The Role of Heat Fluxes and Moist Static Energy in Tropical Convergence Zones

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ABSTRACT

The relationship between monthly mean outgoing longwave radiation (OLR) and the mean moist static energy of the lower troposphere is shown to be similar to the relationship between monthly mean OLR and sea surface temperature over the oceanic regions. The relationship between monthly mean OLR and surface moist static energy shows that the threshold value for the onset of convection is different in continental and oceanic regions. However, the threshold of moist static stability for the troposphere (surface to 400 mb) is the same for oceans and continents. This relationship is consistent with the simple model of the tropical convergence zones proposed by Neelin and Held. The net energy convergence in the troposphere was found to be positive in regions with OLR below 210 W m^{-2}. This result is consistent with the Neelin and Held hypothesis that the necessary but not sufficient condition for the existence of tropical convergence zone is a positive energy convergence in the troposphere.

1. Introduction

The tropical convergence zones (TCZ) play an important role in the energy and moisture budget of the tropical atmosphere. TCZ can be identified from satellite data as regions with low outgoing longwave radiation (OLR) and high albedo. The term intertropical convergence zone (ITCZ) is not used in this paper because if there are two TCZ at the same longitude we cannot classify both as ITCZ. The factors that govern the spatial and temporal variation of TCZ have not been understood completely so far. Miller et al. (1992) have shown that the location of TCZ in the European Centre for Medium-Range Weather Forecasts (ECMWF) model is sensitive to the type of parameterization used for evaporation from oceans. Ji et al. (1994) have shown that the National Meteorological Center (NMC) model that used the Kuo parameterization for convection was not able to accurately simulate the TCZ in the equatorial central Pacific. The interaction between different processes in a general circulation model (GCM) can be quite complex. Hence, it is not easy to decipher the factors that cause the simulation of the TCZ to be poor in a GCM. Zhang (1994) used the simple model of Neelin and Held (1987) to unravel the factors that influenced the variation of TCZ in the Canadian Climate Centre (CCC) model.

Simple models of the tropical atmosphere are useful for a qualitative understanding of the factors that govern the spatial and temporal variation of TCZ. The simple dynamical models proposed by Gill (1980) and Davey and Gill (1987), denoted DG for brevity, have been used extensively to simulate the temporal and spatial variation of TCZ. The location of TCZ in the DG model is determined essentially by the sea surface temperature (SST). In the DG model, heat fluxes from the ocean play no role in the location of TCZ. This is not consistent with the results of Miller et al. (1992), who have shown that the location of TCZ in the ECMWF GCM was sensitive to the parameterization of evaporation from the oceans. Seager (1991) has shown that the deficiency in the DG model is not in the modeling of the dynamics but in the modeling of the thermodynamics. Seager (1991) proposed that TCZ can exist in a region if and only if the surface moist static energy was above a threshold value and assumed that the threshold value will be the same for continents and oceans. The simple thermodynamical model proposed by Neelin and Held (1987) identifies explicitly the role played by radiative and evaporative fluxes in the location of TCZ. Neelin and Held (1987, henceforth NH) have predicted that the TCZ cannot exist if the net energy convergence in the troposphere is negative.

The observed relationship between TCZ and SST has been explored in great detail in recent years. Zangvil

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(1975), Gadgil et al. (1984), and Graham and Barnett (1987) have shown that TCZ are confined to oceanic regions with SST above 300 K. Waliser and Graham (1993) have shown, however, that in regions with SST above 302 K the TCZ occur less often than in regions with SST equal to 302 K. The simple models of the TCZ such as DG and Seager (1991) would not be able to account for the observations of Waliser and Graham (1993). The NH model may be able to account for this since it imposes an additional constraint derived from the energy equation. According to the NH model, TCZ may not exist at the highest SST if the net energy convergence in the troposphere is negative.

In this paper we explore the relationship between TCZ and the moist static energy and heat fluxes in the troposphere. The OLR is used as a proxy for deep convection in the TCZ (e.g., Liebmann and Hartmann 1982; Wang and Rui 1990; Gadgil and Srinivasan 1990). In the Tropics, OLR is governed primarily by cloud-top temperature, so that low OLR values correspond to high cloud tops, which are associated with TCZ. Monthly average OLR has been used in a number of studies of deep convection in the Tropics, for example, by the researchers listed above. We have used the radiation budget data from the Earth Radiation Budget Experiment (ERBE; Barkstrom and Smith 1986), the cloud data from the International Satellite Cloud Climatology Project (ISCCP; Rossow and Schiffer 1991), and the meteorological data from the ECMWF analyses. In section 2 we discuss the data used. In section 3 we discuss the relationship between OLR and moist static energy. In section 4 we discuss the relationship between OLR and heat fluxes in the troposphere. In section 5 we summarize the results.

2. Data

We have used monthly mean OLR and net radiation (in 2.5° × 2.5° grid boxes) during 1986 from ERBE. These data have the advantage that they were analyzed so as to account for diurnal variations of OLR (Barkstrom and Smith 1986). We have used the monthly mean surface radiative fluxes obtained by Darnell et al. (1992). They obtained the radiative heat fluxes at the surface using parameterized radiation models with satellite meteorological data from ISCCP and top of the atmosphere radiative fluxes from ERBE. The moist static energy in the troposphere was calculated from ECMWF analyses. The surface latent and sensible heat fluxes over the oceans were obtained from ECMWF analyses and the aerodynamic formula used by Miller et al. (1992). Although studies of transient features such as wave motions require daily maps (e.g., Srinivasan and Smith 1996), the persistence of TCZ in the Tropics is great enough that monthly average maps can be used to study the heat flux–moist static energy relation.

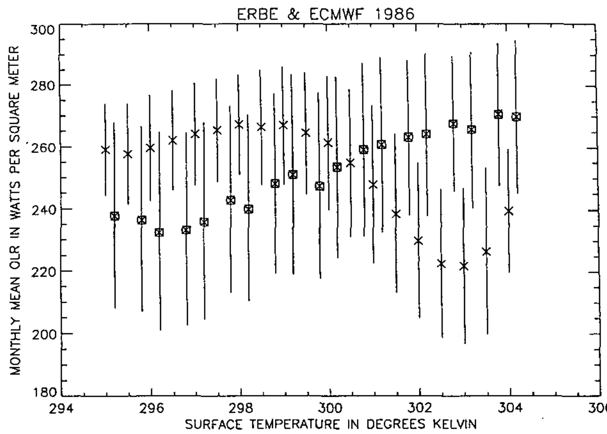

FIG. 1. OLR as a function of surface temperature in 1986. The OLR data was pooled into surface temperature bins from 295 to 305 K with a step of 0.5 K before the mean and standard deviation were calculated. Cross represents oceanic grids and square represents continental grids.

3. OLR and moist static energy

The relationship between OLR and SST in oceanic regions has been explored in great detail by Graham and Barnett (1987), Waliser and Graham (1993), and Zhang (1993). In the simple model proposed by Seager (1991) convection is triggered if the surface moist static energy is above a threshold value. The moist static energy is defined as the sum of sensible heat, potential energy, and latent heat. The relationship between OLR and the surface temperature over continents is quite different from the relationship between OLR and SST over oceans. The relationship between monthly mean OLR and surface temperature over continents and oceans in 1986 is shown in Fig. 1. In all plots in this paper except maps, data from regions between 25°N and 25°S are used. This latitude range excludes the Tibetan plateau, since a low OLR in this region does not indicate the existence of a TCZ. The OLR data were pooled into surface temperature bins from 295 to 305 K in steps of 0.5 K. The mean value in each bin and the standard deviation are shown in this figure. In oceanic regions, the OLR decreases rapidly when the surface temperature increases above 300 K. This is not the case over the continents. This is understandable since there is no simple relationship between surface temperature and surface moist static energy over the continents. In oceanic regions the variation of relative humidity near the surface is much less than the that over the continents, hence there is a unique relationship between SST and surface moist static energy over the oceans. The relationship between OLR and surface moist static energy over continents and oceans is shown separately in Fig. 2 for 1986. All the OLR data were pooled into surface moist static energy bins from 330 to 360 kJ kg^{-1} with a step of 3 kJ kg^{-1}. The mean in each bin and the standard deviation are shown in Fig. 2. OLR decreases rapidly when the surface moist static energy goes above a threshold value. Liebmann and Hartmann (1982) used a monthly mean OLR value of 240 W m^{-2} as a threshold value below which TCZ are present and this value has been used in a number of subsequent studies (e.g., Waliser and Graham 1993). If we assume that TCZ exist in continental and oceanic regions with monthly mean OLR below 240 W m^{-2}, then the threshold value of surface moist static energy is around 340 kJ kg^{-1} for continents and 350 kJ kg^{-1} for the oceans.

Figure 3 shows the relationship between OLR and mean moist static energy of the troposphere (surface to 400 mb). The mean moist static energy of the troposphere (surface to 400 mb) was calculated as an integral over pressure from the surface to 400 mb. The threshold value of the mean moist static energy of the troposphere (for the existence of TCZ) is the same in continental and oceanic regions and is around 335 kJ kg^{-1}. The vertically integrated moist static energy is a better descriptor of the state of the atmosphere in depth than is a surface number, hence it relates better to the OLR. For moist static energy of troposphere greater than 340 kJ kg^{-1}, the OLR increases again. This increase corresponds to the increase of OLR and decrease of cloudiness that was found by Waliser and Graham (1993) for sea surface temperature exceeding 300 K. Zhang has shown that where the SST is high, for example, greater than 30°C, this variability of deep convection is high, so that a significant part if the time there is no deep convection, thus the OLR average increases. Lau and Bua (1996, manuscript submitted to *J. Climate*) have shown that the increase of OLR at

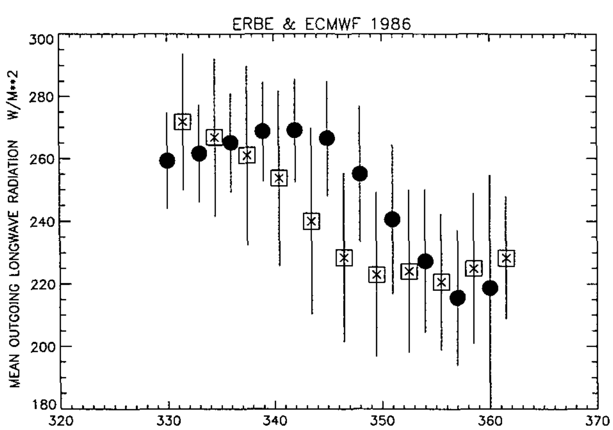

FIG. 2. OLR as a function of surface moist static energy in 1986. The OLR data were pooled into surface moist static energy bins from 330 to 360 kJ kg^{-1} with a step of 3 kJ kg^{-1} before the mean and standard deviation were calculated. Circle represents oceanic grids and square represents continental grids.

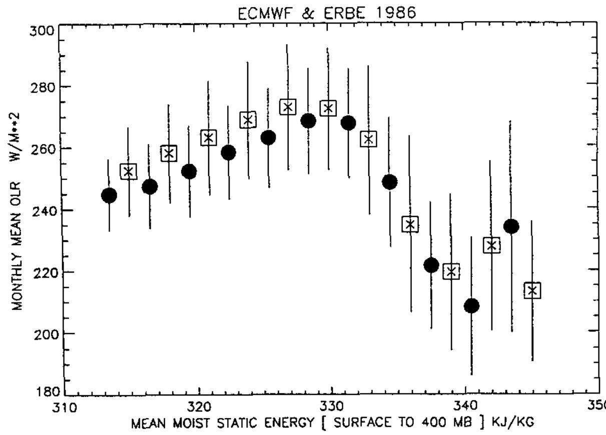

FIG. 3. OLR as a function of moist static energy of the troposphere (surface to 400 mb) in 1986. The OLR data were pooled into moist static energy bins from 315 to 345 kJ kg^{-1} with a step of 3 kJ kg^{-1} before the mean and standard deviation were calculated. Circle represents oceanic grids and square represents continental grids.

high SST is due to the difference in circulation regime over the part of the ocean with high SST. These explanations of the SST–OLR relation will apply also to the moist static energy–OLR relation.

Figures 4a and 4b display OLR and moist static energy contours for January 1986 and July 1986, respectively. These contours of OLR and moist static energy of the troposphere demonstrate again the relation between OLR and moist static energy. Regions with OLR below 240 W m^{-2} occur primarily in regions with moist static energy of troposphere (surface to 400 mb) above 335 kJ kg^{-1}. In most regions the moist static energy contour corresponding to 335 kJ kg^{-1} represents the border between regions of flow convergence and divergence. The threshold values of surface moist static energy were different for oceans and continents, but they are almost same for the mean moist static energy of the troposphere. Although a threshold value of 240 W m^{-2} is used to delineate the TCZ, within this boundary the OLR quickly drops to 210 W m^{-2} or less for regions of spatial extent of 10° or more. In these "core" regions, the TCZ persist during the month.

The relationship between surface moist static energy and the moist static energy of the troposphere (surface to 400 mb) is shown in Fig. 5 for January 1986. When the mean moist static energy of the troposphere (surface to 400 mb) is 335 kJ kg^{-1} the surface moist static energy over the oceans is around 350 kJ kg^{-1}, while the surface moist static energy over the continents is around 340 kJ kg^{-1}. The use of a single threshold value for surface moist static energy suggested by Seager (1991) cannot be justified. Riehl (1979) has shown that the TCZ over the continents are wetter between 500 and 700 mb than the TCZ over the oceans. Hence, for

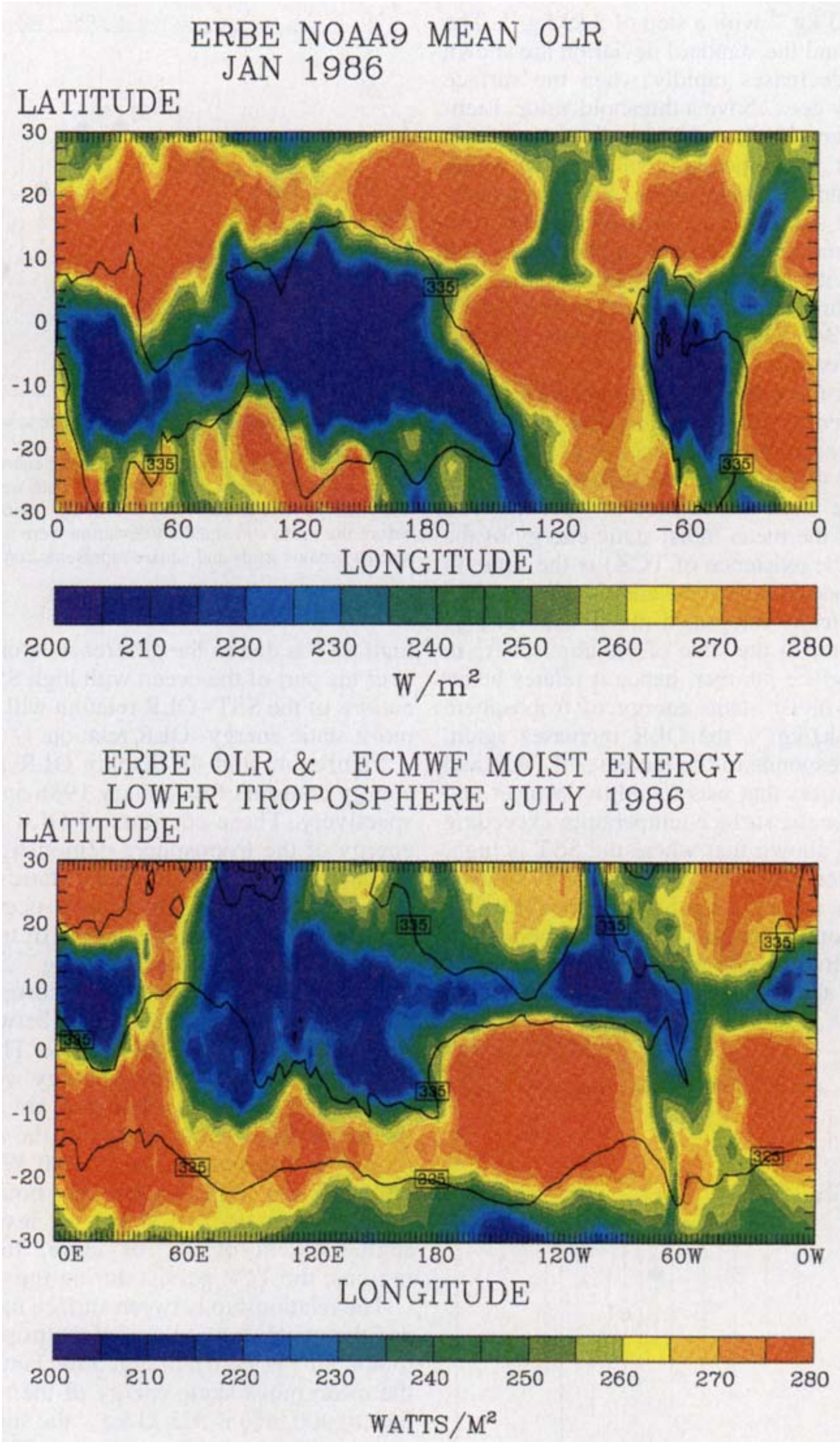

FIG. 4. (a) Monthly mean OLR for January 1986 with contours of moist static energy of the troposphere (surface to 400 mb) superimposed. (b) Monthly mean OLR for July 1986 with contours of moist static energy of the troposphere (surface to 400 mb) superimposed.

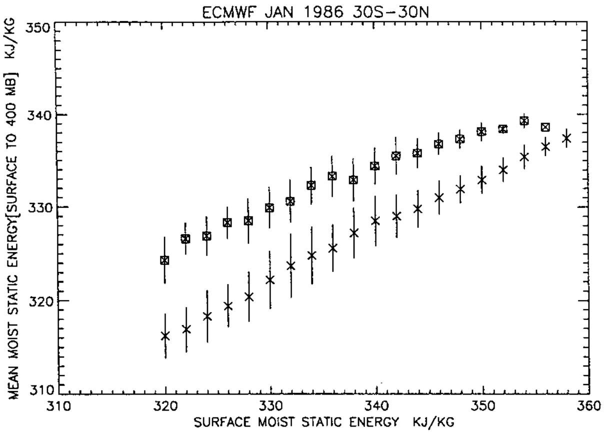

FIG. 5. Moist static energy of the troposphere (surface to 400 mb) as a function of surface moist static energy in January 1986. The moist static energy data of the troposphere were pooled into surface moist static energy bins from 315 to 345 kJ kg^{-1} with a step of 3 kJ kg^{-1} before the mean and standard deviation were calculated. Cross represents oceanic grids and square represents continental grids.

the same surface moist static energy, the mean moist static energy of the troposphere is higher over the continents than over the oceans. However, there is a single threshold value of the mean moist static energy of the troposphere (for both continents and oceans) above which convection occurs.

In the NH model, the mass convergence in the lower troposphere is directly proportional to the net energy convergence in the troposphere and inversely proportional to the gross moist stability. The gross moist stability is defined as the difference between the mean moist static energy of the upper and lower troposphere. The monthly mean moist static energy of the troposphere from the surface to 400 mb is shown in Fig. 6a. The areas of TCZ are marked by monthly mean moist static energy of 337 kJ kg^{-1} or greater. The mean moist static energy of the upper troposphere (400–150 mb) is shown in Fig. 6b and is seen to have a very small spatial variability in the Tropics near 342 kJ kg^{-1}. Thus, the TCZ occur in regions where the monthly mean moist static energy of the troposphere between the surface and 400 mb approaches the upper-troposphere value.

The NH model implies that convection occurs wherever the mean moist static energy of the troposphere (surface to 400 mb) is high and approaches the value in the upper troposphere (400–150 mb), which is supported by Figs. 6a and 6b. This result is quite different from the arguments based on parcel stability that compare the buoyancy of the parcel with that of the environment. The NH model gives primacy to the constraints imposed by energy conservation on the TCZ rather than the usual arguments related to parcel buoyancy. Both the Seager (1991) and the Neelin and Held (1987) models suggest the existence of a moist static energy threshold for the existence of TCZ but for totally different reasons. The NH model does not consider the difference in buoyancy between the parcel and the environment. The basic premise of the NH model is that the moist static energy exported out of the upper troposphere in a TCZ is greater than the moist static energy imported into the lower troposphere in a TCZ. The results we have obtained from ERBE and ECMWF data are consistent with the NH model but not with Seager (1991). The Davey and Gill (1987) model predicts the existence of TCZ wherever SST is high and hence is consistent with Neelin and Held (1987) and Seager (1991) over oceanic regions. The Davey and Gill (1987) model does not provide any information regarding TCZ over continents. Neelin and Held (1987) have argued that the mean moist static energy of the upper troposphere is always higher than the mean moist static energy of the lower troposphere. The NH model asserts that the difference in moist static energy exported out of the upper troposphere and that imported into the lower troposphere must be balanced by the sum of radiative, sensible, and evaporative fluxes. Hence, the NH model introduces an additional constraint for the existence of TCZ. According to the NH model, TCZ can occur if and only if the net energy convergence in the troposphere is positive. This constraint is not invoked in other simple models because they do not have a self-consistent moist static energy budget. The relationship between TCZ and net energy convergence in the troposphere is discussed in the next section.

4. OLR and net energy convergence

The net energy convergence in the troposphere (EC) can be written as

$$EC = RNT - RNB + HL + HS, \qquad (1)$$

where RNT is net radiation at the top of the troposphere, RNB is net radiation at the bottom of the troposphere, HL is latent heat flux, and HS is sensible heat flux.

The net radiation at the top of the troposphere can be assumed to be the same as that at the top of the atmosphere, which is measured by satellite sensors, since the regions above the tropopause are in radiative equilibrium. The net radiation at the bottom of the troposphere is obtained from Darnell et al. (1992). The latent and sensible heat fluxes are computed on the basis of aerodynamic formulas discussed by Miller et al. (1992). Figure 7 shows the net energy convergence in the troposphere as a function of OLR in the Tropics (30°S–30°N) for January 1986. As the OLR decreases, the net energy convergence in the troposphere tends to move toward a positive value. In regions with OLR

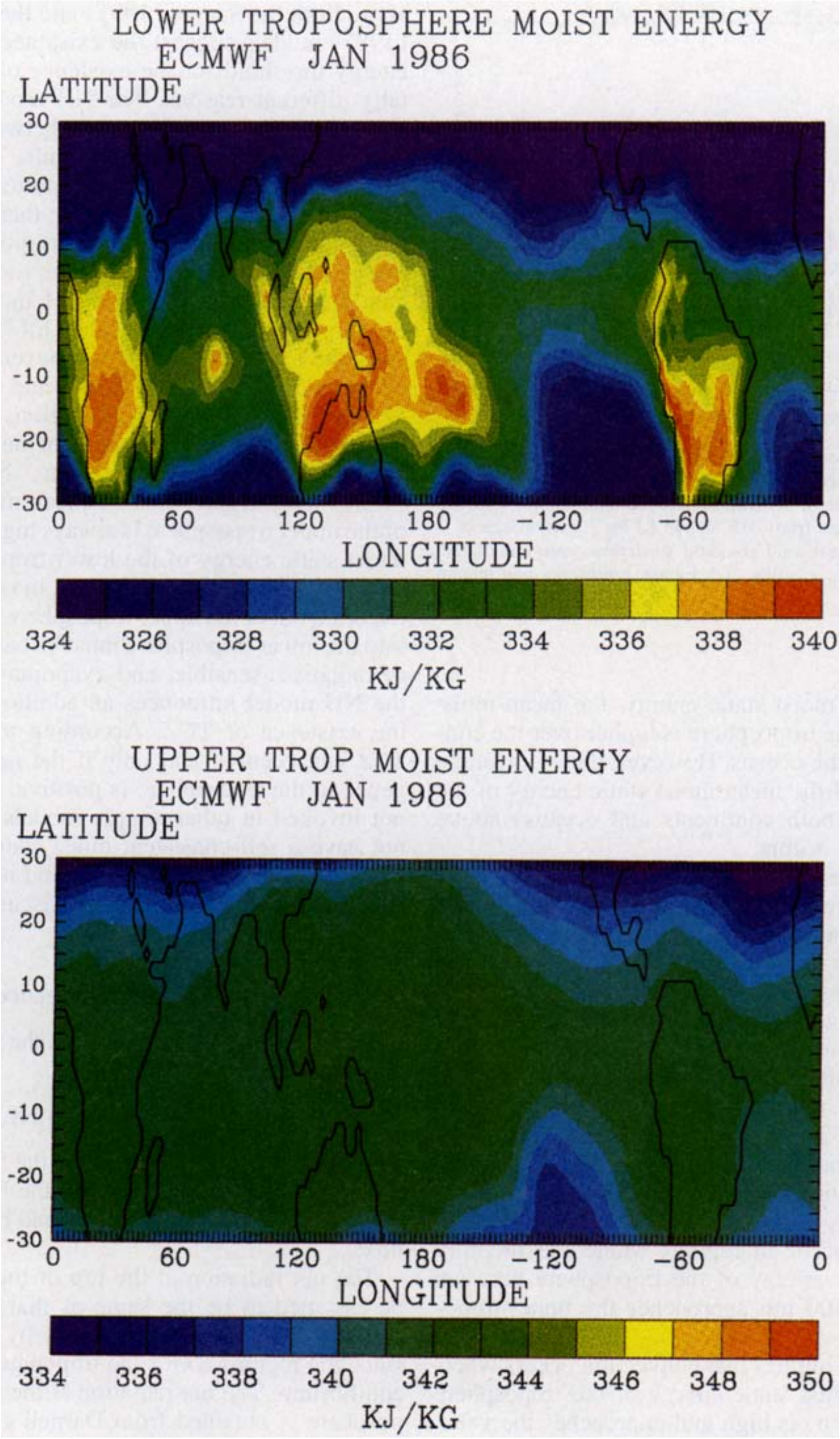

FIG. 6. (a) Monthly mean moist static energy of the troposphere (surface to 400 mb) in January 1986. (b) Monthly mean moist static energy of the upper troposphere (400–150 mb) in January 1986.

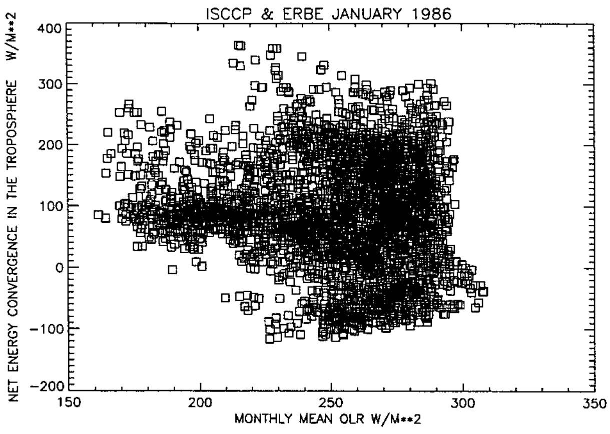

FIG. 7. Net energy convergence in the troposphere as a function of OLR.

below 210 W m^{-2}, the net energy convergence is always positive. These regions were noted in Figs. 4a and 4b as ''core'' regions within which the TCZ are intense throughout the month of observation. This result is consistent with the predictions of the Neelin and Held (1987) model.

There can be large errors in the estimate of evaporative heat flux from the aerodynamic formula. It is not necessary to use the aerodynamic formula to compute the HL and HS over continents. In continental regions, the energy stored in the soil is small over a period of several days and the net radiant heat is transferred to the atmosphere as either sensible heat or as latent heat. Thus, over continents, RNB = HL + HS, so that the net energy convergence in the troposphere is equal to the net radiation at the top of the atmosphere (also called planetary net radiation). This is a better approximation than the estimates of sensible and latent heating. The NH model predicts that the planetary net radiation must be positive over the continental TCZ.

Figure 8 shows the planetary net radiation for the Tropics between 0° and 180° longitude during July 1986. A large area of the tropical Indian and west Pacific Oceans has a net radiation of more than 90 W m^{-2}. Over Africa during the boreal summer, the planetary net radiation is negative everywhere except in the region 5°–15°N, which is the TCZ. A unique feature of the African TCZ is that it does not move poleward of 15°N during the boreal summer. The NH model will attribute the absence of TCZ in the region poleward of 10°N during the boreal summer to the fact that net energy convergence in the troposphere is negative during the boreal summer. Figure 9 shows the temporal variation of OLR and planetary net radiation over Africa (20°–22.5°E) for 1986. Throughout the year the latitudinal excursion of TCZ in this continental region is

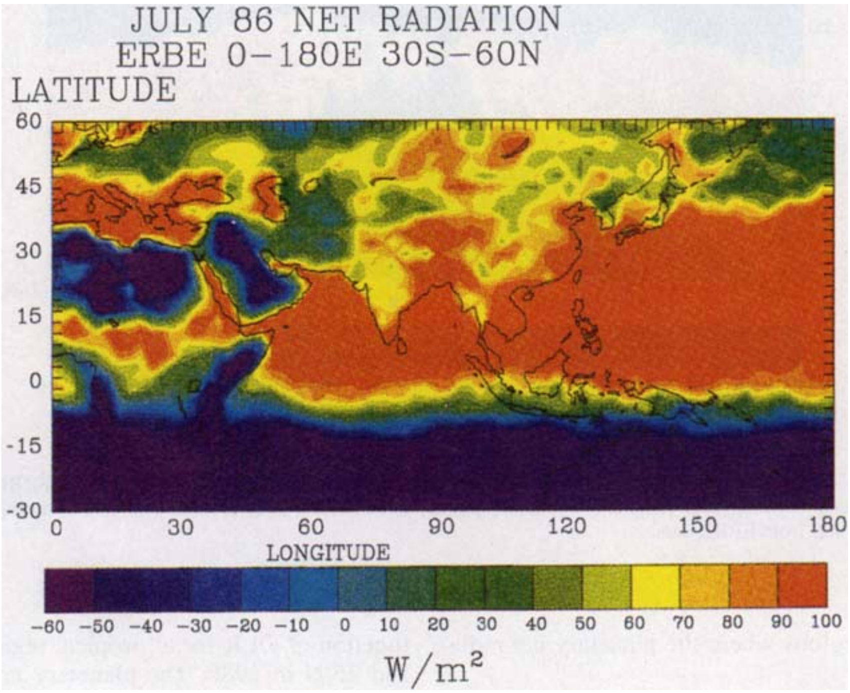

FIG. 8. Monthly mean planetary net radiation in the Tropics in July 1986.

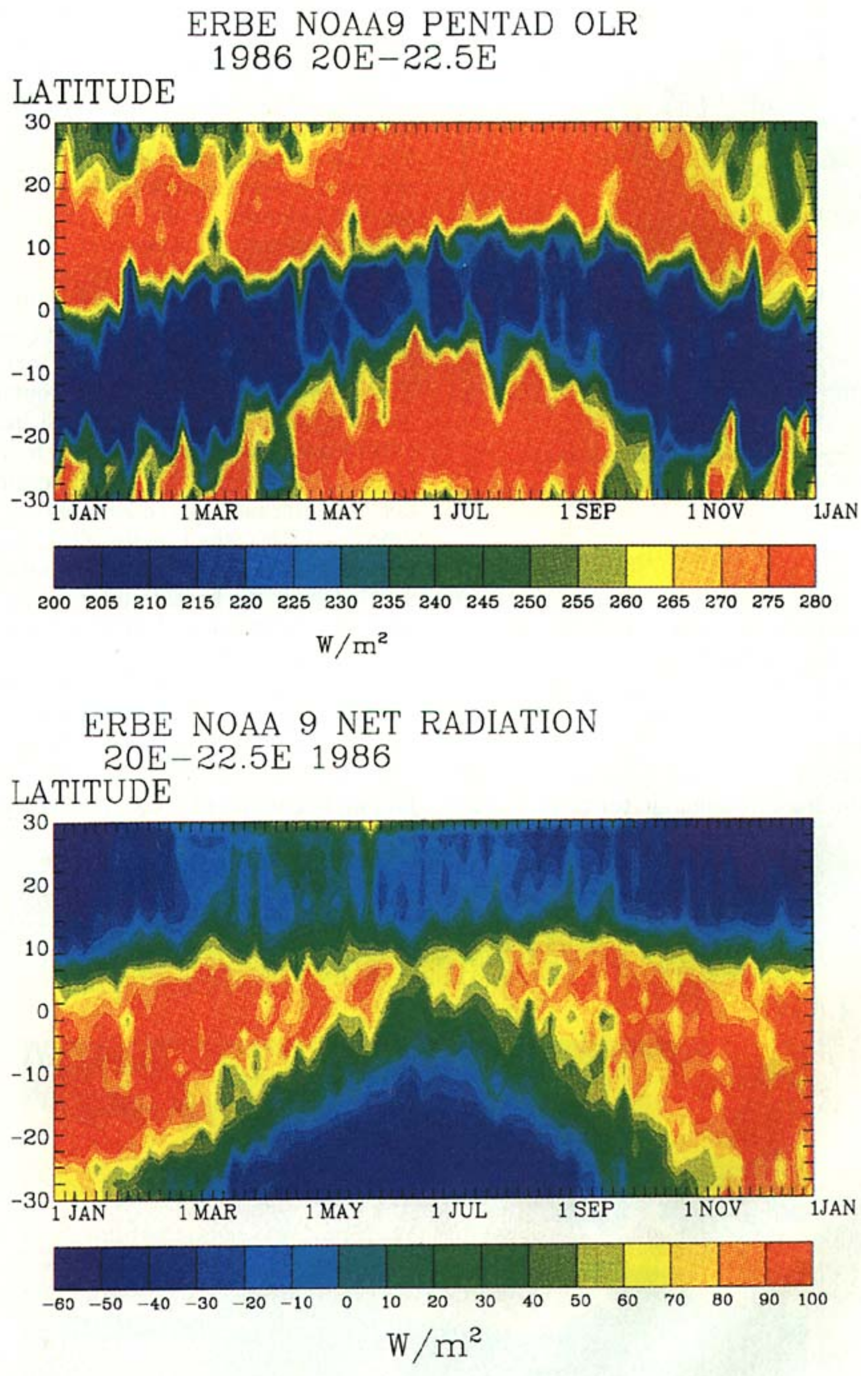

FIG. 9. (a) Temporal variation of pentad OLR in the region 20°–22.5°E during 1986 from ERBE data. (b) Temporal variation of pentad planetary net radiation in the region 20°–22.5°E during 1986 from ERBE data.

restricted to the regions where the planetary net radiation is positive.

Figure 10 shows the net radiation at the top of the atmosphere (also called planetary net radiation) as a function of OLR for all tropical regions between 25°S and 25°N in 1986. The planetary net radiation varies over a wide range at high OLR, but the range of variation decreases as OLR decreases and for OLR below

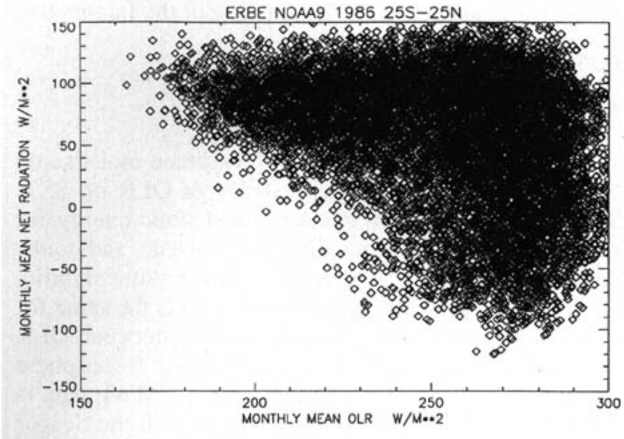

FIG. 10. Monthly mean planetary net radiation as a function of monthly mean OLR for 1986 in the region 25°S–25°N.

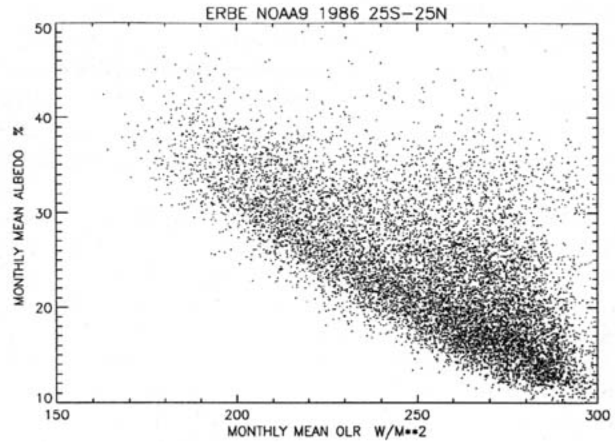

FIG. 11. Monthly mean albedo as a function of monthly mean OLR for 1986.

210 W m^{-2} the net radiation is always positive. The result shown in Fig. 10 includes both continental and oceanic regions. One may be tempted to conclude that net radiation must be positive in regions with low OLR since net radiation is defined as absorbed solar radiation minus OLR. The regions with low OLR have high clouds that tend to reflect solar radiation. Harrison et al. (1990) have shown that the net cloud forcing is close to zero in most of the Tropics and around 50 W m^{-2} in the TCZ. Hence, the planetary net radiation is higher in the TCZ than in the clear regions in the vicinity of the TCZ. The planetary net radiation is positive in the TCZ because the monthly mean albedo never exceeds 50%. The monthly mean albedo (in a 2.5° × 2.5° region) is shown in Fig. 11 as a function of monthly mean OLR from ERBE data for 1986.

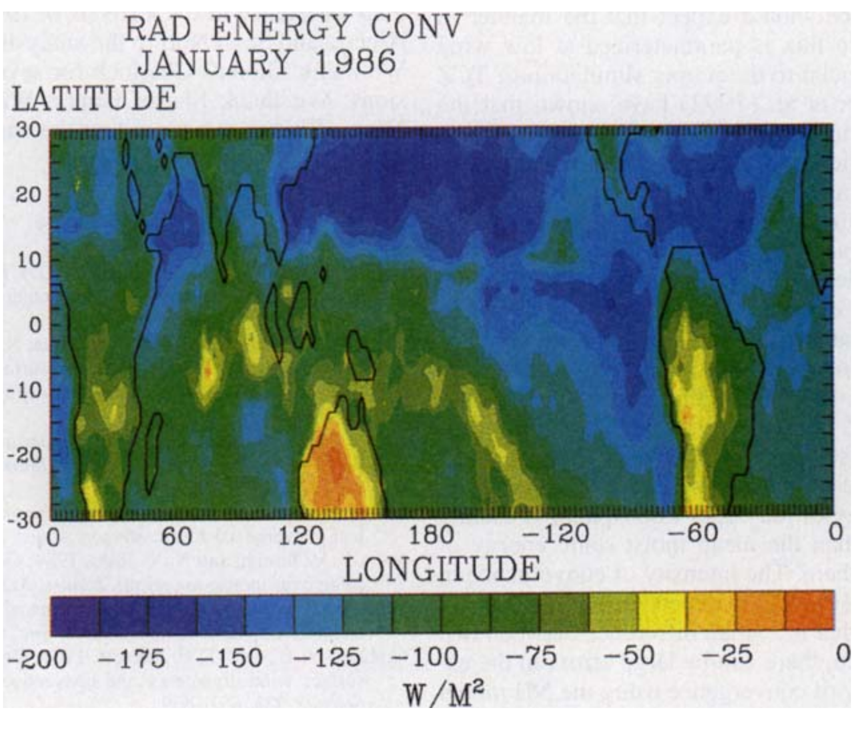

FIG. 12. The net radiative flux convergence in the troposphere in January 1986.

Based on the NH model, we can argue that if the monthly albedo exceeded 50%, the planetary net radiation (and hence net energy convergence) will become negative and hence cannot sustain the TCZ. The results obtained from ERBE data are thus consistent with the argument that TCZ are confined to the regions with positive net energy convergence.

The net radiative flux convergence in the troposphere is the net radiation at the top of the atmosphere, obtained from ERBE, minus the net radiation at the surface from Darnell et al. (1992). The net radiative flux convergence is shown in Fig. 12 for January 1986. In the oceanic region of the Tropics the net radiative flux convergence of the troposphere is between -50 and -200 W m^{-2}. Hence, the NH model shows that the sum of sensible and latent heat flux from the surface must be at least 50 W m^{-2} to ensure that the net energy convergence in the troposphere is positive. Figure 12 shows that the net radiative flux convergence exceeds -50 W m^{-2} only at locations of the TCZ, that is, the South Pacific convergence zone and the Indian Ocean near the equator. Thus, the sensible and latent heat need only contribute 50 W m^{-2} in order for the TCZ to occur. Figure 8 shows that large areas of the tropical oceans have a net radiation of more than 90 W m^{-2} so that there is sufficient radiant flux at the surface to provide the sensible and latent heat fluxes.

If the evaporative flux were to go to zero as wind speed goes to zero, regions with low wind speed could not have positive energy convergence in the troposphere. Hence, one should expect that the manner in which evaporative flux is parameterized at low wind speeds will be crucial to the proper simulation of TCZ in a GCM. Miller et al. (1992) have shown that the location of TCZ in the ECMWF GCM is sensitive to the manner in which the evaporative heat flux was calculated at low wind speeds.

The relationship between OLR, moist static energy, and heat fluxes obtained in this study is consistent with the Neelin and Held (1987) hypothesis regarding the conditions for the existence of the TCZ. One must caution, however, that the NH model cannot be used for the quantitative prediction of the intensity of the TCZ. According to the NH model, the intensity of convergence in the TCZ is inversely proportional to the difference in moist static energy between the upper and lower troposphere. In the convecting regions the mean moist static energy of the upper troposphere is usually 1%–2% higher than the mean moist static energy of the lower troposphere. The intensity of convergence in the TCZ in the NH model is dependent upon a term in the denominator that is a small difference between two large terms. Hence, there can be large errors in the estimate of intensity of convergence using the NH model as a result of the errors in the measurement of specific humidity in the Tropics. NH model is, however, still useful for a qualitative understanding of the factors that govern the existence of TCZ.

5. Conclusions

The dependence of OLR on the surface moist static energy is similar to the dependence of OLR on SST. The threshold value of surface moist static energy for deep convection to occur differs for oceans and continents. However, the threshold of moist static stability for the troposphere (surface to 400 mb) is the same for oceans and continents. The relationship between OLR and moist static energy obtained for the troposphere (surface to 400 mb) from ERBE and ECMWF data is consistent with the NH model but not with the Seager (1991) model. The net energy convergence in the troposphere was found to be positive in regions with low OLR, thereby suggesting that this may be one of the conditions for the existence of TCZ. The net energy convergence in the troposphere will be positive if the sum of latent and sensible heat fluxes is higher than the net radiative cooling of the troposphere, hence the regions wherein the net energy convergence is positive is dependent strongly upon the manner in which the evaporative flux from the ocean is parameterized.

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