

The Partitioning of the Poleward Energy Transport between the Tropical Ocean and Atmosphere

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ABSTRACT

The mass transport in the shallow, wind-driven, overturning cells in the tropical oceans is constrained to be close to the mass transport in the atmospheric Hadley cell, assuming that zonally integrated wind stresses on land are relatively small. Therefore, the ratio of the poleward energy transport in low latitudes in the two media is determined by the ratio of the atmospheric gross static stability to that of the ocean. A qualitative discussion of the gross stability of each medium suggests that the resulting ratio of oceanic to atmospheric energy transport, averaged over the Hadley cell, is roughly equal to the ratio of the heat capacity of water to that of air at constant pressure, multiplied by the ratio of the moist- to the dry-adiabatic lapse rates near the surface. The ratio of oceanic to atmospheric energy transport should be larger than this value near the equator and smaller than this value near the poleward boundary of the Hadley cell.

1. Introduction

The radiative imbalance at the top of the atmosphere, the difference between the net downward shortwave flux and the outgoing infrared flux, is larger than 70 W m^{-2} near the equator when averaged over longitude and season (e.g., Hartmann 1994, Fig. 2.12). In the time mean, this net incoming flux is balanced by energy transport away from the equator, in the atmosphere, and in the oceans. It has proven difficult to determine with any precision how this horizontal energy flux is partitioned. In the presatellite era (e.g., Sellers 1965), the two media were thought to be of roughly equal importance for poleward heat transport out of the deep Tropics, but the surprisingly low albedo of the Tropics observed from space, and the upward reevaluation of the radiative imbalance in the Tropics, led to the picture of oceanic dominance close to the equator (von der Haar and Oort 1973; see also Peixoto and Oort 1992). Despite the remaining uncertainties, the basic picture appears to be robust; Trenberth and Solomon (1994), for example, find no more than $\approx 20 \text{ W m}^{-2}$ being transported out of the equatorial belt (10°N – 10°S) by the atmosphere, using analyses from the European Centre for Medium-Range Weather Forecasts.

In the tropical atmosphere, the ageostrophic mean meridional overturning in the Hadley cell plays the major

role in the horizontal energy transport, although eddy latent heat fluxes are not entirely negligible. In the tropical oceans, the meridional energy transport is evidently dominated by shallow, wind-driven, overturning cells, primarily in the Pacific, that are roughly symmetrical about the equator, rather than by the much deeper asymmetric thermohaline circulation concentrated in the Atlantic (e.g., Haidvogel and Bryan 1992).

In trying to understand the relative importance of the poleward energy transport in the ocean and atmosphere, a useful starting point is the observation that the oceanic mass transport is strongly coupled to the atmospheric mass transport, since the near-surface branches of both the atmospheric and oceanic overturnings can be thought of as the Ekman drift associated with the same surface stress. An estimate for the ratio of the energy transports that follows from this point is presented in the following, after which some consequences of this result for our understanding of tropical dynamics are outlined.

2. Mass transports

The Ekman transport is the mass transport in the boundary layer that produces a Coriolis force that balances the surface stress. Directly from this definition it follows that the oceanic Ekman transport is locally equal and opposite to the atmospheric Ekman transport. Land takes up a modest fraction of the zonal mean surface stress in the Tropics. In a GCM that we have recently examined, $\approx 70\%$ of the stress averaged from 20°N to 20°S is felt by the ocean (see also Peixoto and Oort

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1992, Fig. 11.11). To within this correction factor, the zonally averaged Ekman transport in the tropical oceans is equal in magnitude, and opposite in sign, to the zonally averaged Ekman transport in the atmosphere.

The mass transport in the surface branch of the atmospheric Hadley cell, and, therefore, in the cell as a whole, is well approximated by the Ekman transport. Given the modest size of the surface winds, nonlinear terms in the atmospheric momentum balance near the surface are typically small, and we can additionally assume that mountain torques are small, as are the frictional stresses above the planetary boundary layer. In addition, at the level of approximation at which we work here, there is no need to distinguish between the traditional Hadley cell and the transformed Eulerian mean transport, or the related mass transport within isentropic layers, as the eddy heat fluxes in the Tropics are small.

The oceanic side of the problem is potentially more complex. The zonally averaged meridional flow near the surface will contain the Ekman drift as well as a geostrophically balanced part. In midlatitudes a standard scaling argument (e.g., Pedlosky 1996, p. 14) implies that the geostrophically balanced flow will be larger in magnitude than the Ekman drift. However, the same argument also implies that this dominance disappears as one moves equatorward. Indeed, using a simple layered model of the wind-driven flow between the subtropics and the Tropics, Liu (1994) finds that the net effect of the geostrophic flow in the surface layer is opposite to that of the poleward Ekman drift, but its amplitude is only 15% of the Ekman drift over the bulk of the cell for parameters typical of the Pacific. Liu's model produces circulations that are similar to those found in oceanic general circulation models (GCMs) [Liu et al. (1994); see also McCreary and Lu (1994), and Lu et al. (1998)]. If there is any stress on the equator, Ekman drift loses its relevance for the strength of the overturning within a few degrees of the equator, but the resulting rapid and shallow recirculation, within what Liu et al. (1994) refer to as the "equatorial cell," transports little heat.

Recently, Klinger and Marotzke (2000, hereafter KM), have shown that a simple model of the heat transport by the wind-driven subtropical cells, based on the assumption that the mass transport is the Ekman transport at each latitude, agrees rather well with an idealized ocean GCM. Encouraged by this result, we make the identical assumption, that the mass transported in these shallow subtropical cells can be approximated by the Ekman transport. To the extent that the zonally averaged near-surface flow contains a large geostrophically balanced part, as in the Atlantic, we assume that this does not interact significantly with these shallow overturning mass transports.

Keeping these approximations in mind, the implication is that the zonally averaged mass transport in the wind-driven oceanic cells in the Tropics is close in magnitude to the mass transport in the Hadley cell. In the

annual mean, the atmospheric transport is roughly $6 \times 10^{10} \text{ kg s}^{-1} = 60 \text{ Sv}$ (Peixoto and Oort 1992, Fig. 7.19; using the Sverdrup as a unit of mass transport rather than volume transport, $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1} \times 10^3 \text{ kg m}^{-3} = 10^9 \text{ kg s}^{-1}$). Ignoring stresses on land, this would imply an equatorial upwelling mass flux of $\approx 120 \text{ Sv}$. Removing $\approx 30\%$ of this stress to correct for the presence of land leaves $\approx 85 \text{ Sv}$. Estimates of the Ekman transport in each ocean basin at 10°N and 10°S are provided by KM. They estimate 20 Sv flowing through these boundaries into the equatorial waters of the Atlantic, 58 Sv in the Pacific, and 13 in the Indian, for a total of 91 Sv. These estimates are admittedly sensitive to the latitude at which they are computed and to data inadequacies.

In attempting to reconcile atmospheric and oceanic observations, constraints imposed by estimates of the Hadley cell mass transport should be kept in mind; as briefly discussed in the concluding section of this note, this mass transport is itself constrained by factors that have little to do with uncertain drag coefficients or other details of the boundary layer.

It is the annual mean surface stress and associated mass transports that are of interest here, as we are concerned with the partitioning of the time-averaged poleward energy transport between atmosphere and ocean. The partitioning of the seasonally or interannually varying energy transports in the two media will generally be different.

3. Energy transports

If V is the time- and zonally averaged mass transport in the Hadley cell, the energy transport by the Hadley cell can be set equal to $S_A V$, where we refer to S_A , the ratio of energy transport to mass transport, as the gross atmospheric stability (Held and Suarez 1978; Neelin and Held 1987). One can also set the energy transport in the wind-driven oceanic overturning cells in the Tropics equal to $S_o \xi V$, where S_o is the gross oceanic stability and ξ is the fraction of the total stress that is exerted on the ocean. Combining the atmospheric and oceanic components, the total energy transport is $(S_o \xi + S_A)V = S_T V$, where S_T is the gross stability of the atmosphere-ocean system. It is the close relation between the mass transports in the two media that makes this more than a bookkeeping device. To the extent that the oceanic component dominates the gross stability, $S_o \xi > S_A$, it is the oceanic stability that determines the total Hadley cell energy transport, given the mass transport.

One can set S_o equal to the heat capacity of water per unit mass c_o multiplied by the characteristic temperature difference between the poleward flow in the mixed layer and the return flow beneath. Since the subsurface flow can be assumed to be adiabatic, and can be traced back to subduction in the subtropics, the temperature difference in the vertical should be determined by the meridional temperature difference at the surface

across the Tropics. The three-dimensional structure of the adiabatic part of this circulation can be quite complex, with water moving equatorward along the western boundary, and then flowing eastward in the equatorial undercurrent as it rises, but this need not prevent us from speaking of an effective value of S_o . It is the heat lost along the surface branch of this circulation, rather than the detailed path of the adiabatic return flow, that determines S_o .

It is useful to keep in mind the order of magnitude of the cooling resulting from given energy or mass transports in the tropical oceans. Focusing on the region between 10°N and 10°S , let \mathcal{M} be the sum of the mass transports through the northern and southern boundaries in Sv and let $c_o\delta T$ be the gross stability at these boundaries, with δT measured in degrees Kelvin. The resulting cooling averaged over this tropical belt (in W m^{-2}) will be $0.05\mathcal{M}\delta T$. In order to produce a cooling of 50 W m^{-2} , one requires, for example, the (uncomfortably) large value of 100 Sv (an average of 50 at each boundary) acting on a 10 K temperature difference.

In the tropical atmosphere, the gross stability is the difference in moist static energy between the poleward flow near the tropopause and the equatorward flow in the surface Ekman layer. The moist static energy is $h \equiv c_p T + gz + Lq$, where q is the specific humidity. Since there is little vapor in the flow aloft, the moist static energy aloft is essentially the dry static energy, $c_p T + gz$. We now make the simplifying assumption that the dry static energy aloft is a constant throughout the Tropics. Horizontal temperature variations are small within the free tropical troposphere, but we must also assume that the height of the upper-tropospheric branch of the tropical circulations is horizontally uniform. This appears to be an adequate empirical description of the zonal mean Hadley cell. The implication of these assumptions is that variations in the gross stability of the tropical atmosphere result from horizontal variations in the moist static energy of the boundary layer.

Our simplest picture of the tropical atmosphere is one in which the the dry stability near the tropopause is determined by the level of neutral buoyancy for the most buoyant parcels in the Tropics, those originating in regions where the boundary layer has the largest moist static energy, implying that $S_A \approx 0$ in these regions. Indeed, one can think of this constraint as determining the effective dry static energy of the poleward flow aloft. The gross moist stability, although small, is likely to be positive even in regions that are, on average, strongly convecting (Emanuel et al. 1994), but we shall assume that this value is negligible in the following discussion. (It would play a more important role in the analysis of models in which the ocean does not transport energy.) As one moves away from these convecting regions, S_A increases to the extent that the moist static energy in the surface branch decreases.

The gradient in moist static energy in the surface layer is produced both by temperature gradients and moisture

gradients. If \mathcal{H} is the relative humidity and q_s the saturation specific humidity at the surface pressure,

$$\delta h = c_p \delta T + L\delta(\mathcal{H}q_s) \quad (1)$$

or, if we can ignore variations in \mathcal{H} ,

$$\delta h \approx \left(c_p + L\mathcal{H} \frac{dq_s}{dT} \right) \delta T. \quad (2)$$

If $\mathcal{H} \approx 1$, this expression can be rewritten in terms of the ratio of the dry- to the moist-adiabatic lapse rate:

$$\frac{\Gamma_d}{\Gamma_m} = \frac{c_p + Ldq_s/dT}{c_p}, \quad (3)$$

$$S_A \approx c_p \frac{\Gamma_d}{\Gamma_m} \delta T, \quad (4)$$

where we now interpret δT as the horizontal temperature drop from the strongly convecting regions to the point of interest. The value of the ratio Γ_d/Γ_m is ≈ 3 averaged over the region occupied by the surface branch of the Hadley cell. Therefore, a 2 K temperature drop from the ITCZ, for example, results in a stability of $c_p \times 6$ K. A stability of this value at $\pm 10^\circ$ latitude and a mass transport on each side of the equator of 60 Sv would result in a flux divergence of 10 W m^{-2} within this tropical belt (cf. Peixoto and Oort 1992, p. 334).

This expression overestimates the gross stability gradient somewhat if, instead of saturation, we actually have, say, $\mathcal{H} \approx 0.8$. More importantly, it underestimates the stability gradient if the relative humidity decreases as one moves away from the warmest surface temperatures. A decrease in relative humidity from 0.85 to 0.8 would increase the stability by the same amount as an additional drop of ≈ 2 K with fixed \mathcal{H} .

We define y to be latitude normalized so that it ranges from zero at the equator to unity at the subtropical boundary of the Hadley cell. We suppose also, to keep the discussion concrete, that surface temperatures decrease quadratically away from the equator

$$T_s \approx T_s(0) - \Delta_T y^2, \quad (5)$$

where Δ_T is the total drop from the equator to the subtropical edge of the cell. Then the previous argument leads to the result that

$$S_A(y) \approx c_p \Delta_T \frac{\Gamma_d}{\Gamma_m} y^2. \quad (6)$$

The maximum gross stability in the atmosphere is realized at the subtropical edge of the circulation.

In contrast, within the ocean we expect the gross stability to be small in the subtropics, near the latitude of subduction, and to increase as one moves equatorward. Near the subtropical subduction zone, the temperature of the poleward and equatorward moving waters should be similar. As one moves toward the equator, the temperature of the equatorward branch has not changed, assuming that the interior flow is adiabatic, but the sur-

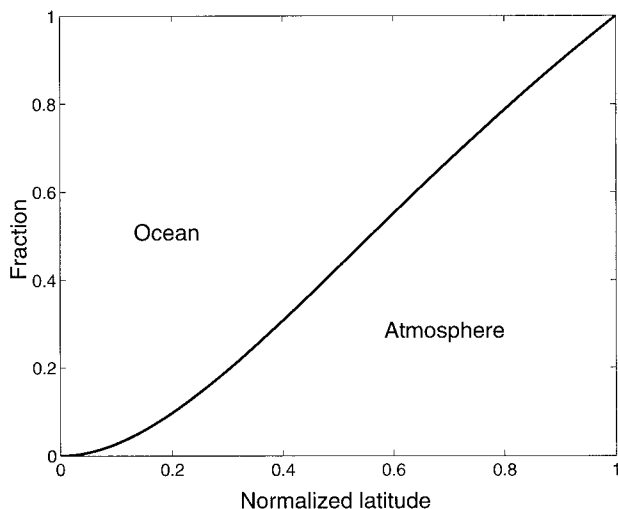


FIG. 1. A plot of the fraction of the poleward energy transport contributed by the atmosphere and by the ocean, assuming that μ defined by (11) is equal to unity. The ordinate is latitude normalized by the latitude of the subtropical boundary of the Hadley cell.

face branch is warmer. Following KM, we assume explicitly that the return flow at depth at a given latitude consists of all of the water subducted poleward of this latitude, with each water mass carrying the temperature of the surface water at the latitude of subduction. The gross stability is then given by

$$\frac{1}{c_o} S_o(y) = T(y) + \frac{1}{V(y)} \int_y^1 \frac{\partial V}{\partial \eta} T(\eta) d\eta, \quad (7)$$

where the first term on the rhs accounts for the surface poleward flow and the second term for the subsurface equatorward flow. Using an integration by parts, and the fact that $V(1) = 0$ by definition, this expression can also be rewritten in the form used by KM

$$\frac{1}{c_o} S_o(y) = \frac{1}{V(y)} \int_{T(1)}^{T(y)} V dT. \quad (8)$$

To proceed we need to prescribe a latitudinal distribution for $V(y)$. A very simple model results from the assumption that the subsidence is uniform throughout the tropical atmosphere, except for a δ -function ITCZ at $y = 0$. This implies that $V(y) \propto 1 - y$. We assume further that the ocean temperatures are the same as those used in the computation of the atmospheric transport, and given by the parabolic expression in Eq. (5). A bit of algebra then provides the estimate

$$\frac{1}{c_o} S_o(y) = \Delta_r \frac{1 + y - 2y^2}{3}. \quad (9)$$

As a consequence of these various approximations, we now have that the energy transport by the atmosphere and oceans is

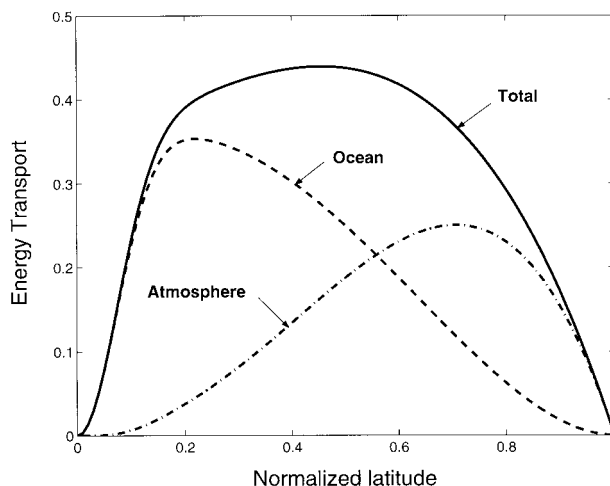


FIG. 2. The predicted shape of the energy transports by ocean, atmosphere, and ocean plus atmosphere, assuming again that μ is equal to unity, and with the shape of the mass transport given by (14). The energy flux is normalized by $\xi c_o V_o \Delta_r$.

$$\begin{aligned} & V(y)(S_A + \xi S_o) \\ &= V(y) \Delta_r \left(c_p \frac{\Gamma_d}{\Gamma_m} y^2 + c_o \xi \frac{1 + y - 2y^2}{3} \right). \end{aligned} \quad (10)$$

It is convenient to define the nondimensional number

$$\mu \equiv \frac{1}{\xi} \frac{c_p}{c_o} \frac{\Gamma_d}{\Gamma_m}. \quad (11)$$

Since $c_o \approx 4c_p$ and $\Gamma_d \approx 3\Gamma_m$, and given the estimate $\xi \approx 0.7$, one has $\mu \approx 1$. The fraction of the total transport carried by the atmosphere at latitude y is

$$\frac{3\mu y^2}{1 + y + (3\mu - 2)y^2} \quad (12)$$

or, assuming that $\mu \approx 1$,

$$\approx \frac{3y^2}{1 + y + y^2}. \quad (13)$$

A plot of this function is provided in Fig. 1. At $y = 1/3$ (corresponding to $\approx 10^\circ$ latitude), the prediction is that the atmosphere transports $3/13 = 23\%$ and the oceans $10/13 = 77\%$ of the energy.

If we plot the energy fluxes themselves, assuming that $V(y) \propto 1 - y$, the divergence of the energy transport in the oceans will have a δ -function at $y = 0$. We can choose to let the mass flux pass through zero smoothly at the origin by setting, for example,

$$V(y) = V_o(1 - e^{-(y/y_o)^2})(1 - y). \quad (14)$$

The result shown in Fig. 2 is drawn with $y_o = 0.1$. As y_o decreases, the width of the region of rapid variation in the oceanic flux near the equator contracts.

4. Implications

This analysis implies that it is difficult to change the partitioning of the poleward energy transport between atmosphere and ocean, as it depends only on physical constants of the system and, rather weakly, on the tropical surface temperatures through the moist-adiabatic lapse rate. On this basis, we should always expect the oceans to dominate the poleward energy flux out of the deep Tropics, but we should also expect the atmosphere to catch up as one approaches the latitude of the maximum Hadley cell energy flux.

The fact that the mass transports in the two media are coupled implies that it is more difficult to change the atmospheric mass transport in a coupled model than in an atmosphere-only model in which the air-sea fluxes are fixed. Suppose that in an ice age climate the extraction of energy from the subtropics by mid-latitude eddies increases. The total Hadley atmospheric + oceanic energy transport must then increase to cool the Tropics and warm the subtropics, so as to avoid a temperature contrast that would imply physically impossible vertical shears in the zonal flow. But any increase in the strength of the atmospheric Hadley cell mass transport is accompanied by an increase in the oceanic counterpart of roughly the same amount. The change in the Hadley cell would be weaker than in an atmosphere-only model, by the ratio $S_A/(S_A + \xi S_O)$ if we can assume a given energy flux requirement, and if we can assume that the gross stabilities do not change. One can equivalently think of an increase in equatorial upwelling as discouraging a more substantial increase in Hadley cell strength. In the ice age scenario, cooler subtropics would likely also increase Δ_T and, therefore, both S_O and S_A , which would further discourage an increase in the strength of the overturning.

There are several ways in which this argument could break down and allow the tropical atmosphere and ocean energy transports to decouple. Changes in stresses on land could play a significant role, thereby decoupling the Ekman mass transports in the two media. The geostrophic component of the surface branch of the mean oceanic overturning could change, thereby decoupling the mass fluxes even if the Ekman fluxes are identical. Finally, even if the mass transports are tightly coupled, the assumption that atmosphere and ocean are both characterized by the same horizontal temperature contrast could break down, given that the subduction in the ocean can be localized, and can occur at temperatures that are not representative of the atmosphere at that latitude. In particular, if the time spent by the water in the surface branch of the oceanic overturning is insufficient for the mixed layer to equilibrate with the atmospheric temperature gradient, one can expect that Δ_T (ocean) will be smaller than Δ_T (atmosphere). The argument above would then overestimate the oceanic contribution to the total flux. It will be of interest to see if the climate

system can take advantage of this degree of freedom when adjusting to changing forcing. It may also be that the atmosphere can free itself from the constraint that its gross stability is simply tied to Δ_T by modifying the gradient of relative humidity. Yet another possibility is that eddy latent heat fluxes can grow in importance in altered climates. Given these numerous potential loopholes, it is best to think of these considerations as providing a framework for discussion and analysis rather than as a convincing theory for the partitioning of the flux.

One can argue that there are more or less independent dynamical constraints on energy transport and mass transport by the atmosphere-ocean system in the Tropics. If the subtropics are essentially devoid of precipitation, it must be the case that the infrared cooling rate in the atmosphere in the subtropics determines the subsidence rate and, therefore, the mass flux in the atmospheric, and oceanic(!), Hadley cells. Energy transports, on the other hand, are also constrained by the need to maintain very small free tropospheric temperature gradients throughout the Tropics. It is likely that the gross stability of the atmosphere-ocean system must adjust so as to satisfy these two constraints simultaneously, much as the atmospheric gross moist stability adjusts in the idealized atmosphere-only models of Satoh (1994) and Fang and Tung (1996). If the atmospheric and oceanic stabilities are both strongly tied to the same temperature gradient, then it is this gradient, Δ_T , that must adjust. If the oceanic or atmospheric stabilities can free themselves from this constraint, a greater variety of responses are possible.

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