



NEMO-GSL v1.0: An integrated hydrodynamic and ice-process model for a deep subarctic lake (Great Slave Lake, Canada)

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Abstract. NEMO-GSL v1.0 is a three-dimensional hydrodynamic and ice configuration of Great Slave Lake (GSL), the deepest lake in North America and a central component of the Mackenzie River system, built on NEMO v5.0 with the SI³ ice module. The configuration targets conditions absent from ocean and temperate large-lake setups: weak, temperature-controlled stratification, seasonal ice cover exceeding seven months annually, and water levels governed by riverine throughflow. Simulation domain construction required a basin-wide bathymetric product, generated by Universal Kriging of digitized nautical charts, sparse soundings and crowd-sourced depths onto a 1 km grid with 81 vertical levels. Adaptations from NEMO defaults include generic length scale $k-\omega$ closure with the Canuto B stability function for weakly stratified freshwater, adjusted SI³ snow conductivity and ice albedo for freshwater ice under prolonged cover, and a modified river routine in which outflow is computed from mean lake level using monthly stage–discharge relationships, with a hydrostatic correction for the ice and snow load applied in winter. The configuration is forced by CaSR v3.2 atmospheric reanalysis and MESH-derived runoff. Evaluation over 1998–2001 against thermistor profiles at six stations, eddy-covariance evaporation, satellite-derived lake surface water temperature (LSWT) and ice phenology yields water-column RMSE of 2.08 °C, LSWT RMSE of 1.68 °C, mixed-layer depth (MLD) MAE of 6.42 m, seasonal evaporation totals within 0.25% of observations, and ice freeze-up and break-up errors of 1 and 2 days, respectively. Forcing sensitivity experiments reveal wind forcing as the dominant control on simulated mixed layer depth. The bathymetric reconstruction methodology, ice–turbulence parameterization protocol, and river-forcing coupling strategy developed here are independent of GSL-specific inputs and are intended for reuse in other data-scarce, ice-covered lakes. Together, these developments are designed to establish a replicable workflow for subarctic lake modeling, extending NEMO’s application domain beyond the contexts of ocean and temperate large lakes.

1 Introduction

The Arctic is warming at a rate that substantially exceeds the global average, with recent analyses covering the period of 1979 to 2021 indicating nearly four times faster warming (Rantanen et al., 2022; Sweeney et al., 2023). Globally, lake surface temperatures are rising, ice cover duration is declining, mixing regimes are shifting, and stratification seasons are lengthening, with cascading consequences for water quality, ecosystem function, and communities that depend on these freshwater resources (Woolway and Merchant, 2019; Woolway et al., 2020). In the Northern Hemisphere, earlier ice break-up accelerates lake sur-



25 face warming during the transitional periods, amplifying warming beyond that of the open water season alone (Li et al., 2022). These changes are especially consequential for large, deep, high-latitude lakes, which until recently were considered buffered from rapid ecological change by their thermal inertia and extensive seasonal ice cover (Woolway et al., 2022; Sharma et al., 2021). Representing these changes numerically requires configurations and parameterizations suited to the distinct physical regime of large, ice-dominated freshwater systems.

30 Great Slave Lake (GSL; Figure 1), located in the Northwest Territories of Canada, is the second-largest lake entirely within Canadian borders and the deepest lake in North America. Located in the middle of the Mackenzie River Basin, it regulates roughly half of the basin's annual runoff to the Arctic Ocean (Gibson et al., 2006). The lake's vast size, complex morphology and seasonal ice cover make it a sensitive indicator of climatic shifts. The basin of GSL represents a quintessential data-sparse northern environment, where the scarcity of long-term observations and the logistical challenges of data collection necessitate
35 new modeling methodologies. Over the past five years, GSL has experienced unprecedented water level fluctuations, reaching historically high levels in the fall of 2020, which resulted in flooding of low-lying areas of Yellowknife, and record lows in fall 2023, which disrupted navigation and transportation of goods along the Mackenzie River. In the North Slave Region, which includes GSL, lake ice thickness and ice cover duration have significantly declined over recent decades (Attiah et al., 2025). The prolonged open water season affects the overall energy balance of the lake and human activities in the region. Rising
40 air temperatures, reduced ice cover, weakening wind speeds, and altered lake stratification have been linked to shifts in algal communities on GSL over recent decades, marking a transition into a new limnological regime unparalleled in the past two centuries (Rühland et al., 2023). A recent community survey among Indigenous governments and organizations around the lake identified climate change vulnerabilities, long-term water temperature monitoring and fish and wildlife health as the highest research priorities (Government of the Northwest Territories, 2024). Furthermore, GSL is downstream of the Alberta oil sands
45 operations, where potential release of industrial effluents in the Peace-Athabasca River system is threatening the water quality, wildlife, human health, local economies, and Indigenous cultural practices deeply connected to these waters (Schindler, 2001; Schindler and Smol, 2006).

Despite their significance, comprehensive hydrodynamic modeling of GSL and similar large, subarctic lakes has been limited. Prior three-dimensional (3-D) simulation of GSL using the Estuary and Lake Computer Model (ELCOM) was restricted
50 to the West Arm during the ice-free periods (Leon et al., 2007), and one-dimensional thermodynamic ice modeling has been applied only at individual sites (Ménard et al., 2002). A primary barrier has been the absence of reliable bathymetric data spanning the entire lake. Simulating the entire basin of GSL requires a model that resolves three-dimensional circulation over a domain spanning 614 m of relief, couples that circulation to a prognostic ice cover through seven months of the year, and can be extended to biogeochemical and transport applications. These requirements are met by The Nucleus for European
55 Modeling of the Ocean (NEMO), an open-source framework supported by an established user community and ongoing development, which includes coupled modules for ocean thermodynamics, sea ice, and biogeochemistry. Its ocean thermodynamics component solves the primitive equations derived from the Navier–Stokes equations, using the hydrostatic and Boussinesq approximations (Madec and the NEMO System Team, 2024). NEMO has been successfully adapted for the Laurentian Great Lakes within the ECCC coupled prediction system, demonstrating its capabilities in reproducing lake water levels, ice cover,



60 surface temperatures, and thermal structure in large freshwater basins (Dupont et al., 2012), and configured for semi-enclosed
 basins with distinct thermohaline characteristics, such as the Baltic and North Seas (Hordoir et al., 2019). However, applying
 NEMO to deep subarctic lakes presents distinct challenges not adequately addressed by existing ocean or temperate Great Lake
 configurations. Subarctic freshwater systems exhibit: (1) weak density stratification driven primarily by temperature rather than
 salinity, reducing the model's reliance on haline stratification physics, with fetch and topographic constraints further altering
 65 the balances governing momentum transfer and vertical mixing; (2) prolonged seasonal ice cover (7+ months in GSL) creating
 a fundamentally different thermal balance from ice-free or transiently ice-covered systems; and (3) water levels and outflow
 governed by riverine throughflow, with the winter ice and snow load modifying the hydrostatic balance, a coupling absent
 from ocean configurations and only weakly represented in temperate large-lake simulations. These conditions require system-
 atic re-evaluation of NEMO parameterizations developed for ocean and temperate freshwater contexts. To our knowledge, no
 70 multi-year, ice-resolving, basin-wide simulation of a deep subarctic lake has been published, nor has NEMO's transferability
 to such systems been demonstrated.

NEMO-GSL v1.0, a configuration of NEMO v5.0 with the SI³ ice module for GSL, is described and evaluated here. Do-
 main construction required a basin-wide bathymetric product, generated by Universal Kriging interpolation of digitized nautical
 charts, sparse depth observations, and community knowledge, as no such product previously existed for the lake; the interpola-
 75 tion workflow is transferable to other data-scarce northern lakes. Configuration development centered on the three components
 identified above: vertical mixing under weak, temperature-controlled stratification, parameterized using a generic length scale
 $k-\omega$ closure; SI³ ice thermodynamics retuned for prolonged freshwater ice cover; and river forcing, in which lake outflow
 is computed from simulated water level through monthly stage–discharge relationships with a hydrostatic correction for the
 winter ice and snow load. The configuration is forced by CaSR v3.2 atmospheric reanalysis and MESH-derived runoff, and
 80 is evaluated over 1998–2001 against thermistor profiles at six stations, eddy covariance evaporation, and satellite-derived lake
 surface water temperature and ice phenology. Sensitivity to atmospheric forcing is assessed through an inter-comparison of the
 CaSR v3.2 and RDRS v2.1 reanalyses and through one-at-a-time perturbation of wind speed and air temperature, identifying
 wind speed as the dominant control on simulated mixed layer depth and evaporation in this configuration. The configuration,
 its input generation scripts, and its documented parameter set are archived for reuse, and are intended to support a climate
 85 change assessment of GSL and to serve as a starting point for NEMO configurations of other deep, seasonally ice-covered
 lakes. Section 2 describes the study area; Section 3 details the model domain, discretization, parameterization choices, forcing,
 and the evaluation datasets and metrics; Section 4 presents evaluation and sensitivity results; and Section 5 discusses model
 skill, and sources of uncertainty.

2 Study area

90 Great Slave Lake (GSL) has a surface area of 28,568 km² and a maximum depth of 614 m measured in Christie Bay in the
 eastern part of the lake (Rawson, 1950; Herdendorf, 1982). The lake's drainage basin (approximately 949,000 km²) covers
 over half of the Mackenzie River's total drainage area (1,780,000 km²), making it a crucial part of the regional hydrology and



energy cycle (Gibson et al., 2006; Rouse et al., 2008b). Communities in the area rely on the lake and the Mackenzie River for navigation, transportation of goods, fishing, recreational activities, and Indigenous cultural practices. GSL spans two distinct hydro-physiographic regions: Boreal Plains and the Canadian Shield, as shown in Figure 1 (Bring et al., 2016). The western portion, located within the plains, features a broad, relatively shallow basin, whereas the eastern section, influenced by the Shield, is characterized by a large number of islands and deep, elongated bays (Carmack and Vagle, 2021). The freshwater supply to GSL is primarily derived from riverine inflow, with the Slave River, formed by the confluence of the Peace and Athabasca rivers, contributing approximately 74% of the total water input. This river drains a basin whose headwaters are in the Rocky Mountains, where snowmelt plays a critical role in determining inflow volume (Elshamy et al., 2025). Smaller rivers from the Canadian Shield and the Boreal Plains contribute around 21% of the inflow, while direct precipitation onto the lake surface accounts for the remaining 5% (Gibson et al., 2006). The large size of the drainage basin results in significant interannual variation in both the amount and timing of inflows (Woo and Thorne, 2003). Outflows from GSL occur through the Mackenzie River, accounting for 96% of the total lake water budget, with the remaining 4% attributed to evaporation, while groundwater contributions are poorly known. The relative impacts of precipitation and evaporation on the lake's water balance are an order of magnitude smaller than those of riverine inflow, underscoring GSL's status as a high-throughflow system whose water levels are largely governed by variability in upstream inputs (Gibson et al., 2006). The lake exhibits strong year-to-year variability in its water balance despite its long mean residence time of 8 years in the West Arm (Rouse et al., 2008a). Rivers flowing in from the Boreal Plains (namely Slave River, Hay River, Buffalo River and Little Buffalo River) bring a substantial amount of suspended and dissolved organic carbon, supporting the conditions for commercial fisheries. The lake is considered dimictic with turnover in the West Arm observed in spring and fall (Schertzer et al., 2008). However, permanent stratification below the depth of 200 m has been observed in the East Arm, with periodic deep-water ventilation through cabbelling during winter convection (Carmack and Vagle, 2021). Due to the lake's geographic location, full ice cover is observed for about 7 months of the year from December to June, with significant interannual variability (Blanken et al., 2003). The combination of a shallow, turbid western basin and deep, optically clear eastern arms creates a domain with strongly heterogeneous light penetration, stratification, and mixing characteristics, posing particular challenges for uniform parameterization in a hydrodynamic model.

Prior limnological studies of GSL have been primarily observational. Rawson carried out the first detailed limnological studies of the lake mid-century in an effort to inform the establishment of commercial fisheries on the lake. His reviews of physical and limnological features of the lake (Rawson, 1950), assessment of fish productivity (Rawson, 1949), fish and plankton species (Rawson, 1951, 1956) have formed the bulk of knowledge of GSL limnology before the turn of the century (Schindler, 2001). Subsequently, several studies into the climatology and thermal regimes of GSL were carried out as part of Mackenzie Global Energy and Water Cycle Experiment (GEWEX) Study (MAGS). Blanken et al. (2008) analyzed direct evaporation measurements taken over the lake over a six year period (1997-2003), finding that about half of the total evaporative water loss took place over 20% of the observation period, mainly driven by periodic incursions of dry cool air over the lake surface, towards the end of the open water season. Wind speed and vapor pressure gradient between the lake surface and the atmosphere served as primary controls in these events. Schertzer et al. (2008) used meteorological and thermal profile

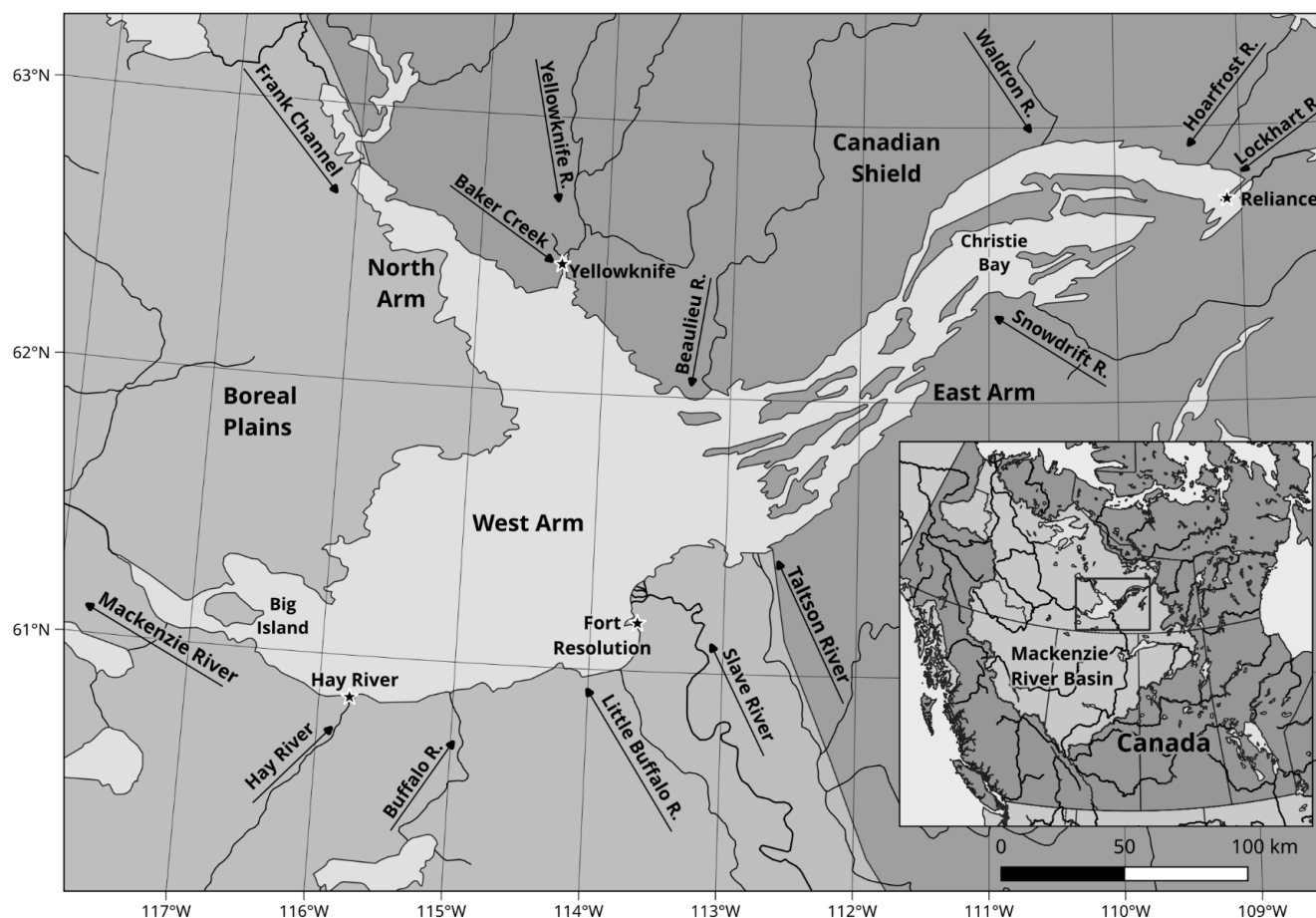


Figure 1. A map of Great Slave Lake showing major tributaries, Mackenzie River, and hydro-physiographic regions.

observations along a transect of the West Arm between 1997 and 2003 to establish the baseline thermal characteristics of GSL. Significant variability between 1998 and 2002 was observed in maximum measured water surface temperatures (21.2 °C and 14 °C, respectively) and duration of open water season (213 days and 174 days, respectively), highlighting the lake's sensitivity to interannual climate variability. Variability in duration of ice cover was found to directly impact the amount of heat absorbed from solar radiation, with early ice break-up typically followed by late ice freeze-up, which in turn prolongs the period with high evaporative fluxes in early winter (Rouse et al., 2003).

Prior efforts at numerical modeling of GSL hydrodynamics have been limited to the West Arm due to a lack of reliable bathymetry for the entire lake. A notable effort using the Estuary and Lake Computer Model (ELCOM) was made by Leon et al. (2007). The study focused on validating the model for integration of a lake component with the Canadian Regional Climate Model (CRCM). The model domain had a horizontal resolution of 2 km with 30 vertical levels and contained the West Arm of GSL, with North Arm and East Arm excluded and treated as closed boundaries. River inflows or outflows were



not considered. Model results were compared to field observations collected during the summer of 2003. Simulated surface temperatures were overpredicted by 0.7 °C in the shallow parts of the lake and underpredicted by 1.1 °C in the deep parts of the lake when forced with observed meteorology. Using the CRCM forcing resulted in surface temperature errors in the region of 15%. Simulated lake circulation showed a counterclockwise gyre in the West Arm, which has been supported by sediment plume observations made by Rouse et al. (2008b). Ménard et al. (2002) simulated the ice phenology of GSL in 3 bays near Yellowknife, Hay River and Reliance using the Canadian Lake Ice Model (CLIMo), a 1-D thermodynamic lake ice model. The model was able to accurately represent ice thickness with a mean error of 7 cm and annual freeze-up/break-up dates with a mean error of 4 days. The adaptation of NEMO for the Laurentian Great Lakes (Dupont et al., 2012) demonstrated the feasibility of applying this framework to large freshwater systems, but no comparable effort has been made for subarctic lakes, where prolonged ice cover and limited observational data impose additional constraints. No prior studies have focused on multi-year simulation of coupled lake ice and hydrodynamics.

3 Configuration overview

The adaptations outlined in Sections 3.2–3.3 represent deliberate configuration choices for subarctic application, addressing three challenges absent from ocean and temperate Great Lake configurations: vertical mixing under weak stratification, ice–water coupling under prolonged ice cover, and freshwater forcing representation. Atmospheric forcing sensitivity is evaluated in Section 4.7

3.1 Lake bathymetry

Digital lake bathymetry is required for a hydrodynamic lake model simulation domain setup. For this study, a digital bathymetry of GSL has been created from nautical charts and additional measurements taken over the central part of the West Arm (CHS, 2004, 1990; Schertzer, 2000). Nautical charts have been digitized from high resolution scans and vectorized to retrieve the lake shoreline from each source. Depth measurements from each chart were manually vectorized and georeferenced using charted coordinates and further adjusted to observed lake shoreline features from satellite imagery. Due to sparse measurements in multiple areas of the lake, assumptions have been made to populate sufficient points for interpolation. These assumptions are informed by approximate depths derived from publicly accessible navigational charts and crowd-sourced bathymetric data available through recreational marine navigation platforms, hereafter referred to collectively as community knowledge. These supplementary data are limited to the North Arm, and the southern and far eastern parts of the East arm. The combined dataset containing georeferenced lake shoreline, 2,646 points of depth observations and 113 points derived from community knowledge is shown in Figure 2. Universal Kriging interpolation with a spherical variogram, selected through testing over a sample of 1,000 points, was used to interpolate the observations to a 1 km horizontal grid. An 80-20 split sampling strategy was used during interpolation to generate and validate the variogram, while the final bathymetry product was prepared using the entire dataset. The quality of interpolation has been assessed using root mean square error (RMSE) for overall accuracy

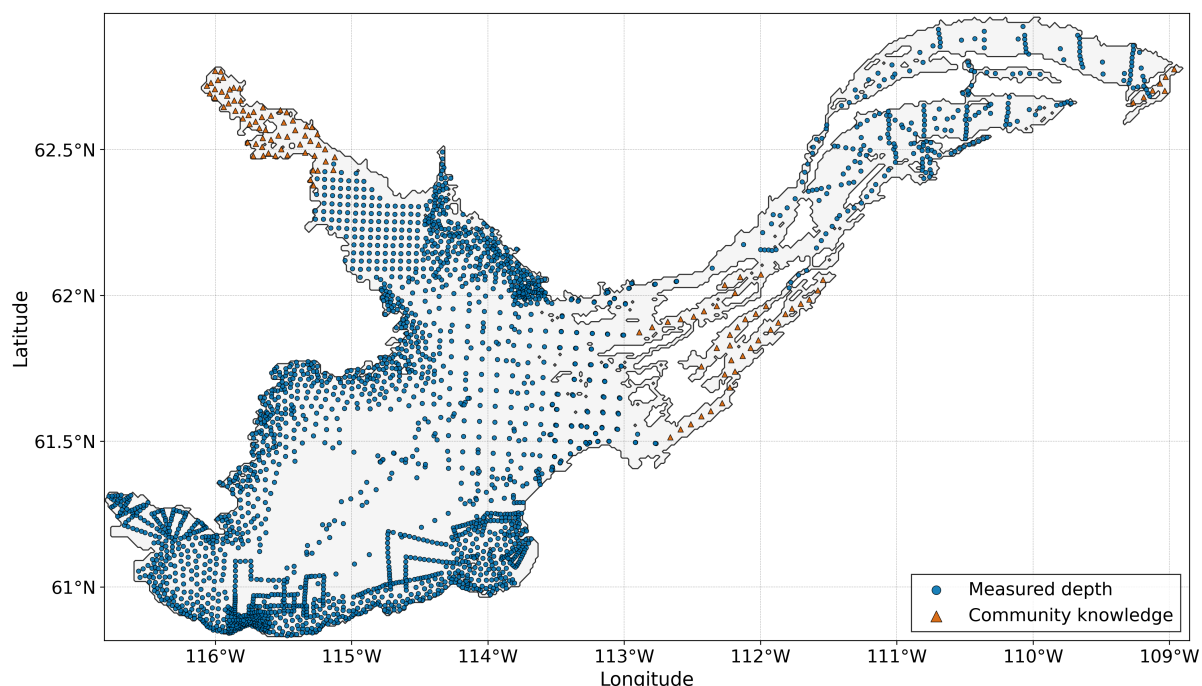


Figure 2. Georeferenced observations of lake depth derived from nautical charts (2,646 points in blue) and assumed depths from community knowledge (113 points in orange) used for GSL bathymetry interpolation.

170 and mean error (ME) to assess interpolation bias. Metrics were calculated using a nearest neighbor approach to identify the nearest interpolated grid point for each observation, and compare the interpolated values to observations used in interpolation.

3.2 Model specification

The GSL hydrodynamics model is set up using NEMO, v5.0. This version includes significant performance optimizations through reduced memory footprint, minimized message passing interface (MPI) communications and most significantly, from the Runge-Kutta third-order (RK3) temporal scheme, improving the time stepping efficiency. The model is compiled on Digital Research Alliance of Canada high-performance computing (HPC) cluster Fir, using available libraries.

180 The following parameterization choices reflect deliberate adaptation of NEMO for subarctic freshwater application and depart substantially from both ocean and existing Great Lakes defaults. The simulation domain uses the newly created lake bathymetry at 1 km horizontal resolution, resulting in a 420×240 rectilinear grid. NEMO initializes each water column from the local bed elevation to the reference surface datum (i.e., initial water surface height $\eta = 0$ m), so the lake is filled to the full depth specified by the bathymetry field. To account for water level changes, the bathymetry is adjusted to the water level recorded by the hydrometric station at Yellowknife (07SB001), at the start of model spin-up period. The bathymetry is made either deeper or shallower to account for deviations from the vertical datum, which is set at 156.27 m above sea level. The



limitation of this approach is that changing water levels do not produce a change in lake surface area. This is a reasonable
 185 assumption for GSL, where inter-annual water level variability of $\sim 1\text{--}2$ m is small relative to the mean depth of ~ 70 m and
 the lake has steep shorelines over much of its perimeter. The simulation domain is produced using NEMO built-in tools with
 a partial step z-coordinate in the vertical direction, and the minimum domain depth set at 5 levels, resulting in an approximate
 minimum depth of 5 m. A total of 81 vertical levels are used with a higher concentration at the surface to accurately resolve
 the mixed layer and thermocline dynamics. The level thickness is 1 m at the surface, gradually stretched to 2.5 m at a depth of
 190 40 m, with the bottom layer in the deepest part of the lake at a thickness of approximately 15.7 m. The partial step provides
 more accurate representation of bathymetry at model boundaries, allowing the bottom cell thickness to vary from the level
 thickness. A minimum partial step thickness equal to the lesser of 4 m and 50 % of the local layer thickness was enforced to
 avoid hydrostatic pressure errors on steep slopes in the deep basin.

The initial conditions for the hydrodynamic model are a uniform water temperature of 4°C throughout the water column
 195 and a uniform absolute salinity of 0.2 g kg^{-1} representative of freshwater. These initial conditions are used at the start of the
 spin-up period on October 1, 1992, and a 5-water-year spin-up period (October 1992 – September 1997) is used to achieve
 quasi-equilibrium conditions, which are then used to initialize the evaluation simulation, starting on October 1, 1997. Closed
 boundaries are defined around the lake basin, with a no-slip lateral boundary condition and no exchanges of mass or energy.
 Bottom friction is set to 0.0025 with non-linear scaling based on flow velocity. The COARE 3.6 bulk forcing algorithm (Edson
 200 et al., 2013), with cool-skin and warm-layer parameterizations, translates atmospheric variables into surface fluxes. Light
 penetration of the water column uses a two band scheme with 78 % of the energy absorbed within a shallow e-folding depth
 of 1.4 m, while the remaining fraction penetrates deeper with an e-folding depth of 7.9 m, defaulting to a Jerlov Type III for
 coastal waters. While this does not represent the spatial differences in turbidity between strongly oligotrophic waters in the East
 Arm and the West Arm, which is affected by sediment brought in by Slave and Hay rivers, it is a compromise made due to lack
 205 of spatially distributed water clarity data. A fourth-order flux-corrected transport (FCT) horizontal tracer advection scheme
 is used (Zalesak, 1979), along with iso-neutral Laplacian diffusion operators for momentum and tracers. The horizontal eddy
 coefficients are adjusted automatically based on local velocity, to avoid excess diffusion. Vertical mixing parameterization is
 critical for representing thermocline sharpness in stratified lakes, and is parameterized using the generic length scale (GLS)
 turbulent closure parameterization (Umlauf and Burchard, 2003, 2005) with a $k\text{--}\omega$ turbulence scheme, the Canuto B stability
 210 function, and background surface roughness and wave action parameterizations adjusted to represent lake surface phenomena.
 Background vertical eddy viscosity and diffusivity values are set to $2 \times 10^{-6}\text{ m}^2\text{ s}^{-1}$ and $2 \times 10^{-7}\text{ m}^2\text{ s}^{-1}$, respectively.
 These serve as baseline values in areas where the eddy viscosity and eddy diffusivity computed through GLS are lower than
 the baseline values. Convection is primarily resolved through GLS parameterization; however, enhanced vertical diffusion
 with a mixing coefficient of $1\text{ m}^2/\text{s}$ is enabled to resolve unstable stratification at the surface, where the mixing length is
 215 limited. The TEOS-10 equation of state is used, which is capable of accurately representing water density at low salinity values
 (Pawlowicz and Feistel, 2012). Astronomic tidal forcing has been excluded from the simulation, as gravitational tides in large
 lakes are negligible. For example, in the Laurentian Great Lakes, spring tidal amplitudes are less than 5 cm and are masked by
 meteorologically driven water level fluctuations (Trebitz, 2006).



The sea ice module (SI³) is enabled to simulate ice thermodynamics with five ice categories, ten ice layers and five snow layers. Ice and snow parameters have been adjusted to achieve accurate representation of lake freeze-up and ice break-up timing. The thermal conductivity of snow was set at $0.25 \text{ W m}^{-1} \text{ K}^{-1}$ and albedo values for melting snow and bare puddled ice have been reduced from default ocean values to 0.72 and 0.49, respectively. Bulk coefficients over ice are calculated using the parameterization by Lüpkes et al. (2012). A 90% salt rejection for new ice formation has been used, resulting in ice salinity of 0.02 g kg^{-1} .

Model parameters were calibrated manually through iterative adjustment of vertical mixing coefficients, snow thermal conductivity, ice albedo, and related parameters, evaluating performance against observations of ice phenology, water temperature profiles, mixed layer depth, and evaporation. The model was run at a 4-minute time step, using 180 cores on a Digital Research Alliance of Canada Fir cluster, requiring approximately 3 hours of wall-clock time per simulated year.

3.3 Forcing

Atmospheric forcing was provided by the ECCO reanalysis product CaSR v3.2 at a 10 km horizontal resolution on an hourly time step. Relevant variables include total precipitation, air temperature, specific humidity, direction-corrected wind forcing vectors, surface pressure, and incoming shortwave and longwave radiation. CaSR v3.2 provides both raw reanalysis precipitation fields and an observation-corrected total precipitation product, but no corresponding observation-corrected snowfall field. To partition the corrected total precipitation into rain and snow, the snow fraction is calculated from the raw reanalysis fields (modeled snowfall divided by modeled total precipitation) and then applied to the observation-corrected total precipitation. This approach preserves observation-constrained total precipitation but inherits the phase partitioning of the reanalysis. Unit conversions are applied to the forcing dataset, and data are regridded to the model grid at 1 km resolution using bilinear interpolation via the climate data operators (CDO) software.

River runoff forcing from the Mackenzie River Basin MESH (Modélisation Environnementale Surface et Hydrologie) simulation was used to determine the amount of inflow through each of the 13 largest tributaries of GSL shown in Figure 1 (Elshamy et al., 2025). The cryospheric-hydrological model was driven by atmospheric forcing from the Regional Deterministic Reforecast System Version 2.1 (RDRS v2.1), an atmospheric reanalysis dataset developed for North America, and the predecessor to CaSR v3.2 (Gasset et al., 2021). As both products share the same underlying reanalysis framework and differ primarily in minor methodological refinements, the use of RDRS v2.1 for hydrological forcing and CaSR v3.2 for lake surface forcing is expected to introduce negligible inconsistency. Most of the inflow values used are derived at gauge locations upstream of the lake inflow locations, except for Slave River and Hay River, where river mouth locations at the lake were used.

Slave River inflows constitute 74% of river contributions to GSL, so accurate discharge values are of high importance. The discharge of the Slave River simulated by the MESH model was compared to discharge observations at hydrometric station 07NB001 (Slave River at Fitzgerald), to evaluate the accuracy of simulated river discharge magnitudes. Simulated annual mean discharge values were 2.7% higher than observed over the four years of evaluation (1998–2001). The simulated discharge was 12.5% lower than observations in 1998, and 16% higher than observations in 1999, with 2000 and 2001 discharge within 4% of observations.



Due to a lack of reliable river temperature observations, inflow temperatures have not been explicitly prescribed in the model. By default in NEMO, the temperature of the receiving grid cell is assigned to river inflows when no discharge temperature is specified. This introduces uncertainty, as river inflow grid cells are located in shallow nearshore regions that experience elevated summer temperatures, effectively warming inflow water rather than representing cooler river inputs. An alternative approach using river temperature climatology derived from hydrometric gauge thermistor records was tested, but introduced the opposite problem: an unrealistic plume of cold water during summer, likely because the gauge thermistors recorded temperatures at depth rather than at the surface. The default approach was retained, as it produced better agreement with thermistor chain and satellite surface temperature observations. For salinity, a uniform value of 0.2 g kg^{-1} has been assigned to all riverine inflows, consistent with the assumed initial lake salinity.

Lake outflow into the Mackenzie River is controlled using a customized river runoff module, representing a methodological innovation necessary for large freshwater systems where continuous observation records are limited, and water-level changes directly affect outflow dynamics. Key modifications include: 1) Salinity treatment at the lake outflow, where ambient salt concentration is removed at outlet cells rather than prescribing a fixed value or using the default 0 g kg^{-1} ; 2) outflow discharge computation is based on the average lake water level and monthly stage-discharge relationships embedded in the modified river routine (Stankevicius et al., 2025). The outflow discharge calculations are conducted on a daily time step, consistent with the temporal resolution of riverine inflow data. In winter, ice formation complicates this process, as the water surface height output by NEMO reflects the water level beneath the ice cover, which does not match the observed surface water levels used for stage-discharge relationships. An additional correction is applied to account for the mass of ice and snow, converting them to a water-level equivalent distributed evenly over the lake surface, thereby approximating the hydrostatic response of the water column to the ice-and-snow load. This correction ensures that winter outflow calculations remain consistent with observed water-level data. Outflow from GSL is distributed across 12 grid cells – 8 to the south of Big Island and 4 to the north – with the confluence downstream of the Big Island forming the Mackenzie River. Higher weights are assigned to the central cells in each channel to emulate a parabolic flow profile. A similar approach is taken at the Slave River delta, where the inflows are distributed across 20 grid cells, with higher weight assigned to the main river channel. The distribution of river inflows and outflows across multiple grid cells in wide rivers and deltas is shown in Figure 3.

3.4 Model evaluation

Model performance was evaluated through a comparison of simulated outputs against four independent observational datasets: vertical temperature profiles from moored thermistor chains, surface evaporative fluxes from eddy covariance measurements over the central part of the lake, and lake surface water temperature (LSWT) and ice cover phenology derived from satellite passive microwave imagery.

Vertical temperature profile observations were collected during the Mackenzie River basin global water and energy experiment (MAGS GEWEX). The data were provided by ECCO and include temperature data for the open-water seasons (June 25th - September 9th) of 1998–2001 at six locations on the West Arm of GSL (Schertzer et al., 2008). The coordinates of each location and respective maximum observation depths are outlined in Table 1 and shown in Figure 3. Observations varied in depth

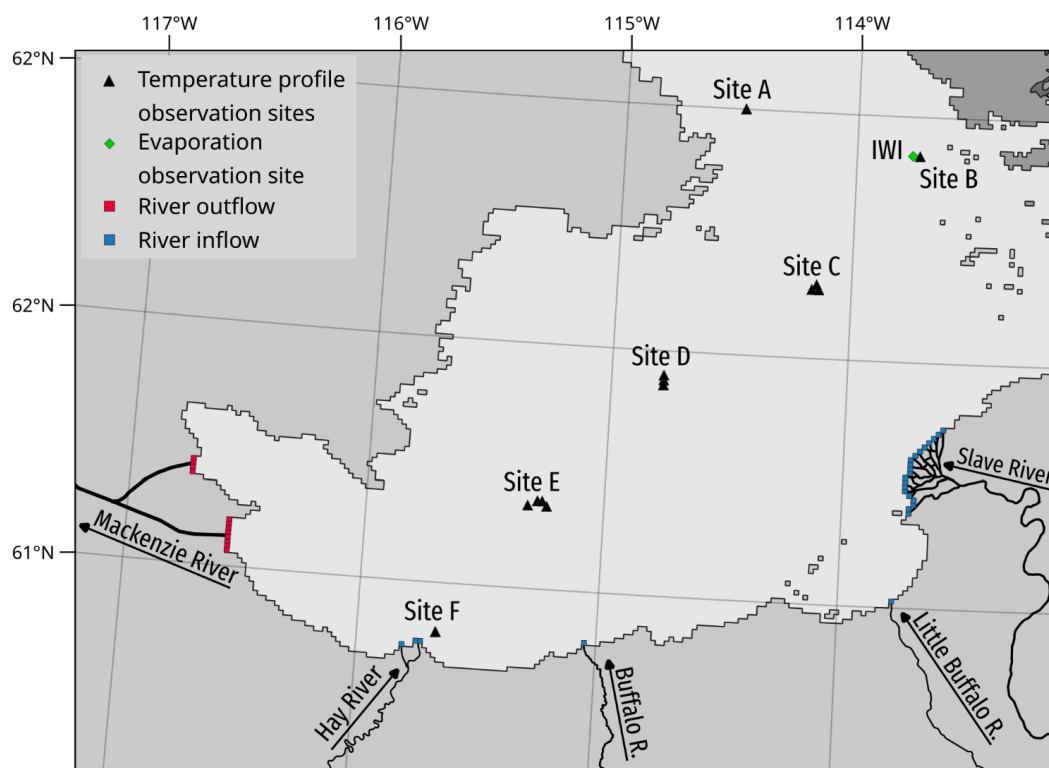


Figure 3. Temperature profile (sites A through F), and evaporation (Inner Whaleback Island) observation locations, and river inflow and outflow configuration.

depending on location, and the spacing of the probes was generally between 2–2.5 m up to a depth of 15 m, 5 m between the depth of 15 m and 30 m, and 10 m at depths greater than 30 m. While observations can be grouped into six nominal locations, the exact deployment coordinates of each buoy varied from year to year within a given site. As a result, both the observed water column depth and the model bathymetry at the corresponding extraction point differed between years for the same nominal site (see Figure 6, where the model-bathymetry dashed line differs slightly between panels). Observed daily mean temperature values at each observation depth throughout the evaluation period were compared to daily mean model outputs at each depth. Model performance was assessed using root mean square error (RMSE) to quantify the quality of overall model performance, giving higher weighting to large outliers and mean error (ME) to assess model bias. Model evaluation metrics were calculated for each of the observation locations over the entire water column by interpolating simulated values to observation depths.

Lake surface water temperature (LSWT) observations for GSL have been sourced from the European Space Agency Lakes Climate Change Initiative v2.1 dataset derived from satellite observations (Carrea et al., 2024). Daily merged products for the open water season (July–September) of each evaluation year (1998–2001) were filtered to include only retrievals with a quality level of 4 (acceptable) or higher, and days with at least 20% spatial coverage of the lake surface. The filtered fields



Table 1. Temperature sampling location coordinates across study years.

Site	Maximum observation depth (m)	1998		1999		2000		2001	
		Longitude	Latitude	Longitude	Latitude	Longitude	Latitude	Longitude	Latitude
A	56 - 60	–	–	–	–	–114.49	62.00	–114.49	62.00
B	50	–	–	–113.76	61.92	–113.73	61.92	–113.76	61.92
C	75 - 100	–114.17	61.65	–114.15	61.65	–114.14	61.64	–114.14	61.64
D	50 - 56	–114.78	61.44	–114.78	61.43	–114.81	61.43	–114.78	61.45
E	50 - 56	–115.32	61.17	–115.28	61.18	–115.25	61.18	–115.24	61.17
F	10 - 12	–	–	–115.67	60.90	–115.67	60.90	–115.69	60.90

were regridded from the native resolution of 1/120 degrees to the NEMO model grid using nearest-neighbor interpolation, and observed LSWT at each grid cell was compared to the corresponding daily mean model skin temperature. The COARE 3.6 bulk forcing algorithm includes a cool-skin parameterization that accounts for the thin thermal boundary layer at the air–water interface, where net longwave and latent heat loss produce a skin temperature slightly cooler than the underlying bulk temperature. Because NEMO applies this correction, its surface temperature output represents a skin-layer estimate and is therefore directly comparable to the satellite-derived skin temperature from the ESA CCI Lakes dataset. Model performance was assessed using RMSE and ME metrics for each image, including only the coincident observation and model output grid cells.

Mixed layer depth (MLD) representation and thermocline sharpness in stratified lakes present significant challenges when applying ocean models to lacustrine systems. Due to the steep vertical temperature gradients characteristic of lake thermoclines, small errors in thermocline depth translate to substantial temperature biases at depth. To evaluate the model’s skill in capturing these critical thermal structure features, MLD is calculated from both observed and simulated temperature profiles and compared using root mean square error (RMSE) and mean error (ME) metrics. Following established practice for MLD determination in stratified water bodies, the mixed layer depth is defined as the depth at which temperature differs by more than 0.5 °C from a near-surface reference temperature. The reference temperature is taken at 2.5 m depth to avoid diurnal surface heating effects while remaining within the actively mixed surface layer. This threshold-based approach provides a robust method for identifying the base of the mixed layer across varying stratification conditions.

Accurate representation of ice phenology is critical to representing the thermal balance of a lake, as increased duration of open water season allows for a longer heating period, and increased thermal capacity within lake waters. The timing of simulated ice cover formation and break-up was validated against satellite-derived observed ice phenology spanning the winter seasons of 1998 to 2001. The observational dataset integrates SMMR, SSM/I, and SSMIS passive microwave imagery to identify dates of complete ice break-up, complete freeze-up, and open water season duration (Cai et al., 2021). In the simulation analysis, complete ice break-up is defined as occurring when the extent of ice cover falls below 5%, while complete freeze-up is



determined when ice cover reaches 95%, to account for residual near-shore ice and model grid cells with partial coverage. This aligns with the approach taken in producing the observational dataset. Model performance is quantified using absolute error, calculated as the difference in day-of-year between observed and simulated freeze-up and break-up dates for each ice season. Total duration of the open-water season is calculated to integrate the ice-on and ice-off dates into a single metric indicative of the total duration of direct interaction between the atmosphere and the water column.

Simulated evaporative fluxes are evaluated against eddy covariance measurements collected during the Mackenzie GEWEX Study (MAGS) field campaign on Inner Whaleback Island (IWI), located near the deepest section of Great Slave Lake's West Arm, and close to water column temperature profile observation Site B, as shown in Figure 3. The observational dataset comprises daily evaporation totals (mm) derived from eddy covariance towers during the open water seasons (June 25–September 10) of 1998 and 1999 (Blanken et al., 2008). Model outputs at the IWI coordinates are compared directly to observations, with RMSE and ME reported for daily evaporation values. Absolute differences summed over each open water season are used to assess the model's skill in simulating latent heat exchange at the lake surface throughout the summer season.

Direct observational data for lake-wide water circulation patterns are not available for GSL. However, the model's demonstrated skill in reproducing thermal structure, ice dynamics, and surface energy fluxes provides confidence that the underlying physics governing momentum transfer and buoyancy-driven flows are represented with good accuracy. The simulated circulation fields therefore offer valuable insight into the relative roles of wind forcing, thermal stratification, and riverine inflows in driving water current patterns across the lake.

To assess the model sensitivity to atmospheric forcing, the model was run with an alternative reanalysis product, the Regional Deterministic Reforecast System v2.1 (RDRS v2.1) (Gasset et al., 2021). The simulation used the same parameters applied to CaSR v3.2 forcing, so differences in performance stem from the differences in atmospheric forcing variable values. The simulation was run for model evaluation period (1998–2001) and RMSE and ME metrics were compared to the baseline results from CaSR v3.2 simulation. Further assessment of model sensitivity to atmospheric forcing was carried out through a one-at-a-time sensitivity check. Scenarios with air temperature and wind speed at baseline values (CaSR v3.2 forcing), and varied by -10% and +10%, were run concurrently for one year from October 1, 2000 to September 30, 2001. For comparability with the wind-speed sensitivity, air temperature was increased by a uniform offset corresponding to 10% of the time-mean magnitude of the dataset air temperature (+1.27 °C) and decreased by the same magnitude (−1.27 °C). The impacts of changing each variable on simulated evaporative fluxes, surface water temperature, and MLD were assessed. Evaporative fluxes were taken as evaporation from the entire lake surface in mm year^{−1} over the simulation period. Lake surface temperature is defined as the mean skin temperature (°C) over the entire lake surface over the simulation period. MLD was determined using the temperature threshold approach detailed earlier in this section.

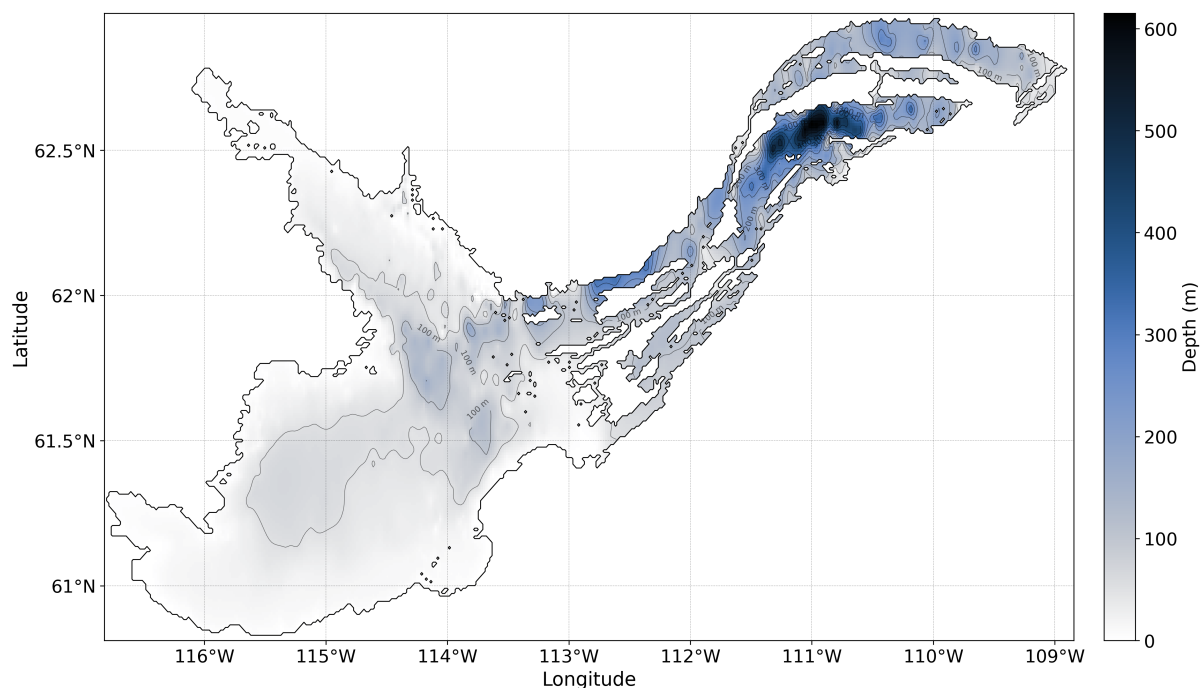


Figure 4. GSL bathymetry generated through Universal Kriging interpolation. Depth contours are shown every 50 m, depth labels every 100 m.

4 Results

4.1 Lake bathymetry

355 The first comprehensive digital bathymetry of GSL represents a novel contribution enabling future model development in this data-scarce region. The simulation domain of the entire GSL was created from a dataset of depth observations and community knowledge and is, to our knowledge, the first basin-wide bathymetric product for GSL. Interpolation onto a 1 km horizontal grid was performed using Universal Kriging with a spherical variogram. The resulting bathymetry is shown in Figure 4. The interpolation resulted in overall RMSE and ME values of 6.19 m and -0.18 m, respectively. The correlation between observed and interpolated depths is shown in Figure 5. The uncertainty in the West Arm arises from dense clusters of observations along the shoreline. In cases where multiple observations fall within a single grid cell, a compromise is made to fit a single best value based on multiple observations of varying magnitudes. Large outliers can be identified as interpolated values of 0, resulting from the grid resolution in the steep gradient regions in close proximity to smaller islands of the lake. Observations in the deep parts of the lake are generally scarce, with a dense cluster in the region of Christie Bay. Steep depth gradients and low density of observations result in large errors in the region. Nonetheless, the resulting bathymetry is deemed acceptable as a simulation domain in light of data limitations.

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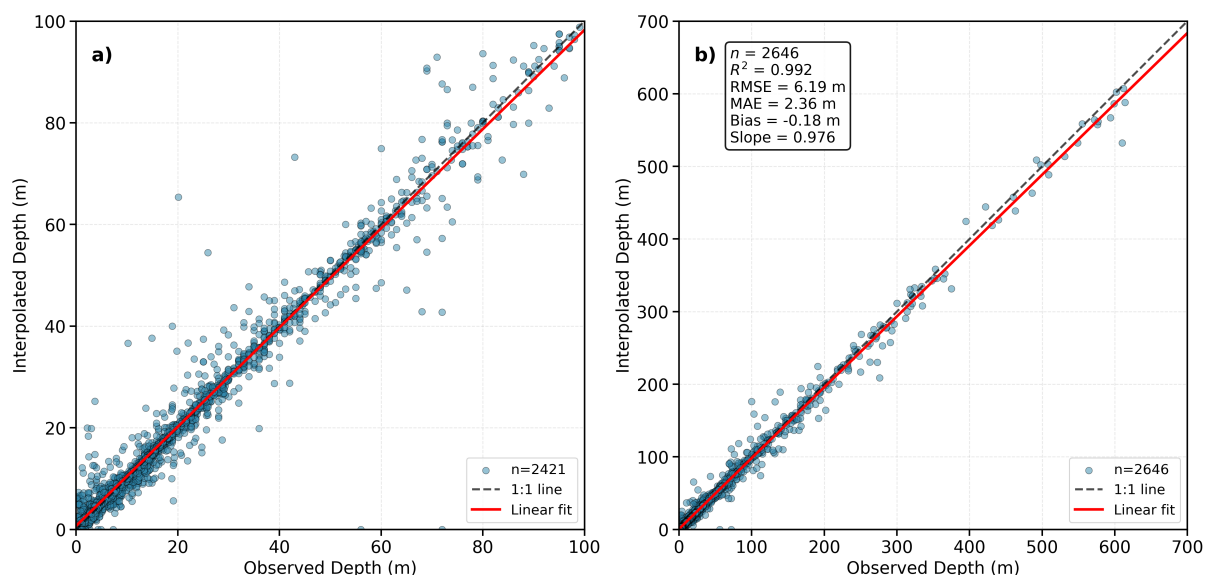


Figure 5. Correlation of observations and interpolated bathymetry values. a) shallow regions (0 m–100 m), b) full depth range.

4.2 Temperature profiles

Following the 5-year spin-up, simulation outputs from the summer months (June 25th - September 10) of the 1998–2001 evaluation period are compared to temperature profile observations. RMSE and ME values are derived for the entire water column at each location and each year of evaluation and are presented in Table 2. Overall RMSE value of 2.08 °C and overall ME value of 0.23 °C are achieved, indicating good skill at reproducing the thermal structure at the observation locations with a slight overprediction bias. A notable outlier is Site C in 1999, with the highest overall RMSE of 2.89 °C and the highest overall ME of 1.99 °C. Furthermore, the shallowest observation location Site F shows consistently high RMSE, and a consistent overprediction bias throughout all years of observation, suggesting difficulty in reproducing temperature profiles in shallow areas of the lake.

A comparison of observed and simulated temperature profile isotherms at the deepest observation location – Site C for all evaluation years is presented in Figure 6. Simulation results show good skill at representing the formation of thermal stratification and mixed layer depth. While the mixed layer depth is overpredicted in 1999 and underpredicted in the year 2000, the simulated thermocline is sharp, though more diffuse than in observations, as indicated by the density of isotherms in the metalimnion. The diffuse thermocline can be attributed to numerical diffusion and model parameterization (Huang et al., 2010). The hypolimnion temperatures, specifically the 4 °C isotherms present at the end of spring turnover are not represented well. This indicates excessive vertical heat transport below the metalimnion, resulting in elevated deep water temperatures. Periodic storms during the open-water season produce strong winds that can disrupt thermal stratification and drive mixing events that penetrate the metalimnion (Schertzer et al., 2003). These events, indicated by the sudden deepening of isotherms are visible



Table 2. Model performance metrics (RMSE and ME) for each observation location across study years.

Site	Observation Depth (m)	1998		1999		2000		2001	
		RMSE	ME	RMSE	ME	RMSE	ME	RMSE	ME
A	56	–	–	–	–	1.63	–0.21	1.20	0.77
B	50	–	–	1.57	0.93	2.50	0.77	2.53	1.28
C	100	1.95	–0.68	2.89	1.99	1.40	–0.53	1.59	–0.20
D	56	2.34	–0.59	2.03	0.94	1.21	0.01	1.53	0.85
E	56	2.07	–0.22	2.54	–0.38	2.51	–1.27	2.70	–1.06
F	12	–	–	2.54	1.67	2.08	0.49	2.56	1.04

in the observed isotherms and are also reproduced by the simulation, although they are less pronounced in some cases. This suggests that further improvements to the wind forcing representation and the vertical mixing parameterization are needed.

Winter water temperature profiles have been collected at multiple locations within the model evaluation period, however these observations started at the depth of 12 m, preventing observation of the full inverse stratification. Site C alone provides continuous water-column temperature observations from June 16th 2000 to January 16th 2001. An isotherm plot for the period shown in Figure 7 reveals that the timing of summer stratification break-up is represented well, starting in early October, with inverse stratification forming in early December. Simulated turnover shows that weak stratification, indicated by the 6 °C isotherm, remains for about 2 weeks, before the full water column is mixed to a uniform temperature. Observed vertical mixing prior to the establishment of winter is characterized by periodic occurrences of instabilities and convective mixing, and the simulation is unable to capture these features, producing consistent deep mixing. The depth of winter inverse stratification is not accurately reproduced, with the model overpredicting deep water temperature, and displaying smooth isotherms, closely packed within the top 60 m in early winter, showing deep-water temperatures warmer than observation.

4.3 Lake surface water temperature

LSWT observations from the ESA Lakes CCI v2.1 satellite-derived dataset were compared to simulated water skin temperature. The coverage of satellite observations varies throughout the evaluation period, all high quality images from July through October with at least 20% coverage of the lake area have been used in a comparison to simulation results. Figure 8 shows the distribution of valid data points with number of observations ranging from 21 to 46 for any given year of evaluation. Highest errors are observed in July of each year, with correlation improving as the season progresses. Across all years, an RMSE of 1.68 °C and an ME of +0.39 °C are achieved, indicating good skill at simulating LSWT. A visual comparison of simulation results and observations reveals sharper temperature gradients in the observation dataset, whereas the simulated temperature distribution is more diffuse. Figure 9 compares simulated and observed LSWT on July 9, 2001, a date with near-complete lake coverage and one of the largest bias magnitudes in the evaluation record. In this case, the model underpredicts LSWT in the southeastern part of the West Arm and in the deepest part of the lake near Christie Bay. Temperatures in the large area near

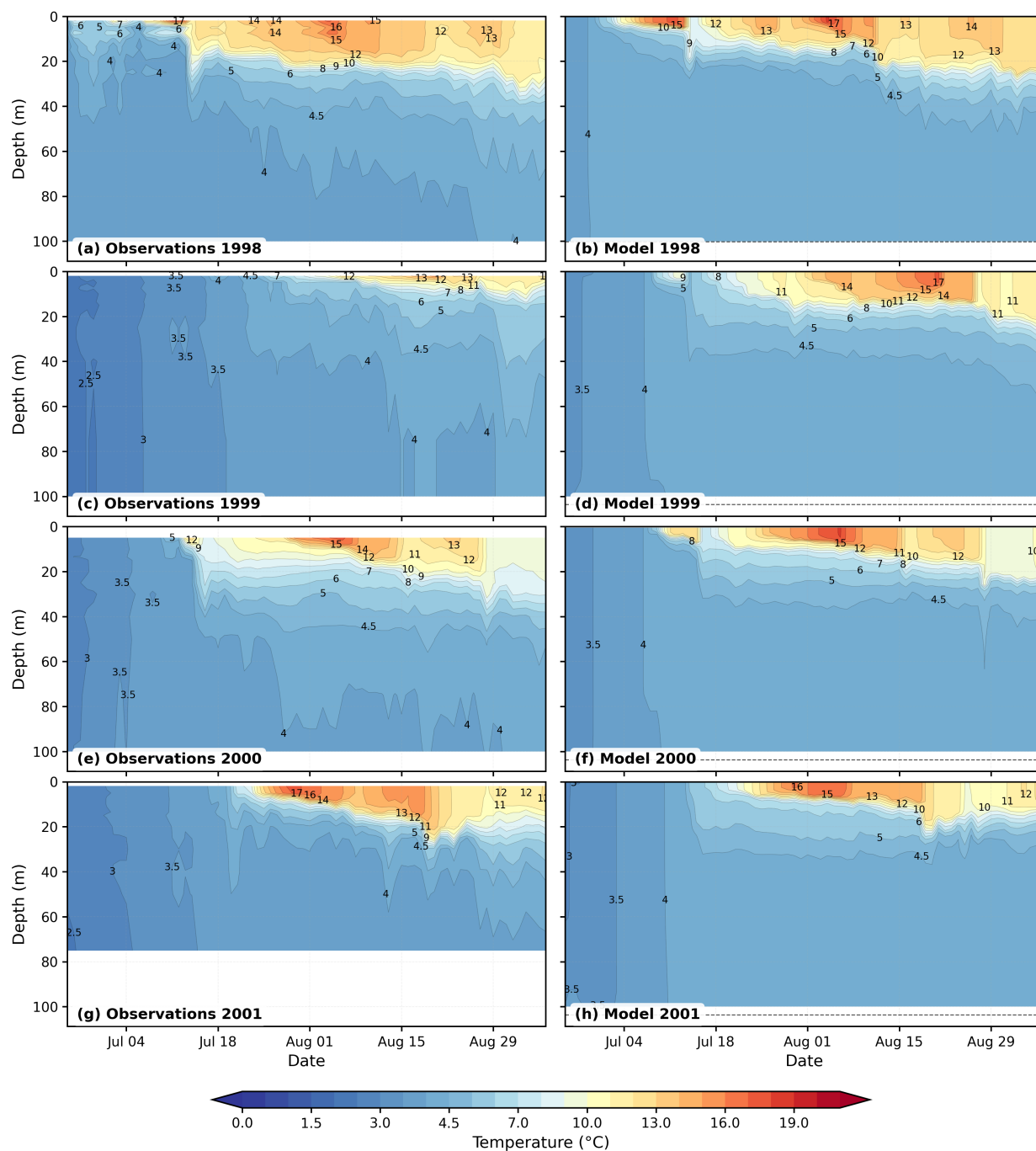


Figure 6. Observed and simulated isotherms during the summer period at Site C. Dashed line on model plots indicates the depth of model bathymetry.

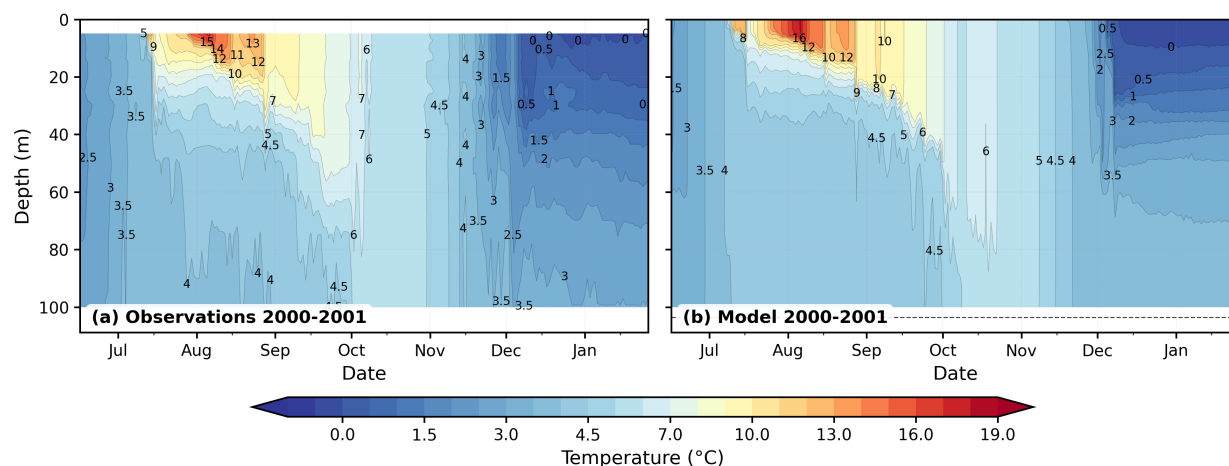


Figure 7. Observed and simulated isotherms during the summer and early winter period at Site C. Dashed line on model plots indicates the depth of model bathymetry.

Slave River delta may be underpredicted due to the approach taken for river temperature (see methodology section). Areas of overprediction are smaller, and generally confined to parts of the West Arm and the northern part of the East Arm. An RMSE of 2.57 °C is achieved with an underprediction bias of 1.31 °C.

4.4 Mixed layer depth

Mixed layer depth (MLD) was calculated from both temperature observations and model outputs on a daily basis using the temperature threshold approach at each monitoring location across the evaluation period. Overall, the model predicts MLD with a mean absolute error (MAE) of 6.42 m and a bias of +0.90 m, though skill varies by location and year. The MLD errors stem primarily from the inaccurate representation of stratification onset timing. Figure 10 compares daily values and 5-day running means at Site C, the deepest monitoring location. In all years, MLD reaches the full observation site depth in late June to early July, marking the spring turnover period. The model captures the onset of thermal stratification to within approximately 7 days, with no systematic early or late bias. Once stratification is established, modeled MLD tracks observations closely, reproducing both the seasonal magnitude and short-term fluctuations. The interannual variability in stratification onset is consistent with the simulated ice break-up timing. Simulated break-up occurs progressively later from 1998 to 2001 (DOY 150 to 165), and this is reflected in correspondingly later establishment of the mixed layer in both observations and model output. Open-water seasons shortened from 205 days in 1998 to 175 days in 2001, and are associated with shallower and less persistent summer mixed layers, as the lake has less time to accumulate heat before fall cooling begins. The exception is 1999, which stands out as a notable outlier. While the simulated break-up date (DOY 161) is later than 1998 (DOY 150), the observed stratification onset is substantially later than modeled and observed MLD remains considerably shallower than in other years. This decoupling between ice-off timing and subsequent thermal stratification suggests that factors beyond break-up date, such as early summer

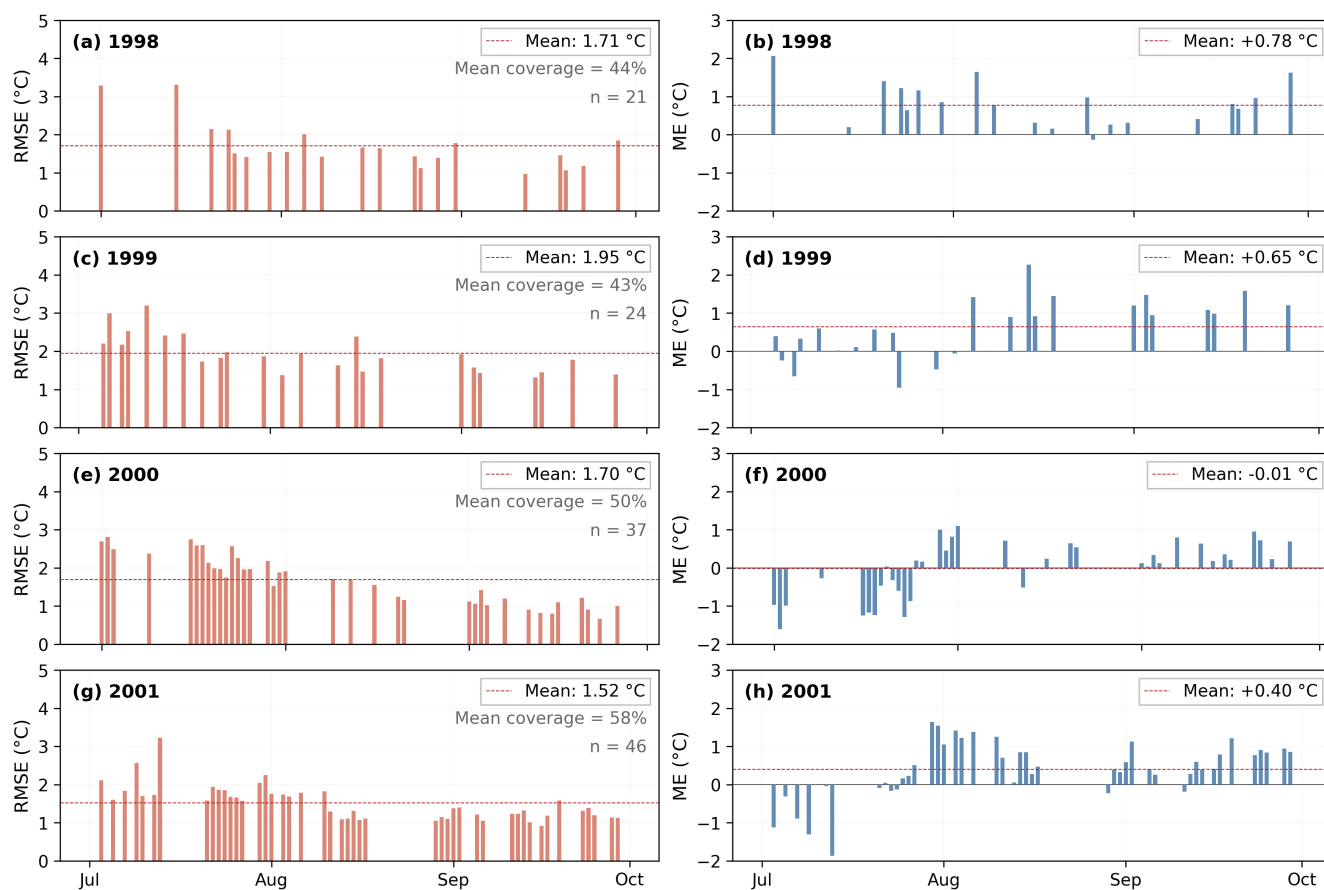


Figure 8. Lake surface water temperature evaluation results, showing the number of observations used in each year, overall performance (RMSE), and simulation bias (ME).

wind forcing or anomalous cloud cover reducing heat input, may have prolonged deep mixing in 1999. The model, driven by atmospheric forcing from a reanalysis that does not resolve convective cloud processes, does not fully capture this trend. This may in part reflect a limitation of the atmospheric forcing, whose 10 km horizontal resolution is an order of magnitude
 430 coarser than the 1 km model grid, meaning that microscale atmospheric variability over the lake, including land–lake breeze circulations and localized convective events, is not resolved in the forcing fields. Such unresolved variability could contribute to the discrepancy between modeled and observed stratification timing in years where local atmospheric conditions deviate from the regional-scale reanalysis.

4.5 Ice cover

435 The model reproduced the timing of ice-on and ice-off with mean absolute errors of 1 day and 2 days, respectively. It is worth noting that ice dynamics, such as floe transport by wind, ridging, and rafting have been excluded from the current model.

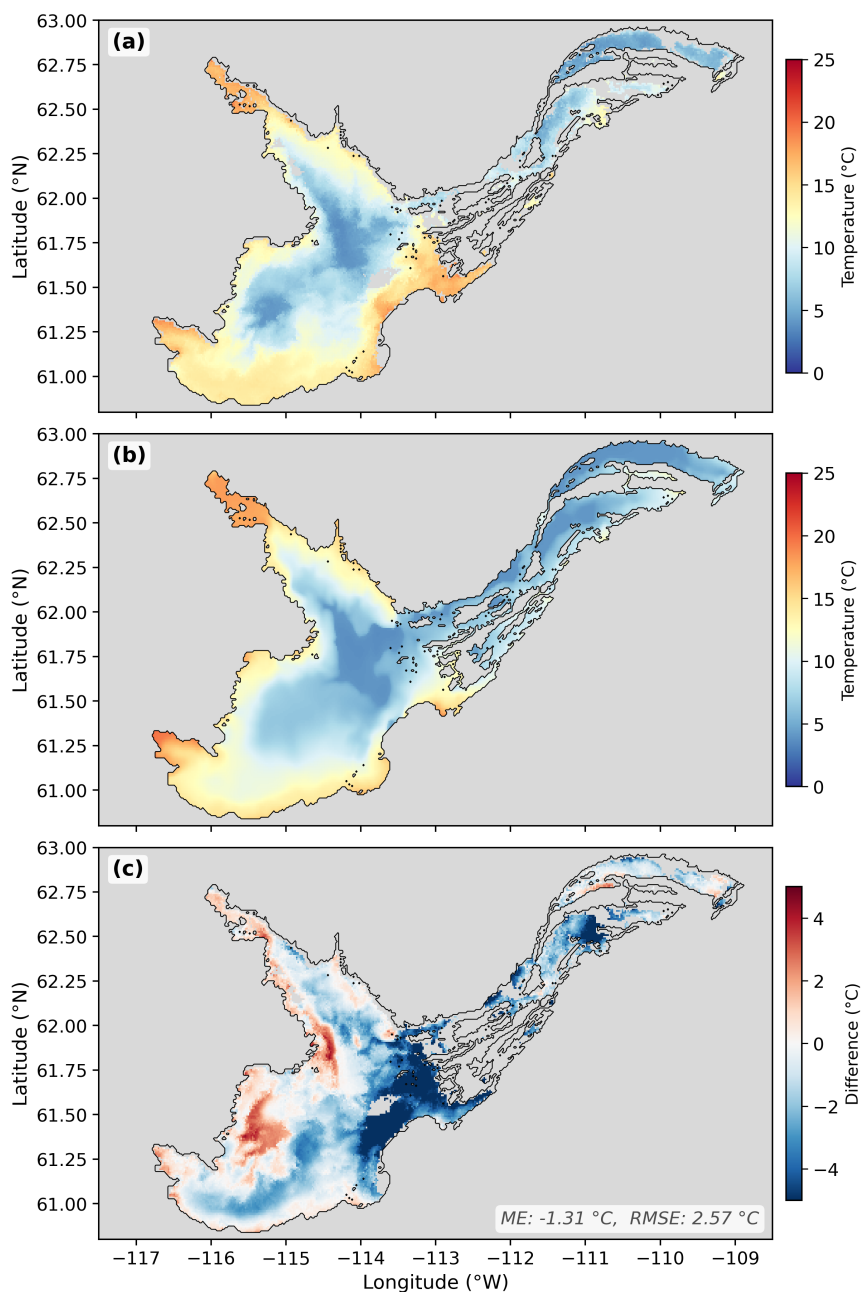


Figure 9. (a) Observed, (b) simulated LSWT of GSL, and (c) the temperature difference between simulation and observation on July 9, 2001.

Including ice dynamics could affect the ice break-up timing through the formation of smaller ice floes, exposed to lateral melting. The simulated freeze-up dates are generally delayed when compared to observations. Thermodynamics are the main driver during ice formation, and the inclusion of ice dynamics is not expected to substantially improve simulated freeze-up

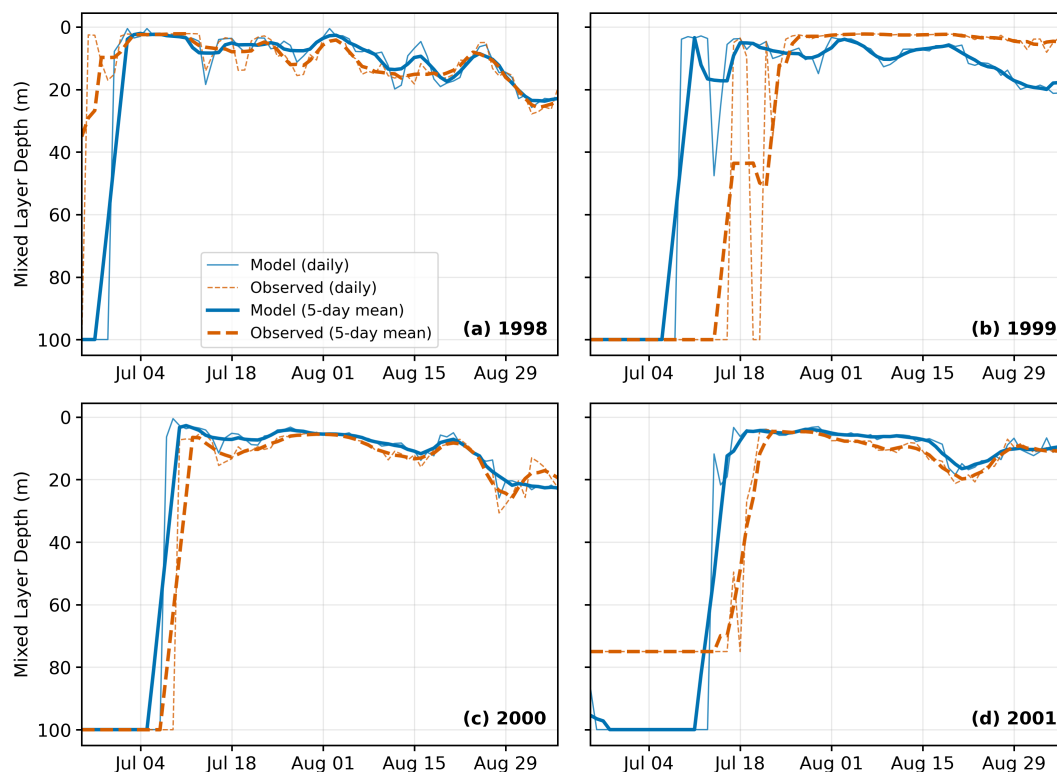


Figure 10. Observed and simulated mixed layer depth during the summer period at Site C throughout the open water seasons of evaluation years (1998–2001). Daily mean values and 5-day running means.

440 dates, which are already biased by only 1 day. The open water season duration is simulated with a mean absolute error of 2 days and is overpredicted in 1998, and underpredicted in 1999 and 2000.

Table 3. Simulated vs. observed ice break-up, freeze-up dates and open water duration.

Year	Ice break-up (DOY)		Ice freeze-up (DOY)		Open water season (days)	
	Observed	Simulated	Observed	Simulated	Observed	Simulated
1998	151	150	354	355	203	205
1999	159	161	346	347	187	186
2000	160	164	345	344	185	180
2001	164	165	339	340	175	175

4.6 Evaporation

Simulated evaporation at the IWI location closely matched observations, with seasonal totals differing by only 0.23% (1998) and 0.10% (1999), as shown in Table 4. Evaporation time series for the full evaluation period are presented in Figure 11. Daily evaporation magnitudes are reproduced by the model with an overall RMSE of 1.41 mm day^{-1} and ME of $-0.004 \text{ mm day}^{-1}$. Over Great Slave Lake, evaporation is characterized by periodic high-intensity events, which the model reproduces with moderate accuracy. During 1998, three of five such events were captured in both timing and magnitude, though the model produced two additional events absent from the observational record. In 1999, major evaporation event timing was represented within a few days; however, simulated magnitudes were systematically underestimated, and one spurious high-evaporation event appeared in the model output. Negative evaporation values, indicating condensation, have not been captured in either year of model evaluation. Although seasonal evaporation totals agree well with observations, reproducing the precise timing of peak events remains challenging. This limitation could be attributed to the use of reanalysis forcing, in which the atmospheric parameters governing stability conditions over the lake, and therefore the intensity of turbulent moisture fluxes, are insufficiently constrained by in situ observations.

Table 4. A comparison of observed and simulated cumulative evaporation over Inner Whaleback Island during the summer seasons of 1998 and 1999

Year	Observed evaporation (mm)	Simulated evaporation (mm)
1998	86.0	85.8
1999	96.1	96.0

4.7 Forcing sensitivity check

To assess sensitivity to atmospheric forcing, the model was run using an alternative reanalysis product – RDRS v2.1. The simulation used the same parameter set, initial conditions, and model spin-up duration as the CaSR v3.2 simulation. The temperature profiles were represented with overall RMSE of $2.39 \text{ }^{\circ}\text{C}$ and ME of $1.42 \text{ }^{\circ}\text{C}$. The MLD is predicted with a negative bias of 3.67 m, in contrast to the bias of 0.9 m observed with the CaSR v3.2 forcing. The total seasonal evaporation is overpredicted by 17.7% in 1998 and 5.7% in 1999. As for the lake ice cover, the break-up and freeze-up timing is reproduced with a mean error of 3 days. The results using RDRS v2.1 show greater errors in all performance metrics than with the CaSR v3.2 dataset, highlighting the model’s dependence on the choice of atmospheric forcing.

A one-at-a-time sensitivity analysis was conducted to further evaluate how variations in atmospheric forcing variables affect simulated LSWT, evaporative fluxes, and MLD, with the results shown in Figure 12. MLD was highly sensitive to wind speed: a 10% increase in wind speed deepened the mixed layer by 26%, while a 10% decrease produced a 23.6% shallower mixed layer. Wind speed also exerted a near-linear but opposite influence on evaporation: a 10% increase reduced evaporative flux by

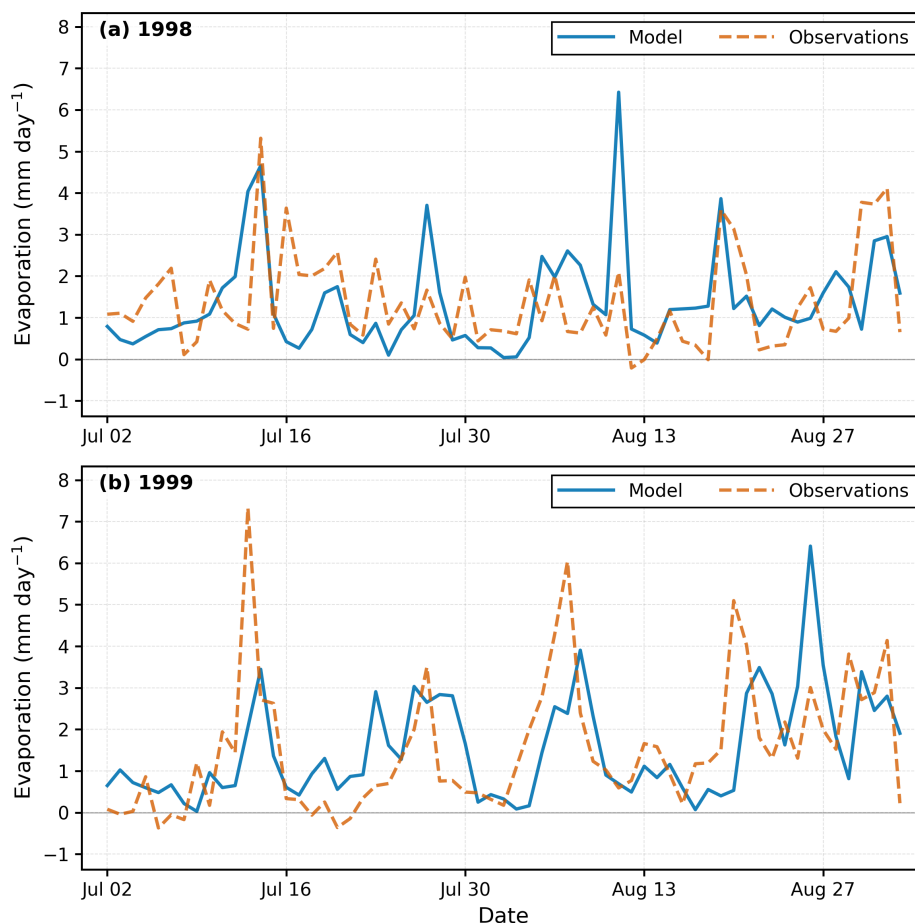


Figure 11. Simulated and observed daily evaporation magnitudes at IWI during the open-water seasons of (a) 1998 and (b) 1999.

9.6%, while a 10% decrease raised it by 10.7%. The effect of wind speed on LSWT was comparatively minor – a 10% increase lowered LSWT by 4.6% (in °C) and a 10% decrease raised it by 5.5%.

Air temperature produced similar-magnitude responses. A +1.27°C offset (equivalent to +10% of the mean magnitude of CaSR air temperature) resulted in a 20% decrease in MLD and increased evaporation by 20.8%, while a –1.27°C offset deepened the mixed layer by 17.8% and reduced evaporation by 21.2%. LSWT responded in the same direction as imposed air temperature change, though with lower intensity, decreasing by 6.7% with cooling and increasing by 6.9% with warming.

4.8 Lake circulation

Since no observational data are available for lake-current verification, a qualitative assessment of model outputs has been performed to evaluate the surface current patterns. Leon et al. (2007) simulated a counterclockwise gyre in the West Arm, a

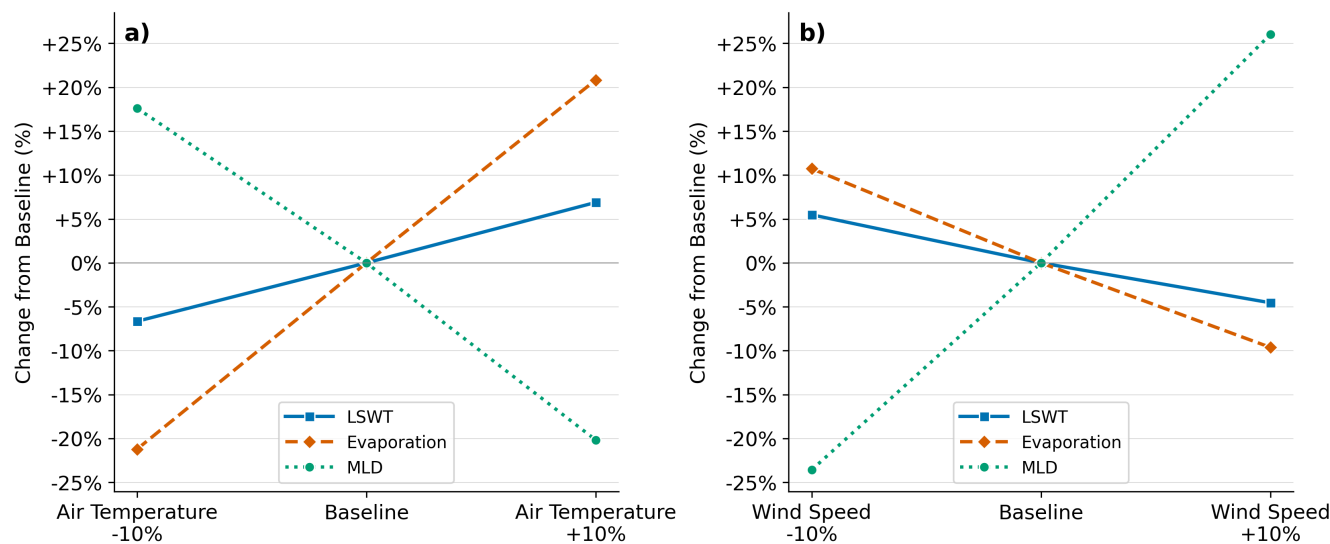


Figure 12. Sensitivity of simulated lake responses to $\pm 10\%$ perturbations in (a) air temperature and (b) wind speed. For comparability with the multiplicative wind speed perturbation, air temperature was perturbed by a uniform offset corresponding to 10% of the time-mean absolute magnitude of the forcing dataset air temperature ($+1.27^\circ\text{C}$ and -1.27°C , respectively). Percentage changes from the baseline simulation are shown for lake surface water temperature (LSWT), evaporation, and mixed layer depth (MLD).

feature corroborated by the sediment plume observations of Rouse et al. (2008b). The surface and depth integrated current distribution over the summer season (July 1 to September 30) is shown in Figure 13 for 1998 and Figure 14 for 2001.

These reveal significant interannual variability in surface currents, depending on the prevailing wind direction. Surface currents in 2001 closely match the depth integrated circulation. Circulation in the East Arm shows surface currents flowing towards the West Arm through the northern passages and into the East Arm through the southern passages. In 1998 strong northwestern winds pushed the warm surface water from the West Arm to the East Arm through the northern passages. Depth integrated flow, however, is neutral to biased towards the West Arm, suggesting an influx of denser, cold water from the East Arm in the form of subsurface currents. A net loss of water from East Arm to the West Arm is expected due to minor river inflows into the East Arm, and the only outflow being the Mackenzie River in the western part of the lake.

In the West Arm, the vast size of the basin makes the Coriolis effect a primary driver for circulation. Large anti-clockwise gyres form along the shores, showing circulation patterns similar to those reproduced by Leon et al. (2007) using ELCOM. Thermal bar formation during early summer can be considered a secondary driver, forming circulation along the shores of the basin during the rapid warming of shallow near-shore areas. The highest flow velocities are observed at the Slave River Delta and outflow into the Mackenzie River. Water inflows from the Slave River delta follow the anticlockwise gyre and flow northward, towards Yellowknife Bay, suggesting that any effluents brought in by the Slave River would be carried towards the city of Yellowknife.

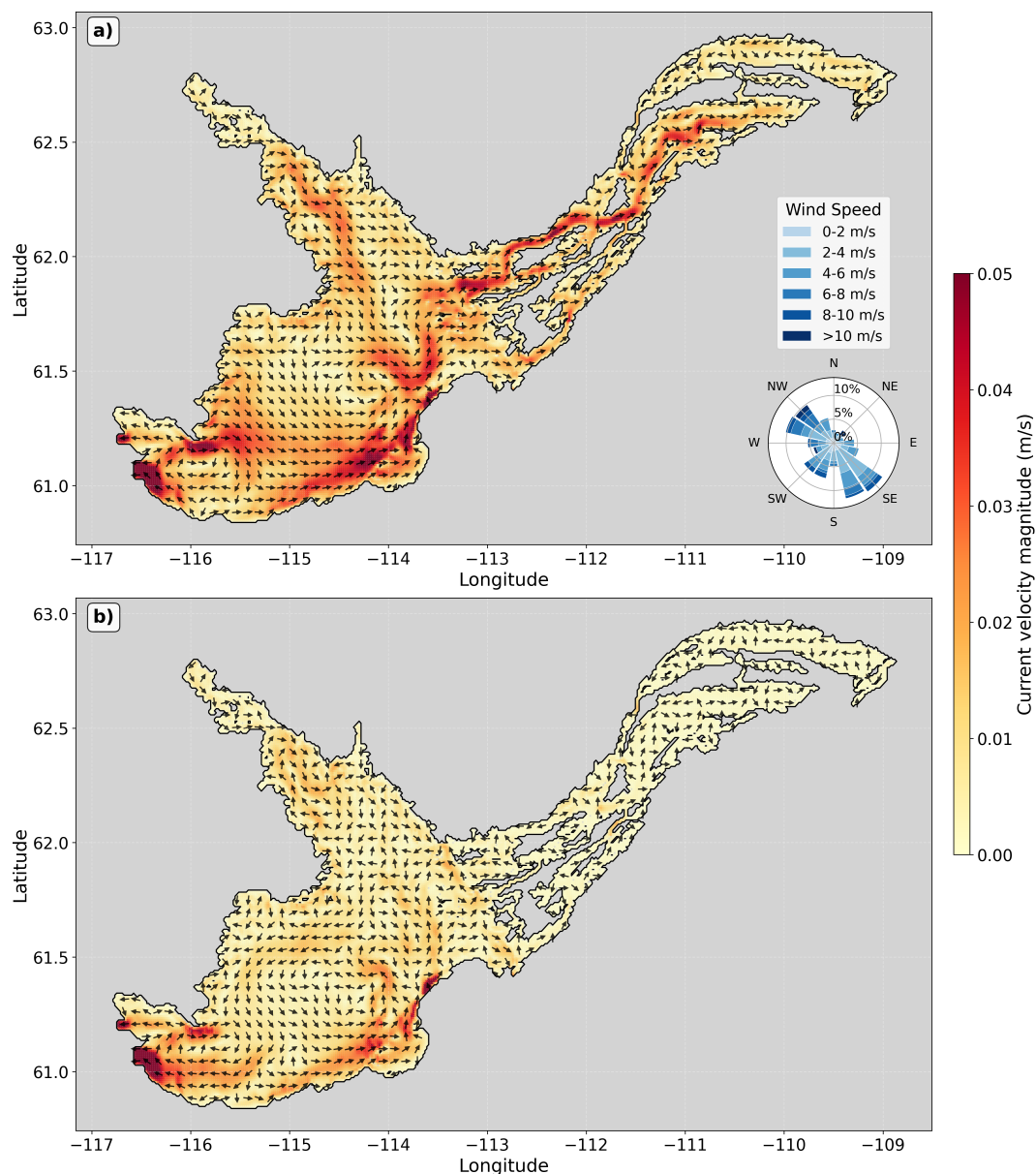


Figure 13. Mean simulated GSL (a) surface currents and (b) depth integrated circulation during the open water period (July 1 - September 30) of 1998. The wind rose shows wind direction and magnitude distribution over the period.

5 Discussion

NEMO-GSL v1.0, a three-dimensional hydrodynamic and ice configuration of Great Slave Lake built on NEMO v5.0 with the SI³ ice module, has been described and evaluated here. Performance was assessed against observations spanning 1998–2001:

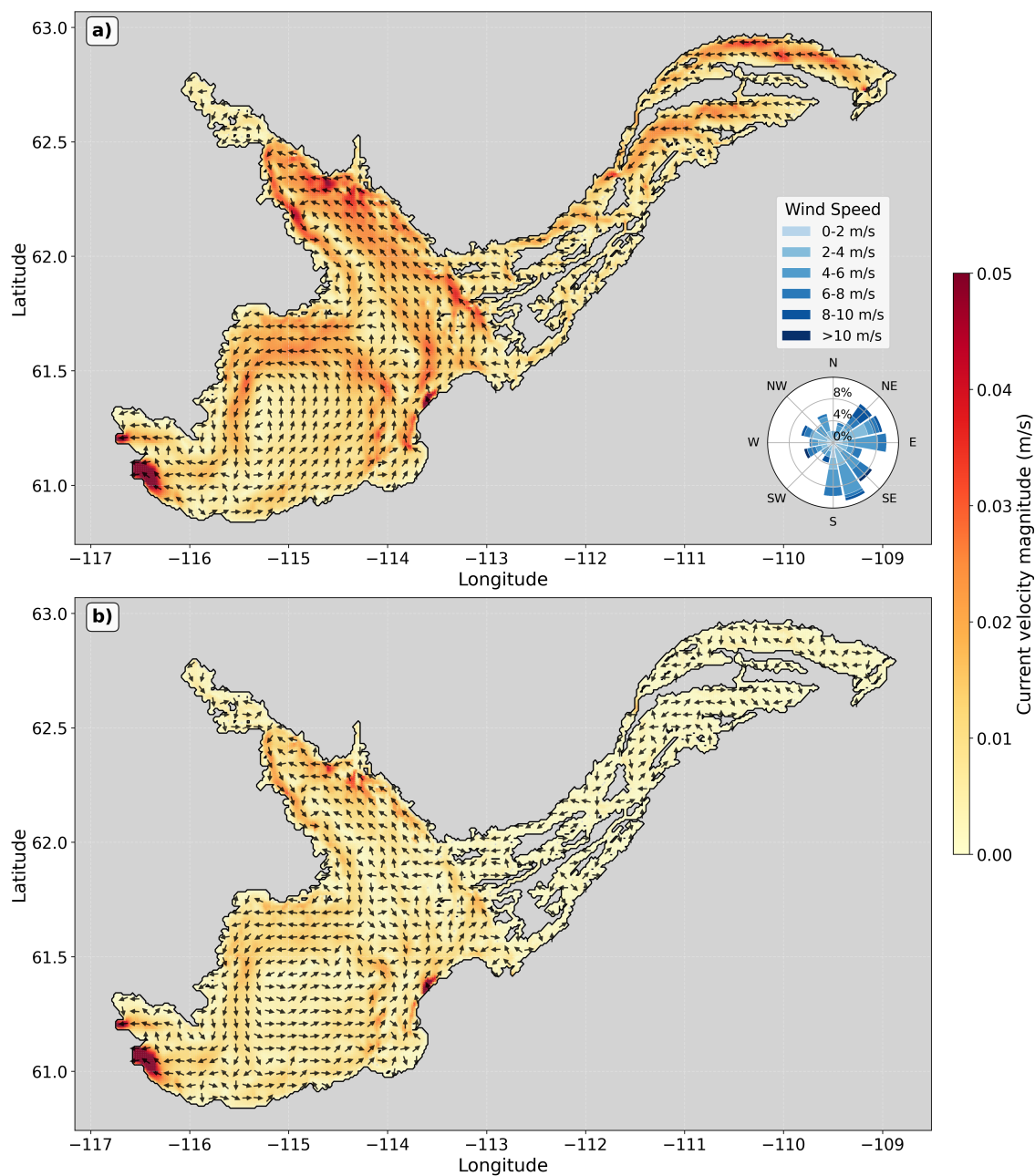


Figure 14. Mean simulated GSL (a) surface currents and (b) depth integrated circulation during the open water period (July 1 - September 30) of 2001. The wind rose shows wind direction and magnitude distribution over the period.

495 temperature profiles at six stations, evaporation at one station, and satellite-derived LSWT and ice freeze-up and break-up



dates. The simulation domain covers the entire lake, generated from available, albeit limited, depth observations, and riverine inflows and outflows derived from a Mackenzie River Basin cryospheric-hydrological model.

5.1 Thermal structure

Water column thermal profiles were reproduced with a mean RMSE of 2.08 °C and ME of +0.23 °C, while lake-wide LSWT was captured with an RMSE of 1.68 °C and an ME of +0.39 °C. These errors are comparable to those reported in studies of the Laurentian Great Lakes (Cannon et al., 2023; Dupont et al., 2012), and are reasonable given the size, depth heterogeneity, and data scarcity of GSL. Model performance varied across both stations and years. Site F, the shallowest station (12 m), exhibited consistently high RMSE values (2.08–2.56 °C) and a persistent warm bias (ME of 0.49–1.67 °C) across all years of available data. The shallow water column at this location amplifies sensitivity to errors in atmospheric forcing, particularly shortwave irradiance and wind-driven mixing, and to local bathymetric inaccuracies, since small absolute errors in modeled depth translate to proportionally larger errors in heat storage per unit area. Among the simulation years, 1999 stands out as the most challenging, with the highest RMSE values at nearly every station (2.89 °C at Site C and 2.54 °C at Site E and Site F). This suggests that the reanalysis atmospheric forcing dataset may not accurately represent conditions over the lake. At a horizontal resolution of 10 km, the atmospheric forcing may be too coarse to resolve sub-grid lake-atmosphere interactions at the 1 km model resolution, especially at lake boundaries.

MLD was assessed using a temperature threshold approach, yielding an MAE of 6.42 m and a positive bias of 0.90 m. The onset of seasonal stratification was captured to within approximately 7 days with no systematic directional bias. Once stratification is established, the model tracks observed MLD closely, reproducing both the seasonal deepening cycle and short-term wind-driven fluctuations. The largest MLD discrepancies occur during the spring transition, when the competing influences of surface heating, wind mixing, and under-ice convective overturning create a sensitive balance that is difficult to capture precisely.

5.2 Ice phenology

Lake ice freeze-up and break-up dates are simulated with mean errors of 1 day and 2 days, respectively, relative to satellite-derived observation. The model also captures the interannual progression in ice timing: observed break-up shifted from DOY 151 in 1998 to DOY 164 in 2001, with the simulation tracking this trend closely (DOY 150 to 165). The corresponding shortening of the open water season — from 203 days to 175 days observed, 205 to 175 simulated — is reflected in progressively later stratification onset and shallower summer mixed layers in later years, demonstrating internal physical consistency between the ice and thermal modules. The current model configuration uses a thermodynamic-only ice model. Implementation of full ice dynamics, including mechanical break-up of the ice pack, floe-size-dependent lateral melting, and wind-driven ice redistribution, could reduce the remaining timing errors, particularly during break-up when melt rates are accelerated by mechanical fragmentation. Observations of ice thickness distribution across the lake would enable a more complete evaluation beyond the phenological dates assessed in this study.



5.3 Evaporation and sensitivity to atmospheric forcing

Over-lake evaporation was modeled with a mean error of 0.17% at the single available observation station. Evaporation is highly sensitive to atmospheric forcing and the duration of the open water season. Sensitivity experiments with $\pm 10\%$ perturbations to air temperature and wind speed revealed that simulated evaporation and MLD are particularly sensitive to wind forcing, with air temperature exerting a moderate influence on both variables. These results underscore the importance of high-quality wind fields for simulating the vertical thermal structure and evaporative losses in large northern lakes, and suggest that wind speed biases in the atmospheric forcing are an important source of uncertainty in the simulated thermocline depth.

5.4 Lake circulation

Simulated currents, while not verified against observations due to the absence of current-meter or drifter data during the evaluation period, show physically consistent spatial patterns driven by geostrophic effects and differential surface heating. Validating the circulation field would require dedicated observational campaigns and would further constrain the model's representation of basin-scale heat redistribution and its influence on spatial variability in ice phenology across the lake.

5.5 Sources of uncertainty

Multiple sources of uncertainty in the current model configuration have been identified, and addressing these could help improve the simulation accuracy. Gaps in the bathymetric data, especially in the central and western portions of the main basin, introduce uncertainty in the modeled circulation and thermal structure. Additional sonar surveys in data-sparse areas would allow the model domain to be refined. The model evaluation period, while providing a meaningful basis for model skill assessment, could be extended to additional years and stations to further strengthen confidence in the model's ability to reproduce interannual variability. Additional temperature profile observation locations, particularly in the East Arm of the lake, would help assess the model performance across the entire lake. Better representation of river inflow temperatures is needed to fully capture lake and ice thermodynamics and understand riverine inflow contributions to lake thermal structure. An influx of warm water from rivers flowing from the south, especially Slave River, could accelerate ice break-up dynamics and reduce errors in the overall energy balance of the lake. Atmospheric forcing quality, particularly wind speed and its spatial variability over the lake, directly controls the simulated MLD and evaporation, as demonstrated by the sensitivity experiments. Refinements to atmospheric reanalysis products could yield improved representation of simulated lake dynamics. The current evaluation of ice cover is limited to freeze-up and break-up timing derived from satellite imagery, which constrains only the temporal endpoints of the ice season. Spatially distributed observations of ice thickness would enable evaluation of the simulated ice mass budget and the spatial heterogeneity of ice growth across the lake's morphologically distinct basins. The present configuration employs a thermodynamic-only ice formulation that grows and melts ice in place without mechanical redistribution. Incorporating prognostic ice dynamics could sharpen the simulated break-up transition.



6 Conclusions

Despite the considerable challenges posed by the remoteness of the Great Slave Lake basin, its complex physiography, and the scarcity of observational data, the NEMO-GSL v1.0 successfully integrated diverse historical, contemporary, and remotely sensed datasets to develop, force, and evaluate a basin-scale hydrodynamic model. The strong agreement with independent observations demonstrates that robust physically based simulations can be achieved even in data-limited northern environments. As such, this work provides both a modeling system for Great Slave Lake and a transferable configuration for advancing process-based investigations of large subarctic lakes worldwide.

The configuration's success, demonstrated through evaluation against multiple independent observations across thermal, ice, and surface energy balance processes, indicates that ocean model physics, with targeted parameterization adaptation, provides an appropriate foundation for subarctic lake simulation. The configuration addresses the most pressing constraints encountered in northern lake modeling through several integrated methodological innovations. A bathymetry interpolation methodology was developed and tested on sparse observational networks, yielding reliable basin-wide depth representations suitable for data-scarce regions. Parameterization choices for turbulence closure have been explicitly tailored to weakly stratified freshwater systems. SI^3 ice thermodynamic tuning for prolonged ice cover reflects the distinct thermal and optical properties of freshwater lake ice under extended seasonal conditions. The customized river module accounts for dynamic stage-discharge relationships and ice-load hydrostatic effects, both critical for large freshwater systems with significant seasonal water-level fluctuations. An atmospheric forcing sensitivity protocol systematically identifies wind dominance as the controlling variable in subarctic lake thermal structure, a finding with direct implications for forcing data selection in northern applications.

Climate-impact pathways of GSL are not unique to the region, but representative of other high-latitude freshwater systems. Longer open-water seasons would increase heat accumulation in the water column, delaying freeze-up and reducing ice thickness through positive feedback. These changes would cascade through thermal stratification strength and mixed-layer depth to affect nutrient transport, primary productivity, and the food-web structure dependent on these physical-biological couplings. The configuration thus provides an operational tool for quantifying these dynamics under future climate scenarios across northern lakes sharing similar physical characteristics.

Beyond the GSL case study, the primary deliverable of this work is a reusable, openly archived NEMO configuration, including bathymetry-generation scripts, forcing-preparation code, and documented parameter choices, rather than a single-site simulation result. For practitioners in northern regions, NEMO-GSL v1.0 eliminates the requirement to configure NEMO from ocean defaults or to invest in new model development for each individual system; it instead provides a starting configuration explicitly designed for subarctic conditions, with documented parameter values, atmospheric forcing sensitivity, and coupling strategies. Explicit protocols for parameterizing weak-stratification freshwater dynamics using ocean-model architecture are established, providing a template for extending NEMO to similar systems. Ice-load correction algorithms necessary for water-level calculations under extended seasonal ice cover are documented and tested, addressing a gap in existing NEMO sea-ice configurations. These contributions establish a transferable workflow for NEMO adaptation to other high-latitude domains,



590 positioning this work as both a practical tool for northern lake modeling and a methodological advance for the broader Earth
system modeling community.

Author contributions. J.S.: bathymetry product development, model implementation and simulation execution, analysis, figures, first draft
and editing; A.P.: supervision, MAGS legacy data coordination, review and editing; Q.Z.: supervision, conceptualization of lake modeling,
turbulence parameterization advice, review and editing; M.E.: MESH runoff forcing and advice on Mackenzie River Basin hydrology, review
595 and editing; J.W.P.: review and editing.

Competing interests. The authors declare that they have no conflict of interest.

Code and data availability. The model configuration, input generation scripts, and analysis code are publicly available at <https://github.com/StanJon/nemo-gsl-gmd-2026> and archived on Zenodo at <https://doi.org/10.5281/zenodo.20513121> (Stankevicius et al., 2026a). Model
inputs, bathymetry, and simulation outputs are archived separately on Zenodo at <https://doi.org/10.5281/zenodo.20512110> (Stankevicius
600 et al., 2026b). NEMO v5.0 is freely available at <https://forge.nemo-ocean.eu> under the CeCILL v2 licence.

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References

- Attiah, G., Scott, K. A., and Kheyrollah Pour, H.: Spatially distributed modeling of lake ice trends and distribution in the North Slave Region, NWT, Canada, *Water Resources Research*, 61, <https://doi.org/https://doi.org/10.1029/2025WR040216>, 2025.
- Blanken, P., Rouse, W., and Schertzer, W. M.: The time scales of evaporation from Great Slave Lake, pp. 181–196, Springer Berlin Heidelberg, Berlin, Heidelberg, ISBN 978-3-540-75136-6, https://doi.org/10.1007/978-3-540-75136-6_10, 2008.
- Blanken, P. D., Rouse, W. R., and Schertzer, W. M.: Enhancement of evaporation from a large northern lake by the entrainment of warm, dry air, *Journal of Hydrometeorology*, 4, 680–693, [https://doi.org/10.1175/1525-7541\(2003\)004<0680:EOEFAL>2.0.CO;2](https://doi.org/10.1175/1525-7541(2003)004<0680:EOEFAL>2.0.CO;2), 2003.
- Bring, A., Fedorova, I., Dibike, Y., Hinzman, L., Mård, J., Mernild, S. H., Prowse, T., Semenova, O., Stuefer, S. L., and Woo, M.: Arctic terrestrial hydrology: A synthesis of processes, regional effects, and research challenges, *Journal of Geophysical Research: Biogeosciences*, 121, 621–649, <https://doi.org/10.1002/2015JG003131>, 2016.
- Cai, Y., Duguay, C. R., and Ke, C.-Q.: Lake ice phenology in the Northern Hemisphere extracted from SMMR, SSM/I and SSMIS data from 1979 to 2020, <https://doi.org/10.1594/PANGAEA.937904>, 2021.
- Cannon, D., Fujisaki-Manome, A., Wang, J., Kessler, J., and Chu, P.: Modeling changes in ice dynamics and subsurface thermal structure in Lake Michigan-Huron between 1979 and 2021, *Ocean Dynamics*, 73, 201–218, 2023.
- Carmack, E. and Vagle, S.: Thermobaric processes both drive and constrain seasonal ventilation in deep Great Slave Lake, Canada, *Journal of Geophysical Research: Earth Surface*, 126, e2021JF006288, <https://doi.org/10.1029/2021JF006288>, 2021.
- Carrea, L., Crétaux, J.-F., Liu, X., Wu, Y., Bergé-Nguyen, M., Calmettes, B., Duguay, C., Jiang, D., Merchant, C. J., Mueller, D., Selmes, N., Simis, S., Spyarakos, E., Stelzer, K., Warren, M., Yesou, H., and Zhang, D.: ESA Lakes Climate Change Initiative: Lake products, Version 2.1, <https://doi.org/10.5285/7FC9DF8070D34CACAB8092E45EF276F1>, 2024.
- CHS: Great Slave Lake, western portion [Map], corrected to notice to mariners: 2022-03-18, 1990.
- CHS: Great Slave Lake, eastern portion [Map], corrected to notice to mariners: 2016-09-30, 2004.
- Dupont, F., Chittibabu, P., Fortin, V., Rao, Y. R., and Lu, Y.: Assessment of a NEMO-based hydrodynamic modelling system for the Great Lakes, *Water Quality Research Journal*, 47, 198–214, <https://doi.org/10.2166/wqrjc.2012.014>, 2012.
- Edson, J. B., Jampana, V., Weller, R. A., Bigorre, S. P., Plueddemann, A. J., Fairall, C. W., Miller, S. D., Mahrt, L., Vickers, D., and Hersbach, H.: On the exchange of momentum over the open ocean, *Journal of Physical Oceanography*, 43, 1589–1610, <https://doi.org/10.1175/JPO-D-12-0173.1>, 2013.
- Elshamy, M., Pomeroy, J. W., Pietroniro, A., Wheeler, H., Abdelhamed, M., and Davison, B.: Land surface hydrological modelling of the Mackenzie River Basin: Parametrization to simulate streamflow and permafrost dynamics, *Journal of Hydrology*, 659, 133–134, <https://doi.org/10.1016/j.jhydrol.2025.133134>, 2025.
- Gasset, N., Fortin, V., Dimitrijevic, M., Carrera, M., Bilodeau, B., Muncaster, R., Gaborit, É., Roy, G., Pentcheva, N., Bulat, M., Wang, X., Pavlovic, R., Lespinas, F., Khedhaouiria, D., and Mai, J.: A 10 km North American precipitation and land-surface reanalysis based on the GEM atmospheric model, *Hydrology and Earth System Sciences*, 25, 4917–4945, <https://doi.org/10.5194/hess-25-4917-2021>, 2021.
- Gibson, J. J., Prowse, T. D., and Peters, D. L.: Hydroclimatic controls on water balance and water level variability in Great Slave Lake, *Hydrological Processes*, 20, 4155–4172, <https://doi.org/10.1002/hyp.6424>, 2006.
- Government of the Northwest Territories: Great Slave Lake community survey results: An overview of key findings, Tech. rep., Water Management and Stewardship Division, Department of Environment and Climate Change, Yellowknife, NT, 2024.



- 645 Herdendorf, C. E.: Large lakes of the world, *Journal of Great Lakes Research*, 8, 379–412, [https://doi.org/10.1016/S0380-1330\(82\)71982-3](https://doi.org/10.1016/S0380-1330(82)71982-3), 1982.
- Hordoir, R., Axell, L., Höglund, A., Dieterich, C., Fransner, F., Gröger, M., Liu, Y., Pemberton, P., Schimanke, S., Andersson, H., et al.: Nemo-Nordic 1.0: a NEMO-based ocean model for the Baltic and North seas, *Geoscientific Model Development*, 12, 363–386, <https://doi.org/10.5194/gmd-12-363-2019>, 2019.
- 650 Huang, A., Rao, Y. R., Lu, Y., and Zhao, J.: Hydrodynamic modeling of Lake Ontario: An intercomparison of three models, *Journal of Geophysical Research: Oceans*, 115, <https://doi.org/10.1029/2010jc006269>, 2010.
- Leon, L. F., Lam, D., Schertzer, W. M., Swayne, D., and Imberger, J.: Towards coupling a 3D hydrodynamic lake model with the Canadian Regional Climate Model: Simulation on Great Slave Lake, *Environmental Modelling & Software*, 22, 787–796, <https://doi.org/10.1016/j.envsoft.2006.03.005>, 2007.
- Li, X., Peng, S., Xi, Y., Woolway, R. I., and Liu, G.: Earlier ice loss accelerates lake warming in the Northern Hemisphere, *Nature Commu-*
 655 *nications*, 13, 5156, <https://doi.org/10.1038/s41467-022-32830-y>, 2022.
- Lüpkes, C., Gryanik, V. M., Hartmann, J., and Andreas, E. L.: A parametrization, based on sea ice morphology, of the neutral atmospheric drag coefficients for weather prediction and climate models, *Journal of Geophysical Research: Atmospheres*, 117, 2012JD017630, <https://doi.org/10.1029/2012JD017630>, 2012.
- Madec, G. and the NEMO System Team: NEMO ocean engine manual, <https://doi.org/10.5281/zenodo.1464816>, 2024.
- 660 Ménard, P., Duguay, C. R., Flato, G. M., and Rouse, W. R.: Simulation of ice phenology on Great Slave Lake, Northwest Territories, Canada, *Hydrological Processes*, 16, 3691–3706, <https://doi.org/10.1002/hyp.1230>, 2002.
- Pawlowicz, R. and Feistel, R.: Limnological applications of the thermodynamic equation of seawater 2010 (TEOS-10), *Limnology and Oceanography: Methods*, 10, 853–867, <https://doi.org/10.4319/lom.2012.10.853>, 2012.
- Rantanen, M., Karpechko, A., Lipponen, A., Nordling, K., Hyvärinen, O., Ruostenoja, K., Vihma, T., and Laaksonen, A.: The Arctic has
 665 warmed nearly four times faster than the globe since 1979, *Communications Earth & Environment*, 3, 168, <https://doi.org/10.1038/s43247-022-00498-3>, 2022.
- Rawson, D. S.: Estimating the fish production of Great Slave Lake, *Transactions of the American Fisheries Society*, 77, 81–92, [https://doi.org/10.1577/1548-8659\(1947\)77\[81:etfpog\]2.0.co;2](https://doi.org/10.1577/1548-8659(1947)77[81:etfpog]2.0.co;2), 1949.
- Rawson, D. S.: The physical limnology of Great Slave Lake, *Journal of the Fisheries Research Board of Canada*, 8a, 3–66,
 670 <https://doi.org/10.1139/f50-001>, 1950.
- Rawson, D. S.: Studies of the Fish of Great Slave Lake, *Journal of the Fisheries Research Board of Canada*, 8b, 207–240, <https://doi.org/10.1139/f50-014>, 1951.
- Rawson, D. S.: The Net Plankton of Great Slave Lake, *Journal of the Fisheries Research Board of Canada*, 13, 53–127, <https://doi.org/10.1139/f56-006>, 1956.
- 675 Rouse, W. R., Oswald, C. M., Binyamin, J., Blanken, P. D., Schertzer, W. M., and Spence, C.: 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:IASVOT>2.0.CO;2](https://doi.org/10.1175/1525-7541(2003)004<0720:IASVOT>2.0.CO;2), 2003.
- Rouse, W. R., Blanken, P. D., Bussièrès, N., Walker, A. E., Oswald, C. J., Schertzer, W. M., and Spence, C.: An investigation of the thermal and energy balance regimes of Great Slave and Great Bear Lakes, *Journal of Hydrometeorology*, 9, 1318–1333,
 680 <https://doi.org/10.1175/2008jhm977.1>, 2008a.



- Rouse, W. R., Blanken, P. D., Duguay, C. R., Oswald, C. J., and Schertzer, W. M.: Climate-lake interactions, pp. 139–160, Springer Berlin Heidelberg, Berlin, Heidelberg, ISBN 978-3-540-75136-6, https://doi.org/10.1007/978-3-540-75136-6_8, 2008b.
- Rühland, K. M., Evans, M., and Smol, J. P.: Arctic warming drives striking twenty-first century ecosystem shifts in Great Slave Lake (Subarctic Canada), North America's deepest lake, *Proceedings of the Royal Society B: Biological Sciences*, 290, 20231252, <https://doi.org/10.1098/rspb.2023.1252>, 2023.
- Schertzer, W. M.: Digital bathymetry of Great Slave Lake, https://publications.gc.ca/collections/collection_2019/eccc/en13-5/En13-5-00-257-eng.pdf, accessed: 2025-04-25, 2000.
- Schertzer, W. M., Rouse, W. R., Blanken, P. D., and Walker, A. E.: Over-lake meteorology and estimated bulk heat exchange of Great Slave Lake in 1998 and 1999, *Journal of Hydrometeorology*, 4, 649–659, [https://doi.org/10.1175/1525-7541\(2003\)004<0649:omaebh>2.0.co;2](https://doi.org/10.1175/1525-7541(2003)004<0649:omaebh>2.0.co;2), publisher: American Meteorological Society, 2003.
- Schertzer, W. M., Rouse, W. R., Blanken, P. D., Walker, A. E., Lam, D. C. L., León, L., and Woo, M.-k.: Interannual variability of the thermal components and bulk heat exchange of Great Slave Lake, in: *Cold Region Atmospheric and Hydrologic Studies. The Mackenzie GEWEX Experience*, pp. 197–219, Springer Berlin Heidelberg, Berlin, Heidelberg, ISBN 9783540749271, 2008.
- Schindler, D. W.: The cumulative effects of climate warming and other human stresses on Canadian freshwaters in the new millennium, *Canadian Journal of Fisheries and Aquatic Sciences*, 58, 18–29, 2001.
- Schindler, D. W. and Smol, J. P.: Cumulative effects of climate warming and other human activities on freshwaters of Arctic and subarctic North America, *AMBIO: A Journal of the Human Environment*, 35, 160–168, [https://doi.org/10.1579/0044-7447\(2006\)35\[160:CEOCWA\]2.0.CO;2](https://doi.org/10.1579/0044-7447(2006)35[160:CEOCWA]2.0.CO;2), 2006.
- Sharma, S., Richardson, D. C., Woolway, R. I., Imrit, M. A., Bouffard, D., Blagrove, K., et al.: Loss of ice cover, shifting phenology, and more extreme events in Northern Hemisphere lakes, *Journal of Geophysical Research: Biogeosciences*, 126, e2021JG006348, <https://doi.org/10.1029/2021JG006348>, 2021.
- Stankevicius, J., Pietroniro, A., Zhou, Q., Elshamy, M., and Pomeroy, W. J.: NEMO-SI³ hydrodynamic and ice model of Great Slave Lake: model configuration and analysis code, <https://doi.org/10.5281/zenodo.20513121>, 2026a.
- Stankevicius, J., Pietroniro, A., Zhou, Q., Elshamy, M., and Pomeroy, W. J.: NEMO-SI³ Great Slave Lake: model inputs, configuration, and outputs (1997–2001), <https://doi.org/10.5281/zenodo.20512110>, 2026b.
- Stankevicius, S., Pietroniro, A., Zhou, Q., and Connon, R.: A climatological approach to predicting water Level of Great Slave Lake, *Canadian Water Resources Journal*, p. 223–235, <https://doi.org/10.1080/07011784.2025.2519131>, 2025.
- Sweeney, B. N., Polvani, L. M., and Singh, H. A.: Internal variability increased Arctic amplification during 1980–2022, *Geophysical Research Letters*, 50, e2023GL106060, <https://doi.org/10.1029/2023GL106060>, 2023.
- Trebitz, A. S.: Characterizing seiche and tide-driven daily water Level fluctuations affecting coastal ecosystems of the Great Lakes, *Journal of Great Lakes Research*, 32, 102–116, [https://doi.org/10.3394/0380-1330\(2006\)32\[102:CSATDW\]2.0.CO;2](https://doi.org/10.3394/0380-1330(2006)32[102:CSATDW]2.0.CO;2), 2006.
- Umlauf, L. and Burchard, H.: A generic length-scale equation for geophysical turbulence models, *Journal of Marine Research*, 61, 235–265, <https://doi.org/10.1357/002224003322005087>, 2003.
- Umlauf, L. and Burchard, H.: Second-order turbulence closure models for geophysical boundary layers. A review of recent work, *Continental Shelf Research*, 25, 795–827, <https://doi.org/10.1016/j.csr.2004.08.004>, 2005.
- Woo, M.-K. and Thorne, R.: Streamflow in the Mackenzie Basin, Canada, Arctic, 56, 328–340, <https://doi.org/10.14430/arctic630>, 2003.
- Woolway, R. I. and Merchant, C. J.: Worldwide alteration of lake mixing regimes in response to climate change, *Nature Geoscience*, 12, 271–276, <https://doi.org/10.1038/s41561-019-0322-x>, 2019.



- Woolway, R. I., Kraemer, B. M., Lenters, J. D., Merchant, C. J., O'Reilly, C. M., and Sharma, S.: Global lake responses to climate change, 720 Nature Reviews Earth & Environment, 1, 388–403, <https://doi.org/10.1038/s43017-020-0067-5>, 2020.
- Woolway, R. I., Sharma, S., and Smol, J. P.: Lakes in hot water: The impacts of a changing climate on aquatic ecosystems, BioScience, 72, 1050–1061, <https://doi.org/10.1093/biosci/biac052>, 2022.
- Zalesak, S. T.: Fully multidimensional flux-corrected transport algorithms for fluids, Journal of Computational Physics, 31, 335–362, [https://doi.org/10.1016/0021-9991\(79\)90051-2](https://doi.org/10.1016/0021-9991(79)90051-2), 1979.