Wintertime Fjord-Shelf Interaction and Ice Sheet Melting in Southeast Greenland

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Key Points:

• Glacier terminus regularly exposed to 0.5 TW of oceanic heat during winter.
• Wind forcing frequency a crucial parameter in determining magnitude of heat exchange.
• Vertical mixing enhances buoyancy-driven overturning circulation.

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Abstract

A realistic numerical model was constructed to simulate the oceanic conditions and circulation in a large southeast Greenland fjord and the adjacent shelf sea region during winter 2007-2008. The major outlet glaciers in this region recently destabilised, contributing to sea level rise and ocean freshening, with increased ocean heat content the most probable trigger. Principal mechanisms for fjord-shelf heat exchange are buoyancy-driven overturning due to freshwater input, and rapid baroclinic exchange events due to wind-driven downwelling on the shelf. The selected study period was wintertime, when freshwater input is minimal and shelf wind events are most frequent, because of the scarcity of wintertime data. As the internal Rossby radius is comparable to the fjord width, it is not apparent a priori whether the fjord dynamics will be influenced by the rotation of the Earth. The model output currents, however, describe a highly three-dimensional system, where the rotation of the Earth is of order-one importance. Along-shelf wind events were seen to drive a rapid baroclinic exchange, mediated by coastally trapped internal waves, which propagate from the shelf to the glacier terminus along the right-hand boundary of the fjord. The terminus was regularly exposed to around 0.5 TW of heat power over the winter season. Wave energy dissipation provoked intense vertical mixing, resulting in a buoyancy flux which strengthened the estuarine overturning circulation. Although the outgoing wave was less energetic and located at the opposite side wall, the fjord did exhibit a resonant response, suggesting the fjords of this scale can also exhibit two-dimensional behaviours. Long periods of moderate wind stress greatly enhanced the cross-shelf delivery of heat towards the fjord, in comparison to stronger events over short intervals. This suggests that the timescale over which the shelf wind field varies is a key parameter in dictating wintertime heat delivery from the ocean to the ice sheet.

1 Introduction

Recent reduction in the Greenland Ice Sheet (GrIS) mass balance has been most profound near its edge, indicative of ocean triggered melting [Rignot and Kanagaratnam, 2006; Nick et al., 2009]. Coastal water temperature has increased contemporaneously [Strange et al., 2013; Khan et al., 2014], however direct contact between the ocean and the GrIS is limited to glacier termini which are typically located within Greenland’s fjords. A thorough understanding of the exchange flows between these fjords and the continental shelf is therefore critical for quantifying the ocean’s impact on the GrIS.

One of the most acutely affected glaciers in the previous two decades is Kangerdlugssuaq Glacier (KG), which terminates at Kangerdlugssuaq Fjord (KF), and is one of the major outlet glaciers of southeast Greenland. KG destabilised in 2004-05, when the rate of discharge suddenly doubled [Bevan et al., 2012], and again in 2016-2017 [Suzanne Bevan 2018, personal communication, 20th April], with re-advance and slowing generally exhibited in the interim period [Khan et al., 2014]. KF is around 80 km long and 6 km wide with a maximum depth of around 900 m and a sill depth of around 500 m. At the mouth, where it widens to around 20 km, the fjord meets the north end of Kangerdlugssuaq Trough (KT), a 600 m deep cross shelf channel (Figure 1). KT is a known pathway for ocean waters from the Irminger Sea [Gelderloos et al., 2017], and intersects the shelf break at its southern end. Here, Atlantic Water (AW, Conservative Temperature $\theta \sim 4.5 - 6.5 ^\circ C$, Absolute Salinity ($S_A$) $\sim 34.9 - 35.2$ g kg$^{-1}$) flows from east to west in a branch of the North Atlantic Current known as the Irminger Current (IC). A second, seasonal pathway for AW water towards KF is north through the Denmark Strait and across the shelf, leading to a warmer AW layer in winter than in summer [Gelderloos et al., 2017]. South of the Denmark Strait, the IC is joined by the East Greenland Current (EGC), which transports Polar Water (PW, $\theta < 0 ^\circ C$, $\sigma_\theta < 27.70$) from the Arctic Ocean. Alongside the EGC, the East Greenland Coastal Current (EGCC) transports PW southwards close to the coast. Dense bottom water, termed Denmark Strait Overflow Water (DSOW, $\theta < 0 ^\circ C$, $34.9 < S_A < 35.2$ g kg$^{-1}$, $\sigma_\theta \geq 27.8$), also enters...
Figure 1. Bathymetry of southeast 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 region here, released over the Denmark Strait sill in periodic boluses [Koszalka et al., 2013].

Due to seasonal sea ice cover, observations of KF hydrography and circulation are biased towards the summer months, when freshwater runoff is strongest, and there is hence a relatively large literature on the buoyancy-driven circulation in Greenland fjords [Sciaccia et al., 2013; Cowton et al., 2015; Carroll et al., 2016]. A recent study by Moon et al. [2017] highlighted the importance of subsurface iceberg melt as a freshwater source in major SE Greenland fjords. This is also apparent in Table 2 of Inall et al. [2014], where the large residual heat loss from PSW is associated with iceberg melting within KF. In winter, when runoff is at a minimum, other factors likely play a primary role in driving circulation. Results from Sermilik Fjord (SF) [Straneo et al., 2010; Jackson et al., 2014; Sutherland et al., 2014], a similarly sized neighbour to KF, indicate that intermediary circulation, a rapid baroclinic exchange regime triggered by along-shelf (with shore to the right) barrier winds, is a significant driver of fjord-shelf exchange. Enhanced wind stress drives coastward flow in the Ekman layer resulting in downwelling of the pycnocline, followed by upwelling once the wind relaxes. Baroclinic exchange flows are generated as the displacement of the pycnocline propagates in-fjord as a subinertial internal wave [Fraser and Inall [2018], hereafter FI18]. As barrier wind events occur predominantly in the winter months [Harden et al., 2011], the capacity for this mechanism to draw warm ocean waters into contact with glacier termini remains uncertain. Modelling studies of KF [Cowton et al., 2016] have found that, while intermediary circulation provokes a rapid baroclinic exchange, heat delivery to the upper fjord is not significant in comparison with values recorded during summer field campaigns [Inall et al., 2014; Sutherland et al., 2014].
Two-dimensional overturning regimes, driven by either runoff or shelf exchange, have been the main focus in previous studies of circulation in KF and SF. However, the recent modelling study by FI18 found that horizontally sheared, geostrophically balanced flows dominate the mean wintertime circulation in KF and facilitate exchange, with the inflowing (outflowing) currents residing against the right-hand (left-hand) boundary looking into the fjord. Furthermore, the subinertial internal waves which drive intermediary circulation were coastally trapped waves (CTWs), with maximum amplitude against the eastern sidewall while propagating up-fjord. FI18 found that the steep topography and high stratification dominated over effects from potential vorticity conservation (placing the system in a high Burger Number regime) so that the waves could be approximated as nondispersive Kelvin waves propagating at the speed of a mode-1 internal wave. Such cross-fjord variability is only prominent in fjords wider than the internal Rossby radius of deformation, \(L_R\), which summer campaigns have estimated to be around 8 km in southeast Greenland [Inall et al., 2014; Sutherland et al., 2014]. KF is approximately 6 km across, and \(L_R = \frac{NH}{f}\) is a linear function of the local stratification so may decrease under winter conditions. The potential for a three-dimensional flow field inside KF introduces complexity to the current understanding, and the implications for fjord-shelf heat exchange are not fully understood.

As well as inducing a dynamical response, barrier winds have been found to make enduring changes to the water column structure in the fjord mouth with considerable implications for subsequent exchange. In an idealised modelling study into barrier wind forcing of the KF/KT system under winter climatological conditions, FI18 found that simulations exposed to wind events exhibited greatly enhanced vertical mixing in the fjord mouth due to subinertial internal wave activity. Transport in KT was also enhanced by barrier wind forcing, and the extent to which cyclonic circulation in KT penetrated the fjord mouth was increased. Together these factors acted to weaken the stratification in the fjord mouth and introduce a more shelf-like water column structure there, an artefact which remained after the dynamical response to wind forcing (i.e. internal wave activity) had decayed. At a later time, dense bottom waters circulating in KT were able to breach the KF sill and cause a deep water renewal event in the fjord, reminiscent of observations of DSOW within KF [Inall et al., 2014]. In model runs where prevailing winds were held constant (without barrier wind events), the mouth, like the KT interior, remained strongly stratified due to the freshening influence of the glacier front, and was resilient to deep-layer exchange with KT.

In this study, we use an adapted version of the model presented in FI18 to study the circulation and exchange in KF during December, January and February (DJF) of 2007-08. While FI18 looked to isolate the effect of barrier wind events against a backdrop of winter climatological conditions through the use of a control run, here we look to place their influence in the context of a realistic reconstruction of a winter season. We focus primarily on shelf exchange processes, with the aim of definitively answering the question “Is there potential for significant wintertime heat exchange between shelf and fjord?”. The model is equipped with a parametrisation of the KG glacier front (a heat sink and freshwater source) which generates output variables for glacial melt rates [Cowton et al., 2015]. We therefore look for correlations between melt rates and various potential drivers of circulation, particularly wind forcing on the shelf.

2 Methods

The model used was the MIT general circulation model, which solves the Boussinesq equations of motion using the finite volume method (MITgcm, Marshall et al., 1997). In this study we also employed the hydrostatic approximation. Integration was performed by the ARCHER UK National Supercomputing Service (http://www.archer.ac.uk). The model grid and bathymetry was constructed exactly as described in FI18, and so is only briefly outlined here.
The model domain covers 66.38–68.5°N, 34.59–28.05°W (Figure 2). It captured KF with a horizontal resolution of 360 m and a vertical resolution of 10 m. The grid spacing increased towards the southern, eastern and western boundaries, so that the resolution on the shelf was relatively coarse with a maximum value of 4 km in the southeast and southwest corners. Bathymetry for the shelf region was extracted from the 30-arcsecond International Bathymetric Chart of the Arctic Ocean (IBCAO). Bathymetry for the fjord interior was collected using a swath on the cruise JR106b to KF [Dowdeswell, 2004]. Initial and boundary conditions were generated using output from the model presented in Gelderloos et al. [2017], which was used to simulate the wider Irminger Sea region for one year beginning 1st June 2007. The availability of this high-resolution forcing data was our motivation for selecting that particular winter for hindcasting.

Wind and air-sea heat flux data were obtained from ERA-Interim 6-hourly and daily reanalysis products [Dee et al., 2011] respectively. Wind stress fields were calculated using the formula from Large and Pond [1981], which were then modified offline to reflect local sea ice cover, as described in FI18, using temporally varying sea ice concentration data obtained from the National Snow and Ice Data Centre (NSIDC). ERA-Interim wind fields have been shown to resolve high-frequency katabatic winds in SE Greenland fjords [Oltmanns et al., 2014], giving confidence that this product is able to adequately capture near-shore wind processes.

The MITgcm iceplume package [Cowton et al., 2015] was employed to incorporate the dynamical and thermodynamical effects of ice-sea interaction at the KG terminus. The package facilitates prescribed subglacial runoff, calculates local melting as a function of the temperature of the adjacent grid cells, and analytically solves the plume equations from Jenkins [2011]. Such a parametrisation eliminates the necessity to run the model in non-hydrostatic mode by distributing resultant water masses at the level of neutral buoyancy. We took advantage of the output variables for glacial melt rates provided by the iceplume package as an opportunity to study correlations between glacial melt and fjord-shelf exchange forcings. However, as we later discuss, the package was designed to describe the influence of the ice on the water, not vice-versa, so we are cautious when interpreting variables related to glacier dynamics.

We employed the $\kappa$-Profile Parametrisation (KPP), introduced by Large et al. [1994], which calculates the vertical mixing coefficient as a function of the local gradient Richardson number in the ocean interior and as a function of the bulk Richardson number in the mixed layer. We used the Leith biharmonic scheme [Leith, 1996] to parametrise horizontal viscosity.

As the model by Gelderloos et al. [2017] was not of sufficient resolution to include KF, the initial conditions within the fjord were horizontally extrapolated from the shelf. A 100-day spin-up period was then carried out, with some runoff (100 m$^3$ s$^{-1}$) prescribed at the KG grounding line during the initial 60 days in order to allow an overturning circulation to develop within the fjord. Wind and boundary forcing were held constant at December 1st values during this period. The model was then integrated forwards using dynamic forcing fields for 91 days, the duration of DJF 2007-08, with a timestep of 5 seconds.

Harden et al. [2011] define a barrier wind event as wind blowing from the northeasterly quadrant, exceeding 20 m s$^{-1}$, and being distinct in time from other such events by 24 h or more. According to this definition, nine barrier wind events occurred on the shelf outside KF in DJF 2007-08, and their occurrences are shown in Figure 3(a) alongside northeasterly component of wind speed (note that this does not necessarily reflect the Harden et al. [2011] definition threshold). The corresponding wind stress, which is usually quadratic in wind speed but becomes cubic when the wind exceeds 11 m s$^{-1}$ [Large and Pond, 1981], is show in Figure 3(b). Barrier wind events occurred less frequently during DJF 2007-08 than is typical during DJF, with the number ranging from 7 to 20
during 1989-2008 [Harden et al., 2011]. Events were generally clustered in time, with four
events taking place in early December (hereafter Cluster A), two around the start of Jan-
uary (Cluster B), two towards the end of January (Cluster C), and one in mid-February
(Cluster D). Two of the wind events, the first in December and first in January, coin-
cided with prolonged periods of strong northeasterly wind stress, and were hence char-
acteristically different to the shorter peaks seen at other times. Both air and sub-surface
water temperatures were anomalously high in comparison with the 1981-2012 mean, though
consistent with other years since 2000 [Khan et al., 2014]. Meridional velocity into KT
at the southern boundary is also an important external driver of dynamical variability,
and is shown in Figure 3(c). Barrier wind events regularly coincided with enhanced in-
flow into the model domain, likely due to the intensification of barotropic currents on
the shelf by along-shore wind stress as described in Nilsen et al. [2016]. Notable excep-
tions arise in late January and early February, however, when enhanced inflow did not
coincide with wind events, indicating that other factors also influence inflow variability.

3 Results and Analysis

We first give some validation of model realism, by way of comparison with in situ
measurements. With wintertime observations of the region scarce, no such data was avail-
able from within the model domain during the period of study. The vertical tempera-
ture structure at 200-300 m depth (Figure 5, Sections 1, 2 and 3) agrees well with the
mooring record by Jackson et al. [2014] in 2009-10. We also utilised summertime obser-
vations from within KF, finding that the cross-sectional temperature structure in the model
is closely comparable to that observed in September 2010 by Inall et al. [2014] below around
250 m. Differences in stratification shallower than this depth are likely attributable to
seasonal variability in freshwater runoff. Some wintertime temperature and salinity data
from the shelf region of the model domain was obtained by an instrumented seal [Trea-
sure et al., 2017] during the 4-5th January 2005. The seal performed 9 dives near the
southern boundary (Figure 2), sometimes exceeding 300 m depth, giving temperature
measurements along its path. Figure 4 shows the resulting temperature field alongside
the corresponding model temperature field for 4-5th January 2008, interpolated onto the
seal’s path. Overall, the model shows generally good agreement with the observations
in terms of stratification structure, thermocline height, and the temperature in the up-
ner and lower layers. The model does not reflect sharp thermocline and subsurface tem-
perature maximum seen at between 100 and 200 m depth in the observations, while the
surface waters (top 50 m) are also colder in the model. This may be due to interannual variability as opposed to model inaccuracy. This agreement with data gives confidence that the model successfully captures the leading-order physics.

The mean flow through six cross-sections of the combined KF/KT system is shown in Figure 5, with mean conservative temperature contours overlaid. Section locations are shown in Figure 2. In KT (Section 6) we see a strongly barotropic flow regime, with inflow (outflow) of around 40 cm s\(^{-1}\) on the right (left) flank looking towards the fjord. In the fjord mouth (Sections 4 and 5) the mean flow is weaker and intensifies with depth, with current cores of around 15 cm s\(^{-1}\) concentrated against side walls at around 400 m depth. Moving in-fjord the currents become weaker still, while retaining the pattern of inflow on the right and outflow on the left. Temperature contours reveal a strong thermocline (which coincides with the pycnocline, not shown) within KF at a mean depth of around 200 m. Absolute geostrophic velocities (not shown), calculated using the sea surface height (SSH) and density fields at each section, are in close agreement with the modelled fields, indicating that the circulation is typically in geostrophic balance to a close approximation.

We computed the overturning streamfunction at each cross-section of the fjord interior (Figure 6), revealing any of the residual overturning circulation within the fjord obscured by the large horizontal variability in the mean flow. At all sections, the time-mean streamfunction displays four local extrema, indicating a complex, multi-layered cir-
Figure 4. In situ temperature (a) along the path of an instrumented seal on 4-5th January 2005 and (b) from the model temperature field interpolated onto the same path, but for 4-5th January 2008.

culation scheme. The strength of overturning increases moving out of the fjord, most markedly between Section 2 and 3.

Motivated by the barotropic nature of the flow in KT, we investigated the sea surface height (SSH) anomaly on Section 6 (relative to the spatio-temporal mean), looking specifically for correspondence between wind forcing and shoreward transport. Figure 7 shows the time evolution of SSH gradients alongside the depth-averaged current (DAC) normal to the section. The surface is generally depressed in the middle of the section and elevated at either side. Barrier winds regularly correspond to a deepening of the central depression, and appear to temporarily hinder the northward DAC on the eastern side while enhancing the southward DACs in the western side. There is a marked discontinuity between the SSH structure in the first half of December and the rest of the simulation. We suspect this is due to either the influence of the erratic wind forcing during of Cluster A (Figure 3) on the Ekman layer, an artefact of the southern boundary condition changing from static to dynamic and the beginning of the simulation, or both.

From density profiles within the fjord, we obtained the horizontal velocity structure associated with normal baroclinic modes of oscillation. The linear, mode-one internal wave speed was $c_1 = 1.1 \text{ m s}^{-1}$, in agreement with Inall et al. [2014]. From this, we computed $L_R = c_1 / f = 8.1$ km and found the resonant seiche period to be $T = 4L/c_1 = 3.4$ days.

We used empirical orthogonal function (EOF) analysis to isolate the statistically dominant modes of variability in the velocity field at each section of the KF interior (Sections 1, 2 and 3). Specifically, we note EOFs featuring a nodal contour corresponding to the zero-crossing in the first normal mode (around 200 m, approximately the mean pycnocline height). This pattern was seen in EOF 1 on Sections 1 and 3, accounting for 31% and 49% of the variance at their respective locations (Figure 8). On Section 2, this class of variability projected onto the second EOF which accounted for 30% of the total variance (the first EOF at Section 2, not shown, accounted for 37% of the variability and was similar in structure through more weakly sheared, with a nodal contour at
around 350 m). In each of these fields, velocities above the pycnocline opposed those below, with strong vertical shear occurring at around 200 m depth. A similar vertical structure was found in the corresponding horizontal normal modes (not shown, methodology in e.g. F18; Sutherland and Straneo [2012]). This pattern of vertical variability is most intense adjacent to the eastern sidewall of the fjord and weakens toward the fjord interior. In Sections 2 and 3 this trend continues to the western side of the fjord, while in Section 1 the pattern reverses west of the fjord centerline and intensifies again towards the western sidewall.

Figure 9 shows a time series of horizontal velocity, \((u, v)\) where \(u\) is the across-fjord component and \(v\) is the along-fjord component, averaged over the Section 2 lower-layer inflow region (defined as \(z < -200 \text{ m}, \tau > 3 \text{ cm s}^{-1}\), Figure 5). Velocities are largely directed along-fjord, regularly alternating in sign. Cross-fjord velocities are maximal during these transitions, but are smaller by an order of magnitude. The largest along-fjord velocities, along with the most frequent sign changes, generally occur in the days immediately following barrier wind event clusters on the shelf.

Figure 10 shows a time series of the model-generated temperature profile 500m from the eastern boundary of Section 2. Quasi-periodic oscillations in the height of the thermocline persist throughout the simulation, although the shape, amplitude and frequency of the waveforms is highly variable. Furthermore, the thickness of the thermocline (defined as \(-0.5 < \Theta < 1.5^\circ\text{C}\)) changes during the simulation, increasing from an initial value of around 50 m to reach almost 200 m, with a subsequent decrease coincident with increasing lower-layer temperature.

Wavelet analysis was used to decompose the velocity variability in frequency space. Similar to Fourier analysis, this method has the added advantage that the amplitude at each basis frequency may vary temporally, allowing a spectral perspective on the model’s
response to either stochastic or externally forced variability on the shelf. We performed
the analysis on the along-fjord component of the Section 2 lower layer inflow (Figure 9),
using a Morlet wavelet basis function (Figure 11; the different basis options are detailed
The most significant harmonic variability occurs with period 2-4 days, consistent with
the predicted resonant seiching period. There is a strong coincidence between barrier wind
activity and excitation of this period band, with the frequency-averaged wavelet power
exceeding the 95% confidence level on 4 occasions (Figure 11(c)), each corresponding to
a barrier wind event cluster. Harmonic variability also occurs with period ∼25 days,
which is broadly consistent with the interval between wind event clusters. However this
period lies largely within the cone of influence (Figure 11(a)), introducing the risk of spu-
rious signals due to edge effects, and does not exceed the 95% confidence interval (Fig-
ure 11(b)).

Defining exchange as

\[ Q = \frac{1}{2} \int \int |v(x,z)| \, dx \, dz \] (1)

where \( x \) and \( y \) are the respective across- and along-fjord coordinates, we calculated
time series of the exchange through each cross-section, shown in Figure 12. In the fjord
mouth (Sections 4 and 5) barrier wind events are commonly followed by spikes in exchange,
particularly following the first wind event of each cluster. The exchange through KT (Sec-
tion 6) appears less sensitive to variability in wind patterns over short timescales. The
maximum correlation between the Section 5 and Section 1 timeseries occurred at a lag
time of 14 hours, with a correlation coefficient of 0.94. The two sections are approximately
55 km apart, indicating that information propagates up-fjord at around 1.1 m s\(^{-1}\), the
predicted mode-one internal wave speed.

Defining advective heat flux as

\[ Q_\Theta = C_p \rho_0 \int \int v(x,z) \Theta(x,z) \, dx \, dz \] (2)

where \( C_p \) is the specific heat capacity of seawater and \( \rho_0 \) is reference density, we
calculated time series of the heat flux through each cross-section, shown in Figure 13.
The mean, standard deviation, and maximum heat flux values of through each section
are shown in Table 1. Barrier wind activity generally results in an oscillating heat flux
signal at all locations, and hence the standard deviation is two orders of magnitude larger
than the mean at each section. The amplitude of the oscillation decreases by around a
factor of 10 between the fjord mouth and the fjord head. The response to each wind event
is inconsistent, differing in amplitude, frequency and number of cycles. For example, the
response to Cluster B is manifest as a relatively low-frequency oscillation, compared to
the responses to Clusters A, C and D. Furthermore, there is evidence of coherent sig-
nal propagation which is not obviously caused by barrier wind forcing. Figure 14 shows
the cumulative time-integral of the heat flux plots shown in Figure 13.

We calculated the internal wave energy flux, \( v'P' \), through each cross-section, where
\( v' \) and \( P' \) are the time-varying deviations from the mean along-fjord velocity and mean
pressure respectively [Nash et al., 2005]. Figure 15 shows the depth-integrated time-mean
energy flux through each sections while Figure 16 shows the depth-varying field in the
fjord interior. These depth-varying fields resemble, somewhat, the corresponding EOFs
in Figure 8. There is a net energy flux into the fjord through all sections, concentrated
on the right-hand side of the fjord at around 300 m depth. From Figure 15 it is evident
that the incoming wave is relatively nondispersive. The significant down-fjord wave en-
ergy flux on the left-hand side of Section 1 indicates that CTWs can propagate around
Table 1. Mean, standard deviation, and maximum heat flux (TW) towards KG through each section.

<table>
<thead>
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<th>Section no.</th>
<th>Mean</th>
<th>σ</th>
<th>Max</th>
</tr>
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<td>0.185</td>
<td>0.540</td>
</tr>
<tr>
<td>2</td>
<td>0.012</td>
<td>1.022</td>
<td>2.275</td>
</tr>
<tr>
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<td>0.014</td>
<td>1.608</td>
<td>3.894</td>
</tr>
<tr>
<td>4</td>
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<td>2.254</td>
<td>7.294</td>
</tr>
<tr>
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<td>2.567</td>
<td>8.580</td>
</tr>
<tr>
<td>6</td>
<td>0.572</td>
<td>15.61</td>
<td>36.00</td>
</tr>
</tbody>
</table>

the fjord head efficiently. However, the outgoing wave energy flux decays quickly moving out fjord, and by the fjord mouth (Section 4) is significantly smaller that the incoming flux. The time-mean energy flux through the fjord mouth (Section 4) was 2.7 MW (directed into the fjord).

Subgrid-scale mixing parameters were calculated on the 300 m depth level, as this depth corresponds to a local maximum in both vertical and horizontal diffusivity which is not related to the surface or bottom boundary layers (Figure 17). The KPP-generated vertical diffusivity, $\kappa_z$, was greatest near the right-hand boundary near the fjord mouth and increased from a background value of around $3 \times 10^{-4} \text{m}^2 \text{s}^{-1}$ in the fjord interior to around $8 \times 10^{-4} \text{m}^2 \text{s}^{-1}$ following barrier wind events. Horizontal diffusivity, $\kappa_h$, which we recovered from the model-generated biharmonic viscosity according to Fox-Kemper and Menemenlis [2008], equation 33, was more uniformly distributed in both space and time, although values were again higher near fjord side walls where the mean values were around $2 \text{m}^2 \text{s}^{-1}$. We investigated the role of shear dispersion, a process whereby an effective horizontal diffusivity is induced by vertical mixing in a vertically sheared flow. Young et al. [1982] estimate that in an oscillating flow, $\frac{1}{2} (\alpha/\omega)^2 \kappa_z$ is generated by shear dispersion, where $\alpha$ is the maximal velocity shear and $\omega$ is the angular frequency of oscillation. From this expression, we found that the mean horizontal diffusivity increased by $0.6 \text{m}^2 \text{s}^{-1}$ due to shear dispersion at 300 m depth, effectively doubling the mean value. Spatial patterns in shear dispersion are inherited directly from those in $\kappa_z$, resulting in a much greater contribution towards horizontal diffusivity at the right-hand boundary where values reached $100 \text{m}^2 \text{s}^{-1}$.

Figure 18(a) shows the time-mean melt pattern on the ice face at the northern boundary of KF. Melting is small in the upper layer and increases with depth, peaking at 350 m where the time-mean melt rate is $0.21 \text{m d}^{-1}$. Melting is also weaker at the lateral boundaries of the ice face so that strong melting is concentrated in the middle of the ice face, where the melt rate reaches a maximum of $1.0 \text{m d}^{-1}$. This is likely due to an implicit dependence on flow speed in the adjacent cells, as faster flow tangent to the terminus will lead to a greater heat supply to the ice. We generated a time series of face-averaged melting over the course of the simulation (Figure 18(b)). Variability in melting occurs on timescales of 2-4 days, corresponding to the dominant period of the internal wave field. We find a correlation coefficient of $r = 0.86$ between time series in face-averaged melt rate and adjacent flow speed, while $r = 0.30$ between melt rate and adjacent temperature. Although parametrised melt rate is explicitly dependent on the temperature adjacent to the ice face [Jenkins, 2011], the range of temperatures in direct contact with the ice is relatively small. Instead, the large changes in flow speed at the head of the fjord make this the dominant control over melting in the model.
The melt rates were spatially integrated to find the total volume melted per unit
time, \( \frac{dV}{dt} \), which was then converted into an effective heat delivery from the ocean to
the ice sheet using

\[
Q_i = \frac{dV}{dt} \rho_i (C_i \Delta T + L_i)
\]  

(3)

where \( \rho_i = 930 \text{ kg m}^{-3} \) is the density of ice, \( C_i = 2100 \text{ J kg}^{-1} \text{ K}^{-1} \) is the spe-
cific heat capacity of ice, \( \Delta T = 10 \text{ K} \) is the temperature below freezing point of the glacier,
and \( L_i = 334,500 \text{ J kg}^{-1} \) is the latent heat of melting ice. We obtain \( Q_i = 1.7 \text{ GW} \)
and \( Q_{i,\text{max}} = 4.6 \text{ GW} \), indicating that over half of the net northward heat supply through
Section 1 (Table 1) goes towards melting ice. That the maximum value is two orders of
magnitude smaller than the maximum advective heat flux through Section 1 highlights
the large temporal variability in \( Q_{\Theta} \).

4 Discussion

4.1 Cross-shelf Transport

The SSH and velocity fields on Section 6 show the mean flow to be barotropic, with
cyclonic circulation (> 50 cm s\(^{-1}\)) supplying shelf waters to the fjord mouth (Video 1, supplementary material). Although the variability in along-KT transport appears rel-
atively unaffected by wind activity, barrier wind events generally coincide with local max-
ima in exchange, \( Q \) (Figure 12, Section 6), and local minima in heat flux, \( Q_{\Theta} \) (Figure
13, Section 6). We interpret this as a first-order response to the offshore barotropic pressure
gradient caused by shoreward Ekman transport. The resulting offshore current super-
poses with the cyclonic pattern in KT, temporarily weakening the inflow on the eastern
side of KT while strengthening the outflow (Figure 7). This provokes a decrease in
northward net heat transport in KT (Figure 14, Section 6).

Following Cluster B, the shoreward heat flux through Section 6 remains positive
throughout the first half of January (Figure 13). This period also corresponds to a small
but sustained increase in cross-shelf exchange at Section 6 (Figure 12). The two wind
events in Cluster B are maxima of long periods of generally increased wind speed, and
are hence different in character to most other barrier wind events during the simulation,
which were typically stronger, shorter gusts with a lifespan of around 2 days. For exam-
ple, the second wind event in Cluster B occurred during the longest uninterrupted spell
of northeasterly wind speeds in excess of 10 m s\(^{-1}\) during the record, which lasted 3.75
days. The wind events during Cluster B were also weaker than many others during the
simulation (Figure 3(a-b)), barely meeting the criteria of 20 m s\(^{-1}\) set by Harden et al.
[2011]. We investigated this further by decomposing the variability of the 10 m north-
easterly wind component (Figure 3(a)) into frequency space, once again using wavelet
analysis, which confirmed that Cluster B coincided with the most significant low-frequency
variability during the record (late December/early January, Figure 19).

Nilsen et al. [2016] describe a mechanism whereby along-shelf winds (with coast
to the right) strengthen the cyclonic circulation within cross-shelf troughs and force the
currents to follow shallower isobaths in order to conserve potential vorticity. This relies
upon quasi-geostrophic theory and hence holds when the wind forcing is steady, or varies
on subinertial timescales. Assuming that the KT system behaves similarly, such a mecha-
nism would explain the period of sustained positive heat flux through KT in the first
half of January, when along-shelf winds were sustained and exhibited low-frequency vari-
ability. With temperatures in northern KT and the fjord mouth region increased, sub-
sequent fjord-shelf exchange would have led to warming of the KT interior, as seen through-
out January in Figure 10.
We therefore assert that, based on the model presented here, the response of the shelf circulation to wind forcing may be partitioned into two contrasting regimes: short (≤ 1 day), strong gusts of along-shore wind act to disrupt cross-shelf transport in KT by altering the barotropic pressure gradient on inertial or superinertial timescales while, conversely, lower-frequency (or sustained) wind forcing provides sufficient time for the cyclonic circulation to adjust to the increased barotropic pressure gradient. In the second case, the enhanced barotropic circulation in KT acts to increase cross-shelf delivery of AW.

4.2 Circulation in the Fjord Interior

The EOF patterns in Figure 8 are symptomatic of CTW activity, due to the intensification on flow variability towards the eastern side. Furthermore, the current time series (Figure 9) showed that, following barrier wind events, the velocity near the eastern side of the fjord broadly follows a highly prolate ellipse, again characteristic of CTW activity. We hence suggest that information about on-shelf wind variability propagates into the fjord interior in the internal wave field via CTWs. Video 1 (supplementary material) gives a qualitative description of the CTW structure, as vertical displacements in the $S_A = 34 \text{ kg} \cdot \text{kg}^{-1}$ isohaline surface (representative of the pycnocline) can be seen propagating from the shelf into the fjord along the right-hand boundary of the fjord mouth. Based on the model presented here, CTWs are the dominant mechanism for exchange in KF during the winter.

The model-generated mean heat flux values (Table 1) were generally consistent with FI18, further constraining estimates of the oceanic contribution to melting at KG during the winter months. The maximal values of 2.2 TW in the mid-fjord (Section 2) and 0.5 TW at the fjord head (Section 1) are greatly in excess of observed values, which were taken in summer. Inall et al. [2014] reports 0.26 TW through an equivalent Section 2, while Sutherland et al. [2014] report 0.003 and 0.19 TW through equivalent Sections 1 and 2 respectively. The high temporal variability, associated with CTW activity, highlights the danger in taking synoptic sections of broad fjords as representative of the mean flow. The strong resemblance between Figure 6 of Inall et al. [2014] and the EOFs shown in Figure 8 leads author MEI to re-emphasise that although in geostrophic balance, the reported heat transport value of 0.26 TW from Inall et al. [2014] should be interpreted as a snapshot of sub-intertial variability around an unknown mean. This consistency between modelling and observational results further validates the model, and also indicates that CTWs influence the KF circulation in summer.

The broadly similar temporal patterns in the volume and heat flux time series at each section (Figures 12 and 13) indicate coherent communication between fjord and shelf. Although changes in shelf temperature are quickly manifest in the fjord mouth and interior, the lag times between sections suggest that information of lower layer inflow/outflow propagates up-fjord in the internal wave field as opposed to anomalous warm or cold patches advecting from KT to the head of KF. The (time-mean) temperature field shows an along-fjord temperature gradient in the lower layer (Figure 5), consistent with the order-of-magnitude decrease in the scale of heat flux variability between the fjord mouth (Sections 4 and 5) and the head of the fjord (Section 1). The decay in the heat-flux signal is hence greater than the decay in wave-energy moving up-fjord (Figure 15). In the mid-fjord, along-fjord advection mediated by the internal wave field is associated with high-frequency variability in the heat content of the water column throughout the simulation (Figure 10). However, it is following the low-frequency Cluster B wind events, when an abundance of AW was present in northern KT, that lower-layer temperatures are seen to increase most significantly and enduringly (Figure 10).

In accordance with intermediary circulation as outlined by Straneo et al. [2010] for SF, barrier winds initially produced a negative heat flux in the fjord interior due to upper-
layer inflow, which model animations reveal to be a redirected branch of the EGCC (Video 2, supplementary material). This is followed by a positive contribution from lower-layer inflow (Video 2, Figure 13), and the expelled water in the upper layer rejoins the cold, coastal current.

Due to the steep topography and high stratification within KF, the CTWs are in a high Burger Number regime and are hence approximated as nondispersive Kelvin waves at a flat wall [FI18, Støylen and Weber [2010]; Inall et al. [2015]]. Wave propagation is therefore focussed strictly along-fjord, driving a sustained oscillation in along-fjord heat flux. The CTWs which mediate the heat exchange dissipate slowly in comparison to the internal waves on the shelf (Figure 13), where the weaker stratification and shallower slope may introduce nonlinear effects and permit greater dispersion.

The opposing pattern found near the western bank of Section 1, EOF 1 (Figure 8) is likely caused by the outgoing wave, propagating outwards after being topographically steered around the head of the fjord. As CTW amplitude decays on an $e$-folding length-scale of $L_R$, amplitudes are still significant at the opposing fjord wall and the incoming and outgoing waves may therefore interact. The inflow variability is dominated by the predicted resonant period band (Figure 11), indicating that this interaction plays a role in determining the timescales for water mass exchange.

While FI18 argued that KF was a broad fjord, evaluation of $L_R$ here suggests KF may be classified as an intermediate case between broad and narrow fjords, and hence displays both broad- and narrow-fjord behaviour. The horizontally sheared mean flow through all cross-fjord sections reaffirms the assertion, made by FI18, that rotational effects are important, as expected in a broad fjord. At the same time, the strong response around the resonant frequency would not be anticipated in the $L_R/W << 1$ regime ($W$ is the fjord width), where the incoming and outgoing waves are spatially distinct and hence cannot interfere significantly. In our case, with $L_R/W \approx 1$, CTW amplitudes are significant at the opposing fjord boundary, so that some resonance is expected (Figure 20).

### 4.3 Mixing in the Fjord Interior

The reversible nature of intermediary circulation (mediated here by CTWs) means that for the process to generate non-zero time-integrated heat flux requires some mixing in the fjord interior (excluding any heat lost to melting at the terminus), while the temporal divergences of the isotherms in Figure 10 imply that periods vertical mixing between the PW and AW layers do occur. The cross-fjord structure of the wave energy flux (Figure 15) shows that the incoming wave dominates over the outgoing wave, indicating that intermediary circulation driven by CTWs is a non-adiabatic process where wave energy is lost to mixing. The net wave energy flux into the fjord implies that 2.7 MW is available for mixing within the interior.

Integrating the buoyancy flux, $\rho c_z N^2$, over the fjord interior, we obtained a time-mean value of 1.25 MW. This represents the mean rate at which the potential energy of the water column increased due to water mass transformation. Given the 2.7 MW of net wave energy into the fjord, this implies a mixing efficiency of 0.46. This exceeds the typical literature value of 0.2 [Gargett, 1984], and is significantly higher than the value of 0.06 proposed for fjords by Stigebrandt [2012]. This high value is a result of the KPP mixing scheme, which has previously been found to be overly diffuse in shallow or coastal regions [Durski et al., 2004]. Nonetheless, we expect that 0.46 gives a reasonable upper bound on the efficiency of internal wave-driven water mass transformation within a SE Greenland fjord.

The increase in the strength of overturning moving out-fjord from the glacier front (Figure 6) implies that water mass transformation in the fjord interior is as significant as that driven by plume dynamics at the terminus. The marked increase in overturning
strength between Sections 2 and 3 is therefore attributed to significant vertical density
flux via diapycnal mixing. The two side fjords in this region provide additional topographic
boundaries for CTWs to follow, increasing the area available for mixing at side walls (Figure 17), and the complex coastline drives mixing around features such as headlands. As
diapycnal mixing is fed by incoming internal wave energy, this result indicates that CTWs
act to increase the overturning circulation. This effect is likely exaggerated by a factor
of two or more as a result of the high mixing efficiency in the model.

4.4 Melting at the Glacier Terminus

Cluster B coincided with the highest melt rates in the simulation (Figure 18). The
high melt rates preceded the large increase in heat content within KF (Figure 14), indi-
cating that they are triggered by increased flow speed due to CTW propagation as op-
posed to increased temperature. Given the close correlation between melt rate and ad-
jacent flow speed in the model, another potentially important factor is the capacity for
CTWs to induce energetic flow in the upper reaches of the fjord. The exchange flows trig-
erged by barrier wind forcing were in general found to decay considerably between the
mid-fjord and the fjord head, while the exchange flows triggered by Cluster B remained
highly significant at Section 1 (Figure 12). Theory shows maximum particle speed to be
linear in amplitude for long waves [Cushman-Roisin and Beckers, 2011], and we there-
fore attribute the strong melting to the large CTW amplitudes during Cluster B (Fig-
ure 10) as opposed to associated low-frequency \(\Omega\) signal which continued throughout
the first half of January.

The strong link between melting and adjacent horizontal velocity in the model is
a result of the increased heat delivery associated with faster flow. This behaviour is not
expected in previous, two-dimensional descriptions of intermediary circulation [Straneo
et al., 2010], where the along-fjord oscillation of warm water does not necessarily replen-
ish the supply of heat energy available to melt ice [Cowton et al., 2016]. The cyclonic
background circulation described in this study, however, suggests there is a steady sup-
ply of heat towards the ice sheet. The lower-layer inflow phase of a CTW therefore draws
water from an upstream heat reservoir which has yet to contact the ice. Furthermore,
as FI18 explained in detail, CTWs lead to an accelerated cyclonic background flow as
a result of Stokes’ Drift [Stokes, 1847].

While the heat delivery to the ice sheet, \(Q_i\), was consistent with the mean advec-
tive heat flux towards the glacier, \(Q_{\Theta}\), the modelled melt rates were two orders of mag-
nitude smaller that the glacial flow speed at KG during 2007-08, which was around 25m d\(^{-1}\)
[Bevan et al., 2012]. Our results therefore appear to suggest that ocean-driven melting
during the winter was not capable of matching the rapid flow speeds observed during this
period. We suspect, however, that our model under-represents the oceanic contribution
to KG frontal ablation. The iceplume package was utilised to provide a heat sink at the
head of the fjord and add a level of realism to hydrography in the far field. As this is
primarily a study of shelf-driven exchange, the model lacks the sophistication to produce
realistic glacier diagnostics. Due to the static ice face geometry, the model cannot ac-
count for the triggering of calving events or instabilities in glacial flow due to ocean-driven
melting at the terminus. The pattern of melting found on the ice face (Figure 18a) would
in reality drive undercutting and hence encourage calving events. Furthermore, the flat
ice face likely does not affect the adjacent flow realistically, as tidewater glacier termini
are typically cracked and uneven over small spatial scales. This may have caused the model
to exaggerate the relative influence of adjacent flow speed over temperature.

5 Summary

A high-resolution numerical model of KF and the adjacent shelf region during win-
ner 2007-08 shows coherent communication between fjord and shelf, with temperature
changes on the shelf able to influence the fjord interior. AW is delivered from the shelf break towards the fjord by the geostrophically balanced cyclonic circulation in KT, which is driven by sea surface tilt. The mean circulation structure in KF is similar, though weaker and with a larger baroclinic contribution, and delivers heat to the glacier terminus due to mean cross-fjord temperature gradients. Water mass transformation due to melting at the glacier front and mixing in the fjord interior adds a buoyancy-driven overturning component to the circulation, although it is the horizontal shear which dominates the mean flow. CTWs, which are instigated by barrier winds on the shelf, emerge as the dominant mode of variability within the fjord and drive greatly enhanced along-fjord volume and heat transport. CTWs also act to enhance both the buoyancy-driven overturning circulation, via diapycnal mixing, and the cyclonic background flow, via Stokes’ Drift. The mechanism has previously been observed in a broad, glaciated fjord in Svalbard and is likely to play a significant role in broad fjords in general.

SE Greenland coastal waters have warmed in recent years, and we have demonstrated here that barrier wind-driven CTWs have likely played a crucial role in communicating this ocean warming to the GrIS. The efficacy of CTWs in delivering heat towards the KG terminus, in a time-mean sense, is highly dependent on the temporal variability of barrier wind forcing. Typically barrier wind events are short and strong, ramping up quickly and exceeding the 20 m s\(^{-1}\) threshold for only ~6 h. However, this class of wind forcing was not found to significantly increase fjord heat content. Long-duration northeasterly wind forcing was found to strengthen the barotropic circulation in KT, increasing AW transport towards the fjord mouth, while provoking low-frequency CTWs which are highly effective at drawing these waters up-fjord. This result points to barrier wind duration, as opposed to strength, as the controlling parameter on the wintertime heat delivery towards the GrIS.

The results indicate that significant oceanic heat (~0.5 TW) is regularly delivered from the shelf to glacier terminus during winter. The CTW exchange process is driven purely by shelf exchange and, although there may be some freshwater runoff in the winter months, this is not a necessary condition for this magnitude of heat exchange. We have encountered strong evidence that the processes occurs in the non-winter months, though it is likely weaker and may be obscured or augmented by increased freshwater-driven overturning. Further research is required to fully understand the interaction between these two circulation schemes.

While the model was able to provide diagnostics for melt rate at the KG terminus, yielding a mean melt rate of 0.21 m d\(^{-1}\) at the centre of the ice face, the simplified parameterisation was unable to describe the glacial impacts of ice-ocean interaction in detail. Coupled ice-ocean models, capturing glacier dynamics, calving, ice face texture and marine icebergs, are needed to significantly further our understanding of the rapid acceleration and retreat of Greenland’s tidewater glaciers.

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Figure 6. Mean overturning streamfunction at Sections 1-4. Positive here indicates up-fjord transport.
Figure 7. Hovmöller diagram showing the SSH anomaly at Section 6 over the course of the simulation. Solid (dashed) black lines denote positive (negative) DAC contours (cm s$^{-1}$), while the bold black lines denotes zero DAC. The bar down either side denote barrier wind activity.
Figure 8. EOF 1 at Sections 1 and 3, and EOF 2 at Section 2, accounting for 30%, 29% and 46% of the velocity variability at Sections 1-3 respectively. Here red opposes blue, while white represents no motion.

Figure 9. Horizontal velocity time series, averaged over the deep layer inflow region of Section 2. The y-axis represents along-fjord velocity (normal to section) and the x-axis shows across-fjord velocity (parallel to section). The grey bars at the surface denote barrier wind events on the shelf.

Figure 10. Temperature profile time series near the eastern end of Section 2. The grey bars at the surface denote barrier wind events on the shelf.
Figure 11. (a) The local wavelet power spectrum from velocity variability at the Section 2 lower layer inflow region, thick black contours enclose regions of 95% confidence or greater while the region below the dashed line is the cone of influence, where we expect edge effects to become important; (b) the Fourier power spectrum, where the dashed line represents the 95% confidence level; (c) Frequency-averaged wavelet power, with the dashed line representing the 95% confidence level. The greyed-out regions denote periods considered barrier wind events.

Figure 12. Volume exchange through each of the standard cross sections of the KF/KT system. Note the different ordinate scales.
Figure 13. Heat flux through each of the standard cross sections of the KF/KT system. Note the different ordinate scales.

Figure 14. Time-integral of heat delivered through each of the standard cross sections of the KF/KT system.
Figure 15. Depth-integrated, time-averaged wave energy flux through each cross-section, with positive values indicating energy flux into the fjord. Note that the x-axis scaling differs between panels.
**Figure 16.** Time-averaged wave energy flux through Sections 1, 2 and 3, with positive values indicating energy flux into the fjord and a dashed line denoting the zero contour.

**Figure 17.** Mean subgrid-scale horizontal diffusivity at 300 m depth (a) from model generated fields and (b) from shear dispersion (vertical diffusivity can be recovered approximately by dividing the shear dispersion values by 2500). (c) Corresponding timeseries of the spatially averaged values.
Figure 18. (a) Time-averaged melt rate simulated at the glacier terminus during DJF 2007-08. (b) Time-series of spatially averaged melting.

Figure 19. The local wavelet power spectrum from the NE component of 10 m wind velocity over the deepest point in KT.
Figure 20. Schematic showing CTW activity in an idealised, Northern Hemisphere fjord where $L_R/W \approx 1$. The neutral pycnocline height is indicated in dashed black. The red (blue) arrow indicates the propagation of incoming (outgoing) wave energy, and is located at the wave amplitude maximum. Solid red (blue) lines indicate the cross-sectional incoming (outgoing) wave envelope. Notice that wave energy decay effectively moves the interference zone (where the red and blue lines intersect) to the left of the fjord centreline. The dashed red lines show the longitudinal structure of the incoming wave at the right hand boundary, with black arrows denoting the associated velocities in the PW and AW layers. Note that these velocities oppose each other and reverse over the course of a wave cycle. The yellow arrows represent shear-driven diapycnal mixing and associated strengthening of the overturning circulation.