Project Summary
Collaborative Research: Critical Layers and Isopycnal Mixing in the Southern Ocean.

Raffaele Ferrari and John Marshall, Massachusetts Institute of Technology
Kevin Speer, Florida State Univeristy

Summary. Climate-scale ocean models unanimously stress the key regulatory function played by the oceanic overturning circulation in the Earth's climate and biogeochemical cycles over decadal and longer time scales. Yet in their quest to resolve many topical climate problems, the models credibility is challenged by their extreme sensitivity to the representation of mixing processes in the Southern Ocean. This peculiarity of model behaviour reflects the unique role of mixing in mediating the vertical and horizontal transports of water masses in the Antarctic Circumpolar Current (ACC), which shape the overturning circulation through their respective impacts on the overturning rate and inter-ocean exchange. The Diapycnal and Isopycnal Mixing Experiment in the Southern Ocean (DIMES) has been recently funded to measure directly eddy mixing along density surfaces in the ACC. Eddy mixing will be measured by releasing a chemical tracer and 150 floats at one vertical level. New theoretical results, not available when the DIMES proposal was written, suggest that eddy mixing rates are strongly enhanced at critical levels in the vertical. The goal of this proposal is to extend the DIMES project and release 50 additional floats at a shallower level to test the hypothesis that critical levels control the rate of upwelling and downwelling of water masses in the Southern Ocean.

Intellectual Merit. Conceptual models of global meridional overturning and numerical predictions for future climate are strongly sensitive to the methods used to represent mixing along and across the ACC. Theory suggest that mixing rates vary greatly in the horizontal and in the vertical. Climate ocean models unanimously stress that model skill is strongly sensitive to these variations. The DIMES project will provide the first direct observations of mixing in the Southern Ocean and will likely deliver a wealth of new information about eddy transport in this part of the ocean. However tracer and float deployments are planned only at one level and will not provide information about the vertical variability of eddy mixing. An addition of 50 floats to be deployed at a shallower level is all that is needed to make sure we do not lose the opportunity to learn about a key aspect of eddy mixing in the Southern Ocean.

Broader Impacts. This proposal has potentially wide impact because it is designed to further our understanding of a central component of the climate system. The proposed work will also contribute to the improvement of mixing and stirring in large-scale ocean models such as the MITgcm. Finally, there is strong educational component through the training of a graduate student and a postdoc, and the development of new curricula to introduce students in the MIT/WHOI Joint Program to the role of the Southern Ocean in the climate system.

Project Description
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1 Introduction

The Meridional Overturning Circulation (MOC) of the ocean is a critical regulator of the Earth's climate and biogeochemical cycles over time scales of decades to millennia (Rintoul et al., 2001b; Sarmiento et al., 2004). Through its action, heat, carbon and other climatically important tracers are distributed around the globe and stored in the deep ocean. Yet in the quest to understand the changing climate system, climate-scale ocean models are confronted by a hurdle: their acute sensitivity to the representation of mixing and eddy processes, particularly in the Southern Ocean (Gregory, 2000; Gnanadesikan et al., 2004). The Southern Ocean is the place where water that sinks in the polar regions of the North Atlantic rises again to the surface with wind, buoyancy, eddy and mixing processes all potentially playing a key role: see, for example, Deacon (1984), Toggweiler and Samuels (28), Speer et al. (2000) and Wunsch and Ferrari (2004) — see Fig. 1. Eddy and mixing processes act to shape the baroclinic structure and transport of the Antarctic Circumpolar Current (ACC) and thus contribute decisively to regulating inter-ocean exchange. It is this unique role of eddies and mixing in mediating the vertical and horizontal transport of water masses in the ACC that underlies the Southern Ocean's importance in the global MOC.

NSF has recently funded the Diapycnal and Isopycnal Mixing Experiment in the Southern Ocean (DIMES) whose primary goal is to measure mixing along and across isopycnals. DIMES includes a tracer release in the Upper Circumpolar Deep Water (UCDW), float releases in the same layer, and measurements of finestructure and microstructure from various platforms. The tracer and floats will be released in the Southeast Pacific, with much of the tracer and many of the floats passing through the Scotia Sea during the 3-year duration of the experiment (Fig. 5).

Recent work strongly suggests that isopycnal mixing rates vary markedly in the vertical (by a factor of 5): see Treguier (1999), Eden (2007), Cerovecki et al. (2008), and Smith and Marshall (2008). The vertical variations appear to be much larger than previously recognized and would induce large changes in the structure and strength of the MOC in the Southern Ocean. Gnanadesikan et al. (2007) also find that mixing rates must be allowed to vary in the vertical and horizontal, if models are to reproduce the observed distributions of physical and biological tracers in the Southern Ocean. DIMES plan is to measure isopycnal mixing by releasing floats and tracers at just one level (isopycnal surface 27.9, nominally around 1300 meters depth). The evidence for vertical variations in isopycnal mixing rates has now come in to sharp focus. It has prompted us to write this proposal to extend the DIMES project to address the question of the vertical variability of isopycnal mixing in the Southern Ocean. There are three components:

1. Deployment of 30 floats from this proposal (plus 20 from DIMES) at a shallower level (isopycnal surface 27.21, nominally around 500 meters depth) on the second deployment cruise of the DIMES project in austral summer 2009/2010.

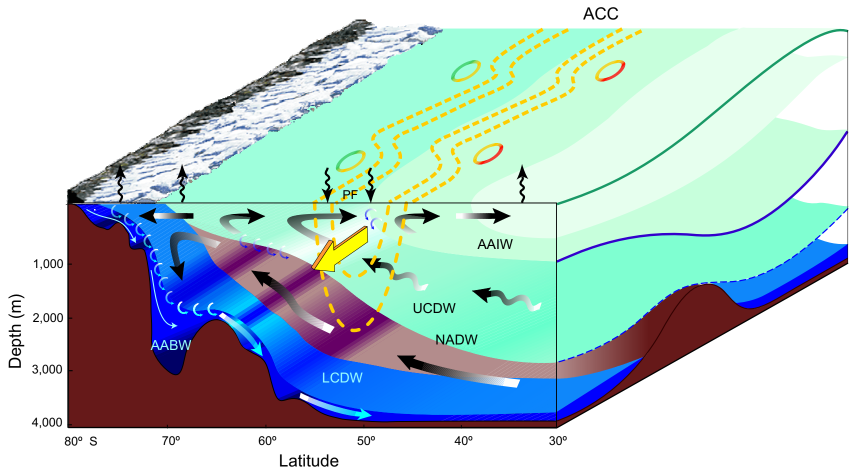

Figure 1: A schematic of the zonal and meridional circulations in the ACC system. Antarctica is at the left side. The curly arrows at the surface indicate the atmospheric buoyancy flux. The curly arrows in the interior represent the transport by geostrophic eddies. Upper Circumpolar Deep Water (UCDW) upwells in the core of the ACC, while Antarctic Intermediate Water (AAIW) is formed north of the ACC (Olbers et al., 2004).

2. Calculation of eddy statistics and eddy mixing coefficients in the DIMES region with output from the Southern Ocean State Estimate (SOSE), a state estimate based on observations and the MIT general circulation model.

3. An investigation of the differences in isopycnal mixing rates as estimated from floats and tracer release experiments. This work will provide the foundations for interpreting the DIMES field observations.

DIMES is the first process study focused on quantifying mixing rates in the Southern Ocean. No such measurements have ever been made in the ACC, so we are bound to learn a great deal about this part of the ocean. However DIMES will be a success only if the new measurements can provide useful constraints on the role of mixing in regulating the MOC and its impact on climate. Modeling studies find the along-isopycnal diffusivity is a key parameter influencing the MOC and its associated water mass properties. The stumbling block is that the traditional approach of using a constant diffusivity coefficient, no matter what value is chosen, seems inconsistent with the observed distributions of physical and biological tracers. Model skill can only be improved by introducing spatially varying diffusivities (e.g. Danabasoglu and Marshall, 2007). The DIMES project is well positioned to quantify the rates of mixing at one vertical level. Given the recent evidence for vertical variability in isopycnal mixing, we propose to complement the DIMES study and address the question of the vertical distribution of isopycnal mixing rates. If this proposal is funded, the DIMES experiment will be in a much better position to provide information necessary to improve the skill of climate models in the Southern Ocean.

2 The Southern Ocean Overturning Circulation

The Southern Ocean is of fundamental importance to the global climate system both through its strong zonal currents and through its weaker meridional circulations (Rintoul et al., 2001b;

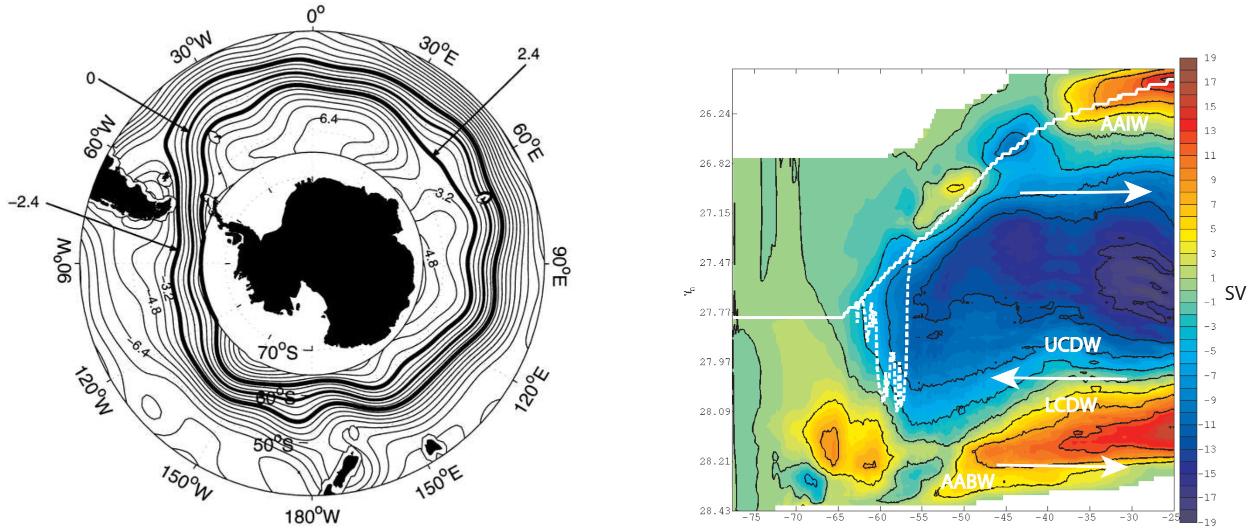

Figure 2: (Left) Climatological surface geostrophic streamfunction in units of 10^4 m^2/s as observed by altimetry (thick lines) and large-scale surface buoyancy structure (thin contours). (Right) The zonally-averaged meridional overturning circulation (MOC) as a function of neutral density. The MOC is computed from the State Estimate Southern Ocean, an ocean model constrained with in-situ and altimetric observations. The white line is the mean mixed layer depth. The white dashed line is the topographic height across Drake Passage.

Wunsch and Ferrari, 2004). First, the eastward flow of the ACC connects the Indian, Atlantic and Pacific Ocean basins. The resulting global circulation redistributes heat and other properties, influencing temperature and rainfall patterns, and allows teleconnections between remote regions. Second, the Southern Ocean renews the world ocean's waters through its MOC. The MOC sets the distribution of the physical and chemical properties of the deep ocean – not just in the Southern Ocean but throughout the world oceans – and controls the rate of ocean uptake of heat and carbon dioxide.

The surface geostrophic streamfunction computed from altimetry is shown in Fig. 2. Buoyancy surfaces extend up to the surface, outcropping around Antarctica. The along-stream current is in thermal wind balance with this interior cross-stream buoyancy gradient. The eastward wind-stress increases equatorwards across the stream to reach a maximum just equatorward of the region of circumpolar ACC flow shown in Fig. 2. The mean air-sea buoyancy flux is out of the ocean around the Antarctic continental shelf, where Antarctic Bottom Water (AABW) sinks into the abyss through convection, and into the ocean north of 60 S providing the buoyancy required to allow Upper Circumpolar Deep Water (UCDW) upwelling around Antarctica to eventually subduct in a much lighter density range, that of Antarctic Intermediate Water (AAIW) and Subantarctic Mode Water (SAMW).

The pattern of the MOC in the Southern Ocean is not well documented. A number of attempts have been made to infer it from observations — see Sloyan and Rintoul (2000), Speer et al. (2000), Karsten and Marshall (2002) and the review of Rintoul et al. (2001a). Here we show a new estimate from an ocean model, the MITgcm, constrained to satellite and in-situ observations [The model is an important component of this proposal and will be described in more detail below.]

The meridional overturning is computed by zonally integrating the meridional velocity within neutral density layers (neutral density is water density after variations due to pressure have been removed). The MOC pattern is of two major separate overturning cells as shown in Fig. 2. The blue cell includes upwelling of UCDW around Antarctica of some 16 Sv, equatorward flow at the surface and subduction of perhaps 20 Sv of AAIW and SAMW just equatorward of the ACC, where a surface eddy driven circulation pushes mixed layer waters poleward. A lower overturning cell is associated with the formation of AABW.

The Southern Ocean MOC cannot be explained solely in terms of large-scale climatological forcing and currents as in the case of mid-latitudes. In the latitude band around Drake Passage, at depths where no topography exists to support zonal pressure gradients, there can be no mean meridional geostrophic flow and the meridional transport is only through eddy motions. The eddy contribution to the meridional transport shown is best illustrated by separating the transport within isopycnal layers into mean and eddy contributions,

$$\underbrace{\overline{vh}}_{\text{Total transport}} = \underbrace{\bar{v}\bar{h}}_{\text{Mean transport}} + \underbrace{\overline{v'h'}}_{\text{Eddy transport}}, \tag{2.1}$$

where v is the meridional velocity, h is the thickness of the isopycnal layer, the overbar denotes a zonal average and primes departures from that average. In the ACC latitude band the mean transport $\bar{v}\bar{h}$ is dominated by a single wind driven cell flowing equatorward at the surface and returning poleward at the ocean bottom, the so-called Deacon cell. This single cell cannot be seen in Fig. 2, because it is largely cancelled by eddy transport due to correlations between the meridional velocity and the isopycnal thickness $\overline{v'h'}$. A theory of Southern Ocean circulation must include both mean and eddy transports (e.g., Marshall and Radko, 2004; Olbers and Visbeck, 2005).

The structure and magnitude of the meridional transport in the Southern Ocean can be understood in terms of the the zonal momentum equation for a density layer (Olbers et al., 2004),

$$-f\overline{vh} = \bar{h}\overline{v'P'} - f\overline{M_E} - \rho_0{}^{-1}\bar{h}\overline{p_x}. \tag{2.2}$$

where f is the Coriolis frequency, $P = (f + v_x - u_y)/h$ is the Ertel potential vorticity, M_E is the wind-driven Ekman transport, p is pressure and ρ_0 a reference density. The overbars denote zonal averages along a neutral density layer. The *isentropic eddy PV flux* is defined as $\bar{h}\overline{v'P'} = \overline{hvP} - \overline{vh}\bar{P}$. Eq. (2.2) holds if the Rossby number of the large scale circulation is small, a condition well satisfied in the ACC, and if diabatic forcing is weak, which is believed to be the case in the UCDW cell. The isopycnal mass transport is driven by the three terms on the right hand side of Eq. (2.2): eddy forcing through an isentropic PV flux*, the wind-driven Ekman transport, and pressure forces acting against bottom ridges or continental margins. At midlatidues, winds and pressure forces dominate the budget. But in the ACC eddy forcing enters at leading order.

There is a rich literature on the effect of the isentropic eddy PV flux on the Southern Ocean circulation (Bryden, 1979; deSzoeke and Levine, 1981; Cessi and Fantini, 2004; Henning and Vallis,

*Schneider (2005) shows that the isentropic Ertel PV flux at the ocean surface becomes a horizontal flux of buoyancy, much like in the quasi-gesotrophic approximation, where the surface flux of quasi-gesotrophic potential vorticity is given by the surface flux of buoyancy.

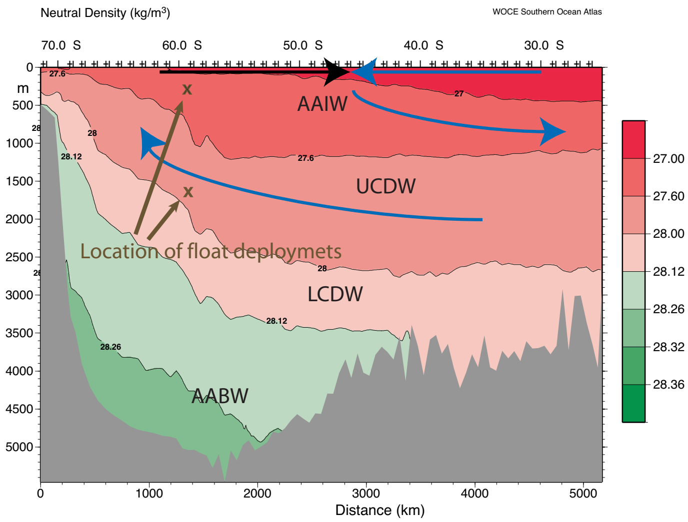

Figure 3: Meridional neutral density section in the Pacific sector of the Southern Ocean along 90 W (WOCE cruise P19). The black line indicates the Ekman transport, $f\bar{M}_E$ in Eq. (2.2). The blue lines sketch the eddy driven transport down the mean gradient of layer thickness, i.e. the PV gradient. The interior flows are along density surfaces because at these depths diabatic processes are weak. Eddies drive upwelling of UCDW and subduction of SAMW/AAIW.

2004; Marshall and Radko, 2006). But the theories require more quantitative underpinning. The goal of this proposal is to use a combination of observations, theory and modeling to quantify the eddy PV flux in the Southern Ocean. We phrase the problem in terms of the isopycnal eddy diffusivity K that relates the isentropic eddy PV flux to the meridional PV gradient averaged over a density layer,

$$\overline{v'P'} = -K\frac{\partial}{\partial y}\left(\frac{\overline{Ph}}{\bar{h}}\right) \approx -K\frac{\beta}{\bar{h}} + K\frac{f}{\bar{h}^2}\frac{\partial \bar{h}}{\partial y}. \tag{2.3}$$

Theory suggests the isentropic eddy PV flux is downgradient, i.e. K is positive (see Rhines and Young, 1982; Plumb and Ferrari, 2005). Hence the sense of the eddy-driven circulation can be inferred from hydrographic measurements of $\bar{h}$. In Fig. 3 we show a WOCE neutral density section from the SE Pacific sector of the ACC that will be sampled during DIMES. In this region the PV gradient is dominated by changes in thickness. The eddy PV flux drives a poleward transport of UCDW and an equatorward flow of SAMW/AAIW (blue arrows in Fig. 3). The interior transport is approximately along density surfaces because at these depths diabatic processes are believed to be weak. There is an additional poleward flow at the surface that acts against the Ekman flow. This eddy transport crosses density surfaces, because the surface waters are exposed to strong atmospheric forcing.

Speer et al. (2000) estimate that with $K = 1000$ m^2/s these gradients give a total UCDW transport of 10 Sv and a weaker SAMW/AAIW transport. But these estimates are very sensitive to the value chosen for K. So what is the value and the vertical structure of K?

2.1 Spatial variations of eddy diffusivities: the critical layer hypothesis

The rate of isopycnal stirring by planetary waves is strongly modulated by the subtropical and polar jets in the atmosphere (Andrews et al., 1987). Marshall et al. (2006) found that eddy diffusivities in the Southern Ocean are similarly modulated by the strong ACC jets. They found

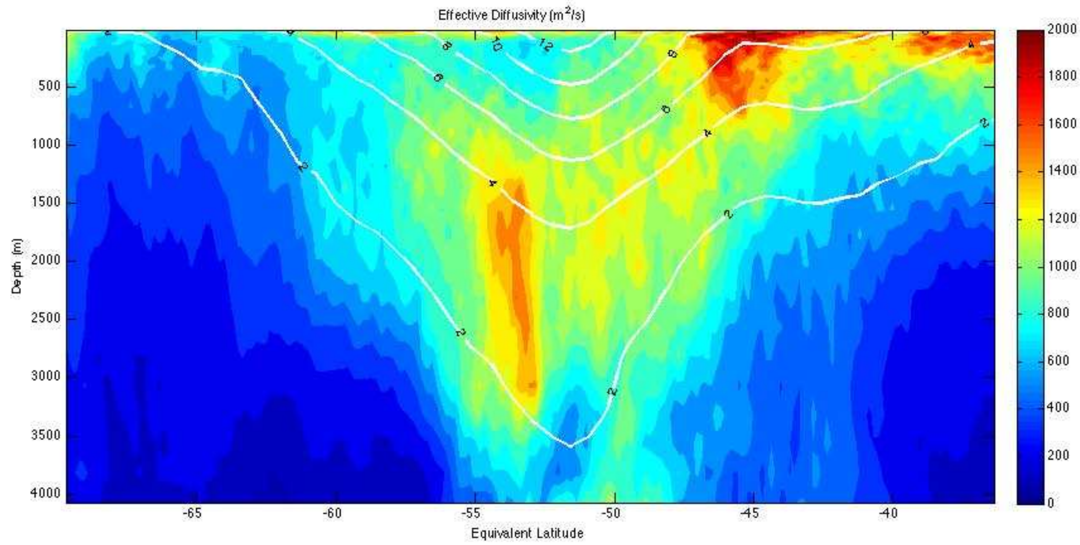

Figure 4: Isopycnal eddy diffusivity in the Southern Ocean, estimated from tracers advected with the geostrophic velocity from an eddy admitting Southern Ocean State Estimate. The figure shows K_{Nak} (see Sec. 3.1) averaged along circumpolar geostrophic contours.

the surface diffusivities are large (2000 m^2/s) on the equatorward flank of the ACC and small (500 m^2/s) at the jet axis. The suppression of eddy diffusivities K at the jet axis has a simple explanation. Altimetric observations show geostrophic eddies in the ACC do not propagate westward, as in midlatitudes. Instead they are swept downstream (eastward) by the strong ACC current as expected if the eddies are the result of baroclinic instability (e.g. Smith and Marshall, 2008). This downstream propagation reduces the efficiency of eddy mixing, because the mean current sweeps the tracer out of the eddy before much filamentation has occurred.

Smith and Marshall (2008) show the mean current speed decays more rapidly with depth than the eddy wavespeed. As a result, at a depth of about 1-2 km, the eddy wavespeed matches the mean current and a critical level develops. Critical levels are regions of enhanced K (see Green, 1970) because eddy perturbations propagate at the same speed as the mean flow and keep stirring the same tracer element achieving strong filamentation. Treguier (1999), Smith and Marshall (2008) and Cerovecki et al. (2008) do find in idealized simulations of the ACC current system K is enhanced at depth in correspondence with critical levels. Furthermore they show the critical levels are deep in the core of the jet (between 1 and 2 km), but they come closer to the surface on the flanks of the jet where the mean flow speed is weak.

In Abernathey et al. (2008) we computed K from an eddy admitting Southern Ocean State Estimate (see Sec. 4.2). Following the approach taken in Marshall et al. (2006), we estimated K by numerically monitoring the lengthening of idealized tracer contours as they are strained by the geostrophic velocity obtained from the state estimate. We found K is indeed enhanced at depth in the core of the ACC and the maximum comes closer to the surface on the flanks of the jet – see Fig. 4 . The region of maximum K tracks very closely the critical levels estimated from linear theory. [We computed the critical levels based on the phase speed of the most unstable linear eigenmodes in the ACC region. See Smith and Marshall (2008) for details.] The figure shows K averaged along circumpolar geostrophic contours. Calculations on small sectors of the ACC show similar patterns with an increase of K by a factor of 4-5 at the critical levels compared to the core of the jet. The absolute values of K are however quite variable from sector to sector.

Comparing the K distribution in Fig. 4 with the sketch of the eddy-driven overturning circulation in Fig. 3, we see the enhancement of K at depth acts to strenghten the volume of UCDW upwelling in the Southern Ocean, while the increase in K on the flanks of the jet acts to strenghten the subduction of SAMW/AAIW. A careful quantification of the patterns of K is essential if we are to estimate the water mass circulation in the Southern Ocean. Present estimates, either based on analysis of observations or numerical simulations, do not account for the strong variations of K. The goal of this proposal is to collect new observations and develop the necessary theory to assess the impact of variations of K on the Southern Ocean MOC.

3 The Diapycnal and Isopycnal Mixing Experiment

The goals of DIMES are to obtain measurements to quantify both along-isopycnal eddy-driven mixing and cross-isopycnal interior mixing in the Southern Ocean. To reveal these processes at work in the ACC, a chemical tracer, trifluoromethyl sulfur pentafluoride (CF_3SF_5), and 75 floats that follow the water along isopycnal surfaces will be released in the ACC near 1300 m depth, 60 S, and 110 W, early in 2009. Vertically profiling floats that measure fine-structure T, S, and velocity within and above the tracer cloud will be released at the same time. The floats and tracer will be carried by the ACC over the relatively smooth bottom of the SE Pacific, spreading both across and along the current as they travel. After a year, the leading edge of the tracer will just start to pass over the ridges of Drake Passage into the Scotia Sea (see DIMES proposal). Another 75 isopycnal floats will be released near the center of the tracer patch at this time. Trajectories of the floats, measured acoustically with an array of sound sources, and the spreading of the tracer will be used to compute the isopycnal eddy diffusivity as explained below. The eddy field, and its vertical structure, will be studied with sea surface height measured by satellite altimeters, and with hydrographic profiles taken from research vessels and from autonomous instruments drifting with the tracer. Turbulent dissipation, from which diapycnal mixing can be estimated, will be measured with ship-based free-falling profilers, special floats drifting with the tracer and floats that profile between the surface and the tracer layer.

The progression of the tracer and the floats across the study region will set the pace and the evolution of the experiment (Fig. 5). The bathymetry beneath the Pacific sector of the ACC immediately to the east of 110 W, where the tracer will be released, is relatively smooth and the eddy energy is relatively weak. In contrast, Drake Passage and the Scotia Sea are rough, and the eddy field there and north of the Falkland Ridge is intense. Hence, in the first stage of the experiment the DIMES project will sample and quantify a region of relatively weak isopycnal and diapycnal mixing in the ACC. Once the tracer and floats pass into and through Drake Passage, the circumpolar current is characterized by high mixing and transport parameters. These two regions are chosen to provide data bounding the extremes of diapycnal and isopycnal mixing.

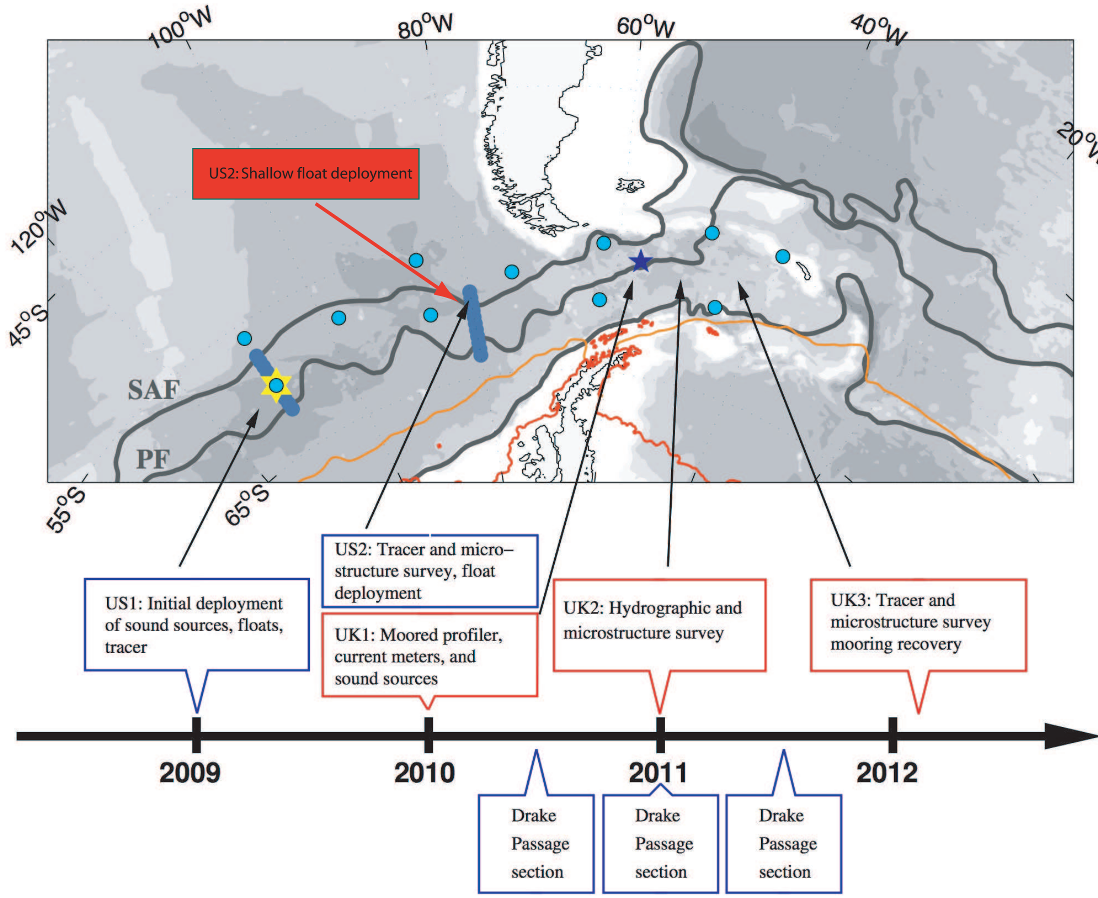

Figure 5: Time line for DIMES field activities. Text boxes at the bottom give the timing of the field activities. The yellow star with cyan dot indicates the tracer deployment location, blue circles show tentative float deployment sites, and cyan dots indicate locations of sound sources. The blue star in the Scotia Sea marks the site of the U.K. mooring array. The gray shading indicates the bathymetry, with 1000-m contour interval; the climatological locations of the Subantarctic Front (SAF) and Polar Front (PF) (Orsi et al., 1995) are labeled; the remaining lines mark the maximum (orange) and minimum (red) seasonal sea-ice extent.

3.1 Estimating isopycnal diffusivities from DIMES observations

Three techniques will be used to estimate K in the DIMES experiment. The first is based on the spreading of a tracer patch (Ledwell et al., 1998; Polzin and Ferrari, 2004). A tracer will be released on a density surface and K will be estimated from the spreading rate of the area occupied by the tracer. The expectation is that, after an initial transient when the tracer becomes streaky and distorted, an equilibrium is reached between the generation of new filaments by eddy stirring and the merging and homogenization of filaments by molecular dissipation. During this equilibrated stage, the tracer area is predicted to increase linearly in time at a rate proportional to the eddy diffusivity,

$$K_{Tracer} = \frac{1}{2}\frac{\mathrm{d}\sigma_y^2}{\mathrm{d}t}, \qquad \sigma_y^2 = \int (y - y_c)^2 c \, \mathrm{d}A, \tag{3.4}$$

where c is the tracer concentration, y is the direction normal to the ACC flow, and y_c is the y-coordinate of the center of mass of the tracer. The tracer will be resampled only 12 months after the injection. While this is too long a period to provide information about variations of K_{Tracer} along the isopycnal surface, it will provide an integrated picture of the eddy stirring across the ACC.

A second method to compute K from tracer distributions has been introduced by Nakamura (1996). It is based on the fact that nondivergent flows are area preserving and only molecular diffusion can change the area enclosed by tracer contours. Turbulent eddies, however, enhance the interface available for diffusion by twisting and folding tracer contours. Nakamura showed in two dimensions the eddy enhancement over the background molecular diffusion is given by,

$$K_{Nak} = \kappa L_{contour}^2 / L_0^2, \tag{3.5}$$

where κ is the background molecular diffusivity, $L_{contour}$ is the observed length of a tracer contour, while L_0 is the minimum (unstrained) length of the contour. This approach is suited to estimate dispersion by geostrophic eddies whose velocity is, to leading order, two dimensional and divergenceless. Nakamura's approach cannot be applied directly to tracer measurements, because it is impossible to accurately measure all the convoluted tracer filaments in a field campaign. Instead, the approach is to advect numerically an idealized tracer using the geostrophic flow from observations and estimate $L_{contour}$ from the resulting tracer distribution. This strategy has proved very useful in atmospheric and oceanographic studies, because if κ is sufficiently small, K_{Nak} does not depend on the numerical dissipation κ (Marshall et al, 2006). The DIMES group will rely on altimteric observations of surface velocities to advect tracers and estimate K_{Nak} at the ocean surface.

Releasing Lagrangian particles is a third way to compute eddy diffusivities. A major component of the DIMES experiment is to release neutrally buoyant floats and to follow their trajectories in order to infer K. Taylor (1921) showed that if the eddy statistics are homogeneous and stationary, the mean square separation of a particle from its starting position is a measure of K. The eddy diffusivity can be expressed in terms of the integral of the autocorrelation of the eddy velocity (the difference between the float velocity and the average regional velocity). For example, the diffusivity in the cross-stream direction can be written as,

$$K_{Taylor} = \frac{1}{2}\frac{\mathrm{d}}{\mathrm{d}t}\left\langle (y(t) - y_0)^2 \right\rangle = \int_0^t \left\langle v(t) v(t') \right\rangle \mathrm{d}t', \tag{3.6}$$

where y and v refer to the eddy displacement and velocity. Once the integral converges (typically ten days in the Southern Ocean, Sallée et al., 2008) the dispersion is statistically equivalent to a diffusive process and K_{Taylor} settles to a constant value. Floats are believed to provide an accurate characterization of isopycnal dispersion and they are often used in oceanographic studies. However, K_{Taylor} is not directly related to the mixing of tracer by eddies and it is not easily interpretable in terms of the K employed in the tracer-gradient relationship in Eq. (2.3).

Two questions must be addresses in order to interpret the various estimates of K.

1) Is the diffusivity estimated from tracers and floats the same as the PV diffusivity necessary to infer the meridonal transport as per Eq. (2.2)? Treguier (1999) and Smith and Marshall (2008) show

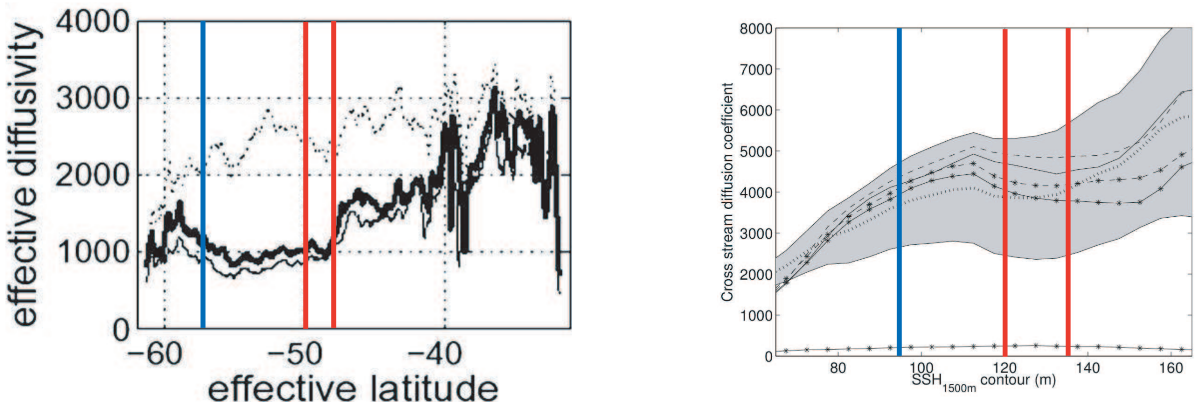

Figure 6: (Left) Two estimates of K_{Nak} in the Southern Ocean for two different values of molecular diffusivity κ (Marshall et al., 2006). The values are averaged along geostrophic contours and the equivalent latitude is the mean latitude of each contour. The blue and red vertical lines represent the Polar Front (PF) and the two Subantarctic Fronts (SAF) bounding the ACC. The dashed line is from a calculation in which the tracer is advected only by eddies with the mean flow set to zero: the suppression of K_{Nak} in the jet core is indeed due to the strong mean flow. The values of K_{Nak} peak on the equatorward flank of the ACC north of the SAF. (Right) Along-stream average of K_{Taylor} for different drifter datasets as a function of sea surface height. Diffusivities are clearly enhanced on the equatorward flank of the ACC fronts. The decrease in K_{Taylor} south of the PF reflects the sudden decrease in eddy activity around Antarctica.

compelling evidence that PV and tracer diffusivities are equivalent. But the analysis is based on idealized flows. We plan to extend their work to observations and simulations of the ACC.
2) Are the three estimates K_{Tracer}, K_{Nak}, and K_{Taylor} equivalent? Diffusivities estimated from drifters and floats are typically much larger than $1000\ \mathrm{m}^2/\mathrm{s}$ (Sallée et al., 2008), the value employed in large-scale ocean models in order to reproduce the observed tracer distributions. In preparation for this proposal, we compared two existing estimates of K_{Taylor} and K_{Nak} at the ocean surface in the ACC latitude band. Sallée et al. (2008) computed K_{Taylor} from satellite-tracked surface drifter data from the Global Drifter Program (Lumpkin and Pazos, 2007). Marshall et al. (2006) calculated K_{Nak} by numerically advecting idealized tracers with the surface geostrophic flow observed by satellite altimetry (Fig. (6). Both approaches show high diffusivities on the equatorward flank of the jet and smaller values in the jet core. However the drifter based estimate return diffusivities four times larger than the tracer based calculation (we checked that the difference is not due to Ekman advection acting on the floats). Large values of K_{Taylor} from drifters and floats have been reported by many authors. (However Fig. 6 represents a worst case scenario because isobaric drifters are less accurate at tracking parcel parcels than isopycnal floats). In this proposal we plan to investigate what estimate of K is most appropriate for estimating PV and tracer fluxes driving the Southern Ocean MOC.

4 Proposed work

4.1 Deployment of floats

The DIMES group proposed to deploy a total of 150 acoustically-tracked isopycnal-following floats. All floats are to be ballasted for the same neutral density surface as the tracer, $\gamma = 27.9$

between the SubAntarctic Fronts (SAF) and the Polar Front (PF). The $\gamma = 27.9$ surface lies in the lower part of the UCDW, i.e., in the lower part of the upper cell of the MOC (Figs. 1 and 3). The floats will be deployed along meridional lines spanning the ACC. Half will be released initially with the tracer and the other half one year after the tracer release as ao to build statistics for the estimate of K_{Taylor}.

The DIMES group had originally proposed to deploy floats at two different levels, but the plan has been successively revised due to high costs of the overall experiment. [The float component was just one of the components that had to be reduced.] Alternatively the group proposed to use the terrigenic Helium-3 distribution in the Scotia Sea to extend estimates of K_{Tracer} to other levels in the upper and lower circumpolar deep water (He-3 is centered at $\gamma = 27.98$, about 400 meters deeper than the float release, but is spread over a depth interval of more than 1000 meters). Altimetric data would be used to estimate K_{Nak} near the surface for a third level. The problem with this approach is that there are large offsets between K estimated from different techniques as we showed above – see Fig. 6. Variations in K can be reliably estimated only by computing K with the same technique at different vertical and horizontal locations.

In order to quantify the vertical variability of K, the best way forward is to augment the present float component of DIMES. Releasing a second tracer at a different level is too costly. The present plan is to deploy all floats on the same isopycnal surface as the tracer, $\gamma = 27.9$, at depths between 1500 and 800m. This is the depth range where K has a subsurface maximum according to Fig. 4. Preliminary work suggests the pattern in Fig. 4 is representative of the Pacific sector between 110 W and 80 W. We propose to deploy 50 additional floats at a shallower level to test the critical layer hypothesis. We expect to find the shallower floats disperse less rapidly than the deeper ones at the PF, while the reverse is true when the floats cross north of the SAF.

We propose to deploy 50 isopycnal-following floats on the isopycnal surface $\gamma = 27.4$ at depths between 200 and 800m (see Fig. 3), i.e. at the SAMW/AAIW level. It is too late to order additional floats for the first deployment cruise in 2009, but there is adequate time for fabrication, preparation, and shipping for deployment on the second deployment cruise in austral summer 2009/2010 (DIMES cruise US2, Fig. 5). The DIMES cruise US2 will span a fairly large sector of the Pacific (Fig. 7), so we have flexibility in choosing the best depolyment strategy in coordination with the DIMES PIs. 30 of the floats will be purchased through this proposal, while the DIMES group has agreed to dedicate 20 of their floats to the shallower level. The RAFOS floats will be of the same type used in the DIMES experiment whose performance has been well-documented (e.g. Rossby et al., 1994; Barth et al., 2004). They will be ballasted in the WHOI facilities and they are expected to follow their designated isopycnal surface with an error of $\pm 0.05\gamma$. The sound sources used to track the floats will be deployed during the first DIMES cruise (blue dots in Fig. 5).

The float trajectories will be used to compute K_{Taylor} along isopycnals $\gamma = 27.4$ and $\gamma = 27.9$. If the critical layer paradigm is correct, when the floats are between the PF and the SAF, K_{Taylor} should be 4-5 times larger at the critical layer depth than in the core of the jet (Fig. 4 smooths the dramatic change in K_{Nak} because it is an along-stream average). How many floats are required to sample reliably these differences? The standard deviation for K_{Taylor} decreases as $n^{-1/2}$, if n is the number of floats (Davis, 1987). It also increases as $t^{1/2}$, which effectively limits the length of the calculation. These error estimates apply only once the velocity field becomes

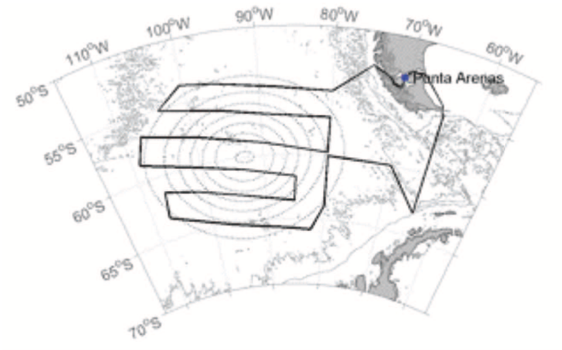

Figure 7: Track (heavy solid line) for the cruise during which we propose to release 50 floats (DIMES cruise US2, see Fig. 5. The ellipses show the expected distribution of the tracer released during DIMES cruise US1 based on 0.1 degree POP model runs at LANL.

decorrelated, typically 10-15 days (Sallée et al., 2008). For the DIMES proposal, POP model floats (from the model's DIMES region) were used to estimate the number of floats required for statistical convergence in the calculation of K_{Taylor}. Initially K_{Taylor} increases rapidly, reflecting vigorous advection by eddies, but settles down after several months to values near 1000-2000 m^2/s. With 150 floats, the 95% confidence limits are approximately ±200 m^2/s, or 20% of the diffusivity. With 50 floats, the errors increase to about 350 m^2/s, in line with the aforementioned $n^{-1/2}$ dependence.

The POP model flow is less energetic than the actual Southern Ocean flow, so it is likely the actual errors will be somewhat larger. However the numbers should not change much, because previous diffusivity calculations from the North Atlantic found good convergence with 30-70 floats (LaCasce and Bower, 2000). We expect diffusivities to increase by a factor of 4-5 at critical levels, a signal much larger than the expected 30% uncertainty for K_{Taylor} from 50 floats.

4.2 Estimate of effective diffusivities from the Southern Ocean State Estimate

SOSE is the first eddy admitting (horizontal resolution 1/6 degree) Southern Ocean State Estimate. The MIT general circulation model is least squares fit to all available ocean observations (satellite and in situ observations, meteorological surface fluxes, and the WOCE Global Hydrographic Climatology). This is accomplished iteratively through an adjoint method. The result is a physically realistic estimate of the ocean state. The estimation period for SOSE is 2005-2006. SOSE is being produced by Matthew Mazloff under the primary guidance of Carl Wunsch and Patrick Heimbach as part of the ECCO-GODAE and ECCO 2 projects. The model domain is from 78 S to 24 S. First guess initial conditions and open boundary conditions are derived from a coarse resolution global state estimate produced as part of the ECCO-GODAE. SOSE uses an atmospheric boundary layer scheme and is coupled to a sea-ice model. The first guess atmospheric state is from NCEP reanalysis. The adjoint method systematically perturbs the atmospheric state and initial conditions, within their uncertainty, to find a model ocean state consistent with the observations.

SOSE has skill in reproducing the mean state of the Southern Ocean, as compared with altimetric and climatological observations, and the eddy variability, as compared with altimetric and ARGO observations. We plan to use SOSE as a working platform to test our ideas about eddy stirring in the Southern Ocean. First we propose to estimate eddy diffusivities with the three approaches described in Sec. 3.1. We will release tracers and particles in the SOSE model and es-

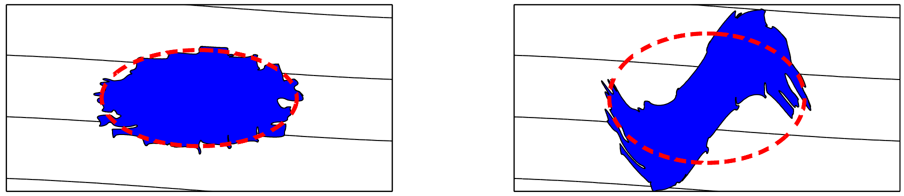

Figure 8: Dispersion of a tracer patch stirred by small-scale eddies. The black contours indicate the direction of the large-scale mean flow. The dashed dotted line is the best fit ellipse to the tracer patch. (a) The tracer is only stirred by small-scale eddies. (b) In addition to small-scale eddies, the tracers has also been deformed by a larger-scale meander of the mean jet.

timate K_{Nak}, K_{Tracer}, and K_{Taylor}. We will study dispersion both in individual 10 degree sectors of the ACC and along the whole ACC. We will study in detail results from the Pacific ACC where we plan to release the floats. This work will provide a framework to put the DIMES observations in the context of the overall ACC dynamics.

4.3 Comparison of diffusivities estimated from tracers and floats

In order to interpret the results of the float deployment, we must understand why estimates of K_{Nak}, K_{Tracer}, and K_{Taylor} can be quite different. The first goal is to determine which approach gives the K that relates the isentropic eddy PV flux to the mean PV gradient in Eq. (2.3). The second goal is to show that the K_{Taylor} obtained from float trajectories can be used to infer the K in Eq. (2.3) and hence to estimate the Southern Ocean MOC.

Let us review the basic assumptions behind the three approaches to estimate K. Nakamura's approach is to diagnose K_{Nak} by identifying the enhancement of diffusion that arises through the effects of eddies stretching and folding tracer contours. In mixing regions tracers are vigorously stretched into complex geometrical shapes with tight gradients, and this leads to large values of K_{Nak}. Tracer geometry in barrier regions, like jets, is usually smooth, creating localized small values of K_{Nak} – so small as to keep the flux minimal despite the often large tracer gradients. Because K_{Nak} is diagnosed directly from tracer fields being dispersed by eddies, it is obviously connected to the diffusivities employed in large-scale ocean models whose purpose is to represent the enhancement of diffusion by unresolved eddies.

The tracer and float methods are based on the spreading rate of the area occupied by tracer patches. In contrast to Nakamura's approach, where complexity of the tracer distribution is key, these methods are only concerned with the area enclosed by the tracer. The growth of the area is estimated by fitting an ellipse around the center of mass of the tracer or float distribution. Garrett (1983) called this ellipse the 'particle domain', meaning the ensemble-average for the area occupied by particles if released within the tracer patch. The expectation is that at long times the growth of the 'particle domain' is linear in time and represents a balance between eddy stirring and diffusion. These diffusivities are not directly related to the mixing properties of eddies and are not easily interpretable in terms of the eddy diffusivities employed in large-scale ocean models.

In our experience the three approaches return identical values if the eddy flow is characterized by a single lengthscale, but the estimates can differ substantially when there is eddy variability on many lengthscales. In the two panels of Fig. 8, a tracer patch is stirred by a small-scale eddies into convoluted filaments that are eventually homogenized by diffusion. In the left panel, the mean flow is uniform and steady and the the growth of the area enclosed by the tracer patch is well captured by the growth of the best fit ellipse. In the right panel, the mean flow has undergone a large-scale meander. The meandering has deformed the tracer patch at a scale much larger than the eddy scale without enhancing the small-scale filaments on which diffusion acts. [This is not quite true in the figure, because the separation between the filament width and the scale of the meander is not as large as in the real ocean.] Nakamura's approach returns the same estimate for the tracer distributions in Fig 8a and b, because the length of the tracer contour is dominated by the small-scale filaments and it is insensitive to the large scale meander. The 'particle domain' approach, instead, returns a higher diffusivity for Fig 8b, because the meridional axis of the best fit ellipse has grown faster. But the large-scale meander does not lead to small scale filamentation and diffusion and should not be included in the estimate of effective diffusivity. In simple test cases we found K_{Taylor} and K_{Tracer} tend to overestimate the rate of small-scale tracer mixing, as perhaps suggested in Fig. 6 [The difference between K_{Taylor} and K_{Nak} for isopycnal floats is expected to be smaller than in Fig. 6 which is based on drifters. Drifters flow at constant depth and do not track very accurately water parcels.]

We propose to compare estimates of K_{Taylor}, K_{Tracer}, and K_{Nak} both from the SOSE simulations and from more idealized flows. The goal is to test our hypothesis that jet meanders on scales larger than the characteristic size of geostrophic eddies account for the differences in the K values reported in the literature and illustrated in Fig. 6. We will also study the relationship among the various estimates of K, with a focus in the Pacific sector where we plan to deploy floats, so as to provide a framework to interpret the observations.

4.4 Estimate of effective diffusivities from an adjoint calculation

Adjoint techniques can be used to estimate K from a coarse model (not eddy admitting) constrained to available observations. The zonal momentum budget in Eq. (2.2) shows the transport in an isopycnal layer is forced by the isentropic eddy PV flux. Ferreira et al. (2005) estimated, using adjoint techniques, the eddy PV forcing that minimizes the departure of a coarse-resolution model from climatological observations of temperature. We propose to extend this approach to test the critical layer hypothesis. We will express K as a function that peaks at critical layers (e.g., Green, 1970; Killworth, 1997) such as,

$$K = K_0 \,/ \big(1 + \alpha(1 - |\boldsymbol{u}|/|c_r|)^2\big) \ , \tag{4.7}$$

where $\boldsymbol{u}$ is the velocity field resolved by the coarse-resolution model and c_r is the first baroclinic mode phase speed, a good proxy for the propagation speed of eddies in the Southern Ocean. This K will appear in the momentum equation multiplied by the PV gradient. Adjoint techniques will be used to compute the parameters K_0 and α that minimize the departure of the coarse-resolution model from all available observations. The spatial variations of the two parameters will provide information on the distribution of critical layers in the ocean.

4.5 Educational component

We propose to support one student (Ryan Abernathey) and one postdoc (John Taylor) through this proposal, because we believe the project would provide tremendous training ground for young scientists. The student and postdoc will benefit through collaborations with the PIs and with our oceanography and meteorology colleagues who have broad expertise in the theoretical issues explored in this project: eddy-driven circulations, PV mixing, and critical layers.

In addition we propose to modify the syllabi of two classes to introduce students to theories of the Southern Ocean. The core curriculum in physical oceanography at MIT still hinges on the classical theories of midlatitude gyre circulation, despite the recent progress in understanding the dynamics of the Southern Ocean. Marshall will present the basic theory of the Southern Ocean circulation in the class on *Steady Circulation of the Ocean*. Ferrari will devote a few weeks to discuss of the effect of eddy motions on the overturning circulations in the ocean and atmospheres in the class on *Geophysical Turbulence in the Ocean and Atmosphere*. The development of such classes is part of our academic work. As part of this proposal, we will post lecture notes and class material in the MIT OpenCourseWare (OCW) website (http://ocw.mit.edu). MIT OCW averages 1 million visits each month and provides a wonderful opportunity to disseminate education around the world.

Work Plan

Year 1: The graduate student will compute K_{Nak} for individual ACC sectors using the SOSE model. The postdoc will develop a Lagrangian code to advect particles with the MITgcm and will estimate K_{Taylor} for the same sectors.
Year 2: 50 floats will be deployed during DIMES cruise US2. The graduate student will use K_{Nak} to compute the overturning circulation of the Southern Ocean. The postdoc will study the differences between K_{Taylor}, K_{Tracer}, and K_{Nak}.
Year 3: The PIs will compare mean flows and eddy statistics from SOSE with the first results from DIMES. The graduate student will compare K_{Nak} estimates from passive tracers and PV distributions. The postdoc will estimate the Souther Ocean PV diffusivity with an adjoint model.

5 Results of Prior NSF Support

Raffaele Ferrari (MIT): Collaborative Research: Interaction of eddies with mixed layers. Ferrari is the leading PI of this collaborative research (10 co-PIs) aimed at developing and testing parameterizations of the interactions of mesoscale eddies with the ocean boundaries. The team has published 24 papers in the last four years. Ferrari is co-author in 10 of them. A detailed list of publications is available at www.mit.edu/~ raffaele/cpt.
John Marshall(MIT): CLIMODE project. Marshall is currently supported in the CLIMODE project which is studying the physics of mode water formation in the subtropical gyre of the North Atlantic. Marshall is carrying out associated theoretical and modeling work addressing the role of upper ocean eddies in water mass transformation. Two papers on this work are in press, and one is published (Marshall, 2005).
Kevin Speer (FSU): Diapycnal and Isopycnal Mixing Experiment in the Southern Ocean. Speer is one of the co-PIs of the recently funded DIMES experiment. DIMES is a US/UK field program aimed at measuring diapycnal and isopycnal mixing in the Southern Ocean (Gille et al., 2007).

References

Abernathey, R., J. Marshall, E. Shuckburgh, and M. Mazloff: 2008, Studies of enhanced isopycnal mixing at deep critical levels in the southern ocean. *J. Phys. Oceanogr.*, in preparation.

Andrews, D. G., J. R. Holton, and C. B. Leovy: 1987, *Middle Atmosphere Dynamics*. Academic Press, San Diego, 489 pp.

Barth, J. A., D. Hebert, A. C. Dale, and D. S. Ullman: 2004, Direct observations of along-isopycnal upwelling and diapycnal velocity at a shelfbreak front. *J. Phys. Oceanogr.*, **34**, 543–565.

Bryden, H. L.: 1979, Poleward heat flux and conversion of available potential energy in Drake Passage. *J. Mar. Res.*, **37**, 1–22.

Cerovecki, I., R. A. Plumb, and W. Heres: 2008, Eddy transport and mixing on a wind and buoyancy driven jet on the sphere. *J. Phys. Oceanogr.*, in press.

Cessi, P. and M. Fantini: 2004, The eddy-driven thermocline. *J. Phys. Oceanogr.*, **34**, 2642–2658.

Danabasoglu, G. and J. Marshall: 2007, Effects of vertical variations of thickness diffusivity in an ocean general circulation model. *Ocean Modelling*, **18**, 122–141.

Davis, R.: 1987, Modeling eddy transport of passive tracers. *J. Mar. Res.*, **45**, 635–666.

Deacon, G.: 1984, *The Antarctic Circumpolar Ocean*. Cambridge University Press, New York, 180 pp.

deSzoeke, R. A. and M. D. Levine: 1981, The advective flux of heat by mean geostrophic motions in the Southern Ocean. *Deep-Sea Res.*, **28**, 1057–1085.

Eden, C.: 2007, Eddy length scales in the north atlantic ocean. *J. of Geophys. Res. - Oceans*, **112**, C06004.

Ferreira, D., J. Marshall, and P. Heimbach: 2005, Estimating eddy stresses by fitting dynamics to observations using a residual-mean ocean circulation model and its adjoint. *J. Phys. Oceanogr.*, **35**, 1891–1910.

Garrett, C.: 1983, On the initial streakiness of a dispersing tracer in two-dimensional and 3-dimensional turbulence. *Dyn. Atmos. Oceans*, **7**, 265–277.

Gille, S. T., K. Speer, J. R. Ledwell, and A. C. N. Garabato: 2007, Mixing and stirring in the southern ocean. *Clivar Variations*, **88(39)**, 382–383.

Gnanadesikan, A., S. M. Griffies, and B. L. Samuels: 2007, Effects in a climate model of slope tapering in neutral physics schemes. *Ocean Modelling*, **16**, 1–16.

Gnanadesikan, A. J., P. Dunne, R. M. Key, K. Matsumoto, J. L. Sarmiento, R. D. Slater, and P. S. Swathi: 2004, Oceanic ventilation and biogeochemical cycling: Understanding the physical mechanisms that produce realistic distributions of tracers and productivity. *Global Biogeochem. Cycles*, **18**, GB4010.

Green, J. S. A.: 1970, Transfer properties of the large-scale eddies and the general circulation of the atmosphere. *Quart. J. Royal Meteor. Soc.*, **96**, 157–185.

Gregory, J. M.: 2000, Vertical heat transports in the ocean and their effect on time-dependent climate change. *Clim. Dyn.*, **16**, 501–515.

Henning, C. C. and G. K. Vallis: 2004, The effects of mesoscale eddies on the main subtropical thermocline. *J. Phys. Oceanogr.*, **34**, 2428–2443.

Karsten, R. and J. Marshall: 2002, Constructing the residual circulation of the antarctic circulation current from observations. *J. Phys. Oceanogr.*, **32**, 3315–3327.

Killworth, P.: 1997, On the parameterization of eddy transfer. Part I: Theory. *J. Mar. Res.*, **55**, 1171–1197.

LaCasce, J. H. and A. Bower: 2000, Relative dispersion in the subsurface north atlantic. *J. Mar. Res.*, **58**, 863–894.

Ledwell, J. R., A. J. Watson, and C. S. Law: 1998, Mixing of a tracer in the pycnocline. *J. of Geophys. Res. - Oceans*, **103**, 21499–21529.

Lumpkin, R. and M. Pazos: 2007, Measuring surface currents with surface velocity program drifters: the instruments, its data and some recent results. *Lagrangian analysis and prediction of coastal and ocean dynamics*, Cambridge University Press, New York.

Marshall, J.: 2005, Climode: a mode water dynamics experiment in support of clivar. *Clivar Variations*, **3(2)**, 8–14.

Marshall, J. and T. Radko: 2004, Residual mean solutions for the Antarctic Circumpolar Current and its associated thermohaline circulation. *J. Phys. Oceanogr.*, **33**, 2341–2354.

— 2006, A model of the upper branch of the meridional overturning of the southern ocean. *Progress In Oceanography*, **70**, 331–345.

Marshall, J. C., E. Shuckburgh, H. Jones, and C. Hill: 2006, Estimates and implications of surface eddy diffusivity in the southern ocean derived from tracer transport. *J. Phys. Oceanogr.*, **36**, 1806–1821.

Nakamura, N.: 1996, Two-dimensional mixing, edge formation, and permeability diagnosed in an area coordinate. *J. Atmos. Sci.*, **53**, 1524–1537.

Olbers, D., D. Borowski, C. Wolker, and J. Wölff: 2004, The dynamical balance, transport and circulation of the antarctic circumpolar current. *Antarctic Science*, **16**, 439–470.

Olbers, D. and M. Visbeck: 2005, A model of the zonally-averaged stratification and overturning in the southern ocean. *J. Phys. Oceanogr.*, **35**, 1190–1205.

Orsi, A. H., T. Whitworth, and W. D. Nowlin: 1995, On the meridional extent and fronts of the antarctic circumpolar current. *Deep-Sea Res. - Part I*, **42**, 641–673.

Plumb, R. A. and R. Ferrari: 2005, Transformed eulerian-mean theory. i: Non-quasigeostrophic theory for eddies on a zonal mean flow. *J. Phys. Oceanogr.*, **35**, 165–174.

Polzin, K. and R. Ferrari: 2004, Isopycnal dispersion in natre. *J. Phys. Oceanogr.*, **34**, 247–257.

Rhines, P. B. and W. R. Young: 1982, Homogenization of potential vorticity in planetary gyres. *J. Fluid Mech.*, **122**, 347–367.

Rintoul, S. R., C. Hughes, and D. Olbers: 2001a, The antarctic circumpolar current system. *Ocean Circulation and Climate*, G. Siedler, J. Church, and J. Gould, eds., Academic Press, New York, 271–302.

Rintoul, S. R., C. W. Hughes, and D. J. Olbers: 2001b, The antarctic circumpolar current system. *Ocean Circulation and Climate: Observing and Modelling the Global Ocean*, G. Siedler, J. Church, and J. Gould, eds., Academic Press, London, 171–302.

Rossby, T., J. Fontaine, and E. C. Carter: 1994, The f/h float - measuring stretching vorticity directly. *Deep-Sea Res. - Part I*, **41**, 975–992.

Sallée, J., K. Speer, R. Morrow, and R. Lumpkin: 2008, An estimate of lagrangian eddy statistics and diffusion in the mixed layer of the southern ocean. *J. Phys. Oceanogr.*, **34**, submitted.

Sarmiento, J. L., N. Gruber., M. Brzezinski, and J. P. Dunne: 2004, High-latitude controls of thermocline nutrients and low latitude biological productivity. *Nature*, **427**, 56–60.

Schneider, T.: 2005, Zonal momentum balance, potential vorticity dynamics, and mass fluxes on near-surface isentropes. *J. Atmos. Sci.*, **62**, 1884–1900.

Sloyan, B. M. and S. R. Rintoul: 2000, Estimates of area-averaged diapycnal fluxes from basin-scale budgets. *J. Phys. Oceanogr.*, **30**, 2320–2341.

Smith, K. S. and J. Marshall: 2008, Evidence for deep eddy mixing in the southern ocean. *J. Phys. Oceanogr.*, in press.

Speer, K., S. R. Rintoul, and B. Sloyan: 2000, The diabatic deacon cell. *J. Phys. Oceanogr.*, **30**, 3212–3222.

Taylor, G. I.: 1921, Diffusion by continuous movements. *Proc. London Math. Soc.*, **20**, 196–211.

Toggweiler, J. R. and B. Samuels: 28, On the ocean's large-scale circulation near the limit of no vertical mixing. *J. Phys. Oceanogr.*, **1998**, 1832–1852.

Treguier, A. M.: 1999, Evaluating eddy mixing coefficients from eddy-resolving ocean models: A case study. *J. Mar. Res.*, **57**, 89–108.

Wunsch, C. and R. Ferrari: 2004, Vertical mixing, energy, and the general circulation of the oceans. *Ann. Rev. Fluid Mech.*, **36**, 281–314.