

A climatology of seasonal freshwater accumulation as a basis for defining the freshwater budget in Baffin Bay and on the Labrador shelf[☆]

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ABSTRACT

The western gateway linking the Arctic to the North Atlantic, encompassing Baffin Bay, the Labrador Shelf, and the West Greenland Shelf (called BBLS hereafter), plays an important role in regulating the freshwater budget of the subpolar North Atlantic. This region channels locally and remotely sourced freshwater towards the Labrador Sea, where it influences deep water formation and ultimately enters the lower limb of the Atlantic Meridional Overturning Circulation (AMOC). Recent changes in the BBLS freshwater budget are impacted by enhanced Greenland ice sheet retreat, reduced seasonal sea ice melting, and increasing Arctic Ocean freshwater content. Separating seasonal freshwater accumulation from persistent interannual anomalies of remote origin remains a major challenge, as overlapping time scales can obscure climatologically relevant variability and hinder our understanding of freshwater influences on coastal dynamics, deep water formation, and AMOC variability. Here, we construct a high-resolution weekly climatology of Freshwater Content Change (FCC) in the BBLS, independent of a subjective salinity reference. FCC, in our formulation, represents the actual freshwater volume gained in the region from April to October through regional seasonal processes. Totalling 3509 km³, the summer-integrated FCC is in close agreement with independent estimates of the local freshwater budget. The climatology reveals strong regional contrasts in the seasonal evolution of freshwater accumulation, reflecting the combined influences of sea ice melt, glacier runoff, and remotely advected freshwater signals entering through the Canadian Arctic Archipelago and along the West Greenland pathway. The FCC framework provides a physically interpretable description of seasonal freshwater accumulation in this key Arctic–subpolar gateway and a basis for investigating low-frequency freshwater anomalies and their propagation toward the Labrador Sea and the wider subpolar North Atlantic.

1. Introduction

The subpolar North Atlantic (SPNA), comprising the Irminger Sea, Labrador Sea, Iceland Basin, and Rockall Trough, is an essential region of the ocean, where warm and saline subtropical Atlantic waters meet with cold and fresher water of Arctic and subpolar origin. The interaction between water masses of these two types, involving their mixing and transformation, results in exchanges of heat, salt, and materials (e. g., nutrients, carbon), and the formation of intermediate and deep

waters. The formation of intermediate and deep water in the SPNA contributes to the lower limb of the Atlantic Meridional Overturning Circulation (AMOC), and by extension the Global Ocean Conveyor Belt (Haupt and Seidov 2007), thus affecting both the regional and global climate.

At the transition between the Arctic and the SPNA, the region confined to Baffin Bay to the north, the West Greenland Shelf to the east, and the Labrador Shelf to the south and west, plays a crucial role for the oceanic climate and ecosystem as both a pathway and receiving basin for

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freshwater of local and remote origin. We, hereafter, name this domain the Baffin Bay-Labrador Shelf region (BBLs). The local freshwater is generated mainly by sea ice melting in and around Baffin Bay and off the Labrador coast, as well as by river runoff and freshwater input from the Greenland Ice Sheet and glaciers (Cuny et al. 2005; Curry et al. 2011, Curry et al. (2014); Bamber et al. 2018). The remote freshwater originates from the Arctic, flowing via the Fram Strait towards the West Greenland shelf, and via the Canadian Arctic Archipelago into the Baffin Bay (Fig. 1a) (Dickson et al. 2007; Haine and Coauthors, (2015); Carmack et al. 2016; Tesdal and Haine 2020; Serreze et al. 2006). In particular, freshwater from the Beaufort Gyre is released through these two pathways into the SPNA with a periodicity of 5 to 10 years (Zhang et al. 2021).

The circulation in the BBLs is cyclonic, with the northward-flowing West Greenland Current (WGC) on the eastern side of Baffin Bay, and the Baffin Island Current on its western side (Fig. 1b). The Labrador Current (LC) is the extension of the Baffin Island Current, and it also receives waters from a westward-flowing branch of the WGC and the Hudson Bay

outflow. The LC flows southward from 60°N, with its inshore branch flowing along the North American coast as far as the mid-Atlantic bight (~37°N). Its offshore branch follows the shelf break, coming into close contact and interacting with the warm and saline North Atlantic Current south of the Grand Banks of Newfoundland (Fig. 1b). The currents through the BBLs cyclonic gyre transport increasingly fresher waters as freshwater additions by sea ice melt, glacier runoff, and other land runoff occur along their pathway (e.g., Mertz et al. 1993; Benetti et al. 2016). The freshwater flowing on the shelf interacts with and ultimately spreads to the interior Labrador Sea (Dukhovskoy et al. 2021; Luo et al. 2016; Gillard et al. 2016) via the local currents, eddies, and Ekman transport (Schulze Chretien and Frajka-Williams 2018; Fratantoni and McCartney 2010; Lilly et al. 2003; Reverdin et al. 2003).

Variability in freshwater input from sea ice melt, glacier runoff, and river discharge can influence the AMOC through its impact on Labrador Sea Water (LSW) formation. LSW is produced by intense winter convection in the Labrador Sea (Clarke and Gascard 1983; Lazier et al. 2002; Yashayaev 2007a; Yashayaev and Loder 2016; Yashayaev 2024;

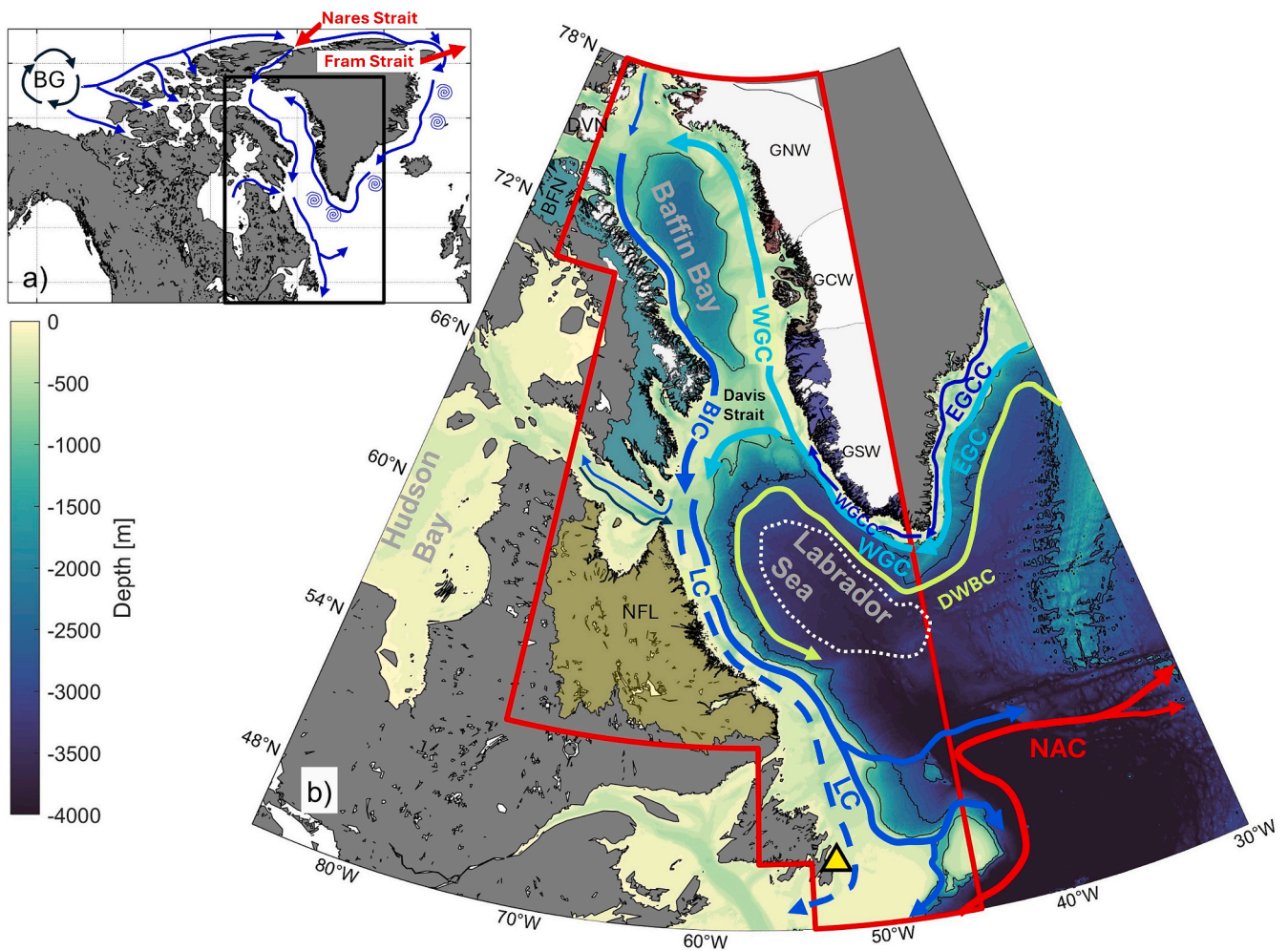


Fig. 1. (a) Schematic view of the freshwater pathways (blue lines) in the Western Arctic and Subpolar North Atlantic. BG refers to the Beaufort Gyre. The small spirals indicate eddy shedding that contributes to the spreading of the freshwater to the interior. The black square indicates the region drawn in (b). (b) Close-up view of the subpolar North Atlantic with our study region enclosed within the red shape. The arrows indicate the main currents in the regions which are mentioned in this paper, counterclockwise: Deep Western Boundary Current (DWBC), East Greenland Current (EGC), East Greenland Coastal Current (EGCC), West Greenland Current (WGC), West Greenland Coastal Current (WGCC), Baffin Island Current (BIC), Labrador Current (LC), with its inshore branch indicated by the blue dotted line, North Atlantic Current (NAC). The white-dashed line in the Labrador Sea indicates the region where deep convection happens and where the Labrador Sea Water is formed. In the land areas, the colored (white) shaded regions indicate the different land (ice-covered) regions used in our energy balance (mass balance) model, with their acronyms written in black, counterclockwise: Southwest Greenland (GSW), Central West Greenland (GCW), Northwest Greenland (GNW), Devon Island (DVN), Baffin Island (BFN), Newfoundland and Labrador (NFL). The yellow triangle marks the position of Station 27 by the St. John's harbor, used to validate our salinity and freshwater climatologies. The 1000-m and 2500-m isobaths are represented by thin solid black lines. The map in (b) was generated using the colormap scheme of Thyng et al. (2016).

Yashayaev and Zhang 2025), incorporating surface waters strongly affected by freshwater input and sea ice processes (Frew et al. 2000). Variations in stratification induced by freshwater input can substantially affect deep winter convection and thereby influence subpolar overturning variability (Thornalley et al. 2018; Yang et al. 2016; Yashayaev and Dickson 2008; Boessenkool et al. 2007). Although the direct role of LSW formation in controlling the AMOC has recently been debated (Lozier et al. 2019), subsequent studies argued that Labrador Sea convection remains an important component of AMOC variability (Le Bras 2023; Yashayaev 2024), while also playing a major role in carbon uptake and sequestration in the subpolar North Atlantic. In 2023, Yashayaev (2024) documented a complete shutdown of LSW formation in 2023 associated with an anomalously large Arctic origin freshwater input, likely dominated by enhanced sea ice melt. Yashayaev (2024) further suggested that this event may represent a precursor of a major freshwater release from the Beaufort Gyre, which has accumulated anomalously large freshwater volumes since the early 2000 s (Jahn and Laiho 2020; Haine 2020; Proshutinsky and Coauthors, (2019)).

The upper layer of the BBLS is largely impacted by the strong seasonal cycle of the freshwater input due to sea ice melt, freshwater runoff from the Greenland Ice Sheet and glaciers, iceberg calving, continental runoff, and precipitation (e.g., Reverdin et al. 1997). Increased freshwater input from the Greenland Ice Sheet and glaciers in both western and eastern Greenland constitutes major anomalies in this cycle (Dukhovskoy et al. 2021, 2019; Luo et al. 2016; Gillard et al. 2016). Moreover, the region is also affected by the advection of locally or remotely generated anomalous signals. One of the most remarkable documented freshening events was the Great Salinity Anomaly (Dickson et al. 1988). Initiated in the Greenland Sea in the late 1960 s, it spread across the entire SPNA. There were other upper-layer freshening events of weaker magnitude reported in the literature (Reverdin et al. 1997; Belkin et al. 1998; Dickson et al. 1988; Reverdin 2010; Holliday et al. 2020; Biló et al. 2022; Fox et al. 2022).

The coexistence of a strong seasonal cycle with major locally and remotely driven anomalies requires an accurate description of the seasonal freshwater input and its influence on salinity. If such a climatology of freshwater input existed, both the anomalies of the seasonal cycle and the longer, persistent anomalies such as the Great Salinity Anomaly (GSA) could be better defined and more accurately quantified. Freshwater content is an informative variable commonly employed for quantifying the meltwater amount and fluxes, and several studies provide its estimates in the region (Tesdal and Haine 2020; Carmack et al. 2016; Haine et al. 2015; Serreze et al. 2006; Curry et al. 2011, 2014; Dickson et al. 2007). However, the use of a subjectively defined salinity reference makes freshwater content not entirely appropriate for distinguishing the various freshwater sources or defining a precise seasonal cycle. More generally, it raises conceptual difficulties regarding the quantification of oceanic freshwater variability (Schauer and Losch, 2019). For example, the commonly accepted reference value of 34.8 does not properly reflect the freshwater input in the shelf regions on the western side of the BBLS, where the depth-average salinity is below 33.5. Furthermore, it becomes unclear how such reference should be adjusted in a changing climate, imposing trends on all oceanographic variables, and presumably the related references.

In this work, we introduce an alternative approach, involving an objectively defined variable, the freshwater content change (FCC), to quantify the freshwater input occurring in the BBLS predominantly throughout spring and summer. Unlike the traditional salinity-reference-based expressions of freshwater content, FCC represents the actual amount of freshwater gained or lost over a specified period from various freshwater sources, computed without the use of a reference salinity. Since FCC quantifies the cumulative freshwater input from all active sources, it can naturally serve as a closing term in freshwater balance equations. Furthermore, as FCC can be placed in the context of specific spatial (local-to-regional) and temporal (weekly-to-decadal) constraints, it also allows for the assessment of the impacts of changing

inputs on different spatiotemporal scales.

Using all available salinity observations, we developed an iterative methodology that accounts for spatial and temporal aliasing to construct a weekly climatology of FCC (Section 3), and carefully validated its results (Section 4). The resulting climatology allows us to characterize the seasonal evolution of freshwater accumulation and heat content within this key Arctic–subpolar gateway, and to relate the observed regional contrasts to the combined influences of sea ice melt, glacier runoff, and freshwater pathways. These results are also compared with the climatology of the local freshwater forcing derived by a summation of components from independent sources (Section 5). These independent estimates include observed sea ice volume and sea ice fluxes, modelled glacier runoff and calving, river runoff, and precipitation minus evaporation. Also, to support and enhance the interpretation of the FCC results, a climatology of heat content change (HCC) was estimated following a similar approach from ocean temperature data. Together, the FCC and HCC climatologies provide a framework for extracting anomalous signals from *in situ* observations, and investigating low-frequency Arctic freshwater anomalies and their propagation toward the subpolar North Atlantic.

2. Data

In this section, we describe the oceanographic data used to construct the freshwater and heat content change climatologies, as well as the data used to evaluate the different components of the freshwater budget. These components are the freshwater input contributions by (1) local sea ice melt, (2) glacier melt (hereafter referring collectively to surface melting from the Greenland Ice Sheet and peripheral glaciers), glacial discharge, river runoff, and precipitation minus evaporation (hereafter P–E), and (3) sea ice volume fluxes entering the study domain.

2.1. Salinity and temperature data for the freshwater and heat content change climatology

The freshwater and heat content change climatologies were derived from weekly salinity and temperature fields estimated across Baffin Bay and the shelf and slope regions around the Labrador Basin, bounded by 48°N–78°N, 45°W–80°W (excluding Hudson Bay and the St. Lawrence River estuary), and the 2500 m isobath (Fig. 1). The 2500 m isobath was chosen to focus the analysis on the shelf and slope regions that are most directly influenced by seasonal freshwater input, while limiting the influence of the deep convection regions of the Labrador and Irminger Seas. The climatologies cover April to October, a period that captures the full cycle of freshening and heat increase from the onset of sea ice melt in spring to late summer. Owing to the scarcity of hydrographic observations during the ice-covered season, the present climatologies focus on the freshening phase of the seasonal cycle (April–October).

Historical conductivity-temperature-depth (CTD) data were compiled from the World Ocean Database 2023 (WOD23, Mishonov et al. 2024), supplemented by data from the Marine Environmental Data Section (MEDS) of Canada's Department of Fisheries and Oceans (DFO, 2025a). Data were selected from February to October across the entire study region, with February–March data used to generate initial estimates at the start of the climatological period (Supplementary Text 1). Profiles were depth-averaged over 5–300 m, or 5 m to the bottom in shallower areas. Profiles with insufficient vertical coverage near the surface or bottom were discarded, as were profiles acquired in water depths less than 10 m. Of the 95,258 profiles originally available, 16,746 were discarded for insufficient vertical coverage or shallow water depth, and a further 3551 profiles were flagged as outliers and removed following an initial quality check (Supplementary Text 1), yielding a final dataset of 74,961 salinity and 74,803 temperature profiles spanning from 1950 to 2024. The bi-monthly distribution of the salinity data used for the climatology is shown in Fig. 2.

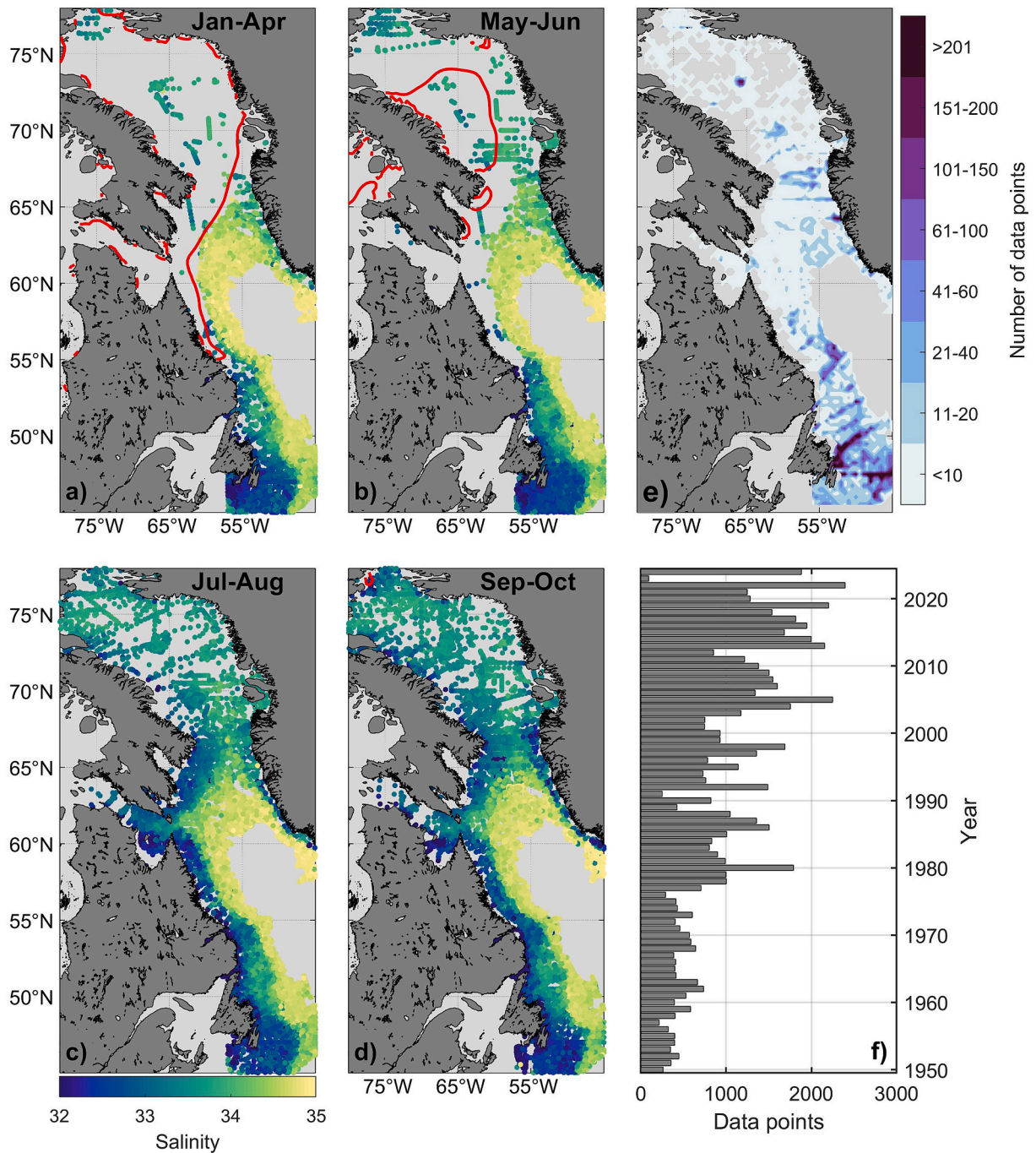


Fig. 2. Distribution of 5–300 m depth-average salinity over the study domain after the initial data preselection for (a) January–April, (b) May–June, (c) July–August, and (d) September–October. (e) Number of data points within 0.5° longitude $\times$ 0.25° latitude boxes. (f) Number of data points per year. The red lines in panels (a–d) represent the 80 % sea-ice concentration contour corresponding to each period.

2.2. Freshwater budget from independent data sources

In this work, we compare the FCC estimates derived from salinity to freshwater inputs quantified from independent data sources. These sources are described in the following subsections.

2.2.1 Sea ice thickness and sea ice melt

The sea ice melt volume was estimated as the difference between the maximum winter and minimum summer ice volume within any given region of interest. The monthly sea ice thickness from 1979 to 2023 was obtained from the Pan-Arctic Ice Ocean Modeling and Assimilation System (PIOMAS; Zhang & Rothrock, 2003) with a 50-km grid spacing.

Prior to calculating ice volume, sea-ice thickness was multiplied by sea-ice concentration derived from passive microwave observations (SMMR, SSM/I, and SSMIS) provided by the National Snow and Ice Data Center (DiGirolamo et al. 2022) at 25-km resolution and interpolated onto the PIOMAS grid. Averaged over 1979 – 2023 and over the full study domain, the resulting sea ice melt contribution totals 1158 km^3 .

The PIOMAS ice volume estimates fall within the range of other available sea ice thickness products. Averaged over 2011–2020 and $63\text{--}78^\circ\text{N}$ — the common domain shared by all products — the maximum winter ice volume equals 865 km^3 from PIOMAS, compared to 629 km^3 from SMOS-CryoSat-2 (Ricker et al. 2017) and 944 km^3 from the Proxy Sea Ice Thickness product (Glissenaar et al. 2023). These estimates

should therefore be interpreted as first-order values, owing both to uncertainties in PIOMAS sea ice thickness estimation (Schweiger et al. 2011) and to the fact that sea ice advection is not explicitly accounted for. Nevertheless, while uncertainties affect the absolute ice volume estimates, the relative seasonal evolution and spatial patterns central to our study are less sensitive to these uncertainties.

2.2.2 NUMERICAL SIMULATIONS AND REANALYSIS DATA FOR THE LAND AND GLACIER RUNOFFS AND P-E

The freshwater budget over land and ice was estimated from the glacier energy-mass balance model GLIMB (Fujita and Ageta 2000; Fujita and Sakai 2014; Kondo and Fujita 2026) for ice-covered terrain, and an energy balance model developed by Fujita and Sakai (2014) for river runoff over ice-free terrain. Both discharge models are forced by ERA5 daily data (0.25° resolution), with air temperature estimated from pressure level atmospheric temperature and geopotential height (Khalzan et al. 2022). While ERA5 exhibits a documented warm bias over Arctic sea ice during winter and early spring (up to 1–2 °C; Batrak and Müller 2019), this bias is substantially reduced during summer — the season most relevant for melt estimates — and the precipitation calibration procedure described below partially compensates for systematic forcing biases. For ice-covered regions, precipitation calibration parameters were adjusted at 0.5° resolution to reproduce the observed long-term glacier mass balance derived from satellite data (Hugonnet and Coauthors, (2021)), following Sakai et al. (2015) and Sakai and Fujita (2017). The models then calculated surface energy and mass balance at 50-m elevation intervals. Additional details regarding the calibration procedure and forcing uncertainties are provided in Supplementary Text 3.

To calculate the freshwater input to Baffin Bay, we determined the contributing drainage basins (Fig. 1) from catchment analysis based on the 500-m resolution ETOPO2022 digital elevation model. Glacier masks were obtained from the Randolph Glacier Inventory version 6 (Consortium RGI, 2017), while Greenland basins were taken from IMBIE (The IMBIE Team, 2020); both datasets were merged to generate the ice masks used in the model. Hypsometry (area-elevation distribution) was then prepared separately for ice-covered and ice-free surfaces at 50-m elevation intervals on the 0.25° calculation grid. Daily freshwater input was integrated from nominal Day 90 to Day 300, consistent with the duration of the FCC climatology (Section 3). The resulting discharge estimates are summarized for the entire study region in Table 1, while basin-scale values are provided in the supplementary materials (Table ST1).

The models do not calculate “ice discharge”, i.e., freshwater input in the form of icebergs, primarily originating from the Greenland ice sheet. Instead, the ice discharge values used in this study (Table 1) were obtained from the dataset of Mankoff et al. (2021), together with the reported 10 % uncertainty estimate (Mankoff et al. 2020). Monthly ERA5

precipitation and evaporation fields were used to estimate precipitation minus evaporation (P-E) over the ocean surface. Because precipitation falling onto sea ice accumulates as snow before entering the ocean upon melting, P-E over sea ice is not directly represented as immediate ocean freshwater input in ERA5. However, since our budget focuses on the April–October period when sea ice progressively retreats and P-E is dominated by open-ocean conditions, the impact of this limitation is expected to be small.

All datasets described in this section cover and are averaged over 1965–2022, except for iceberg discharge (1986–2022). We assume that these climatological periods represent the overall 1950–2024 period covered by the FCC climatology. Uncertainties in Table 1 are derived from ERA5 uncertainty estimates for oceanic P-E, and from ERA5 P-E forcing uncertainties integrated over land for ice melt and river runoff estimates.

To assess the quality of the GLIMB model output, we compared the climatological seasonal cycle of ice melt and land runoff with the largely independent product of Igneczi and Bamber (2025), which routes the RACMO2 regional climate model surface mass balance output to coastal outflow points. The comparison, described in detail in Supplementary Text 3 and shown in Supplementary Fig. S2, reveals broadly similar seasonal cycles and annual totals across the five regions examined, with total runoff (ice melt + land runoff) differing by less than 1 % overall (599 km³ in our product vs. 600 km³ in Igneczi and Bamber, 2025), although small regional differences remain.

2.2.3 ice-volume flux estimates

(i) *Satellite-derived estimates of ice-volume flux through Nares Strait and the Canadian Arctic Archipelago.*

We use a widely established methodology to estimate the ice area and volume flux transported from the western Arctic Ocean to Baffin Bay between October 2016 and May 2024 following previous studies (e.g., Agnew et al. 2008; Howell and Brady 2019; Moore et al. 2023; Howell et al. 2024; Kwok et al. 2010; Komarov et al. 2026). Overlapping SAR image pairs from RADARSAT-2, Sentinel-1, and the RADARSAT Constellation Mission (RCM) were acquired across the width (L) of the Nares Strait (139 km), Jones Sound (48 km), and Lancaster Sound (83 km), and resampled to 200 m spatial resolution. For each SAR image pair, sea ice motion was estimated using the Environment and Climate Change Canada Automated Sea Ice Tracking System (ECCC-ASITS; Howell et al. 2022). Sea ice motion estimates were interpolated to a 30 km buffer region at each gate and then sampled at 5 km intervals. The sea ice area flux (F_A) was calculated using the following equation:

$$F_A = \sum c_i u_i \Delta x \quad (1)$$

where, Δx is the spacing along each gate (i.e., 5 km), u_i is the ice motion normal to the flux gate at the i^{th} location and c_i is the sea ice concentration determined from the Canadian Ice Service ice charts (Tivy et al. 2011). The sea ice area flux values are subsequently summed for each month.

The monthly uncertainty (σ_{FA}) in F_A can be estimated following (Kwok and Rothrock 1999) by assuming errors in sea ice motion are additive, uncorrelated, and normally distributed using the following equation:

$$\sigma_{FA} = \frac{\sigma_u L}{\sqrt{N_s}} \cdot \sqrt{N_D} \quad (2)$$

where σ_u is the error in SAR-derived ice motion (0.43 km; Komarov and Barber 2014); L is the width of the gate; N_s is the number of samples across the gate, and N_D is the number of SAR image pairs per month (~30). Accordingly, the monthly σ_{FA} estimates for Nares Strait, Jones Sound, and Lancaster Sound are 62 km², 40 km², and 48 km², respectively.

The volume flux (F_V) was obtained by multiplying the monthly sea ice area flux by the monthly CryoSat-2 sea ice thickness from Landy et al.

Table 1

Seasonal freshwater budget of the major freshwater sources contributing to freshwater content change (FCC) in Baffin Bay and the Labrador Sea. Uncertainties are shown in brackets. An approximate conservative upper bound for the total uncertainty was obtained by linear summation of the individual component uncertainties, excluding the Fram Strait ice-flux contribution, whose reported spread represents interannual variability rather than a formal uncertainty estimate.

Components	Volume (km ³)	%
Sea Ice	1158 (±116) ¹	34
Arctic Ice flux (CAA)	103 (±6) ²	3
Fram Strait Ice flux (1/3rd)	650 (±170) ³	19
P-E Ocean	448 (±26)	13
River runoff (P-E Land)	520 (±51) ⁴	15
Glacier melt	436 (±48)	13
Ice discharge (Iceberg calving)	116 (±17) ⁵	3
Total	3431 (±264)	100

(2022). The uncertainty (σ_{FV}) in volume flux can be estimated following Kwok and Rothrock (1999) using the following equation:

$$\sigma_{FV} = \sqrt{(F_A \sigma_h)^2 + (h \sigma_{FA})^2} \quad (3)$$

where h is the ice thickness and σ_h is the uncertainty in thickness is taken from Landy and Coauthors, (2022). The resulting monthly σ_{FV} estimates for Nares Strait, Jones Sound, and Lancaster Sound are 3 km³, 1 km³, and 2 km³ respectively.

For the freshwater budget, we integrated the near-daily volume flux values between March and October, assuming an average one-month transit time between the gates and the study region. The March–October ice volume flux integrated across the three gates and averaged over the eight-year period totals 103 km³, approximately half of the annual transport.

(ii) Estimates of ice-volume flux through Fram Strait.

Estimating the ice volume flux through Fram Strait is necessary because sea ice meltwater advected along East Greenland likely contributes to the freshwater input in our study region. The Fram Strait ice export exhibits a strong seasonality, with maximum values between December and March (Spreen et al. 2020; Sumata et al. 2022). Sea ice drifts southward at a speed of around 8 km day⁻¹ between 80°N and 70°N (Martin & Wadhams, 1999), and most of it melts before reaching this latitude. South of 70°N, the East Greenland Coastal Current and East Greenland Current (Fig. 1b) transport the resulting meltwater southward at velocities of ~0.5 m s⁻¹ (Sutherland & Pickart, 2008). Based on these transport speeds, a freshwater signal originating from the Fram Strait should reach the region south of Greenland (~2500 km downstream) in approximately six months. To estimate this contribution to the freshwater budget, we integrated the Fram Strait ice volume flux from November to May, using the 1992–2014 monthly mooring-based estimates of Spreen et al. (2020), yielding a total of 1951 km³. This volume is much larger than the ~50 km³ cumulative decrease in liquid freshwater transport estimated from De Steur et al. (2018).

However, only a fraction of this ice-derived freshwater is expected to reach our study region, as part of the signal is redistributed during its southward transport, through recirculation within the Greenland and Iceland Seas or advection toward the central Labrador Sea (Dukhovskoy et al. 2016; Foukal et al. 2020). Additional freshwater loss likely occurs through shelf-basin exchange near Cape Farewell, where the coastal current shifts offshore and interacts with the shelfbreak flow (Lin et al. 2018; Karpouzoglou et al. 2023). These processes contribute to the uncertainty in estimating the fraction of the Fram Strait ice export ultimately reaching our domain.

3. Methods

3.1. Construction of an unbiased climatology

The objective of this study is to construct a climatology with a weekly resolution describing the seasonal cycle of freshwater and heat content changes across the study region. In this context, the climatology represents the regular seasonal cycle, from which interannual anomalies can be subsequently separated into year-specific (irregular seasonal) and longer-term (interannually persistent) components. The climatologies presented here do not include the November–March period, during which freshwater content generally decreases due to processes such as sea ice formation, brine rejection, and vertical mixing. The scarcity of hydrographic observations across most of the study domain during winter, resulting from sea ice cover and challenging observational conditions, motivated our focus on the freshening phase of the seasonal cycle.

In climatological studies, the term “climatology” is often used interchangeably with a simple average of available observations. However, the World Meteorological Organization (WMO) defines a climatological normal as the average over a sufficiently long period, typically

30 years (WMO, 2017). In the ocean, variability on multi-decadal timescales is significant, and we adopt here a minimum averaging period of 50 years. More importantly, in the context of observational oceanography, a climatology should account for the effects of the strong spatiotemporal inhomogeneity of the in-situ measurements. In this study, we therefore distinguish between a “mean field”, defined as a direct average of observations within spatial bins, and a “climatology”, defined as the long-term seasonal cycle after correction or compensation for spatiotemporal sampling biases.

Such biases arise from inadequate sampling, which in this study refers to the uneven distribution of observations in both space and time. Sampling density varies within a season (e.g., ice-covered versus ice-free conditions) and across decades (e.g., evolving sampling strategies and observing systems). In addition, horizontal gradients and low-frequency variability can introduce systematic deviations when data acquired at different locations and in different seasons, years, or decades are combined. As a result, direct averaging of all available data may not provide an accurate estimate of the mean seasonal cycle.

To address these issues, we adopt an iterative approach based on the methodology of Yashayaev and Zveryaev (2001), in which the different sources of variability are estimated, removed, and reevaluated sequentially through a course of iterations until convergence is achieved. At each iteration, four components are evaluated:

- the background spatial variability (spatial bias),
- the weekly mean seasonal cycle,
- outliers, and
- low-frequency variability (temporal bias).

After each iteration, the estimated biases and outliers are removed from the data, except the one that will be reevaluated at the next step. This procedure is repeated until convergence of the corrected values. The components are treated in decreasing order of magnitude, from large-scale spatial variability to lower-amplitude temporal variability.

The general structure of the method is illustrated in Fig. 3. A concise description of each step is provided below, while the full implementation details are given in Supplementary Text 1. The same procedure is applied to both salinity and temperature to derive freshwater and heat content change climatologies, respectively. As the method relies on spatial interpolation and correlation scales, the mapping approach used in this study is first described in the next section.

In this study, the analysis is performed separately for three depth ranges (0–50 m, 50–150 m, and 150–300 m) in order to better resolve the vertical structure of freshwater and heat content changes. The methodology described below is applied independently to each depth layer. For clarity, the 0–50 m layer is used as an illustrative example throughout this section, unless otherwise specified. The full 0–300 m freshwater content change is then obtained by combining the contributions from the three layers as described in Section 3.8. The estimation of uncertainties associated with the FCC climatology is described in Supplementary Text 2, with the resulting uncertainty maps shown in Supplementary Fig. S2.

3.2. Use of a distance and topography-constrained mapping and its decorrelation scales

The climatology is estimated at selected points of a 0.25° latitude × 0.5° longitude grid at weekly resolution across the study domain. To maximise the use of the limited observational dataset, we take advantage of the strong spatiotemporal correlation of hydrographic properties over shelves and slopes, as expected in weakly stratified polar and subpolar regions. We use the weighted-mean scheme of Mensah and Ohshima, 2023, hereafter referred to as the Shrinking-Stretching Integrated Constraint (SSCI) mapping. This approach includes both an along-isobath distance constraint and a relative depth-difference constraint. The latter represents a dynamical topographic constraint reflecting the ability of the water column to shrink or stretch to satisfy potential vorticity conservation, a key property of weakly stratified

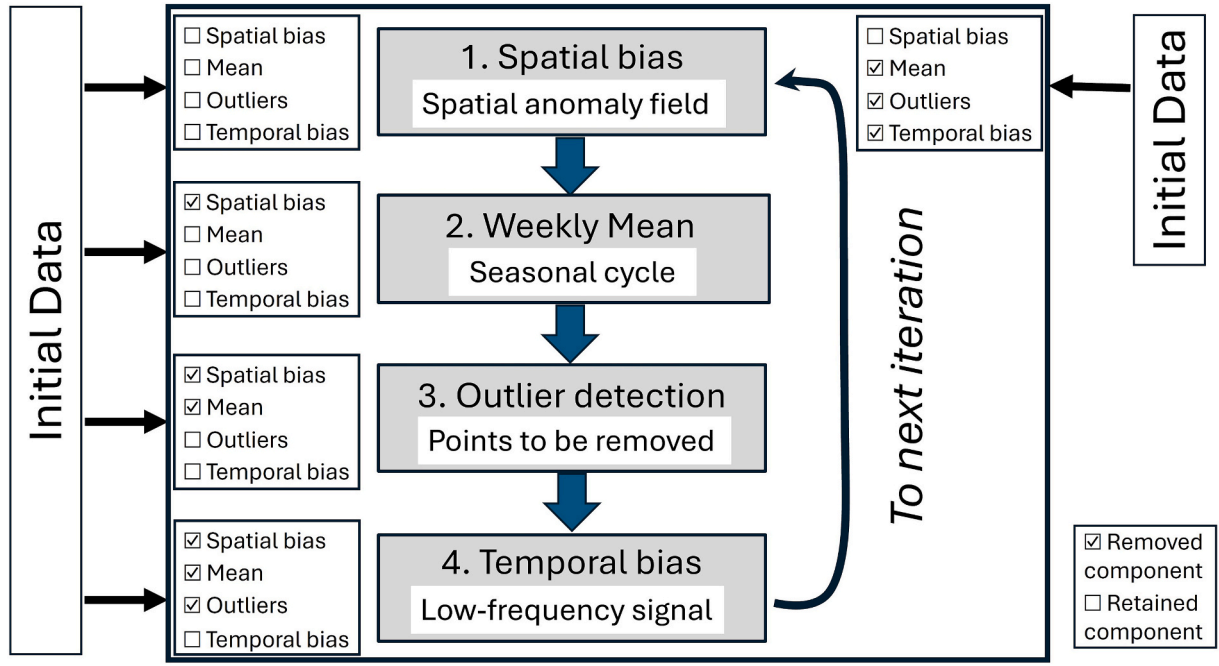


Fig. 3. Schematic of the procedure to generate a weekly climatology representing the unbiased seasonal cycle used in this study. This figure is adapted and modified from Yashayaev and Zvereyev 2001.

flows. Assuming that observations remain correlated over timescales shorter than one month, a temporal constraint is included as follows:

$$W(r, \phi, t) = \exp \left[-\left(\frac{r}{L_r}\right)^2 - \left(\frac{\Phi}{L_\phi}\right)^2 - \left(\frac{t}{L_t}\right)^2 \right] \quad (4)$$

where r is the distance between the grid point and the observation, t is the time difference (days) between the observation date and the center of the weekly bin (D90, D97, ..., D304), and $\Phi = \Delta h/h$ represents the relative depth difference, where h is the bottom depth at the grid point and Δh is the bottom depth difference between the observation and grid points. L_r , L_ϕ , and L_t are the decorrelation scales. Following Mensah and Ohshima, 2023, we estimated $L_r = 158$ km and $L_\phi = 0.51$, while L_t was set to 30 days. The temporal term is included when estimating weekly values, but omitted when the mapping is used solely for spatial interpolation or for constructing low-frequency fields (Sections 3.5 and 3.7).

The SSCI mapping is used both to estimate weekly salinity values at selected locations (Section 3.5) and to interpolate the resulting fields onto the full 0.25° latitude $\times$ 0.5° longitude grid.

3.3. Selection of analysis locations and spatial interpolation

To ensure that the climatology is estimated only where the observations provide sufficient support, we restrict the analysis to a subset of grid locations that satisfy criteria for hydrographic homogeneity and data availability, and subsequently interpolate the results onto the full grid.

Grid points associated with strong horizontal gradients are first excluded, based on the background salinity field (Section 3.4). Locations where the horizontal salinity gradient exceeds 0.01 km^{-1} are removed, as they correspond to persistent frontal regions where local averaging is less appropriate. Among the remaining grid points, we retain only those with sufficient observational coverage to resolve the seasonal cycle. Within the core freshening period (days 150–270), we require that at least nine weekly bins contain a minimum of 50 observations within one spatial decorrelation scale (L_r) and topographic constraint (L_ϕ), and that at least 15 distinct years are represented within the same limits. To improve coverage in shallow coastal regions, a secondary selection is

performed using relaxed spatial limits ($1.5 L_r$, $1.5 L_\phi$), retaining a limited number of additional locations.

The resulting candidate locations are then subsampled based on the horizontal decorrelation scale to reduce redundancy. Locations are retained at an approximate spacing of 1.25° in latitude throughout the domain, with a longitude spacing of 1.5° south of 66°N and 2° north of 66°N . In regions of strong topographic gradients, additional locations are retained to better resolve along-slope variability. This procedure yields a set of 272 analysis locations across the domain. The distribution of the selected analysis locations is shown in Supplementary Fig. S1. The weekly climatology is estimated at each selected location following Sections 3.4–3.6, and the resulting fields are interpolated onto the full grid using the spatial component of the SSCI mapping scheme (Section 3.2). To limit extrapolation, interpolated values are computed only where at least one selected location lies within $0.5 L_r$ of the target grid point.

3.4. Background spatial gradient

The background spatial gradient (referred to here as the spatial bias) is estimated on a 10×10 km grid as the difference between the local depth-averaged salinity $\bar{S}_a$ and a reference salinity S_{ref} (see Supplementary Text 1 for details):

$$S_{bias} = \bar{S}_a - S_{ref} \quad (5)$$

The reference salinity S_{ref} is determined as an isohaline of the 0–300 m depth-averaged salinity field present throughout most of the study domain; here, $S_{ref} = 33.5$. This single reference is used for all three vertical layers, and the resulting freshwater content changes are not sensitive to its exact value. This field is also used to identify regions of strong horizontal gradients, which are excluded from the analysis in Section 3.3. The resulting spatial bias field is shown in Fig. 4c.

3.5. Estimation of the weekly mean and seasonal cycle

This step consists of averaging individual hydrographic profile data within prescribed spatial and temporal limits. The input data are depth-

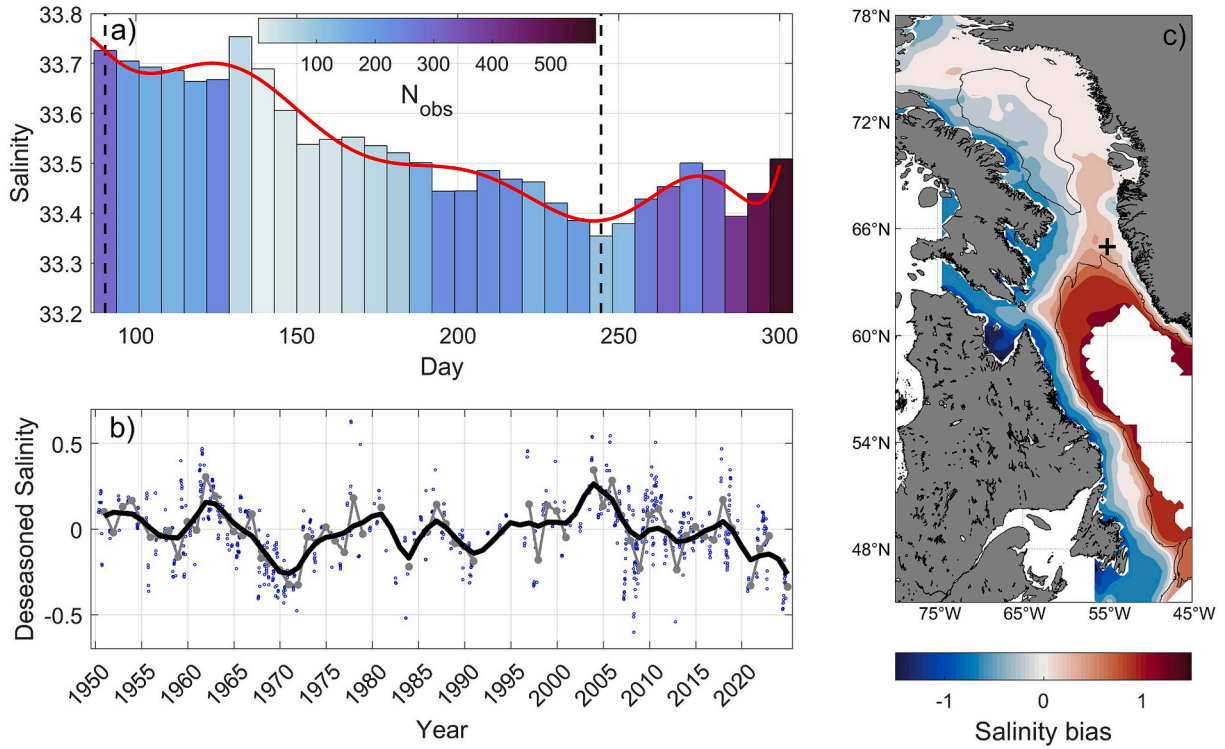


Fig. 4. Illustration of the iterative process at the third iteration for the 50–150 m layer. (a) Weekly mean salinity at a grid point located off western Greenland at 65°N, 55°W, bottom depth 620 m. Colors indicate the number of observations located within $2 L_r$ (308 km) and $2 L_\phi$ (1.02, corresponding here to a bottom-depth range of 0–1252 m) contributing to each weekly estimate. (b) Deseasoned weekly (dots) and yearly (gray line) salinity anomalies between 1950 and 2024. The thick black line represents the 5-year moving-average used to estimate the low-frequency variability (temporal bias). (c) Spatial salinity bias field relative to a reference salinity of 33.5, shown only to illustrate the background spatial gradient and not affecting the final FCC estimates. The black cross marks the grid point shown in (a) and (b), and the thin black contour the 1000 m isobath. Weekly values in (a) were estimated after removing the temporal and spatial biases in (b) and (c) from individual salinity observations according to their year and location.

average salinity profiles corrected for the spatial bias (Section 3.4). At each selected location, salinity values are estimated by averaging observations within the spatial decorrelation scales L_r and L_ϕ , and within a temporal window of ± 14 days centered on predefined weekly bins between year days 90 and 300 (D90–D300, 1 April to 27 October).

The whole seasonal cycle of salinity S_{clim} can then be approximated by fitting a smooth polynomial function. A 9th-order polynomial is used to represent the seasonal cycle while smoothing high-frequency variability. This approximated salinity $S_{clim}(w)$ is the climatology that is used as a reference for all further calculations. The climatology FCC_{clim} is estimated from S_{clim} as follows:

$$FCC_{clim}(w + 0.5) = \frac{H}{S_{clim}} (S_{clim}(w + 1) - S_{clim}(w)) \quad (6)$$

where H is the thickness of the layer (or the local bottom depth if shallower) and $S_{clim}(w)$ is the weekly climatological salinity and $\overline{S_{clim}}$ denotes the temporal mean of the climatological salinity over the freshening season. The beginning (D_{start}) and end (D_{end}) days of the freshening season are defined as the days of the maximum and minimum S_{clim} , respectively (Fig. 4a). Following Eq. (6), the FCC for the whole freshening season is obtained by integrating the FCC in time between D_{start} and D_{end} . S_{clim} and FCC_{clim} constitute the regular cycle of salinity and FCC, respectively.

3.6. Outlier detection

Salinity data are first selected within one decorrelation scale L_r and L_ϕ from the analysis locations. The weekly climatological mean and spatial bias are then subtracted from each data point. Data falling outside the range defined by the mean ± 5 median absolute deviations

are excluded. This procedure is applied at all selected grids.

3.7. Low-frequency variability estimation

At each selected location, a weekly salinity time series $S_{year}(w, y)$ is constructed over the full 1950–2024 period. The annual anomaly is then defined as the yearly-averaged difference between the weekly salinity values and the climatological seasonal cycle:

$$S_{an}(y) = \overline{S_{year}(w, y)} - \overline{S_{clim}(w)} \quad (7)$$

The anomaly includes both the year-specific seasonal cycle and longer-term signals. The latter are isolated as a low-frequency component $S_{LF}(y)$, obtained by applying a 5-year triangular moving average of S_{an} (Fig. 4b). To obtain a spatially coherent representation, S_{LF} is mapped annually using the spatial component of the SSCI mapping (Eq. (4), without the temporal constraint). The resulting field defines the temporal bias $S_{biasT}(y)$, which is removed at each iteration.

3.8. Layer combination and definition of the seasonal freshwater input

As stated in Section 3.1, the climatological analysis is performed independently for the 0–50 m, 50–150 m, and 150–300 m layers. This choice is motivated by the vertical structure of FCC, with preliminary tests revealing systematic breaks in the cumulative FCC profile near 50 m and 150 m depth. Approximately 60 % of the total FCC occurs within the 0–50 m layer, ~80 % within the 0–150 m layer, and nearly all within the 0–300 m layer. Resolving these layers separately also allows for a more physically meaningful definition of the freshening season and total seasonal freshwater input. Note that, because many profiles do not include measurements within the upper 5 m, the upper-layer average

was calculated from 5 to 50 m and treated as representative of the 0–50 m layer.

The freshening season start date D_{start} is defined as the earliest onset among the three layers at each grid point, while the end date D_{end} corresponds to the latest end date. This approach captures the full duration of freshening, including cases where the signal develops asynchronously between layers.

The total seasonal FCC (Section 5.1) at each grid point is then estimated by integrating the FCC contributions from the three layers between D_{start} and D_{end} . Only positive FCC contributions are included at each time step, preventing freshwater gain in one layer from being artificially compensated by salinification in another due to subsurface water-mass variability or vertical salt redistribution. The resulting total

seasonal FCC (Fig. 8a) therefore represents the cumulative positive freshwater input throughout the water column and constitutes the primary metric used in the freshwater budget analysis (Section 5.6).

4. Validation

4.1. Number of iterations

An accurate climatology should minimize the residual between individual salinity observations and the climatological estimate at the corresponding time and location of acquisition, with both spatial and temporal biases being corrected. To evaluate this, we calculate the residual salinity at each iteration as follows:

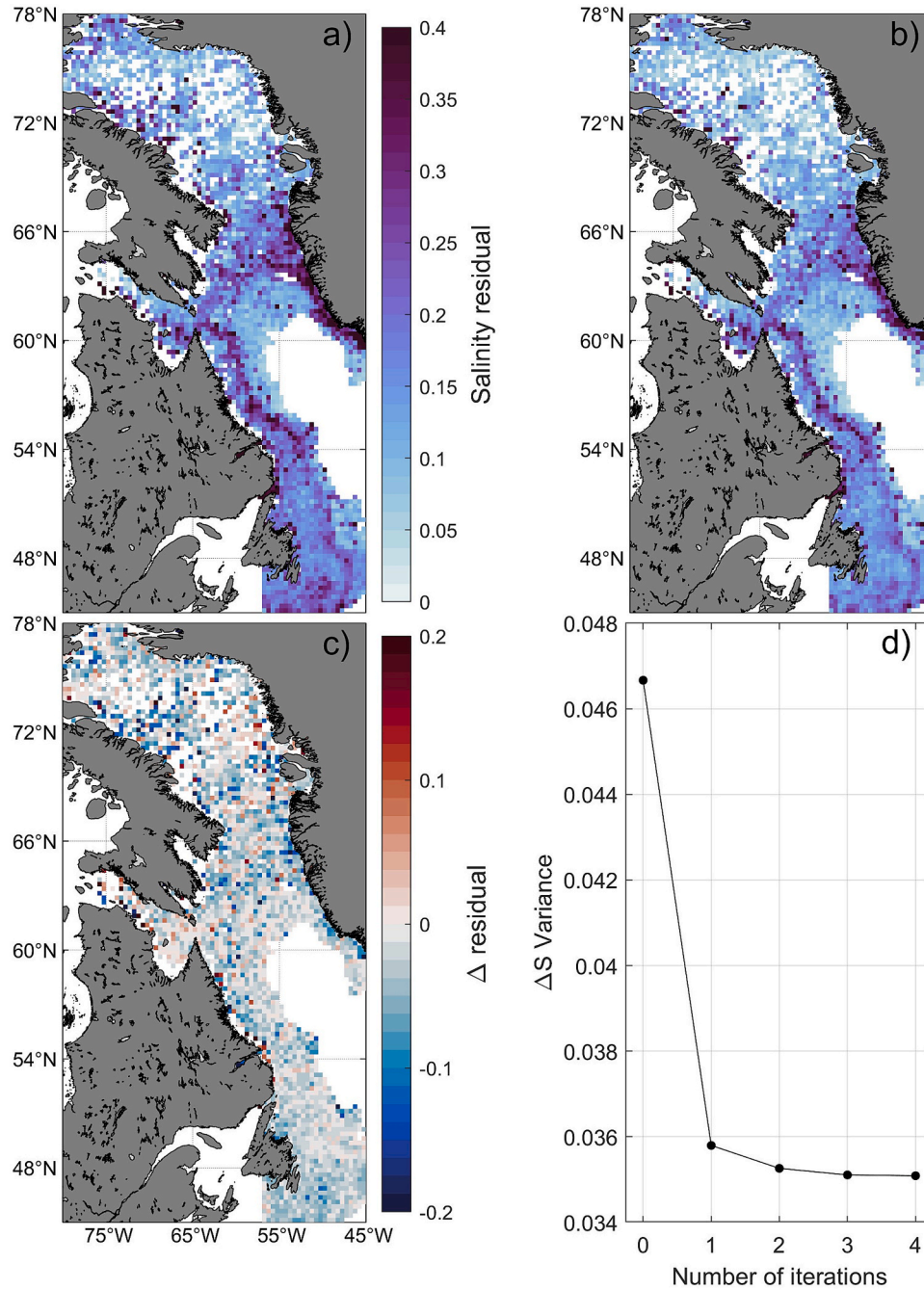


Fig. 5. Three-layer integrated salinity residuals (raw salinity minus climatology and biases) at the (a) first, and (b) third iteration. (c) Difference between the residuals shown in (b) and (a). All values were spatially averaged within 0.5° longitude $\times$ 0.25° latitude grid cells. (d) Variance of the salinity residuals as a function of iteration number.

$$S_{res}(i, n) = S_{raw}(i) - S_{clim}(i, n) - S_{biasS}(i, n) - S_{biasT}(i, n) \quad (8)$$

where n is the iteration number, and i indexes each individual data point S_{raw} or its interpolated location for the gridded variables S_{clim} , S_{biasS} , and S_{biasT} . For each observation, the climatology S_{clim} and the temporal bias S_{biasT} were extracted at the nominal day and actual date of acquisition of the raw data, respectively. The distribution of the absolute value of the residual salinity at the first guess (iteration 0) and the third iteration are shown in Fig. 5a-b, and the difference in these residual salinities is shown in Fig. 5c.

The accuracy of the climatology over the full domain at each n

iteration was then evaluated as the variance of the residual $\sigma_{res}^2(n)$, and the iteration process was repeated until the difference between $\sigma_{res}^2(n)$ and $\sigma_{res}^2(n-1)$ became negligible. We found that σ_{res}^2 stabilizes after three iterations, as illustrated in Fig. 5d. Through this process, a total of 717 points out of 74,961 were excluded at the 0–50 m layer, with the number of outliers at each iteration decreasing from 469 at the first iteration to 28 at the third iteration. The difference between the residual of the third iteration and the first guess (Fig. 5c) further shows that the iteration process leads to the reduction of the residual salinity throughout the whole domain.

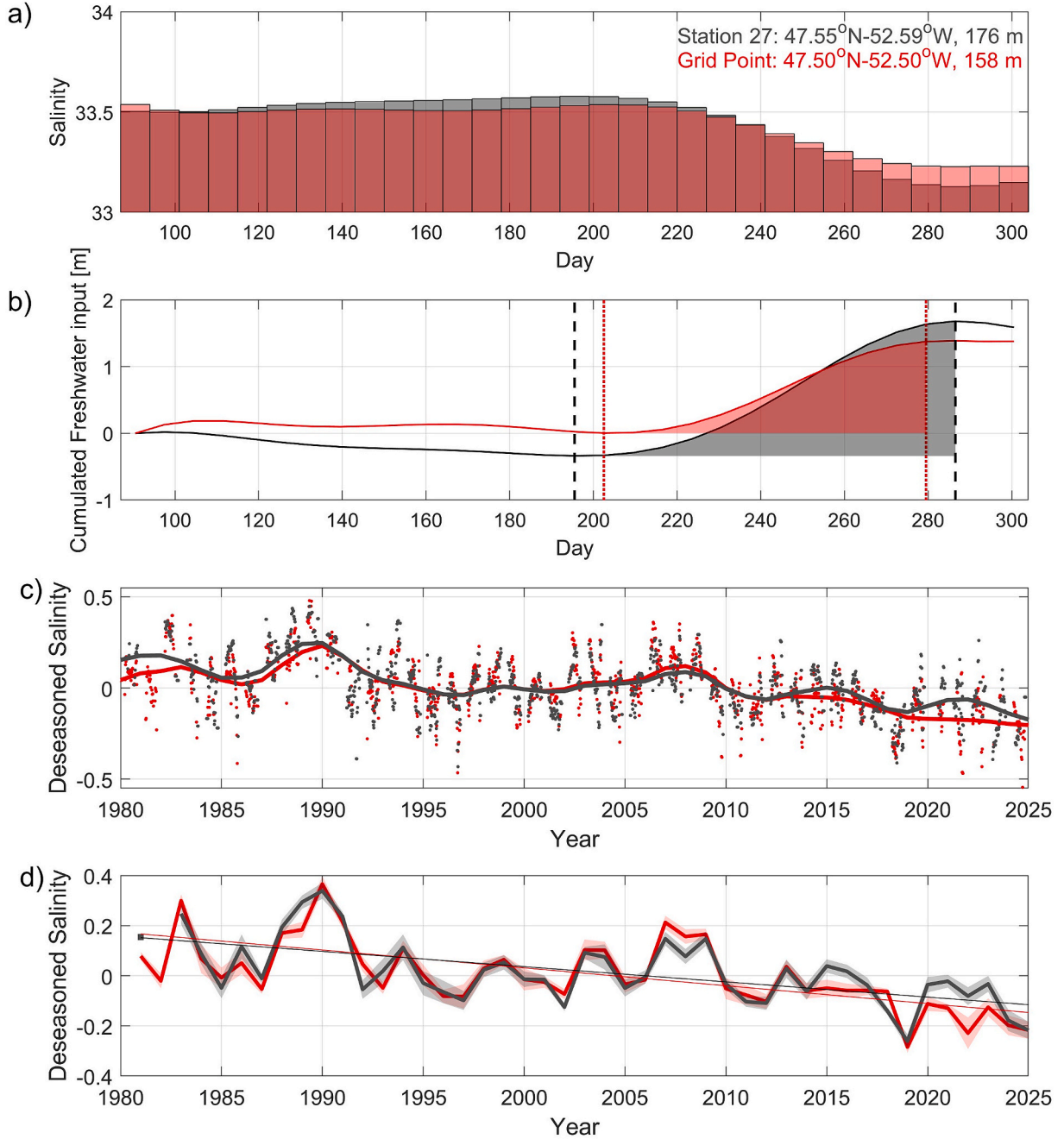


Fig. 6. (a) Salinity climatology at Station 27, estimated from observations acquired at the station only (gray bars), and at the closest grid point with comparable bottom depth estimated from all data available within 1.5 decorrelation length scales, but using only half of the Station 27 profiles (Experiment 2, red bars). (b) Cumulative freshwater input derived from the salinity climatology at Station 27 (gray shading) and at the corresponding grid point (red shading). Gray and red dashed lines indicate the beginning and end of the freshening season, respectively. (c) Weekly binned deseasoned salinity anomalies (dots) between 1950 and 2024 and their 5-year moving average (thick lines) at Station 27 (gray) and the corresponding grid point (red). (d) Yearly averaged deseasoned salinity anomalies (thick lines) with ± 2 standard error envelopes (shading). The 1980–2024 trends are represented as thin black and red lines.

4.2. Comparison with salinity data from a monitoring station off St. John's harbor, Newfoundland

To complement the validation, we estimated a climatology of salinity and freshwater input using a salinity dataset obtained from DFO's MEDS (Section 2.1), consisting of 2210 salinity profiles acquired at Station 27, located 7 km off St. John's harbor, Newfoundland (47.55°N, 52.59°W, 176 m depth), between 1980 and 2024 (DFO, 2025b). This climatology is hereafter referred to as the validation climatology. The data used in the validation dataset and the gridded climatology are only partially independent. Of the 2210 validation profiles, 874 are not included in the dataset used to construct the gridded climatology. Conversely, the climatological values at the grid point closest to Station 27 were estimated from 10,566 profiles, of which only 1336 (13 %) were acquired at Station 27. Overall, approximately 40 % of the validation profiles and 87 % of the profiles contributing to the gridded climatology are independent. The validation dataset benefits from frequent, fixed-location observations, with one profile every ~ 8 days on average between year days 90 and 300. This allows the estimation of a climatology without spatial interpolation or bias correction, and provides a robust reference for evaluating the mapping technique and the spatial and temporal bias correction procedures.

To assess the sensitivity of the gridded climatology to local data availability, we performed two additional experiments in which the number of observations near Station 27 was artificially reduced. In Experiment 2, the number of nearby profiles was reduced from 1336 to 892, corresponding to an average temporal resolution of ~ 16 days. In Experiment 3, the number was further reduced to 670 profiles, corresponding to ~ 24 days. In all cases, the number of observations outside the immediate vicinity of Station 27 was kept unchanged, and for Experiments 2 and 3, the gridded climatology was estimated only from observations acquired between 1980 and 2024.

The resulting climatologies are shown in Fig. 6 for Experiment 2, which is representative of all experiments. The three experiments yield very similar results in terms of the timing and shape of the seasonal cycle, as well as the overall interannual variability. The timing of the freshening season is consistent across all experiments, with differences of less than one week. The interannual variability is also well reproduced, with a root-mean-square difference of ~ 0.055 between the validation climatology and the degraded experiments. In particular, the smoothed low-frequency variability exhibits similar temporal evolution across all datasets (Fig. 6c–d), supporting the robustness of the mapping approach even under substantially reduced local sampling conditions.

At the same time, several systematic differences exist between the validation and gridded climatologies. The amplitude of the seasonal freshwater content change (FCC) is consistently lower in the gridded climatology (1.35–1.41 m) than in the validation climatology (2.02 m), likely reflecting the distinct spatial representativeness of the two datasets. The validation climatology captures highly local variability at Station 27, influenced by processes such as river discharge and local sea ice melt, whereas the gridded climatology integrates variability over a broader surrounding region extending further offshore.

The long-term salinity trends also differ slightly between the two datasets, with the gridded climatology exhibiting a somewhat stronger freshening trend over 1980–2024 (~ -0.07 per decade) than the Station 27 validation series (~ -0.06 per decade). This difference becomes substantially more pronounced after ~ 2007 , when the gridded climatology freshens at a rate of ~ -0.17 per decade compared to ~ -0.11 per decade for the local Station 27 record. Together with the reduced seasonal FCC amplitude in the gridded product, this behavior suggests that part of the observed variability may reflect larger-scale freshwater redistribution processes that are not fully captured by the local validation dataset alone.

Near Station 27, both the independent validation climatology and the gridded climatology show an apparent onset of the depth-integrated freshening season in late June, whereas the corresponding 5–50 m

climatologies (not shown) indicate a much earlier onset in early April. This agreement demonstrates that the contrasting seasonal behaviour between the depth-integrated and near-surface layers is consistently captured by both datasets. This reflects compensating seasonal signals between the surface and subsurface layers, the physical interpretation of which is discussed further in Section 5.2.

5. Climatologies of seasonal freshwater and heat content changes

5.1. Freshwater and heat input during the freshening season

The distributions of the freshening season onset and duration are displayed in Fig. 7a–b, while the cumulated FCC and HCC throughout the freshening season are presented in Fig. 8a–b. The cumulative seasonal FCC totals 3619 km^3 across the study domain. The FCC and HCC maps reveal two broad regimes of freshwater and heat input, which are discussed in detail in the following sections.

The first regime is characterized by positive FCC associated with negative or near-zero HCC, and encompasses most of Baffin Bay and the shelf region between $\sim 60^\circ\text{N}$ and $\sim 65^\circ\text{N}$ along Davis Strait and the northern Labrador Shelf. Within this regime, however, the physical drivers differ markedly between sub-regions, as revealed by the difference between the freshening season onset and the sea ice melt onset (Δday , Fig. 9c). In central Baffin Bay between 71°N and 75°N , the freshening season begins within ~ 30 days of the local sea ice melt onset, suggesting that local sea ice melt is the primary driver of freshening there (Section 5.3). In contrast, the region between $\sim 60^\circ\text{N}$ and $\sim 65^\circ\text{N}$ exhibits a Δday of 60–120 days, indicating that freshening begins long after local sea ice melt. The vertical structure analysis presented in Section 5.2 instead suggests that freshening there is primarily driven by the advection of cold, fresh Arctic-origin waters from central Baffin Bay.

The second regime encompasses the West Greenland shelf and the southern Labrador shelf and slope, where large FCC gains ($>3 \text{ m}$, Fig. 8a) coincide with strongly positive HCC (Fig. 8b). This combination — freshwater input concurrent with heat gain — is characteristic of regions where freshwater arrives already warmed, either through the advection of melted sea ice from remote sources such as the Fram Strait (Section 5.4) or through contributions from precipitation and river runoff. The strongly positive HCC along the southern Labrador slope additionally reflects the influence of warm, saline Irminger Water at depth, as discussed in Section 5.2.

In most regions, the freshening season lasts four to six months (Fig. 7b). The bi-weekly FCC and HCC distributions (Figs. 12 and 13) further reveal additional regional processes, including the outflow of freshwater from Hudson Bay (Section 6) and possible glacier melt influence (Section 5.5).

5.2. Vertical structure of the seasonal freshening signal

The three-layer decomposition of the FCC and HCC climatologies (Figs. 10–11) reveals that the seasonal freshening is not vertically uniform across the study domain. We highlight here the main features of this vertical structure, focusing on regions where vertical contrasts are most pronounced and physically informative. A comprehensive analysis of these three-dimensional signals is beyond the scope of this study and will be addressed in future work.

A robust and spatially coherent feature is that the onset of the freshening season is progressively delayed with depth (Fig. 10a–c). The 0–50 m layer typically freshens first, generally between April and early May, while the 50–150 m and 150–300 m layers follow with delays ranging from several weeks to a few months depending on the region. This vertical lag is consistent with the progressive redistribution of surface-driven freshening into deeper layers through vertical mixing and diffusion, superimposed on the slower evolution of the deeper layers resulting from advective and other transport processes. The duration of

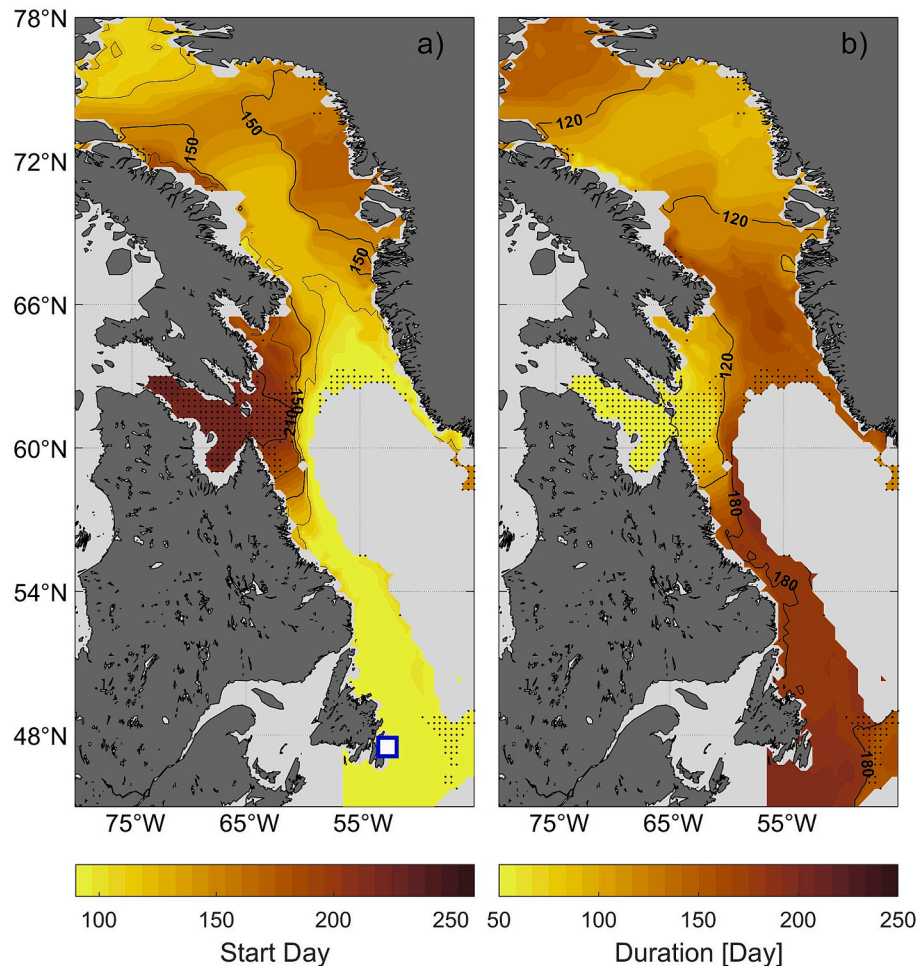


Fig. 7. Distribution of the freshening season (a) starting day, and (b) duration. In (a) the contour lines have an interval of 30 days, with thicker lines at 60-day intervals. The white square in (a) marks the position of Station 27. Black dots indicate locations where the estimated uncertainty of the cumulative seasonal FCC exceeds 50 % of the FCC estimate.

the freshening season (Fig. 10d–f) generally decreases with depth, reflecting both the delayed onset and the shorter persistence of freshening signals at depth.

The strongest vertical contrast is observed on the southern Labrador Shelf south of $\sim 58^\circ\text{N}$. Here, the 0–50 m layer freshens as early as day ~ 90 –100 and exhibits a long freshening season of ~ 160 –200 days (Fig. 10a, 10d), whereas the 150–300 m layer does not begin freshening until day ~ 220 –260 and has a duration of only ~ 40 –60 days (Fig. 10c, 10f). This large vertical separation—approaching five months—indicates strongly surface-intensified freshening, likely driven by sea ice melt and river runoff, while deeper layers remain under the influence of warmer and more saline waters. Consistently, Fig. 11c shows negative FCC south of $\sim 52^\circ\text{N}$ in the 150–300 m layer, while Fig. 11f shows positive HCC over the same region, reflecting the presence of warm, saline Irminger Water along the southern Labrador slope. Near Davis Strait, the negative FCC signal weakens toward the shelf while HCC remains positive, although with reduced amplitude in shallower regions, suggesting progressive mixing between Irminger Water and colder, fresher Baffin Bay waters toward the shelf.

In northern Baffin Bay north of $\sim 75^\circ\text{N}$, near the exit of Nares Strait, the 150–300 m layer also shows a delayed onset of freshening ($\sim$ day 180–200; Fig. 10c) and a relatively short duration (~ 60 –80 days; Fig. 10f). This is accompanied by negative FCC and weakly positive HCC (Fig. 11c, 11f). The coexistence of salinification and warming at depth, in a region where the upper layer freshens early and strongly, is consistent with the southward advection of Canada Basin Atlantic Water

through Nares Strait (Münchow et al. 2016; Barut et al. 2025), which carries relatively warm and saline subsurface waters.

In central Baffin Bay, the 150–300 m layer exhibits negative HCC concurrent with positive FCC (Fig. 11c, 11f), a combination that warrants further consideration. Two mechanisms may explain this pattern. A seasonal intensification of the cyclonic gyre (Shan et al. 2024) would induce isopycnal doming, raising warm and saline Atlantic-origin waters from below ~ 200 m into the 150–300 m layer (Huang et al. 2024). This process would be expected to produce positive HCC and negative FCC, opposite to the observed pattern. Conversely, the inflow of colder, fresher Arctic waters through Lancaster Sound peaks in summer, with volume transport reaching ~ 1 Sv (Prinsenbergh and Hamilton, 2005; Peterson et al. 2012), introducing water masses distinct from those of Atlantic-origin (Münchow et al. 2015). The observed combination of negative HCC and positive FCC is consistent with this Arctic inflow signal, suggesting that it may dominate over isopycnal doming during the summer freshening season. The southward extension of the negative HCC signal along western Davis Strait (Fig. 11f) further supports downstream advection toward the Labrador Shelf.

5.3. Importance of local sea ice melt in the freshening

With a maximum winter volume of 1158 km^3 estimated from PIOMAS, local sea ice within the study region (Fig. 9a) represents a major freshwater source. In addition, approximately 103 km^3 of sea ice is transported into Baffin Bay between March and October (Section 2.2.3).

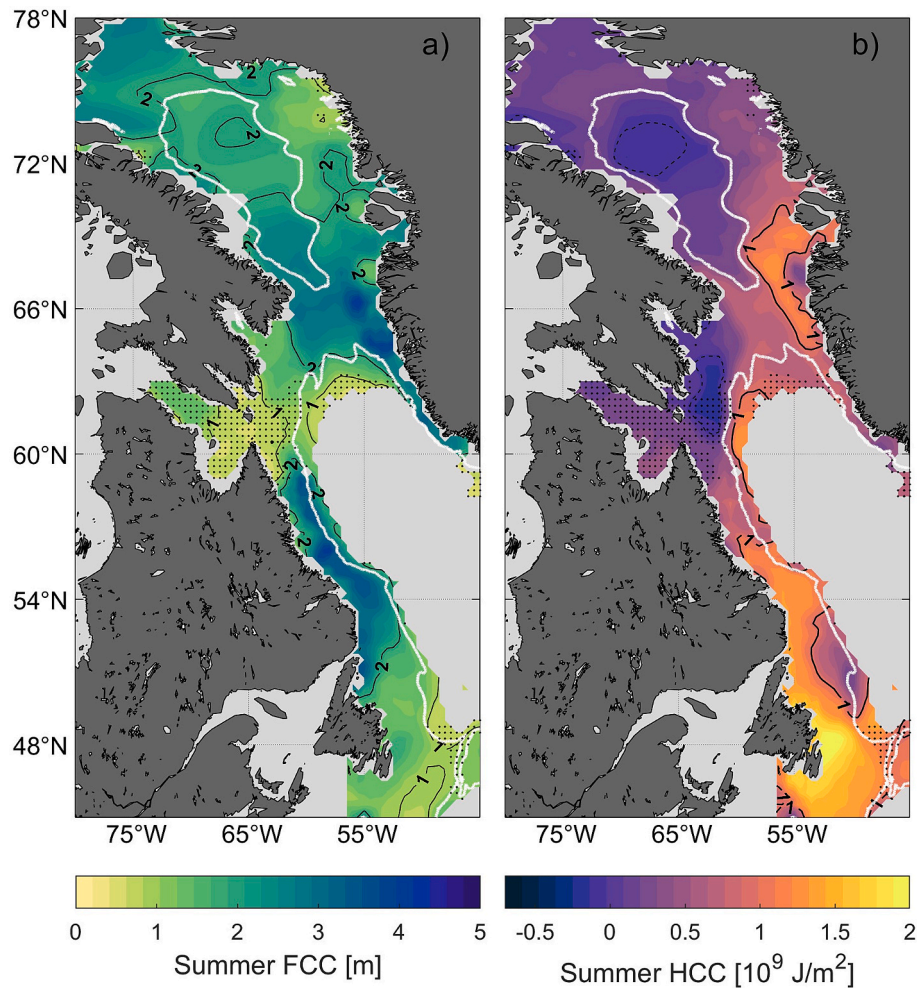


Fig. 8. Distribution of the (a) FCC and (b) HCC integrated during the freshening season (from the freshening season starting day in Fig. 7a, for the duration indicated in Fig. 7b). The white contour line represents the 1000 m isobath. Black dots indicate locations where the estimated uncertainty of the cumulative seasonal FCC exceeds 50 % of the FCC estimate; these locations should be interpreted with caution.

The thickest seasonal ice occurs in central Baffin Bay between 71°N and 75°N, corresponding with the region of high FCC and weak net heat gain or slight heat loss identified in Section 5.1. Sea ice melt in this area begins between early and mid-May (Fig. 9b), less than 30 days before the onset of the freshening season (Fig. 9c), consistent with a strong contribution of local sea ice melt to the freshwater budget. This interpretation is further supported by the rapid increase in FCC observed between mid-May and mid-July in this region (Fig. 12d–h), concurrent with near zero or weakly negative HCC (Fig. 13d–h).

To further investigate the relationship between sea ice melt, FCC, and HCC, we examined bi-weekly climatological time series of sea ice concentration (SIC; Fig. 14a), FCC (Fig. 14b), and HCC (Fig. 14c) in central Baffin Bay (71–75°N, 60–72°W), separated into three depth layers (0–50 m, 50–150 m, and 150–300 m). The increase in FCC broadly coincides with the rapid decline in SIC between year days ~ 140 and ~ 210, consistent with local sea ice melt being a dominant freshwater source during this period.

The evolution of upper-layer HCC suggests two distinct thermodynamic regimes associated with the seasonal evolution of SIC. During the early melt season, while SIC remains above ~ 80 %, upper-layer HCC stays near zero despite increasing FCC. This behavior likely reflects a near balance between atmospheric heat input and the latent heat consumed by melting beneath a largely continuous ice cover. As SIC decreases below ~ 80 % around year day 170, marking the transition toward a marginal ice zone, upper-layer HCC rapidly becomes positive

and reaches a maximum near year day 210, indicating that atmospheric and oceanic heat input increasingly exceeds latent heat consumption. The subsequent decline in HCC after year day 210 is consistent with the seasonal reduction in atmospheric heat input from late summer onward. After year day 250, upper-layer HCC becomes persistently negative, whereas the 50–150 m and 150–300 m layers stabilize or become weakly positive, consistent with autumn vertical mixing redistributing heat from the surface into the subsurface layers.

The deeper layers display a markedly different behavior during the melt season itself. In the 150–300 m layer, HCC becomes negative between year days ~ 150 and ~ 210 while FCC remains positive throughout. As discussed in Section 5.2, this combination supports the advection of colder Arctic-origin waters entering through Lancaster Sound and propagating southward through Baffin Bay and Davis Strait.

Overall, these results highlight the dominant role of local sea ice melt in driving the seasonal freshwater cycle in central Baffin Bay, while also demonstrating that the associated heat content response depends strongly on sea ice concentration and vertical layer. Together, the FCC and HCC climatologies reveal a transition from an early melt regime dominated by latent heat consumption beneath compact ice cover to a later regime characterized by enhanced upper-ocean heat gain in the marginal ice zone, while deeper layers remain influenced by advected Arctic-origin waters.

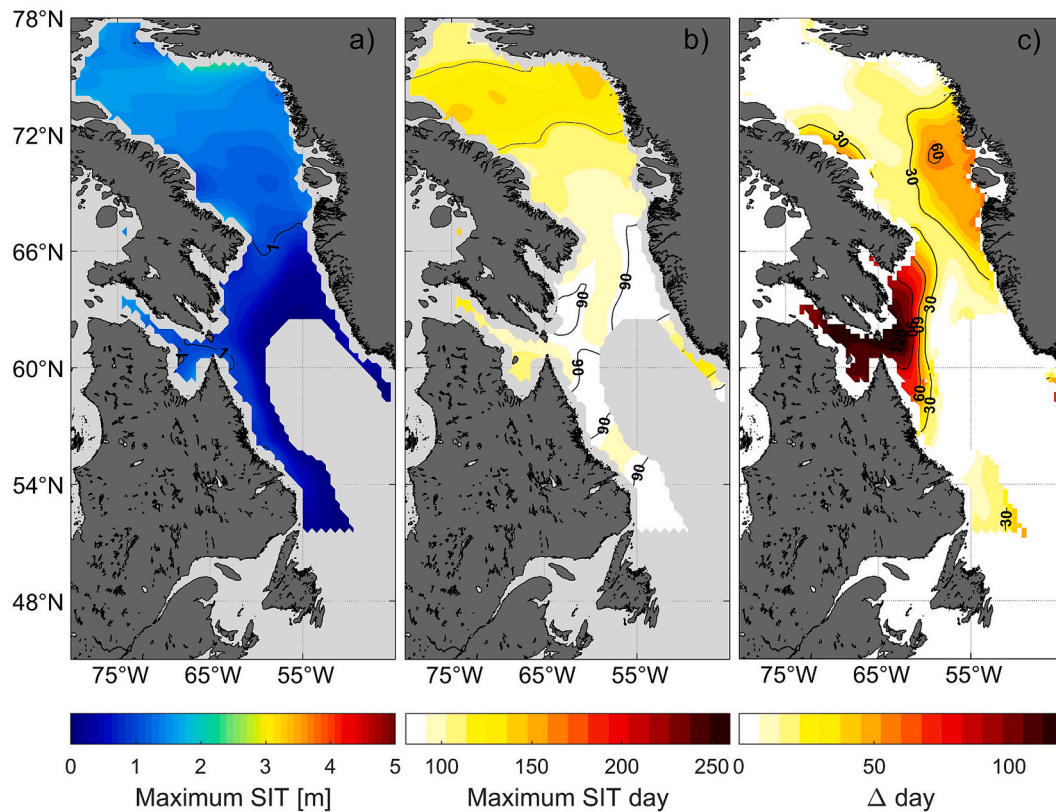


Fig. 9. Distribution of (a) the 1979–2023 averaged maximum Sea Ice Thickness from PIOMAS, (b) the average day of the maximum (or day of sea ice melt onset), and (c) the difference between the onset day of freshening (Fig. 7a) and the onset day of sea ice melting (panel b), with warmer colors indicating a lag of the freshening.

5.4. Advection of ice meltwater from remote locations

High FCC and HCC occur along the West Greenland shelf between the southern tip of Greenland and $\sim 69^\circ\text{N}$, following the pathway of the West Greenland Coastal Current (WGCC, Fig. 1b). As the downstream continuation of the East Greenland Coastal Current, the WGCC transports into our domain a combined freshwater signal of Arctic-origin sea ice meltwater, liquid Arctic freshwater, and East Greenland runoff (Håvik et al. 2017; Sutherland and Pickart 2008; Lin et al. 2018). Among these contributions, the Fram Strait sea ice export is by far the largest, with an integrated ice volume flux of 1951 km^3 between November and May (Spren et al. 2020), approximately one order of magnitude larger than the East Greenland runoff (Bamber et al. 2012; Mernild et al. 2008) and the sea ice produced locally east of Greenland.

We propose that this Fram Strait ice export represents a major contributor to the seasonal FCC variability along the southwestern Greenland shelf and slope. Here, the southwestern Greenland shelf is defined as the region between $58^\circ\text{N} - 69^\circ\text{N}$ inshore of the 500 m isobath, and the slope as the area between $62^\circ\text{N} - 66^\circ\text{N}$, deeper than 500 m (see Fig. 12 for the location of this isobath). A strong similarity exists between the seasonal evolution of the FCC in these regions and that of the Fram Strait ice volume flux (Fig. 15). In both cases, the time series exhibit elevated values lasting about three months, followed by an abrupt decline. The FCC variability lags the Fram Strait ice export by approximately 5–6 months, consistent with the advection timescale estimated from the mean current velocity and the distance between the Fram Strait and southern Greenland (Section 2.2.3).

By the time this freshwater signal reaches our domain, most of the sea ice has already melted during its southward transport along East Greenland. Consequently, the freshwater entering the southwestern Greenland shelf is expected to have already gained substantial atmospheric heat, consistent with the strongly positive HCC observed in this

region (Fig. 8b, Fig. 13d–j). Along the slope region, FCC increases only moderately (dashed-gray line in Fig. 15), whereas HCC continues to rise substantially throughout the melt season (Fig. 13d–m). This contrast suggests that the slope south of Davis Strait may act as an important regional heat reservoir for the BLS system.

The cumulative FCC over the southwestern Greenland shelf and slope (668 km^3) represents approximately one third of the Fram Strait ice volume flux, confirming our earlier assumption (Section 2.2.3) that a large fraction of the exported ice-derived freshwater does not ultimately reach our study region. This attenuation likely results from multiple redistribution processes occurring during the southward transport, including mixing with liquid Arctic freshwater and East Greenland runoff within the EGC system (Håvik et al. 2017), shelf-basin exchange near Cape Farewell (Lin et al. 2018), and recirculation within the Greenland and Iceland Seas or advection toward the central Labrador Sea by eddies and currents (e.g., Dukhovskoy et al. 2016; Foukal et al. 2020). This interpretation is also consistent with recent observations highlighting the strong spatial redistribution of freshwater anomalies across the North Atlantic (Reverdin et al. 2025).

5.5. Glacier and river runoff

Glacier runoff has two main components, namely surface melt and iceberg calving, both of which contribute to the FCC variability observed in our climatology. Here, we focus on the coastal region off Central West Greenland (GCW, $68^\circ\text{N} - 71^\circ\text{N}$, inshore of the 1000 m isobath), downstream of Jakobshavn Glacier (69°N , 49.5°W), which has the highest calving rate among Greenland's marine-terminating glaciers, with typical values between 26.5 km^3 and 50 km^3 per year (Joughin et al. 2014; Joughin et al. (2004)), and represents one of the regions most likely to exhibit a detectable glacier-ocean interaction signal in our dataset.

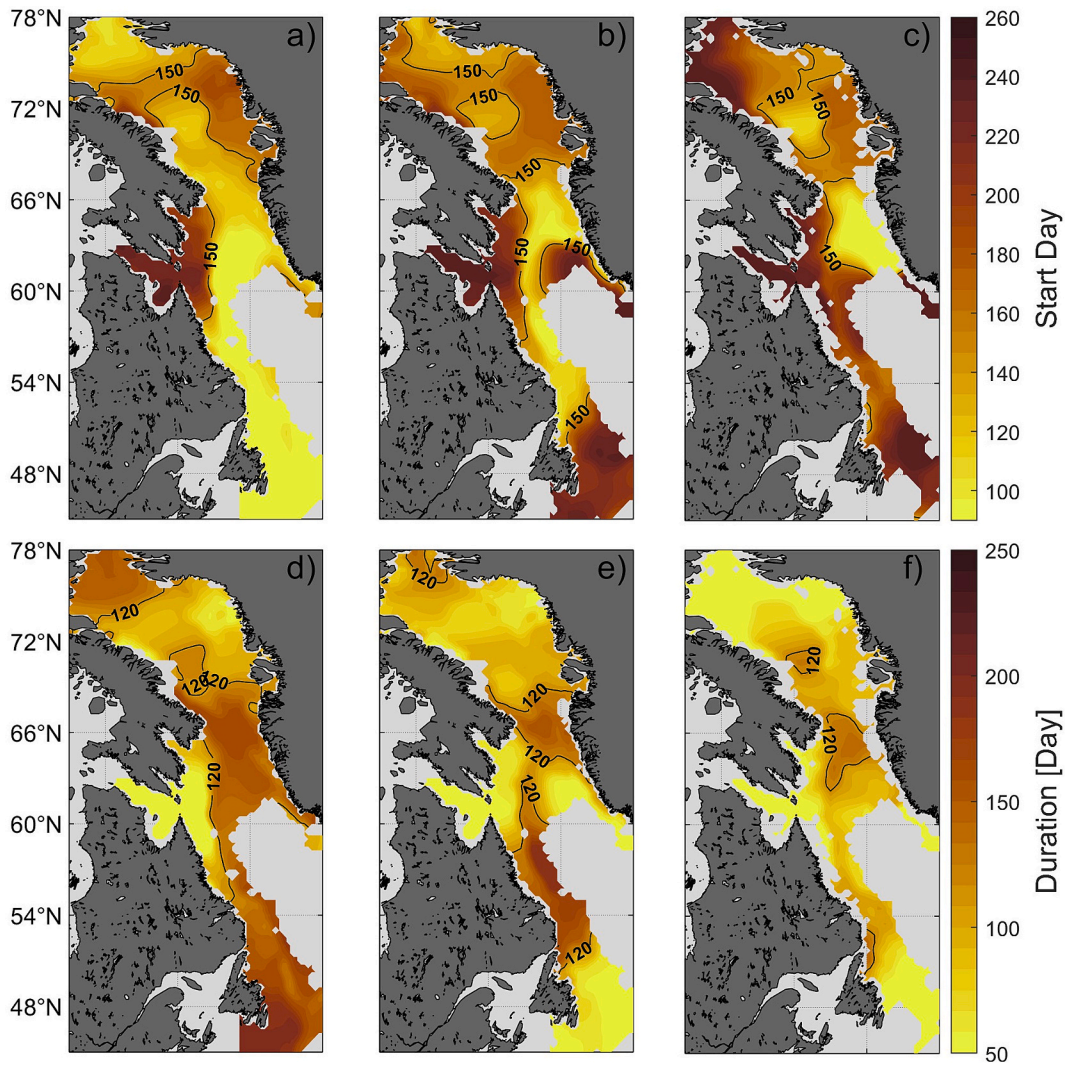


Fig. 10. Spatial distribution of the timing of the freshening season derived from the three-layer FCC climatology. (a–c) Start day of the freshening season for the 0–50 m, 50–150 m, and 150–300 m layers, respectively. (d–f) Corresponding duration of the freshening season (in days) for each layer. The start day is defined as the first day when the cumulative freshwater content anomaly becomes persistently positive, and the duration is defined as the time interval between the start and end of the freshening season. Contours indicate selected reference values to highlight large-scale spatial patterns.

The seasonal cycle of FCC in the upper layer (0–50 m) broadly follows the combined glacier surface melt and ice discharge estimated from the GCW model results (Fig. 16a–b), with both curves exhibiting a clear summer peak centered around year days ~ 200 –210. The similarity between the seasonal cycles suggests that the upper-layer FCC captures the seasonal glacier-influenced freshwater variability in the GCW region. However, the FCC signal cannot be attributed solely to direct glacier meltwater input, since the exported glacial signal is strongly modified by entrainment and mixing with oceanic waters before reaching the shelf (Beaird et al. 2018). Despite the complexity inherent to the modification of the glacier meltwater signal, the cumulative FCC in the GCW region between Day 140 and Day 280 is 80 km^3 , comparable in order of magnitude to the integrated glacier discharge from the GCW model over the same period (100 km^3). Meanwhile, the upper layer HCC is positive throughout summer and declines toward zero or weakly negative values in autumn (Fig. 16c), broadly consistent with the seasonal cycle of atmospheric heat input rather than a strong local ice-melting signal at the surface.

The deeper layers exhibit more complex behavior (Fig. 16b–c). In the 50–150 m layer, the FCC seasonal cycle broadly resembles that of the upper layer but appears delayed by several weeks, possibly reflecting downward redistribution of the surface freshwater anomaly through

mixing and diffusion. The similarity between the 0–50 m and 50–150 m HCC cycles, with the latter similarly delayed, is consistent with this hypothesis. In the 150–300 m layer, FCC and HCC exhibit an approximate inverse relationship, with positive FCC concurrent with negative HCC during mid-summer and the opposite pattern in autumn. Compared to the upper layer, these deeper signals appear more strongly influenced by processes associated with the seasonal variability of Atlantic-origin waters along the West Greenland slope and current system (Münchow 2016). In particular, the increase in HCC during late summer and autumn is consistent with an enhanced influence of relatively warm Atlantic-origin waters at depth, potentially associated with seasonal changes in isopycnal displacement, stratification, and mixing. A more definitive attribution would require targeted process studies beyond the scope of the present study.

5.6. Freshwater budget

The sum of the positive FCC from April to October, excluding the Hudson Strait, is 3509 km^3 (see Section 6 for the rationale about the exclusion of the Hudson Strait from this analysis). In comparison, the freshwater budget established from the various observations, reanalysis, and model data totals 3431 km^3 (Table 1). The close agreement between

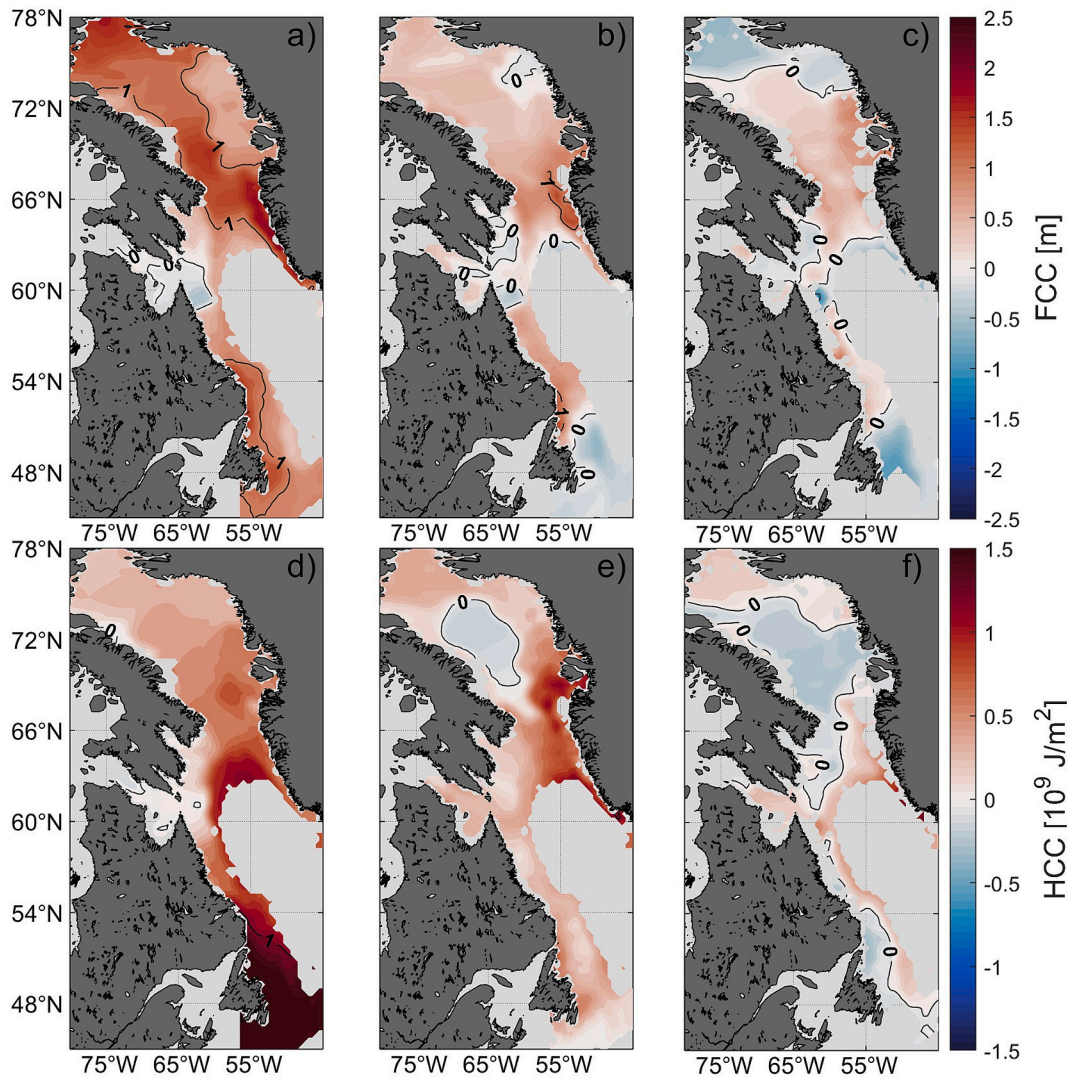


Fig. 11. Vertically partitioned freshwater and heat content changes integrated over the surface-defined freshening season. (a–c) Cumulated freshwater content change (FCC) in the 0–50 m, 50–150 m, and 150–300 m layers, respectively, computed over the freshening period defined from the 0–50 m layer. (d–f) Corresponding integrated heat content change (HCC) in each layer over the same period. Positive FCC indicates freshening, while positive HCC indicates heat gain. The use of a common integration period allows for direct comparison of the vertical structure of freshwater and heat content changes across layers.

these two largely independent estimates suggests that the climatology captures the dominant seasonal freshwater input to the region reasonably well. At the same time, both estimates remain subject to important uncertainties. For the FCC climatology, the cumulative uncertainty estimated from the mapped seasonal uncertainty fields amounts to approximately 880 km^3 over the study domain, corresponding to about 25 % of the cumulative FCC estimate. Likewise, the independent freshwater budget depends on several assumptions and external datasets, including the choice of sea ice product and the fraction of the Fram Strait ice export ultimately reaching the study region.

In Table 1, the Fram Strait ice volume flux (1951 km^3) was conservatively scaled by one third, corresponding approximately to the $\sim 668 \text{ km}^3$ of FCC identified along the southwestern Greenland shelf and slope that could be clearly associated with this signal (Section 5.4; Fig. 15). However, this scaling factor remains an educated estimate rather than a directly constrained quantity, and part of the Fram Strait freshwater signal may spread more diffusely across the domain without producing a distinct FCC signature. Nevertheless, given the uncertainties associated with both the FCC climatology and the independent freshwater budget components, the close agreement between the two estimates suggests that the dominant seasonal freshwater contributions to the region are

captured reasonably well.

The largest contribution to the freshwater budget is sea ice melt. The combined contribution of local sea ice melt and scaled ice volume flux through the Fram Strait and the Canadian Arctic Archipelago represent 1911 km^3 , or 56 % of the total budget (Table 1). Glaciers represent the second-largest contribution, with a total of 552 km^3 (16 %) from surface glacier melt (including the Greenland Ice Sheet and peripheral glaciers) and iceberg calving. River runoff and precipitation minus evaporation over the ocean contribute 15 % and 13 % of the local freshwater budget, respectively. Fig. 17 summarizes the main freshwater sources, pathways, and redistribution processes discussed throughout this study. The schematic highlights both the relative importance of the major freshwater contributions and the strong transformation of freshwater signals during their transport through the BBLS region, particularly along the Greenland and Labrador margins.

In the estimates listed in Table 1, we did not account for either the iceberg calving or liquid runoff from East Greenland, whose yearly value totals $440\text{--}650 \text{ km}^3$ (Mernild et al. 2008; Bamber et al. 2018; Marson et al. 2021). Although the EGC system transports East Greenland runoff together with Arctic-origin freshwater and sea ice meltwater (Håvik et al. 2017; Sutherland and Pickart 2008), we expect the seasonal runoff

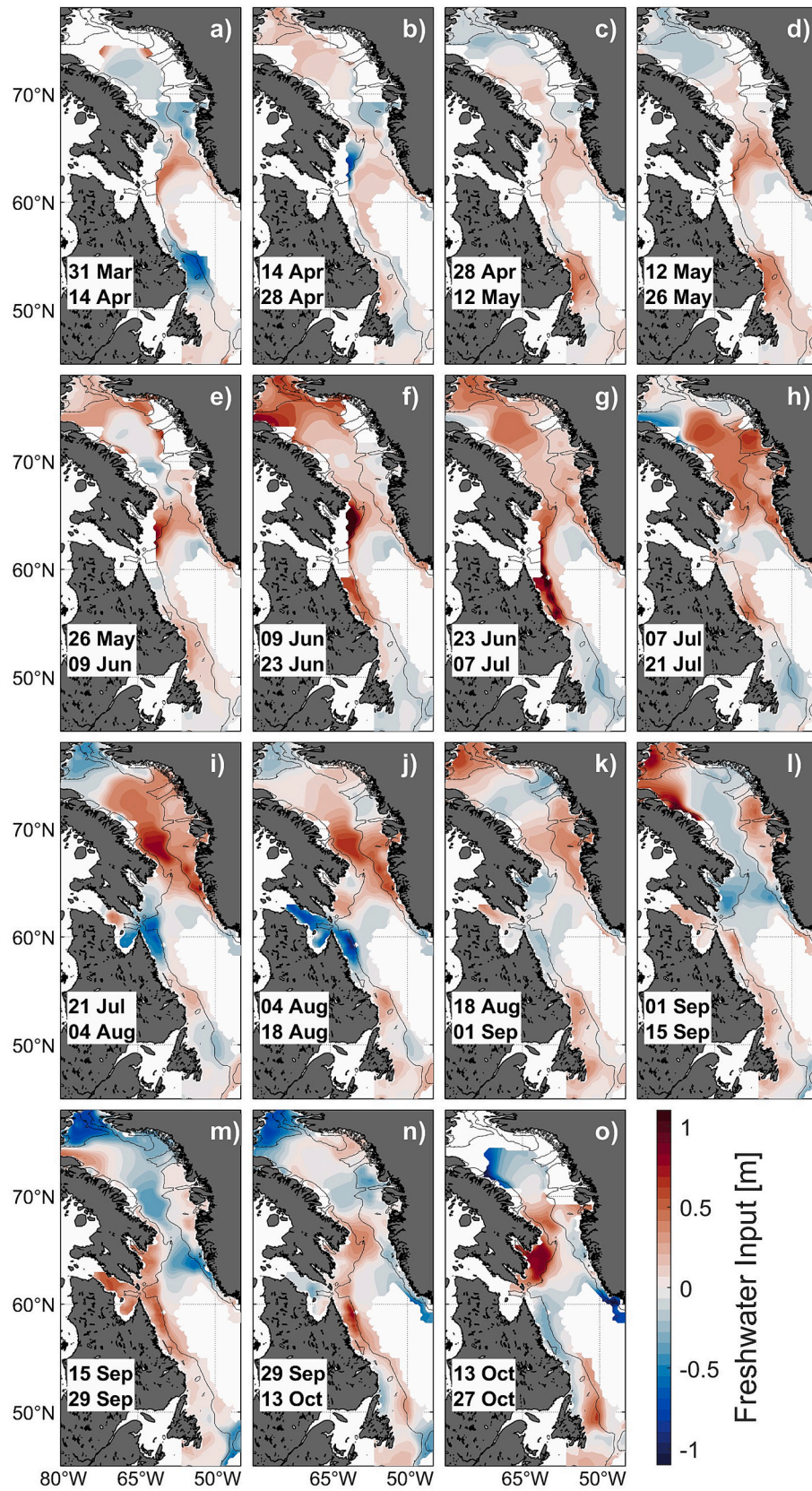


Fig. 12. Bi-weekly distribution of FCC from April to October. The thin black line represents the 500 m isobath.

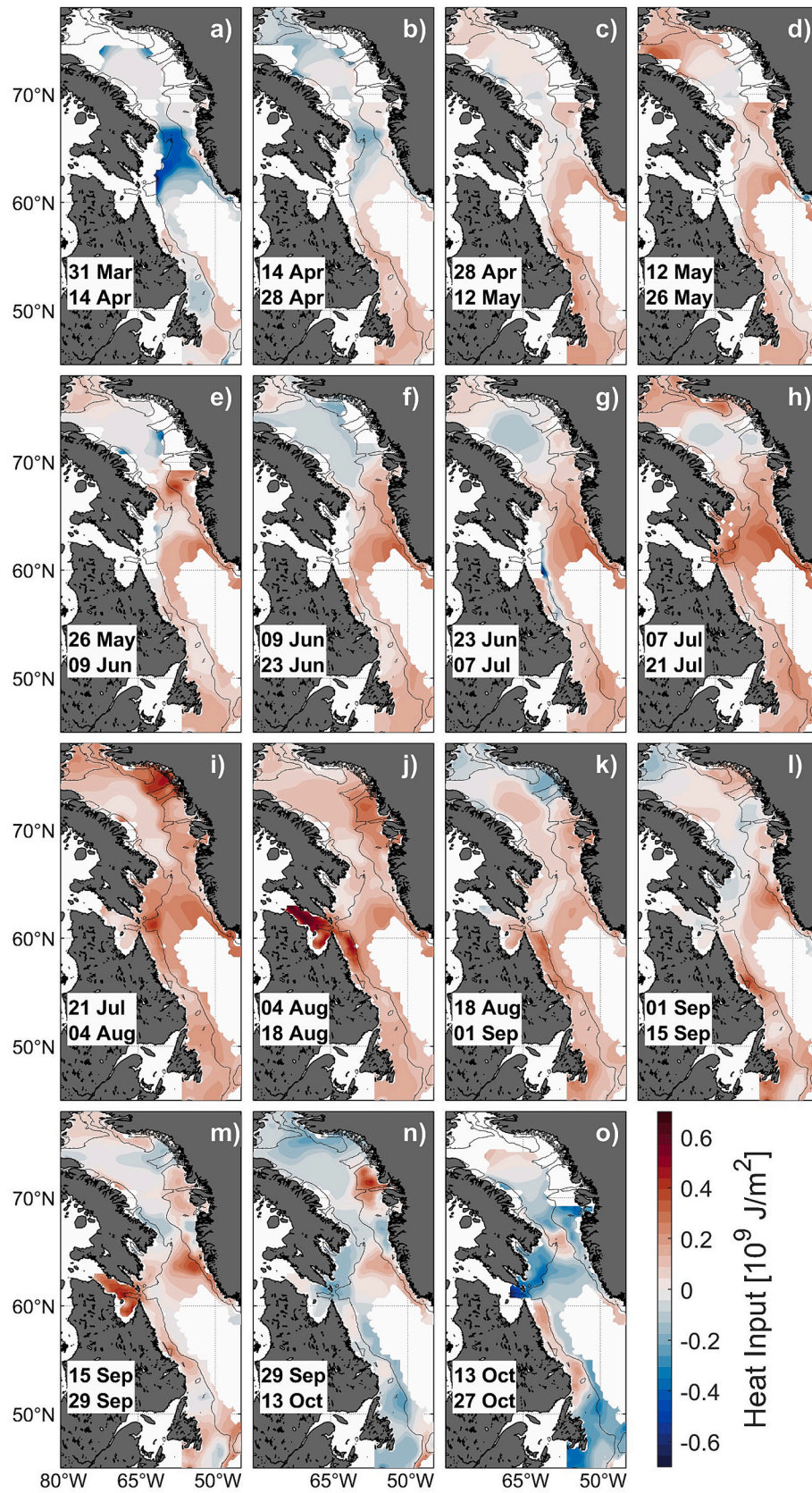


Fig. 13. Bi-weekly distribution of HCC from April to October.

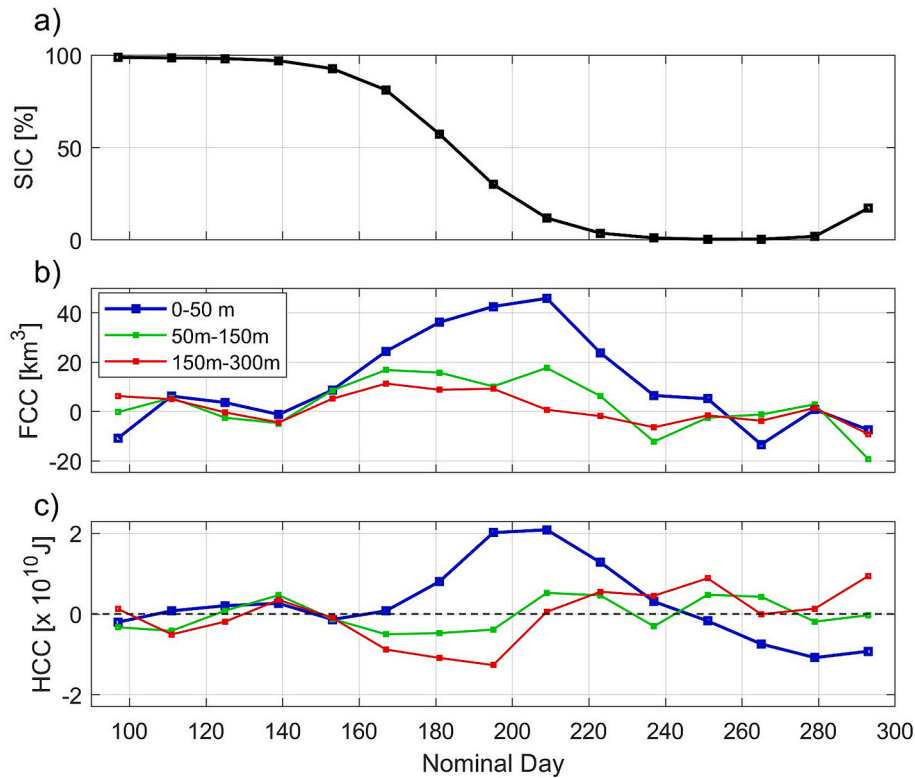


Fig. 14. Bi-weekly climatological evolution of (a) sea ice concentration (SIC), (b) freshwater content change (FCC), and (c) heat content change (HCC) in central Baffin Bay (71–75°N, 60–72°W). FCC and HCC are shown separately for three depth layers: 0–50 m (blue), 50–150 m (green), and 150–300 m (red).

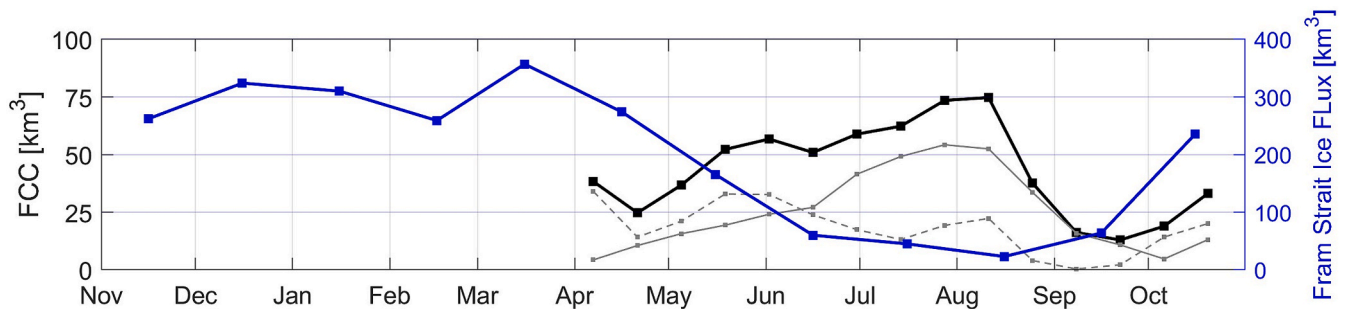


Fig. 15. Bi-weekly positive FCC off Southwestern Greenland (black line) and monthly Fram Strait Ice Flux (blue line, Spreen et al. 2020). The solid and dashed gray lines represent the FCC integrated over the Southwestern Greenland shelf (inshore of 500 m) and slope region (offshore of 500 m), respectively.

signal to be substantially attenuated before reaching our study region because of the long transport distance, the high residence time of freshwater within the deep East Greenland fjords, and redistribution processes along the EGC pathway (Section 5.4). Considering these processes and the six-month duration covered by our climatology, East Greenland runoff likely represents only a minor contribution to the seasonal FCC variability observed in our region.

Notes: ¹ Data available from 1979 to 2023; the uncertainty value is based on the $\sim 10\%$ PIOMAS sea-ice volume uncertainty estimated for late winter conditions by Schweiger et al. (2011), but is not specific to our study region. ² Data available from 1917 to 2024. ³ Data from Spreen et al. 2020 available from 1992 to 2014 scaled to 1/3rd based on our FCC results (Section 5.4); values in brackets indicate the standard deviation of yearly values rather than an uncertainty estimate. ⁴ Out of 520 km³, 356 km³ is contributed by Newfoundland and Labrador, excluding the St. Lawrence River. ⁵ Data from Mankoff et al. (2021), from 1986 to 2022, between April and October.

6. Discussion and conclusion

The freshwater content change (FCC) climatology presented in this study allowed us to describe with great detail the processes of seasonal input of freshwater via sea ice melt, river or ice discharge, or advection of fresh waters throughout the Baffin Bay, the west Greenland Shelf, and the Labrador Shelf. The freshwater input by these regional and seasonal processes is 3509 km³, with an estimated mapped uncertainty of approximately 25 %, against a conservative estimate of 3431 km³ from independent sources (Section 5.6, Table 1). The FCC estimated for our study region exceeds the liquid freshwater outflow from the Arctic Ocean through the Fram Strait (2800 km³, Haine et al. 2015). While the estimates by Haine et al. (2015) are referenced to a salinity reference of 34.8 and thus cannot be compared directly with our FCC values, the similar order of magnitude between these two variables highlights the relative importance of regional and seasonal processes in the freshwater budget of the subpolar North Atlantic.

Our results underscored the essential role of sea ice melt in the

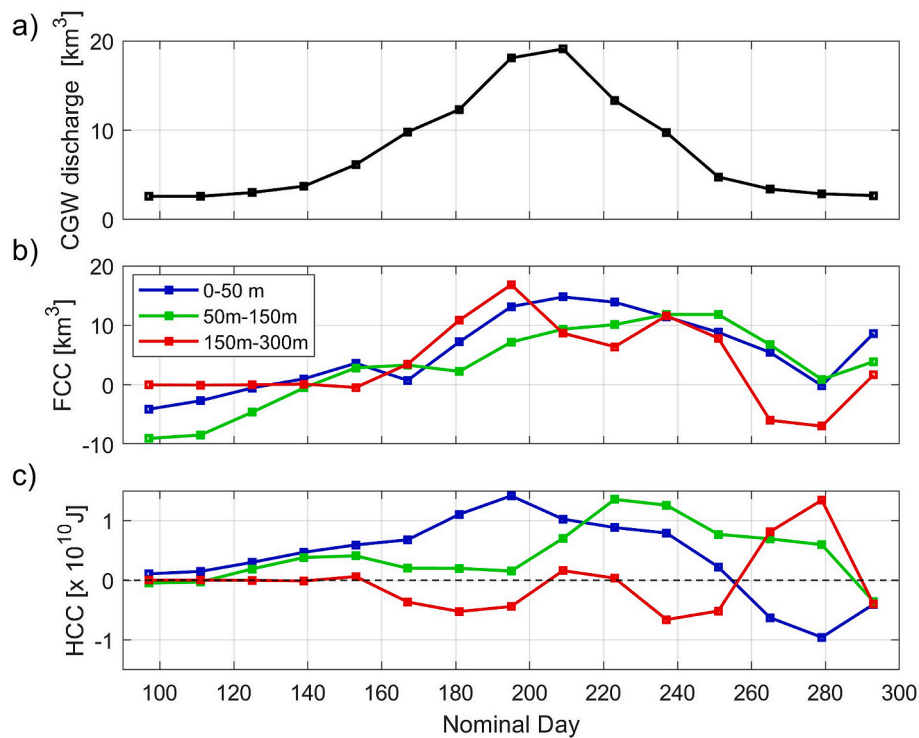


Fig. 16. (a) Bi-weekly climatological combined glacier surface melt and ice discharge for the Central West Greenland (GCW) region, estimated from the energy-mass balance model of Fujita and Sakai (2014). (b) Bi-weekly climatological FCC integrated over three depth layers (0–50 m, blue; 50–150 m, green; 150–300 m, red) in the GCW coastal region (68°N–71°N, inshore of the 1000 m isobath). (c) Same as (b) but for HCC.

freshwater budget, with local sea ice melt and the Fram Strait and Canadian Arctic Archipelago ice volume fluxes accounting for at least 56 % (1911 km³, Table 1) of all seasonal freshwater input in our region. The bi-weekly maps and targeted FCC and HCC time series (Figs. 12–16) allowed us to identify the progression of sea ice melt in Baffin Bay and its relationship with the seasonal evolution of heat content, quantify a freshwater input of 668 km³ along the southwestern Greenland coast and slope—likely associated with the Fram Strait ice volume flux—and evaluate the impact of glacier meltwater runoff offshore of Jakobshavn Glacier. The combined analysis of FCC and HCC in three vertical layers further revealed strong vertical contrasts in the freshwater and heat content variability, distinguishing surface-intensified freshening driven by local sea ice melt from deeper signals more strongly influenced by Arctic-origin waters and Atlantic-origin circulation along the Davis Strait and southwest Greenland slope. The interpretations proposed in this study were supported through local comparisons with a land and ice runoff model, as well as estimates of sea ice volume and flux.

The quality of the results presented in this study owes largely to the careful data processing adopted to produce the climatology, i.e., the estimation of weekly fields of salinity by iterations while accounting for the spatiotemporal biases inherent to ocean observations. To underscore the importance of using such a methodology, we also generated two comparison datasets of weekly FCC mean fields using a simplified procedure in which neither iterations nor estimations of the spatiotemporal biases were made. These simplified datasets, described in Supplementary Section 2, yield considerably larger errors (Supplementary Fig. S4 vs. Fig. 5) than our dataset, and larger FCC estimates (6597 km³ and 8976 km³ depending on the mapping scheme used) compared to our dataset and the freshwater budget (Table 1). These datasets also compare less favorably against the validation climatology at Station 27 (Supplementary Fig. S6 vs. Fig. 6). We thus conclude that to produce reliable climatologies of FCC and HCC at such high temporal resolution, it is essential to account for the spatiotemporal biases and improve iteratively the quality of the estimates (Fig. 5).

Our freshwater budget (Table 1) does not include the freshwater outflow from Hudson Bay through Hudson Strait, despite Hudson Bay representing a major freshwater reservoir. Previous studies showed that freshwater exits primarily through the southern part of Hudson Strait, while Arctic-origin inflow occurs through the northern side, with substantial cross-strait exchange between the two branches (Straneo & Saucier, 2008; Granskog et al. 2009; Florindo-López et al. 2020; Ridenour et al. 2021). Although the annual freshwater export through Hudson Strait was estimated at ~ 900 km³ in 2004–2005 using a salinity reference of 33 (Straneo & Saucier, 2008), most of the outflow occurred in November–December, outside the April–October climatological period considered here, and its seasonality may vary substantially depending on freshwater residence times within Hudson Bay (Granskog et al. 2009). For these reasons, Hudson Strait and the associated Hudson Bay river runoff were excluded from the summer freshwater budget estimate. Nevertheless, both Straneo & Saucier (2008) and Florindo-López et al. (2020) reported strengthening freshwater export in late summer and early autumn, broadly consistent with the elevated FCC observed in Hudson Strait during August–September in our climatology (Fig. 12k–m), totaling ~ 180 km³. This signal likely contributes to the freshwater anomalies observed over the northern Labrador Shelf approximately one month later (Fig. 12l–o). However, the relatively sparse hydrographic coverage and elevated uncertainty in Hudson Strait prevent a more quantitative interpretation.

The high spatiotemporal resolution of our climatology will be very valuable to study the interannual anomalies of FCC, especially regarding the separation of the anomalies' seasonal and aseasonal contributions. This issue is crucial as the advection of freshwater signals with periods greater than a season, like the GSA, is a major source of anomaly that has the potential to significantly change the freshwater content in this region (Häkkinen, 1993; Yashayaev, 2007b; Yashayaev et al. 2015). The possibility of separating signals of remote origin from the local anomaly of the seasonal cycle will help us to improve our understanding of the sources, propagation pathways, and the interaction between these

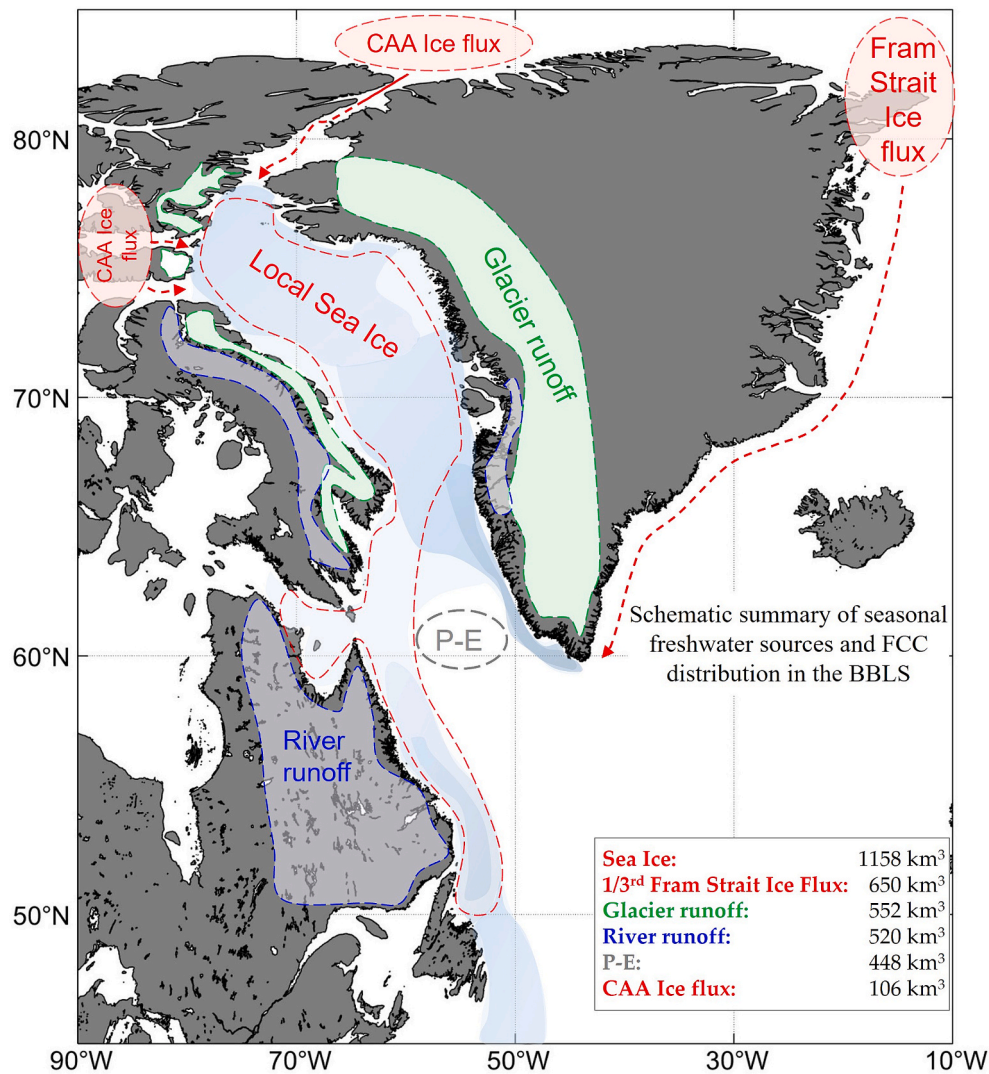


Fig. 17. Schematic summary of the major seasonal freshwater sources and associated freshwater content change (FCC) distribution within the Baffin Bay–Labrador Sea (BBLs) system. The blue shading represents the climatological FCC distribution inferred from the hydrographic analysis, while the dashed contours indicate the approximate regions influenced by the different freshwater forcing components: local sea ice melt and remote sea ice fluxes (red), glacier runoff and iceberg calving (green), river runoff (blue), and precipitation minus evaporation (P–E; gray). Red arrows indicate the main pathways of Arctic sea ice transport through the Canadian Arctic Archipelago (CAA) and from the Fram Strait toward the West Greenland shelf. The freshwater volumes associated with each forcing component are summarized in the legend, while their estimated uncertainties are provided in Table 1.

long-period anomalies and seasonal anomalies, at time scales ranging from interannual to multi-decadal.

While the analysis of long-term trends and interannual anomalies will be the subject of future work, the validation analysis presented in Section 4 already highlights the importance of separating seasonal and low-frequency variability. At Station 27, a significant freshwater increase is observed after removal of the seasonal cycle, accompanied by a marked acceleration since 2007. Notably, the cumulative freshwater anomaly over the 2007–2024 period is comparable in magnitude to the local seasonal freshening cycle itself (1.40 m). This result underscores both the importance of accurately defining the seasonal cycle to avoid aliasing and the magnitude of the freshwater changes that have occurred in the region over the past two decades.

The analysis of FCC anomalies may help better understand two important long-term freshwater signals affecting the BBLs region: the increasing freshwater input from Greenland glaciers and ice sheet melting, and the possible release of freshwater accumulated within the Beaufort Gyre. The annual freshwater input from the Greenland Ice Sheet and surrounding glaciers is approximately 200 km³ yr⁻¹ larger than in the 1990 s (Bamber et al. 2012; Bamber et al. 2018),

corresponding to a cumulative anomaly of nearly 5000 km³ since 1994. Although the signature of East Greenland runoff likely remains weak within our domain (Section 5.5), its long-term anomaly should be detectable in the FCC anomaly field. Meanwhile, the Beaufort Gyre has accumulated anomalously large freshwater since the early 2000 s (Proshutinsky et al. 2019; Haine, 2020), exceeding the typical 5–10-year accumulation-release cycle (Zhang et al. 2021). The rate at which this freshwater is released may strongly influence the Labrador Sea freshwater budget and deep convection, with potential implications for the AMOC (Yashayaev and Dickson, 2008; Jahn and Holland, 2013; Thorndalley et al. 2018). A precursor of this signal may already have reached the central Labrador Sea through increased Fram Strait ice export in 2023 (Yashayaev, 2024).

More broadly, the present methodology also offers an alternative perspective on the long-standing question of how freshwater variability should be quantified in the ocean. Following the discussion of Florindo-López et al. (2020) regarding the conceptual relationship between freshwater-content and salinity-budget approaches, the FCC framework developed here estimates the actual temporal evolution of freshwater within the water column by integrating observed salinity

changes over objectively determined freshening periods, rather than defining freshwater relative to an arbitrary fixed salinity reference. In this sense, the approach remains closely connected to the salinity-budget perspective advocated by Schauer and Losch (2019) while still yielding freshwater-content estimates that can be directly compared with independent freshwater forcing estimates. As emphasized by Florindo-López et al. (2020), uncertainties in freshwater transport and storage estimates remain strongly linked to both observational limitations and methodological choices. The agreement obtained here between FCC estimates and independent freshwater forcing estimates suggests that accounting explicitly for seasonal aliasing and temporal sampling may help reconcile some of the differing perspectives between freshwater-content and salinity-budget approaches.

Declaration of generative AI and AI-assisted technologies in the manuscript preparation process

During the preparation of this work, the author(s) used OpenAI ChatGPT and Anthropic Claude Sonnet for English editing and for improving the clarity, structure, and presentation of the manuscript. The author(s) reviewed and edited the output as needed and take full responsibility for the content of the published article.

CRediT authorship contribution statement

Vigan Mensah: Writing – original draft, Validation, Methodology, Investigation, Funding acquisition, Formal analysis, Conceptualization. **Igor Yashayaev:** Writing – review & editing, Writing – original draft, Methodology, Investigation, Conceptualization. **Koji Fujita:** Writing – review & editing, Writing – original draft, Methodology, Formal analysis, Data curation. **Stephen E.L. Howell:** Writing – review & editing, Writing – original draft, Methodology, Data curation. **Miho Ikeda:** Writing – review & editing, Conceptualization. **Kay I. Ohshima:** Writing – review & editing, Funding acquisition.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.pocan.2026.103812>.

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Supplementary Information for

A climatology of seasonal freshwater accumulation as a basis for defining the
freshwater budget in the Baffin Bay and Labrador shelf

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Supplementary Text 1: Technical details for the implementation of the iterative methodology

This section provides the implementation details of the methodology described in Section 3 of the main text.

1. Data preselection

Quality control was applied to the hydrographic dataset using a combination of automated density inversion detection and systematic visual inspection.

For the automated screening, vertical density gradients were computed for each profile and evaluated against a set of thresholds ($\text{kg m}^{-3} \text{ m}^{-1}$): -0.005 , -0.01 , -0.03 , -0.05 , and -0.10 . Profiles were flagged and removed if the number of density inversions exceeding these thresholds surpassed predefined limits of 30, 15, 5, 3, and 1 occurrences, respectively. These thresholds were chosen conservatively to minimize false rejections while ensuring the removal of profiles exhibiting large or numerous unstable density gradients.

Automated screening was complemented by systematic visual inspection. Profiles were examined sequentially by year of acquisition and by nominal day of year (in two-day increments), allowing identification of profiles inconsistent with the expected seasonal or interannual hydrographic structure.

Flagged profiles typically corresponded to instrumental issues, including faulty sensors or conductivity cell thermal-mass effects producing salinity spikes or hysteresis. In rare cases, profiles associated with extreme mesoscale events not representative of the local hydrographic variability were also excluded. Following this procedure, approximately 3–5% of the profiles were removed from the dataset.

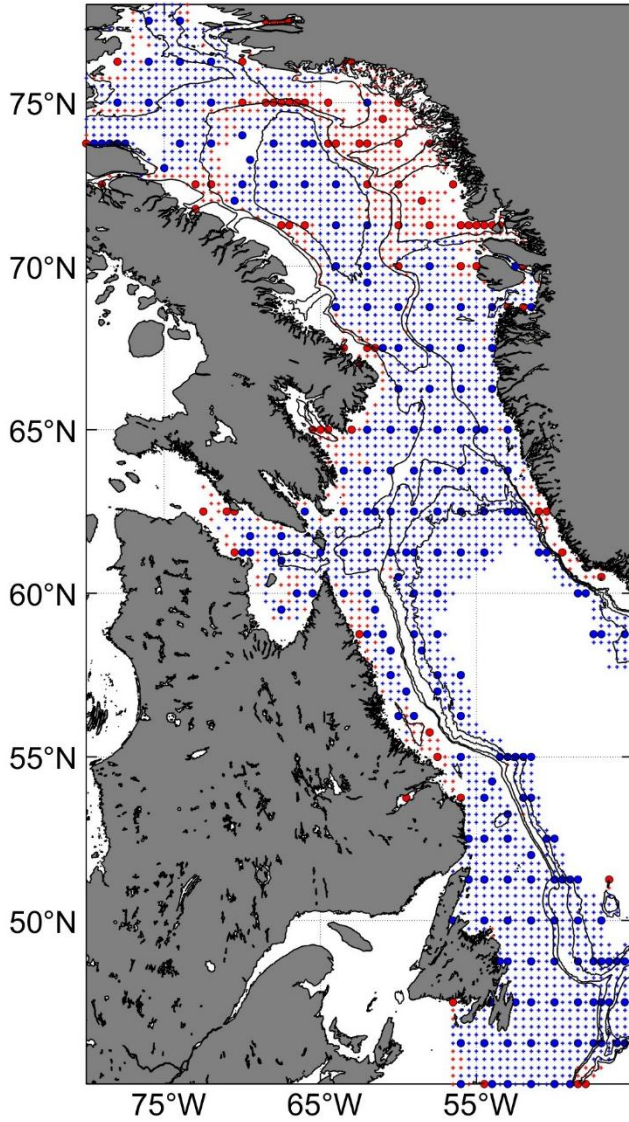


Fig. S1. Distribution of the selected analysis locations used for the construction of the weekly climatology. Small crosses indicate locations fulfilling the selection criteria, with blue crosses corresponding to points satisfying the criteria within one spatial decorrelation scale (L_r) and topographic constraint (L_ϕ), and red crosses corresponding to points satisfying the relaxed criteria within $1.5 L_r$ and $1.5 L_\phi$. Circles indicate the 272 subsampled analysis locations retained based on the horizontal decorrelation scale in order to reduce spatial redundancy. Thin black contours represent the 500 m, 1000 m, 2000 m, and 3000 m isobaths, shown to illustrate the slope steepness across the domain.

2. Spatial bias estimation

The background spatial gradient (spatial bias) was estimated on a 10×10 km grid covering the study domain as

$$S_{bias}(i) = \overline{S}_y(i) - S_{ref} \quad (\text{SE1}),$$

where $\overline{S}_y(i)$ is the mean salinity at grid cell (i), and S_{ref} is the reference salinity.

Because observations are more numerous in mid-summer than during the rest of the season, a direct average of all available salinity values would be biased toward summer conditions. To reduce this effect, weekly mean salinities $S_w(i, t)$ were first estimated for each week (w) (D90, D97, ..., D304) by averaging salinity data located within 25 km of the grid cell and within ± 14 days of the target week. The mean salinity $\overline{S}_y(i)$ was then obtained by averaging these weekly values, provided that at least eight weekly bins were available.

The reference salinity S_{ref} was defined as an isohaline of the 0–300 m depth-averaged salinity field present throughout most of the domain. In this study, $S_{ref} = 33.5$.

3. Weekly bin construction

At each selected analysis location, weekly salinity values were estimated by weighted averaging of the available observations using the SSCI mapping scheme described in Section 3.2 of the main text. The weighted mean was computed as

$$S_{week}(i, w) = \frac{1}{\sum W} \sum W(n) \cdot S_{cor}(n) \quad (\text{SE2}),$$

where $W(n)$ is the SSCI weight, $n = 1, \dots, N$ indexes the observations included in the bin, and S_{cor} is the bias-corrected salinity:

$$S_{cor}(n) = S_{raw}(n) - S_{bias}(n) - S_{biasT} \quad (3),$$

where $S_{bias}(n)$ and $S_{biasT}(n)$ are the spatial and temporal biases interpolated at the location of observation n . At the initial iteration S_{biasT} was set to 0.

For each weekly bin, observations were first selected within ± 7 days of the bin center and within $1.5 L_r$ and $1.5 L_\phi$. If fewer than 50 observations were available, the selection criteria were progressively relaxed as follows:

- ± 10 days, $1.75 L_r$, $1.75 L_\phi$

- ± 14 days, $2 L_r$, $2 L_\phi$
- fallback condition: at least 25 observations within ± 14 days and $0.5 L_r$, without applying the depth-difference constraint.

If fewer than 25 observations were available after the fallback condition, no weekly value was estimated for that bin. For bins occurring during the sea-ice season, observations were allowed to extend up to 60 days before the bin center, based on the assumption that the water column may undergo limited modification under winter sea-ice cover. Accordingly, observations from February to October were used to construct the climatology for the April–October period.

4. Outlier detection

Outliers were identified from salinity data located within one decorrelation scale L_r and one L_ϕ of each selected analysis location. For each time series, the weekly climatological mean, spatial bias, and temporal bias were subtracted from the raw salinity values:

$$S_{cor}(n) = S_{raw}(n) - S_{biasS}(n) - S_{clim}(n) - S_{biasT}(n) \quad (4),$$

where all correction terms were interpolated at the position of the raw observation. At the initial iteration, S_{biasT} was set to 0. Observations falling outside the mean ± 5 median absolute deviations were excluded.

5. Low-frequency variability

At each selected analysis location, a weekly salinity time series was constructed for each year over 1950–2023 using the same weighted-mean framework as for the climatological weekly bins, except that a minimum of 5 observations was required to estimate a yearly weekly value.

Yearly anomalies were then computed relative to the climatological seasonal cycle. A yearly mean anomaly was retained only if at least 7 weekly anomaly values were available in that year. The low-frequency component was then estimated by applying a 5-year triangular moving average to the yearly anomaly time series.

Finally, these low-frequency anomaly values were spatially mapped using the spatial component of the SSCI scheme in order to obtain the temporal bias field used in the iterative procedure.

Supplementary Text 2. Uncertainty estimates of freshwater content change (FCC)

Uncertainty estimates were evaluated for both the seasonal freshwater content change (FCC) fields and the integrated FCC time series presented in this study. These uncertainty estimates should be interpreted as uncertainty envelopes associated with the observational and methodological limitations of the FCC reconstruction, rather than as strict statistical confidence intervals.

1. Weekly salinity uncertainty

At each selected location and weekly bin, salinity was estimated as a weighted average of the contributing observations using the spatial weighting function defined in Eq. (4). Two complementary sources of uncertainty were considered.

First, the local spread of the observations contributing to each weekly bin was estimated from a weighted standard deviation:

$$\sigma_{\text{bin}} = \sqrt{\frac{\sum_i w_i (S_i - \bar{S})^2}{\sum_i w_i - \frac{\sum_i w_i^2}{\sum_i w_i}}},$$

where S_i are the contributing salinity observations and w_i their associated spatial weights.

Second, a residual uncertainty associated with the climatological reconstruction was estimated from the root-mean-square residual between individual observations and the climatological salinity field:

$$\sigma_{\text{clim}} = \sqrt{\langle (S_i - S_{\text{clim}})^2 \rangle}.$$

The total salinity uncertainty was estimated as:

$$\sigma_S = \sqrt{\sigma_{\text{bin}}^2 + \sigma_{\text{clim}}^2}$$

2. Seasonal FCC uncertainty

The seasonal FCC corresponds to the salinity difference between the beginning and end of the freshening season:

$$\text{FCC}_{\text{season}} = \frac{H}{S} [S(D_{\text{end}}) - S(D_{\text{start}})],$$

where H is the layer thickness (or local bottom depth in shallow regions).

Assuming independent uncertainties between the beginning and end of the freshening season, the uncertainty associated with the seasonal FCC estimate is:

$$\sigma_{\text{FCC}} = \frac{H}{S} \sqrt{\sigma_S^2(D_{\text{start}}) + \sigma_S^2(D_{\text{end}})}.$$

The resulting uncertainty field is shown in Supplementary Fig. S2.

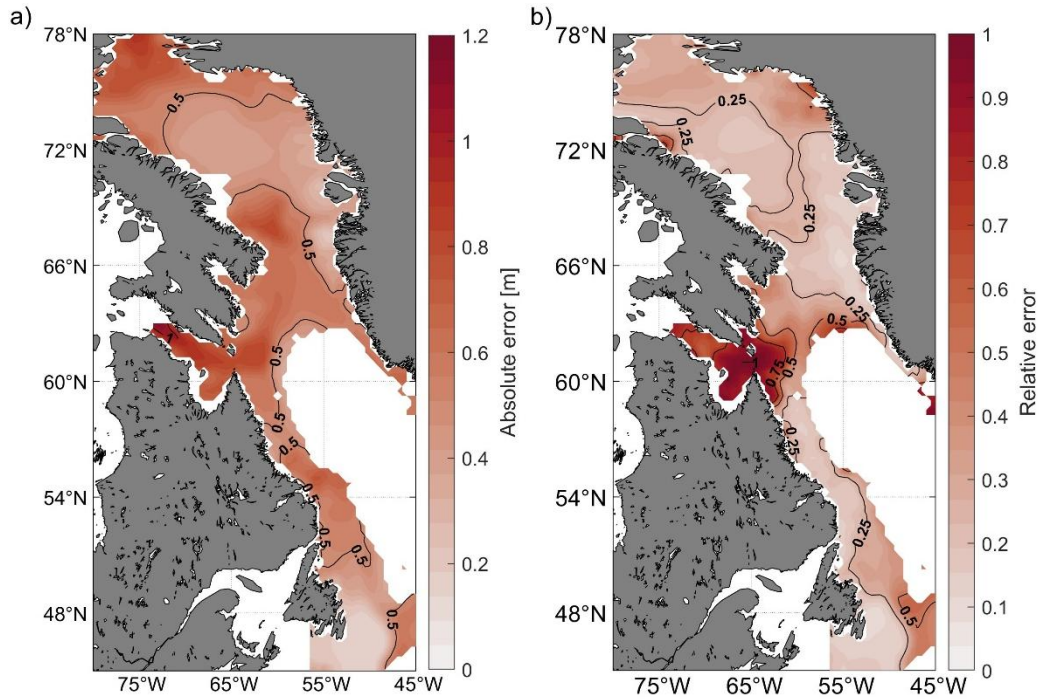


Fig. S2. Spatial distribution of the uncertainty associated with the cumulative seasonal FCC estimates.

(a) Absolute uncertainty of the cumulative FCC (m), estimated from the mapped seasonal uncertainty fields integrated between the beginning and end of the freshening season. Black contours indicate the 0.5 m absolute uncertainty level. (b) Relative uncertainty, defined as the ratio between the absolute uncertainty and the cumulative seasonal FCC. Black contours indicate relative uncertainty levels of 0.25,

0.5, and 0.75. Elevated uncertainties are primarily observed in regions of sparse hydrographic sampling and/or weak cumulative FCC, particularly in Hudson Strait and parts of northwestern Baffin Bay.

3. FCC volume uncertainty

Integrated FCC volumes were obtained by spatially integrating FCC over the selected study regions. Because the selected points were subsampled approximately at the horizontal decorrelation scale, neighboring selected locations were considered approximately independent spatial samples.

The uncertainty associated with the integrated FCC volume was therefore estimated as:

$$\sigma_F = \sqrt{\sum_i (\sigma_{\text{FCC},i} A_i)^2},$$

where A_i is the area represented by each selected location.

Supplementary Text S3: Comparison of the GLIMB energy-mass balance model with an independent estimate of land and glacier runoff

To calibrate the GLIMB in the ice-covered terrain, and the Energy balance models in ice-free terrain, we adopted the simulated freshwater inputs from the ice-covered surface with calibrated precipitation and from the ice-free surface with non-calibrated precipitation, respectively. To discuss the validity of this procedure, we compared the results (discharge and precipitation) with vs. without the calibration (Fig. S3). For the ice-free terrain, we first compared the results yielded from the monthly ERA5-based precipitation minus evaporation, and those of the simulation without the precipitation calibration (Figs. S3a-c). The precipitation data from the two sources are identical (Fig. S3b) and the slight differences in discharge or river runoff (Fig. S3a) thus result from the moderate discrepancy in the evaporation results from the two sources (Fig. S3c). Further, the comparison between the simulated results in their calibrated and non-calibrated version (Fig. S3d-f) revealed that the calibrated discharge data were two to four times larger than the non-calibrated results (Fig. S3d). Since the evaporation data exhibit little sensitivity to the calibration (Fig. S3f), the discharge responds mostly to the precipitation calibration (Fig. S3d vs. Fig. S3e). The largest contribution to the river runoff is that of the Newfoundland and Labrador region (NFL in Fig. 1), with 301 km^3 for the non-calibrated data vs. 1100 km^3 for the calibrated data. Since discharge data estimated from in-situ data by Dery and others (2016) in this same region represents 190 km^3 , we decided to adopt the simulated discharge without precipitation calibration as the freshwater input from the ice-free terrain.

On the other hand, the precipitation calibration did not affect the discharge from the glacier (Fig. S3g). This seems counterintuitive because the GLIMB model should yield more negative mass balance than the satellite-based observational data if the lower ERA5 precipitation was used directly. Thus, the model should yield more meltwater as discharge. Conversely, as the precipitation calibration resulted in an increase in precipitation, the meltwater discharge should be suppressed. The insignificant difference between the results with/without precipitation calibration is due to the compensation of the suppressed meltwater

by the precipitation falling as rainwater during the summer season. We thus decided to use the precipitation-calibrated version of the GLIMB model results for our study.

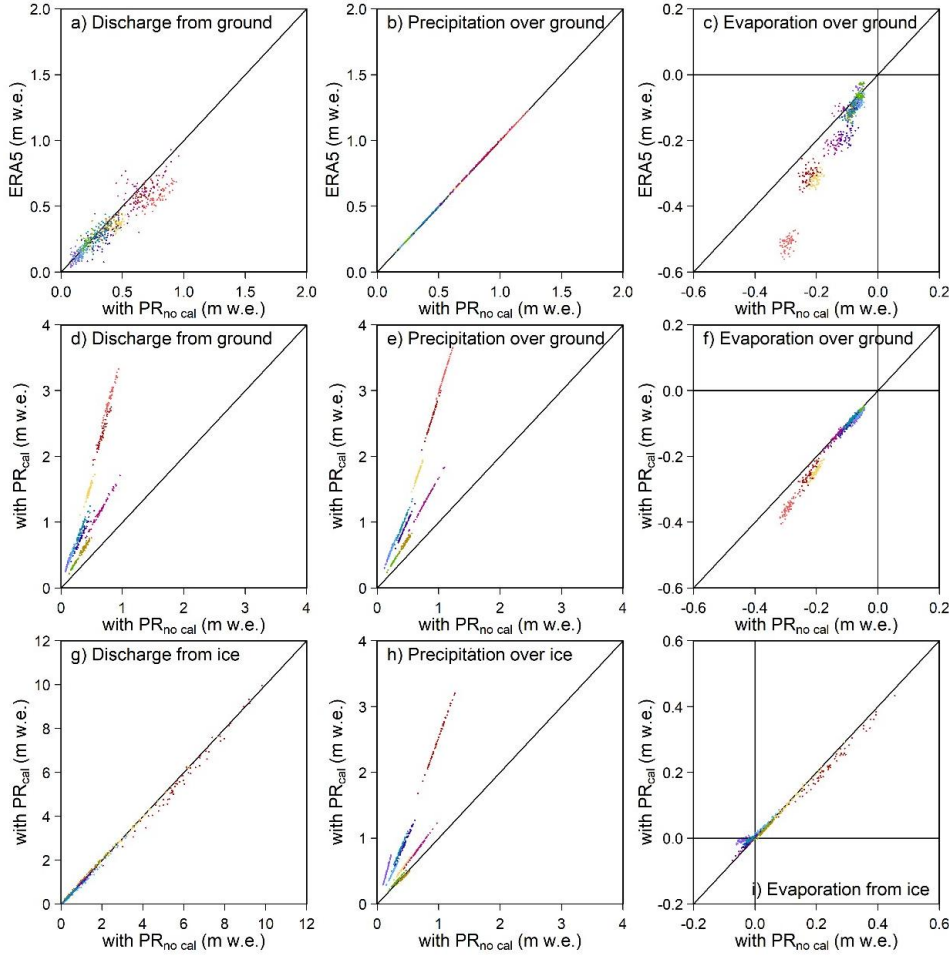


Fig. S3. Comparisons of discharge (left panels, a, d, g), precipitation (center panels, b, e, f), and evaporation (right panels, c, f, i) between (a-c) the simulation without precipitation calibration and the use of ERA5 data, (d-f) between the simulations with vs. without precipitation calibration for the ice-free terrain, and (g-i) for the glacier surface.

Supplementary Text S4: Comparison of the GLIMB model with an independent land and ice runoff product

We compared the climatological seasonal cycle of ice melt (glacier runoff) and land runoff from our GLIMB-based model with the independent product of Igneczi and Bamber (2024) for five of the regions within our study domain: Devon Island (DVN), Baffin Island (BFN), Southwest Greenland (GSW), Central West Greenland (GCW), and Northwest Greenland (GNW). Newfoundland and Labrador (NFL) was excluded from this comparison as it falls outside the spatial coverage of the Igneczi and Bamber (2024) product. The two products differ substantially in methodology: our approach uses the GLIMB glacier energy-mass balance model (Fujita and Sakai 2014) forced directly by ERA5, with precipitation parameters calibrated at 0.5° resolution against observed long-term glacier mass balance (Hugonnet et al. 2021), and explicitly simulates the surface energy and mass balance at 50-m elevation intervals. In contrast, Igneczi and Bamber (2024) route surface mass balance output from the RACMO2 regional climate model to coastal outflow points using a DEM-based drainage basin delineation, without explicit energy balance simulation at the glacier surface. Given these methodological differences, the overall agreement between the two products is remarkable.

The seasonal cycles of ice runoff (black curves, Fig. S4) show close agreement in both amplitude and timing across all five regions, with no systematic bias. The ice runoff totals over the five regions sum to 436 km^3 in our product and 478 km^3 in Igneczi and Bamber (2024), a difference of approximately 9%. For land runoff (green curves, Fig. S4), the agreement is generally good in DVN, GSW, GCW, and GNW, both in terms of seasonal cycle shape and amplitude. The main exception is BFN, where our product yields a substantially larger land runoff than Igneczi and Bamber (2024), with an earlier and sharper peak around day ~ 155 compared to the more moderate and later peak in the Bamber product. The cause of this discrepancy is not entirely clear but may reflect differences in the treatment of tundra drainage and permafrost hydrology between the two approaches, or differences in basin delineation for this region. The land runoff totals over the five regions sum to 163 km^3 in our product and 122 km^3 in Igneczi and Bamber (2024), with the BFN discrepancy

accounting for most of this difference. Despite these regional differences, the total runoff (ice melt + land runoff) across the five regions is nearly identical: 599 km³ in our product versus 600 km³ in Igneczi and Bamber (2024), a difference of less than 1%. This near-perfect agreement in the aggregate total, despite compensating regional differences, provides confidence in the overall reliability of our runoff estimates.

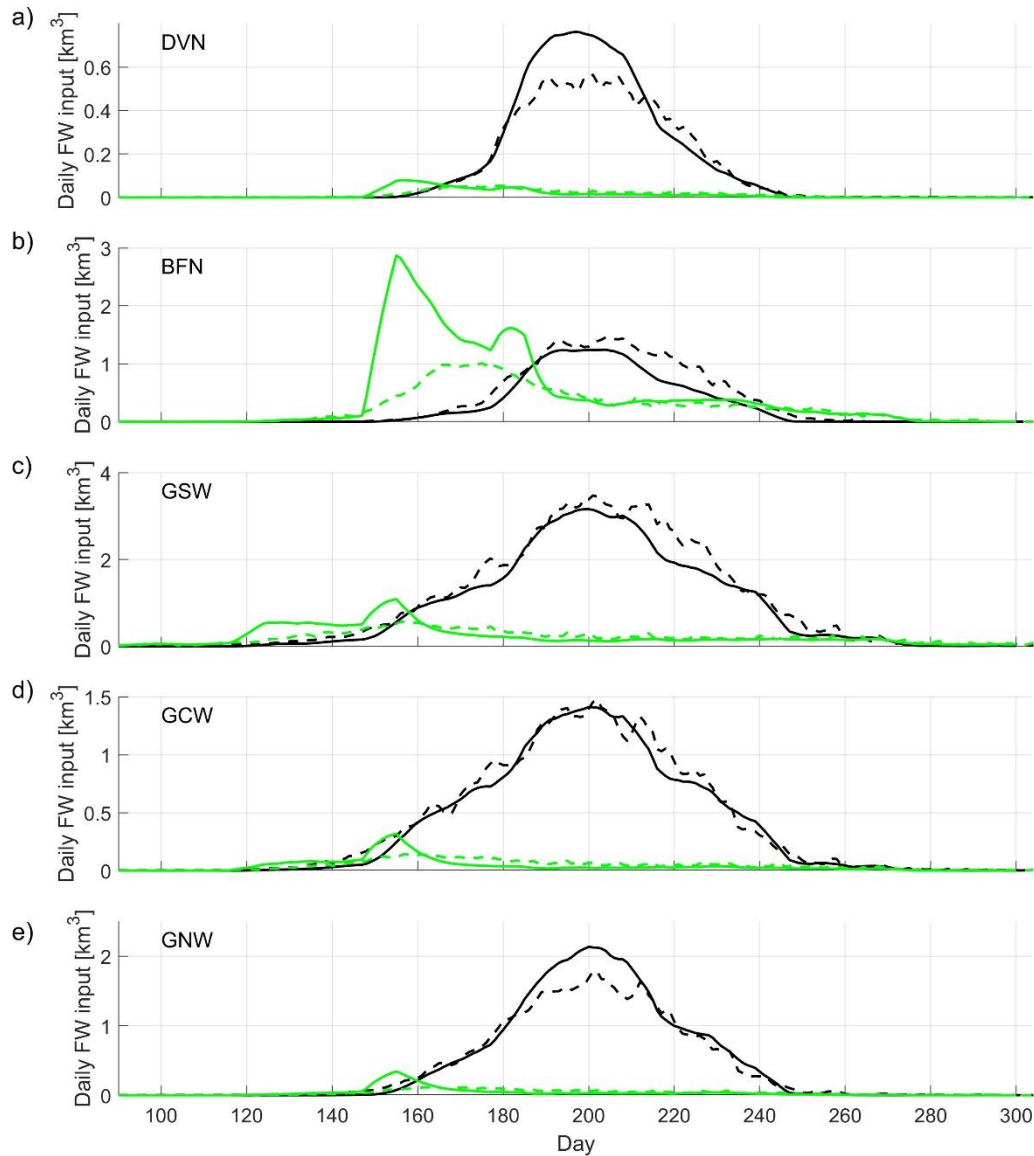


Fig. S4. Comparison of the climatological daily freshwater input from the GLIMB energy-mass balance model (solid lines) and the independent product of Igneczi and Bamber (2024) (dashed lines) for ice runoff (black) and land runoff (green), for five regions within the study domain: (a) Devon Island (DVN), (b)

Baffin Island (BFN), (c) Southwest Greenland (GSW), (d) Central West Greenland (GCW), and (e) Northwest Greenland (GNW). The climatological period was set to that of the GLIMB model (1965–2022) in both cases.

Supplementary Text S5: Evaluation of salinity and FWI climatologies using a simplified methodology

Datasets of ocean temperature and salinity observations are characterized by uneven sampling in space and time that can introduce numerous biases to the mean field. These biases cause large errors in the freshwater or heat input estimates, which are estimated from the derivative of salinity and temperature. To circumvent this issue, we quantified the spatial bias generated by uneven sampling across salinity fronts and the temporal bias caused by uneven sampling within a season and natural anomalies with a period greater than a season, such as the Great Salinity Anomalies (section 3, Figure 5). These biases were removed when estimating the weekly mean field of salinity, and several iterations consisting of weekly mean estimate, outlier detection, and spatial and temporal biases estimate, were conducted to obtain the results shown in this study (Section 3).

To underscore the importance of such a procedure, we also generated two comparison datasets of weekly FWI using a simplified procedure. Hereafter, the dataset shown in the main text of this paper is named the Standard Dataset, and the simplified versions are named Class1 and Class2. In these versions, weekly mean fields of salinity were calculated by mapping salinity data following a careful data selection similar to that used in this study (Supplementary Text 1). However, the two datasets were made without iteration and detection of the spatial and temporal biases. For Class1, we used the Shrinking-Stretching Integrated Constraint mapping scheme by Mensah and Ohshima (2023), also used in our standard dataset. This scheme has distance ($L_r=158$ km), relative depth difference ($L_{phi}=0.51$), and time ($L_T=30$ days) as constraints (Eq. 1). For Class2, we used a weighted-mean scheme with only distance and time as constraints. The summer-integrated results of these two datasets are shown in Fig. S5. Error maps are presented in Supplementary Figs. S6 whereas a comparison of Class1 and Class2 with the data at Station 27 is shown in Fig. S7 and Fig. S8, respectively.

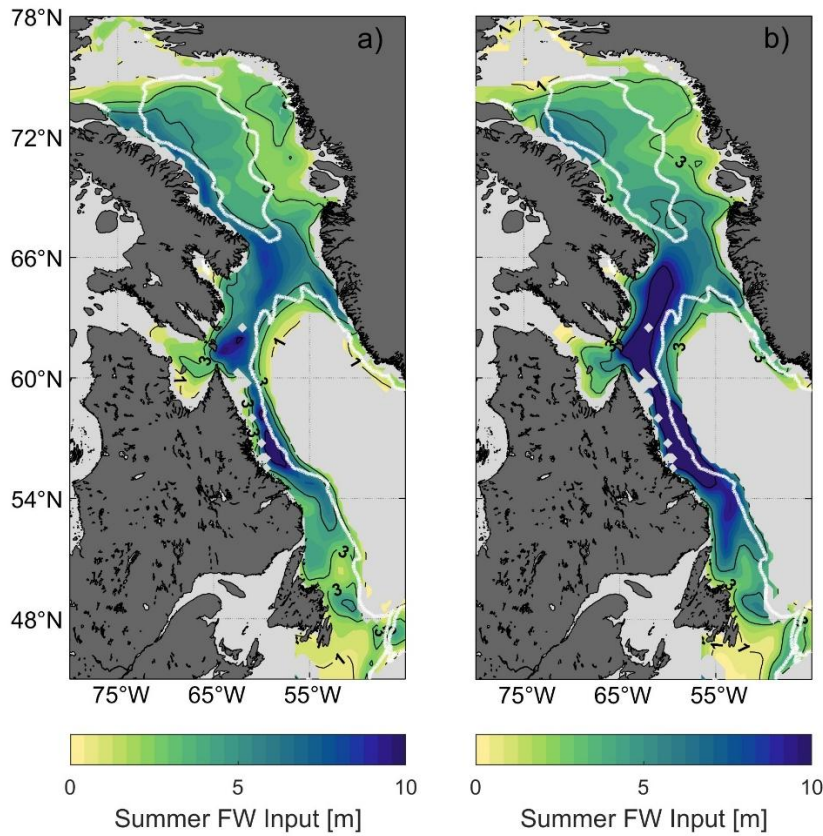


Fig. S5. Distribution of the FWI integrated during the freshening season for the simplified climatologies (a) Class1 (using the SSCI weighting scheme with distance, topographic, and time constraint) and (b) Class2 (using a weighting scheme with only a distance and topographic constraint). The white contour line represents the 1000 m isobath. Note the difference in color scale (0-10 m) compared with Fig. 8a (0-5 m).

The Class1 (Fig. S5a) and Class2 (Fig. S5b) FWI climatologies present a general pattern similar to that of the standard climatology in Fig. 8a. They differ however by their larger values, especially off Baffin Island, in the Davis Strait and off the Labrador coast north of 54°N. These large FWI are likely not to be accurate as the variance of the residual between the climatology and the raw salinity data for Class1 and Class2 is about three times that of the Standard dataset (Supplementary Fig. S6d). The larger values of FWI in Class1 and Class2 are mostly located on either side of the salinity fronts (see Fig. S6b-c vs. Fig. 7c for the location of fronts), and we propose that the absence of the spatial bias correction in these datasets generates artificially high FWI. In some areas located far away

from salinity fronts or/and where data are numerous, the residual salinity is equivalent in Class1 and Class2 vs. the standard datasets (Supplementary Fig. S6).

The residual salinity data around Station 27 (47.5°N, 52.6°W) also present a relatively small residual difference with the Standard dataset (Fig. S6e-f). Thus, the detailed comparison of the simplified climatologies vs. the validation dataset at Station 27 (Figs. S7-S8) should represent the situation where these climatologies perform at their best. Although the salinity climatologies of Class1 (Fig. S7a) and Class2 (Fig. S8a) compare well with that of Station 27, the time series of anomalies present a much larger standard deviation (Fig. S7d-S8d). The RMS difference between the yearly averaged salinity anomaly of Class1 and Class2 and the validation dataset at Station 27 is 0.102 and 0.086, respectively, whereas it is only 0.055 for the standard dataset (see Fig. 6d). The interpretation of the interannual anomalies using Class1 or Class2 could be hindered by the significant differences with the actual anomalies (Figs. S7c, S8c). It is noteworthy that even at a location where data is widely available and the mapping schemes perform the best, the results of the simplified datasets are less reliable than those of the standard dataset.

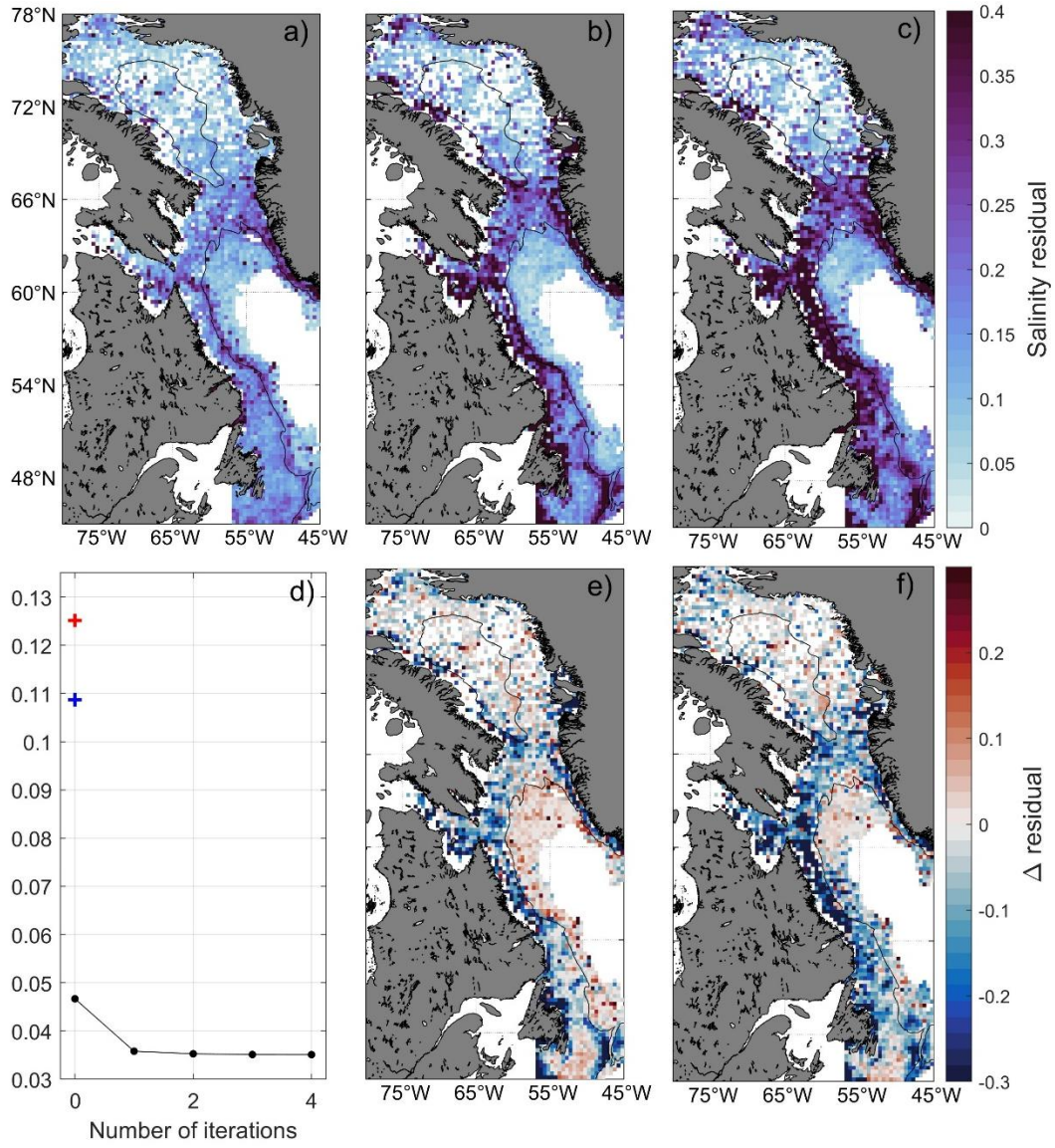


Fig. S6. Salinity residual (raw salinity minus climatology and biases) at the (a) 5th iteration of our standard dataset, (b) for the Class1 and (c) Class2 simplified climatologies. (d) Variance of the salinity residuals of the standard climatology as a function of the number of iterations. The blue and red crosses indicate the variance of the residuals for Class1 and Class2, respectively. (e) Difference between the data in (b) and (a), and (f) difference between the data in (c) and (a), and (f). Blue color indicates a lower residual for the standard dataset compared to the simplified climatologies.

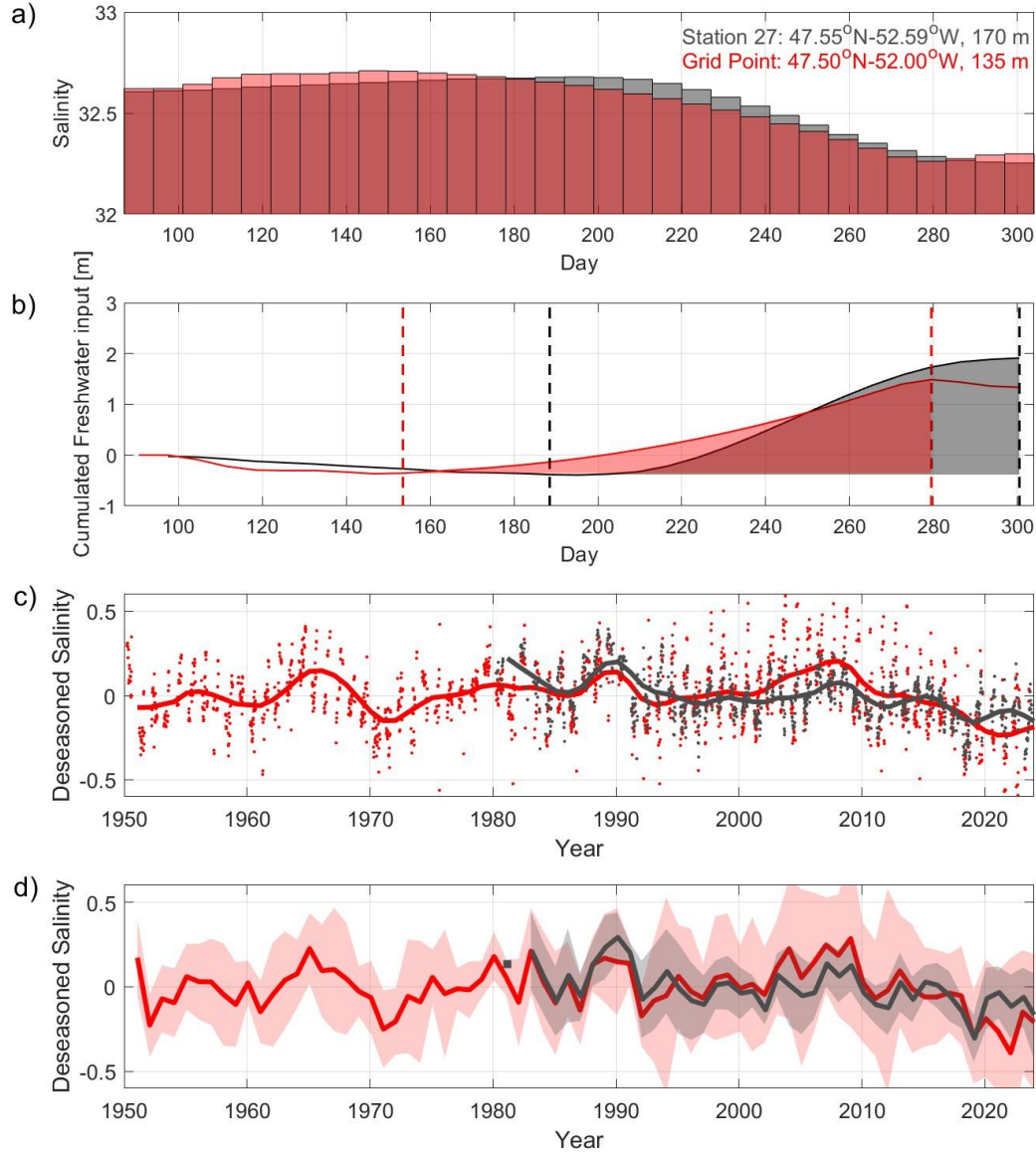


Fig. S7. (a) Salinity climatology at Station 27, estimated only from points acquired at this station obtained residual (gray bars), and simplified salinity climatology Class1 at the closest grid point with comparable bottom depth. (b) Cumulated freshwater input estimated from the salinity climatology at Station 27 (gray shade) and the corresponding grid point (red) for Class1. The gray and red dashed lines represent the beginning and end of the freshening season in each case. (c) Time series of weekly binned deseasoned salinity (dots) between 1950 and 2023 and their 5-year moving average value (thick solid line) at Station 27 (gray) and the corresponding grid point for Class1 (red). (d) Time series of yearly averaged deseasoned salinity (thick solid lines) with their two standard deviation envelopes (shading).

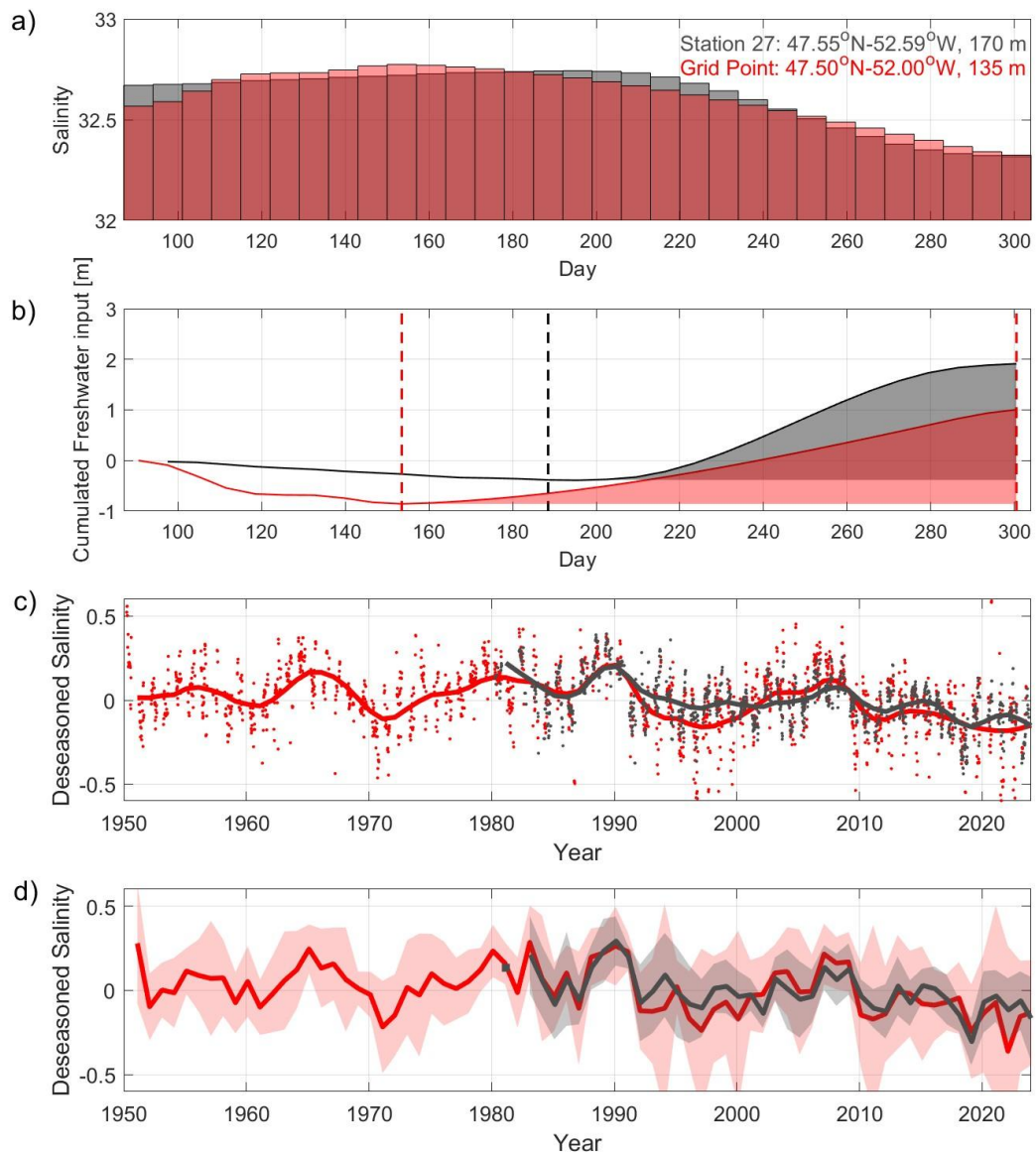


Fig. S8. Same as Fig. S7 but using Class2 (red lines and shading) as the comparison dataset.

Supplementary Tables TS1: Glacier melt and river runoff for each individual basins around Baffin Bay and the Labrador Shelf

Table TS1: Outputs of river runoff and glacier melt for each individual basin around Baffin Bay and the Labrador Shelf (see the basin limits in Fig. 1). NFL stands for Newfoundland and Labrador, BFN for Baffin Island, DVN for Devon Island, GNW for north-western Greenland, GCW for center-west Greenland, and GSW for southwestern Greenland. Daily values were integrated from day 90 to day 300, and averaged for the period 1965-2022.

Components	Freshwater volume (River runoff) (km ³)	Freshwater volume (Glacier melt) (km ³)
NFL	356.8	0.1
BFN	95.3	53.5
DVN	2.7	28.9
GNW	7.8	94.6
GCW	8.6	78.1
GSW	48.7	181.0
Total runoff	519.9	436.2