Response of an Antarctic Peninsula Fjord to Summer Katabatic Wind Events

ØYVIND LUNDESGAARD, BRIAN POWELL, MARK MERRIFIELD,^a AND LISA HAHN-WOERNLE

University of Hawai'i at Mānoa, Honolulu, Hawaii

PETER WINSOR

University of Alaska Fairbanks, Fairbanks, Alaska

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ABSTRACT

Fjords along the western Antarctic Peninsula are episodically exposed to strong winds flowing down marine-terminating glaciers and out over the ocean. These wind events could potentially be an important mechanism for the ventilation of fjord waters. A strong wind event was observed in Andvord Bay in December 2015, and was associated with significant increases in upper-ocean salinity. We examine the dynamical impacts of such wind events during the ice-free summer season using a numerical model. Passive tracers are used to identify water mass pathways and quantify exchange with the outer ocean. Upwelling and outflow in the model fjord generate an average salinity increase of 0.3 in the upper ocean during the event, similar to observations from Andvord Bay. Down-fjord wind events are a highly efficient mechanism for flushing out the upper fjord waters, but have little net impact on deep waters in the inner fjord. As such, summer episodic wind events likely have a large effect on fjord phytoplankton dynamics and export of glacially modified upper waters, but are an unlikely mechanism for the replenishment of deep basin waters and oceanic heat transport toward inner-fjord glaciers.

1. Introduction

Due to their origin as submerged glacial valleys, most fjords are enclosed by steep topography that acts as a barrier for cross-fjord winds. As a result, local surface momentum transfer occurs predominantly in the along-fjord direction. High-latitude fjords often experience particularly strong down-fjord wind forcing during buoyancy-driven air flows known as *katabatic winds* (Argentini et al. 2003; Nielsen et al. 2014). The energetic surface forcing during katabatic events can elicit a strong response in fjord waters, including the export of sea ice (Johnson et al. 2011; Oltmanns et al. 2014) and upper-ocean water masses (Svendsen et al. 2002), and

upwelling and inflow of deep waters (Cottier et al. 2010; Spall et al. 2017). Katabatic winds at high latitudes typically occur when cool and dense air masses from continental ice sheets descend along the topographic gradient and flow out over the ocean (Manins and Sawford 1979; Renfrew 2004). Katabatic winds have long been recognized as a characteristic feature of the surface wind field near Antarctica (Ball 1957; Parish 1988), and they play a particularly important role in the near-surface climate of the Antarctic Peninsula, where the complicated topography strongly influences the wind field (van Wessem et al. 2015, 2016). Nowacek et al. (2011) attributed the dominant, rotationally modified surface circulation observed in a large bay on the western Antarctic Peninsula (WAP) to a katabatic forcing episode, but otherwise little is known about how extreme down-fjord forcing events affect WAP fjords.

The ocean response to along-axis wind forcing in fjords and estuaries has been studied extensively (e.g., Hansen and Rattray 1966; Svendsen and Thompson 1978; Klinck et al. 1981). In narrow geometries, surface stress creates an along-wind transport in the top ocean layer, which is balanced by a deep inflow. Down-axis winds can act to enhance the mean buoyancy-driven

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^a Current affiliation: Scripps Institution of Oceanography, La Jolla, California.

Corresponding author: Øyvind Lundesgaard, oyvind.lundesgaard@gmail.com

circulation and modify the stratification in shallow estuaries (Geyer 1997; Chen and Sanford 2009).

The dynamical response of a high-latitude, two-layer fjord to down-fjord katabatic wind events was explored through numerical model experiments in Spall et al. (2017). For broader fjords, the authors found that the main dynamical response is a balance between surface forcing and the horizontal pressure gradient. In such cases, the overall fjord response was found to be well represented by a nonlinear, inviscid two-layer model (Farmer 1976).

Rotational effects may be expected to play a role in the baroclinic dynamics of “dynamically broad” fjords where the first internal deformation radius L_1 is equal to or smaller than the fjord width. Due to their geometry and relatively weak stratification, Arctic fjords typically fall within this category (Cottier et al. 2010), and it is likely also the case for many WAP fjords. Rotational deflection tends to concentrate surface outflow near the coast on one side of broad fjords (Svendsen 1995; Ingvaldsen et al. 2001; Svendsen et al. 2002), and baroclinic instability of the resulting lateral shear was shown by (Carroll et al. 2017) to complicate the structure of the outflow in a model fjord. In addition, diverging Ekman transport near the coast can generate a cross-fjord pattern of upwelling and downwelling that may result in an estuarine-like residual geostrophic circulation in the along-fjord direction (Cushman-Roisin et al. 1994).

Several studies have observed that down-axis wind pulses enhance the water exchange with the exterior ocean, reducing the upper-layer residence time (Geyer 1997) in shallow estuaries and increasing the inflow of oceanic deep waters (Moffat 2014) in silled fjord. Spall et al. (2017) estimated from numerical model results that katabatic wind events in the Sermilik Fjord, Greenland, could flush out 17%–35% of the upper layer and 7%–15% of the lower layer, respectively. Since relatively warm deep waters are often present outside glacio-marine fjords (Cottier et al. 2010; Straneo et al. 2012), deep inflow can drive heat transport toward temperature-sensitive tidewater glaciers (Rignot et al. 2010; Jenkins 2011; Straneo et al. 2011; Cook et al. 2016). This can in turn generate glacier melt and retreat (e.g., Sutherland and Straneo 2012; Inall et al. 2014). Several studies have found that local, along-fjord winds may drive import of warm deep waters in glacio-marine fjords (Nilsen et al. 2008; Moffat 2014; Spall et al. 2017; Sundfjord et al. 2017), but the efficiency of this process is not well constrained and has not been documented in WAP fjords.

Fjords and bays along the WAP host productive marine pelagic ecosystems (Garibotti et al. 2003; Ducklow et al. 2007). Phytoplankton are concentrated in the euphotic zone, which is most directly impacted by surface forcing. A strong upper-ocean response to wind forcing

could conceivably impact the ecosystem both by advecting phytoplankton out of the fjord and by replenishing euphotic zone water masses with more nutrient-rich water through local upwelling or lateral exchange with the outer ocean.

This study explores the response of a high-latitude fjord to strong, episodic down-fjord wind stress forcing through a series of idealized numerical model experiments based on a wind event observed in Andvord Bay in December 2015. Our approach does not include buoyancy effects, and focuses solely on the effects of wind stress forcing. Particular attention is given to changes in fjord water masses and exchange with the exterior ocean, and how these processes depend on wind strength and duration, stratification and fjord geometry. We discuss the role of such wind events in ventilation of fjord waters, and in the oceanography of WAP fjords in general.

2. Data and numerical experiments

a. Observations

As part of the NSF-funded FjordEco program, oceanographic observations were conducted in Andvord Bay, a glacial fjord on the northern WAP. Measurements were collected in Andvord Bay between 23 November and 22 December 2015, during the LMG15-10 cruise on the R/V *L. M. Gould*. Underway near-surface salinity was obtained from a Sea-Bird SBE45 microthermosalino-graph connected to a flow-through seawater system with intake at 5–7-m depth. Underway winds were measured by a Gill WindObserver II ultrasonic anemometer mounted on the main mast of the ship. Down-fjord wind speed was taken as the component of the wind vector directed toward the NW, which corresponds approximately to the orientation of the main axis of Andvord Bay. Underway measurements were processed into 1-min data (Smith 2017a) using precruise sensor calibrations. The data were further low-pass filtered using both hourly and 6-hourly running boxcar functions. Only measurements collected inside Andvord Bay proper (as determined by the cruise navigational record) were included in this study.

A total of 105 conductivity–temperature–depth (CTD) profiles were collected within Andvord Bay proper during the LMG15-10 cruise (Smith 2017b). CTD profiles were collected using a SeaBird SBE 911 + instrument package. CTD data were processed into 1 dB bins using SeaSoft processing software and precruise calibration coefficients obtained within 1 year of the cruise. Only downcasts were used in this study. Additional CTD profiles were collected in Andvord Bay in April 2016 during the NBP16-03 cruise

FIG. 1. Wind and near-surface salinity measured by the R/V *L. M. Gould* within Andvord Bay in 2015. Orange arrows indicate hourly averaged wind vectors (down-fjord direction is approximately toward the northwest). Blue lines indicate salinity measured by the underway flow-through system. Thick and thin lines show 6-hourly and hourly running averages, respectively.

(Smith 2018) and in March 2017 during the LMG17-02 cruise (Kohut 2019).

Starting near 10 December 2015, sustained strong winds were observed blowing out of the fjord over at least 3 consecutive days. Hourly averaged wind speeds as high as 25 m s^{-1} were observed by the shipboard wind sensors (Fig. 1). The wind event coincided with a significant increase in the surface salinity of the near-surface fjord waters, as observed by the shipboard thermosalinograph. The observations from this event were used to motivate and design the model experiments detailed in this study.

b. Oceanographic context

The most thorough reviews of the oceanography of WAP fjords to date are found in Domack and Williams (1990) and Domack and Ishman (1993). In these papers, Andvord Bay is described as lacking strong buoyancy forcing, including the submarine outflow of surface glacial meltwater, which is an important forcing mechanism in many other glacial fjords (e.g., Straneo et al. 2011; Motyka et al. 2013). Instead, Domack and Williams (1990) describe the circulation in Andvord Bay as sluggish and dominated by the slow outward drift of subsurface plumes generated by submarine melt, and possibly a weak, cyclonic geostrophic gyre circulation in the upper fjord waters. Furthermore, drifter studies from (Niiler and Hu 1990; Zhou et al. 2002) have shown that typical surface-layer residence times in fjords and bays the Gerlache Strait area (Fig. 2a) are on the order of two months, suggesting that exchange between WAP fjords and the outside ocean is generally very slow.

Average water temperature in Andvord Bay was near or less than 0°C below the surface-forced layer during all three FjordEco cruises, and always decreased with depth (Figs. 3b,c). Ocean-driven ice melt is a strong function of ambient water temperature above the freezing point

FIG. 2. (a) Overview map showing the location of Andvord Bay on the Antarctic Peninsula. Schematics of model grid used in (b) Main Scenario and (c) Andvord Bay runs; (b) has a uniform depth of 400 m, and (c) shows 100-m-depth contours. Red arrows show the spatial envelope used for the wind stress forcing. Edges with no coastline are open model boundaries.

(Holland et al. 2008), which is approximately -1.9°C for seawater near the surface. Therefore, submarine melt in Andvord Bay is likely small compared to Arctic glacial fjords, where interior fjord water masses can reach several degrees above 0°C (Saloranta and Svendsen 2001; Straneo et al. 2012). Moreover, the cooling with depth likely prevents a positive feedback where melting drives upwelling of relatively warm waters along the glacier face, and subsequent additional melt. Yet, the subsurface plumes described in Domack and Williams (1990) indicate that submarine release of buoyant water does occur in Andvord Bay, although this process does not appear to drive a strong circulation in the fjord.

FIG. 3. (a) Hourly averaged wind (positive down-fjord) from shipboard measurements within Andvord Bay in 2015 (orange), model down-fjord wind speed (blue), and model down-fjord wind stress (black). (b) Profiles of salinity (red) and temperature (blue) from CTDs in Andvord Bay before the wind event in December 2015. Thick lines show the average profiles prescribed as model initial and boundary conditions. (c) Profiles of average salinity (red) and temperature (blue) from CTDs in Andvord Bay during NBP16-03 (April 2016, solid lines) and LMG17-02 (March 2017, dashed lines).

The importance of wind forcing relative to buoyancy forcing is often estimated using the Wedderburn number (Geyer 1997; Sutherland et al. 2014):

$$W = \frac{\tau_x L}{\beta_s \Delta_x S g H_u}. \quad (1)$$

Here, τ_x is the along-fjord wind stress, L the fjord length, g the gravitational acceleration, β_s the haline contraction coefficient, and $\Delta_x S$ the average along-fjord salinity difference averaged over an upper-ocean layer of height H_u . Taking H_u as 80 m, and estimating $\Delta_x S$ from CTD profiles collected in the inner fjord and the outside ocean during the NBP16-03 cruise (when no major katabatic event occurred), we obtain $\Delta_x S \sim 5 \times 10^{-3}$, giving $W > 10$ for $\tau_x > 0.15 \text{ N m}^{-1}$. Although W is more appropriate for steady forcing than for wind episodes (Spall et al. 2017), this indicates that buoyancy-driven circulation is unlikely to be important during major wind events.

Tides along the WAP are weak compared to the eastern side of the Antarctic Peninsula (e.g., Padman et al. 2018). The CATS2008 tidal model [an update to the model described by Padman et al. (2002)], which has been shown to correspond well with observations from the WAP (e.g., Oreiro et al. 2014), yields an annual maximum tidal range of 1.87 m outside Andvord Bay, and tidal current amplitudes of $\sim 1 \text{ cm s}^{-1}$ in what corresponds to the outer fjord. Andvord Bay is a short ($\sim 20 \text{ km}$) fjord with a deep ($> 300 \text{ m}$) opening to the outside ocean, so tidal currents within the fjord are expected to be weak.

In this study, we focus on the summer season where phytoplankton blooms occur, and assume that the sea

ice fraction is negligible. Time lapse imagery from cameras mounted in Andvord Bay during 2016 (Truffer 2018) shows that the fjord was largely free of sea ice from January to April, and only covered by fast ice during 1–2 months starting in early August.

An automatic weather station mounted outside Andvord Bay over a period of 15 months during the FjordEco period (Lazzara et al. 2012; data available through the AMRC website) registered 12 instances where daily average winds out of the fjord exceeded 10 m s^{-1} (Fig. S1, supplemental material). Only on one occasion, in January 2017, were the measured wind speed and duration comparable to those observed in December 2015; most of the remaining events were weaker and did not last for much more than one day. Based on this record, we suggest that Andvord Bay is likely exposed to ~ 10 minor katabatic events annually, while major events like the one shown in Fig. 1 are relatively rare, occurring once a year or less.

It should be noted that overall, little is known about the physical oceanography of WAP fjords, Andvord Bay included. More comprehensive observational or modeling studies would be needed to determine the efficiency of tidal or mean flow in driving water exchange, or to study the effects of wind events in cases where the sea ice fraction is substantial. The aim of the numerical experiments in this study is not to provide a full dynamical representation of Andvord Bay, but rather to study the effects of strong summer wind events in a simplified, analogous model system.

c. Numerical model

We performed a series of idealized numerical experiments using the Regional Ocean Model System (ROMS),

a free-surface, terrain following model commonly used in coastal studies (Shchepetkin and McWilliams 2005). The model experiments were conducted on a $40\text{ km} \times 30\text{ km}$ f -plane model grid with 100 m uniform horizontal resolution and 30 s -levels in the vertical. Layer thickness varied from $<1.5\text{ m}$ near the surface and 11 m at the bottom to a middepth maximum of 29 m .

The model was initialized with horizontally homogeneous vertical temperature and salinity profiles. The same profiles were applied as boundary conditions at the open model boundaries. All experiments were initialized at rest, with no flow prescribed at the boundaries. A sponge layer increased horizontal diffusivity and viscosity linearly by a factor of 10 over the outer 5 km toward the open boundaries.

Vertical mixing was parameterized using the conventional k - ϵ turbulence model (Launder and Spalding 1983). Background vertical momentum viscosity and diffusivity were set to $10^{-6}\text{ m}^2\text{ s}^{-1}$. We used the ROMS fourth-order scheme for tracer advection, and ROMS's third-order upstream advection scheme for momentum. Viscosity was parameterized using a small Laplacian coefficient scaled by the local deformation rate (Smagorinsky 1963), using the conventional choice of 2.2 for Smagorinsky coefficient c_s (e.g., Griffies and Hallberg 2000). Explicit background horizontal diffusivity was set to $2\text{ m}^2\text{ s}^{-2}$ for all tracers.

d. Experiments

1) MAIN SCENARIO

The main experiment discussed in this study (hereafter “Main Scenario” experiment) was designed to simulate the wind event observed in Andvord Bay in December 2015. The bathymetry in this experiment was that of a highly idealized fjord inset in a coastline running along the north–south direction (Fig. 2a). The width and length of the fjord were 5 and 19 km respectively, and ocean depth was set to 400 m throughout the entire domain including the ocean outside the fjord. The Coriolis parameter f was set to -1.81 cpd , corresponding to 64.75°S latitude.

Initial and boundary temperature and salinity conditions were prescribed based on smoothed average profiles from 17 CTD casts taken in Andvord Bay between 5 and 9 December 2015, as part of the FjordEco program (Fig. 3b). To avoid spurious values in the deep ocean due to varying CTD depth ranges, temperature was set to a fixed value (-0.915°C) everywhere below 300 m . Salinity was set to increase at a fixed rate of $2.5 \times 10^{-5}\text{ m}^{-1}$ below 300 m , giving a weak but stable stratification consistent with the observations from Andvord Bay. With this setup, the first baroclinic deformation radius was calculated to be 2.9 km .

Water masses in the model were tagged using passive numerical dyes in order to trace the pathways of water

masses during the model run. All water outside the fjord ($x < 20\text{ km}$) was initialized with an ocean water (OW) tracer used to study the evolving distribution of ocean water in the fjord. To examine the exchange of the near-surface water masses where marine organisms are concentrated, we introduced a second Upper Fjord Water (UFW) tracer within the fjord ($x > 20\text{ km}$). The UFW was initialized in the upper 35 m , a typical extent of the euphotic zone on the coastal WAP during moderate bloom conditions (Vernet et al. 2008).

The idealized model wind time series was prescribed as a sinusoidal bell function between the model start time t_0 and a later time t_1 , and set to zero beyond. Surface wind stress τ_{wind} was computed from the model wind time series $\mathbf{U}$ using the conventional bulk formula (e.g., Gill 1982), where τ_{wind} is proportional to $|\mathbf{U}|^2$ below 10 m s^{-1} and to $|\mathbf{U}|^3$ above (Large and Pond 1981).

Model winds were scaled to approximately match the observed event, with maximum wind speed 18 m s^{-1} and corresponding maximum wind stress amplitude 0.645 Pa (shown in Fig. 3a). While the entire wind event was approximated with a 5-day sinusoidal envelope ($t_1 = 5\text{ days}$), 98% of the integrated wind stress was applied within a 3-day window, 56% within a single day.

The wind stress forcing was applied uniformly over the surface area of the fjord (Fig. 2a). Outside of the fjord, the forcing amplitude was set to decay linearly over 5 km westward, with a 1-km decay in the y direction on each side in order to avoid extreme surface stress gradients.

2) OTHER EXPERIMENTS

The ocean response to wind events more generally may be expected to vary as a function of forcing, stratification and fjord geometry. To more fully understand the effects of the winds, we explored the changes in ocean response and water mass exchange under variations of key parameters of the idealized model (summarized in Table 1). Unless explicitly stated, the configuration of these additional experiments was otherwise identical to that of the Main Scenario.

The strength and duration of the surface forcing were varied by scaling and stretching the bell-shaped wind function (“Forcing Amplitude” and “Forcing Duration” experiments). Changes in wind speed are further exaggerated in wind stress due to the nonlinearity of the bulk formula; the time period containing 95% of the integrated wind stress is listed alongside the forcing duration ($t_1 - t_0$) in Table 1. Sensitivity to the choice of horizontal viscosity and vertical mixing parameters was tested (“Mixing and Viscosity” experiments), and we conducted experiments without forcing, with f set to zero, and with wind stress applied across the entire spatial domain (“Limit Case” experiments).

An experiment was also performed using the idealized two-layer profile described in Spall et al. (2017), with an

TABLE 1. Overview of idealized experiments and relevant diagnostic parameters.

Experiment	Parameters	$\mathbf{U}_{\max}^a$ (cm s ⁻¹)		$\Delta\eta_{\max}^b$ (cm)		UFW ^c (%)		OW: Fjord ^d (%)		OW: 0–50 m (%)		OW: 50–150 m (%)		OW: >150 m (%)	
		Out	In	Along	Cross	Fraction		Mouth	>5 km in	Mouth	>5 km in	Mouth	>5 km in	Mouth	>5 km in
Main Scenario Width	$L_1 = 2.7$ km	-37.2	14.9	-2.0	-1.7	19.1		18.7	11.3	75.3	72.2	19.5	9.3	7.3	0.1
	1 km	-44.8	19.1	-2.0	-0.5	5.8		21.9	12.2	84.9	82.2	13.9	7.8	12.7	0.1
	3 km	-42.1	21.0	-2.0	-1.1	7.8		20.0	14.0	87.1	85.4	17.1	13.5	8.0	0.1
	7 km	-35.5	12.3	-2.0	-2.0	36.2		21.1	9.4	54.2	50.8	24.3	11.3	13.5	0.5
Two-layer Forcing Amplitude (maximum)	9 km	-32.6	11.3	-2.1	-2.2	48.4		21.3	12.1	43.0	42.2	27.3	16.6	14.8	4.5
	$L_1 = 8.4$ km	-23.6	3.8	-1.0	-0.2	77.3		10.3	2.0	20.7	6.6	18.4	4.8	5.1	0.1
	$10 \text{ m s}^{-1}/0.14 \text{ Pa}$	-15.8	8.3	-1.1	-0.8	63.5		6.9	2.5	30.3	19.5	9.4	0.5	1.3	0.0
	$14 \text{ m s}^{-1}/0.33 \text{ Pa}$	-27.7	9.5	-1.6	-1.3	32.8		13.3	6.5	50.9	43.8	18.3	4.2	3.9	0.0
Forcing Duration ($t_1 - t_0$ / period of 95% of integrated wind stress)	$22 \text{ m s}^{-1}/1.12 \text{ Pa}$	-48.3	24.2	-2.3	-1.9	6.6		26.7	18.7	92.0	91.1	36.0	28.1	10.2	0.9
	1 day / 0.5 days	-41.1	9.7	-1.5	-1.8	63.5		7.4	2.4	28.4	18.1	9.3	0.6	2.5	0.0
	3 days / 1.5 days	-40.6	16.4	-1.9	-1.8	30.4		13.1	7.5	57.0	51.5	13.2	4.6	4.5	0.0
	7 days / 3.6 days	-40.1	14.7	-2.1	-1.5	10.7		22.5	14.3	83.0	79.7	24.6	15.4	9.9	1.0
Mixing and Viscosity	9 days / 4.6 days	-38.4	14.8	-2.0	-1.4	12.8		24.3	16.9	87.8	85.1	32.1	21.9	8.8	1.5
	$\nu_H = 10 \text{ m}^2 \text{ s}^{-1}$	-38.7	14.3	-2.0	-1.7	29.5		13.0	9.8	62.0	59.4	12.0	9.5	3.7	0.1
	$c_S = 4$	-37.5	14.3	-2.0	-1.7	20.2		18.2	11.9	76.7	72.1	19.9	12.0	6.1	0.1
	$A_V = 10^{-5} \text{ m}^2 \text{ s}^{-1}$	-37.2	15.0	-2.0	-1.7	19.2		18.4	11.3	75.3	72.6	18.9	9.1	7.1	0.1
Limit Case	Zero forcing	0.0	0.0	0.0	0.0	100.0		0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
	$f = 0$	-47.1	27.7	-1.9	-0.0	3.1		26.3	16.7	77.2	72.7	20.7	12.1	18.6	7.1
	Full domain wind	-40.3	26.0	-1.6	-1.8	10.8		31.1	21.0	31.8	30.3	24.4	12.4	33.9	22.6
Andvord Bay	—	-37.2	12.4	-1.4	-1.3	32.4		32.3	17.3	48.5	47.0	33.8	17.0	25.8	6.9

^a Maximum/minimum cross-fjord-averaged along-fjord velocity at 5 km into the fjord.^b Maximum difference in sea surface anomaly η . “Along” represents the cross-fjord-averaged $\Delta\eta$ from entrance to head: from $x = 20$ to 39 km (from $x = 20.5$ to 34.5 km in the Andvord Bay experiment). “Cross” represents $\Delta\eta$ from the south to north coasts averaged from $x = 20$ to 35 km (from $x = 20.5$ to 34.5 km in the Andvord Bay experiment).^c Fraction of UFW retained in the fjord 5 days after the cessation of the event ($t = 10$ days in all except the Forcing Duration runs).^d Fraction of OW within a given depth range 5 days after the cessation of the event. Shows fraction within the fjord and inward of 5 km beyond the mouth.

upper layer with a salinity of 31 transitioning to a lower layer with a salinity of 32.5 around 150-m depth (“Two-layer” experiment). Additional experiments were also performed using the observed average stratification from the two latter FjordEco cruises (Fig. 3c).

The fjord width was varied between 1 and 9 km, thus spanning the range from “dynamically narrow” to “dynamically broad” fjords (“Width” experiments). Finally, we performed a simulation with a model of the Andvord Bay bathymetry and coastline as a comparison to the idealized cases (“Andvord Bay” experiment). The model grid was rotated to align the main axis of the fjord with the x direction (Fig. 2b), and the geographic labels used here (e.g., “northern coast”) correspond to the rotated grid. Horizontal resolution, number of vertical layers, wind forcing function, and initial and boundary conditions were identical to in the Main Scenario. The spatial envelope of the wind forcing was modified to resemble the wind pattern typically observed in Andvord Bay, with winds emanating from the southern embayment.

3. Results

a. Dynamical response to down-fjord wind events

1) NEAR-SURFACE CURRENTS

We first consider the response of the near-surface layer, taken as the upper 35 m of the water column. The surface forcing rapidly spins up an energetic response in this layer, shown in Fig. 4. During peak forcing, strong Ekman deflection of the wind-driven currents concentrates the outflow in a band along the southern edge of the fjord. At its strongest, westward surface velocity within this band at the fjord mouth exceeds 80 cm s^{-1} . Outside the fjord, the flow continues to be deflected toward the southern boundary of the model domain.

The increasing winds set up and maintain strong pressure gradients, with sea surface depressions to the east and north. These gradients persist as the wind forcing subsides, and drive a surface return flow into the fjord that appears around $t = 4$ days, concentrated along the northern coast (Fig. 4b). The spatial structure of the currents also becomes increasingly complex in this phase. In particular, three bands of flow toward the northeast appear in the middle of the fjord, separated by a zonal distance of 5–8 km. The pattern resembles that of three cyclonic vortices, each with a radius similar to the width of the fjord.

This complex flow pattern largely eliminates the cross-fjord pressure gradient. However, a sea surface depression

FIG. 4. Snapshots of currents averaged over the upper 35 m at successive times of the Main Scenario experiment. Blue markers show $x = 25$ km.

of $\sim 1.9 \text{ cm}$ still exists in the inner fjord at $t = 4$ days. The along-fjord pressure gradient is equalized as the near-surface inflow propagates into and clockwise around the fjord, consistent with the behavior of a baroclinic Kelvin wave (Fig. 4c). The propagation speed of the signal along the fjord walls is approximately 22 cm s^{-1} , similar to the gravity wave speed of the second baroclinic mode (19 cm s^{-1}). After the passage of the wave, the near-surface horizontal pressure gradient is greatly diminished, and the surface flow in the fjord subsides, with the notable exception of a geostrophically balanced cyclonic eddy present in the middle of the fjord (Fig. 4d).

To explore the dynamics in greater detail, we examine the time-varying terms of the zonal momentum equation, equivalent to

$$\underbrace{\frac{\partial u}{\partial t}}_{\text{Acceleration}} = \underbrace{-u \frac{\partial u}{\partial x} - v \frac{\partial u}{\partial y}}_{\text{Horizontal advection}} - \underbrace{w \frac{\partial u}{\partial z}}_{\text{Vertical advection}} + \underbrace{fv}_{\text{Coriolis}} - \underbrace{\rho^{-1} \frac{\partial p}{\partial x}}_{\text{Pressure gradient}} + \underbrace{F_D}_{\text{Vertical viscosity}} + \underbrace{F_H}_{\text{Horizontal viscosity}}, \quad (2)$$

where u, v , and w are the Cartesian velocity components, ρ is density, and p is pressure. The two right-most terms represent the vertical and horizontal eddy viscosity.

The first-order dynamic balance in the near-surface fjord is between surface forcing (transferred down into the water column through vertical eddy viscosity) and the zonal pressure gradient resulting from the sea surface depression inside the fjord (Fig. 5a). The sum of these two terms is directed out of the fjord during peak forcing, but changes sign around day 3, as the forcing subsides while the pressure gradient remains. However, the dynamics are more complex than this approximate balance, and both Coriolis and advective terms are of significant magnitude (Figs. 5d,g). These secondary terms counteract the dominant balance, thus acting to limit the acceleration of the water. Horizontal friction plays a negligible role in the laterally integrated response of the fjord.

The cross-fjord momentum balance is entirely dominated by the geostrophic terms, which track closely at all depths throughout the event (see Fig. S2 in the online supplemental material). The evolution of the cross-fjord pressure gradient mirrors that of the along-fjord currents with an initial two-layer vertical structure transitioning into a three-layer one.

2) VERTICAL STRUCTURE

The flow has a two-layer structure during the active forcing phase (Fig. 6a). An outflow concentrated above 70-m depth ($3.7 \times 10^4 \text{ m}^3 \text{ s}^{-1}$ at $x = 25 \text{ km}$, $t = 2.5 \text{ days}$) is nearly balanced by a deep inflow below, with a small ($8.0 \text{ m}^3 \text{ s}^{-1}$) net volume flux out of the fjord. The upper-layer outflow is strongly surface-intensified, while the deep inflow is distributed throughout the deep waters, with a cross-fjord average maximum at $\sim 95\text{-m}$ depth. As the wind forcing weakens, a distinctly different vertical pattern emerges. One day after peak forcing, flow below 200 m has reversed and is directed outward. In addition, a mean inflow of up to 7 cm s^{-1} develops between 50 and 100 m at this time.

As the forcing subsides entirely, the vertical structure changes once again. By $t = 5 \text{ days}$, the along-fjord currents are stacked in three layers, with outflow between 50 and 200 m and inflow above and below. After $t = 9 \text{ days}$, the layered structure has largely dissipated, and the currents are instead dominated by a weak ringing at a frequency of approximately 2 cpd, close to the local inertial frequency $|f| = 1.8 \text{ cpd}$ (see Fig. S3 in the online supplemental material).

Near 100-m depth, direct stress forcing has little impact (Fig. 5b). Instead, the pressure gradient resulting from the sea surface displacement is the dominant forcing term. Counteracted by a combination of the Coriolis and vertical advection terms, the pressure gradient drives the weak inflow during the active forcing (Figs. 5e,h). Near $t = 4 \text{ days}$ the pressure gradient changes sign, and a strong

FIG. 5. Terms of the along-fjord momentum equation [Eq. (2)] in the Main Scenario experiment as a function of time at select depths. All terms are defined as positive out of the fjord. Each term has been integrated horizontally across the fjord and between $x = 25$ and 35 km . (a)–(c) Vertical friction (orange), pressure gradient (blue), and the sum of the two (black), (d)–(f) the remaining terms, and (g)–(i) acceleration. The peak and the end of the surface forcing ($t = 2.5$ and 5 days , respectively) are indicated with dashed vertical lines.

outward acceleration occurs before the terms settle into geostrophic balance approximately 1 day later.

Below 300 m, the dominant momentum balance is between the pressure gradient and the zonal acceleration, with the Coriolis term playing a secondary role (Figs. 5c,f,i). As the overlying gradients in density and sea surface evolve, the pressure gradient changes sign several times, accounting for the flow reversal in the bottom layer and resulting three-layer structure.

3) CROSS-FJORD STRUCTURE AND UPWELLING

The velocity structure is not uniform across the fjord, as shown in cross-fjord transects of along-fjord velocity during various phases of the response (Figs. 6b–e). During peak forcing, the surface outflow is intensified in the southern half of the fjord, while the deep inflow is stronger toward the north, consistent with rotational deflection to the left of the flow direction. In addition, an inflowing subsurface jet (up to 10 cm s^{-1}) is present near the southern coast near 50-m depth.

As the forcing relaxes, the evolving deep outflow is concentrated along the southern edge, where it extends from 100-m depth to the bottom (Fig. 6c). By $t = 5 \text{ days}$, the surface inflow has become restricted to the upper 50 m, and the midwater outflow and bottom inflow are

FIG. 6. Along-fjord velocity in the Main Scenario experiment (negative is blue out of the fjord). (a) Along-fjord currents averaged across the fjord at $x = 25$ km as a function of depth and time. Wind stress shown on top. (b) Cross sections at $x = 25$ km at successive times of the model run. The view is out of the fjord, with x -axis distance increasing toward the north. The color scale is saturated at $\pm 10 \text{ cm s}^{-1}$; velocities beyond this range are indicated by white contour lines spaced by 10 cm s^{-1} .

both strongly intensified toward the southern coast. By $t = 10$ days, the response has died down with the exception of the residual geostrophic eddy, which extends down to ~ 100 m.

In the active forcing phase, the fjord displays a distinct pattern of downwelling along the southern coast, and upwelling to the north and in the inner fjord. Surface salinity increases by as much as 0.6 near the northern coast and the head of the fjord during peak forcing as isohalines originally located at 60–80-m depth outcrop at the surface (Fig. 7a). In the relaxation phase, the cross-fjord surface density gradients are quickly equalized (Fig. 7b), followed by downwelling and inflow throughout the fjord that returns the upper ocean to near its initial state.

b. Impact on fjord waters

The bulk of the water mass initially occupying the upper 35 m (UFW) is quickly exported out of the fjord along the southern coast and into the exterior ocean. By the time of maximum forcing ($t = 2.5$ days), 70% of the UFW has exited the fjord, and 5 days later this has further increased to 80%. Of the UFW that does remain in the fjord, only 54% is located above 35 m, suggesting that much of the near-surface water gets mixed downward in the course of the event. Most of the exported UFW becomes entrained in circulation outside the fjord and exits through the model boundaries in the southwestern quadrant of the model domain.

FIG. 7. Snapshots of surface-layer salinity at successive times of the Main Scenario experiment.

The exported UFW is almost entirely replaced by water masses from the outside ocean (OW, Fig. 8). OW begins to enter the upper waters of the fjord as the surface circulation reverses, and most of the import occurs rapidly around $t = 4$ days, when the inflow is at its strongest. By $t = 10$ days, 75% of the fjord waters above 50 m consist of OW.

Between 50- and 150-m depth, the OW fraction at $t = 10$ days is reduced to 20%. The majority of the exchange in this depth range is restricted to the area around the fjord entrance (dashed line in Fig. 8); in the inner 14 km of the fjord ($x > 25$ km), the water between 50 and 150 m consists of only 9% OW.

Almost no exchange with the external ocean occurs in the deep waters of the inner fjord. The weak, deep inflow during the active forcing stage brings external water into the outer reaches of the fjord, shown as an eastward displacement of the OW gradient from its initial location at the mouth (Fig. 8b). As the deep flow subsequently reverses, the gradient moves back toward the west (Fig. 8c). Only 7% of the fjord water below 150 m is replaced by OW by $t = 10$ days, and the import is restricted to the outer few kilometers.

An alternative method of quantifying the deep import of external water is to integrate the inward volume flux across a cross-sectional area at the mouth of the fjord. Figure 8e shows the deep volume exchange estimated, first, by integrating the OW concentration

FIG. 8. (a)–(d) Cross-fjord averaged OW fraction at successive times of the Main Scenario experiment. Location of the fjord entrance is indicated with a dashed line. (e) Fractional volume exchange below 150-m depth in the Main Scenario experiment computed as the integral of positive volume flux across the fjord mouth (blue) and the volume of water tagged as OW present inside the fjord (orange). Surface forcing is indicated with the thin black line.

below 150 m and, second, by integrating the volume flux through the cross-sectional area below 150 m. In the first phase of the wind response, the deep flow is directed into the fjord, and the two methods agree closely. However, the volume flux approach fails to capture the re-export of imported OW occurring as the forcing subsides. Additionally, a deep cyclonic circulation develops near the mouth of the fjord after $t = 5$ days. While this circulation cell contributes little to import of OW, it results in a second increase in inward volume flux, which gradually decays as the eddy dies out. As a result, the two methods therefore diverge after $t = 3$ days, and the total exchange below 150 m that has occurred at $t = 10$ days is 4.1 times higher when calculated from integrated positive volume flux (30.0%) than when integrating the amount of OW present below 150 m (7.3%).

In total, 7.1 km^3 , or 18% of the total fjord water volume, is composed of OW by $t = 10$ days (7.5 days after peak forcing). Of the imported OW, 3.2 km^3 makes it more than 5 km into the fjord, corresponding to 11% of the fjord volume in the inner 14 km. Roughly half (3.6 km^3) of the total exchange occurs in the upper 50 m, where OW is found throughout the length of the fjord by $t = 10$ days.

c. Dependence on forcing, geometry, and other parameters

The overall characteristics of the oceanic response are conserved throughout the range of variational experiments. Changes in the ocean response and water mass exchange in the various cases are summarized in Table 1, and briefly discussed in this section.

1) FORCING DURATION AND AMPLITUDE

Increasing the maximum wind speed to 22 m s^{-1} increases the amount of UFW exported from the fjord at $t = 10$ days to 93% of the initial volume, compared to 81% in the Main Scenario (maximum wind speed 18 m s^{-1}). The deep waters of the inner fjord remain relatively unaffected by OW. Export and exchange are similarly reduced when the wind stress amplitude is decreased. However, even at a maximum wind speed of 10 m s^{-1} (a reduction in maximum wind stress by nearly 80% from the Main Scenario), 35% of the UFW is exported from the fjord.

Lengthening the duration of the applied wind stress forcing increases the upper-layer flushing, although the efficiency saturates or even decreases slightly beyond a forcing envelope of 7 days (which corresponds to applying 95% of the wind stress within 3.6 days). A shorter forcing duration decreases the flushing efficiency, but even after a drastic reduction, the wind forcing has a significant effect on the surface layer. When the envelope window is shortened to one day (thus applying 95% of the wind stress within 12 h), the forcing still causes more than 35% of the UFW to be flushed out of the fjord.

Net exchange below 150-m depth remains similar when the duration of the forcing is extended. However, at the longest duration (9 days envelope, 95% wind stress within 4.6 days), there is a significant increase in the OW penetrating beyond 5 km into the fjord, to 1.5% from 0.1% in the Main Scenario. However, this inflow does not extend into the innermost 10 km, where the OW fraction below 150 m is negligible.

In the case where surface forcing is applied throughout the entire ocean domain, the exported UFW fraction

at $t = 10$ days increases to 89% from 81% in the Main Scenario. This is also the experiment with the greatest import of OW below 150 m (34%, versus 7% in the Main Scenario), and the only case in which OW penetrates into the deep waters near the head of the fjord. The deep inflow occurs along the northern coast and in the later stages of the experiment, after the main fjord response has died down. The inflow is associated with a southward deep current, which develops along the outside coast and is partially deflected into the fjord. This deep coastal current is absent in the other experiments, and is likely a result of adjustment to forcing applied in the outer ocean.

2) ROTATION AND GEOMETRY

In the case where f is set to zero, the maximum value of the cross-fjord averaged flow speed increases significantly (47 cm s^{-1} , relative to 37 cm s^{-1} in the Main Scenario). The exported UFW fraction after the event increases to 97% (from 81%). The overall import of external water also increases without rotation, although OW does not penetrate into the deep waters of the inner fjord. The effects of rotation are also evident in experiments varying the fjord width, and thereby the ratio between horizontal scale and deformation radius (Fig. 9). The flushing (fractionally) of UFW decreases as the fjord becomes wider, from 94% exported UFW in the 1-km-width case to 52% in the 9-km case. In the wider cases, more UFW remains along the southern edge as the outflow subsides. This water is advected back into the inner fjord and/or entrained in the rotational circulation cells that develop in the upper ocean after the wind event. Conversely, in the 1-km-wide fjord, the great majority of UFW is immediately exported during the active forcing phase.

Inflow of deep water extends further into the fjord when the width is increased, and 5% OW is found in the inner 14 km below 150-m depth at $t = 10$ days for the 9-km width case. However, the OW fraction is still near zero in the inner 10 km of the fjord.

3) STRATIFICATION

Replacing the Andvord Bay stratification with the two-layer profile used by Spall et al. (2017) significantly changes the exchange flow in the model. The import of OW is reduced by 45%, largely due to decreased exchange in the upper layer. The UFW flushing is also greatly diminished; the fraction exported from the fjord at $t = 10$ days decreases from 81% to 23%. While the bulk of UFW is advected out of the fjord during the active forcing in the Main Scenario, it is here rapidly mixed vertically throughout the unstratified upper layer, down to the layer interface near 150-m depth. Along-fjord flow speeds are also notably reduced in the two-layer case.

FIG. 9. Distribution of UFW above 35 m in the Width experiments as a function of time for fjord widths of (a)–(c) 1, (d)–(f) 5, and (i)–(l) 9 km. The black line on top of each figure shows the average in the y direction. This metric does not include UFW, which has been mixed or advected below 35 m but remains within the fjord.

Using the stratification from the two latter FjordEco cruises (not shown) only results in a slight decrease in UFW flushing efficiency relative to the Main Scenario (from 81% export to 75% and 77%, respectively).

4) VISCOSITY AND VERTICAL MIXING

Model results are practically unchanged when the background vertical mixing coefficient A_V is increased by an order of magnitude compared to the Main Scenario. The same is true for a moderate increase in horizontal viscosity (increasing the Smagorinsky parameter c_s from 2.2 to 4). In contrast, applying a high, fixed Laplacian viscosity ($\nu_H = 10 \text{ m}^2 \text{ s}^{-1}$) changes the character of the dynamical response significantly. In particular, the small-scale flow features around $t = 4$ days are largely suppressed, and tracer and velocity distributions become smoother. The net result is a reduction in UFW flushing efficiency (70% UFW exported) and the overall exchange with the exterior ocean (13% OW in the fjord at $t = 10$ days versus 19% in the Main Scenario).

d. Andvord Bay experiment and comparison with observations

1) ANDVORD BAY EXPERIMENT

The experiment using realistic bathymetry for Andvord Bay shares the main characteristics of the idealized experiments. An outflow develops during active forcing,

FIG. 10. (a)–(d) Horizontal currents and (e)–(h) percentage of original upper fjord water averaged over the upper 35 m at successive times of the Andvord Bay experiment.

with average currents of $30\text{--}40\text{ cm s}^{-1}$ in the upper 35 m in a band along the southern coast of the outer fjord (Fig. 10). The strong outflow flushes the bulk of UFW into the outside strait as well as into the small inlet south of the fjord. The outflow is followed by a reversal of the surface flow, concentrated along the northern side and extending into the inner fjord. Although strong cross-fjord gradients are present during the active forcing phase, there is no development of evenly spaced fjord-scale vortices as in the idealized Main Scenario, likely due to the effects of variable topography and bottom-slope drag.

After the main response, a coastal Kelvin wave propagates along the northern coast and into the northern embayment, where it largely dissipates. At $t = 6$ days, a clockwise circulation pattern dominates the surface currents in the main body of the fjord. Some residual flow remains during the later stages of the experiment. In particular, a cyclonic eddy is situated in the outer part of the fjord, and two smaller circulation cells are present farther inside, but these largely dissipate by $t = 10$ days.

Although the majority of the upper layer is flushed out, a smaller fraction of UFW (68%) is exported from the fjord compared to the idealized Main Scenario experiment (81%). This is likely a result of the geometry of the coastline outside the fjord. Whereas there is an open path to the southwest in the idealized case, the coastline partially blocks this outflow in the Andvord Bay experiment. As a result, much of the exported water remains directly outside the fjord, and is reimported along the northern edge as the surface flow reverses.

The overall amount of external water entering the fjord is larger than in the idealized case (32% OW at

$t = 10$ days). A significantly larger amount of the water exchange occurs below 150 m (26% OW below 150 m in the Andvord case, 7% in the Main Scenario). This inflow only occurs in the outer region of the fjord, while the inner reaches including the inner basins remain largely unaffected by water from the external ocean (0.5% OW below 150-m depth inward of $x = 30$ km).

2) OBSERVED WATER MASS CHANGES

Successive CTD profiles before and after the December 2015 event show a distinct increase in salinity above 100 m (Fig. 11). This increase is not balanced by a decrease at depth, indicating an exchange of water masses rather than a vertical redistribution due to wind-driven mixing. The salinity increase is greatest at the surface, but it is apparent down to 80-m depth in all profiles.

Surface salinity in CTD measurements after the wind event is 34.0 or greater, an increase of 0.3–0.4 compared to before the event. During the active forcing, near-surface salinity measured by the shipboard flow-through system (Fig. 1) reaches 34.3, before decreasing as the wind subsides. The pattern is consistent with upwelling in the fjord during the event, and subsequent downwelling as the pressure gradients cease to be balanced by surface wind stress. The process does not seem to be entirely reversible, and salinity remains elevated after the event.

An interesting feature observed in CTD profiles from the northern embayment is the increase in surface-layer temperature θ after the wind event (θ – S diagram in Fig. 11c), where near-surface waters warm up by approximately 0.5°C to $+0.25^\circ\text{C}$. In general, there is a weak upper-ocean cooling toward the head of Andvord

FIG. 11. (a)–(c) Main axes show salinity profiles from CTD measurements before (black) and after (orange) the wind event on 11–14 Dec 2015. (bottom right) Locations of the salinity profiles within Andvord Bay. Smaller inset panels show potential temperature θ from the same profiles as a function of salinity. Density contours are overlaid with increments of 0.2 kg m^{-3} .

Bay, and the warm anomaly may be an indication of net heat transport toward the inner fjord during the wind event. It could also be a reflection of spatial variability of temperature in the area.

The Andvord Bay model experiment reproduces the rapid salinity increase observed in the upper ocean during the active forcing (Fig. 12). However, model upper-layer salinity reverts back to significantly lower values after the event than is the case in the observations. While average salinity in the upper 35 m of the model fjord is below 34.1 after the event, corresponding values calculated from individual CTD profiles are all around 34.2. The model experiments suggest that the upper fjord waters are largely replaced by water masses from outside the fjord (Fig. 8d). While the model was initialized with horizontally uniform water properties, any lateral gradients between Andvord Bay and the Gerlache Strait at the time of the December 2015 wind event would result in the advection of water masses with different properties into the fjord. This suggests that the upper fjord waters did not revert to their original state due to the replacement of the upper fjord water mass with higher-salinity water originating outside the fjord.

FIG. 12. Average salinity in the upper 35 m of the water column: average salinity in the Andvord Bay model experiment (blue) and CTD casts within Andvord Bay in December 2015 (orange).

4. Discussion and conclusions

This study shows the response of a highly simplified, ice-free model fjord to an idealized wind forcing event. The results presented here should therefore not be interpreted as representative of the average circulation in WAP fjords, or of the response of fjords to katabatic forcing events during periods of substantial sea ice cover. However, given the high amplitude of the response and the lack of known strong background forcing mechanisms, it is likely that the dynamics of open fjords are dominated by the wind response during such events. This is supported by the results from Carroll et al. (2017), who found that strong down-fjord wind forcing dominates the exchange flow even in the presence of tidal and buoyancy forcing. As such, the behavior of the model fjord can still provide useful insight into how local down-wind forcing events may affect WAP fjords.

a. Dynamical response

In the model experiments, strong episodic down-fjord wind forcing initially generates a two-layer exchange flow, where a vigorous surface outflow in the directly wind-forced layer overlies a weak inflow below. In the relaxation phase, the flow reverses as a result of the strong pressure gradients present in the fjord as the forcing subsides. The flow also becomes more complex, with the overall velocity structure eventually transitioning from two to three opposing layers. A three-layer flow structure is not uncommon in fjords, in particular as a feature of the background, tidally driven flow (Valle-Levinson et al. 2014). The transition from two to three layers may be explained by a framework of vertical normal modes, as the wave speed associated with the first baroclinic mode is twice that of the second mode. A trapped, propagating two-layer signal should therefore be able to exit the fjord faster than a three-layer one.

Due to the weak stratification, the first baroclinic radius of deformation (3–4 km) is smaller than the width of typical WAP fjords. As a result, rotational effects are important in all phases of the fjord response of the numerical experiment. The surface flow is strongly deflected by rotation, and the along-fjord currents are highly asymmetric at all depths. Substantial cross-fjord flow also develops, as a result of Ekman deflection of the outflow. This generates strong cross-fjord property gradients in the surface waters during the active forcing, consistent with the response described by [Cushman-Roisin et al. \(1994\)](#) for wind driven upwelling in a broad, infinite channel. The cross-fjord density gradient is in phase with the along-fjord currents, and suggests a tight geostrophic balance in the cross-fjord direction.

A spatially periodic cross-fjord flow pattern develops within the upper model fjord in the relaxation phase of the wind event. A previous study of wind-forced fjord currents found a similar pattern associated with baroclinic instability of the flow ([Carroll et al. 2017](#)). The exact evolution of the instability is dependent on both viscosity and fjord geometry. A complex surface flow also evolves in the relaxation phase of the more realistic wind fjord experiment, but no spatially periodic flow structure appears in this case. Details of the flow are highly dependent on the choice of parameterization of horizontal viscosity, which has limited empirical basis in small-scale coastal environments.

b. Water exchange

A key hypothesis motivating this study was that down-fjord winds might drive deep heat flux into the inner fjord and thereby impact glacier melt rates. Since deep water temperature typically increases offshore along the Peninsula ([Costa et al. 2008](#); [Martinson et al. 2008](#)), wind-driven import could affect WAP glaciers, which are believed to be highly sensitive to deep-water temperatures ([Cook et al. 2016](#)).

The present study offers little support for this hypothesis as the inflow of external water masses into the inner reaches of the fjord is minimal throughout the experiments. The exchange that does occur is generally restricted to the outermost region of the fjord, and none of the experiments suggest that local wind forcing can generate import of external water toward the head of the fjord where tidewater glaciers are typically located. These results do not apply to fjords with high or full sea ice cover, where wind-driven export of sea ice could potentially impact water exchange through increased sea ice formation and subsequent convection.

While deep flushing is largely inefficient, down-axis wind events provide an effective mechanism for flushing

the upper waters of the fjord. Of the water initially located in the upper 35 m, 81% is exported from an idealized fjord when applying surface wind stress comparable to a wind event observed Andvord Bay. Even in experiments with short duration or weak amplitude, the upper layer is significantly modified.

The observed increase in upper-ocean salinity after the December 2015 is not balanced by a freshening of the deeper waters, suggesting that exchange, rather than vertical mixing, is the agent of the upper-ocean change. Upper-ocean salinity also increases in the model experiments; however, the observed upper-ocean change persists for weeks, while the model fjord nearly reverts to its initial state within a few days of the peak forcing. There are several conceivable reasons for this discrepancy. First, the model is initialized with horizontally uniform salinity, so that lateral exchange in the model fjord only replaces the fjord waters with external waters of identical salinity. In reality, lateral exchange in the upper ocean may bring in water masses with other properties, likely denser and saltier than the native upper waters of Andvord Bay. Second, the model wind forcing is only applied in and directly outside the fjord. However, observations and continuity arguments suggest that wind forcing during down-fjord events in Andvord Bay also extends to the Gerlache Strait. Larger scale wind forcing could lead to coastal upwelling in the region outside the fjord, reducing the horizontal pressure gradients driving the return flow while also making the returned waters more saline.

In an experiment conducted with an unstratified top layer extending down to 150-m depth, the fraction of upper water exported from the fjord is reduced to less than 25%. In this case, surface water masses are quickly mixed down to the bottom of the unstratified layer, and at the same time the maximum outflow speeds are greatly reduced compared to the case with a slight freshening near the surface. This suggests that even a weak near-surface stratification plays a key role in concentrating the forcing into the rapid acceleration of the top layer, while in the unstratified case the energy is immediately dispersed over a much larger volume.

Katabatic wind events have the potential to impact bloom dynamics in the fjord, as phytoplankton biomass inside the fjord would likely be significantly decreased by the effective export of the bulk of the euphotic zone water volume. On the other hand, exchange with the external ocean could also be a mechanism for replenishing the nutrients of depleted upper fjord waters. Future studies of phytoplankton and nutrient measurements performed from this specific event will provide insight into the details of the biogeochemical significance of such events, as well as useful validation of the results of this study.

It has previously been shown that along-fjord winds can enhance the outflow of subglacial discharge plumes in Arctic fjords (Carroll et al. 2017). Although the water exchange in the present study is surface-intensified, wind events likely act to increase the dispersal of sediment-laden, neutrally buoyant “cold plumes” that are often found around 50–150-m depth near the inner glaciers of Andvord Bay and the surrounding area (Domack and Ishman 1993; Rodrigo et al. 2016). These plumes are a potential source of dissolved iron (Bown et al. 2017; Annett et al. 2017), and wind events may therefore drive export of trace metal export from the inner fjord to the outer ocean.

Wind events like the one explored in this study are typically associated with larger-scale atmospheric systems, and along-axis winds likely occur in many bays and fjords along the WAP simultaneously. As a result, the coastal ocean of the WAP could be exposed to episodic wind-driven pulses of glacially modified water from the coast.

c. Modeling wind events in fjords

The two-layer, estuarine-like flow structure generated in the model fjord during the active wind forcing is qualitatively consistent with classical models for surface-forced, nonrotating fjords and estuaries (Hansen and Rattray 1966; Farmer and Osborn 1976; Geyer 1997). However, in the relaxation phase following the active forcing, the flow structure grows increasingly complex, with large cross-fjord gradients and a transition to a three-layer vertical structure. Furthermore, the complex zonal momentum balance in the fjord suggests that two-layer, linear, nonrotating models like those presented by Farmer (1976) or Klinck et al. (1981) cannot be expected to adequately capture the physics of the fjord response. In particular, both the nonlinear and Coriolis terms act to oppose the dominant balance between wind stress and pressure gradient, thus slowing down the overall exchange flow.

While previous studies have suggested that episodic wind events may drive significant exchange of deep water in fjords (Moffat 2014; Spall et al. 2017), the present study suggests that the import of water into the deep fjord is minimal. We attribute this difference to a difference in methodology: this study adopts a tracer-based approach to estimate the exchange with the external ocean, as opposed to the more conventional method of integrating the volume flux across a cross-section of the fjord (e.g., Jackson et al. 2014; Spall et al. 2017). Direct comparison of the two methods suggests that the latter method overestimates the amount of exchange that occurs in the deep layer of the fjord by a factor of 4 as it fails to account for the re-export of

imported water (see Fig. 8e). It should be noted that our study does not extend to fjords with shallow sills, and the effects on deep water flushing could be different in such cases.

The difference between the two methods is not as great in the case of the upper layer or other situations where outflow precedes inflow, since the geometry in this case is much less favorable to the reimport of exported water. However, significant upper-ocean reimport of water can occur in cases with complicated topography outside the fjord, such as in the semirealistic Andvord Bay experiment in this study. While the integrated volume flux approach is a reasonable approach for low-frequency flows like steady wind-driven circulation or true estuarine-like circulation, tracer-based methods are better suited to quantify exchange for frequently reversing flows deep-silled fjords. Furthermore, a tracer approach allows the tracking of the distribution of imported water masses, in this case revealing that deep ocean water does not enter the inner fjord.

While fjords and estuaries are often modeled as two-layer systems (e.g., Svendsen and Thompson 1978; Klinck et al. 1981; Spall et al. 2017), the observed stratification in Andvord Bay is continuous and extends to the surface. The dynamical importance of near-surface stratification to the response of fjords to wind forcing has previously been pointed out in both Klinck et al. (1981) and Carroll et al. (2017), and this study shows that the quantitative effect on upper-ocean water exchange is also significant. Using a continuous stratification based on hydrographic observations changes the character of the ocean response, as upper-ocean stratification inhibits the downward transfer of momentum, and instead concentrates the energy into a shallow upper layer. When the prescribed model stratification is replaced with a two-layer one, surface currents are significantly dampened and the net export of water masses located within 35 m of the surface waters is reduced by ~70% compared to the continuously stratified case. The coastal WAP is characterized by a shallow summer mixed layer (Vernet et al. 2008), and it is likely that the presence of upper-ocean stratification greatly increases the flushing effects of wind events on phytoplankton blooms in WAP fjords.

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