Thermobaric Processes Both Drive and Constrain Seasonal Ventilation in Deep Great Slave Lake, Canada

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Abstract The contrasting roles of seasonal stratification and seasonal ventilation are among the most important attributes of lakes in defining their geochemistry, biology, and response to climate change. The physical processes that regulate these processes are especially significant in very deep lakes, where the joint effect of temperature and pressure on water density becomes important. Here we report on a year-long time series (2019–2020) of temperature from a mooring deployed in seasonally ice-covered and dimictic Great Slave Lake (614 m depth). Because the temperature of maximum density \( T_{MD} \) decreases with increasing pressure, once \( \sim 4°C \) water begins to sink, it is no longer at its depth-specific maximum density and additional processes are required to drive convection and convective mixing. The key physical mechanism governing deep-water renewal is the conditional thermobaric instability. We show here that (a) different temperature dynamics control the quantity and timing of convective renewal in the upper \( \sim 200 \) m versus deep-water renewal below; (b) fall and spring deep-water renewal events are asymmetric and linked to weather during the brief time of ice forming (melting) and surface temperatures cooling (warming) through \( 4°C \); (c) short term variability (days to weeks) related to the length of time between surface waters passing through \( T_{MD} \) and ice formation (fall) and melt (spring) shapes the subsequent thermal structure; and (d) the penetration of solar radiation through ice in early spring drives penetrative convection in the upper layer, deepening the mixed-layer, and this likely affects the timing of spring phytoplankton production.

Plain Language Summary Seasonal water stratification and exchange of heat, water and gases with the overlying atmosphere are important attributes of lakes in defining their geochemistry, biology and response to climate change. The physical processes that regulate these are especially important in very deep lakes, like Great Slave Lake (GSL) (614 m depth). Here we use data from a yearlong mooring of thermistors in GSL (2019–2020) to investigate the physical processes that are important and show that these processes likely affects the timing of spring phytoplankton production in the lake.

1. Introduction

Approximately 50% of the world’s reservoir of liquid freshwater is contained in the large deep lakes located in the temperate latitudes between \( \sim 45 \) and \( 65° \)N (see Herdendorf, 1982; Hutchinson, 1957; Messager et al., 2016); for this reason alone, the dynamics of such lakes deserve serious attention in a warming world. Of fundamental importance to lakes is their seasonal pattern of stratification (the vertical layering of water bodies) and ventilation (the convective renewal of subsurface waters), for it is these deep convection processes that most strongly control the internal redistribution of scalar properties such as heat, dissolved oxygen, nutrients, and anthropogenic pollutants (Boehrer & Schultze, 2008). A common assumption is that when dimictic lakes cool or warm through the temperature of maximum density at surface pressure (3.984°C at zero salinity) they convectively mix completely from top to bottom. It has long been recognized, however, that bottom temperatures in most deep, temperate-latitude lakes remain below \( \sim 4°C \) throughout the year (Bennett, 1978; Carmack & Farmer, 1982; Johnson 1964, 1966; Strom, 1945; Wright, 1931; Yoshimura, 1936a, 1936b). These deep layers are made stable by the fact that \( T_{MD} \) decreases with depth at a rate of 0.021°C per 100 m (Eklund, 1963, 1965). Farmer and Carmack (1981) noted that since the \( T_{MD} \) decreases with depth, an inverse stratified lake can be made unstable if the interface at the base of the upper mixed layer is lowered to a depth where its temperature matches that of maximum density at that depth. Above this depth, the interface is stable and wind must work against buoyancy during cooling; below this depth, the interface becomes gravitationally unstable and free convection can occur. Once this conditional mechanism, the so-called thermobaric instability, is triggered, water can sink to the bottom unless yet colder waters closer to the \( T_{MD} \) curve are encountered at some greater depth. A signature characteristic of such lakes, noted by Johnson (1966), and later modeled by Kay (2001), is the formation of a deep temperature...
maximum ($T_{max}$) in winter, or mesothermal layer, which separates the upper, seasonally ventilated layer from the permanently stratified layer below. In deep lakes the commonly used term, turn over, must therefore take into account the actual sequence of physical processes governing the seasonally ventilated upper layers versus the permanently stratified deep layers below. In the following we define convection above the thermobaric horizon as "seasonal convection", convection below the horizon as "deep-water renewal"; and "ventilation" as the collective processes by which waters that have recently been in contact with the surface are injected into the interior and then exported away from the source. Further, more comprehensive reviews of processes that influence mixing and convection in lakes are given by Imboden and Wüst (1995) and Bouffard and Wüst (2019).

While the theory and concept behind the thermobaric instability process is relatively straightforward, case studies demonstrating the mechanisms of whole lake ventilation in deep temperate-latitude lakes are limited. In Lake Baikal (1,621 m deep; Weiss et al., 1991) these include thermobaric instabilities (Carmack & Weiss, 1991; Schmid et al., 2008; Wüst et al., 2005), thermal bars (Shimaraev et al., 1993), hydrothermal springs (Kipfer et al., 1996), and bottom boundary layer mixing (Ravens et al., 2000). In both Crater Lake (594 m deep; Crawford & Collier, 1997, 2007) and Quesnel Lake (511 m deep; Laval et al., 2012) thermobaric instability triggered by wind-forced displacement of isotherms and internal seiching has been proposed. In Shikotsu Lake (363 m deep) Boehrler et al. (2008) noted that a distinct step in dissolved oxygen concentration near mid-depth is evidence that differing mixing regimes dominate the upper and lower layers of the lake. In a comparative study of five long Norwegian lakes, all deeper than 200 m, Boehrler et al. (2013) found that lake length was a strong determinant in setting bottom temperature; as discussed below, this is supporting evidence for the generation of thermobaric instabilities by internal seiching and/or wind-driven downwelling at the ends of the lakes. Of especial importance to the annual cycle of dimictic lakes is whether or not an ice cover forms annually, and the morphometric constraints of depth and size, as discussed in detail by Yang et al. (2020).

Great Slave Lake (614 m, Hoffman, 1987) in the Northwest Territories, the focus of our study, is classified as a seasonally ice-covered, dimictic lake, in that its surface temperatures ($T_{surf}$) pass through the temperature of maximum density ($T_{MD}$) relative to surface pressure (4°C) twice yearly. Rawson (1950) remains the most recent publication on the physical limnology of Great Slave Lake. In the following we address the question of what physical processes control the annual cycle of lake convection and ventilation in a deep, seasonally ice-cover lake. We briefly describe the geologic setting and morphological characteristics of Great Slave Lake and methods used in this study (Section 2). We then present data from CTD profiles taken from a temperature mooring deployed in Christie Bay, the deepest section of the lake and interpret these data in terms of lake ventilation processes (Section 3). We then apply these results, with reference to existing literature, to the development of conceptual models of possible deep ventilation mechanisms within the context of the annual thermal cycles (Section 4), and finally present an outlook on the response of the system’s geochemistry and biology to annual cycles and climate forcing (Section 5).

2. Methods

2.1. Study Area Description and Background

Great Slave Lake, which lies within the Mackenzie River Basin in Canada (Figure 1), is the deepest lake in North America, and the sixth deepest lake on Earth. It straddles the geologic boundary between the Canadian Shield (Mesoproterozoic ~1,27 Ga to Archean ~2,65 Ga; see Hoffman and Hall, 1993) to the east and the much younger sedimentary layers forming the Central Plains to the west. The deepest part of Great Slave Lake is in Christie Bay, where the lake bottom is 614 m deep. Christie Bay is ~500 m below present-day sea level; its morphology is shaped by faults associated with a 2 Ga transform structure (Hoffman, 1987).

Rawson (1950) described the deep, narrow East Arm as a fundamentally distinct system from the broader and shallower main basin of the adjoining Great Slave Lake. We too distinguished the very deep Christie Bay in the East Arm from other regions of Great Slave Lake, including the moderately deep West Basin (maximum depth 163 m), and the island-studded, archipelago-like passage joining the two basins (maximum depth ~320 m and estimated sill depth ~100 m). (Figures 1 and 2c). Other components of the system (e.g., the shallow North Arm, maximum depth ~ 50 m) and deep Hearne Channel and McLeod Bay (maximum depths ~300 m) were not considered. Further, we divide Christie Bay into its upper (~200 m) layer that is seasonally mixed and its lower (~200 m to bottom), with 200 m considered the transition zone (bold-dashed line, Figure 2c).
2.2. Data Collection

Temperature and conductivity profiles (CTD) were collected across Christie Bay along a north-south transect where the lake is at its deepest on 21 October 2019 using a Seabird Electronics SBE19plus (temperature resolution and accuracy of >0.0001°C and ±0.005°C; conductivity resolution and accuracy of 0.5 μS cm⁻¹ and ±5 μS cm⁻¹) (Figure 2). The measured conductivity profiles were converted to specific conductance ($\kappa_4$ (μS cm⁻¹ at 4°C)) using

$$\kappa_4 = \frac{\kappa}{1 + 0.0304 \Delta T}$$  \hspace{1cm} (1)

where $\Delta T = T - 4$ is the temperature anomaly from 4°C (Boehrer et al., 2009).
Second, to document the annual temperature cycle of Christie Bay, a full depth mooring (CB1) was deployed near the deepest position shown on Canadian Hydrographic Service Chart 6341 (62.574°N; 111.010°W (Figure 2)).

Mooring data were collected with internally recording thermistors (9 RBR Solo T at 33, 84, 135, 186, 288, 390, 492, 543, and 594 m and 1 RBR Duet TD at 12 m) with a resolution of 0.001°C and an accuracy of 0.003°C. Sensors were attached individually to mooring lines between bottom anchors and the near-surface flotation buoy. The RBR Duet TD, which also recorded depth, was positioned at the shallowest depth to detect changes in lake surface level. Measurements were recorded at 2 s intervals throughout the year. The mooring was deployed on 20 October 2019, and recovered on 22 September 2020 by employing a looped line to lasso the subsurface flotation buoy. Dates of ice formation and ice break-up were visually determined using NASA Worldview EOS-DIS imagery.

Figure 2. (a) Map of Great Slave Lake showing the location of the mooring (black diamond) and the outline of the larger scale map shown in (b) (black rectangle). (b) Shows the locations of CTD stations in October 2019. (c) Shows the thalweg profile of the lake along a line drawn to join the deepest points along the length of the lake thus joining Christie Bay in the East Arm and the main lake near the Mackenzie River outflow (shown as a black dotted line in (a). (d) Shows a south-north depth profile across Christie Bay passing through its deepest point. The locations of the thermistor mooring and the 2019 CTD stations are also shown as vertical lines. Vertical lines in (c) denote boundaries between the West Basin (WB), the Archipelago Islands (AI) and Christie Bay (CB). The sill between CB and the AI is also indicated, although the exact sill depth is uncertain. The blue dashed horizontal line in (c) denotes the depth of the thermobaric horizon, as described in the text.
3. Observations

3.1. Vertical Temperature Structure Across the Bay

Subsequent to the deployment of the mooring, temperature and conductivity profiles were taken by CTD across Christie Bay, on 21 October 2019 (Figure 3). Also plotted, for comparison, are temperature profiles taken with reversing thermometers by Rawson (1950) in July–August 1946 and 1947. Also shown is the temperature of maximum density ($T_{MD}$), temperature of maximum stability proposed by Eklund (1965) ($T_{MS_E}$), and 4°C are shown with dashed gray lines. Horizontal lines show temperature ranges observed near the bottom in Quesnel Lake by Laval et al. (2012), and in Great Bear Lake by Johnson (1966). (b) Specific conductance ($\kappa_4$) profiles for the 20 October 2019 CTD casts shown in (a).

We plotted the transect results with temperature profiles taken by Rawson (1950), using reversing thermometers in July–August 1946 and 1947 as comparison to our data (Figure 3a). These measurements were congruent with our own (Figure 3a). The temperature profiles from July 1947 showed the lake to still be in inverse stratification above the 200 m inflection point. Below this depth profiles were virtually unchanged from the previous summer. These observations support the claim that the lake maintains a weak but permanent stratification below the thermobaric horizon at roughly 200 m (cf. Boehrer et al., 2008). Also plotted for comparison are historical...
The temperature data collected by the full depth mooring was plotted as isotherm depth versus time plot from moored recording thermometers in Christie Bay from 21 October 2019 to 22 September 2020: (a) isotherms plotted as temperature, and (b) isotherms plotted as temperature minus temperature of maximum density (T–T_{\text{MD}}), where T_{\text{MD}} is at the respective depths. The white rectangle denotes the wintertime T_{\text{max}} layer and corresponding wintertime T–T_{\text{MD}} minimum layer.

**Figure 4.** Isotherm depth versus time plot from moored recording thermometers in Christie Bay from 21 October 2019 to 22 September 2020: (a) isotherms plotted as temperature, and (b) isotherms plotted as temperature minus temperature of maximum density (T–T_{\text{MD}}), where T_{\text{MD}} is at the respective depths. The white rectangle denotes the wintertime T_{\text{max}} layer and corresponding wintertime T–T_{\text{MD}} minimum layer.

Deep-water temperatures from Great Bear Lake (Johnson, 1966), and Quesnel Lake (Laval et al., 2012) (Figure 3a). Note all curves are close to or slightly to the right of the maximum stability proposed by Eklund (1965; Figure 3a).

### 3.2. Annual Thermal History (Dates of Ice Cover and Ventilation)

The temperature data collected by the full depth mooring was plotted as isotherm depth versus time plot from moored recording thermometers in Christie Bay over the year (Figure 4). Both temperature (T) (Figure 4a), and observed temperature minus the temperature of maximum density at the corresponding depth (T–T_{\text{MD}}; Figure 4b) were plotted over depth. These plots clearly show the development of a cold upper layer between the dates when the surface temperatures transition through 4°C to cool in November and then warm again in July (Figure 4). Between these dates of inverse stratification, a persistent mid-depth temperature maximum (T_{\text{max}}) (Figure 4a) and a corresponding (T–T_{\text{MD}})_{\text{min}} (Figure 4b) formed between interpolated depths of roughly 160 and 260 m. Waters below these features remained permanently stratified throughout the year. This boundary between the upper, seasonally mixed layer and the lower, thermobarically stratified layer will further be referred to as the thermobaric horizon.

Data from individual recording thermometers on the mooring, depicting T and T–T_{\text{MD}} through time (Figure 5), at both full (Figures 5a and 5c) and finer (Figures 5b and 5d) temperature scales, helped better determine the dates of the thermobaric horizon formation. In the fall, the lake surface temperatures cooled through 4°C and entered inverse stratification on 3 November 2019. Development of inverse stratification in the upper water column continued as ice began to form on Christie Bay on 22 November, finally reaching full ice cover on 7 December 2019 (Figures 5a and 5c). Convective cooling was evident in the 135 m temperature recorder, while waters below the sensor at 186 m remained permanently (e.g., thermobarically) stratified (Figures 5b and 5d). Warming in spring began with penetration of solar radiation through the ice in mid-May, and continued through the date of open water around 26 June 2020. Upper water column temperatures warmed through 4°C on 15 July, followed by the onset of summer stratification (Figures 5a and 5c). For convenience in climate applications, we define the time interval between the passages of surface temperatures through 4°C as the length of winter (also see Carmack et al., 2014).

In this case the length of winter was determined to be 257 days, and the corresponding length of full ice cover was 201 days. We emphasize, however, that these dates apply only to Christie Bay, as ice breakup comes earlier in the main lake, especially off the north-flowing Slave River, and later in McLeod Bay, the last region on Great Slave Lake for break-up (cf. Rawson, 1950).

In contrast to the upper water column, temperatures below ~186 m did not cool and warm through the temperature of maximum density, but rather remained thermobarically stratified throughout the year (Figures 5, 6b and 7b). Permanent stratification and increased static stability below this depth was shown by the increase in T–T_{\text{MD}} with depth. Importantly, these deep-water records did show enhanced temperature fluctuations in the approximately two-month period following ice cover formation in winter and warming through 4°C in spring (Figures 4 and 5). These fluctuations, demonstrate evidence of active ventilation by lateral intrusions, spreading either along isolines of density, or as a near-bottom density plume.

### 3.3. Fall Ventilation of the Upper Layer

At the time of the mooring deployment on 21 October 2019, the upper water column was fully mixed to the depth of the 33 m sensor, from which lower temperatures progressed rapidly with depth (Figure 6a). Wind-forced turbulent mixing first warmed, then fully entrained water to 84 m on 26 October, and reached the 135 m sensor on 30 October 2019. The entire upper 135 m subsequently passed through 4°C between 31 October and 1 November. In contrast, throughout this period the water temperature at 186 m remained relatively constant, after which a
progressively deepening mixed-layer reached this depth on 4–5 November (Figure 6). The overall cooling rate during this period was ~0.1°C day⁻¹. That the observed mixing and cooling through fall was progressively offset in time with depth confirmed the importance of downward, vertically driven, free and forced convection.

On 6 November the upper water column to 186 m was homogenous to within 0.15°C while the temperature of waters below 186 m remained relatively unchanged, with temperature fluctuations on the order of 0.2°C. However, from 7 November the upper water column entered inverse stratification (fall transition, Figure 6a). As this progressed, and the lake became inversely re-stratified, sensors 33 m at first, and then 88 m, again displayed internal seiche-like temperature variability. Re-stratification continued through 22 November, as ice began to form on Christie Bay, and ended 7 December when Christie Bay was fully frozen (Figure 6a). This 30-day time between fall transition and full ice cover emerged as the critical period of fall ventilation, during which wind was able to act directly on the underlying and weakly stratified water. The fall transition thus progressed in two stages: first the vertically driven stage of convective mixing followed by restratification in the upper ~200 m until complete ice cover, and then, as described further in Section 3.5, a laterally intrusive stage affecting the water column below ~200 m. Throughout, these two layers are vertically separated at the thermobaric horizon.
3.4. Spring Ventilation of the Upper Layer

The progression of events through spring ventilation was very distinct from that of fall ventilation in several important ways (Figures 6b–6d). Spring ventilation was initiated as solar radiation first penetrated the snow and ice cover, warming the underlying water towards the $T_{MD}$, and providing potential energy for convection in a diurnally active and deepening mixed layer. This process has previously been described by Farmer (1975a), Farmer and Carmack (1981), Mironov et al. (2002), and Ulloa et al. (2018). Between 1 May and 27 June 2020, the upper mixed-layer progressively deepened, successively entraining underlying layers, and warming the mixed layer at a rate of 0.036°C day$^{-1}$ (Figure 6b). Diurnally forced convection was evidenced during this time (Figure 6c). On 27 June, coincident with the initial ice break-up, presumed increases in solar radiation and more efficient wind-forced mixing, the rate of warming doubled to 0.070°C day$^{-1}$. Warming continued at this rate until 20 July.
whereby the upper layers warmed through 3.66°C and began the process of restratification (Figure 6d). As seen during the fall, the 24-day time period between the loss of full ice cover and the initiation of re-stratification emerged as the critical period of spring ventilation. During this time both solar radiation and the wind were able to act directly on the underlying inversely and weakly stratified water column, effectively driving penetrative convection. At this time the bottom temperature was 3.54°C (Figures 4 and 7). Importantly, 3.66°C was determined to be the temperature of maximum density at a depth of 160 m, while 3.54°C is the temperature of maximum density at 220 m (Figures 4 and 7a). We thus could identify 160 and 220 m as the upper and lower boundaries of the thermobaric horizon. Above this layer a seasonal, vertically driven turnover took place in spring 2020, while below, renewal required lateral intrusions initiated by waters at least as cold as 3.54°C, which were depressed to a depth of 220 m, and thus triggering the thermobaric instability. Such displacement was not directly observed at the CB1 mooring, which was located at mid-basin, and so we concluded was likely to have occurred at the edges or ends of the lake, in areas where internal seiche heights are amplified or where wind-driven coastal downwelling can occur (cf. Schmid et al., 2008). However, confirmation of this supposition must await additional mooring deployments in the deeper regions (>220 m) at the ends of Christie Bay.

Figure 7. Expanded scale plots deep water temperatures for (a) T versus time for several depths, (b) T–T_{MD} versus time, and (c) corresponding fluctuations (dT/dt°C/day) from 288 to 594 m depth. Note, in (c), each plot of dT/dt is shifted up by 0.025°C day^{−1}. The red dashed line in (a) illustrates T = 3.567°C between 20 October 2019 and 1 February 2020, dropping to 3.54°C until 1 August 2020 when it dropped to 3.485°C.
3.5. Deep Ventilation

Ventilation of waters below the thermobaric horizon was distinct in timing and process from the seasonal pattern described above (T and T – T_{MD} in Figures 7a and 7b, respectively). Deep waters first warmed, beginnings on 2–5 December owing to wind-driven, downward mixing of warmer upper-layer waters near the date of freeze up in Christie Bay, followed by cooling through mid-December and January. Throughout this time substantial fluctuations in T of order 0.05°C day^{-1} were observed (Figure 7c). These fluctuations reflect periods of active ventilation, and were interpreted here as a signature resulting from active ventilation by lateral intrusions. In fall the amplitudes of these fluctuations were large at mid-depth and decreased with increased depth, while in spring the opposite was true, with the largest fluctuations occurring near the bottom (Figure 7c). Active ventilation of waters below the thermobaric horizon occurred under full ice cover, and lagged waters above the thermobaric horizon by approximately a month. This process also extended over two months longer for the deeper waters. The sharp drop of ~0.04°C at 594 m and at 543 m in early February suggests the arrival of an intrusive, near bottom plume. Throughout the remainder of winter, deep temperatures were relatively quiescent (Figure 7). An abrupt cooling of 0.05°C in early August, subsequent to ice break-up, signaled the start of spring ventilation. As in the case of fall ventilation, spring ventilation in the deep layer lags that the upper layer by over a month, and begins only after the removal of the ice cover and exposure to wind forcing (Figures 4–7).

3.6. Heat Budget

Schertzer et al. (2003) and Rouse et al. (2003) combined extensive meteorological and in-lake data to compute heat budgets for the west basin of Great Slave Lake; we compliment their work by calculating the heat content in Christie Bay. The net effect of surface heat exchange is stored within the lake: the corresponding heat content of the lake per unit area, H, can be calculated from

\[ H = \int_{d_1}^{d_2} \rho c_p T(z) dz, \]  

where \( \rho = 1.000 \text{ kg m}^{-3} \) is the water density and \( c_p = 4184 \text{ J kg}^{-1} \text{ K}^{-1} \) is the specific heat of water and d1 and d2 are the upper and lower depth limits of interest. In Equation 2 we take 0°C as our reference temperature, but we acknowledge that the freezing temperature decreases by 0.0753°C per 100 m, so that the freezing temperature at the bottom of Great Slave Lake is ~0.451°C. Therefore, we can also calculate the heat content per unit area relative to \( T_{MD} \) \( H_2 \) as

\[ H_2 = \int_{d_1}^{d_2} \rho c_p (T(z) - T_{MD}(z)) dz. \]  

Heat content calculated using Equations 2 and 3 from the mooring data was calculated for the upper layer (12–186 m, Black curves Figure 8) and lower layer (186–594 m; gray curves Figure 8). The annual range in heat content H (Equation 2) for the upper layer was 1.79 \times 10^9 \text{ J m}^{-2}, derived from a winter (December 2019) minimum value of 1.61 \times 10^9 \text{ J m}^{-2} at the time of full ice cover and a summer (September 2020) maximum value of 3.40 \times 10^9 \text{ J m}^{-2}. In fall the transition from positive to inverse stratification (surface passing through 4°C) occurred on 7 November (Figure 8), after which heat loss progressed at a rate of ~3 \times 10^7 \text{ J m}^{-2} \text{ day}^{-1}. The sharp dip and recovery in heat content seen immediately prior to full ice cover is likely an artifact of internal seiching and relaxation (Figure 8). Subsequent to full ice cover (after 7 December) the heat content was smooth but with a slight upwards slope until mid-May, when solar radiation began to penetrate the ice cover and warm the underlying water. Following ice break-up and prior to the upper layer starting to restratify, the heat content changed at a relatively constant rate of 0.2 \times 10^6 \text{ J m}^{-2} \text{ day}^{-1} (Figure 8).

In contrast, the range in heat content below 200 m was much smaller, ranging between a minimum of 6.09 \times 10^9 \text{ J m}^{-2} observed in September 2020 and a maximum of 6.22 \times 10^9 \text{ J m}^{-2} just prior to full ice cover in December (Figure 8). This yielded an annual range of 0.13 \times 10^9 \text{ J m}^{-2}, which is an order of magnitude smaller than that of the upper layer. In fall, large fluctuations in heat content (of order 5 \times 10^7 \text{ J m}^{-2}) were observed between the date of surface water cooling through \( T_{MD} \) and late January. Then, in the spring between the time of initial break-up and the upper layer warming through \( T_{MD} \) the heat content of the lower layer dropped sharply.
indicative of deep ventilation (Figure 8). We therefore conclude that the ventilation patterns in the upper 200 m are largely de-coupled from those in the lower layer (Figure 10).

4. Discussion

4.1. The Vertical/Lateral Distinction

We found the convective renewal processes in deep Christie Bay could be partitioned into distinct fall and spring forcing periods, and subsequently into upper- and lower-layer responses. Above the thermobaric horizon, at roughly 200 m, the fall and spring ventilation were driven vertically by free and wind-forced convection. Complete ventilation (turnover) occurred twice yearly, as observed in classic dimictic lakes, and scalar properties (e.g., heat, dissolve oxygen, nutrients, etc.) are redistributed twice yearly. The continual deepening of isotherms in fall, and restratification in spring both exhibit phase-locked patterns, with shallow events preceding deeper events, indicative of vertical mixing. This process, however, only extended to the thermobaric horizon, below which additional processes involving downwelling are required.

Waters below the thermobaric horizon near 200 m decrease monotonically in temperature with depth and remain permanently stratified. A vertical potential temperature profile of uniform 4°C is neutrally stable, as is a vertical temperature profile that follows the $T_{\text{MD}}$ curve; it follows that a profile of maximum stability should exist between. (Note that at temperatures where the coefficient of thermal expansion $\alpha = 0$ the difference between in situ and potential temperature is small (cf. Farmer, 1975b; Wüest et al., 2005)). Eklund (1965) argued that maximum static stability is attained when the slope of the temperature profile lies mid-way between 4°C and $T_{\text{MD}}$ ($T_{\text{MS}}$).
This stability curve, along with historical data from other deep lakes, including Great Bear Lake (Johnson, 1966), Quesnel Lake (Laval et al., 2012) and the 2019 profiles from Christie Bay are shown in Figure 3. All temperature-depth profiles are close to or slightly to the right of the maximum stability proposed by Eklund (1965). The tightness of observed deep water temperatures between the two lines of zero stability has been noted by Boehrer et al. (2013), but he also ascribes the importance of lake length and associated processes over maximum depth in determining bottom water temperatures. This envelope suggests that thermobaric stratification likely constrains the rate of warming of deep water under climate change.

4.2. Conceptual Models of Deep Ventilation

Effective ventilation of deep waters must occur in Christie Bay, as oxygen data collected by Rawson (1950) reveal percent saturation values above 80%. The arrival of laterally spreading intrusions is seen in the Christie Bay temperature record at depths below about 200 m. For accurate modeling to follow, it is necessary to identify the source of such intrusions. Two mechanisms may explain the deep ventilation observed in Christie Bay. The first hypothesizes that wind stress is sufficient to depress the colder upper layer to its thermobaric compensation depth at one end of the basin. For this we calculate the Weijlderburn number, We, which is the ratio between the surface layer depth, $h_s$, and the longitudinal tilt of the equilibrium thermocline for an applied wind stress

$$W_e \sim \frac{g \Delta \rho h_s^2}{L u_f^2},$$ (4)

where the wind shear velocity, $u_f$, is the square root of the wind stress divided by water density ($1,000 \text{ kg m}^{-3}$), $L$ is the wind fetch (taken here as the east-west dimension of Christie Bay, ~80 km), and $g' = g \Delta \rho$ is the reduced gravity with $g = 9.81 \text{ m s}^{-2}$ being the acceleration of gravity and $\Delta \rho = \rho_2 - \rho_1$ with $\rho_1$ and $\rho_2$ being the water densities in the upper and lower layers, respectively. Importantly, We decreases and wave amplitude increases, as either $L$ increases or $\Delta \rho$ decreases. Winds measured at Lutselk'e village along the south shore of Christie Bay (Figures 1 and 2) are of order 6–10 m s$^{-1}$ during fall overturn. Taking a two-layer approximation for temperature structure on 1 December relative to $h_1 = 200m$, then $T_1 = 2.0^\circ \text{C}$; $T_2 = 5.8^\circ \text{C}$; $g' = 3.12 \times 10^{-4} \text{ m s}^{-2}$ (with density calculated from Chen & Millero, 1986). Applying Equation 4 and a wind speed of 10 m s$^{-1}$ shows that internal displacements well exceeding 100 m are theoretically possible at lake ends, sufficient to trigger the thermobaric instability (Figure 9a). An analogous situation may occur through wind-driven downwelling at the lake's boundaries (cf. Schmid et al., 2008). In either case, the newly ventilated waters are likely formed near the perimeter or ends of the basin, sinks, and then spread laterally as gravity flows into the interior.

We propose an additional mechanism for deep-water renewal based on the joint effects of two nonlinearities in the equation of state for fresh water. The first is the differential compressibility of water (cold water is more compressible than warm water) which results in the thermobaric instability; the second is differential contraction on mixing (the cabbeling instability, cf. Eckel, 1948; Fofonoff, 1956; Foster, 1972) which drives the well-known thermal bar circulation. The classic thermal bar occurs during spring (fall) transition when shallower near-shore waters warm (cool) through 4°C more rapidly than deeper offshore waters. This results in a convergent frontal zone where offshore and onshore waters mix and contract to form surface water at 4°C, which subsequently sinks (Kay et al., 1995; Rogers, 1965). A similar phenomenon, the riverine thermal bar, can occur at spring and fall transition when incoming river waters warm (spring) and cool (fall) more rapidly than the lake, again resulting in a convergent frontal zone maintained by contraction on mixing (Carmack, 1979). This vertical flow draws down upper layer waters from both the colder and warmer sides of the sinking plume. In deep lakes, however, the classic thermal bar circulation may be modified by differential compressibility. Because $T_{\text{mb}}$ decreases with depth, the “core” of the vertically sinking water within the thermal bar plume will shift progressively to colder temperatures with depth as the plume descends (cf. Kay, 2001). In some cases, waters on the cold side of the convergent flow can eventually be depressed below their thermobaric compensation depth, thus triggering the thermobaric instability and initiating deep convection (Figure 9b). We term this the thermobaric thermal bar. The depth to which this mechanism reaches will also depend on the temperature gradient below the thermobaric horizon. We also note that all river-derived waters sinking below the basin outflow sill depth must be balanced by an upwards velocity, in the manner of a filling box (cf. Killworth and Carmack, 1979).
It is noteworthy that the deep and bottom temperatures measured by Rawson (1950) over 70 years earlier were virtually unchanged from present day measurements presented for this study despite climate change over the intervening years, and are within the range of annual variability observed in 2019–2020 (Figure 3). Similar results were noted by Boehrer et al. (2008) in their comparison of data collected in deep, caldera lakes in Japan between 1915 and 1931 with contemporary measurements. These observations suggest that thermobaric processes constrain the response of deep ventilation processes to climate warming.

Figure 9. Conceptual models of deepwater ventilation by (a) wind-driven downwelling or internal seiche set-up in fall, and (b) convection within the proposed thermobaric thermal bar in spring. Block vertical arrows depict surface heat flux. Dashed lines at ~200 m depict the thermobaric horizon.
4.3. Under Ice Considerations

The ice-covered period in Christie Bay currently extends over roughly two thirds of the year, and so under-ice processes are important to observe if reliable models are to be validated (cf. Ulloa et al., 2018) and chemical-biological consequences understood (Hampton et al., 2017; Yang et al., 2020). Especially significant is that ice cover currently remains on the East Arm well past the date of summer solstice, which means that much of the annual supply of solar radiation is either reflected, used to melt snow and ice, and/or used to drive penetrative convection. The latter, penetrative convection, begins in late spring, when the snow overlying the ice thins sufficiently for solar radiation to penetrate the ice. Our observations show that a progressively deepening upper layer forms until the ice melts and the upper layer is warmed to the $T_{MD}$ corresponding to the depth of interfacial entrainment; in our case, the depth of the thermobaric horizon. In Christie Bay in 2020 this began on 1 May when mixed-layer deepening was first observed, continued through to 27 June when ice began to break up, and ceased on 20 July when the upper layers warmed through 3.66°C and began the process of restratification (Figure 6). Potential bio-geochemical consequences of under-ice processes are discussed below.

5. Outlook

Rawson (1950), in his opening paragraph, states:

Great Slave Lake occupies a place of special interest among the large lakes of the world. It lies in a region for which limnological studies are as yet almost lacking; its cold climate and great size and depth place it close to the ultimate in oligotrophic condition; and it had in 1944, the further attraction of an unexploited fish population.

Rawson's goal was not only to estimate the lake's fisheries potential, but also to understand the fundamental physical conditions and processes that regulate biological productivity. Over 70 years later we support Rawson's view and speculate on processes by which seasonal ice cover, convection and restratification may affect primary
productivity and thus fisheries potential. We also concur with Hampton et al. (2017) that serious deficiencies exist in our knowledge of the ecology of lakes with seasonal ice cover.

Based on our observations of physical processes we suggest that light-limited primary production in very deep lakes will experience two growing windows during the spring convection and restratification period. Both growing stages are extensions of the critical depth hypothesis of Sverdrup (1953) which posits that spring phytoplankton blooms occur when the convective mixed-layer is shallower than a critical depth where growth by photosynthesis exceeds loss by respiration, generally at some multiple of the 1% light level. Both stages further presume sufficient nutrients exist following winter ventilation. The first begins when sufficient snow melts to allow solar radiation to penetrate the ice and into the water below and, dependent on phytoplankton species occupying the euphotic zone, slows or ends when convection deepens the mixed layer below the critical depth. Initially, under-ice convection can trigger early growth of phytoplankton (Jewson et al., 2009; Vehmaa and Salonen, 2009; Yang et al., 2017, 2020), as the convective thermals mix nutrients close to the ice and keep phytoplankton in the photic layer (Kelley, 1997). In the analogous ocean case, Backhaus et al. (2003) used model results and field observations to argue that well-structured convective flows, rather than turbulence, accounts for both the transport of phytoplankton resting stages to the surface throughout late winter, and for the early phases of the spring bloom, thus allowing active chlorophyll at greater depths than would be predicted by the Sverdrup model. Differential warming leading into shallower waters may also generate cellular, lateral circulation patterns which may also aid in maintaining phytoplankton within the euphotic zone (Kenney, 1996). Then, subsequent to breakup but prior to surface water warming past $T_{MP}$, incoming solar radiation will drive increasingly deeper mixing, reaching to the depth of the thermobaric horizon. This will reduce prolonged exposure to light, potentially removing phytoplankton from the mixed layer (i.e., export losses) and suppressing primary production (Bennett, 1978). The second growing stage is presumed to begin when surface waters warm past $T_{MP}$ and begin to re-stratify, thus setting conditions for a second bloom. Under-ice time series of solar radiation, chlorophyll fluorescence and dissolved oxygen (a by-product of net primary productivity) are required to understand this distinction.

From the perspective of geochemical responses, observations by Rawson (1950) and later summarized by Brunskill (1986) demonstrate the importance of viewing the entire lake in light of its different regions and reaches, as well as its source rivers (Figure 1) as each of these regions may exhibit widely differing properties. In particular, the differing geochemical properties of river source waters in the Great Slave Lake result in major water property differences between the West Basin and East Arm. For example, rivers draining the central plains carry much higher sediment and total dissolved solids than those draining the Canadian Shield. Hence the waters of the West Basin are far more turbid that those of the East Arm, thus impacting the underwater light climate. From a macrolimnological perspective, waters with higher concentrations of total dissolved solids tend to support higher levels of primary productivity, a concept resulting in the so-called morphoeconomic index (Northcote and Larkin, 1956; Ryder, 1982). Thus, we expect large regional differences in oligotrophic states, carrying capacity, and response to acidification across the lake. For Christie Bay in particular, observations collected over the annual cycle illustrate periods of ventilation by lateral intrusions that extend the “active” period of lake ventilation processes well after full ice cover (Figures 7 and 10). The time-lag between the deep-water renewal below the thermobaric horizon and the waters above effectively extends the period of oxygen renewal in deep waters, with possible implications for nutrient regeneration, oxygen depletion, and the remineralization cycle in the sediments (Hampton et al., 2017). Additional observations of the annual cycles of nutrients, dissolved oxygen, and pH will lend more insight into the impacts of seasonal ventilation processes on the Great Slave Lake ecosystem.

The unique value of lakes as sentinels of climate change has been noted by many authors (for example see Adrian et al., 2009; Livingstone, 2003; Williamson et al., 2008, 2009), while O’Reilly et al. (2015) emphasize the importance of landscape, morphometry and other climate filters to a given lakes’ response to climate warming. Similar responses related to upper layer warming, increased stratification, and mixed layer deepening responses that have been noted present in the ocean (Sallée et al., 2021). Carmack et al. (2014) examined deep dimictic lakes lying along a South-to-North climate gradient in Western Canada, and proposed six metrics that describe the annual thermal cycle of dimictic lakes: (a) summer maximum and winter minimum mean temperatures, (b) the dates of maximum and minimum heat content, (c) the lengths of summer and winter stratification, (d) the timing of the spring and fall turnovers, (e) the timing of the onset and break-up of ice cover, and (f) the rate of response to atmospheric heating and cooling (in MJ m$^{-2}$ day$^{-1}$). We speculate that the response of each of these metrics will be very different for the eastern and western regimes of Great Slave Lake. We also hypothesize that thermobaric
deep-water ventilation and considerations of maximal deep water stability act to constrain secular climate change. Support for this was seen in the comparisons between deep temperatures in Christie Bay in 1947 versus 2019, as well as deep temperatures in Great Bear Lake (Johnson, 1965, 1966) and Quesnel Lake (Laval et al., 2012), where deep temperatures lie midway between 4 °C and the T_{MD} profile (Figure 3a).

The observations presented here, based on a single deep-water mooring and lake-wide data collected over 70 years ago, provide new insight into the annual thermal cycle of deep, seasonally ice-covered high-latitude lakes (Figure 10). However, they also leave many questions unanswered which require further investigations. Sufficient along-lake profile data are required to determine whether the conditions for thermobaric instability could be accounted for by vertical displacements elsewhere in the lake. The relative roles of dissolved solids and suspended particles on the ventilation process should be determined. Additional measurements should include chlorophyll fluorescence, dissolved oxygen, pH, turbidity and nutrients to characterize the biogeochemical context of the lake and inform further study of the food web and trophic dynamics. Under-ice CTD, solar radiation and fluorescence data collected before ice break-up are required to better define the onset of stratification and associated biological responses. The annual cycle of snow and ice cover (timing and thickness) should be monitored for each region of the lake (e.g., Figure 2; West Basin, North Arm, Archipelago Islands, McLeod Bay), and used to ground-truth satellite observations (cf. Howell et al., 2009). Further observations in other regions of the lake, especially over a longer time, are required to confirm that the key mechanisms driving the formation, maintenance, and evolution of stratification observed in Christie Bay are applicable elsewhere in Great Slave Lake. Finally, similar observations from deep lakes laying north of Great Slave Lake (e.g., Great Bear Lake and Hazen Lake) are required to evaluate individual lake responses to climate change along a south-north climate gradient. Major changes may be expected should cold deep lakes transition to colder (cold monomictic) or warmer (ice-free dimictic or warm monomictic) mictic states, highlighting again the need for the monitoring of high-latitude lakes along climate gradients.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

The 2019–2020 data set used in the present analysis is available in the Arctic Research Foundation’s database at articfocus.org/database (https://doi.org/10.48760/vkr5-n231 and https://doi.org/10.48760/vrde-1494).

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References

Adrian, R., M ’Oléry, C., Zagarëse, H., Baines, S. B., Hessen, D. O., Keller, W., et al. (2009). Lakes as sentinels of climate change. Limnology & Oceanography, 54(6), 2283–2297. https://doi.org/10.4319/lo.2009.54.6_part_2.2283
Backhaus, J. O., Hégésth, E. N., Wehde, H., Irigoién, X., Hatten, K., & Logemann, K. (2003). Convection and primary production in winter. Marine Ecology Progress Series, 251, 1–14. https://doi.org/10.3354/meps251001
Bennett, E. B. (1978). Characteristics of the thermal regime of Lake Superior. Journal of Great Lakes Research, 4, 310–319. https://doi.org/10.1016/s0380-1337(78)72200-8
Boehrer, B., Fukuyama, R., & Chikita, K. (2008). Stratification of very deep, thermally stratified lakes. Geophysical Research Letters, 35, L16405. https://doi.org/10.1029/2008GL034519
Boehrer, B., Fukuyama, R., Chikita, K., & Kikukawa, H. (2009). Deep water stratification in deep caldera lakes Ikeda, Toya and Shikotsu. Limnology, 10, 17–24. https://doi.org/10.1007/s10070-008-0257-1
Boehrer, B., Golmen, L., Løvik, J. E., Rahn, K., & Klaveness, D. (2013). Thermobaric stratification in very deep Norwegian freshwater lakes. Journal of Great Lakes, 39, 690–695. https://doi.org/10.1016/j.jglr.2013.08.003
Boehrer, B., & Schultzze, M. (2008). Stratification of lakes. Reviews of Geophysics, 46, RG2005. https://doi.org/10.1029/2006RG000210
Bouffard, D., & Wüst, A. (2019). Convection in lakes. Annual Review of Fluid Mechanics, 51, 189–215. https://doi.org/10.1146/annurev-fluid-010518-040506
Brunskill, G. J. (1986). Environmental features of the Mackenzie system. In R. B. Davis, & K. F. Walker (Eds.), The ecology of river systems (pp. 435–471). Dr W. Junk Publishers.
Carmack, E. C. (1979). Combined influence of inflow and lake temperatures on spring circulation in a riverine lake. Journal of Physical Oceanography, 9, 422–434. https://doi.org/10.1175/1520-0485(1979)009<0422:ciolt>2.0.co;2
Carmack, E. C., & Farmer, D. M. (1982). Cooling processes in deep, temperate lakes: A review with examples from two lakes in British Columbia. Journal of Marine Research, 40(suppl), 85–111.
Carmack, E. C., Vagle, S., Morrison, J., & Laval, B. (2014). Space-time proxy for climate change in deep lakes in the Canadian cordillera: Seasonality along a latitudinal climate gradient. Journal of Great Lakes Research, 40(3), 608–617. https://doi.org/10.1016/j.jglr.2014.06.007
Carmack, E. C., & Weiss, R. F. (1991). Convection in Lake Baikal: An example of the thermobaric instability. In P.-C. Chu, & J.-C. Gascard (Eds.), Deep convection and deep water formation in the ocean (pp. 215–228). Elsevier. https://doi.org/10.1016/b978-08/700069-2
Chen, C. T. A., & Millero, F. J. (1986). Thermodynamic properties for natural waters covering only the limnological range. *Limnology & Oceanography, 31*(3), 657–662. https://doi.org/10.4319/lo.1986.31.3.0657

Crawford, G. B., & Collier, R. W. (1997). Observations of a deep-mixing event in Crater Lake, Oregon. *Limnology & Oceanography, 42*, 299–306. https://doi.org/10.4319/lo.1997.42.2.0299

Crawford, G. B., & Collier, R. W. (2007). Long-term observations of deepwater renewal in Crater Lake, Oregon. *Hydrobiologia, 574*, 47–68. https://doi.org/10.1007/s10750-006-0245-3

Eckel, O. (1948). Über die Mischungsarbeit von stabil geschichteten Wassermassen. *Archiv für Meteorologie, Geophysik und Bioklimatologie, 1*, 264–269. https://doi.org/10.1007/bf02247177

Eklund, H. (1963). Fresh water: Temperature of maximum density calculated from compressibility. *Science, 142*, 1457–1458. https://doi.org/10.1126/science.142.3598.1457

Eklund, H. (1965). Stability of lakes near the temperature of maximum density. *Quarterly Journal of the Royal Meteorological Society, 142*, 1457–1458.

Farmer, D. M. (1975a). Penetrative convection in the absence of mean shear. *Quarterly Journal of the Royal Meteorological Society, 101*, 869–891. https://doi.org/10.1002/qj.49710143011

Farmer, D. M. (1975b). Potential temperatures in deep freshwater lakes. *Limnology & Oceanography, 20*, 634–635. https://doi.org/10.4319/lo.1975.20.4.0634

Farmer, D. M., & Carmack, E. C. (1981). Wind mixing and restratification in a lake near the temperature of maximum density. *Journal of Physical Oceanography, 11*, 1516–1533. https://doi.org/10.1175/1520-0485(1981)011<1516:WMARIA>2.0.CO;2

Fofonoff, N. P. (1956). Energy transformations in the sea. *Fisheries Research Board of Canada. Manuscript. Report. Series* (Vol. 109).

Foster, T. D. (1972). An analysis of the cabbeling instability in sea water. *Journal of Physical Oceanography, 2*, 294–301. https://doi.org/10.1175/1520-0485(1972)002<0294:AAOTCI>2.0.CO;2

Hampton, S. E., Galloway, A. W. E., Powers, S. M., Ozersky, T., Woo, K. H., Batt, R. D., et al. (2017). Ecology under lake ice. *Ecology Letters, 20*, 98–111. https://doi.org/10.1111/eclet.12699

Herdendorf, C. E. (1982). Large lakes of the world. *Journal of Great Lakes Research, 8*, 379–412.

Hoffman, P. & Hall, L. (1993). *Slave Craton and Environments* (Vol. 2559). GSCOF. doi.org/10.1017/s0022112095004265

Hoffman, P. F. (1987). Continental Transform Tectonics—Great Slave Lake Shear Zone (Ca. 1.9 Gu): Geology (Vol. 15, pp. 785–788). https://doi.org/10.1130/0091-7613(1987)015<785:CTTGSL>2.0.CO;2

Jewson, D. H., Granin, N. G., Zhdanov, A. A., & Gnatovsky, R. Y. (2009). Effect of snow depth on under-ice irradiance and growth of *Aulacoseira baicalensis* in Lake Baikal. *Aquatic Ecology, 43*, 673–679. https://doi.org/10.1007/s10238-009-9267-2

Johnson, L. (1965). Physical and chemical characteristics of Great Bear Lake. *Journal of the Fisheries Research Board of Canada, 32*, 1971–1337. https://doi.org/10.1139/f75-20.4.0634

Johnson, L. (1966). Temperature regime of deep lakes. *Science, 144*, 1336–1337.

Johnson, L. (1968). Physical and chemical characteristics of Great Bear Lake. *Journal of the Fisheries Research Board of Canada, 32*, 720–730. https://doi.org/10.1139/t75-20.4.0634

Kay, A. (2001). Thermobaric flow. *Dynamics of Atmospheres and Oceans, 34*, 263–289. https://hdl.handle.net/2134/679

Kay, A., Kuiken, H. K., & Merkin, J. H. (1995). Boundary layer analysis of the thermal bar. *Journal of Fluid Mechanics, 303*, 253–278. https://doi.org/10.1017/s0022112095004265

Kelley, D. (1997). Convection in ice-covered lakes: Effects on algal suspension. *Journal of Physical Oceanography, 27*(7), 201–217. https://doi.org/10.1175/1520-0485(1997)027<201:CTICIE>2.0.CO;2

Kipfer, R., Aeschbach-Hertig, W., Hofer, M., Hofmann, R., Imboden, D. M., Baur, H., et al. (1996). Bottomwater formation due to hydrothermal activity in Frolikha Bay, Lake Baikal, eastern Siberia. *Geochimica et Cosmochimica Acta, 60(6)*, 961–971. https://doi.org/10.1016/0016-7037(95)00448-3

Laval, B. E., Morrison, J., Vagle, S., Potts, D., James, C., Foreman, M., et al. (2012). The joint effects of thermal, riverine and wind forcing on dimictic fjord lakes: A case study. *Journal of Great Lakes Research, 38*(3), 540–549. https://doi.org/10.1016/j.jglr.2012.06.007

Livingstone, D. M. (2003). Impact of secular climate change on the thermal structure of a large temperate central European lake. *Climatic Change, 57*, 205–225. https://doi.org/10.1023/A:102219503144

Messager, M., Lehner, B., Grill, G., Nedeva, I., & Schmitt, O. (2016). Estimating the volume and age of water stored in global lakes using a geo-statistical approach. *Nature Communications, 7*, 13603. https://doi.org/10.1038/ncomms13603

Mironov, D., Terzhevik, A., Kirillin, G., Jonas, T., Malm, J., & Farmer, D. (2002). Radiatively driven convection in ice-covered lakes: Observations, scaling, and a mixed layer model. *Journal of Geophysical Research, 107*(C4), 3032. https://doi.org/10.1029/2001JC000892

Northcote, T. G., & Larkin, P. A. (1956). Indices of productivity in British Columbia lakes. *Journal of the Fisheries Research Board of Canada, 13*(4), 515–540. https://doi.org/10.1139/f56-032

O’Reilly, C. M., Sharma, S., Gray, D. K., Hampton, S. E., Read, J. S., Rowley, R. J., et al. (2015). Rapid and highly variable warming of lake surface waters around the globe. *Geophysical Research Letters, 42*, 10773–10781. https://doi.org/10.1002/2015GL066235

Rawson, D. S., Kocics, O., & Wiest, A. (2000). Small-scale turbulence and vertical mixing in Lake Baikal. *Limnology & Oceanography, 45*(1), 159–173. https://doi.org/10.4319/lo.2000.45.1.0159

Rawson, D. S. (1950). The physical limnology of Great Slave Lake. *Journal of the Fisheries Research Board of Canada, 8*, 3–66. https://doi.org/10.4319/lo.1950.45.1.0159

Rogers, K. G. (1965). The thermal bar in the Laurentia great lakes. *Proceedings of the 8th conference on Great lakes research* (pp. 358–363). Lakes Div, Univ. Michigan.

Rouse, W. R., Oswald, C. M., Binyamin, J., Blanken, P. D., Schertzger, W. M., & Spence, C. (2003). Interannual and seasonal variability of the surface energy balance and temperature of central Great Slave Lake. *Journal of Hydrometeorology, 4*, 720–730. https://doi.org/10.1175/1525-7541(2003)004<0720:iavote>2.0.co;2
Ryder, R. A. (1982). The morphoedaptic index–Use, abuse and fundamental concepts. *Transactions of the American Fisheries Society, 111*, 154–164. [https://doi.org/10.1577/1548-8659%281982%29111%3C154:tmiaaf%3E2.0.co;2](https://doi.org/10.1577/1548-8659%281982%29111%3C154:tmiaaf%3E2.0.co;2)

Sallée, J.-B., Pellichero, V., Akhoudas, C., Pauthenet, E., Vignes, L., Schmidko, S., et al. (2021). Summertime increases in upper-ocean stratification and mixed-layer depth. *Nature, 591*, 592–598. [https://doi.org/10.1038/s41586-021-03303-x](https://doi.org/10.1038/s41586-021-03303-x)

Schmittner, W. M., Rouse, W. R., Blanken, P. D., & Walker, A. E. (2003). Over-lake meteorology and estimated bulk heat exchange in 1998 and 1999. *Journal of Hydrometeorology, 4*, 649–659. [https://doi.org/10.1175/1525-7541%282003%29004%3C0649:oameeb%3E2.0.co;2](https://doi.org/10.1175/1525-7541%282003%29004%3C0649:oameeb%3E2.0.co;2)

Schmid, M., Buinev, N. M., Granin, N. G., Sturm, M., Schurter, M., & Wiest, A. (2008). Lake Baikal deepwater renewal mystery solved, *Geophysical Research Letters, 35*, L09605. [https://doi.org/10.1029/2008gl033223](https://doi.org/10.1029/2008gl033223)

Shimaraev, M. N., Granin, N. G., & Zhdanov, A. A. (1993). Deep ventilation of Lake Baikal waters due to spring thermal bars. *Limnology & Oceanography, 38*, 1068–1072. [https://doi.org/10.4319/lo.1993.38.5.1068](https://doi.org/10.4319/lo.1993.38.5.1068)

Strøm, K. M. (1945). The temperature of maximum density for fresh waters. *Geofysiske Publikasjoner, 16*, 1–14.

Sverdrup, H. (1953). *On conditions for the Vernal Blooming of Phytoplankton* (Vol. 18, pp. 287–295). *Journal du Conseil/Conseil Permanent International pour l'Exploration de la Mer.* [https://doi.org/10.1093/icesjms/18.3.287](https://doi.org/10.1093/icesjms/18.3.287)

Ulloa, H., Wiest, A., & Bouffard, D. (2018). Mechanical energy budget and mixing efficiency for a radiatively heated ice-covered waterbody. *Journal of Fluid Mechanics, 852*, R1. [https://doi.org/10.1017/jfm.2018.587](https://doi.org/10.1017/jfm.2018.587)

Vehmaa, A., & Salonen, K. (2009). Development of phytoplankton in Lake Pääjärvi (Finland) during under-ice convective mixing period. *Aquatic Ecology, 43*, 693–705. [https://doi.org/10.1007/s10452-009-9273-4](https://doi.org/10.1007/s10452-009-9273-4)

Weiss, R. F., Carmack, E. C., & Koropolov, V. M. (1991). Deep water renewal and biological production in Lake Baikal. *Nature, 349*, 665–669. [https://doi.org/10.1038/349665a0](https://doi.org/10.1038/349665a0)

Williamson, C. E., Dodds, W., Kratz, T. K., & Palmer, M. A. (2008). Lakes and streams as sentinels of environmental change in terrestrial and atmospheric processes. *Frontiers in Ecology and the Environment, 6*, 247–254. [https://doi.org/10.1890/070140](https://doi.org/10.1890/070140)

Williamson, C. E., Saros, J. E., Vincent, W. F., & Smol, J. P. (2009). Lakes and reservoirs as sentinels, integrators, and regulators of climate change. *Limnology & Oceanography, 54*(6, part 2), 2273–2282. [https://doi.org/10.4319/lo.2009.54.6_part_2.2273](https://doi.org/10.4319/lo.2009.54.6_part_2.2273)

Wright, S. (1931). Bottom temperatures in deep lakes. *Science, 74*, 413. [https://doi.org/10.1126/science.74.1921.413](https://doi.org/10.1126/science.74.1921.413)

Wüst, A., Ravens, T. M., Granin, N. G., Kocsis, O., Schurter, M., & Sturm, M. (2005). Cold intrusions in Lake Baikal—Direct observational evidence for deep water renewal. *Limnology & Oceanography, 50*(1), 184–413. [https://doi.org/10.4319/lo.2005.50.1.0184](https://doi.org/10.4319/lo.2005.50.1.0184)

Yang, B., Wells, M. G., Li, J., & Young, J. (2020). Mixing, stratification, and plankton under lake-ice during winter in a large lake: Implications for spring dissolved oxygen levels. *Limnology & Oceanography, 65*, 2713–2729. [https://doi.org/10.1002/lo.11543](https://doi.org/10.1002/lo.11543)

Yang, B., Young, J., Brown, L., & Wells, M. (2017). High frequency observations of temperature and dissolved oxygen reveal under-ice convection in a large lake. *Geophysical Research Letters, 44*, 12218–12226. [https://doi.org/10.1002/2017GL075373](https://doi.org/10.1002/2017GL075373)

Yoshimura, S. (1936a). A contribution to the knowledge of deep water temperatures of Japanese lakes. Part 1. Summer temperature. *Japanese Journal of Astronomy and Geophysics, 13*, 2.

Yoshimura, S. (1936b). Deep water temperatures of lakes of Japan in winter. *Sea Air, 15*, 195–208.