Mechanisms controlling net air-sea heatflux over the Southern Ocean

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Abstract

The role of large-scale atmospheric eddies and the mean meridional circulation (the Ferrell cell) in explaining the observed net heating of the ocean in the latitude band of the Antarctic Circumpolar Current (ACC) is examined in the NCEP-NCAR reanalyses. Emphasis is placed on processes limiting evaporative cooling rather than on those promoting radiative heat gain. It is found that the moistening and drying of low levels induced by eddy and Ferrell cell moisture transports is small ($\sim 0.1 - 0.2 S v$, $1 S v = 10^9 kgs^{-1}$) in comparison of the moistening caused by surface evaporation ($\sim 1.5 S v$). As a result, it is suggested that evaporative cooling over the ACC is controlled by vertical moisture exchange between the free-troposphere and the boundary layer rather than by lateral moisture fluxes.

From the perspective of large-scale ocean atmosphere interactions, it is argued that a warm and moist midlatitude jet and the dynamical oceanic response to the associated surface windstress (inducing upwelling of cold water on the poleward flank of the ACC and its subsequent equatorward advection) are the two key ingredients required to explained net ocean heat gain over the ACC.

1 Introduction

The global ocean circulation is associated with the transformation of warm waters to cold waters in poleward flowing currents. To maintain a steady
state, the reverse transformation, from cold to warm, must occur (Walin, 1982). The relative importance of air-sea fluxes and interior diffusive processes in this transformation is still uncertain and much debated. Indeed it is one of the most central and long-standing questions in physical oceanography, as reviewed in Wunsch and Ferrari (2004). One possible scenario — sketched in Fig. 1 — is that sinking in northern Atlantic polar latitudes triggered by heat loss, is balanced by warming at the sea surface of the Southern hemisphere due to air-sea interaction, with interior mixing playing a secondary role (see for example Toggweiler and Samuels, 1998; Gnanadesikan, 1999; Marshall and Radko, 2003).

Of crucial importance for the surface diabatic mechanism depicted in Fig. 1 is that large-scale ocean-atmosphere interactions can sustain net heating of Southern Ocean waters. Inspection of net surface heat flux and sea level height indeed suggests that there is net surface heating of the ocean along the path of the Antarctic Circumpolar Current (ACC). The heating can be clearly seen in Fig. 2, which displays the net surface heat flux from the NCEP-NCAR reanalysis (Kalnay et al., 1996) as an annual average over the 1995-2004 period. Various estimates based on ‘bulk-formulae’ (Taylor et al., 1978; Grist and Josey, 2003) and global ocean inversions (e.g., Ganachaud and Wunsch, 2000) support the view that there is Southern Ocean heating, although the precise magnitude and geographical extent shown in Fig. 2 has considerable uncertainty.

From the point of view of the oceanographer, Southern Ocean heating is no surprise and is understood as a required feature of the meridional circulation of the Southern Ocean, the so-called “diabatic Deacon cell” (Speer
et al., 2000). Water parcels upwell in the Antarctic divergence south of the ACC, move equatorward in the upper Ekman layer - see Fig. 1- and thus experience a temperature increase as they cross the ACC front. This requires net surface heating. Simple budgets reveal that the magnitude of the heating is consistent with the strength of the circulation and the temperature change across the ACC temperature front (see, e.g, Marshall, 1997; Speer et al., 2000). As pointed out in Hoskins (1985), net surface heating of isotherms experiencing an eastward mechanical stress (as do outcropping isotherms within the ACC) is also expected from potential vorticity considerations.

Despite the soundness of these arguments, it is still somewhat a surprise to observe surface heating of the ocean at latitudes as far south as 40°−60°S. Taylor et al. (1978) suggest that this can be understood as the result of warm, moist advection of air by weather disturbances from the subtropics over cold ocean waters. Speer et al. (2000) emphasized the zonal asymmetry of the heating, with most of it being concentrated over the Atlantic-Indian sector where the ACC is displaced equatorward to the largest extent (Fig. 2). Again, cold waters under warm air is invoked to explain the heating.

An obvious criticism of these arguments is that, typically, sensible and latent heat flux actually cool the ocean surface and it is only the net radiative component of the surface flux which can provide heating (Csanady, 2001). To some extent this can be reconciled with the above studies in that the presence of warm, moist air above cold water helps reduce the intensity of (sensible and evaporative) cooling and thus favors warming. The role of baroclinic eddies might, however, not only be limited to low level heating and moistening through meridional transports. Indeed, baroclinic eddies also transport heat
and moisture upward which might have a cooling and drying effect on the surface layer. If this effect dominates, the net impact of atmospheric eddies on the ocean-to-atmosphere heat flux would then be to cool the surface by enhancing evaporative (and sensible) cooling.

It is the purpose of this paper to examine the role of large-scale atmospheric eddies, and their associated mean meridional circulation (the Ferrell cell), on the net ocean-to-atmosphere heat flux. Surprisingly, we will show that the key to explaining net Southern Ocean heating does not lie so much in understanding which of the drying/cooling and moistening/warming effects of eddies dominate. Rather, we will show that large-scale moisture transport terms (whether due to eddies or to the Ferrell cell) are small at low levels when compared to surface evaporation. A more indirect role for atmospheric eddies in the Southern Ocean heating problem is instead suggested: (i) through their control of the atmospheric vertical temperature and moisture structure—a warmer/moister air column leading locally to a weaker drying of the surface by turbulent entrainment and hence a reduced surface evaporation (ii) through their driving of surface winds and the associated oceanic Ekman mass transport and upwelling—a colder ocean mixed layer leading to reduced surface evaporation.

The paper is structured as follows. In Section 2, we analyze the various components of the net surface heat flux over the Southern Ocean and show that net heating essentially reflects a small residual between radiative heating and evaporative cooling. In Section 3 we investigate the mechanisms controlling surface evaporation by computing the moistening and drying of low levels associated with the advection of moisture by the mean meridional
circulation and by the eddy moisture flux divergence. In section 4 we pro-
vide further support for the controlling mechanism suggested in section 3
through an investigation of the time variability of surface evaporation over
the Southern Ocean. Conclusions are presented in Section 5. Unless other-
wise mentionned, all data used in this study come from the NCEP-NCAR
reanalysis (Kalnay et al., 1996) over the 1995-2004 period.

2 Net surface heating over the southern ocean

Net surface heating of the ocean \((Q_{net}, \text{positive if directed in to the ocean})\)
is written as the difference between net surface radiation \((R)\) and the sum
of the latent heat flux \((L_v E, L_v \text{ being the enthalpy of vaporization of water})\)
and sensible heat flux \((S)\),

\[
Q_{net} = R - (L_v E + S).
\]  

Fig. 3 presents an estimate of the zonal average of each of these terms.
All three components typically increase moving toward the equator. Net
surface radiation (continuous gray) reaches values of \(\sim 140 \text{ Wm}^{-2}\) at \(20^\circ S\)
and is directed into the atmosphere south of \(70^\circ S\), mostly as a result of zero
incoming shortwave radiation during the polar winter season (Nakamura and
Oort, 1983). Evaporation (thick black line) is weak south of \(70^\circ S\) and reaches
values in excess of \(120 \text{ Wm}^{-2}\) at \(20^\circ S\). Sensible heat fluxes (thin black line)
warm the surface poleward of \(65^\circ S\), because the surface is much colder than
the air above, but cools the ocean equatorward of this latitude, reaching a
magnitude of \(25 \text{ Wm}^{-2}\) at \(20^\circ S\).
Over the ocean (equatorward of 60°S), the balance (1) is dominated by the compensation between heating through surface radiation and cooling through surface evaporation. The net surface heat flux is indicated by the dashed gray curve in Fig.2. It shows a broad region of net surface heating of the ocean in the 40°S – 60°S band, roughly delimiting the meridional excursions of the ACC seen in Fig.1. This warming band is flanked by net cooling on the equatorial side and approximately zero heating on the poleward side (as is to be expected over the continent).

Over the 40°S – 60°S band, the mean value of each term in Eq.(1), weighted by surface area, is found to be \( R = 67 \, \text{Wm}^{-2} \), \( L_v E = 46 \, \text{Wm}^{-2} \) and \( S = 6 \, \text{Wm}^{-2} \) yielding a net heating \( Q_{net} \simeq 15 \, \text{Wm}^{-2} \). With a surface area of 5.7 \( 10^{13} \text{m}^2 \), this amounts to a net spatially integrated heating of about 0.9 PW (where 1 PW = 10^{15} W).

As emphasized in the introduction, one cannot be confident in the precise value of the surface heating over the Southern Ocean because various datasets yield different estimates. However, the above numbers are suggestive of a minor role for sensible cooling, allowing (1) to be simplified thus:

\[
Q_{net} \simeq R - L_v E
\]

This result is well known for subtropical and equatorial latitudes where SST is so high that evaporative cooling overwhelms sensible cooling. It is less obvious in mid-to-high latitudes. In the North Atlantic, for example, the presence of cold continental air masses blowing over the ocean, fundamentally changes the nature of air-sea interactions and both evaporative and sensible cooling are comparable (see Isemer and Hasse, 1987). The 40°S – 60°S
latitude band is essentially a water world, however, keeping sensible cooling to a minimum.

As a result of the simplification (2), the Southern Ocean heating problem can be phrased as two subproblems: (i) at fixed evaporative cooling, one wishes to understand which processes favour/limit net radiative heat gain (ii) at fixed net radiative flux, one wishes to understand which processes favour/limit evaporative cooling. We restrict ourselves to the second aspect in the following, and focus especially on the impact of atmospheric eddies on the low level moisture budget. It is clear that a full understanding of why net surface radiation should dominate over evaporative cooling also requires the consideration of (i). As a result, our study should only be considered as a first step towards a better understanding of Southern Ocean atmosphere-ocean interaction.

3 Low level moisture budget

To estimate which dynamical processes control surface evaporation over the $60^\circ S - 40^\circ S$ band, we consider the Eulerian, zonally averaged, equation for specific humidity $q$ in spherical and pressure coordinates,

$$\frac{\partial \bar{q}}{\partial t} + \frac{\bar{v}}{a} \frac{\partial \bar{q}}{\partial \phi} + \bar{\omega} \frac{\partial \bar{q}}{\partial p} + \frac{1}{a \cos \phi} \frac{\partial \bar{q}' q' \cos \phi}{\partial \phi} + \frac{\partial \bar{\omega} q'}{\partial p} = \chi$$

(3)

In this equation, an overbar denotes a zonal average, $a$ is the Earth radius, $v$ is meridional velocity, $\omega$ is (pressure) vertical velocity, $\phi$ denotes latitude, and $p$ pressure. The variable $\chi$ includes sources and sinks of moisture associated with small scale turbulence, phase changes and, in particular, evaporation of
seawater from the ocean surface. We estimate below the advection of specific humidity by the mean circulation and the eddy moisture flux divergence appearing in (3) over the Southern Hemisphere, focusing especially on a control volume defined by the $60^\circ S - 40^\circ S$ band (where net ocean heating is observed) and the $1000 mb - 800 mb$ layer (where the bulk of the Ferrell cell mass transport is found – see below). Each estimate is produced by averaging the calculations carried over ten consecutive Southern Hemisphere winters $^1$ using daily data from the NCEP-NCAR reanalyses (June through August, 1995-2004).

### 3.1 Moistening and drying by eddies and the Ferrell cell

The (Eulerian) mean meridional circulation and the zonally averaged moisture distribution are shown in Fig. 4. The prominent feature over the latitude band displayed ($70^\circ S - 20^\circ S$) is the thermally indirect (ascent over cold regions, descent over warm regions) Ferrell cell, with relatively moist air on its equatorward flank and relatively dry air on its poleward flank (thick contours). The total mass transport in the wintertime Ferrell cell is about 40 $Sv$ ($1 Sv = 10^9 kgs^{-1}$), with the bulk of the poleward flow being found over the $1000 mb - 800 mb$ layer. The specific humidity contrast across the Ferrell cell is on the order of a few $g/kg$, depending on the pressure level considered.

From the sense of circulation of the cells in Fig. 4, we expect the Ferrell $^1$The cold season was chosen because it is during this time that surface evaporation is largest.
cell to moisten low levels in midlatitudes since it advects moist air poleward over this region. We also expect a drying by the Ferrell/Hadley cell around \( \simeq 30^\circ S \), as a result of advection of dry air from aloft. A full calculation of moisture advection by the mean meridional circulation (first term underbraced in (3), expressed as a specific humidity tendency in \( g \ kg^{-1}/day \)) is provided in Fig. 5a. The expected low level moistening by the Ferrell cell is seen as the dashed contours over \( 60^\circ S - 40^\circ S \), with an amplitude of \( \simeq 0.2 \ g \ kg^{-1}/day \). As we will emphasize below, this number is small in comparison to surface evaporation. The mean meridional circulation drives a larger drying (continuous contours, \( \simeq 0.5 \ g \ kg^{-1}/day \)) over a thicker layer (1000mb – 500mb) in the subtropics.

The divergence of the eddy moisture flux (second term underbraced in (3)) is shown in Fig. 5b, with the same contouring interval as in Fig. 5a. At a given latitude, one typically observes a drying of low levels (continuous contours) with a magnitude reaching 0.6 \( g \ kg^{-1}/day \) at low subtropical levels and a moistening of upper levels (dashed contours) of weaker amplitude but acting over a thicker layer. This dipolar pattern reflects the upward transport of moisture by baroclinic eddies. The poleward transport of moisture by baroclinic eddies is indicated (at a constant pressure) by a tendency for drying in the subtropics and moistening in the extra-tropics, in particular close to the 850mb layer.
3.2 The impact of eddies and the Ferrell cell on the low level moisture budget

From an inspection of Fig. 5a,b we expect that, when integrated over the control volume \((1000\,mb - 800\,mb/60^\circ\,S - 40^\circ\,S)\) there is going to be significant compensation between Ferrell cell and eddy moistening/drying. Indeed, a careful estimate of the moisture budget terms from the reanalyses data (in “flux form” rather than the advective form (3)) produces the following numbers\(^2\),

\[
\begin{align*}
    \int_{60^\circ\,S}^{40^\circ\,S} \int_{775\,mb}^{1000\,mb} 2\pi a \frac{\partial (\bar{\varphi} \cos \phi)}{\partial \varphi} \frac{dp}{g} d\phi &\simeq -0.17 \, Sv \quad \text{(moistening)} \quad (4) \\
    \int_{60^\circ\,S}^{40^\circ\,S} \int_{775\,mb}^{1000\,mb} 2\pi a \frac{\partial (\bar{\omega} q) \cos \phi}{\partial \phi} \frac{dp}{g} d\phi &\simeq -0.18 \, Sv \quad \text{(moistening)} \quad (5) \\
    \int_{60^\circ\,S}^{40^\circ\,S} \int_{775\,mb}^{1000\,mb} 2\pi a^2 \cos \phi \frac{\partial (\bar{\omega} q)}{\partial p} \frac{dp}{g} d\phi &\simeq +0.18 \, Sv \quad \text{(drying)} \quad (6) \\
    \int_{60^\circ\,S}^{40^\circ\,S} \int_{775\,mb}^{1000\,mb} 2\pi a^2 \cos \phi \frac{\partial (\bar{\omega} q)}{\partial p} \frac{dp}{g} d\phi &\simeq -0.06 \, Sv \quad \text{(drying)} \quad (7)
\end{align*}
\]

with a net residual (the sum of all the above) of \(-0.23 \, Sv\) (moistening).

Eddies are seen indeed to have the moistening effect introduced in section 1 through an horizontal convergence of moisture transport across \(60^\circ\,S - 40^\circ\,S\) \((-0.18\,Sv)\). This effect is however largely cancelled through a larger upward eddy moisture transport at \(775\,mb\) than at \(1000\,mb\), yielding a net divergence (drying) of \(+0.18\,Sv\).

Irrespective of various cancellations, the most important result of this analysis is that all terms, even before any cancellation is taken into account,

\(^2\)Data were provided at the 850\,mb and 700\,mb levels, so the integrals were evaluated at \(775\,mb = (850 + 700\,mb)/2\) rather than at 800\,mb.
are small compared to surface evaporation,

$$\int_{60^\circ S}^{40^\circ S} 2\pi a^2 \cos \phi E d\phi \simeq +1.55 \text{ Sv (moistening)}$$  \hspace{1cm} (8)

The reason for the relative weakness of the eddy terms can be seen “geometrically” in Fig. 5b, in that the zero divergence line of eddy moisture flux (heavy black line in Fig. 5b) straddles the control volume. Further analysis indicates that this feature primarily reflects that the vertical moisture flux by the eddies peaks along approximatively the $\bar{q} = 3 g/kg$ contour. As for the mean meridional circulation, a simple scaling readily explains its small contribution to the low level moisture budget ($-(0.17 + 0.06) = -0.23 \text{ Sv}$).

Typically, an amount $\Psi_F \simeq 40 \text{ Sv}$ of air is brought in the $1000mb - 800mb$ layer by the Ferrell cell on its equatorward flank and the same amount leaves this layer at its poleward flank (Fig. 4). The difference in specific humidity between the air entering and leaving at low level is $\simeq 2 - 3 g/kg$ (Fig. 4), which yields a net moisture transport by the Ferrell cell $\mathcal{M}_F$ of amplitude

$$\mathcal{M}_F = \Psi_F \Delta_F q \simeq 40 \text{ Sv} \times 2.5 g/kg = 0.1 \text{ Sv}$$  \hspace{1cm} (9)

Were the Ferrell cell to vanish, the associated drying of the $1000mb - 800mb$ layer would be equivalent to a minor reduction in surface evaporation of about $0.1/1.5 \leq 10\%$. Put differently, the Ferrel cell would have to strengthen by one order of magnitude, at fixed specific humidity, to represent as large a moisture source for low levels as surface evaporation.

From the above we conclude that at low levels over $60^\circ S - 40^\circ S$, the moisture budget (3) reduces, in steady state, to

$$\bar{\chi} \simeq 0$$  \hspace{1cm} (10)
Large scale transport terms, whether associated with eddies or the mean meridional circulation, have little direct impact on low level $\bar{q}$. The simplified budget (10) describes a vertical balance between moistening of low levels by surface evaporation and drying by upward transport of moisture through the 800mb surface, presumably associated with boundary layer turbulence and entrainment of free-troposphere air into low levels.

4 Discussion

To bring further support for the vertical balance suggested in section 3, we have investigated in the NCEP-NCAR reanalyses the changes in atmospheric moisture distribution associated with changes in surface evaporation over the 60°S − 40°S band. Our aim is to seek a relationship between enhanced (reduced) evaporation and a locally drier (moister) free tropospheric column. The results of this analysis are displayed in Fig. 6, as a regression of monthly anomalies (departures from the mean seasonal cycle) of zonally averaged zonal wind $\bar{U}$ (Fig. 6a), temperature $\bar{T}$ (Fig. 6b) and specific humidity $\bar{q}$ (Fig. 6c) onto a monthly index of surface evaporation. The latter was constructed by averaging $\bar{E}$ over 60°S − 40°S. Note that to enhance statistical significance the regression analysis was carried out over a period of 25 years (1980-2004). The sign and amplitude of Fig. 6 is chosen so that each panel displays changes in atmospheric variables associated with one standard deviation reduction in surface evaporation over the 60°S − 40°S band. In Fig. 6a we observe a poleward shift of the westerly belt (the climatological mean jet is centred at 50°S − not shown), with $\bar{U}$ anomalies reaching amplitude of $\approx 1\ ms^{-1}$. 
Consistent with this poleward displacement of the jet and the associated anomalous subsidence and adiabatic warming over the $60^\circ S - 40^\circ S$ band (not shown), Fig. 6b indicates a warming of the atmospheric column over $60^\circ S - 40^\circ S$ with an amplitude of $\approx 0.3K$. Such relationships between anomalous $\bar{U}$ and $\bar{T}$ are reminiscent of the changes in atmospheric circulation associated with the Southern Annular mode (Thompson and Wallace, 2000). Indeed, reduced surface evaporation in middle latitudes is usually found associated with a positive phase of the annular modes (see Verdy et al. (2006) for the Southern Ocean and Cayan (1992) for the North Atlantic).

The changes in $\bar{q}$ associated with a reduced evaporation over $60^\circ S - 40^\circ S$ (Fig. 6c) consist of a moistening of the atmospheric column over that latitude band. This is consistent with the idea that changes in $\bar{q}$ induce a reduction in evaporation. If reduced evaporation were to drive $\bar{q}$ changes, a drier, rather than a moister, atmospheric column would result. Note also that changes in surface windspeed are small when averaged over $60^\circ S - 40^\circ S$ (Fig. 6a indicates dipolar anomalies in $\bar{U}$ centred at $50^\circ S$) so that, from the perspective of bulk-formulae, they could not account for the reduced surface evaporation. Further analysis indicates that a pattern very similar to Fig. 6c is obtained if $\bar{q}$ changes (denoted by $\delta \bar{q}$) are predicted using the approximate relation (not shown),

$$\delta \bar{q}(\phi, P) \approx RH_o(\phi, P) \times \delta q_{sat}(\bar{T}, P)$$

in which $q_{sat}$ is the saturation specific humidity and $RH_o$ is the climatological mean relative humidity at a given latitude and pressure. In other words, an increase in specific humidity of a westerly belt column can be simply un-
derstood as resulting from a warmer column, the (zonally averaged) relative humidity within the $60^\circ S - 40^\circ S$ latitude band not changing significantly when the jet moves.

In summary, on monthly and interannual timescales (those possibly resolved with a 25-yr record), changes in surface evaporation over $60^\circ S - 40^\circ S$ are driven by meridional shifts of the jet and its associated warming/moistening and cooling/drying of free tropospheric columns over that latitude band. This result is entirely consistent with the control of surface evaporation by local, vertical exchanges of moisture between the free troposphere and the surface invoked in section 3. More work is needed to assess whether it is the moister/drier air entrained which is responsible for the change in surface evaporation or if it is the change in static stability associated with a warmer/colder column aloft which reduces the entrainment velocity and hence surface evaporation.

5 Conclusion: Southern Ocean heating as a coupled Ocean-Atmosphere problem

Our diagnostic study suggests that the main cooling process at the Southern Ocean surface is evaporative cooling. For a given net radiative heating at the surface, the Southern Ocean will experience net heat gain if the evaporative cooling is sufficiently weak. Attempts at explaining this reduction as a result of meridional advection of moist air from lower latitudes by large-scale atmospheric eddies at low levels fail. The reason for this conclusion is
that atmospheric eddies also dry low levels as a result of upward moisture transport. A study of the moisture budget in the NCEP-NCAR reanalysis suggests the moistening and drying effect of eddies have comparable strength at low levels.

Instead it is found that large scale transport terms (eddies and/or mean meridional circulation) are small at low levels compared to the moisture source associated with surface evaporation. The reason lies in the spatial structure of the (zonally averaged) eddy moisture flux, whose zero divergence line, in a meridional-height plane, straddles low levels over $60^\circ S - 40^\circ S$. As for the mean meridional circulation (the Ferrell cell), its mass transport would have to increase by one order of magnitude for it to become a low level moisture source of similar strength as surface evaporation. As a result, surface evaporation must primarily be balanced by turbulent upward moisture transport in the boundary layer, i.e. by the vertical exchange of moisture between the interior troposphere and the surface. This picture is confirmed by the analysis of temporal changes in surface evaporation over the 1980-2004 period, which are observed to arise as a local response to the drying/moistening of the atmospheric column caused by meridional shifts of the westerly belt.

This atmospheric control of surface evaporation is, however, clearly not the sole mechanism involved in net Southern Ocean heating and we would like to emphasize the coupled (Ocean-Atmosphere) dynamics at work. From bulk-formulae (e.g., Gill, 1982), we express the heating of the ocean $Q_{net}$ as

$$Q_{net} \simeq R - L_v \rho_a C_D |U| (q_{sat}(T_s) - q_a) \quad (12)$$

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in which $R$ is the net radiative energy gain, $L_v$ is the enthalpy of vaporization of water, $T_s$ the sea surface temperature, $q_{sat}$ is the saturation specific humidity, $q_a$ the low level atmospheric specific humidity, $\rho_a$ air density, $C_D$ a drag coefficient, $|U|$ surface windspeed and we have neglected sensible cooling (see section 2). At fixed $R$, net heating is favoured if evaporation can be reduced. Over the westerly belt in which surface windspeeds are high, this can only arise as a result of weak moisture contrast $q_{sat}(T_s) - q_a$. This is achieved in two ways. First, and this is the atmospheric control emphasized in section 4, by having a relatively warm midlatitude jet and high relative humidity in the storm-track. For a given strength of turbulent entrainment, more moist air will then be entrained in the boundary layer and $q_a$ will consequently be relatively high. Second, through ocean dynamics. The response of the ocean to the eastward mechanical windstress imposed by the atmospheric jet is an equatorward mass flux in the upper layer (the Ekman mass transport) with its associated upwelling of cold water on the poleward flank of the ACC. This upwelling brings cold water to the surface which tends to favor low $q_{sat}(T_s)$. We speculate that slow changes in the temperature of the upwelled water, perhaps linked to changes in the characteristics of North Atlantic Deep Water, will introduce slow changes in surface evaporation, illustrating an oceanic control on surface evaporation.

Finally, we would like to briefly return to the broader perspective, put forward in section 1, of the global ocean circulation and the role of air-sea fluxes and interior mixing in the dynamics of its overturning circulation. Net heating is one way to increase the buoyancy of surface waters over the Southern Ocean and provide a mechanism for balancing the densification
occurring over warm western boundary currents such as the Gulf Stream. Such lightening might also be provided by precipitation and/or ice melt. Indeed, the net heating of about 15 $W m^{-2}$ found over the $60^\circ - 40^\circ S$ latitude band is comparable to a net freshwater flux of about 0.8 $Sv$, which is of the same order as the net precipitation estimated by Ganachaud and Wunsch south of $30^\circ S$ ($\approx 0.8 \pm 0.9Sv$; Ganachaud and Wunsch, 2003). Both surface heating and freshening are thus probably important in converting dense waters to light waters in the Southern Ocean.

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**Figure 1:** Schematic of a possible interpretation of the ocean’s overturning circulation in the meridional-height plane. The volume of warm water (shaded) is maintained, in steady state, by a balance between conversion of warm to cold water through surface cooling over western boundary currents such as the Gulf Stream in the North Atlantic, and the reverse conversion of cold to warm waters through surface heating over the upper branch of the Deacon cell in the Southern Ocean.

**Figure 2:** Long term mean net air-sea heat flux over the Southern ocean (contoured every 20 $Wm^{-2}$, positive when heating -continuous-, negative when cooling -dashed) from the NCEP-NCAR reanalysis. The mean sea surface height (gray contours, every 10 cm) from the Topex-Poseidon altimeter over 1992-2002 is superimposed. Only the surface height contours that pass through Drake Passage are shown.

**Figure 3:** Zonal mean annual net surface radiation ($R$), evaporative flux ($L_vE$), sensible heat flux ($S$) and the net surface heat flux $Q = R - (L_vE + S)$. All fluxes are in $Wm^{-2}$.

**Figure 4:** Wintertime mean Eulerian mass transport streamfunction (thin continuous lines when clockwise, dashed when anticlockwise, $CI = 10 Sv$) and zonally averaged specific humidity (thick contours, $CI = 1g/kg$). The streamfunction was set to zero at 1000mb.

**Figure 5:** Wintertime mean (a) advection of specific humidity by the
mean meridional circulation (b) eddy moisture flux divergence. Continuous (dashed) contours indicate drying (moistening) with a contour interval of 0.1 g kg\(^{-1}\)/day. The zero contour is shown as heavy black line. Note that the divergence value at 962.5 mb was extrapolated to the 1000 mb level.

**Figure 6:** Regression of monthly anomalies of zonally averaged (a) Zonal wind (CI = 0.2 m/s) (b) Temperature (CI = 0.1 K) (c) Specific humidity (CI = 0.01 g/kg) onto an index of surface evaporation over 60\(^\circ\) S – 40\(^\circ\) S. All data were linearly detrended prior to computing the regression. Continuous contours indicate positive values, dashed negative.
Schematic of a possible interpretation of the ocean’s overturning circulation in the meridional-height plane. The volume of warm water (shaded) is maintained, in steady state, by a balance between conversion of warm to cold water through surface cooling over western boundary currents such as the Gulf Stream in the North Atlantic, and the reverse conversion of cold to warm waters through surface heating over the upper branch of the Deacon cell in the Southern Ocean.
Long term mean net air-sea heat flux over the Southern ocean (contoured every 20 W m^{-2}, positive when heating -continuous-, negative when cooling -dashed) from the NCEP-NCAR reanalysis. The mean sea surface height (gray contours, every 10 cm) from the Topex-Poseidon altimeter over 1992-2002 is superimposed. Only the surface height contours that pass through Drake Passage are shown.
Zonal mean annual net surface radiation ($R$), evaporative flux ($L_v E$), sensible heat flux ($S$) and the net surface heat flux $Q = R - (L_v E + S)$. All fluxes are in W m$^{-2}$. 

$Q_{net} = R - (L_v E + S)$
Fig. 4

Wintertime mean Eulerian mean mass transport streamfunction (thin continuous lines when clockwise, dashed when anticlockwise, CI = 10 Sv) and zonally averaged specific humidity (thick contours, CI = 1 g/kg). The streamfunction was set to zero at 1000mb.
Fig. 5

Wintertime mean (a) advection of specific humidity by the mean meridional circulation (b) eddy moisture flux divergence. Continuous (dashed) contours indicate drying (moistening) with a contour interval of 0.1 g/kg per day. The zero contour is shown as heavy black line. Note that the divergence value at 962.5mb was extrapolated to the 1000mb level.
Regression of monthly anomalies of zonally averaged (a) Zonal wind (CI = 0.2 m/s) (b) Temperature (CI = 0.1K) (c) Specific humidity (CI = 0.01 g/kg) onto an index of surface evaporation over 60S-40S. All data were linearly detrended prior to computing the regression. Continuous contours indicate positive values, dashed negative.