Influence of Barrier Wind Forcing on Heat Delivery Toward the Greenland Ice Sheet

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Abstract
A high-resolution numerical hydrodynamic model of Kangerdlugssuaq Fjord and the adjacent southeast Greenland shelf region was constructed in order to investigate the dynamics of fjord-shelf exchange. Recent studies have suggested that rapid exchange flows, driven by along-shelf barrier wind events, are the dominant agent of exchange between fjord and shelf. These events are prone to occur during the winter, when freshwater forcing is minimal and observations of the fjord interior are scarce. Subglacial freshwater discharge was held at zero, so that any buoyancy-driven overturning circulation was driven by melting alone. The model described a geostrophically balanced background flow transporting water masses between the fjord mouth and the glacier terminus, indicating that rotational effects are of order-one importance. Barrier wind events were found to trigger coastally trapped internal wave activity within fjord, temporarily enhancing exchange and vertical mixing, and causing warm water to oscillate in the along-fjord direction. These internal waves were also found to enhance the background flow via Stokes’ drift. Heat delivery through the fjord mouth was smaller than that recorded in summer observations, however the system is more effective at delivering this heat to the head of the fjord. There exists the potential for wintertime melting at the ice-ocean interface to be significant to the same order as summertime melting.

1. Introduction

1.1. Background
The glaciated fjords which permeate the Greenland coastline act as gateways for heat delivery from the ocean to the Greenland Ice Sheet (GrIS; Hanna et al., 2009; Inall et al., 2014; Rignot & Kanagaratnam, 2006; Straneo et al., 2010). Due to significant freshwater input, these regions are typically cooler and less saline than waters on the continental shelf. A fjord’s ability to deliver heat to the GrIS, or freshwater to the open ocean, is dependent on the fluid dynamics of the fjord interior. The circulation is governed in general by a variety of forcing mechanisms, which vary in importance both seasonally and between fjords due to differences in local setting. The dominant drivers of circulation in a Greenland fjord are freshwater input, wind stress, tides, sea ice formation/melting, and shelf exchange (Cottier et al., 2010). Freshwater enters the fjord via iceberg calving, direct melting at the glacier-ocean interface, surface runoff, and subglacial discharge (Cottier et al., 2010). In broad fjords (with respect to the internal Rossby radius of deformation, $L_R$), we expect the Coriolis effect to significantly influence the dynamics, giving rise to an across-fjord aspect to the circulation which would otherwise be negligible (Inall & Gillibrand, 2010).

Over the last two decades, the GrIS has undergone mass loss at an unprecedented rate, accounting for ~16% of global sea level rise during this period (Khan et al., 2014). Thinning of the ice sheet has been most pronounced near the coast, with many major outlet glaciers observed to accelerate and retreat over this period (Bevan et al., 2012; Nick et al., 2009; Rignot & Kanagaratnam, 2006). These changes to the GrIS coincide with an increase in water temperature around southern Greenland, indicating that they are triggered by oceanic heating (Hanna et al., 2009; Straneo & Heimbach, 2013). On the east coast, during 2000–2005, interannual changes in glacial dynamics are observed and equatorward of 69°N (Seale et al., 2011), which corresponds to a strong latitudinal gradient in ocean temperature, further evidence of ice-ocean interaction as the principal driver of GrIS reduction.
1.2. Objective

The objective of this paper is to better understand the wintertime dynamics of a large Greenlandic fjord and, by extension, large glaciated fjords in general. The study is focused on the interaction between fjord and shelf driven by along-shelf winds, and the additional effects in broad fjord systems due to the Earth’s rotation. We tackled this problem by designing a realistic numerical hydrodynamic model of the fjord and adjacent shelf region, allowing the heat delivery toward the glacier front to be quantified directly under a variety of wind forcing scenarios. The paper is structured as follows: the remainder of section 1 describes the oceanographic setting and outlines gaps in the current understanding, providing the rationale for conducting winter-specific modeling experiments. Section 2 details the model grid, bathymetry, initial conditions, and forcing fields, before explaining the experimental configuration. In section 3, the model output is presented, revealing a mean flow field strongly influenced by the Earth’s rotation in all model experiments. Wind forcing is shown to be linked with internal wave activity, which is seen to greatly impact along-fjord volume and heat exchange. Also, due to increased advection and vertical mixing, the water column structure in the fjord mouth is altered following wind forcing, with implications for subsequent exchange. In section 4, we interpret the dynamical response to wind forcing as coastal trapped wave activity and discuss the implications for heat delivery toward the GrIS during winter in the context of the existing literature on high-latitude fjord-shelf exchange. Our key outcomes are summarized in section 5.

1.3. Setting

Kangerdlugssuqq Glacier (KG) in southeast (SE) Greenland is the second largest outlet glacier of the GrIS by discharge (Enderlin et al., 2014) and drains around 4% of the GrIS by area (Bevan et al., 2012). KG drains into Kangerdlugssuqq Fjord (KF, Figure 1), which is around 75 km in length, 6 km wide and 900 m deep, and oriented at approximately 340° from north (Cowton et al., 2016; Inall et al., 2014; Sutherland et al., 2014a). The entrance of KF opens onto a relatively broad region of the SE Greenland continental shelf. A deep trough (Kangerdlugssuqq Trough, KT hereafter) crosses the shelf here, running southward from the KF entrance to intersect the shelf break at the northern boundary of the Irminger Basin. KT has a maximum depth of around 650 m, while the typical shelf depth is around 300 m. It is separated from KF by a relatively deep sill of around 480 m. In 2004–2005, KG experienced profound acceleration and retreat, before subsequently reverting to a stable state (Bevan et al., 2012). This sudden change has been linked to increased water temperatures around SE Greenland at that time (Hanna et al., 2009). In order to understand these changes in glacial dynamics, both in KG and throughout Greenland, it is therefore necessary to quantify the up-fjord heat flux associated with the various drivers of fjordic circulation.

Figure 1. Bathymetry of SE Greenland seas, with the locations of KF and SF indicated, along with the pathways of the IC, DSO, EGC, and EGCC and the model domain shown in black.
The SE Greenland shelf generally displays seasonal variation in water column structure. Polar Surface Water (PSW), a relatively cold, light water mass ($\Theta < 0^\circ C, \sigma_\theta < 27.70$; Sutherland & Pickart, 2008), occupies the upper layer. It is transported into the region via the East Greenland Current (EGC), a geostrophically balanced slope current carrying water southward from the Arctic Ocean along the east Greenland continental shelf break (Figure 1). A portion of the PSW may be warmed by the atmosphere and freshened by ice melt during transport, giving rise to the seasonal variant Warm Polar Surface Water (PSWw, $\Theta > 0^\circ C, \sigma_\theta < 27.70$; Inall et al., 2014). The EGC also carries the colder, deeper lying Polar Intermediate Water (PIW, $\Theta < 0^\circ C, \sigma_\theta > 27.70$) from the Arctic Ocean (Sutherland & Pickart, 2008). These three Arctic origin water masses, PSW, PSWw, and PIW, are collectively termed Polar Water (PW), and generally occupy the upper 75–300 m of the water column (Bacon et al., 2014). On the shelf is the East Greenland Coastal Current (EGCC), a second source of PW into the region (Bacon, 2002). This high speed, coastally steered surface current is thought to be highly seasonal (Bacon et al., 2014) due to the influence of freshwater from the eastern GrIS and increased wind forcing in winter. The EGCC is not observed upstream of the KF entrance and is thought to be fed, at least in part, by a branch of the EGC which is directed landward by KT before continuing southward along the coast (Sutherland & Pickart, 2008). The freshening influence of the KF system may also play a part in the establishment of the EGCC at the head of KT. Below PW lies Atlantic Water (AW, $\Theta \sim 4.5–6.5^\circ C, S_\alpha \sim 34.9–35.2$; Sutherland & Pickart, 2008), which is advected into the region via the Irminger Current (IC), a branch of the North Atlantic Current which circulates cyclonically around the boundary of the Irminger Basin after entering the basin west of the Reykjanes Ridge. The IC is the primary oceanic source of heat into the region. South of the Denmark Strait, the IC and EGC merge into a composite flow, denoted the IC/EGC front hereafter, consisting of PW overlaying AW and following the shelf break. The IC/EGC front is prone to meandering and eddy shedding (Magaldi et al., 2011), giving rise to variability in the water mass structure on shelf and, crucially, allowing AW to flow westward of the shelf break, such that water mass exchange between the Irminger Basin and the shelf is episodic in nature. In a SE Greenland regional model by Gelderloos et al. (2017), a branch of the IC is seen to circulate into KT at the shelf break while, in summer, a second branch is found to go north through the Denmark Strait and across the shelf toward KT, thus advecting AW toward the mouth of the KF. The densest water mass in the region is Denmark Strait Overflow Water (DSOW, $\Theta < 0^\circ C, 34.9 < S_\alpha < 35.2, \sigma_\theta \geq 27.8$). DSOW is released into the region in periodic boluses, which pass through the Denmark Strait and cascade down into Irminger Basin with a frequency of around 2–5 days (Koszalka et al., 2013). Modeling results suggest the DSOW can circulate in KT (Koszalka et al., 2013) and cross onto the shelf (Magaldi et al., 2011) and has been observed inside KF below sill depth (Inall et al., 2014).

Along-fjord temperature and salinity gradients are introduced by freshwater input at the terminus of KG and at 13 smaller outlet glaciers surrounding the fjord (Cowton et al., 2016; Inall et al., 2014). Inall et al. (2014) observe glacial melt water (GMW, $\sigma_\theta < 24.0$) in KF interior in September, alongside the aforementioned oceanic water masses. Although there is a dearth of observations from close to glacier termini, both modeling and theoretical studies (Cowton et al., 2015; Jenkins, 2011; Sciascia et al., 2013; Xu et al., 2012) suggest that stratification breaks down at the glacier terminus due to the buoyant plume dynamics associated with subglacial discharge and melting. In the classical picture of estuarine circulation, surface runoff at the glacier terminus drives an overturning cell which draws shelf water into the fjord via deep layer inflow. In the deep-silled fjords of SE Greenland, freshwater injected at depth forms a buoyant plume, and the entrainment of ambient fjord water can lead to a portion of this plume finding neutral buoyancy at the PW/AW interface rather than at the surface (Carroll et al., 2016; Cowton et al., 2016; Inall et al., 2014). Summer observations of KF confirm this, with outflow of GMW observed both at the surface and the pycnocline, and compensatory inflow in both the PW layer and the AW layer (Inall et al., 2014).

1.4. Exchange During Winter
An additional mechanism for the advection of heat into deep-silled fjords, termed intermediary circulation, has been recently identified in Sermilik Fjord (SF; Jackson et al., 2014; Straneo et al., 2010), a similarly sized neighbor to KF. Exchange flows are triggered by along-coast barrier wind events, which drive strong baroclinic exchange flows as a result of downwelling on the shelf. As the pycnocline on the shelf is initially depressed by the action of northerly coastal wind, inflow first occurs in the upper layer, with outflow in the lower, as the circulation within the fjord acts to match the stratification to that on the shelf. Once the wind relaxes, the pycnocline on the shelf upwells to a neutral position and the reverse process occurs, resulting
in inflow in the warm lower layer. The mechanism was first described by Klinck et al. (1981) and has been observed to generate rapid exchange flows between SF and the adjacent shelf (Straneo et al., 2010), and estimates of the volume transport associated with intermediary circulation are an order of magnitude greater than that from freshwater-driven circulation, both in SF and KF (Sutherland et al., 2014a). While these estimates place intermediary circulation as the most significant driver of exchange in the major SE Greenland fjords, uncertainties remain over the time-averaged contribution toward heat transport.

Barrier winds are strong, northeasterly winds occurring on the SE Greenland shelf, generally confined to the winter months (when freshwater runoff and subglacial discharge are minimal), occurring on average once per week during December–February (Harden et al., 2011). Wintertime hydrographic observations of the KF interior are limited to a single mooring location from winter 2009 to 2010 (Jackson et al., 2014), leading to large uncertainties in the extent to which intermediary circulation contributes toward melting at the KG terminus. Regional modeling results suggest that heat supply toward the KF entrance is greater in winter than in summer (Gelderloos et al., 2017).

There have been a number of recent numerical modeling studies aimed specifically at assessing the relative importance of wind-driven and freshwater-driven exchange, which typically employ idealized topography in order to maximize generality (Carroll et al., 2016, 2017; Cowton et al., 2015; Spall et al., 2017). There is, however, no firm consensus on the significance of heat delivery from the ocean to the GrIS during winter.

A realistic numerical modeling study of KF by Cowton et al. (2016), which simulated the effect of barrier winds on KF by altering the stratification at the fjord mouth, found that while intermediary circulation is effective at bringing shelf-resident water into the fjord interior, this water does not penetrate into the upper reaches of the fjord so as to influence glacial stability. While the result is highly illuminating, the horizontally uniform boundary fields fail to account for any cross-fjord variability in the fjord mouth. The circulation at the northern end of KT is influenced by rotational effects (Koszalka et al., 2013), and the nature of the interaction between barrier wind-induced downwelling and the dynamics in the fjord mouth remains poorly understood.

The large outlet fjords of SE Greenland have previously been presumed narrow in comparison to $L_r$ (Christoﬀersen et al., 2011; Sciascia et al., 2013), estimated by Sutherland et al. (2014a) to be $\sim$9 km, such that cross-fjord variability may be neglected. Recent results from KF and SF, however, demonstrate cross-fjord variability in the hydrography and circulation of these fjords (Inall et al., 2014; Sutherland et al., 2014b). This gives rise to the potential for the geostrophically balanced circulation in KT extending to the KF interior, acting as an uninterrupted pathway transporting heat between the continental shelf and KG.

2. Methods

A high-resolution modeling study was carried out in which the interaction of KF with the adjacent shelf seas was simulated under typical winter conditions. The model was forced using mean winter wind fields as a control case, and this was compared to simulations forced using wind fields designed to emulate typical barrier wind events. Glacial melt was simulated using a subgrid-scale parametrization, thus providing a heat sink and freshwater source at the ice-ocean interface. Sea ice was parametrized through alteration of the wind fields. This work used the ARCHER UK National Supercomputing Service (http://www.archer.ac.uk).

2.1. The MITgcm

Model simulations were run using The Massachusetts Institute of Technology general circulation model (MITgcm; Marshall et al., 1997), a highly versatile fluid dynamics simulator designed to operate on a wide range of scales in both the atmosphere and the ocean. It is built around solving the incompressible Boussinesq equations using the finite-volume method on a curvilinear coordinate system. These equations are in general nonhydrostatic, however the model may be configured to utilize the hydrostatic approximation, eliminating any vertical momentum within the fluid and simplifying the numerical integration. The grid spacing may be varied, or “telescoped,” throughout the model domain, both in the horizontal and the vertical, and partial depth, or “shaved,” cells help to deal with steep or complex bottom topography. A more comprehensive explanation of the inner workings of the MITgcm is given in Adcroft et al. (2004).
2.2. Grid
The model grid is of size $240 \times 240 \times 100$ cells ($lon \times lat \times z$). A telescopic grid was employed so that the relatively small-scale dynamics in and around KF could be captured in necessarily high resolution, while minimizing the number of computationally expensive grid cells allocated to the wider shelf sea region. The maximum resolution of approximately $360 \text{ m} \times 360 \text{ m}$ is used throughout KF in the northern part of the domain. Moving east or west (south) of this region, the zonal (meridional) grid spacing increases linearly until the domain boundary, such that the southwestern and southeasternmost cells are the largest, with a resolution of approximately $4 \text{ km} \times 4 \text{ km}$. The domain spans $66.38^\circ \text{N}–68.5^\circ \text{N}$, $34.59^\circ \text{W}–28.05^\circ \text{W}$ and $1,000 \text{ m}$ depth, with a constant vertical resolution of $10 \text{ m}$.

2.3. Bathymetry
Raw bathymetry data for the shelf region came from the International Bathymetric Chart of the Arctic Ocean (IBCAO), and had a resolution of 30 arc sec (the width of KT is around $1^\circ$). Although sufficiently detailed in the open ocean, this data set failed to properly capture the full depth and complex geometry of the KF interior. Swath bathymetry data gathered on the cruise JR106b to KF (Dowdeswell, 2004) were used in this region. The data were interpolated onto the model’s horizontal grid in the KF region, with the IBCAO data used everywhere else. The swath data did not cover the northernmost portion of KF, so a region of idealized bathymetry was created whereby a U-shaped cross section was replicated for eight cells along the fjord, adjacent to the domain’s northern boundary. The cells along the northern boundary itself were then mapped to zero in order to represent an idealized, vertical glacier front. The model bathymetry is shown in Figure 2.

2.4. Initial and Boundary Conditions
Open boundary conditions (OBCs) were generated using a larger domain MITgcm model. Barrier winds are thought to strongly influence stratification throughout the shelf sea region (Jackson et al., 2014) and are by no means localized around the KF/KT system (Harden et al., 2011). Barrier wind forcing on the model domain used here should not, therefore, be restricted to wind stress at the surface but also include changes in the OBCs. In order to tackle this, a larger domain was designed which was subject to the same atmospheric forcing as the main model so that, using a one-way nesting, output could be used to generate dynamic OBCs for the main model. The OBCs hence reflected changes in the water column symptomatic of the large scale atmospheric forcing. Tidal signal was not included in OBC forcing since tides in the region are weak ($\sim 1 \text{ cm s}^{-1}$; Inall et al., 2014). Free-slip conditions were assigned to all material side boundaries, with no-slip conditions used at the seabed. This choice is made due to the aspect ratio of the grid cells: side

Figure 2. (a) The main model bathymetry and (b) a zoom of the fjord interior. Cross-fjord sections are shown and numbered in yellow, the thalweg section is shown in dashed black. Section 1 is considered the upper fjord, sections 2 and 3 are termed midfjord, sections 4 and 5 are termed fjord mouth while section 6 crosses KT.
friction will have only a small effect on a 360 m diameter parcel of water, whereas bottom friction will play a much more significant role in the dynamics of a 10 m high column of water.

The larger domain is of size $144 \times 96 \times 130$ cells ($\text{lon} \times \text{lat} \times z$). Telescoping was used so that the region of interest, centered around KT, has a resolution of $4.5 \text{ km} \times 7.5 \text{ km}$ while, moving away from this region, the cell size increases linearly toward the boundary. This relatively coarse grid is not sufficient to capture KF but is able to capture the shelf dynamics in sufficient detail to provide boundary fields for the main model. The horizontal grid covers $59.56^\circ \text{N} - 70.39^\circ \text{N}$ and $44.4^\circ \text{W} - 8.16^\circ \text{W}$, and is designed such that the main model domain lies entirely within the high-resolution region. It is $4,000 \text{ m}$ deep, with the vertical resolution varying from $10 \text{ m}$ in the surface layer to $100 \text{ m}$ for the bottom $2,000 \text{ m}$. Bathymetry data came from GEBCO and initial temperature and salinity fields came from Polar Science Center Hydrographic Climatology (PHC3.0; updated from Steele et al. (2001)) monthly climatology data. From these data, December, January, February (DJF) winter mean temperature and salinity fields were generated. Initial velocities were zero everywhere, as was the free surface height anomaly, $\eta$. The initial conditions were held constant at the boundary in all large domain simulations. A sponge layer was used to relax points near the boundary toward these values, with a thickness of four cells and relaxation times $\tau_{\text{inner}} = 10 \text{ days}$ and $\tau_{\text{outer}} = 1 \text{ day}$. A balancing routine was used to ensure that the net flow across all boundaries summed to zero, with free-slip conditions used at all material boundaries. Atmospheric forcing fields (wind stress and net heat flux) were relaxed linearly to zero at the boundaries over a four cell wide zone. The larger domain was not designed to facilitate melting or runoff as there are no fjords or glaciers included and was therefore unable to capture the freshening influence of KG or any other of SE Greenland’s tidewater glaciers. Snapshots of $\Theta, S, u,$ and $v$ from day 50 onward were mapped to the main model grid, including KF interior, to provide initial and boundary conditions. This was done to ensure that the conditions on the shelf had settled into a balanced, dynamically consistent circulation pattern, with the IC and EGC well established, along with a geostrophically balanced circulation pattern established in KT. The initial water column inside the fjord closely matched that on the shelf, even below sill depth, due to the nearest-neighbor extrapolation method used.

### 2.5. Freshwater Input

Freshwater-driven circulation is an observed feature of KF (Inall et al., 2014), and of arctic fjords in general (Mortensen et al., 2011). This circulation is strongest in summer, when the fjord has a continual supply of freshwater due to glacier surface melt. Even during the winter months, however, glacier/iceberg melt water will drive some estuarine flow regime so long as sufficiently warm water is drawn into contact with the glacier front. It is therefore possible that when intermediary circulation is strongest, some freshwater-driven circulation is still active in the fjord. The interaction between these two circulation schemes is complex and likely nonlinear. It is therefore advantageous to incorporate a realistic freshwater parametrization into our numerical representation of the system, even though our study is focused primarily on shelf-driven circulation. The MITgcm “iceplume” package (Cowton et al., 2015) was developed to overcome the problem of subgridscale plume dynamics at the ice-ocean interface in high-latitude numerical models. The package allows selected coastal grid locations to be considered ice and will modify the conditions in the adjacent fluid cells so as to replicate the influence of a buoyant plume, as described by Jenkins (2011) although half-conical in shape. Subglacial discharge may be prescribed by the user, while melting is calculated as a function of adjacent water temperature and pressure (Jenkins, 2011). This scheme circumvents the need to resolve plume dynamics numerically, such that the model may be run in hydrostatic mode. In the model presented here, plume parametrization cells were placed along the northern boundary to the fjord, representing the KG glacier front. The model does not include freshwater input from any of the smaller output glaciers of the KF system nor does it account for the contribution due to melting of icebergs which, for deep-keeled icebergs, can be as significant as melting at the glacier terminus (Enderlin et al., 2016; Enderlin & Hamilton, 2014; Inall et al., 2014).

### 2.6. Atmospheric Forcing

Harden et al. (2011) identify two distinct locations where barrier wind events most commonly occur, Denmark Strait North (DSN; 67.7°N, 25.3°W) and Denmark Strait South (DSS; 64.9°N, 35.9°W). In order to characterize these events, they generated 96 h composite wind fields which represent a typical event at each location during DJF. These composite wind fields were used as forcing fields for the model presented here. Upon examining both fields, only DSN fields were selected for model input as they produced higher
produce a dimensionless coefficient capturing the dilation effect. The 2 by changing the coefficient of the exponential term from 0.07 to 0.085. The following expression is there-

tic Ocean, Martin et al. (2014) find that the momentum transfer between air and ocean, words, we expect both a dilation and rotation of wind stress fields. In a modeling study of sea ice in the Arc-

2.7. Sea Ice

The Kangerdlugssuaq region is subject to seasonal ice cover. Barrier winds are a predominantly winter phe-

nomenon, and sea ice greatly influences the interaction between the wind and the ocean (Martin et al., 2014; Rabinovich et al., 2007; Wadhams, 2000). MITgcm contains a package in which ice can grow and melt at the free surface, and flow subject to drag forces from both the ocean and the atmosphere. The package requires, as both initial and boundary conditions, sea ice thickness, snow thickness, sea ice coverage, and velocity and is designed to work alongside prescribed precipitation. This was deemed excessive for the study presented here, which is not designed to study sea ice dynamics. The MITgcm sea ice package was therefore eschewed in favor of a simpler approach, whereby sea ice is represented only through its effect on wind forcing.

Monthly mean sea ice coverage data were obtained from the National Snow and Ice Data Centre (NSIDC), ranging 1979–2014. From these fields, the DJF winter climatology fields for SE Greenland were generated using only data from 1989 to 2008, the time interval from which the composite wind stress fields were taken. These data were then interpolated onto both the large model grid and the main model grid. The wind stress fields were then modified to reflect the presence of sea ice. The effect of sea ice on wind stress is tensorial: the magnitude of momentum transfer is altered due to the differing drag coefficients of ice and water, while the direction changes due to an additional Ekman-like effect from the layer of ice. In other words, we expect both a dilation and rotation of wind stress fields. In a modeling study of sea ice in the Arctic Ocean, Martin et al. (2014) find that the momentum transfer between air and ocean, \( \tau_{ao} \), initially increases with sea ice concentration, \( c \), peaking around \( c = 0.8 \) before dropping off rapidly. They describe this behavior using the simple model:

\[
\tau_{ao}(c) = 0.05c - 0.07e^{-20(1-c)} + 0.035 
\]

In the fjordic regions of SE Greenland, we presume that the momentum transfer is effectively zero where sea ice concentration \( c = 1 \). This is because, unlike in the Arctic Ocean, \( c = 1 \) here indicates fast ice, which imposes a rigid lid. To account for this, equation (1) was altered such that \( \tau_{ao}(1) = 0 \). This may be achieved by changing the coefficient of the exponential term from 0.07 to 0.085. The following expression is therefore used:

\[
\tau_{ao}(c) = 0.05c - 0.085e^{-20(1-c)} + 0.035
\]

This matches the original closely, differing only when the exponential term begins to dominate at large \( c \). It was further assumed that the maximum deflection to the right, \( \phi_{max} \), would coincide with maximum ice cover, whereas no direction change is expected in open water. We used a linear sliding scale between these two extremes to prescribe the rotation as a function of \( c \). All wind stress fields were therefore altered to account for sea ice cover using the formula \( \tau' = \Gamma(c) \cdot \tau \) with

\[
\Gamma(c) = \frac{\tau_{ao}(c)}{\tau_{ao}(0)} \begin{pmatrix} \cos(c\phi_{max}) & \sin(c\phi_{max}) \\ -\sin(c\phi_{max}) & \cos(c\phi_{max}) \end{pmatrix}
\]

where \( \tau_{ao}(c) \) is the expression given in equation (2), normalized here by the open water value, \( \tau_{ao}(0) \), to produce a dimensionless coefficient capturing the dilation effect. The \( 2 \times 2 \) matrix captures the rotation. The value \( \phi_{max} = 25^\circ \) was used as the maximum angle of deflection, taken from Wadhams (2000). This value was taken from observations in the open ocean, and subsequent research (Rabinovich et al., 2007) has indicated that in coastal regions the value may be lower. The maximum angle of deflection, \( \phi_{max} \), is never achieved in practice as this occurs when \( c = 1 \) and therefore when \( \tau' = 0 \). Inside the fjord, \( c = 1 \) at all times
and in all simulations, so that direct wind forcing on the KF interior is eliminated and the effects of the barrier winds are felt only through changes on the shelf. Air-sea heat flux data from ERA-Interim accounts for local ice cover, so no modification of those fields was required.

2.8. Subgridscale Mixing
The $\kappa$-Profile Parameterization (KPP; Large et al., 1994) is a subgridscale vertical mixing parametrization available as a package for MITgcm and was implemented in all simulations presented in this paper. KPP is comprised of several mixing schemes, each representing a distinct mechanism for mixing. It treats mixing in the surface boundary layer and mixing in the ocean interior as separate problems with the boundary layer depth calculated at each grid location as a function of the bulk Richardson number. Boundary layer mixing is dependent on surface forcing fields while, in the ocean interior, shear-driven mixing is calculated as a function of the local Richardson number, with both the vertical mixing coefficient and its derivative constrained by continuity at the interface.

In the horizontal, Leith biharmonic viscosity was utilized (Leith, 1996). This regime is proportional to the relative vorticity gradient and dependent on local grid resolution, with scaling according to Griffies and Hallberg (2000).

2.9. Experimental Design
A crucial problem in designing the experiments was cultivating conditions within the fjord representative of wintertime, due to the paucity of observations from nonsummer months. Prior to experimental runs, the model was subjected to a 100 day spin-up period with mean wind DJF forcing held constant. Subglacial discharge of 100 m$^3$ s$^{-1}$ (an order of magnitude smaller than estimates by Sole et al. (2012) and Cowton et al. (2016) for peak summer runoff into the entire KG basin) was simulated at the fjord’s northern boundary for 60 days, long enough establish quasi-stable buoyancy-driven overturning and realistic along-fjord temperature gradients. The fjord was then left to adjust for the remaining 40 days in the absence of subglacial discharge. From this state, the model was then further integrated in three distinct configurations. In the control case (Run-0), the prevailing DJF wind conditions were held constant throughout. This was compared to two alternative setups, designed to reflect (a) a single DSN wind event (Run-1) and (b) two wind events in succession (Run-2). In this way, we were able to isolate the effect of a single barrier wind event, and also to gauge the cumulative effect of successive winds events. Both Run-1 and Run-2 were subject to a 4 day wind event beginning at the start of the integration, while in Run-2 a second wind event was imposed at 10 days. All three runs ended after 25 days. The model runs are detailed in Table 1.

### Table 1

<table>
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<th>Day</th>
<th>Run-0</th>
<th>Run-1</th>
<th>Run-2</th>
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<td>Prevailing wind</td>
<td>Prevailing wind</td>
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3. Results and Analysis

3.1. Initial State
Figure 3a shows the sea surface temperature field after the 100 day spin-up period, which was used as the starting point for all subsequent experimental runs, nested within the high-resolution portion of the larger domain. The IC can be seen circulating around the head of the Irminger Basin, with a branch circulating north into the Denmark Strait. The eastern edge of KT appears to represent a boundary in surface water temperature, with warmer water residing to the west and a particularly strong temperature front between KT and the area of shallower topography to the east. The supply of heat along KT toward the KF mouth region is not spatially uniform, with pulses of anomalously warm water ($\Theta > 2.5^\circ$C) interspersed with cooler regions ($\Theta \approx 2^\circ$C). Surface temperature and salinity generally decreases approaching the coast...
indicating that the model has captured the EGCC, as was confirmed by taking velocity sections normal to the coastline (not shown).

Temperature and salinity sections along the thalweg of the combined KF/KT system (Figures 3b and 3c) reveal a stark difference in stratification structure between fjord and shelf. In KT, the upper 150 m is well mixed. The warmest waters are found at around 300 m, though stratification is relatively weak in both temperature and salinity. In contrast, the upper 200 m of the water column inside KF is characterized by a sharp vertical salinity gradient and a negative vertical temperature gradient. This warm layer is thickest and warmest in the fjord mouth. As a result of the imposed subglacial discharge and any melting at the KG terminus during the spin-up period, the fjord interior has become strongly salt stratified in the surface layer, with the pycnocline shallower and sharper than in KT. Cold, dense waters are found isolated below sill depth, both in KF and in the deeper, central region of KT. There is no clear evidence of the four-layer overturning circulation scheme observed by Inall et al. (2014), indicating that, after switching off subglacial discharge, melting alone was unable to sustain such a vertical hydrography structure. The temperature structure in the mid-fjord is comparable with that observed by Jackson et al. (2014) during winter 2009, where the temperature maximum was often located at 200–300 m. The small depth range observed by Jackson et al. (2014), along with the high variability in thermocline height recorded within that range, make it difficult to validate the model using this data set, but the general structure is consistent.

Viewing the model’s water properties in T-S space (Figure 4) sheds further light on the hydrographic structure of the model domain. We include profiles, taken from the larger model, in the Denmark Strait, the Irminger Current and on the shelf northeast of the main domain, which we take to represent the source waters to the main model. This helps illustrate how the relative influence of each of these three source waters, the heaviest (DSOW), warmest (AW), and freshest (PW) inflowing waters, respectively, dictates the conditions at a given location. In the upper layer of the fjord interior the freshening effect of the plume

Figure 3. State of model after the 100 day spin-up period. Sea surface conservative temperature from the main model, nested within the larger model, is shown in (a), with the thalweg marked by a dashed black line. (b) Vertical profiles in conservative temperature, absolute salinity, and σt from the deepest point in KF (section 3, solid lines) are compared to corresponding profiles taken from the deepest point in KT (section 6, dashed lines). Conservative temperature and absolute salinity section taken along the thalweg shown in (c), with the locations of the standard cross sections 1–6 shown in dashed yellow. Note the different temperature scale from that used in Figure 3a.
parametrization at the head of the fjord is clearly seen, matching the
gradient of the mixing line between AW and GMW. This may be con-
sidered another water mass source, albeit supplied via glacial modifi-
cation rather than advection, since the freshest waters toward the
head of the fjord lie outside the convex hull of the three source water
masses (red, blue, and green in Figure 4). The along-fjord temperature
gradient in the warm layer is highlighted by selecting profiles from
the fjord mouth and near the KG terminus. The presence of PW is also
greater in the mouth, where we see a straight mixing line running
from the temperature maximum toward the PW source, than near KG,
where all the water in the upper layer falls between the temperature
maximum and the freshest waters on the surface. The surface water in
the mouth is slightly warmer than the PW directly below it due to
upward heat transport by the plume dynamics encapsulated in the
iceplume package, and subsequent estuarine outflow. Surface waters
in KT have similar properties to intermediate depth waters inside KF.
PSWw, as found in KF by Inall et al. (2014) and Sutherland et al.
(2014a), is cooler and less abundant in our model due to the pre-
scribed winter conditions, and without injected freshwater the model
does not generate surface waters as fresh and light as those observed
in summer. We also note that deepest waters in KT are marginally
heavier than those inside the fjord, indicating that a deep water
renewal event could occur were deep layer (below sill depth)
exchange permitted.

3.2. Mean Flow

The normal component of the mean flow (time averaged over the full
25 day run) through six cross sections (Figure 2) for Run-0 is shown in
Figure 5. The circulation displays horizontal variability in all sections
with inflow (outflow) on the right (left), a behavior only permissible in
broad fjords. We calculated \( L_K \) to be \( \sim 3 \) km in the midfjord (section 2), firmly placing KF as a broad fjord
under the stratification conditions in the model. In KT (section 6), the outflow is strongly barotropic while
the inflow intensifies with depth such that strong currents of up to 20 cm s\(^{-1}\) sit against the eastern flank.
In the mouth of KF (sections 4 and 5), the lower layer circulation structure matches that of KT, with an
inflowing current core residing against the eastern boundary below 200 m. In the KF interior (sections 1–3),
we see a strong cross-fjord velocity shear throughout most of the water column, with inflow on the right,
with the surface dominated by outflow in the upper and midfjord (sections 1 and 2).

Corresponding geostrophic velocities for Run-0 are shown in Figure 6. The vertical shear structure was
obtained for each density profile pair via the thermal wind relation and each velocity profile was then refer-
cenced using sea surface slope. These fields show excellent agreement, in circulation strength and structure,
with those obtained from dynamical output variables, indicating that the circulation in both KT and KF is
typically in geostrophic balance (as is later confirmed via direct evaluation of the momentum terms). The
sloping isopycnals in the mouth (sections 4 and 5) and, in particular, KT (section 6), indicate that the outflow
is driven by a barotropic pressure gradient, and the lower layer inflow by thermal wind.

In Run-2, which was subject to two barrier wind events, the mean circulation was more intense that from
Run-0, and with near identical structure (Figure 7). The corresponding geostrophic velocities (not shown)
are once again in excellent agreement with model section-normal velocities. The Run-1 mean flow (not
shown) has similar structure again, and sits between Run-0 and Run-2 in intensity. The action of barrier
winds therefore appears to strengthen preexisting background circulation both in KF and KT. This effect is
later shown to be caused by Stokes’ drift (Stokes, 1847).

We quantified the exchange through each section by considering the volume flux due to the inflow only,
which is valid so long as there is no net volume transport. This definition is therefore imprecise in sections 5
and 6, which are not closed, however it does provide values for the northward transport through KT, the

Figure 4. State of model after the 100 day spin-up period in T-S space. The
data from the fjord interior, as well as two profiles taken in the near the fjord
entrance and near the glacier terminus, are highlighted. Water mass definitions
are indicated in black text. To indicate the properties of the source waters to
the region, single profiles were also taken from the larger domain from the
Irminger Current west of Iceland, the Denmark Strait saddle point, and the shelf
north of KT.
predominant pathway for IC water into the mouth of KF. Exchange was found to increase with barrier wind forcing. For example, the mean exchange through the fjord mouth (section 4) was 0.14, 0.25, and 0.30 Sv in Run-0, Run-1, and Run-2, respectively. The mean exchange values for all sections are listed in Table 2.

3.3. Temporal Variability

Figure 8 shows time series of the exchange through KT (section 6), the fjord mouth (section 4), and the mid-fjord (section 2) from each of the three simulations. The exchange through KT is relatively steady, at around 1 Sv. Barrier winds are seen to enhance the exchange there but changes are not highly significant compared to the synoptic variability in the control run. The initial barrier wind event in Run-1 caused an approximate trebling of the exchange through the fjord mouth, ramping up and down over around 2 days, before increasing again to around twice the background value. The Run-1 signal reverted back to closely match the Run-0 signal after around 10 days. In the midfjord (Figure 8a) the signal due to the same barrier wind event was manifested as a sinusoidal signal, with three pronounced peaks reaching around four times the background value before dying away. By splitting the signal into northward transport both above and below 250 m, it is evident that the peaks in exchange correspond to times when upper layer velocities opposed those in the lower layer, a pattern which alternated in time. After the oscillations finished, the Run-1 exchange signal remained around double the background signal until around day 18, consistent with the intensification of the mean flow with increasing barrier wind activity. The second wind event in Run-2 had a similar initial effect on exchange in both locations, driving a spike in exchange through the mouth followed by a sustained enhanced exchange, while driving an oscillating exchange pattern in the midfjord. However, at later times in Run-1 and, to a greater extent, Run-2, the exchange through the fjord entrance (section 4) was found to greatly exceed the background (Run-0) value (caused, as we later discuss in more detail, by dense shelf waters spilling over the sill and cascading down into the fjord bottom). The oscillating signal reemerged in the midfjord (section 2) in Run-2, but was neither as pronounced or sustained as following the signal due to the initial wind event.
We looked to better understand the oscillating nature of the exchange in the midfjord via direct evaluation of the terms in the momentum equations. Figure 9 shows time series of the pressure gradient, Coriolis force and advective terms from the core of the mean inflow in the midfjord (section 2) for both Run-0 and Run-1. The viscous term was omitted as it was smaller by more than two orders of magnitude. In Run-0, the lateral pressure gradient (Figure 9a) demonstrates stochastic variability, fluctuating around the mean value and sometimes changing sign, and is consistently balanced by the Coriolis force, further confirmation that the modeled circulation within KF was geostrophically balanced. The wind event in Run-1 excited a large oscillation in the pressure gradient with period 3.5 days, twice that of the exchange signal. Again, the Coriolis term always opposed to the pressure gradient, even under sign reversal, indicating that this low-frequency oscillation was slow enough that geostrophic balance was maintained. We see a tendency for growth of the lateral pressure gradient (and Coriolis force) superimposed onto the oscillating signal, such that the system does not revert to the Run-0 case after the wind-induced oscillation dies away. This increase is consistent with the stronger mean circulatory flow found in model runs subject to wind forcing. The advective term is small in comparison throughout. We also see an oscillation emerge in the along-fjord pressure gradient after wind forcing (Figure 9b) which is around one-quarter the amplitude of the lateral oscillation and a quarter-period out of phase, such that the two components together describe an elliptical trajectory. The along-fjord Coriolis term is also small, as lateral velocities are generally small, and there is less clear symmetry with the pressure gradient term. Advection is of comparable magnitude to the other two terms in the along-fjord component and, from around day 7 onward, appears to be driven by the pressure gradient. This suggests that geostrophically balanced wave motions act to enhance along-fjord advection.

The oscillation of the pressure gradient corresponded with a wave in the halocline (here we used the 34.7 g/kg isohaline as a proxy) with an amplitude of around 70 m at the eastern boundary (30 m at the western boundary) in the midfjord (section 3), dissipating to 40 m at the eastern boundary of the upper fjord (section 1). The first peak took 10 h to propagate from section 4 to section 3 (15 km), 15 h from section 3 to section 2 (22.5 km), and 30 h from section 2 to section 1 (45 km).
3 to section 2 (21 km), and 15 h from section 2 to section 1 (17 km) giving propagation speeds of approximately 42, 39, and 32 cm s\(^{-1}\), respectively. Figure 10 shows the state of the thalweg section in Run-1 compared to Run-0 after 9 days of simulation. While internal waves are visible in the isohalines in both plots, wave amplitudes are greater in Run-1. Strikingly, in the aftermath of barrier wind forcing, cold, dense waters are found immediately outside KF, forming a sharp temperature and salinity front between the fjord mouth and the fjord interior.

We computed the most significant empirical orthogonal functions (EOFs) of the section 2 velocity field for each model run (Figures 11a–11c). EOF-1 accounts for 35%, 52%, and 59% of the variance in Run-0, Run-1, and Run-2, respectively. In all cases, EOF-1 is dominated by vertical shear with the upper layer velocity opposing that in the lower layer and appears to reflect the first normal mode (or baroclinic mode) of oscillation. However, some cross-fjord structure is present in EOF-1, with greater variability against the eastern side than against the western side (where internal wave amplitudes are greater).

We found a strong link between barrier wind forcing and the temporal patterns in the EOF-1 coefficient. Figure 11d shows time series of the EOF-1 coefficients for each run. The different model output fields project onto slightly different eigenbases, however the three EOF states appear similar enough that we proceed under the assumption that they are all broadly representative of a first normal mode. When information of the barrier wind event, which took place on the shelf over the first 4 days and peaked on day 2, propagates into the midfjord region it sets up an oscillation in the EOF-1 coefficient. Consistent with Figure 8a and the response in SF described by Straneo et al. (2010), we initially see inflow in the upper layer and outflow at depth, before the behavior reverses. Reconstructed EOF-1 velocities peak

![Figure 7. Normal component of mean flow through each section for Run-2. Section 6 has a separate velocity scale as current speeds were higher there. Isotherms are denoted by solid black contours. All sections are viewed looking up-fjord with red (blue) denoting fjord into (out of) the page.](image)

Table 2

<table>
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<tr>
<th>Section no.</th>
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<th>Run-2</th>
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</tr>
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around 25 cm s$^{-1}$ in the upper layer. The system then oscillates with period 3.5 days, corresponding to the waves in the pycnocline, with extrema in EOF-1 coefficient corresponding to maxima in exchange (Figures 8 and 11). On day 13, we see the influence of the second barrier wind event on Run-2, manifested once again as an oscillation in the EOF-1 coefficient, this time with marginally lesser amplitude. In each case, the excitation dies away around 8–10 days after the initial divergence from the control run, 10–12 days after peak wind stress on the shelf.

As EOFs are purely statistical in nature and have no inherent connection to physical processes, we follow the common approach (e.g., Sutherland & Straneo, 2012) of directly comparing basis EOFs to normal modes, which are calculated from linear profiles in $N$, the buoyancy frequency. For this comparison to be meaningful requires that the horizontal variability in the EOFs is small in comparison to the vertical variability. While EOF-1 meets this criterion, higher EOFs have a more complex structure with significant spatial variability both vertically and horizontally (Figures 12b–12d). By horizontally collapsing the velocity field at section 2 in Run-2 onto a single, time-dependent profile, we proceeded to calculate one-dimensional EOFs and compared them to the time average of the first two normal modes calculated from the corresponding density profiles. The first EOF profile is consistent with the first normal mode in that we have velocities in the upper layer opposed to those in the lower layer, with a zero crossing at around 300 m depth (Figure 12e). This gives confidence that EOF-1 is dynamically meaningful and representative of first-order baroclinic oscillations in KF.

The propagation speed of a first-order internal wave based on the initial density structure at section 2 is 40 cm s$^{-1}$, highly consistent with the estimates based on time lags between cross sections. At later times this value does decrease slightly as the stratification is weakened via shear-driven vertical mixing. We note that dynamic modes calculated here are not necessarily fully correct for the type of wave we see since the

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**Figure 8.** Time series of exchange through (a) section 2, (b) section 4, and (c) section 6. In Figure 8a, the up-fjord volume transport in the upper layer (above 250 m, dotted red line) and the lower layer (below 250 m, dashed line) for Run-2 are also shown in red. The plot at the top of the figure shows the northeasterly component of wind stress (Pa) over the shelf. The grayed-out regions denote the periods during which barrier wind forcing was switched on.
dynamic mode theory used assumes an infinite flat bottom. Nonetheless, they offer useful insight and, as will be seen in the discussion, have the equivalent dispersion relation to the internal waves observed.

### 3.4. Heat Delivery

The temperature contours in Figure 5 show that in Run-0 the warm layer at the head of KF (sections 1 and 2) was generally slightly thicker on the eastern side. This side of the fjord corresponds with the inflowing

![Figure 9](image.png)

**Figure 9.** Time series of (a) across-fjord and (b) along-fjord components of the pressure gradient, Coriolis and advection terms of the momentum equation taken at the mean inflowing core in section 2. Dashed lines indicate Run-0 while solid lines indicate Run-1. Note the differing y axis scaling between the two plots. The grayed-out region denotes the period during which barrier wind forcing was switched on.

![Figure 10](image.png)

**Figure 10.** Conservative temperature and absolute salinity along the thalweg section in (a) Run-0 and (b) Run-1 taken after 9 days of simulation. Labels at the top of the figure reflect how each section is referred to in the text.
mean velocities, indicating a positive net heat flux through fjordic sections due to the background circulation. However, at the mouth of KF and in KT (sections 4–6) the outflow is warmer than the inflow, driving a net cooling of the waters outside the KF mouth. We see a similar pattern in Run-2 (Figure 7).

Heat flux calculations through closed sections of KF (sections 1–4) were carried out on the native model grid, eliminating interpolation error. They were chosen to be equivalent to those shown in Figure 2, but oriented east-west rather than normal to the thalweg. Heat flux through sections 5 and 6 were calculated using the thalweg-normal sections shown in Figure 2.

In Run-0, the northward heat transport through KT (section 6) is highly variable on a time scale of around 10 days, with a mean heat flux of 0.47 TW, and maximum and minimum values of 3.1 and $-2.1$ TW, respectively. This is consistent with the results of Koszalka et al. (2013), who find that exchange between ocean and shelf in this region is episodic in nature. Negative mean values were obtained in the fjord mouth area, with $-0.27$ and $-0.01$ TW at sections 5 and 4. The negative section 4 result is somewhat misleading, however, as the time-integrated value is highly dependent on the integral limits. An increased presence of cold water in the last 2 days of simulation diminished the value greatly, and the mean flux discounting this period was 0.01 TW. The peak value through section 4 was 0.22 TW. We obtained positive time-averaged heat fluxes in the fjord interior, with $-0.06$ to $0.10$ TW in the midfjord (section 2).

Figure 13 shows the heat flux anomaly toward KG in KT (section 6), the fjord mouth (section 4), and the mid-fjord (section 2) attributable to barrier wind forcing. The anomaly due to the first (second) wind event was calculated by subtracting the Run-0 (Run-1) value from the Run-1 (Run-2) value. Barrier wind forcing boosted the northward heat delivery through KT, with two large spikes, each representative of an additional $-5$ TW, occurring immediately after the first wind event and two smaller spikes ($-2$ TW) after the second. In
KF and the fjord mouth (Figure 13), we see the expected behavior after the first wind event, whereby the barrier wind event induced an initially negative heat flux anomaly due to the shoreward velocities in the upper layer. This was followed by an oscillating signal featuring two distinct peaks. Once the oscillatory signal dies, the Run-1 heat flux displays a sustained negative signal at the fjord mouth (section 4). The mean heat flux toward KG through each section in the first 12 days of Run-0 and Run-1 is shown in Table 3. We chose to integrate up to 12 days to ensure that we captured the entire duration of the dynamical response to wind forcing: after this time, the oscillating EOF-1 signal in Run-1 had been damped out and closely matched that in Run-0 (Figure 11d). In cross sections of the fjord interior (sections 1–3), the mean value is found to be 47% (0.0006 TW), 132% (0.010 TW), and 97% (0.014 TW) greater than in the control run, respectively.

Figure 13 also shows the heat flux anomaly attributable to the second barrier wind event in Run-2. In the fjord mouth (section 4), the initial behavior is similar to that induced by the first wind event. However, this oscillation dies after a single positive peak, as opposed to the two peaks from the first case. In the midfjord (section 2), the anomaly is positive initially, unlike the anomaly due to the first wind event, reaching 0.1 TW before later dropping to ~0.2 TW, with no clear oscillating pattern. The heat flux though section 4 was found to be positive in the control run in the aftermath of the second wind event (Table 4). This signal is negative in Run-1 and Run-2, as seen in Figure 13b. This is explained by Figure 14 which shows the along-fjord temperature and $\sigma_0$ distribution in all simulations after 22 days, 12 days after the onset of the second wind event. The extent of warm water in the fjord mouth, as well as the thickness of the warm layer inside KF, has been greatly diminished by barrier wind forcing. The density contours above the sill have risen up such that the cross-sectional area for dense water ($\sigma_0 > 27.725$) exchange has increased. Cold, dense waters have become abundant in KT in all three model runs, however in Run-2 and, to a lesser extent, Run-1, these waters were able to cross the sill and cascade down into the bottom of KF. The mean heat flux through all sections in Run-2, Run-1 and Run-0 between day 10 and day 22 is shown in Table 4. The values in KF and the fjord mouth were greatly diminished in Run-2 compared to Run-1, indicating that a second wind event triggered a greater influx of cold, dense waters to KF. Furthermore, the values were systematically lower in

![Figure 12](image-url)
Run-1 than Run-0 during this period, even though the two runs were subject to identical wind stress during this time. This means that the capacity for fjord-shelf exchange remains altered as a consequence of previous wind events long after any dynamical response has dissipated.

3.5. Vertical Mixing

We can better understand the enduring effects of barrier wind forcing on the stratification structure in KF by studying the vertical mixing output from the KPP package. Figure 15 shows the spatial mean of the vertical mixing coefficient, \( \kappa_z \), in the fjord mouth, taken from cells in the halocline (34.7 < \( S_a < 34.8 \)), and which were over 100 m from both the surface and the seabed. The signal is therefore representative of Richardson number dependent, shear-driven mixing between layers without the direct effects of the wind or bottom friction. In Run-1, we see an increase in \( \kappa_z \) during the initial barrier wind event in the first 4 days, initially displaying an oscillatory signal similar to that seen in the midfjord exchange plot (Figure 8a). Enhanced vertical mixing, with respect to Run-0, continues until the end of the simulation, peaking around day 17. In Run-2, we again see the \( \kappa_z \) signal increase again immediately following the second wind event, although any oscillating signal is not as pronounced, peaking on day 21. In Run-1 and Run-2, the strong mixing toward the end of the simulation is likely driven not by internal waves but by dense waters cascading over the sill and spreading along the fjord bottom as a gravity current (as seen in Figure 14).

4. Discussion

4.1. Background Circulation

That the mean flow through each section of the fjord interior displays horizontal shear, with inflow on the right and outflow on the left, is
symptomatic a fjord broader than $L_R$ and with dynamics hence influenced by the earth’s rotation. Furthermore, the good agreement between the mean flow field and the mean geostrophic flow in the KF interior (Figures 5 and 6), together with symmetry between the Coriolis and pressure gradient terms in the momentum equation (Figure 9), shows that the circulation in this region is typically in geostrophic balance. While previous studies of KF have generally focused on overturning circulation driven by either freshwater input or shelf exchange (Cowton et al., 2016; Jackson et al., 2014; Sutherland et al., 2014a), the cross-fjord variability in our model is the most striking feature of the mean flow and, in terms of exchange, is more significant than any vertically sheared regime. The study by Cowton et al. (2016), which focused solely on the fjord interior, did not report significant cross-fjord velocity variability in their modeling study of KF. This suggests that such circulation may be generated as an extension of the geostrophic circulation around the head of KT. The mean flow through section 5 in Run-0 (Figure 5) shows that this connectivity between the KT and KF circulation is strongest in the lower layer. Our model suggests that the fjord mouth is a complex, dynamically three-dimensional region, the circulation and hydrography of which have order-one consequences on the dynamics of the interior. By prescribing horizontally uniform boundary forcing at the fjord mouth, Cowton et al. (2016) may therefore have failed to capture this interaction. As the internal Rossby radius increases with the local stratification, it is also possible that the winter conditions in our model allowed for rotational effects to become significant, while the highly stratified summer conditions used by Cowton et al. (2016) did not. The geostrophically balanced circulation scheme observed in the model requires no external driver, such as wind or freshwater, and will continue to deliver heat up-fjord so long as a positive temperature gradient between the head of the fjord and the entrance is sustained.

<table>
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<th>Section no.</th>
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Table 4: Mean Heat Flux (TW) Toward KG Through Each Section During Day 10–22 of Simulation in Run-0, Run-1, and Run-2.

Figure 14. Conservative temperature and $\sigma_0$ along the thalweg section in (a) Run-0, (b) Run-1, and (c) Run-2 taken after 22 days of simulation.
Our value of \( \sim 0.01 \) TW through the fjord mouth is two orders of magnitude smaller than the 1 TW value measured there during summer 2009 (Sutherland et al., 2014a), and even our peak value of 0.22 TW was significantly smaller. As we move into the fjord, however, the agreement between our model and the existing literature grows stronger. We obtained a value of 0.01 TW in the midfjord, compared with values of 0.19 TW (Sutherland et al., 2014a) and 0.26 TW (Inall et al., 2014) from observation, while the time-averaged heat flux value of 0.0018 TW through section 1 in Run-0 is in excellent agreement with the value of 0.003 TW given by Sutherland et al. (2014a) for the same location. The fact that synoptic heat flux measurements from summer systematically exceeded our maximum modeled values indicates that the capacity for large heat delivery from the shelf to KF is significantly reduced in winter. However, the good agreement near the fjord head indicates that, in winter, heat is delivered from the midfjord to the glacier terminus more efficiently than in summer. We postulate that while the multilayer circulation seen in summer (Inall et al., 2014; Sutherland et al., 2014a) is efficient at drawing warm shelf waters into the fjord, the heat content in the inflowing layers is depleted due to entrainment via shear-driven mixing with the colder outflows both above and below. In contrast, in the horizontally sheared circulation generated by our model, the inflowing and outflowing current cores reside against opposite sides of the fjord, minimizing the opportunity for entrainment.

4.2. Conditions at the Mouth

The fjord mouth was colder after barrier wind forcing and the stratification was weaker (Figure 14), better matching the stratification on the shelf than in the fjord interior. It is important to understand the response on the shelf in order to give context to the results from the KF interior.

In Run-2, we found the time-mean inflow on the eastern side of KT and the fjord mouth (sections 5 and 6) to be strongly barotropic (Figure 7) whereas in the control run the inflow intensified toward the bottom, indicating a significant baroclinic contribution (Figure 5). This is exactly analogous to findings from West Spitsbergen, where increased barrier wind stress enhanced the AW-influenced barotropic, cross-shelf trough currents (Nilsen et al., 2016). Furthermore, these currents were found to follow shallower contours on shelf, as the barotropic pressure gradient has shifted coastalward, and the extent to which they penetrated the fjord mouth was hence increased. This is reflected in Figure 14, where shelf waters are seen to circulate inside the fjord mouth in Run-1 and Run-2, while in Run-0 the fjordic water properties of KF remain well established in the fjord mouth and extend well beyond.

The time-averaged flow in KT was strengthened with wind forcing (Figure 7), and an abundance of cold, dense water was transported along KT and across the fjord mouth. This had the effect of temporarily isolating the waters in KF from those on the shelf and, later, cooling the head of KT (Figure 10).

We also found that barrier winds greatly enhance vertical mixing due to increased vertical shear. This effect was strongest in the fjord mouth, where the internal wave field was most energetic. With prescribed subglacial discharge set to zero, melting alone was insufficient to reestablish the stratification so far from the KG.
terminus after intense mixing. Iceberg melt is a process missing from the model which would help reestablish buoyancy away from the terminus.

The changing conditions at the fjord mouth had profound effects on interactions between fjord and shelf. This is evidenced by the contrasting responses between the three model runs to the presence of dense waters outside the fjord mouth in the latter days of the experiment. Although the transport of cold, dense waters along KT was increased by barrier wind forcing, the region underwent net cooling in all runs. In Run-0, where the stratification inside KF was largely unchanged as no barrier winds were induced, the fjord is resilient to changes on the adjacent shelf and the warm layer remains well established in the mouth (Figure 14). The horizontal and vertical extent of this layer is seen to be reduced after the barrier wind events in Run-1 and Run-2, and the interface between fjord-like stratification and the shelf-like stratification is located further inside KF. This makes the fjord susceptible to deep water renewal, and toward the end of the model run dense shelf waters are seen to circulate north of the sill and flood the fjord bottom. This result suggests an additional mechanism for fjord-shelf exchange, through which barrier wind events intermittently replenish the waters below sill depth. It is likely that in summer, with nonzero subglacial discharge acting to increase stratification within KF, the system would have reverted more quickly back to the control case after barrier wind forcing.

4.3. Coastal Trapped Waves

In the absence of both subglacial discharge and runoff, KF was dynamically a two layer system, as reflected by the first baroclinic mode accounting for over half of the variability in velocity in Run-1 and Run-2. This is in contrast to observations from KF in summer where subglacial discharge creates a multilayer stratification introducing higher order dynamical modes which dominate the variability (Sutherland et al., 2014a). In our model, the dynamical response to barrier wind forcing was broadly consistent with that proposed by Straneo et al. (2010) for SF, in that subinertial waves in the pycnocline drive opposing currents in the upper and lower layers. However, the nature of the intermediary circulation is not fully consistent with previous descriptions based on KF or SF.

We find a subinertial, geostrophically balanced internal wave propagating up-fjord, intensified toward the eastern fjord wall. The pressure gradient term was found to follow an ellipse, with the major axis parallel to the fjord wall. This behavior is symptomatic of coastal trapped wave (CTW) activity. CTWs are a class of mixed gravity/rotational internal wave of subinertial frequency where the Coriolis force is balanced against a coastal boundary, decaying laterally on an e-folding lengthscale of one Rossby radius (Allen, 1975). The Burger Number, \( Bu = (NH/L)^2 \) where \( H/L \) is the topographic slope and \( f \) is the Coriolis parameter, represents the relative effect of stratification compared to potential vorticity conservation over sloping terrain. The high values obtained in KF (40 < \( Bu < 160 \), with values increasing moving up-fjord) indicates that stratification dominates over slope effects here, such that the CTWs can be closely approximated as internal Kelvin waves bounded by a vertical wall. This is further evidenced by the close match between the observed phase speed and phase speed predicted by dynamical mode analysis, as Kelvin waves are nondispersive and therefore have a constant phase speed over the wavenumber spectrum.

The asymmetry afforded by broad fjords prevents the resonant seiching motions described in SF (Sutherland & Straneo, 2012), as up-fjord and down-fjord CTWs have maximum amplitudes on opposite sides of the fjord. Due to energy dissipation, the up-fjord wave dominates the variability in comparison to any potential reflected down-fjord wave motion (Figure 11), and it is not clear that the signal can propagate around the head of the fjord to produce an outgoing wave. CTWs in fjords may therefore be considered a special class of intermediary circulation where rotational effects give rise to additional physics.

CTWs have previously been observed in West Spitsbergen (Inall et al., 2015), where they were also found to drive large intermediary exchange flows between fjord and shelf. Furthermore, the cross-fjord geostrophic velocity section from KF in Inall et al. (2014, Figure 6) showed lower layer velocity opposing that in the upper layer, and current intensification toward the coast. This was previously considered a snapshot of a balanced flow, potentially steady over a longer time scale than that of CTWs, which were not examined as a candidate explanation for the observed geostrophic flow structure. In light of the results found here, we reinterpret this signal as more likely a manifestation of CTW activity. Furthermore, Sutherland et al. (2014a, Figure 8d) shows a geostrophic velocity field which is consistent with CTW propagation, calculated based
on observations of KF in summer 2009. Model output animations by Carroll et al. (2017) (supporting information) shows CTW behavior in a broad fjord with idealized topography.

### 4.4. Stokes’ Drift

We attribute the strengthening of the geostrophically balanced background currents seen in Run-1 and Run-2 to Stokes’ drift (Stokes, 1847), a process whereby the fluid through which a wave propagates is accelerated in the direction of propagation. CTWs are known to produce this effect (Weber & Ghaffari, 2014; Wunsch, 1973), and this behavior has been suggested as a residual flow driver in Arctic fjords in West Spitsbergen (Inall et al., 2015). From Inall et al. (2015, equation 11), we find a Stokes’ drift of $O(1 \text{ cm s}^{-1})$ at section 2, which is consistent with the mean flow enhancement from the model (Figures 5 and 7). This is a secondary mechanism through which barrier winds can enhance exchange. Unlike intermediary circulation mediated by CTWs, where the time-averaged exchange cancels, the Stokes’ drift drives a net enhancement of the background flow. It is reliant on the lateral asymmetry of CTWs and, hence, exclusive to broad fjords. This phenomenon, together with the strengthened barotropic trough current, helps to explain why the exchange in Run-1 remained enhanced after the internal wave activity had ceased. We speculate that the large capacity for CTWs to advect shelf waters into the fjord interior via intermediary circulation may be significant in initiating a two-part process whereby these waters are then distributed to upper fjord by the enhanced geostrophic circulation of the fjord interior.

### 4.5. Volume and Heat Exchange

CTW activity was seen to dominate the KF dynamics in the days following a barrier wind event on the shelf and gave rise to enhanced exchange through all sections. In particular, the mean exchange through all sections of the KF mouth and interior (sections 1–5) was approximately doubled in Run-2, which was subject to two wind events in 25 days, compared to Run-0, which was subject to none. Barrier wind events occur with frequency around one per week during the winter months, more frequently than in Run-2, suggesting that KF has the capacity for significant wintertime exchange.

Given a positive temperature gradient moving from fjord to shelf, we anticipate the enhanced exchange due to intermediary circulation will increase the heat content of the fjord interior. However, the supply of IC water to the mouth of KF via KT is not steady, with Run-0 heat flux through KF (section 6) varying on a time scale of over a week. The conditions required at the fjord mouth for large positive heat flux were hence not always in place during wind events. The single barrier wind event in Run-1 was found to enhance the time-integrated heat delivery toward KG through all closed sections of the fjord (Table 3). However, the temperature outside the fjord mouth decreased during the simulation, so that the enhanced exchange following the second wind event in Run-2 resulted in negative heat flux toward KG. The temperature of the Irminger Basin and SE Greenland shelf has been anomalously high in the last two decades (Khan et al., 2014), while the PHC3.0 climatology data used to initialize the model reflect the conditions in the latter half of the twentieth century. It is therefore quite possible that the use of hydrographic conditions reflecting a more modern shelf sea state would have resulted in larger temperature gradients between fjord and shelf and would hence have yielded larger heat flux values.

When looking to quantify the capacity for barrier winds to drive up-fjord heat flux, we therefore chose to focus on the aftermath of the wind event in Run-1. While the percentage increase was large, the time-averaged heat flux through the midfjord following the initial barrier wind event was, at ~0.02 TW, still an order of magnitude smaller than the literature values recorded in summertime observations (Inall et al., 2014; Sutherland et al., 2014a) or summertime simulations (Cowton et al., 2016). Near KG (section 1), the percentage increase was smallest, however the value of 0.0018 TW approximately matches the value of 0.003 TW measured there by Sutherland et al. (2014a) in summer 2009. This reaffirms that, although heat delivery through the mouth and midfjord is significantly smaller in winter than in summer, the heat energy brought into direct contact with KG may be comparable. The doubling of this value in the 12 days after the first barrier wind event would likely have consequences for glacier stability, given the frequency of these events during DJF.

### 4.6. Wider Implications

While increased AW content of West Spitsbergen fjords has been linked to changing along-shelf wind patterns in during the last two decades (Nilsen et al., 2016), the number of barrier wind events per winter...
season has remained stable over this period (Harden et al., 2011). This indicates that, while barrier winds would have driven exchange between the fjord and the shelf during the period while ocean temperatures increased, they did not trigger the sudden glacier acceleration and retreat observed across SE Greenland in the mid-2000s.

It is likely that CTWs are a major mechanism for exchange in broad fjords in general. The summertime observations of CTWs in SE Greenland Inall et al. (2014) and West Spitsbergen Inall et al. (2015) indicate that they are not a winter-specific phenomenon, while the results of Carroll et al. (2017) indicate that they can be generated by sill-tide interactions. While the strong, along-shelf winds characteristic of SE Greenland in winter are a trigger for CTW driven exchange, this is not a necessary condition and we thus anticipate this behavior in broad fjords elsewhere. For example, Petermann Fjord in northwest Greenland, Isfjorden in West Spitsenbergen, and SF all have a rotational influence (Johnson et al., 2011; Nilsen et al., 2016; Sutherland et al., 2014b) and have all been acutely affected by ocean warming. CTWs present a candidate mechanism for the effective transport of heat into these regions.

4.7. Evaluation

Though velocity sections from summer also display strong cross-fjord variability (Inall et al., 2014), it is not clear the extent to which these geostrophically balanced currents represent the year-round mean flow. Freshwater input was restricted to ocean-driven melt at the glacier terminus, and hence we do not see the strong buoyancy-driven circulation or complex stratification structure observed in summer field campaigns (Inall et al., 2014; Sutherland et al., 2014a). As winter observations in KF are very limited it is difficult to validate the model, leading to uncertainties in how accurately it recreates typical wintertime conditions. Additional mooring data from the KF interior, particularly during barrier wind times, would be hugely beneficial for this purpose and for designing future modeling studies of wintertime exchange.

The model presented here was designed to be as simple as possible while capturing the leading-order processes. We therefore omitted some features which could be of significance. For example, icebergs have been identified as a major contributor to the GrIS freshwater flux (Enderlin et al., 2016; Moon et al., 2018), brine release due to sea ice formation during winter is known to exert influence on the stratification within high-latitude fjords (Cottier et al., 2010), and the tides have been found to generate exchange and vertical mixing at the sill in many fjords (Inall & Gillibrand, 2010); all processes we were unable to quantify here.

With processes on the shelf (which dictate conditions at the fjord mouth) varying on time scales of over a week, the 25 day simulations were too short to expose KF to a comprehensive range of external conditions during barrier wind forcing. Furthermore, the climatology data used to initialized the model may not be representative of the modern state of the SE Greenland shelf. The wind fields used to force the model featured a typical barrier wind event focused on the region of shelf east of KT (Harden et al., 2011). It is likely that a barrier wind event focused over KT itself would drive a stronger exchange. We are therefore not in a position to claim full knowledge of the processes governing wintertime exchange. A modeling study running for an entire winter season, exposing the system to a range of wind forcing scenarios and capturing the evolution of the fjord-shelf interaction over a range of time scales, would allow us to better evaluate the mean wintertime heat exchange.

5. Summary

We have reported here on a numerical model of Kangerdlugssuaq Fjord (KF) and the adjacent shelf to study the physics of fjord-shelf exchange under typical wintertime conditions. We have identified an additional mechanism for along-fjord circulation, whereby geostrophically balanced background currents advect water along the fjord. This was the dominant feature of the mean circulation, and was of the same order as intermediary circulation with regards to heat delivery through the upper fjord.

Barrier wind forcing had a profound effect on the dynamics in the ~8 days following peak wind stress, provoking a highly energetic intermediary circulation. We find, in accordance with wintertime observations of KF by Jackson et al. (2014), that the barrier wind signal propagates rapidly (~ 40 cm s^{-1}) from the shelf to the upper fjord. This was, however, a different flavor of intermediary circulation to that previously discussed for SE Greenland fjords: the barrier wind signal propagated up-fjord as coastal trapped waves (CTWs), bounded by the eastern side wall. Under this regime, which likely features in broad fjords in general,
incoming and outgoing CTWs are spatially distinct, prohibiting resonant seiching motions and enhancing the geostrophic background circulation via Stokes’ drift.

CTWs were found to be highly effective at increasing heat transport through the fjord mouth and midfjord, but the values were still small in comparison with observations taken during the summer. The average heat flux through the inner part of the fjord was, however, closely comparable with summertime observations, and approximately doubled in the period during which CTWs were active. Hence, barrier winds are likely a significant factor in oceanic heat delivery toward the GrIS at KF and other broad fjords. Although it is unlikely that barrier wind activity was the trigger for the rapid changes to KG in the past two decades, our results suggest that heat delivery through the upper fjord can be as significant in winter as in summer.

The stratification inside the fjord was found to be different in nature to that of the adjacent shelf, and we identified a transition zone in the fjord mouth 80–100 km from the model’s glacier terminus. This adds a subtlety to the notion of waves in the pycnocline propagating from shelf to fjord, as the pycnocline on the shelf does not necessarily coincide in depth, strength or nature with that in the melt water-freshened KF. As well as driving CTW activity, barrier wind events were found to weaken the stratification within the fjord mouth resulting in a northward shift in the interface between the two stratification types. This was due to a combination of enhanced vertical mixing in the fjord mouth and strengthening of barotropic shelf currents in KT. Shelf waters were subsequently able to circulate more easily in the fjord mouth after barrier wind forcing, and cold, dense shelf waters were able to cascade over the sill into the bottom of KF in the final days of Run-1 and Run-2. In contrast, in the control run the fjord-like stratification remained well established in the fjord mouth and above the sill, isolating the dense waters in the fjord from those on shelf.

We operated under the assumption that runoff and subglacial discharge are negligible in winter, however this is hard to verify due to the dearth of wintertime observations. Regardless, we found that subglacial discharge is not required in order for the background circulation to deliver heat toward KG.

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