Why is There an Evaporation Minimum at the Equator?*

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ABSTRACT

At all longitudes oceanic evaporation rates are lower on the equator than at latitudes to the north and south. Over the oceanic cold tongues this is related to the presence of cold water and divergence of heat by the ocean circulation. Herein is investigated why there is also a minimum over the Indo-Pacific warm pool. Model results confirm the recent suggestion of Sobel that deep convective clouds over the warm pool reduce the amount of solar radiation coming into the ocean that the evaporation has to balance. The results also confirm that this is only a partial explanation. Less evaporation over the warm pool than in the trade wind regions is also caused by an interaction between the ocean heat transport and the distribution of surface wind speeds. Low wind speeds over the warm pool reduce the latent heat flux and increase the SST, and stronger wind speeds in the off-equatorial regions of the Tropics increase the latent heat flux and cool the SST. Consequently, the wind speed distribution increases the meridional temperature gradient and increases the poleward ocean heat transport. Low latent heat fluxes over the warm pool can be sustained because the incoming solar radiation is partially offset by ocean heat flux divergence. Large values under the trade winds are sustained by ocean heat flux convergence. Climate models are used to show that, in equilibrium, wind speeds can only influence the latent heat flux distribution through their coupling to the ocean heat transport. In the presence of ocean heat transport, advection of moisture in the atmospheric boundary layer from the subtropics to the equator also increases the evaporation under the trade winds, but this has a much smaller effect than the wind speed or the cloud–radiation interactions.

1. Introduction

At all longitudes, evaporation from the ocean is less at the equator than at tropical latitudes to the north and south (Fig. 1). The lowest values are over the cold waters of the east Pacific cold tongue but even over the Indo-Pacific warm pool, the warmest ocean waters of the planet, the latent heat flux is 60 W m^{-2} less than that between 10° and 15° of latitude immediately north and south. This paper seeks to explain why.

Lower levels of evaporation over the warmest waters than over cooler waters nearby is surprising given the exponential dependence of the saturation specific humidity of air at the ocean surface, q_s, on its temperature, T_s, which closely follows the sea surface temperature (SST). Using the bulk formulation the latent heat flux Q_{LH} is estimated from

$$Q_{LH} = \rho c_E LU(1 - \widehat{RH})q_s(T_s), \tag{1}$$

where ρ is the air density, c_E is an exchange coefficient, L is the latent heat of vaporization, and U is the surface wind speed. The value $\widehat{RH}$ is equal to RH[$q_s(T_a)$/$q_s(T_s)$], where RH is the relative humidity, and $q_s(T_a)$ is the saturation specific humidity, of near-surface air. Here, $\widehat{RH} \sim$ RH because T_a and T_s are typically separated by less than 1K and we will use RH from now on. If the wind speed and RH are constant then the latent heat flux would increase with SST as $q_s(T_s)$ does.

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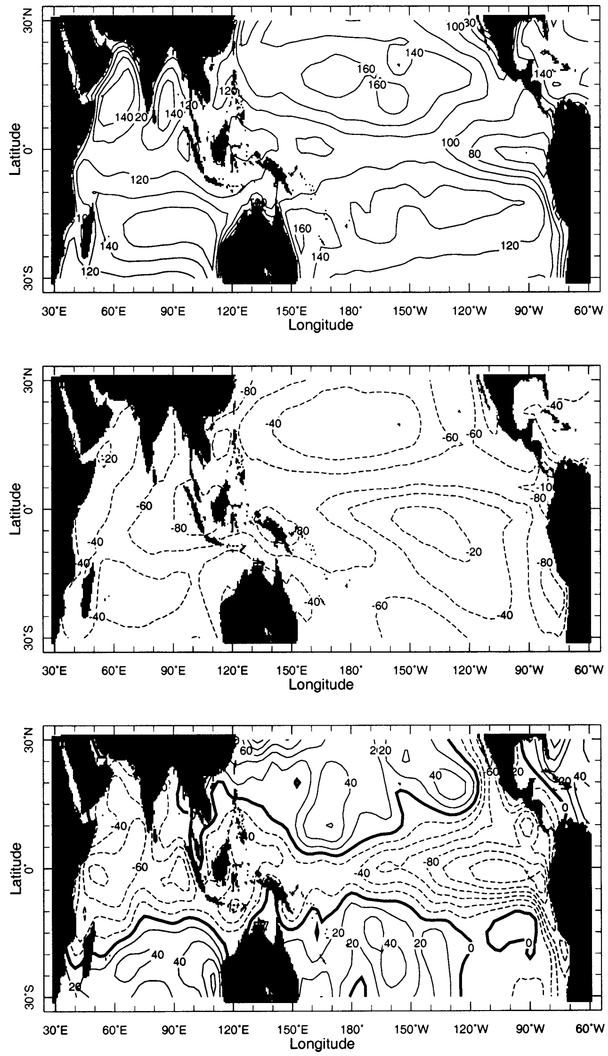

FIG. 1. (a) The NCEP–NCAR annual mean latent heat flux (Kalnay et al. 1996), (b) ERBE annual mean shortwave cloud forcing, and (c) Trenberth annual mean net surface heat flux, all in W m^{-2}.

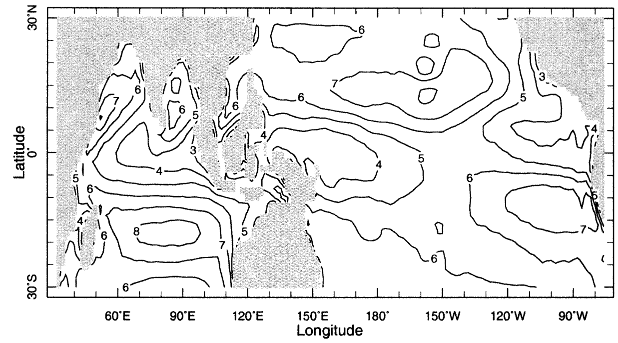

FIG. 2. The NCEP–NCAR annual mean surface wind speed in m s^{-1}.

Over the cold tongue low levels of evaporation are caused by cooling from upwelling and ocean heat flux divergence. In contrast, the low levels of evaporation over the west Pacific warm pool are usually explained in terms of the collocated wind speed minimum (Zhang and McPhaden 1995; Peixoto and Oort 1992) (Fig. 2) and/or higher relative humidity (e.g., McGregor and Nieuwolt 1998). Sobel (2003) correctly points out that the wind speed explanation neglects the requirement for energy balance in the ocean mixed layer that can be written as

$$\mathrm{OP} + Q_{\mathrm{SW,sfc}} = \rho c_E L U(1 - \mathrm{RH}) q_s(T_s) + Q_{\mathrm{LW,sfc}} + Q_{\mathrm{SH}}. \qquad (2)$$

Here, OP is the ocean heat flux convergence, $Q_{\mathrm{SW,sfc}}$ is the net downward surface solar radiation, Q_{LH} is the latent heat flux, $Q_{\mathrm{LW,sfc}}$ is the net surface longwave radiation and Q_{SH} is the sensible heat flux, these three defined positive upward. Sobel assumes that OP is small in the warm pool and that Q_{SH} is small everywhere. In addition he assumes that spatial variations of $Q_{\mathrm{LW,sfc}}$ are small. In this case the spatial variations, denoted by $(\cdot)^*$, are related by

$$Q^*_{\mathrm{SW,sfc}} = \rho c_E L[U(1 - \mathrm{RH}) q_s(T_s)]^*. \qquad (3)$$

The net surface solar radiation is reduced over the warm pool by highly reflective deep convective clouds (Ramanathan and Collins 1991). This is seen in Fig. 1b, which shows the shortwave cloud forcing at the surface (defined here as the observed flux minus the clear sky flux) according to the Earth Radiation Budget Experiment (ERBE) data of Li and Leighton (1993). According to the simplified surface budget of Eq. (3) the low evaporation over the warm pool occurs because it has to balance less incoming solar radiation. In this case the variations of the latent heat flux have to match those of the net surface solar radiation regardless of the wind speed. Since $q_s(T_s)$ is a maximum over the warm pool, if the wind speed was not a minimum RH would have to be a maximum.

As discussed by Sobel, referring to Neelin and Held (1987), convection and precipitaton require a net vertical energy flux into the atmosphere column. According to Sobel, convection and precipitation can persist as cloud cover lowers the latent heat flux into the atmosphere column because there is a compensating reduction of longwave radiation to space by the clouds. At the ocean surface the reduced evaporation is balanced by reduced surface solar radiation. Further, in the absence of cloud absorption, the reduction of surface solar radiation by increased cloud cover equals the increase of solar radiation reflected to space. Putting all these equalities together, Sobel's explanation meets the additional observational constraint that the shortwave and longwave cloud forcings at the top of the atmosphere cancel (Ramanathan et al. 1989). However, there are reasons for thinking that this is not the end of the story.

In all fairness Sobel was primarily considering the variations along the equator, but if we extend our field of vision we see that the spatial variations in shortwave cloud forcing at the surface (Fig. 1b) are too small to explain the variations in latent heat flux. For example, because of clouds, the ocean under cloudy skies in the off-equatorial regions of the Tropics receives some 40 W m^{-2} more solar radiation than in the warm pool. This difference will be opposed by the lesser clear sky radiation (because of the greater distance from the equator). The latent heat flux away from the equator is, however, up to 60 W m^{-2} greater than in the warm pool. Also, at each longitude, there is a latent heat flux minimum on the equator, which would not be predicted by the shortwave cloud forcing.

However, the spatial variations of the latent heat flux do closely match the pattern of the annual mean net surface heat flux out of the ocean [shown in Fig. 1c using data from Trenberth et al. (2001)] which, in equilibrium, equals the convergence of heat by the ocean. In Fig. 3 we show scatterplots of the latent heat flux and the convergence of heat by the ocean using different data sources. The relationship is closest when using both latent and net surface heat flux from the same dataset (Fig. 3b) but is anyway clear: where the ocean diverges heat (around the equator) the latent heat flux is small, where it converges heat (poleward of about 15°N and S) the latent heat flux is large. In this view the warm pool latent heat flux is unexceptional: with its 20–40 W m^{-2} ocean heat flux divergence it sits in between the very low values of the cold tongue, where the ocean heat export is larger, and the large values of the extratropics where the ocean converges heat. The relationship between the latent heat flux and the ERBE surface shortwave cloud forcing is weaker (Fig. 3c).

In the current paper we will use atmosphere and ocean models to examine the relative roles of cloud–radiative feedbacks and ocean heat transport in determining the distribution of latent heat flux in the Indo-Pacific Oceans. We begin, in section 3, with a series of model experiments with an ocean general circulation model (GCM), coupled to a simple model of the atmospheric mixed layer, designed to test the impact of spatial variations in cloud cover and surface wind speed on the spatial distribution of the latent heat flux. We also examine the impact of moisture advection in the atmosphere. In section 4 we discuss how, because the ocean moves heat, the wind speed distribution actually is important in maintaining the equatorial minimum of latent heat flux. However, atmospheric processes are also important—both the reduction of solar radiation by deep convective clouds and, to a lesser extent, the advection of humidity from cooler regions to warmer regions. In section 5 we examine the spatial pattern of latent heat flux in an atmosphere general circulation model coupled to a mixed layer ocean that neglects ocean heat transport thus isolating the impact of radiative feedbacks. Our conclusion that the equatorial evaporation minimum is explained by an interplay between the atmosphere and ocean is offered in section 6.

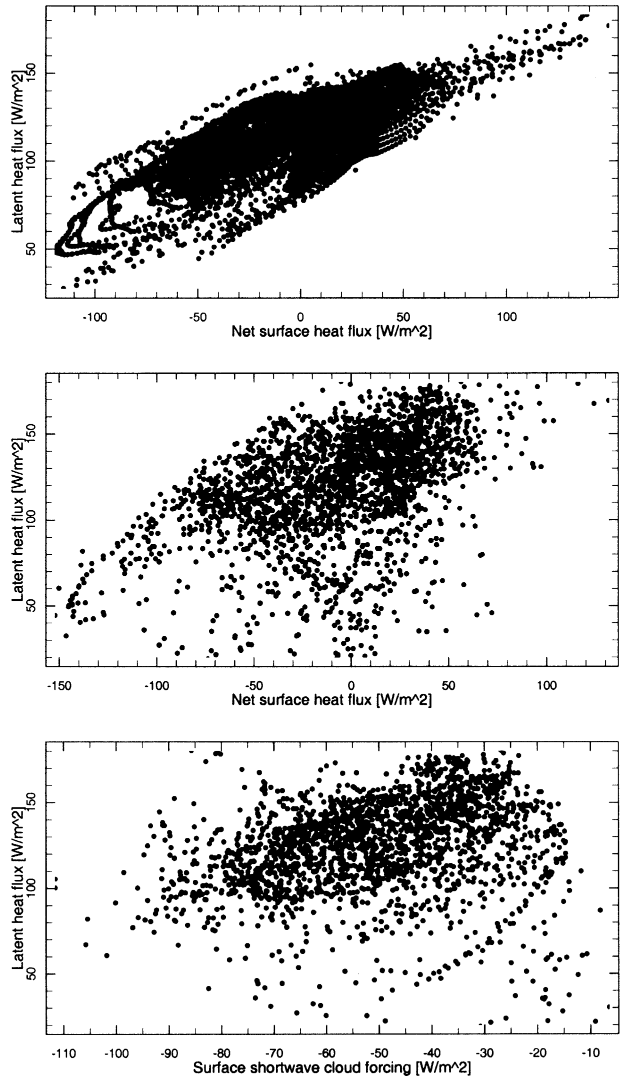

FIG. 3. Scatterplots of the annual means of the net surface heat flux, which equals the ocean heat flux convergence, vs the latent heat flux for the Indian and Pacific Oceans between 30°N and 30°S. (a) The data of da Silva et al. (1994) and (b) NCEP–NCAR reanalysis latent heat flux and the net surface heat flux of Trenberth et al. (2001). (c) Scatterplot of the annual mean NCEP latent heat flux vs the annual mean ERBE shortwave cloud forcing at the surface for the same domain.

2. Ocean model description

The ocean GCM is the reduced gravity, primitive equation, sigma-coordinate model of Gent and Cane (1989). The vertical structure of the ocean model consists of a mixed layer and 14 layers below according to a sigma coordinate (Murtugudde et al. 1996). The mixed layer depth and the thickness of the lowest sigma layer are computed prognostically and the remaining layers are computed diagnostically such that the ratio of each sigma layer to the total depth below the mixed layer is held to its prescribed value. The model uses an A-grid structure with a resolution of 1/2° latitude and 1/2° longitude. The model domain extends from 30°N to 30°S and from 32°E to 76°W. Fourth-order central differences are employed in the horizontal with second-order central differences in the vertical. The Lorenz (1971) N-cycle scheme is used for time integration and a high-order, scale-selective Shapiro filter provides horizontal friction. The model Indonesian throughflow (ITF) is computed as the westward flow between Australia and Java in the top 400 m (Murtugudde et al. 1998) and is found to reproduce the seasonal to interannual variability of the ITF and its effects on the dynamics and thermodynamics of the Indo-Pacific region.

The model mixed layer is based on a hybrid vertical mixing scheme (Chen et al. 1994) that combines the advantages of the traditional bulk mixed layer model of the Kraus and Turner (1967) type with the dynamic instability model of Price et al. (1986). Complete hydrology has been added to the model (Murtugudde and Busalacchi 1999) with freshwater forcing treated as a natural boundary condition (Huang 1993). The UNESCO equation of state is used for computing buoyancy from salinity and temperature. A sponge layer is utilized at the north and south boundaries of the model domain, with the model damped toward the climatological values of Levitus and Boyer (1994).

Surface heat fluxes are computed by coupling the ocean GCM to an advective atmospheric mixed-layer (AML) model (Seager et al. 1995) as reported in (Murtugudde et al. 1996). The AML represents either a dry convective layer or the mixed layer that underlies shallow marine clouds. Within the mixed layer, the air temperature and air humidity are determined by a balance between surface fluxes, horizontal advection by imposed winds, entrainment from above the mixed layer, horizontal diffusion and, for temperature, radiative cooling. Once the air temperature and humidity are determined, surface sensible and latent heat fluxes (and thus evaporation for the freshwater forcing) can be calculated in terms of the ocean model SST and the imposed winds. To compute the longwave radiative heat loss from the surface we use a standard bulk formula and observed cloud cover (Seager and Blumenthal 1994). To compute the surface solar radiative forcing, we use the formula of Reed (1977) and a specified cloud cover. For the control case the cloud cover is taken from observed values estimated from the International Satellite Cloud Climatology Project (ISCCP) data (Rossow and Schiffer 1991). To test the role of cloud cover, in other experiments we replace the ISCCP cloud cover in the Reed formula with a constant cloud cover in time and space. In the computation of the net surface heat flux, only the cloud cover and winds are specified. In the experiments performed we examine how the latent heat flux distribution is influenced by the cloud and wind speed distribution. The quantities that the ocean has some direct control over, that is, air temperature and humidity, are modeled internally in terms of the SST, the winds, and the values of the air temperature and air humidity at the continental margins.

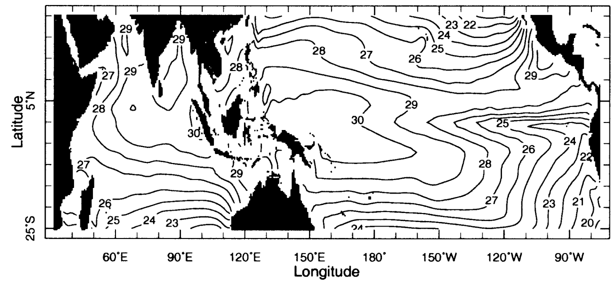

FIG. 4. The annual mean SST (°C) for the control case with realistic spatially and temporally varying cloud cover and wind speed.

3. Results from the ocean model experiments

We conduct a number of experiments with the ocean GCM–AML model, beginning with a control simulation and then testing how the distribution of latent heat flux is influenced by spatial variations of cloud cover, wind speed, and relative humidity. Each experiment is forced by climatological surface winds, wind speeds and stresses from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalyses (Kalnay et al. 1996) and integrated for 25 yr. The results shown are from the last year of the integration.

a. The case with realistic spatially and temporally varying cloud cover and wind speed

First we conducted an experiment with realistic surface forcing using climatological ISCCP cloud cover for the 1983–91 period and spatially and temporally varying wind speeds in the surface flux calculation. The model simulation of SST (Fig. 4) is quite realistic with a well-defined Indo-Pacific warm pool. The latent heat flux is shown in Fig. 5a. At the equator, and at all longitudes, it is less than at higher latitudes. In the longitudinal domain of the warm pool, the model simulates lower latent heat flux over the warmest waters than over the cooler waters north and south. This is correct although the spatial variations are weaker than observed.[1] We now proceed to test how this distribution arises.

b. The case of uniform cloud cover

To test the extent to which alteration of the net surface solar radiation by clouds impacts the latent heat flux distribution we replaced the ISCCP cloud cover with a uniform fractional cloud cover of 0.59 (the time and

[1] The weaker spatial variation may be due to the fact that the spatial variations of wind speed in the NCEP–NCAR reanalysis are too small. For example, they are weaker than those in the European Centre for Medium-Range Weather Forecasts reanalysis.

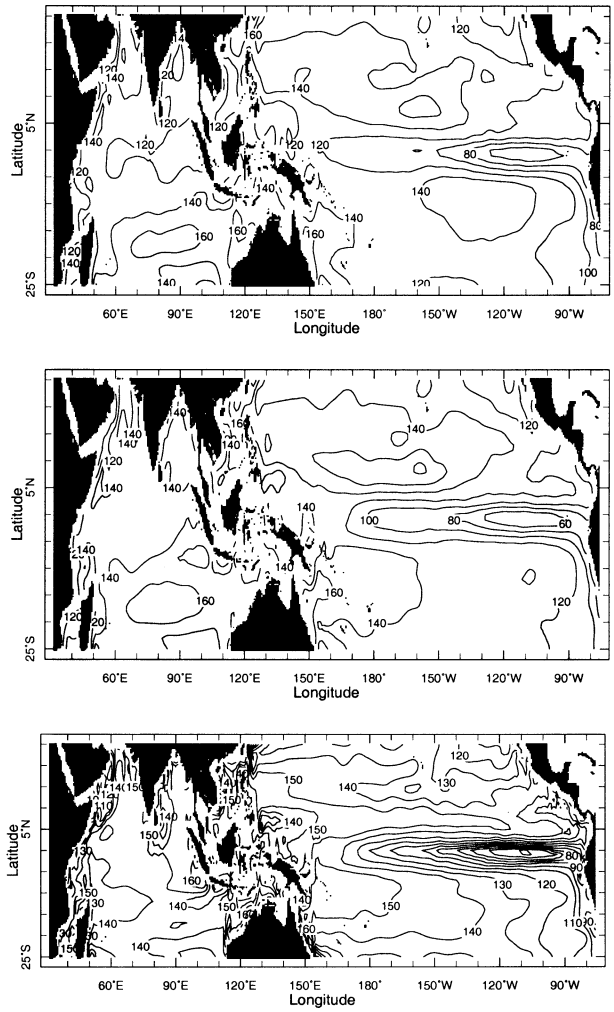

FIG. 5. Annual means of (a) the latent heat flux in W m^{-2} for the control case with variable cloud cover and wind speed, (b) the case with uniform fractional cloud cover of 0.59, and (c) the case with uniform cloud cover and wind speed. (c) Note that the contour interval is half that in (a) and (b).

area mean of the ISCCP cloud cover over the tropical Indo-Pacific domain). Reduced cloud cover causes the Indo-Pacific warm pool to warm (not shown, but see Seager et al. 1988). However, as shown in Fig. 5b, the latent heat flux over the Indo-Pacific warm pool remains lower than over cooler waters north and south even though this minimum is less marked. In this case the net surface solar radiation has a maximum on the equator so the evaporation minimum there must be sustained by the wind speed minimum and/or spatial variations in the modeled surface relative humidity. Because the evaporation minimum is less marked than in the case with realistic cloud cover, this confirms the suggestion of Sobel (2003) that reduced net surface solar radiation assists creation of the evaporation minimum. That the minimum remains at all confirms that Sobel's explanation is only partial.

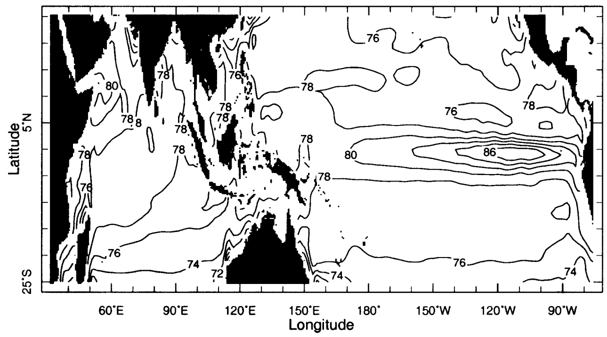

FIG. 6. The annual mean percent relative humidity for the case with uniform fractional cloud cover of 0.59 and uniform wind speed.

c. The case of uniform cloud cover and uniform wind speed

In this case we retain the uniform cloud cover but also replace the time- and space-varying wind speed with a uniform wind speed of 6.5 m s^{-1}, thus removing the influence of wind speed variations on the latent heat flux. The wind stresses applied to the ocean were not changed. Strengthening of the winds over the Indo-Pacific warm pool causes cooling, while weakening of winds under the trade winds causes warming, with the net effect that the warm pool broadens relative to both of the previous two cases (not shown).

In this case, with both uniform cloud cover and wind speed, the evaporation still remains less over the Indo-Pacific warm pool than over areas to the north and south (Fig. 5c). However the minimum is now very weak (note that the contour interval in Fig. 5c is half that in Figs. 5a or 5b). Since it now cannot be due to either a wind speed minimum or a solar radiation minimum it must be caused by spatial variations in relative humidity.

According to the AML model of Seager et al. (1995) used to compute the surface fluxes, the latent flux is given in steady state by

$$\rho c_E L U[q_s(T_s) - q_a] = \rho L h \underline{u} \cdot \nabla q_a + \rho c_E L U \mu q_a, \quad (4)$$

where q_a is the specific humidity in the AML. The first right-hand side term represents moisture advection within the atmosphere mixed layer and the second right-hand side term parameterizes turbulent mixing at the top of the AML and $\mu = (1/\delta - 1)$ where δ is the relative humidity for local equilibrium (see section 3d). The mixing term removes humidity from the AML and increases evaporation where the humidity is high. On the other hand the advection of moisture increases evaporation in the trade winds where the winds blow from drier air to moister air, but has less impact near the equator in the Indo-Pacific warm pool region where winds are light. This causes the latent heat flux to be larger away from the equator. This is confirmed by looking at the modeled annual mean relative humidity for this experiment (Fig. 6). The relative humidity is lower in the trade wind regions, where there is strong dry

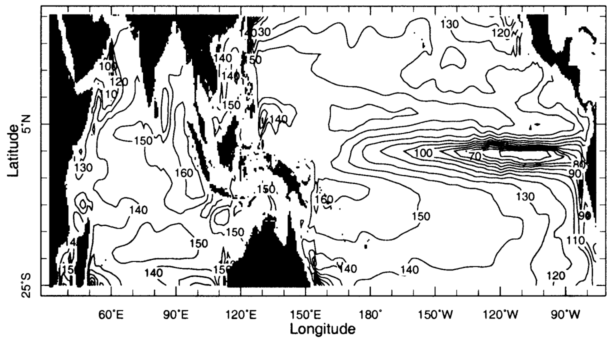

FIG. 7. The annual mean latent heat flux in W m^{-2} for the case with uniform fractional cloud cover of 0.59, uniform wind speed, and local boundary layer equilibrium.

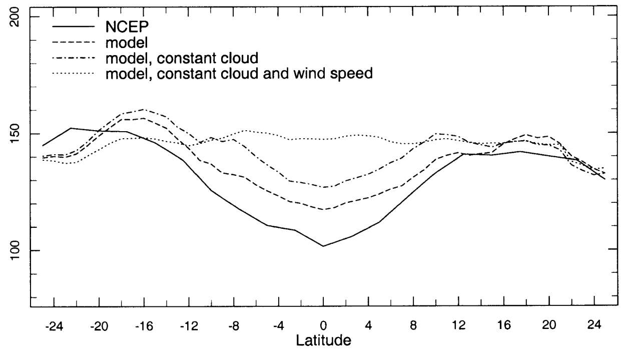

FIG. 8. The annual mean latent heat flux in W m^{-2}, zonally averaged over the Indo-Pacific warm pool between 60°E and the date line for observations from NCEP, the ocean GCM − AML model with realistic cloud cover and wind speeds, the case with uniform cloud cover, and the case with uniform cloud cover and wind speed.

advection, and higher over the Indo-Pacific warm pool (and over the cold tongue where moister air advects in from warmer surrounding areas). The spatial variation of relative humidity is not very large, which is why the spatial variations of the latent heat flux in this experiment are also weak.

d. The case of uniform cloud cover, uniform wind speed, and local boundary layer equilibrium

As our final ocean model experiment we simplify the atmosphere even further by removing advection, by setting $\underline{u} \cdot \nabla q_a = 0$ in Eq. (4), thus imposing a local boundary layer equilibrium between evaporation and entrainment of drier air at the top of the atmosphere mixed layer. The AML model retains a diffusion (over the open ocean only) of specific humidity and temperature to hold in check a possible positive feedback between the AML and the OML (see Murtugudde et al. 1996). This introduces some nonlocality but, putting this aside, according to Eq. (4) the relative humidity becomes $\delta = (1 + \mu)^{-1}$ and the latent heat flux becomes

$$Q_{\mathrm{LH}} \propto (1 - \delta)q_s(T_s), \tag{5}$$

where the proportionality holds because of the uniform wind speed. In this case the latent heat flux can only increase with SST and $q_s(T_s)$. This is indeed the case as shown in Fig. 7. The lack of strong trade winds north and south of weaker winds, and the lack of equatorward advection of moisture, causes the area of warm water to expand well away from the equator. The latent heat flux maxima shown in Fig. 7 are over the warmest waters of this deformed SST field.

e. The relative contribution of cloud–solar radiation interactions and wind speed to the evaporation minimum over the warm pool

The evaporation minimum over the cold tongue is no surprise and is caused by upwelling, cooling of the surface ocean, advection and horizontal mixing of warmer and moister air from surrounding regions, and strong poleward divergence of heat by ocean currents. Over the warm pool, reduced solar radiation and weak winds are the direct causes of the evaporation minimum. In Fig. 8 we show latitudinal transects, averaged over the Indo-Pacific warm pool between 60°E and the date line, of the observed latent heat flux from NCEP–NCAR reanalysis and the latent heat flux from three experiments: the control case, the case with uniform cloud cover and the case with both uniform cloud cover and wind speed. As mentioned before, the model slightly underestimates the strength of the equatorial minimum. However, on the basis of these results, the impact of the wind speed minimum on the latent heat flux minimum is twice the size of the radiative impact. This proportionality should not be taken too seriously being dependent on the choice of bulk formula in the ocean GCM–AML model. However, the conclusion to be drawn is that the wind speed minimum is at least a significant factor, and perhaps the most significant one, that through its interaction with the ocean heat transport causes an evaporation minimum over the warmest waters.

4. On the importance of the ocean heat transport

Three processes have been identified that contribute to the small latent heat flux over the warm waters of the Indo-Pacific warm pool. These are reduced surface solar radiation, the minimum in wind speed, and the advection of humidity in the atmospheric boundary layer. Of these three the latter is the least important while the first two appear to be of comparable significance. Reduction of the surface solar radiation reduces the amount of latent heat flux required to bring balance into the net surface flux. This mechanism does not require ocean heat transport to be effective. In contrast a minimum in the wind speed does not change the other terms (mainly solar radiation) in the net surface heat budget that the latent heat flux has to balance. In the absence of ocean heat transport a wind speed minimum will cause the SST to be higher so that the latent heat flux

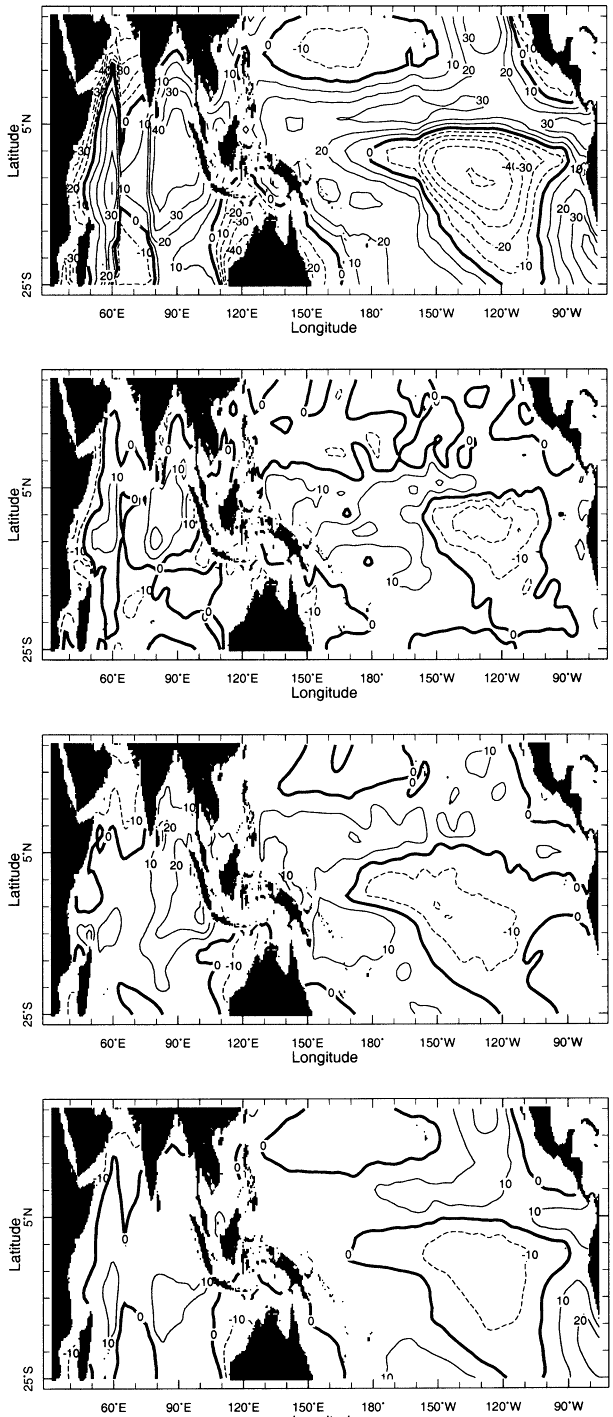

FIG. 9. Change in the annual means of (a) net surface solar radiation, (b) the net surface heat flux (equal to the ocean heat flux divergence), (c) the latent heat flux, and (d) the net surface longwave flux, all in W m^{-2}, for the case with uniform fractional cloud cover of 0.59 minus the case with ISCCP cloud cover.

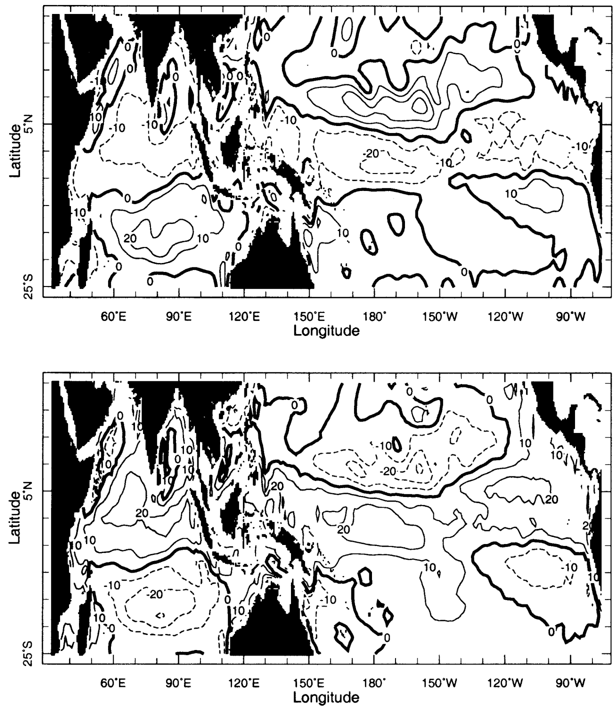

FIG. 10. Change in the annual means of (a) the net surface heat flux (equal to the ocean heat flux divergence), and (b) the latent heat flux for the case with uniform wind speed minus the case with realistic varying wind speed, both with uniform fractional cloud cover of 0.59. Units are W m^{-2}.

can still attain the value needed to balance the other terms in the net surface heat flux. Matters are different when the net surface heat flux balances ocean heat flux divergence or convergence. In that case, a wind speed minimum *can* cause reduced latent heat flux. Low winds over the warmest waters will cause ocean warming but the ocean heat divergence from the region will also increase, such that balance is restored with a smaller latent heat flux. For the same reason atmospheric advection can only cause a minimum latent heat flux over the warmest waters in the presence of ocean heat export.

In Fig. 9 we show the difference in net surface solar radiation for the case with uniform cloud cover minus the case with varying cloud cover together with the change in the net surface heat flux (which is of the sign that it equals the change in ocean heat flux *divergence*), the change in latent heat flux, and the change in net surface longwave radiation. Reducing the cloud cover over the Indo-Pacific warm pool causes 20–30 W m^{-2} more solar radiation to pass into the ocean. This heating is balanced by a 10 W m^{-2} increase in ocean heat flux divergence, 10 W m^{-2} increase in latent heat flux, and the remainder by increased net surface longwave loss (which, although a smaller term, cannot justifiably be neglected).

Figure 10 shows the change in ocean heat flux divergence and latent heat flux for the case of shifting to a uniform wind speed minus the case with realistic wind speeds. In this case there is no change in solar radiation and the change in net surface longwave radiation is tiny because the cloud cover remains the same. Consequently, increased latent heat flux over the warm pool, and reduced latent heat flux in the trade wind regions, is balanced by reduced ocean heat flux divergence in the warm pool and reduced convergence under the trades, that is, reduced poleward heat transport. This result makes clear that stronger wind speeds in the core of the trade wind regions (between 10° and 20°N and S) than in the equatorial region explains a significant fraction of the total tropical Pacific Ocean heat transport [as can easily be inferred from Eq. (2)].

The case of a change in deep convective cloud cover is worth considering. Increased deep convective cloud cover reduces the solar radiation entering the warmest regions of the tropical oceans and will cause a reduction in ocean heat export from that region. However, since the deep convective cloud cover does not impact the net radiation balance at the top of the atmosphere, the total—atmosphere plus ocean—export from the region must be the same. Consequently, the atmosphere heat export must increase, shifting the partitioning of the total transport towards the atmosphere. This disagrees with the suggestion of Held (2001) that the partitioning cannot alter because of the couplings between the mean meridional overturnings of the atmosphere and ocean and between the SSTs and the air temperature and humidity. One resolution is that our reasoning is incomplete because we fail to take into account how the atmosphere's mean meridional overturning will respond to the cooling of the warmest SSTs and also fail to account for the ocean overturning's subsequent response. Another possible resolution is that the ocean heat transport is a sum of transports by a variety of processes, of which the wind-driven mean overturning is just one, and that the partioning may be able to change (Hazeleger et al. 2003). More work is needed to resolve this issue.

The ocean heat transport variations also help explain how deep convection persists even as the local evaporation is decreased by the cloud cover. When the cloud cover is greater, the reduced surface solar radiation is partially offset by a smaller ocean heat flux divergence. Consequently, the latent heat flux does not need to be reduced by as much as Eq. (3) implies. The latent heat flux reduction is therefore less than the reduction by the clouds of the outgoing longwave radiation, and the net vertical energy flux into the atmosphere column is *increased* as the cloud cover is increased. This allows convection to persist.

5. The latent heat flux minimum in an atmosphere–ocean model that neglects ocean heat transport

The ocean model results highlighted the role that ocean heat transport plays in creating low levels of latent heat flux in the warm pool region. However, the surface solar method of creating the minimum could work in the absence of ocean heat transport. We tested this by examining the latent heat flux in an atmosphere GCM coupled to a uniform-depth ocean mixed layer in which the ocean heat transport is set to zero and in which, therefore, the SST is determined as that required for the annual mean net surface heat flux to be zero (Clement and Seager 1999). The atmosphere GCM is the Goddard Institute for Space Studies model used by Clement and Seager (1999).[2]

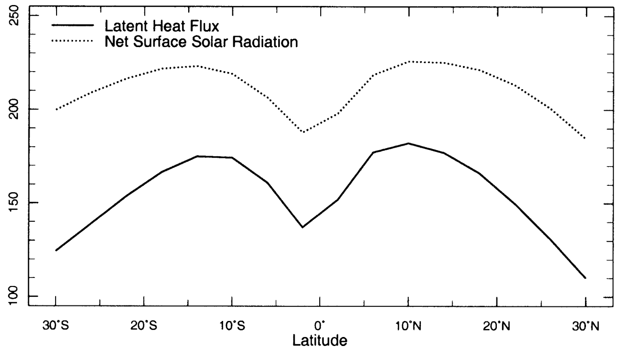

FIG. 11. The annual mean net surface solar and latent heat flux in W m^{-2} from an atmosphere GCM coupled to a mixed layer ocean in which the ocean heat transport is set to 0.

When the ocean heat transport is removed the deep convection in the Tropics collapses onto, or close to, the equator. Because there is no ocean heat transport Eq. (3) holds (if the change in net surface longwave radiation is small enough) and the variations within the Tropics of the latent heat flux must balance those of the net surface solar radiation regardless of the wind speed. In Fig. 11 we show latitudinal transects of the zonal mean, averaged over ocean grid points only, of surface latent heat flux and net surface solar radiation. There is an equatorial minimum in both and of the same magnitude. Clearly in this case, which is closest to the conceptual situation of Sobel (2003), the evaporation minimum is caused by the solar radiation minimum induced by the deep convective cloud cover. Because in this unrealistic case all the deep convection occurs in a narrow strip around the equator, the equatorial latent heat flux is 60 W m^{-2} less than to the north and south, which is as large as the observed variation. The results confirm that under some circumstances—in this case, unreal ones—cloud radiation feedbacks alone can cause a collocation of an evaporation minimum with maxima of SST and precipitation.

[2] We obtained very similar results performing the same experiment with the National Center for Atmospheric Research Community Climate Model (Kiehl et al. 1998), run with T42 resolution and 18 vertical layers.

6. Conclusions

It has been shown that there are three ways in which lower rates of evaporation over warm equatorial waters than over cooler waters to the north and south can be sustained. These are as follows:

1) Extensive deep convective cloud cover over the warmest water reduces the amount of surface solar radiation that the latent heat flux needs to balance (Sobel 2003).
2) Lower wind speeds over the warmest water reduces the latent heat flux and, because it also causes the ocean to warm, can be balanced by increased ocean heat flux divergence.
3) Advection of drier air from the subtropics to the warm equatorial regions reduces the relative humidity off the equator, increasing evaporation, and increases it over the warmest waters, reducing evaporation.

Each of these common explanations is only partial. Of the three it is the first two that are most important. The first one works in the presence or absence of ocean heat transport. Reduced net surface solar radiation over the warmest waters explains about half of the observed minima. The wind speed distribution is only important in the presence of ocean heat transport. However, given that the tropical oceans move as much heat as the atmosphere (Trenberth et al. 2001), the wind speed distribution explains at least half of the spatial variation of latent heat flux and has an equally large impact on the ocean heat transport. These results confirm the suggestions of numerous previous investigators that the wind speed is part of the reason why there is low evaporation over the warmest waters. However, our results make clear, as it was not before, that the wind speed can only have this impact because of ocean heat flux divergence from the warmest regions. To our knowledge the manner in which the atmosphere and ocean coupling allows an evaporation minimum at the equator, despite the rather obvious processes involved, has not been advanced before.

In this work we have imposed changes in wind speed and cloud cover on an ocean GCM and also analyzed results from a coupled atmosphere–ocean mixed layer model in which the strength of the ocean heat transport was specified. As such we have glossed over the true extent to which the tropical atmosphere and ocean are coupled. The work brings attention to the links between the ocean heat transport, cloud cover, and wind speed that give rise to low evaporation from warm equatorial regions. However it should also be recognized that the spatial variations of the cloud cover and wind speed are themselves determined through coupled atmosphere–ocean processes. Exactly how this happens remains an interesting problem for future study.

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