The partitioning of poleward heat transport between the atmosphere and ocean

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Submitted to Journal of Atmospheric Sciences
Abstract:

Observations of the poleward heat transport of the Earth ($H$) suggest that the atmosphere is the primary transporting agent polewards of 30°, that oceanic ($H_O$) and atmospheric ($H_A$) contributions are comparable in the equatorial belt, and that ocean transport dominates in the deep tropics.

To study the partition we express the ratio $H_A/H_O$ as

$$\frac{H_A}{H_O} = \frac{\Psi_A}{\Psi_O} \times \frac{C_A \Delta \theta_A}{C_O \Delta \theta_O}$$

where $\Psi$ (with subscripts $A$ and $O$ denoting, respectively, atmosphere and ocean) is the meridional mass transport within $\theta$—layers (moist potential temperature for the atmosphere, potential temperature for the ocean) and $C \Delta \theta$ ($C$ being the specific heat) is the change in energy across the circulation defined by $\Psi$.

We argue that the observed partition of heat transport between the atmosphere and ocean is a robust feature of the Earth climate and reflects two limits:

(i) dominance of atmospheric mass transport in mid-to-high latitudes ($\Psi_A \gg \Psi_O$ with $C_A \Delta \theta_A \sim C_O \Delta \theta_O$ and hence $H_A/H_O \gg 1$)

(ii) dominance of oceanic energy contrast in the tropics ($C_O \Delta \theta_O \gg C_A \Delta \theta_A$ with $\Psi_A \sim \Psi_O$ and hence $H_A/H_O \ll 1$).

Motivated by simple dynamical arguments, these ideas are illustrated through diagnosis of the NCEP-NCAR reanalysis, and long simulations of an ocean model and a coupled atmosphere-ocean model of intermediate complexity.
1 Introduction

Fig.1a shows an estimate of the atmospheric \( H_A \), black) and oceanic \( H_O \), grey) heat transport from data published by Trenberth and Caron (2001), based on NCEP and ECMWF atmospheric reanalyses (continuous and dashed curves, respectively). Both atmospheric products suggest that (Fig.1b), poleward of about 30°, the atmospheric contribution to the total poleward heat transport amounts to roughly 90% of the total. As one approaches lower latitudes, however, both contribute in roughly equal amounts, although a precise estimate is made difficult by the fact that both \( H_O \) and \( H_A \) become small in the tropics (note the large spread between the dashed and continuous curves in Fig. 1b equatorward of 10°).

An obvious question is whether there is a simple physical explanation for the partitioning suggested in Fig.1b. An important step towards this goal was recently made by Held (2001), who represented the oceanic and atmospheric heat transport as the product of an overturning mass transport streamfunction \( \Psi \) and an energy contrast \( C \Delta \theta \),

\[
H_{A,O} = \Psi_{A,O} C_{A,O} \Delta \theta_{A,O} \tag{1}
\]

where the subscripts \( A, O \) denote atmospheric or oceanic values, and \( C \) and \( \Delta \theta \) being, respectively, the heat capacity and difference in potential temperature across the upper and lower branches of the overturning circulation. Such a decomposition is commonly carried out in oceanography (e.g., Talley, 2003). Using Eq.(1), the ratio of atmospheric to oceanic heat transport becomes

\[
\frac{H_A}{H_O} = \frac{\Psi_A}{\Psi_O} \times \frac{C_A \Delta \theta_A}{C_O \Delta \theta_O} \tag{2}
\]
Figure 1: (a) Estimates of oceanic ($H_O$, grey) and atmospheric ($H_A$, black) heat transport in PW ($1PW = 10^{15}W$). (b) Relative contribution of ocean (grey) and atmosphere (black) to the total energy transport $H = H_O + H_A$ using (a). Continuous curves correspond to NCEP-based estimates, while dashed curves correspond to ECMWF-based estimates. The ratios in (b) were not plotted in the deep Tropics ($2^\circ S - 2^\circ N$) where $H_A + H_O$ vanishes.
In the tropics, where transfer by quasi-two-dimensional atmospheric eddies is largely absent, Held (2001) argued that oceanic and atmospheric mass transports are close to one another and set by the Ekman meridional mass transport. He went on to propose that the relative magnitude of $C \Delta \theta$ in Eq.(1) must be responsible for the partitioning. To rationalize this, Held suggested that $C \Delta \theta$ is essentially a measure of the oceanic and atmospheric stratification in the tropics. Oceanic ‘stratification’ is high in the deep tropics where upwelling of cold fluid to the warm surface on the equator forms a pronounced thermocline. In the atmosphere, by contrast, tropical deep convection acts to return the fluid to marginal stability in which $\theta_A$, the moist potential temperature, has weak vertical gradients. The state of affairs is schematized in Fig.2. Hence, as Held (2001) emphasized, dominance of ocean over atmospheric heat transport is expected in the deep tropics.

It is not straightforward to carry this argument to mid-to-high latitudes where it is well established that atmospheric transient eddies cannot be ignored when estimating meridional mass transports. Thus we might expect $\Psi_A$ and $\Psi_O$ to decouple moving away from the tropics, and the role of stratification in setting the partitioning becomes much less clear.

In this paper, we attempt to estimate oceanic and atmospheric heat transports globally, motivated by the decomposition Eq.(1). To do so, we will estimate the meridional mass transports within constant energy, or potential temperature layers (defined by moist static energy in the atmosphere and potential temperature in the ocean), which will allow us to estimate $\Psi$ and $C \Delta \theta$ directly from data. Our main conclusion is that the partitioning seen in Fig.1b can be simply understood as two limits of Eq.(1): a mass trans-
Figure 2: Schematic of the distribution of atmospheric moist potential temperature ($\theta_A$) and oceanic potential temperature ($\theta_O$) as a function of latitude and height (black contours).
port ‘$\Psi$ limit’ in mid-to-high latitudes where $\Psi_A \gg \Psi_O$ and the atmospheric contribution to $H$ dominates; an energy contrast ‘$C \Delta \theta$ limit’ in the tropics where $C_A \theta_A \ll C_O \theta_O$ and the oceanic contribution to $H$ overwhelms that of the atmosphere.

The paper is set out as follows. Sections 2 and 3 present an estimate of $\Psi_A$ and $\Psi_O$ from, respectively, the NCEP-NCAR reanalyses and a long simulation of an ocean circulation model. The application of these estimates to the partitioning problem will be presented in section 4. In section 5 we propose simple scalings to understand the mass transport and energy contrast limits highlighted above. Motivated by these simple arguments, in Section 6 we analyze a coupled climate model of intermediate complexity run in an aquaplanet geometry. A partitioning which is very similar to the modern observational record is found, suggesting that it is likely to be a robust feature of earth’s climate. We summarize and conclude in section 7.

2 Atmospheric heat transport

2.1 Meridional mass transport within moist potential temperature layers

We use daily estimates of temperature $T$, geopotential $\Phi$ and specific humidity $q$ from the NCEP-NCAR reanalysis (Kalnay et al., 1996) on a $2.5^\circ \times 2.5^\circ$ grid and 17 pressure levels. The central diagnostic quantity is the moist potential temperature, $\theta_A$, defined by the relation

$$C_A \theta_A = C_A T + L_v q + \Phi$$  (3)
where $L_v$ is the latent heat of vaporization and $C_A$ is the specific heat capacity of dry air. Note that $\theta_A$ is just a scaled version of the widely used moist static energy. Fig. 3 shows the zonal mean distribution of $\theta_A$ (dashed contours) which can be compared to the potential temperature $\theta$ (continuous contours). The most striking feature is the homogeneity of $\theta_A$ in the tropics, which is consistent with the tropical atmospheric lapse rate being close to that of a moist adiabat (Xu and Emanuel, 1989). Another interesting tropical feature is the presence of a low to mid-level minimum of $\theta_A$, associated with the subsidence of dry air from aloft. In mid-latitudes, one observes that $\theta_A$ contours slope more steeply than $\theta$ contours due to the water vapor loading term in Eq.(3). Note that the shading in Fig. 3 indicates the zonal mean distribution of Ertel’s potential vorticity (PV) — the 3 PV unit contour (the dashed white line) will be used below to define the position of the tropopause.

On a given day, the meridional mass transport within an air column is partitioned into a set of $\theta_A$—layers with a resolution $\delta \theta_A = 5K$ (finer $\theta_A$ grids were also considered but the results were found to be insensitive to this choice). For each $\theta_A$—layer the total meridional transport across a given latitude was then computed by summing the contributions over each longitude. Note that to enforce mass conservation, the long term mean northward mass transport $M$ (in $kgs^{-1}$ per $\delta \theta_A$), summed over all temperature classes, was set to zero for each latitude. From $M$, at each latitude $\phi$ we compute a mass transport streamfunction $\Psi_A$ (in $kgs^{-1}$) using the definition:

$$\Psi_A(\phi, \theta_A) = -\int_{\theta_A^{min}}^{\theta_A} M(\phi, \theta_A') d\theta_A'$$  

(4)
Figure 3: Zonal mean distribution of $\theta_A$ (dashed contours) and $\theta$ (continuous contours) for the period from May 1st 2003 to September 30th 2003. The contour interval is $10K$. Also shown (shading) is the zonal mean distribution of Ertel’s PV, in $PVU$ ($1 \text{ PVU} = 10^{-6} \text{Km}^2 \text{s}^{-1} \text{kg}^{-1}$). The tropopause (defined as the $3 \text{ PVU}$ contour) is indicated by the thick dashed white line.
where we imposed that $\Psi_A$ vanishes for all latitudes at some low temperature $\theta_A^{\text{min}}$. With the sign convention adopted, a positive $\Psi_A$ indicates clockwise circulation in the $(\phi, \theta_A)$ plane. An appendix sets out details of the mass transport calculations.

Fig. 4 shows our estimate of $\Psi_A$ for the time period spanning May 1st 2003 to September 30th 2003 (Fig.4a, Southern Hemisphere winter) and for the time period spanning November 1st 2002 to March 31st 2003 (Fig.4b, Northern Hemisphere winter). In the Southern hemisphere, one observes a single equator-to-pole circulation in both seasons, with poleward mass transport at high $\theta_A$ and equatorward mass transport at low $\theta_A$. The wintertime cell is the most intense, reaching a maximum of more than 200$Sv$ at about 40° of latitude. Note that we have ‘redefined’ a Sverdrup (the traditional oceanographer’s unit of volume transport) to represent a mass transport, i.e. $1Sv \equiv 10^9 kgs^{-1}$ \(^1\). In the Northern Hemisphere, the summertime result (Fig.4a) again shows a single equator-to-pole cell with a maximum of about 100$Sv$ near 30° of latitude. In winter, however, there is indication of two separate ‘centers of action’ in Fig.4b near 20° and 50° of latitude, each with an amplitude of about 140$Sv$. Anticipating somewhat what is to follow, we note that meridional mass transports of 100 – 200 $Sv$ are significantly larger than the typical 20$Sv$ observed in the ocean. It is suggested below that this

\(^1\)This choice is made for ready comparison to oceanic mass transports. The density of seawater is always close to $10^3 kgm^{-3}$, so that a Sverdrup of $10^6 m^3 s^{-1}$ represents a mass transport of $10^6 kgs^{-1}$. Accordingly, a mass transport of, say, $20 \times 10^6 kgs^{-1}$ is the same as a volume transport of $20 \times 10^6 m^3 s^{-1}$, i.e. ‘volume Sverdrups’ and ‘mass Sverdrups’ are equivalent to one-another.
Figure 4: Meridional mass transport streamfunction within $\theta_A$ layers (contoured every $25\,Sv$ where $1\,Sv = 10^9\,kg\,s^{-1}$; positive when clockwise — see arrows; zero contour omitted; extrema indicated on the plot) for (a) Southern and (b) Northern Hemisphere winter of 2003. The $x$—axis denotes latitude and the $y$—axis moist potential temperature (in $K$). The upper (lower) grey curve is the zonally average $\theta_A$ at the tropopause (ground).
is indeed the major reason why the atmospheric contribution to $H$ dominates over that of the ocean in mid-to-high latitudes.

In Fig.4, we also plot (grey lower curve) the zonally averaged mean surface $\theta_A$. It is readily seen that most of the equatorward mass transport occurs in layers with temperatures colder than the mean (zonally average) surface $\theta_A$. In midlatitudes this is because — see Held and Schneider (1999) — the equatorward mass transport is achieved by cold air outbreaks, i.e. transient features of the general circulation. In the tropics however, there is a different reason: a picture very similar to Fig. 4 can be obtained (in the $20^\circ S - 20^\circ N$ latitude band) if seasonal mean, rather than daily fields, of meridional velocity and moist potential temperature are used (not shown). In the tropics, equatorward mass transport at a lower $\theta_A$ than that of the ground occurs because of the presence of a minimum of $\theta_A$ at mid-to-low levels (see Fig.3): the equatorward mass transport is shared by high (surface) and low (lower to mid-tropospheric levels) $\theta_A$ layers, thereby being at a lower average $\theta_A$ than the surface.

The upper grey curve in Fig.4 is the value of $\theta_A$ on the tropopause. It is seen that the latter gives a good estimate of the upper boundary of the cells for all seasons and both hemispheres.

2.2 Further discussion of atmospheric mass transport

A striking result in the above analysis is that the Hadley and midlatitude eddy-driven cells tend to be ‘joined’ in to a single cell (Fig.4a,b). This is a pronounced feature of our diagnostics, even more so than in previous es-
timates of meridional mass transport streamfunction within ‘dry’ potential temperature layers — the so-called residual circulation (e.g., Karoly et al., 1997; Held and Schneider, 1999).

To understand the origin of this difference, we display in Fig.5 the analog of Fig.4 but for a calculation in which we have set $L_v = 0$ in (3), thereby only considering the contribution of dry static energy to the heat transport. One then observes a much more pronounced ‘two cell’ structure in each winter hemisphere, with the tropical (Hadley) cell dominating. The differences between Figs.4 and 5 can only arise because of moisture effects. They simply reflect that the moisture and dry static energy are both transported poleward in mid-to-high latitudes, and thus add to one another to create the vigorous overturning cell in midlatitudes seen in Fig.4, while they oppose each other in the Tropics, thereby reducing the pronounced Hadley cell component in Fig.4.

A second important aspect we wish to discuss is whether the midlatitude circulation in Fig.4a,b can indeed be interpreted as an overturning cell in the meridional–height plane. To address this issue, we check against observations that $\Psi_A$ is well approximated by $\Psi_A^{TEM}$, the streamfunction advecting the zonal mean $\theta_A$ in a Transformed Eulerian Mean (TEM) procedure (Andrews et al., 1987) applied to $\theta_A$. This is written as

$$\Psi_A \simeq \Psi_A^{TEM} \equiv \Psi_A^{Eul} + \Psi_A^{Stokes}$$

(5)

where $\Psi_A^{Eul}$ is the Eulerian mean mass streamfunction and $\Psi_A^{Stokes}$ is the quasi-Stokes mass streamfunction (e.g., Hoskins, 1983; Plumb and Ferrari,
Figure 5: Same as Fig. 4 but for the meridional mass streamfunction within dry static energy layers. The $y$-axis now denotes an equivalent potential temperature $\theta_e$ (in K) defined by $C_A \theta_e = C_A T + \Phi$. 
2004) defined as

$$
\Psi_A^{Stokes} = \frac{L_x}{g} \frac{\overline{v^' \theta^'_A}}{\partial \theta_A / \partial p}
$$

(6)

In Eq. (6), $L_x$ is the length of a latitude circle, $g$ is gravity, $p$ pressure, and $\overline{v^' \theta^'_A}$ is the meridional eddy flux of moist potential temperature (primes denote departures from zonal mean, itself denoted by an overbar).

Fig. 6 compares $\Psi_A$ from the direct calculation (Fig. 6a, reproduced from Fig. 4a) and an estimate of each term in the rhs of Eq. (5) for the Southern Hemisphere winter of 2003. Both the Eulerian (Fig. 6c) and the quasi-Stokes (Fig. 6b) streamfunctions were computed in the latitude-pressure plane, then mapped onto the latitude–$\theta_A$ plane using zonal mean, wintertime averaged $\theta_A$ profiles. It is seen that in midlatitudes, the quasi-Stokes contribution alone is a very good description of the intensity and meridional/$\theta_A$ scales of $\Psi_A$ (as before, grey curves indicate the $\theta_A$ of the tropopause and the ground). In midlatitudes, the Eulerian contribution — the Ferrell cell — is a thermally indirect cell, of much smaller intensity. Toward the tropics, however, the Eulerian contribution — the wintertime Hadley cell — dominates the total mass transport and provides a good approximation to $\Psi_A$.

That the approximation Eq. (5) holds gives support to the interpretation of $\Psi_A$ as a true overturning cell in the meridional-height plane.3 This interpretation can also be understood more physically by noting that $\Psi_A^{Stokes}$

\footnote{Note that only transient eddies were included in the computation of $\overline{v^' \theta^'_A}$ in (6). The contribution of steady eddies was found to be negligible except close to the Antarctic continent, where the contribution was noisy (not shown).}

\footnote{The good agreement between $\Psi_A$ and $\Psi_A^{Tout}$ is not unexpected for these $\theta_A$ layers which do not intersect the ground (see McIntosh and McDougall, 1996).}
in Eq. (5) is an approximation to the Stokes drift associated with baroclinic waves (e.g., Wallace, 1978).

3 Oceanic heat transport

The oceanic analog to moist static energy is simply the potential temperature, owing to several simplications associated with the thermodynamics of sea water (Warren, 1999). We thus present an estimate of the meridional mass transport within potential temperature ($\theta_o$) layers from a long simulation of the MIT General Circulation Model (GCM; Marshall et al., 1997). The model is a state of the art GCM forced by seasonally varying surface winds and buoyancy forcing. It was run on the ‘cubed sphere’ grid (Adcroft et al., 2004) at coarse resolution (15 vertical levels and $C32 \simeq 2.8^\circ \times 2.8^\circ$ in the horizontal). More information about the simulation can be found in the appendix.

From the last 50 years of a thousand-year integration, we have estimated a time mean meridional velocity ($v$) and potential temperature field, as well as a time mean bolus velocity ($v^*$) introduced by the Gent and McWilliams scheme (Gent and McWilliams, 1990). At each grid point, the water column is partitioned into a set of $\theta_o$–layers (with a resolution of $\delta \theta_o = 1 K$) and the meridional mass transport ($v + v^*$) within each layer is computed. Note that this procedure neglects time dependent effects which, in this simulation, are limited to the seasonal cycle (no interannual variability in the surface forcing). The error thereby introduced in the annual mean was estimated by comparing the ocean heat transport computed from time mean $v + v^*$ and
Figure 6: Various estimates of the mass streamfunction within $\theta_A$ layers (a) direct (b) using Eq.(6) (c) the Eulerian mean. Contour interval is $20Sv$ ($1\; Sv = 10^9 kgs^{-1}$). Grey curves as in Fig.4. See text for details.
\( \theta_o \) with that obtained from the time mean of the product \((v + v^*)\theta_o\). It was found to be negligible. Further details of the numerical procedure are discussed in the appendix.

Fig. 7 shows the mass streamfunction within \( \theta_o \)-layers for the global ocean, calculated from Eq.(4) in the same way as in the atmosphere. One observes prominent shallow cells at warm temperatures \((\theta_o = 15^\circ C - 30^\circ C)\), flanked by ‘cold’ \((\theta_o \leq 10^\circ C)\) deep cells at higher latitudes. The intensity of the warm and cold cells is comparable \((\simeq 30 \ Sv)\). The poleward flow in both cold and warm cells typically occur at a temperature which is larger than the zonally average surface temperature (shown by the thick median grey curve in Fig. 7 — the warmest and coldest surface temperature are also indicated as the upper and lower grey curves). This is the oceanic analog to the observation that in the atmosphere, equatorward flow occurs at a temperature lower than the surface owing to the effect of transient eddies (see Section 2). In the ocean, it is a consequence of the three dimensional character of the steady circulation depicted in Fig. 7, with advection in warm western boundary currents and their interior extension being a key aspect. As we will see below, \( \Psi_o \) differs from basin to basin. Overall, the features in Fig. 7 are significantly more complicated than its two-dimensional atmospheric counterpart in which there is a single overturning cell in each Hemisphere — see Fig.4.

Analysis of the mean thickness of the \( \theta_o \)-layers reveals (not shown) that the warm cells occupy a volume of fluid with high stratification (thin layers) and can be identified with the mass circulation within the ventilated thermoclines of the Southern and Northern Hemispheres. This is further confirmed by the fact that their poleward extension is well predicted by the position of
the midlatitude zero windstress curl lines (not shown).

The ‘cold cells’ are associated with thick $\theta_o$—layers (weak stratification) and correspond to the circulation of North Atlantic Deep Water and Antarctic Bottom Water (not shown). Note that the traditional Eulerian mass streamfunction displays more modest ($\simeq 10 - 15Sv$) cold cell mass transports.

Further information about $\Psi_o$ is provided in Fig.8, where the calculation was repeated separately for the Indo-Pacific (Fig.8a) and Atlantic (Fig.8b) basins. A simple partitioning is thereby obtained, in which the two warm (symmetric) cells originate in the main from the Indo-Pacific basin and the (asymmetric) Northern cold cell from the Atlantic basin.

4 The partitioning of heat transport

We now combine oceanic and atmospheric mass streamfunctions within their respective potential temperature layers. In doing so (Fig. 9), we have mapped the $y$—axis of Figs.4 and 7 on to an energy axis, which is $C_A\theta_A$ for the atmosphere and $C_O\theta_O$ for the ocean.

The first striking feature of Fig.9 is that, as anticipated in Section 2, the intensity of the oceanic cell is much weaker than its atmospheric counterpart (both are annual averages — see caption of Fig.9). Even at $20^\circ$, where $\Psi_o$ reaches its maximum, the atmospheric mass transport is roughly four times that of the ocean. It is only within the deep tropics that the two transports are comparable. This is further demonstrated in Fig.10 (continuous black curve) which shows the ratio of the maximum of $\Psi_o$ and that of $\Psi_A$, com-
Figure 7: Annual mean mass streamfunction within $\theta_o$ layers (contoured every 5 $Sv$, continuous when clockwise; zero contour omitted). The $x$–axis denotes latitude and the $y$–axis potential temperature $\theta_o$ (in degree Celsius). The middle grey curve is the zonally average surface $\theta_o$ while the upper and lower grey curves indicate the maximum and minimum surface $\theta_o$ as a function of latitude.
Figure 8: Same as Fig.7 but for (a) the Indo-Pacific and (b) Atlantic basins.
Figure 9: Annual mean atmospheric (black) and oceanic (grey) mass streamfunction within constant energy layers. The contour interval is 10 Sv, dashed when circulating anti-clockwise. The $y-$axis is an energy coordinate ($C\theta$) in units of $10^4 \text{J} \text{kg}^{-1}$. The oceanic cells are the same as shown in Fig.7 (annual mean) while the atmospheric cells are a rough annual estimate obtained by averaging Fig.4a and Fig.4b.
puted from Fig.9 at each latitude. Note that since both $\Psi_O$ and $\Psi_A$ become small close to the equator, the mass ratio shown in Fig.10 is rather noisy at low latitudes.

The second important feature seen in Fig. 9 is that the thickness of the cells in energy space are comparable. To estimate this more precisely, we have used Fig. 9 and computed the change in $C_O \theta_O$ and $C_A \theta_A$ (hereafter denoted as $C_O \Delta \theta_O$ and $C_A \Delta \theta_A$, respectively) across a reference contour of mass transport. The latter was chosen to be 10% of the overall maximum of $\Psi_O$ ($0.1 \times 32 \text{ Sv} \simeq 3 \text{ Sv} \text{ in the ocean}$) and $\Psi_A$ ($0.1 \times 143 \text{ Sv} \simeq 14 \text{ Sv} \text{ in the atmosphere}$). The ratio $C_O \Delta \theta_O / C_A \Delta \theta_A$ is plotted in Fig.10 (grey). As one approaches the tropics, where $\Delta \theta_A$ is small, the ratio diverges. Conversely, on moving to high latitudes where oceanic potential temperature variations are small (weak stratification), the ratio approaches zero. In midlatitudes, $C_O \Delta \theta_O / C_A \Delta \theta_A$ is about unity, the differences in heat capacity ($C_O / C_A \simeq 4$) being compensated for by a larger temperature difference across the atmospheric cell ($\Delta \theta_A \simeq 40 K$ compared to $\Delta \theta_O \simeq 10 K$).

The product of the continuous black ($\Psi$ ratio) and grey ($C \Delta \theta$ ratio) curves allows an estimate of the respective contribution of mass transport and energy contrast to the partitioning of heat transport. It is shown as the black dot-dashed line in Fig.10, and falls between the black and grey curves. Indeed, the energy contrast tends to favor the ocean as the major contributor to the total heat transport up to latitudes of $40^\circ - 50^\circ$. However, the mass transport strongly favors the atmosphere as the dominant contributor to the total heat transport at almost all latitudes. The net result of these two competing effects (dot-dashed curve) is consistent with the partitioning
Figure 10: The ratio $C_O \Delta \theta_O / C_A \Delta \theta_A$ (grey), $\max(\Psi_O) / \max(\Psi_A)$ (continuous black), and their product (dot-dashed), all computed from Fig. 9. A ratio of unity is indicated by the horizontal dashed line.
shown in Fig.1b, with the atmosphere dominating the total heat transport at latitudes poleward of about 20°.

Finally, we briefly comment on the interpretation of the $C \Delta \theta$ term as a measure of stratification. As can be seen from Fig.4, in both the tropics and midlatitudes, $C_A \Delta \theta_A$, the typical change in moist static energy across $\Psi_A$, is significantly different (twice as large) from the tropopause to ground $\theta_A$ difference. Similarly in the ocean, it is readily seen from Fig.7 that $C_O \Delta \theta_O$ for the warm cells is significantly different (by about 20—30%) than a (zonally averaged) surface to thermocline bottom temperature difference. One must be careful, then, when estimating the energy contrast across the oceanic and atmospheric circulations appropriate for computation of meridional energy transport.

5 Simple scalings

Our diagnostics suggest that the observed ratio $H_A/H_O$, which is larger (smaller) than unity poleward (equatorward) of roughly 20°, can be understood as a consequence of two simple limits of Eq.(2):

(i) dominance of atmospheric mass transport in mid-to-high latitudes ($\Psi_A \gg \Psi_O$ with $C_A \Delta \theta_A \sim C_O \Delta \theta_O$ and hence $H_A/H_O \gg 1$)

(ii) dominance of oceanic energy contrast in the tropics ($C_O \Delta \theta_O \gg C_A \Delta \theta_A$ with $\Psi_A \sim \Psi_O$ and hence $H_A/H_O \ll 1$).

The second limit — also invoked by Held (2001) — is probably the easiest to understand since it simply invokes very small energy contrasts within the
tropical atmosphere. This is a consequence of deep convection returning the atmosphere to a moist adiabatic lapse rate and dynamical adjustments forbidding significant horizontal potential temperature gradients within a few tens of degree off the equator. On the other hand, having larger midlatitudes atmospheric meridional mass transport is, at first sight, somewhat counter-intuitive when one thinks of the much larger density of the ocean compared to the atmosphere. We now attempt to rationalize limit (i) using simple dynamical scalings.

Our starting point is to view the mass circulations $\Psi_A$ and $\Psi_O$ as overturning cells in the meridional-height plane. This simplification is justified by Fig.6 for the atmosphere, in which the quasi-Stokes advection provides a good description of $\Psi_A$ in midlatitudes (see Section 2.2). It is probably less so in the ocean because of a more fundamental three dimensional structure of $\Psi_O$ (see Section 3). Nevertheless we will limit ourselves to this simple picture. For such an overturning circulation, a simple estimate of the mass transport is given by (for either ocean or atmosphere)

$$\Psi = V \times M \times L$$  \hspace{1cm} (7)

where $V$ (in $m s^{-1}$) is a typical meridional velocity, $M$ (in $kg m^{-2}$) is the mass carried per unit area and $L$ (in $m$) is the zonal length over which the circulation extends. All terms are estimated for either the poleward or the equatorward branch of the circulation. Since the ocean has a much greater density than the atmosphere, the $M$ term in Eq.(7) will always be larger for an oceanic compared to an atmospheric cell. Conversely, meridional velocities tend to be much larger in the atmosphere than in the ocean. Since the
zonal lengths $L$ are, however, of similar order in both ocean and atmosphere (they are exactly the same in the aquaplanet geometry considered in Section 6), there must be a strong compensation of ‘mass’ and ‘velocity’ effects in Eq.(7). This implies that a simple scaling based on the latter equation will be problematical since it must account for two large compensating terms.

To overcome this problem, we instead compare atmospheric and oceanic mass transports to the Ekman mass transport ($\Psi_{Ek}$), which is the limit case in which $M$ and $V$ effects in Eq.(7) exactly balance. This choice is motivated by the fact that the Ekman mass transport is a traditional oceanographic scale and that it is also a measure of the Eulerian mass transport in the atmosphere.

Consider first the atmosphere. In the quasi-geostrophic approximation, the zonally average (denoted by an overbar), steady, zonal momentum budget reads, in the transformed Eulerian mean form $^4$,

$$-f \nabla_{res} \frac{\Delta P}{g} I_x$$  \hspace{1cm} (8)

where $\nabla_{res}$ is the (residual) meridional velocity in the upper branch of the thermally direct circulation, $f$ the Coriolis parameter, and $\nabla_{qf}$ is the eddy flux of quasi-geostrophic potential vorticity. Using the definitions

$$\Psi_A = \nabla_{res} \frac{\Delta P}{g} I_x$$ \hspace{1cm} (9)

$$\Psi_{Ek} = \frac{\tau_x}{f} I_x$$ \hspace{1cm} (10)

where Eq.(7) has been used with $V = \nabla_{res}, L = I_x$ (the length of a latitude circle) and $M = \Delta P/g$ (related to the pressure thickness of the upper branch

$^4$this relationship is exact in isentropic coordinates in the limit of small Rossby number, but for simplicity we limit ourselves to quasi-geostrophic framework.
of the cell), and $\tau_x$ is the surface zonal windstress, Eq. (8) allows one to scale the ratio $\Psi_A/\Psi_{Ek}$ as

$$\frac{\Psi_A}{\Psi_{Ek}} = \frac{\overline{u'q'} \Delta P}{\tau_x g}. \quad (11)$$

This expression emphasizes that the atmospheric mass transport is eddy driven, increasing as eddy PV fluxes increase, this dependence being somewhat weakened by a likely simultaneous increase in surface windstress, the denominator in (11). Observations suggest that in midlatitudes the following scales are appropriate: a thickness $\Delta P \simeq 10^5 Pa$ (see Fig. 6), a stress $\tau_x \simeq 0.1 Pa$ and quasi-geostrophic PV fluxes $\overline{u'q'} \simeq 2.10^{-5} ms^{-2}$ (see Edmon et al., 1980)\(^5\), yielding a ratio $\Psi_A/\Psi_{Ek} \simeq 2$.

Another way to interpret this result is to recognize that (neglecting relative vorticity fluxes)

$$\overline{u'q'} = g \frac{\partial \tau_{eddy}}{\partial p}$$

where $\tau_{eddy}$ is an eddy stress defined by $\int \overline{u'q'}/N^2$ (see the discussion in Marshall and Radko, 2003). Thus, from Eq. (11)

$$\frac{\Psi_A}{\Psi_{Ek}} = \frac{\Delta P}{\tau_x} \frac{\partial \tau_{eddy}}{\partial p} \simeq \frac{\tau_{eddy}}{\tau_x}.$$  

Our previous scaling suggests that interior eddy stress are typically of the same order or larger than surface windstress.

Finally, one can interpret the ratio $\Psi_A/\Psi_{Ek} \geq 1$ as reflecting a larger thermally direct component than an indirect component (the so-called Ferrel

\(^5\)This estimate was obtained from Eliassen and Palm flux divergence maps averaged between winter and summer. We note that a very similar number is used if a downgradient PV flux ($\overline{u'q'} = -K\overline{\beta}$) is used with diffusivity $K = 10^6 m^2 s^{-1}$ and $\overline{\beta}$ taken as the planetary PV gradient at 45° of latitude.
cell), since the strength of the Ferrel cell roughly scales with the Ekman transport.

Let us now consider the ocean. For simplicity, let us further restrict ourselves to the case of wind-driven circulation. In a channel geometry like that of the Southern Ocean, various studies have suggested that the net meridional mass transport $\Psi_O$ is less than the Ekman mass transport $\Psi_{Ek}$ owing to the effect of mesoscale baroclinic eddies $\Psi_{Eddy}$ (see, for example, the model of Marshall and Radko (2003) and the recent review by Olbers et al., 2004):

$$|\Psi_O| = |\Psi_{Ek} + \Psi_{Eddy}| \leq |\Psi_{Ek}|$$  \hspace{1cm} (12)

Diagnostic studies suggest a considerable cancellation in Eq.(12) — see, for example, Karsten and Marshall (2002). Thus, for channel geometries, we expect

$$\frac{\Psi_A}{\Psi_O} = \frac{\Psi_A}{\Psi_{Ek}} \times \frac{\Psi_{Ek}}{\Psi_O} = \frac{\Psi_A}{\Psi_{Ek}} \frac{1}{1 - |\Psi_{Eddy}/\Psi_{Ek}|} \gg 1 \text{ (channel geometry)}$$  \hspace{1cm} (13)

if $\Psi_A/\Psi_{Ek} \approx 2$ (as estimated above) and invoking the near cancellation of $\Psi_{Eddy}$ and $\Psi_{Ek}$.

A more relevant scaling for a gyre geometry — i.e. for the ‘warm cells’ in Fig.7 — can be readily obtained by assuming that the mass they carry is simply that pumped down at the base of the Ekman layer. Accordingly, we write $\Psi_O \approx \rho W_{Ek} L_{x yre} L_{ygr}$ where $\rho$ is the density of seawater, $W_{Ek}$ the Ekman pumping velocity and $L_{x yre}, L_{ygr}$ are the zonal and meridional scales of the subtropical gyre, respectively. Assuming further that the meridional scale of the surface windstress is equal to $L_{ygr}$, we obtain

$$\frac{\Psi_A}{\Psi_O} = \frac{\Psi_A}{\Psi_{Ek}} \times \frac{\Psi_{Ek}}{\Psi_O} \approx \frac{\Psi_A}{\Psi_{Ek}} \times \frac{L_x}{L_{ygr}} \gg 1 \text{ (gyre geometry)}$$  \hspace{1cm} (14)
since $\Psi_A/\Psi_{Ek} \simeq 2$ and the length of a latitude circle $L_x$ is larger than the zonal scale $L_x^{\text{gyre}}$ of a subtropical gyre. Note that this scaling of the mass transport by the warm cells is essentially that of Klinger and Marotzke (2000).

Both Eqs. (14) and (13) predict a dominance of the atmospheric mass transport. Despite its much smaller density, the atmosphere is able to develop sufficiently strong meridional flows over large distances in midlatitudes that its mass transport considerably exceeds that of the ocean.

6 Partitioning in a coupled climate model

The scalings proposed in the previous section are rather fundamental and do not rely on a detailed representation of geometry and orography, etc. One thus expects that the heat transport partitioning seen in the observations should also be found in the presence of very different continental configurations. We briefly explore this possibility now by analyzing the partitioning of heat transport in a long simulation of a coupled model of intermediate complexity run in the absence of continents (aquaplanet geometry) but with an active dynamical ocean.

The coupled model consists of a 5-layer primitive equation atmosphere with simplified physical representations (Molteni, 2003), coupled to the MIT ocean model as described in Marshall et al (2004). The atmosphere and ocean are run on the same ‘cubed sphere’ grid (see Adcroft et al, 2004) at a horizontal resolution of C32 ($\simeq 2.8^\circ \times 2.8^\circ$). The coupled system also includes a simple thermodynamic sea-ice model. The results shown below are based
on the last 500 years of the simulation.

Fig.11 displays the contribution of the atmosphere (black) and the ocean (grey) to the total heat transport \( H_A/H \) and \( H_O/H \), respectively) in the aquaplanet coupled model in a format similar to Fig.1b. One observes (continuous curves) that again a simple partitioning is found, with the atmosphere dominating poleward of 20° and the ocean dominating equatorward of 20°. For reference, we have also indicated the corresponding curves from observations (see caption of Fig.11) as broken black and grey curves. As expected from the general arguments outlined in Section 5, the similarity between the two sets of curves is striking. Even in a world very different from our own, the partitioning of heat transport between the atmosphere and remains, to first order, the same.

7 Summary and conclusions

A diagnostic study of the partitioning of the total poleward heat transport between the ocean and atmosphere was presented. The analysis was conducted in a coordinate system consisting of latitude and potential temperature \( \theta \) (energy). In this coordinate system, the poleward heat transport by both the ocean or the atmosphere scales as \( H = \Psi C \Delta \theta \), where \( \Psi \) represents the meridional mass transport within \( \theta \)-layers (moist potential temperature for the atmosphere, potential temperature for the ocean) and \( C \Delta \theta \) scales the energy difference across the circulation defined by \( \Psi \). The diagnostic was applied to NCEP-NCAR reanalyses NCEP-NCAR reanalyses and a long simulation of an ocean model.
Figure 11: Same as Fig. 1b but for the coupled model in an aquaplanet geometry (continuous curves). As a reference, observed ratios are also indicated (broken curves). The latter were computed from the observations presented in Fig.1, averaging the NCEP and ECMWF oceanic and atmospheric heat transports.
Both the oceanic ($\Psi_O$) and atmospheric ($\Psi_A$) meridional mass transports display remarkable simplicity in energy coordinate. In the atmosphere, a single equator-to-pole cell appears in each hemisphere, a feature outlined in previous studies of the ‘residual circulation’, but which is even more pronounced in the moist potential temperature coordinate system employed here. The oceanic streamfunction shows a more distinct hemispheric asymmetry in both ‘warm’ wind-driven cells and ‘cold’ buoyancy driven cells.

Two simple limits emerge from our diagnostics. In mid-to-high latitudes, the intensity of the atmospheric meridional mass transport dominates that of the ocean ($\Psi_A \sim 100 \, Sv$ but $\Psi_O \leq 30 \, Sv$ where $1 \, Sv = 10^9 kgs^{-1}$). With such a dominant mass transport, the atmosphere is the main contributor to the total poleward heat transport ($H_A/H_O \sim \Psi_A/\Psi_O \gg 1$). In the tropics however, it is the energy contrast which sets the partitioning, and the ocean becomes the major contributor to the total poleward heat transport ($H_A/H_O \sim C_A \Delta \theta_A/C_O \Delta \theta_O \ll 1$).

We have argued that these two limits reflect fundamental, robust dynamics of the ocean-atmosphere system. The dominance of the atmospheric meridional mass transport in midlatitudes is a consequence of the efficiency of baroclinic eddies at driving a thermally direct circulation. The strength of the latter is more than twice that of the (indirect) Eulerian mean circulation (the Ferrell cell), which roughly scales the wind driven ocean meridional mass transport. The tendency of the atmosphere to follow a moist adiabat and display weak potential temperature gradients within a few tens of degrees off the equator is chiefly responsible for the second limit.

As a consequence of these basic ideas, we have further argued that the
observed partitioning is a robust feature of the Earth climate, unlikely to change significantly in a different configuration of continents, as has happened on geological timescales. Some support for this claim was obtained from a long simulation of an intermediate complexity climate model run in an aquaplanet geometry (no continents at all!).

The buoyancy driven component of the oceanic meridional mass transport — the thermohaline component — has been neglected from our simple scalings. Inspection of Fig.1a suggests, however, a north-south asymmetry in $H_O$ which is a likely signature of this component. Our main point here is that the thermohaline circulation only introduces relatively small changes in $H_O$ — of the order of a few tens of a PW in mid-to high latitudes — small compared to $H_A \simeq 5\ PW$. Note, however, that fluxes of such magnitude may be very important in maintaining the mean climate and in the dynamics of climate variability.

Combined with the study of Stone (1978), who argued that the total (ocean+atmosphere) heat transport is essentially set by the planetary albedo, the solar constant and the radius of the Earth, our results suggest that the oceanic and atmospheric heat transport might themselves change rather modestly in very different climate states. In other words, climate variability may be associated with only small departures from fixed, background, $H_A, H_O$ curves. These ideas will undoubtedly soon become testable as oceanic and atmospheric reanalyses products improve, and a vast hierarchy of coupled ocean- atmosphere models, in various geometries, become available for analysis.
Acknowledgements: This work was supported by a grant from NOAA’s Office of Global Programs as part of Atlantic CLIVAR. Jean-Michel Campin was of tremendous help in analyzing the result of the Ocean GCM.
Appendix: specific details of the mass transport calculation within $\theta$ classes

Atmosphere

Daily outputs from the NCEP-NCAR reanalysis (Kalnay et al., 1996) were used to compute moist potential temperature $\theta_A$ and meridional velocity $v$ at each nominal pressure level. First, on a given day, $\theta_A$ and $v$ were interpolated onto a finer, regular ($\delta P = 10 \, mb$) pressure grid, in order to provide a better resolution in temperature class. Layers with pressure greater than the surface pressure for that day (the latter also being taken from the NCEP-NCAR reanalysis) were excluded from the calculation since they physically would lie below the ground. The meridional mass transport within the finer grid column ($v\delta P/g$ for a given pressure layer, where $g$ is gravity) was subsequently repartitioned as a function of moist potential temperature class, with a resolution of $\delta \theta_A = 5K$. Finer and coarser resolutions were tested for the Southern Hemisphere winter period of 2003 and found to produce very similar results. We thus used this resolution for all calculations presented here, including meridional mass transport within dry potential temperature layers.

Ocean

We use yearly outputs from a thousand year long integration of the MIT Ocean GCM (Marshall et al., 1997) run on the ‘cube sphere’ in z-level coordinates (Adcroft et al., 2004) with a realistic geometry. The model was forced by a seasonally varying climatology of surface windstress and net surface heat flux and evaporation minus precipitation (the simulation is the same as that
discussed in Marshall et al., 2004). The model has no surface mixed layer but a simple convective adjustment scheme. Surface temperature and salinity are restored to observed climatology with a timescale of 2 months and 2 years, respectively. There is no sea-ice model, but the temperature is maintained above freezing. A vertical diffusivity of $K_v = 3 \times 10^{-5} m^2 s^{-1}$ was used for temperature and salinity. The Gent and McWilliams (1990) scheme is used together with an isopycnal diffusion (Redi tensor formalism) coefficient set to $K_\rho = 800 m^2 s^{-1}$.

The procedure to obtain the meridional mass transport within a set of $\theta_O$ (potential temperature) layers is almost identical to that used for the atmosphere (replacing the finer pressure grid by a finer geometric height grid with $\delta z = 10 m$). Only one change was made: since the ocean model predicts a transport within each layer, $v$ was kept the same when interpolating onto the finer $z$-grid (yielding a staircase-like profile, as in the raw model outputs). This procedure exactly preserves the (vertically integrated) meridional mass transport at each grid point, but only approximately conserves the (vertically integrated) $\theta_O$ transport at each grid point. The error in the latter was found to never exceed 1%, however.

Finally, we emphasize that in both the ocean or atmosphere, the mass transport streamfunction $\Psi$ computed from the meridional mass transport within $\theta$-layers according to Eq. (4) is different from the mass circulation within $\theta$-layers, i.e. the diabatic circulation (e.g., Townsend and Johnson, 1985). Estimating the latter would require the calculation of both meridional and upward mass flux within a set of $\theta$-layers, whereas we only considered the meridional mass flux. One must therefore be cautious when interpreting
the ‘cross’ $\theta$ flows in Figs. 4, 5, 7 and 8 as truly diabatic in origin.
References


