Winter thermal structure of Lake Superior

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Abstract

Moorings equipped with thermostats that span the water column were deployed at up to seven locations throughout Lake Superior from October 2005 through May 2013. This year-round, multi-year, multi-location, full water-column record of the thermal structure reveals significant inter-annual and spatial variability in Lake Superior’s winter heat content, thermocline depth, and phenology. There is a stark contrast in thermal structure between the cold, icy winter of 2009, during which strong negative stratification formed, and the much warmer winter of 2012, during which the stratification was much weaker. Significant inter-annual and spatial variability was also observed in ice cover, which influenced heat content. Ice cover significantly inhibits heat flux between the lake and the atmosphere, and spatial variability in ice cover translates into spatial variability in end-of-winter heat content. This is found to be preserved through the spring warming season, and is strongly correlated with variability in the timing of the onset of summer stratification, with regions that have warmer end-of-winter water columns stratifying earlier than regions with colder end-of-winter water-columns. Observed spring warming rates appear to depend not on mooring depth, but on regionally averaged depth; this suggests the existence of a lateral mixing mechanism.

A primary focus of physical limnology is on understanding the thermal structure of lakes on a wide range of temporal and spatial scales. Much of the available data on this topic reflects the ease with which certain types of data are collected: the bulk of available temperature data are taken at or near the surface of lakes, and (at least for mid- to high-latitude lakes) most of the data are focused on milder seasons, because of the difficulty of collecting winter data, especially in partially ice-covered lakes. Although winter conditions in smaller ice-covered lakes are reasonably well-represented in the literature (Ellis et al. 1991; Jonas et al. 2003; Forrest et al. 2008), winter structure in larger lakes has received relatively scant attention, often consisting of a small number of data points taken from ships of opportunity (Bennett 1978; Assel 1986). Especially in the Laurentian Great Lakes, very little has been written on winter or spring thermal structure. However, it has been shown (Rodgers 1987; Austin and Colman 2007) that winter thermal structure, including the extent of ice in a given winter season, plays a significant role in determining the thermal character of large lakes throughout the subsequent year. Winter and spring thermal structure and the role of ice thus remain important topics.

Recent studies have investigated the thermal characteristics of large lakes using a variety of methods, including direct, in situ measurements (Verburg et al. 2003; Coats et al. 2006; Hampton et al. 2008), remote sensing techniques (Schneider et al. 2009; Schneider and Hook 2010), and other sources, such as historical records of ice cover (Magnuson et al. 2000). Collecting data in large lakes requires overcoming challenges not encountered when making measurements on land or in small lakes, and few time-series or other observations go back more than a few decades, (McCormick and Fahnenstiel 1999; Verburg et al. 2003; Vollmer et al. 2005). Phenomena such as surface waves and seasonal ice cover make access to the open waters of large lakes difficult during the winter. This has led to a seasonal bias in many long-term lake records, with greater availability of data from summer months than from winter. In addition, most observations of lake temperature are taken at or near the lake surface, with much less data available on subsurface thermal structure.

Although a fair amount of attention has been focused on the nearshore waters of the Keweenaw Peninsula (Smith 1973; Churchill et al. 2003), observations of thermal structure in the open waters of Lake Superior are surprisingly rare. The earliest observations of open water temperature in Lake Superior that we are aware of are observations taken by the U.S. Lake Survey in 1871, summarized by Nichols (1881), but are all summer data. As far as more modern data are concerned, compilations of year-round bathythermograph survey data were considered by both Bennett (1978) and Assel (1986), demonstrating the dimictic nature of Lake Superior and establishing the magnitude of the lake’s heat budget. Currently, regular monitoring efforts of Superior’s open waters include conductivity–temperature–depth (CTD) surveys conducted by the Environmental Protection Agency (EPA) in the late spring and summer since 1992, data from which have been used by modeling groups to ground-truth models (Bennington et al. 2010; White et al. 2012), and National Data Buoy Center surface-water temperature data, which have been used to study inter-annual variability in summer water temperatures (Austin and Colman 2007). However, neither of these data sets provides any winter temperature data.

This manuscript consists of a description of a year-round temperature data set from Lake Superior, which spans multiple locations and years. The measurements are of high temporal resolution and span nearly the entire water column, presenting unique opportunities to investigate the
thermal structure of the lake and the mechanisms that influence it. The year-round nature of the data set offers a depiction of the lake during seasons where very few data have been available, including through periods of ice cover. In addition, Lake Superior is composed of somewhat distinct eastern and western basins, and having data from multiple geographic locations during the study period provides insight into spatial variability throughout the lake. The strength of this data set lies not in revealing long-term trends, because this is not possible in the context of an 8 yr data set. Instead, the record is useful in elucidating interannual variability in the thermal structure of the lake, and for better understanding linkages between seasons.

There are two primary results from this work. Spatial and temporal variability in ice cover has a direct effect on the winter thermal structure of the lake as a whole, which plays a significant role in thermal structure throughout the following year. With several years of year-round data, we are able to characterize the extent to which variability in winter conditions, as communicated through the lake’s thermal structure, is ‘lost’ during the spring warming season, affecting the timing of the onset of summer stratification. Spring warming rates at a given location appear to be a function not of the local depth, but of regionally averaged depth, suggesting the influence of a lateral mixing mechanism. The moored thermistor data presented here addresses a major gap in our understanding of the thermal structure of large lakes.

Methods

The Lake Superior mooring array has consisted of one or more moorings deployed continuously since 2005 (Fig. 1). The Western Mooring (hereafter, WM), first deployed in 2005, along with the Eastern (EM) and Central (CM) moorings, each first deployed in 2008, form the core mooring array that is in service as of 2014. Beginning in autumn 2008, this core mooring array was supplemented by a combination of one or more outer moorings: the Far Western (FWM), Northern (NM), Southern (SM), and Far Eastern moorings (FEM). A subset of the moorings has also been equipped with Acoustic Doppler Current Profilers, discussed in Austin (2013).

Each mooring was equipped with 10–16 thermistors (depending on the local water depth), which were distributed throughout the water column. Thermistors were spaced more closely near the surface. Because of Coast Guard restrictions on moorings with no surface signature, the uppermost thermistor on each mooring was located at a depth of 10 m below the water surface. The bottommost thermistor for most moorings was located ~ 5 m above the lake bottom, with only the mooring anchor and acoustic release positioned below.

The placement of the core mooring array (WM, CM, and EM) was chosen to coincide with the locations of the National Data Buoy Center (NDBC) buoys, with the Western, Central, and Eastern moorings deployed within 2 km of NDBC buoys 45006, 45001, and 45004, respectively. These NDBC buoys record near-surface (1 m depth) temperature readings, which can be used to fill-in surface temperatures for the core mooring deployment over the summer ice-free season, when the NDBC buoys are in service. The emphasis of this data set, however, is on winter thermal structure. In the winter, the top several thermistors (over many tens of meters) would typically suggest an isothermal upper layer, so we assume that the uppermost thermistor on each mooring provides a reasonable estimate for the surface temperature. However, this could conceivably lead to error in the estimate of the heat content in persistent cold, windless conditions.

A variety of thermistor models have been used throughout the deployments, including Brancker Research (RBR) TR-1000, TR-1050, TR-1060, and TD-2050 sensors, and Seabird Electronics SBE-39 and SBE-39P sensors. The TD-2050 and SBE-39P models contain pressure sensors in addition to temperature sensors, which are used to verify...
that thermistors are deployed at the designed depth. Every mooring is equipped with pressure sensors at 10 m and 100 m depth. Measurement frequencies are dictated by the storage capacity of the instruments, with the TR-1000 and SBE sensors taking measurements every 10 min (and a response time of several minutes), and the TR-1050, TR-1060, and TD-2050 sensors taking measurements every 1 min (all with response times on the order of 3 s). All temperature data considered in this paper have been averaged over hour intervals. This reduces noise in the data and establishes a uniform time scale among moorings and thermistors. Thermistors were periodically recalibrated, with pre-calibration drifts typically $< 10$ mK, and post-calibration accuracy of $\sim 2$ mK.

Once deployed, moorings remain in continuous operation, being briefly removed once or twice per year to retrieve data and replace batteries. In autumn 2009, a faulty batch of mooring wire was used, causing the Far Western, Western, and Central moorings to collapse during that deployment period, when the wire corroded. As a consequence, there is a gap in temperature data at these moorings from winter 2009 through their spring 2010 redeployment, which is the only significant gap in thermistor data. Additionally, NDBC buoy data are not available at the Western Mooring location in 2007, so summer surface temperature estimates (and derived products, such as heat content) are not available at the Western Mooring for summer 2007.

Lake Superior ice-cover time-series were extracted from Interactive Multisensor Snow and Ice Mapping System (IMS) data, available through the National Ice Center (www.natice.noaa.gov, 2008). The IMS data are derived from satellite imagery, and are available at both 24 km and 4 km nominal spatial resolutions ($\sim 576$ km$^2$ and 16 km$^2$ per pixel, respectively), and at a daily temporal resolution. The data sets are created manually by a satellite analyst, relying first on visual imagery, and using microwave imagery (which is not inhibited by cloud cover) and data from previous days to estimate ice cover where cloud cover is present. These data depict the presence or absence of ice, with no information regarding ice thickness or density. The 24 km IMS snow and ice data set is available from February 1997 to present, and the 4 km data set is available from February 2004 to present. Lake Superior ice data were extracted from the 4 km data set, which contains a 6042 grid-point image of the lake. In addition, local ice-cover data were extracted using a 10 km radius around each mooring. The 10 km radius incorporates 21 grid points surrounding each mooring, with the intent of reducing noise in the data and providing a more meaningful estimate of ice cover in the vicinity of the mooring than would a single point. The results are not sensitive to the averaging radius.

To put these recent data in a broader temporal context, historical ice-cover data were obtained from the Great Lakes Ice Atlas, which is a compilation of ice-cover data from various data sources for the Great Lakes, beginning in 1973 (Assel 2003, 2005; Wang et al. 2012). Using these data, winter average lake-wide ice cover (December through May) was calculated for Lake Superior over this historical period.

Heat content is a metric used to quantify the relative amount of thermal energy within the water column, and its time derivative is used to quantify the rate of change occurring within the water column. If the system were purely one-dimensional (i.e., no horizontal advection, neglecting heat fluxes through the sediments) then this rate of change should be equivalent to the surface heat flux. Heat content was estimated using $H = \sum_{t=1}^{n} \rho w c_p (T_i - T_{MD}) \Delta t_i$, where $H$ is the heat content of the water column in joules per square meter, relative to the temperature of maximum density $T_{MD}$ (Chen and Millero 1986), $\rho w$ is the density of water in kilograms per cubic meter, $c_p$ is the heat capacity of water (taken to be a constant $4180$ J kg$^{-1}$ K$^{-1}$), $T_i$ are the measured temperatures in degrees Celsius, and $\Delta t_i$ is the vertical thickness assumed to be represented by each thermistor. Defining heat content relative to $T_{MD}$ makes heat content a useful proxy for estimating shifts in phenology. When heat content is rising during the spring, the time at which the heat content crosses the threshold of $0$ J m$^{-2}$ represents the time at which the water column warms above $T_{MD}$ and, thus, roughly the time at which the water column will begin to stratify. Additionally, the average annual heat-content cycle was determined using a Levenberg–Marquardt non-linear least-squares approach (Glover et al. 2011) to solve for coefficients to the equation with form $H_{ANN} = H_0 + C \cos(2\pi/1y)t + \phi$, where $t$ is time in years, $H_{ANN}$ is the average annual cycle of heat content, and $H_0$, $C$, and $\phi$ are fitted parameters, representing the annual average heat content with respect to $T_{MD}$, the magnitude of the annual signal, and the phase of the annual signal, respectively.

The latent heat associated with ice formation and melting was not accounted for in heat content calculations. It is relatively small, compared with the fluctuations in heat content occurring within the water column. Estimates of open lake ice thickness on Lake Superior are sparse, but Nghiem and Leshkevich (2007) measured open lake ice thicknesses on the order of 0–30 cm during winter 1997. The upper end of this range corresponds to a latent heat contribution on the order of $-1 \times 10^8$ J m$^{-2}$, while interannual and spatial variability in heat content are on the order of $1 \times 10^9$ J m$^{-2}$, and the annual signal is on the order of $2.5 \times 10^9$ J m$^{-2}$.

Results

The Lake Superior thermistor record shows a clear dimictic (Lewis 1983) temperature pattern throughout the lake, with periods of strong positive stratification each summer, and periods of weaker negative stratification each winter (Fig. 2). This contrasts with some recent limnology texts (Mackie 2004; Dodds and Wiles 2010; Mackie and Claudi 2010), which classify Superior as ‘warm monomictic.’ Temperatures near the bottom of the water column remain between 3° and 4°C throughout most of the year. At the peak of summer stratification, hourly averaged temperatures near the surface reach $\sim 19$–23°C, depending on the year. In the winter, surface temperatures fall below 1°C most years. Between these periods of stratification, during both spring warming and autumn cooling, the lake
experiences transitional mixing periods, during which it is effectively isothermal. This dimictic cycle is observed every year and at every mooring location in the data set.

The Western Mooring time-series is shown as an example of the data collected from the mooring array (Fig. 2). In this figure, a representative mid-depth thermistor (36–40 m depth), a bottom thermistor (120–180 m depth), and a surface thermistor (1 m depth, from nearby NDBC buoy 45006) are plotted for the duration of the Western Mooring’s deployment history. (The range in depths for the middle and bottom thermistors is due to thermistors being deployed at varying depths throughout multiple deployments.)

Over the study period, significant inter-annual and spatial variability is observed in the lake’s winter thermal structure. Of particular note is the contrast between the strong winter stratification observed during the winter of 2009, and the very weak winter stratification observed during the winter of 2012 (Fig. 3). In 2009 at the Western Mooring, for example, temperatures in the upper 40–50 m of the water column fell to roughly 0.1°C for the majority of the winter season, from late January through late March, while temperatures below that remained ~ 2.5–3.5°C. Temperatures near 0°C were sustained to even greater depths, ~ 75 m, at the Central Mooring throughout the winter of 2009; however, the duration of this relatively strong stratification was shorter at this mooring than at the Western Mooring, lasting from early February through late March. Below-average temperatures were also observed in the upper water column of the Eastern Mooring in 2009. However, unlike the Western and Central moorings, upper water-column temperatures at the Eastern Mooring only briefly approached 0°C, during early March.

In stark contrast, during the 2012 winter, all three mooring locations experienced significantly higher temperatures and much weaker negative stratification. The Eastern Mooring, in particular, barely stratified, with a maximum temperature difference of < 1.0 K across the entire water column. Likewise, the maximum temperature difference was < 1.5 K at the Central Mooring, and < 2.0 K at the Western Mooring. In 2009, all three moorings recorded temperature differences across the water column of ~ 3.0 K.

In addition to substantially colder temperatures in the upper water column in 2009 than in 2012, thermocline depths were much deeper in 2012 than in 2009. This difference was not trivial, with the thermoclines observed more than 50 m deeper at all three moorings in 2012 than in 2009. At the Eastern Mooring in 2009, where stratification was weakest, the water column was roughly isothermal for approximately the top 150 m of the water column, with only the bottom 50 m deviating slightly from that temperature.

There was very little inter-annual variability in hypolimnion temperatures during summer stratification, but there was some variability in hypolimnion temperatures during the winter. At the Western Mooring, for example, summer average temperatures at the bottom of the water column
were consistently 3.8 ± 0.1°C, consistent with the temperature of maximum density of water ($T_{MD}$) at that depth (Chen and Millero 1986), which is roughly 3.6°C. However, during winter-stratified periods, seasonally averaged hypolimnion temperatures ranged from 2.9°C (in 2007) to 3.4°C (in 2009) over the study years. Although this year-to-year variability in averaged winter hypolimnion temperature was only 0.5 K, it is worth noting, because it corresponds to a significant difference in water-column heat content. The consistent summer hypolimnion temperatures suggest that a stable thermocline forms relatively rapidly once the temperature of the water column nears $T_{MD}$ each year, thereby preventing incoming heat energy from being mixed through the bottom portions of the water column. During the winter, the greater variability suggests that it takes longer for a stable density gradient to form, allowing the water column to remain vertically mixed and isothermal to temperatures significantly below $T_{MD}$, while heat loss continues from throughout the entire water column, likely due to more energetic wind forcing in the autumn.

The average annual heat-content cycle was calculated at the Western Mooring, based on the complete history of thermistor data at that mooring (Fig. 4). Based on this calculated average heat-content cycle, the water column at the Western Mooring reached an average annual minimum heat content (relative to $T_{MD}$) of $-1.3 \times 10^9$ J m$^{-2}$ in late March and an average annual maximum heat content of $1.1 \times 10^9$ J m$^{-2}$, attained in late September. This is consistent with Bennett’s (1978) estimates of the spring heat income (heat gained in the spring beyond $T_{MD}$) and annual heat income (the difference between maximum and minimum heat content). From our observations, the water column has an average spring heat income of $1.3 \times 10^9$ J m$^{-2}$, and an annual heat income of $2.4 \times 10^9$ J m$^{-2}$, consistent with Bennett’s estimates of a spring heat income of roughly $1.5 \times 10^9$ J m$^{-2}$ and an annual heat income of $2.7 \times 10^9$ J m$^{-2}$.

Although the maximum temperature differences from $T_{MD}$ were significantly smaller in the winter, significantly more heat was lost from the water column below 0 J m$^{-2}$ than is gained above 0 J m$^{-2}$, due to the much greater thickness of the surface mixed layer in the winter. The interannual variability of heat content has some unusual characteristics: the year with the lowest winter heat content (2007) is not characterized by particularly high ice cover, and the year with anomalously high ice cover (2009) is not marked by particularly low heat content. In general, it appears that ice cover and minimum heat content are not related. Further, it appears that the highest observed summer heat contents (2009) follow the year with the highest ice cover.

Historical ice-cover data from the Great Lakes Ice Atlas show that between 1973 and 2012, there has been a clear decline in seasonal ice-cover averages (Fig. 5). There was, in general, much more ice cover in the 1970s and 1980s than there has been since the late 1990s. This is consistent with findings that the duration of ice cover is decreasing on lakes and rivers in the northern hemisphere (Magnuson et al. 2000). Of most significance to this study, 2009 was the only year of the study period where there was substantial ice cover on Lake Superior. The 2009 winter experienced the highest average ice cover of any year since 1996, with ice cover comparable to levels typical of the 1970s and 1980s. It was also the only year during the study in which the entire lake received ice cover, albeit briefly, from 02 to 06 March. The remainder of study years experienced low levels of ice cover, especially when considered in this long-term context. In particular, 2006, 2010, and 2012 received very little ice cover. In these years of especially low ice cover, ice was generally confined to bays and the perimeter of the lake basin.

There was significant spatial variation throughout the lake, in terms of the timing and duration of ice formation. Ice formation generally follows a similar progression each
Fig. 4. Water-column heat content at WM for 2005–2012, and least-squares fitted annual cycle based on the same data.

Fig. 5. Lake Superior average winter (01 December–01 May) ice cover. The year corresponds to the January–May portion of the winter. Light bar data are from the Great Lakes Ice Atlas; dark bar data are from the Ice Mapping System data set.
year, with ice cover progressing further in high-ice years than lower-ice years. The bays and basin perimeters are first to freeze. As ice cover extends into the open-water portions of the lake, the western basin of the lake generally freezes earlier than the eastern basin, with the deep part of the eastern basin being the last part of the lake to freeze. Ice-out patterns are more variable, but ice in the eastern basin of the lake generally melts earlier than ice in the western basin. As a result, the western portions of the lake generally experience a longer ice-covered season than the eastern portions.

Of the three core moorings, the Western Mooring experienced ice cover most often (Fig. 6). During the study period, the Western Mooring location became ice covered in 2007, 2008, and 2009. With the exception of a small amount of ice cover in the vicinity of the Central Mooring in 2007, 2009 was the only year in which the region around the Central and Eastern moorings iced over. The duration of this ice cover was shorter around the Central Mooring than at the Western Mooring, and was shorter yet around the Eastern Mooring. No moorings received local ice cover in 2005, 2006, 2010, or 2012; and only a minimal amount of ice was briefly present around the Western Mooring in 2011. During the study period, the Western Mooring, without exception, experienced the greatest amount of local ice cover, earliest ice formation, and longest ice duration each year. Meanwhile, the Eastern Mooring experienced the least amount of ice cover, latest ice formation, and shortest ice-covered seasons. One clear weakness of this data set is the lack of high-ice years, limiting our insight into variability among icy years.

During 2009, the Western Mooring was the first of the core moorings to ice over, and the Eastern Mooring was the last. Once the Western Mooring iced over on 23 January, it remained consistently ice-covered through 06 April—a period of 73 d. The Central Mooring was ice-covered for a period of 64 d, from 27 January to 01 April. Ice cover at the Central Mooring was less consistent than at the Western Mooring, with the percent of local ice cover dropping below 100% on several occasions throughout the winter season. The Eastern Mooring received only sporadic ice cover, which occurred toward the end of winter, between 02 March and 21 March. During this 19 d period, local ice cover at the Eastern Mooring dropped to 0% for about half of the period.

Discussion

Although the relationship between meteorology and ice cover is complex, it is reasonable to hypothesize that to first-order, high ice-cover years correspond to years with anomalously low air temperatures. However, the relationship between ice cover and the heat content of the water column is considerably less straightforward. Ice cover can reduce the heat input from shortwave radiation through increased albedo, which would result in lower water-column heat content. However, ice cover also creates a barrier at the lake-atmosphere boundary layer, thereby insulating the lake surface from the overlying atmosphere, which has the potential to inhibit turbulent fluxes, thus potentially reducing heat loss. These mechanisms are not well-characterized because of the paucity of lake temperature data available from deep ice-covered lakes. The presence of the three core moorings, each with varying degrees of ice cover throughout the winter, allows for characterization of these effects across multiple points throughout the lake, and will provide some insight into the extent to which spatial variability in heat content can occur within the lake.

During the winter of 2009, ice cover was clearly observed to inhibit the rate of change of heat content, presumably by reducing the surface heat flux (Fig. 7). This effect was most dramatic at the Western Mooring, where seasonal ice cover was greatest. Prior to ice formation, heat loss was occurring at a relatively fast rate. However, immediately following ice formation, this heat loss was inhibited to the point that
heat content remained relatively constant throughout the remainder of the ice-covered season. At the Central Mooring, where ice cover is less consistent, a similar effect is observed, but is less pronounced than at the Western Mooring. The water column at the Central Mooring continues to lose heat after ice forms, but at a significantly reduced rate. In contrast, heat flux appeared uninhibited during this timeframe at the Eastern Mooring, where ice cover was brief and sporadic.

To quantify this relationship, the average rate of change of heat content (equivalent to surface heat flux in the absence of horizontal advection) was estimated using a linear regression over the 4 week periods preceding and following the date of first ice formation for each of the three core mooring locations (Fig. 7). At the Western Mooring, the rate of change prior to ice formation was $-343 \pm 3$ W m$^{-2}$. Here and in other estimates, the uncertainty reflects both the uncertainty in our estimation of heat content (due to the discrete nature of the thermistor array) and uncertainty in the linear fit. After ice formed, the rate of change of heat content was reduced to $-336 \pm 4$ W m$^{-2}$, a reduction of $\sim 80\%$. At the Central Mooring, change in heat content was $-106 \pm 15$ W m$^{-2}$ before ice and $-124 \pm 20$ W m$^{-2}$ after ice, suggesting no statistically significant change in the rate of cooling of the water column.

Assuming these changes are due to changes in surface heat flux, the changes observed before and after ice formation cannot necessarily be attributed entirely to the presence of ice, because meteorological conditions and timing within the annual shortwave radiation cycle would also influence these rates. However, the dramatic and statistically significant reductions observed at the Western and Central moorings, occurring at the time of local ice formation, demonstrate that ice clearly has an important influence on heat flux between the lake and the atmosphere. Although it is clear that ice has a significant effect on the surface heat flux, the overall effect on the heat content of the water column is going to be strongly dependent on the timing of the ice cover. Ice cover earlier in the season might block heat loss, whereas ice cover later in the season might block heat gain.

The evidence that ice cover at the Western Mooring resulted in a warmer spring water column and an earlier onset of stratification is contrary to some recent work. Specifically, Austin and Colman (2007) found that years of high ice cover are correlated with years with later summer stratification times. Although the way in which to reconcile these conclusions is not currently clear, and thus remains an important next step, there is a mechanistic distinction between the findings of each. Here, we show that ice cover can act as a mechanism for inhibiting heat loss through sensible and latent fluxes, while Austin and Colman (2007) discuss how ice cover increases albedo and reduces heat gain through shortwave radiation. During periods of ice cover, these mechanisms are, in effect, competing, and it is unknown what is significant in determining the relative magnitude of each. Which mechanism dominates may depend on the timing of the ice cover; in particular, whether it is early-season or late-season ice. It remains to better understand what initiates the formation of ice on large lakes; specifically, to determine how a unit of heat removed from the surface of a lake (below $T_{MD}$) is partitioned into deepening the surface mixed layer, cooling the mixed layer, and the formation of ice. With only 1 yr of under-ice data available at this time of this writing, collecting additional under-ice data will be essential toward examining these mechanisms. Hopefully, data collected in the unusually high-ice winter of 2013–2014 will provide additional insight into heat exchange processes in high-ice years.

These local variations in heat content are not mixed away over the course of the winter. The Western Mooring
had a significantly higher heat content at the end of the winter season than did the Central or Eastern moorings (which both continued to experience some level of heat loss throughout the winter). The Central Mooring, which experienced a significantly reduced heat flux in the presence of ice cover, had slightly higher heat content at the end of the winter than the Eastern Mooring. Spatial variability in ice cover acted as a mechanism for spatial variability in the end-of-winter heat content.

Between the winter and summer stratified seasons, the water column vertically mixes and warms isothermally until the water column reaches a temperature near $T_{MD}$ and begins to stratify. The timing of this event plays an important role in the thermal characteristics of the following summer (Austin and Colman 2007). The date on which the water column will stratify is a function of both the initial heat content of the water column and the rate at which the water column warms. However, these parameters are not independent. All other factors being equal, a colder water column should warm at a faster rate than a warmer water column, due to equilibrative heat flux components, which are proportional to the air–water temperature difference. Therefore, variability in winter heat content is inevitably reduced during spring warming, because a lower initial heat content (and hence lower surface-water temperature) will result in a more positive heat flux. The relative significance of the variability in heat content compared with the variability of spring warming rates will ultimately determine the extent to which winter variability is preserved.

To examine this question, we will initially focus on the spring warming period following the winter of 2009, during which ice cover played a role in driving significant spatial variability in end-of-winter heat content (Fig. 8A). During this warming period, there is some variability between moorings in warming rates, which may be attributable, in part, to this variability in heat content. Specifically, the Western Mooring had the warmest water column, and warmed more slowly (226 W m$^{-2}$) than the Central and Eastern moorings (255 W m$^{-2}$ and 244 W m$^{-2}$, respectively). This variability in warming rates, however, was insufficient to overcome the variability in end-of-winter heat content over the length of the warming period. The Western Mooring stratified significantly earlier than the other two core moorings in 2009. The Western Mooring had warmed to a heat content of 0 J m$^{-2}$ by 27 June, while the Central and Eastern moorings did not reach a heat content of 0 J m$^{-2}$ until 08 July and 16 July, respectively. Based on this proxy, the Western Mooring began its summer stratified season roughly 2 weeks earlier than the other two moorings.

A similar analysis was conducted using multiple years of data from the Western Mooring, in order to assess interannual variability in warming rates (Fig. 8B). As with the analysis of spring warming rates at the core moorings in 2009, the variability in warming rates observed at the Western Mooring over multiple years may be, in part, attributable to differences in water-column temperature. The year with the warmest water column (2006) had the slowest warming rate (207 W m$^{-2}$), while the year with the coldest water column (2007) had the fastest warming rate (251 W m$^{-2}$). Again, however, over the duration of the spring warming period, this variability was insufficient to overcome the variability in end-of-winter heat content, and years with warmer end-of-winter water columns generally stratify earlier than years with colder water columns.

This result is consistent with the fact that shortwave radiation is the dominant heat-flux component during the spring warming period. Shortwave radiation is independent of water-column temperature, so it is of similar magnitude from year to year, with variability due primarily to variability in cloud cover. Lofgren and Zhu (2000) estimate the annual heat flux cycles for each of the five Laurentian
Great Lakes, and show sensible and latent heat flux to be a relatively small proportion of the total heat flux during the spring warming period. Values of clear-sky shortwave radiation calculated for the months comprising the warming period (~350–400 W m\(^{-2}\)) are roughly consistent with this, once clouds and surface albedo (Payne 1972) are taken into account.

The relationship between end-of-winter heat content, defined as the heat content on 01 May (the results here are not sensitive to the choice of this date), and the timing of summer stratification is consistent over multiple moorings and years. There is a statistically significant correlation \(r^2 = 0.86, p < 0.05\) between end-of-winter heat content and the timing of the onset of summer stratification at the core moorings (Fig. 9). Among the core moorings, the earliest stratification time was observed in 2010, when the Eastern Mooring stratified on 04 June. The latest observed stratification was at the Eastern Mooring in 2009, which stratified on 16 July, which was a difference of 42 d over the course of the study period, among the core moorings. As such, end-of-winter heat content is shown to be a strong predictor of the timing of summer stratification over a wide range of conditions. This is consistent with the findings of Rodgers (1987), who demonstrated a correlation between winter water-column temperatures and the timing of summer stratification onset in Lake Ontario.

Although the relationship between end-of-winter heat content and stratification onset time is quite strong at the core moorings, it weakens when outer moorings are included (Fig. 9). End-of-winter heat content remains an important factor in determining the time at which the water column will stratify; the \(r^2\) value for this correlation is 0.38 \((p < 0.05)\). However, the fact that this correlation is significantly weaker than when only core moorings are considered suggests that other factors, in addition to end-of-winter heat content, are important in determining the timing of summer stratification onset when considering both outer and more central portions of the lake.

The weaker correlation between end-of-winter heat content and stratification onset is due, in part, to significant spatial variability in estimated warming rates (Fig. 10). There is a greater amount of variability in warming rates when outer moorings are included, and this variability is no longer insignificant compared with the variability in heat content. In all years for which there is mooring thermistor data, the three core moorings warmed at rates below 260 W m\(^{-2}\), while the outer moorings warmed at rates above 260 W m\(^{-2}\), with the exception of the Northern Mooring in 2010, which warmed at a rate of 255 W m\(^{-2}\). In particular, warming rates at the Southern and Far Eastern moorings, for which there are 2 yr of data each, exceeded 330 W m\(^{-2}\) both years, including a warming rate of 415 W m\(^{-2}\) at the Southern Mooring in 2011.

This variability in observed warming rates is too large to be explained in terms of spatial variability in surface heat flux. Although spatial variability in meteorological forcing,
Fig. 10. Spring warming rate at all moorings in all available years. The spring period is defined as beginning 01 May and ending on the first calendar day with a daily average heat content $\geq 0 \text{ J m}^{-2}$ (a proxy for the onset of summer stratification).

Fig. 11. Spring warming rates in 2011 at all moorings, adjusted to a regionally averaged depth over a range of averaging radii.
such as cloud cover, can result in significant short-term variability in surface heat flux, it is unlikely that significantly different meteorological conditions persist among moorings throughout their entire spring warming periods. Also, the differences in apparent warming rates are consistent from year to year. In addition, spatial variability in end-of-winter heat content is not observed in the data, so the variability in warming rates must be due to something other than the initial temperature of the water column. There is no apparent correlation between a mooring’s warming rate and other potentially influential properties of the mooring location, including water-column depth, distance from shore, or latitude.

Some of the spatial variability in warming rates can likely be explained by horizontal mixing in regions of strongly varying bathymetry. Coincidentally, the deployment depths of all of the outer moorings are significantly greater than their average regional depth. In the absence of horizontal mixing, a location that is anomalously deep will warm more slowly than a shallow location. However, these bathymetry induced differences in temperature appear to get smoothed out, so that the effective warming rate is ultimately related not to the depth of a mooring location, but related to the average depth in a region around the mooring. Without this smoothing, one might expect the details of the bathymetry of the lake to be reflected in late spring surface-water temperatures.

In the case where the temperature is uniform over the water column, the rate of change of heat content can be estimated using $dH/dt = \rho_v c_p \langle dT \rangle D$, where $D$ is the local depth and $T$ the vertically uniform temperature. If, instead, an average regional depth is used, but the temperature retained, the adjusted rate of change of heat content is $dH'/dt = \rho_v c_p \langle dT \rangle D'$ where $H'$ is the adjusted heat content and $D'$ is a regionally averaged depth. It remains to be determined what is meant by ‘regionally averaged.’ Testing this for a range of averaging radii (Fig. 11), we find that the adjusted warming rate at all of the outlying moorings decreases significantly as the averaging radius is increased, with the outer moorings showing more reasonable rates of warming for radii on the order of tens of kilometers. It does not appear that there is a single value that works at all sites, and the mechanism by which this horizontal mixing occurs is unknown.

The data and results presented here are significant because they provide insight into the behavior of the lake and its response to climatic forcing during a time of year during which very little is known about the structure of the water column below the surface. In addition, these results demonstrate the interconnected nature of the physical processes examined, and that conditions experienced by the lake during one season can be articulated in the lake’s thermal structure in such a way that their effects extend into subsequent seasons.

Spatial variability in heat content imposed by ice is preserved through spring and affects the timing of summer stratification. This is important, because the timing of summer stratification has significance to biogeochemical processes, such as oxygen and nutrient distribution. It is likely that the variability in winter heat content may have implications even farther into the subsequent year, and that future research may be able to expand upon these findings to better characterize the lake’s phenology. Understanding the mechanisms through which the lake responds to variable climatic forcing, such as the relationship between ice and heat flux presented here, will be essential to predicting how Lake Superior, and other large lakes, will respond to climate change.

Acknowledgments

We thank the crew of the R/V Blue Heron for their important contributions to the collection of the data used in this manuscript, as well as two anonymous reviewers whose comments have resulted in a significantly improved manuscript.

Observations were supported by National Science Foundation division of Ocean Sciences grant OCE-0825633.

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