Closing the meridional overturning circulation through Southern Ocean upwelling

(Review)

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November 2010

Abstract

The Meridional Overturning Circulation (MOC) of the ocean plays a central role in climate and climate variability by storing and transporting heat, freshwater and carbon around the globe. Historically, the focus of research has been on the MOC in the North Atlantic basin, because the Atlantic is a primary site of deep convection triggered by buoyancy loss to the atmosphere, inducing large volumes of surface water to sink to depth. It is becoming increasingly clear, however, that return paths to the surface through Southern Ocean upwelling, driven by westerly winds encircling Antarctica, are a key part of the MOC puzzle. It is now thought that Southern Ocean upwelling has an importance that rivals the Atlantic downwelling branch for our understanding of climate and climate variability, because it controls the rate at which the ocean reservoirs of heat and carbon communicate with the surface. Here we review what is known about the upwelling branch of the MOC, our dynamical understanding of it and its possible role in paleoclimate, the present climate and climate change.

1 Introduction

'It's easy to see how convection gets water down in to the interior of the ocean.
The challenge is figuring out where and how that water gets back to the surface.'

Quote from the late Melvin Stern on the porch of Walsh Cottage, Woods Hole, during the 1996 Geophysical Fluid Dynamics Summer Program on Open Ocean Deep Convection.

An early conceptual model of the MOC of the ocean is that of the 'filling box' (Baines & Turner, 1969, Hughes & Griffiths, 2006, Wahlin & Cenedese, 2006). The 'box' represents the ocean basin. The 'filling' process is conversion in the polar oceans of light upper layer water to more dense deeper water by convection and mixing in the open seas and in shelf and bottom boundary layer processes. These dense waters rise within the basins and ultimately, after a circuitous route which we discuss here, flow back toward the sinking regions to close the circulation meridionally.

In today's climate, the conversion to dense water takes place primarily in the northern North Atlantic and around Antarctica in the Southern Ocean (see reviews by Killworth, 1983; Marshall and Schott, 1999). As sketched schematically in Fig.1, two meridional overturning cells emanate from these formation regions: an Upper Cell associated with sinking to mid-depth in the northern North Atlantic, and a Lower Cell associated with sources of abyssal water around Antarctica. Although the polar source regions for these overturning cells have been identified (see the recent review of the North Atlantic downwelling branch of the MOC by Lozier, 2010), detailed upwelling pathways back to the surface and the underlying controlling physical mechanisms, have long been debated and are just now coming in to clearer focus.

The main theme of this review is the central role of the Southern Ocean in the upwelling branch of the MOC. It appears that a significant portion of the water made dense in sinking regions ultimately returns to the surface in nearly-adiabatic pathways along tilted isopycnals that rise from depth toward the surface around Antarctica. These tilted surfaces mark the great density difference between the subtropics and polar regions associated with the planet's largest current, the Antarctic Circumpolar Current (ACC). They connect the interior ocean to the sea surface enabling fluid which has sunk to depth at high latitudes to return to the surface without the need to invoke large, and unobserved, levels of interior vertical mixing in the ocean's thermocline. An appreciation of the central role of the Southern Ocean in closing the MOC largely solves the 'missing mixing' paradox which has remained a theme of oceanographic literature since the work of Munk (1966). In that paradox, all dense water was assumed to upwell through the thermocline to close the overturning circulation. To do so, strong vertical mixing is required within the interior of the ocean (Munk and Wunsch, 1998). However, strong mixing is only observed in the weakly stratified abyss in association with rough topography (Polzin et al, 1997; Toole et al, 1997; Ledwell et al., 2000; Kunze et al, 2006) and is generally small within the thermocline (Ledwell et al, 1998, 2010).

Broecker (1987, 1991) adumbrated a circulation schematic in which polar sinking is balanced by upwelling from depth through the thermocline of the North Pacific, as sketched in his famous conveyor belt analogy, inspired from earlier descriptions by Gordon (1986). However, there is growing and compelling evidence from observations, models and theoretical considerations, that a primary return path to the surface occurs in the Southern Ocean. Ekman upwelling around Antarctica driven by strong westerly winds has long been implicated as a primary driver of Southern Ocean overturning and meridional flow (Sverdrup, 1933). Recent modeling studies point to a global influence of this control (Doos and Webb, 1994; Toggweiler and Samuels, 1998; Webb and Suginohara, 2001; Ferreira et al, 2010). With such upwelling (further aided by boundary entrainment; Hughes and Griffiths, 2006) it is no longer necessary to assume enhanced levels of interior mixing to account for the return branches of the MOC, except in deep ocean basins topographically isolated from the ACC. Diverse mixing processes such as internal gravity wave breaking and hydraulic effects, operate close to rough and complex bottom topography of ocean basins where it can be difficult even to distinguish the interior from the boundaries.

In this review, within the larger theme of Southern Ocean circulation, we will focus on the observational basis for the upwelling branch of the MOC, our theoretical understanding of it, and the implications of these new insights for the role of the Southern Ocean in climate, climate variability and climate change. We conclude by presenting an updated conveyor schematic which we believe more faithfully captures ocean circulation.

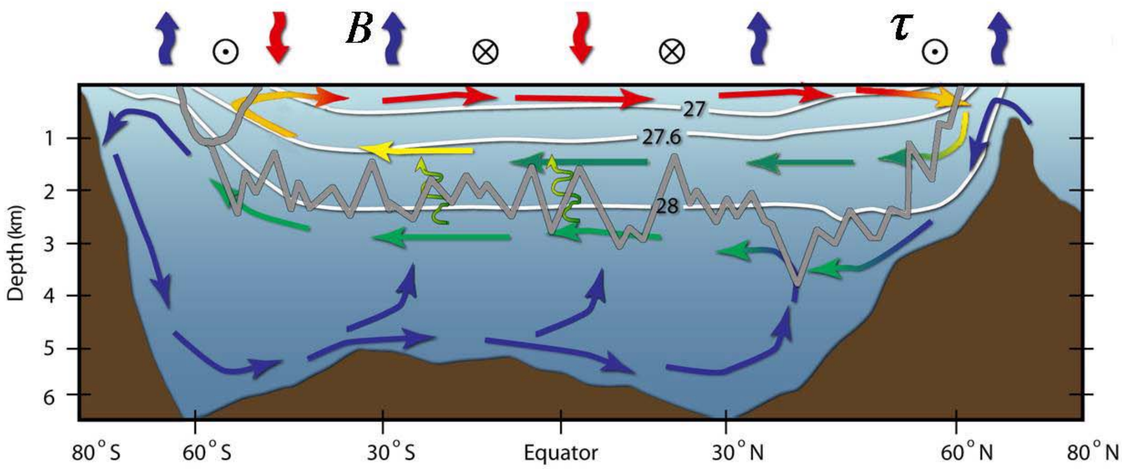

Figure 1: A schematic diagram of the Upper Cell and Lower Cell of the global meridional overturning circulation emanating from polar seas, together with the depth of key neutral density surfaces and ocean bathymetry. Grey lines indicate roughly the depth of the Mid-Atlantic Ridge and the Scotia Ridge (just downstream of Drake Passage) in the Southern Ocean (see Fig. 2). Low-latitude, wind-driven shallow cells are not indicated. The neutral density surface $\gamma^n = 27.6$ is the rough divide between the two cells. General patterns of surface buoyancy flux, B, (red downward) and loss (blue upward) are indicated, together with the broad pattern of zonal winds, τ, ($\odot$, westerlies; $\otimes$, easterlies). Colors schematically indicate the relative density of water masses: warmer mode and thermocline waters (red), upper deep waters (yellow), deep waters including North Atlantic Deep Water (green), and bottom waters (blue). Bottom water is supplied mainly from southern sources. The Upper Cell loses buoyancy in the northern hemisphere, which is largely balanced by air-sea buoyancy gain in the Southern Hemisphere. The Lower Cell loses buoyancy around Antarctica and regains it through abyssal mixing below topographic ridge crests. Mixing processes associated with topography are indicated by the vertical squiggly arrows. Westerly winds blowing around Antarctica, together with buoyancy fluxes, are instrumental in drawing water up toward the surface in the Southern Ocean. This schematic is a highly simplified representation of a three-dimensional flow illustrated more completely in Fig.6.

2 Inferences from observations

Key features of the circulation and hydrography of the Southern Ocean are shown in Fig.2. In Fig.2a the climatological positions of the Sub-Antarctic Front and Polar Front are marked, with the thickness of the lines representing the variance in the latitudinal positions of the fronts. They indicate regions of pronounced meridional temperature (and other property) gradients and strong zonal currents (green arrows) within the ACC. The fronts and associated currents show evidence of topographic steering, often being collocated with gaps in the topography. The Polar Front roughly marks the northern boundary of the sea-ice influence, but ice concentration is low approaching the ACC. South of the Polar Front, sea-ice grows and decays with the light summer and dark winter seasons, with large consequences for ocean physics and biology. The positions of the summer and winter ice edge, although not always well defined, are also marked in Fig.2a.

Fig.2b shows hydrographic sections of temperature (T) and salinity (S) crossing the ACC along World Ocean Circulation Experiment section SR3 running from Tasmania to Antarctica (see Rintoul and Sokolov, 2001). Property horizons rise toward the surface moving south along the sections, following surfaces of constant density (note the white contour on the sections marking the $\gamma^n = 27.6$ neutral density surface). The zonal current is in thermal wind balance with the tilted density surfaces: mean deep currents are small, building up over the water column to mean surface speeds of $0.3 \longrightarrow 0.4\,\mathrm{m\,s^{-1}}$ shown by the arrows in Fig.2a. Fig.2c shows an entire circumpolar T section, along S4, very roughly coincident with the position of the winter ice edge. We see a thick layer of warm fluid extending from a depth of $1 \longrightarrow 2\,\mathrm{km}$ up toward the surface. Water right at the surface around Antarctica is colder, but since it is fresh due to the melting of sea-ice in the summer, the water column remains stably stratified. The distribution of salinity shows high values at mid-depth due to the influence of salty North Atlantic Deep Water (NADW), with lower concentrations above and below. This pattern has long been suggestive of a two–cell meridional overturning circulation as sketched in Fig.1 with incoming deep, relatively fresh water, feeding an Upper and Lower Cell.

A number of attempts have been made to map the pattern of meridional overturning circulation in the Southern Ocean from hydrographic observations like those summarized in Fig.2. Among the first to observe and make use of salinity and other property distributions, Deacon (1937) sketched a southward deep flow and northward shallow flow across the ACC. Sverdrup (1933) had proposed a broad circumpolar cross-ACC flow or overturning, which was quantified in a conceptual model by Wyrtki (1960), linking the observed density field and wind field to infer southward ageostrophic flow and upwelling. Subsequent studies elaborated the large-scale circulation within the Southern Ocean and exchanges with other basins through descriptions of water mass distributions (Callahan, 1972; Jacobs, 2004; Jacobs and Georgi, 1977; McCartney, 1977; Whitworth et al, 1998) and major fronts (Nowlin and Klinck, 1986; Orsi et al., 1995; Pollard et al., 2002).

More recently, box inverse methods (Wunsch, 1996; Ganachaud and Wunsch, 2000) have been used to provide basin-scale estimates of mass and property transports by exploiting basic conservation principles applied over boxes, such as those delineated by the hydrographic sections shown in Fig.2a, and assumptions about the statistics of property distributions (e.g. Sloyan and Rintoul, 2001; Talley et al, 2003; Lumpkin and Speer, 2007). Zonal-

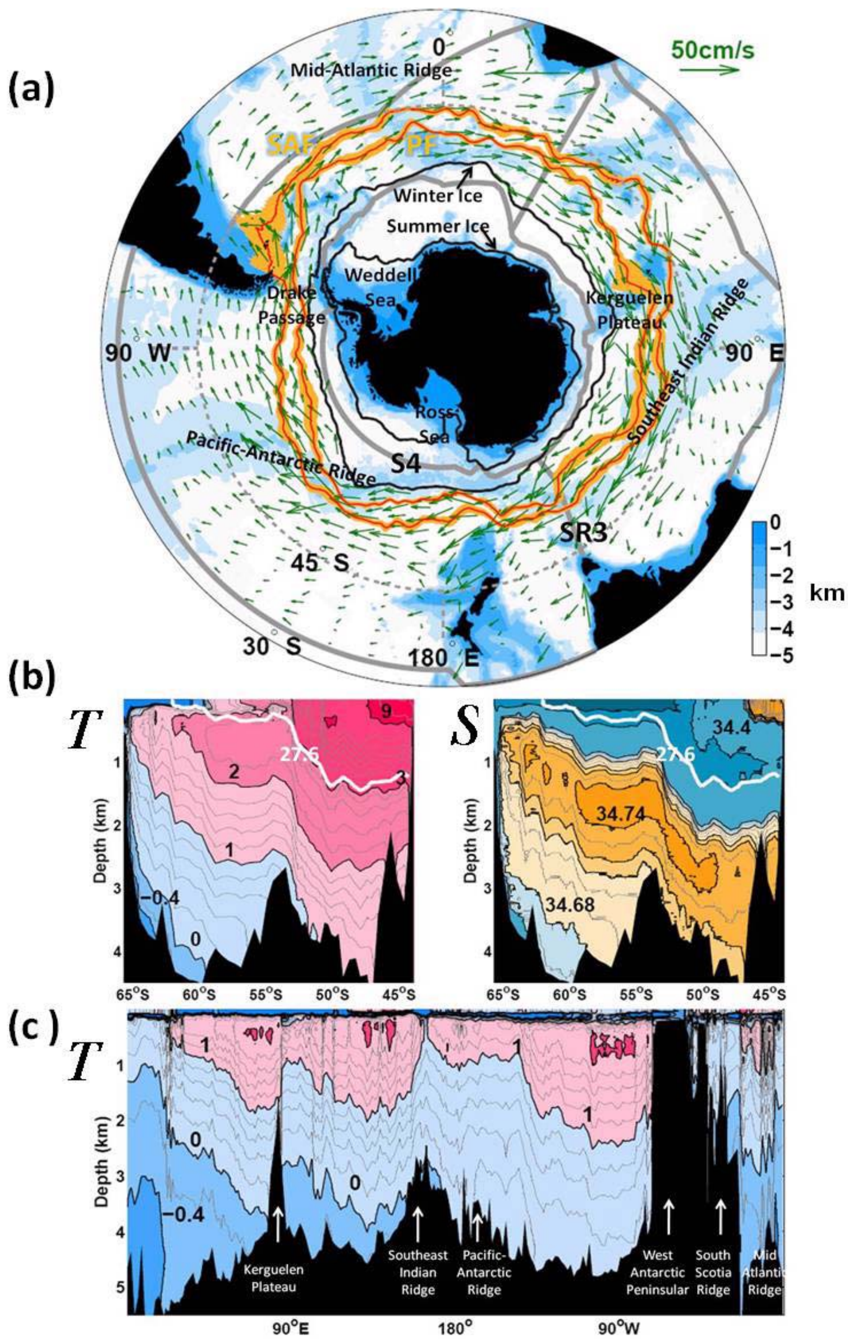

Figure 2: (a) Climatological position of the Subantarctic Front (SAF) and Polar Front (PF) are marked in orange with the thickness of the line representing the variance in the latitudinal position. The green arrows indicate the observed speed and direction of surface ocean currents as measured by drifters floating at a depth of 15 m (note the scale in the upper right). The depth of the ocean is color coded in blue: major topographic features are labelled. The black lines mark the summer (min) and winter (max) extent of sea ice. The position of key hydrographic sections are marked by the thick grey lines. (b) T and S sections along the SR3 section (145 °E) cutting across the ACC from Tasmania toward Antarctica. Black contours are labelled in °C (for T) and psu (for S). The thick white line is the 27.6 potential density surface. (c) T along the circumpolar transect, S4, whose position is marked on (a).

average estimates have also been deduced from air-sea fluxes (Speer et al., 2000) and from observations of surface winds and satellite altimetry, making use of residual-mean theory (Karsten and Marshall, 2002; Marshall et al. 2006). Tracer observations have been exploited in greater detail to determine the explicit dependence of overturning in the Southern Ocean on mixing coefficients (Zika et al., 2009), showing that low diapycnal mixing is consistent with plausible circulation patterns and strengths. Many details are uncertain since data is relatively sparse in the Southern Ocean, the circulation is not well known close to the continent within the ice pack (Jenkins and Jacobs, 2008), and the air-sea fluxes that force the ocean remain poorly constrained by observations and models. Nevertheless, certain robust features of the circulation emerge, as we now describe.

The inversion of Lumpkin and Speer (2007; hereafter LS07 — see Fig.3) shows that two global-scale counter-rotating meridional cells dominate the overturning circulation and represent distinct circulation regimes, as schematised in Fig.1. The Upper Cell is fed both from the northern Atlantic, where buoyancy loss triggers convection and sinking in the marginal seas (forming various components of NADW) and also from below by deep diapycnal upwelling. The convergence of flow at intermediate depths is roughly balanced by upwelling in the Southern Ocean, induced by the strong, persistent westerly winds that blow over the Southern Ocean (Fig.4a). Surface buoyancy fluxes associated with freshwater and heat gain (Warren et al, 1996 ; Toole, 1981, Speer et al 1997; Fig.4b), convert upwelling water to less dense Subantarctic Mode Water (SAMW). In contrast, the Lower Cell is fed by dense water formation processes around Antarctica, principally in the Ross and Weddell Seas, forming Antarctic Bottom Water (AABW); it is the result of a balance between buoyancy loss by air-sea fluxes (Fig.4b) and sea-ice export (Fig.4c) around Antarctica and buoyancy gain by abyssal mixing. Only the overall transports are represented in this view; the detailed mechanisms that control the fluxes of mass into and out of upwelling or sinking regions are not considered (see e.g. Lozier, 2010, for a review of the Atlantic sinking region physics).

The $\gamma^n = 27.6$ neutral density surface, outcropping south of the Polar Front all the way around Antarctica (see the white line in Fig.4a, b), marks the average division between the Upper and Lower Cells within the Southern Ocean, as indicated in the schematic in Fig.1 and the dotted line in Fig.3. Water crossing $\sim 30\,^\circ$S in the Atlantic roughly in the range $27 \longrightarrow 27.6$ enters the Southern Ocean and rises up to the surface where it is exposed to surface buoyancy gain (Fig.4b) and northward Ekman transport induced by the westerly winds: see the zonal wind-stress in Fig.4a and the consistently northward (Ekman) component of the surface currents driven by it in Fig.2a. Waters entering the Southern Ocean from the Atlantic at greater densities upwell and outcrop near the Antarctic continent, and are transformed into dense bottom water. Indeed LS07 estimate that about 75% of NADW enters the Lower Cell and is transformed to even denser bottom-water classes. Some becomes Antarctic Bottom Water (AABW). The remainder enters the Indian and Pacific as circumpolar deep water where abyssal mixing transforms it to lighter water ($\gamma^n \lesssim 27.6$). It then reenters the Upper Cell returning again southward and upward to the surface.

Is the return flow to the surface associated with interior mixing? Note that the abyssal ocean is stratified, albeit weakly so, and thus interior mixing must be acting vertically to diffuse properties that are initially set by high-latitude processes and carried in to the abyss by sinking fluid (see Munk and Wunsch, 1998). However, such mixing apparently occurs primarily near topographic ridges, the tops of which are marked in Figs.1 and 3. This

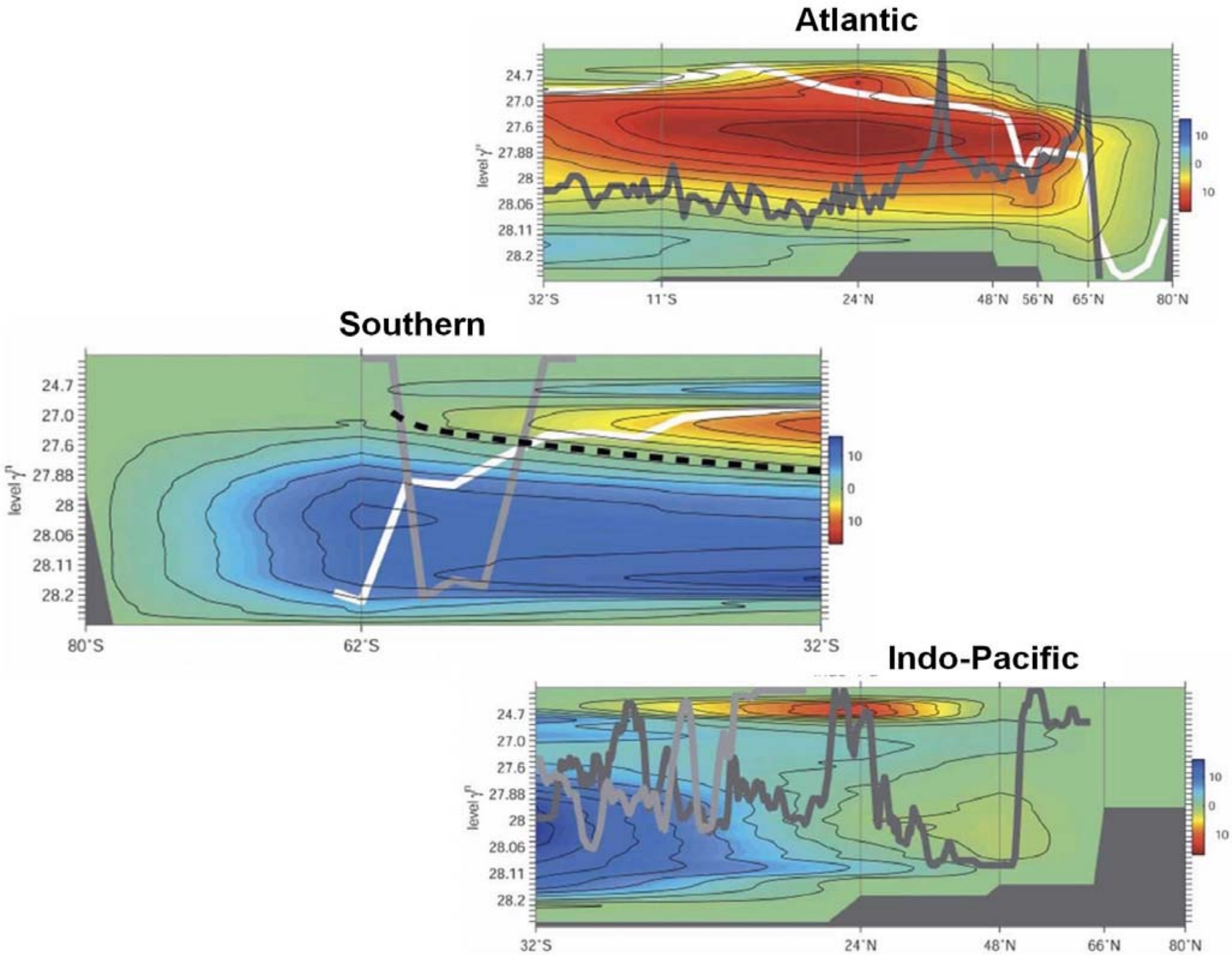

Figure 3: Meridional Overturning Circulation of the (top) Atlantic, (middle) Southern and (bottom) Indo-Pacific Oceans contoured every 2 Sverdrups ($1\mathrm{Sv} = 10^6\,\mathrm{m}^3\,\mathrm{s}^{-1}$), plotted as a function of latitude and density. The dashed line in the southern ocean divides the Upper and Lower cells and roughly corresponds to the $\gamma^n = 27.6$ surface, which outcrops in the latitude range of Drake Passage, marked on Fig.4. Gray lines indicate the crest of the Mid-Atlantic ridge in the Atlantic and major bathymetric features of the Pacific (dark) and Indian (light) Ocean basins and the Scotia Ridge in the Southern Ocean. The thick white line represents the interior density to which surface forcing effects are felt. Modified from Lumpkin and Speer (2007).

fundamentally diabatic part of the overturning circulation facilitates upward transfer of water in the Lower Cell to mid-depths within the World Ocean basins (see Ito and Marshall, 2006, for a conceptual model of the Lower Cell and the role of mixing in sustaining it). Once water reaches mid-depths, however, the outcropping density field of the Southern Ocean provides a quasi-adiabatic route directly to the surface and plays a central role in both the Upper and Lower cells of the MOC. Although some mixing is at work on the upwelling branch (note the slight drift of the southward flow toward lighter densities in Fig.3 above topography) the rates are small (see LS for a discussion). Importantly, upwelling from depth directly through the thermocline, due to elevated levels of diapycnal mixing, is not observed.

To complete the picture, note that isopycnal outcropping also occurs in high northern latitudes. What is to stop water from upwelling along density surfaces there too? In fact, the isopycnal pathway is not confined to the Southern Ocean and some intermediate depth water does indeed re-enter the sinking regions of the northern Atlantic via northward flow in the wind-driven circulation and entrainment into dense overflows. This flow is not easily identified since much water mass modification happens along its pathway. Estimates suggest that a substantial fraction of the overturning in the northern North Atlantic is the entrained component (Lumpkin and Speer, 2003), with the bulk coming from overflows and waters made dense by surface convection in the marginal seas. Thus, various pathways are important and the ocean is, by its nature, a complex three dimensional system. However, the central role of the Southern Ocean quasi-adiabatic pathway has emerged from the observational evidence of global overturning.

3 Dynamics of the Southern Ocean and its upwelling branch

The Southern Ocean is driven by surface fluxes of momentum and buoyancy induced by the strong, predominantly westerly winds that blow over it and the freezing phenomena close to the continent (see the review of Rintoul et al, 2001). Zonal wind-stress (Fig.4a) induces upwelling polewards of the zonal surface wind maximum and downwelling equatorwards of the maximum. This directly wind-driven circulation, known as the 'Deacon Cell', acts to 'overturn' isopycnal surfaces supporting the thermal wind current of the ACC and creating a store of available potential energy. This is the fundamental mechanical drive of the Southern Ocean.

Air-sea buoyancy fluxes generate dense water near the continent (Fofonoff, 1956; Gill, 1973) and lighten the surface layers within the ACC (see Fig.4b). The dense water sinks and tends to draw in warmer, saltier water from the surrounding ocean; however, rather than being fed from the surface the sinking is fed from deeper layers due to the outcropping density surfaces. The notion of a dividing isopycnal (27.6) between the two cells is somewhat rough, but adequately distinguishes these two buoyancy regimes. Furthermore, rising warmer deep water (clearly evident in Fig.2b, c) melts ice both on the shelf and in the open ocean, controlling to some degree the northern extent of the cryosphere. This perhaps also accounts for the marked coincidence between the 27.6 outcrop and the winter ice edge.

The overturning of buoyancy surfaces by winds is balanced, we believe, through the baroclinic instability of the thermal wind currents, an instability that acts to flatten out the buoyancy surfaces and to transport mass polewards. The resulting highly energetic eddies (analogous to atmospheric weather systems) have a scale of $\sim 100\,\mathrm{km}$ and are a ubiquitous feature of the circumpolar flow. From an energetic perspective, much of the potential energy imparted to the ocean by the mechanical tilting of isopycnals is extracted by eddies flattening them out (Gill, Green and Simmons, 1975; Marshall et al, 2002; Marshall and Radko, 2003). Indeed the Southern Ocean is a principal region where energy is imparted to the ocean by the wind (Wunsch, 1998) and it is here that we observe the most widespread mesoscale eddy field in the ocean, extending all the way along the path of the ACC and throughout the depth of the water column.

Fundamental to the role of eddies in the Upper Cell is their ability to transport mass, buoyancy and potential vorticity (related to angular momentum) across the axis of the ACC (McWilliams et al, 1973; Marshall, 1981; Johnson and Bryden, 1989; Gnanadesikan, 1999; Hallberg and Gnanadesikan, 2001; Karsten et al, 2002; Marshall and Radko, 2003, 2006; Bryden and Cunningham, 2003; Olbers and Visbeck, 2005). Indeed, the upper meridional cell in the Southern Ocean cannot be understood without invoking eddy transport: it is part of a 'residual' circulation in which lateral eddy fluxes largely balance the wind-driven circulation (Danabasoglu and McWilliams, 1995). This can be shown by decomposing the mass transport in a density layer of thickness h in to mean and eddy contributions to define what is known as the 'residual flow',

$$\underbrace{v_{res}\bar{h}}_{\substack{\text{Residual transport}\\ \psi_{res}}} = \overline{vh} = \underbrace{\bar{v}\bar{h}}_{\substack{\text{Mean Transport}\\ \overline{\psi}}} + \underbrace{\overline{v'h'}}_{\substack{\text{Eddy transport}\\ \psi^{*}}} . \tag{1}$$

Here the overbar denotes a time and streamwise average and primes departures from that average: ψ is a streamfunction (with units of $\mathrm{m}^3\,\mathrm{s}^{-1}$ i.e. in Sverdups) to represent the flow in the meridional plane. The residual flow is of central importance because it is this flow, rather than the Eulerian mean, that advects properties — tracers — in a turbulent ocean. The distinction between ψ_{res}, $\overline{\psi}$ and ψ^* is crucial to our understanding of the overturning circulation in the Southern Ocean (Marshall and Radko, 2003, 2006; Ito et al, 2004; Olbers and Visbeck, 2005).

The dynamical processes at work are shown in the numerical results presented in Fig.5 from a very high resolution 'process model' of a portion of the ACC. The wind induces a pattern of upwelling and downwelling, an analogue of the Deacon Cell represented by $\overline{\psi}$ in the diagram, tilting up buoyancy surfaces. This is largely balanced by the circulation associated with eddies, ψ^*, acting to return them to the horizontal. Note that both $\overline{\psi}$ and ψ^* *cross* mean buoyancy surfaces. However, the upwelling branch of residual-mean flow, ψ_{res}, is directed *along* mean buoyancy surfaces and is therefore not associated with elevated levels of diapycnal mixing. The upwelling feeds an Upper Cell and, in this open channel, a Lower Cell, similar to the data inversion shown in Fig.3 and schematized in Fig.1. These cells are associated with air-sea buoyancy fluxes (indicated at the top of Fig.5) with a cooling, warming, cooling pattern evident in the data (reminiscent of Fig.4b, c).

The dynamical balances in the model and in the Southern Ocean can be described in terms of a streamwise force balance for a density layer of mean thickness $\overline{h}$ (see, e.g. Marshall, 1997; Greatbatch, 1998; Olbers et al, 2004):

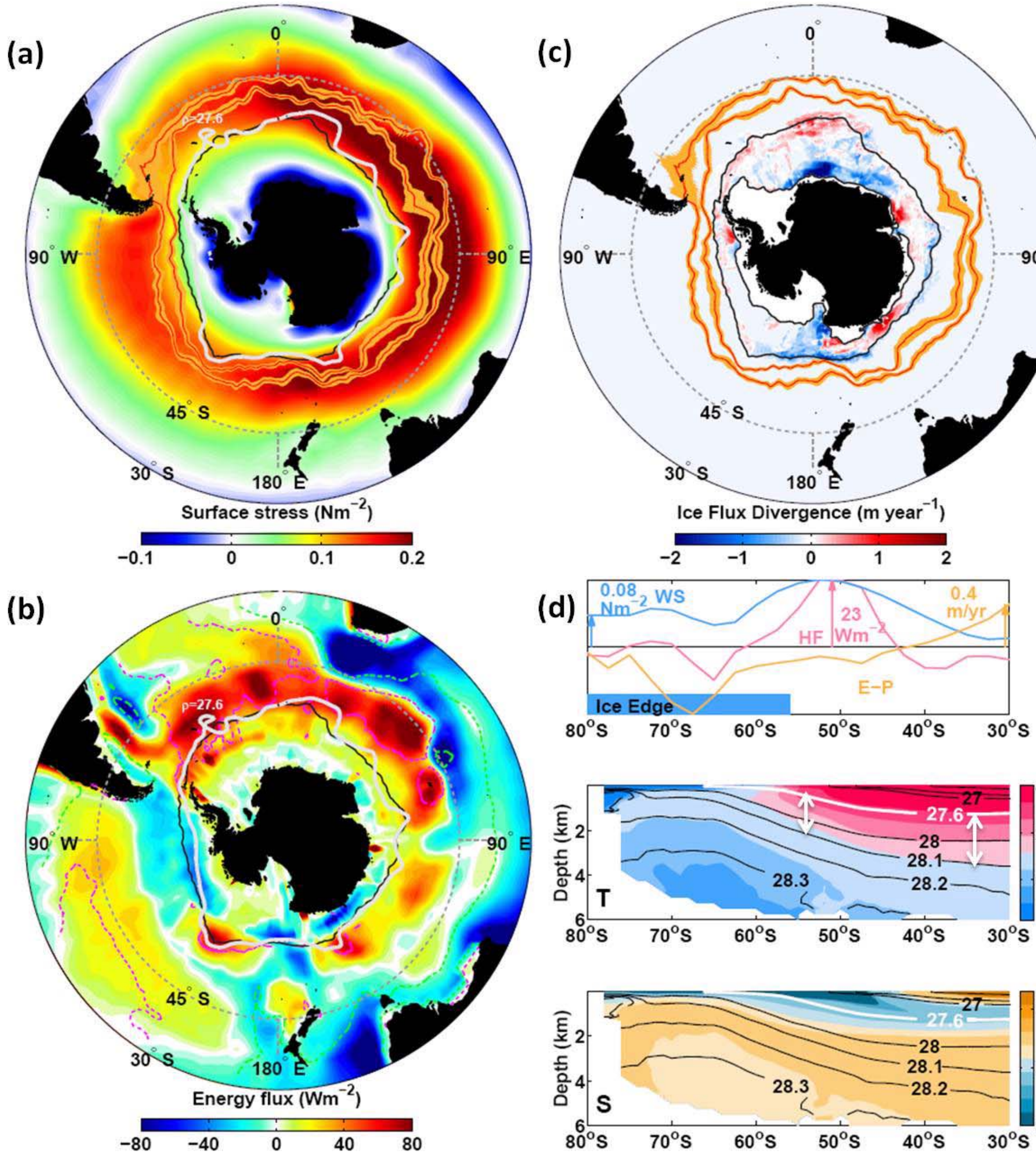

Figure 4: (a) The mean zonal wind stress, τ_x (color scale in units of Nm^{-2}) for the period 1980–2000 from the National Centers for Environmental Prediction (NCEP) reanalysis. Subantarctic and Polar fronts are marked in orange as in Fig.2. The winter ice edge is marked by the black line and the $\gamma^n = 27.6$ outcrop by the white line. (b) The NCEP mean net (radiative + sensible + latent) air–sea heat flux for the same period in Wm^{-2}, including contributions from E-P expressed as a pseudo-heat flux. Blue indicates regions where the heat flux is out of the ocean, and yellow-orange represents regions where it is directed in to the ocean. As in (a), the black and white lines mark the position of the winter ice edge and the 27.6 outcrop. (c) An estimate of the freshwater flux contribution from sea ice manufacture and export is plotted in units of myr^{-1}: red shading indicates buoyancy gain by the ocean and blue shading indicates buoyancy loss. The position of the summer and winter sea ice edge is marked in black. The area within the summer ice edge is shaded in white. The Subantarctic and Polar front positions are again marked in orange. (d) Circumpolar zonal-average T and S (scale on right) from the CARS2009 atlas (courtesy of CSIRO) with potential density contoured in black. The corresponding meridional profiles of wind-stress, E-P, and heat flux are plotted above computed from the NCEP reanalysis. The vertical white

$$-\rho_0 f\ \overline{vh} \ = \ \overline{F}_{eddy} + \ \overline{F}_{wind} - \frac{\Delta P}{L_x}, \tag{2}$$

expressing a balance between (i) the Coriolis force (ii) eddy 'form drag' forces related to the isopycnal eddy flux of potential vorticity $\overline{F}_{eddy} = \rho_0 \overline{h}^2 \overline{v'Q'}$, where $Q = (f + \zeta)/h$ is the Ertel potential vorticity, f is the planetary vorticity due to Earth's rotation (the Coriolis parameter) and ζ is the vertical component of the relative vorticity (iii) the driving force due to the prevailing winds $\overline{F}_{wind}$ and (iv) the (back) pressure gradient force, $\frac{\Delta P}{L_x}$, due to the intersection of the layer with bottom ridges and/or continental margins and L_x is the distance around Antarctica. The balance in Eq.(2) holds as long as the Rossby number of the large scale circulation is small, a condition well satisfied in the ACC system. The meridional mass flux, $\overline{vh}$, Eq.(1), is driven by the three terms appearing on the right hand side of Eq.(2). In midlatitudes, winds and pressure forces dominate the budget. But in the upper levels of the ACC where zonal flow is unblocked, the eddy forcing term enters at leading order, balancing in part the wind forcing, and resulting in a meridional residual flow.

Using the momentum budget in (2), we can now infer the MOC pattern in terms of hydrographic measurements of the density field. Hypothesizing that eddies flux Q down its large-scale gradient (Rhines and Young, 1982), v_{res} is an inevitable and rather general consequence of eddies acting on the geometry of the isopycnal surfaces. Beneath the direct influence of the wind and above topographic ridges, Eq.(2) implies that

$$\overline{vh} \ = -\frac{\overline{h}^2}{f}\overline{v'Q'} = \frac{\overline{h}^2}{f}K\frac{\partial \overline{Q}}{\partial y} = -K\frac{\partial \overline{h}}{\partial y}, \tag{3}$$

where the mean Q gradient is evaluated (neglecting the meridional variation of f — Marshall et al, 1993), on an isopycnal surface of mean thickness $\overline{h}$, K is an isopycnal eddy diffusivity and y is a coordinate increasing northwards. In Fig.4(d) we present the zonal-average potential density section with $\overline{T}$ and $\overline{S}$ superimposed. In this region the mean $\overline{Q}$ gradient is dominated by the changes in thickness of a layer between two density surfaces marked by the vertical white arrows in the figure. The thickening of layers moving northward implies, from Eq.(3), a southward and upward volume transport — a 'thickness diffusion' as in the closure of Gent and McWilliams (1990) — directed along sloping isopycnals, just as inferred from observations (Fig.3) and seen in the eddying model (Fig.5). This is the primary mechanism returning water from mid-depth to the surface. The implied volume flux is of magnitude $L_x K \partial\overline{h}/\partial y$ where $L_x \simeq 20,000\,\mathrm{km}$, $\partial\overline{h}/\partial y$ is the meridional thickness gradient, typically 500 m in 1000 km, and K measures the vigor with which the eddy field smooths out thickness gradients. Speer et al (2000) estimate that if $K = 1000\ \mathrm{m^2 s^{-1}}$, typical of eddy diffusivities at mid-depth, such gradients imply a residual southward water mass transport of roughly 10Sv, less than the Ekman transport (see estimates in Marshall et al, 2006) but comparable to the upwelling branch observed in Fig.3.

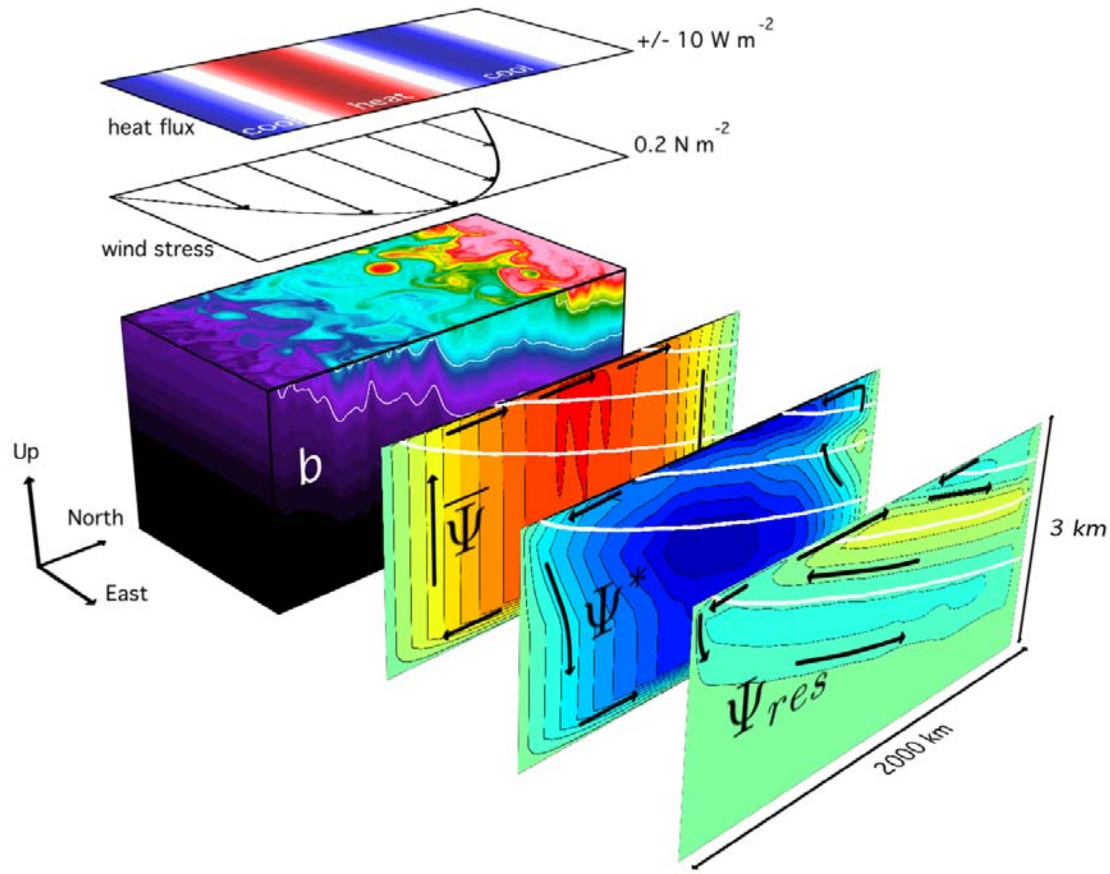

Figure 5: Results from numerical simulations of an eddying re-entrant channel showing (i) the wind and surface heat fluxes driving the channel flow (overlaid above) (ii) an instantaneous 3D snapshot of the model's buoyancy field with two buoyancy surfaces picked out in white, undulating in concert with the mesoscale eddy field and (iii) time-mean overturning cells $\overline{\psi}$, ψ^* and ψ_{res} (computed as defined in Eq.1) with time-mean buoyancy surfaces plotted in white. Antarctica is imagined to be on the left. The model is that of Marshall et al (1997a,b) run at a horizontal resolution of 4 km over a 1000 km by 3000 km domain. The driving wind and air-sea buoyancy fluxes are indicated above. The cooling, warming, cooling pattern on moving out from Antarctica are arranged to be reminiscent of Fig.4b, and are responsible for the Upper and Lower residual overturning cells marked on the far right of the figure. Courtesy of Ryan Abernathey, MIT.

4 Implications for role of Southern Ocean in climate and climate change

The North Atlantic MOC has long been considered a major control of the climate system driven by winds and excess salt left behind in the Atlantic as the result water vapor export (Broecker el al., 1985; Broecker, 1991) and whose strength is sensitive to fresh water discharges from adjacent continents caused by glacial melt (see, for example, recent discussions and reviews in Marotzke, 2000; Rahmstorf, 2002 and Alley et al, 2003). Huybers and Wunsch, 2010, however, argue that the dominant role of the North Atlantic is overemphasized in the literature. Indeed, along with the growing realization of the importance of the upwelling branch of the MOC, the Southern Ocean is now taking center stage in discussions of processes that drive modern and ancient climate variability.

Proxy data show that for a period of at least the last 800Kyr, Antarctic temperatures have covaried with atmospheric CO_2 although the relationship may not be causal. Moreover, it seems likely that marine processes operating in the Southern Ocean have played a central role in setting atmospheric CO_2 levels on glacial to interglacial timescales (see the review by Sigman and Boyle, 2000; Sigman et al, 2010). Biogeochemical box models indicate the strong sensitivity of atmospheric CO_2 levels to perturbations of high latitude surface ocean processes (see Sarmiento and Toggweiler, 1984; Knox and McElroy, 1984; Siegenthaler and Wenk, 1984; Toggweiler et al, 2003). The precise mechanisms are an active area of research. Many hypotheses invoke variations in the efficiency of communication between the carbon reservoir of the deep ocean and the surface (Sigman and Boyle, 2000). Accumulation of CO_2 in the deep, cold, sluggish ocean, occurs during glacial periods when, it appears, the ocean was more salty and more stratified (Adkins et al, 2002). Communication between the interior ocean and the surface may have been less efficient than today due mainly to reduced residual upwelling rates but also suppressed mixing. Release of abyssal CO_2 into the atmosphere may be indicative of increased exchange between the deep and the surface, and could have contributed to, for example, the transition out of the last ice age (deglaciation). Skinner et al (2010), for example, show that during the last glacial period, deep water circulating around Antarctica was perhaps more than twice as old relative to the atmosphere than it is today. During deglaciation, the dissipation of this old and presumably CO_2-enriched deep water, may have played an important role in the punctuated rise of atmospheric CO_2 (Anderson et al, 2009; Anderson and Carr, 2010).

Communication between the deep and surface in the modern ocean is governed by the upwelling branch of the MOC, but how it operated in glacial periods is unclear. Proxy data suggest that the winter sea-ice edge in glacial times was at the position of today's Polar Front (which is essentially locked in position by the interaction of the ACC with topography), and the summer edge was close to where the winter ice edge is today (Crosta et al., 1998; Gersonde et al, 2005). Sea-ice in glacial periods was perhaps more strongly packed and upwelling reduced in strength, changing the relative amount of deep global recirculation and hence deep water age.

A number of mechanisms have been proposed in which the southern ocean plays a central role in glacial cycles. Here we mention four, all of which address factors that control the communication rates between the interior carbon reservoir of the ocean and the surface, set by the upwelling branch of the MOC. They are not independent of one another and elements of each may have been at work.

1. sea-ice cover. Stephens and Keeling (2000) proposed that increased sea ice cover in glacial times would reduce the outgassing of CO_2 from the upwelling branch of the MOC thereby reducing the concentration of atmospheric CO_2.

2. the 'bipolar seesaw' (Broecker, 1998; Stocker,1998; Severinghaus, 2009). Here it is imagined that changes in the southern hemisphere are caused by a cessation of the MOC in the northern hemisphere, induced by freshwater release due to glacial melt in the north Atlantic/Arctic. Heat remains in the southern hemisphere due to reduced northward heat transport, melting back sea ice and allowing winds to more efficiently drive air-sea exchange between deep and surface waters.

3. southern hemisphere 'westerly wind shifts' (Toggweiler, et al, 2006; Toggweiler and Russell, 2008; Toggweiler, 2009; Anderson et al, 2009). Here it is supposed that in cold climates the surface westerlies may have been significantly equatorward (perhaps $\sim 10^\circ$) of their present position and so not 'aligned' with Drake Passage. Thus they may have been largely balanced by the pressure gradient term in Eq.(2) implying reduced residual meridional flow and upwelling rates. It is argued that as the climate warmed the surface westerlies shifted southward toward their current location thus turning on the upwelling branch of the MOC and enhancing communication between the abyss and the surface, releasing CO_2 to the atmosphere and so accelerating the warming.

4. a reorganization of air-sea buoyancy fluxes in glacial times. The pattern of ψ_{res} seen in, e.g. Fig.5, is connected diagnostically to the surface buoyancy balance ($\psi_{res}\partial\overline{b}/\partial y = B_s$: Walin, 1982; Marshall, 1997). Watson and Naveira-Garabato (2006) argue that the much colder atmospheric temperatures of glacial times and/or the reduced hydrological forcing, diminished the surface buoyancy fluxes, B_s, over the Southern Ocean, with a concomitant decrease in residual upwelling resulting from an increase in eddy fluxes, further compensating the Ekman driven Deacon Cell.

The Southern Ocean and its upwelling MOC branch are also central to our understanding of how the climate is responding to anthropogenic forcing. In today's climate the dominant mode of climate and atmospheric variability in the extra-tropical southern hemisphere is the Southern Annular Mode (SAM, Rogers and Van Loon, 1982; Thompson and Wallace, 2000). In recent decades SAM has shown a marked upward trend, the likely result of ozone depletion and (a lesser effect) anthropogenic global warming (Thompson and Solomon, 2002; Marshall, 2003, Polvani, personal communication). Strengthening of SAM and associated surface wind stress have been invoked to postulate enhancement in the strength, isopycnal slope and eddy heat fluxes of the ACC and thence the MOC (Hall and Visbeck, 2002; Webb and deCuevas, 2007; Hogg et al, 2008), thus affecting the Earth's climate system through changes in heat and carbon sequestration. Although changes in the slope of density surfaces in the ACC cannot yet be detected (Boening et al., 2008), ocean observations do indicate a freshening of Antarctic Intermediate Water (Wong et al., 1999) and a substantial warming of the upper 1000 m of the Southern Ocean at all depths (Gille, 2002, 2008) which may be linked to atmospheric forcing (Banks and Bindoff, 2003; Fyfe et al., 2007). Modeling studies and theory, however, suggest that eddy transport in the ACC (ψ^* — see Eq.1) can readily compensate for changes in Ekman transport ($\overline{\psi}$) leading to little change in the strength of the MOC (ψ_{res}): — see Henning and Vallis, (2005), Hallberg and Gnanadesikan (2006); Farneti et al, (2010a,b), Treguier et al (2010). Some of these studies use coupled models allowing two-way interaction between the atmosphere and ocean. However, representing ocean eddy mass transport in coarse coupled models depends not just on highly uncertain eddy parameterizations but also on their ability to capture stratification (Doney et al, 1998; Speer et al, 2000b).

The Southern Ocean is also the primary region where anthropogenic CO_2 enters the ocean from the atmosphere (Caldeira and Duffy, 2000; Sabine et al., 2004). It is subsequently incorporated into mode and intermediate waters. Isolated from the atmosphere, it may be transported equatorwards within the global MOC (Sarmiento et al., 2004, Mignone et al., 2006; Marinov et al, 2006). Observations suggest that the outgassing of natural CO_2 from the interior ocean has increased in the last twenty years (see le Quéré et al., 2007), offsetting the anthropogenic source. Some studies argue that this may be linked to an increase in the westerly winds blowing over the southern ocean (Lovenduski et al., 2008), whilst other studies question this relationship (Zickfeld et al., 2008). The net (natural + anthropogenic) CO_2 flux depends on the strength of the wind, upwelling, and on the mixed-layer cycle of carbon and nutrients and is thus directly related to the dynamics of the MOC's residual upwelling branch. Climate models predict that the wind stress over the Southern Ocean is likely to increase and shift slightly poleward (Meehl et al., 2007) under global warming and ozone depletion, although there remains some uncertainty (Thompson and Solomon, 2002). The consequence of this increase may be a reduction in the efficiency of the Southern Ocean sink of CO_2 and thus a possibly higher level of stabilization of atmospheric CO_2 on a multicentury time scale.

5 Conclusion and updated schematic of global circulation

On the global scale, water that sinks in high latitude convection is ultimately balanced by a return to the surface primarily in the Southern Ocean upwelling branch of the MOC. Here eddies transfer mass across the ACC and up to the surface layers. A new overturning schematic which places due emphasis on the upwelling branch is presented in Fig.6. To construct such a diagram much of the richness of the ocean's three-dimensional circulation structure has to be ignored, and only an overall, idealized representation of the cells can be presented (for a review of the rationale and history of overturning schematics, see Richardson, 2008). It is based on the quantitative results and qualitative ideas that have been developed in recent decades, and draws upon the many individual studies that were made as part of larger experimental programs, such as the World Ocean Circulation Experiment (see Siedler et al, 2001). The notion of deep water rising toward the surface to replace water blown north at the surface or dense water sinking along the bottom in the Southern Ocean is not new, as Sverdrup et al's (1942) own description clearly shows. However, what is new is the global quantification of these flows and property fluxes, the compelling dynamical theory put forth in explanation, and our growing realization of the importance of the upwelling system in the global climate. The new diagram links these old and new ideas and brings the Southern Ocean to the forefront of the meridional overturning circulation.

The link between upwelling and mesoscale eddy fluxes places a large burden on climate models as the eddy fluxes are computationally difficult to obtain and parameterizations of the fluxes may not always be faithful, especially in a changing climate. Will rising deep water bring natural CO_2 to the surface and overwhelm the Southern Ocean anthropogenic sink? Will changing winds accelerate the melting of Antarctic sea-ice and glaciers? These questions cannot be answered in isolation, but depend on detailed observations and models of an interacting ocean-atmophere-cryophere system. Nevertheless, important strides are being made in the physical and biogeochemical components of climate models and the development of observing networks, which will improve our understanding of the Earth system and reveal the deeper connections and long-term consequences of increasing CO_2 and other anthropogenic effects. While the northern hemispheric sinking has received so much attention as the axis of climate change, the upwelling branch in the Southern Ocean is now recognized as a vital component of our climate system, and an equally important agent of global impact.

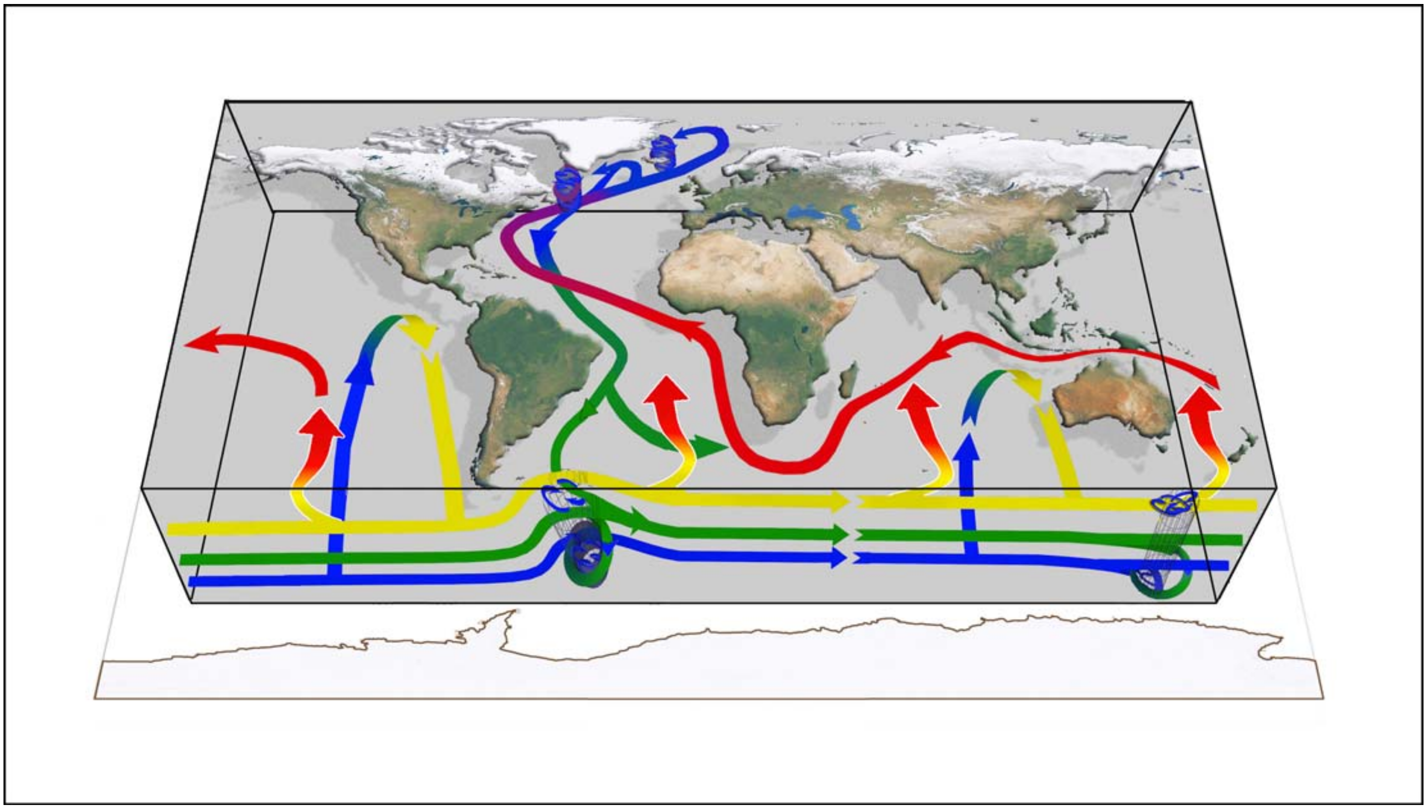

Figure 6: A cartoon of the ocean's global meridional overturning circulation illustrating the asymmetry between the Atlantic, Pacific, and Indian basins and between the northern and southern hemispheres. Colors schematically indicate the relative density of water masses: warmer mode and thermocline waters (red), upper deep waters (yellow), deep waters including North Atlantic Deep Water (green), and bottom waters (blue). The Upper Cell is fed predominantly by broad diapycnal upwelling at mid-depth across the major ocean basins (rising blue-green-yellow arrows). Upwelling to the surface occurs mainly around Antarctica along density surfaces (rising yellow-red arrows) with wind and eddy processes playing a central role. Much of this upwelled water is converted into mode waters, and, after entering various thermocline layers it continues to evolve, circulating in the ACC and subtropical regions, eventually re-supplying the upper branch of the global cell in the Atlantic with relatively warmer waters. In the northern North Atlantic warm water is converted to colder waters in the Subpolar Gyre and eventually becomes dense enough to sink under the thermocline in the polar seas and Labrador Sea convection regions (blue arrows). The dense water formed by convection in the Atlantic flows southward broadly in the deep branch of the overturning cell (green arrow), underneath the thermocline before joining the Circumpolar Current system. Below that, in turn, is the dense bottom water spreading from Antarctica feeding the upwelling branch of the Lower Cell. Convection regions mixing and cooling water masses in the vertical are indicated in the Weddell and Ross Seas around Antarctica and the Labrador and Greenland-Iceland Seas in the northern North Atlantic. A zonal-average view of this complex pattern is shown in Fig.1.

6 Acknowledgements

We would like to thank our many colleagues for discussions and comments during the writing of this review: in particular, Ryan Abernathey, J-M Campin, David Ferreira, Raf Ferrari, Mick Follows, Peter Huybers, David Marshall, Robbie Toggweiler, Ross Tulloch and Andy Watson. Thanks also to Nico Wienders, Ross Tulloch and Russ Windman for help in preparation of the figures.

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