Why the ITCZ Is Mostly North of the Equator

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ABSTRACT

Although the distribution of sunshine is symmetrical about the equator, the earth's climate is not. Climatic asymmetries are prominent in the eastern tropical Pacific and Atlantic Oceans where the regions of maximum sea surface temperature, convective cloud cover, and rainfall are north of the equator. This is the result of two sets of factors: interactions between the ocean and atmosphere that are capable of converting symmetry into asymmetry, and the geometries of the continents that determine in which longitudes the interactions are effective and in which hemisphere the warmest waters and the intertropical convergence zone are located. The ocean–atmosphere interactions are most effective where the thermocline is shallow because the winds can readily affect sea surface temperatures in such regions. The thermocline happens to shoal in the eastern equatorial Pacific and Atlantic, but not in the eastern Indian Ocean, because easterly trade winds prevail over the tropical Atlantic and Pacific whereas monsoons, with a far larger meridional component, are dominant over the Indian Ocean. That is how the global distribution of the continents, by determining the large-scale wind patterns, causes climatic asymmetries to be prominent in some bands of longitude but not others. The explanation for asymmetries that favor the Northern rather than Southern Hemisphere with the warmest waters and the ITCZ involves the details of the local coastal geometries: the bulge of western Africa to the north of the Gulf of Guinea and the slope of the western coast of the Americas relative to meridians. Low-level stratus clouds over cold waters are crucial to the maintenance of the asymmetries.

1. Introduction

Although the time-averaged solar radiation has a maximum precisely at the equator, the earth's climate has asymmetries relative to the equator that are most pronounced in the eastern tropical Pacific and Atlantic Oceans. Figure 1 shows that in those regions the intertropical convergence zone, where cloudiness of convective origin and rainfall have maxima, is several hundred kilometers north of the equator. The surface winds converge onto the ITCZ so that both zonal and meridional components of the winds have asymmetries. The ITCZ is located over the warmest surface waters that are associated with an eastward oceanic current, the North Equatorial Countercurrent, that has no counterpart south of the equator (Philander 1990). These climatic asymmetries must be a consequence of the asymmetries of the continents relative to the equator. But which aspects of the continental geometries favor asymmetries in some longitudes rather than others, and why is the Northern rather than Southern Hemisphere favored with the warmest surface waters and the heaviest rainfall? The answer to these riddles is composed of two parts: The first involves physical instabilities and feedbacks that amplify an initial perturbation to a symmetric state so that that state becomes asymmetric. This part of the argument favors neither hemispheres; the instabilities can cause the ITCZ to be either north or south of the equator. The second part of the argument involves specific aspects of the continents that determine why the instabilities and feedbacks are effective in some regions and not others and why they keep the ITCZ north rather than south of the equator. The results in this paper demonstrate how both the global distribution of the continents and the details of local coastal geometries in certain tropical regions influence the climatic asymmetries.

Previous steady-state studies of the location of the ITCZ assumed axisymmetry. Pike (1971) showed that cold surface waters at the equator, associated with oceanic upwelling, prevents the ITCZ from being located there. Charney (1971) and Waliser and Somerville (1994) argue that the off-equatorial position of the ITCZ depends on a positive feedback between atmospheric heating associated with moist deep convection and boundary-layer convergence of moisture that feeds the convection and is also intensified by the convection. This feedback, which is most effective in the band of latitudes from 4° to 12°, displaces the ITCZ from the

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equator even when sea surface temperatures have a maximum at the equator. This paper is concerned primarily with the reasons for the off-equatorial ITCZ in the eastern tropical Pacific and Atlantic where sea surface temperature has a minimum at the equator. The surface temperature pattern is regarded as part of the puzzle that needs to be explained so that the focus is on interactions between the ocean and atmosphere in a world that has east–west inhomogeneities.

From a meteorological perspective the sea surface temperature pattern strongly influences the location of the ITCZ—air tends to rise over the warmest surface waters—and, hence, determines the surface winds because those winds converge onto the ITCZ. It is evident in Fig. 1c that where the ITCZ happens to be north of the equator, the surface winds near the equator are northward. Such winds cause upwelling and elevate the thermocline to the south of the equator, cause downwelling and deepen the thermocline to the north of the equator, and thus maintain a sea surface temperature pattern that has cold water at and south of the equator and warm water north of the equator (Moore and Philander 1976). The winds are therefore both a cause and a consequence of the sea surface temperature pattern. This circular argument suggests that interactions between the ocean and atmosphere determine the phenomena under discussion. Such interactions, which have been studied extensively in connection with the Southern Oscillation and El Niño, permit a variety of ocean–atmosphere modes. [See Philander (1990) and Neelin et al. (1994) for reviews.] In some of them, sea surface temperature changes are caused primarily by fluctuations in the depth of the thermocline because of a horizontal redistribution of warm surface waters. The timescales of those modes depend on the travel time of certain planetary waves in the oceans. In some of the other ocean–atmosphere modes, divergent surface currents alter sea surface temperatures, and the dominant timescales are determined by advective processes. The latter modes, which are referred to as ''slow sea surface temperature'' modes, were investigated by Neelin (1991) for the case of symmetry about the equator and by Chang and Philander (1994) for the antisymmetric case. Chang and Philander demonstrate, in a model, how the antisymmetric modes can convert symmetric conditions into the observed asymmetric climate states. The rate at which these antisymmetric modes amplify initial perturbations and destroy symmetric conditions decreases as the depth of the thermocline increases. The modes are therefore ineffective in the Indian Ocean and western tropical Pacific where the thermocline is deep, but are effective in the eastern tropical Pacific and Atlantic where the thermocline is shallow. To explain the prominence of climatic asymmetries in some regions but not others, it is therefore necessary to explain why the thermocline is at different depths in different regions.

The reason for spatial variations in the depth of the tropical thermocline is the different wind systems that prevail over the different ocean basins. If there were no continents, on a water-covered globe easterly trade winds would prevail everywhere in the Tropics. The influence of continents is so profound in the Indian Ocean, which is almost land locked, that cross-equatorial monsoons are more prominent than the trades in that sector. Over the Atlantic and Pacific Oceans, where the continents are less influential, the trade winds prevail. In those oceans the westward winds drive the warm surface waters westward, causing the thermocline to shoal in the east. In the Indian Ocean easterly winds are practically nonexistent at and north of the equator (Fig. 1d) so that there is little east–west variation in the depth of the thermocline. The global distribution of continents, which strongly influences the large-scale wind systems in the Tropics and thereby determines where the thermocline is shallow, is hence a primary cause for the eastern tropical Pacific and Atlantic to be the favored regions for air–sea interactions that can create climatic asymmetries. Additional feedbacks reinforce the asymmetries. One is related to the release of latent heat in the ITCZ. It intensifies the rising motion so that the moist low-level winds that converge onto the ITCZ are also intensified (Waliser and Somerville 1993). Another feedback comes into play when the meridional sea surface temperature gradients create atmospheric pressure gradients that intensify the winds (Lindzen and Nigam 1990); the stronger winds create stronger meridional temperature gradients. Those gradients are further magnified when evaporation cools the ocean surface in regions where the winds are most intense, away from the ITCZ (Xie and Philander 1994; Xie 1994).

The air–sea interactions and the feedbacks just mentioned favor neither hemisphere with the warmest surface waters and the ITCZ. To determine why the warmest waters are north rather than south of the equator in the eastern tropical Atlantic and Pacific, we use an atmospheric GCM (described in section 2 of this paper) to explore which aspects of the continents create modest asymmetries in the winds, asymmetries that ocean–atmosphere interactions and other feedbacks can amplify. Next we proceed with coupled ocean–atmosphere GCMs to calculate the amplification but discover that an additional feedback involving low-level stratus clouds is of crucial importance. Section 3 discusses those clouds before we present the results from coupled models in section 4.

2. The models

The atmospheric model used in this study is the GFDL R30 GCM and is essentially the one described by Manabe and Hahn (1981). Meteorological variables are represented by spherical harmonics with a rhomboidal truncation at wavenumber 30. The cor-

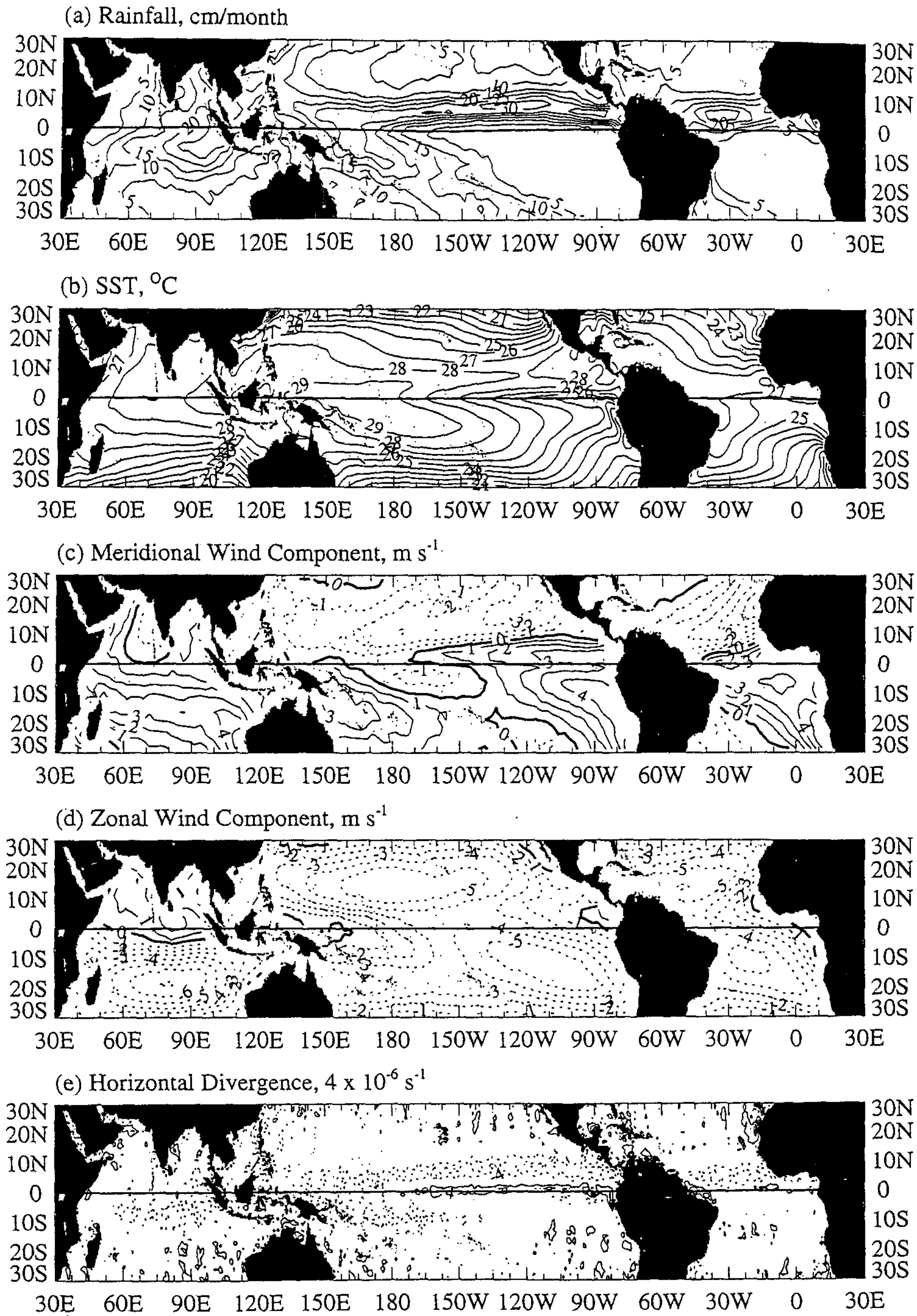

FIG. 1. Time-averaged distributions of (a) rainfall estimated by Spencer (1993) using measurements recorded by the Microwave Sounding Unit on NOAA polar orbiting satellites; the contour interval is 5 cm mo^{-1}. (b) Sea surface temperatures from Reynolds and Smith (1994); The contour interval is 1°C. (c) Meridional and (d) zonal wind components at a height of 10 m from Halpern et al. (1993); the contour interval is 1 m s^{-1}. Northward and eastward winds are indicated by solid lines.

responding Cartesian grid has a horizontal resolution of approximately 2.25° in latitude and 3.75° in longitude. The model has 14 vertical levels; the lowest 4 levels are within the boundary layer (below 1500 m). It has a convective adjustment parameterization following Manabe and Strickler (1964) and a cloud prediction scheme described by Wetherald and Manabe (1988).

The model, when forced with seasonally varying solar radiation (that has no diurnal variation), reproduces the seasonal cycle, that of the surface winds and precipitation for example, realistically when the observed sea surface temperature patterns are specified as lower-boundary condition. However, the model does not offer a realistic simulation of low-level stratus clouds.

In section 3 we force the atmospheric model with idealized sea surface temperatures—they are symmetrical about the equator and correspond to those along the date line—and regard the model as a tool to explore how different continental geometries can affect the winds. In the various numerical experiments the forcing function is the annual mean solar radiation so that the seasonal cycle is excluded. All land surfaces are assumed to be flat. (We comment on the role of mountains in the last section of this paper.)

The oceanic model used in this study is the GCM known as MOM (Modular Ocean Model of GFDL). It has a longitudinal resolution of 1° and a variable latitudinal grid that has its finest resolution, of 33 km, between 10°S and 10°N. The model has 27 levels in the vertical with 10 of those in the upper 100 m. The Richardson-number-dependent scheme used for vertical mixing is the one proposed by Pacanowski and Philander (1981). This model is essentially the same as the one used by Philander et al. (1992) as the oceanic component of their coupled model that simulates the Southern Oscillation. The atmospheric model provides the surface heat and momentum fluxes for the ocean; the oceanic model provides sea surface temperatures for the atmosphere.

Initial conditions for the ocean are generated by starting with zero currents and the observed temperature and salinity fields (Levitus 1982) and by then driving the ocean model for three years with specified surface winds and heat fluxes. The two sets of boundary conditions for the ocean used in this study are obtained from the atmospheric model by running it with specified sea surface temperatures and continental geometries. Those temperatures and geometries are idealized for one set of experiments and are realistic in another. (See the end of section 3 for further details.) Once the oceanic and atmospheric GCMs are coupled, they exchange information once per day. They run in the coupled mode for three years while being forced with the annual-mean solar radiation. The results to be described are the time averages of the last year.

3. The effects of continental geometries

This section explores which aspects of continental geometries can cause modest climatic asymmetries to be amplified into the observed asymmetries by the air–sea interactions and various feedbacks mentioned in the introduction. In the calculations, made with the atmospheric GCM, the forcing function is the annual-mean solar radiation, and the specified sea surface temperatures are symmetrical about the equator, are independent of longitude, and correspond to those observed in the Northern Hemisphere along the date line. Hess et al. (1993) report similar calculations that explore the effects of different parameterization of convection. On a water-covered globe they find that the position of the ITCZ is over the warmest water, even when the sea surface temperature gradient is weak, provided the model employs the convective adjustment scheme. Our model employs that scheme and would have an ITCZ over the warmest waters, with all fields symmetrical about the equator, if continents were absent. Figure 2 shows the modifications introduced by the presence of different continental geometries. In Fig. 2a, coastlines coincide with circles of latitude and lines of longitude so that the "Indian" and "Atlantic" Oceans but not the "Pacific" have continental asymmetries. Over the tropical Pacific the winds are essentially symmetrical about the equator and the zone of lowest surface pressure and maximum rainfall (not shown) is over the warmest water at the equator. It is intriguing that the winds over the tropical Indian Ocean are easterly because in reality they are, on the average, westerly at and to the north of the equator (Fig. 1d). Either the geometrical details and orography of the Indian subcontinent strongly influence the time-averaged winds or the seasonal fluctuations are of central importance to those winds. These possibilities will be pursued on another occasion.

Whereas conditions are essentially symmetric about the equator in the Pacific and Indian Oceans, slight asymmetries are evident in the eastern "Atlantic." For example, the minimum in surface pressure is displaced northward and, over the eastern part of the basin, the winds are cross equatorial. The explanation is obvious: the West African bulge to the north of the Gulf of Guinea. The land north of 5°N approximately attains a surface temperature so much higher than that of the ocean south of it that winds similar to land–sea breezes or the monsoons come into play. Even though the annual-mean solar radiation is symmetric about the equator, the winds are cross equatorial toward the landmass, as is evident in Fig. 2a. In these simulations the specified sea surface temperatures are symmetrical about the equator and constant, but we can anticipate that, in a coupled ocean–atmosphere model, the southerly component of the winds will cause coastal upwelling along southwestern Africa. The resultant north–south sea surface temperature gradient, with warm surface waters north of and cold waters south of the equator, will intensify the winds and amplify the climatic asymmetries relative to the equator.

The climatic asymmetries of the eastern tropical Pacific do not have as simple an explanation as those of the Atlantic. One possibility is that the global distribution of continents, rather than local details of coastal

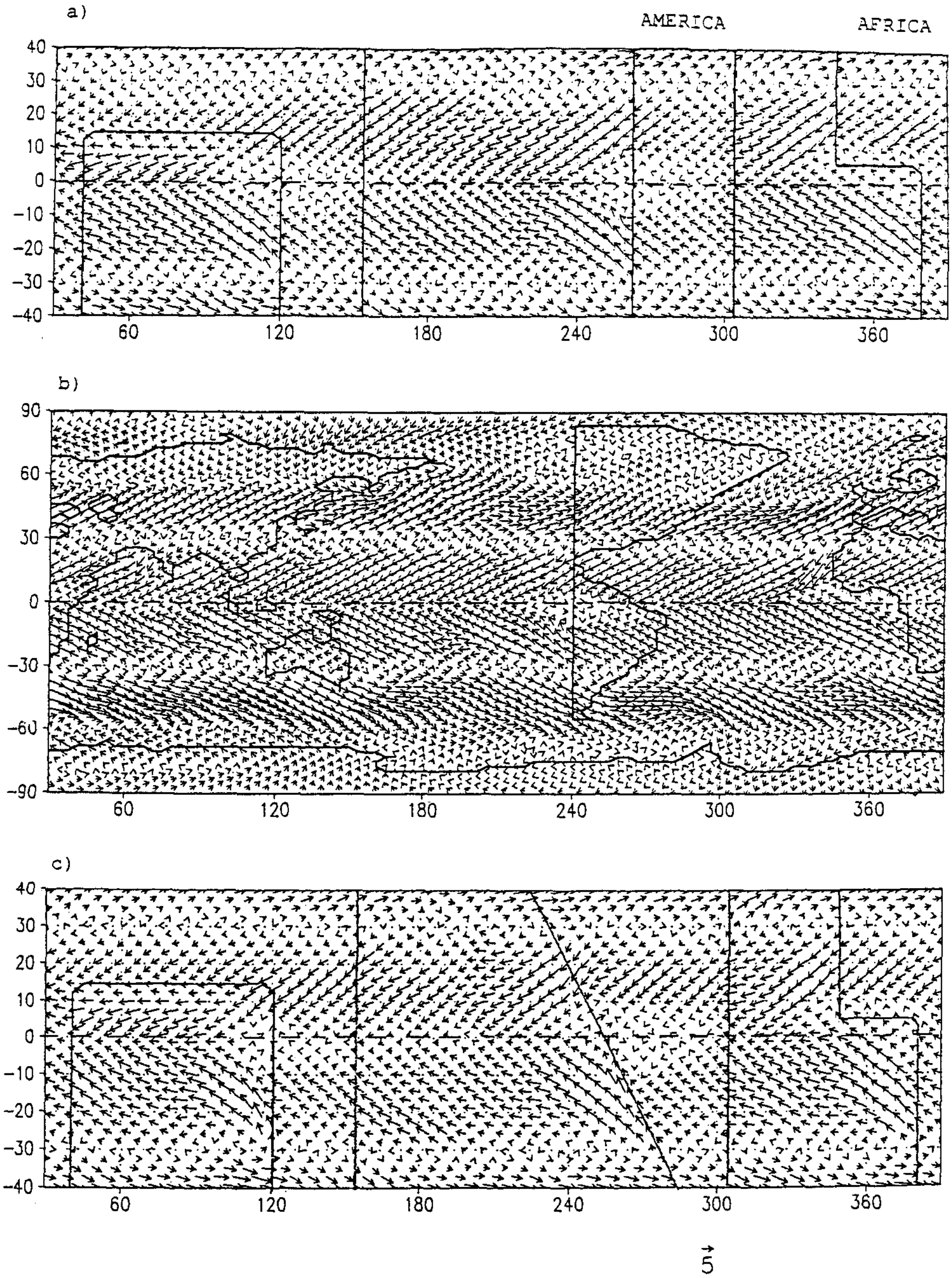

FIG. 2. Surface wind vectors as calculated by an atmospheric GCM in which the specified sea surface temperatures vary only with latitude and correspond to those observed along the date line. The three panels correspond to simulations with different idealized coastal geometries. The length of the arrows represents the surface wind speeds, and their directions are those of the winds. The arrow at the bottom corresponds to an eastward wind of 5 m s^{-1}.

geometry, contribute to asymmetries relative to the equator. Air–sea interactions and various feedbacks can then amplify those modest asymmetries in certain regions, the eastern tropical Pacific for example. The most important global factor is the far larger land area in the Northern Hemisphere relative to that in the Southern Hemisphere.

We retain that factor and eliminate possible asymmetries due to local coastal geometry by changing the western coast of the Americas so that it coincides with a line of longitude. The eastern coast is also distorted, as shown in Fig. 2b, so as to preserve (approximately) the land area in each latitude. As is evident in Fig. 2b, there is no discernable asymmetry in the surface winds off the (straight) western coast of the Americas. There are anticyclonic winds around high pressure zones to the north and south of the equator in the eastern tropical Pacific. The high pressures are over the water that is colder than the land. The associated southeast and northeast trades meet along the equator. This result remains unchanged when sea surface temperatures to the south of 20°S are lowered. (Larger amounts of cloudiness over the southern oceans, associated with the greater expanse of water there, contribute to lower sea surface temperatures but, in this model, has little effect on the Tropics.) Apparently the climatic asymmetries of the eastern tropical Pacific are not attributable to the greater land area of the Northern Hemisphere. (It is conceivable that the inclusion of a seasonal cycle could change this result.)

Next we explore whether the local coastal geometry, specifically the inclination of the coast to meridians, can contribute to asymmetries. When the western coast of the Americas is altered to become a straight line inclined to meridians (Fig. 2c), then comparison of the wind distributions shown in Figs. 2a,c indicates that the change in coastal geometry has very little effect on the coastal and offshore winds. The anticyclonic systems are hardly affected so that the minimum of the surface pressure (and the position of the ITCZ) remains essentially on the equator. From a meteorological point of view the change in coastal geometry is of little consequence. From an oceanographic point of view, however, a significant change has occurred: The trade winds north of the equator are now perpendicular to the coast and those south of the equator are parallel to the coast. We can anticipate that the winds will induce alongshore oceanic currents that are more intense south than north of the equator. Such alongshore currents are associated with offshore Ekman drift and, hence, with coastal upwelling. The resultant lowering of sea surface temperatures will be more pronounced south than north of the equator because of the inclination of the coast to meridians. In other words, the ocean is of central importance to climatic asymmetries in the eastern tropical Pacific. We explore this possibility by using the coupled model described in section 2.

Results from the relatively simple coupled ocean–atmosphere model of Chang and Philander (1994) indicate that ocean–atmosphere interactions can rapidly amplify modest initial perturbations into realistic climatic asymmetries relative to the equator in regions where the thermocline is shallow. In our more complex coupled GCM, the winds of Fig. 2c produce asymmetries in sea surface temperature, but they remain disappointingly and unrealistically modest. The initial oceanic state was generated by driving the ocean model with the winds of Fig. 2c until a state of equilibrium was obtained. The coupled run proceeded from there. Next we tried as initial conditions a realistically asymmetric state by first obtaining winds and heat fluxes from the atmospheric model when realistic sea surface temperatures and continental geometries are specified. Then the ocean model is forced with those heat and momentum fluxes. Finally, the coupled run starts from realistic initial conditions for the ocean and atmosphere, but gradually the realistic asymmetries are degraded as sea surface temperatures south of the equator in the eastern Pacific and Atlantic increase. An example of such a simulation will be presented in section 5 (see Fig. 7a). The principal flaws of these simulations are sea surface temperatures that are far too high in the southeastern tropical Pacific (and Atlantic) Oceans. Temperatures are so high that deep convection occurs over those regions. As a consequence there is a tendency for each hemisphere to have an ITCZ. At the moment, this problem—surface waters that are too warm off the coast of Peru—afflicts not only our coupled model but practically all coupled GCMs. Mechoso et al. (1994) review the performance of some 11 different coupled GCMs and find that, in essentially all available coupled models, the winds parallel to the South American coast create sea surface temperatures slightly, but not realistically, lower than those to the north of the equator. The cold surface waters due to coastal upwelling are too confined to the coast for the asymmetry relative to the equator to be amplified by the ocean–atmosphere interactions mentioned earlier. These model deficiencies could partially be attributed to inadequate simulation of coastal currents and the horizontal spreading of the cold water in the upwelling zones. Apparently the relatively simple coupled models in which asymmetries readily appear [those of Chang and Philander (1994) and of Xie (1994) for example] exaggerate the importance of air–sea interactions. If we accept that the coupled GCMs are more realistic in the role they assign to ocean–atmosphere interactions, then those coupled GCMs must be missing an important feedback that complements the air–sea interactions and that significantly lowers sea surface temperatures to the south of the equator. In the next section we propose that the missing feedback involves stratus clouds.

4. The tropical stratus clouds

The atmospheric circulation in the Tropics corresponds primarily to direct thermal circulations in which moist air rises over the regions of maximum surface temperatures. Two examples include the Walker circulation, whose rising branch is over the Maritime Continent of the western tropical Pacific, and the Hadley circulation, whose rising branch is the ITCZ over the band of high sea surface temperatures to the north of the equator in the Pacific and Atlantic. The principal regions of subsiding air include those parts of the eastern tropical Pacific and Atlantic that have low sea surface temperatures and high sea level pressures. In the regions of subsidence, the surface winds evaporate water vapor from the ocean, but the atmospheric inversion

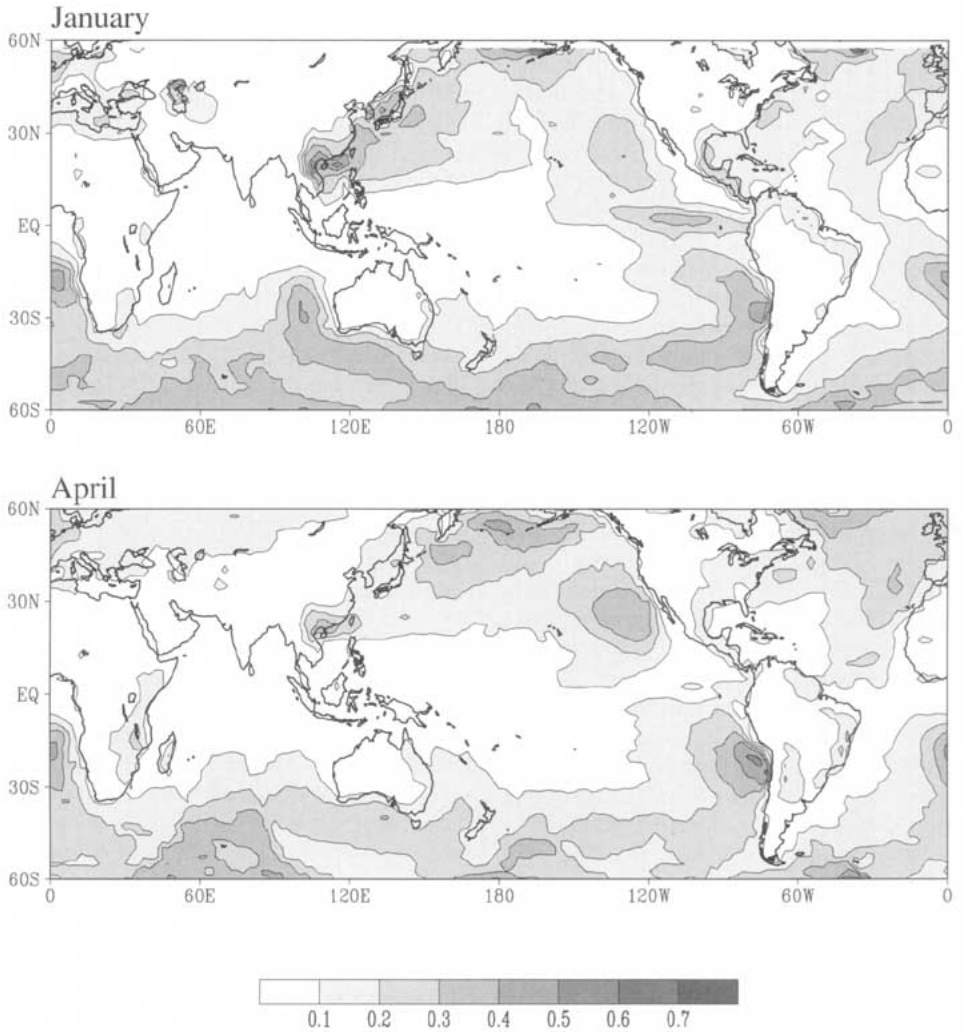

FIG. 3. The fractional stratus cloud cover, on the basis of ISCCP climatology, for January, April, July, and October; contour interval is 0.1.

prevents the moist air from rising to significant elevations. A thin layer of stratus clouds form at the base of that inversion. Radiative (longwave) cooling just below the cloud tops drives convection in a layer below the clouds and causes entrainment that maintains the inversion. [See Klein and Hartmann (1993) for a recent discussion of these clouds.] In the discussion of clouds in climate models, the focus is usually on deep convective clouds because they represent important heat sources for the atmosphere. Stratus clouds, from a strictly atmospheric point of view, are of secondary significance, important primarily because they increase the surface albedo. However, they are so reflective—they can reflect more than 30% of the incident solar radiation—that they are of prime importance in phenomena that involve ocean–atmosphere interactions. They both influence and depend on sea surface temperatures because those temperatures affect the atmospheric temperature profile and, hence, the inversion that controls the thickness of the cloud layer. This means that the clouds participate in a positive feedback: lower sea surface temperatures strengthen the atmospheric inversion and hence favor more stratus clouds, which lower the temperatures even further. Such feedbacks affect the seasonal variations in sea surface temperatures and cloudiness in those regions where stratus clouds are prominent.

In Fig. 3 stratus clouds are seen to be prominent off the western coasts of the Americas and Africa. The data in that figure are from the Stage C1 dataset produced by the International Satellite Cloud Climatology (ISCCP) for the 1983–90 period. The details of the ISCCP cloud detection algorithms are documented by Rossow and Schiffer (1991). Cloud pixels corresponding to cloud tops below 680 mb and optical thicknesses exceeding 3.6 are assumed to represent stratus clouds.

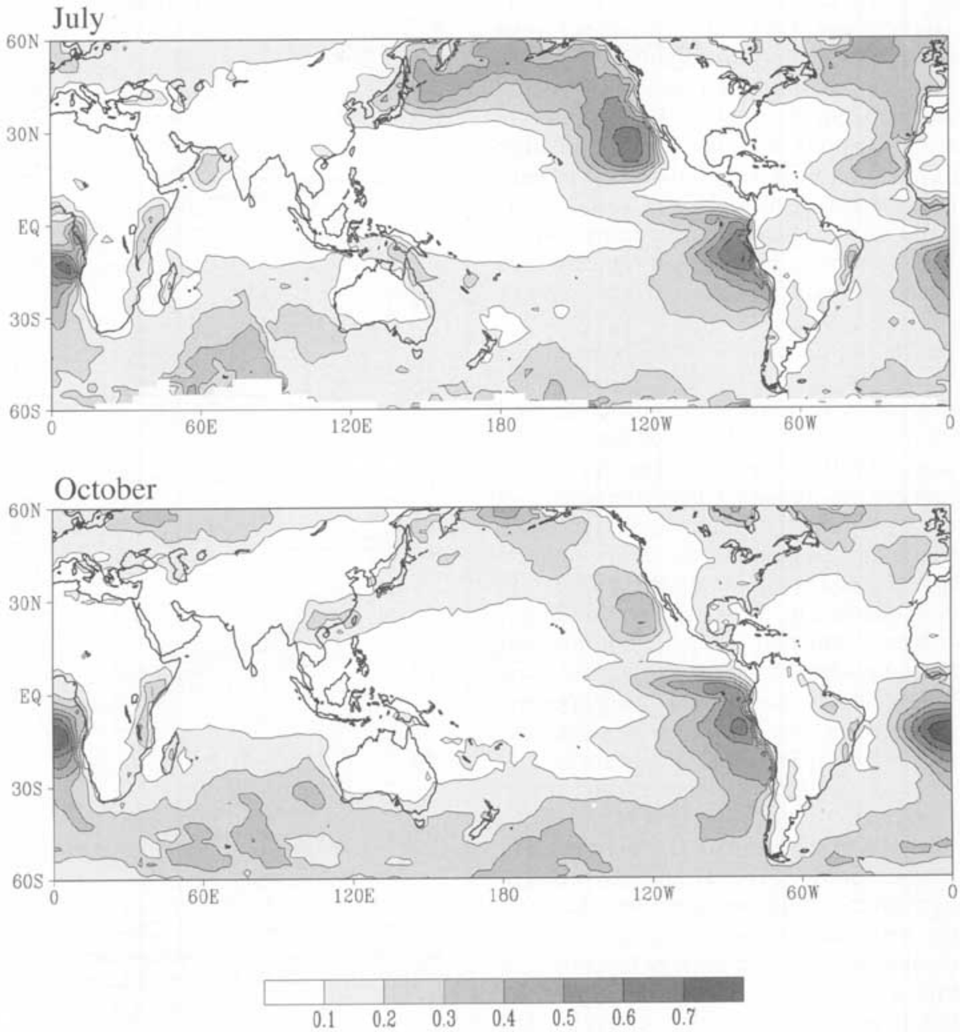

FIG. 3. (*Continued*)

The fractional cover by stratus is obtained by dividing the number of pixels identified as stratus cloud elements by the total number of pixels being analyzed.

Rapid progress is being made in the simulation of stratus decks in studies with very high resolution two-dimensional models. A major challenge is the transition from stratocumulus to trade cumulus as the winds advect air in the marine boundary layer from regions of low sea surface temperature and high cloud fraction toward warmer waters with much lower cloud fraction (Bretherton 1991; Bretherton 1996, manuscript submitted to *J. Atmos. Sci.*). Such issues are dealt with in a very crude manner, or not at all, in climate models because of their coarse vertical resolution; the layers are so thick that the relative humidity is never sufficiently large in a layer for clouds to form. These models, with sea surface temperatures specified, do however succeed in reproducing reasonably well the conditions that lead to the appearance of the stratus clouds: realistic large-scale fields of subsidence and temperature inversions that determine where and when the clouds form. Confirmation that those are key factors comes from a comparison of (time averaged) maps of atmospheric stability near the surface (not shown) and maps of the stratus cloud cover, such as the one in Fig. 3. The regions of high stability essentially coincide with the regions of stratus cloud cover. A further indication that the atmospheric stability is indeed a key parameter that controls the extent of cloud cover comes from the seasonal fluctuations of these clouds. At first it appears that those fluctuations have a surprising feature: the clouds are at a maximum during the local summer off California, but during the local winter off the coasts of Ecuador, Peru, and Angola. Sea surface temperatures alone cannot dictate the cloud cover because those temperatures are at a maximum during summer and at a minimum during winter. The explanation for the cloudiness off California in summer, in spite of relatively high sea surface temperatures, is the strong vertical stability of the atmosphere at that time, as shown on Fig. 4. Close to the equator, the temperatures aloft, at 850 mb say, do not have much of a seasonal variation so that surface temperatures control the temperature profile. Matters are different at higher latitudes, such as those of California. There the seasonal change in temperatures aloft exceeds that at the ocean surface. Off the coast of California local sea surface temperatures are close to their maximum in July, but temperatures aloft are also at a maximum and so is the stability of the atmosphere.

To parameterize stratus clouds in an atmospheric GCM it is necessary to establish, on the basis of measurements, correlations between the cloud cover and atmospheric variables such as subsidence and vertical stability. [Slingo (1980) pioneered this approach.] Correlations between the seasonal changes in low-level clouds (as given by the ISCCP dataset) and a measure of atmospheric stability [defined as the temperature at 850 mb (from NMC analysis) minus the SST (from COADS)] are shown in Fig. 5. These correlation coefficients have been computed using long-term-averaged values of stratus cloud fraction and atmospheric stability for the 12 calendar months and, hence, portray the relationships between the climatological annual cycles in these two variables. The correlations are seen to be high in the regions that have extensive cloud cover. However, the correlations are also strongly positive over parts of the western equatorial Pacific that have little stratus cloud cover (see Fig. 3). This is a problem associated with measurements from satellites. Because of their vantage point, above the clouds, they underestimate low-level clouds beneath another layer of high clouds. Such high clouds frequently cover much of the western equatorial Pacific. A negative correlation thus exists between the satellite-derived high- and low-cloud covers over this region. The large positive correlations between atmospheric stability and stratus clouds over the western equatorial Pacific are hence partially attributable to the anticorrelations between high and low clouds, and between high-level cloudiness and atmospheric stability.

An additional factor that needs to be considered is the effect of subsidence. Maps of the subsidence, estimated from the divergence of the surface winds as measured by the Active Microwave Instrument on *ERS-1* (the European Remote Sensing Satellite), confirm that, in the Tropics, the regions of stratus cloud cover and subsidence coincide (Halpern et al. 1994). It may seem desirable, in a regression formula for the clouds, to include a term that represents the effects of subsidence. However, the correlations between variations in subsidence (from *ERS-1* data) and cloudiness are poor. In other words, subsidence influences where stratus clouds form but does not strongly influence the thickness of the cloud layer. We therefore use subsidence (as inferred from the wind divergence) only to restrict

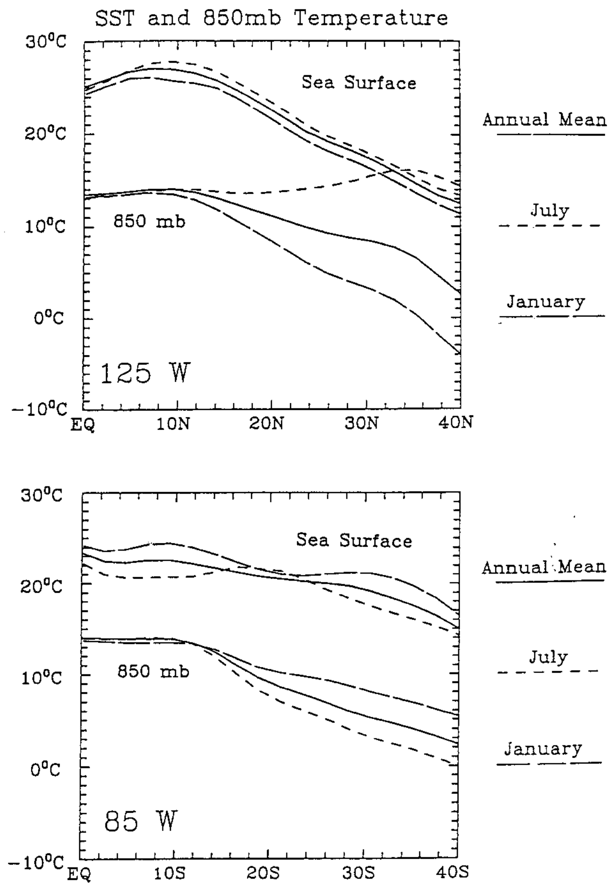

FIG. 4. Sea surface and 850-mb temperatures along 125°W (upper panel) and 85°W (lower panel). The solid lines show the annual-mean values; the dashed lines show averaged values for January and July. The surface data are from COADS; the 850-mb data correspond to the climatology of the NMC analyses.

the regions where stratus clouds can form. That restriction eliminates unrealistic stratus clouds over the western tropical Pacific.

On the basis of the results described above we developed a simple empirical model for stratus cloud cover C, whose value is between zero and one. Cloud cover enters the model only to reduce the shortwave radiation (SW) into the ocean:

$$SW = SW^*(1 - 0.62C),$$

where SW^* is the shortwave radiation at the top of the atmosphere. [This formula is taken from Reed (1977). It is, however, worth noting that Norris and Leovy (1994) recently reported a lower sensitivity of SW to C.] Our parameterization of the clouds is strictly a flux correction for the ocean. The low-level clouds do not appear in the atmospheric GCM in any form and therefore do not affect the atmospheric radiation budget. This implies an absence of interactions in which radiative cooling near the top of the clouds can promote an increase in cloudiness.

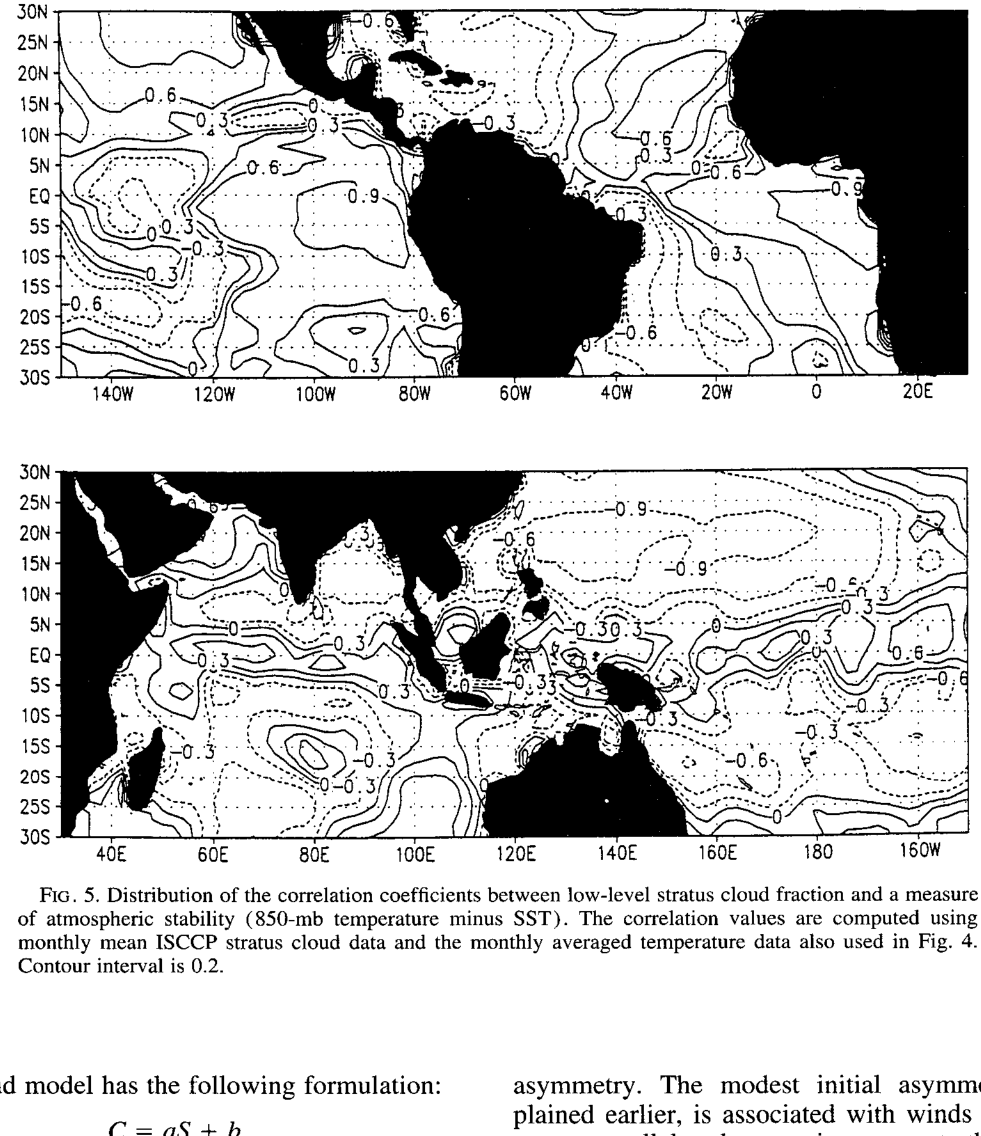

FIG. 5. Distribution of the correlation coefficients between low-level stratus cloud fraction and a measure of atmospheric stability (850-mb temperature minus SST). The correlation values are computed using monthly mean ISCCP stratus cloud data and the monthly averaged temperature data also used in Fig. 4. Contour interval is 0.2.

The cloud model has the following formulation:

$$C = aS + b.$$

In this regression formula the cloudiness C depends on the temperature difference S between the sea surface and 850 mb. The constants a and b have the values $a = 0.031$ and $b = 0.623$. The straight line is the best fit to the data shown in Fig. 6. The stratus clouds are allowed only over the ocean and only in regions where there is descending motion at a height of 600 mb.

5. Effects of stratus clouds in a coupled ocean–atmosphere model

We now examine the impact of the stratus cloud parameterization as formulated in section 4 on the performance of a coupled model. A more detailed description of this model and the experimental procedure have been given at the end of sections 2 and 3. The introduction of stratus clouds cannot create an asymmetry about the equator; it can only amplify a preexisting asymmetry. The modest initial asymmetry, as explained earlier, is associated with winds whose component parallel to the coast is stronger to the south than to the north of the equator (see Fig. 2c). It is therefore not surprising that the coupled model, when the coastal geometry corresponds to that of Fig. 2a, gives results that are symmetrical about the equator in the Pacific even when stratus clouds are taken into account. When a realistic coastal geometry is used, with the western coast of the Americas sloping relative to meridians, then the presence of clouds enables the same model to produce the realistic asymmetries relative to the equator shown in Fig. 7. Comparison of sea surface temperature patterns in the model without (Fig. 7a) and with (Fig. 7b) clouds indicates that it is primarily the southeastern tropical Pacific that cools off, to such an extent that convection and rainfall over that region is now inhibited. It is also noteworthy that the SST at the warm pool in the western equatorial Pacific in Fig. 7b is cooler than that in Fig. 7a, possibly as a result of the strong surface winds in the experiment that incorpo-

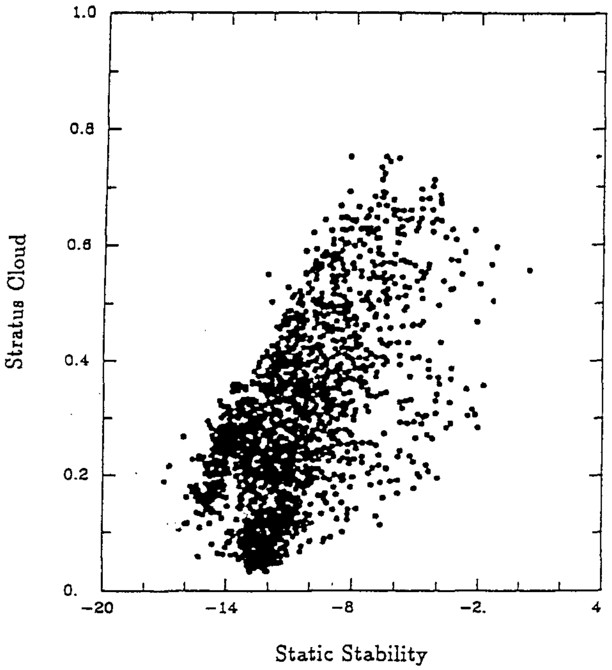

FIG. 6. Scatter diagram for low-level stratus cloud versus a measure of the atmospheric stability (850-mb temperature minus SST). The data are from the regions where the correlation between the static stability and the stratus cloud is larger than 0.6 and where the mean vertical velocity between 400 and 700 mb (in a simulation of the seasonal cycle with the GFDL R30 model) is downward. The empirical formula in section 4 is derived from these samples.

rates stratus clouds (see Fig. 7c). The strengthened circulation in the latter run affects the underlying SST through enchanced convection (thus leading to a reduction in the incident shortwave radiative flux), or through increased evaporation (Hartmann and Michelsen 1993).

The parameterization of stratus clouds, though on the whole successful, has shortcomings. This is evident in Fig. 8, which shows the degree to which the clouds reduce the flux of heat into the ocean. In addition to certain coastal zones, the clouds affect the neighborhood of the equator too. The magnitude of the decrease in radiative flux as simulated in this model is larger than observational estimates of cloud forcing by Ramanathan et al. (1989) and Harrison et al. (1990) by about 30%–40%, probably as a result of the exaggerated sensitivity of shortwave energy to cloudiness in the Reed (1977) parameterization used in section 4. This excessive flux reduction and winds that are too strong (see Fig. 7c) are the main reasons why the simulated sea surface temperatures (Fig. 7b) are lower than in reality. The argument, of course, is circular because sea surface temperature gradients that are too large cause the winds to be too intense, and those winds in turn induce upwelling that creates large temperature gradients. The problems with the model cannot be remedied by paying exclusive attention to stratus cloud parameterization. For instance, boundary-layer processes could result in spatial displacements between the stratiform clouds and the underlying cold tongue (Deser et al. 1993), thus leading to feedbacks between SST and cloudiness that are quite different from those considered here.

The oceanic currents in the model and the associated thermal structure of the ocean, shown in Fig. 9, are realistic in structure. The coupled model succeeds in reproducing the complex system of eastward and westward flowing tropical currents, including the eastward North Equatorial Countercurrent, which is north of the equator and has no counterpart south of the equator. The speeds of the simulated currents are also realistic.

The results from the coupled model with stratus clouds, shown in Figs. 7, 8, and 9, proved to be insensitive to modest changes in the stratus cloud parameterization. For example, a 10% increase in the value of the coefficient b (in the expression for the cloud cover C) from 0.623 to 0.7 had essentially no effect on the results. In particular, the problem with excessive clouds near the equator still exists. There is clearly a need for improvements to the model, including an improved parameterization of stratus clouds. That will be attempted in a future study with a coupled model that takes into account the seasonal variations in solar radiation.

6. Discussion

Climatic asymmetries are prominent in the eastern tropical Pacific and Atlantic Oceans because the shallowness of the thermocline in those regions permits ocean–atmosphere interactions and other feedbacks that amplify modest initial perturbations and destroy symmetry. The crucial question is why those perturbations are such that the warmest waters are north rather than south of the equator. The results presented here suggest that certain aspects of the local coastal geometries are of central importance: the bulge of the African continent to the north of the Gulf of Guinea in the Atlantic and the slope of the western coast of the Americas relative to meridians. In an atmospheric model with specified sea surface temperatures that are symmetrical about the equator, the surface winds are slightly asymmetrical about the equator in the eastern tropical Pacific and Atlantic. The slight asymmetries of the winds force the ocean to have sea surface temperature patterns with asymmetries: surface waters that are slightly cooler south than north of the equator off the western coasts of equatorial Africa and South America. In principle, air–sea interactions ought to amplify those asymmetries, but they fail to do so in our coupled ocean–atmosphere model until we include a parameteriza-

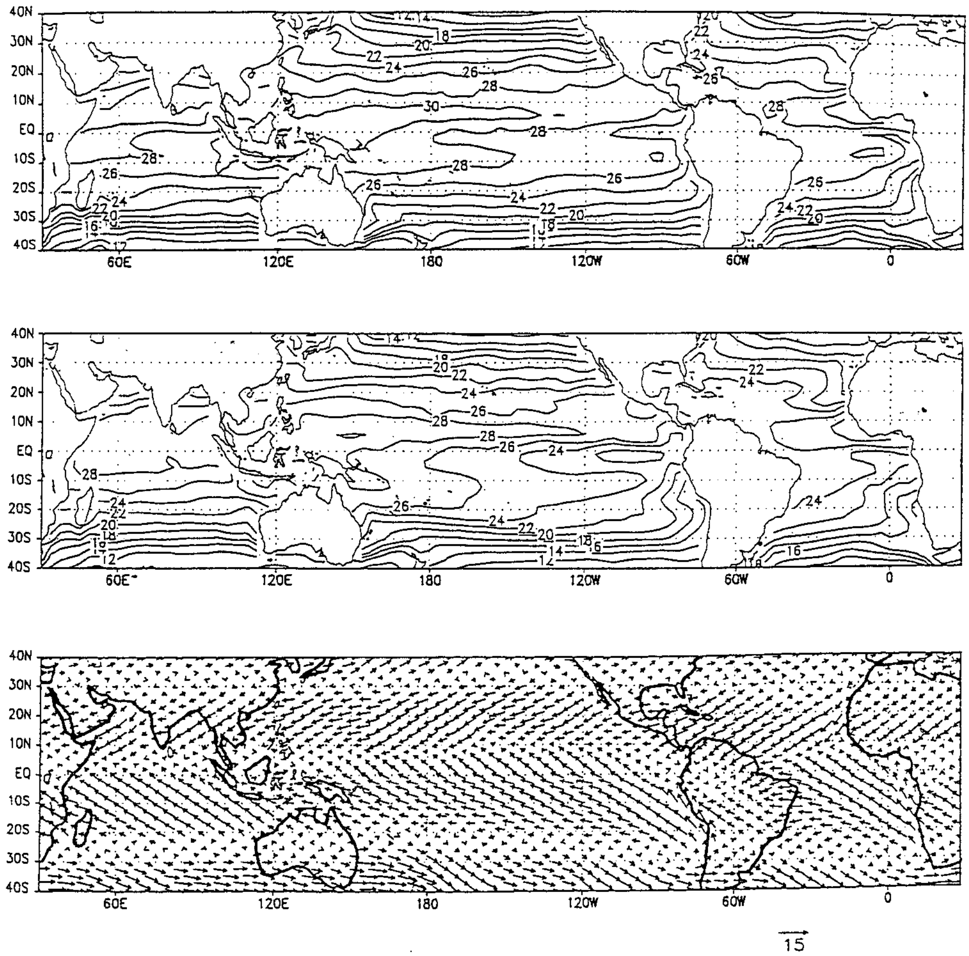

FIG. 7. Simulated sea surface temperature in a coupled ocean–atmosphere model (a) without stratus clouds and (b) with stratus clouds. Contour interval: 2°C. The surface winds associated with the pattern in (b) are shown in (c); the arrow at the bottom corresponds to an eastward wind of 15 m s^{-1}.

tion of the low-level stratus clouds that form over cold tropical waters in regions of subsiding air. It is very likely that positive feedbacks involving stratus clouds—their formation over cold water promotes the lowering of sea surface temperatures—contribute to the magnitude of the observed climatic asym-

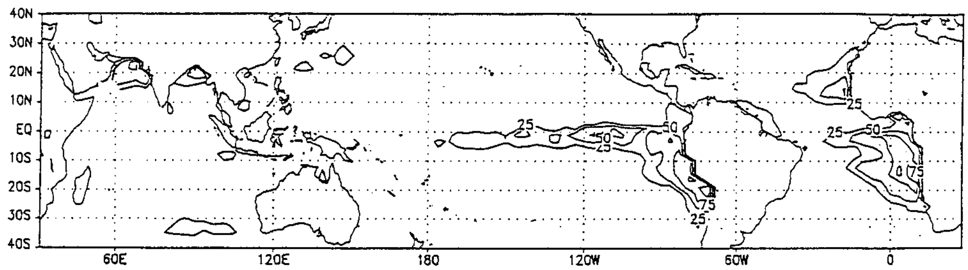

FIG. 8. The extent to which the parameterized stratus clouds reduce the heat flux into the ocean. Contour interval: 25 W m^{-2}.

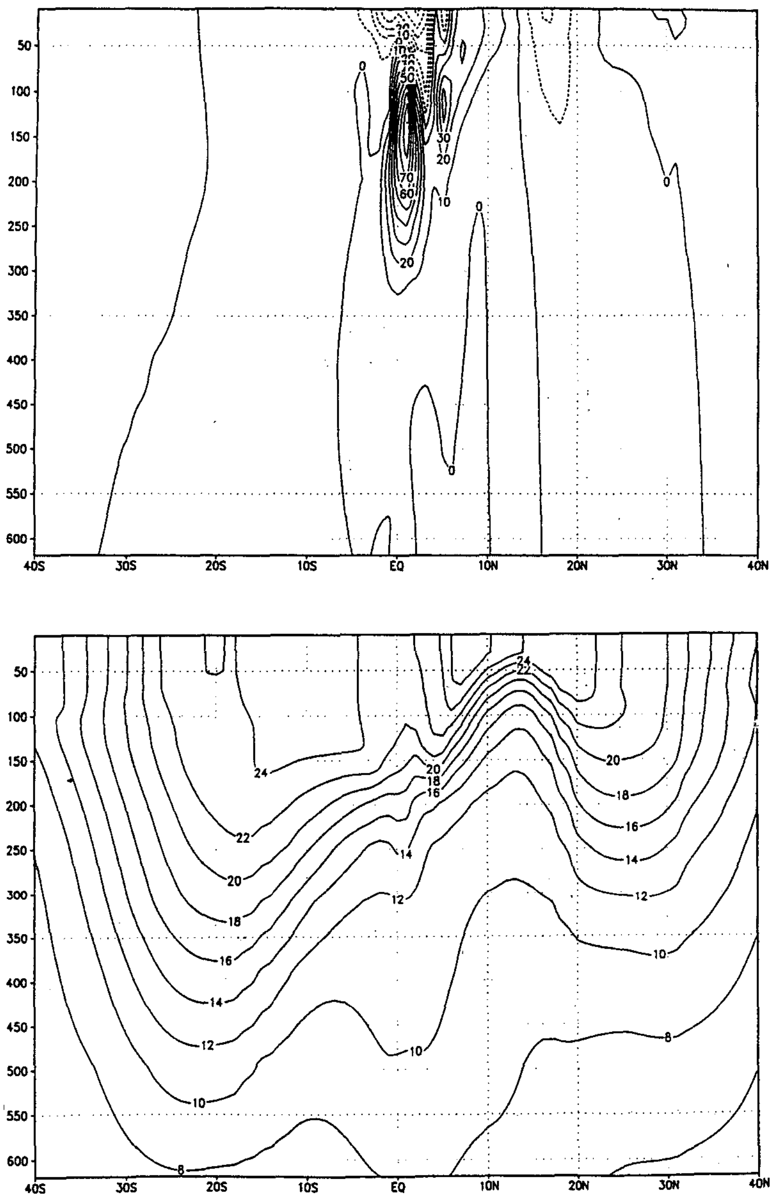

FIG. 9. The zonal currents (top panel) and temperature (bottom panel) as a function of latitude and depth along 140°W, as simulated by the coupled model. Contour intervals are 10 cm s^{-1} and 2°C.

metries. However, further numerical experiments are necessary to determine exactly how much they contribute. Those experiments will have to take into account factors neglected in this paper, factors such as boundary-layer processes, mountains, and the seasonal cycle.

D. Battisti (1994, personal communication) has investigated the effect of mountains in the NCAR Community Climate Model (CCM1 T42) when sea surface temperatures are specified to be symmetrical about the equator. When the mountain slopes relative to meridians, then the winds to the west of the mountain are such that they would favor oceanic upwelling and lower sea surface temperatures south of the equator. Battisti treats the surface of the mountain as a water surface and prescribes the lapse rate of the temperature on the mountain so as to represent primarily a mechanical forcing. The relative importance of the land–sea contrast (included in this study) and of the Andes (in Battisti's study) in creating an asymmetry of the western coast of the Americas remains to be evaluated. It is noteworthy that atmospheric and oceanic conditions off the coast of Angola in the Atlantic are similar to those off the coasts of Peru and Chile, even though the one coast has towering mountains and the other none.

Efforts to understand atmospheric and oceanic phenomena usually rely on theoretical studies with a hierarchy of models. The most complex models are capable of the greatest realism, but their results are difficult to analyze and explain. It is therefore important to have simpler models that by excluding certain processes, sacrifice realism, but in return allow detailed analyses and yield physical insight into the retained processes. The simpler models therefore assist with the interpretation of measurements and data from complex models. This study demonstrates that the complex models can identify unexpected shortcomings of the simpler models. For example, the simple coupled model of Chang and Philander (1994) suggests that ocean–atmosphere interactions are sufficient to create climatic asymmetries relative to the equator. This study reveals that, in a model sufficiently complex to assign an appropriate weight to air–sea interactions, those interactions are inadequate, and some additional processes are needed.

The results in this paper describe the time-averaged atmospheric and oceanic condition in response to the time-averaged solar radiation. The seasonal cycle has been suppressed so that the question of the effect of seasonal forcing on time-averaged conditions has gone begging. This matter will be addressed in subsequent studies.

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