary production fueled by greater light penetration.

Overall these findings support the idea that the reactivity of autochthonous organic matter favors mineralization and that sediment carbon sequestration is primarily supported by allochthonous organic matter sources (Sobek et al., 2009). The difference in percent organic matter content between the deep sediments was much smaller than between the shallow sediments, suggesting that the sediment organic matter content of the deep lake sediments will converge in time. Thus, the interaction between organic matter source and mineralization rate in these lakes and other shallow low arctic lakes may normalize carbon burial over long time periods and minimize the effect of differences in autochthonous production.

The preservation of organic matter in sediments may also be affected by environmental conditions that prevent the breakdown of otherwise labile material. Previous work has shown that the absence of oxygen (or limited oxygen exposure time) can limit sediment

organic matter mineralization (Capone and Kiene, 1988; Hartnett et al., 1998; Sobek et al., 2009). Oxygen is used as both an electron acceptor (Capone and Kiene, 1988; Bastviken et al., 2003) and during the initial hydrolysis of complex organic molecules with oxidase and peroxidase enzymes (Kristensen and Holmer, 2001; Bastviken et al., 2003; Burdige, 2007). The need for oxygen in the initial hydrolysis of complex molecules means that the oxygen sensitivity of the mineralization rate often decreases with increasing organic matter lability (Kristensen, 2000; Kristensen and Holmer, 2001; Lehmann et al., 2002).

Previous studies have found as much as 900% more DIC production from refractory diatom cultures exposed to oxygen relative to anoxic incubations (Kristensen and Holmer, 2001). Less refractory sources of organic matter typically show less dramatic differences and Andersen (1996) and Hansen and Blackburn (1991) found 25% and 57% greater DIC flux in oxic relative to anoxic intact marine cores respectively. Using eutrophic lake sediments, Bastviken et al. (2003) found 28% greater DIC flux in oxic relative to anoxic conditions. The overall 60% increase in DIC flux in oxic relative to anoxic conditions from the sediments of the low arctic lakes in my study compares favorably to the lower end of this range. However, the presence of oxygen significantly increased the flux of DIC from the sediments of both lakes and there was a clear interaction between the effect of oxygen and the lability of the sediment.

Between the sediment sources there were significant differences in the effect of oxygen on DIC production. In the sediments from lake E–4, the deeper sediments had 641% more DIC production in the presence of oxygen. This level of oxygen sensitivity is in the same range (300% to 900%) as cultures of refractory algal and terrestrial (hay) organic matter (Kristensen and Holmer, 2001) and suggests that the organic matter in the deep sediments of lake E–4 is diagenetically old (Kristensen, 2000; Burdige, 2007; Sobek et al., 2009). The surface sediments of lake E–4 were much less sensitive to oxygen and only showed 120% increase in mean DIC production in oxic conditions. However this increase is still consistent with diagenetically aged material, such as the 127% increase in DIC flux under oxic conditions observed by Bastviken et al. (2003) for allochthonous lake sediments,

and the 150% increase reported by Hulthe et al. (1998) for marine shelf sediments.

In the shallow sediments of lake S–3 the mean DIC flux from the oxic treatments was 40% greater than the anoxic treatments. Hulthe et al. (1998) found that DIC production from marine sediment slurries differed between oxic and anoxic by only -21 to 35% when fresh organic matter was added. The similarity of the effect in the shallow sediments from lake S–3 supports the conclusion that these sediments contain diagenetically young (presumably autochthonous) organic matter.

The behavior of the deep sediments from lake S–3 was not consistent with the idea that the mineralization of refractory organic matter is limited by oxygen. The deep sediments from lake S–3 incubated under anoxic conditions produced more DIC than the oxic sediments. Highly labile organic matter can be mineralized at similar rates under oxic and anoxic conditions (Kristensen, 2000; Burdige, 2007) but it is unlikely that such high lability explains this observation. The surface sediments from lake S–3 show an increase in DIC flux when exposed to oxygen indicating that some of the organic matter in the shallow sediments is sufficiently refractory to be oxygen sensitive. If the anoxic DIC flux from the deep sediments was controlled entirely by the structural lability of the sediments, then the amount of labile organic matter would have had to increase with burial, and there is no reason to believe that the sediments would gain labile organic matter during burial. ^{210}Pb analysis of the S–3 sediments shows no evidence of mixing down to the depth of the deep sediment (10 cm) in this experiment (Chapter 5). The DIC flux from the anoxic, deep sediments had more variability than the other treatments from lake S–3. This variability suggests a less consistent process in this treatment despite the constant incubation conditions. At present I do not have an explanation for the unexpected behavior of the sediments in this treatment.

Methane production was two orders of magnitude lower than DIC production and essentially only observed in the anoxic shallow sediments. The relative amount of organic matter mineralized through methanogenesis therefore appears to be minor. In productive lakes, between 43% (Lojen et al., 1999) to greater than 80% (Ogrinc et al., 1997;

Vreča, 2003) of the overall gaseous C flux may be derived from methanogenesis but the low methane production (relative to DIC production) observed in the anoxic treatments indicates that methanogenesis is not a major component of organic matter mineralization in the unproductive lakes in this study. Significant methane production by the oxic treatments would not be expected but the lack of methane production in the deep anoxic sediments further indicates the refractory nature of the buried organic matter. The greater methane production in the sediments from lake S–3 relative to lake E–4 was consistent with the idea that overall, more organic matter is being mineralized in lake S–3.

3.5 Conclusions

The mineralization of sediment organic matter was significantly affected by the source of the organic matter and the presence of oxygen. Overall, differences in the structural lability of the organic matter (i.e., the source of the sediments) had the greatest effect on DIC production. The largest differences in DIC production rates were observed in sediments derived from different organic matter pools; either different lakes or different depths within a lake. Therefore, the organic matter preservation in these lakes appears to be primarily facilitated by the structural lability of the organic matter inputs to the sediments. Future changes in autochthonous inputs are not likely to have large impacts on long-term organic matter storage but increased inputs of allochthonous material would likely increase storage.

Exposing the sediments to oxygen increased the mineralization of sediment organic matter and the relative effect of oxygen was generally greatest in the most refractory sediments. These findings suggest that changing environmental conditions that would expose stored organic matter to oxygen (e.g., increased sediment resuspension or lake drying), would result in mineralization losses and the relative effect would be greater in more refractory organic matter pools.

Table 12: The sediment mass and organic matter content (% OM) of the slurries.

Lake	Treatment	g dry sediment ml^{-1}	%OM
E–4	Anoxic Deep	0.020	29.2
	Anoxic Shallow	0.032	32.6
	Oxic Deep	0.019	29.8
	Oxic Shallow	0.034	32.1
S–3	Anoxic Deep	0.008	35.2
	Anoxic Shallow	0.017	63.6
	Oxic Deep	0.008	36.5
	Oxic Shallow	0.017	63.4

Table 13: Results of repeated measures ANOVA of the DIC flux normalized to organic matter mass.

Source	*df*	df_{denom}	F	p
Lake	1	85	337.8	<0.0001
Julian	1	85	16.4	0.0001
Treatment	3	19	81.2	<0.0001
Lake * Julian	1	85	3.6	0.062
Lake * Treatment	3	85	44.3	<0.0001
Julian * Treatment	3	85	4.9	0.004
Lake * Julian * Treatment	3	85	0.7	0.526

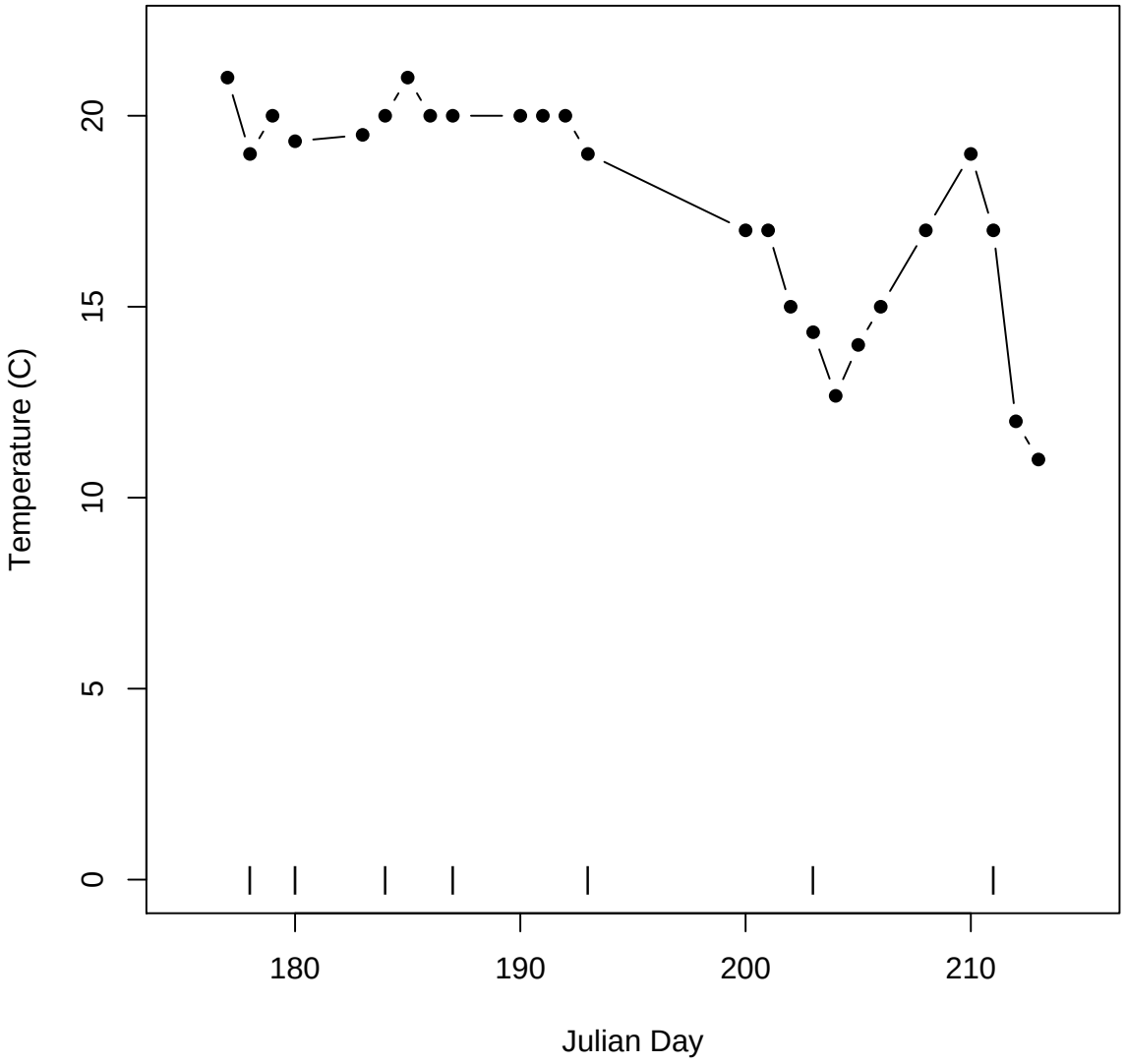

Figure 5: Mean temperature (°C) by Julian day during the sediment slurry experiment. The vertical hash–marks along the X–axis indicate dissolved inorganic carbon (DIC) sampling events.

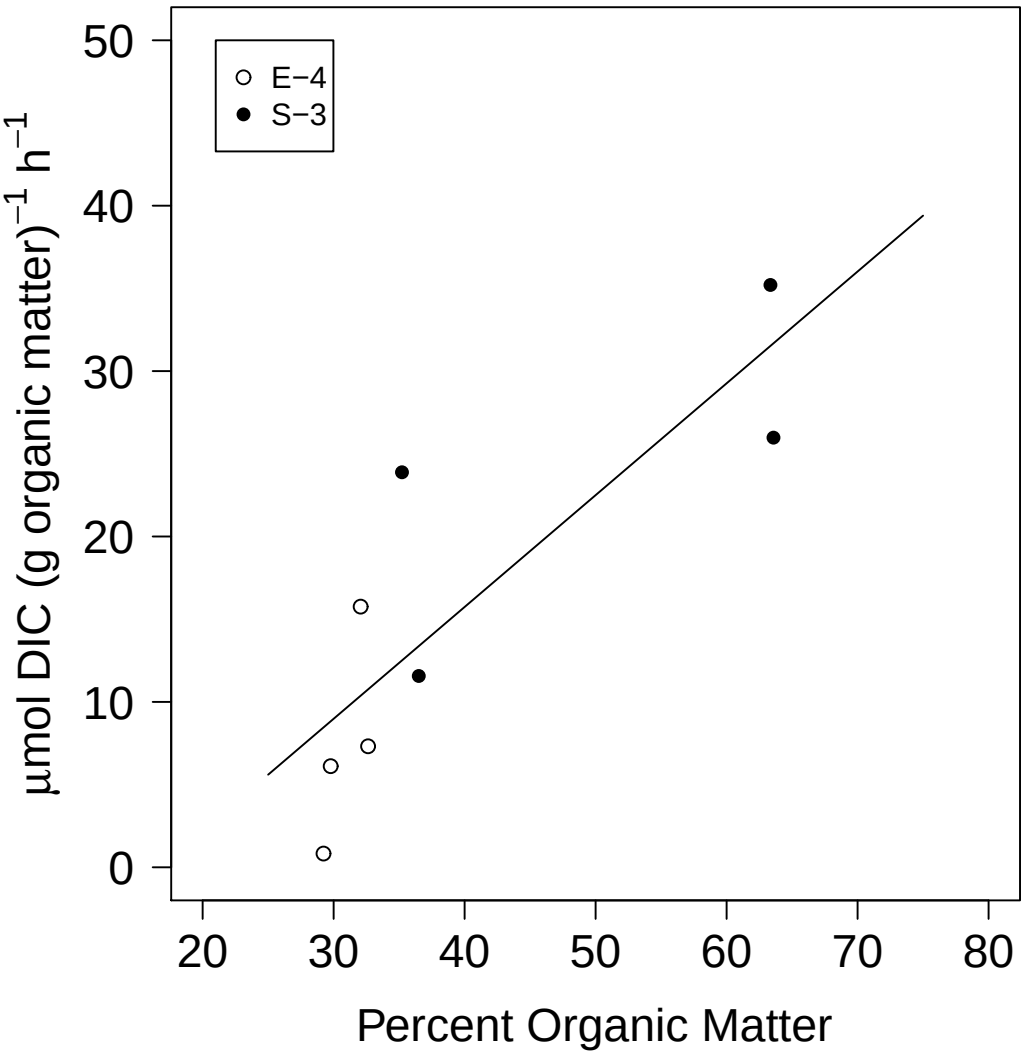

Figure 6: The relationship between dissolved inorganic carbon (DIC) flux and the percent organic matter content of the slurries in each lake. The line is best fit least squares regression through all of the points.

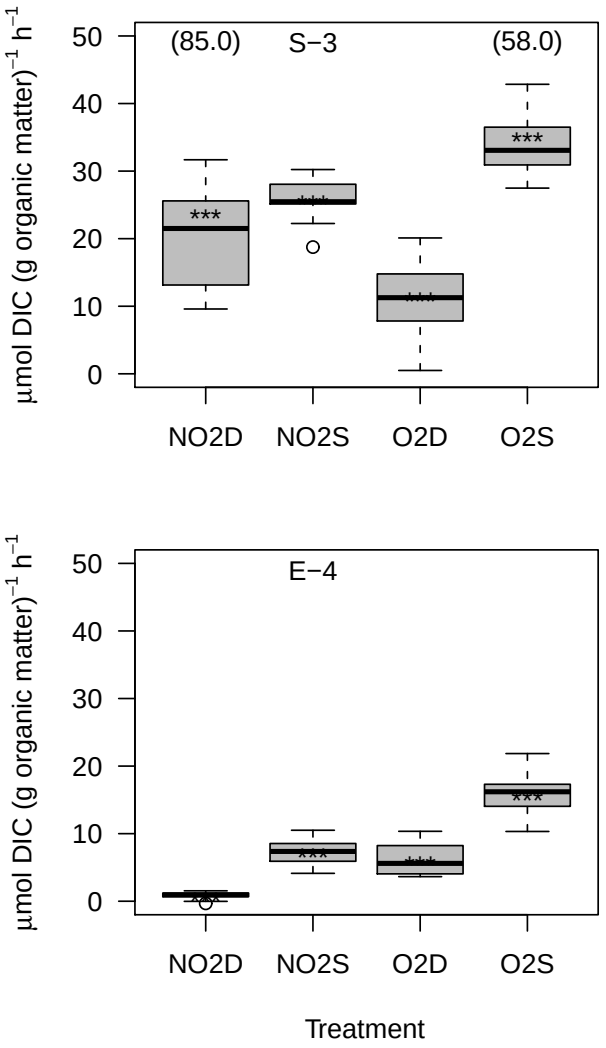

Figure 7: A boxplot of the flux of dissolved inorganic carbon (DIC) in each of the treatments in lakes S–3 and E–4. The treatments are sediments from 9–10 cm without oxygen (NO2D), surface (1–2 cm) sediments without oxygen (NO2S), sediments from 9–10 cm with oxygen (O2D), and surface (1–2) sediments with oxygen (O2S). Boxes represent the upper and lower quartiles, whiskers indicate 1.5 X the interquartile range. The median of the data is represented by the horizontal bar within the box and the mean is represented by "***". Any values that fall outside of the range of the whiskers are shown as open points. The numbers in parentheses are the values of outlying points that fall above the scale of the axis.

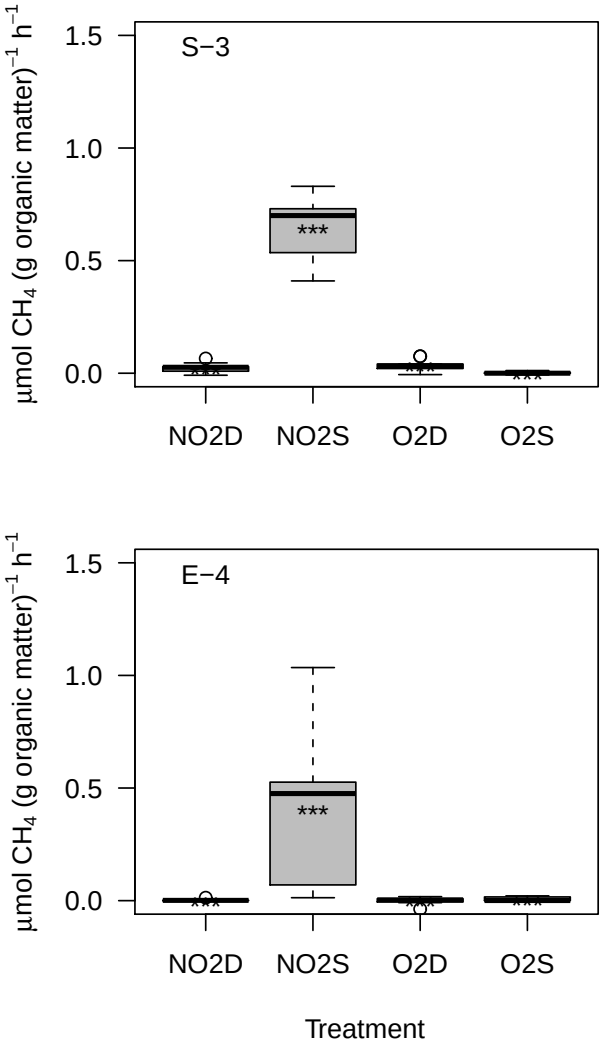

Figure 8: A boxplot of the flux of methane from each of the treatments in lakes S–3 and E–4. The treatments are sediments from 9–10 cm without oxygen (NO2D), surface (1–2 cm) sediments without oxygen (NO2S), sediments from 9–10 cm with oxygen (O2D), and surface (1–2)sediments with oxygen (O2S). Boxes represent the upper and lower quartiles, whiskers indicate 1.5 X the interquartile range. The median of the data is represented by the horizontal bar within the box and the mean is represented by "***". Any values that fall outside of the range of the whiskers are shown as open points.

4 The effect of light attenuation on thermal stratification in arctic lakes: Implications for sediment organic matter processing

4.1 Introduction

Lake sediments are an important component of global organic matter cycling. Globally, lakes are estimated to retain between 0.03 to 0.07 Pg of organic carbon per year in their sediments, which represents 30 to 60% of oceanic storage (Cole et al., 2007). Furthermore, Holocene lentic sediments contain 25 to 50% of the total terrestrial organic carbon (Cole et al., 2007). The fate of organic matter in lake sediments can therefore have implications for the carbon cycling at scales greater than would be implied by their total surface area. Chapters 2 and 3 show that increased oxygen availability and temperature significantly increase the mineralization of sediment organic matter in arctic lakes. Therefore, factors affecting the temperature and oxygen distribution of a lake will have implications for sediment organic matter sequestration.

Thermal stratification is the main factor controlling the distribution of temperature and oxygen in lakes. Stratified lakes contain steep and persistent temperature and oxygen gradients which alter biogeochemistry and trophic dynamics (Wetzel, 2001). Since sediments are affected by the physical properties of the overlying water, the depth and intensity of lake stratification also affects sediment processes. In stratified lakes, the sediments of the epilimnion are exposed to relatively warm and oxygenated conditions which elevate organic matter mineralization rates. In the hypolimnion, the sediments are exposed to relatively cold temperatures and are isolated from inputs of atmospheric oxygen, which lowers organic matter mineralization rates (Granéli, 1978; Pace and Prairie, 2005;

Sobek et al., 2009). The depth of the thermocline determines the proportion of the sediment area within the epilimnion and hypolimnion, and therefore the relative availability of two principal factors determining the fate of sequestered sediment organic matter (i.e., temperature and oxygen).

The attenuation of light energy by suspended and dissolved compounds in the lake water is one of the principal factors affecting thermocline depth (Fee et al., 1996; Houser, 2006; Caplanne and Laurion, 2008) and therefore the distribution of temperature and oxygen in lakes. There is evidence that dissolved organic carbon (DOC) inputs to freshwater systems may change under future climate scenarios due to changes in vegetation and hydrology (Forsberg, 1992; Schindler et al., 1997; Clair et al., 1999; Evans et al., 2005) which would change lake transparency and indirectly affect organic matter storage in lake sediments by changing the proportion of sediments exposed to conditions favoring organic matter mineralization (i.e., warm and oxygenated) vs. storage (cold and anoxic). In this study I evaluate the effect of light attenuation on thermocline depth in two separate lake groups during 2006, 2007, and 2008. I apply the results of these surveys to a well–studied lake to predict the impact that changes in lake clarity would have on the proportion of the lake sediments exposed to relatively warm, oxygenated water and therefore the storage potential of sediment organic matter.

4.2 Materials and Methods

4.2.1 Lake Surveys

On 7 August, 2006, I sampled a total of 18 lakes in the region to the east of the Toolik Lake Field Station (Fig 9). At each lake, a temperature and light (photosynthetic photon flux density; PPFD) profile was collected using a YSI Model 85 multiparameter water quality meter (temperature) and LiCor LI-192SA underwater 2π quantum sensor with a Li-Cor LI-250 Quantum Meter (PPFD). In some lakes the profiles for both temperature and PPFD were collected with a Hydrolab, Data Sonde 5. Light and temperature measurements were collected at 0.5 m intervals from the water surface to the lake bottom or to a point

where the lake reached a constant hypolimnetic temperature with increasing depth. The thermocline depth was defined as the depth with the greatest change in temperature from the preceding depth. The light attenuation coefficient (K_d) in the lake was determined as the slope of the natural log of the PPFD with depth.

A second survey was conducted on 7 August, 2007. Twelve lakes were sampled to the north of the Toolik lake field station (Fig. 9). Temperature and light profiles were collected and analyzed using the same methods as in 2006. Surface water samples for dissolved organic carbon (DOC) concentration was collected in 20 ml borosilicate glass scintillation vials. Each sample was filtered through a 0.45 μm polypropylene (PP) filter, acidified with 500 μl of 1N HCl and stored at 4^o C until analyzed on a Schimadzu TOC–V Total Carbon Analyzer for DOC.

The final survey was conducted on 12 July, 2008. I resampled 15 of the same lakes that were sampled in 2006. Temperature and light profiles were collected using the same methods as in 2006 and 2007. A 1 L sample of surface water was collected from each lake in an amber high density polyethylene (HDPE) sample bottle from which, 2 duplicate 20 ml sub–samples were filtered through a 0.45 μm PP filter and analyzed for chromophoric dissolved organic matter (cDOM) using a Turner Designs 10–AU fluorometer (Clark et al., 2004). Following the cDOM measurement, the samples were acidified with 100 μl of 1 N HCl and analyzed on a Schimadzu TOC–V Total Carbon Analyzer for DOC. The remaining 960 ml of sample were used to determine Chlorophyll a (Chl a) concentration. Chl a concentration was measured fluorometrically (Turner Designs Model TD-70 Fluorometer) from duplicate samples of filter–trapped (Whatman GF/F) particulate matter extracted for 24 h in a buffered 90% acetone solution (Welschmeyer, 1994).

Lake areas and maximum depths were determined as part of a group sampling effort and lake watershed areas were calculated using a digital elevation model and the hydrology toolset in ArcMap GIS software (ESRI, 2006). Monthly rainfall was calculated from data collected by the Arctic LTER program (Shaver, 2006, 2008).

4.2.2 Statistics and Calculations

The relationship between the thermocline depth and K_d was investigated with simple regressions for each year. Variables were natural log transformed when appropriate to improve linearity. Dissolved organic carbon, cDOM, and Chl a concentrations were standardized as Z–scores (Gotelli and Ellison, 2004) to facilitate direct comparison of the regression slopes. Differences in watershed:lake area between the lakes sampled in 2007 and 2008 were compared using a Kolmogorov-Smirnov test. Dissolved organic carbon, cDOM, and Chl a concentrations, as well as K_d, lake area, and Julian day were compared using Pearson's correlations. Analyses were performed in JMP (JMP, Ver. 4.0.4. SAS Institute Inc., Cary, NC, 1989–2007) or R (R Development Core Team, 2009).

4.2.3 Case Study

The impact of the relationship between K_d and thermocline depth on the area of sediments exposed to epilimnetic (i.e., warm and oxygenated) water was evaluated using data from lake GTH 91. Lake GTH 91 (68.62°N; -149.47°W) is a 2.5 ha oligotrophic lake with a maximum depth of 10 m. Temperature, irradiance and dissolved oxygen were sampled using a YSI Model 85 water quality meter and LI-192SA underwater 2π quantum sensor with a LiCor 250 Quantum Meter as described above. The bathymetry of the lake was mapped by combining a lake perimeter measurement determined with a Tremble Geo Explorer with sonar transects collected with a Garmin GPSMAP 180. The lake bottom profile was extrapolated from the sonar measurements using a triangulated irregular network to convert the observed depths into 1 m contour lines in ARC–GIS (ESRI, 2006). Surface areas and volumes of each depth interval were calculated using the 1 m contour intervals. The surface area of the sediments above the thermocline was estimated from the hypsographic curve for the lake.

To model how changes in lake clarity would affect the percent sediment area above the thermocline in lake GTH 91, I recalculated the thermocline depth and percent sediment area above the thermocline following the addition or subtraction of 0.5 m^{-1} from K_d on each

date using the relationship between thermocline depth, K_d, surface area, and Julian day identified by the multiple regression described above. The magnitude of the hypothesized addition or reduction in K_d (0.5 m^{-1}) is based on the mean K_d observed from the survey lakes (0.51 m^{-1}; see Results). The addition of this value therefore represents an approximate doubling of the observed K_d.

4.3 Results

4.3.1 Lake Surveys

Light attenuation coefficients (K_d) ranged from 0.21 to 0.97 m^{-1} across all years and had a mean (± 1 SD) of 0.51 (0.2) m^{-1} (Tables 14 and 15). The mean K_d in the lakes sampled in 2008 was significantly greater ($F_{2, 44} = 6.4$, $p = 0.003$) than the mean K_d of the 2006 or 2007 samples (Fig. 10). K_d was significantly correlated with lake surface area in 2008 ($r = -0.68$, $p = 0.003$) and Julian day ($r = -0.47$, $p = 0.001$) but not with lake surface area in 2006 ($r = -0.35$, $p = 0.152$) or 2007 ($r = 0.12$, $p = 0.716$). Thermocline depth ranged 3.5 to 11.5 m across all years (Tables 14 and 15) and was significantly related to K_d (Fig. 11, Table 16).

The DOC concentration ranged 4.2 to 6.0 mg L^{-1} in the lakes sampled in 2007, which was significantly greater ($F_{1, 27} = 38.6$, $p < 0.001$) and less variable than the range of DOC concentrations (1.3 to 4.2 mg L^{-1}) in the 2008 survey (Tables 14 and 15). Dissolved organic carbon concentration was significantly related to K_d in the 2008 survey ($r^2 = 0.62$, $p = 0.0002$) but not in the 2007 survey ($r^2 = 0.17$, $p = 0.185$) (Fig. 12). There was no significant difference in the watershed:lake area between the lakes sampled in 2007 and 2008 ($p = 0.775$)

During the 2008 survey, DOC and cDOM concentrations were significantly and highly correlated ($r = 0.94$, $p < 0.0001$). Chl a concentrations were also significantly correlated with DOC concentration ($r = 0.52$, $p = 0.031$) and cDOM concentration ($r = 0.51$, $p = 0.034$). The significant correlations prevent the partitioning of the individual impact of each factor on K_d in a single model. Separate regressions show similar effects of the

standardized (Z–score) variables on K_d (Fig. 13). Chromophoric dissolved organic matter concentration explained the most variation in K_d ($r^2 = 0.78$), while Chl a concentration explained the least variation in K_d ($r^2 = 0.34$) and DOC concentration was intermediate ($r^2 = 0.62$) (Fig. 13). Estimating light attenuation due to Chl a concentration alone using a Chl a specific attenuation coefficient of 0.016 m^{-1} (mg Chl a)$^{-1}$ (Bannister, 1974) produced values varying from 0.006 and 0.034 m^{-1}, which accounted for only 1.25 to 4% of observed variation in K_d.

4.3.2 Case Study

Thermocline depth ranged between 2.0 and 5.5 m during the times that lake GTH 91 was sampled in 2006 and 2008. Over the same time period, K_d ranged from 0.56 to 1.04 m^{-1}. The model containing only K_d, surface area, and Julian day without interactions (Table 17) explains 92% of the temporal variation in thermocline depth in lake GTH 91. A regression fit to the actual data also explains 92% of the temporal variation in thermocline depth and shows a 0.088 m d^{-1} increase in thermocline depth. The modeled data estimates a 0.086 m d^{-1} increase in thermocline depth and skews the thermocline depth an average of 12.4% deeper than the observed data (Fig. 14).

During the period sampled, the percent sediment area above the thermocline varied from 44 to 75%. Following an increase in K_d of 0.5 m^{-1} there would be a median decrease in the percent sediment area above the thermocline of 30% overall. A decrease in K_d of 0.5 m^{-1} would result in a median increase in percent sediment area above the thermocline of 14% (Fig. 15). The epilimnetic water of GTH 91 was an average ($\pm$ 1 SD) of 7.2 ($\pm$ 2.0) °C warmer (Fig. 16) and had an average of 2.5 ($\pm$ 1.2) mg L^{-1} more dissolved oxygen than the hypolimnetic water (Fig. 17).

4.4 Discussion

4.4.1 Lake Surveys

The results of the surveys clearly show that lakes with greater light attenuation develop shallower thermoclines. Although the relationship is correlative and therefore cannot indicate conclusively that light attenuation determines thermocline depth, the relationship between water clarity and thermocline depth has been observed in other freshwater systems. In small lakes (< 500 ha), water clarity is considered a principal factor controlling thermocline depth (Fee et al., 1996; Mazumder and Taylor, 1994; Houser, 2006). Perhaps the strongest evidence supporting a causal relationship between light attenuation and thermocline depth is the deepening of the epilimnion that follows increases in clarity due to other factors (e.g., pH or filter-feeder community shifts) (Bukaveckas and Driscoll, 1991; Yu and Culver, 2000).

Light attenuation results from the absorption of photons by dissolved and particulate matter suspended in the lake water (Wetzel, 2001). The lakes in the region of this study contain very low levels of suspended particulates (K. Fortino, pers. obs.) and therefore, as in other systems, light attenuation would likely result from chromophoric DOC (Fee et al., 1996; Houser, 2006; Caplanne and Laurion, 2008). The ability of DOC concentration to predict K_d differed between the 2007 and 2008 surveys. In 2007, DOC was a poor predictor of K_d and there was less light attenuation per unit of DOC than in the 2008 survey. In 2008, DOC concentration explained 62% of the variation in K_d and was very highly correlated with the cDOM concentration. Chromophoric dissolved organic matter was not measured in 2007 but it is likely that the poor relationship between DOC and clarity in 2007 survey is due to a decoupling of the bulk DOC concentration from the chromophoric component of the DOC.

The lakes sampled in the different years were in distinct lake groups so it is possible that there are catchment-scale differences in the relative proportion of cDOM and non-chromophoric DOC delivered to the lakes. However, non-chromophoric (and therefore, less light attenuating) DOC is commonly derived from algal exudates or the photobleaching

of cDOM within the lake (Zepp, 2003). Although Chl a was not measured in the 2007 lakes, algal biomass tends to be regionally very low and it is unlikely that the source of the non–chromophoric DOC is algal exudates in these systems. The most likely source of the non-chromophoric DOC in the 2007 survey lakes is the photobleaching of cDOM. Photobleaching rates are variable between lakes but color loss can be as high as 19% d^{-1} (Reche et al., 1999). Lakes with greater lake-water residence time tend to have greater photobleaching of the cDOM (Hargreaves, 2003). Relative differences in water residence time can be estimated by comparison of watershed area relative to lake volume. I do not have measures of lake volume for the survey lakes, but lake volume is well correlated with lake surface area in those lakes in the region for which bathymatric data exist ($r = 0.98$, $n = 20$; K. Fortino unpub. data). Therefore, watershed:lake area should provide an estimate of relative water residence time. There is no difference in watershed:lake area between the lakes in the 2007 and 2008 surveys and thus, no simple morphometric prediction of greater water residence times and photobleaching in the 2007 survey lakes. Alternatively, hydrologic differences between the years may have resulted in differences in water residence times and photobleaching potential. The June and July rainfall totals in 2007 were 34 and 55% respectively of those for 2008 (Shaver, 2007, 2008). The lower rainfall during 2007 should have increased the water residence time of the lakes and may have increased DOC photobleaching.

In the 2008 survey DOC, cDOM, and Chl a all have a nearly identical relationship with K_d, however the correlation between the factors prevents the statistical partitioning of their individual effects. Unlike in 2007, bulk DOC is highly correlated with K_d and cDOM, suggesting that photobleaching was likely minor and that both variables reflect allochthonous organic matter (OM) inputs. The relationship between Chl a and K_d is significant but estimation of the light attenuation specific to Chl a shows that the observed Chl a concentrations could only account for a small amount ($< 5\%$) of the measured K_d. Overall, these results suggest that K_d is likely controlled by variation in the input and photobleaching of allochthonous OM.

4.4.2 Case Study

A regression using K_d, Julian day, and lake surface area (without interaction) accurately predicted the observed changes in thermocline depth in lake GTH 91. The model skews the thermocline depth an average of 12.4% deeper than the observed data but it is not clear what factors may be contributing to this bias. The observed data show that as the thermocline increases throughout the summer, the sediment area above the thermocline increases from slightly less than half to more than three fourths of the total sediment area indicating that thermocline depth is a major factor determining the temperature and oxygen conditions of the sediments.

Previous incubations of sediment cores from lake GTH 91 and other regional lakes showed that sediment organic matter mineralization (measured as sediment oxygen demand) declined between a 1.8 and 2.2 mmol O_2 m^{-2} d^{-1} per mg L^{-1} drop in oxygen concentration (Chapter 2). Similarly log sediment organic matter mineralization declined an average of 0.22 (mmol O_2 m^{-2} d^{-1}) per reduction in log degree C. Combining these estimated effects with the observed mean differences in temperature and oxygen concentration between the epilimnion and hypolimnion in lake GTH 91 predicts 6 to 7 mmol O_2 m^{-2} d^{-1} greater oxygen consumption (and therefore organic matter mineralization) in the sediments of the epilimnion relative to those in the hypolimnion. These estimates are conservative since they are based on sediment oxygen demand in dark incubations. Photosynthetic oxygen production in the epilimnion could increase the differences between the epilimnion and hypolimnion even further. The proportion of epilimnetic sediment area and therefore the proportion of the lake with a greater organic matter mineralization rate in the sediments is determined by thermocline depth. Since cDOM concentration is a major factor controlling thermocline depth, the loading of cDOM has an important indirect effect on the mineralization and storage of organic matter in the lakes sediments.

4.5 Implications and Conclusion

The arctic is expected to experience dramatic changes as a result of human induced climate warming (Chapin et al., 2005; Sturm et al., 2005). One likely outcome is that the increase in terrestrial primary production that results from warming and shrub expansion (Myneni et al., 1997; Jia et al., 2003) will result in increased organic matter export from the landscape to the lakes (Neff and Hooper, 2002). Warming may also directly stimulate soil DOC production and increase the export of DOC from the landscape (Freeman et al., 2001; Worrall and Burt, 2004).

A decrease in organic matter export from the landscape is also possible. Soil organic matter could be lost as CO_2 if soil carbon mineralization is stimulated to a greater degree than primary production (Shaver et al., 1992). Reduced precipitation and the resultant decline in soil moisture and runoff would also reduce the amount of DOC exported from the landscape (Schindler et al., 1997; Tranvik and Jansson, 2002; Blodau et al., 2004; Clark et al., 2005). However, drying can stimulate the production of soil DOC (Freeman et al., 2001; Worrall et al., 2004) and may lead to an increase in DOC export even in the absence of shrub expansion. Although it is unclear as to the direction, the ongoing alterations in arctic climate will almost certainly alter the delivery of DOC to lakes.

Changes in K_d substantially alter the amount of sediments exposed to the relatively warm, oxygenated waters of the epilimnion. A decrease in cDOM loading from the watershed would increase thermocline depth and result in an overall increase in the area of epilimnetic sediments with greater organic matter mineralization rates and would reduce organic matter burial efficiency. Conversely, an increase in cDOM loading would reduce thermocline depth and the area of epilimnetic sediments with greater organic matter mineralization rates thereby increasing the burial efficiency of the lake.

Concomitant with these effects on burial efficiency would be the other impacts of changes in cDOM (and more generally OM) loading from the watershed. Allochthonous inputs are an important source of organic matter to the sediments of oligotrophic lakes (Wetzel, 2001), so changes in OM loading will alter the delivery of organic matter to

the sediments simultaneous with the alterations in the processing of that organic matter. In the case of increased OM loading from the watershed, the lake will experience both an increase in the amount of organic matter inputs and a reduction in the proportion of epilimnetic lake sediments with reduced burial efficiency due to cDOM. Therefore it is possible that the net effect of these two processes could buffer the system against increased organic matter mobility from the watershed.

Table 14: Description of lakes surveyed during 2006 and 2008. Z_{max} is the maximum depth of the lake (m), Area is the surface area of the lake (ha), WS Area is the area of the lakes watershed (ha), K_d is the light attenuation coefficient in m^{-1}, Thermocline Z is the depth of the thermocline (m), DOC is the concentration of dissolved organic carbon in mg L^{-1}, cDOM is the concentration of chromophoric dissolved organic matter in quinone sulfate units (QSU), and Chl a is the concentration of chlorphyll a in μg L^{-1}. A dash indicates that the lake was not stratified at the time of the sampling. A "ns" indicates that the parameter was not sampled.

Lake	Z_{max}	Area	WS Area	Year	K_d	Thermocline Z	DOC	cDOM	Chl a
GTH 30	21.4	6.8	526.9	2006	0.87	5.0	ns	ns	ns
				2008	0.86	4.0	4.2	67.6	0.84
GTH 31	12.0	2.2	114.0	2006	0.58	6.0	ns	ns	ns
				2008	0.97	3.5	3.9	59.6	0.76
GTH 32	15.7	12.9	568.0	2006	0.80	6.0	ns	ns	ns
				2008	0.70	4.5	3.6	48.7	1.1
GTH 33	13.3	4.2	635.9	2006	0.76	5.0	ns	ns	ns
				2008	0.77	4.0	3.7	46.55	1.07
GTH 34	17.4	3.6	782.6	2006	0.80	5.5	ns	ns	ns
				2008	0.83	4.5	3.5	47.9	0.92
GTH 57	21.6	30.0	228.5	2006	0.25	9.5	ns	ns	ns
				2008	0.39	5.5	2.3	15.5	0.46
GTH 58	16.2	4.1	103.2	2006	0.39	7.5	ns	ns	ns
				2008	0.61	3.5	4.1	45.1	1.3
GTH 65	16.5	4.3	141.2	2006	0.38	6.0	ns	ns	ns
				2008	0.62	4.0	3.1	46	0.91
GTH 66	25.9	16.6	202.9	2006	0.30	7.0	ns	ns	ns
				2008	0.54	4.0	2.1	20.9	0.46
GTH 68	27.4	77.1	478.7	2006	0.26	11.5	ns	ns	ns
				2008	0.30	5.5	1.3	5.9	0.42
GTH 69	14.3	6.1	17.1	2006	0.37	6.0	ns	ns	ns
GTH 71	18.3	17.1	66.0	2006	0.23	9.0	ns	ns	ns
				2008	0.47	5.5	1.4	3.9	0.55
GTH 73	13.7	9.9	213.9	2006	0.46	–	ns	ns	ns
				2008	0.68	4.5	2.3	17.5	0.97
GTH 74	11.3	5.9	1295.1	2006	0.27	–	ns	ns	ns
				2008	0.65	4.5	2.8	33.9	0.64
GTH 92	10.4	5.6	842.0	2006	0.23	7.5	ns	ns	ns
				2008	0.62	4.0	1.7	25	0.80
GTH 96	9.2	8.2	504.8	2006	0.21	9.5	ns	ns	ns
				2008	0.85	4.5	3.3	46.7	2.1

Table 15: Description of lakes surveyed during 2007. Z_{max} is the maximum depth of the lake (m), Area is the surface area of the lake (ha), WS Area is the area of the lake's watershed (ha), K_d is the light attenuation coefficient (m^{-1}), Thermocline Z is the depth of the thermocline (m), and DOC is the concentration of dissolved organic carbon (mg L^{-1}).

Lake	Z_{max}	Area	WS Area	K_d	Thermocline Z	DOC
GTH 10	11.3	0.1	11.7	0.37	5.5	4.5
GTH 13	11.3	19.8	112.8	0.49	4.5	5.5
GTH 16	9.8	6.7	82.4	0.39	6.0	4.7
GTH 18	15.2	13.6	495.7	0.41	6.5	4.2
GTH 19	10.7	4.3	16.1	0.37	6.5	4.2
GTH 20	18.3	1.2	39.1	0.46	5.5	6.0
GTH 21	13.7	5.1	373.3	0.39	6.0	4.6
GTH 23	9.5	1.3	518.5	0.42	5.5	5.0
GTH 24	9.1	3.3	937.1	0.57	5.0	4.6
GTH 76	12.2	6.2	53.3	0.41	7.5	4.4
GTH 80	10.1	2.1	73.2	0.44	6.0	4.5
GTH 81	10.1	12.7	62.6	0.41	6.5	4.3

Table 16: Results simple regression analysis of the K_d and thermocline depth in the 2006, 2007, and 2008 survey lakes.

Year	Source	*df*	SS	F	p
2006	K_d	1	28.1	17.20	0.001
	Error	14	22.9		
2007	K_d	1	2.48	5.58	0.040
	Error	10	4.44		
2008	K_d	1	3.79	15.31	0.001
	Error	15	3.71		
		Estimate	r^2		
2006	Intercept	9.76	0.55		
	K_d	-5.77			
2007	Intercept	9.45	0.36		
	K_d	-8.25			
2008	Intercept	6.20	0.51		
	K_d	-2.67			

Table 17: Results of the multiple regression analysis (without interaction terms) of the factors affecting thermocline depth in the survey lakes.

Source	*df*	SS	F	p
Model	3	91.20	35.40	< 0.0001
Error	41	35.21		
	Estimate	Std. Err.	t	p
Intercept	-7.08	3.22	-2.20	0.03
K_d	-3.08	0.87	-3.53	0.001
Area	0.04	0.01	3.76	0.001
Julian Day	0.067	0.014	4.85	<0.0001

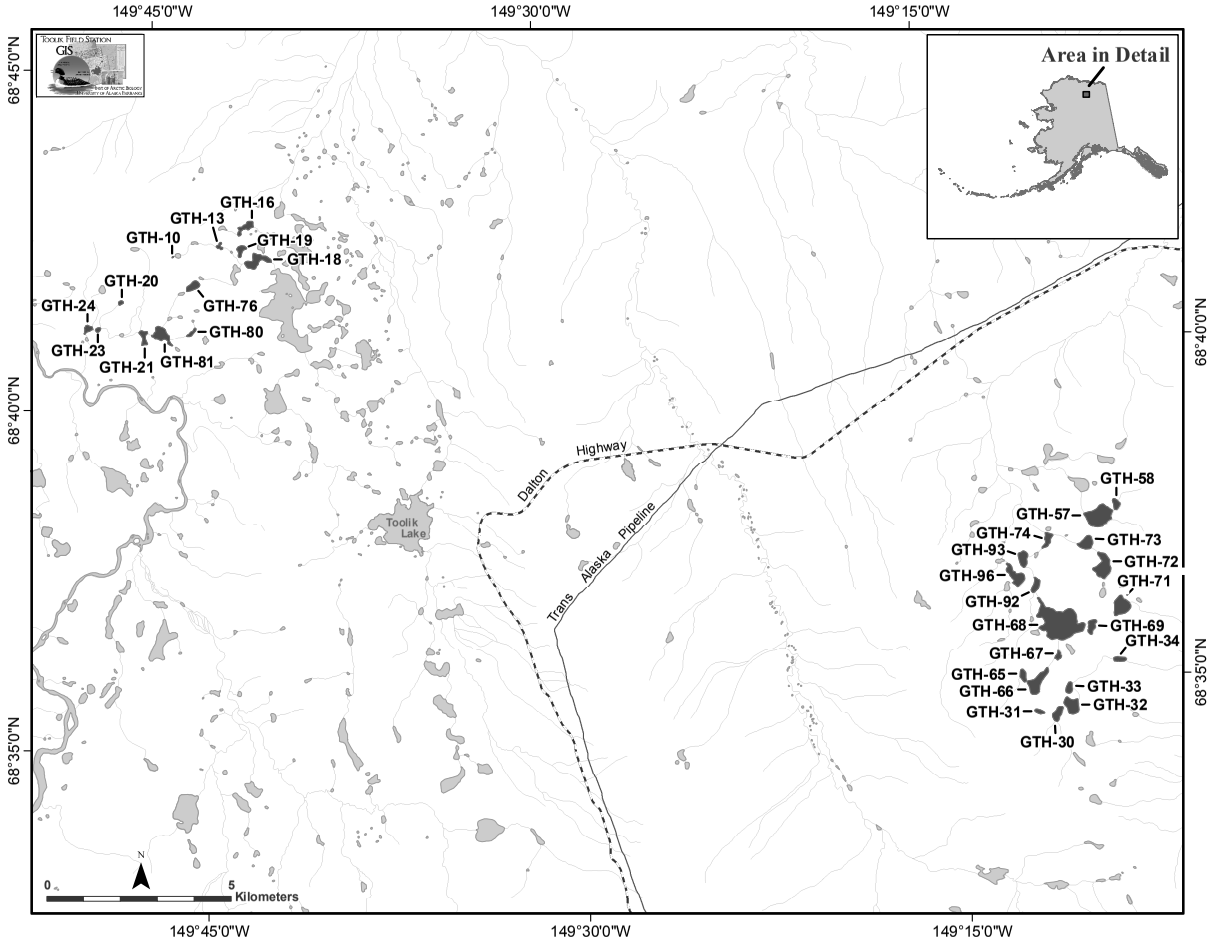

Figure 9: A map of the lakes used in the survey of light attenuation and themocline depth.

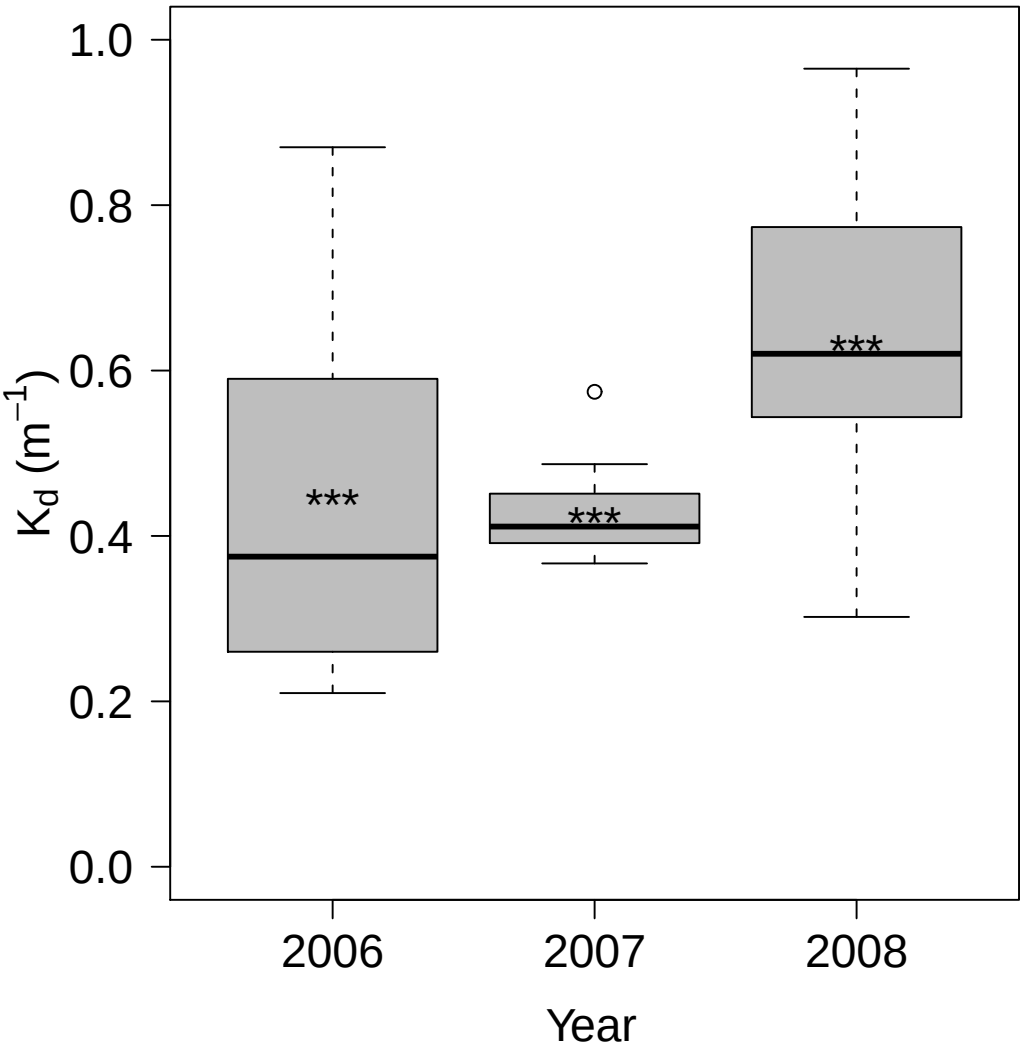

Figure 10: Light attenuation coefficient (K_d) by year. Boxes represent the upper and lower quartiles, whiskers indicate the full extent of the data or 1.5 X the interquartile range. The median of the data is represented by the horizontal bar within the box and the mean is represented by "***". Any values that fall outside of the range of the whiskers are shown as open points.

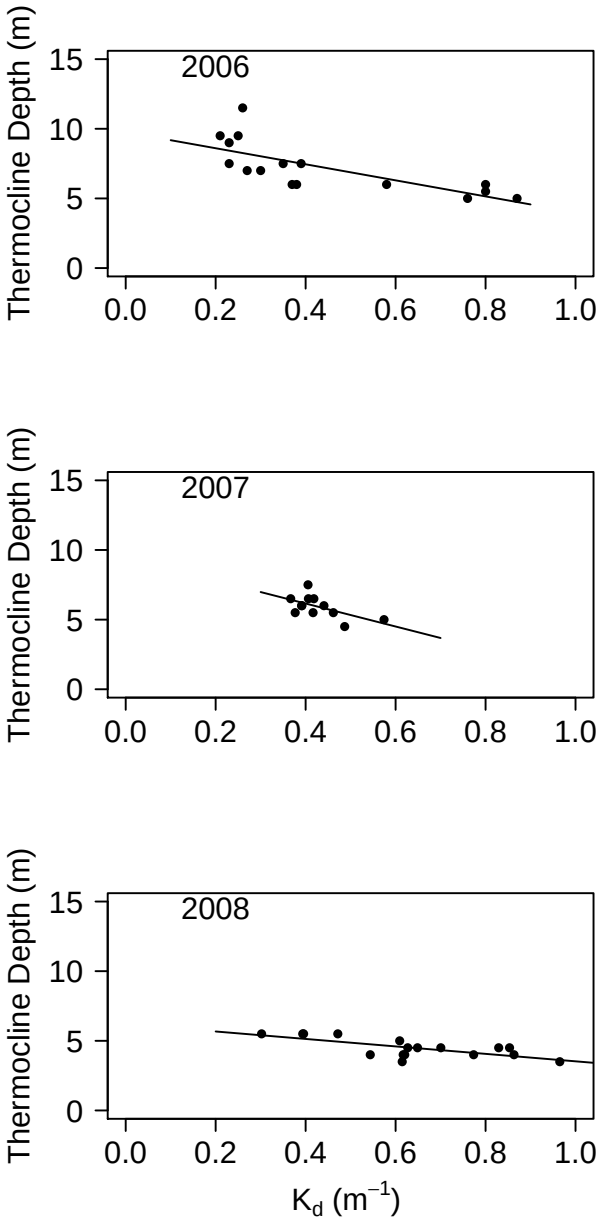

Figure 11: The relationship between thermocline depth and light attenuation coefficient (K_d) in surveys conducted between 2006 and 2008. The line represents the best fit least squares regression.

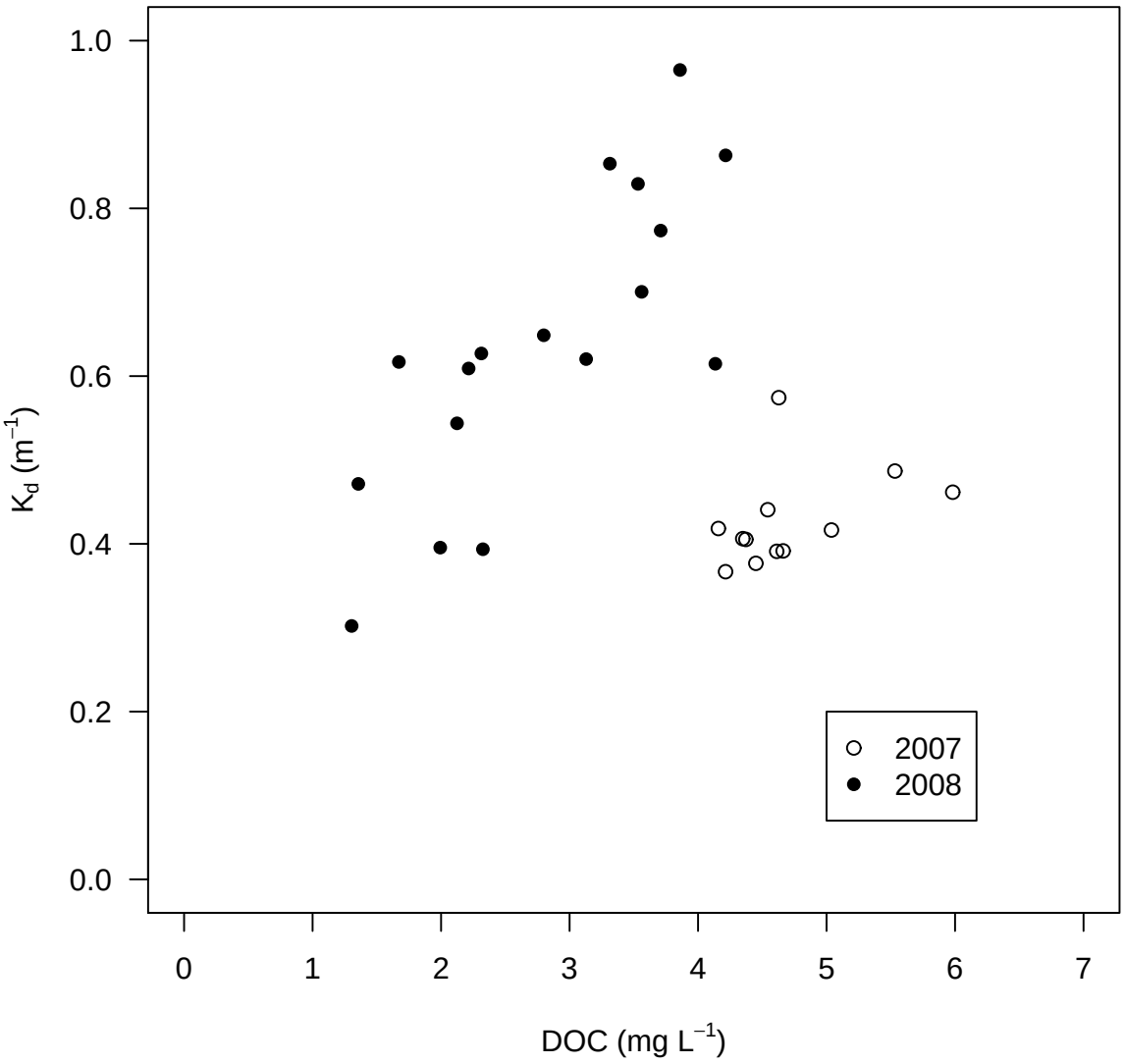

Figure 12: The relationship between the dissolved organic carbon (DOC) concentration and light attenuation coefficient (K_d) in the lakes sampled in 2007 and 2008.

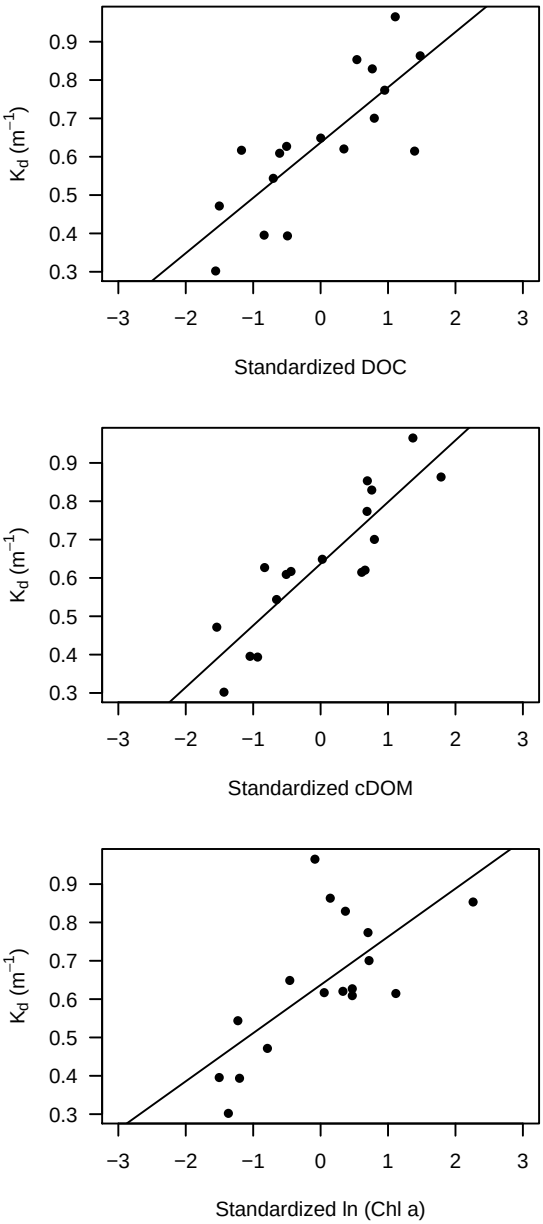

Figure 13: The relationship between the light attenuation coefficient (K_d) and Z–score standardized dissolved organic carbon (DOC), chromophoric dissolved organic matter (cDOM), or natural log transformed chlorophyll a (Chl a) concentrations in the lakes sampled in 2008.

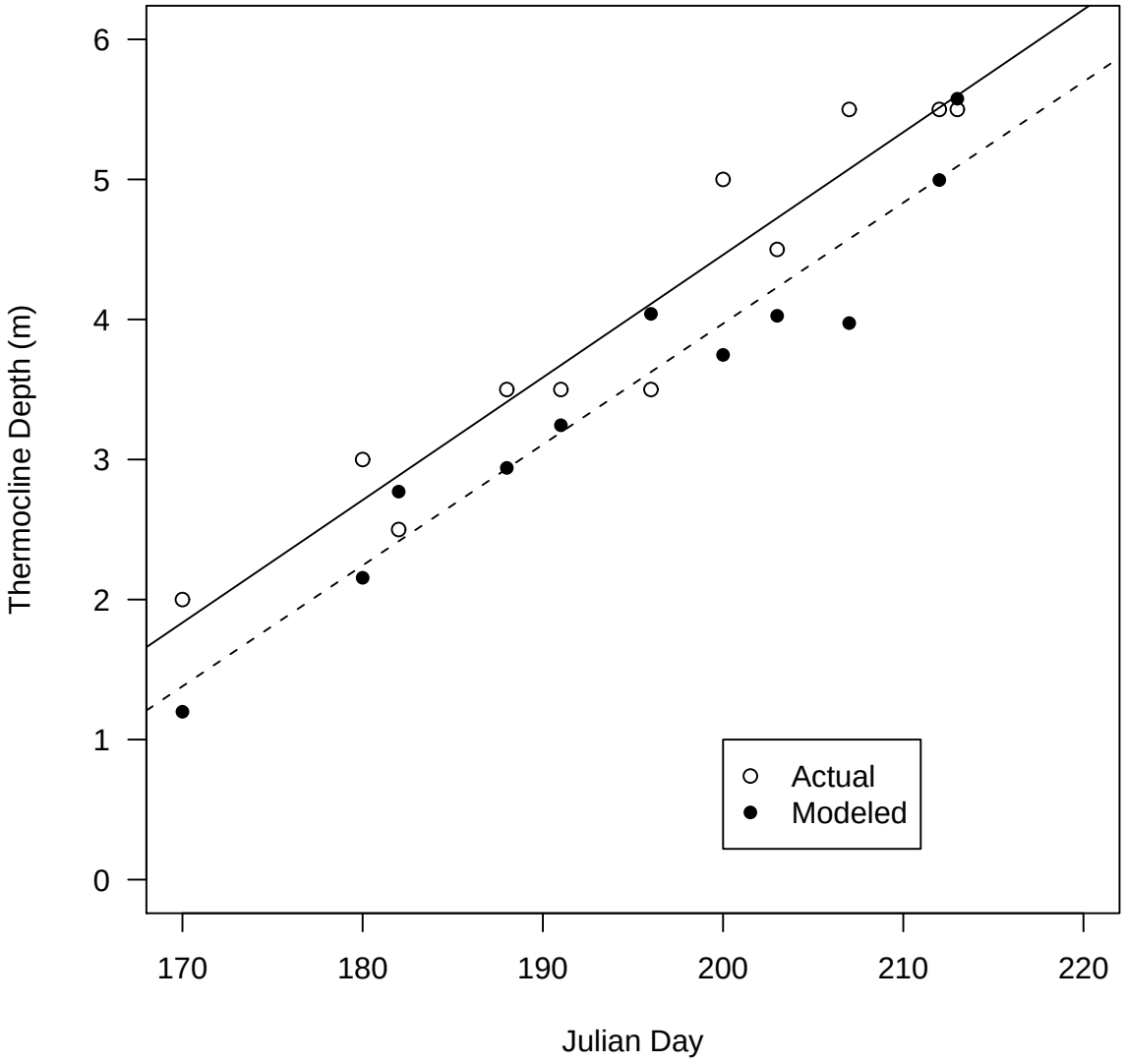

Figure 14: A plot of the temporal variation in actual (open circle, solid line) and modeled (solid circle, dashed line) thermocline depth in lake GTH 91. The lines show the least squares fit to the data.

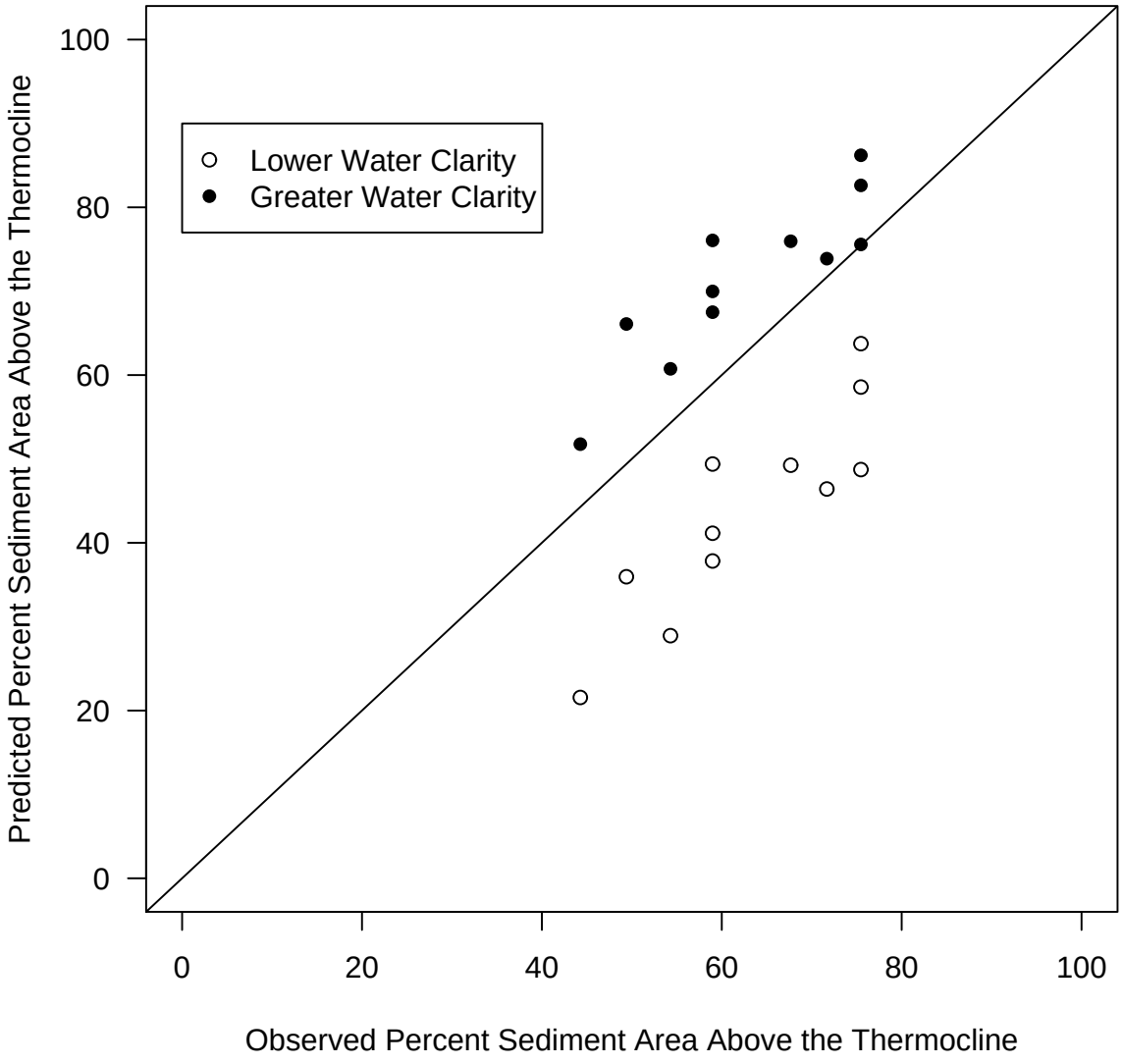

Figure 15: Percent sediment area above the thermocline predicted with either an increase (solid circles) or decrease (open circles) in K_d of 0.5 m^{-1} by the observed percent sediment area above the thermocline. The diagonal line indicates a 1:1 change. Points above the line indicate where the thermocline depth is predicted to increase relative to the current conditions and points below the line indicate where the thermocline depth is predicted to decrease relative to the current conditions.

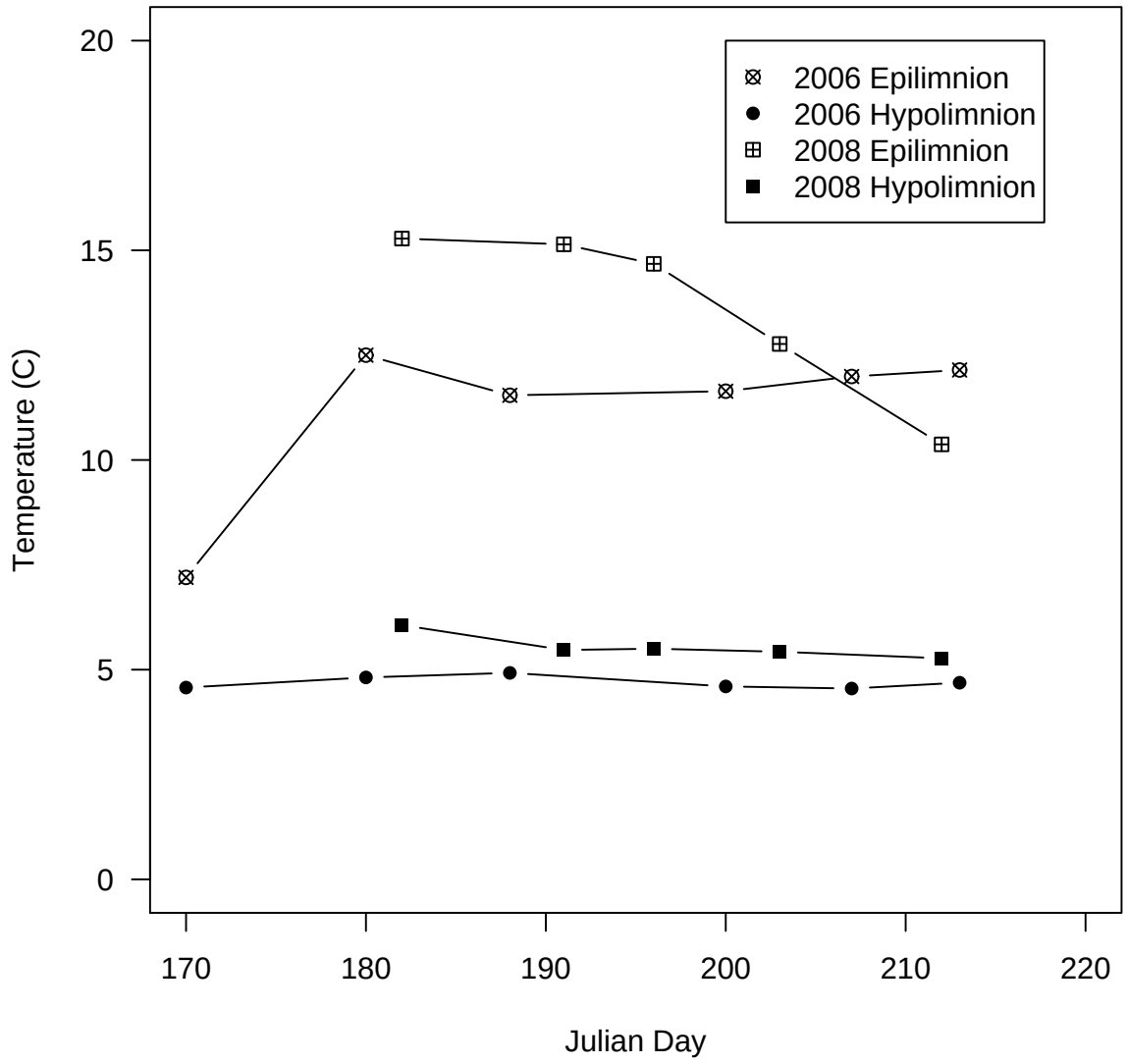

Figure 16: Mean temperature of the epilimnion and hypolimnion of lake GTH 91 during the summers of 2006 and 2008.

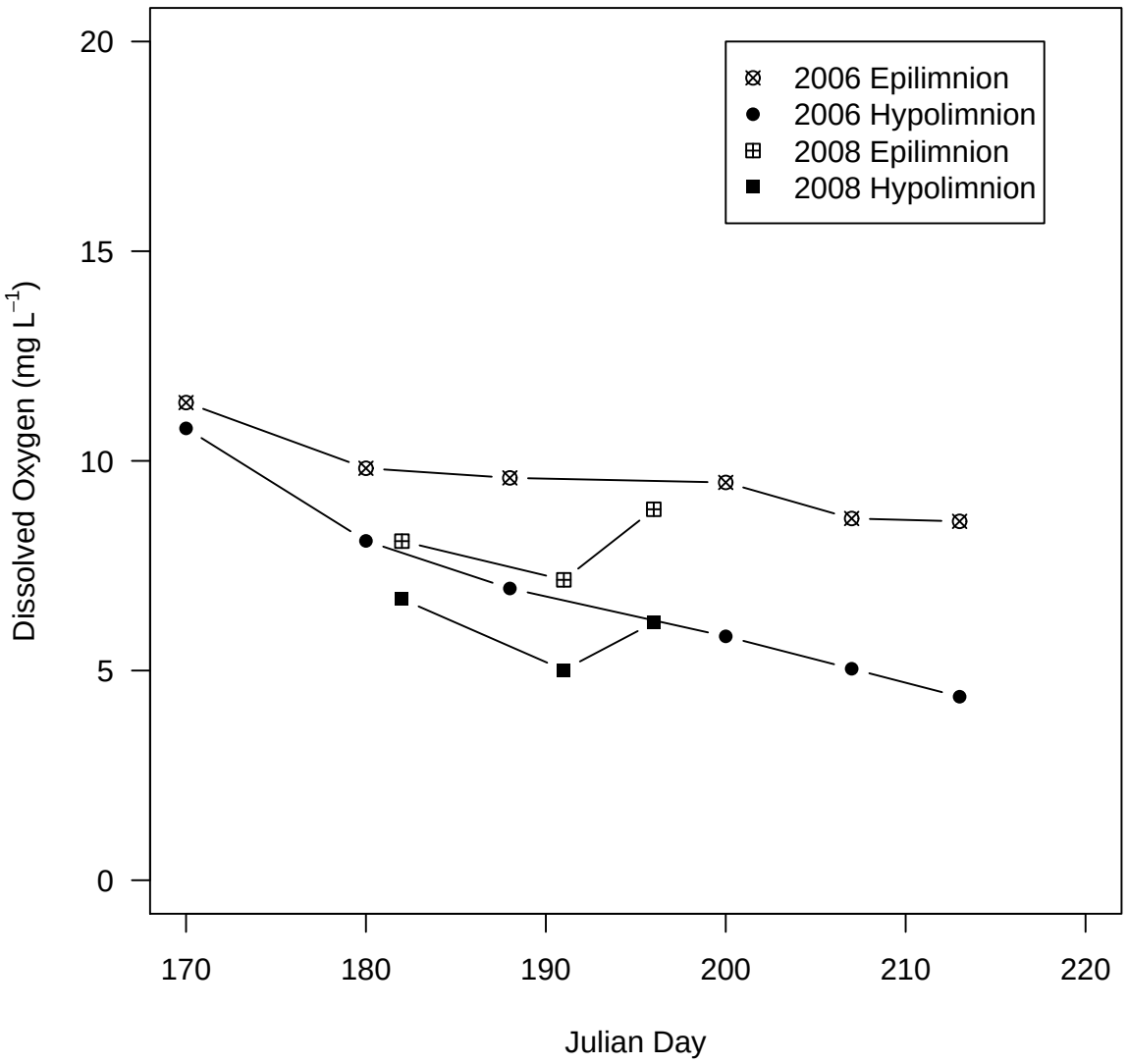

Figure 17: Mean dissolved oxygen concentration of the epilimnion and hypolimnion of lake GTH 91 during the summers of 2006 and 2008.

5 The distribution of sediment organic matter in arctic lakes

5.1 Introduction

The anthropogenic alteration of the global carbon cycle through forest clearing and the burning of fossil fuels has highlighted the need to understand the distribution and fate of organic carbon in the world's ecosystems. Cole et al. (2007) estimate that globally, lakes store between 0.03 and 0.07 Pg of organic carbon per year in their sediments. Thus, despite covering $< 3\%$ of the earth's land surface (Downing et al., 2006) lakes represent hot spots of organic matter storage.

Changes in the amount of organic matter in an ecological system such as a lake result from the balance of organic matter inputs and losses. Gross primary production and detrital import increase the amount of organic matter in the system, while ecosystem respiration, organic matter export, and non-biological oxidation remove organic matter (Lovett et al., 2006). In lake sediments, the losses due to non-biological oxidation and fluvial export are likely minimal. Thus the organic matter content of lake sediments is determined principally by benthic gross primary production plus detrital imports (terrestrial and pelagic), minus total sediment respiration.

In the small shallow systems that dominate the worldwide distribution of lakes (Downing et al., 2006), benthic primary production is often a large component of whole lake primary production (Stanley, 1976a; Vadeboncoeur et al., 2002, 2008; Whalen et al., 2008; Ask et al., 2009; Karlsson et al., 2009). Benthic net primary production can substantially increase the sediment organic matter content of shallow lakes and ponds (Stanley, 1976b). In shallow eutrophic lakes, high phytoplankton production shades the lake bottom and limits benthic primary production (Vadeboncoeur et al., 2003) while simultaneously fu-

eling large phytodetritus exports to the sediments. However meso- to oligotrophic lakes, lacking high phytoplankton settling fluxes, receive most of their organic matter inputs from the settling of organic particles that wash into the lake from the watershed (Molot and Dillon, 1996). Thus in lake ecosystems, the addition of organic matter to the sediments will depend on the relative importance of factors that control the rate of benthic and pelagic primary production and the rate of organic matter input from the watershed.

Light is the principal factor limiting the amount of benthic primary production in lakes (Stanley, 1976b; Bjork-Ramberg, 1983; Hansson, 1992; Vadeboncoeur et al., 2001; Ask et al., 2009; Karlsson et al., 2009) and benthic production is not often directly affected by nutrient inputs to the lake (Bjork-Ramberg, 1983; Vadeboncoeur et al., 2001). The factors controlling the input of organic matter to the sediments are related to the source of the material. Inputs of phytodetritus typically are restricted by the nutrient limitation of the phytoplankton. Fertilization of an oligotrophic lake in the same region as the present study resulted in substantial stimulation of phytoplankton production and the elevated input of phytodetritus to the sediments (O'Brien et al., 2005). The factors controlling the delivery of organic material from the watershed to the lake are more diverse and are related to the soil characteristics and the hydrology of the watershed (Forsberg, 1992; Freeman et al., 2004; Worrall and Burt, 2004; Worrall et al., 2004).

The accumulation of sediment organic matter via the above mechanisms is constantly being countered by heterotrophic respiration of organic substrates. Sediments are a dynamic venue for the mineralization of organic matter, and respiration can greatly reduce the accumulation of sediment organic matter (Stanley, 1976a; Ask et al., 2009). Over geologic time scales only a very small proportion of deposited organic matter will ultimately escape mineralization (Burdige, 2007). However the rate of sediment organic matter decomposition is limited by temperature, the availability of electron acceptors (notably oxygen), and the lability of the organic matter substrate (Capone and Kiene, 1988; Canfield, 1994; Burdige, 2007). The most rapid mineralization rates typically occur under aerobic conditions, on labile substrates, and at warmer temperatures, conditions not found in most lake

sediments (Capone and Kiene, 1988).

Integration of the above processes indicates that the organic matter content of a given sediment sample will reflect its production, deposition and mineralization history. This history is the result of factors that vary across spatial and temporal scales. Understanding the scale at which the factors affecting organic matter storage in lake sediments are operating will allow for more informed predictions regarding how these processes will be affected by natural and anthropogenetically driven variation.

Lakes make up an important component of the ecology and biogeochemistry of the Arctic (Hobbie et al., 1980) and are likely to play a significant role in how arctic ecosystems respond to anticipated climate changes (Tranvik et al., 2009). I surveyed the organic matter content of the upper 10 cm of sediment in the shallow and deep portions of lakes in the Alaskan Arctic. Combining measurements of sediment organic matter content and its loss with sediment depth and measurements of associated environmental variables (e.g., depth, water temperature, irradiance, dissolved oxygen), I evaluate the questions: 1) What factors correlate with the organic matter content of arctic lake sediments and with the loss of organic matter with sediment age? and 2) At what spatial scales do patterns in sediment organic matter content and loss rate manifest?

5.2 Materials and Methods

5.2.1 Core Sampling and Sediment Collection

Sediments were collected using a K–B style gravity corer from 23 lakes in the region of the Toolik Lake Biological Station (Fig. 18). In 2007, all cores were sectioned into 1 cm increments in the field. Each sediment section was homogenized and transferred to a preweighed 20 ml plastic scintillation vial. Two cores each were collected from a single "shallow" and "deep" location in each lake. The relative designations of "shallow" and "deep" refer to samples collected at the shallowest depth with sufficient sediments for coring and the deepest location in the lake. If the shallowest depth suitable for coring and the maximum depth of the lake were similar, only a single sample was collected and was

designated “shallow” or “deep” based on the sample depth relative to the depth of the other lakes in the survey.

In 2008 lakes E–4, S–3 and GTH 91 were sampled in the same manner as the lakes surveyed in 2007 except that the sediments were collected into a 15 ml glass centrifuge tube following slicing. The porewater was extracted from these sediments via centrifugation (1000 or 2000 rpm for 30 min) and the sediments were transferred to glass 20 ml scintillation vials. All sediments were dried at 40 – 60° C for at least 48 h or 105° C for 12 h. The percent organic matter in the sediments was determined as the percent mass lost after 4 h at 550° C (Wetzel and Likens, 2000).

5.2.2 Environmental and Spatial Variables

Depth profiles of temperature and dissolved oxygen were collected using either a YSI Model 85 multiparameter water quality meter or Hydrolab, Data Sonde 5. All profiles began just below the air-water interface and measurements were collected in 0.5 m intervals to the deepest point in the lake. Photosynthetic photon flux density (PPFD) was measured in 0.5 m intervals using a LI-192SA underwater 2π quantum sensor with a Li–Cor LI-250 quantum meter. The percent of the PPFD reaching the sediments at each depth (hereafter, percent surface irradiance) was estimated using the light attenuation coefficient calculated as the slope of the natural log of PPFD with depth. Lake watershed areas were calculated using a digital elevation model and the hydrology toolset in ArcMap GIS software (ESRI, 2006).

The glacial geology of each lake in the survey was determined from the map in Hamilton (2002). Lakes were assigned to either the Itkillik glacial drift (id) or the Sagavanirktok glacial drift (sd). All of the lakes in the Itkillik drift are on the phase II drift which occurred between 25 and 11.5 kyr. Lakes E–2 and E–pond are on the phase I drift which has an age of 120 to 55 kyr. The older Sagavanirktok surface is between 780 and 125 kyr (Hamilton, 2002). Two lakes could not be clearly assigned to one of the above categories. Hamilton (2002) shows lake S–3 on subglacial meltwater deposits associated with the Itkillik drift

so this lake was included with the younger id lakes. Finally lake GTH 110 occurs partially on the sd surface and partially on solifluction deposits (Hamilton, 2002) but was grouped with the sd lakes.

5.2.3 ^{137}Cs Analysis

Sediment accumulation rates were determined for lakes E–4, S–3, and GTH 91 using ^{137}Cs analysis. Two sediment cores were collected from the deepest location in each lake using a K-B style sediment corer. The upper 10 cm of the cores were sectioned in 1 cm intervals as described above. The sediments from each section were homogenized and dried at 40 – 60^{o} C for at least 48 h. The ^{137}Cs analysis was performed at the Department of Environmental Science, Policy, and Geography, University of South Florida. Briefly, the ^{137}Cs activity was measured from the 661.66 keV gamma peak, using an intrinsic germanium detector coupled to a multi-channel analyzer (Princeton Gamma-Tech HPGe). Activity was calculated by multiplying the counts per minute by a factor (determined from standard calibrations) that includes the gamma-ray intensity and detector efficiency. Identical geometry was used for all samples.

The peak ^{137}Cs activity was taken to represent sediments that were deposited in 1963, thus the sediment deposition rate (cm y^{-1}) could be estimated from the amount of sediment above the peak ^{137}Cs activity. The mass accumulation rate (g cm^{-2} y^{-1}) was determined using the product of the mean density of the solid portion of the sediment, the sedimentation rate in cm y^{-1} and unity minus the mean porosity of the core (Berner, 1980). The mean density of the solids in the core was assumed to be 2.0 g cm^{-3}.

5.2.4 Statistics and Calculations

The mean percent organic matter content of the sediments (hereafter, mean percent organic matter) was calculated by averaging the percent organic matter in each sediment slice across the entire 10 cm core. The rate of sediment organic matter loss with sediment depth was estimated by fitting a linear model (least squares) to the change in percent sediment

organic matter with depth. The slope of this relationship (percent organic matter cm^{-1}) was scaled to the age of the sediments (percent organic matter y^{-1}) by multiplying the slope of the percent organic matter by depth relationship by the average sediment accumulation rate (cm y^{-1}) in lakes S–3 and E–4 calculated using the ^{137}Cs dates. The percent organic matter of the sediments at the sediment–water interface (hereafter surface percent organic matter) was defined by the y–intercept of the linear model of percent sediment organic matter and sediment depth.

The differences between the loss of organic matter with age in the shallow and deep samples and all comparisons between the old (sd) and young (id) landscapes was evaluated with a Kruskal Wallis test due to non–homogeneity of variance between the groups. All other comparisons were made using analysis of variance (ANOVA). The relationships between the organic matter variables (i.e., loss of organic matter with age, mean percent organic matter, and surface percent organic matter) and environmental variables (i.e., the lake depth from where the core was collected, percent surface irradiance, water column dissolved oxygen concentration, and water temperature) were explored using pairwise Pearson's correlations. Any comparisons with a correlation coefficient greater than 0.3 were tested for significance.

Mantel tests were used to assess the correlation between the Euclidean distance matrices of the organic matter variables and the geographic distance between the lakes (Euclidean distance in decimal degrees). Differences in the mean and surface percent organic matter as well as the loss of organic matter with age for lakes on the different aged land surfaces (i.e., sd or id) were assessed with separate ANOVAs for each response variable.

For the lakes with sediment oxygen demand data (Lakes E–4, S–3, and GTH 91), I estimate the time required to mineralize the observed difference in organic matter between the uppermost and lowermost slices of the 10 cm cores based on the mean sediment oxygen demand from the temperature dependence experiment described in Chapter 2. I calculated the loss of organic matter mass in each core slice by multiplying the organic matter mass of the uppermost slice by the proportion of the percent organic matter in the surface

slice remaining at each depth in each core. The fraction of the original organic matter remaining was calculated as the percent organic matter at a given depth divided by the percent organic matter in the surface slice. I estimated the days required to mineralize the difference in organic matter between the uppermost and lowermost slices, assuming that the organic matter was either all autochthonous or all allochthonous. In both cases I used a sediment oxygen demand of 15.35 mmol O_2 m^{-2} d^{-1} (i.e., the mean sediment oxygen demand measured in the temperature dependence experiment from Chapter 2). For the assumption of all autochthonous organic matter I used a DIC:O_2 flux ratio of 0.77 and a 106:16:1 C:N:P (Torgersen and Branco, 2007). For the assumption that all of the organic matter was allochthonous, I used a DIC:O_2 flux ratio of 0.98 and a 790:7.6:1 C:N:P (Torgersen and Branco, 2007). All analyses were performed in R (R Development Core Team, 2009)

5.3 Results

Shallow samples were collected from 20 lakes and deep samples were collected from 13 lakes. The mean (± SD) depths of the shallow and deep samples were 2.4 (0.7) and 6.7 (2.9) m respectively. The mean percent organic matter of the sediments ranged from 17.2 to 68.9% and the surface percent organic matter ranged from 15.9 to 76.7%. The surface percent organic matter and the mean percent organic matter of the same sample were nearly perfectly correlated ($r = 0.97$, $p = < 0.001$) (Fig. 19). The average mean or surface percent organic matter did not differ significantly between the shallow and deep samples (Fig. 20). Both the mean and surface percent organic matter showed correlations greater than 0.3 with percent surface irradiance and dissolved oxygen concentration in the water column. Examination of the relationships indicated that the positive correlations were being driven primarily by interactions in the shallow samples. Thus, separate linear models were constructed for the two depths (Fig. 24). In the shallow samples, the dissolved oxygen concentration ranged from 6.3 to 13.3 mg L^{-1} and explained 51% and 42% of the variation in the mean ($F_{1, 9} = 9.32$, $p = 0.01$) and surface ($F_{1, 10} = 7.36$, $p = 0.02$) percent organic

matter respectively. The dissolved oxygen concentration of the deep samples ranged from 0.7 to 8.1 mg L^{-1} and had no relationship with the mean ($F_{1, 8} = 0.09$, $p = 0.77$) or surface ($F_{1, 8} = 0.60$, $p = 0.46$) percent organic matter. The percent surface irradiance of the shallow samples ranged from 0.44 to 57% and explained 55% of the variation in both mean ($F_{1, 10} = 12.24$, $p = 0.006$) and surface ($F_{1, 11} = 13.49$, $p = 0.004$) percent organic matter. (Fig. 24). Very little surface irradiance reached the deep samples (range = 0 to 12%; median = 0.05%) and irradiance had no impact on the variation in deep mean ($F_{1, 10} = 1.78$, $p = 0.212$) or surface ($F_{1, 10} = 4.39$, $p = 0.06$) percent organic matter.

Due to the lack of suitable conditions to collect samples at both shallow and deep locations in all lakes, samples from both depths were collected in only 11 lakes (42% of the total). Variation in the percent organic matter of the deep samples was significantly and positively correlated with variation in the percent organic matter of the shallow samples from the same lake for both the mean ($r = 0.75$, $p = 0.012$) and surface ($r = 0.71$, $p = 0.015$) percent organic matter (Fig. 25).

The percent organic matter loss with age ranged from –0.59 to 0.06% organic matter per year. Sediments showing no significant loss of organic matter with sediment depth occurred in 25% and 38% of the shallow and deep samples respectively. The r^2 of the relationships with significant slopes ranged from 0.22 to 0.96 with a median of 0.67. Percent organic matter loss with age was significantly correlated with the r^2 of the model ($r = –0.54$, $p = 0.001$) and there was no significant difference between the percent organic matter loss with age in the shallow or deep samples ($F_{1, 31} = 3.12$, $p = 0.09$; Fig. 26). The loss of percent organic matter with age was positively and significantly correlated with the dissolved organic carbon (DOC) concentration of the water ($r = 0.46$, $p = 0.046$; Fig. 27). The loss of organic matter with age was negatively correlated with the percent surface irradiance but the correlation was only nearly significant ($r = -0.38$, $p = 0.062$; Fig. 27). The correlation between surface percent organic matter and loss of organic matter with age was not significant ($r = -0.31$, $p = 0.081$; Fig. 27).

The distance matrices based on mean or surface percent organic matter were positively

correlated with the geographic distance between the lakes (Table 18). There was no significant correlation between the distance matrix based on the loss of organic matter with sediment age and the geographic distance between the lakes. Watershed or lake area was not significantly correlated with the mean or surface percent organic matter or the loss of organic matter with sediment age.

In both the shallow and deep samples the mean and surface percent organic matter in the sediments was significantly greater in the lakes on the younger (id) landscape than in in the lakes on the older (sd) landscape (Figs. 21 and 22). There was no significant difference in the loss of percent organic matter with sediment age between the two aged surfaces in either the shallow or deep samples (Fig. 23). The mean (± SD) percent surface irradiance reaching the shallow sediments of the lakes on the younger surface (id) was 24.5% (± 17.0), which was significantly greater than the mean 5.9% (± 4.8) reaching the shallow sediments of the lakes on the older surface ($p = 0.012$).

Assuming an average sediment oxygen consumption rate of 15.35 mmol O_2 m^{-2} d^{-1} and all the organic matter was autochthonous (see Methods), the mean (± SD) time required to mineralize the organic matter lost from the cores was 234 (92), 218 (151), and 173 (90) days in lakes E–4, S–3, and GTH 91 respectively. Assuming the organic matter was all allochthonous produced estimates of 210 (83), 197 (136), and 156 (81) in lakes E–4, S–3, and GTH 91 respectively. The mean sediment accumulation rate calculated from the ^{137}Cs analyses was 52.0 g m^{-2} y^{-1} in lake E–4 and 36.0 g m^{-2} y^{-1} in lake S–3. Using the peak ^{137}Cs activity, the age of the sediments at the base of the 10 cm cores were estimated to be 77 and 120 years for lakes E–4 and S–3 respectively. ^{137}Cs activity was observed down to the base of the 10 cm cores in lakes E–4 and S–3 indicating complete sediment mixing. The peak ^{137}Cs activity was in the uppermost slice of the core from lake GTH 91 making it impossible to determine the age of the core. The mean porosity of lakes E–4 and S–3 was 0.96 (0.02) and 0.97 (0.02) respectively.

5.4 Discussion

There is substantial variation in the mean and surface percent organic matter of the sediments among lakes in the survey. Mean and surface percent organic matter are nearly perfectly correlated and therefore appear to represent two measures of the same sediment property. In the following discussion I refer simply to percent organic matter when both measures produce the same patterns. The range of mean sediment percent organic matter found in my study (17.2 – 68.9%) is toward the high end of the 1 – 62% range of sediment organic matter content observed in subarctic lakes (Granéli, 1978; den Heyer and Kalff, 1998; Aberg et al., 2007) but greater than the 10 – 25% sediment organic matter content reported from other arctic lakes (Livingstone et al., 1958; Cornwell and Kipphut, 1992).

Sediment percent organic matter appears to be primarily a lake-scale property where whole lakes have overall greater (or lesser) sediment percent organic matter than other lakes. There was a significant positive correlation between the percent organic matter of the shallow and deep sediments of an individual lake and depth did not explain any of the variation in percent organic matter variation across lakes. Similarity in percent organic matter content of the sediments was correlated with distance between lakes, suggesting that the factors affecting percent organic matter content vary at a spatial scale greater than that of a lake.

In most lakes, sediment organic matter content is closely tied to factors related to photosynthesis (i.e., benthic primary production) and differences in sediment organic matter appear to result from differences in the amount of light limitation on benthic primary production (Bjork-Ramberg, 1983; Hansson, 1992). Consistent with this concept, I found that sediment percent organic matter was greater in sediments with higher percent surface irradiance and dissolved oxygen in the overlying water but only in the illuminated (i.e., shallow) sediments. The correlations between light, oxygen and organic matter suggest that variation in sediment percent organic matter results from variation in benthic primary production.

The correlation between geographic distance and percent organic matter in the sedi-

ments suggests that there are landscape-scale influences on the factors that are affecting the organic matter content of the sediments. There was significantly greater percent organic matter of the sediments from lakes on the younger landscape surface (id) suggesting that the age of the landscape may influence sediment organic matter. Given that variation in the percent organic matter of the sediments appears to be related to differences in benthic photosynthesis, this finding suggests that lakes on younger landscape surfaces may support higher benthic photosynthesis. Benthic photosynthesis has been shown to be limited buy light availability in the region of this study (Whalen et al., 2006) and the lakes on the younger landscape have significantly greater light penetration to the sediments. Therefore it is likely that differences associated with the age of the watershed influence light availability and therefore benthic primary production. The mechanism driving the differences in light attenuation between the two landscape surfaces is unknown, however Hamilton (2002) identifies significant geological and vegetational differences between the surfaces.

A similar geographic correlation is not seen with differences in the loss of organic matter with sediment age, nor is there a significant difference in the loss of organic matter with age between the different aged land surfaces. These combined results suggest a more local (i.e., within-lake) control of this process. The presence of the landscape-scale patterns in the organic matter content of the lake sediments despite no similar pattern in the loss of organic matter with age suggests that variation in the organic matter content of the lakes is principally driven by organic matter inputs rather than mineralization processes.

The percent organic matter of the sediments below the photic zone appears to reflect benthic primary production in the illuminated portions of the lake. There was a significant positive correlation between the organic matter content of the shallow and deep sediments and in most of the lakes the percent organic matter content of the deep sediments was greater than or nearly equal to the percent organic matter of the shallow sediments. Given the evidence that benthic primary production is a principal source of organic matter to these lakes, these results suggest that organic matter produced in the photic zone is re-

distributed to the deeper portions of the lake. Of the mechanisms that can redistribute sediments within a lake, there are essentially 6 that are likely to affect small shallow systems such as these: intermittent complete lake mixing, epilimnetic mixing, mass wasting from slumping, random sediment redistribution, and wave erosion of littoral sediments (Hilton et al., 1986). Of these mechanisms, all but epilimnetic mixing and random sediment redistribution result in focusing or the movement of sediments from the littoral to deeper portions of the lake (Hilton et al., 1986). I do not have the specific data needed to separate these mechanisms and they all may be occurring to various degrees where sediment organic matter is greater in the deeper sediments. An additional mechanism that is likely to be operating in the stratified lakes is that greater organic matter mineralization in the epilimnetic sediments results in greater loss of organic matter in the shallow portions of the lake. Hilton and Gibbs (1985) found that the differences in the carbon content of sediments at different depths in a shallow lake in England were due to greater summer mineralization rates in the shallow portions of the lake. My findings that sediment organic matter mineralization rates are affected by temperature and the availability of oxygen (Chapters 2 and 3) suggest that the shallow sediments of stratified lakes in the region of this study would have higher organic matter mineralization rates as well.

In lakes with high percent organic matter (approx. $> 50\%$) in the shallow sediments the pattern described above was reversed and the shallow sediments had greater organic matter content than the deep sediments. This difference is likely the result of a build-up of organic matter in the illuminated sediments due to benthic primary production that exceeds the transfer of organic matter to the dark portions of the lake by focusing. It is not clear why these sediments are not redistributed as in the other lakes. One possibility is that the accumulation of benthic algal biomass is sufficient to impede the resuspension of the sediments (Holland et al., 1974; Paterson, 1989).

The majority of shallow (75%) and deep (62%) sediments sampled showed a significant loss of organic matter with sediment depth (i.e, age). The median r^2 of the models is 0.67, indicating that a linear model does an acceptable job of describing these relationships.

Unlike the factors controlling sediment organic matter content, the factors controlling the loss of organic matter from the sediments (i.e., mineralization) appear to vary primarily at a within-lake scale.

There was no correlation between the spatial distance between the lakes and the loss of organic matter with sediment age and the landscape surface did not affect rate of organic matter loss with age in the sediments. In other words, lakes did not cluster according to the loss of organic matter with sediment age at a landscape scale, suggesting that the factors controlling the loss of sediment organic matter with age are operating at smaller spatial scales. As far as I am aware, my study is the only multi-lake evaluation of sediment organic matter loss with age in the Alaskan low arctic, but greater variation in sediment metabolic processes within lakes relative to differences among lakes has been shown before (Hobbie et al., 1980; den Heyer and Kalff, 1998). On a much larger scale, and not analyzed explicitly, Sobek et al. (2009) appear to show no relationship between location and burial efficiency in a global survey of lakes.

The observed percent organic matter of a given lake's sediment reflects both the production (i.e., benthic gross primary production and allochthonous input) and the loss (i.e., community respiration) of organic matter (Lovett et al., 2006). Sediment percent organic matter was significantly correlated with the location of the lake on the landscape so that lakes that were geographically closer together were similar in sediment percent organic matter. This correlation suggests that the factors controlling sediment percent organic matter are varying at the landscape-scale. The loss of sediment percent organic matter with age was not correlated with geographic location, suggesting that the factors controlling organic matter mineralization are not controlled by factors varying at a landscape-scale but vary primarily at the within-lake scale. Since the variation in sediment percent organic matter is not on the same scale as the variation in sediment organic matter mineralization, it appears that sediment organic matter content is primarily a function of organic matter production and not loss. This may be the situation in many small lakes. In shallow arctic ponds Stanley (1976a) found that 97% of benthic net primary production

was buried as detritus and the ponds accumulate 500 to 1000 mg C m^{-2} y^{-1}. Similarly, Ask et al. (2009) found that for 81% of benthic biomass accumulation was derived from autochthonous material in 4 small, oligotrophic, subarctic lakes.

Overall the loss of organic matter with age was more weakly related to the evaluated environmental factors than the percent organic matter of the sediments. The mineralization of sediment organic matter in the lakes in the region appears to be primarily controlled by temperature and the availability of oxygen over short time scales (hours to days; Chapter 2) and the lability of the organic matter over longer time scales (years; Chapter 3). Although my data do not provide detailed analysis of sediment characteristics, the marginally positive relationship between the loss of organic matter with age and the percent irradiance reaching the sediments suggests that sediments produced under high light conditions are more rapidly mineralized. The high mineralization rate appears to be more related to the source and not the quantity of organic matter since the correlation between the loss of organic matter with age and the percent organic matter in the sediments was weaker. The source of the labile organic material is likely benthic algal production (Stanley, 1976a). Therefore the source of the organic matter (in this case, benthic primary production) and not simply the quantity of organic matter has an impact on sediment organic matter storage (Sobek et al., 2009). The mechanism underlying the negative correlation between the DOC concentration of the water and the loss of organic matter with age is less clear but may indicate that greater DOC is associated with lower sediment organic matter lability. Jonsson et al. (2003) found that sediment respiration was positively correlated with DOC concentration in lakes in northern Sweden suggesting that under some circumstances DOC can be associated with greater loss of organic matter from the sediments.

The lack of a significant change in percent organic matter with sediment depth in 25% and 38% of the shallow and deep sediment samples, respectively, is likely the effect of sediment mixing obscuring patterns resulting from mineralization. Lake sediments are subject to mixing from bioturbation (Stanley, 1976b) as well as physical processes (Hilton et al., 1986; Larsen and MacDonald, 1993). Evidence of mixing was observed in the

sediments of two of the shallow lakes and it is likely that the sediments of all the lakes mix to varying degrees. Even in the lakes where significant losses of organic matter with sediment depth were observed (a pattern attributed to mineralization), the sediments with the weakest relationship between sediment depth and percent organic matter (i.e., low r^2) had the smallest losses. Such a pattern would result if the sediments are being partially homogenized due to mixing. The greater proportion of deep samples without significant changes in percent organic matter with depth does not likely reflect the greater potential for these sediments to mix but rather the overall lower rates of organic matter loss with age. These less dramatic patterns would be more easily obscured by smaller amounts of sediment mixing.

The sedimentation rates estimated in the 3 shallow lakes in the study ranged between 36 and 54 g m^{-2} y^{-1} and are between the rate of 27 g m^{-2} y^{-1} estimated for Toolik Lake (Cornwell and Kipphut, 1992) and the range of sedimentation rates (44 – 180 g m^{-2} y^{-1}) observed in other shallow arctic lakes (Hermanson, 1990). There is a large discrepancy between the amount of organic matter found in the sediments and the amount of organic matter predicted to be in the sediments based on the average sediment organic matter mineralization rates measured in Chapter 2 and the age of the sediments. The sediment organic matter mineralization rate measured in Chapter 2 predicts that it would take approximately 200 d to mineralize the organic matter lost between the surface and 10 cm slices of the sediments from lakes E–4, S–3 and GTH 91 regardless of the origin of the organic matter. However, the ^{137}Cs analysis estimates that the sediments at 10 cm in lakes E–4 and S–3 are 77 and 112 years old, suggesting that if the sediments were undisturbed that same amount of organic matter was lost over decades.

The use of the mean mineralization rate from the experiments in Chapter 2 clearly does not accurately model all of the variables affecting the loss of organic matter from the sediments via mineralization. However the magnitude of the discrepancy between the estimated and observed loss of organic matter cannot simply be attributed to the crudeness of the estimate. There was a 41 fold difference between the minimum and maximum min-

eralization rate observed in all of the core incubation experiments in Chapter 2. However there is an approximately 116 fold difference between the observed and predicted time for organic matter loss. Therefore, even assuming the lowest observed mineralization rate, I would predict less organic matter than what is observed in the sediments.

It is likely that the mineralization rates observed in Chapter 2 reflect mainly the aerobic mineralization of freshly deposited (i.e., most labile) material at the sediment-water interface. The mineralization rate under these conditions is likely much greater than the average mineralization rate reflected in the loss of percent organic matter with sediment depth (Burdige, 2007). The 1 cm sediment sampling resolution was likely insufficient to observe the rapid loss of organic matter at the sediment-water interface even if it were not disrupted by sediment mixing. Furthermore, any mixing of the sediments would continually bury labile organic matter into depths with conditions less favorable to rapid mineralization, while at the same time transporting older, more refractory organic matter into the surface sediments where mineralization proceeds more rapidly (Meyers and Ishiwatari, 1993). Therefore the loss of organic matter observed over decadal scales in the core profiles represents the rate that organic matter is lost from the system due to the integrated processes of deposition, mineralization, and mixing.

5.5 Conclusions

The organic matter content of these arctic lake sediments varies primarily at a whole-lake scale and appears to be controlled by the production and not the loss of organic matter from the sediments. Recent work has highlighted the importance of sediment heterotrophic processes in lake carbon cycling (Algesten et al., 2005; Kortelainen et al., 2006), however the net sediment organic matter production and mineralization in these lakes appears to be driven primarily by autotrophic dynamics. Lakes with the greatest light penetration had the highest organic matter content despite the fact that sediments receiving the most light also trended toward the greatest loss of organic matter with sediment age. Furthermore, the positive relationship between sediment percent organic matter and photosynthetic

indicators (i.e., light and oxygen) suggests that differences in sediment organic matter stocks among lakes is due to the accumulation of benthic net primary production. If these patterns are representative of low arctic lakes, then climate mediated changes to arctic lake carbon cycling may be driven by factors that alter benthic primary production, namely lake transparency.

Table 18: Results of Mantel tests of the correlation between the geographic distance between the lakes and the difference between lakes in mean or surface percent organic matter or the loss of organic matter with sediment age. The Shallow samples were collected in the shallowest water with sufficient sediments for coring and the deep samples came from the deepest part of the lake. The Lower CL and the Upper CL are the upper and lower limits of the 95% confidence interval.

Factor	Mantel's r	p	Lower CL	Upper CL
Shallow				
Mean percent organic matter	0.55	0.001	0.46	0.67
Surface percent organic matter	0.51	0.002	0.40	0.65
Loss of organic matter with age	0.07	0.68	0.01	0.17
Deep				
Mean percent organic matter	0.47	0.03	0.26	0.64
Surface percent organic matter	0.62	0.02	0.40	0.77
Loss of organic matter with age	–0.14	0.48	–0.26	-0.002

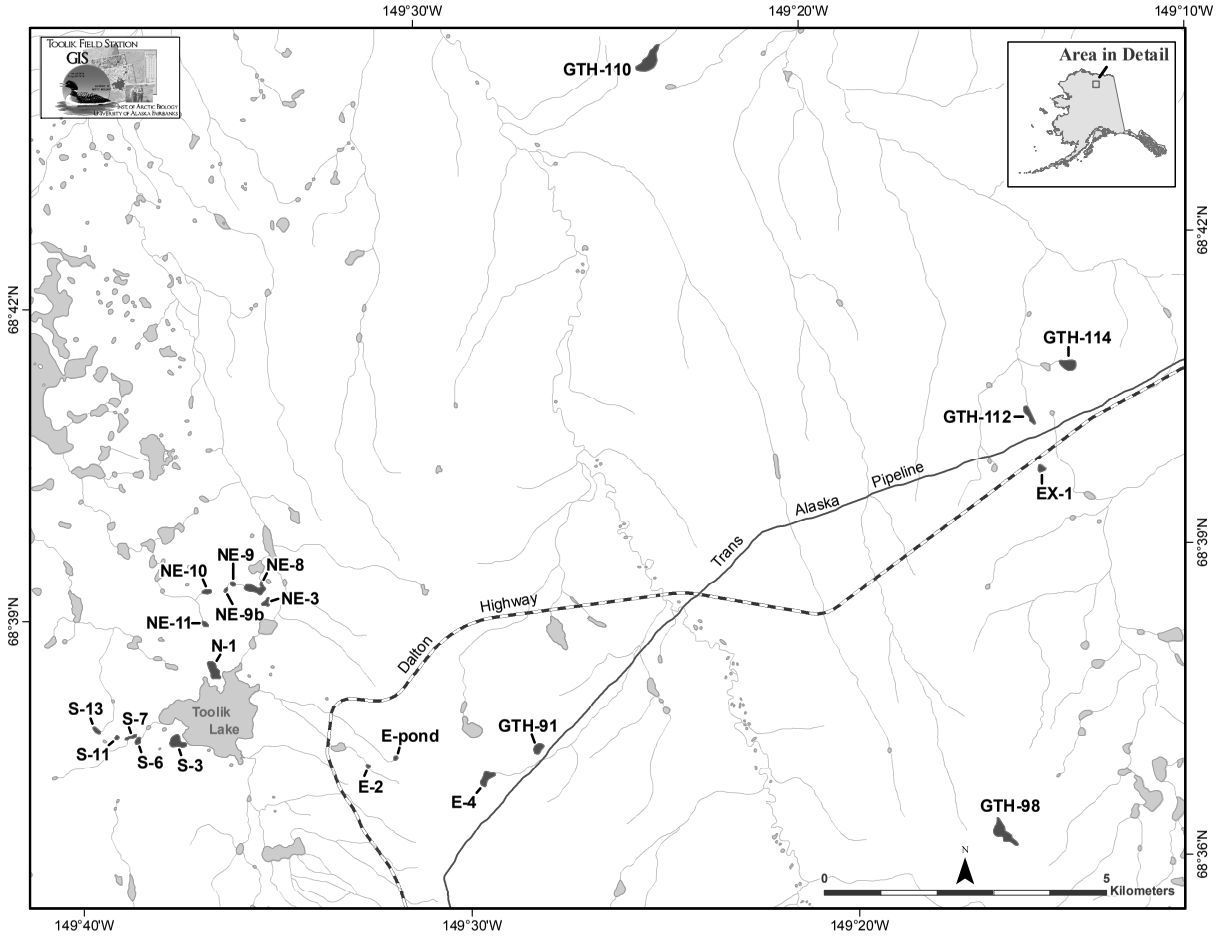

Figure 18: Location of the lakes used in the survey of sediment organic matter content. Lake GTH 156 is not shown but is located approximately 12.9 km north of Toolik Lake.

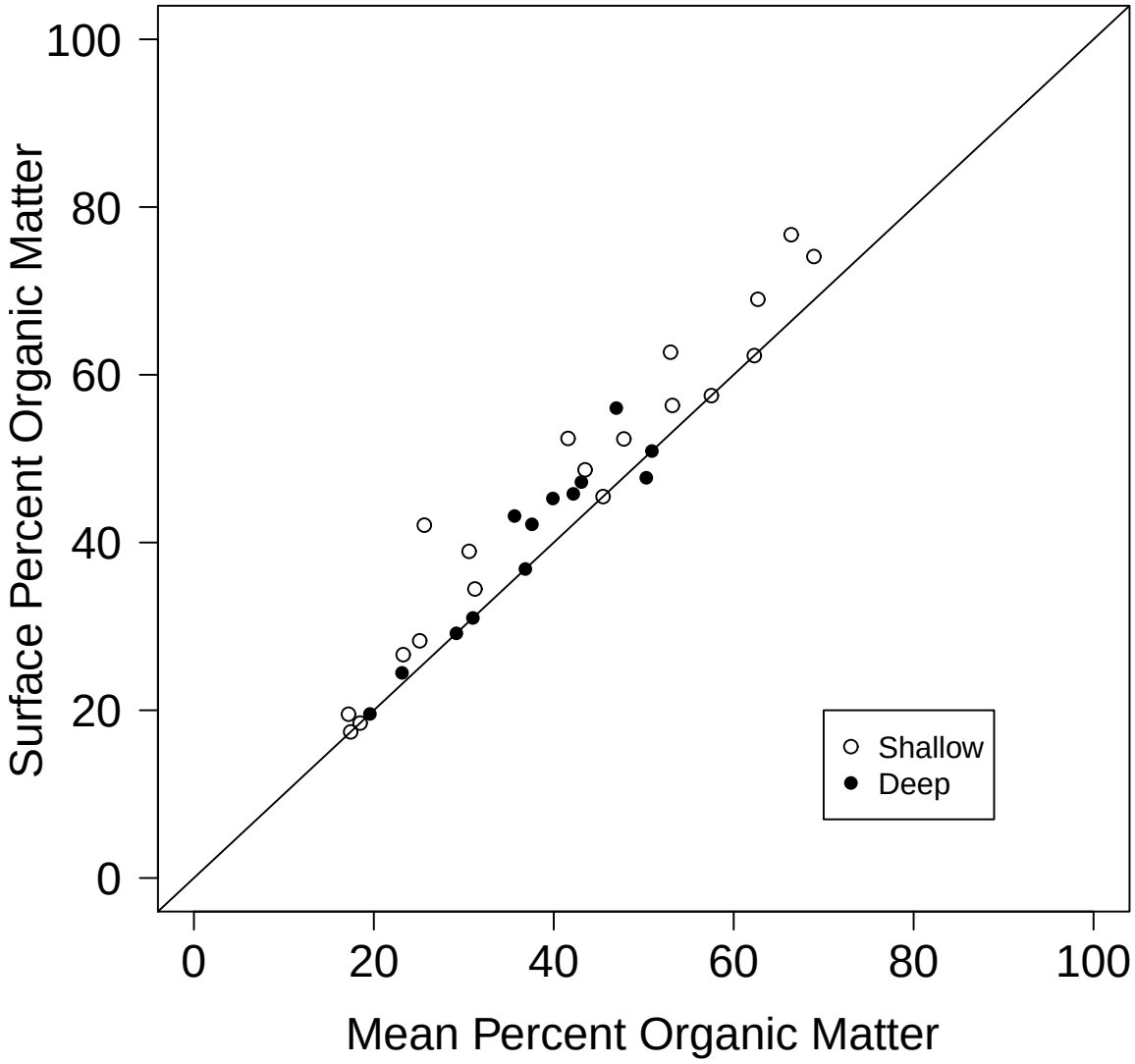

Figure 19: The relationship between the surface percent organic matter in the sediments and the mean percent organic matter in the sediments. The line represents the 1:1 relationship indicating the greater overall percent organic matter content of the surface samples.

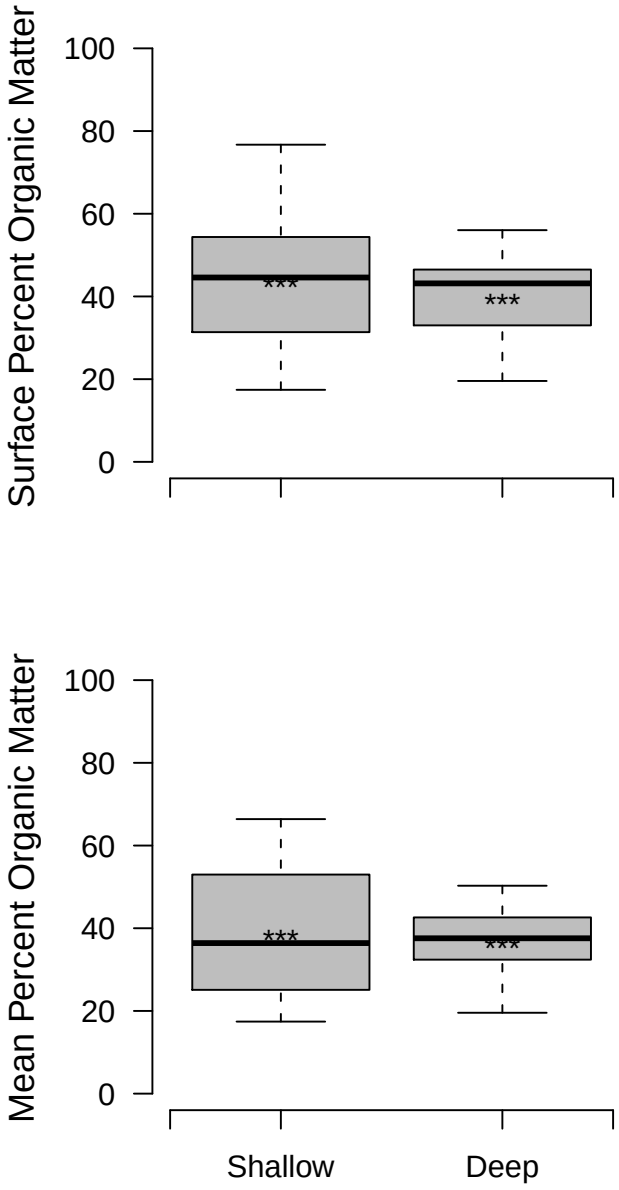

Figure 20: Mean and surface percent organic matter in the shallow and deep portions of the lake. The median value is shown as the horizontal bar within the box and the mean is indicated by "***". The edge of the box indicates the upper and lower quartiles and the whiskers indicate the greatest and least values in the data.

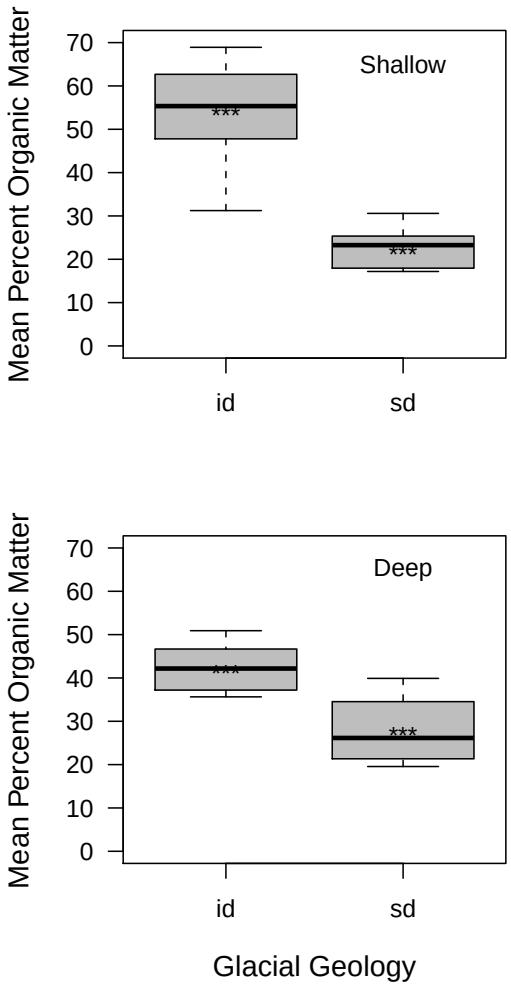

Figure 21: Difference in the mean percent organic matter in the sediments on different glacial drift surfaces. Shallow and deep refer to the relative depth of the sediment samples within the lake. The glacial drift surfaces are defined by Hamilton (2002). Here, id refers to the younger Itkillik drift and sd refers to the older Sagavanirktok drift. The median value is shown as the horizontal bar within the box and the mean is indicated with "***". The edges of the box indicate the upper and lower quartiles and the whiskers indicate the highest and lowest values in the data. Any points exceeding 2 standard deviations of the mean are shown as open circles.

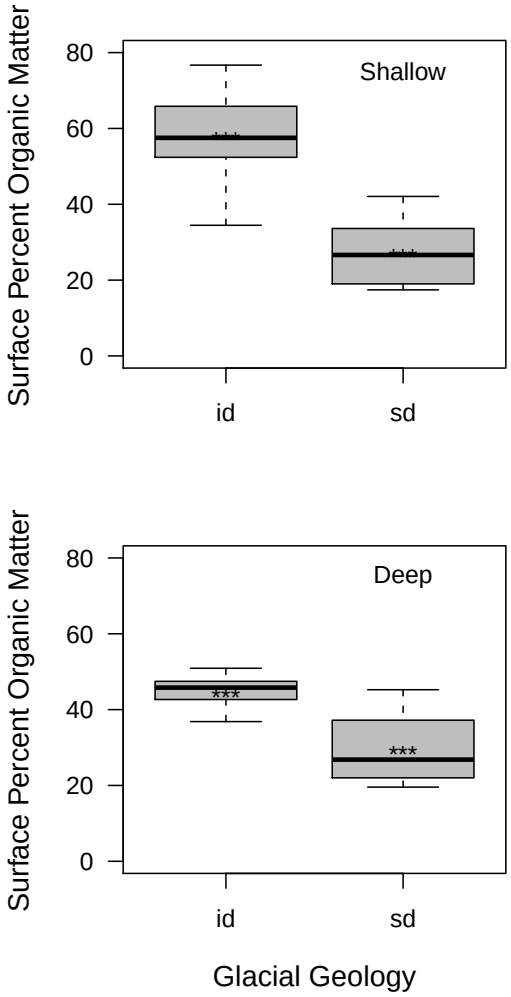

Figure 22: Difference in the surface percent organic matter in the sediments on different glacial drift surfaces. Shallow and deep refer to the relative depth of the sediment samples within the lake. The glacial drift surfaces are defined by Hamilton (2002). Here, id refers to the younger Itkillik drift and sd refers to the older Sagavanirktok drift. The median value is shown as the horizontal bar within the box and the mean is indicated with "***". The edges of the box indicate the upper and lower quartiles and the whiskers indicate the highest and lowest values in the data. Any points exceeding 2 standard deviations of the mean are shown as open circles.

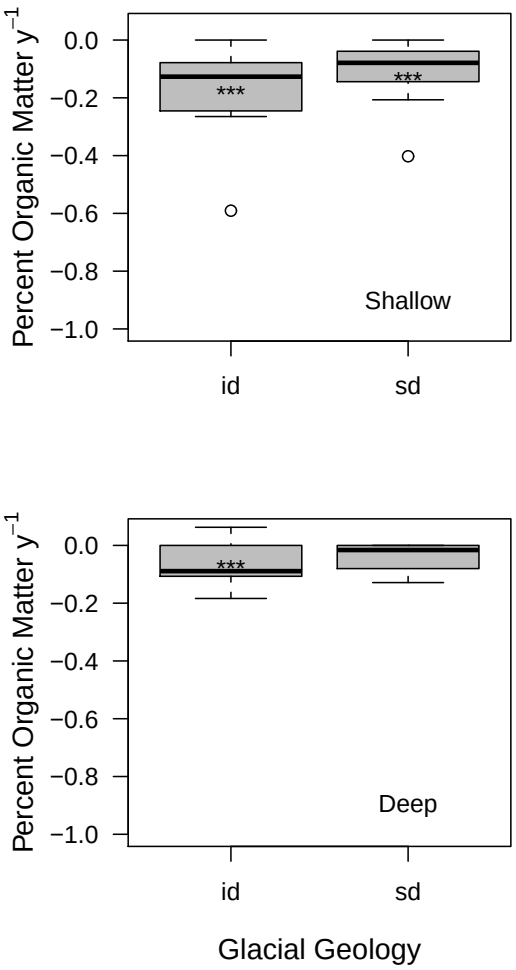

Figure 23: Difference in the loss of percent organic matter with age in the sediments on different glacial drift surfaces. Shallow and deep refer to the relative depth of the sediment samples within the lake. The glacial drift surfaces are defined by Hamilton (2002). Here, id refers to the younger Itkillik drift and sd refers to the older Sagavanirktok drift. The median value is shown as the horizontal bar within the box and the mean is indicated with "***". The edges of the box indicate the upper and lower quartiles and the whiskers indicate the highest and lowest values in the data. Any points exceeding 2 standard deviations of the mean are shown as open circles.

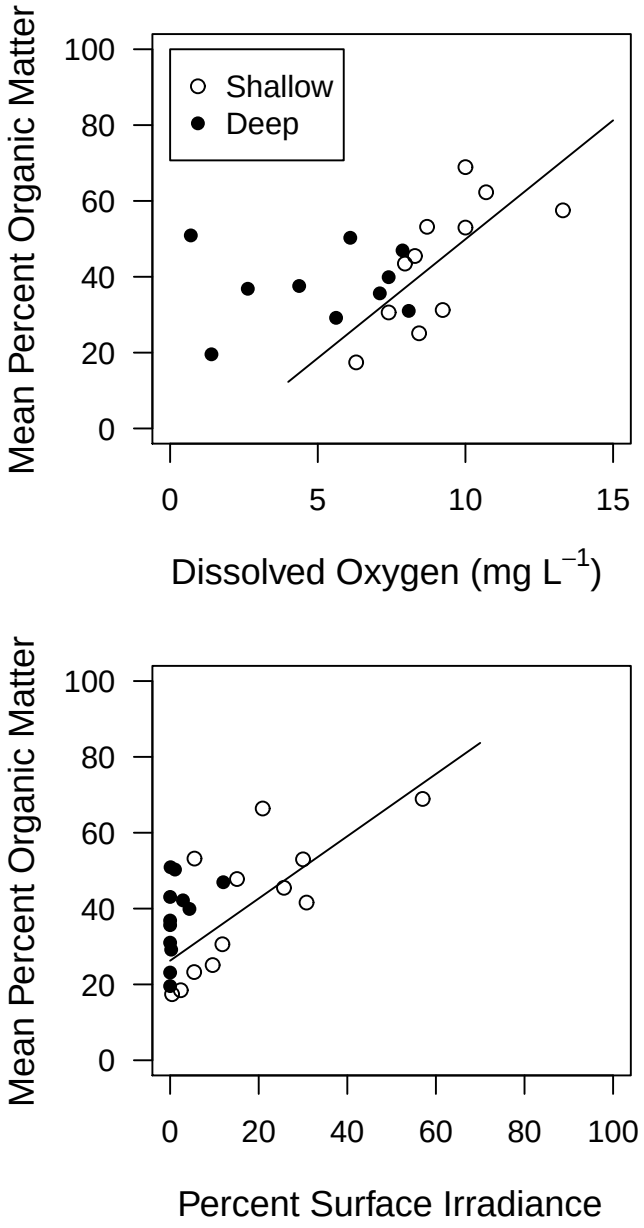

Figure 24: The relationship between the mean or surface percent organic matter in the sediments and the dissolved oxygen concentration of the water overlying the sediments or the percent of surface irradiance reaching the sediments.

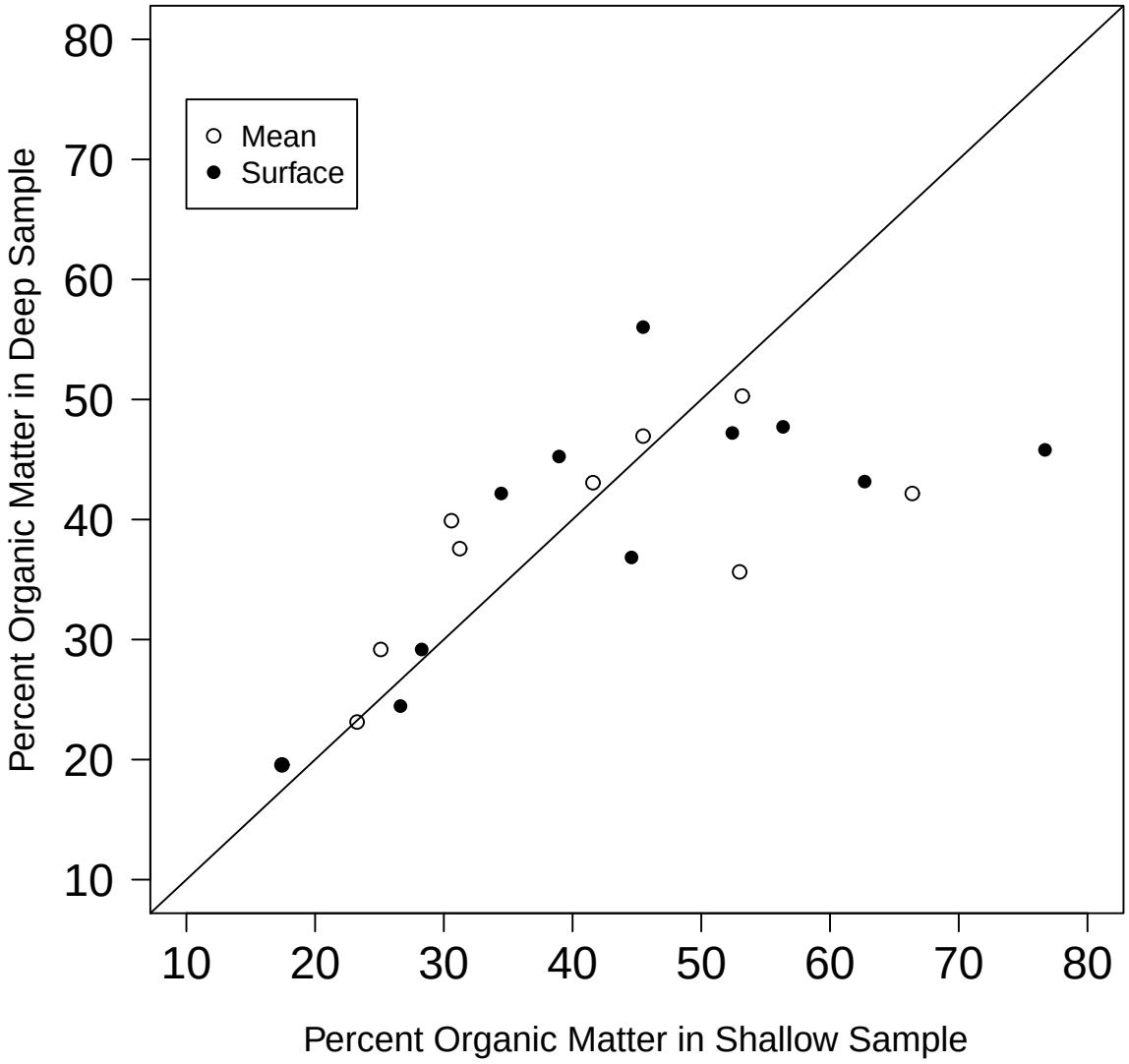

Figure 25: Mean or surface percent organic matter of the "deep" sample by the mean or surface percent organic matter of the "shallow" sample. Each point represents a single lake and the line indicates a 1:1 realtionship.

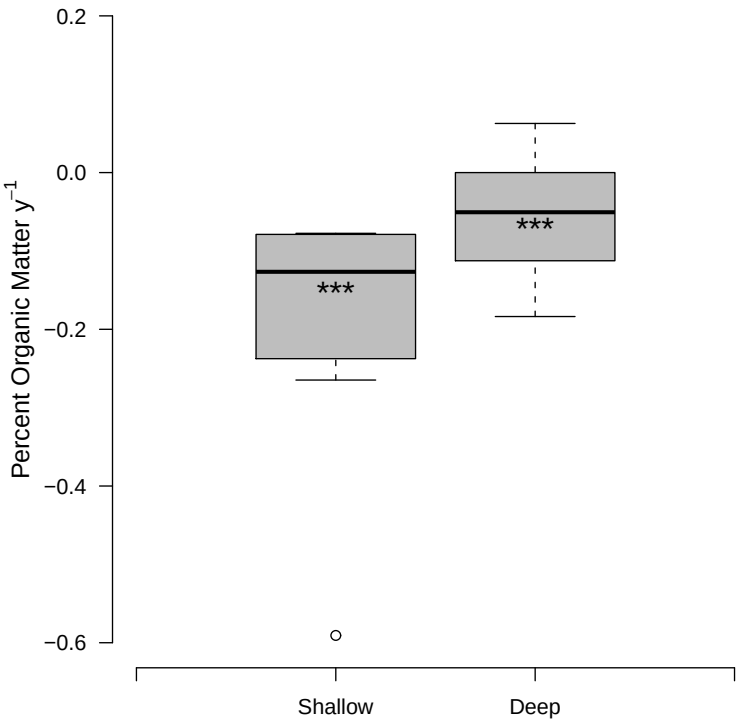

Figure 26: The loss of percent organic matter with sediment age in the shallow and deep portions of the lake. The median values is shown as the horizontal bar within the box and the mean is indicated with "***". The edge the box indicates the upper and lower quartiles and the whiskers indicate the greatest and least values in the data. Any points exceeding 2 standard deviations of the mean are shown as open circles.

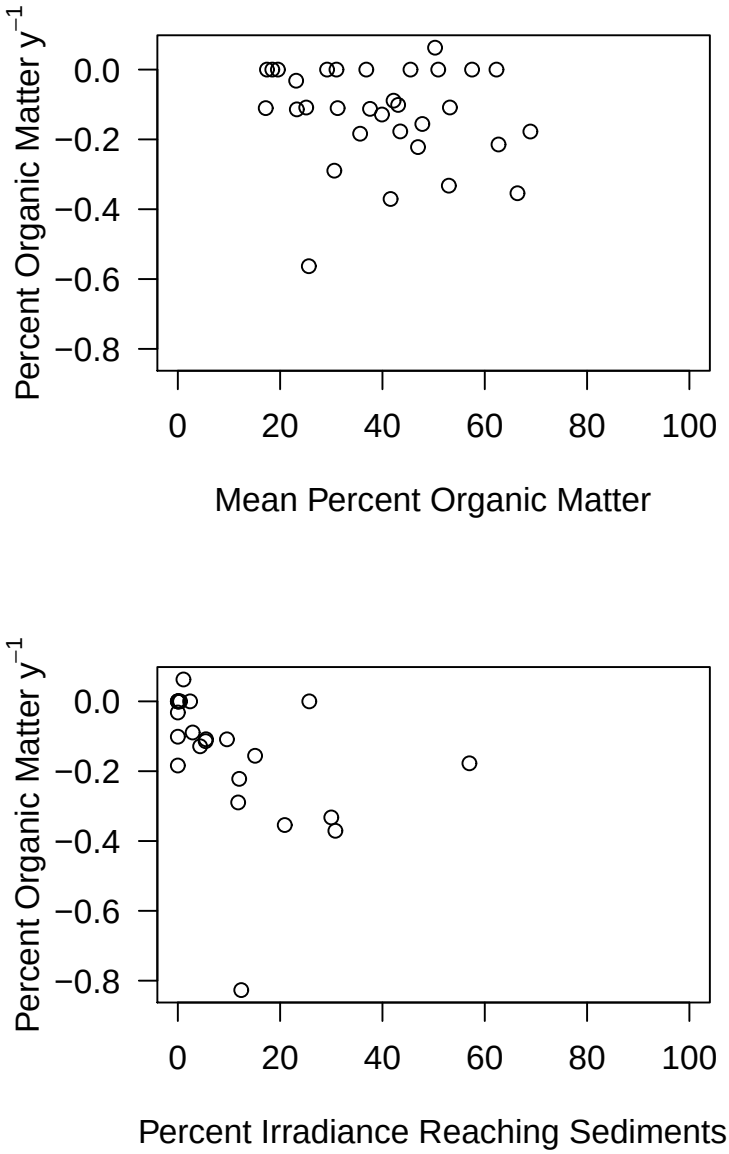

Figure 27: Correlation between the loss of organic matter with age and surface percent organic matter, the percent of surface irradiance reaching the sediments and the dissolved organic carbon (DOC) concentration of the water.

6 Conclusion: The influence of light and organic matter loading on sediment organic matter processing in arctic lakes

The recent realization that lentic sediment accumulation contributes significantly to global carbon cycling (Cole et al., 2007) has highlighted the need to understand the factors affecting the sequestration of organic carbon in lakes. Most lakes simultaneously show a net loss of inorganic carbon through the degassing of CO_2 from the water column, and a net accumulation of organic matter in the sediments (Wetzel, 2001; Cole et al., 2007). Therefore with respect to landscape carbon cycling, lakes function as both a source of inorganic carbon (i.e, CO_2) to the atmosphere and a sink for organic carbon in the sediments. The magnitude of organic matter sequestration is primarily controlled by limitations to sediment respiration (Capone and Kiene, 1988; Canfield, 1994; Burdige, 2007; Sobek et al., 2009).

The results of my dissertation research indicates that the light environment of the lake directly or indirectly alters nearly all of the factors controlling the balance between the production and mineralization of sediment organic matter (Fig. 28). Light transmission was the principal factor related to differences in the percent organic matter of the sediments among lakes. Lakes with the greatest light transmission had sediments with the highest percent organic matter (even in those sediments not actually illuminated). This relationship appears to be due to limitations on epipelic production (i.e., organic matter production) in lakes with less light. Although relationships between irradiance and benthic primary production are well established (Hansson, 1992; Vadeboncoeur et al., 2003; Ask et al., 2009; Karlsson et al., 2009), it is striking that in the lakes I surveyed, whole-lake sediment organic matter content appears to be driven by organic matter production in

surficial sediments and is light limited.

In addition to the effect of light on organic matter production, my results show that light transmission affects the rate of sediment organic matter mineralization through indirect control of oxygen availability, sediment organic matter lability, and temperature. The transmission of light to the sediments indirectly alters the availability of oxygen and the lability of sediment organic matter by limiting benthic photosynthesis (Fig. 28). I found a significant correlation between the rate of organic matter loss from sediments and the amount of light received by those sediments. This is likely due to increases in oxygen availability (Epping and Jörgensen, 1996) and organic matter lability due to primary production (Stanley, 1976b). As I experimentally increased irradiance levels the oxygen deficit of the overlying water produced by mineralization declined due to the production of oxygen by photosynthesis. Furthermore, although I do not have direct evidence of the increase in organic matter lability with increasing light, the sediments collected from a lake with greater light penetration supported faster mineralization rates than those of a similar lake when incubated under identical conditions. This result indicates that the sediments differed in their inherent lability and this difference was correlated with the availability of light.

Light transmission can affect the mineralization rate of sediment organic matter at the whole-lake scale through its influence on the depth of thermal stratification. Among the lakes that I surveyed, those with greater light penetration had significantly deeper epilimnia and a larger proportion of the lake sediments were therefore overlain with epilimnetic water that was relatively warm and capable of exchanging oxygen with the atmosphere. As a result of this relationship, a greater proportion of the sediments in lakes with greater light transmission will be warmer and less susceptible to hypoxia, conditions that will increase sediment organic matter mineralization rates (Hargrave, 1969; Granéli, 1978).

A principal factor affecting the light transmission is the delivery of organic matter from the watershed to the lake (Kirk, 1994; Branco and Kremer, 2005). Although allochthonous inputs can directly add (presumably refractory) organic matter to the sediments (Fig. 28),

DOC loading has been variously shown to correlate with reduced (Algesten et al., 2005) and elevated (Jonsson et al., 2003) sediment organic matter mineralization rates. Nonetheless, my results suggest that for the lakes in my study, allochthonous inputs principally affect sediment organic matter processing through the attenuation of light and the direct and indirect effects described above rather than the direct input of organic matter (Fig. 28).

The combination of the above interactions indicates that although the principal factors affecting sediment organic matter mineralization rate vary at the within-lake scale, the direct and indirect influence of light on the production and mineralization of sediment organic matter suggests that arctic lake organic matter processing is sensitive to changes in light environment. Variation in the factors that control light transmission, such as alterations to arctic watersheds that change the export of light attenuating organic matter from the landscape will alter the way that arctic lakes respond to and feedback into the regional and global carbon cycle.

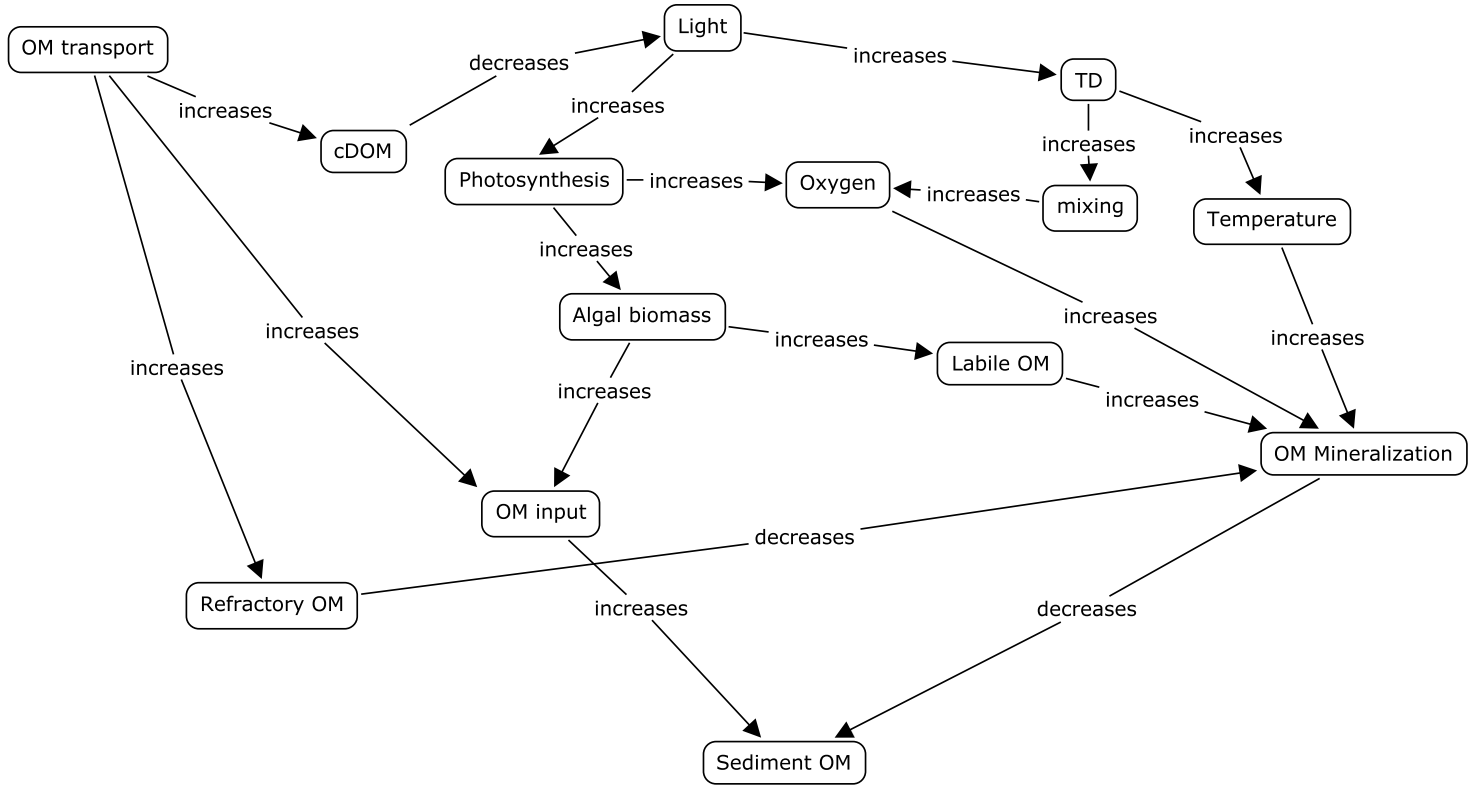

Figure 28: Conceptual model of the effect of light on sediment organic matter processing. OM refers to organic matter, TD is thermocline depth, and cDOM is chromophoric dissolved organic matter.

Appendices

A Temperature Experiment Data

Table 19: Data collected during the experiment evaluating the effect of temperature on sediment oxygen demand. Lake is the lake from which the sediment cores were collected. Date is the date of the core collection. Core is the specific ID of the core tube used in the incubation. uE is the irradiance in μE m^{-1} s^{-1}. Depth.m is the depth in the lake from which the cores were collected in m. Temp.C is the temperature of the incubation ($\pm$ 1° C). T0.O2.mgL is the inital oxygen concentration in the core in mg O_2 L^{-1}. T0.O2.mmolL is the initial oxygen concentration of the core in mmol O_2 L^{-1}. Incub.h is the duration of the incubation in h. Water.Vol.ml is the volume of the water overlying the sediments in ml. T1.flux is the flux of oxygen into the sediments between T0 and T1 in mmol O_2 m^{-2} d^{-1}. T2.flux is the flux of oxygen into the sediments between T1 and T2 in mmol O_2 m^{-2} d^{-1}. sum.flux is the flux of oxygen into the sediments between T0 and T2 in mmol O_2 m^{-2} d^{-1} calculated by adding the change in O_2 between T0 and T1 to the change in O_2 between T1 and T2.

Lake	Date	Core	uE	Depth.m	Temp.C	T0.O2.mgL	T0.O2.mmolL	Incub.h	Water.Vol.ml	T1.flux	T2.flux	sum.flux
E4	2-Aug-06	Y1	0	4	7	8.00e+00	2.50e-01	1.22e+01	1.37e+02	1.55e+01	8.30e+00	1.04e+01
E4	2-Aug-06	Y6	0	4	7	8.78e+00	2.74e-01	1.22e+01	1.46e+02	1.34e+01	1.18e+01	1.23e+01
E4	2-Aug-06	Y9	0	4	7	7.82e+00	2.44e-01	1.22e+01	1.37e+02	4.41e+00	1.37e+01	1.10e+01
E4	2-Aug-06	Y4	0	4	7	7.91e+00	2.47e-01	1.22e+01	1.37e+02	3.41e+00	1.20e+01	9.51e+00
E4	2-Aug-06	Y2	0	4	7	8.21e+00	2.56e-01	1.22e+01	1.37e+02	1.89e+01	1.12e+01	1.35e+01
E4	2-Aug-06	Y8	0	4	7	8.23e+00	2.57e-01	1.22e+01	1.37e+02	1.73e+01	1.14e+01	1.31e+01
E4	2-Aug-06	Y3	0	4	7	8.32e+00	2.60e-01	1.22e+01	1.37e+02	1.88e+01	8.09e+00	1.12e+01
E4	2-Aug-06	Y7	0	4	7	1.04e+01	3.27e-01	1.22e+01	1.55e+02	5.93e+01	1.00e+01	2.45e+01

Data collected during the experiment evaluating the effect of temperature on sediment oxygen demand.

Cont.

Lake	Date	Core	uE	Depth.m	Temp.C	T0.O2.mgL	T0.O2.mmolL	Incub.h	Water.Vol.ml	T1.flux	T2.flux	sum.flux
E4	2-Aug-06	Y5	0	4	7	8.88e+00	2.78e-01	1.22e+01	1.37e+02	3.63e+00	1.80e+01	1.38e+01
E4	2-Aug-06	B9	0	4	2	8.36e+00	2.61e-01	1.22e+01	1.46e+02	7.22e+00	1.36e+01	1.17e+01
E4	2-Aug-06	B5	0	4	2	9.73e+00	3.04e-01	1.22e+01	1.37e+02	4.33e+01	1.05e+01	2.01e+01
E4	2-Aug-06	B8	0	4	2	8.87e+00	2.77e-01	1.22e+01	1.28e+02	1.45e+01	-1.07e+01	-3.31e+00
E4	2-Aug-06	B4	0	4	2	7.41e+00	2.31e-01	1.22e+01	1.46e+02	2.14e+01	-5.48e+00	2.43e+00
E4	2-Aug-06	B6	0	4	2	7.21e+00	2.25e-01	1.22e+01	1.37e+02	5.72e+00	7.39e+00	6.90e+00
E4	2-Aug-06	B2	0	4	2	7.29e+00	2.28e-01	1.22e+01	1.37e+02	1.69e+01	9.18e+00	1.14e+01
E4	2-Aug-06	B3	0	4	2	7.34e+00	2.29e-01	1.22e+01	1.37e+02	5.53e+00	4.41e+00	4.74e+00
E4	2-Aug-06	B7	0	4	2	8.19e+00	2.56e-01	1.22e+01	1.46e+02	9.31e+00	1.35e+01	1.23e+01
E4	2-Aug-06	B1	0	4	2	7.51e+00	2.35e-01	1.22e+01	1.46e+02	1.59e+00	1.09e+01	8.17e+00
E4	4-Aug-06	B5	0	4	12	8.74e+00	2.73e-01	1.31e+01	1.37e+02	1.97e+01	1.38e+01	1.54e+01
E4	4-Aug-06	B6	0	4	12	8.00e+00	2.50e-01	1.31e+01	1.55e+02	4.58e+00	1.03e+01	8.74e+00
E4	4-Aug-06	B3	0	4	12	8.19e+00	2.56e-01	1.31e+01	1.37e+02	-5.52e+00	2.76e+01	1.85e+01
E4	4-Aug-06	B4	0	4	12	7.67e+00	2.40e-01	1.31e+01	1.37e+02	1.07e+01	1.37e+01	1.29e+01
E4	4-Aug-06	B8	0	4	12	7.80e+00	2.44e-01	1.31e+01	1.37e+02	1.36e+01	1.14e+01	1.20e+01
E4	4-Aug-06	B2	0	4	12	7.90e+00	2.47e-01	1.31e+01	1.37e+02	5.12e+00	1.80e+01	1.45e+01
E4	4-Aug-06	B9	0	4	12	7.59e+00	2.37e-01	1.31e+01	1.46e+02	7.55e+00	1.05e+01	9.73e+00
E4	4-Aug-06	B1	0	4	12	8.63e+00	2.70e-01	1.31e+01	1.37e+02	3.98e+01	1.76e+01	2.37e+01
E4	4-Aug-06	B7	0	4	12	7.99e+00	2.50e-01	1.31e+01	1.37e+02	9.65e+00	1.05e+01	1.03e+01

Data collected during the experiment evaluating the effect of temperature on sediment oxygen demand.

Cont.

Lake	Date	Core	uE	Depth.m	Temp.C	T0.O2.mgL	T0.O2.mmolL	Incub.h	Water.Vol.ml	T1.flux	T2.flux	sum.flux
E4	4-Aug-06	Y3	0	4	17	7.84e+00	2.45e-01	1.31e+01	1.46e+02	2.14e+01	1.47e+01	1.65e+01
E4	4-Aug-06	Y1	0	4	17	7.00e+00	2.19e-01	1.31e+01	1.37e+02	2.29e+01	1.14e+01	1.45e+01
E4	4-Aug-06	Y8	0	4	17	8.29e+00	2.59e-01	1.31e+01	1.37e+02	5.32e+01	3.36e+01	3.90e+01
E4	4-Aug-06	Y6	0	4	17	6.93e+00	2.17e-01	1.31e+01	1.37e+02	1.57e+01	1.20e+01	1.30e+01
E4	4-Aug-06	Y5	0	4	17	7.77e+00	2.43e-01	1.31e+01	1.37e+02	3.21e+01	1.81e+01	2.19e+01
E4	4-Aug-06	Y2	0	4	17	6.81e+00	2.13e-01	1.31e+01	1.37e+02	1.57e+02	-4.05e+01	1.36e+01
E4	4-Aug-06	Y4	0	4	17	7.34e+00	2.29e-01	1.31e+01	1.37e+02	1.61e+01	1.50e+01	1.53e+01
E4	4-Aug-06	Y9	0	4	17	7.29e+00	2.28e-01	1.31e+01	1.37e+02	2.71e+01	1.42e+01	1.77e+01
E4	4-Aug-06	Y7	0	4	17	6.54e+00	2.04e-01	1.31e+01	1.46e+02	1.14e+01	1.18e+01	1.17e+01
GTH91	19-Jun-06	B6	0	4	12	9.05e+00	2.83e-01	1.20e+01	1.59e+02	1.25e+01	1.60e+01	1.41e+01
GTH91	19-Jun-06	B8	0	4	12	9.10e+00	2.85e-01	1.20e+01	1.46e+02	2.58e+01	2.06e+01	2.34e+01
GTH91	19-Jun-06	B7	0	4	12	8.13e+00	2.54e-01	1.20e+01	1.60e+02	1.19e+01	9.07e+00	1.06e+01
GTH91	19-Jun-06	B3	0	4	12	9.95e+00	3.11e-01	1.20e+01	1.66e+02	2.63e+01	1.05e+01	1.90e+01
GTH91	19-Jun-06	B1	0	4	12	9.52e+00	2.97e-01	1.20e+01	1.64e+02	1.82e+01	2.07e+01	1.94e+01
GTH91	19-Jun-06	B9	0	4	12	9.61e+00	3.00e-01	1.20e+01	1.66e+02	1.63e+01	2.05e+01	1.82e+01
GTH91	19-Jun-06	B5	0	4	12	9.27e+00	2.90e-01	1.20e+01	1.57e+02	2.01e+01	1.20e+01	1.64e+01
GTH91	19-Jun-06	B4	0	4	12	8.92e+00	2.79e-01	1.20e+01	1.66e+02	1.95e+01	1.41e+01	1.70e+01
GTH91	19-Jun-06	B2	0	4	12	9.37e+00	2.93e-01	1.20e+01	1.73e+02	2.04e+01	1.94e+01	1.99e+01
GTH91	19-Jun-06	B10	0	4	12	9.64e+00	3.01e-01	1.20e+01	1.73e+02	2.57e+01	2.53e+01	2.55e+01

Data collected during the experiment evaluating the effect of temperature on sediment oxygen demand.

Cont.

Lake	Date	Core	uE	Depth.m	Temp.C	T0.O2.mgL	T0.O2.mmolL	Incub.h	Water.Vol.ml	T1.flux	T2.flux	sum.flux
GTH91	19-Jun-06	Y1	0	4	17	8.54e+00	2.67e-01	1.20e+01	1.64e+02	2.75e+01	1.14e+01	2.00e+01
GTH91	19-Jun-06	Y8	0	4	17	8.97e+00	2.80e-01	1.20e+01	1.60e+02	3.20e+01	1.34e+01	2.34e+01
GTH91	19-Jun-06	Y5	0	4	17	8.48e+00	2.65e-01	1.20e+01	1.57e+02	2.68e+01	1.50e+01	2.13e+01
GTH91	19-Jun-06	Y3	0	4	17	8.46e+00	2.64e-01	1.20e+01	1.68e+02	2.75e+01	1.04e+01	1.95e+01
GTH91	19-Jun-06	Y7	0	4	17	8.18e+00	2.56e-01	1.20e+01	1.68e+02	2.08e+01	6.70e+00	1.43e+01
GTH91	19-Jun-06	Y10	0	4	17	7.76e+00	2.43e-01	1.20e+01	1.55e+02	1.96e+01	1.53e+01	1.76e+01
GTH91	19-Jun-06	Y6	0	4	17	8.64e+00	2.70e-01	1.20e+01	1.60e+02	2.02e+01	1.25e+01	1.66e+01
GTH91	19-Jun-06	Y4	0	4	17	8.56e+00	2.67e-01	1.20e+01	1.55e+02	2.36e+01	7.76e+00	1.63e+01
GTH91	19-Jun-06	Y2	0	4	17	8.41e+00	2.63e-01	1.20e+01	1.60e+02	2.46e+01	1.85e+01	2.18e+01
GTH91	19-Jun-06	Y9	0	4	17	8.00e+00	2.50e-01	1.20e+01	1.59e+02	2.60e+01	1.44e+01	2.06e+01
GTH91	21-Jun-06	B3	0	4	2	9.45e+00	2.95e-01	6.92e+00	1.60e+02	-1.45e-01	1.57e+01	4.46e+00
GTH91	21-Jun-06	B9	0	4	2	9.29e+00	2.90e-01	6.92e+00	1.62e+02	1.35e+01	1.45e+01	1.38e+01
GTH91	21-Jun-06	B2	0	4	2	1.05e+01	3.27e-01	6.92e+00	1.64e+02	1.16e+01	2.37e+01	1.51e+01
GTH91	21-Jun-06	B4	0	4	2	9.41e+00	2.94e-01	6.92e+00	1.64e+02	1.27e+01	-6.76e-01	8.84e+00
GTH91	21-Jun-06	B8	0	4	2	8.93e+00	2.79e-01	6.92e+00	1.50e+02	3.75e+00	9.11e+00	5.30e+00
GTH91	21-Jun-06	B10	0	4	2	9.12e+00	2.85e-01	6.92e+00	1.68e+02	1.28e+01	9.96e+00	1.20e+01
GTH91	21-Jun-06	B1	0	4	2	9.53e+00	2.98e-01	6.92e+00	1.66e+02	1.02e+01	1.86e+01	1.26e+01
GTH91	21-Jun-06	B6	0	4	2	9.43e+00	2.95e-01	6.92e+00	1.62e+02	1.52e+01	6.55e+00	1.27e+01
GTH91	21-Jun-06	B7	0	4	2	9.64e+00	3.01e-01	6.92e+00	1.66e+02	1.46e+01	1.68e+00	1.09e+01

Data collected during the experiment evaluating the effect of temperature on sediment oxygen demand.

Cont.

Lake	Date	Core	uE	Depth.m	Temp.C	T0.O2.mgL	T0.O2.mmolL	Incub.h	Water.Vol.ml	T1.flux	T2.flux	sum.flux
GTH91	21-Jun-06	B5	0	4	2	8.10e+00	2.53e-01	6.92e+00	1.69e+02	1.05e+01	1.36e+01	1.14e+01
GTH91	21-Jun-06	Y5	0	4	9	9.00e+00	2.81e-01	6.92e+00	1.64e+02	1.20e+01	1.57e+01	1.31e+01
GTH91	21-Jun-06	Y6	0	4	9	9.19e+00	2.87e-01	6.92e+00	1.64e+02	1.49e+01	1.28e+01	1.43e+01
GTH91	21-Jun-06	Y2	0	4	9	8.36e+00	2.61e-01	6.92e+00	1.62e+02	1.76e+01	8.44e+00	1.50e+01
GTH91	21-Jun-06	Y7	0	4	9	8.69e+00	2.72e-01	6.92e+00	1.66e+02	1.77e+01	1.21e+01	1.61e+01
GTH91	21-Jun-06	Y3	0	4	9	9.61e+00	3.00e-01	6.92e+00	1.64e+02	1.56e+01	1.05e+01	1.42e+01
GTH91	21-Jun-06	Y1	0	4	9	7.99e+00	2.50e-01	6.92e+00	1.57e+02	1.67e+01	1.03e+01	1.49e+01
GTH91	21-Jun-06	Y8	0	4	9	8.09e+00	2.53e-01	6.92e+00	1.64e+02	1.63e+01	1.52e+01	1.60e+01
GTH91	21-Jun-06	Y4	0	4	9	8.81e+00	2.75e-01	6.92e+00	1.59e+02	2.20e+01	-2.20e-01	1.55e+01
GTH91	21-Jun-06	Y9	0	4	9	8.23e+00	2.57e-01	6.92e+00	1.57e+02	1.81e+01	7.73e+00	1.51e+01
GTH91	21-Jun-06	Y10	0	4	9	8.24e+00	2.57e-01	6.92e+00	1.64e+02	1.18e+01	1.47e+01	1.27e+01
S3	8-Aug-06	B2	0	4	12	8.95e+00	2.80e-01	1.26e+01	1.37e+02	1.47e+01	1.64e+01	1.59e+01
S3	8-Aug-06	B1	0	4	12	9.09e+00	2.84e-01	1.26e+01	1.37e+02	2.38e+01	1.52e+01	1.79e+01
S3	8-Aug-06	B5	0	4	12	9.00e+00	2.81e-01	1.26e+01	1.37e+02	3.29e+01	1.45e+01	2.03e+01
S3	8-Aug-06	B7	0	4	12	8.76e+00	2.74e-01	1.26e+01	1.37e+02	1.53e+01	1.45e+01	1.48e+01
S3	8-Aug-06	B3	0	4	12	8.58e+00	2.68e-01	1.26e+01	1.46e+02	1.92e+01	1.54e+01	1.66e+01
S3	8-Aug-06	B9	0	4	12	9.19e+00	2.87e-01	1.26e+01	1.46e+02	2.18e+01	1.59e+01	1.77e+01
S3	8-Aug-06	B4	0	4	12	8.77e+00	2.74e-01	1.26e+01	1.37e+02	1.95e+01	2.11e+01	2.06e+01
S3	8-Aug-06	B8	0	4	12	8.23e+00	2.57e-01	1.26e+01	1.28e+02	1.61e+01	1.55e+01	1.57e+01

Data collected during the experiment evaluating the effect of temperature on sediment oxygen demand.

Cont.

Lake	Date	Core	uE	Depth.m	Temp.C	T0.O2.mgL	T0.O2.mmolL	Incub.h	Water.Vol.ml	T1.flux	T2.flux	sum.flux
S3	8-Aug-06	B6	0	4	12	8.64e+00	2.70e-01	1.26e+01	1.46e+02	2.76e+01	1.46e+01	1.87e+01
S3	8-Aug-06	Y3	0	4	17	8.76e+00	2.74e-01	1.26e+01	1.46e+02	2.45e+01	1.79e+01	2.00e+01
S3	8-Aug-06	Y1	0	4	17	9.06e+00	2.83e-01	1.26e+01	1.37e+02	2.70e+01	1.66e+01	1.98e+01
S3	8-Aug-06	Y9	0	4	17	8.00e+00	2.50e-01	1.26e+01	1.37e+02	2.49e+01	1.97e+01	2.13e+01
S3	8-Aug-06	Y2	0	4	17	7.88e+00	2.46e-01	1.26e+01	1.37e+02	1.92e+01	2.08e+01	2.03e+01
S3	8-Aug-06	Y6	0	4	17	8.43e+00	2.63e-01	1.26e+01	1.37e+02	2.66e+01	1.53e+01	1.89e+01
S3	8-Aug-06	Y4	0	4	17	7.80e+00	2.44e-01	1.26e+01	1.37e+02	3.10e+01	1.32e+01	1.88e+01
S3	8-Aug-06	Y8	0	4	17	8.48e+00	2.65e-01	1.26e+01	1.37e+02	3.35e+01	1.62e+01	2.17e+01
S3	8-Aug-06	Y5	0	4	17	7.72e+00	2.41e-01	1.26e+01	1.37e+02	4.03e+01	1.48e+01	2.28e+01
S3	8-Aug-06	Y7	0	4	17	7.94e+00	2.48e-01	1.26e+01	1.46e+02	3.69e+01	1.62e+01	2.27e+01
S3	11-Aug-06	Y7	0	4	8	8.73e+00	2.73e-01	1.18e+01	1.46e+02	5.98e+00	2.00e+01	1.51e+01
S3	11-Aug-06	Y6	0	4	8	1.01e+01	3.14e-01	1.18e+01	1.37e+02	2.20e+01	1.39e+01	1.67e+01
S3	11-Aug-06	Y8	0	4	8	9.18e+00	2.87e-01	1.18e+01	1.37e+02	1.44e+01	1.08e+01	1.21e+01
S3	11-Aug-06	Y4	0	4	8	8.97e+00	2.80e-01	1.18e+01	1.46e+02	1.93e+01	1.41e+01	1.59e+01
S3	11-Aug-06	Y1	0	4	8	9.24e+00	2.89e-01	1.18e+01	1.46e+02	2.19e+01	1.17e+01	1.53e+01
S3	11-Aug-06	Y3	0	4	8	9.50e+00	2.97e-01	1.18e+01	1.37e+02	1.69e+01	1.77e+01	1.74e+01
S3	11-Aug-06	Y2	0	4	8	9.11e+00	2.85e-01	1.18e+01	1.37e+02	1.15e+01	1.25e+01	1.22e+01
S3	11-Aug-06	Y5	0	4	8	8.59e+00	2.68e-01	1.18e+01	1.37e+02	1.69e+01	1.30e+01	1.43e+01
S3	11-Aug-06	Y9	0	4	8	9.18e+00	2.87e-01	1.18e+01	1.37e+02	2.20e+01	1.55e+01	1.77e+01

Data collected during the experiment evaluating the effect of temperature on sediment oxygen demand.

Cont.

Lake	Date	Core	uE	Depth.m	Temp.C	T0.O2.mgL	T0.O2.mmolL	Incub.h	Water.Vol.ml	T1.flux	T2.flux	sum.flux
S3	11-Aug-06	B7	0	4	2	9.18e+00	2.87e-01	1.18e+01	1.37e+02	1.36e+01	1.41e+01	1.39e+01
S3	11-Aug-06	B2	0	4	2	9.05e+00	2.83e-01	1.18e+01	1.37e+02	2.65e+01	8.82e+00	1.50e+01
S3	11-Aug-06	B5	0	4	2	9.51e+00	2.97e-01	1.18e+01	1.37e+02	1.90e+01	1.38e+01	1.56e+01
S3	11-Aug-06	B8	0	4	2	8.86e+00	2.77e-01	1.18e+01	1.37e+02	6.49e+00	1.96e+01	1.50e+01
S3	11-Aug-06	B9	0	4	2	9.47e+00	2.96e-01	1.18e+01	1.37e+02	1.42e+01	1.31e+01	1.35e+01
S3	11-Aug-06	B1	0	4	2	8.43e+00	2.63e-01	1.18e+01	1.37e+02	9.21e+00	1.70e+01	1.43e+01
S3	11-Aug-06	B4	0	4	2	9.42e+00	2.94e-01	1.18e+01	1.37e+02	8.40e+00	1.31e+01	1.15e+01
S3	11-Aug-06	B6	0	4	2	9.08e+00	2.84e-01	1.18e+01	1.37e+02	1.59e+01	1.18e+01	1.32e+01
S3	11-Aug-06	B3	0	4	2	8.95e+00	2.80e-01	1.18e+01	1.46e+02	1.01e+01	1.69e+01	1.45e+01

B Oxygen Availability Experiment Data

Table 20: Data collected during the experiment evaluating the effect of oxygen concentration on sediment oxygen demand. Lake is the lake from which the sediment cores were collected. Date is the date of the core collection. Core is the specific ID of the core tube used in the incubation. uE is the irradiance in μE m^{-1} s^{-1}. Temp.C is the temperature of the incubation ($\pm$ 1^{o} C). Depth.m is the depth in the lake from which the cores were collected in m. O2.Time.Step is the sampling interval. mgO2.L is the O_2 concentration of the overlying water during that sample interval in mg O_2 L^{-1}. mmolO2.L is the O_2 concentration of the overlying water during that sample interval in mmol O_2 L^{-1}. Inc.Time.h is the duration of each sample interval in h. Cum.Time.h is the cumulative incubation time since the beginning of the experiment in h. O2.flux.mmol is the flux of O_2 into the sediments during that sample interval in mmol O_2 m^{-2} d^{-1}. Outlier shows values of O2.flux.mmol that were removed from the analysis. NA indicates missing data.

Lake	Date	Core	uE	Temp.C	Depth.m	O2.Time.Step	mgO2.L	mmolO2.L	Inc.Time.h	Cum.Time.h	O2.flux.mmol	Outlier
E4	4-Aug-06	B5	0	12	4	T1	8.74e+00	2.73e-01	4.95e+00	4.95e+00	1.97e+01	NA
E4	4-Aug-06	B6	0	12	4	T1	8.00e+00	2.50e-01	4.95e+00	4.95e+00	4.58e+00	NA
E4	4-Aug-06	B3	0	12	4	T1	8.19e+00	2.56e-01	4.95e+00	4.95e+00	-5.52e+00	NA
E4	4-Aug-06	B4	0	12	4	T1	7.67e+00	2.40e-01	4.95e+00	4.95e+00	1.07e+01	NA
E4	4-Aug-06	B8	0	12	4	T1	7.80e+00	2.44e-01	4.95e+00	4.95e+00	1.36e+01	NA
E4	4-Aug-06	B2	0	12	4	T1	7.90e+00	2.47e-01	4.95e+00	4.95e+00	5.12e+00	NA
E4	4-Aug-06	B9	0	12	4	T1	7.59e+00	2.37e-01	4.95e+00	4.95e+00	7.55e+00	NA
E4	4-Aug-06	B1	0	12	4	T1	8.63e+00	2.70e-01	4.95e+00	4.95e+00	3.98e+01	NA
E4	4-Aug-06	B7	0	12	4	T1	7.99e+00	2.50e-01	4.95e+00	4.95e+00	9.65e+00	NA

Data collected during the experiment evaluating the effect of oxygen concentration on sediment oxygen demand. Cont.

Lake	Date	Core	uE	Temp.C	Depth.m	O2.Time.Step	mgO2.L	mmolO2.L	Inc.Time.h	Cum.Time.h	O2.flux.mmol	Outlier
E4	4-Aug-06	B7	0	12	4	T4	5.34e+00	1.67e-01	1.81e+01	4.33e+01	6.00e+00	NA
E4	4-Aug-06	B7	0	12	4	T5	4.52e+00	1.41e-01	1.05e+01	5.38e+01	4.51e+00	NA
E4	4-Aug-06	B6	0	12	4	T3	6.88e+00	2.15e-01	7.08e+00	2.52e+01	1.99e+01	NA
E4	4-Aug-06	B3	0	12	4	T3	5.34e+00	1.67e-01	7.08e+00	2.52e+01	1.44e+01	NA
E4	4-Aug-06	B4	0	12	4	T3	5.77e+00	1.80e-01	7.08e+00	2.52e+01	1.93e+01	NA
E4	4-Aug-06	B5	0	12	4	T3	6.38e+00	1.99e-01	7.08e+00	2.52e+01	9.39e+00	NA
E4	4-Aug-06	B2	0	12	4	T3	5.73e+00	1.79e-01	7.08e+00	2.52e+01	9.13e+00	NA
E4	4-Aug-06	B9	0	12	4	T3	6.27e+00	1.96e-01	7.08e+00	2.52e+01	1.83e+01	NA
E4	4-Aug-06	B1	0	12	4	T3	5.00e+00	1.56e-01	7.08e+00	2.52e+01	1.45e+01	NA
E4	4-Aug-06	B8	0	12	4	T3	6.03e+00	1.88e-01	7.08e+00	2.52e+01	1.15e+01	NA
E4	4-Aug-06	B1	0	12	4	T5	3.23e+00	1.01e-01	1.05e+01	5.38e+01	5.77e-01	NA
E4	4-Aug-06	B6	0	12	4	T2	7.87e+00	2.46e-01	1.31e+01	1.81e+01	1.03e+01	NA
E4	4-Aug-06	B3	0	12	4	T2	8.47e+00	2.65e-01	1.31e+01	1.81e+01	2.76e+01	NA
E4	4-Aug-06	B7	0	12	4	T3	6.48e+00	2.02e-01	7.08e+00	2.52e+01	2.04e+01	NA
E4	4-Aug-06	B5	0	12	4	T2	7.92e+00	2.47e-01	1.31e+01	1.81e+01	1.38e+01	NA
E4	4-Aug-06	B2	0	12	4	T2	7.73e+00	2.42e-01	1.31e+01	1.81e+01	1.80e+01	NA
E4	4-Aug-06	B9	0	12	4	T2	7.34e+00	2.29e-01	1.31e+01	1.81e+01	1.05e+01	NA
E4	4-Aug-06	B4	0	12	4	T2	7.26e+00	2.27e-01	1.31e+01	1.81e+01	1.37e+01	NA
E4	4-Aug-06	B8	0	12	4	T2	7.26e+00	2.27e-01	1.31e+01	1.81e+01	1.14e+01	NA

Data collected during the experiment evaluating the effect of oxygen concentration on sediment oxygen demand. Cont.

Lake	Date	Core	uE	Temp.C	Depth.m	O2.Time.Step	mgO2.L	mmolO2.L	Inc.Time.h	Cum.Time.h	O2.flux.mmol	Outlier
E4	4-Aug-06	B4	0	12	4	T4	4.72e+00	1.48e-01	1.81e+01	4.33e+01	9.37e+00	NA
E4	4-Aug-06	B5	0	12	4	T4	5.93e+00	1.85e-01	1.81e+01	4.33e+01	9.87e+00	NA
E4	4-Aug-06	B1	0	12	4	T2	6.94e+00	2.17e-01	1.31e+01	1.81e+01	1.76e+01	NA
E4	4-Aug-06	B7	0	12	4	T2	7.62e+00	2.38e-01	1.31e+01	1.81e+01	1.05e+01	NA
E4	4-Aug-06	B1	0	12	4	T4	4.28e+00	1.34e-01	1.81e+01	4.33e+01	7.69e+00	NA
E4	4-Aug-06	B8	0	12	4	T4	5.46e+00	1.71e-01	1.81e+01	4.33e+01	9.36e+00	NA
E4	4-Aug-06	B6	0	12	4	T4	5.88e+00	1.84e-01	1.81e+01	4.33e+01	6.92e+00	NA
E4	4-Aug-06	B3	0	12	4	T4	4.60e+00	1.44e-01	1.81e+01	4.33e+01	9.99e+00	NA
E4	4-Aug-06	B6	0	12	4	T5	5.02e+00	1.57e-01	1.05e+01	5.38e+01	8.05e+00	NA
E4	4-Aug-06	B3	0	12	4	T5	3.18e+00	9.93e-02	1.05e+01	5.38e+01	9.40e+00	NA
E4	4-Aug-06	B2	0	12	4	T4	5.31e+00	1.66e-01	1.81e+01	4.33e+01	1.01e+01	NA
E4	4-Aug-06	B9	0	12	4	T4	5.33e+00	1.67e-01	1.81e+01	4.33e+01	9.39e+00	NA
E4	4-Aug-06	B2	0	12	4	T5	3.84e+00	1.20e-01	1.05e+01	5.38e+01	1.48e+01	NA
E4	4-Aug-06	B9	0	12	4	T5	4.06e+00	1.27e-01	1.05e+01	5.38e+01	9.60e+00	NA
E4	4-Aug-06	B4	0	12	4	T5	3.39e+00	1.06e-01	1.05e+01	5.38e+01	1.23e+01	NA
E4	4-Aug-06	B5	0	12	4	T5	4.48e+00	1.40e-01	1.05e+01	5.38e+01	6.37e+00	NA
E4	4-Aug-06	B8	0	12	4	T5	4.10e+00	1.28e-01	1.05e+01	5.38e+01	9.57e+00	NA
GTH91	26-Jul-06	Y10	0	9	4	T1	8.46e+00	2.65e-01	5.93e+00	5.93e+00	1.76e+01	NA
GTH91	26-Jul-06	Y7	0	9	4	T1	8.89e+00	2.78e-01	5.93e+00	5.93e+00	1.60e+01	NA

Data collected during the experiment evaluating the effect of oxygen concentration on sediment oxygen demand. Cont.

Lake	Date	Core	uE	Temp.C	Depth.m	O2.Time.Step	mgO2.L	mmolO2.L	Inc.Time.h	Cum.Time.h	O2.flux.mmol	Outlier
GTH91	26-Jul-06	Y10	0	9	4	T3	6.63e+00	2.07e-01	7.28e+00	2.39e+01	1.40e+01	NA
GTH91	26-Jul-06	Y4	0	9	4	T3	7.28e+00	2.27e-01	7.28e+00	2.39e+01	1.12e+01	NA
GTH91	26-Jul-06	Y3	0	9	4	T3	7.09e+00	2.22e-01	7.28e+00	2.39e+01	1.07e+01	NA
GTH91	26-Jul-06	Y3	0	9	4	T2	8.18e+00	2.56e-01	1.06e+01	1.66e+01	1.21e+01	NA
GTH91	26-Jul-06	Y3	0	9	4	T1	8.95e+00	2.80e-01	5.93e+00	5.93e+00	1.51e+01	NA
GTH91	26-Jul-06	Y4	0	9	4	T1	9.16e+00	2.86e-01	5.93e+00	5.93e+00	1.01e+01	NA
GTH91	26-Jul-06	Y4	0	9	4	T2	8.65e+00	2.70e-01	1.06e+01	1.66e+01	1.49e+01	NA
GTH91	26-Jul-06	Y8	0	9	4	T4	6.19e+00	1.93e-01	5.70e+00	2.96e+01	1.38e+01	NA
GTH91	26-Jul-06	Y10	0	9	4	T2	7.59e+00	2.37e-01	1.06e+01	1.66e+01	1.09e+01	NA
GTH91	26-Jul-06	Y7	0	9	4	T2	8.08e+00	2.52e-01	1.06e+01	1.66e+01	1.35e+01	NA
GTH91	26-Jul-06	Y8	0	9	4	T3	6.81e+00	2.13e-01	7.28e+00	2.39e+01	1.10e+01	NA
GTH91	26-Jul-06	Y2	0	9	4	T1	8.89e+00	2.78e-01	5.93e+00	5.93e+00	1.23e+01	NA
GTH91	26-Jul-06	Y4	0	9	4	T4	6.63e+00	2.07e-01	5.70e+00	2.96e+01	1.19e+00	NA
GTH91	26-Jul-06	Y3	0	9	4	T4	6.49e+00	2.03e-01	5.70e+00	2.96e+01	6.17e+00	NA
GTH91	26-Jul-06	Y10	0	9	4	T4	5.83e+00	1.82e-01	5.70e+00	2.96e+01	1.06e+01	NA
GTH91	26-Jul-06	Y8	0	9	4	T1	9.18e+00	2.87e-01	5.93e+00	5.93e+00	-7.67e+00	NA
GTH91	26-Jul-06	Y8	0	9	4	T2	9.59e+00	3.00e-01	1.06e+01	1.66e+01	2.98e+01	NA
GTH91	26-Jul-06	Y8	0	9	4	T5	5.62e+00	1.76e-01	1.09e+01	4.05e+01	7.33e+00	NA
GTH91	26-Jul-06	Y4	0	9	4	T5	6.68e+00	2.09e-01	1.09e+01	4.05e+01	7.68e+00	NA

Data collected during the experiment evaluating the effect of oxygen concentration on sediment oxygen demand. Cont.

Lake	Date	Core	uE	Temp.C	Depth.m	O2.Time.Step	mgO2.L	mmolO2.L	Inc.Time.h	Cum.Time.h	O2.flux.mmol	Outlier
GTH91	26-Jul-06	Y3	0	9	4	T5	6.29e+00	1.96e-01	1.09e+01	4.05e+01	7.84e+00	NA
GTH91	26-Jul-06	Y7	0	9	4	T3	6.86e+00	2.14e-01	7.28e+00	2.39e+01	NA	1.09e+02
GTH91	26-Jul-06	Y5	0	9	4	T1	9.13e+00	2.85e-01	5.93e+00	5.93e+00	1.82e+01	NA
GTH91	26-Jul-06	Y6	0	9	4	T1	8.74e+00	2.73e-01	5.93e+00	5.93e+00	1.58e+01	NA
GTH91	26-Jul-06	Y9	0	9	4	T1	9.35e+00	2.92e-01	5.93e+00	5.93e+00	1.87e+01	NA
GTH91	26-Jul-06	Y9	0	9	4	T4	6.77e+00	2.12e-01	5.70e+00	2.96e+01	8.41e+00	NA
GTH91	26-Jul-06	Y9	0	9	4	T5	6.45e+00	2.02e-01	1.09e+01	4.05e+01	4.00e+00	NA
GTH91	26-Jul-06	Y2	0	9	4	T2	8.27e+00	2.58e-01	1.06e+01	1.66e+01	1.41e+01	NA
GTH91	26-Jul-06	Y5	0	9	4	T2	8.19e+00	2.56e-01	1.06e+01	1.66e+01	9.99e+00	NA
GTH91	26-Jul-06	Y6	0	9	4	T2	7.94e+00	2.48e-01	1.06e+01	1.66e+01	1.08e+01	NA
GTH91	26-Jul-06	Y10	0	9	4	T5	5.43e+00	1.70e-01	1.09e+01	4.05e+01	6.82e+00	NA
GTH91	26-Jul-06	Y9	0	9	4	T3	7.58e+00	2.37e-01	7.28e+00	2.39e+01	1.36e+01	NA
GTH91	26-Jul-06	Y7	0	9	4	T4	NA	NA	5.70e+00	2.96e+01	NA	-1.11e+02
GTH91	26-Jul-06	Y2	0	9	4	T4	6.66e+00	2.08e-01	5.70e+00	2.96e+01	1.20e+01	NA
GTH91	26-Jul-06	Y9	0	9	4	T2	8.39e+00	2.62e-01	1.06e+01	1.66e+01	9.01e+00	NA
GTH91	26-Jul-06	Y6	0	9	4	T4	6.71e+00	2.10e-01	5.70e+00	2.96e+01	1.40e+01	NA
GTH91	26-Jul-06	Y6	0	9	4	T5	6.11e+00	1.91e-01	1.09e+01	4.05e+01	9.09e+00	NA
GTH91	26-Jul-06	Y5	0	9	4	T5	6.49e+00	2.03e-01	1.09e+01	4.05e+01	9.51e+00	NA
GTH91	26-Jul-06	Y5	0	9	4	T4	6.53e+00	2.04e-01	5.70e+00	2.96e+01	2.92e+00	NA

Data collected during the experiment evaluating the effect of oxygen concentration on sediment oxygen demand. Cont.

Lake	Date	Core	uE	Temp.C	Depth.m	O2.Time.Step	mgO2.L	mmolO2.L	Inc.Time.h	Cum.Time.h	O2.flux.mmol	Outlier
GTH91	26-Jul-06	Y5	0	9	4	T3	7.30e+00	2.28e-01	7.28e+00	2.39e+01	1.32e+01	NA
GTH91	26-Jul-06	Y6	0	9	4	T3	6.98e+00	2.18e-01	7.28e+00	2.39e+01	5.69e+00	NA
GTH91	26-Jul-06	Y2	0	9	4	T3	6.99e+00	2.18e-01	7.28e+00	2.39e+01	6.49e+00	NA
GTH91	26-Jul-06	Y7	0	9	4	T5	5.56e+00	1.74e-01	1.09e+01	4.05e+01	6.02e+00	NA
GTH91	26-Jul-06	Y2	0	9	4	T5	6.16e+00	1.93e-01	1.09e+01	4.05e+01	4.61e+00	NA
S3	8-Aug-06	B3	0	12	4	T1	8.58e+00	2.68e-01	5.78e+00	5.78e+00	1.92e+01	NA
S3	8-Aug-06	B6	0	12	4	T1	8.64e+00	2.70e-01	5.78e+00	5.78e+00	2.76e+01	NA
S3	8-Aug-06	B9	0	12	4	T1	9.19e+00	2.87e-01	5.78e+00	5.78e+00	2.18e+01	NA
S3	8-Aug-06	B5	0	12	4	T1	9.00e+00	2.81e-01	5.78e+00	5.78e+00	3.29e+01	NA
S3	8-Aug-06	B7	0	12	4	T1	8.76e+00	2.74e-01	5.78e+00	5.78e+00	1.53e+01	NA
S3	8-Aug-06	B5	0	12	4	T2	7.37e+00	2.30e-01	1.26e+01	1.84e+01	1.46e+01	NA
S3	8-Aug-06	B7	0	12	4	T2	8.03e+00	2.51e-01	1.26e+01	1.84e+01	1.46e+01	NA
S3	8-Aug-06	B4	0	12	4	T1	8.77e+00	2.74e-01	5.78e+00	5.78e+00	1.95e+01	NA
S3	8-Aug-06	B8	0	12	4	T1	8.23e+00	2.57e-01	5.78e+00	5.78e+00	1.61e+01	NA
S3	8-Aug-06	B4	0	12	4	T2	7.83e+00	2.45e-01	1.26e+01	1.84e+01	2.12e+01	NA
S3	8-Aug-06	B8	0	12	4	T2	7.42e+00	2.32e-01	1.26e+01	1.84e+01	1.56e+01	NA
S3	8-Aug-06	B3	0	12	4	T2	7.71e+00	2.41e-01	1.26e+01	1.84e+01	1.55e+01	NA
S3	8-Aug-06	B1	0	12	4	T2	7.92e+00	2.47e-01	1.26e+01	1.84e+01	1.53e+01	NA
S3	8-Aug-06	B2	0	12	4	T4	5.72e+00	1.79e-01	1.60e+01	4.12e+01	7.21e+00	NA

Data collected during the experiment evaluating the effect of oxygen concentration on sediment oxygen demand. Cont.

Lake	Date	Core	uE	Temp.C	Depth.m	O2.Time.Step	mgO2.L	mmolO2.L	Inc.Time.h	Cum.Time.h	O2.flux.mmol	Outlier
S3	8-Aug-06	B1	0	12	4	T4	5.75e+00	1.80e-01	1.60e+01	4.12e+01	9.33e+00	NA
S3	8-Aug-06	B6	0	12	4	T2	7.38e+00	2.30e-01	1.26e+01	1.84e+01	1.39e+01	NA
S3	8-Aug-06	B9	0	12	4	T2	8.19e+00	2.56e-01	1.26e+01	1.84e+01	1.60e+01	NA
S3	8-Aug-06	B3	0	12	4	T4	5.13e+00	1.60e-01	1.60e+01	4.12e+01	1.05e+01	NA
S3	8-Aug-06	B9	0	12	4	T4	5.73e+00	1.79e-01	1.60e+01	4.12e+01	6.09e+00	NA
S3	8-Aug-06	B5	0	12	4	T4	5.30e+00	1.65e-01	1.60e+01	4.12e+01	1.07e+01	NA
S3	8-Aug-06	B7	0	12	4	T4	5.73e+00	1.79e-01	1.60e+01	4.12e+01	7.05e+00	NA
S3	8-Aug-06	B6	0	12	4	T4	5.22e+00	1.63e-01	1.60e+01	4.12e+01	8.87e+00	NA
S3	8-Aug-06	B2	0	12	4	T3	6.51e+00	2.03e-01	6.83e+00	2.52e+01	1.54e+01	NA
S3	8-Aug-06	B4	0	12	4	T4	4.84e+00	1.51e-01	1.60e+01	4.12e+01	9.11e+00	NA
S3	8-Aug-06	B8	0	12	4	T4	4.48e+00	1.40e-01	1.60e+01	4.12e+01	6.40e+00	NA
S3	8-Aug-06	B1	0	12	4	T1	9.09e+00	2.84e-01	5.78e+00	5.78e+00	2.38e+01	NA
S3	8-Aug-06	B3	0	12	4	T3	6.20e+00	1.94e-01	6.83e+00	2.52e+01	2.14e+01	NA
S3	8-Aug-06	B1	0	12	4	T3	6.32e+00	1.97e-01	6.83e+00	2.52e+01	1.17e+01	NA
S3	8-Aug-06	B5	0	12	4	T3	5.87e+00	1.83e-01	6.83e+00	2.52e+01	1.21e+01	NA
S3	8-Aug-06	B7	0	12	4	T3	6.50e+00	2.03e-01	6.83e+00	2.52e+01	1.50e+01	NA
S3	8-Aug-06	B6	0	12	4	T3	6.03e+00	1.88e-01	6.83e+00	2.52e+01	1.70e+01	NA
S3	8-Aug-06	B9	0	12	4	T3	6.60e+00	2.06e-01	6.83e+00	2.52e+01	1.78e+01	NA
S3	8-Aug-06	B4	0	12	4	T3	5.58e+00	1.74e-01	6.83e+00	2.52e+01	1.51e+01	NA

Data collected during the experiment evaluating the effect of oxygen concentration on sediment oxygen demand. Cont.

Lake	Date	Core	uE	Temp.C	Depth.m	O2.Time.Step	mgO2.L	mmolO2.L	Inc.Time.h	Cum.Time.h	O2.flux.mmol	Outlier
S3	8-Aug-06	B8	0	12	4	T3	5.69e+00	1.78e-01	6.83e+00	2.52e+01	2.16e+01	NA
S3	8-Aug-06	B2	0	12	4	T1	8.95e+00	2.80e-01	5.78e+00	5.78e+00	1.47e+01	NA
S3	8-Aug-06	B2	0	12	4	T2	8.25e+00	2.58e-01	1.26e+01	1.84e+01	1.64e+01	NA
S3	8-Aug-06	B2	0	12	4	T5	4.85e+00	1.52e-01	7.35e+00	4.85e+01	1.03e+01	NA
S3	8-Aug-06	B1	0	12	4	T5	4.59e+00	1.43e-01	7.35e+00	4.85e+01	6.76e+00	NA
S3	8-Aug-06	B5	0	12	4	T5	3.95e+00	1.23e-01	7.35e+00	4.85e+01	1.00e+01	NA
S3	8-Aug-06	B7	0	12	4	T5	4.88e+00	1.53e-01	7.35e+00	4.85e+01	1.07e+01	NA
S3	8-Aug-06	B3	0	12	4	T5	3.90e+00	1.22e-01	7.35e+00	4.85e+01	1.46e+01	NA
S3	8-Aug-06	B9	0	12	4	T5	5.06e+00	1.58e-01	7.35e+00	4.85e+01	1.35e+01	NA
S3	8-Aug-06	B4	0	12	4	T5	3.73e+00	1.17e-01	7.35e+00	4.85e+01	6.35e+00	NA
S3	8-Aug-06	B8	0	12	4	T5	3.71e+00	1.16e-01	7.35e+00	4.85e+01	1.49e+01	NA
S3	8-Aug-06	B6	0	12	4	T5	4.20e+00	1.31e-01	7.35e+00	4.85e+01	1.57e+01	NA

C Light Availability Experiment Data

Table 21: Data collected during the experiment evaluating the effect of light availabilty on sediment oxygen demand in the sediments of lakes E – 4 and S – 3. Lake is the name of the lake from which the cores were collected. Position identifies the location in the incubation chamber where the core was incubated. I.um. is the amount of light reaching the core in μE m^{-2} s^{-1}. Zm is the depth within the lake were the cores were collected. Temp is the temperature of the incubation chamber during the experiment. Cflux is the flux of dissolved inorganic carbon into the sediments in μmol C m^{-2} d^{-1}. Oflux is the flux of oxygen out of the sediments in μmol C m^{-2} d^{-1}. Oflux.pos is the flux of oxygen into the sediments in μmol C m^{-2} d^{-1}. FinalO2 is the oxygen concentration of the water overlying the core as the conclusion of the incubation in mg O_2 L^{-1}. NA indicates missing data.

Lake	Position	I.um.	Zm	Temp	Cflux	Oflux	Oflux.pos	FinalO2
E-4	14	0.000	3.0	15	-20.66	-27.086	27.086	4.10
E-4	15	0.000	3.0	15	-31.87	-20.347	20.347	4.00
E-4	16	0.000	3.0	15	-20.01	-23.481	23.481	4.61
E-4	9	0.498	3.0	15	-28.31	-26.675	26.675	4.59
E-4	11	5.312	3.0	15	-16.12	-16.461	16.461	5.49
E-4	13	18.260	3.0	15	-1.11	-6.096	6.096	7.80
E-4	10	19.920	3.0	15	-3.95	-9.928	9.928	7.21
E-4	4	34.030	3.0	15	-2.48	0.289	-0.289	9.63
E-4	8	34.860	3.0	15	3.17	-5.709	5.709	8.66
E-4	3	46.480	3.0	15	-7.61	-11.688	11.688	8.15
E-4	7	51.460	3.0	15	18.40	5.568	-5.568	10.85
E-4	1	89.640	3.0	15	11.21	2.302	-2.302	10.65
E-4	5	102.920	3.0	15	12.12	1.367	-1.367	10.07
E-4	6	111.220	3.0	15	18.59	12.083	-12.083	12.47
E-4	2	190.900	3.0	15	11.08	1.580	-1.580	10.45
E-4	12	249.000	3.0	15	19.60	6.377	-6.377	11.22
GTH 91	d2	0.266	3.0	10	NA	-4.489	4.489	NA

Data collected during the experiment evaluating the effect of light availabilty on sediment oxygen demand in the sediments of lakes E – 4 and S – 3. Cont.

Lake	Position	I.um.	Zm	Temp	Cflux	Oflux	Oflux.pos	FinalO2
GTH 91	26	0.315	3.0	10	NA	-22.567	22.567	NA
GTH 91	d3	0.564	3.0	10	NA	-17.455	17.455	NA
GTH 91	27	0.564	3.0	10	NA	-12.197	12.197	NA
GTH 91	26	1.328	3.0	11	NA	-17.227	17.227	NA
GTH 91	21	1.328	3.0	16	NA	-18.183	18.183	NA
GTH 91	23	1.328	3.0	16	NA	-14.762	14.762	NA
GTH 91	27	1.328	3.0	16	NA	-22.576	22.576	NA
GTH 91	17	1.494	3.0	16	NA	-21.547	21.547	NA
GTH 91	20	1.660	3.0	16	NA	-17.763	17.763	NA
GTH 91	5	1.826	3.0	16	NA	-20.969	20.969	NA
GTH 91	25	1.992	3.0	11	NA	-13.222	13.222	NA
GTH 91	d1	1.992	3.0	10	NA	-19.046	19.046	NA
GTH 91	8	2.158	3.0	16	NA	-15.807	15.807	NA
GTH 91	0	2.324	3.0	16	NA	-15.769	15.769	NA
GTH 91	24	2.656	3.0	11	NA	-15.411	15.411	NA
GTH 91	19	4.648	3.0	11	NA	-16.160	16.160	NA
GTH 91	23	5.312	3.0	10	NA	-22.677	22.677	NA
GTH 91	16	6.308	3.0	11	NA	-15.997	15.997	NA
GTH 91	12	9.960	3.0	11	NA	-12.881	12.881	NA
GTH 91	13	13.280	3.0	11	NA	-14.083	14.083	NA
GTH 91	19	21.580	3.0	10	NA	-7.874	7.874	NA
GTH 91	4	23.240	3.0	11	NA	-18.883	18.883	NA
GTH 91	13	39.840	3.0	10	NA	7.141	-7.141	NA
GTH 91	12	41.500	3.0	10	NA	-0.756	0.756	NA
GTH 91	5	66.400	3.0	10	NA	-1.674	1.674	NA
GTH 91	0	146.080	3.0	10	NA	13.138	-13.138	NA

Data collected during the experiment evaluating the effect of light availabilty on sediment oxygen demand in the sediments of lakes E – 4 and S – 3. Cont.

Lake	Position	I.um.	Zm	Temp	Cflux	Oflux	Oflux.pos	FinalO2
S-3	14	0.000	3.2	15	-21.63	-29.581	29.581	2.59
S-3	15	0.000	3.2	15	-19.19	-32.802	32.802	1.49
S-3	16	0.000	3.2	15	-26.95	-22.530	22.530	4.12
S-3	9	0.498	3.2	15	-25.59	-23.246	23.246	4.43
S-3	11	5.312	3.2	15	-7.37	-7.706	7.706	6.90
S-3	13	18.260	3.2	15	1.72	-16.422	16.422	5.92
S-3	10	19.920	3.2	15	-8.78	-19.136	19.136	4.61
S-3	4	34.030	3.2	15	-8.21	-11.045	11.045	8.04
S-3	8	34.860	3.2	15	-10.43	-26.236	26.236	3.96
S-3	3	46.480	3.2	15	-25.55	-25.309	25.309	3.88
S-3	7	51.460	3.2	15	-16.88	-23.877	23.877	3.91
S-3	1	89.640	3.2	15	21.57	12.661	-12.661	12.69
S-3	5	102.920	3.2	15	3.69	-7.125	7.125	8.73
S-3	6	111.220	3.2	15	32.58	11.462	-11.462	13.03
S-3	2	190.900	3.2	15	18.70	16.339	-16.339	13.28
S-3	12	249.000	3.2	15	36.26	18.532	-18.532	13.97

D Organic Matter Lability Experiment Data

Table 22: Data collected during the experiment evaluating dissolved inorganic carbon (DIC) production in sediment slurries from arctic lake sediments. Lake is the lake from which the sediments were collected. Julian is the Julian day on which the incubating slurries were sampled. Treatment is the coded treatment designation where NO2D is anoxic and deep sediments (9 to 10 cm in the core), NO2S is anoxic and surface sediments (1 to 2 cm in the core), O2D is oxic and deep sediments, and O2S is oxic and surface sediments. Temp is the temperature of the incubation room in ° C. Rep designates the replicate vial of each treatment. T1.flux is the DIC production from the sediments during the first 3 hour sampling period in μmol DIC (g organic matter)$^{-1}$ h^{-1}. T2.flux is the DIC production from the sediments during the second 3 hour sampling period in μmol DIC (g organic matter)$^{-1}$ h^{-1}. Mean.flux is the mean of T1.flux and T2.flux in μmol DIC (g organic matter)$^{-1}$ h^{-1}. perc.water is the proportion of water in the sediment slurry. g.dry.sed is the mass of the dry sediment in the slurry in g. ash.mass is the mass of the sediment following 5 h at 550° C in g. g.OM is the mass of the component of the dry sediment in g, calculated as g.dry.sed - ash.mass. perc.OM is the proportion of organic matter in the sediment slurry. g.OM.ml is the concentration of organic matter in the slurry in g organic matter ml^{-1}.

Lake	Julian	Treatment	Temp	Rep	T1.flux	T2.flux	Mean.flux	perc.water	g.dry.sed	ash.mass	g.OM	perc.OM	g.OM.ml
E-4	178	NO2D	19	A	1.04e-01	2.67e-01	1.85e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	178	NO2D	19	B	1.71e-01	1.83e-01	1.77e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	178	NO2D	19	C	-1.77e-02	3.71e-01	1.77e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	178	NO2S	19	A	1.60e+00	1.77e+00	1.68e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	178	NO2S	19	B	1.58e+00	1.84e+00	1.71e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02

Data collected during the experiment evaluating dissolved inorganic carbon (DIC) production in sediment slurries from arctic lake sediments. Cont.

Lake	Julian	Treatment	Temp	Rep	T1.flux	T2.flux	Mean.flux	perc.water	g.dry.sed	ash.mass	g.OM	perc.OM	g.OM.ml
E-4	178	NO2S	19	C	1.71e+00	1.65e+00	1.68e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	178	O2D	19	A	1.51e+00	1.28e+00	1.40e+00	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	178	O2D	19	B	1.68e+00	1.40e+00	1.54e+00	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	178	O2D	19	C	1.28e+00	1.20e+00	1.24e+00	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	178	O2S	19	A	2.41e+00	2.79e+00	2.60e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	178	O2S	19	B	2.60e+00	2.78e+00	2.69e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	178	O2S	19	C	2.82e+00	3.13e+00	2.97e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
S-3	178	NO2D	19	A	1.52e+00	3.05e+00	2.29e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	178	NO2D	19	B	1.31e+00	3.38e+00	2.34e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	178	NO2D	19	C	2.76e+01	2.28e+00	1.50e+01	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	178	NO2S	19	A	7.34e+00	8.99e+00	8.17e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	178	NO2S	19	B	6.80e+00	9.25e+00	8.03e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	178	NO2S	19	C	7.78e+00	8.41e+00	8.10e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	178	O2D	19	A	8.30e-01	3.28e+00	2.06e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	178	O2D	19	B	1.40e+00	1.75e+00	1.57e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	178	O2D	19	C	1.02e+00	1.56e+00	1.29e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	178	O2S	19	A	9.33e+00	1.01e+01	9.74e+00	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
S-3	178	O2S	19	B	8.49e+00	8.92e+00	8.71e+00	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
S-3	178	O2S	19	C	9.52e+00	1.01e+01	9.78e+00	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02

Data collected during the experiment evaluating dissolved inorganic carbon (DIC) production in sediment slurries from arctic lake sediments. Cont.

Lake	Julian	Treatment	Temp	Rep	T1.flux	T2.flux	Mean.flux	perc.water	g.dry.sed	ash.mass	g.OM	perc.OM	g.OM.ml
E-4	180	NO2D	20	A	1.73e-01	2.77e-01	2.25e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	180	NO2D	20	B	1.81e-01	1.36e-01	1.58e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	180	NO2D	20	C	9.63e-02	1.76e-01	1.36e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	180	NO2S	20	A	1.07e+00	1.62e+00	1.35e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	180	NO2S	20	B	9.13e-01	1.48e+00	1.20e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	180	NO2S	20	C	1.07e+00	1.75e+00	1.41e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	180	O2D	20	A	6.11e-01	4.93e-01	5.52e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	180	O2D	20	B	5.20e-01	5.63e-01	5.42e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	180	O2D	20	C	6.31e-01	5.89e-01	6.10e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	180	O2S	20	A	2.33e+00	2.76e+00	2.55e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	180	O2S	20	B	2.62e+00	2.95e+00	2.79e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	180	O2S	20	C	2.68e+00	2.85e+00	2.77e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
S-3	180	NO2D	20	A	3.49e+00	6.77e+00	5.13e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	180	NO2D	20	B	3.34e+00	6.69e+00	5.02e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	180	NO2D	20	C	3.59e+00	7.57e+00	5.58e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	180	NO2S	20	A	6.34e+00	7.80e+00	7.07e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	180	NO2S	20	B	7.31e+00	7.85e+00	7.58e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	180	NO2S	20	C	6.42e+00	9.75e+00	8.09e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	180	O2D	20	A	9.41e-01	2.14e+00	1.54e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03

Data collected during the experiment evaluating dissolved inorganic carbon (DIC) production in sediment slurries from arctic lake sediments. Cont.

Lake	Julian	Treatment	Temp	Rep	T1.flux	T2.flux	Mean.flux	perc.water	g.dry.sed	ash.mass	g.OM	perc.OM	g.OM.ml
S-3	180	O2D	20	B	1.02e+00	1.30e+00	1.16e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	180	O2D	20	C	1.20e+00	1.42e+00	1.31e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	180	O2S	20	A	1.04e+01	1.12e+01	1.08e+01	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
S-3	180	O2S	20	B	1.01e+01	1.12e+01	1.06e+01	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
S-3	180	O2S	20	C	1.18e+01	1.35e+01	1.26e+01	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
E-4	184	NO2D	20	A	1.57e-02	3.63e-01	1.90e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	184	NO2D	20	B	7.80e-02	1.29e-01	1.04e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	184	NO2D	20	C	1.82e-01	7.81e-02	1.30e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	184	NO2S	20	A	8.09e-01	1.29e+00	1.05e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	184	NO2S	20	B	9.00e-01	9.57e-01	9.28e-01	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	184	NO2S	20	C	-7.27e+00	9.27e+00	1.00e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	184	O2D	20	A	8.60e-01	6.58e-01	7.59e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	184	O2D	20	B	1.41e+00	1.02e+00	1.21e+00	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	184	O2D	20	C	1.52e+00	1.16e+00	1.34e+00	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	184	O2S	20	A	1.52e+00	1.79e+00	1.66e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	184	O2S	20	B	2.07e+00	1.80e+00	1.93e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	184	O2S	20	C	2.19e+00	1.84e+00	2.02e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
S-3	184	NO2D	20	A	2.46e+00	5.12e+00	3.79e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	184	NO2D	20	B	2.45e+00	5.30e+00	3.88e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03

Data collected during the experiment evaluating dissolved inorganic carbon (DIC) production in sediment slurries from arctic lake sediments. Cont.

Lake	Julian	Treatment	Temp	Rep	T1.flux	T2.flux	Mean.flux	perc.water	g.dry.sed	ash.mass	g.OM	perc.OM	g.OM.ml
S-3	184	NO2D	20	C	2.20e+00	5.79e+00	3.99e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	184	NO2S	20	A	6.99e+00	9.06e+00	8.03e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	184	NO2S	20	B	7.69e+00	1.11e+01	9.37e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	184	NO2S	20	C	7.49e+00	9.52e+00	8.51e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	184	O2D	20	A	1.91e+00	2.23e+00	2.07e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	184	O2D	20	B	1.85e+00	2.51e+00	2.18e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	184	O2D	20	C	2.27e+00	2.97e+00	2.62e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	184	O2S	20	A	8.37e+00	1.12e+01	9.80e+00	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
S-3	184	O2S	20	B	1.02e+01	9.70e+00	9.93e+00	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
S-3	184	O2S	20	C	1.10e+01	1.00e+01	1.05e+01	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
E-4	187	NO2D	20	A	1.02e-01	2.25e-01	1.64e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	187	NO2D	20	B	1.91e-01	3.12e-03	9.70e-02	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	187	NO2D	20	C	1.19e-01	8.44e-02	1.02e-01	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	187	NO2S	20	A	7.21e-01	1.69e+00	1.20e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	187	NO2S	20	B	8.50e-01	1.91e+00	1.38e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	187	NO2S	20	C	9.36e-01	1.47e+00	1.20e+00	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	187	O2D	20	A	1.04e+00	9.35e-01	9.89e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	187	O2D	20	B	8.54e-01	8.14e-01	8.34e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	187	O2D	20	C	9.09e-01	8.42e-01	8.75e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03

Data collected during the experiment evaluating dissolved inorganic carbon (DIC) production in sediment slurries from arctic lake sediments. Cont.

Lake	Julian	Treatment	Temp	Rep	T1.flux	T2.flux	Mean.flux	perc.water	g.dry.sed	ash.mass	g.OM	perc.OM	g.OM.ml
E-4	187	O2S	20	A	2.33e+00	2.54e+00	2.43e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	187	O2S	20	B	2.84e+00	2.73e+00	2.78e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	187	O2S	20	C	3.58e+00	3.42e+00	3.50e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
S-3	187	NO2D	20	A	2.39e+00	5.26e+00	3.82e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	187	NO2D	20	B	2.22e+00	5.35e+00	3.79e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	187	NO2D	20	C	3.20e+00	2.36e+00	2.78e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	187	NO2S	20	A	8.64e+00	8.93e+00	8.78e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	187	NO2S	20	B	9.80e+00	9.42e+00	9.61e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	187	NO2S	20	C	7.43e+00	1.17e+01	9.57e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	187	O2D	20	A	3.34e+00	4.00e+00	3.67e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	187	O2D	20	B	1.99e+00	2.00e+00	1.99e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	187	O2D	20	C	3.66e+00	3.67e+00	3.66e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	187	O2S	20	A	1.33e+01	1.38e+01	1.36e+01	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
S-3	187	O2S	20	B	1.22e+01	1.24e+01	1.23e+01	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
S-3	187	O2S	20	C	1.71e+01	1.96e+01	1.84e+01	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
E-4	193	NO2D	19	A	1.22e+00	-1.23e+00	-3.16e-03	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	193	NO2D	19	B	-3.08e-01	3.58e-01	2.53e-02	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	193	NO2D	19	C	0.00e+00	-9.34e-02	-4.67e-02	9.81e-01	1.96e-02	1.39e-01	5.74e-02	2.92e+01	5.74e-03
E-4	193	NO2S	19	A	5.71e-01	9.00e-01	7.35e-01	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02

Data collected during the experiment evaluating dissolved inorganic carbon (DIC) production in sediment slurries from arctic lake sediments. Cont.

Lake	Julian	Treatment	Temp	Rep	T1.flux	T2.flux	Mean.flux	perc.water	g.dry.sed	ash.mass	g.OM	perc.OM	g.OM.ml
E-4	193	NO2S	19	B	4.81e-01	8.65e-01	6.73e-01	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	193	NO2S	19	C	4.54e-01	9.59e-01	7.06e-01	9.70e-01	3.17e-02	2.14e-01	1.03e-01	3.26e+01	1.03e-02
E-4	193	O2D	19	A	2.51e+00	-1.29e+00	6.11e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	193	O2D	19	B	5.91e-01	6.01e-01	5.96e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	193	O2D	19	C	4.85e-01	6.33e-01	5.59e-01	9.81e-01	1.91e-02	1.34e-01	5.69e-02	2.98e+01	5.69e-03
E-4	193	O2S	19	A	2.36e+00	2.40e+00	2.38e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	193	O2S	19	B	3.43e+00	8.16e-01	2.12e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
E-4	193	O2S	19	C	3.64e+00	1.77e+00	2.71e+00	9.70e-01	3.36e-02	2.28e-01	1.08e-01	3.21e+01	1.08e-02
S-3	193	NO2D	19	A	9.69e-01	3.02e+00	2.00e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	193	NO2D	19	B	1.12e+00	2.25e+00	1.69e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	193	NO2D	19	C	-4.68e-02	4.11e+00	2.03e+00	9.93e-01	7.61e-03	4.93e-02	2.68e-02	3.52e+01	2.68e-03
S-3	193	NO2S	19	A	6.90e+00	9.02e+00	7.96e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	193	NO2S	19	B	6.62e+00	1.15e+01	9.05e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	193	NO2S	19	C	2.05e+00	9.87e+00	5.96e+00	9.83e-01	1.70e-02	6.21e-02	1.08e-01	6.36e+01	1.08e-02
S-3	193	O2D	19	A	8.51e+00	-2.96e+00	2.78e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	193	O2D	19	B	-2.49e+00	2.67e+00	9.03e-02	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	193	O2D	19	C	9.41e+00	-2.08e+00	3.67e+00	9.92e-01	8.63e-03	5.48e-02	3.15e-02	3.65e+01	3.15e-03
S-3	193	O2S	19	A	9.34e+00	1.12e+01	1.03e+01	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02
S-3	193	O2S	19	B	1.19e+01	9.06e+00	1.05e+01	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02

Data collected during the experiment evaluating dissolved inorganic carbon (DIC) production in sediment slurries from arctic lake sediments. Cont.

Lake	Julian	Treatment	Temp	Rep	T1.flux	T2.flux	Mean.flux	perc.water	g.dry.sed	ash.mass	g.OM	perc.OM	g.OM.ml
S-3	193	O2S	19	C	8.66e+00	1.07e+01	9.68e+00	9.84e-01	1.70e-02	6.22e-02	1.07e-01	6.33e+01	1.07e-02

E Survey of Thermocline Depth and Light Extinction Data

Table 23: Data collected for the analysis of the relationship between light attenuation and thermocline depth. Lake is the lake from which the data were collected. Year is the year of the survey. Kd is the light attenuation coefficient in m^{-1}. TD is the thermocline depth in m. MaxZ is the maximum depth of the lake in m. EpiZ is the depth of the epilimnion on the date of the survey in m. HypoZ is the depth of the hypolimnion on the date of the survey in m. Area is the surface area of the lake in ha. DOC is the dissolved organic carbon concentration in the surface water in mg L^{-1}. TN is the total nitrogen concentration in the surface water in mg L^{-1}. cDOM is the concentration of chromophoric dissolved organic matter in the surface water in quinone sulfate units. Chl is the concentration of chlorophyll a (Chl a)in the surface water in μg Chl a L^{-1}. perc.I.at.TD is the percentage of surface irradiance reaching the thermocline depth. EpiT is the mean of all the temperature measurements in the epilimnion in ° C. HypoT is the mean of all the temperature measurements in the hypolimnion in ° C. LakeT is the mean of all of the temperature measurements collected from the lake in ° C. Julian is the Julian day of the survey. NA indicates missing data.

Lake	Year	Kd	TD	MaxZ	EpiZ	HypoZ	Area	DOC	TN	cDOM	Chl	perc.I.at.TD	EpiT	HypoT	LakeT	Julian
GTH30	2006	0.870	5.0	21.4	5.0	7.0	6.825	NA	NA	NA	NA	1.291	12.0	5.88	9.67	219
GTH31	2006	0.580	6.0	12.0	3.5	6.5	2.185	NA	NA	NA	NA	3.081	12.1	4.88	9.31	219
GTH32	2006	0.800	6.0	15.7	5.5	8.0	12.904	NA	NA	NA	NA	0.823	12.7	5.10	10.55	219
GTH33	2006	0.760	5.0	13.3	5.0	7.5	4.285	NA	NA	NA	NA	2.237	12.5	5.07	10.00	219
GTH34	2006	0.800	5.5	17.4	5.0	5.7	3.583	NA	NA	NA	NA	1.228	12.3	5.09	9.32	219
GTH57	2006	0.250	9.5	21.6	8.5	11.0	29.959	NA	NA	NA	NA	9.301	12.4	5.33	10.78	219

Data collected for the analysis of the relationship between light attenuation and thermocline depth. Cont.

Lake	Year	Kd	TD	MaxZ	EpiZ	HypoZ	Area	DOC	TN	cDOM	Chl	perc.I.at.TD	EpiT	HypoT	LakeT	Julian
GTH58	2006	0.390	7.5	16.2	6.0	8.5	4.083	NA	NA	NA	NA	5.366	12.6	5.03	10.60	219
GTH65	2006	0.380	6.0	16.5	5.5	8.0	4.308	NA	NA	NA	NA	10.228	12.2	4.95	10.29	219
GTH66	2006	0.300	7.0	25.9	6.5	8.5	16.630	NA	NA	NA	NA	12.246	12.6	5.73	10.05	219
GTH67	2006	0.590	NA	7.0	NA	NA	2.605	NA	NA	NA	NA	NA	NA	NA	NA	219
GTH68	2006	0.260	11.5	27.4	9.0	11.5	77.079	NA	NA	NA	NA	5.029	12.3	6.66	10.68	219
GTH69	2006	0.370	6.0	14.3	6.0	7.5	6.126	NA	NA	NA	NA	10.861	12.8	5.24	9.78	219
GTH71	2006	0.230	9.0	18.3	8.0	10.5	17.107	NA	NA	NA	NA	12.619	12.6	6.70	10.40	219
GTH73	2006	0.460	NA	13.7	NA	NA	9.865	NA	NA	NA	NA	NA	NA	NA	NA	219
GTH74	2006	0.270	7.0	11.3	6.0	8.0	5.862	NA	NA	NA	NA	15.107	12.8	6.50	10.43	219
GTH92	2006	0.230	7.5	10.4	7.0	7.5	5.565	NA	NA	NA	NA	17.817	11.9	8.38	11.04	219
GTH93	2006	0.350	7.5	9.2	7.0	8.0	8.152	NA	NA	NA	NA	7.244	12.8	8.62	11.86	219
GTH96	2006	0.210	9.5	10.0	9.0	NA	13.546	NA	NA	NA	NA	13.601	13.0	9.30	12.72	219
GTH10	2007	0.377	5.5	11.3	5.0	8.5	0.120	4.45	0.597	NA	NA	12.588	14.3	5.25	11.85	219
GTH13	2007	0.487	4.5	11.3	3.5	6.0	19.817	5.53	0.633	NA	NA	11.185	14.1	4.83	10.95	219
GTH16	2007	0.392	6.0	9.8	4.5	8.5	6.673	4.66	0.425	NA	NA	9.541	14.7	4.70	11.49	219
GTH18	2007	0.418	6.5	15.2	5.5	9.0	13.647	4.16	0.429	NA	NA	6.594	15.1	5.20	11.93	219
GTH19	2007	0.367	6.5	10.7	5.5	9.0	4.300	4.21	0.425	NA	NA	9.216	15.0	5.67	12.15	219
GTH20	2007	0.462	5.5	18.3	4.5	7.0	1.154	5.98	0.724	NA	NA	7.896	14.7	5.25	12.44	219
GTH21	2007	0.391	6.0	13.7	5.0	7.5	5.071	4.61	0.589	NA	NA	9.569	14.9	7.17	11.98	219
GTH23	2007	0.416	5.5	9.5	4.0	6.5	1.293	5.04	0.819	NA	NA	10.119	14.5	7.17	12.24	219

Data collected for the analysis of the relationship between light attenuation and thermocline depth. Cont.

Lake	Year	Kd	TD	MaxZ	EpiZ	HypoZ	Area	DOC	TN	cDOM	Chl	perc.I.at.TD	EpiT	HypoT	LakeT	Julian
GTH24	2007	0.574	5.0	9.1	4.5	6.5	3.284	4.63	0.648	NA	NA	5.661	14.6	6.83	11.95	219
GTH76	2007	0.405	7.5	12.2	6.5	9.5	6.150	4.37	0.386	NA	NA	4.785	15.2	6.00	12.85	219
GTH80	2007	0.441	6.0	10.1	5.0	7.5	2.146	4.54	0.554	NA	NA	7.106	14.9	5.88	11.81	219
GTH81	2007	0.406	6.5	10.1	5.5	7.5	12.695	4.35	0.527	NA	NA	7.134	15.1	8.27	12.80	219
GTH 30	2008	0.863	4.0	12.4	3.0	5.5	6.825	4.21	NA	67.60	0.835	3.166	13.6	5.81	9.27	194
GTH 31	2008	0.965	3.5	12.0	2.0	5.0	2.185	3.86	NA	59.60	0.755	3.413	13.0	5.68	9.20	194
GTH 32	2008	0.701	4.5	15.7	4.0	5.5	12.904	3.56	NA	48.65	1.077	4.276	14.5	6.41	10.66	194
GTH 33	2008	0.773	4.0	13.3	3.0	5.0	4.285	3.71	NA	46.55	1.070	4.532	14.2	5.88	9.76	194
GTH 34	2008	0.829	4.5	17.4	3.5	5.5	3.583	3.53	NA	47.90	0.922	2.396	13.6	5.80	10.24	194
GTH 57	2008	0.394	5.5	21.6	4.0	7.0	29.959	2.33	NA	15.45	0.460	11.477	14.2	6.33	10.45	194
GTH 58	2008	0.615	3.5	16.2	1.0	4.5	4.083	4.13	NA	45.05	1.285	11.632	16.4	5.14	9.10	194
GTH 65	2008	0.620	4.0	16.5	2.0	5.0	4.308	3.13	NA	46.00	0.907	8.364	14.1	5.29	9.58	194
GTH 66	2008	0.544	4.0	25.9	3.5	5.0	16.630	2.12	NA	20.85	0.455	11.363	13.8	6.37	9.71	194
GTH 68	2008	0.302	5.5	27.4	5.0	7.0	77.079	1.30	NA	5.89	0.427	18.984	12.3	6.36	9.49	194
GTH 71	2008	0.471	5.5	18.3	4.5	8.0	17.107	1.36	NA	3.85	0.552	7.478	13.9	6.66	10.90	194
GTH 72	2008	0.396	5.5	36.6	4.5	7.5	19.696	1.99	NA	13.30	0.403	11.358	14.3	5.97	10.80	194
GTH 73	2008	0.627	4.5	13.7	3.0	6.5	9.865	2.31	NA	17.45	0.965	5.952	15.6	7.08	11.96	194
GTH 74	2008	0.649	4.5	11.3	1.5	5.5	5.862	2.80	NA	33.85	0.640	5.398	16.0	6.13	10.84	194
GTH 92	2008	0.617	4.0	10.4	1.0	4.5	5.565	1.67	NA	24.95	0.802	8.479	14.2	6.90	9.81	194
GTH 93	2008	0.609	5.0	9.2	4.0	6.5	8.152	2.21	NA	23.60	0.965	4.757	15.2	6.14	11.66	194

Data collected for the analysis of the relationship between light attenuation and thermocline depth. Cont.

Lake	Year	Kd	TD	MaxZ	EpiZ	HypoZ	Area	DOC	TN	cDOM	Chl	perc.I.at.TD	EpiT	HypoT	LakeT	Julian
GTH 96	2008	0.853	4.5	9.0	4.0	6.0	13.546	3.31	NA	46.65	2.135	2.150	14.9	6.71	11.31	194

F Sediment Organic Matter Survey Data

Table 24: Data collected during a survey of organic matter in the sediments of arctic lakes near Toolik Lake. Lake is the name of the lake from which the sediment cores were collected. Year is the year that the cores were collected. Depth is the relative depth within the lake where "epi" indicates the shallowest depth suitable for coring and "hypo" indicates the deepest point in the lake. Z is the depth from where the cores were collected in m. Temp is the temperature of the water in the location where the cores were collected in ^{o} C. DO is the dissolved oxygen concentration of the water in the location where the cores were collected in mg O_2 L^{-1}. PAR is the percentage of the irradiance immediately below the water surface reaching the sediment-water interface. DOC is the dissolved organic concentration of the water in the location where the core was collected. R^2 is the R^2 of the least squares linear model describing the loss of organic matter with sediment depth. Slope is the slope of the least squares linear model describing the loss of organic matter with sediment depth in (percent organic matter) cm^{-1}. Surf.OM is the estimated percent organic matter at the sediment water interface based on the y-intercept of the least squares linear model of the loss of percent organic matter with sediment depth or if the model was not significant, the mean percent organic matter of the core. Mean.OM is the mean percent organic matter content of the core. SD is the standard deviation of mean.perc.OM. Notes are comments regarding the samples. Loss is perc.OM.slope rescaled to the age of the sediments in (percent organic matter) y^{-1}. NA indicates missing data.

Lake	Year	Depth	Z	Temp	DO	PAR	DOC	R^2	Slope	Surf.OM	Mean.OM	SD	Notes	Loss
E-2	2007	epi	2.20e+00	1.44e+01	9.23e+00	NA	NA	2.2e-01	-7.17e-01	3.45e+01	3.12e+01	4.48e+00		-7.89e-02
E-2	2007	hypo	4.80e+00	6.80e+00	4.37e+00	NA	NA	7.2e-01	-1.02e+00	4.22e+01	3.76e+01	3.56e+00		-1.13e-01
E pond	2007	epi	2.10e+00	1.56e+01	7.95e+00	NA	NA	8.4e-01	-1.15e+00	4.87e+01	4.35e+01	3.70e+00		-1.26e-01
EX 1	2007	epi	2.50e+00	1.04e+01	NA	2.40e+00	1.11e+01	1.7e-01	0.00e+00	1.85e+01	1.85e+01	2.00e+00	non-linear	0.00e+00
GTH 110	2007	epi	2.00e+00	NA	NA	NA	7.70e+00	7.2e-01	-7.17e-01	1.95e+01	1.72e+01	1.91e+00		-7.89e-02
GTH 112	2007	epi	2.00e+00	1.36e+01	6.30e+00	4.40e-01	1.15e+01	1.7e-01	0.00e+00	1.74e+01	1.74e+01	6.85e-01		0.00e+00
GTH 112	2007	hypo	5.00e+00	8.60e+00	1.40e+00	0.00e+00	2.11e+01	1.1e-01	0.00e+00	1.96e+01	1.96e+01	9.32e-01		0.00e+00
GTH 114	2007	epi	2.40e+00	1.31e+01	8.43e+00	9.60e+00	8.39e+00	4.7e-01	-7.06e-01	2.83e+01	2.51e+01	3.03e+00	non-linear	-7.77e-02
GTH 114	2007	hypo	6.50e+00	6.50e+00	5.62e+00	2.40e-01	6.21e+00	4.0e-02	0.00e+00	2.92e+01	2.92e+01	8.73e-01		0.00e+00
GTH 156	2007	epi	2.00e+00	1.45e+01	8.29e+00	2.57e+01	6.10e+00	0.0e+00	0.00e+00	4.55e+01	4.55e+01	4.02e+00		0.00e+00
GTH 156	2007	hypo	4.00e+00	1.43e+01	7.87e+00	1.20e+01	6.10e+00	9.6e-01	-2.02e+00	5.60e+01	4.69e+01	6.06e+00		-2.22e-01

Data collected during a survey of organic matter in the sediments of arctic lakes near Toolik Lake. Cont.

Lake	Year	Depth	Z	Temp	DO	PAR	DOC	R^2	Slope	Surf.OM	Mean.OM	SD	Notes	Loss
GTH 98	2007	epi	2.40e+00	NA	NA	NA	5.50e+00	4.1e-01	-3.66e+00	4.21e+01	2.56e+01	1.68e+01	poor residuals	-4.02e-01
N-1	2007	epi	2.00e+00	1.40e+01	1.00e+01	3.00e+01	5.03e+00	9.0e-01	-2.16e+00	6.27e+01	5.30e+01	6.72e+00		-2.37e-01
N-1	2007	hypo	1.29e+01	3.50e+00	7.10e+00	0.00e+00	5.03e+00	7.0e-01	-1.67e+00	4.32e+01	3.56e+01	5.89e+00		-1.84e-01
NE-10	2007	epi	2.60e+00	1.06e+01	8.70e+00	5.50e+00	NA	5.4e-01	-7.05e-01	5.64e+01	5.32e+01	2.82e+00		-7.75e-02
NE-10	2007	hypo	4.00e+00	7.90e+00	6.10e+00	1.10e+00	NA	5.4e-01	5.70e-01	4.77e+01	5.03e+01	2.29e+00		6.27e-02
NE-11	2007	epi	2.00e+00	1.51e+01	1.00e+01	5.70e+01	1.09e+01	7.7e-01	-1.15e+00	7.41e+01	6.89e+01	3.87e+00		-1.27e-01
NE-3	2007	epi	3.50e+00	6.60e+00	1.33e+01	NA	7.82e+00	8.0e-02	0.00e+00	5.75e+01	5.75e+01	4.90e+00		0.00e+00
NE-8	2007	epi	1.50e+00	NA	NA	NA	8.63e+00	4.5e-01	-1.39e+00	6.90e+01	6.27e+01	6.15e+00		-1.53e-01
NE-9	2007	epi	2.00e+00	1.73e+01	1.07e+01	NA	1.05e+01	2.0e-02	0.00e+00	6.23e+01	6.23e+01	3.56e+00	outliers?	0.00e+00
NE-9b	2007	hypo	7.00e+00	3.56e+00	7.00e-01	1.00e-01	9.67e+00	1.1e-01	0.00e+00	5.09e+01	5.09e+01	3.31e+00		0.00e+00
S-10	2007	hypo	5.10e+00	6.00e+00	8.08e+00	0.00e+00	NA	0.0e+00	0.00e+00	3.10e+01	3.10e+01	6.64e+00		0.00e+00
S-11	2007	epi	4.90e+00	6.20e+00	7.30e+00	1.24e+01	NA	6.2e-01	-5.37e+00	4.46e+01	NA	NA	poor residuals, nonlinear	-5.91e-01
S-11	2007	hypo	1.09e+01	4.00e+00	2.63e+00	0.00e+00	NA	7.0e-02	0.00e+00	3.68e+01	3.68e+01	7.07e+00		0.00e+00
S-6	2007	epi	2.00e+00	1.71e+01	NA	3.08e+01	7.58e+00	8.8e-01	-2.41e+00	5.24e+01	4.16e+01	7.56e+00		-2.65e-01
S-6	2007	hypo	7.20e+00	5.70e+00	NA	0.00e+00	7.85e+00	8.8e-01	-9.20e-01	4.72e+01	4.31e+01	2.88e+00		-1.01e-01
S-7	2007	epi	2.70e+00	1.60e+01	NA	1.51e+01	3.80e+00	6.7e-01	-1.01e+00	5.24e+01	4.78e+01	3.65e+00		-1.11e-01
S-3	2008	epi	2.00e+00	9.10e+00	NA	2.09e+01	NA	8.1e-01	-2.30e+00	7.67e+01	6.64e+01	7.41e+00		-2.53e-01
S-3	2008	hypo	5.50e+00	9.00e+00	NA	2.90e+00	NA	5.9e-01	-8.10e-01	4.58e+01	4.22e+01	3.07e+00		-8.91e-02
E-4	2008	epi	2.00e+00	1.01e+01	7.40e+00	1.18e+01	NA	6.5e-01	-1.88e+00	3.90e+01	3.06e+01	6.59e+00		-2.07e-01

Data collected during a survey of organic matter in the sediments of arctic lakes near Toolik Lake. Cont.

Lake	Year	Depth	Z	Temp	DO	PAR	DOC	R^2	Slope	Surf.OM	Mean.OM	SD	Notes	Loss
E-4	2008	hypo	4.00e+00	1.02e+01	7.40e+00	4.36e+00	NA	2.8e-01	-1.17e+00	4.52e+01	3.99e+01	6.49e+00	low r2 due to a few outliers - rel is linear and slope accurate	-1.29e-01
GTH 91	2008	epi	3.00e+00	1.05e+01	NA	5.43e+00	NA	8.4e-01	-7.40e-01	2.66e+01	2.33e+01	2.39e+00		-8.14e-02
GTH 91	2008	hypo	9.90e+00	4.60e+00	NA	0.00e+00	NA	2.7e-01	-2.90e-01	2.45e+01	2.31e+01	1.69e+00	non-linear, greater slope between 0-4cm then somewhat pos slope btwn 4-9cm	-3.19e-02

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