

The Effect of Cloud Cover on the Meridional Heat Transport: Lessons from Variable Rotation Experiments

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ABSTRACT

The question “What determines the meridional heat transport (MHT)?” is explored by performing a series of rotation-rate experiments with an aquaplanet GCM coupled to a slab ocean. The change of meridional heat transport with rotation rate falls into two regimes: in a “slow rotating” regime (rotation rate $< 1/2$ modern rotation) MHT decreases with increasing rotation rate, whereas in a “fast rotating” regime (rotation rate $\geq 1/2$ modern rotation) MHT is nearly invariant. The two-regime feature of MHT is primarily related to the reduction in tropical clouds and increase in tropical temperature that are associated with the narrowing and weakening of the Hadley cell with increasing rotation rate. In the slow-rotating regime, the Hadley cell contracts and weakens as rotation rate is increased; the resulting warming causes a local increase in outgoing longwave radiation (OLR), which consequently decreases MHT. In the fast-rotating regime, the Hadley cell continues to contract as rotation rate is increased, resulting in a decrease in tropical and subtropical clouds that increases the local absorbed shortwave radiation (ASR) by an amount that almost exactly compensates the local increases in OLR. In the fast-rotating regime, the model heat transport is approximately diffusive, with an effective eddy diffusivity that is consistent with eddy mixing-length theory. The effective eddy diffusivity decreases with increasing rotation rate. However, this decrease is nearly offset by a strong increase in the meridional gradient of moist static energy and hence results in a near-constancy of MHT. The results herein extend previous work on the MHT by highlighting that the spatial patterns of clouds and the factors that influence them are leading controls on MHT.

1. Introduction

The atmosphere–ocean system transports energy poleward, balancing the energy surplus in the tropics with the deficit in the extratropics. In the modern climate, the net annual-mean meridional heat transport (MHT) peaks at 35° latitude in both hemispheres, with atmospheric transport contributing about 80% in the Northern Hemisphere and about 90% in the Southern Hemisphere (Trenberth and Caron 2001; Fasullo and Trenberth 2008). Numerical modeling studies show that the MHT tends to stay remarkably invariant even for very different climates, including for Last Glacial Maximum boundary conditions and modern boundary conditions with quadrupled CO_2 (Manabe and Bryan 1985; Donohoe and Battisti 2012), and for very different

ocean heat transport (Magnusdottir and Saravanan 1999; Rose and Ferreira 2013).

Based on the one-dimensional energy-balance equation, Stone (1978) argued that the magnitude of the annual mean total MHT is insensitive to the details of dynamics of the atmosphere–ocean system, and this insensitivity is due to 1) the high efficiency of the dynamical transport mechanisms, 2) the negative correlation between the local planetary albedo and the outgoing radiation to space in the extratropics, and 3) a robust structure in the large-scale meridional profiles of absorbed shortwave radiation (ASR) and outgoing longwave radiation (OLR). As a caveat, Stone also noted that “the precise cancellation of the structure terms may not hold if the structure and dynamics of the atmosphere–ocean system change from those of current state” (Stone 1978, p. 133).

One obvious way to test Stone’s (1978) argument and explore the mechanisms that control MHT is to perform

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numerical experiments that alter the atmospheric dynamics, while keeping the overall geometry fixed. A straightforward way to achieve this is by altering Earth's rotation rate. Williams (1988) performed some early simulations with different rotation rates but did not focus on MHT. Vallis and Farneti (2009) studied the change of MHT with Earth's rotation rate and found that MHT decreases with increasing rotation rate. However, the study used a gray atmosphere (i.e., no clouds and constant longwave emissivity; Frierson et al. 2006) and therefore omits the potentially important impact that cloud changes might have on the global energy budget (e.g., Trenberth et al. 2009).

In this paper, we explore the question "What determines the meridional heat transport?" by modifying Earth's rotation rate and allowing for the concomitant change in clouds and water vapor distribution that we hypothesize are fundamental to determining MHT. This is done by performing experiments with an aquaplanet atmospheric general circulation model coupled to a slab ocean, changing the planet's rotation rate between 1/16 and 4 times the present-day value. We find that in this range of rotation rates, and for our model, the change of MHT with rotation rate falls into two regimes: a slow-rotating regime, in which MHT decreases with increasing rotation, and a fast-rotating regime, in which MHT stays relatively invariant. But to a large extent, MHT stays relatively constant with increasing rotation: the maximum change is about 30% of the ensemble mean MHT. The constancy of MHT in the fast-rotating regime is not a result of a constancy in ASR or OLR with rotation rate, as assumed by Stone (1978); on the contrary, there are large changes in the meridional structure of both ASR and OLR. However, the changes almost completely offset each other, so MHT remains approximately constant. The top-of-the-atmosphere radiation patterns are associated with the changing width and strength of the Hadley cell, and are by themselves sufficient to determine the response of MHT to changing rotation rate. We also evaluate how eddies adjust in order to achieve the MHT in each experiment, and whether the eddies' activities scale with metrics such as the Rhines length that depend on rotation rate.

2. Experiments and methods

We use the Geophysical Fluid Dynamics Laboratory Atmospheric Model, version 2.1 (GFDL AM2.1; Anderson et al. 2004) in aquaplanet configuration coupled to a 2.4-m slab ocean without sea ice component. The atmosphere model includes a prognostic cloud scheme in which cloud microphysics are parameterized according to Rotstayn (1997) and cloud fraction is parameterized following Tiedtke (1993) [see Anderson

et al. (2004) for more details]. The model is run at a horizontal resolution of 2° latitude $\times$ 2.5° longitude with 24 levels. Insolation is set to its annual-mean value at each latitude. We prescribe in all experiments a zonally and hemispherically averaged ocean heat flux (Q flux) to the slab ocean as in Rose and Ferreira (2013), which features a maximum poleward transport of 2.2 PW at 15° latitude and a zero global mean. Seven experiments are performed, in which the rotation rate is set to 1/16, 1/8, 1/4, 1/2, 1, 2, and 4 times the present-day value, respectively. Results of each experiment are presented in terms of the relative rotation rate Ω , the rotation rate divided by Ω_E , where $\Omega_E = 7.292 \times 10^{-5} \text{ rad s}^{-1}$ is the present terrestrial value.

3. Results

Figure 1 shows the meridional profile of surface temperature. As the rotation is increased, the surface temperature increases equatorward of 40° latitude and stays constant or even decreases in the polar regions (except as Ω changes from 1/16 to 1/8), resulting in an increase in the equator-to-pole temperature gradient. This feature is also seen in previous studies using different models (Williams and Holloway 1982; Geisler et al. 1983; Del Genio and Suozzo 1987; Williams 1988; Jenkins 1996; Navarra and Boccaletti 2002), indicating that it is a robust feature across models. That the tropics warm as the rotation rate is increased is often attributed to a reduction in the poleward energy transport. However, additional experiments (described in section 6) show that tropical warming is due to increasing absorbed solar radiation via reduction in tropical clouds; the warming is further amplified by the water vapor feedback.

Consistent with the surface temperature, the meridional gradient of surface moist static energy also increases with increasing rotation (Fig. 2). Note that the degree of increase in the surface moist static energy gradient is much greater than that in the surface temperature gradient because of exponentially larger moisture loading in the tropics as a result of the Clausius–Clapeyron relationship. As will be discussed in section 5, this strong increase in the gradient of surface moist static energy compensates the weakening of the eddies and keeps the MHT relatively unchanged.

In the control experiment ($\Omega = 1$), the Hadley cell extends from the equator to 30° latitude, approximating well the observed Hadley cell (Fig. 2) (Dima and Wallace 2003). As rotation increases, the Hadley cell becomes narrower and weaker (Fig. 2). The decrease in the width of the Hadley cell with increasing rotation is expected from the heuristic Held–Hou model (Held and Hou 1980; Held 2000).

The narrowing and weakening of the Hadley cell is also associated with a weakening of the midlatitude eddies

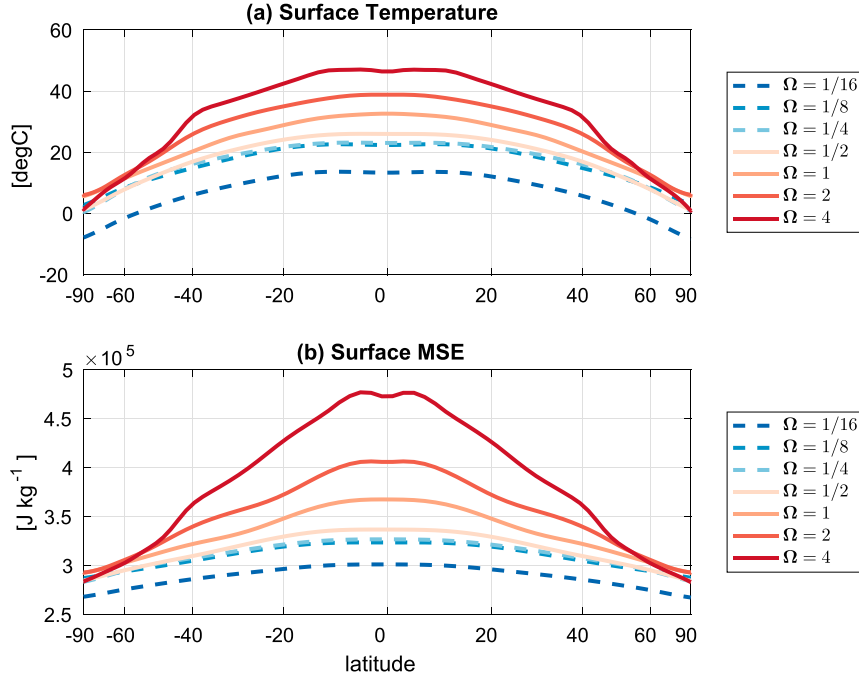


FIG. 1. (a) Surface temperature ($^{\circ}\text{C}$) for various values of Ω . Note that the displacement of latitude is area weighted; hence, distance along the abscissa is linearly proportional to area covered. (b) As in (a), but for surface moist static energy (J kg^{-1}).

because the midlatitude eddy momentum fluxes contribute significantly to the width and strength of Hadley cell: the strength of the meridional overturning circulation is proportional to the divergence of the eddy momentum flux (Schneider 1984; Hess et al. 1993; Walker and Schneider 2006; Vallis 2006). Figure 3 shows the eddy momentum flux¹ at 300 hPa. Averaged over the width of the Hadley cell, the eddy momentum flux is a positive contribution to the strength of the Hadley cell for all but the $\Omega = 1/16$ case (Table 1). A measure of the relative importance of the eddy momentum fluxes to the strength of the Hadley circulation is given by the ratio of the average of the eddy momentum flux divergence over the Hadley cell to the Hadley cell strength, and referred to as the “eddy efficiency.” Both the absolute amplitude of the eddy momentum flux divergence and the eddy efficiency are near extrema for the modern-day rotation rate and greatly decay for both higher and lower rotation rates.

The maximum poleward MHT in the $\Omega = 1$ experiment using AM2.1 is 5.4 PW at 35° latitude (Fig. 4a), which is indistinguishable from that observed. The latitude of the maximum MHT increases somewhat with

rotation rate: from about 25° for $\Omega = 1/16$ to 38° for $\Omega = 4$. With increasing rotation rate, MHT falls into two regimes (Fig. 4a): a “slow rotating” regime, $1/16 \leq \Omega < 1/2$, in which MHT decreases slightly with increasing rotation, and a “fast rotating” regime, $\Omega \geq 1/2$, in which MHT stays relatively unchanged with increasing rotation. This two-regime feature of MHT is evident for both maximum MHT (MHT_{max}) (Fig. 4b) and the MHT averaged between 20° and 60° latitude (Fig. 4c). In the following analysis, we will be using MHT_{max} as a metric of MHT to understand its change with rotation.

4. Understanding changes in MHT in terms of radiation

Our analysis of the changes in the MHT_{max} uses the diagnostics developed in Donohoe and Battisti (2012). In an equilibrium climate, MHT_{max} is equal to the net radiative surplus integrated over the tropics or, equivalently, the net radiative deficit integrated over the extratropics:

$$\text{MHT}_{\text{max}} = 2\pi R^2 \int_0^{x(\text{ASR}=\text{OLR})} (\text{ASR} - \text{OLR}) dx \quad (1a)$$

$$= -2\pi R^2 \int_{x(\text{ASR}=\text{OLR})}^1 (\text{ASR} - \text{OLR}) dx, \quad (1b)$$

¹ The eddy momentum flux is defined as the zonal mean time mean of the product of u' and v' , where u and v are the zonal and meridional wind components and prime denotes the deviation from the zonal and time mean.

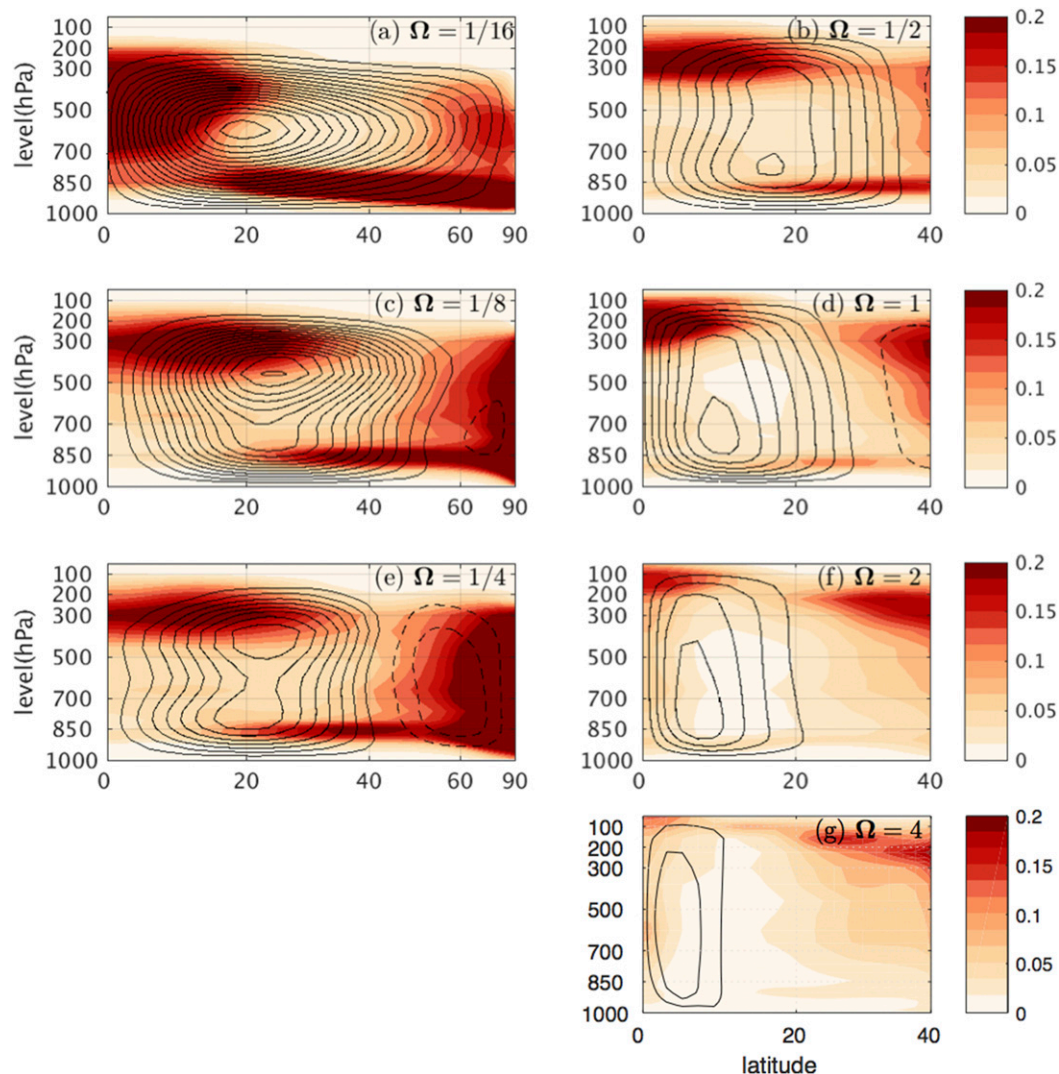


FIG. 2. Cloud fraction (shading) overlaid with meridional streamfunction (contours) for each experiment; contours start from $\pm 20 \times 10^9 \text{ kg s}^{-1}$, with interval of $20 \times 10^9 \text{ kg s}^{-1}$ for all experiments except for $\Omega = 1/16$ (contours start from $\pm 40 \times 10^9 \text{ kg s}^{-1}$, with interval $40 \times 10^9 \text{ kg s}^{-1}$). Note the displacement of latitude is area weighted. The latitude range in (b), (d), (f), and (g) is 0° – 40° latitude because in the fast-rotating regime ($\Omega \geq 1/2$) clouds in the high latitudes are saturated and so changes do not affect the energy balance (see Figs. 8b,d).

where x is the sine of latitude (Fig. 5). Since an equilibrium climate achieves global radiative equilibrium (i.e., the globally integrated ASR is equal to that of OLR), we can subtract the global average of ASR and OLR from the right-hand side to yield

$$\text{MHT}_{\max} = 2\pi R^2 \int_0^{x(\text{ASR}'=\text{OLR}'=0)} (\text{ASR}' - \text{OLR}') dx \quad (2a)$$

$$= -2\pi R^2 \int_{x(\text{ASR}'=\text{OLR}'=0)}^1 (\text{ASR}' - \text{OLR}') dx, \quad (2b)$$

where primes denote deviations from the global average and it has been assumed that $x(\text{ASR}' = 0) = x(\text{OLR}' = 0)$. Equations (2a) and (2b) can be combined to give

$$\begin{aligned} \text{MHT}_{\max} = & \frac{1}{2} \left[2\pi R^2 \int_0^{x(\text{ASR}'=\text{OLR}'=0)} (\text{ASR}' - \text{OLR}') dx \right. \\ & \left. - 2\pi R^2 \int_{x(\text{ASR}'=\text{OLR}'=0)}^1 (\text{ASR}' - \text{OLR}') dx \right] \quad (3a) \\ \approx & \text{ASR}^* - \text{OLR}^*, \quad (3b) \end{aligned}$$

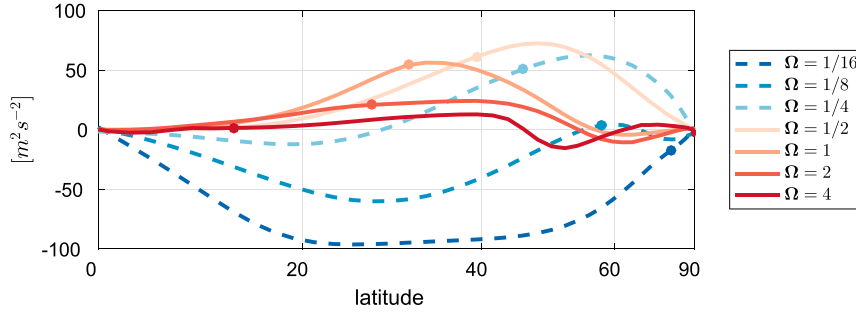


FIG. 3. The eddy momentum transport measured by the zonal and temporal mean of $u'v'$ at 300 hPa, where u' and v' are the deviation of the zonal wind and meridional wind from their time and zonal mean, respectively ($\text{m}^2 \text{s}^{-2}$) for various values of Ω . The dot on each line that is of the same color as the line it resides on denotes the poleward edge of the Hadley cell (i.e., the width of the Hadley cell). Note that the displacement of latitude is area weighted.

where

ASR*

$$\equiv \frac{1}{2} \left[2\pi R^2 \int_0^{x(\text{ASR}'=0)} \text{ASR}' dx - 2\pi R^2 \int_{x(\text{ASR}'=0)}^1 \text{ASR}' dx \right] \quad (4)$$

and

OLR*

$$\equiv \frac{1}{2} \left[2\pi R^2 \int_0^{x(\text{OLR}'=0)} \text{OLR}' dx - 2\pi R^2 \int_{x(\text{OLR}'=0)}^1 \text{OLR}' dx \right]. \quad (5)$$

Note that ASR* is the surplus of ASR in the tropics relative to the global mean, and OLR* is the surplus of OLR relative to the same global mean (or equivalently, ASR* and OLR* are the deficit of ASR and the deficit of OLR in the extratropics relative to the global mean, respectively). Thus, Eq. (3b) is a compact representation of the relationship that must exist between the poleward energy transport and the spatial patterns of ASR and OLR. The terms ASR* and OLR* are fundamental to the energetic gradients that must be accommodated by transport. A graphical illustration of the meaning of ASR* and OLR* is provided in Fig. 5 for the observed climate. The decomposition into ASR* and OLR* provides insight into processes acting via shortwave and longwave fluxes. Changes in the MHT_{max} from one experiment to another can thus be understood in terms of changes in ASR* and/or OLR*. In the following text, we will use this method to analyze the different behaviors of MHT as a function of rotation in each regime.

It is worth noting that the near equality in Eq. (3b) holds exactly when the meridional nodes of ASR' and

OLR' are collocated, which is true for the fast-rotating regime ($\Omega \geq 1/2$) but not for the slow-rotating regime ($\Omega < 1/2$). Nonetheless, even for the slow-rotating regime, this method gives a fair estimate of the change of MHT_{max} with rotation rate: ASR* – OLR* underestimates MHT_{max} by only 3%–16%, depending on rotation rate (cf. Figs. 4b and 6c). The results shown below are not sensitive to the small changes in the latitude of the nodal points associated with the changes in rotation rate: the same results are found when the nodal point from the $\Omega = 1$ experiment is used to estimate ASR* and OLR* for each of the experiments.

TABLE 1. The Hadley cell strength vs the mean divergence of the eddy momentum flux. As shown below, the width of the Hadley cell L_H is determined as the first latitude poleward of the maximum absolute value of the Hadley cell streamfunction at which the mass flux streamfunction at the sigma level of its extremum above sigma level = 0.7 is 10% of its extremal value, following Walker and Schneider (2006). The Hadley cell strength is defined as the maximum absolute value of the mass flux streamfunction. The mean divergence of the eddy momentum flux (fourth column) is the meridional average of the divergence of eddy momentum flux from the equator to L_H , that is, $\int_0^{L_H} [d(u'v')/d\theta] \cos\theta d\theta / \int_0^{L_H} \cos\theta d\theta$, where $u'v'$ is the eddy momentum flux, and θ is latitude. Eddy efficiency is defined as the mean divergence of the eddy momentum flux divided by the strength of the Hadley cell.

Ω	L_H (lat)	Hadley cell strength ($10^{11} \text{ kg s}^{-1}$)	Mean divergence of eddy momentum flux	
			(10^{-5} m s^{-2})	Eddy efficiency ($10^{-16} \text{ m s}^{-1} \text{ kg}^{-1}$)
1/16	74°	7.4	−0.22	0.005
1/8	58°	3.2	0.047	0.02
1/4	46°	1.8	1.1	0.57
1/2	39°	1.43	1.4	0.97
1	31°	1.41	1.5	1.03
2	27°	1.0	0.68	0.59
4	13°	0.6	0.084	0.54

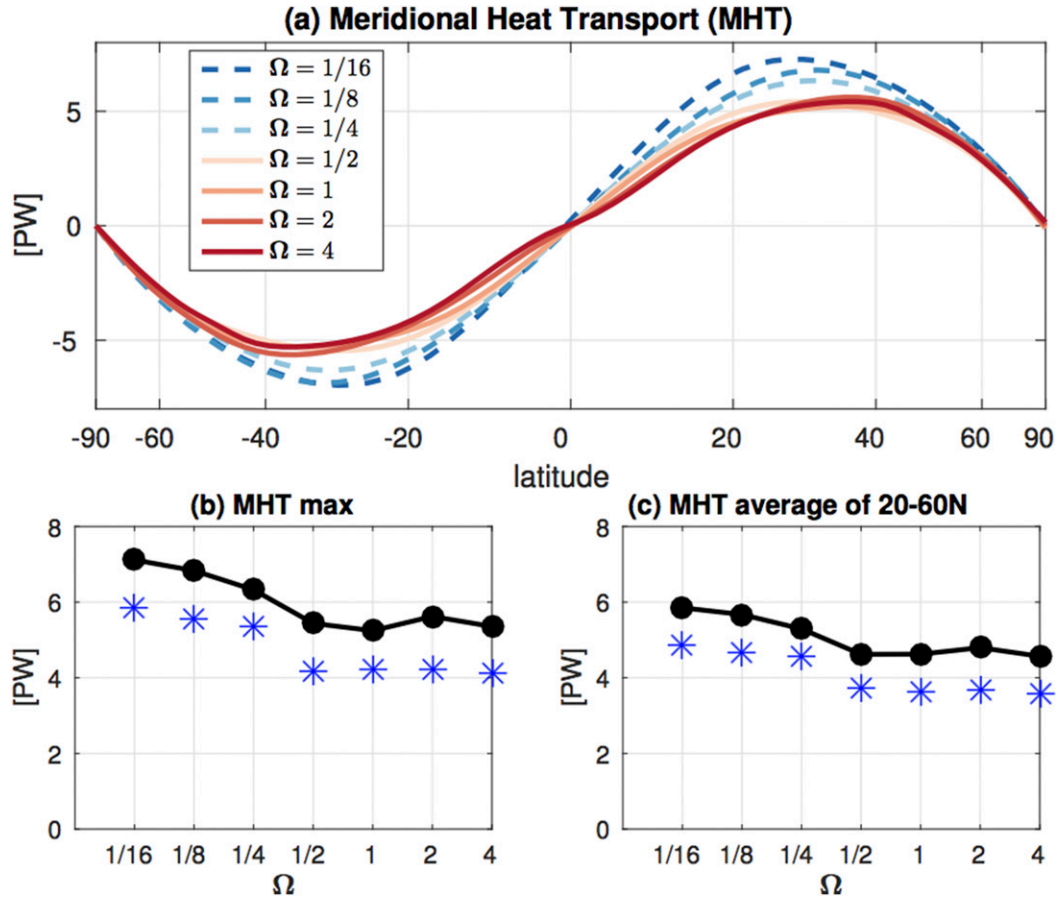


FIG. 4. (a) The total meridional energy transport (PW) for various values of Ω . Note the displacement of latitude is area weighted. The peak values of (b) the meridional energy transport and (c) the transport averaged between 20° and 60° latitude. The uncertainty in MHT_{max} is typically ± 0.2 PW. Blue asterisks denote the experiments without Q flux.

a. The fast-rotating regime

In the fast-rotating regime ($\Omega \geq 1/2$), both OLR^* and ASR^* increase with rotation rate and they increase by approximately equal amounts (Figs. 6a,b); as a result, MHT_{max} is largely insensitive to rotation rate.

To understand why both ASR^* and OLR^* increase with increasing rotation rate, we further partition ASR^* and OLR^* into the contribution from the clear-sky ($\text{ASR}_{\text{clr}}^*$ and $\text{OLR}_{\text{clr}}^*$) and the contribution from the clouds ($\text{ASR}_{\text{cld}}^*$ and $\text{OLR}_{\text{cld}}^*$). Similar to ASR^* and OLR^* [cf. Eqs. (4) and (5)], $\text{ASR}_{\text{clr}}^*$ and $\text{OLR}_{\text{clr}}^*$ are defined to be

$$\text{ASR}_{\text{clr}}^* = \frac{1}{2} \left[2\pi R^2 \int_0^{x(\text{ASR}'_{\text{clr}}=0)} \text{ASR}'_{\text{clr}} dx - 2\pi R^2 \int_{x(\text{ASR}'_{\text{clr}}=0)}^1 \text{ASR}'_{\text{clr}} dx \right] \quad (6)$$

and

$$\text{OLR}_{\text{clr}}^* = \frac{1}{2} \left[2\pi R^2 \int_0^{x(\text{OLR}'_{\text{clr}}=0)} \text{OLR}'_{\text{clr}} dx - 2\pi R^2 \int_{x(\text{OLR}'_{\text{clr}}=0)}^1 \text{OLR}'_{\text{clr}} dx \right]. \quad (7)$$

The terms $\text{ASR}_{\text{cld}}^*$ and $\text{OLR}_{\text{cld}}^*$ are given by

$$\text{ASR}_{\text{cld}}^* = \text{ASR}^* - \text{ASR}_{\text{clr}}^* \quad (8)$$

and

$$\text{OLR}_{\text{cld}}^* = \text{OLR}^* - \text{OLR}_{\text{clr}}^*. \quad (9)$$

Thus $\text{ASR}_{\text{clr}}^*$ and $\text{OLR}_{\text{clr}}^*$ represent the equator-to-pole gradient of the absorbed clear-sky shortwave radiation and clear-sky outgoing longwave radiation, respectively. The term $\text{ASR}_{\text{cld}}^*$ describes the equator-to-pole gradient

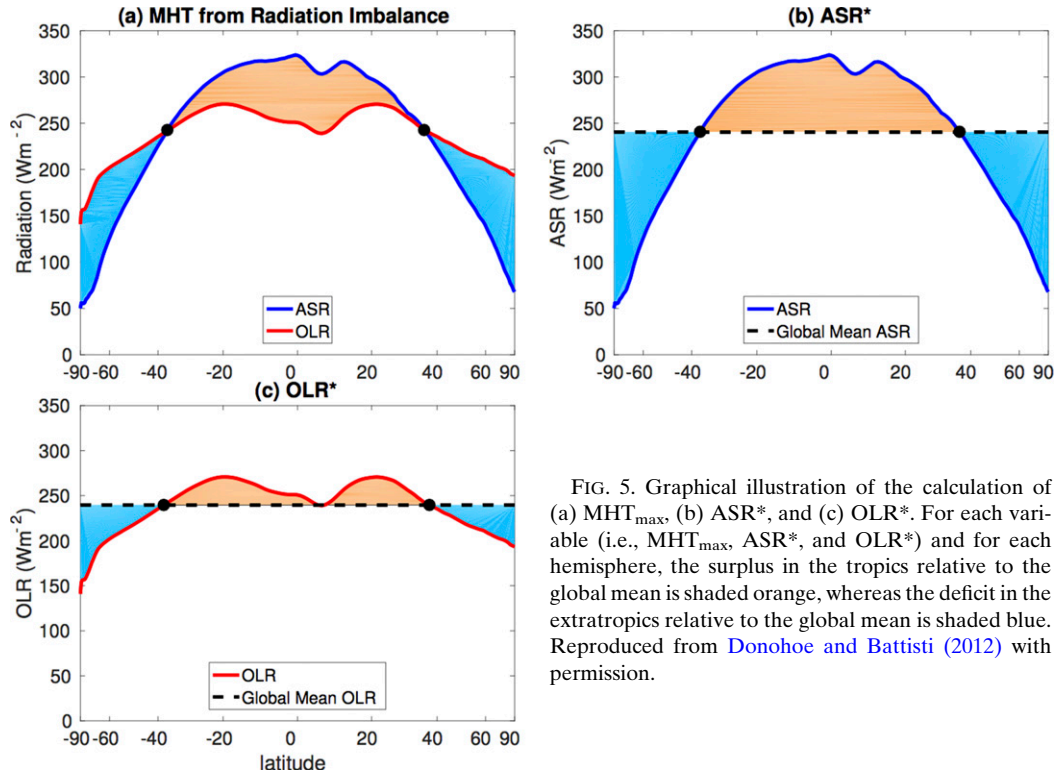


FIG. 5. Graphical illustration of the calculation of (a) MHT_{max} , (b) ASR^* , and (c) OLR^* . For each variable (i.e., MHT_{max} , ASR^* , and OLR^*) and for each hemisphere, the surplus in the tropics relative to the global mean is shaded orange, whereas the deficit in the extratropics relative to the global mean is shaded blue. Reproduced from [Donohoe and Battisti \(2012\)](#) with permission.

of ASR_{cld} , the shortwave cloud forcing (i.e., the negative of the shortwave reflected by clouds). Similarly, OLR_{cld}^* describes the equator-to-pole gradient of OLR_{cld} , the longwave cloud forcing (i.e., the negative of the longwave trapped by the clouds).

In the fast-rotating regime, ASR_{clr}^* is essentially invariant of rotation rate (Fig. 7a). Therefore, the increase of ASR^* with increasing rotation rate is due predominantly to an increase in ASR_{cld}^* (cf. Figs. 6a and 7c). The change in ASR_{clr}^* and ASR_{cld}^* can be understood by examining the spatial patterns of ASR_{clr} and ASR_{cld} (Figs. 8a,b). Because of the unchanging geometry and solar constant, ASR_{clr} remains nearly constant with rotation rate at each latitude; for all the experiments, ASR_{cld} remains negative, as it must be by definition. As the rotation is increased, ASR_{cld} is less negative equatorward of 40° latitude, indicating that less shortwave is reflected by clouds. This results in a decrease in the equator-to-pole gradient of shortwave reflected, or equivalently, an increase in the equator-to-pole difference of ASR_{cld} ($=2ASR_{cld}^*$).

The reduction in the shortwave reflected in the tropics, in turn, is related to a reduction in the high cloud amount from the equator to 20° latitude and a reduction in the low cloud amount from 20° to 40° latitude (Figs. 2b,d,f,g); both are associated with the change in the Hadley cell. As shown in Fig. 2, the tropical high

clouds are associated with the ascending branch of the Hadley cell and the subtropical low clouds are associated with the sinking branch; both types of clouds reduce with the weakening and shrinking of the Hadley cell with increasing rotation. Poleward of 40° latitude, ASR_{cld} stays relatively unchanged because cloud changes are constrained to be over the polar cap where the insolation is weak and clouds are prevalent in all experiments.

The increase in OLR^* with the increasing rotation rate is primarily due to an increase in OLR_{clr}^* and secondarily to an increase in OLR_{cld}^* (Figs. 7b,d). The increase in OLR_{clr}^* with increasing rotation rate is mainly due to an increase in clear-sky OLR equatorward of 20° latitude, consistent with the surface warming in the tropics as rotation is increased (cf. Figs. 8c and 1). The OLR_{cld} is a positive contribution to OLR^* : as the rotation rate is increased, less longwave is trapped by clouds in the tropics; therefore, OLR_{cld}^* increases. The reduction in the trapped longwave is associated with the reduction in the tropical high cloud amount (Figs. 2b,d,f,g), which is related to the weakening and narrowing of the Hadley cell (Fig. 2).

b. The slow-rotating regime

In the slow-rotating regime ($\Omega < 1/2$), ASR^* is nearly invariant with rotation, but OLR^* increases as the

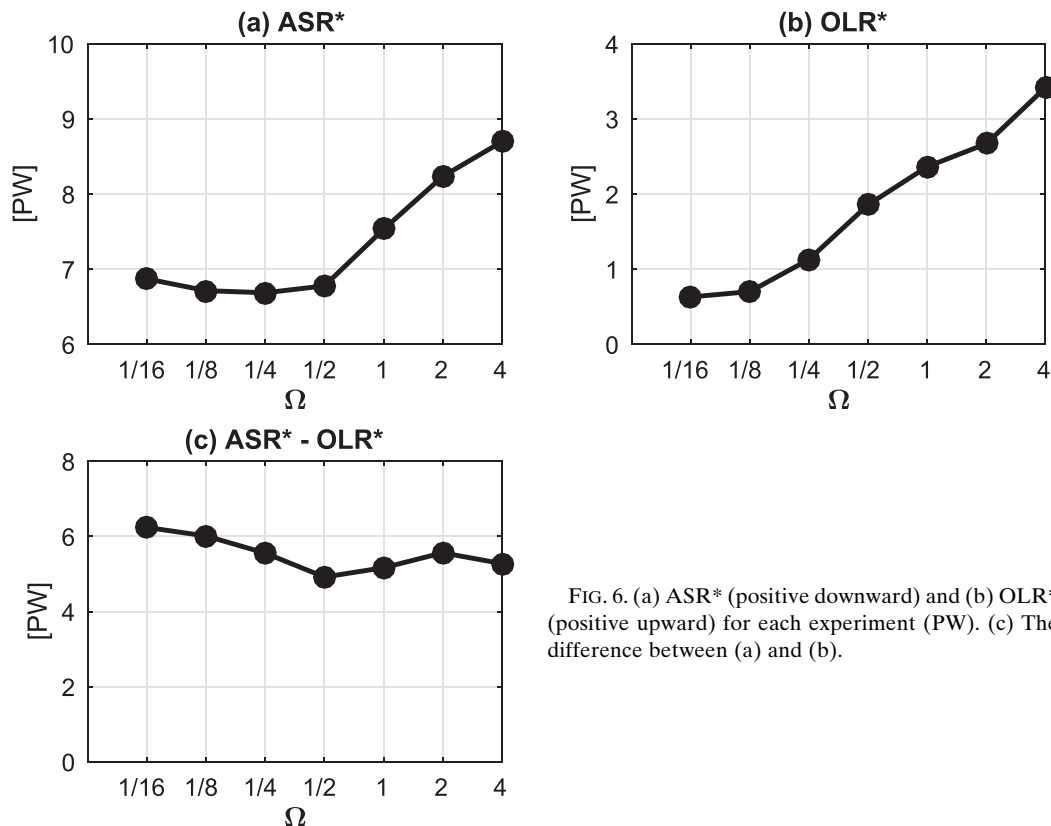


FIG. 6. (a) ASR* (positive downward) and (b) OLR* (positive upward) for each experiment (PW). (c) The difference between (a) and (b).

rotation is increased, resulting in a decrease in MHT_{max} with increasing rotation (Fig. 6); OLR^* increases with rotation primarily as a result of an increase in $\text{OLR}_{\text{clr}}^*$ (Figs. 7b,d).

As in the fast-rotating regime, the increase in $\text{OLR}_{\text{clr}}^*$ with increasing rotation rate is mainly due to an increase in clear-sky OLR in the tropics. Note that the temperature change in the slow-rotation regime is nearly uniform globally, resulting in a larger increase in clear-sky OLR in the tropics than in the polar regions via the nonlinear Planck response (Figs. 9c,d). The increase in $\text{OLR}_{\text{clr}}^*$ with increasing rotation rate is due to a decrease in longwave trapping by clouds in the tropics, which in turn is related to the decrease in tropical high cloud amount associated with the weakening of the Hadley cell (Figs. 2a,c,e).

Note that ASR^* does not change with the rotation in the slow-rotating regime because neither $\text{ASR}_{\text{clr}}^*$ nor $\text{ASR}_{\text{cld}}^*$ changes (Figs. 7a,c). Also, $\text{ASR}_{\text{cld}}^*$ remains nearly invariant because as rotation rate increases, $\text{ASR}_{\text{cld}}^*$ increases the same amount equatorward of the nodal point (roughly 35° latitude) as it does poleward of the nodal point (Fig. 9b). The increase in ASR is predominantly due to the reduction in the low clouds (Figs. 2a,c,e) that stems from the weakened subsidence associated with the sinking branch of the Hadley cell (note that the Hadley cell is much wider in the slow-rotating regime and extends

to the pole in the $\Omega = 1/16$ experiment). This relationship between the changes of clouds and the changes of the Hadley cell is the same as that of the fast-rotating regime. Although there are large changes in tropical high clouds and mid-to-high clouds in the polar regions, they do not impact the shortwave, and hence ASR^* , significantly.

5. The connection with dynamics

For $\Omega \geq 1/4$ the MHT is accomplished predominantly by the eddies (Table 2). It might be expected that eddy heat transport would depend on the geometry and intensity of the eddies, which in turn depend on rotation rate via eddy metrics [see Barry et al. (2002) for an important review], such as the Rhines scale (which scales as the inverse of the square root of the meridional gradient of the Coriolis parameter) or the maximum Eady growth rate (which scales as the Coriolis parameter). Turbulence mixing theory suggests that eddy diffusivity scales as VL , where V and L are eddy velocity and meridional length scales, respectively (e.g., Vallis 2006). Figures 10a and 10b show V and L area averaged between 30° and 60° latitude. Following Barnes and Hartmann (2012), V is taken to be the root-mean-square of the instantaneous (6 hourly) 850-hPa meridional wind averaged over each latitude band and

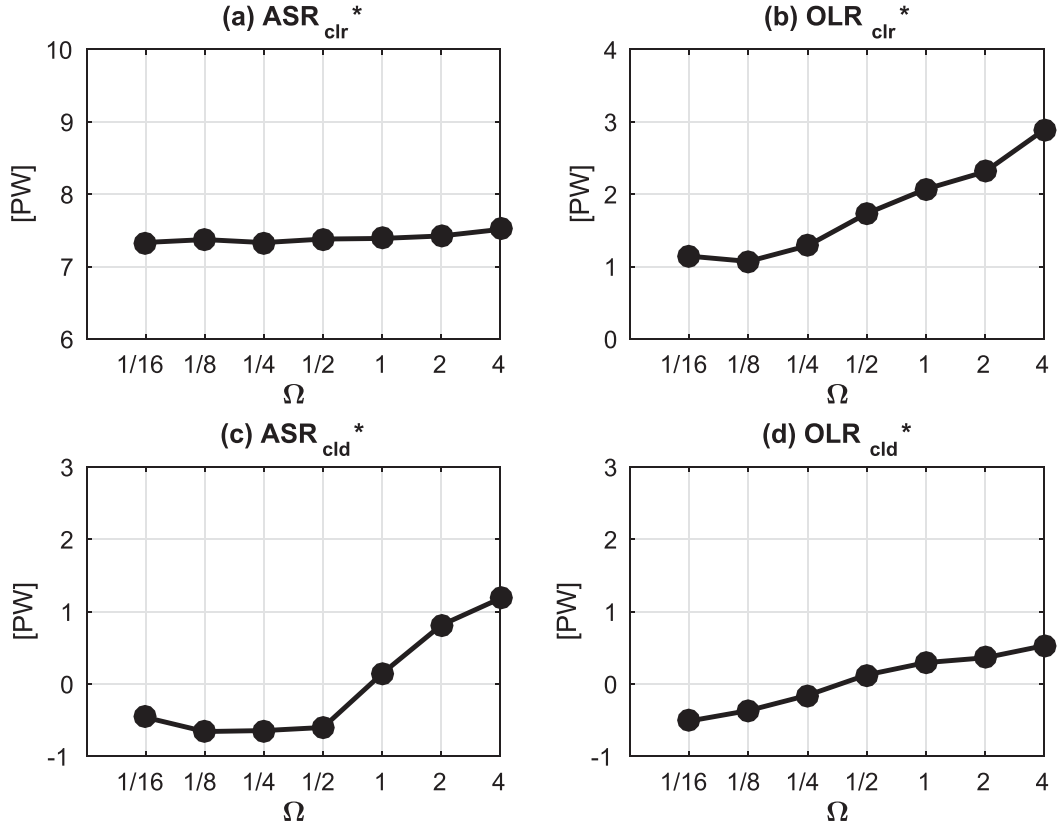


FIG. 7. (a) ASR_{clr}^* and (b) OLR_{clr}^* for each experiment (PW). (c),(d) As in (a),(b), but for ASR_{cld}^* and OLR_{cld}^* , which are defined as $ASR^* - ASR_{clr}^*$ and $OLR^* - OLR_{clr}^*$, respectively.

over time V_{RMS} , and L is defined as the meridional distance over which the autocorrelation in the instantaneous meridional wind decays by a factor of e . Also shown in Figs. 10a and 10b, respectively, are a velocity scale developed by Barry et al. (2002)² and the Rhines β scale, defined as $(2V/\beta)^{1/2}$, where V is, as in the present study, the root-mean-square of meridional velocity.

The term V weakly depends on the rotation rate: it decreases only slightly as the rotation rate is increased (except for $\Omega = 4$) (Fig. 10a). Across all of the experiments, there is a qualitative similarity between V and the velocity scale developed by Barry et al. (2002). The implication is that the weak dependence

of V on rotation is because the increase in β is partly offset by the increase in baroclinicity with increasing rotation rate. The term L decreases with increasing rotation rate, almost exactly following the change of the Rhines β scale (Fig. 10b), consistent with Barnes and Hartmann (2012). The suggested scaling for eddy diffusivity, VL , decreases monotonously with increasing rotation, mainly associated with the decrease in L (Fig. 10c). This implies that the eddies are less effective at transporting heat as the rotation is increased.

Several recent studies have shown atmospheric heat transport (AHT) in GCMs can be emulated by simple downgradient diffusion of near-surface moist static energy (e.g., Flannery 1984; Frierson et al. 2007; Hwang and Frierson 2010; Jansen and Ferrari 2015; Rose et al. 2014; Roe et al. 2015):

$$AHT = -2\pi \cos\theta \frac{p_s}{g} D_{eff}(\theta) \frac{dh}{d\theta}, \quad (10)$$

where θ is latitude, p_s/g is mean mass per unit area (taken to be a constant, 10^4 kg m^{-2}), $D_{eff}(\theta)$ is the effective diffusivity as a function of latitude, and

² Barry et al.'s (2002) velocity scale is $V = V_b \propto (a \bar{T}_y q / T_0)^{2/5} (2/\beta)^{1/5}$, where a is the radius of Earth, $\bar{T}_y$ and T_0 are the 1000–200-hPa vertically integrated meridional gradient of zonal mean temperature and the mean temperature area averaged over the midlatitudes (taken to be 30°–60° latitude), respectively, and, following Lapeyre and Held (2003), q is the average heating rate poleward of the latitude of maximum AHT. The arbitrary constant is chosen such that V_b matches V_{RMS} at $\Omega = 1$.

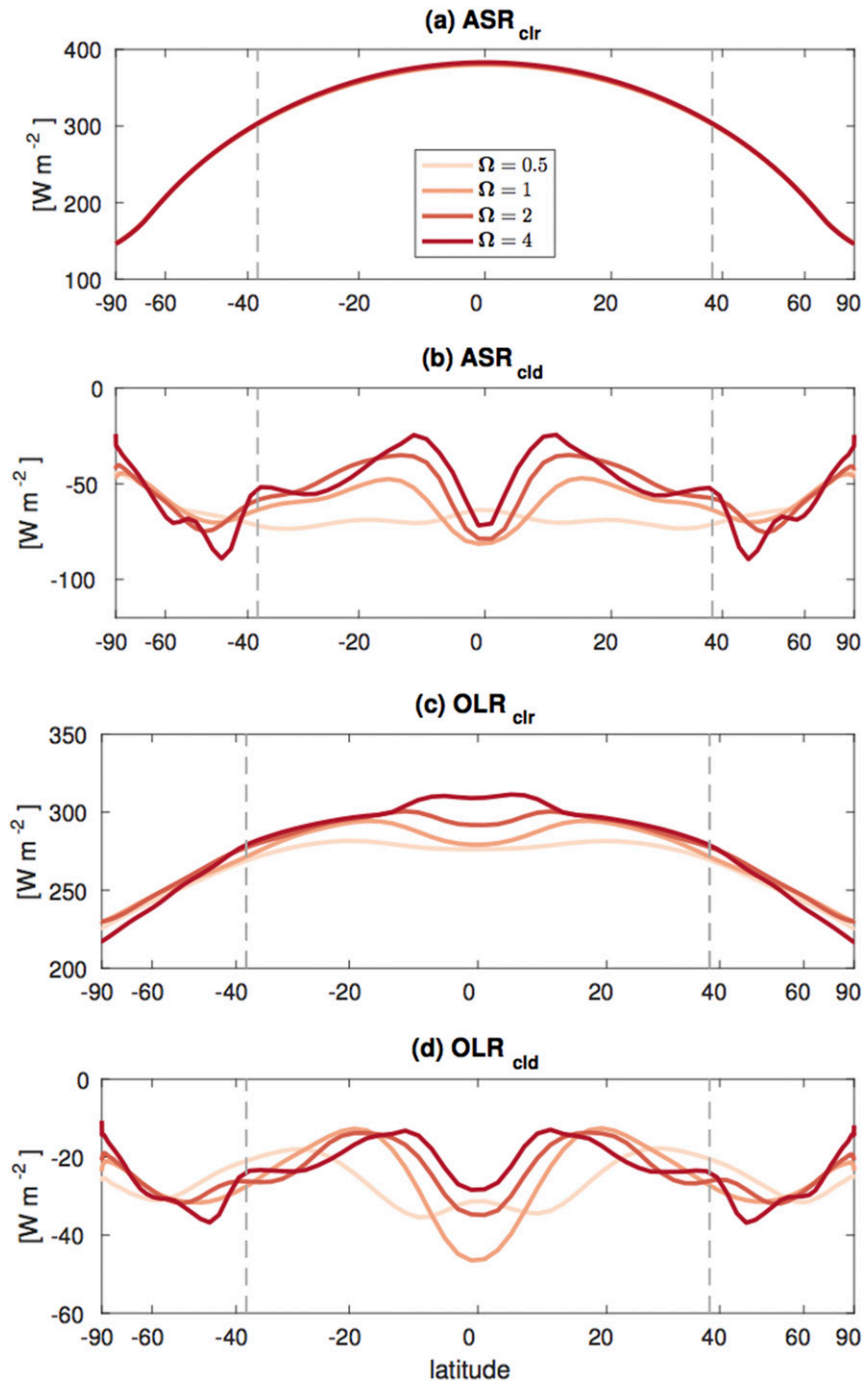


FIG. 8. (a) ASR_{clr} and (b) ASR_{cld} for experiments in the fast-rotating regime ($W m^{-2}$). Note the displacement of latitude is area weighted. (c),(d) As in (a),(b), but for OLR. The gray dashed lines indicate location of the nodal point of $\Omega = 1$. Note the different scales for the y axes in each panel. The results illustrate that the biggest changes are equatorward of the nodal point.

$h = c_p T + L_v q$ is the near-surface moist static energy (where c_p is specific heat at constant pressure, T is temperature, L_v is the latent heat of vaporization, and q is specific humidity). Also, D_{eff} can be diagnosed from model output using Eq. (10). Figure 10c presents a scatterplot

of D_{eff} versus VL as a function of rotation rate. Note that D_{eff} decreases by approximately tenfold as Ω increases from 1/16 to 4. There is an impressive, near-linear association between D_{eff} and VL over a 64-fold variation in rotation rate; the

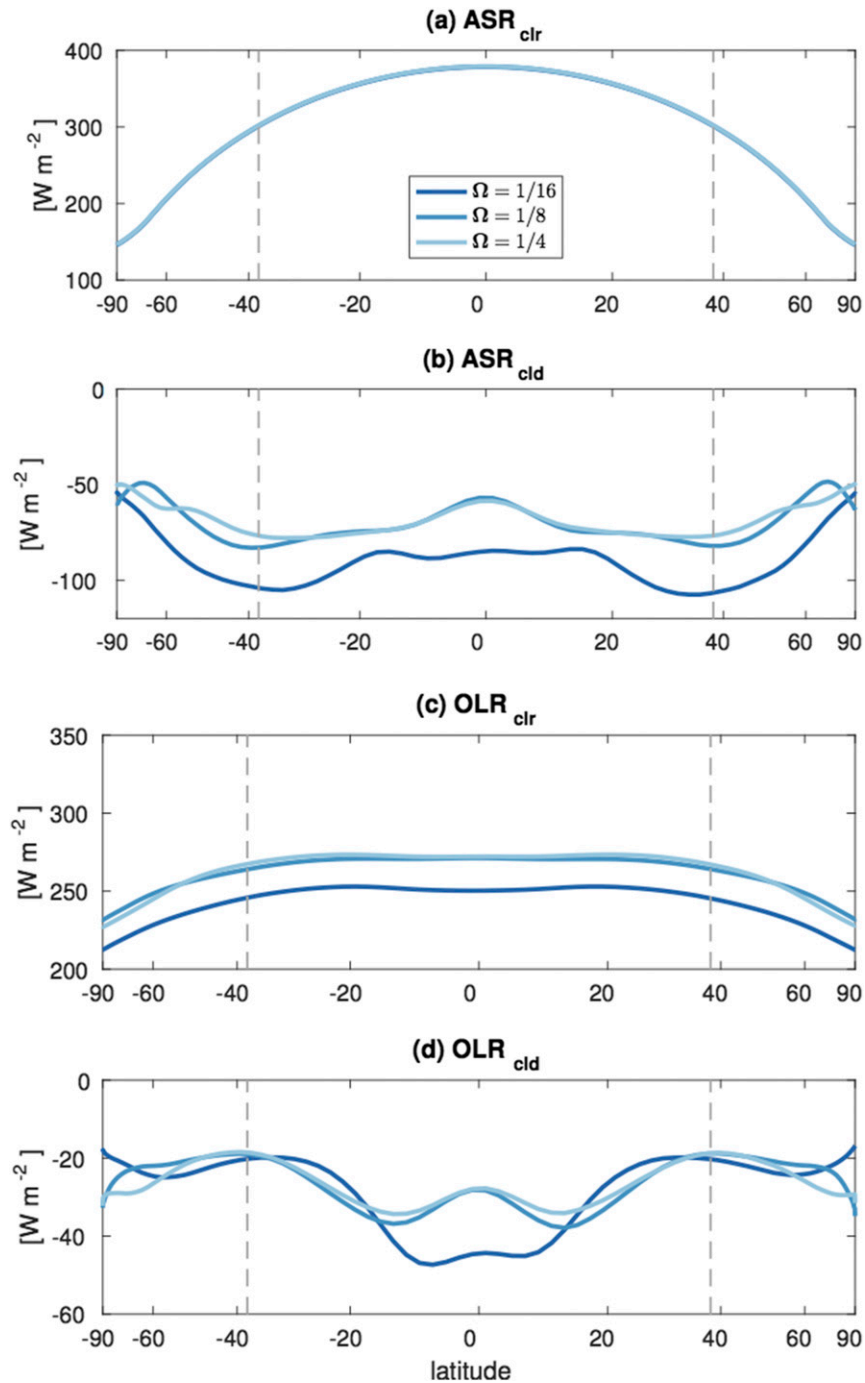


FIG. 9. As in Fig. 8, but for experiments in the slow-rotating regime.

correlation coefficient between these two exceeds 0.99. The results thus show that the model heat transport is indeed approximately diffusive, with an effective diffusivity that is consistent with eddy mixing-length theory. This is even true in the slow-rotating regime where the eddies are not the dominant transport mechanism. The results are also consistent

with the idea that eddies are less effective at transporting heat at higher rotation rates.

Despite the tenfold decrease in D_{eff} , AHT and hence MHT (because the ocean heat transport is fixed in this series of experiments) change by no more than about 30% across this range (Table 2). The reason is that the decrease in D_{eff} is nearly offset by the strong increase in

TABLE 2. Decomposition of AHT into the contribution by mean meridional circulation (MMC) and the contribution by eddies. The contribution by MMC is calculated as $\int_{-\pi/2}^0 \int_0^{2\pi} \left(\int_{p_s}^0 [v][h] dp \right) \cos\theta d\lambda d\theta$, where v is the meridional wind, h is the moist static energy, p is pressure, p_s is surface pressure, θ is latitude, λ is longitude, and square brackets denote zonal and temporal mean. The eddy energy transport is defined as the residual of the MMC transport from the total AHT estimated as $\int_{-\pi/2}^0 \int_0^{2\pi} \left(\int_{p_s}^0 v h dp \right) \cos\theta d\lambda d\theta$. In the calculation of total AHT, the vertical integral of vh is calculated at the model grid at each time step. As shown below, $\text{lat}_{\max}$ denotes the latitude where AHT calculated from the meridional integral of the net heat flux into the atmosphere reaches its maximum value.

Ω	$\text{lat}_{\max}$	MMC transport (PW)	Eddy transport (PW)	MMC + eddy (PW)	AHT _{max} (PW)
1/16	33°	4.8	1.1	5.9	5.8
1/8	35°	2.9	2.8	5.7	5.5
1/4	35°	1.4	3.6	5.0	5.1
1/2	37°	0.2	3.8	4.0	4.1
1	41°	−0.6	4.8	4.2	4.3
2	39°	0.0	4.5	4.5	4.6
4	39°	−0.1	4.2	4.1	4.5

the gradient of moist static energy associated with the tropical warming at high rotation rates (Fig. 1b).

6. Discussion

It is evident from the above analysis that the regime feature of MHT with increasing rotation rate is related to the regime behavior of $\text{ASR}_{\text{cld}}^*$: it is nearly constant in the slow-rotating regime but increases with increasing rotation in the fast-rotating regime. Why does $\text{ASR}_{\text{cld}}^*$ change differently with rotation rate in these two regimes,

given that the patterns of change in cloudiness that are essential to the change in $\text{ASR}_{\text{cld}}^*$ are the same for both regimes? It is related to the mean distribution of tropical and subtropical clouds that are associated with the Hadley cell. In both regimes, the tropical high clouds are associated with the rising branch of the Hadley cell and the subtropical low clouds are associated with the sinking branch of the Hadley cell. In the slow-rotating regime, the subtropical low clouds decrease on both the equatorward and the poleward sides of the nodal point (roughly 36° latitude) (Figs. 2a,c,e) because the Hadley cell extends

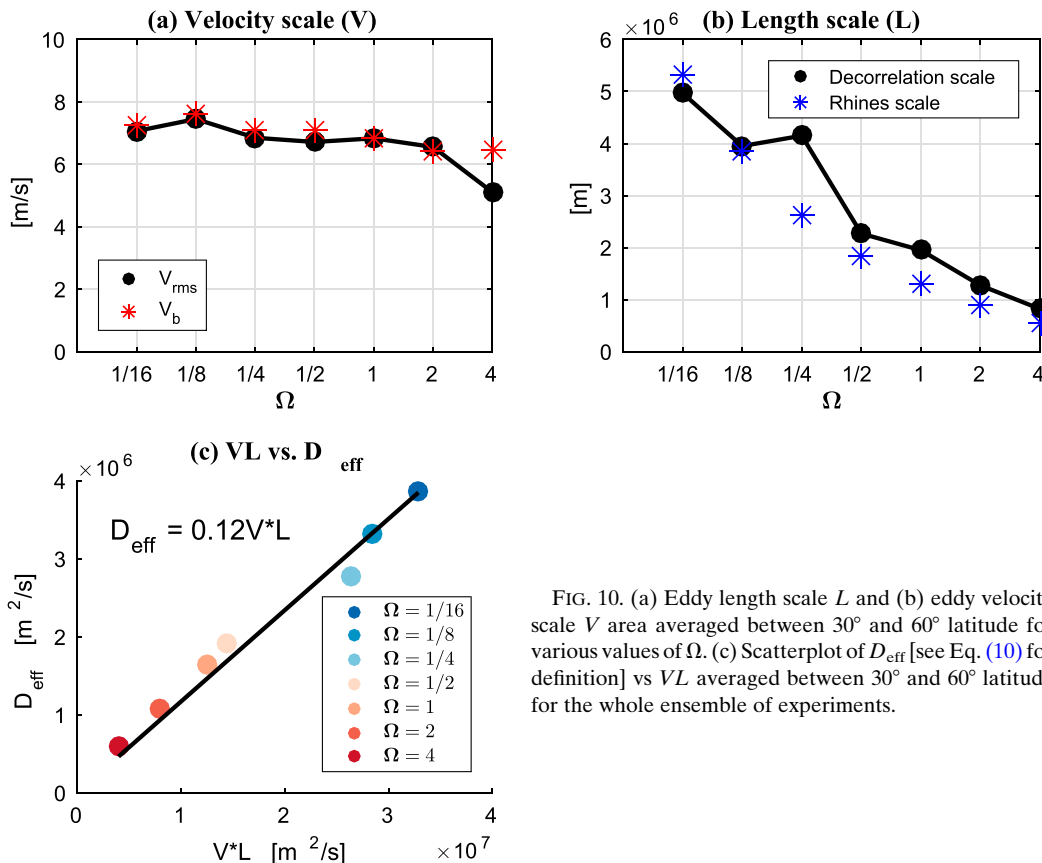


FIG. 10. (a) Eddy length scale L and (b) eddy velocity scale V area averaged between 30° and 60° latitude for various values of Ω . (c) Scatterplot of D_{eff} [see Eq. (10) for definition] vs $V \cdot L$ averaged between 30° and 60° latitude for the whole ensemble of experiments.

poleward of the nodal point, resulting in a near-zero change in the equator-to-pole gradient of ASR_{cld} , that is, ASR_{cld}^* . On the contrary, in the fast-rotating regime the reduction in tropical high clouds and the reduction in subtropical low clouds are both equatorward of the nodal point (Figs. 2b,d,f,g) so they work in a concerted way to reduce the shortwave reflected and hence cause a net increase in ASR_{cld}^* as rotation rate increases.

In terms of dynamics, the two-regime nature of MHT is related to the Clausius–Clapeyron relationship, which increases the meridional gradient of surface moist static energy more dramatically in the fast-rotating regime (Fig. 1b), compensating for the decrease in eddy diffusivity (Fig. 10). In the slow-rotating regime, however, because the increase in the gradient of surface moist static energy is not sufficient to compensate the decrease in eddy diffusivity, the MHT decreases with increasing rotation rate. Our results contrast with those of Vallis and Farneti (2009), who found that the MHT decreases monotonously with increasing rotation rate. The primary reason for this difference is the choice of atmosphere models: Vallis and Farneti (2009) use a gray atmosphere with fixed spatial distribution of clouds; as a result, ASR^* cannot change with rotation rate. In contrast, in our experiments changing ASR^* causes changes in tropical temperature and moist static energy, which in turn affect the meridional gradient and hence the flux of moist static energy.

Another striking feature revealed in this study is that the eddy diffusivity scales linearly with the effective eddy diffusivity (Fig. 10c). The linearity implies an important constraining relationship between the dynamics and the radiation.

Our results above suggest that changes in clouds are fundamental to the insensitivity of MHT to rotation. To further test the role of clouds, we repeated the experiments with a gray radiation model in which clouds and longwave emissivity are prescribed. We found that when clouds and water vapor feedback are omitted, the tropical warming is much smaller, ASR^* stays constant, and OLR^* increases with increasing rotation for $\Omega < 1/2$ and remains relatively unchanged for $\Omega \geq 1/2$. This conclusion is further supported by two additional sets of experiments, increasing the rotation rate from $\Omega = 1/16$ to $\Omega = 4$: in the first set of experiments, SST was fixed to be that from the control ($\Omega = 1$) experiment using the slab ocean; in the second set of experiments, the rotation rate is held fixed at $\Omega = 1$, but the SST is prescribed to be that from the varying rotation experiments using the slab model. Comparing the results from these two sets of experiments to those using the slab model, we found that (i) the change in clouds and hence in ASR^* is primarily due to the changes in the Hadley cell as a result of changes in rotation (the tropical ASR increases with

rotation in the variable rotation–fixed SST experiments but not in the variable SST–fixed rotation experiments) and (ii) the change in Hadley cell is predominantly determined by change in the rotation without changing SST [the Hadley cell contracts and weakens with increasing rotation in the variable rotation–fixed SST experiments, but strengthens and widens slightly in the set of variable SST–fixed rotation experiments (whereby the prescribed SST is marched from the low to the high rotation rate solution), in contrast to what is seen in the experiments using slab model].

In the present study, we used the same fixed Q flux for all of the experiments; if we had used a dynamic ocean model the ocean heat transport would also change with changes in rotation rate. However, using the fixed Q flux is reasonable in the sense that MHT_{max} is dominated by the atmospheric heat transport and therefore should be relatively insensitive to changes in ocean heat transport. We also performed another set of rotation rate experiments in which no Q flux is used. For this set of experiments, the behavior of MHT, ASR^* , and OLR^* with rotation is the same as discussed in sections 3 and 4 (Figs. 4b,c).

The change of MHT with rotation rate is independent of the model resolution. To better resolve the eddies in the fast rotation rate experiments, we performed another set of experiments in which the horizontal resolution is doubled (from $2.0^\circ \times 2.5^\circ$ to $1.0^\circ \times 1.25^\circ$). For each rotation rate, MHT, ASR^* , and OLR^* differ by only 2%–10% when the model resolution is doubled (not shown). The changes in cloudiness, circulation, radiation, and so on that underlie the changes in ASR^* and OLR^* (and hence MHT) are also independent of the model resolution used. The only notable difference between the sets of high- and low-resolution experiments is that the former set features double intertropical convergence zones (ITCZs), whereas the latter set features a single ITCZ on the equator.

We also repeated our experiments and analysis using the NCAR aquaplanet CAM4 coupled to a slab ocean to examine the model dependence of our results. We found that the dynamics and the radiative fluxes in NCAR CAM4 change with rotation in exactly the same way as those in GFDL AM2.1: as the rotation is increased, the Hadley cell contracts and weakens, and the associated tropical and subtropical clouds are reduced; as a result, ASR increases in the tropics, the tropical temperature increases, the clear-sky OLR increases, and less longwave is trapped by clouds. Hence, just as in the experiments with the AM2 coupled to a slab ocean, both ASR^* and OLR^* increase as rotation rate increases in the CAM4 plus slab model, leading to modest change in MHT (figures not shown). Encouragingly, other studies also have found that tropical temperatures increase and

subtropical clouds decrease as rotation rate is increased [e.g., Jenkins et al. (1993), using the NCAR Community Climate Model (CCM); Navarra and Boccaletti (2002), using ECHAM4; and Salameh et al. (2017), using ECHAM6]. Together, our results and the aforementioned studies give us confidence that as rotation rate increases, changes in clouds will cause changes in ASR that will tend to be compensated by changes in OLR, resulting in only modest changes in MHT relative to those in ASR and OLR, although the degree of compensation might be sensitive to the parameterization of cloud processes, ice microphysics, and the parameterization of the PBL.

7. Summary

The atmosphere–ocean system transports energy meridionally from the equator toward the poles. The magnitude of this meridional heat transport (MHT) is such as to achieve a balance between the pattern of absorbed shortwave radiation (ASR) and the pattern of outgoing longwave radiation (OLR) at the top of the atmosphere. In this study we have sought to understand what controls the magnitude of MHT by varying Earth's rotation rate in a series of numerical experiments. The essential result is that the changes in rotation rate cause changing patterns of clouds. The resulting changes in ASR* and OLR* control changes in MHT. In this regard our results are consistent with a recent study that demonstrated that the large (~ 2 PW) spread in MHT_{max} among CMIP3 GCMs for the modern climatology is related to differences in their representation of ASR* associated with difference in cloud shortwave forcing (Donohoe and Battisti 2012).

Consistent with basic theory and several other modeling studies, we find that an increasing rotation rate causes a narrowing and weakening of the Hadley cell. Changes in MHT are linked to the accompanying changes in both tropical high and subtropical low clouds. For the GFDL AM2.1 model (and for the NCAR CAM4 model), we find two distinct regimes: (i) a slowly rotating regime (rotation less than half present-day value) in which, with increasing rotation rate, high tropical clouds disperse leading to less trapping of OLR, more local accommodation of ASR, and hence a reduction in MHT (about 30% going from $\Omega = 1/16$ to $\Omega = 1/2$); and (ii) a fast rotating regime (rotation greater than half present-day value) in which, with increasing rotation rate, there are closely offsetting changes in OLR* and ASR* (the latter mediated by changes in low subtropical clouds), such that MHT remains nearly invariant ($<5\%$ change going from $\Omega = 1/2$ to $\Omega = 4$).

While the detailed response of MHT to rotation rate will of course depend on the GCM used, the basic cloud

response to increasing rotation is found in several different models. The fundamental point is that changing cloud patterns are first-order controls on ASR* and OLR* and hence MHT. Our results echo the principal idea laid out in Stone (1978) of the importance of top-of-the-atmosphere fluxes in setting MHT, but when cloud adjustments are included there is a potential for greater variation in MHT than he recognized. In units of watts per meter squared, the magnitudes of cloud adjustments with rotation (Figs. 7c and 8b) are comparable to, or exceed, the magnitude of the climatological ocean heat uptake (e.g., Hartmann 2015); thus, just in energetic terms, cloud variations are arguably more important than ocean transport variations in setting MHT. This is also borne out in a study evaluating the range of MHT among CMIP3 GCMs in modern climatology (Donohoe and Battisti 2012).

It is worth noting that although we attributed the insensitivity of the MHT to rotation rate to changes in the top-of-the-atmosphere fluxes, this does not mean that dynamics plays no role. On the contrary, dynamics is essential. The changes in clouds that are fundamental to the changes in top-of-the-atmosphere fluxes are largely determined by changes in the Hadley cell, which, as reported in section 6, are overwhelmingly due to the changing rotation rate.

Our study also suggests that the change in MHT_{max} , if there is any, is most sensitive to changes in the zonal-mean distribution of clouds in the tropics. In speculating about MHT in past climates, this implies that to get a large change in MHT_{max} one has to change the tropics dramatically, for example, by moving land into the tropics (e.g., Pangaea during the Triassic Period, perhaps). In this regard, the glacial fluctuations of the Pleistocene have primarily involved changes in the high latitudes, and so one would not expect much change in its MHT_{max} .

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