Transverse variability of flow and density in a Chilean fjord

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Abstract

Measurements of velocity and density profiles were made to describe the transverse structure of the flow in Aysen Fjord, Southern Chile (45.2°S and 73.3°W). Current profiles were made with a 307.2 kHz acoustic doppler current profiler (ADCP) during 20 repetitions of a cross-fjord transect during one semi-diurnal tidal cycle. The transect had a ~ 320° orientation, 3 km length, and its bathymetry consisted of two channels, one on the southern side (230 m depth) and the other on the northern side (180 m depth), separated by a bank ca. 65 m depth, which was located ~ 1 km from the northern coast. Density measurements to a maximum depth of 50 m were made at the extremes of each transect repetition and over the bank. Also, a total of nine CTD stations that covered the surroundings of the bank was sampled 2 days prior to the ADCP sampling. During the sampling period the mean flow showed a three layer structure that was consistent with up-fjord wind-induced exchange: a thin (~8 m) weak outflow close to the surface due to river discharge; a layer of inflow (down-wind) underneath attributed to the effect of wind-stress; and a deep compensatory outflow due to the barotropic pressure gradient set up by the wind. The bank caused the strongest wind-induced net inflows and outflows to be shifted to the channels and also disrupted the three-layer structure. Also, the strongest tidal current amplitudes were located over the bank. The near-surface flow and density distributions suggested that the transverse dynamics were ageostrophic in a layer above ~ 50 m. This was indicated by the wind-induced shifting of the location of salt water intrusion from the northern side to the southern side of the fjord. Density measurements also suggested an alternation to quasigeostrophic conditions in this upper layer during calm winds. Below this layer the dynamics remained quasigeostrophic.

Keywords: Acoustic current meters; Ageostrophic flow; South America; Chile; Aysen; Fjords; Aysen Fjord; Physical oceanography; ADCP; Momentum balance; Physics of estuaries; Fjords oceanography

1. Introduction

Fjords are high latitude estuaries that are usually long relative to their width, relatively steep sided and deep, normally possess one or more sills, and usually feature a river discharging at their
most upstream reaches (their head). As in other types of estuaries, studies in fjords have focused particularly on the understanding of the longitudinal and vertical variability of the flow, while the transverse distributions have received relatively little attention. The works of Freeland et al. (1980) and Farmer and Freeland (1983) provide extensive reviews and abundant references about the circulation in fjords, confirming that the preferential interest has been given to the longitudinal and vertical dimensions.

The hydrodynamics of fjords are usually assumed to have an along-fjord momentum balance in which the pressure gradient is balanced by friction or advective accelerations (in constrictions), and a lateral geostrophic balance, i.e., between Coriolis accelerations and pressure gradient accelerations (Dyer, 1997). The geostrophic approximation has been demonstrated to be valid in fjords of British Columbia (Cameron, 1951) and in Juan de Fuca Strait (Tully, 1958). However, wind stress may frequently become relevant to the lateral dynamics. Wind stress may also reverse the classical mean velocity profile in estuarine systems. For instance, year-long series of current measurements in a coastal-plain estuary have shown that, at least during 21% of the time, inflow was observed on the surface and outflow near the bed (Elliot, 1974). This effect may not be very different in fjords. Their steep coastal walls may confine the wind flow, thus enhancing the along-fjord intensity of wind and hence the wind-induced flow. The reversal of the surface layer flow by up-fjord winds has been well documented in British Columbia (Pickard and Rodgers, 1959; Farmer, 1976; Farmer and Osborn, 1976) and Norway (Svendsen and Thompson, 1978; Svendsen, 1980). In some cases wind has been postulated to be one of the main forcing agents for deep water renewal (Gade, 1973; Gade and Edwards, 1980). Some efforts have been made to model the wind-driven circulation in fjords (Klinck et al., 1981; Krauss and Brugge, 1991). The across-fjord variations of the flow, nevertheless, are still a matter that requires more scrutiny. The objective of this study is to document the lateral structure of the flow and density fields near the mouth of a Chilean fjord. The objective is pursued with measurements of velocity and density profiles obtained at a cross-fjord transect over a bank located near the mouth of Aysen Sound in southern Chile, South America.

1.1. Study area

Aysen Sound (Fig. 1b) is a fjord-like inlet with a length of about 65 km measured between its mouth (station C in Fig. 1) and its head (station J in Fig. 1) and an average width of about 6.5 km. The average depth in the longitudinal section (stations A–J in Fig. 1) is 217 m, with maximum values of 360 m in the outer basin at the mouth (C) and 350 m in the inner basin (H–I). It has a sill of about 125 m depth located to the west and east-southeast of Colorada Island (station E in Figs. 1 and 3a), and banks of 60, 90 and 90 m at stations F, G and H, respectively (Fig. 1). The orientation of the fjord in its first 30 km from the mouth is in the SW–NE direction. The orientation changes in the following 25 km to the NW–SE direction, and to the SW–NE direction in the last 10 km (toward the head). As Aysen Sound is glacially carved and has a fjord-like morphology, henceforth it will be referred as Aysen Fjord.

Waters of oceanic origin reach the fjord chiefly from the north through the Meninea constriction (Fig. 1b). Modified Subantarctic waters of salinity 32.0 may fill the deep basin (210 m deep, letter A in Figs. 1b and c) to the south of the Meninea constriction according to Silva et al. (1995). Because of this constriction, along with the shoalings (~50 m) at Middle Passage (B) and to the south of Elisa Peninsula in Fig. 1b, and also because of the narrow channels that connect the study area with the ocean 50 km to the west, the circulation in Aysen Fjord is believed to be little affected by remote forcing generated on the continental shelf (outside the Inland Sea). To the south of Costa Channel the system receives fresh water inputs from small rivers and glaciers and its exchanges with the open ocean are strongly reduced by the presence of narrow (~1 km width) and shallow (~50 m) channels. Owing to this insulating morphology, the main forcing agents in the vicinity of the mouth of the Aysen Fjord are postulated to be tidal forcing from the coastal ocean, local wind stress, and river discharge. The
The main characteristics of these forces in the region are outlined below.

The tidal regime throughout the inland sea is mixed, predominantly semidiurnal. Outside of Aysen Fjord, four months of sea level measurements made during spring 1998 in the vicinity of Meninea constriction and at the entrance to Costa Channel have shown M₂ tidal amplitudes of 76.33 and 81.67 cm, respectively, for these locations (Fierro et al., 1999). They have also revealed mean tidal ranges during spring tides of 2.46 and 2.64 m at the same locations. Inside Aysen Fjord, similar ranges of 2.2 m in Puerto Pérez and 2.7 m at the head of the fjord (C, Fig. 1b) also have been reported for this region (SHOA, 1993). Following Stigebrandt (1977), the amplitude of the tidal current \( u \) in the bank region should be about 0.12 m/s, using \( u = a Y w / A \), where \( w \) is the frequency of the semidiurnal tide \( (2\pi/12.42 \text{ h}) \), \( a \) is the mean amplitude in water level at Puerto Pérez, \( Y \) is the horizontal area of the fjord \( (350 \text{ km}^2) \), and \( A \) is the cross sectional area at the bank \( (0.5 \text{ km}^2) \). As confirmed by this study, tidal currents were in the range 0.1–0.2 m/s. These are similar to the wind-induced currents in the fjord.

According to information provided by the Meteorological Service of the Chilean Navy (SMA), the wind regime in this region of the Inland Sea is dominated by southerly and southwesterly winds during spring and summer (October–March), and northerly and northwesterly winds during fall and winter (April–September). Westerly and southwesterly winds may favor wind-driven, up-fjord surface circulation at the mouth and in the outer half of the fjord, which would decelerate the surface layer mean outflow. North and northwesterly winds could be important for wind-driven circulation in the inner half of
the fjord, as the low land profile at the northern region of Puerto Perez may favor winds coming from that direction.

The main source of fresh water to the fjord is provided by the Aysen River (Fig. 1). Similar to many other Andean and trans-Andean rivers, the Aysen River exhibits two annual peaks of discharges at the head corresponding to the autumn increase in precipitation (April–June) and the summer melting of snow. Sampling for this study was carried out in October, close to the snow melting peak in November. The mean annual river discharge is estimated to be 620 m$^3$/s (Niemeyer and Cereceda, 1984) and the mean total annual precipitation (1968–1999) for stations around the study area fluctuates between 1100 and 2100 mm (SMA, personal communication).

In the transition zone at the mouth between C and E (Fig. 1), a strong horizontal salinity gradient at the surface ($\sim$ 12.5/10 km) has been observed in March (Silva et al., 1995), September (Muñoz et al., 1992) and November (Silva et al., 1997). This feature suggests the presence of a tidal intrusion front in the region of confluence between the brackish layer coming from the fjord and the intrusion of salty water from the west. As salinity is representative of the density field in this region, this will be used hereafter when describing density patterns. A satellite image obtained in February 1993 (Fig. 2) also suggests the shape of the front, which might represent the tidal intrusion going into the fjord on the northern side and the brackish water tending to leave the fjord on the south.

This study concentrated on the bank near the mouth located in the box marked F in Fig. 1b and had the objective of examining the transverse variability of the flow associated with this bathymetric feature. The working hypothesis was that the transverse dynamics would reflect geostrophic conditions: inflow of saltier water on the northern side and outflow of fresher water on the southern side, as suggested by Muñoz et al. (1992) and also by the satellite image in Fig. 2.

2. Data collection and processing

Current velocity and density measurements were obtained between October 14 and 17, 1998 in the vicinity of the mouth of Aysen Fjord, in order to describe the transverse structure of density and flows in that region. Current profiles were obtained along a $\sim$ 3-km-long cross-fjord transect (Fig. 3) using RD Instruments 307.2 kHz towed acoustic doppler current profiler (ADCP) during a complete semidiurnal tidal cycle on October 16. The across-fjord transect was traversed 20 times during the 12h of data collection. The sampling started at 13:00 h and it had a time interval between repetitions of about 33 min. The ADCP was mounted looking downward on a catamaran 1.2 m in length, which was towed at approximately 2.5 m/s on the starboard side of the local boat “Cordillera I” on October 16. The ADCP recorded ensembles of eight pings averaged over 30 s. Bin size was 4 m and the first velocity bin was centered at $\sim$ 8 m depth. The standard deviation of
velocity data were removed following the procedure explained by Valle-Levinson and Atkinson (1999). The semidiurnal tidal signal (represented by the $M_2$ constituent with a period of 12:42 h) was separated from the subtidal signal of the observed flow components using sinusoidal least-squares regression analysis (e.g. Lwiza et al., 1991). The subtidal signal represented the mean over the 12 h of observation.

The current measurements were complemented with density profiles obtained with a Sea-Bird SBE-19 conductivity–temperature–depth (CTD) recorder from the Chilean Navy patrol craft “Rano Kau” on October 14. Data processing was followed according to the manufacturer’s software. The pressure, temperature and conductivity measurements were aligned, data with pressure reversals were removed and the data were averaged to 1 m bins. Profiles were measured at the end of each transect repetition and over the bank in the center of the transect (Fig. 3b). Maximum depth of CTD profiles was 50 m owing to restrictions of the CTD’s pressure sensor. On October 14, 2 days prior to the ADCP sampling, 12 CTD stations (numbered 9–19, Fig. 3b) were sampled during $\sim$3 h of flood. The CTD sampling on October 14 provided a general idea of the horizontal distribution of the density field in the vicinity of the bank. This is explained in the context of the velocity profiles in the next section.

The closest meteorological stations to the study area were Guafo Island and Raper Cape, located in the coastal region at 225 km to the northwest and 250 km to the southwest, respectively. Owing to this long distance and to the orography in the fjord region, they were not representative of the wind conditions in the study area. Wind data were collected onboard during the period of sampling. Winds were calm during the first 3 h of sampling, and then they started to increase reaching maximum values of about 7 m/s in the up-fjord direction (southwesterly winds) at the end of the sampling.

A right-handed coordinate system was adopted for which $y$ was positive to the north and $x$ was positive to the east. It follows that the along-fjord and across-fjord components of the velocity were given by $u$ and $v$, respectively. The data were
rotated counter clockwise 60° to an along- (u-flow) and an across- (v-flow) channel coordinate system. This angle was oriented to the direction of greatest variability of the tidal currents and of weakest across-channel tidal flows. It follows that the x- and y-axis were oriented to 030° and 300°, respectively.

3. Results

This section presents the across-fjord distribution of the subtidal and tidal flows for the observation period. It also describes the horizontal salinity distribution, time series of salinity profiles, time evolution of the near-surface salinity across the fjord, and the time evolution of vertical sections of salinity across the fjord. These salinity representations help explain the observed subtidal flow patterns discussed below.

3.1. Subtidal and tidal flows

The along-fjord mean u-flow over the period of observations (Fig. 4) suggests a three-layer structure that included a thin (<8 m) surface weak outflow close to the surface (not entirely resolved by the ADCP measurements), a region of net inflow immediately underneath, and a compensatory outflow farther below. Vertical profiles of
current velocities at selected sites of the transect showed that the average outflow within a layer extrapolated to the surface from \( \sim 8 \) m depth was 0.03 m/s. Below this unresolved outflow, a 30–40 m thick layer showed up-fjord velocities of 0.1 m/s, most likely due to the dominant southwesterly winds during the experiment. The interior outflow layer of 60–100 m thickness had typical velocities of 0.05 m/s and was most likely a compensatory flow from the pressure gradient set up by the wind in the along-estuary direction.

The along-fjord mean flow exhibited transverse variability related to the bathymetry. Upper layer current magnitudes were weaker over the channel north of the bank than to the south. This might be attributed to the fact that the northern channel is shallower and narrower than the southern channel. Over the bank, frictional effects were evident in the decreasing magnitudes of the mean current relative to that on either side and by the fact that most of the water column showed up-fjord net flow, i.e., the near-bottom compensatory outflow was not developed in this region of the cross-section. The effects of the steep bottom also showed that horizontal friction might be playing a role in the dynamics, as the greatest outflow magnitudes were observed in the center of both channels at mid-depth. Throughout the section, the zero isotach at about 50 m exhibited downward slope toward the north, which was contrary to that expected from a geostrophic balance in a two-layered density driven flow, where the zero isotach would slope upward toward the north. The sign of the slope of the zero isotach indicated that the lateral momentum balance in the mean flow could not be explained by geostrophy. Below 50 m, nevertheless, where the flow was westward, the current was forced to the left by the Coriolis effect, resulting in a thicker layer at the south, i.e. the zero isotach was higher in the water column at the south. The observed slope suggested then that the compensatory flow below 50 m was consistent with geostrophy. It appears then that the observed distribution consisted of a frictional surface layer and a quasigeostrophic interior.

The transverse flows (Fig. 5a) were directed northward above \( \sim 50 \) m and throughout the northern channel, and southward only within the southern channel. The interaction of the flow going toward the north and toward the south over the bank suggested recirculations over this feature. On both sidewalls of the bank, flow tended to travel away from the bank. Thus, the transverse circulation indicated that effects of advection could be important around the bank. The strongest northward flow was located at the surface over the southern channel, far from the effects of the frictional effects of the sidewalls. Even though the velocities were in the range of the standard deviation (\( \pm 0.03 \) m/s), the distribution could not be dismissed as it showed a coherent pattern. The mean flow is depicted by flow vectors in Fig. 5b, which shows the tendency of the vectors to follow the bathymetry around the bank. This figure also shows the uppermost (8 m) layer going down-fjord on the northern side, and up-fjord on the southern side.

The amplitude of the semidiurnal along-fjord current (Fig. 6) also showed transverse variability that featured maximum tidal currents of \( \sim 0.2 \) m/s over the bank. This distribution contrasted those in coastal plain estuaries where bottom friction causes the greatest magnitudes to appear in the deepest part of the cross-section (Li and Valle-Levinson, 1999). Nonetheless, the transverse distribution observed in the Aysen Fjord was consistent with observations of greatest flows over sills or constrictions in response to the Bernoulli effect (e.g. Farmer and Armi, 1999), where bottom friction plays a secondary role in shaping the flows. Therefore, the tidal flows in this area should be mostly responding to the pressure gradient induced by tidal elevations and advective accelerations produced by the abrupt changes in morphology.

### 3.2. Tidal and subtidal salinity

In general, both density and salinity profiles showed a pycnocline in the first \( \sim 8 \) m depth with typical salinities between 15 and 25 in the surface layer and 30 in the deeper layer through 50 m depth, which was the maximum depth of CTD casts. Other observations in this area (Silva et al., 1995, 1997) showed consistent distributions with
those observed here and salinities that increased up to 31.2 in the deepest part of the fjord.

Fig. 7 shows the horizontal salinity distributions from stations 8 to 19 near the surface and 15 m depth on October 14, 2 days before the ADCP measurements. The time between the first and last station was about 2.7 h. Above the pycnocline (at 1 m), a fresh water tongue was evident over the northern side, while saline water intruded from the northern channel into the center of the fjord. Horizontal salinity gradients across the fjord seemed to be at least as large as along-fjord gradients in this particular instance. Below the pycnocline (at 15 m), where the horizontal and vertical salinity gradients were greatly reduced, the horizontal salinity distribution showed fresh water over the northern side and salty water on the southwest. This distribution also suggested the influence of wind-induced upwelling before the ADCP measurements, because the greatest salinities were found along the southern side. Also worth noting is the separation of the flow suggested by the shape of the 29.30 isohaline around the bank in station 15 (Fig. 7), where salty
water bifurcated around this feature (based on just three points across the fjord). This provided further evidence of the influence of the bank in modifying the flow. Evidence of fresher surface water on the northern side relative to the southern side was also suggested by time series of vertical salinity profiles at stations 4a, 6 and 5a (Fig. 3) taken during the ADCP measurements (Fig. 8). In this case, the surface salinity was lower over the northern side than on the southern side. This distribution was contrary to that expected from the earth’s rotation influence but could be related to local upwelling over the southern coastline of the fjord, as is also suggested by the across-fjord tilt of the zero isotach (Fig. 4) and the across-fjord flow (Fig. 5). Also worth noting is the change in the depth of the 28 isohaline during the tidal cycle, which is associated with the along-fjord velocity changes during this period. Below 8 m, the depth resolved by the measurements, velocities may be attributed to the combined effect of tide and wind. In the southern station, the down-fjord flow is less dominant than in stations in the center and north, probably attributed to the combined effect of up-fjord wind (Fig. 4) and low tidal amplitude shown in that side (Fig. 6). Some tidal straining effects are apparent in the three stations in the sense that the buoyant layer thickens during ebb and thins during flood.

The transverse structure of the near-surface salinity field throughout the tidal cycle (Fig. 9) showed that, as wind started to blow from the southwest at about 17:30 h and persisted during the rest of the experiment, high salinity waters began to appear over the southern portion of the section, another suggestion of upwelling. In accordance with the effect of the earth’s rotation, the salty water due to inflow at that depth was expected to be located over the northern side of the fjord as observed between 14 h and 17:30 h. As southwesterly winds persisted, high salinity waters covered the whole width of the fjord at 24 h. Low salinities observed over the bank at about 21 h were probably due to the effect of fresh water convergence at 2 m depth, by the combined effect of up-fjord wind and starting of flood. The effect of the wind was also appreciable in the vertical sections of the CTD stations 4a, 6 and 5a (Fig. 10) on the first three repetitions, where all isohalines tilted down toward the south side during calm wind, as expected from a typical geostrophic balance in the transverse direction for down-fjord flow. As winds picked up during the fourth repetition, isohalines of 22, 24, 26 and 28 rose progressively in repetitions 4, 5, 7 and 8, respectively. During repetition 8 every isohaline lower than 29 in the surface layer was higher on the southern side than on the northern side. Between

Fig. 6. Across-fjord distribution of the along-fjord M2 tidal current amplitude obtained from the ADCP measurements. Highest amplitudes were observed over the bank. Contour interval is 2 cm/s. Data in the shading region near bottom were not considered in the analysis due to side lobe effects. White region near bottom was out of the profiling range of the ADCP.
The salinity distributions discussed in this subsection showed a consistent message. Southwesterly winds in the lower Aysen Fjord seem to produce upwelling along the southern coast and downwelling along the northern coast above 50 m depth, which transiently invalidates the geostrophic approximation for the transverse dynamics. Also during these wind episodes, the saline intrusion shifts from the northern to the southern side.

4. Discussion

Wind forcing seemed to be the main driving mechanism of the mean flow during the sampling period as the mean flow was up-fjord near the surface, below 8 m depth. Similar profiles to those presented in Fig. 3, resulting from up-fjord wind driven circulation, have been observed in the fjords of Norway (Svendsen and Thompson, 1978; Svendsen, 1980) and British Columbia (Farmer, 1976). This wind episode also may have induced upwelling, as suggested by the zero isotach and the pycnocline both tilting downward toward the north, and by the presence of more saline water in the south than in the north. Evidence of fresher water on the northern side than on the southern side could be a transient wind-induced phenomenon, as salinity horizontal fields observed in the region of Colorado Island by Muñoz et al. (1992) showed more saline water over the northern side during calm winds. This was also observed during calm winds at the beginning of our experiment.

The bank affected the across-channel distribution of subtidal flow as it shifted the location of both the strongest near-surface inflow and the strongest mid-depth outflow to the channels. It also masked the three-layer mean flow that was observed in the channels. Evidences of flow bifurcation induced by the bank were observed in the horizontal salinity distribution below the pycnocline. The effect of the bank in shaping the flows was also observed in the secondary circulation patterns, as flow tended to travel away of the sidewalls and suggested recirculations over the summit. These effects induced by the bank suggest that advective accelerations and friction could be

Fig. 7. Horizontal salinity distribution from stations 8 to 19 at surface and 15 m depth on October 14. Isohalines are drawn in white and station positions are printed in black. Positions of CTD stations made on October 16 (4a, 6, 5a) are also drawn for reference. The area around Colorado Island has been masked because of lack of measurements there.
important for the across-fjord momentum balance during strong wind episodes.

In order to establish the relative contribution of advection and friction to the across momentum balance, the Coriolis term and the advective and friction (vertical and horizontal) terms were compared. In the across-channel direction, the subtidal momentum balance for transects 1–4 can then be given by

$$\frac{\partial}{\partial t} (q v) = q_y + \frac{\partial}{\partial z} (\frac{q}{z^2}) + \frac{\partial}{\partial y} (\frac{q}{y^2}) = \frac{1}{\rho} (\nabla p)$$

(1)

where $v(\partial v/\partial y)$ represents the advective terms (the other two could not be evaluated from the sampling strategy), $fu$ is the Coriolis term ($f = 1.03 \times 10^{-3}$), $1/\rho (\partial p/\partial y)$ is the pressure gradient term, where $\rho$ is the seawater density, $A_z (\partial^2 v/\partial z^2)$ is the vertical frictional term, $A_h (\partial^2 v/\partial y^2)$ is the horizontal frictional terms and brackets ($\langle \rangle$) denote tidal averages. The coefficients $A_z$ and $A_h$ denote the vertical and horizontal eddy viscosities. These were the terms of the complete momentum balance that could be reliably approximated, with exception of the pressure gradient term.

For the vertical eddy viscosity, $A_z$, we first consider wind stresses of the order of 0.03 Pa, for an average wind of 2.5 m/s during the tidal cycle. If $\tau = \rho A_z \frac{\partial u}{\partial z}$ then $A_z$ may be given by $A_z = \frac{\tau}{(\rho \partial u/\partial z)}$. Considering $\rho = 1024 \text{ kg/m}^3$ (the average of the full set of density data) and $\partial u/\partial z = 3.5 \times 10^{-3} \text{ s}^{-1}$ (averaged from distribution of Fig. 4), an estimate of $A_z$ from wind stresses will be $0.0027 \text{ m}^2/\text{s}$. The relationship proposed by Csanady (1975), $A_z = u_f^2/200f$ for estuaries of large depth, where $u_f$ is the frictional velocity ($\tau/\rho)^{1/2}$, provides an estimate of $A_z$ produced by bottom friction. This solution gives typical values in the order of $A_z = 0.0004 \text{ m}^2/\text{s}$.

Fig. 8. Temporal variability of the salinity (contours) and flow profiles (vectors) from stations 4a, 6 and 5a on October 16. Along-fjord ADCP measurements for those stations are overplotted. Broken line corresponds to the zero ADCP velocity contour.
Subtracting this from 0.0027 (as shown by Wong, 1994; Csanady, 1975) gives \( A_z = 0.0023 \text{ m}^2/\text{s} \). Therefore, we used a constant eddy viscosity coefficient of 0.002 m\(^2\)/s. As expected, bottom friction played a secondary role in the dynamics and most of the vertical friction was attributed to the wind stress.

Similar empirically derived forms of the magnitude of the horizontal eddy viscosity in tidal channels are less widely used. Tee (1976) used a value of \( A_h = 100 \text{ m}^2/\text{s} \) for an energetic tidal channel (currents of ~5 m/s), Ianniello (1979) used estimates ranging from 1 to 10 m\(^2\)/s for tidal channels, and Jones and Elliot (1996) used estimates of 0.5–10 m\(^2\)/s in the parametrization of friction in rivers. Following the method proposed by Ianniello (1979) to estimate the width of the channel that makes horizontal friction important for the momentum balance, we used an estimate of \( A_h = 1 \text{ m}^2/\text{s} \), according to the velocities and horizontal lengths observed in this system.

The absolute value of the ratios \( \langle v(\partial v/\partial y) \rangle / \langle fu \rangle \), \( \langle A_z(\partial^2 v/\partial z^2) \rangle / \langle fu \rangle \) and \( \langle A_h(\partial^2 v/\partial y^2) \rangle / \langle fu \rangle \) is shown in Fig. 11. These ratios showed that the non-linear accelerations, vertical friction term and horizontal friction term were comparable to Coriolis accelerations during the sampling period, as most of the magnitudes of the sectional means were about 1. The distributions showed that values >1 were located around the bank and close to the sidewalls and bottom, where advective and frictional terms were expected to be larger than the Coriolis term. There was also a thin layer of values >1 at about 50 m depth across the entire fjord, which may be explained by the position of the zero velocity in Fig. 4, where the Coriolis term tends to zero.

Analogously, the along-fjord momentum balance appears to undergo transition from a balance between pressure gradient and advection during calm conditions, reminiscent of a Bernoulli-type flow, to a balance that also includes friction during...
wind events. The terms in the along-fjord momentum balance may be given by

\[
\frac{\langle v (\partial u / \partial y) \rangle}{q} - \langle fu \rangle = - \frac{1}{\rho} \langle \partial p / \partial x \rangle \\
+ \langle A_z (\partial^2 u / \partial z^2) \rangle + \langle A_y (\partial^2 u / \partial y^2) \rangle. 
\]  

Again, with the exception of the pressure gradient term, these were the terms that could be reliably estimated from the measurements. The ratios of the advective term, vertical friction term and horizontal friction term to Coriolis term (not plotted here) showed absolute values of the

Fig. 10. Across-fjord vertical salinity sections at different times (looking into the fjord). Persistent southwesterly winds starting after repetition 3 induced upwelling, as reflected by the progressive rise of the isohalines on the southern side. Time corresponds to the beginning of each repetition. Maximum depth plotted is 20 m, which was where the most appreciable changes appeared.
sectional means of 1.4, 2.6, and 1.3, respectively. This revealed that, in the along-fjord momentum balance, vertical and horizontal friction were also as important as advection.

The contribution of advection to the momentum balance is supported by the distribution of the tidal current amplitudes. The greatest values were observed over the shallow areas as expected from a Bernoulli-type balance. In other systems, where bottom friction is influential, the transverse distribution of tidal current amplitudes follows the bathymetry with greatest values appearing in

Fig. 11. Ratios of absolute values of the tidally averaged advective term, vertical friction term and horizontal friction term to the tidally averaged Coriolis acceleration. Darker tones denote values below 1 and white denotes values over 1. The absolute value of the sectional mean is shown at the upper right corner of each panel. The advective, vertical friction and horizontal friction terms were comparable to Coriolis accelerations (contours shown: 0.5, 1, 5, 10, 15, etc.). Data in the shading region near bottom were not considered in the analysis due to side lobe effects. White region near bottom was out of the profiling range of the ADCP.
the region where the depths are greatest (Li and Valle-Levinson, 1999).

With respect to the frictional influences, further investigations are required to reveal the main frequencies of the wind stress in the region. If a strong diurnal (24 h) period is found, as observed by Pickard and Rodgers (1959), Farmer (1976) and Farmer and Osborn (1976) in fjords of British Columbia, the system could be affected by high-frequency (>1 cycle/day) fluctuations in the across-fjord momentum balance.

In summary, wind forcing seemed to be the main driving mechanism of the mean flow during the sampling period as the mean flow was up-fjord near the surface (below 8 m depth). The bank had salient effects on the across-channel distribution of subtidal flow by shifting the location of strongest near-surface inflow and strongest mid-depth outflow to the channels, by masking the three layer vertical structure, by shifting the flow direction away from the bank, by inducing recirculation at its top, and by causing a bifurcation of the flow below the pycnocline. Evidences of wind-induced upwelling observed during the sampling period indicated that the across-fjord momentum balance likely switches from quasigeostrophic during calm winds to a balance that in the surface is influenced by non-linear effects from wind stress and advection) during wind episodes, and remains quasigeostrophic in the interior.

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