Tree-ring isotopes adjacent to Lake Superior reveal cold winter anomalies for the Great Lakes region of North America

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Tree-ring carbon isotope discrimination ($\Delta^{13}C$) and oxygen isotopes ($\delta^{18}O$) collected from white pine (Pinus strobus) trees adjacent to Lake Superior show potential to produce the first winter-specific paleoclimate reconstruction with inter-annual resolution for this region. Isotopic signatures from 1976 to 2015 were strongly linked to antecedent winter minimum temperatures ($T_{min}$), Lake Superior peak ice cover, and regional to continental-scale atmospheric winter pressure variability including the North American Dipole. The immense thermal inertia of Lake Superior underlies the unique connection between winter conditions and tree-ring $\Delta^{13}C$ and $\delta^{18}O$ signals from the following growing season in trees located near the lake. By combining these signals, we demonstrate feasibility to reconstruct variability in $T_{min}$, ice cover, and continental-scale atmospheric circulation patterns ($r \geq 0.65$, $P < 0.001$).

Trees growing in cold environments do not directly record winter conditions in their tree-rings because they are dormant during this period. Isotopic signals imprinted upon tree-ring cellulose can serve as robust climate proxies that are conventionally known to record growing season conditions. Indeed, the stable isotope composition of tree-ring cellulose has been shown to reflect canopy-integrated leaf responses to environmental drivers that are further modified by downstream ecophysiological processes. More specifically, in temperate, near-coastal locations, the primary environmental signals recorded by tree-ring carbon and oxygen isotopes include warm season temperature, vapor pressure deficit, irradiance, and cloud cover. Therefore, to our knowledge, no tree-ring signals of any type have been linked to winter temperatures or ice cover of lakes or oceans via well-understood meteorological and ecophysiological phenomena. To overcome this limitation, herein we first describe aspects of coastal “lake-effect” climate, and show how climate conditions adjacent to Lake Superior during spring and early summer are seasonally-lagged and driven by antecedent winter conditions. Thereafter we demonstrate how isotopic signals fixed in the central portion of each tree-ring appear to record winter temperature anomalies for the Great Lakes region of North America.

Lake Superior is the largest freshwater lake in the world by surface area and the third largest by volume. The resulting thermal mass produces “lake-effect” climate conditions that strongly regulate minimum and maximum air temperatures ($T_{min}$ and $T_{max}$, respectively) and other meteorological variables near the lake-land-atmosphere interface. Ice forms around the shallow perimeter of Lake Superior every winter, but the lake surface rarely freezes over completely due to the large heat storage capacity of the lake. Winter area ice cover generally peaks near the first week of March at about 40%, but thin ice forms and melts earlier compared to lags of approximately two and four weeks for medium and thicker ice, respectively. Inter-annual variation in peak ice cover across thickness classes is controlled primarily by winter minimum temperatures (Supplementary Fig. 1b).
Data from NOAA buoys deployed since 1979 reveal that Lake Superior water surface temperatures and over-lake air temperatures during summer are inversely related to ice coverage during the previous winter (Supplementary Fig. 1c). Lake-cooled air temperatures occur near the shoreline and at some distance inland, as demonstrated by spatial patterns of significant correlations between antecedent winter $T_{\text{min}}$ and warm-season $T_{\text{max}}$ during June, July and August (Supplementary Fig. 1d). The inland-advection of cool lake breezes is driven in part by greater overland warming and convection that can modify air temperature regimes up to 40 km from one of the Great Lakes. Moreover, it is notable that inter-annual variability in surface water temperatures is amplified compared to regional air temperatures, corresponding to decreases in ice cover of 0.49% yr$^{-1}$ over the same period (Supplementary Fig. 2). Amplified water temperature variability has been thought to represent ice albedo-effects interacting with the timing of stratification of lake water temperatures.

Climate responses of tree-ring carbon and oxygen isotopes are widely acknowledged to record growing season conditions, and this includes climate conditions along the US West Coast. The climate of the US West Coast is influenced by ocean temperatures, but compared to Lake Superior, ocean temperature variability is muted and influenced primarily by coastal upwelling and large-scale atmospheric and ocean interactions. Due to the strong variability in Lake Superior water temperatures, we hypothesized that by sampling stable isotope signals fixed in cellulose during the spring and early summer, trees will have recorded winter season conditions rather than growing season conditions as have been demonstrated in other coastal locations. Here we test this hypothesis using tree-ring carbon and oxygen isotopes collected from white pines (Pinus strobus L.) growing 2.3 km distance from the Lake Superior shoreline, at a location where strong summer lake-effect air temperature gradients have been documented using a high-density network of temperature sensors.

**Methods**

We sampled five dominant or codominant white pine trees 2.3 km from the Lake Superior shoreline. Each tree was approximately 130 years in age and was located on the southern shoreline of Rush Lake, on the property of the Huron Mountain Club (Fig. 1). Three 12 mm-diameter cores were collected from each tree. Tree cores were mounted, surfaced and visually cross-dated. Thereafter rings were measured with a linear encoder (Velmex Inc., Bloomfield, NY) and MeasureJ2X software (Voortech Consulting) and statistically cross-dated against local and regional near-lake cloud cover data were summarized from airports within 7 km of Lake Superior.

Likewise, regional near-lake cloud cover data were summarized from airports within 7 km of Lake Superior. We obtained residuals, from a linear regression relating summer $T_{\text{max}}$ to winter $T_{\text{min}}$ using data summarized from airports within 7 km of Lake Superior. Amplified water temperature variability has been thought to represent ice albedo-effects interacting with the timing of stratification of lake water temperatures.

Isotopic composition is expressed using "delta" notation as $\delta^{13}C$ or $\delta^{18}O = (R_{\text{sample}}/R_{\text{standard}} - 1) \times 1000$, where $R^{13}C$ or $R^{18}O$ is the molar ratio of heavy to light isotopes and $R_{\text{standard}}$ is Vienna Pee Dee Belemnite (VPDB) or Vienna Standard Mean Ocean Water (VSMOW), respectively. Wood was ground to a powder, heat-sealed within a polyester filter bag (ANKOM Technology, Macedon, NY), extracted to $\alpha$-cellulose, and weighed out on a microbalance. Carbon isotope ratios were obtained using standard high temperature combustion in a vario-Pyrocube elemental analyzer interfaced with IsoPrime/Elementar IsoPrime 100 gas phase isotope ratio mass spectrometer (IsoPrime Ltd., Manchester, UK) at the Roden laboratory at Southern Oregon University. Oxygen isotope ratios were determined by pyrolyzing $\alpha$-cellulose in an elemental analyzer (TC/EA, IsoPrime/Elementar vario-Pyrocube) and analyzing the resulting gas with an isotope ratio mass spectrometer (IsoPrime100) at the Dawson laboratory at UC-Berkeley. At these laboratories, the long-term precision is less than 0.1‰ for $\delta^{13}C$ and 0.2‰ for $\delta^{18}O$. $\delta^{13}C$ data were converted to $\Delta^{13}C$ following Farquhar (1989) as: $\Delta^{13}C = (\delta^{13}C_{\text{cellulose}} - \delta^{13}C_{\text{plant}})/(1 + \delta^{13}C_{\text{plant}}/1000)$ where $\delta^{13}C_{\text{cellulose}}$ was estimated from McCarroll and Loader (2004) through 2003 and merged seamlessly with a similarly smoothed record of annual $\delta^{13}C_{\text{cellulose}}$ from Mauna Loa, Hawaii (http://scrippsco2.ucsd.edu/data/atmospheric_co2/mlo) for 2004 to 2015.

Daily resolution temperature data, from buoys and meteorological stations near the lake were from the United States National Ocean and Atmospheric Association National Buoy Data Center (http://www.ndbc.noaa.gov/maps/WestGL.shtml) and the Canadian Government (http://climate.weather.gc.ca/). Daily resolution cloud cover data were from the NOAA Great Lakes Environmental Research Laboratory (https://www.glerl.noaa.gov/). The Big Bay and Marquette, MI meteorological stations are 15 and 55 km to the southwest of the tree-ring collection site, respectively. Both stations are located <0.1 km from Lake Superior. After summarizing temperature data by month at each site, missing and/or incomplete monthly values for either record were gap-filled using linear regressions developed between the two sites. Hereafter we define this site-averaged data set to be the "regional near-lake" mean. Likewise, regional near-lake cloud cover data were summarized from airports within 7 km of Lake Superior.
Figure 1. Images of variation in areal ice cover and cloud cover for the Lake Superior region noting the location of tree core collection and the nearest two meteorological stations. MODIS images, courtesy of the Space Science and Engineering Center (SSEC), University of Wisconsin–Madison and NASA (http://ge.ssec.wisc.edu/modis-today/), showing a year with extreme low winter ice cover on 3/11/2012 (a), extreme high winter ice cover on 3/11/2014 (b), and low cloud cover surrounding Lake Superior during the early growing season on 6/3/2010 (c). Higher resolution images (Map data: Google, Landsat/Copernicus) show the location of tree core collection on the south shoreline of Rush Lake, Michigan and the approximate locations of the nearest two meteorological stations at Big Bay and Marquette, Michigan (d,e).
Marquette and Big Bay (data not shown), to produce a warm-season temperature record independent of the effect of winter $T_{\text{min}}$ which are hereafter referred to as $T_{\text{max}}^\ast$.

Weekly ice cover estimates were obtained from the Canadian Ice Service (http://iceweb1.cis.ec.gc.ca/IceGraph/page1.xhtml). Thin, medium, and thick ice cover were averaged into a single value per year across weeks 7–12, 9–14 and 11–16, respectively, with the week of January 1$^\text{st}$ defined as week zero. This procedure combines ice cover signals while accounting for lags in the formation of medium and thick ice (Supplementary Fig. 1a).

Winter cold outbreaks across the Midwest, southern Canada and Eastern United States have been associated with low atmospheric pressure or geopotential height (HGT) anomalies, generally centered between Lake Superior and the eastern edge of Hudson Bay, Canada$^{35,36}$. The fluctuation of such winter “stationary waves” in the upper atmosphere modulates surface air temperature across the Great Lakes Region. Indeed, the “Polar Vortex” event of the 2013–14 winter easily set a new record of areal ice cover on Lake Superior since records have been kept. This included a peak coverage of 100%, with some ice remaining into early June visible from MODIS images (Fig. 1a). The winter climate over North America and particularly the Great Lakes is affected by mean-state stationary atmospheric waves, characterized by a pair of circulation features in the upper troposphere: A high-pressure ridge over western North America and a low-pressure trough over the Hudson Bay area$^{38}$, referred to as the dipole$^{36,37}$. To characterize this continental-scale circulation pattern, we obtained HGT anomaly data at 250 hPa from the two centers of this dipole: over the Northeastern Pacific Ocean and the eastern edge of Hudson Bay, Canada, respectively. The difference between these anomalies has been called the “dipole index”$^{35,36,39}$ and the HGT anomaly values forming this dipole index were derived from data with a 2.5° latitude $\times$ longitude resolution from the National Centers for Environmental Prediction/National Center for Atmospheric Research Reanalysis$^{40}$. Linear regression and multiple linear regression models were constructed and compared using the “lm” or “glm” packages in the R statistical computing environment version 3.4.3$^{41}$.

### Results

Winter $T_{\text{min}}$ the strongest driver of ice formation on Lake Superior, differs greatly from year to year, as demonstrated by MODIS images from early March that contrast ice conditions during the relatively warm winter of 2011–2012 (Fig. 1a) versus the cold and so-called “Polar Vortex” winter of 2013–2014 (Fig. 1b). In spring and summer, the thermal inertia of Lake Superior causes the lake surface to be much cooler than adjacent land surfaces, resulting in subsidence of air passing over the lake and advection of cool lake breezes inland while stabilizing the surrounding atmosphere. These effects can be visualized collectively through a snapshot of cloud cover surrounding the perimeter of Lake Superior, but essentially no cloud cover over the lake (Fig. 1c).

White pine tree-ring $\delta^{18}O$ and $\Delta^{13}C$ chronologies were characterized by series inter-correlation of 0.76 and 0.54, respectively, while expressed population signals$^{42}$ were 0.94, and 0.85, respectively. These characteristics are a hallmark of chronologies that have a robust common signal driven by climatic variability. To determine the strongest climate drivers of $\Delta^{13}C$ and $\delta^{18}O$, multiple regression analyses were conducted that included local (winter $T_{\text{min}}$ and summer $T_{\text{max}}^\ast$ from the two closest meteorological stations), regional (summer cloud cover and lake-wide peak ice cover $\%$), and large-scale variables (winter HGT and winter dipole index) for the current year (Table 1).

In contrast to most other studies of tree-ring temperature signals, only winter season variables were found to be significant and thereafter retained in the final multiple regression models for white pine MW $\Delta^{13}C$ and $\delta^{18}O$ (Table 1). Ice cover and local winter $T_{\text{min}}$ values had the strongest influence on the MW $\Delta^{13}C$ and $\delta^{18}O$, respectively, whereas HGT and dipole index also described significant variation in the same models (Table 1). The modeling process was then inverted, with the goal of identifying targets for climate reconstruction utilizing MW $\Delta^{13}C$ and $\delta^{18}O$.

All winter-season variables were strongly predicted by individual isotope chronologies and their dual-isotope combinations (Fig. 2). For the variables included in climate reconstructions (Fig. 2), variance inflation factors were low, $<2$, indicating that multicollinearity did not contribute to over-fitting the model. Across all variables there were no significant relationships with summer $T_{\text{max}}^\ast$ or summer near-lake cloud cover (Table 2). It is notable for interpreting these results that summer $T_{\text{max}}^\ast$ is the summer temperature record after having removed the impact of winter $T_{\text{min}}$ due to the lagging response of lake-effect climate (See Methods). This means that stable isotope variation did not reflect variability in regional summer climate conditions that were independent of from lake-effect climate variability. The strength of the relationships noted in Table 2 and that between ice cover and the dipole index increased only weakly through time (Supplementary Fig. 3). These results clearly indicated that,

| Variable | Model | Adjusted $R^2$ | Model vs predictand correlation | P-value |
|----------|-------|---------------|---------------------------------|---------|
| $\Delta^{13}C$ | Full model | 0.26 | 0.63 | 0.0169 |
| $\Delta^{13}C$ | Ice cover + HGT | 0.33 | 0.62 | 0.0006 |
| $\delta^{18}O$ | Full model | 0.30 | 0.65 | 0.0082 |
| $\delta^{18}O$ | Winter $T_{\text{min}}$ + Dipole Index | 0.30 | 0.58 | 0.0006 |

Table 1. Results from multiple regression analyses predicting inter-annual variation in $\Delta^{13}C$, $\delta^{18}O$ from current year climate data. The “Full model” includes all terms; Winter $T_{\text{min}}$, Summer $T_{\text{max}}$, Near-lake $T_{\text{min}}$, Near-lake cloud cover, Peak ice cover, Geopotential height (HGT) and Dipole index. Results from an alternative model, retaining only significant ($P < 0.05$) predictor variables are also listed. For models considering more than one predictor variable the absolute value of model vs predictand correlations are given. See Methods for definitions of variables.
for this species, spring and early summer conditions near the lakeshore were dominated by the influence of winter season climate signals.

Compared to models using current and antecedent year isotopic signals (Table 2), when only current year signals were considered, climate variables that operate at local to regional scales, such as winter Tmin and ice cover, were predicted somewhat better compared to larger-scale influences on winter atmospheric circulation, such as winter HGT and the winter dipole index (Supplementary Table 1). Current year \( \Delta^{13}C \) predicted variability in all of the winter climate signals reasonably well, with little additional explanatory power provided by current year \( \delta^{18}O \) for either of the winter HGT or dipole index (Supplementary Table 1). However, predictions of Tmin and peak ice cover were substantially improved by including \( \Delta^{13}C \) of the subsequent year, and predictions of HGT and the winter dipole index improved substantially by including \( \delta^{18}O \) of the subsequent year (cf. Table 2 and Supplementary Table 1). Atmospheric pressure patterns and attendant “waviness” of the jet stream produces winter weather conditions at a given location but these large scale phenomena do not directly affect isotopic signatures. Likewise, ice cover on Lake Superior is largely a product of winter Tmin (Supplementary Fig. 1b). Therefore, ultimately, winter temperatures drove tree-ring isotopic variation over time, with spatial signatures of this relationship being strongest over south-central Canada, the upper peninsula of Michigan and Northern Wisconsin (Supplementary Fig. 4).

Figure 2. White pine \( \Delta^{13}C \) & \( \delta^{18}O \) chronologies collected at the Huron Mt. Club, Michigan used to predict inter-annual variation in winter minimum temperatures (Tmin) (a), Lake Superior peak ice cover (b), regional geopotential height (HGT) centered near Hudson Bay, Canada (c) and the dipole index defining differences between HGT centers over the Northeast Pacific Ocean and Hudson Bay (d). Winter Tmin represents the mean values from Big Bay and Marquette, Michigan averaged across the previous November to the current March and reconstructed values. As in Table 2, current and subsequent year \( \Delta^{13}C \) & \( \delta^{18}O \) values were used to predict each variable for each year except for 2015, in which only current year isotope data were available.
According to carbon isotope theory, $\Delta^{13}C$ is primarily controlled by the ratio of photosynthetic assimilation rate to stomatal conductance ($A/g_\text{c}$), which results in two physiological processes which can be modified by climate variability. Tree-ring $\Delta^{13}C$ is often strongly influenced by $g_\text{c}$, as indicated by tree-ring $\Delta^{13}C$ relationships with soil moisture deficits and/or atmospheric drought stress. However, $g_\text{c}$ is unlikely to constrain $\Delta^{13}C$ near Lake Superior during spring and early summer because the air is cool and humid, the soil should still be wet from snowmelt, and precipitation $>5$ mm occurs on about 50% of days. Indeed, monthly Palmer Drought Severity Index showed no significant relationships with MW $\Delta^{13}C$ (Supplementary Table 2). Theory relating $\Delta^{13}C$ to $A/g_\text{c}$ suggests the primary remaining variable that could potentially explain inter-annual variation in MW $\Delta^{13}C$.
Scientific Reports | (2019) 9:4412 | https://doi.org/10.1038/s41598-019-40907-w

Figure 3. Anomalous patterns of winter (Nov-Feb) 250 hPa eddy geopotential height (HGT, without the zonal mean) regressed against the observed peak ice cover for Lake Superior from Fig. 2b (a), the predicted dipole index from Fig. 2d (b), and the PNA index obtained from the Climate Prediction Center/NOAA, as shadings (c). The winter-mean stationary wave eddies are overlaid in all three panels as white contours, constructed from the 1981–2010 climatology.

is how late spring and early summer air temperatures affect A, as influenced by antecedent winter conditions. Near Marquette, average air temperatures during June and July include a T min of 11.8 °C and Tmax of 22.4 °C. A responds positively across this range of air temperatures, peaking near 20 °C for most temperate species 52,53. Therefore, close to the lake, Δ13C signals from early in the growing season are fixed in the MW component of tree-rings in association with frigid previous winters. This is caused by near-lake temperatures during this period being inversely proportional to ice cover such that A is constrained by cooler Tmax adjacent to the lake26 and more strongly so after winters with colder Tmin.

Tree-ring δ18O is primarily influenced by source water isotopic signatures and the degree to which leaf water is either enriched by evaporation or influenced by atmospheric water vapor2–4,14,51. Source water isotopic signals near Lake Superior surely contributed to some of the inter-annual variation in the tree-ring δ18O patterns observed due to how temperatures influence precipitation δ18O54–56. Monthly temperatures can explain about 67% of the variation in monthly-resolution precipitation δ18O data from 1995 to 2015 located 180 km North across Lake Superior from our collection site at Sibley, Ontario (F. Longstaffe, unpublished data). White pine MW δ18O also had a strong relationship (R2 = 0.47, P < 0.001) with the same Sibley precipitation δ18O data centered on the late winter to early spring period over the previous two years (Supplementary Fig. 5). Finally, Sibley precipitation δ18O averaged across the previous November to current March was significantly correlated with HGT and the Dipole index (r > 0.49, P < 0.027, data not shown), thereby linking winter atmospheric circulation, precipitation δ18O and tree-ring δ18O. The apparent integration of late winter precipitation δ18O signals across two previous years rather than just one likely reflects the time it takes for winter snow to melt and move down the slope of the north face of Huron Mountain, to the base of the mountain where seeps and springs commonly occur along the shoreline of Rush Lake near where the white pine trees were sampled for this study. Hence, this link to conditions across two previous winters for tree-ring δ18O may not be as strong at sites with less varied local topography. On the other hand, at other sites that lack a strong influence of lateral soil water flow, we would expect the current year precipitation δ18O signals to have a relatively stronger influence on tree-ring δ18O.

Lake-effect precipitation is known to affect large spatial gradients in water isotopes, across distances of 50 kilometers or more in lower Michigan57. However, there is only a very weak relationship between winter Tmin and winter precipitation amounts from data compiled across all near-lake stations including the years 1900–2015.
In contrast, the amount of lake-effect precipitation or phases of the PNA appear to have had little influence and indirectly via leaf water enrichment (lake-effect lags in lake temperature due to previous winter conditions).

Another influence on inter-annual variation in tree-ring δ18O surely derives from leaf water δ18O dynamics during the spring and early summer. Indeed, the dominant control over tree-ring δ18O was air temperatures during the early growing season, which were in turn strongly influenced by antecedent winter Tmin and other unmeasured winter conditions producing variation in winter ice cover (Table 2). This pattern can largely be attributed to a reduction in leaf temperature and leaf water evaporative enrichment by cold and humid air adjacent to the lake, particularly under conditions where lake breezes advect cool air inland. Overall, tree-ring δ18O formed during the spring and early summer is influenced by winter temperatures via source water (late winter precipitation δ18O) and indirectly via leaf water enrichment (lake-effect lags in lake temperature due to previous winter conditions). In contrast, the amount of lake-effect precipitation or phases of the PNA appear to have had little influence on tree-ring δ18O.

Agreement between winter-mean stationary wave eddies and the wave-train patterns produced by regression maps of HGT with Lake Superior ice cover and the dipole index predicted from tree-ring Δ13C and δ18O (Fig. 3a,b) contrasts sharply with long-term mean atmospheric circulation patterns associated with the PNA pattern (Fig. 3c). Physically, this translates to the dipole pattern reflecting amplification or attenuation of the mean stationary atmospheric wave pattern99, whereas the PNA describes atmospheric circulation patterns that are comparatively shifted in space.

The potential for using tree-ring isotopes to better understand past variation in winter conditions is strongly evident from the data and analyses presented here. Both carbon and oxygen isotopic signals fixed in the MW of white pines reflected winter conditions through the lagging lake-effect climate near the shore of Lake Superior interacting with well-known ecophysiological drivers of how trees and leaves record environmental influences. Hence, there appears to be strong potential for reconstructing winter temperatures or atmospheric circulation patterns influencing this region using an intra-annual tree-ring sampling and a dual isotope approach. These same climate signals are also highly correlated with winter ice cover for Lake Superior, which raises the possibility of better understanding past ice dynamics that are a major determinant of annual evaporation and associated lake levels for Lake Superior100 and can also affect the limnological metabolism of this important ecosystem101.

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Acknowledgements

We gratefully acknowledge the support of the Huron Mountain Wildlife Foundation for providing travel funds, lodging, and access to the Huron Mountain Club property. We also thank two anonymous reviewers who provided comments that have greatly improved this manuscript. This is Western’s Laboratory for Stable Isotope Science Contribution #371. S.L.V. was supported in part by the National Science Foundation Paleo Perspectives on Climate Change awards AGS-1003050 and AGS-1003601 and the Utah Agriculture Experiment Station Project 1304.
Author Contributions
S.L.V. conceived of the research, collected data, conducted analyses and wrote the paper. S.-Y.S.W. conducted analyses and contributed to writing the paper. T.E.D. and J.S.R. oversaw stable isotope analyses and contributed to writing the paper. C.J.S. contributed to writing the paper. F.J.L. and A.A. contributed precipitation stable isotope data and contributed to editing the paper.

Additional Information
Supplementary information accompanies this paper at https://doi.org/10.1038/s41598-019-40907-w.

Competing Interests: The authors declare no competing interests.

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