The Role of the Ocean in the Seasonal Cycle of the Hadley Circulation

AMY C. CLEMENT

Rosenstiel School of Marine and Atmospheric Science, University of Miami, Miami, Florida

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ABSTRACT

The influence of ocean heat transport on the seasonal cycle of the Hadley circulation is investigated using idealized experiments with a climate model. It is found that ocean heat transport plays a fundamental role in setting the structure and intensity of the seasonal Hadley cells. The ocean's influence can be understood primarily via annual mean considerations. By cooling the equatorial regions and warming the subtropics in a year-round sense, the ocean heat transport allows for regions of SST maxima to occur off the equator in the summer hemisphere. This leads to large meridional excursions of convection over the ocean and a seasonal Hadley circulation that is strongly asymmetric about the equator. The broadening of the latitudinal extent of the SST maximum and the convecting regions by the ocean heat transport also weakens the annual mean Hadley circulation in a manner that is consistent with simpler models. The results are discussed in the context of prior studies of the controls on the strength and structure of the Hadley circulation. It is suggested that a complete understanding of the seasonal Hadley circulation must include both oceanic and atmospheric processes and their interactions.

1. Introduction

The mean meridional atmospheric circulation in the Tropics is dominated by the well-known Hadley circulation, which was first described by Hadley (1735). His depiction of the annual mean circulation consisted of rising motion on the equator and subsidence in the subtropics and has been confirmed with global radiosonde data and multiple reanalysis products (Oort and Yienger 1996; Waliser et al. 1999). From month to month, however, the Hadley circulation can look quite different from the annual mean. During the solstices, there is a closed circulation that extends from deep in the summer hemisphere to the winter hemisphere subtropics (which we will refer to as the "cross-equatorial cell"), and there is a cell that resides within the summer hemisphere (which we will refer to as "the summer cell") that is significantly weaker than the cross-equatorial cell, as shown in Figs. 1a,b.

The simplest explanation for the seasonal zonal mean meridional circulation shown in Figs. 1a,b is that it arises in response to atmospheric heating in the summer hemisphere. Gill (1980) showed that a meridional overturning circulation is expected as the baroclinic response to a heating north of the equator with sinusoidal vertical structure. A series of studies using axisymmetric models showed that the main features of the Hadley circulation, including its seasonality, can be simulated under the constraints of thermal wind, angular momentum balance, and zero heat flux across the boundaries of the cell (Schneider and Lindzen 1977; Schneider 1977, 1987; Held and Hou 1980; Lindzen and Hou 1988; Hou and Lindzen 1992; Walker and Schneider 2005). While these prior studies form the basis for the present understanding of the Hadley circulation, they are designed to solve for the atmospheric response to a given heating distribution. Hence, they do not address the processes that give rise to the spatial structure and seasonal evolution of the atmospheric heating. The focus of this paper will be on those processes.

Some previous studies have investigated the processes behind the seasonal cycle of the Hadley circulation. For example, Cook (2003) used atmospheric GCM simulations to evaluate the role of heating over the continents on the seasonal Hadley cells. Results showed that the influence of the continents double the intensity of the cross-equatorial cell and halve the intensity of the summer cell. Qualitative support for this large role of the continents in setting the seasonal structure of the Hadley cells comes from an analysis of the

Corresponding author address: Dr. Amy Clement, RSMAS/MPO, 4600 Rickenbacker Causeway, Miami, FL 33149.
E-mail: aclement@rsmas.miami.edu

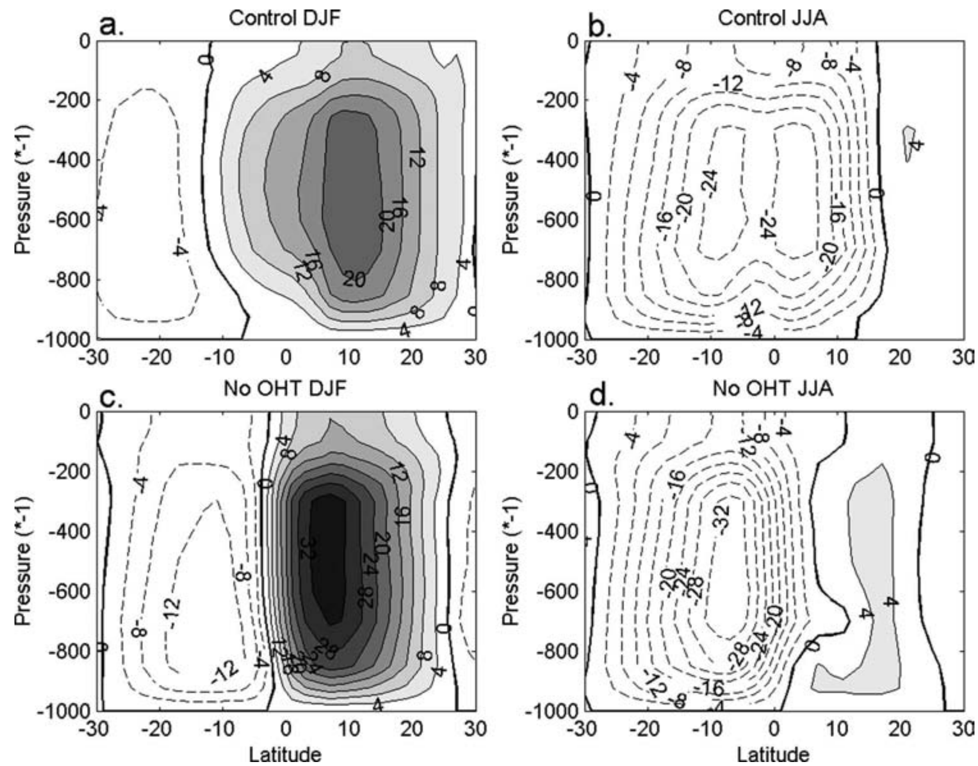

FIG. 1. The streamfunction of the zonal mean meridional atmospheric mass overturning (10^{10} kg s^{-1}) simulated by the GFDL atmosphere model: the (a) December–February (DJF) and (b) JJA mean circulation for the Control run, and the (c) DJF and (d) JJA mean circulation for the No OHT run. The sign convention is that positive values indicate clockwise circulation. Positive values are shaded. Contour interval is 4×10^{10} kg s^{-1}. For hPa pressures, multiply the vertical coordinate values by -1.

climatology of the atmospheric circulation. Trenberth et al. (2000) showed that the upper-tropospheric divergence field is highly zonally asymmetric and is dominated by monsoonal regions. The seasonal cycle of this field, they argued, largely explains the seasonal cycle of the Hadley circulation. Dima and Wallace (2003) found that most of the seasonal cross-equatorial mass flux occurs at the longitudes of the large landmasses. Together, these studies suggest that processes that lead to atmospheric heating over the continents can largely explain the features of the seasonal Hadley cells.

While monsoons represent a large seasonal heating of the atmosphere, most of the seasonal atmospheric convection in the Tropics actually takes place over the ocean, as illustrated with the observed zonal mean precipitation (Figs. 2a,b), suggesting an important role of the ocean in the seasonal Hadley circulation. The ocean processes that influence where these zones of atmospheric convergence occur have been studied for the annual mean (e.g., Kiehl 1998; Xie and Saito 2001; Clement et al. 2005) and are well known for interannual migrations, especially during El Niño–Southern Oscillation events (see Chang et al. 2006 for a review). However, the role that ocean processes play in the seasonal migrations of the oceanic convective regions, and hence in the seasonal Hadley circulation, is not well known.

Here it is argued that ocean dynamics play a fundamental role in setting the structure and intensity of the seasonal Hadley circulation. Idealized experiments with a climate model are used to isolate the effect of ocean dynamics on the seasonal Hadley circulation, as described in section 2. Results from the model experiments are presented in section 3. A final section will provide a discussion of the model results in the context of previous explanations for the seasonal Hadley circulation.

2. Methodology and climate model experiments

The atmosphere model used here is a developmental version of the new Geophysical Fluid Dynamics Laboratory (GFDL) atmospheric general circulation model. It is a B-grid model with 2.0° latitudinal and 2.5° longitudinal resolution and 18 vertical levels. Moist convection is handled with a modified relaxed Arakawa–Schubert (RAS) mass flux scheme (Moorthi and Suarez

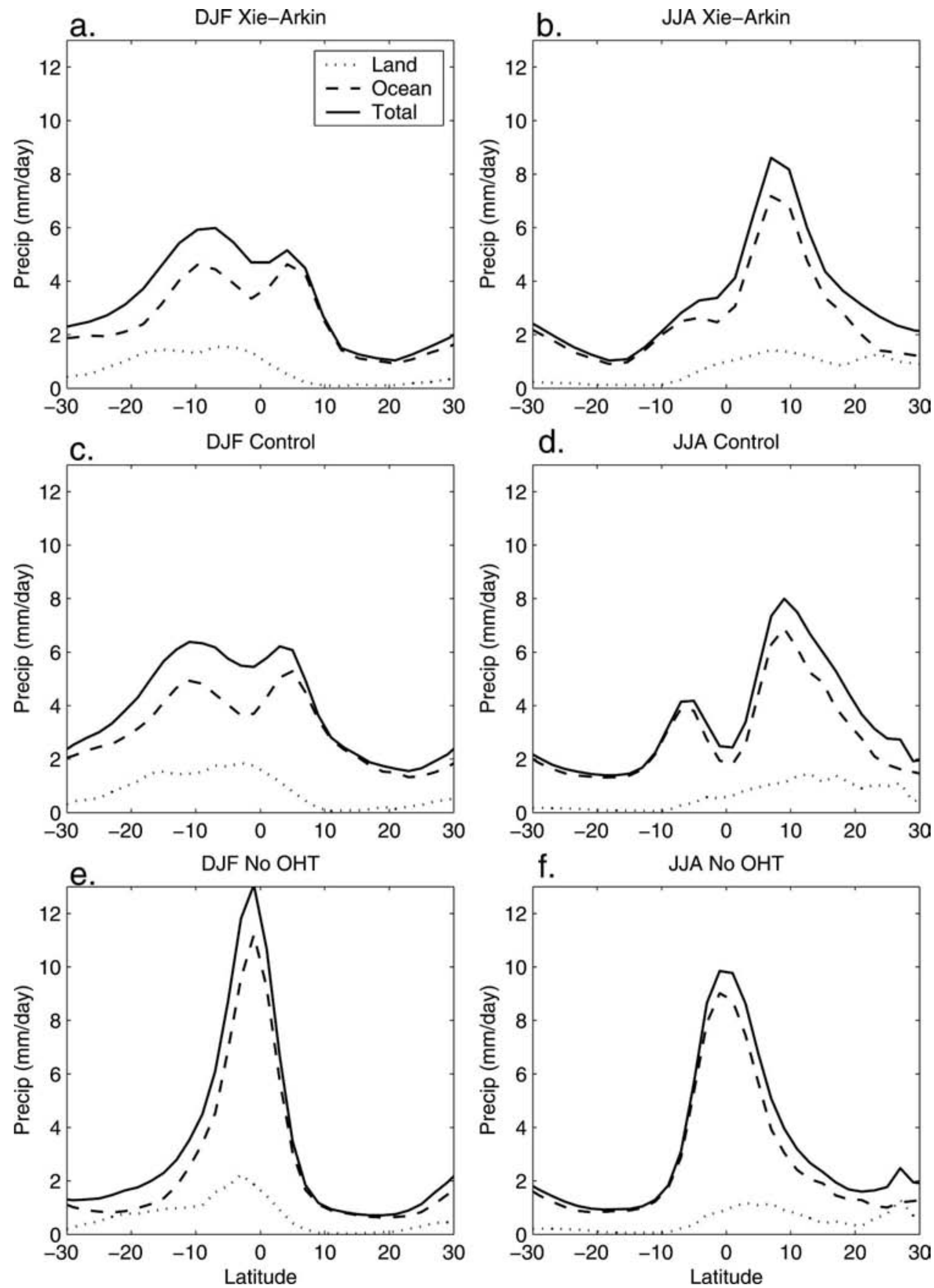

FIG. 2. The zonal mean climatological precipitation (mm day^{-1}) from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (Xie and Arkin 1997) averaged over land points (dotted), ocean points (dashed), and all points (solid) for (a) DJF, (b) JJA, (c) the model Control run DJF, (d) the model Control run JJA, (e) No OHT DJF, and (f) No OHT JJA.

1992). The version of the GFDL model is the same as that used in Winton (2003). The simulation of the seasonal Hadley circulation (Figs. 1a,b) is similar in structure to and within the range of magnitudes of reanalysis products (Mitas and Clement 2005). The model also simulates well the observed meridional structure of the zonal mean precipitation field, as well as its distribution between land and ocean points, albeit with a more pronounced double intertropical convergence zone (ITCZ) in July–August (JJA; Fig. 2d), which is a common problem in climate models.

Two simulations with this model will be used to evaluate the effects of ocean heat transport on the seasonal Hadley circulation—one with and one without ocean heat transports. In the simulation in which ocean heat transports are included, the atmosphere model is forced with the observed climatological SSTs and run until the model climate is stable (referred to hereafter as "Control"). This simulation includes the effects of ocean processes, but only implicitly. A heat flux correction, commonly referred to as the Q-flux, can be derived such that if the atmosphere model were rerun and coupled to a mixed layer ocean in which this Q-flux is added to the SST equation, the resulting SST would be close to the observed (e.g., as described in Kiehl et al. 1998). The Q-flux contains the effects of ocean dynamics (including wind-driven and thermohaline components) and model biases, although it will be argued here that it is dominated by the effects of ocean heat transport (as in Winton 2003).

The second experiment is one in which the atmosphere model is coupled to a mixed layer ocean of uniform 50-m depth, but the Q-flux is set to zero at all ocean points [referred to hereafter as "No OHT" (no ocean heat transport)]. In this case, the SST comes into balance with the atmosphere locally and the model is run until the annual mean net surface heat flux is zero everywhere. Results from this have previously been used to quantify the effects of ocean heat transport on the global mean temperature (Winton 2003; Herweijer et al. 2005).

An important test of whether the difference between these two simulations represents the effect of ocean heat transport on the climate and not simply model errors is whether the Q-flux derived from the Control simulation is realistic. The ocean heat transport is notoriously difficult to quantify from observations (Trenberth and Caron 2001; Jayne and Marotzke 2001; Ganachaud and Wunsch 2000, 2003). Trenberth and Caron (2001) suggest that the best way to calculate it is by using the net top-of-atmosphere radiative fluxes observed from satellites together with the atmospheric heat transports from reanalyses products to estimate the net heat flux at the ocean surface. The mixed layer temperature equation shows how the surface heat flux is related to the divergence of ocean heat transport:

$$\rho c_p h \frac{dT}{dt} = Q_{\text{net}} - D, \tag{1}$$

where the left-hand side is the heat storage in the mixed layer and the terms on the right-hand side are the net surface heat flux (the sum of the net radiative heat flux, latent heat flux, and sensible heat flux) and the divergence of heat transport by the ocean (or for the model, the Q-flux). In the annual mean, the left-hand side vanishes, so the net surface heat flux is balanced by the divergence of ocean heat transport. Trenberth and Caron's divergence estimate is shown with the model annual mean Q-flux in Fig. 3. There is good agreement both in magnitude and the spatial structure between the estimate based on observations (top panels) and the model (bottom panels), which supports our interpretation of the difference between the Control and No OHT as representing the effect of ocean heat transport on the seasonal Hadley circulation. These also generally agree with estimates based on a combination of inverse models and ocean data (Ganachaud and Wunsch 2000, 2003). As can be seen in Fig. 3, the ocean heat transport divergence primarily affects the SST in upwelling zones (where ocean circulation cools the SST) and in western boundary current regions (where ocean circulation warms the SST). The model underestimates the warming of the ocean in the western boundary current region, which is apparent in the zonal mean (right-hand panels of Fig. 3). This means that the model atmosphere does gain enough heat from the ocean in those regions, which could be related to errors in the radiative and/or turbulent heat fluxes at the ocean surface. However, it will be argued below that such errors do not affect the main conclusions of this paper. Note that because the Q-flux represents the sum of all ocean processes, we cannot distinguish between the effects of, for example, the thermohaline versus the wind-driven components of the ocean circulation in a quantitative way. However, we will offer some qualitative discussion of the specific ocean processes with the most importance to the seasonal Hadley circulation.

There is also a large seasonally varying component to the ocean heat transport divergence and Q-flux. This can be calculated from Eq. (1) with knowledge of the seasonal temperature changes and the seasonal net surface heat flux, assuming a mixed layer depth, h, which we will take to be 50 m. The seasonal patterns (shown in Fig. 4) are similar between observations and the model in that there is divergence of heat in the summer hemisphere (negative values) and convergence in the winter hemisphere (positive values), and this also agrees with estimates from numerical ocean models forced by observed surface winds and heat fluxes (Bryan 1982; Jayne and Marotzke 2001; not shown). The model overestimates the seasonal cycle in ocean heat transport divergence, particularly in the subtropics and midlatitudes, which is due to errors in the surface heat flux. However, differences between the model and observations do not affect the main conclusions of the paper, as discussed below.

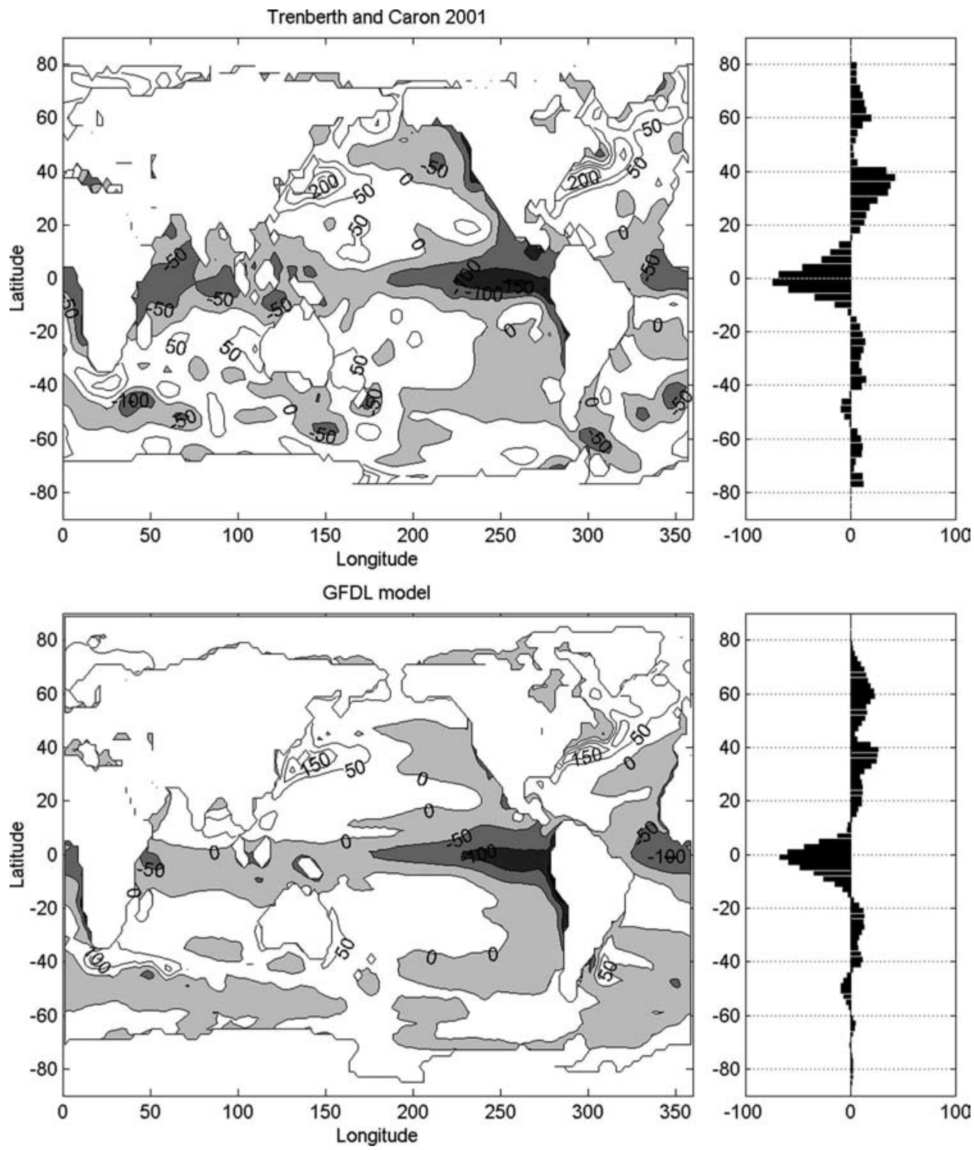

FIG. 3. (top left) The annual mean ocean heat transport divergence (W m^{-2}) estimated by Trenberth and Caron (2001) and (top right) zonal mean of this field. (bottom left) The annual mean net surface heat flux (W m^{-2}) in the Control run and (bottom right) zonal mean of this field. Contour interval is 50 W m^{-2} and negative values (which correspond to cooling by ocean dynamics) are shaded. Zonal mean values are scaled by the cosine of the latitude.

3. Influence of ocean dynamics on the seasonal Hadley circulation

Atmospheric convection over the ocean generally occurs over the warmest water. Considering local mixed layer ocean processes alone, one would expect that the SST maximum and convection would simply follow the maximum solar heating seasonally, with some lag due to the finite heat capacity of the mixed layer. The processes that determine the SST, however, are both local and nonlocal, particularly in the Tropics [Eq. (1)]. Convergence and divergence of heat by ocean dynamics significantly alters the geographical pattern of tropical SST from what it would be on the basis of local processes alone (Clement et al. 2005). Hence, it stands to reason that ocean heat transport would also affect the spatial structure and evolution of atmospheric convection and circulation.

The difference between the Control and No OHT simulations shows that the ocean heat transport influences both the structure and the intensity of the seasonal Hadley circulation (Fig. 1). Inclusion of the ocean

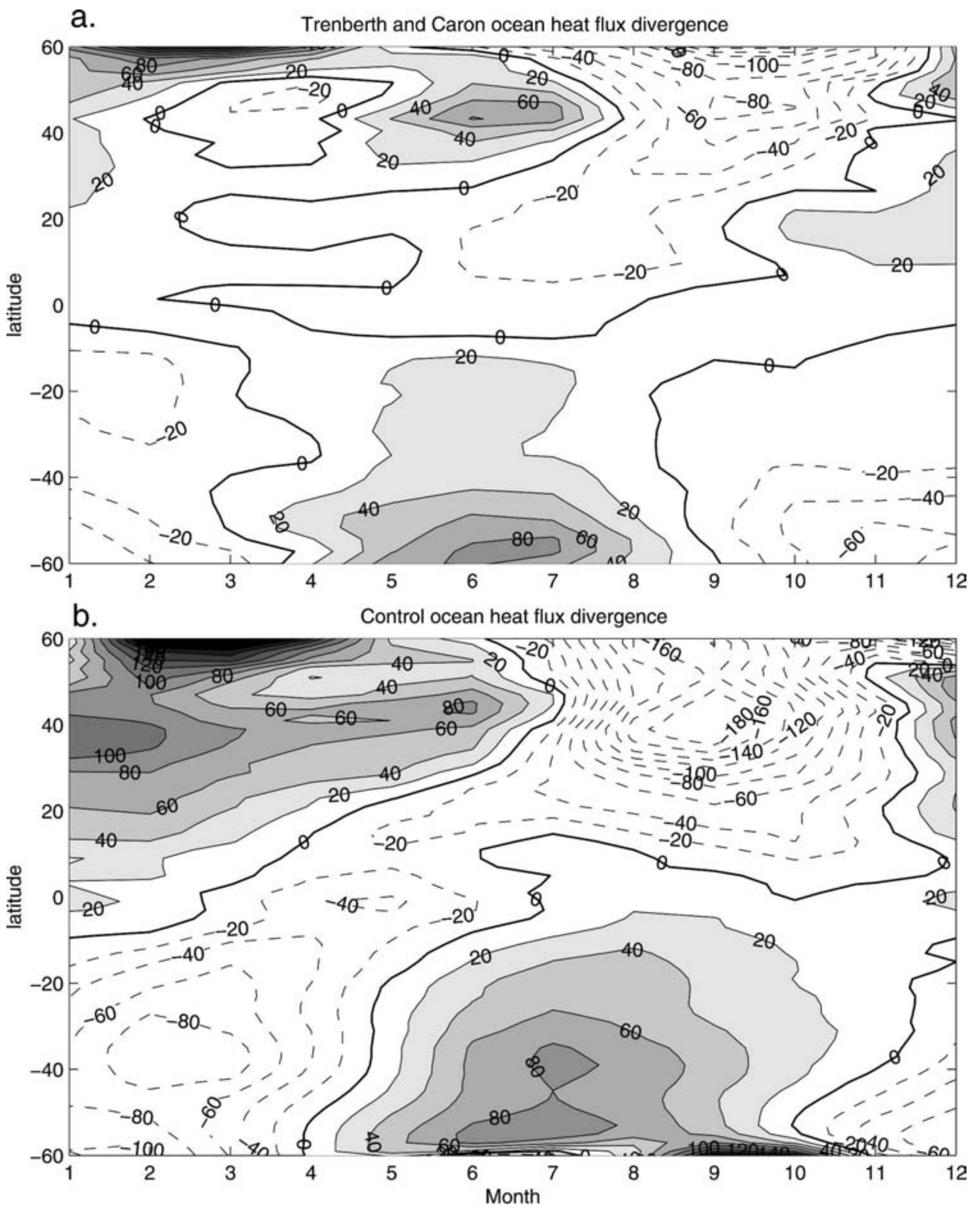

FIG. 4. (a) The zonal mean seasonal ocean heat transport divergence (W m^{-2}) derived from Trenberth and Caron (2001). (b) Same as in (a), but derived from the control run with the GFDL model. The annual mean (from Fig. 3) is removed in both cases. Description of how the seasonal heat transport is derived is described in the text. Positive values indicate convergence and are shaded.

heat transport results in a significantly more asymmetric cross-equatorial cell; the ascending branch of the cell moves deep into the summer hemisphere to 10°–15° relative to the No OHT case (Figs. 1c,d), where it is confined to within 5° latitude. Inclusion of the ocean heat transport also weakens the strength of both the cross-equatorial and summer cells; the cross-equatorial cell is weaker by about 30% in both seasons and the summer cell is essentially eliminated.

To understand how the ocean influences the seasonal Hadley circulation, let us first consider the impact of ocean heat transport on the seasonally varying SSTs. Figure 5 shows the zonal mean SST in the Control and No OHT experiments as a function of calendar month. In the Control run (Fig. 5a), the region of highest SST moves well into the summer hemisphere. Temperatures exceeding 27.5°C, the temperature above which convection generally occurs in the modern climate, extend to 10°S and 15°N in the Southern and Northern Hemi-

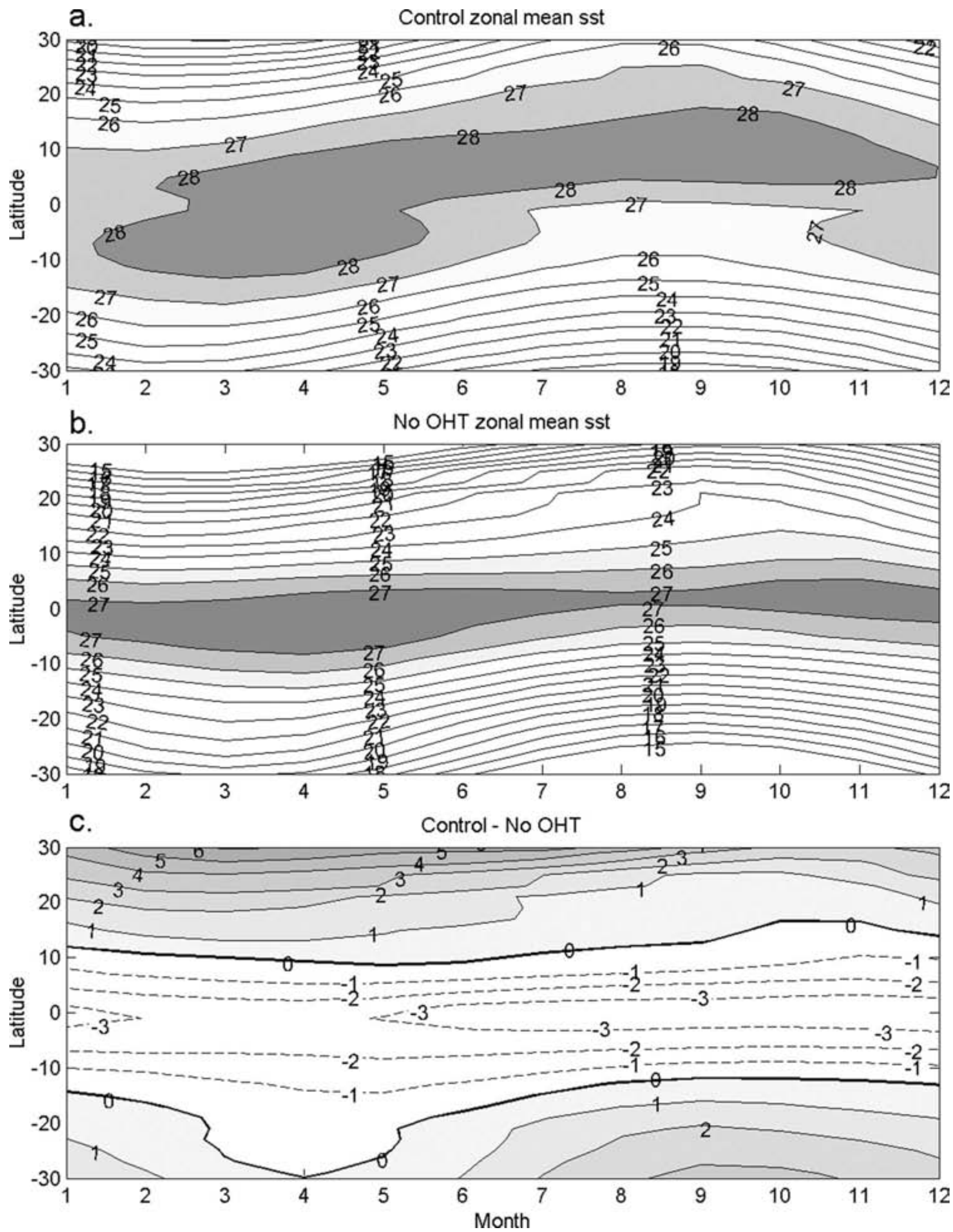

FIG. 5. The zonal mean SST for the (a) Control simulation and (b) No OHT simulation as a function of calendar month and latitude. The contour interval is 1°C and SSTs within 2° of the maximum SST are shaded. (c) The Control minus No OHT SSTs. The contour interval is 1°C and positive values are shaded. Note that in (c), a tropical mean warming between the Control and No OHT of 3.5°C has been removed in order to emphasize the influence of SST on the meridional structure of SST. This warming has been explained in Winton (2003) and Herweijer et al. (2005) as mainly due to a decrease in the marine stratocumulus cloud cover when ocean heat transports are included.

sphere summers, respectively. This corresponds with the regions of ascent (Fig. 6a), which form the upward branch of the Hadley circulation. In the No OHT case, the regions of maximum SST (Fig. 5b) and ascent (Fig. 6b) are tightly confined to the equator. As such, the main regions of oceanic convection move only slightly off the equator seasonally, consistent with the streamfunction and precipitation for that simulation.

The difference between the Control and No OHT SST (Fig. 5c) shows that ocean heat transport influences the SST in both an annual mean and a seasonal sense. To illustrate these different effects, let us return

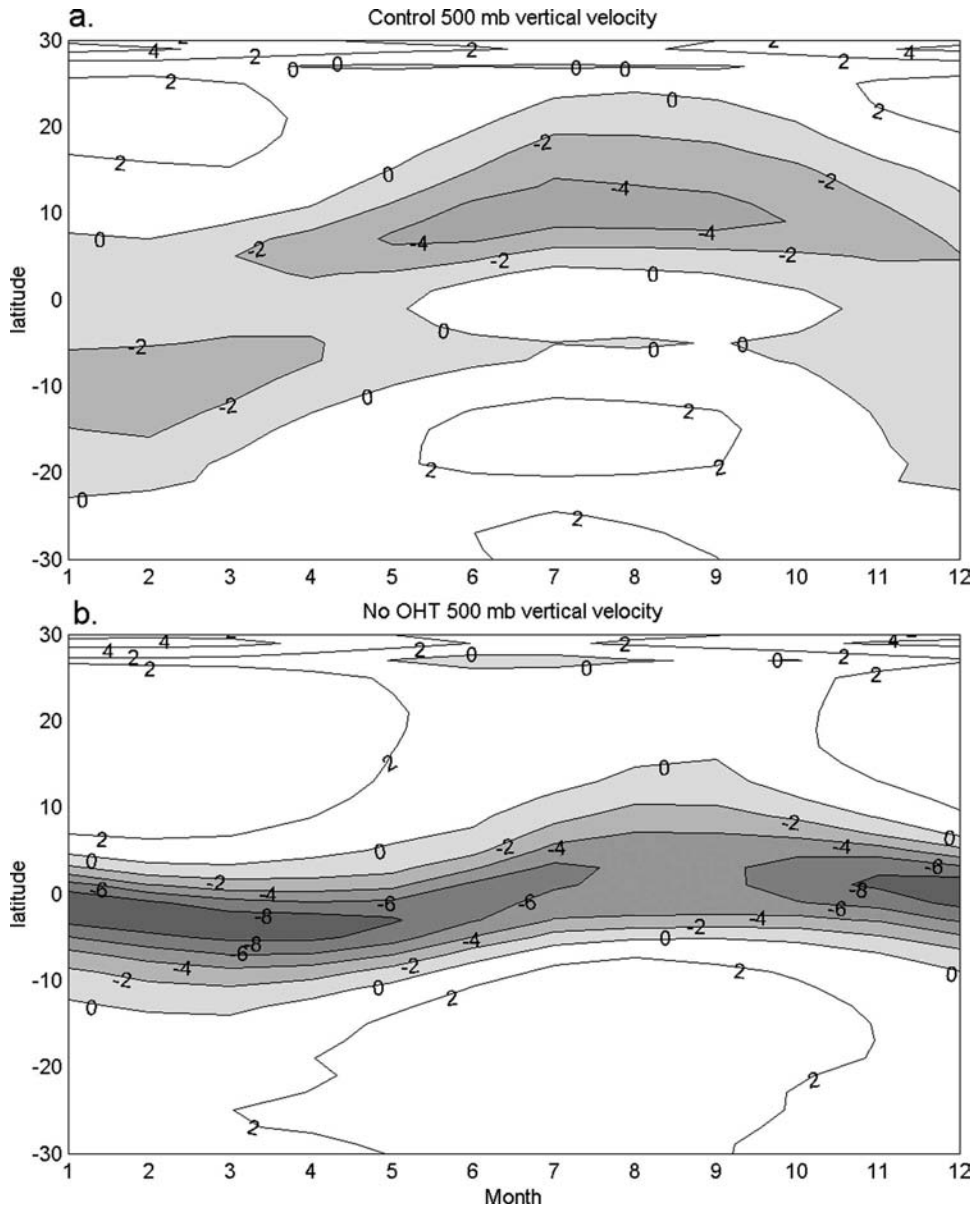

FIG. 6. The zonal mean 500-mb vertical velocity (10^{-2} Pa s^{-1}) for the (a) Control simulation and (b) No OHT simulation as a function of calendar month and latitude. The contour interval is 2×10^{-2} Pa s^{-1}. Negative regions (upward motion) are shaded.

to the mixed layer heat budget [Eq. (1)]. For the No OHT run, since D is zero everywhere, in the annual mean the net surface heat flux must also be zero. The addition of ocean heat transport in the Control run is primarily balanced by a change in the evaporation, so that in regions where the ocean diverges heat (e.g., the equatorial Pacific cold tongue), the evaporation is reduced (Seager et al. 2003). The resultant effect on SST can be illustrated with a simple calculation. The sensitivity of evaporation to surface temperature in the Tropics is approximately 12 W m^{-2} K^{-1}, assuming no change in drag coefficient or surface relative humidity (Hartmann and Michelsen 1993). Thus, if the model Q-flux (Fig. 3, bottom panels) is simply scaled by 12 W m^{-2} K^{-1} and is added to the No OHT SST (Fig. 5b), the result is an estimate of the effect of the annual mean ocean heat transport divergence on SST (with essentially no atmospheric feedbacks). The result is shown in Fig. 7a. Because the ocean circulation cools the equatorial region and warms the subtropics (Fig. 5c), it has the important effect of allowing the SST maximum to move off of the equator seasonally. There

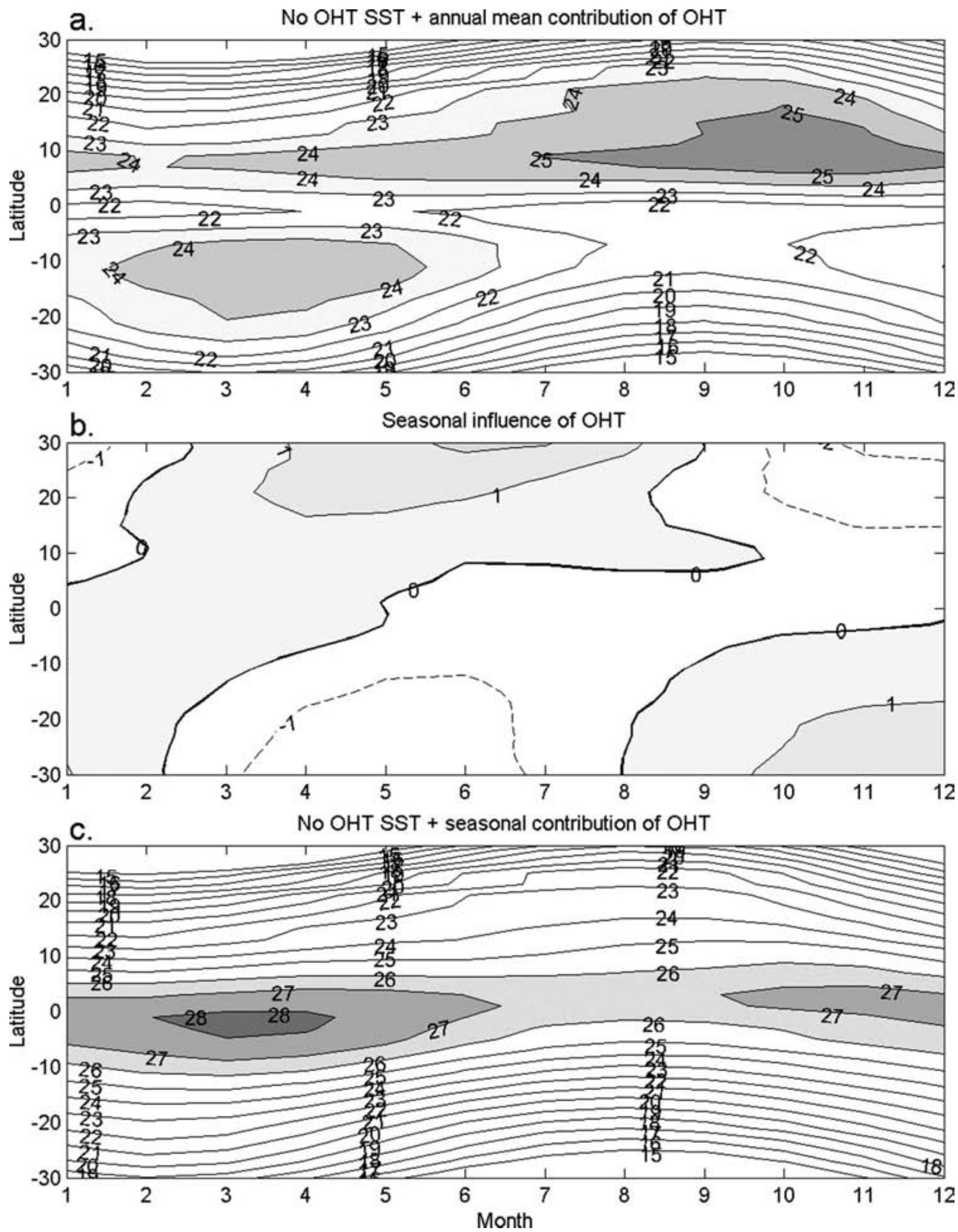

FIG. 7. (a) The No OHT SST plus the contribution of the SST by annual mean ocean heat transport. The contour interval is 1°C and SSTs within 2° of the maximum SST are shaded. (b) The calculated contribution to SST by seasonal ocean heat transport. The contour interval is 1°C and positive values are shaded. (c) Same as in (a) but for seasonal ocean heat transport.

are, of course, differences between the SST calculated this way (Fig. 7a) and the Control run SST (Fig. 5a) due to atmospheric feedbacks involving, for example, clouds and moisture convergence (Herweijer et al. 2005; Clement et al. 2005). However, it is clear that the signal of interest here, the off-equatorial maximum of SST, can be explained as the direct influence of the annual mean ocean heat transport divergence.

The annual mean ocean heat transport in the Tropics involves both Ekman and Sverdrup dynamics (Jayne and Marotzke 2001; Hazeleger et al. 2004; Boccaletti et al. 2004; Clement et al. 2005). Easterly wind stress in the Tropics drives equatorial upwelling and poleward Ekman drift across the entire basin, which transports warm water poleward. Also under easterly winds, the equatorial thermocline develops an east–west tilt. As such, in the eastern ocean basins, the upwelled subsurface water that replaces the water advected poleward

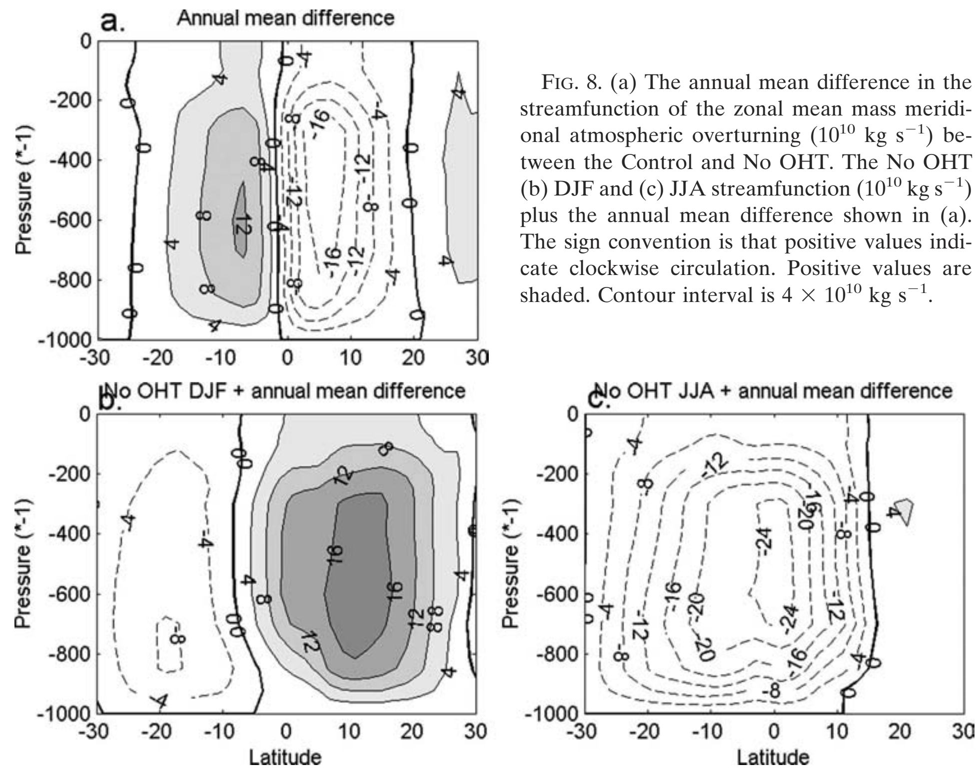

FIG. 8. (a) The annual mean difference in the streamfunction of the zonal mean mass meridional atmospheric overturning (10^{10} kg s^{-1}) between the Control and No OHT. The No OHT (b) DJF and (c) JJA streamfunction (10^{10} kg s^{-1}) plus the annual mean difference shown in (a). The sign convention is that positive values indicate clockwise circulation. Positive values are shaded. Contour interval is 4×10^{10} kg s^{-1}.

by Ekman drift is relatively cold. The shallow equatorial thermocline is fundamental to the ocean's ability to transport heat; cooling by upwelling in regions where the thermocline is shallow is compensated for by reduced surface evaporation, so the ocean gains heat from the atmosphere in those regions. The heat gain must be balanced in the annual mean by a poleward export of heat by the ocean from the Tropics (Boccaletti et al. 2004).

In addition to the annual mean transport, there is large seasonal variability in the ocean heat transport divergence both in the model and observations (Fig. 4). Jayne and Marotzke (2001) showed that this variability is dominated by a seasonally reversing Ekman flow rather than by surface temperature gradients. Those authors also showed that on a seasonal time scale, the ocean heat flux convergence is primarily balanced by the storage of heat in the mixed layer. The effect of seasonal ocean heat transport (defined as the departure from the annual mean) on SST can thus be calculated by integrating Eq. (1) over time with a balance between the left-hand side and D, the result of which is shown in Fig. 7b, assuming a fixed 50-m mixed layer depth. If this SST perturbation is then added to the No OHT SST (Fig. 5b), it is clear that the seasonal ocean heat transport divergence has little effect on the location of maximum SST (cf. Fig. 7c with Fig. 5b). This is because the seasonal ocean heat transport primarily affects the SST in the subtropics, not in the Tropics where the SST is high and where atmospheric convection occurs. Because of the assumption of a fixed-depth mixed layer depth, it is likely that the calculation of the SST response is an upper limit. Seasonal deepening of the mixed layer can offset temperature change in the storage term. This implies that perhaps the ocean heat transport does even less to the seasonal SST (and hence the Hadley circulation) than we have estimated here, supporting the interpretation that most of the influence is primarily through the annual mean component.

Next let us consider how the ocean heat transport influences the strength of the seasonal Hadley circulation. Previous studies have shown that the annual mean Hadley circulation is weakened by about 50% with the inclusion of ocean heat transport in climate models of many different vintages (Clement and Soden 2005), including the Manabe climate model (Manabe et al. 1979), Goddard Institute for Space Studies (GISS) model II (Hansen et al. 1983), Community Climate Model version 3 (CCM3; Kiehl et al. 1998), as well as the model used here (shown in Fig. 8a). It is the annual mean influence of the ocean heat transport that affects the seasonal Hadley circulation strength. This can be shown by simply adding the annual mean difference in the streamfunction between the Control and No OHT

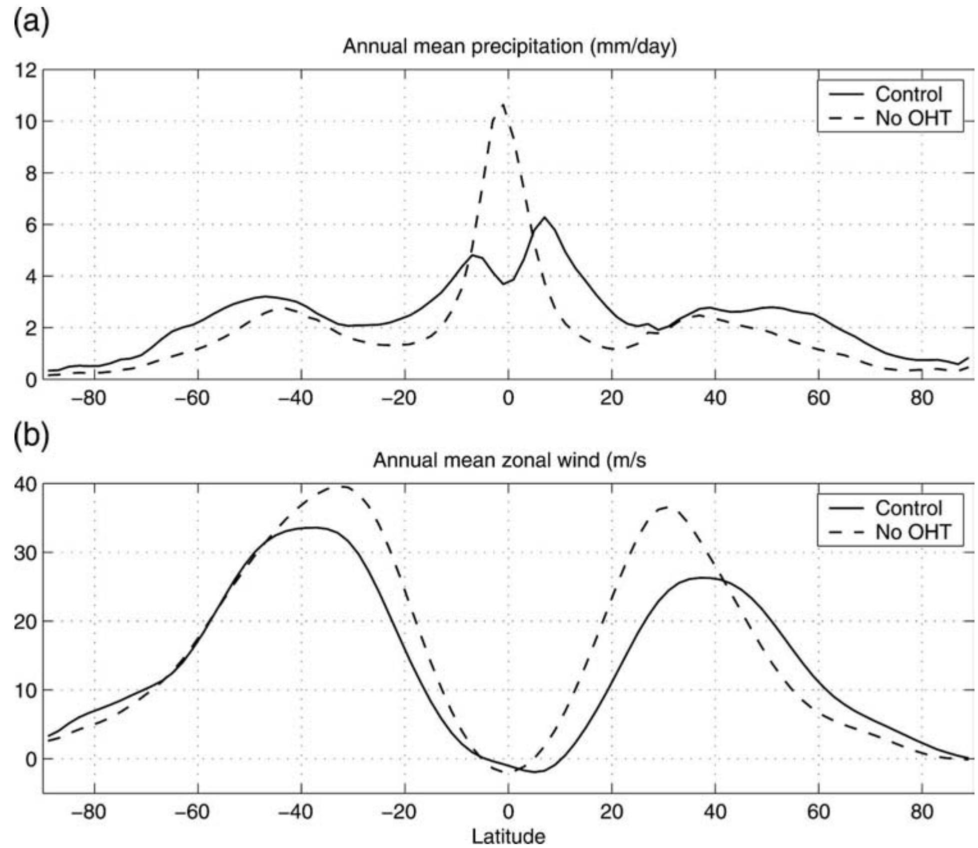

FIG. 9. The (a) annual mean, zonal mean precipitation (mm day^{-1}) and (b) zonal mean zonal wind at 200 mb (m s^{-1}) for the Control (solid line) and No OHT (dashed line).

(Fig. 8a) to the seasonal Hadley cells of the No OHT run (Figs. 1c,d). The result is shown in Figs. 8b,c. Comparing this with Figs. 1a,b, it is clear that the effect of the ocean heat transport on the seasonal Hadley cell intensity can be largely explained by the annual mean adjustment. This is consistent with the results presented above showing that the primary effect of the ocean heat transport on the SST, in particular the seasonal location of the SST maximum, is via the annual mean component. While the ocean transports a large amount of heat seasonally, the impact of this seasonal heat transport on the seasonal Hadley cell intensity appears to be minimal. The ocean moves heat from the summer hemisphere to deep into the winter hemisphere, but because of the heat capacity of the mixed layer, that heat is released into the atmosphere several months later, that is, in the equinoctial seasons (Fig. 7b). The impact of the seasonally varying ocean heat transport on SST is primarily in the subtropical latitudes, where the ocean has less ability to affect atmosphere convection because of the generally colder temperatures.

Because the effect of the ocean on the seasonal Hadley circulation is via this annual mean contribution, we can confine our discussion to the causes of the change in the annual mean Hadley circulation (Fig. 8a). Prior studies with simplified models can serve as a guide to explaining this result. Schneider (1984) and Hou and Lindzen (1992) showed that narrowing the atmospheric heating strengthens the Hadley circulation. Figure 9a shows clearly that the addition of OHT expands the region of convection from a narrow band centered on the equator to a broad band of convection. The zonally averaged model of Schneider (1984) also predicts that a narrowing of the convecting region will move the subtropical jets equatorward, which is consistent with the GCM results as shown in Fig. 9b. The results shown here are also consistent with prior GCM results. Hoskins et al. (1999) showed that the strength of the Hadley circulation depends sensitively on the curvature of the meridional SST structure. Those authors performed idealized experiments with specified SST patterns in which they found that the flatter the SST distribution, the weaker the Hadley circulation. Our results are consistent with these and suggest that the curvature of the SST can be related to physical processes, namely the ocean heat transport, which flattens the meridional SST distribution and weakens the cell. While there are other possible controls on the strength of the Hadley circulation (e.g., eddy heat and momentum fluxes), both simplified models and the GCM experiments show a consistent relationship between the width of the convective region and the strength of the Hadley circulation. Hence, it appears that by broadening the region of convection, the ocean heat transports weaken the Hadley circulation in the annual mean. This annual mean perturbation affects the strength of the seasonal Hadley cells in a simple additive way, weakening both the cross-equatorial and summer hemisphere cells.

4. Summary and discussion

Results from idealized experiments with a climate model suggest that ocean heat transport plays a fundamental role in setting the structure and intensity of the seasonal Hadley circulation. The ocean's influence can be understood primarily via annual mean considerations. By lowering equatorial SSTs and raising the off-equatorial SST in the annual mean, ocean heat transport allows the SST maximum and convection to move off the equator seasonally. This creates a cell that extends from deep in the summer hemisphere into the winter hemisphere subtropics. With regard to the strength of the cell, by broadening the region of maximum SST and the latitudinal extent of the convecting region of the atmosphere, again in an annual mean sense, ocean heat transport weakens the strength of the Hadley circulation in a manner consistent with that predicted by simple models. Seasonally, the ocean does move heat from the summer hemisphere to the winter hemisphere, where it is released to the atmosphere several months later. However, the heat release occurs at latitudes too high to significantly affect the Hadley circulation strength or intensity.

Of course, these results are subject to the caveat that they are based on model simulations, and models contain biases. However, the implied ocean heat transport in the model compares favorably with observations, suggesting that its influence is well represented for the simulation of the modern climate. Also, the fact that the change in annual mean Hadley circulation strength by the inclusion of ocean heat transports is similar in quite different models (Clement and Soden 2005) suggests that the annual mean influence of the ocean is not particularly model dependent. In other words, in all models, ocean heat transport cools the equatorial regions and warms the subtropics, which is essentially how the ocean affects the seasonal Hadley circulation.

The interpretation of the results presented here presupposes that the ocean influences the Hadley circulation primarily within the Tropics. However, because the ocean heat transport is set to zero everywhere in the global ocean, there are also changes in the extratropical climate between the Control and No OHT. These extratropical changes could influence the Hadley circulation through heat, moisture, and momentum fluxes between the subtropics and the midlatitudes (e.g., as described in Trenberth and Stepaniak 2003 or Walker and Schneider 2006). In the experiments performed here, we cannot distinguish the influence on the Hadley circulation of the ocean heat transport within the Tropics from that outside the Tropics. However, there are several pieces of evidence that support our interpretation of the primary control coming from the tropical oceans. First, it is only in the Tropics that the ocean and atmospheric heat transport are comparable; by 40° latitude, the ocean transports about 4–5 times less heat poleward than the atmosphere (Trenberth and Caron 2001). Hence, while the removal of the ocean heat transport is an enormous perturbation in the Tropics, the perturbation in the extratropics is small. Seager et al. (2002) have argued that the primary influence of the extratropical ocean on the atmospheric circulation is in the local storage and release of heat over the seasonal cycle and that the effect of ocean heat transport is of secondary importance for the extratropical atmospheric circulation. Second, results are consistent with simpler models (Schneider 1984; Hou and Lindzen 1992) and GCM experiments (Hoskins et al. 1999) that directly test the effects of changes in the tropical forcing. Thus, while it is plausible that conditions in the extratropics would affect some details of the Hadley circulation, to a first order, the changes are consistent with changes in the tropical oceans.

Discussion of results in the context of prior explanations of the seasonal Hadley circulation

Dima and Wallace (2003) have previously noted that while the seasonal Hadley circulation appears to be dominated by a single solsticial cell, there is actually an annual mean component, symmetric about the equator, that is present year-round and comparable in magnitude to the solsticial, asymmetric component. Our model results lend some physical insight into the presence of this year-round symmetric circulation. That is, the ocean and atmosphere transport heat poleward all year-round. The seasonal Hadley circulation is a perturbation to this that gives rise to a seasonally reversing cross-equatorial transport but does not alleviate the requirement to remove heat from the Tropics in all seasons. By relieving the atmosphere of some of the burden of carrying out this heat transport through its annual mean's influence, the ocean weakens the year-round symmetric component of the Hadley circulation. Also, the heat converged in the subtropics all year-round by ocean dynamics allows the SST maximum to move off the equator and the circulation becomes seasonally asymmetric with respect to the equator. In other words, the model results suggest that there is a trade-off between these different components of the circulation; when the annual mean component is reduced by the inclusion of ocean heat transport, it is done so in favor of a more asymmetric (solsticial) cell (cf. Fig. 1).

Similar conclusions were drawn by Lindzen and Hou (1988), but with some different outcomes. Those authors showed that displacing the maximum atmospheric heating from the equator results in a circulation that is asymmetric about the equator. However, one of the main conclusions of that study was that the farther the heating moves into the summer hemisphere, the stronger the circulation. Here the opposite was found; when ocean heat transport is included, the maximum heating moves into the summer hemisphere, but the cell is weakened. The results suggest that the reason for this discrepancy is that the width of the atmospheric heating is, in this case, a stronger control on the strength of the cell than the latitudinal displacement. Studies subsequent to Lindzen and Hou (1988) showed that the sensitivity of the cell strength to latitudinal displacement in the axisymmetric model can be different depending on assumptions about stationarity (Fang and Tung 1999) and boundary conditions (Walker and Schneider 2005). Further work in a consistent model framework is necessary to provide a more complete picture of the relative influences of different factors on the strength of the Hadley circulation.

Cook (2003) argued that continents also have a strong effect on the seasonal Hadley circulation. Experiments with a general circulation model that are analogous to those shown here were performed in which continental surfaces were removed to evaluate their effect. Those results showed that the inclusion of continental surfaces broadens the atmospheric heating and moves it deeper into the summer hemisphere, just as the ocean heat transport does. However, the inclusion of the continent in Cook's simulations had the effect of *strengthening* the cross-equatorial cell and weakening the summer cell, while our results showed a weakening of both cells. Although the effect of the continents appears to be consistent with the Lindzen and Hou (1988) results, Cook (2003) points out that the underlying mechanisms of this relationship are not included in the axisymmetric model. This, along with the results shown here, underscores the point that it is essential to understand the mechanisms that give rise to the seasonal Hadley circulation in order to understand its behaviors.

The results shown here also raise the issue of what the relative roles are of the ocean and land surfaces in setting the structure and strength of the seasonal Hadley circulation. Looking back at Figs. 1c,d, it is clear that even without ocean heat transport, there are some significant seasonal asymmetries about the equator. In particular, the summer hemisphere cells are considerably weaker than the cross-equatorial cells in both seasons, but most notably in JJA. Cook (2003) has argued that this weakening of the summer cell is due to monsoonal circulations that shift mass out of the summer subtropics. Our results support this interpretation. The zonal mean ascent (Fig. 6b) hardly shifts off of the equator seasonally without ocean heat transport. Axisymmetric theory would predict a fairly symmetric cell in response to this heating pattern (Lindzen and Hou 1988), unlike the full GCM response (Fig. 1d). It is presumably the inclusion of zonally asymmetric processes (i.e., monsoonal circulations due to continental surfaces) that can explain the strong reduction in the strength of the JJA summer cell in the GCM (Figs. 1c,d), as described by Cook (2003). Ocean heat transport further reduces the summer cell by weakening the annual mean circulation (Fig. 8a), so that it becomes almost nonexistent in JJA. On the other hand, ocean heat transport and land surfaces appear to have opposing effects on the strength of the cross-equatorial cell, with the inclusion of ocean heat transport weakening it and the inclusion of continents strengthening it. The processes by which these occur are entirely different; the ocean's influence is via the broadening of the convective region while the effect of the continents is via angular momentum constraints (Cook 2003). Overall, the model results shown here, together with the previous model results of Cook (2003), suggest that ocean dynamical processes have as important (albeit different) an effect on the seasonal Hadley circulation as monsoonal processes over land. Thus, the notion of the Hadley circulation being essentially a monsoonal phenomenon (Trenberth et al. 2000; Dima and Wallace 2003) does not appear to be correct. The relative influences of ocean and land on the seasonal Hadley circulation should be evaluated within the same modeling framework.

Finally, and perhaps most importantly, the results presented here imply that the Hadley circulation must be thought of as a coupled phenomenon. On one hand, the model experiments demonstrate that ocean heat transport has a strong influence on the structure and intensity of the seasonal Hadley circulation. On the other hand, the poleward ocean heat transport is obviously driven in part by the Hadley circulation through surface wind stress. The resulting balance sets the partitioning of poleward heat transport between the ocean and atmosphere and suggests that the Hadley circulation characteristics are related to this partitioning. In the model experiments shown here, the partitioning was essentially specified; in the Control run it is about equal between the ocean and atmosphere, and in the No OHT run it is 100% in the atmosphere. Are other configurations possible? There is currently no complete theory for what sets the partitioning, and little understanding of how it may change with climate change. Held (2001) has argued that the transport of mass by the atmosphere and the wind-driven circulation in the ocean must be approximately the same and that the resultant partitioning of energy transport can be qualitatively tied to quantities that are not likely to vary. However, there is a component to the ocean heat transport that is not directly wind driven and is related instead to the ocean thermal structure. In particular, Boccaletti et al. (2004) have argued that the mean depth of the equatorial thermocline is key to the ocean's ability to gain and transport heat and that that depth is set by global heat budget constraints. Philander and Fedorov (2003) have argued, for example, that at times in the past when the high latitudes and the deep ocean were much warmer than present, the equatorial thermocline was significantly deeper. This is consistent with a weaker ocean heat transport by the arguments of Boccaletti et al. (2004), which could alter the partitioning between the ocean and atmosphere and hence the Hadley circulation. This illustrates that a complete understanding of the seasonal Hadley circulation and its sensitivity to climate change must involve both ocean and atmospheric processes and their interactions in a global context.

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