Ocean Circulation Connecting Fram Strait to Glaciers off North-East Greenland: Mean Flows, Topographic Rossby Waves, and their Forcing

Andreas Münchow*

*Corresponding author address: Andreas Münchow, College of Earth, Ocean, and Environment, University of Delaware, Newark, DE 19716.

E-mail: muenchow@udel.edu

University of Delaware, Newark, Delaware

Janin Schaffer

Alfred Wegener Institute, Bremerhaven, Germany

Torsten Kanzow

Alfred Wegener Institute, Bremerhaven, Germany
ABSTRACT

From 2014 through 2016 we instrumented ∼80 km wide Norske Trough near 78° N latitude that cuts across the 250 km wide shelf from Fram Strait to the coast. Our measurements resolve a ∼10 km wide bottom-intensified jet that carries 0.27 ± 0.06 Sv of warm Atlantic water from Fram Strait towards the glaciers off North-East Greenland. Mean shoreward flows along the steep canyon walls reach 0.1 $m s^{-1}$ about 50 m above the bottom in 400 m deep water. The same bottom-intensified vertical structure emerges as the first dominant empirical orthogonal function that explains about 70-80% of the variance at individual mooring locations. We interpret the current variability as remotely forced wave motions that arrive at our sensor array with periodicities longer than 6 days. Coherent motions with a period near 20 days emerge in our array as a dispersive topographic Rossby wave that propagates its energy along the sloping canyon towards the coast with a group speed of about 63 km per day. Amplitudes of wave currents reach 0.1 $m s^{-1}$ in the winter of 2015/16. The wave is likely generated by Ekman pumping over the shelfbreak where sea ice is always mobile. More than 40% of the along-slope ocean current variance near the bottom of the canyon correlates with vertical Ekman pumping velocities 180 km away. In contrast the impact of local winds on the observed current fluctuations are negligible. Dynamics appear linear and Rossby wave motions merely modulate the mean flow.
1. Introduction

Warm Atlantic waters accelerate the melting of Greenland glaciers (Holland et al. 2008; Straneo and Heimbach 2013; Johnson et al. 2011; Mayer et al. 2018). While the details of ocean melting vary from glacier to glacier (Jackson et al. 2014; Carroll et al. 2016; Washam et al. 2019), ocean processes contribute to retreating glaciers as heat is advected across continental shelves (Sutherland et al. 2014; Inall et al. 2014; Jackson et al. 2014). We currently do not know how the heat from the deep Atlantic Ocean crosses shallow continental shelves to reach Greenland’s coastal glaciers. Presenting new observations from North-East Greenland, we demonstrate that subsurface jets over sloping bottom topography advect heat through canyons towards marine terminating glaciers in North-East Greenland such as Nioghalvfjerdsbrae (Mayer et al. 2000; Schaffer et al. 2017) and Zachariæ Isstrøm (Mouginot et al. 2015).

Submarine canyons often enlarge across-shelf exchange of mass, buoyancy, and vorticity at both polar (Münchow and Carmack 1997) and mid-latitudes (Freeland and Denman 1982). A cyclonic flow over a canyon on the shelf off Vancouver Island, Canada, for example, caused high algae blooms near the surface due to a convergence of the near-bottom frictional Ekman layer and attendant upwelling of nutrient-rich waters into the upper layer (Freeland and Denman 1982). Allen (1996) and Allen and Hickey (2010) simulated the physics of canyon upwelling while Hickey and Banas (2008) explained high productivity off the northern US-Canadian West Coast by interactions of the wind-driven California Current, buoyancy-driven Strait of Juan de Fuca outflow, and canyon dynamics. Coastal Greenland, too, features wind- and freshwater-driven currents (Sutherland and Pickart 2008; Münchow 2016; Hävik et al. 2017) along with prominent canyon systems (Arndt et al. 2015) that amplify across-shelf exchanges (Sutherland and Cenedese 2009). Off North-
East Greenland such exchange moves subsurface warm offshore water towards coastal glaciers via Norske Trough (Schaffer et al. 2017) which cuts across a broad continental shelf.

Bourke et al. (1987) first described surveys of the continental shelf off North-East Greenland in 1979 and 1984 from coastal glaciers to Fram Strait near 79° N latitude. Belgica Bank emerged as a shallow submarine shoal between 78-80° N latitude, Norske Trough as a 400 m deep canyon to the south and west of Belgica Bank (Fig. 1). Wadhams et al. (2006) noted that the 20-80 km wide canyons were poorly mapped. Nevertheless, early investigators defined water masses, estimated geostrophic velocities (Bourke et al. 1987), and suggested northward flow along the coast (Schneider and Budéus 1995). Shelf waters near 79° N latitude differed from those found farther south near 77° N latitude which led Budéus et al. (1997) to conclude that "... there is no one-directional through-flow of deeper waters in the trough system ...". More recent studies such as Richter et al. (2018) provide synoptic velocity observations that describe a largely barotropic Arctic outflow north of 79° N which transforms into a more baroclinic boundary current near the shelfbreak off North-East Greenland that flows south past Denmark Strait as the East Greenland Current (Håvik et al. 2017).

The first direct ocean current measurements from the shelf and canyon systems of North-East Greenland were reported by Johnson and Niebauer (1995) and Topp and Johnson (1997) during a 1992/93 experiment. Two moorings near 80° N (Topp and Johnson 1997) and survey data sketched out a clockwise circulation around Belgica Bank. These early measurements did not always resolved the internal Rossby radius $R_i \sim 10$ km which scales stratified flows on a rotating earth (Gill 1982). This scale is similar to the width of sloping topography which is small relative to the 80-km wide canyon. Our 2014-16 mooring and survey data resolve ocean currents at this scale across both slopes of the wide canyon.
Sea ice always covers the ocean off North-East Greenland (Reeh et al. 2001). During the 20th century the continental shelf featured large areas of fast-ice throughout the year (Hughes et al. 2011). The general clockwise circulation around Belgica Bank (Budéus and Schneider 1995) under the fast-ice then created a polynya to the north of the fast-ice whose biological productivity functioned as a carbon sink (Yager et al. 1995). More recently, satellite imagery shows that this landfast ice breaks up for about two months in summer of the 21st century (Sneed and Hamilton 2016). Our observations from 2014 through 2016 thus describe ocean currents during both summer and winter seasons when the coastal sea ice is mobile and landfast, respectively.

We use a diverse set of data to describe how ocean currents, temperature, sea ice, and winds are distributed in time and space on the vast continental shelf off North-East Greenland. Ocean current moorings across Norske Trough form the focus of our study that was designed to quantify dominant vertical and horizontal scales. These scales relate to dynamics that must be resolved to adequately describe across-shelf exchange of heat and buoyancy. We quantify the hypothesis that topographic Rossby waves contribute to the temporal variability of bottom-intensified currents (Rhines 1970; Pickart and Watts 1990) and that a spatially variable wind field over the slope of Fram Strait generates such waves with 20 day periods. Alternative generation mechanism exist and we speculate about spatially coherent oscillations with periods of 6 and 11 days.

2. Study area, Data, and Methods

Norske Trough connects deep Fram Strait in the east to coastal glaciers in the west. The trough starts as broad topographic depression near longitude 10° W, narrows to 80 km near Isle de France at latitude 77° N, and narrows farther to 30 km adjacent to a chain of islands (Fig. 1). The canyon cuts across the 300 km wide continental shelf (Arndt et al. 2015).
In June 2014 we deployed seven moorings to measure ocean currents and water properties in Norske Trough which serves as a potential pathway from the deep ocean to coastal glaciers off North-East Greenland such as Nioghalvfjerdsbræ and Zachariæ Isstrøm (Fig. 1). We recovered all instruments more than 2 years later in August 2016 (Table 1). Each mooring included 75 kHz Acoustic Doppler current profilers (ADCP) of Teledyne RD Instruments.

We employed two distinct mooring designs (Table 1): Two ADCPs contain batteries inside the pressure housing that also contains electronics and transducer assemblies (Design-A). The two ADCPs were mounted inside a cage shackled to a standard mooring line containing additional instruments. The ADCP can pitch, roll, and swivel on the mooring line. Tilt sensor measure these mooring motions. In contrast, five ADCPs have their transducers placed 3-m above the bottom atop a flat steel buoy (Design-B). External battery cases, acoustic releases, SeaBird SBE37sm Micocats, and additional steel buoys are all mounted within a single stiff frame that can pitch and roll, but not swivel (Münchow and Melling 2008). No additional mooring lines or sensors are above the ADCP transducer heads. The torsionally rigid design minimizes compass errors, if one knows the fixed orientation of the instrument on the sea floor. Appendix-A describes data processing details to move data from both mooring designs into a common earth-referenced coordinate system.

Bottom-mounted ADCP depend on scatterers in the water column to reflect sound back to the transducers to estimate the Doppler shift in the transmit frequency $F_0 = 75 \text{ kHz}$ in vertical bins $D = 8 \text{ m}$. Our broadband ADCPs use two encoded pulses with an ambiguity velocity $V_a = 1.75 \text{ m s}^{-1}$ and beam angle $\theta = 30 \text{ deg}$. Random single-ping velocity errors thus are (RDI 1996):

$$\sigma = \frac{150 \cdot V_a}{\pi} \cdot \left[ \frac{(1/R^2 - 1) \cdot 2C \cdot \cos(\theta)}{F_0 \cdot D} \right]$$

(1)
For our 2-pulse system the ideal correlation $R$ is 0.5, speed of sound $C$ is $1440 \, m/s^{-1}$, and the standard deviation $\sigma$ becomes $0.096 \, m/s^{-1}$. This uncertainty is reduced as $M^{-1/2}$ for $M$ independent pings. The ADCPs of Design-A sampled $M=40$ acoustic pings every two hours while the ADCPs of Design-B sampled 120 velocity pings every half hour. We thus estimate measurement uncertainties of 0.015 and $0.009 \, m/s^{-1}$ for Design-A and B, respectively.

Scatterers appear intermittently during the winter below the sea ice causing gaps in the data. We fill gaps shorter than a day by interpolating velocities in time for each vertical bin. First, we replace all 'bad' velocity values with tidal predictions that we determine from the 'good' values. The tidal fit includes 4 semi-diurnal ($M_2$, $S_2$, $N_2$, $K_2$) and 3 diurnal ($K_1$, $O_1$, $P_1$) components.

Second, for every daily interval with more than 8 hourly values, we interpolate missing values for each 24 hour segment after the tidal predictions are removed. The first step fills all gaps with a small albeit deterministic current oscillation. The second step fills gaps shorter than 16 hours with a biharmonic spline using only detided 'good' velocity estimates. The tides are added back after the spline interpolation. This procedure removes all data gaps. In a final step, the time series are filtered with a Lanczos raised cosine filter with a half-power point near 34 hours and a window width of 75 hours to remove tidal and inertial oscillations (Walters and Heston 1982).

These filtered values are then subsampled into daily values that represent subtidal ocean current variability.

Estimating uncertainty for speed and direction of ocean currents, we use 95% confidence levels (Table 2). Speed estimates are assumed to originate from a Gaussian distribution where velocity vectors have a decorrelation time scale $T_d$ that varies between 4 and 12 days depending on location. We determine $T_d$ from a lagged auto-covariance function that we integrate to its first zero crossing following procedures introduced by Kundu and Allen (1976). Uncertainty in directions are 95% confidence limits from directional statistics that use the von Mises’ probability distribution on a
unit circle from which a mean resultant length $R$ of current vectors on a unit circle (Mardia 1972) are used.

During mooring deployment in June 2014 and recovery in August 2016 we measured ocean water properties with a SeaBird 911+ conductivity-temperature-depth (CTD) sensor package that was calibrated at the factory a few month prior and after each expedition (Kanzow et al. 2017; Schaffer et al. 2017). Salinity estimates were compared at sea against measured salinities from discrete water samples via salinometer.

Atmospheric observations are available from the Danish Meteorological Institute at Station Nord ($81.6^\circ$ N, $16.7^\circ$ W), Hendrik Krøyer Holme Island ($80.7^\circ$ N, $13.7^\circ$ W), and Danmark Havn ($76.8^\circ$ N, $18.7^\circ$ W). Fig. 1 shows that both Station Nord and Danmark Havn are located near mountainous terrain and wind vectors from these stations do not necessarily represent conditions on the continental shelf, however, more representative daily wind observations are available from an automated weather station on uninhabited Hendrik Krøyer Holme Island. For the 822 days of our study period from 1 June, 2014 through 31 August 2016 we have 695 daily wind observations at Hendrik Krøyer Holme Island. These wind and air temperature data are accessed from NOAA’s National Centers for Environmental Information as hourly values (https://www.ncei.noaa.gov/data/) that we bin into daily averages. The observations compare favorably to estimates from the European Center for Medium-Range Weather Forecasting ERA-Interim Reanalysis products (Dee et al. 2011) of 10-m wind vectors and 2-m air temperatures (not shown).

Remotely sensed surface conditions originate from daily observations of microwave SSM/I and optical MODIS satellites, respectively. We use SSM/I sea ice concentrations data from NSIDC archives at 25 km resolution (Steffen and Schweiger 1991) while MODIS data are surface reflectance at 865 nm obtained from NASA’s Goddard Space Flight Center at 250 m resolution.
3. Sea Ice and Atmospheric Conditions

In polar regions the distribution of sea ice in time and space determines how the atmosphere forces the ocean in at least two ways. Mobile sea ice can enhance the frictional coupling and thus momentum flux between atmosphere and the ocean, because the roughness elements of ridged sea ice are larger than that of an ice free sea surface (Lübkes and Birnbaum 2005; Martin et al. 2014; Schulze and Pickart 2012). In contrast when densely packed sea ice becomes less mobile, then it can shut down local wind forcing completely as it does in winter over coastal areas of the North-East Greenland continental shelf (Hughes et al. 2011). Hence we next provide a description of sea ice and winds to reveal the seasonality that impacts our ocean current observations.

Sea ice covers most of the North-East Greenland shelf for most the year. Fig. 2a shows a time series of daily sea ice concentrations. We quantify the sea ice cover by averaging daily SSM/I passive microwave data over an area that includes our mooring locations (Fig. 1). From October through June in 2015 and 2016 we find that about 85% of our study area is covered by sea ice (Fig. 2a). The sea ice minimum occurs the last two weeks in August when only about 20% of our area is covered by mobile sea ice (Fig. 3). Ice-cover diminishes slowly for 2 months in June and July as a result of sea ice melting and dispersal, but it reforms rapidly within 2 weeks in October of 2014 and within 4 weeks in September and October of 2015.

Visualizing the spatial sea ice distribution, we show in Fig. 3 two MODIS Terra images. The first image on 15 June 2014 coincides with the mooring deployment in partially open water with many large and mobile floes. About 80 km to the north and west, however, the ice cover is 100% and immobile, that is, landfast sea ice covers a coastal area about 100 km wide and 250 km long between 78° and 80° N latitude. We indicate this area by the yellow contour in Fig. 3 that we also use as an estimate of maximal extent of the landfast sea ice. In years past, this landfast ice
survived the entire summer (Hughes et al. 2011), however, this is no longer the case (Hughes et al. 2011; Sneed and Hamilton 2016). The image of 30 August 2014 depicts a small remnant of this landfast sea ice in the form of a roughly 80 km long and 30 km wide segment near 79° N and 14° W (Fig. 3). Most of the shelf area in August, however, is covered by mobile sea ice. Unlike winter, the summer sea ice is mobile as it is advected and dispersed by winds and ocean currents while it melts in response to solar heating.

Air temperatures over the ocean range from -30 °C in October to May and reach +10 °C in July to August (Fig. 2c). We estimate seasonal air temperatures by fitting observations against $T_{air}(t) = T_0 + T_1 \cos(\omega t - \phi)$ where $\omega$ represents the known annual frequency and $T_1$ and $\phi$ the amplitude and phase of the annual cycle. Minimizing the least squares between data and fit, we find that the mean temperature is $T_0 = -10.2 \pm 0.6$ °C that varies seasonally by $T_1 = 10.8 \pm 1.2$ °C. The phase indicates temperature maxima on about 25th July. Uncertainties are 95% confidence limits assuming Gaussian noise and a decorrelation time scale of 4 days.

Observed wind vectors on Hendrik Krøyer Holme Island appear largely in-phase with air temperatures. Largest winds to the south coincide with minimal air temperatures on about 25th of January. Seasonal wind variations have an amplitude of $4.2 \pm 1.8 \text{ m s}^{-1}$ about the mean of $3.5 \pm 0.9 \text{ m s}^{-1}$ along the north-south principal axis of variation (Fig. 2b). Note that observed winds during the cold season frequently exceed $20 \text{ m s}^{-1}$ over the generally land-fast sea ice on the continental shelf. Strong but short-lived warming events occur in winter such as in Nov.-2014, Apr.-2015, and Feb.-2016 when winds from the north reverse and bring warm air from open Fram Strait waters in the south-east.
4. Hydrography

Schaffer et al. (2017) reviews all available hydrography for the entire shelf area to note subsurface temperature maxima in the 800 km long trough system between 76° and 80° N latitude. Averaging data into 100 km bins along the trough system below 200-m depth, they find waters during the 2000-16 period to be about 0.5 °C warmer than those observed during the 1979-99 period. In contrast, we here focus on the vertical and across-slope structure of the temperature and salinity fields. Figs. 4 and 5 show 2014 and 2016 summer snapshots of the local hydrography, respectively, along with measurement locations of both CTD survey and ADCP mooring data labeled M10 through M70 (Table 1).

In both 2014 and 2016 we find the warmest waters below 250 m depth adjacent to the north-eastern rim of the trough system. In June of 2014 (Fig. 4) potential temperatures reach values above +1.5 °C almost across the entire section at depths below 300 m, but temperatures reach +1.7 °C for a CTD cast at km-75. The salinity distributions are almost level across the section except for a 5-km wide area over the north-eastern slope at km-85 where we will find the core of the subsurface jet. Maximal ocean temperatures in August of 2016 are +1.8 °C near 300 m depth (Fig. 5). The subsurface temperature maximum appears as a 15 km wide and 50 m thick intrusion that is attached to the north-eastern rim of the canyon. Furthermore, we find upward sloping salinity contours near km-80 within which the local temperature maximum is embedded. Dynamically this is consistent with a northward geostrophic flow below 150 m depth relative to a weaker flow at the surface. Temperature time series within about 3 m off the bottom (not shown) indicate that these patterns persist throughout the year, that is, seasonal variations are small and negligible.
5. Ocean Circulation

a. Mean flow and volume flux

The data from this array of seven ocean current moorings describe a jet over the north-eastern slope of Norske Trough whose depth- and time-averaged representation we show in Fig. 3 and list with uncertainties in Table 2. Fig. 6 shows the record-mean velocity along the canyon for the three north-eastern mooring sites M50, M60, and M70. The flow constitutes a 10 km wide bottom intensified jet towards coastal glaciers in the north-west. The velocity maximum exceeds 0.08 m s$^{-1}$ near 300 m depth and diminishes towards both the surface and the channel center (Fig. 6). This bottom-intensified flow is the main circulation feature to dominates both the time mean and the time-domain variability.

Spatial patterns coherent to all records from all 7 mooring locations reveal that more than 36% of the total variance is contained within the principal pattern (not shown) that has its maximal value 150 m below the surface at mooring sites M50, M60, and M70 over the north-eastern slope of Norske Trough. The data from the south-western 4 moorings contribute little to this mode. Hence we subsequently focus on the spatial and temporal variability of this north-eastern slope current that advects warm waters of Atlantic origin across the continental shelf of North-East Greenland.

Fig. 7 shows the volume flux $Q(t)$ of Atlantic influenced waters that we define as those with salinities larger than 34 psu. This isohaline is almost level across the section at 150-m depth (Figs. 4 and 5). Estimating volume transport, we generate regular grids of daily velocity $u(x,z,t)$ that we integrate vertically from the bottom at $H(x)$ to 150 m depth and horizontally from $x_1 = 75$ km to $x_2 = 90$ km, e.g.,

$$Q(t) = \int_{x_1}^{x_2} \int_{-H(x)}^{-150} u(x,z,t) \, dz \, dx.$$  \hspace{1cm} (2)
The record-mean volume flux $Q_0$ becomes $0.27 \pm 0.06$ Sv ($10^6 m^3 s^{-1}$). Uncertainties are 95% confidence limits assuming an integral time scale of about 11 days. Also shown in Fig. 7 are the 2-day low-pass and 19-21 day band-pass filtered data. Volume flux values including the temporal mean range from -0.4 to +1.4 Sv for the low-pass and from +0.0 to +0.6 Sv for the band-pass data. Positive flux is to the north-west towards coastal glaciers. Next we detail vertical and lateral structures of this along-slope current.

b. Vertical Velocity Variations

In Fig. 8 we show current vector time series at the core of the slope jet at five selected depths between 39 m and 335 m below the generally ice-covered surface. The flow below 150 m depth is remarkably steady and to the northwest towards the coast. Current reversals are few and short-lived. Seasonal variations emerge and their amplitudes determined from a harmonic analysis (not shown) increase from less than $0.015 ms^{-1}$ near the bottom to almost $0.06 ms^{-1}$ at 143 m (not shown). Strongest seasonal flow to the north occurs in December/January and weaker flow in June/July. Both the time mean flow and variations align along an axis close to 319 degrees clockwise from true North that we subsequently select as the along-canyon direction. We will next quantify mean, seasonal, and residual sub-inertial currents. The latter we will decompose into dominant empirical orthogonal modes.

Fig. 9 depicts the record-mean vertical profile at M60 separated into along-canyon and across-canyon components. Within about 150 m of the surface both along- and across-channel speeds are about $0.04 ms^{-1}$ with uncertainties of about $0.02 ms^{-1}$. The positive across-channel velocity is a flow to the west which is dynamically consistent with the geostrophic adjustment to a frictional Ekman-like response to winds generally blowing from the north, e.g., Fig. 2b. At depths below 150 m the across-channel velocity component vanishes while the along-channel component in-
creases to a maximum of $0.10 \pm 0.01 \, \text{m s}^{-1}$ near 325 m depths. Below this depth currents along the topography reduce towards the bottom with a small across-channel component down-slope that is dynamically consistent with a frictional Ekman-like response also, but the down-slope flow here is imposed by the along-channel current above.

The temperature structure on both 10-June-2014 and 16-August-2016 reveal 150-m thick surface layers that consists of uniformly cold (-1.6 °C) waters where salinity stratifies the water (Figs. 4 and 5) and the time mean flow is dominantly across topography. Below this halocline layer we find a thermocline layer between 150 and 300 m depth that has a constant temperature gradient of about 2.8 °C per 150 m. Vertical current shear is largest within this thermocline layer as we approach both the vertical temperature and velocity maximum below 300-m depth. Here the waters are uniformly warm, (+1.3 °C) and salty (34.8 psu). The core of the the bottom-intensified jet thus advects heat from Fram Strait across the shelf towards the coast.

c. Time-dependent Currents

Removing the mean currents at each depth, we next describe principal patterns of variability in the along-channel direction with empirical orthogonal functions (EOF) as these constitute the most efficient presentation of variability (Davis 1976). More specifically, EOF analyses quantify the distribution of variance among a set of mutually uncorrelated patterns in space and how these spatial patterns vary in time. We decompose 45 time series of along-channel velocity observations $u(z,t)$ at M60 (see Fig. 9 for locations) into $M=45$ modes via the linear transformation

$$u(z_i,t) = \sum_{n=1}^{M} A_n(t) \cdot \theta_n(z_i)$$

(3)
where $A_n(t)$ and $\theta_n(z_i)$ denote the temporal amplitude and vertical pattern of the n-th EOF mode, respectively. We scale $A_n(t)$ and $\theta_n(z_i)$ such that $A_n(t)$ is dimensionless with variance 1 and $\theta_n(z_i)$ has units of cm s$^{-1}$. Appendix-B explains the details.

From 45 distinct eigenvectors $\theta_n(z)$ we find that $\theta_1$, $\theta_2$, and $\theta_3$ dominate as they explain about 95% of the total velocity variance. Fig. 10 shows these three dominant modes and we limit our discussion to them. The first mode explains 77% of the variance and its vertical structure resembles the time-mean bottom-intensified flow. This first EOF has a peak velocity of about 0.08 m s$^{-1}$ at 180-m depth with strong vertical shear towards zero above this depth and little shear towards the bottom (Fig. 10a). The flow within 100-m of the surface does not contribute much to this EOF which contrasts with the next two modes. Mode-2 reflects a balanced exchange flow for which surface and bottom currents are anti-correlated, that is, a flow towards the coast at the surface coincides with a bottom flow in the opposite direction and vice versa. This exchange mode captures about 15% of the total variance. An intensified surface flow in the upper 100-m of the water column is described by mode-3, however, this mode contains only 4% of the total variance and thus can be neglected along with higher modes not shown.

We note that the time scales of the bottom-intensified flow $A_1$, the exchange flow $A_2$, and the surface intensified flow $A_3$ differ. Fig. 11 presents auto-spectral estimates of the temporal variations $A_1(t)$ and $A_2(t)$ that for $A_1$ reveal three clear spectral peaks near 6, 11, and 20 day periodicities. Signals with these time scales each contain more than 15% of the total variance and represent the principal time scales at which the bottom-intensified jet varies about its mean. The second mode $A_2$ has a single distinct peak near 20 days and contains most of its variance at periodicities larger than 60 days (0.016 cycles per day or cpd below). We speculate that seasonally variable sea-ice and winds causes these variations expressed in $A_2$ and $\theta_2$. Identical analyses for the data from adjacent upslope and downslope moorings M70 and M50 reveal similar results (not shown).
find similar vertical flow structures and time scales of variability from EOF analyses. We next
investigate the across-canyon correlation of these flow structures with a linear systems analysis
that will reveal wave-like properties of observed current fluctuations.

d. Topographic Rossby Wave

We estimate spectral properties from 2-year time series at M50 and M60 at 309-m and 303-m
depth, respectively, because these two time series are both within the core of the warm subsurface
jet (Fig. 6). A frequency-domain linear system $Y = H * X$ uses M60 as input $X$ with an auto-
spectrum $G_{xx}$ and M50 as output $Y$ with an auto-spectrum $G_{yy}$. The auto-spectra are real functions
of frequency $f$ that estimate the kinetic energy per frequency interval. In contrast the transfer
function $H = |H|e^{-j\Theta_{xy}}$ is complex and thus carries both amplitude and phase information. The
phase is estimated from the complex cross-spectrum $G_{xy} = \left|G_{xy}\right|e^{-j\Theta_{xy}}$ (Bendat and Piersol 1986).
We use $G_{xy}$ to aid in the interpretation of oscillatory motions with enhanced kinetic energies near
periodicities of 20, 11, and 6 days (Fig. 11). More specifically, across-slope phase relations and
principal axes of variations implicate topographic Rossby waves (Pickart and Watts 1990) as a
possible explanation of the observed current oscillations. We discuss wave properties below after
we quantify the phase propagation across the slope of signals that are coherent at M60 and M50.

Fig. 12 shows the phase $\Theta_{xy}(f)$ for $f \leq 0.3$ cpd. We find $\Theta_{xy} > 0$, that is, oscillations occur first
at M60 before the appear at M50. For example, at 0.05 cpd a phase $\Theta_{xy} = 30$ degrees indicates
that a signal takes 40 hours to reach M50 from M60. This corresponds to phase speed $c_p =
(1/T)(360/\Theta_{xy})(\Delta s/cos\Delta) = 3.2$ km/day for a wave with period $T=20$ days, site separation $\Delta s=5.3$
km, and the angle $\Delta = 1$ degrees between the mooring line (39T) and the orientation of the wave
number vector (Pickart and Watts 1990). The implied wavelength is $\lambda = c_p \cdot T = 63$ km.
How much of the variance of this 63-km across-slope scale at location M60 relates in this “linear wave fashion” to the variance at location M50? Fig. 12 suggests about 70%, because the coherence
\[ \Gamma^2 = \left| G_{xy} \right|^2 / \left( G_{xx} G_{yy} \right) \]
estimates the fraction of the input variance related linearly to the output variance. The coherence \( \Gamma^2 \) reaches almost 0.7 at the relevant frequencies between 0.01 and 0.3 cpd (Fig. 12). This value is far above the 95% confidence level of 0.15. We next compare these coherent observations with predictions from linear wave theory.

The dispersion relation of linear internal waves on a rotating plane in the presence of a bottom slope relates the wave frequency \( \omega = 2\pi/T \) to the wave number \( K = 2\pi/\lambda \) as

\[
\omega(K) = -\frac{N \cdot \alpha \cdot \sin \phi}{\tanh(K \cdot L_D)}
\]

(4)

that defines topographic Rossby waves (Gill 1982; Rhines 1970). Here \( N \) is the stability or Brunt-Väisälä frequency (0.005 s\(^{-1}\)), \( D \) is a vertical scale of motion (400-m), \( \alpha \) is the bottom slope (0.03), and \( \sin \phi = k/K \) indicates the orientation of the wave number vector \( \vec{k} = (k,l) \) with magnitude \( K = \sqrt{k^2 + l^2} \) and components \( k \) along and \( l \) across the bottom topography. The parameter \( L_D = N \cdot D / f_0 = 14 \) km is the internal Rossby radius of deformation and \( f_0 \) is the local Coriolis parameter (0.00014 s\(^{-1}\)).

The waves are transverse, that is, they have \( \vec{k} \cdot \vec{u} = 0 \) where \( \vec{u} \) is flow velocity associated with the wave (Rhines 1970). Thus we can interpret the angle \( \phi + \pi/2 \) as the orientation of the current ellipse (Pickart and Watts 1990; Harden and Pickart 2018). The minus sign in Equation 4 indicates that waves propagate with shallower water to the right in the northern hemisphere, that is, the wave moves its perturbations along \( \vec{k} \) across topography from shallow to deep water. The restoring force is the vortex tube stretching over a sloping bottom in the presence of both rotation and stratification (Zhao and Timmermans 2018).
For the observed wavelength of 63 km and period of 20 days the dispersion relation Equation 4 predicts the angle $\phi = 179$ degrees that is the direction of the phase velocity relative to upslope. We can compare this prediction against the orientation of the observed current ellipse at this frequency, because the wave number is always perpendicular to the current vector. We estimate current ellipses from band-pass filtered velocity data and list both observed and estimated wave properties in Table 3 for waves with periods near 20, 11, and 6 days.

Fig. 13 shows our co-ordinate system that we orient along (x) and across (y) the local topography. We also show current ellipses near 220 m depth at three locations across the slope. The deviation of the semi-major axis of length $R_{maj}$ from the along-slope direction constitutes the observational estimate of the angle $\phi_{obs}$ measured from the across-slope direction. Analytical predictions of $\phi$ from the dispersion relation, e.g., $\phi_{ana}$ agree with those derived directly from the observed, band-passed flow field, e.g., $\phi_{obs}$ (Table 3). The phase velocity vector is almost perpendicular to the topography, that is, the principal axis of current oscillations is almost along topography. This is a characteristic of short and strongly stratified Rossby waves where $tanh(KL_D) \sim 1$ (Pickart and Watts 1990). The bottom slope traps wave energy, that is, velocity variance near the bottom exceeds that near the surface. Velocity profiles vary as $cosh(\mu z)$ in the vertical $z$-coordinate (Rhines 1970) where $\mu^{-1} = D / (KL_D)$ ranges from about 200 to 300 m for our waves (Table 3). Furthermore, we find, as expected from linear wave theory, that the angles $\phi_{obs} = \phi_{obs}(\omega)$ rotate clockwise with increasing $\omega$, i.e., higher frequency motions propagate perturbations with more of an along-slope component than lower frequency motions. We note, however, that all these angles are small as the along-slope component of the wavenumber $k$ is much smaller than the across-slope component $l$, that is, $k \ll l$. 

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6. Forcing

For stratified topographic Rossby waves with \( K \cdot L_D \gg 1 \) the dispersion relation simplifies to

\[ \omega(K) \approx -N \cdot \alpha \cdot \sin \phi. \]

From this approximation we find the group velocity vector

\[ \vec{c}_g = (\partial_k \omega, \partial_l \omega) = c_0 \cdot (-\cos \phi, \sin \phi) \] (5)

where \( c_0 = (N \cdot \alpha \cdot \cos \phi / K) \) and \( \cos \phi = l / K \) (Harden and Pickart 2018). Fig. 13 shows \( \vec{c}_g \) for the 19.3 day oscillation. The speed of the energy flux vector \( \vec{c}_g \) is 63 km/day and it is directed into the Norske Trough system from offshore Fram Strait towards the coastal glaciers in the north-west. Hence we must seek the energy source for our observed wave motions to the south-east of our study area where the trough system broadens and merges with the slope of the outer continental shelf and Fram Strait. Here the ice cover is mobile all year and allows for efficient momentum transfer from the atmosphere to the ocean.

Our observed topographic Rossby waves depend on vertical density stratification \( N \) and a sloping bottom \( \alpha \), e.g., \( \omega = \omega(N, \alpha) \) that both provide the restoring mechanism for wave perturbations via vortex tube stretching in a baroclinic fluid. The curl of the surface wind stress perturbs the density field via an Ekman pumping velocity \( w_e \) (Gill 1982)

\[ w_e = \left( \frac{\partial_x \tau^{(y)} - \partial_y \tau^{(x)}}{\rho_0 \cdot f_0} \right) \] (6)

and this density perturbation, we postulate, then travels as the topographic Rossby wave whose properties we list in Table 3. Appendix C describes details of our wind stress estimation in ice-covered seas that use ERA interim wind and SSM/I sea ice products at daily and 25-km scales.

Fig. 14 shows a snapshot of the spatial wind variability during a particular strong event on 17th December of 2015 when cyclonic winds reach 20 \( ms^{-1} \) in Fram Strait just seaward of the 1000-m isobath. In contrast, near our mooring location wind speeds are less than 5 \( ms^{-1} \). The vertical upwelling is maximal near the continental slope while it is close to zero at our mooring.
site. Bottom-intensified ocean currents at the time are two standard deviations above the mean as expressed in the first EOF at M60 (Fig. 10).

Searching more systematically for a source location of the topographic Rossby wave generation, we correlate our ocean current time series with Ekman pumping time series at all locations. The correlation is done for signals in the narrow [19, 21] day band in the frequency domain where we find enhanced kinetic energy (Fig. 11). The correlation then becomes $\Gamma^2(x,y)$ for $\omega = \text{const.}$ that we map in Fig. 15. While statistically significant coherences $\Gamma^2$ are found scattered throughout the study area, we indeed find the largest coherence $\Gamma^2 > 0.4$ in a 100 km long and 50 km wide band over the shelf break near the 1000 m isobath where Norske Trough connects to Fram Strait.

We will next take a closer look at the data from this location to quantify the distribution of Ekman pumping and how it correlates with velocity fluctuations that include our topographic Rossby waves.

The Ekman pumping velocity $w_e(x = 8^\circ W, y = 76.5^\circ N, t)$ contains enhanced variance at periodicities near 20, 11, 6, and 4.5 days as shown by the power spectral density (Fig. 16). The spectral peaks are similar to those we find in the spectra of bottom currents in Norske Trough (Fig. 11) that correlate across the slope with $\Gamma^2 \sim 0.7$ (Fig. 12). Estimating the coherence of bottom currents in Norske Trough and Ekman pumping near the shelf break about 180 km to the east, we find that the correlated variance at the 20 day time scale exceeds 40%. This is well above the 99% confidence level $\Gamma^2_{99} = 0.23$ for 40 degrees of freedom. Note also that the correlation of the signals with frequencies between 0.15 and 0.07 cycles per day (7-14 days) falls below this significance level. Currents at these frequencies thus are not correlated to Ekman pumping at a 99% level of confidence. Hence we suggest that only low-frequency bottom currents in Norske Trough with periods near 19-21 and 5-6 days are forced remotely by the wind-stress curl near sloping topography of Fram Strait. The kinetic energy generated by the wind stress curl propagates with the group
velocity of a baroclinic Rossby wave along the sloping topography to cause along-slope velocity oscillations at 20 day periodicities.

Furthermore, the correlated variance suggest a linear input-output system: Bottom currents lag the Ekman pumping by about 70 degrees or about 4 days at a 20-day period. This lag translates to a speed of 45 km/day for the 180 km distance between these locations. Visualizing this phase lag, we show in Fig. 17 band-pass filtered time series of Ekman pumping at two select locations (shown in Fig. 14) that we each overlay with similar band-pass-filtered bottom velocity at M60. Prominent bottom velocity oscillations with amplitudes near 0.1 m s\(^{-1}\) emerge in November of 2015 that persist through February of 2016. The visual correlation is strong at the shelfbreak for the \(w_e\) at 7\(^\circ\) W (bottom panel of Fig. 17) while it is weak at the mooring locations near 15\(^\circ\) W (top panel of Fig. 17). The winter 2015/16 Ekman pumping near the 1000 m isobath of Fram Strait has an amplitude of about 0.4 m/day from which our linear system analysis gives a gain (not shown) of 19 (cm/s)/(m/day) for both the 20 and 6 day oscillations. Positive \(w_e\) (downwelling) in Fram Strait leads negative along-slope bottom velocities (landward) in Norske Trough. Note, however, that we present gain, phase, and coherence in the frequency domain and we thus weight all data from the entire 945 day long time series equally.

7. Discussion and Conclusions

We introduced ocean current time series from a single array of 7 moorings that each contained a vertically profiling ADCP that returned data from June 2014 through August 2016. The data described the flow field within an 80 km wide canyon off North-East Greenland that connects Fram Strait offshore with tidewater glaciers inshore. The array sampled a single section across Norske Trough with distinct circulations on each side of the sloping canyon. We focused on a
bottom-intensified jet over the north-eastern slope that advects warm waters of Atlantic origin from Fram Strait across the continental shelf via Norske Trough.

Sub-tidal flows exceed 0.2 $m s^{-1}$ and are remarkably steady especially below 150 m depth where they flux a volume of 0.27 ± 0.06 Sv towards coastal glaciers. Currents are highly correlated both vertically and across the north-eastern slope. Maximal mean values occur 50-100 m above the bottom in 400 m deep water. They are directed along the slope towards the coast in the north-west. A single dominant EOF mode captures about 77% of the variance at each of three mooring locations that describes the sub-surface jet. Jet amplitudes average about 0.1 $m s^{-1}$, but amplitudes twice this value occur 2-3 times during the winter of 2014/15 and 2015/16.

Variability about the seasonal cycle is minimal in summer (Fig. 10) when the sea ice cover is mobile, but local winds are weak. In contrast variability is maximal during December and January when the sea ice is less mobile or even landfast over much of the coastal continental shelf, but both winds and wind-stress curl are maximal, especially seaward from our measurement locations.

Spectral analyses of temporal EOF amplitudes suggest that variance of the subsurface jet has distinct peaks at 20, 11, and 6 day time scales. Furthermore, we find that about 70-75% of the ocean current variance at these time scales correlates across the sloping topography. This high correlation gives stable estimates of the phase relation of oscillatory signals. More specifically, correlated signals propagate from shallow into deep water with speeds ranging from about 3.2 km/day near 20 days to 6.8 km/day near 6 days periodicities (Table 3).

We interpret these propagation speeds as phase velocities of linear topographic Rossby waves (Pickart and Watts 1990; Harden and Pickart 2018) that have wave lengths between 93 km (6.5 days) and 131 km (19.3 days). Comparing estimated angles of the wave vector relative to the orientation of the topography against those predicted from linear Rossby wave theory (Rhines 1970), we find excellent agreement even though both observed and predicted angles are small,
about 1-4 degrees from upslope. The observations also reveal smaller angles for lower frequencies which is a property predicted by linear theory that is not always observed consistently (Pickart and Watts 1990; Kanzow and Zenk 2014). Some of our agreements with theory are likely fortuitous, however, because random errors impact our estimates of the orientation of the wave number vector.

For example, we do not know the orientation of the topography well enough to ensure phase propagation with shallow water on the right as required in the northern hemisphere.

Furthermore, cross-spectral estimates have phase uncertainties larger than 2 degrees and non-steady wave scattering may occur along irregular smaller scale topography. Nevertheless, our quantitative wave estimates are consistent within the array and agree well with linear wave theory that predicts a group velocity of about 63 km/day along the slope towards the coast. This suggests a wave generation source seaward from our mooring site. We identify the generation region as the area of high correlation of wind stress near the surface and ocean currents near the bottom. More than 50% of the Ekman pumping variance correlates with the along slope bottom currents more than 180 km away within the sloping canyon topography.

The dynamics appear linear, because Rossby numbers $Ro = \delta U / (f_0 \cdot \delta y) \ll 1$. Here we use the lateral velocity differences $\delta U$ across the jet ($\sim 0.08 \text{ m s}^{-1}$) over a distance $\delta y$ ($\sim 11 \text{ km}$) to find the lateral shear of about $0.05 \cdot f_0$ where $f_0$ is the local Coriolis parameter. The ratio of relative ($\xi \sim \delta U / \delta y$) to planetary ($f_0$) vorticity is thus small ($Ro \ll 1$) and nonlinear inertial terms in the momentum equations become negligible (Gill 1982). We neglect small variations of across-slope velocity $\delta V$ along the slope. For small Rossby number the potential vorticity

$$PV = \frac{f_0 + \xi}{D} \approx \frac{f_0}{D} \quad (7)$$

where $D$ is the vertical scale of motion. For unforced motions PV is conserved and the flow follows contours of $f_0 / D$. Such flows are said to be topographically steered.
For barotropic motions $D$ is the water depth $H$ and the steady flow follows depth contours on an $f$-plane. In contrast, our steady (mean) velocity observations are bottom-intensified below the main pycnocline (Fig. 6). Hence our flow is baroclinic and $D$ is the layer thickness below the pycnocline. The flow can move into shallower or deeper water, but in order to conserve PV, the pycnocline moves vertically in unison to keep $D$ constant on an $f$-plane. In steady state the sloping isopycnals will be geostrophically balanced, that is, a bottom-intensified flow moving over a sill into deeper water will be turned or steered to keep shallow water on the right rather than moving into deeper water.

We speculate that our bottom-intensified mean flow is topographically steered and enters Norske Trough near $76.5^\circ$ N from Fram Strait to the north of $77.5^\circ$ N latitude (see Fig. 1). Håvik et al. (2017) report similar waters, density structures, and velocities inshore of the shelfbreak East Greenland Current (EGC). Fig. 18 compares temperature and salinities along this proposed path from northern Fram Strait near $77.5^\circ$ N inshore of the 400-m isobath to our mooring section in the west via the entrance of Norske Trough in the east near $76.5^\circ$ N latitude. The core of Atlantic Waters are the temperature maxima near the $27.9 \text{ kg m}^{-3}$ potential density anomaly. Over the upstream continental slope this water has a temperature of $3 ^\circ$C while at the eastern entrance to Norske Trough in the south its temperature maximum has cooled to $2.5 ^\circ$C. At our mooring section in the west, we find this maximum near $1.9 ^\circ$C. The densities are almost identical which suggests mixing with cooler and fresher waters. Coachman and Barnes (1963) presents similar data to describe the circulation of Atlantic Waters in the Arctic Ocean by tracing the core of this water.

In summary, we posit that our observations describe a topographically steered mean flow near the 400-m isobath, however, the heat of waters that enter Norske Trough and move it towards coastal glaciers do not originate from the EGC. Instead, they originate from the upper continental slope of Fram Strait inshore from the EGC.
The small Rossby numbers motivated our search for remotely forced baroclinic Rossby waves to explain the temporal variations of the observed bottom intensification of currents over the slope at a set of discrete periodicities. We find statistically significant phase propagation of ocean current variability from shallow to deep water. These temporal oscillations modulate the mean flow at periodicities of 20, 11, and 6 days. Properties of these oscillations are dynamically consistent with the dispersion relation of topographic Rossby Waves where the bottom slope provides the main restoring mechanism in our rotating and stratified fluid. These propagating waves can be generated by any disturbance of the density field. A time-dependent meandering ambient flow such as the EGC can generate perturbations that propagate as topographic Rossby Waves (Maslowski 1996). The upwelling or downwelling caused by an eddy crashing into sloping bottom topography, too, can generate topographic waves (LaCasce 1998). While we cannot offer a specific wave generation process for the 11 day oscillation, we describe both a generation mechanism and location for our 20 and 6 day waves via Ekman pumping over the upper slope of Fram Strait.

The evidence presented supports the hypothesis that the wind-stress curl over the topographic slope of Fram Strait generates baroclinic perturbations that reach our study area where we detect them as bottom-intensified low-frequency current oscillations. Similar oscillations may exist farther into Norske Trough and perhaps even 79N Glacier. It is unclear, however, if the decreasing canyon width and/or its smooth change of orientation will affect the topographic Rossby waves.

The proposed Rossby wave generation differs from the upwelling and downwelling off East Greenland described by Håvik and Våge (2018). Using the same ERA-Interim wind data as we do here, Håvik and Våge (2018) find that coastal upwelling and downwelling favorable winds cause across-shelf currents near the bottom that change the density below the East Greenland Current at 68 °N latitude. Analyzing wind climatology, Håvik et al. (2017) suggest that this process diminishes in importance north of 72 °N latitude.
Within 200 km of Denmark Strait Harden and Pickart (2018) interpret moored ocean current observations across sloping topography off northwest Iceland as topographic Rossby waves. These waves are represented by the same dispersion relation that we use to describe current oscillations within Norske Trough. Off Iceland the dominant Rossby wave has a period of \( \sim 4 \) days, wavelength of \( \sim 60 \) km, and an offshore phase speed of \( \sim 17 \) km/day. The corresponding onshore group velocity of \( \sim 36 \) km/day suggests an offshore generation region. Harden and Pickart (2018) hypothesize that a meandering East Greenland Current (Hâvik et al. 2017) or aspiration of dense waters moving towards the sill of Denmark Strait 200 km away (Harden et al. 2016) may also contribute to the Rossby wave generation. Hence while their observed wave is similar to ours, the generation mechanism differs.

Our analysis bolsters a hypothesis that a carefully designed experiment can test: Ekman pumping caused by the wind stress curl generates topographic Rossby waves over the strongly sloping topography of the Fram Strait shelf break that propagate into canyons that cut across wide shelves off north-east Greenland. The experimental design to test this hypothesis consists of an array of bottom-mounted current meters spaced along the slope of the Norske Trough system from Fram Strait to 79N Glacier. Similar physics may impact observable ocean current variability elsewhere near sloping topography.

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international scientists, engineers, and technicians that included Jonathan Poole and Mathias Monsees. We also thank Captains and crews of R/V Polarstern during PS85 and PS100 as well as Drs. Benjamin Rabe and Wilken Jon von Appen who supported mooring deployment and recoveries in a safe, efficient, and enjoyable fashion. Drs. Paul Dodd and Laura de Steur of the Norwegian Polar Institute helped with recovery and deployment of M71 and M72, respectively, in 2015. Two reviewers provided detailed and constructive feedback that improved both text and figures.

APPENDIX A

Compass Heading and Calibration

Ocean current sensors generally use a flux gate compass to reference direction. While the magnetic field strength vector is constant at a fixed time, it is its horizontal component that turns the “compass needle” as the sensor package changes orientation. This horizontal component decreases with decreasing distance from the magnetic North Pole. In 2015 this pole was about 1100 km from our study area where the horizontal component was only 7230 ± 130 nT for the 2014-16 period of our observations. Furthermore, the local declination $D_0$ is 15.9 ± 0.8 degrees W and its drift $D_1$ is about 0.46 degrees per year. We thus convert the magnetic compass heading $M$ to true North $T = M + D_0 + D_1 \times t$ where time $t$ is measured in years. We extracted numerical values from the World Magnetic Model as distributed by NOAA’s National Centers for Environmental Information (https://www.ngdc.noaa.gov/geomag).

Our moored ocean current sensing array contains two different mooring designs that are listed in Table 1. Design-A places the ADCP package on a mooring line that changes orientation as a result of ocean current drag on the entire >200 m mooring line that accommodates additional sensors and buoyancy elements. The ADCP is contained in a stainless steel frame and its batteries are
de-gaussed. This minimizes compass bias, but the sensor package changes orientation by ±180 degrees in response to variable drag on the mooring.

In contrast, Design-B places the ADCP transducers atop a steel float 3-m off the seabed on a semi-rigid frame that is allowed to pitch and roll, but cannot change orientation. Fig. A1 shows 25-month long time series of pitch, roll, heading and pressure at M20. It shows a stable mooring platform with small pitch and roll near 0.5 degrees each with deviations from these means always smaller than 0.4 degrees. The magnetic compass starts at about 86.5 degrees and declines at the predicted rate $D_1 = 0.46$ degrees per year caused by the steady movement of the earth’s magnetic pole. Furthermore, we identify five spikes caused by geomagnetic storms of charged solar particles that also cause intense Northern Lights. Observations of these solar flares are distributed by the National Geophysical Data Center (https://www.ngdc.noaa.gov/stp). The storms cause intermittent compass fluctuations of up to 4 degrees. Our records thus contain a random noise <1 degrees. The compass on Design-A moorings contain the drift, solar perturbations, and random fluctuations as well.

Münchow and Melling (2008) introduced this torsionally-rigid mooring Design-B in Nares Strait where magnetic compasses are unreliable. The problem then becomes to determine the (unknown) constant offset or bias of the instrument package that sits fixed on the ocean floor. In Nares Strait 0.5 m s$^{-1}$ strong and rectilinear tidal currents provided a natural co-ordinate system in the 30 km wide channel, however, no such current or channel exist in Norske Trough and a different method is developed here to determine the unknown constant offset.

More specifically, we employ complex correlation analysis of our observed vector time series with a reference time series. The reference vector time series originates from barotropic tidal predictions made for the entire Arctic Ocean by Padman and Erofeeva (2004) that use 8 discrete tidal constituents. Our observation-based vector time series results from a tidal harmonic analysis
of vertically averaged currents that use the same 8 discrete tidal constituents as the model predicts.

We thus quantify the constant offset angle from observed tidal current vectors \( w_1 = u_1 + j \cdot v_1 \) with predicted tidal current vectors \( w_2 = u_2 + j \cdot v_2 \), where "\( j \)" represents the complex unit vector. The complex correlation \( c_{12} = (u_1 + j \cdot v_1) \cdot (u_2 - j \cdot v_2) \) can be decomposed into a magnitude and an orientation in the complex plane. The orientation angle rotates \( w_2 \) into \( w_1 \) to result in maximal correlation. Table 1 includes this angle at which input and output time series of current vectors result in a maximal correlation.

We estimate directional uncertainties to be less than 0.5 degrees, because moorings M50 and M71 result in offset angles of 0.6 and 0.1 degrees, respectively, following the above complex correlation. These Design-A moorings do not require compass corrections as their absolute directions are known. Furthermore, the largest uncertainty probably results from uncertain tidal currents in the barotropic Padman and Erofeeva (2004) model that requires accurate bottom depths. Table 1 lists both actual and model depths: at M50 the model depth of 453 m agrees well with the actual depth of 456 m. In contrast, a large depth discrepancy exists at M71 where the model uses 347 m when the actual depth is only 250 m. Nevertheless, tidal predictions and observations agree within 0.1 degrees in direction.

APPENDIX B

Empirical Orthogonal Functions (EOF)

EOF analyses organize serial observations at multiple locations \( u(t_k, x_j) \) as a set of mutually uncorrelated serial variations \( A_n(t_k) \) where the spatial dependence is expressed via a structure function \( \theta_n(x_j) \). The transformation from the vector space of observations \( u(t_k, x_j) \) to the vector space of
uncorrelated EOF modes is linear, that is,

\[ u(x_j, t) = \sum_{n=1}^{M} A_n(t) \cdot \theta_n(x_j) \]  

(B1)

Both \( \phi_n(x) \) and \( A_n(t) \) are orthogonal in the sense that

\[ \sum_{i=1}^{M} \phi_n(x_i) \cdot \theta_m(x_i) = \delta_{n,m} \]  

(B2)

\[ 1/K \sum_{k=1}^{K} A_n(t_k) \cdot A_m(t_k) = \lambda_n \cdot \delta_{n,m} \]  

(B3)

Here \( \delta_{n,m} \) is the Kronecker Delta which takes on a value of 1 for \( n = m \) and 0 for \( n \neq m \).

This heuristic description of EOFs describes a formal eigenvector problem for the cross-covariance matrix \( R_{i,j} = \sum_{k=1}^{K} u_i(x_i, t_k) \cdot u_j(x_j, t_k) \), that is,

\[ \sum_{j=1}^{M} R_{i,j} \cdot \theta_m(x_i) = \lambda_m \cdot \theta_m(x_i) \]  

(B4)

where \( \theta_m \) is the eigenvector of \( R_{i,j} \) for the eigenvalue \( \lambda_m \). The sum of all eigenvalues \( \sum_{m=1}^{M} \lambda_m = \sum_{m=1}^{M} R_{m,m} \) represents the total variance which we use to normalize the eigenvalues. In Equation-B1 we thus scale \( A_n(t) \) and \( \theta_n(x) \) such that the eigenvector \( \theta_n(x) \) carries units of \( cm s^{-1} \) while the eigenfunction \( A_n(t) \) carries no units and has a variance of 1 for all modes \( n \).

APPENDIX C

Wind-stress in ice-covered seas

The curl of the surface wind stress generates vertical motion that deforms the ocean density field via Ekman pumping \( w_e \), i.e.,

\[ w_e = (\partial_x \tau^{(y)} - \partial_y \tau^{(x)}) / (\rho_0 \cdot f_0) \]  

(C1)

We estimate the wind stress \( \tau = (\tau^{(x)}, \tau^{(y)}) \) from the wind vector \( \vec{U} = (U, V) \) as

\[ \vec{\tau} = \rho_{air} \cdot C_D \cdot (U^2 + V^2)^{1/2} \cdot \vec{U} \]  

(C2)
with the density of air \( \rho_{\text{air}} = 1.32 \text{kg/m}^3 \) and a drag coefficient \( C_D \). We use daily surface wind data provided by the European Centre for Medium-Range Weather Forecast as the ERA-interim product (Dee et al. 2011) to estimate wind-stress, its curl, and related Ekman pumping to construct time series and maps of \( \omega_r(x,y,t) \). A spatially variable ice cover such as shown in Fig. 3 will result in spatially variable drag coefficients \( C_D \) that reflect different efficiencies of the momentum transfer from the atmosphere to the ocean (Martin et al. 2014). We here implement the parameterisation introduced by Lübkes and Birnbaum (2005) that includes a form drag \( C_f(A) \) as well as skin drag at ice-ocean \( (C_{\text{ice}} = 1.89 \times 10^{-3}) \) and air-ocean \( (C_{\text{ocean}} = 1.25 \times 10^{-3}) \) interfaces, e.g.,

\[
C_D = C_f(A) + A \cdot C_{\text{ice}} + (1 - A) \cdot C_{\text{ocean}},
\]

where \( A \) is the fractional sea ice cover that we take from daily SSM/I imagery at 25 km resolution (Steffen and Schweiger 1991).

The parameterisation assumes freely-moving sea ice that we do not always encounter on the continental shelf where sea ice becomes land-fast in coastal and shallow areas (Hughes et al. 2011). We thus arbitrarily set \( C_D = 0 \) when more than 98% of a pixel is covered by sea ice \( (A < 0.98) \).

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Table 1. Mooring locations and records. Design-B moorings average 120 pings per ensemble every 30 minutes while Design-A moorings average 40 pings every 2 hours. Both mooring designs have 8-m vertical bin sizes. Heading is the magnetic compass correction from a complex (vector) correlation of tidal currents against model predictions at depth $H_{model}$ with $r^2$ the fraction of the correlated variance. Mooring M70 was deployed 2014-15 as M71 to be redeployed 2015-16 as M72 with a 30 hour gap on 29/30 August 2015 between the two records.

Table 2. Basic statistics of depth-averaged currents. Mean directions and ellipse orientations are in degrees positive counter-clockwise from true east. The degrees of freedom (dof) are determined as the ratio of the record length $T=792$ days and the integral decorrelation time scale $T_d$. Mean flows are not significantly different from zero at 95% confidence at M10 and M20 for speed. Uncertainty in directions are 95% confidence limits from directional statistics that use dof and a mean resultant length $R$ of current vectors on a unit circle (Mardia 1972).

Table 3. Estimated Topographic Rossby Wave properties. The wave period $T$ is estimated from the frequency-domain location of kinetic energy maxima. Cross-spectra estimate phase lags at these periods that are used to estimate phase speed $C_p$, group velocity $C_g$, wave length $\lambda$, and, from the dispersion relation, the orientation $\phi_{ana}$ of the wave number vector relative to the upslope direction (38T). Positive angles are counter-clockwise. A second independent estimate of this angle is $\phi_{obs}$ which is derived from band-passed filtered, time-domain estimation of the ocean current ellipse orientations. Additional ellipse parameters are semi-major $R_{maj}$ and minor $R_{min}$ axes.
TABLE 1. Mooring locations and records. Design-B moorings average 120 pings per ensemble every 30 minutes while Design-A moorings average 40 pings every 2 hours. Both mooring designs have 8-m vertical bin sizes. Heading is the magnetic compass correction from a complex (vector) correlation of tidal currents against model predictions at depth $H_{\text{model}}$ with $r^2$ the fraction of the correlated variance. Mooring M70 was deployed 2014-15 as M71 to be redeployed 2015-16 as M72 with a 30 hour gap on 29/30 August 2015 between the two records.

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<td>77.4653</td>
<td>303</td>
<td>305</td>
<td>32</td>
<td>B</td>
<td>33.4</td>
<td>0.880</td>
</tr>
<tr>
<td>M40</td>
<td>15.4420</td>
<td>77.7092</td>
<td>369</td>
<td>374</td>
<td>41</td>
<td>B</td>
<td>-48.0</td>
<td>0.890</td>
</tr>
<tr>
<td>M50</td>
<td>14.6478</td>
<td>77.9252</td>
<td>456</td>
<td>453</td>
<td>49</td>
<td>A</td>
<td>-0.6</td>
<td>0.751</td>
</tr>
<tr>
<td>M60</td>
<td>14.5020</td>
<td>77.9632</td>
<td>402</td>
<td>409</td>
<td>45</td>
<td>B</td>
<td>45.6</td>
<td>0.668</td>
</tr>
<tr>
<td>M71</td>
<td>14.3102</td>
<td>77.9975</td>
<td>250</td>
<td>347</td>
<td>24</td>
<td>A</td>
<td>0.0</td>
<td>0.772</td>
</tr>
<tr>
<td>M72</td>
<td>14.3102</td>
<td>77.9975</td>
<td>250</td>
<td>347</td>
<td>24</td>
<td>A</td>
<td>-11.8</td>
<td>0.779</td>
</tr>
</tbody>
</table>
TABLE 2. Basic statistics of depth-averaged currents. Mean directions and ellipse orientations are in degrees positive counter-clockwise from true east. The degrees of freedom (dof) are determined as the ratio of the record length $T=792$ days and the integral decorrelation time scale $T_d$. Mean flows are not significantly different from zero at 95% confidence at M10 and M20 for speed. Uncertainty in directions are 95% confidence limits from directional statistics that use dof and a mean resultant length $R$ of current vectors on a unit circle (Mardia 1972).

<table>
<thead>
<tr>
<th>Name</th>
<th>km</th>
<th>Mean Speed</th>
<th>Direction</th>
<th>$R$</th>
<th>Major/Minor Axis</th>
<th>Orientation</th>
<th>dof</th>
</tr>
</thead>
<tbody>
<tr>
<td>M10</td>
<td>7</td>
<td>0.8±0.9</td>
<td>180±180</td>
<td>0.11</td>
<td>8.4/4.2</td>
<td>64</td>
<td>140</td>
</tr>
<tr>
<td>M20</td>
<td>14</td>
<td>1.4±1.5</td>
<td>-106±70</td>
<td>0.15</td>
<td>7.1/4.3</td>
<td>78</td>
<td>90</td>
</tr>
<tr>
<td>M30</td>
<td>22</td>
<td>1.8±1.5</td>
<td>-68±40</td>
<td>0.25</td>
<td>7.0/4.5</td>
<td>81</td>
<td>73</td>
</tr>
<tr>
<td>M40</td>
<td>46</td>
<td>0.7±0.7</td>
<td>-158±50</td>
<td>0.14</td>
<td>4.9/4.1</td>
<td>71</td>
<td>143</td>
</tr>
<tr>
<td>M50</td>
<td>76</td>
<td>2.4±1.1</td>
<td>137±20</td>
<td>0.35</td>
<td>6.9/4.1</td>
<td>137</td>
<td>157</td>
</tr>
<tr>
<td>M60</td>
<td>82</td>
<td>6.7±1.2</td>
<td>142±7</td>
<td>0.69</td>
<td>9.2/3.8</td>
<td>138</td>
<td>209</td>
</tr>
<tr>
<td>M70</td>
<td>87</td>
<td>4.1±1.3</td>
<td>156±17</td>
<td>0.41</td>
<td>9.2/4.9</td>
<td>131</td>
<td>167</td>
</tr>
</tbody>
</table>
TABLE 3. Estimated Topographic Rossby Wave properties. The wave period $T$ is estimated from the frequency-domain location of kinetic energy maxima. Cross-spectra estimate phase lags at these periods that are used to estimate phase speed $C_p$, group velocity $C_g$, wave length $\lambda$, and, from the dispersion relation, the orientation $\phi_{ana}$ of the wave number vector relative to the upslope direction (38T). Positive angles are counter-clockwise. A second independent estimate of this angle is $\phi_{obs}$ which is derived from band-passed filtered, time-domain estimation of the ocean current ellipse orientations. Additional ellipse parameters are semi-major $R_{maj}$ and minor $R_{min}$ axes.

<table>
<thead>
<tr>
<th>Period $T$</th>
<th>$T^{-2}$</th>
<th>$C_p$ (km/day)</th>
<th>$C_g$ (km/day)</th>
<th>$\lambda$ (km)</th>
<th>Decay $\mu^{-1}$</th>
<th>$\phi_{ana}$ (deg)</th>
<th>$\phi_{obs}$ (deg)</th>
<th>$R_{maj}/R_{min}$</th>
<th>Scale</th>
</tr>
</thead>
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<tr>
<td>days</td>
<td>-</td>
<td>km/day</td>
<td>km/day</td>
<td>m</td>
<td>deg.</td>
<td>deg.</td>
<td>-</td>
<td>days</td>
<td></td>
</tr>
<tr>
<td>6.5</td>
<td>0.73</td>
<td>6.8</td>
<td>44</td>
<td>93</td>
<td>197</td>
<td>175.9</td>
<td>175.9</td>
<td>6</td>
<td>3.5 to 7</td>
</tr>
<tr>
<td>11.2</td>
<td>0.76</td>
<td>6.0</td>
<td>67</td>
<td>140</td>
<td>298</td>
<td>177.8</td>
<td>177.0</td>
<td>8</td>
<td>7 to 14</td>
</tr>
<tr>
<td>19.3</td>
<td>0.69</td>
<td>3.2</td>
<td>63</td>
<td>131</td>
<td>280</td>
<td>178.7</td>
<td>177.5</td>
<td>8</td>
<td>14 to 28</td>
</tr>
</tbody>
</table>
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Fig. 1. Map of Northeast Greenland with 2014-16 mooring array near Isle de France across Norske Trough. Red triangles place weather data from Station Nord (81.2 N), Henrik Krøyer Holme (80.5° N), and Denmark Havn (76.9° N). Gray box indicates area of mooring locations (red symbols) with local along-slope (308°T) and across-slope (218°T) orientation indicated by black axes. Black contours indicate 250-m and 300-m depth on the continental shelf (light blue) and 1000, 1100, and 1200-m water depths at the shelf break of Fram Strait (dark blue). White contour indicates maximal offshore extent of landfast sea ice in 2014. Yellow symbols are CTD station locations. Insert at top right shows our study area off Greenland as a red dot while the insert at bottom right details locations of 7 moorings in red and 12 CTD cast from 2016 in yellow. 46

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