Flow and Mixing Around a Glacier Tongue: A Pilot Study

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Abstract

A glacier tongue floating in the coastal ocean presents a significant obstacle to the local flow and influences oceanic mixing and transport processes. Here acoustic Doppler current profiler and shear microstructure observations very near to a glacier tongue side-wall capture flow accelerations and associated mixing. Flow speeds reached around twice that of the ambient tidal flow amplitude and generated vertical velocity shear as large as 3×10^{-3} s^{-1}. During the maximum flow period turbulent energy dissipation rates reached a maximum of 10^{-5} $m^2 s^{-3}$, around three decades greater than local background levels. This is in keeping with estimates of the gradient Richardson Number which dropped to ~1. Associated vertical diffusivities estimated from the shear microstructure results were higher than expected using existing parameterization, possibly reflecting the proximity of the glacier.

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1. **Introduction**

Quantifying ice-ocean interaction, especially at the small-scale, is a major challenge in high-latitude earth system sciences (e.g. Sirevaag et al. 2010; Rignot et al. 2010) where relatively long timescales and complex thermohaline and pressure effects interact with cryogenic topography that is continually changing. Oceanic mixing in polar waters includes different facets relative to that which occurs at lower latitudes. For example, rotational effects are large, there is little dynamic influence due to temperature, the effect of being close to the freezing temperature has a controlling influence on behaviour and there are the additional frictional boundary effects of the frozen upper surface (McPhee 2008).

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Glacier (or ice) tongues add additional complexity to coastal ice-ocean interaction. These features, formed by glacier outflows into the coastal ocean, can extend many tens of kilometers from shore (Frezzotti 1997) and be many hundreds of meters thick in places. Such cryogenic structures significantly influence local circulation and mixing (Jacobs et al. 1981; Legresy et al. 2004). In the case of flow around a glacier tongue, the mixing processes are then a result of the tidal and circulatory flows past a bluff body. The closest relevant work is that concerned with flow around headlands and islands (e.g. Edwards et al. 2004), except of course with a glacier tongue the flow can pass under the obstacle as well as around.

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Ice-ocean interaction processes are important in a situation like southern McMurdo Sound, Antarctica where Haskell Strait forms an oceanic connection between the western Ross Sea and the cavity beneath the combined Ross and McMurdo Ice Shelves (Figure 1). The Sound acts as a conduit for both ice shelf waters and warmer

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Ross Sea waters (Robinson et al. 2010). The fate of these waters is dependent on transport and mixing processes in the region. This exchange influences sea ice growth, which in turn affects climate processes over large space and time scales (Hellmer 2004; Dinniman et al. 2007; Robinson et al. 2010).

The Erebus Glacier Tongue (EGT, Figure 2) in southern McMurdo Sound has been observed to influence local vertical stratification through formation of diffusive-convective layering (Jacobs et al. 1981). It has also been suggested that such glacier/ice tongues generate local sources of supercooled water (e.g. Debenham 1965; Jeffries and Weeks 1992) – water cooled at depth to the in situ freezing temperature but then advected to shallower depths where it is colder than the new insitu freezing point. This supercooled water remains liquid until it encounters nucleation sites at which point it forms frazil/platelet ice (e.g. Dmitrenko et al. 2010). Ice crystals grow so large in McMurdo Sound (>250 mm diameter) that the largest of the frazil crystals are referred to as platelets (Smith et al. 2001). This phenomenon is readily observed in the region (Gough et al. 2011; Robinson, 2011) including during the very first of such measurements by Edward Nelson in 1911 during Scott's second expedition (Deacon, 1975).

The EGT provides an ideal site to examine the degree of variability of, and interaction between, small-scale mixing and advective processes. Much work has been developed relating to the cryomechanics and the effect of the ocean on the glacier tongue (see Squire et al. 1994 and papers therein). As well as providing evidence for the effect of the glacier tongue on local oceanography, these new data also have relevance for flow and melting processes at the face of ice shelves (Rignot et al. 2010). The objective here

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is to present new shear microstructure observations providing evidence of elevated ocean mixing rates near the tip of a glacier tongue. The Discussion then explores the (4.1) water column structure and kinematics, (4.2) mixing, (4.3) mechanisms for enhancement of local supercooling and (4.4) generalization of the results beyond the EGT and how an expanded study might extend the present measurements.

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2. Methods

In November 2009 we conducted exploratory oceanographic measurements within 30 m of the sidewall of the EGT at a station called Microstructure Camp (MSC). Whilst brief, the sampling was timed such that we captured a diurnal tidal cycle of turbulence data during the fastest tidal flows along with several tidal cycles of upper water column velocity data. Tidal elevation data were recorded at Scott Base near Cape Armitage, 17 km to the south (**Figure 2**) and provide a spring-neap context for the experiment.

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2.1 Location

The EGT divides the surface waters of the majority of Erebus Bay from the Dellbridge Islands and Cape Evans (Figure 2). At the time of sampling, the glacier tongue was 12 km long, 2 km wide and 300 m thick at the grounding line (Figure 3). Based on earlier work (Debenham 1965; Holdsworth 1982; DeLisle et al. 1989) the glacier tongue had a basal slope of ~0.02 along its main axis, and the tip was ~50 m thick (Figure 3b). Note there are few bathymetric data available in the area and none to the east of Big

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Razorback Island (Figure 2, Figure 3a). Water depths near its tip are around 400 m (Jacobs et al. 1981) but highly variable due to the complex geology of the region (Figure 3a). At the time of writing the most recent calving of the EGT was in March of 1990 when a 3.5 km section broke away. Similar events were known to have occurred in 1911 and at some point during the 1940s (Robinson and Haskell, 1990).

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2.2 Acoustic Doppler current profiler

A 300 kHz acoustic Doppler current profiler (ADCP - RDI Workhorse) was deployed through a 650 mm diameter hole in the 2.3 m thick first year ice at the MSC for a four day period (**Figure 4**) ~1 km shoreward from the tip on the north side of the EGT (**Figure 3**a). The ADCP was mounted just beneath the sea ice at a depth of 2.5 m with the first measurements starting 4 m below this. The ADCP used the same hole as the microstructure profiler, with the ADCP being held to one side. Because of judicious beam orientation there were no obvious effects in the ADCP data due to the profiler (whereas we have experienced substantial interference in similar applications in other experiments). The sampling recorded two-metre thick velocity bins every five minutes. Good quality data were typically resolved down to depths of 70 m and as much as 120 m. This is good penetration for this type of instrument in these waters where we have previously observed far shallower penetration due to lack of suitable scatterers (Leonard et al., 2006; Stevens et al. 2006).

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Velocity shear magnitude Sh_A was resolved from the ADCP vertical derivative of horizontal velocity components u and v (east-west and north-south respectively) so that

$$Sh_A = \left[\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right]^{1/2} .$$

A magnetic declination offset of 144 degrees was included. Compass testing did not indicate any inconsistencies due to the near-vertical magnetic field.

2.3 Microstructure profiler

Turbulence properties were quantified using a Rockland VMP500 (Victoria, Canada) microstructure loose-tethered free-fall profiler with dual shear sensors (Macoun and Lueck, 2004). This enabled estimation of the turbulent energy dissipation rate ε (units of m^2s^{-3} equivalent to W kg^{-1}). We have used this device in such conditions previously (Stevens et al., 2009). The main physical modification to the profiler involved the use of a circular drag brush rather than the standard pair of square drag elements as provided by the manufacturer. This allowed the instrument to pass through smaller holes (~ 500 mm) that possible with the square elements. Two tradeoffs were that (i) there was more vibration than with the square brushes and (ii) that the round brushes promoted rotation (as many as 10 full rotations in 250 m) during profiling that was not reversed on recovery. This latter effect began to seriously affect the cable integrity until we took action to counter the rotation by reversing the brushes as well as leaving the profiler to freely spin at the end of each profile.

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As described in Stevens et al. (2009), energy dissipation rates were resolved from the dual shear probe profiler using standard techniques (Prandke 2005). The profiles were segmented into five m-thick bins that were overlapped by 50%. Analysis first corrected for profiler vibration, identified the reliable section of the spectrum by comparing with the vibration spectrum derived from accelerometers, and then isolated a dissipation spectrum. The tail beyond this vibration limit was substituted with the tail of a Nasmyth model spectrum (Roget et al. 2006). The dissipation rate was then calculated with the integral $\varepsilon = 7.5\nu \int_k S_m dk$ (Prandke 2005) where *k* is the wave number and S_m is the shear spectrum. The noise floor in terms of ε was around 3x10^{-10} m^{2}s^{-3}. This was, however, not a fixed quantity as it depended on a number of variables like cable influence that were not exactly the same in every profile. Thirty eight profiles were recorded, with a total profiled distance of over 11 km.

While the profiles penetrated to ~ 300 m depth, only the upper 120 m are considered here in order to focus on the influence of the glacier. Due to operational constraints there were a number of gaps in the profiling, most notably a 2.5 hr gap. Fortunately, from the perspective of identifying dominant events, this gap fell during a period of slower flow.

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2.4 Conductivity, temperature depth measurements

Conductivity-temperature –depth (CTD) profiles were acquired using a SBE (Seabird Electronics, USA) temperature (SBE 3) and conductivity (SBE 4) sensor pair mounted on the VMP500. These sensors were un-pumped so as to not affect the shear data through the creation of vibration, so the spatial resolution was ~1 m. The profiler was kept in the water continuously so that the entire package remained at the ambient temperature removing thermal inertia start-up effects. The buoyancy frequency squared was calculated from the vertical derivative of density $\partial\rho/\partial z$ so that $N^2 = (g/\rho_0)(\partial\rho/\partial z)$

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where ρ_0 is a reference (background) density and g is gravitational acceleration. The gradient Richardson Number, $Ri_{gr} = N^2 / Sh_A^{\ 2}$ provides a measure of shear-induced instability whereby values substantially less than unity are likely to be or become unstable. Values of Ri_{gr} where N^2 was less that 10^{-8} s^{-2} were rejected (around 8% of data values).

3. Observations

3.1 Sea-ice conditions, regional circulation and tides

At the time of sampling McMurdo Sound sea ice cover was in retreat after a decade of record coverage due to the large icebergs of the early 2000's (Robinson et al. 2010). Pack ice and open water were present off Cape Royds some 20 km to the north of the sampling location. Much of the Sound from this line, down to a little south of the EGT was first year ice (Figure 1) with multiyear fast ice confined to the western shores and Erebus Bay. The tip of the EGT extends into 2.3 m thick first year ice beyond the fast ice by a few km. There was a residual multiyear ice bridge of ice at least 4 m thick running between the EGT and Big Razorback Island (**Figure 3**).

Available data on near-surface residual circulation are sparse but suggest a predominant northward current near the south of Erebus Bay (**Figure 2**), at least during winter (Leonard et al. 2006; Robinson 2011; Mahoney in prep.). The primarily diurnal tides in the region are around 1 m in peak-peak elevation (Goring and Pyne, 2003) with a

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14 day spring-neap cycle (Figure 4a-b). Tidal flows some 5 km to the SW of the EGT reached amplitudes of ~ 0.20 m s^{-1} (Stevens et al. 2006) with residuals of 0.05 m s^{-1}.

3.2 Velocity data

While the tides typically generate maximum flows of ~ 0.15-0.20 m s^{-1} in the Sound-proper (Barry and Dayton 1988; Leonard et al. 2006; Stevens et al. 2006; Robinson, 2011), at the MSC station we saw flow magnitudes reach 0.40 m s^{-1} (**Figure 4** c-d). The currents appear moderately cyclical (**Figure 4**c-d) but careful examination reveals the faster flows in each cycle were not phase-locked to the tide. The short duration of the dataset precluded any co-spectral analysis. There was a consistent weak westward flow. The faster flows occurred on the rising/high tide and were variable in magnitude and direction. The strongest flow was quite short in duration in each tidal period and so we refer to it here as a pulse, in an otherwise moderately quiescent water column.

A progressive vector diagram (**Figure 5**) that separated near-surface and deeper than the local depth of the EGT (~50 m) shows that there were some clear differences in apparent trajectory. There was partial flow rectification around day 323 with the deeper flow reversing whilst the shallower flow did not. This diagram shows the westward trend clearly along with the anti-clockwise sense of the rotation.

Focusing on the period when there were concurrent microstructure and CTD data (323.6-324.6) enables closer examination of the pulse. During this day (this was the strongest measured flow during the sampled period) the pulse was located on the rising phase of the tide (323.85-324.05) and, at the observation point at least, this was seen as

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flow acceleration initially at depths > 40 m (**Figure 6**) at which time shallower flows were towards the EGT suggesting a wake recirculation in the vertical. The pulse then expanded to fill the entire measured depth before decaying. The main body of the pulse lasted ~4 hours. It was followed by apparent oscillations whereby there were moderate flow accelerations for short periods spaced ~70 minutes apart. Directionally, the flow within the pulse was directed to the north-east whilst the flow before and after varied but with a net westerly direction. Flow magnitudes other than during the pulse were comparable with locations away from the coast (Stevens et al. 2006; Robinson, 2011).

The flow pattern included relatively strong vertical flows that reached up to 10 cm s^{-1} (**Figure 6**c). The main body of the pulse was preceeded by a downwards flow at the start, largely upwards flow during the pulse and finally with a strong increase at the cessation of the pulse. The vertical flows were reduced in magnitude near the surface. The post-pulse oscillations contained both up and down flows.

Backscatter amplitude variations as seen here (**Figure 6d**) often provide a qualitative picture of a tracer field (Leonard et al 2006; Stevens et al 2006). It is a difficult property to interpret as it is an integral whereby the value at a given depth is a function of the water column through which the acoustic beams must pass twice. Hence, a low acoustic backscatter at say 80 m might indicate few reflectors at that depth or a high degree of flow attenuation above the sample volume. There were three periods in the 24 hours when the backscatter was reduced sufficiently enough to affect the velocity signal to noise ratio adversely so that the deepest depth of good quality data would shoal. These reductions were pre- and post-pulse, as well as a sustained period occurring 12 hours after the pulse. This might be a diurnal cycle in scattering due to biological

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modulation as, despite the 24 hour daylight, a variation in signal can persist in the region (pers. comm. A. Mahoney; Leonard et al. 2010) or due to variations in platelet/frazil ice concentration (Stevens et al. 2006). Our sampling coincided with a period when the pulse was essentially around midnight local time so it was co-located temporally with what should be biologically-induced maximum data return (i.e. good signal to noise). Furthermore, this diurnal variation was not nearly as strongly apparent on other days. Thus, the drop in backscatter may have instead been a response entirely to the stronger flow. Curiously, at the very end of the data shown in **Figure 6**d there was a period when there was a local minimum in backscatter between 20 and 40 m depth. This effect was not seen in earlier measured tidal periods when the flow was weaker.

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3.3 Temperature and salinity

A diurnal cycle of CTD data revealed the very small thermal (8 mK) and salinity (0.04 PSU) ranges (**Figure 7**). The potential temperature and salinity fields remained quite constant except for a cool/fresh flow before and after the pulse which itself was warmer and fresher than background levels. The net dynamic result (**Figure 7**d) was a decrease in density just prior to the pulse when the local fluid was heavier than background levels. The trailing oscillations were then lighter, especially near the surface. The remainder of the diurnal period slowly increased in potential density with a lighter surface layer. The temperatures were well above the in situ freezing point (**Figure 7**b).

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3.4 Stability and turbulence

ADCP velocity shear magnitude Sh_A was typically in the order of 5x10^{-4} s^{-1}, although during the main pulse flow it reached six times this value i.e. 3x10^{-3} s^{-1} (**Figure 8a**). At the same time, the stratification persisted during the strongest section of the pulse flow as the N^2 was greatest during and just after the flow pulse (**Figure 8b**) when lighter fluid appeared in the upper 40 m of the water column and N^2 reached 10^{-5} s^{-2}. During the period of post-pulse oscillation N^2 was at detectable limits of ~10^{-8} s^{-2}. These values bracket other comparable observations in the Sound (Stevens et al. 2009) where N^2=3x10^{-6} s^{-2} was observed over the ridge running off Cape Armitage.

The turbulent energy dissipation rate ε reached a maximum of 10^{-5} m^{2}s^{-3} (**Figure 8c**). Such high dissipation rates were mostly confined to the upper 40 m of the measured water column for a period of ~ four hours. The post-pulse oscillations were aliased in the profiling so that any elevated ε events associated with these brief accelerations were not necessarily captured. After the post-pulse oscillations finished, the ε dropped to ~ 10^{-7} m^{2}s^{-3} and then, after the final backscatter minimum (time 324.5; midday 21 Nov. 2009), it fell, at least at depths greater than the thickness of the EGT, to the noise floor of the instrument.

4. Discussion

4.1 Kinematics and water column structure

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The present observations were at a time of extremely weak stratification. Of the three field seasons we have worked in the area these conditions were the most weakly stratified (**Figure 9**). Indeed even CTD profiles a week prior to the present data indicated substantially stronger stratification. Furthermore, the profiles were also the warmest and saltiest observed at this location. Wintertime results (Mahoney et al., in prep) from a site several km to the south of the present field site in Erebus Bay showed how fronts could pass through the observation site changing the local density structure markedly in only a few hours. These data end around a month prior to our measurements and at the end of their record their temperatures were around 0.02 ^{o}C colder than our observations – more in keeping with previous CTD work (**Figure 9**). Sound-scale analysis show how McMurdo Sound is still adjusting and recovering from the large iceberg residences of 2002-2005 (Robinson, 2011). The general trend is of a decreasing salinity – the opposite to that observed here. This suggests local water column properties might be quite different to that throughout the sound due to mixing induced by the complex topography and the presence of substantial glacial ice. Further to this we speculate that the T and S variation seen closely tied to the flow might be a direct manifestation of the flow-glacier interaction. Cold fresh water might pool in the vicinity of the glacier and then be washed away during the pulse.

Flow at the field site was likely a combination of tidal and Sound circulation flows that are affected by the islands, local bathymetry and the glacier tongue itself. There is unlikely to be direct wind influence as open water was 20 km to the north. The net flow was weakly to the north-west with the fastest flows (albeit for brief periods) in a north easterly direction. Sound scale circulation is believed to be southward on the east

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side of the Sound (Barry and Dayton 1988; Dinniman et al. 2007; Robinson 2011) although some authors suggest northward flow is possible (Lewis and Perkin 1985). It's worth noting that there has been no large scale survey undertaken that can answer this with any temporal perspective. Thus, we suggest our observations show the influence of tidal rectification whereby the close presence of topography results in the tide being asymmetric. Our results could be influenced purely by the presence of the EGT itself or the combined influence of it and Tent Island to the north.

The local maximum depth of the EGT (~50-60 m) was not readily apparent in any of the data. One might reasonably have expected a step in stratification or a strongly consistent velocity shear feature. That neither of these were apparent is possibly indicative of the fact that the underside of the glacier tongue is not level either along, or across, it's main longitudinal axis (Holdsworth, 1982). Even the relatively weak local stratification will retard vertical flows, so that the influence of the glacier tongue closer to the grounding line might still have a strong influence at the observation position thus spreading the depth of the effect on properties.

In the case of the near-EGT flow, the response of the flow past the headland will manifest itself in a manner depending on the stratification. In the present situation the post-pulse oscillations were comprised of accelerations around 70 minutes apart. There are a range of relevant frequencies to consider (Albrecht et al., 2006). Here N^2 varied from 10^{-8}-10^{-5} s^{-2} implying timescales of 17 hrs and 33 minutes respectively so it does cover the observed 70 minute period. Consideration of vertical velocity spectra both near the surface and deeper (**Figure 10**) shows the deeper water exhibiting greater periodic energy at frequencies <20 cpd. This suggests there is greater energy finding a pathway

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through the internal wave spectrum beneath the tongue. This is sensible as (i) vertical flows will attenuate close to horizontal surfaces, (ii) the fast ice boundary potentially generates a more homogenous surface layer less able to support internal waves and (iii) internal waves persist at depths other than that at which the obstacle exists (e.g. mountain lees waves).

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The Strouhal number $St=fL/u$ (f is shedding frequency, L is a body lengthscale either vertical or horizontal, and u is velocity) is in the range 0.03-0.3 (assuming f=2x10^{-4} s^{-1}, L=50-500 m and u=0.3 m s^{-1}). This intermediate flow regime suggests variable flow with buildup and release of vorticies (Sobey, 1982). Thus, on the inward flood tide into the Dellbridge Islands embayment, vorticies are possibly shed off the tip of the EGT. On the return tide the decaying waves are advected away from the EGT. While progressive vector diagrams like **Figure 5** should be interpreted with care it does illustrate the distances material might move in a few tidal cycles.

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Not only does melting of glacier and ice shelf walls have a strong influence on the ice dynamics (Rignot et al. 2010; Olbers and Hellmer 2010), it can also result in highly variable stratification near the wall, including double diffusive layering (Jacobs et al. 1981). Whilst the present study was initially motivated by the search for diffusive convection as observed in the Jacobs study, it is beyond the scope of the present work to compare the magnitudes of shear-induced and diffusion-induced mixing as we have no complementary observations of the latter situation. It seems highly likely that the pulse feature would sweep away layering structure generated by double diffusion. However, the pulse is transient and so for the majority of the time conditions were relatively quiescent and thus conducive to the accumulation of melt water. Hence, further

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exploration of the melting processes needs to incorporate temporal variation in the ambient conditions at timescales less than a tidal period.

4.2 Mixing rates

The energy dissipation rate (ε) rose - albeit for localized moments in time and presumably space - as much as three orders of magnitude above what might be regarded as normal background levels. Edwards et al. (2004) noted how local turbulent buoyancy flux scaled with ε and that it is likely maximum very close to a headland. Recent modeling efforts (Dinniman et al. 2007; Reddy et al. 2010) that encompassed the region were at scales that are unlikely to be instantaneously influenced even by this degree of variability. Evidence from elsewhere suggests such sub-grid scale mixing processes can be influential at the Sound scale (e.g. Xing and Davies 2010).

The Ri_{gr} estimated here is rather intermittent as it is the ratio of two derivative properties so that noise and uncertainty is amplified. However, the depth-time distribution of Ri_{gr} does have some consistent structure (**Figure 8**d) so that in the core of the pulse Ri_{gr}<1, implying mixing due to shear is likely and this was consistent with the dissipation rate observations. At times away from the pulse feature, stratification persisted more strongly so that N^2=10^{-5} s^{-2} and background shear was more like Sh_A^2=3x10^{-8} s^{-2}. Under these conditions Ri_{gr}~300, implying mixing was unlikely.

Efforts continue to determine the functionality of vertical mixing in response to the balance captured in Ri_{gr} (e.g. Fer, 2006; Zaron and Moum 2009). The K_ρ estimates presented here use the Shih et al. (2005) estimate for K_ρ $\left(=2\nu\left(\langle\varepsilon\rangle/\nu\left(\langle N^2\rangle\right)\right)^{1/2}\right.$ where <>

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denote ensemble averages) which are, like the comparison made in Stevens et al. (2009), somewhat higher than the Osborn method (Osborn, 1980).

In order to relate diffusivities to stability Fer (2006) suggests modification of the background levels in the Pacanowski and Philander (1981) formulations for vertical diffusivities of momentum and buoyancy (K_m and K_ρ). The modifications were based on observations in an Arctic fjord and so likely relevant here due to the proximity to topographic variations. The parameterizations are given as

$$K_m = \frac{5.5\times10^{-3}}{(1+5Ri_{gr})^2} + K_{m0} \qquad (1)$$

$$K_\rho = \frac{K_m}{(1+5Ri_{gr})} + K_{\rho 0} \qquad (2)$$

where background levels are given by K_{m0}=1.3x10^{-3} m^{2}s^{-1} and $K_{\rho 0}$=1.3x10^{-4} m^{2}s^{-1}. There are trends between the estimate of the Ri_{gr} and the measured dissipation and inferred vertical diffusivity K_ρ (**Figure 11**). The diffusivity is somewhat higher than the model determined by Fer (2006) – especially in the Ri_{gr}=1-10 band.

The Fer (2006) model uses the Osborn method to determine (1) and (2) which it notes is an upper bound (see Zaron and Moum 2009). The difference possibly lies in the likely small spatial scales. The observed N^2 may not be particularly representative of the conditions that spawned the mixing. It is useful to consider the relationship between displacement and energy dissipation. Assuming a timescale of ~35 minutes associated with the half-period of the oscillations (i.e. the period motion will be in a particular direction) it is possible to place the dissipation rate estimates on a displacement diagram (**Figure 12**). It is clear that the eastward flowing periods generated the strongest dissipation rates and the scales of the "action at a distance" were comparable with the

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distance to the tip of the glacier tongue. Thus, the observations downstream might potentially be the result of direct flow-obstacle interaction. There are strong similarities between the velocities, diffusivity and dissipation rates observed here and by Edwards et al. (2004). Finescale modeling might elucidate the distribution and longevity of such mixing hot spots.

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4.3 Local supercooling

A number of authors have inferred that glacier/ice tongues might be generators of supercooled water (Debenham 1965; Jeffries and Weeks 1992). Such conditions arise when seawater in contact with ice at depth is cooled to the in situ freezing point. If, at some later time it rises in the water column, the increase in local freezing temperature may result in the fluid actually being colder than this temperature. This substantially increases its propensity to freeze. While this is unlikely in the particular relatively warm period observed here, there have been previous times even at the same time of year when it has been possible as shown in CTD data of **Figure 9**. Thus, the presently observed kinematics are relevant to the supercooling problem. The ADCP data do show a weak upward flow for most of the time. Furthermore, as the post-pulse oscillations had a half-period timescale of 35 min and a vertical speed of ~ 0.05 ms^{-1} then the individual events might be expected to advect water ~ 100 m vertically in this time. This displacement represents a -0.075 ^{o}C change in the freezing point and so highly influential if the local water is very cold to begin with.

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Platelet growth rates are very difficult to quantify as they inherently depend on their initial conditions and observations have difficulty in separating advection versus actual growth. However, Leonard et al. (2006) observed cylinders of platelets 30 cm in diameter growing on near-surface lines after periods as short as three days. Consequently if, as in the EGT situation, water advected up from beneath the tongue were immediately returned to depth it is unlikely there would be any effect. On the other hand, if as the observations suggest (**Figure 5**), the flow were rectified in some way then platelet growth might ensue. Similar temporally variable generation possibilities exist in the internal waves observed at Cape Armitage 17 km to the south of the EGT (Robinson, 2011). Thus, these highly localized features could potentially act as "ice factories".

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4.4 Generalization, conclusions and future work

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Sharp headlands strongly influence local flow through rectification of tides (e.g. Signell and Geyer 2991; Edwards et al. 2004). Although the EGT's finger-like morphology is at the extreme end of the headland aspect ratio scale and glacier tongues are located at the surface rather than the bed, we expect there is much in common in terms of oceanic response to tidal flow. The headland work of Edwards et al. (2004) identified the split between skin and form drag and that the majority of the diapycnal turbulent buoyancy flux occurred close by the headland. This supports the contention that more turbulence data very close to the glacier wall would help to identify the dynamic importance of such features.

When considering other floating glaciers, the EGT has quite a large aspect ratio (length/width) when compared to the Drygalski Ice Tongue (and the Mertz and Ninnis Glacier Tongues prior to their recent break ups). At the same time, the EGT is smaller in absolute scale by a factor of ten from these giant glacier tongues. Possibly the EGT is atypical in that it is protected from currents and waves by the Dellbridge Islands and the generally low tidal flows in the area. Consequently, the relative scale of separated flow to that of the tongue is likely much larger with the EGT. However, vertical scales are greater for the big ice tongues so that vertical flows may be even stronger. Where the present data are of relevance to both the larger ice tongues and indeed ice shelf fronts is at the small scale. The EGT observations should be relevant wherever there are comparable dramatic changes in topography and glacier-induced buoyancy fluxes.

The work here identified the complex flows that persist around glacier tongues. It demonstrated that the resulting mixing can be quite substantial. The data demonstrate the small temporal scales of variability in turbulence, the strong and fluctuating vertical flows, and the very large dissipation rates and diffusivities possible right next to the ice wall. Future work on the topic should seek to resolve spring-neap differences in mixing as well as relate the local mixing to the far-field transport in the Sound. Furthermore, headland studies (Edwards et al. 2004) have demonstrated the effectiveness of spatial mapping of currents. However, such mapping is difficult in the presence of fast ice requiring underwater vehicle technology (e.g. Hayes and Morison 2002; Doble et al. 2009). Modelling needs to progress at two scales. Over long timescales continual elevated mixing at a sub-grid location will influence circulation. Intermediate scales of regional modeling would elucidate this influence whilst finescale modeling would

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A subjective series of sketches (**Figure 11**) shows the likely flow evolution around the EGT. The tidally-driven flow is highly rectified and also variable with depth and so generates quite specific mixing events in time and presumably space. It would be useful to translate this description to other systems. ¶

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generate a valuable picture of where and how the mixing is generated. This modeling would need to be informed by velocity, scalar and turbulence measurements in a wide range of background stratification regimes.

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FIGURE Captions

Figure 1 Satellite image (SAR) captured on 19 October 2009 showing McMurdo Sound located in the south western Ross Sea connected to the Ross Ice Shelf cavity via Haskell Strait beneath the McMurdo Ice Shelf. The greyscales give an indication of open water and pack ice off Cape Royds and the first year and multiyear ice in the Sound. The inset shows the locale within the western Ross Sea.

Figure 2 ASTER satellite image of south east McMurdo Sound including the Erebus glacier tongue (EGT), the Dellbridge Islands (DI) including Tent Island (TI) and Big Razorback Island (BRI), Erebus Bay (EB), Cape Evans (CE). Cape Armitage (CA), Haskell Strait (HS), Scott Base (SB) and the microstructure field camp (MSC). The dashed box shows the locale for Fig. 3a.

Figure 3 (a) The sampling locale including the microstructure field camp (MSC), around 1000 m east of the tip of the EGT. The grey lines highlight the known 100 m contour whilst the blue dashed line shows a tongue of multi year fast ice (i.e. ice connected to the coast or glacier) extending out from the EGT to Big Razorback Island. (b) A cross-section of the EGT simplified from DeLisle et al. (1989).

Figure 4 The tidal elevation (a) at Scott Base is expanded (b) to highlight a four day segment within which our ADCP (blue box in a) and turbulence profiling (green box in all panels) were recorded. Easterly and northerly currents, roughly approximating along

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and across glacier directions) are shown for depths of (c) 11 and (d) 71 m. The grey bar was a period of no data return from the 71 m depth range.

Figure 5 Horizontal velocity data as a progressive vector diagram showing 11 (black) and 71 (blue) m depths. The start of each day is marked.

Figure 6 ADCP data products. The ADCP resolved (a) eastward velocity, (b) westward velocity, (c) upward velocity w, and (d) backscatter amplitude. The approximate local depth range of the glacier is shown as horizontal dashed lines. The tidal elevation is shown above panel (a).

Figure 7 CTD scalar data products showing (a) potential temperature, (b) degree of supercooling (in situ temperature-local freezing temperature), (c) salinity and (d) potential density. The approximate local depth range of the glacier is shown as horizontal dashed lines. The tidal elevation is shown above panel (a).

Figure 8 Calculated properties including (a) ADCP-resolved shear Sh_A and VMP profiler-resolved (b) buoyancy frequency squared N^2, (c) turbulent energy dissipation rate ε and (d) the Ri_{gr}. The approximate local depth range of the glacier is shown as horizontal dashed lines. The tidal elevation is shown above panel (a).

Figure 9 (a) Potential temperature and (b) salinity profiles at various locations within 500 m of the tip of the EGT from the Oct.-Nov. period during 2005, 2008 and 2009, plus a single representative example from the present microstructure data (2009-ms). The

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Figure 10 Spectra of ADCP-derived vertical velocities from depths 9-17, 35-43 m and 59-67 m. The spectra are shown scaled by the frequency vector to provide an area-preserving form.

Figure 11 (a) Dissipation rate as a function of Ri_{gr} showing individual estimates (hollow squares), bin-log-space-averaged (solid squares) and bars showing standard deviation of estimates for 0.15 log(Ri_{gr}) spaced bins. (b) The same plot format for the diffusivity K_ρ estimate and including the Fer (2006) estimate of vertical diffusivity of mass (dashed line).

Figure 12 Dissipation rate as a function of pseudo "horizontal displacements" X and Y using a timescale of 35 minutes.

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do not yield any strong differences associated with direction of rotation except for (i) a weak emphasis on clockwise motion at lower frequencies and (ii) a significant difference at around 20 cpd. This latter difference might be associated with the "ringing" timescale seen in the ADCP data (**Figure 3**). When comparing spectra from the two depths, the only obvious difference is the different slopes in the 1-24 cpd bands,