Novel, direct observations of ice on Lake Superior during the high ice coverage of winter 2013-2014

Daniel Titze^a and Jay Austin^{b,*}

^aLarge Lakes Observatory, University of Minnesota-Duluth, Duluth, Minnesota

^bLarge Lakes Observatory and Department of Physics and Astronomy, University of Minnesota-Duluth, Duluth, Minnesota

*Corresponding author: jaustin@d.umn.edu

Abstract

During the 2013-2014 winter, the Lake Superior basin experienced record low air temperatures and record high ice cover. We present observations from three subsurface moorings which provided a novel view of the ice cover during this extreme winter. Each mooring carried a set of thermistors, and two sites had ADCPs, providing a rare glimpse into the subsurface characteristics and behavior of a large, partially ice-covered lake, including ice drift velocity, keel size and abundance, surface gravity wave suppression, and the remobilization of ice crystals from below the ice sheet. Ice drift velocities were well-correlated with nearby wind velocities, with ice speeds approximately 4% of the wind speed and 20° to the right of the wind direction. Passing ice keels displaced the top subsurface float at the western mooring to depths of 6 m on a regular basis, and up to 11 m during one event. Anomalously large signal return strength at two ADCPs occurred sporadically at roughly the same time at two geographically separate locations, but only during periods of ice cover. The distribution of signal strength with depth suggests that these events are actually sounds from the ice cover itself, rather than reflectors in the water. Broadly speaking, ice cover on Lake Superior has more in common with oceanic ice cover than with ice cover on small lakes.

Keywords: Ice; Lake Superior; moorings, ice drift

Introduction

For mid- to high latitude lakes, ice is acknowledged as playing a vital role in a wide range of lake processes, from under-ice thermal structure (Titze and Austin 2014), the timing of summer stratification (Austin and Colman 2007), and how lakes interact with the atmosphere, including lake-effect precipitation (Cordeira and Laird 2008) and evaporation (Blanken et al. 2011). Regardless of the acknowledged importance of ice, many surface and subsurface processes of seasonally ice-covered large lakes are not well understood, due in large part to the difficulty of access during the winter months, especially on large lakes where ice may not be fixed in place.

Characteristics of ice cover, such as drift velocities and draft depths, as well as water column properties below the ice, have been extensively studied in the world's oceans (McPhee 2008, Leppäranta 2008, Ekman 1905), made possible by the availability of ice-hardened vessels, including some custom built for science, and the political and economic importance of ice-covered regions of the oceans. In freshwater systems, recent reviews of seasonally ice-covered lakes (Bengtsson 2011, Kirillin et al. 2012) and the literature that they discuss focus in large part on relatively small lakes, which tend to have land-fast ice cover; that is, lakes that freeze over completely, and whose ice is typically fast to the shore, resulting in very low water velocities below the ice and a relatively quiescent thermal structure. The imbalance in direct research of ice on small and large lakes is likely simply a matter of the logistics of access to these systems. A small number of large lakes are acknowledged to have drifting ice (Kirillin et al. 2012), such as Lake Ladoga (Burda 1999) and Lake Peipsi (Wang et al. 2006), though evidence for drifting ice on these lakes is inferred from remote sensing or modeling simulations. The single published observation of ice drift in the Laurentian Great Lakes of which we are aware is from a data report

(Campbell et al 1987), in which three transponders were deployed onto the surface of Lake Erie during the winter of 1984, showing generally eastward drift, consistent with the wind field. Lake Baikal, while large and deep, is in a sufficiently cold climate that it freezes over completely each year, and the freezing process takes from days to weeks (Shimarev et al. 1994), largely isolating the water column from direct meteorological forcing. This is in contrast to most other large, partially ice-covered lakes, for which the formation and movement of ice is dynamic throughout the entire season.

Literature addressing ice on the Laurentian Great Lakes (hereafter “the Great Lakes”) focuses almost entirely on remotely sensed data (Assel 2003, 2005; Wang et al. 2012a, Leshkevich and Ngheim 2013), consisting largely of measures of areal coverage, or modeling studies (Wang et al 2010, Fujisaki 2012, White et al. 2012). Of these three, Fujisaki (2012) includes some direct ice-thickness observations made between 1966 and 1979 (Sleator 1995), but these are all made very close to shore. The last review of Great Lakes ice research (Bolsenga 1992) addresses a wide variety of topics, but confirms the fact that there are very few direct observations of ice on the Great Lakes away from coastal, shallow settings. Surprisingly, there are few direct, in situ observations of ice or water column physics during ice cover on large lakes, the Great Lakes being no exception.

A recent paper (Titze and Austin 2014) discussed some aspects of winter and spring conditions in Lake Superior. It compared thermal conditions in a low ice year (2012) and a high ice year (2009) using data from moored instrumentation. They made observations of heat content and the impact of ice cover on rates of change of heat content, examined inter-annual variability in heat content and its relationship to ice cover and the timing of summer stratification, as well as

a careful analysis of warming rates in the spring. The current manuscript builds on these efforts, but is distinct from it in that it focuses largely on the characteristics of the ice itself.

Over the last few decades, a decline in ice cover has been observed in water bodies throughout the Northern Hemisphere (Magnuson et al. 2000, Benson et al. 2012) and has been well-documented in Lake Superior (Wang et al. 2012b, Van Cleave et al. 2014). The decades-long decline of ice cover on Lake Superior and other lakes represents a significant shift in the character these lakes. With fewer high-ice years occurring in recent decades, opportunities to collect data on ice cover in large lakes, as well as role ice plays in influencing various lake-climate processes, are becoming increasingly less common.

In contrast, during the 2013-2014 winter, the Great Lakes region experienced unusually cold temperatures and high ice cover (Clites et al. 2014, Gronewold 2015). Lake Superior in particular experienced average winter air temperatures 4°C below the long-term average, and was essentially entirely ice covered for the months of February and March, far in excess of observed coverage in previous years. This resulted in a rare opportunity to study the characteristics and behavior of the ice. In this manuscript, we present moored under-ice observations of ice drift velocity, ice keel depth and distribution, surface gravity wave suppression, and natural under-ice acoustics from Lake Superior during the cold 2013-2014 winter. Data are examined in the context of regional meteorology to analyze the manner in which lake ice responds to wind forcing. A subsequent manuscript will consider seasonal linkages and inter-annual variability, placing this extreme winter in a broader temporal context. These analyses provide a unique glimpse at the characteristics of ice on a large lake, and begin to address a gap in our understanding of the influence that ice has on surface and subsurface lake processes. In addition, as more numerical models of lakes are including ice processes, direct in-

lake observations will be important in order to ground-truth not just areal ice coverage, but other aspects of ice dynamics as well.

Methods

Three subsurface moorings were deployed in Lake Superior during the winter of 2013-2014 (Figure 1) referred to as the Western Mooring (WM), Central Mooring (CM), and Eastern Mooring (EM). WM, CM, and EM are located in distinct basins of the lake, and are positioned within 2 km of the locations of National Data Buoy Center (NDBC) buoys 45006, 45001, and 45004, respectively (Figure 1), which are typically deployed from April-November. Moorings have been deployed at the WM site continuously since September 2005, and CM and EM have each been deployed continuously since June 2008. Moorings are recovered, data recovered, instruments replaced, and moorings redeployed once or twice per year. Some data gaps exist due to instrument or mooring failure. For more information on the array, see Austin (2013) and Titze and Austin (2014).

The configuration of moorings over the 2013-2014 winter is provided in Table 1. Each mooring was equipped with a set of thermistors, which are spaced more closely in the upper portion of the water column, where the strongest temperature gradients occur. The thermistors used throughout the project have consisted of Brancker Research models TR-1000, TR-1050, TR-1060, TD-2050, and TR-SOLO, and Seabird Electronics models SBE-39 and SBE-39P. The interval at which the thermistors can take measurements over the course of the deployment is limited by their storage capacity and response time. A measurement frequency of 10 min was used for the TR-1000 and SBE sensors, which have response times of several minutes, while a

measurement frequency of 1 min was used for TR-1050, TR-1060, and TD-2050 sensors, which have response times of approximately 3 s, and a measurement frequency of 10 s was used for TR-SOLO sensors, which have response times of approximately 1 s. All thermistors were periodically recalibrated, with pre-calibration drifts generally less than 10 mK, and post-calibration accuracy of approximately 2 mK. Moorings do not extend to the surface, so that they remain safely below the draft of shipping vessels and surface ice. Due to steep gradients in bathymetry in certain regions of the lake, which can make it difficult to deploy moorings at the designed depth, WM was inadvertently deployed approximately 5 m shallower than intended during the 2013-2014 winter deployment. Thermistors at around 10m and 100m depth are also equipped with pressure sensors.

In addition to thermistors, some moorings have been, at times, equipped with Acoustic Doppler Current Profilers (ADCPs). During winter 2013-2014, WM and EM were equipped with upward-looking 300 kHz RDI Workhorse ADCPs, deployed at nominal 80 m depth, to measure currents in the upper portion of the water column. A bin size of 2 m and a measurement frequency of 20 min were used during this deployment, and the compass of each ADCP was calibrated on land immediately prior to being deployed.

Although ADCPs were initially intended to measure currents throughout upper 80 m of the water column, a clear signal was obtained off the bottom of the ice during periods of ice cover at WM, producing observations of ice movement. This technique has been used in the oceanographic community (Belliveau et al. 1990, Shcherbina et al. 2005, Björk et al. 2008). Velocity time-series were estimated for both the ice and the upper water column. Ice velocities were calculated using the average of the two 2-m bins at the top of the water column with the highest intensity signal return, which is indicative of a reflective interfacial boundary. In the

absence of ice, measurements in these bins are contaminated by the presence of surface gravity waves and are disregarded. The vertical resolution and range were not sufficient to make estimates of ice thickness.

In order to estimate the depth of the pressure sensors, the local air pressure needs to be subtracted from the measured in-situ pressure. To do this, we used air pressure as measured at the nearby CMAN station DISW3 (Figure 1). This air pressure time-series p_{air} was subtracted from mooring pressure sensor measurements p_{sensor} , and sensor depth z_{sensor} was obtained using the hydrostatic equation:

$$z_{sensor} = \frac{p_{sensor} - p_{air}}{\rho g} \quad (1)$$

where ρ is the water density (taken to be a constant 1000 kg m^{-3}) and g is gravitational acceleration (9.80 m s^{-2}).

Wind velocity time series were used to analyze the relationship between ADCP velocities and overlying wind velocities at WM. When comparing wind velocities with ADCP velocities, hourly wind data were interpolated onto the 20 minute ADCP time grid. Because NDBC open water buoys are not deployed during the ice cover period, a land-based station was used to estimate offshore winds at WM. Wind velocity time-series were obtained from station DISW3, and were compared to wind velocities at NDBC Buoy 45006 (located during the ice-free season at WM) during periods in which both stations are operating simultaneously. To evaluate the extent to which DISW3 wind velocities can be used to predict 45006 wind velocities, separate regressions were performed between the two stations for the 01 May through 31 August periods for each of the 10 years leading up to the 2013-2014 winter (2004-2013).

These ten seasons of regressions show a directional offset between the stations of $2^\circ \pm 4^\circ$, with a linear regression slope of 1.01 ± 0.03 (r^2 ranged from 0.71 to 0.79). This demonstrates that wind direction measurements between the two stations are statistically indistinguishable, and that wind direction measurements from DISW3 can be used as a direct proxy for wind direction at WM. The regressions for wind speed have a linear slope of 0.57 ± 0.03 and an intercept of $1.6 \text{ m s}^{-1} \pm 0.1 \text{ m s}^{-1}$ (r^2 ranged from 0.44 to 0.53), so that $u_{WM} \approx 0.57u_{DISW3} + 1.6 \text{ m s}^{-1}$. The significant difference in wind speeds between the two stations can likely be explained, at least in part, by the fact that the anemometer at DISW3 is located at an elevation approximately 29 m higher than at 45006. However, although wind speeds at WM cannot be directly taken from land-based DISW3, the consistent relationship over multiple seasons suggests that an empirical transfer function can be used to adequately estimate wind speed at WM using measurements from DISW3 (this relies on the assumption that the relationship between winds at these two sites does not have a strong seasonal dependence).

Ice cover data was obtained from the National Ice Center (NIC) Ice Mapping System (IMS) datasets (National Ice Center 2008) in order to help determine the times during which ice cover was present above each of the three moorings. The IMS data is available at 4 km grid spacing and daily temporal resolution, dating back to February 2004. IMS maps are manually produced by a satellite analyst, using primarily visual imagery, but incorporating microwave imagery and data from previous days when cloud cover is present. The IMS data are binary, indicating only whether ice was present or absent in each grid point, and do not give any information regarding ice thickness or the fraction of ice cover within a grid point. Local ice cover time-series were developed for each of the three moorings by averaging the 21 IMS grid points within a 10 km radius of the mooring.

Results and Discussion

Temporal and spatial extent of ice cover in 2014. The above average lake-wide ice cover observed in 2014 is due both to enhanced spatial and temporal coverage of ice on the lake. The ice cover year 2014 had historically early ice-on dates and historically late ice-off dates, as defined by the first and last days of the winter season with lake-wide ice cover above 5%. The ice-on date for the 2014 winter was 19 December, which is 24 days earlier than the 10-year average ice-on date of 12 January. Further, with an ice-off date of 22 May, ice cover in 2014 extended 41 days later into the year than the 10-year average of 11 April. Therefore, the 2014 winter ice season was over two months longer than the average season over the last 10 years.

During the 2014 winter (December 2013-May 2014), all three mooring locations experienced similarly high amounts of ice cover (Figure 2), with seasonally averaged ice covers of 100 days at WM, 106 days at CM, and 96 days at EM, which are the highest values observed at these locations within the IMS dataset. However, given that ice cover in the eastern portion of the lake has historically been much lower than in the western portion of the lake, the increase in ice cover around EM in 2014 represents a more remarkable anomaly in ice extent than that around WM. Therefore, the above-average lake-wide ice cover observed in 2014 can be attributed to an increase in ice cover throughout the lake, but particularly to unusually high ice cover in the eastern basin of the lake. The onset of summer stratification was also significantly delayed in 2014 (not shown), compared to previous years throughout the mooring record, consistent with Austin and Colman (2007). In 2014, the average day of stratification onset at WM, CM, and EM, as defined by the first calendar day during which near-surface temperatures

were consistently above 3.98°C for the entire 24-hour period, was 03 August, about a full month later than the 10-year average of 04 July. EM did not stratify until 15 August, which is the latest stratification time observed at a mooring throughout their deployment history. In contrast, the average date of stratification at the three moorings following the warm 2012 winter was 07 June. Therefore, over the course of the mooring deployment history, there is a range in stratification times of more than 2 months, which likely has important implications for lake ecology throughout subsequent seasons.

Water column thermal structure. The temperature time series from the mooring array provide a detailed view of the thermal structure of the water column throughout these extreme conditions (Figure 2). At all three mooring locations, deep surface mixed layers with temperatures just above 0°C persisted throughout most of the winter season. At WM, a layer of water with a temperature of approximately 0.1°C was observed from late January through mid April. The depth of this cold surface layer fluctuated throughout the winter season. It extended to depths between 60 and 80 m by the end of January, deepened to depths of greater than 100 m in early February, rose to between 80 and 100 m deep from late February through mid March, and to between 60 and 80 m from mid March through mid April. This slow, progressive shoaling of the winter thermocline actually represents an increase in heat content, which suggests that there is some sort of large-scale adjustment of the thermal field in the lake beneath the ice over the course of the season. Given the partially ice-covered nature of Lake Superior, wind-driven circulation is likely an important factor in this large-scale redistribution of heat. This is in contrast to smaller lakes, which tend to freeze over completely, thereby significantly limiting the influence of both surface heat fluxes and wind-driven momentum fluxes (Kirillin et al. 2012).

Similar deep surface mixed layers with temperatures of approximately 0.1°C were observed at CM and EM (Figure 2). Unlike at WM, where the depth of the surface layer was greatest in early February, the cold surface layer deepened throughout the winter at CM and EM. At CM, this approximately 0.1°C layer deepened from approximately 60 m in early February to between 80 m and 100 m in early April. As the water column at CM began to warm in early May, the surface mixed layer deepened to below 100 m. At EM, the surface mixed layer extended deeper than 40 m by late February, deeper than 60 m by early March, and deeper than 100 m for most of April. The surface layer was also slightly colder at EM than at WM or CM, dropping consistently below 0.05°C from mid-March through early April. Bottom temperatures were warmest at CM, ranging from approximately 3.3°C to 3.5°C throughout the negatively stratified season, and coldest at EM during the same period, ranging from 2.6°C to 3.3°C . At WM, the bottom temperatures were approximately 2.9°C to 3.4°C .

One noteworthy aspect of the thermal structure under the ice is the fact that ice does not appear to form until the surface mixed layer reaches temperatures below approximately 0.1°C , at which point the surface mixed layer is very deep, extending to at least 60m at all three sites. This is in marked contrast to the behavior of some smaller lakes, which can form ice soon after the lake cools below the temperature of maximum density (Ellis et al 1991, Jonas et al. 2003). In these lakes, the temperature of the bulk of the water column is just below the temperature of maximum density, and a thin molecular sublayer is located just below the bottom of the ice. Even in Lake Baikal, where the surface water cools to near 0°C before ice formation, significant stratification forms over the course of the season in the meter below the bottom of the ice (Aslamov et al. 2014). The difference in behavior between the large and small lakes is likely due to the difference in fetch and the subsequent wave climate; the difference between Superior and

Baikal may be a consequence of ice on Baikal being largely landfast, whereas ice on Superior will be shown to be predominantly drifting. Another potential source of this difference could be differences in snow cover, though we do not have data to verify this.

Ice Drift. The drift of ice, while extensively studied in the world oceans (Leppäranta 2008), has been largely unstudied in large lakes, such as Lake Superior, due in large part to the fact that few lakes are large enough to support drifting ice (Kirillin et al. 2012). Observed ice speeds from an arbitrary two-week period (Figure 3a) at WM tends to be on the order of tens of centimeters, with velocities observed up to roughly 0.6 ms^{-1} . Large spikes in the time series are likely due to brief periods of open water when a lead passed over the mooring site, and are not reliable data due to contamination by surface gravity waves. There are also periods when the velocity drops to nearly 0 ms^{-1} for many hours or even days at a time. This suggests that even in the open waters of Lake Superior (the closest land to this site is roughly 50km distant), internal stresses in the ice can occasionally be important to the force balance, bringing the ice to rest. The fact that we observe ice drift all season except during a few exceptional events in an extremely heavy ice year suggests that drifting ice is a typical feature for Lake Superior in winters with ice formation.

Using progressive vector diagrams (“PVDs”; Emery and Thomson 2001), ice velocities are shown to follow the same general patterns as wind velocities (Figure 3,4). From the time of ice formation (21 January) through the beginning of March, ice velocities are in a predominantly eastern direction, with a slight southern component for most of January, and a slight northern component for most of February. As with wind direction, there is no predominant direction of

ice movement throughout March, and ice movement is generally to the south-southwest throughout April. The consistently eastern flow of ice at WM location during the first half of the ice-covered season suggests there are regions of ice formation and ice accumulation in the western basin of the lake. Specifically, these data suggest that from late January through early February, ice must be continuously formed along the western shore of the lake, blown to the east, and likely accumulated along the Keweenaw Peninsula, to the east of WM.

Ice velocities were compared to wind velocities during times when the ice speed was greater than or equal to 0.05 m s^{-1} and less than 0.6 m s^{-1} , which corresponds to periods when ice is predominately free drifting, with lower ice speed values corresponding to periods during which the ice field was locking up, and higher values likely corresponding to periods during which there were leads in the ice field above the ADCP, thereby contaminating the return signal with noise from surface waves. Data points were formed by binning the 20-minute ADCP ice data onto the 1-hour NDBC wind time grid, and points with ice speeds greater than 0.6 m s^{-1} were discarded because they are not believed to be true measurements of ice drift.

Ice velocities in 2014 were reasonably well-correlated with wind velocities, in terms of both speed ($r^2 = 0.42$; $p < 0.01$) (Figure 5) and direction ($r^2 = 0.93$; $p < 0.01$) (Figure 6). The linear regression for direction shows that ice direction is directly proportional to wind direction, with measured ice directions approximately 20° clockwise of wind direction. This is consistent with Ekman dynamics (Pond and Pickard 1983), and with previous measurements of this offset in oceanographic settings (Leppäranta 2008). The linear regression for ice speed demonstrates a clear relationship between ice speed to wind speed in which ice speeds are, on average, approximately 4.3% those of the wind. This strong relationship can be used to understand the force balance on the ice sheet. The shear stress on a surface can be estimated as

$$\tau^s = \frac{1}{2} C_D(Re) \rho U^2 \quad (2)$$

where τ^s is the stress, C_D is a drag coefficient which is a function of the Reynolds number Re , ρ is the density of the fluid, and U is the velocity of the fluid relative to the surface. If we assume that, to first order, $U_A \gg U_I \gg U_W$, where U_A is the air velocity, U_I the ice velocity, and U_W the water velocity, then the drag of the wind on the surface of the ice is

$$\tau_A = \frac{1}{2} C_D(Re_A) \rho_A U_A^2 \quad (3)$$

where ρ_A is the density of air, and the drag of the water on the bottom of the ice sheet is

$$\tau_W = \frac{1}{2} C_D(Re_W) \rho_W U_I^2 \quad (4)$$

where ρ_W is the density of the water. If it is assumed that the force balance is dominated by the top and bottom stresses (i.e. there are no internal stresses, and the Coriolis force on the ice itself is small compared to the stresses), and the wind and ice velocities are roughly in the same direction, then the two stresses can be set equal to each other. This yields

$$\frac{1}{2} C_A \rho_A U_A^2 = \frac{1}{2} C_W \rho_W U_I^2 \quad (5)$$

This can be rearranged to form the Nansen number Na (Leppäranta 2008) which yields the ratio of the ice velocity to the wind velocity:

$$\frac{U_I}{U_W} = \sqrt{\frac{C_A \rho_A}{C_W \rho_W}} \equiv Na \quad (6)$$

Because the Reynolds numbers for the two flows can be shown to be similar, and the drag coefficients for Reynolds numbers in the relevant range are a weak function of the Reynolds number, we can assume for the sake of simplicity that the drag coefficients are the same, which yields

$$\frac{U_I}{U_A} \approx \sqrt{\frac{\rho_A}{\rho_w}} = \sqrt{\frac{1.3 \text{ kg m}^{-3}}{1000 \text{ kg m}^{-3}}} = 0.036 \quad (7)$$

This is reasonably close to the observed ratio of ice speed to wind speed of 0.043. Literature values for the Nansen number include 0.017 for Arctic ice and 0.035 for Antarctic ice (Leppäranta 2008). A similar analysis in the Bothnian Bay (Björk et al 2008) showed that internal stresses became increasingly important as the ice thickened. While no other published observations of this ratio are known to exist in freshwater systems, a modeling study (Wang et al. 2010) found wind speed to ice speed ratios on the order of 0.02 for Lake Erie; their result may be smaller due to the thicker ice found on Erie, which is more likely to become fast to the shore. The distribution of the ratio of ice speed to wind speed (Figure 7a) shows that the response of the ice is not isotropic, and in fact lower ice speed to wind speed ratios are observed during periods of NW and SE winds. These correspond to the directions in which land is closest; this suggests that internal stresses do play some role in the force balance. Likewise, a plot of when the ice is locked up (defined here as having a velocity lower than 0.05 ms^{-1}) (Figure 7b) shows that the ice can be stationary at wind stress levels as high as 0.15 N m^{-2} , which corresponds to wind speeds of just over 9 m s^{-1} .

This can be used to make a very rough estimate of the magnitude of the ice thickness.

The critical point for breakage is

$$\tau_A \ell \approx \sigma_Y = c_s P^* h \quad (8)$$

Where τ_A is the imposed surface stress, ℓ is the distance over which that stress is imposed, $c_s \sim 1$ is a calibration factor, $P^* \sim 30$ kPa is the yield strength of ice, and h the thickness (Kirillin et al. 2012). In order for the ice to lock up, the thickness needs to exceed:

$$h > \frac{\tau_A \ell}{c_s P^*} = \frac{(0.15 \text{ Nm}^{-2})(5 \times 10^4 \text{ m})}{(1)(30 \text{ kPa})} \sim 0.25 \text{ m} \quad (9)$$

This estimate is roughly consistent with previous modeling results (e.g. White et al. 2012).

Ice keels. While the moorings were not purposely equipped with instrumentation designed to monitor ice in 2014, the pressure sensors on the moorings, intended only to verify the depth of deployment, fortuitously captured observations of passing ice formations. As ice keels deeper than the subsurface mooring floats passed over the moorings, these pressure sensors documented the depth to which the top of the mooring was depressed. In particular, due to the inadvertent shallow deployment of WM over the 2014 winter, ice keels deeper than about 5 m were detectable, as this was the depth of the top mooring float, atop which the upper pressure sensor was fixed.

For this deployment, WM was equipped with two TD-2050 sensors, nominally at 6 m and 97 m, which collected data at one-minute intervals (Figure 8). The slow drift in sensor depths over the course of the season corresponds to seasonal changes in lake level, based on observations from the Grand Marais, Minnesota station (National Oceanic and Atmospheric Administration's National Ocean Service 2014). During times outside of the ice-covered period,

defined by the first and last days of the season in which a 10 km region around the mooring shows complete ice cover (21 January through 30 April), the signal from the upper sensor contains significant ‘symmetric noise’, which is qualitatively indicative of surface gravity waves (though significant aliasing effects exist; these are not true measurements of wave height). Due to the relatively short wavelengths of these waves, a pressure effect is not observed at the deeper sensor (Pond and Pickard 1983). Surface waves are observed into the beginning of the ice-covered period, during which ice is expected to be thinner and patchier. As the season progresses and more cohesive ice forms, this wave-induced noise diminishes, is absent throughout the remainder of the ice-covered period, and is observed again shortly after the offset of ice in the spring.

While the mooring is under ice, frequent quick spikes in pressure readings are observed, going only in the positive (deeper) direction (Figure 8). These jumps in sensor depth can be attributed to passing ice keels, as corroborated by corresponding smaller jumps in the depth of the lower pressure sensor. As deep ice keels pass through the mooring location, the top float of the mooring is pulled in the direction of ice movement, generating an angular displacement throughout the mooring cable. This angular displacement increases as the ice passes over the mooring, until the top float passes under the ice keel and the mooring returns to its vertical position. It follows, then, that vertical displacement of the sensors would be greatest near the surface and decrease roughly proportionally with depth, as was observed in WM pressure sensor records. The pressure record from the upper sensor shows that ice keels of about 5 to 8 m depth occurred frequently at WM throughout the winter 2014 season, and that an ice keel deeper than 11 m was observed in early March (Figure 8c).

The top floats at CM and EM were located at about 12 m depth during the 2014 winter, with pressure sensors just below, at 13 m, so only very deep ice keels could be detected. A vertical displacement of approximately 1.4 m was observed at the upper sensor at CM on 17 March, and a vertical displacement of approximately 0.2 m was observed at EM on 06 March, both of which were corroborated by corresponding smaller displacements in the deeper sensors (Figure 9). These displacements demonstrate the presence of one observed ice keel deeper than 12 m at EM, and one ice keel deeper than 13 m at CM during the winter of 2014.

To put the abundance of these keels in perspective, WM was depressed to 6 m or more for roughly 0.3% of the ice-on time period (Figure 10). If we assume that this percentage also corresponds (at least in order of magnitude) to the areal fraction of the lake where the ice was thicker than 6 m at any given time, this would suggest that at any given time, over 250 km² of the lake was composed of ice keel of 6 m or greater thickness. If we integrate the abundance curve shown in Figure 10 and assume that the rest of the lake is covered with ice of thickness 0.25m, this would suggest that roughly 15% of the ice volume of the lake is in keels. This is likely an underestimate because it does not take into account keels of less than the detectable 5.5 m thickness, which are likely to be more abundant than thicker keels. Additionally, the width of these ice keels can be estimated using ice velocity as estimated from the ADCP from immediately before and after the ice keel passes, in conjunction with the duration of the displacement. In the case of the deepest WM ice keel, observed on 10 March (Figure 8c), ice velocities (U_i) were approximately 5 cm s⁻¹ both before and after the mooring was displaced, and the duration (Δt) of the displacement was approximately 1 hour. These values can be used to make a first-order estimate of the horizontal size of these keels (L), using the simple scaling:

$$L = U_I \Delta t \quad (10)$$

While the orientation of the keel relative to its velocity would be necessary to accurately characterize its dimensions, these numbers demonstrate an upper limit of approximately 180 m on the width of that particular deep ice keel.

Natural acoustics from the ice sheet. During the period of ice cover, large acoustic events as recorded by the ADCP's Return Signal Strength Indicator (RSSI) occurred at both the WM and EM sites at roughly the same time (Figure 11). These events only occurred during periods of ice cover. Similar events have been observed in other Great Lakes ADCP records, again only during periods of ice cover (Nathan Hawley, pers. comm.). Our initial interpretation of this was that some sort of event was causing the water column to be inundated with acoustic scatterers, possibly in the form of ice crystals, as similar phenomena have been observed in oceanographic settings (Leonard et al. 2006, Dieckmann et al., 1986), although a generation mechanism for crystals in this scenario is not clear. However, we now believe these events to be a natural acoustic signal, originating from either compression or fracturing of the ice sheet itself. The distribution of unadjusted RSSI as a function of depth (Figure 12) shows that while the signal drops off with distance from the ADCP, as would be expected for sound emitted from the ADCP and reflected off of objects in the water column, during these energetic acoustic events, the signal is roughly uniform through the water column, suggesting that the source of the sound is something other than reflected acoustic energy from the ADCP. This raises the intriguing possibility of using passive acoustic techniques to study the behavior of ice sheets on large lakes.

In fact, little if any attention has been paid to natural acoustics in the Great Lakes, regardless of the source of the signal.

Conclusion

With the ice-covered season representing a significant and relatively under-studied portion of the year on the Great Lakes and with ice cover acknowledged to affect a variety of physical, climatic, and biogeochemical lake processes, it is fundamentally important that the dynamics of lake ice are understood. The observations of ice characteristics presented in this paper work toward that understanding, and are the first of their kind from the Great Lakes. When examined in the context of wind velocity, for most of the season ice drift at WM was shown to move at approximately 4% of the wind speed and about 20° clockwise of wind direction, which is consistent with measurements of free-drifting ice in the oceans. These observations are especially significant because they demonstrate that even during the 2013-2014 winter, which was characterized by record-high ice cover, ice in the open-water portions of the lake was predominately free-drifting. However, under certain wind conditions, there were periods on the order of 1-2 days when the ice field did, in fact, become sufficiently concentrated to lock up at least 50 km out from the shore, demonstrating that internal stresses in the ice field at times dominate the dynamical balance. A simple scaling using the yield strength of ice along with estimated wind stress facilitates a novel first-order estimate of ice thickness, and suggests that the minimum ice thickness would need to be on the order of 0.25 m for these lockup events to occur.

Data from pressure sensors at WM, CM, and EM demonstrate that surface waves are suppressed in the open water portions of Lake Superior in the presence of ice cover, and keels with depths upwards of 11 m were observed to displace the top floats of the moorings. Deep keels were observed at all three open water locations on the lake, and are believed to account for a significant portion of the total ice volume on the lake, likely more than 15%. Finally, RSSI data from multiple ADCPs revealed sporadic high acoustic energy events during periods of ice cover, not from sound reflected off of scatterers in the water column, but ambient sound from the compression and fracturing of the ice sheet itself.

Given the limitations of remotely sensed data, from which the vast majority of our knowledge of ice on the Great Lakes is derived, the results discussed herein begin to fill the large gap in our understanding of the ice cover on the Great Lakes. In particular, the observations of ice movement demonstrate that ice dynamics on Lake Superior fall within a regime that is characterized by a balance between free-drifting conditions and conditions in which internal stresses in the ice field dominate. As such, in order to understand and accurately model this balance, a careful and thorough characterization of ice cover on the Great Lakes is critical. We believe the observations presented in this paper represent an important step toward that understanding, and hope they are of immediate use to those studying and modeling these systems. Using these methods and results as a baseline, our understanding of ice characteristics would further benefit from additional, more focused observational campaigns that build upon these important results.

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Tables

Table 1. Mooring configuration over the 2013-2014 winter

Mooring	Location	Bottom Depth (m)	Thermistor Depths (m)	Pressure Sensor Depths (m)	ADCP Depth (m)
Western Mooring	47° 19.02' N 89° 48.52' W	182	5.5, 12, 17, 22, 27, 37, 47, 57, 78, 97, 122, 147, 177	5.5, 97	77
Central Mooring	48° 01.39' N 87° 46.01' W	255	13, 15, 20, 25, 30, 40, 50, 60, 80, 100, 200, 250	13, 94	--
Eastern Mooring	47° 32.18' N 86° 34.26' W	213	13, 15, 20, 25, 30, 40, 50, 60, 80, 100, 150, 207	13, 104	80

Figure Legends

Figure 1. Map of subsurface moorings and meteorological stations, showing the locations of the three subsurface moorings (Western Mooring [WM], Central Mooring [CM], Eastern Mooring [EM]), three NDBC buoys (West Superior [45006], Mid Superior [45001], East Superior [45004]), and the coastal CMAN NDBC meteorological station used (Devils Island, WI [DISW3]).

Figure 2. Water temperature as a function of time and depth for the (A) WM, (B) CM, and (C) EM. The 0.2°C isotherm is shown by a bold white line, and dashed vertical lines depict the beginning and end of the ice-covered period, as defined by the first and last days of the season for which the IMS dataset shows 100% local ice cover. Black tick marks along the y axes represent the depths of thermistors at each of the moorings.

Figure 3. (a) Ice cover speed (velocity magnitude) and (b) wind speed at WM during February 2014.

Figure 4. Progressive vector diagrams visualizing the predominant velocities of (a) wind and (b) the upper water column (~10-20m depth). The bold lines indicate the period during which WM was under ice cover, as defined by the period between the first and last day for which the IMS dataset shows 100% local ice cover around the mooring.

Figure 5. The linear relationship between ice drift speed and wind speed is presented. Data points (dots) were formed by binning the 20-minute ADCP ice data onto the 1-hour NDBC wind time grid. A least-squares linear regression was performed on these data points and is plotted (bold dashed line) along with the 95% prediction interval (dotted lines). Data were also binned into 0.5 m s^{-1} bins by wind speed (circles).

Figure 6. The relationship between the direction of ice drift and the direction of wind speed is shown. Directions on the axes represent the direction in which the ice or wind is headed. Data points (dots) were formed by binning the 20-minute ADCP ice data onto the 1-hour NDBC wind time grid. A least-squares linear regression was performed on these data points and is plotted (bold dashed line) along with the 95% prediction interval (dotted lines). The solid $y=x$ line is the line points would follow if ice traveled in the same direction as wind. Data were also binned into 15° bins by wind speed (circles) to demonstrate that the mean signal is linear in nature.

Figure 7. (a) The ratio of ice speed to wind speed as a function of direction. (b) Wind stress and corresponding direction for times when ice was free-drifting (light dots) and locked up (dark dots).

Figure 8. Pressure sensor depth from the (a) upper and (b) lower pressure sensors at WM. (c) Pressure sensor depth from the upper sensor on 10 March 2014.

Figure 9. Pressure sensor records from CM (for (a) upper and (b) lower) and from EM (for (c) upper and (d) lower).

Figure 10. The fraction of the ice-covered period (January 21 through April 30) for which ice keels were present at WM is shown for the detectable depth-range of ice keel depths.

Figure 11. Range-corrected RSSI from (a) WM and (b) EM for a two-week period in February, 2014. Hot colors correspond to high returned signal strength, cool colors to low signal strength.

Figure 12. Unadjusted RSSI from WM on 2 February during a high acoustic energy event, and from 8 February during a quiet period.

Figure 1

Figure 2

Figure 3

Figure 4

(a) Wind PVD

(b) Ice PVD

Figure 5

Figure 6

Figure 7

radius = wind stress (N m^{-2}); angle = direction wind stress is applied

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12