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Deep cloud clusters in the tropics

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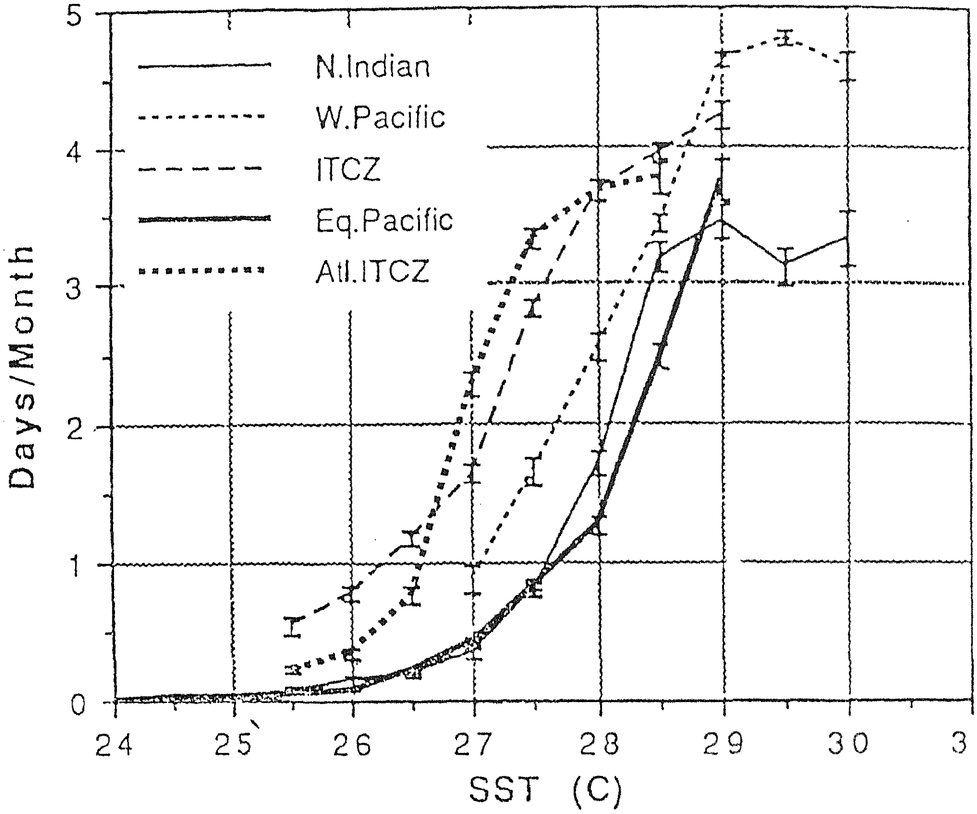
Abstract

Deep cloud clusters (DCC) play an important role in the transfer of energy and moisture from the surface to the upper troposphere of the earth. The factors which govern the existence of DCC over oceanic regions have been examined by many investigators and they have shown that the necessary condition for the existence of DCC in oceanic regions is that the sea surface temperature (SST) must be above 28 C. In this paper the factors which govern the existence of DCC in both continental and oceanic regions has been examined. It has been shown that the necessary condition for the existence of DCC in continental (oceanic) regions is that the surface moist static energy must be above 350 kJ/kg (340 kJ/kg). The difference in the moist static energy threshold between the oceans and the continents has been attributed to the differences in the vertical profile of moist static energy. If the mean moist static energy of the troposphere from surface to 400 millibar is used (instead of the surface moist static energy) then the threshold for the existence of DCC is 335 kJ/kg and is the same for both oceans and continents. The total energy convergence in the troposphere has been shown to be an important constraint which could reduce the presence of DCC in the warmest oceans or continent. It has been shown that there can be several routes to desertification when vegetation is removed from a continental region.

1. Introduction

Deep cloud clusters appear prominently in the daily satellite imagery of the tropics. Deep cloud clusters contain clouds with mean cloud top height above 8 kilometres. In the satellite imagery these clouds have monthly mean Outgoing Longwave Radiation (OLR) below 240 W/m^2. In the visible region the albedo (reflectivity to solar radiation) of DCC is above 50 %. They have spatial scales spanning the range from a few hundred to several thousand of kilometres. When the axis of DCC is oriented in the east-west direction, it is called the Inter-tropical Convergence Zones (ITCZ). Earlier ITCZ were also called Inter-tropical Front (ITF) or Inter-tropical Discontinuity (ITD) because it was believed that a large discontinuity in temperature and humidity must exist across the ITCZ. This assumption was based on the experience in the mid-latitudes where a large discontinuity in temperature and humidity is found across the cold polar front. In the tropics, there is, however, no discontinuity in temperature across the ITCZ. The term ITCZ was coined for the east-west oriented DCC to highlight the fact that the convergence of moisture in the ITCZ occurs primarily in the meridional direction and hence the moisture will converge from both northern and southern tropics. If there are two east-west oriented DCC at the same longitude then one of the DCC may not converge moisture from both hemispheres. Hence the term Tropical Convergence Zone (TCZ) may be more appropriate than ITCZ. The DCC occupy not more than 5% of the area of the tropics at any instant but they are essential for



(a) HRC

FIG. 1. The variation of the mean number of highly reflective cloud days (HRC) with sea surface temperature (SST) in different tropical oceanic regions (from Waliser et al , 1993).

the transfer of energy and moisture from the atmospheric boundary layer near the surface of the earth to the upper troposphere (Riehl, 1979).

What are the dynamical and thermodynamical factors which govern the existence of DCC? Gadgil *et al.* (1984) showed that the necessary (but not sufficient) condition for the existence of DCC in the oceanic regions is that the sea surface temperature (SST) must be above 28°C. They obtained cloudiness from the satellite data and SST from ship data in the northern Indian ocean for the period 1966–72. Graham and Barnett (1987) extended this result to all tropical oceanic regions. Recently, Waliser *et al* (1993) have examined the relationship between DCC and SST in greater detail by using satellite data for the period 1971–1990. Waliser *et al* (1993) have shown that the highly reflective clouds (HRC) data set was the most reliable indicator of the existence of DCC. The relationship between the number of highly reflective cloud days (HRC days) per month and SST obtained by Waliser *et al.* (1993) is shown in Fig. 1 for Indian, Atlantic and Pacific Oceans. We find that HRC occur when the SST is above a certain threshold and this threshold can be different for different oceans. Note that when the SST is above 29°C, the number of HRC

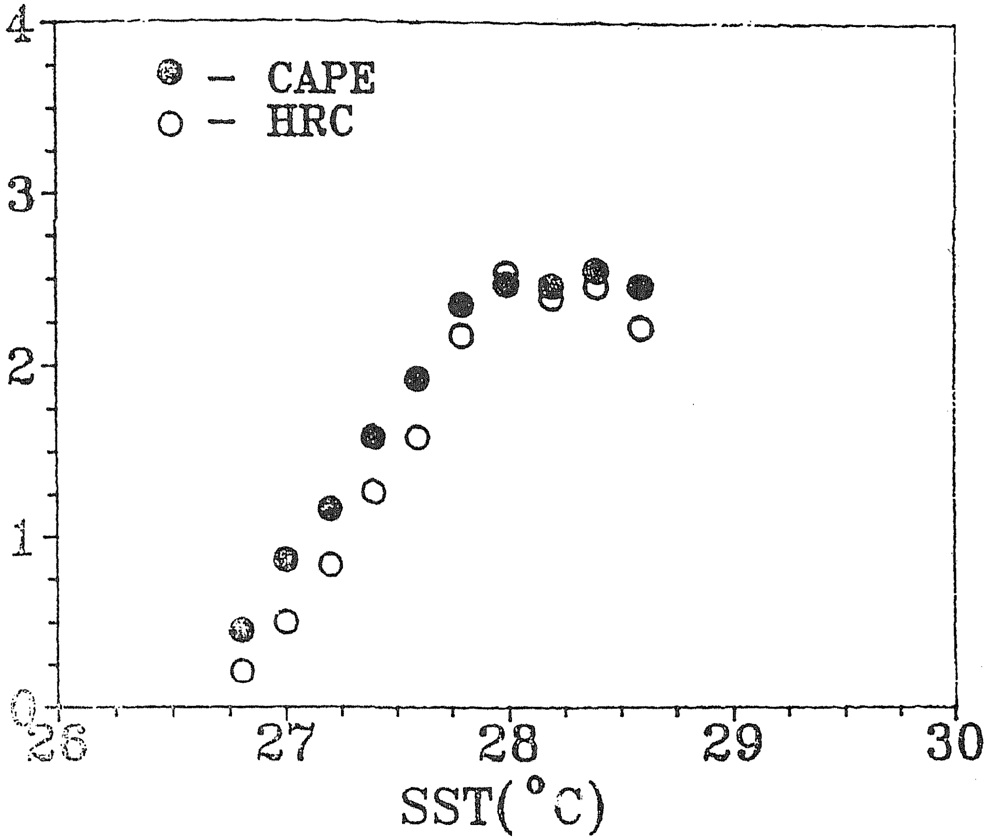


FIG. 2. The variation of the mean number of highly reflective cloud days (HRC) and monthly mean convective available potential energy (MCAPE) with sea surface temperature (SST) at Grantley Adams (from Bhat *et al.*, 1996).

days doesn't increase but reaches a plateau. Bhat *et al.*, (1996) have shown that the non-linear relationship between HRC and SST can be attributed to the non-relationship between the monthly mean Convective Available Potential Energy (MCAPE) and SST (see Fig. 2). The relationship between DCC and the surface parameters over the tropical continental regions have not been examined as closely as the relationship between DCC and SST in oceanic regions.

In this paper, we examine the factors governing the existence of DCC in the tropical continental regions and compare it with factors governing the existence of DCC in tropical oceanic regions. In section 2, we indicate the data used in the analysis. In section 3, we discuss the relationship between DCC and moist static energy in the continental regions which was discovered recently by Srinivasan and Smith (1996). In section 4, we show that the simple model for the TCZ proposed by Neelin and Held (1987) is useful to understand the factors governing the existence of DCC. The energetics of the DCC is discussed in section 5. In section 6, we provide a summary.

2. Data

In this paper the monthly mean data from the ECMWF (European Center for Mediumrange Weather Forecasting) analyses and NCEP (National Center for Environmental Prediction) re-analyzes and Earth Radiation Budget Experiment (ERBE) have been used. All the data are on 2.5^0 by 2.5 (latitude by longitude) grid and the analysis of the data is confined to the region 30 S to 30 N. When Outgoing Longwave Radiation (OLR) data is used we restricted the analysis domain to 25 N to 25 S since a low OLR over Tibet is not an indicator of DCC.

3. DCC over continents

The simple relationship between DCC and sea surface temperature that was obtained for oceanic regions (Fig 1) is not seen between DCC and surface temperature in continental regions. In Fig. 3, we have shown the relationship between OLR and surface temperature . In this figure, OLR data has been pooled into 0.5 K temperature bins. The mean and standard deviation in each bin is shown separately for continental and oceanic regions in Fig. 3. There is a decrease in monthly mean OLR when the SST goes above 300 K in oceanic regions. In continental regions OLR increases with increase in surface temperature although the increase is not monotonic. This is not surprising since there is no simple rela-

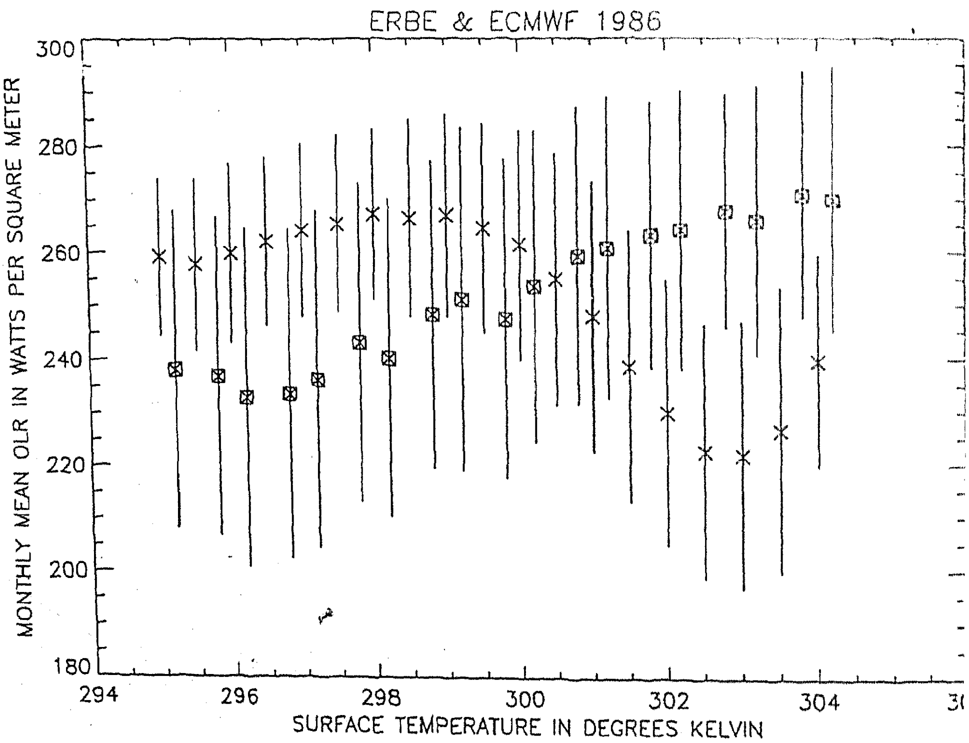


FIG. 3. The variation of monthly mean outgoing longwave radiation (OLR) with surface temperature for oceanic and continental regions in 1986 from ERBE data (from Srinivasan and Smith, 1996).

tionship between specific humidity and surface temperature over the continents while there is a strong correlation between specific humidity at the surface and sea surface temperature in oceanic regions (Wang and Li, 1993). Hence it would be necessary to use a surface parameter in continental region which takes into account the effect of surface temperature and surface specific humidity over the continents. The surface parameter we have chosen is the surface moist static energy since it is an indicator of the stability of the surface air (Seager, 1991). In Fig. 4 we have plotted monthly mean OLR versus the surface moist static energy. The OLR data was pooled into surface moist static energy bins from 330 kJ/kg to 360 kJ/kg in steps of 3 kJ/kg. The mean OLR in each bin and the standard deviation is shown in the Fig. 4. We find that the relationship between OLR and surface moist static energy in continental region is very similar to the relationship between OLR and SST over oceanic region. When OLR is plotted versus surface moist static energy we note that the surface moist static energy threshold at which the OLR decreases rapidly is different for oceanic and continental regions. If we assume that DCC can exist in regions with monthly mean OLR below 240 W/m^2, then the surface moist static energy threshold for continents (oceans) is around 340 (350) kJ/kg. Why should there be a difference in the surface moist static energy threshold for the continental and oceanic regions? This could on account of the differences in vertical profile of moist static energy in continental and

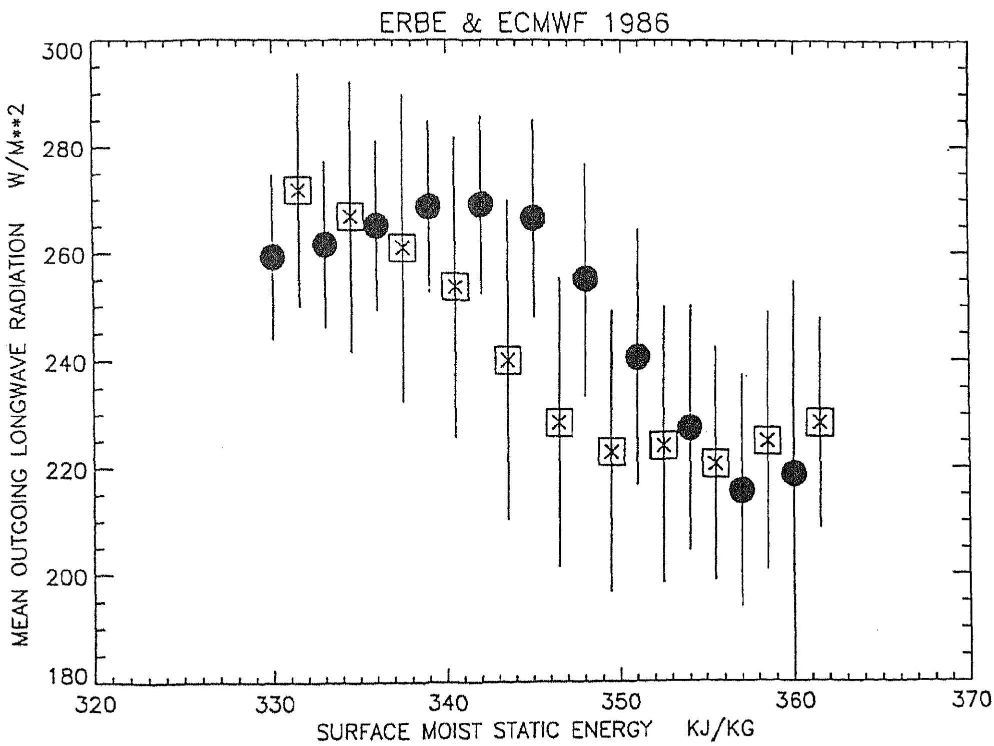


FIG. 4. The variation of monthly mean outgoing longwave radiation (OLR) with surface moist static energy for oceanic and continental regions in 1986 from ERBE data and ECNWF analyses (from Srinivasan and Smith, 1996). Solid circle ocean and square land.

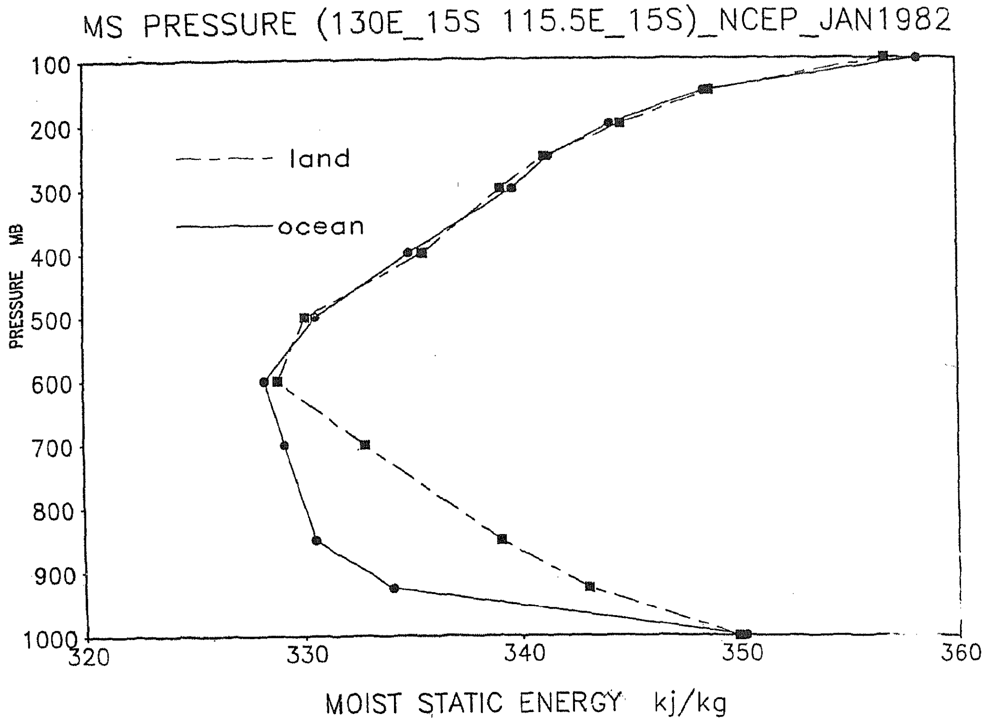


FIG. 5. The vertical variation of moist static energy (kJ/kg) in an oceanic and continental region having the same surface moist static energy from NCEP renalyses.

oceanic regions. Riehl (1979) has shown that the mid troposphere is wetter in the continental ITCZ than in the oceanic ITCZ. In Fig. 5, we have shown the vertical profile of moist static energy in two grids in the southern hemisphere (one oceanic and the other continental). The two grids have the same moist static energy at the surface but the moist static energy of the lower troposphere in the continental region is much higher than in the oceanic region. This means that the mean moist static energy of the lower troposphere in continental region will be higher than the mean moist static energy of the lower troposphere in oceanic regions even if the surface moist static energy in these regions is same. Hence we should expect that the threshold moist static energy for triggering of convection should occur at lower surface moist static energy in continental regions when compared to oceanic regions. Why is the lower troposphere wetter over continents than over oceans. One reason could be the strong diurnal variation of convection in continental regions. The clouds in the continental regions can penetrate deeper into the lower troposphere in the afternoon and hence moisten the lower troposphere more effectively than in oceanic regions where there is hardly any diurnal variation of SST. The relationship between OLR and mean moist static energy of the lower troposphere (surface to 400 mb) is shown in Fig. 6. We find the oceanic and continental regions show a similar trend and hence the moist static energy threshold is same in both cases. If we assume that DCC can exist in regions with monthly mean OLR below 240 W/m^2 then the threshold value of

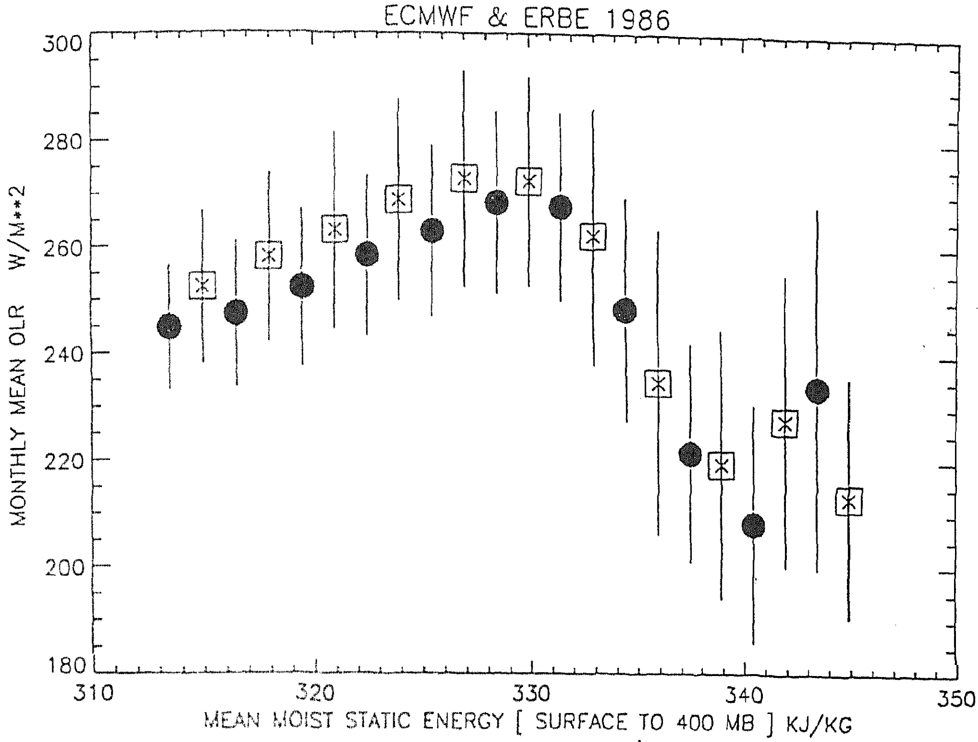


FIG. 6. The variation of monthly mean outgoing longwave radiation (OLR) with the mean moist static energy of the lower troposphere (surface to 400 mb) for oceanic and continental region in 1986 from ERBE data (from Srinivasan and Smith, 1996) Solid circle ocean. Square land

mean moist static energy of the lower troposphere is same for both oceans and atmosphere and equal to 335 kJ/kg. The relationship between surface moist static energy and the mean moist static energy of the lower troposphere is shown in Fig. 7 for continental and oceanic regions. We find that for the same surface moist static energy the continental regions have a higher mean moist static energy of the lower troposphere than the oceanic regions. In order to understand the implication of the results obtained above we need a simple model for the DCC. This is discussed in the next section.

4. Simple model for DCC

The factors which govern the location of DCC in nature and in General Circulation Models (GCM) of the atmosphere can be quite complex and subtle. Miller *et al.*, (1992) have shown that the location of DCC in the European Center for Medium Range Forecasting (ECMWF) model was sensitive to the manner in which the evaporation from the ocean (at low wind speeds) was parameterized. This result is puzzling since it is well known that the location of DCC in oceanic regions is governed mainly by SST and large scale convergence of moisture and not by local evaporation of moisture. The simple dynamical model of Gill (1980) and Davey and Gill (1987) assumes that the location of DCC over

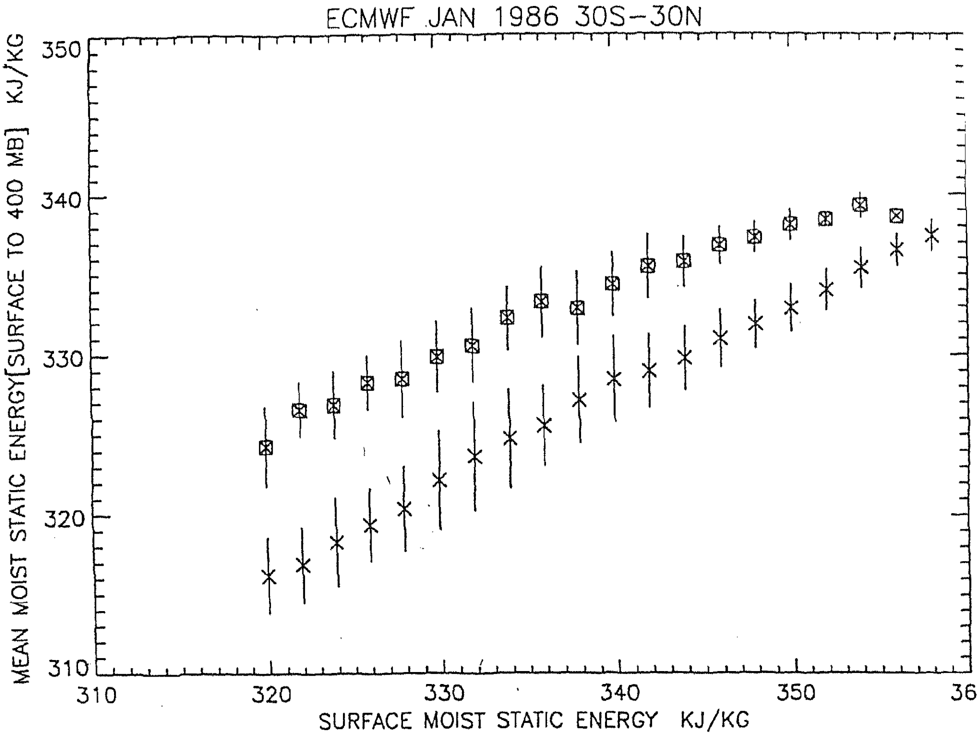


FIG. 7. The variation of the mean moist static energy of the lower troposphere (surface to 400 mb) with surface moist static energy for oceanic and continental regions in 1986 from ERBE data (from Srinivasan and Smith, 1996). Square land. Crosses ocean.

oceans is dependent upon the sea surface temperature only. The Davey and Gill (1987) model (DG henceforth) is simple and elegant but the location of DCC in this model is determined solely by the SST and not by the heat fluxes from the ocean. Hence the DG model cannot explain why the location of DCC in the ECMWF model is sensitive to the parameterization of surface evaporation. Seager (1991) has shown that the deficiency in the Gill (1980) model is not in the modeling of the dynamics but in the modeling of the thermodynamics.

In this section we discuss the Neelin and Held (1987) model (NH henceforth) in which the role of thermodynamics in the location of DCC has been stressed. In the NH model it is assumed that the DCC can be represented by a two-layer troposphere in which convergence occurs in the lower troposphere and divergence in the upper troposphere (Fig. 8). The two-layer structure assumed by NH is similar to that of DG model. In the NH model the equation governing the conservation of moist static energy is integrated from the surface to the top of the troposphere. The advection, transient and eddy terms in the moist static energy equation were neglected. The neglect of the advection terms can be justified since the meridional and zonal gradient of temperature, humidity and geopotential are small in the tropics. The neglect of transient terms

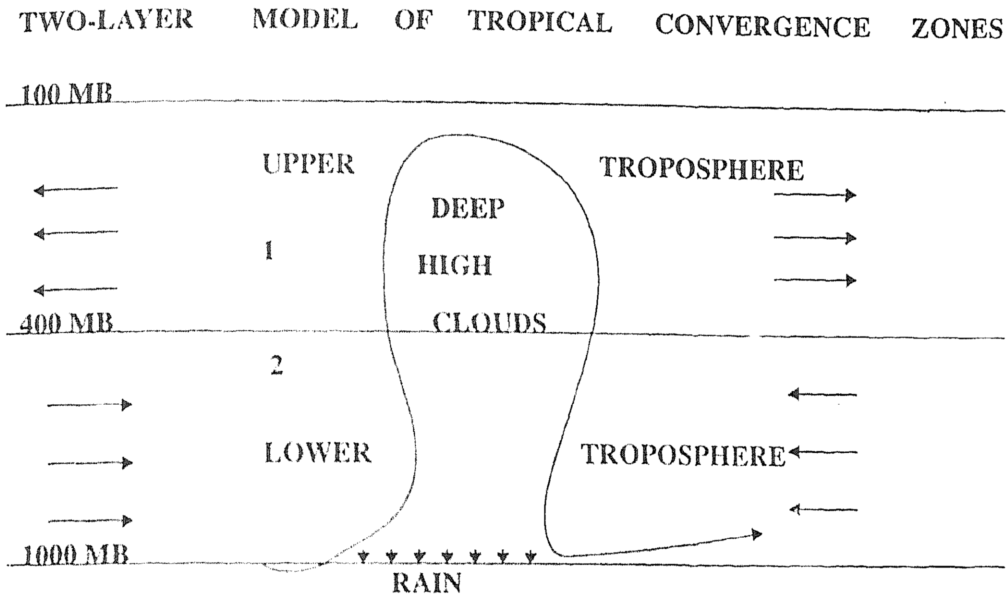


FIG. 8. The schematic diagram for the Neelin and Held (1987) model of the tropical convergence zones.

cannot be justified so easily. The integrated moist static energy equation can be written as follows:

$$F_N - F_B = C^*(m_1 - m_2) \tag{1}$$

In the above equation, F_N is the net radiative flux at the top of the troposphere, F_B is the net energy flux at the bottom of the troposphere, C is the convergence in the lower troposphere and m_1 and m_2 are the vertically averaged moist static energy in the upper and lower troposphere respectively. We define (F_N-F_B) as the net energy convergence in the troposphere. One advantage of using the equation for the conservation of moist static energy is the absence of latent heating term which is the dominant term in the energy conservation equation. The equation for the conservation of moist static energy is obtained by adding the equations for the conservation of dry static energy and moisture. When these two equations are added the latent heating term in the dry static energy conservation equation cancels with the moisture sink term in the moisture conservation equation. Hence the latent heating does not appear in the expression for the net energy convergence in the troposphere. Equation (1) shows that the energy convergence in the troposphere must be equal to the difference between the moist static energy exported out of the upper troposphere and the moist static energy imported into the lower troposphere in the DCC. The mean moist static energy of the lower troposphere (m_2) must be *less* than or equal to the mean moist static energy of the upper troposphere (m_1) on the monthly mean scale. If this is not true there will large-scale overturning of the two layers and the mixing between the two layers will ensure that m_1 approaches m_2 . Therefore (m_1-m_2) has been termed gross moist stability (GMS) by Neelin and Held (1987). Note that GMS is not directly related to the stability of a parcel lifted from the surface. GMS is related to stability of two adjacent layers but not to the stability of a parcel lifted from the surface. If GMS is negative the lower layer will be unstable and hence there will a large-scale rapid overturning of fluid between the lower troposphere and upper troposphere. Hence we should expect that GMS will positive when it is calculated on a monthly mean basis Thus the concept of stability enshrined in GMS is not the same as the concept of the stability of parcel lifted from the surface that is used in the study of cloud dynamics. The stability of a parcel depends both upon the temperature and humidity of the parcel as well as the environment. The gross moist stability (GMS) defined in the NH model is, however, dependent upon environmental profile of temperature and humidity and is not concerned with conditions within a parcel lifted from the surface. The concept of GMS is thus more akin to concept of stability of two adjacent layers than stability of a parcel.

Neelin and Held (1987) wrote the above equation in the following form to highlight the factors which govern the existence of DCC :

$$C = (F_N - F_B)/(GMS) \tag{2}$$

According to NH hypothesis, DCC will occur in regions with low GMS. This can be verified by examining the relationship between GMS and precipitation in the NCEP reanalyses. This is shown in Figure 9. We find the regions with high precipitation are confined to regions with low GMS. There are regions with low GMS in which there is no or small precipitation. We will show in the next section that in these regions the net energy convergence in the troposphere is small or negative. Since GMS is always positive, moisture

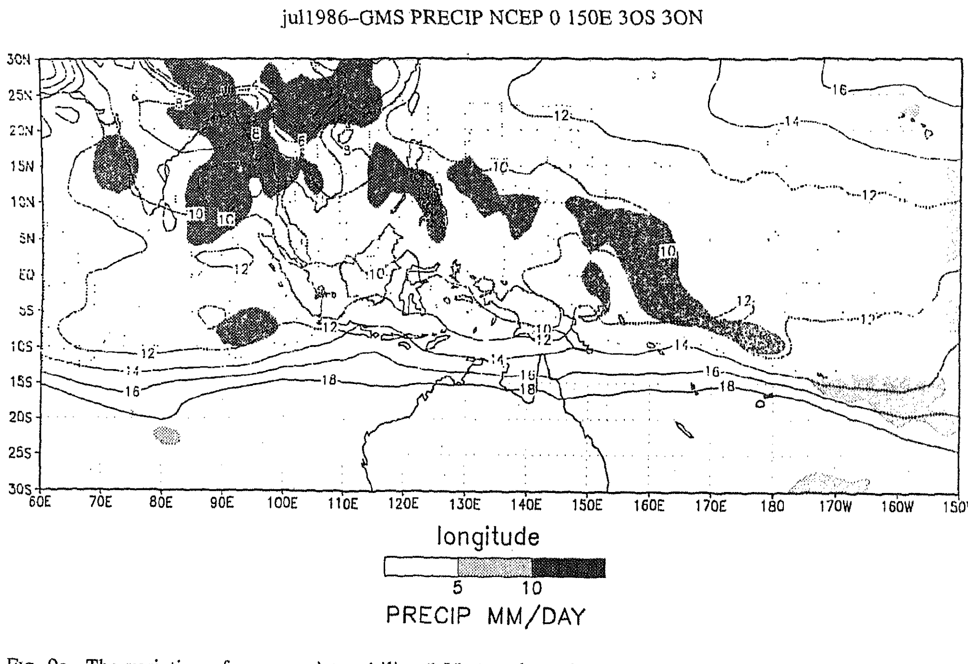


FIG. 9a. The variation of gross moist stability (kJ/kg) and precipitation (mm/day) in July 1986 in the Indo-Pacific region.

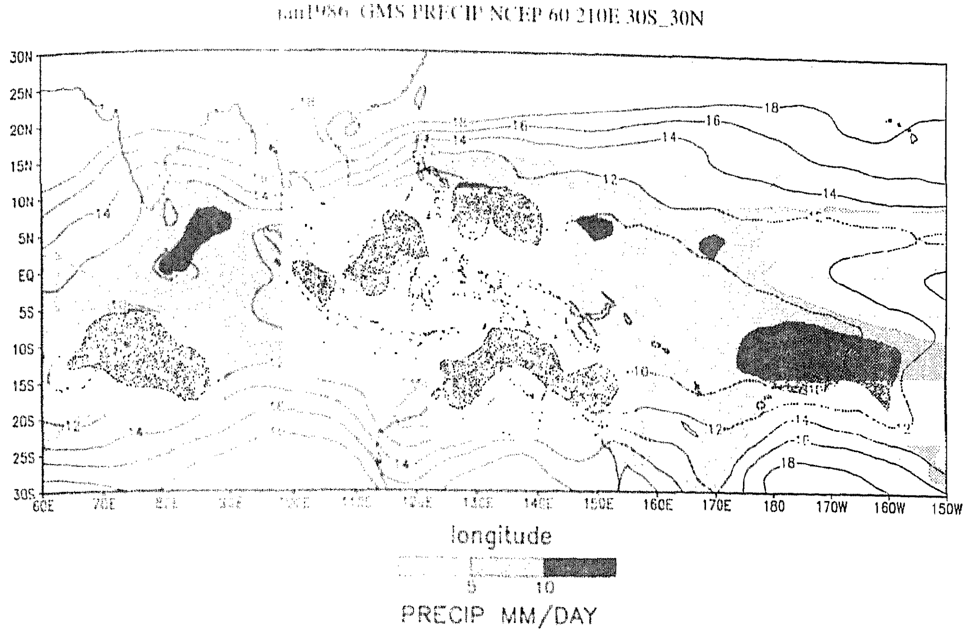


FIG. 9b. Variation of gross moist stability (J/kg) and precipitation (mm/day) in January 1986 in the Indo-Pacific region

convergence can occur in the lower troposphere if and only if the net energy convergence in the troposphere is positive. Hence DCC cannot occur in regions where the net energy convergence in the troposphere is negative. The main difference between continental and oceanic regions is the inability of the former to store much energy in the soil and the ability of the latter to store large amount of energy in the surface mixed layer of the ocean. Since the continental regions cannot store much energy, the net convergence of energy in the troposphere can be approximated quite well by the planetary net radiation at the tropopause. Since the regions above tropopause are in radiative equilibrium, the planetary net radiation at the tropopause is equal to the net radiation at the top of the atmosphere measured by satellites such as Earth Radiation Budget Experiment (ERBE). The NH model predicts that in continental regions DCC cannot exist if the net radiation at the top of the atmosphere is negative. The NH model is superior to other simple models of the tropical atmosphere because it makes a prediction which can be easily verified using satellite data alone. We have indicated that in the satellite data regions with low monthly mean OLR can be classified as DCC. In Fig. 10, the monthly mean OLR (in a 2.5° by 2.5° region) is plotted versus the monthly mean net radiation (in the same region) from ERBE data for all regions in the tropics in the latitude range 25° N to 25° S. We have, deliberately, excluded the Tibetan plateau because a low OLR in this region does not necessarily indicate the presence of deep clouds. We find the regions where OLR is below 200 W/m^2 the planetary net radiation is always positive. In this figure both oceanic and continental regions have been included. If DCC exists in an oceanic region the net radiation at the top of the atmosphere must be higher than that over the continent in order to account for the energy

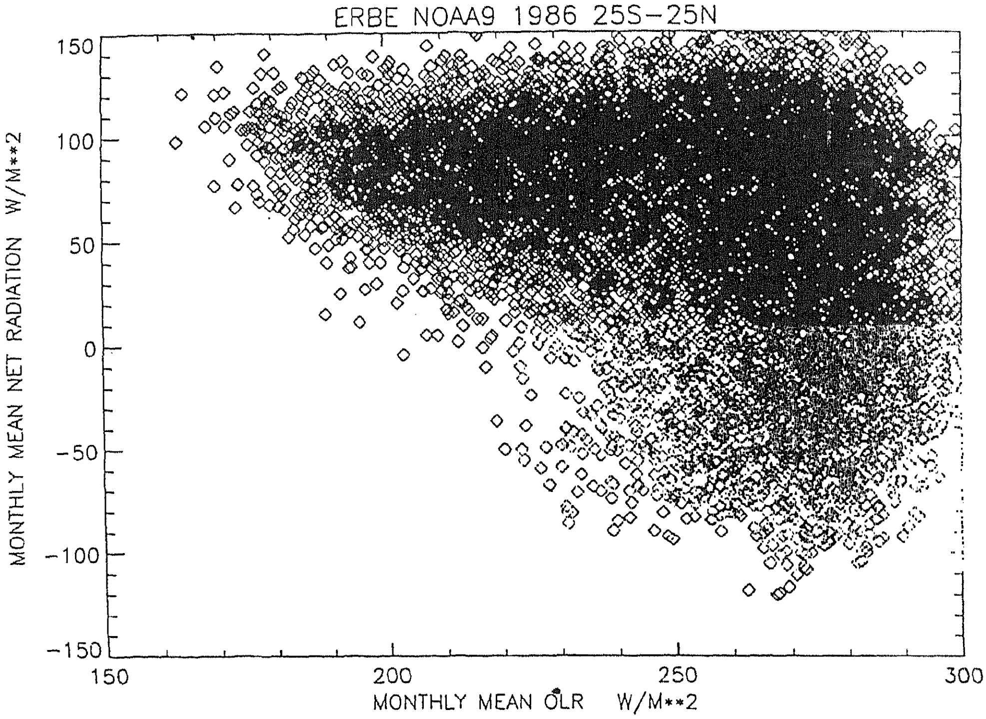


FIG. 10. The variation of monthly mean net radiation with monthly mean outgoing longwave radiation (OLR) in January 1986 from ERBE data (from Srinivasan and Smith, 1996).

absorbed in the oceanic surface mixed layer which is not immediately available to the atmosphere. The relationship between OLR and net radiation at the top of the atmosphere obtained from ERBE data is consistent with the predictions made by the NH model and hence must be considered as posterior justification for the assumptions made in the NH model.

5. Energetics of DCC

The NH model provides new physical insight regarding the energetics of the DCC. We find that net energy available in the troposphere is equal to the net moist static energy exported out of the DCC. If the moist static energy of the lower and upper troposphere are equal then there is no constraint on the amount of mass that converge in the DCC. In the DCC, the moist static energy of the upper troposphere is usually slightly higher than that of the lower troposphere and this small difference imposes a strong constraint on the amount of mass that can converge in the lower troposphere. If the net energy available in the troposphere is negative (which is usually the case in the winter hemisphere) then there can be no convergence of mass in the lower troposphere and hence DCC cannot exist in such regions.

The net energy available in the troposphere is usually positive in the summer hemisphere. There could, however, be some remarkable exceptions. One of them is the Sahara

Desert. Srinivasan and Smith (1996) have shown that the planetary net radiation (and hence the net energy convergence in the troposphere) is negative in the Saharan region. This could be on the reasons why the DCC are never observed in Africa in the region north of 15° N latitude. The ERBE data indicates that in the Boreal summer the region north of 15° N is the only region in the northern hemisphere which has negative planetary net radiation (Srinivasan and Smith, 1996). This unusual feature of the Sahara desert is on account high albedo of the desert surface and the low relative humidity of the air . Hence the reflected solar radiation and the outgoing longwave radiation (OLR) are both large in this region. Charney (1975) had argued that the removal of vegetation can lead to an increase the surface albedo and thus cause the net energy available in the troposphere to be negative in summer. This will mean that DCC cannot occur in that region and hence will lead to desertification. The NH model suggests two more routes to desertification. The continental regions cannot store much energy in the soil and hence, on a monthly mean scale, the net radiation incident on the continental surface is equal to the latent and sensible heat fluxes from the surface. Hence the term F_B in equation 2 can be set equal to zero. Therefore equation 2 can be written ,for continental regions, as

$$C = F_N/GMS = \{S(1-\alpha)-OLR\}/GMS \quad (3)$$

In the above equation S is the solar energy incident at the tropopause , α is the albedo of the earth-atmosphere system, OLR is the outgoing longwave radiation at tropopause, and

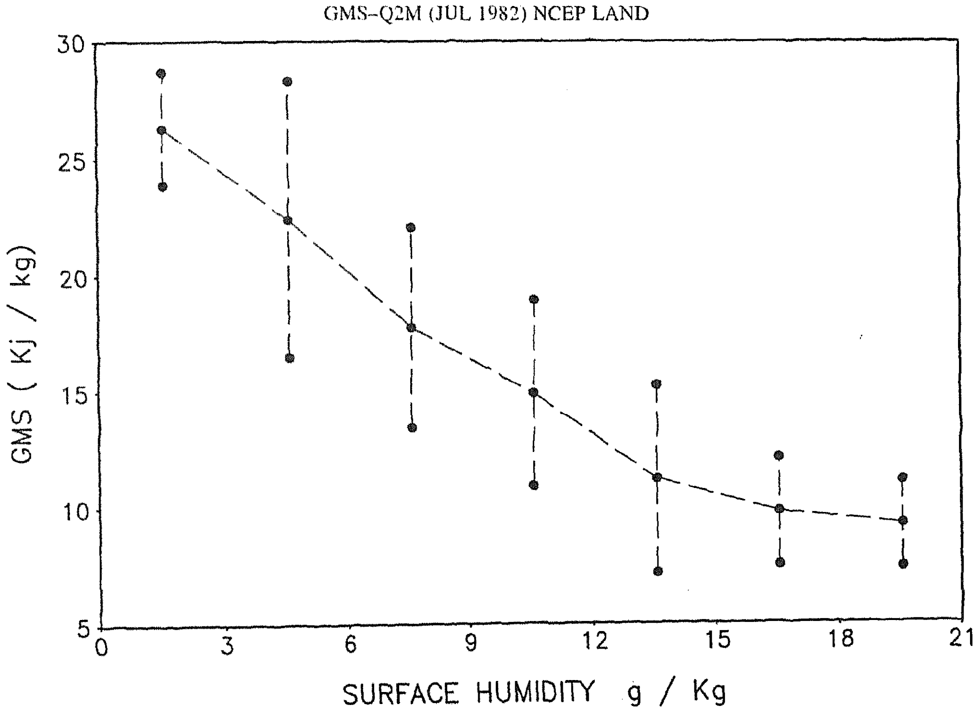


FIG. 11. The variation of gross moist stability with surface specific humidity.

GMS is the gross moist stability. According to equation 3 there three ways in which convergence can decrease in the lower troposphere. The first is on account of increase in albeo when vegetation is removed. This is the route to desertification suggested by Charney (1975). The NH model suggests two other routes. When vegetation is removed, less moisture is pumped to the atmosphere from beneath the surface. This will lead to the reduction of the specific humidity at the surface. The reduction of specific humidity at the surface will result in an increase in GMS (see figure 11) and OLR. The removal of vegetation will increase reflected solar radiation and OLR and hence cause a reduction of F_N (*i.e.*, numerator of equation 3). These three factors can act together and cause a dramatic reduction in the convergence of moisture in the lower troposphere. The three different routes to desertification is shown schematically in figure 12.

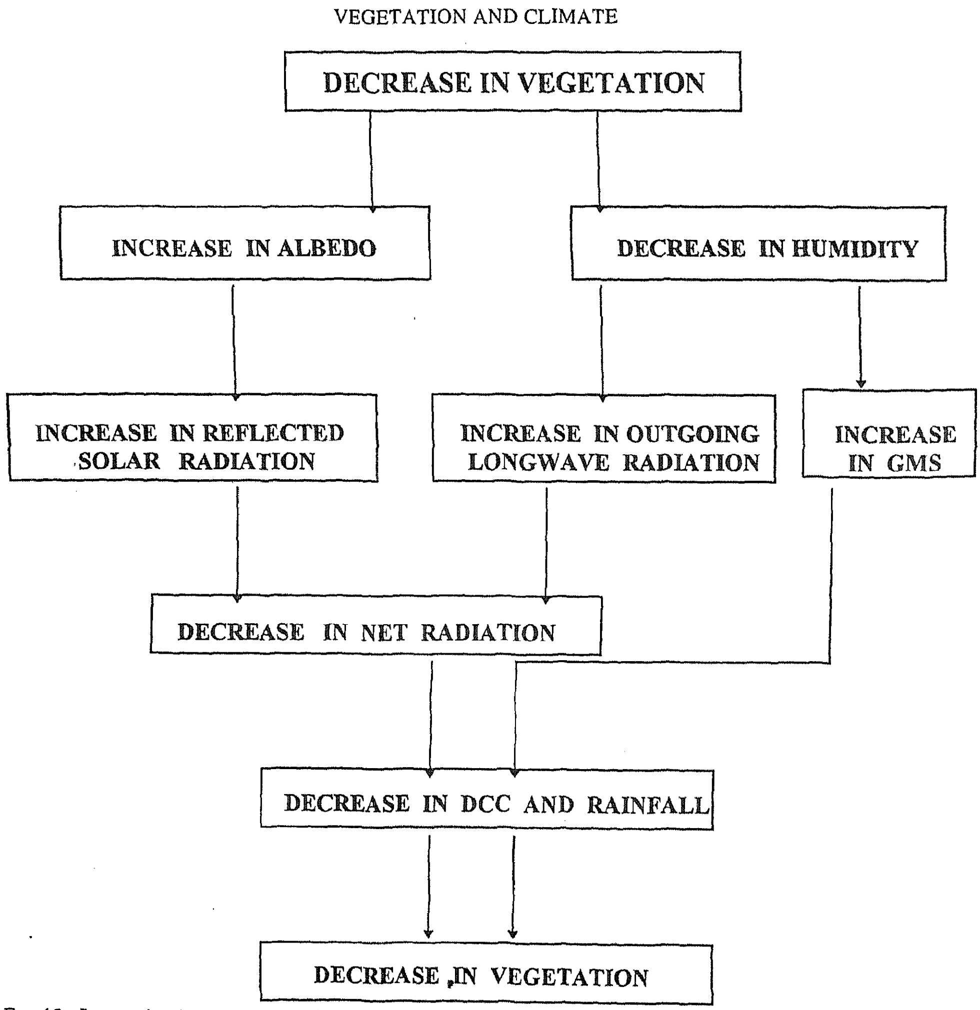


FIG. 12. Interaction between vegetation and climate.

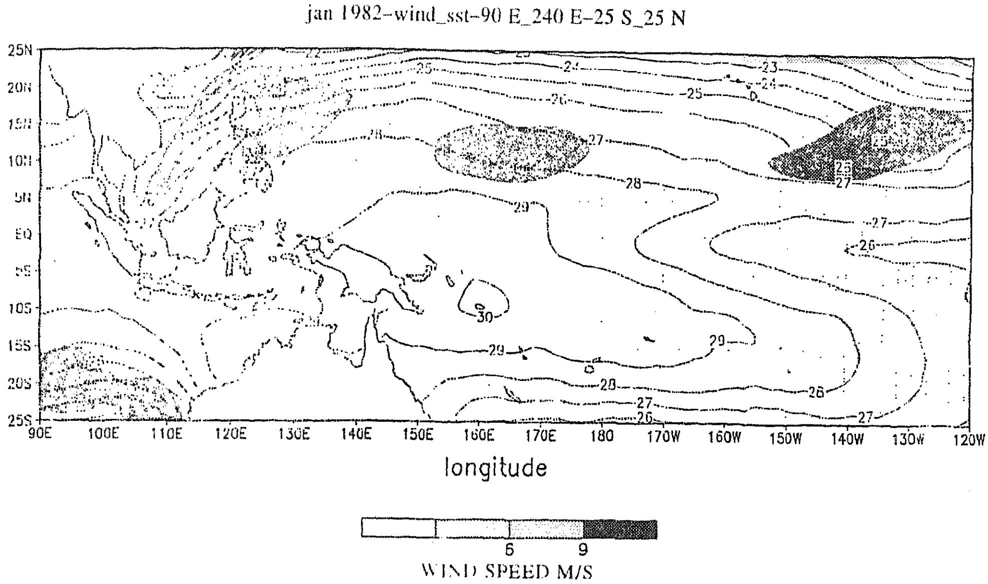


FIG. 13. The variation of sea surface temperatures and surface wind speed during January 1982 from NCEP reanalyses. Contours indicate sea surface temperature in degree celsius.

The mechanism for desertification proposed here is quite different from that proposed by Charney (1975) because it doesn't depend upon the increase in surface albedo alone. We have shown that desertification can be caused by different mechanisms. One is the increase in albedo on account of the removal of vegetation and the other is the drying of the troposphere due to removal of vegetation because less moisture is pumped from beneath the soil. The drying of the troposphere can cause reduction of rainfall because of reduction of net radiation (through an increase in OLR) and increase in GMS. On the other hand, if a desert region is planted with vegetation, the albedo decreases and more moisture is pumped from beneath the soil and hence the relative humidity in the troposphere increases. This will increase planetary net radiation and reduce GMS and hence increase the probability of DCC appearing over that region. Once the DCC appears over that region it will lead to a positive feedback mechanism and hence can result in an increase in the amount of vegetation.

Another region in the tropical summer hemisphere which is remarkable is the warm pool of the west Pacific. In this region the SST is always above 28°C and hence satisfies the condition necessary for the existence of DCC. This region is, however, not always covered by DCC. Waliser and Graham (1993) have shown that DCC occur less often in regions with SST above 29°C in the warm pool of the west Pacific ocean than in regions with SST between 28°C and 29°C. This surprising result can be explained by the NH model by invoking the fact that the net energy available in the troposphere must be positive. In the warm pool of the west Pacific the surface wind speed can be lower than 3 m/s in some regions (Fig. 13). In such regions the latent heat flux can be small and hence may not be able to overcome the radiative cooling of the troposphere. In such regions, the net

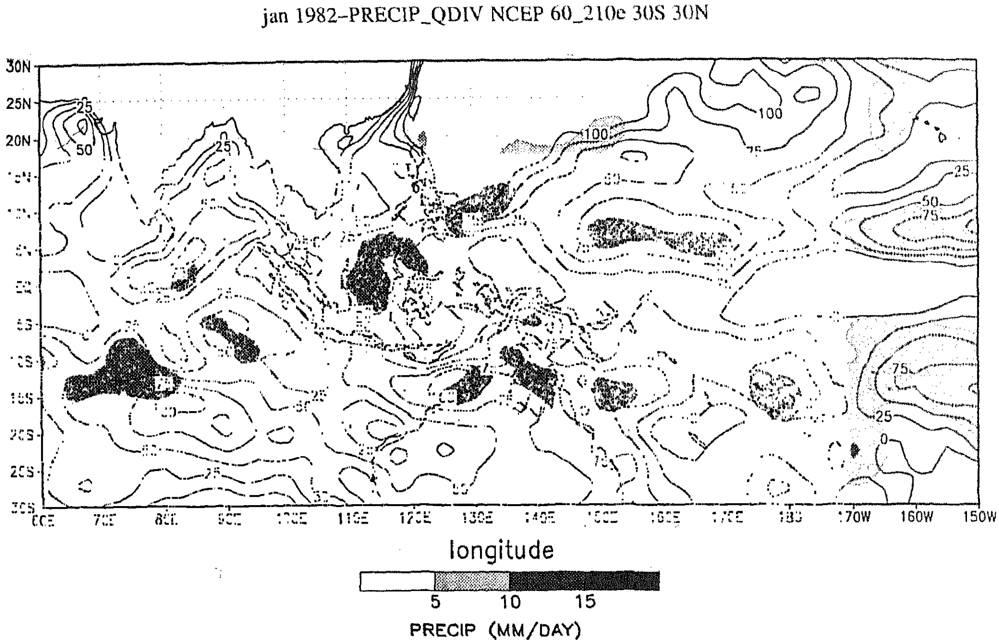


FIG. 14. The variation of the energy convergence in the troposphere and precipitation during January 1982 from NCEP renalyses. Contours indicate energy convergence in W/m^2.

energy convergence in the troposphere can be negative and hence DCC cannot exist in such regions although the SST is above 28°C. The net energy convergence in the troposphere and precipitation are shown in Fig. 14 for the Indo-Pacific region in January 1982. We find that precipitation is low or zero in regions with negative tropical energy convergence. This result highlights the crucial role played by the energy availability in the troposphere on the existence of DCC in regions which are otherwise favourable for the presence of DCC.

The NH model highlights the role of energetics and thermodynamics on the location of DCC. In this model the vertical structure of the moist static energy determines the gross moist stability (GMS). This term is, however, quite different from the concept of Convective Available Potential Energy (CAPE) used in the study of cloud dynamics. This is because GMS is not directly related to the stability of a parcel lifted from the surface of the ocean or continent. The convective available potential energy (CAPE) can be computed as an integral of the difference between the virtual temperature (of a parcel lifted from the surface) and the environment (Bhat *et al.* 1996). On the other hand, the gross moist stability is obtained from the vertical integration of the environmental moist static energy profile in the converging and diverging regions separately. Bhat *et al.* (1996) have shown that there is a linear relationship between HRC and CAPE based on monthly mean soundings (called MCAPE by Bhat *et al*, 1996). What is the relationship between MCAPE defined by Bhat *et al.* (1996) and gross moist stability (GMS) defined by Neelin and Held (1987)? In Fig. 15 the contours of MCAPE and GMS are shown for oceanic regions in July 1986

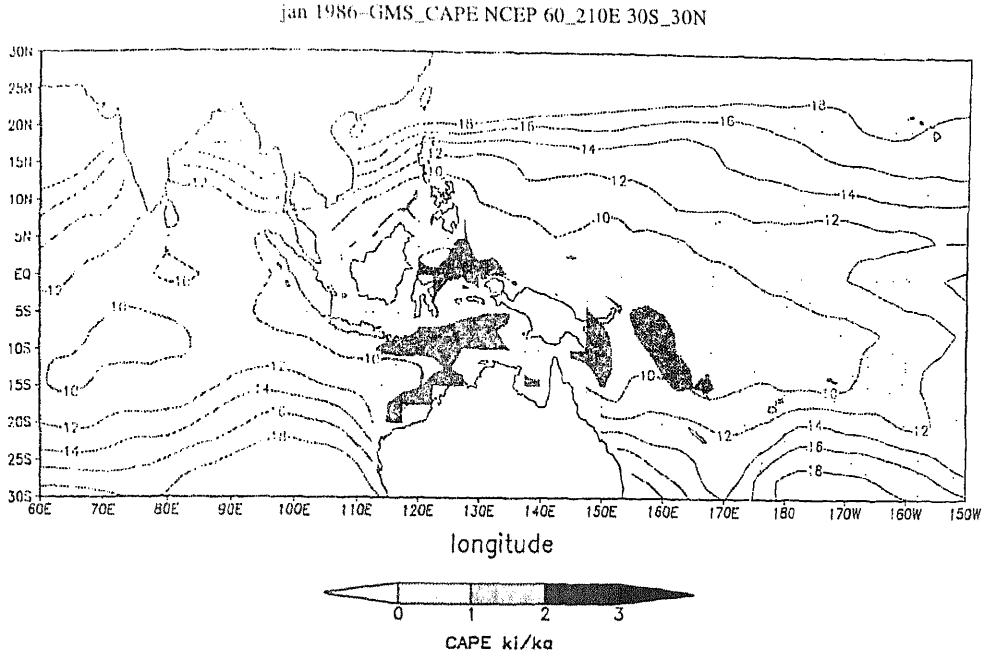


FIG. 15. The variation of gross moist stability (GMS) and monthly mean convective available potential energy (MCAPE) in oceanic regions during January 1986 from NCEP renalyses. Contours show GMS in Kj/Kg.

based on NCEP reanalyses . We find that the pattern of variation of MCAPE and GMS are very similar. Thus the factors which contribute to a decrease in GMS cause an increase in CAPE. The strong coupling between GMS and precipitation is seen in Fig. 16 where the monthly mean GMS is plotted against monthly mean precipitation for all regions in the tropics in the month of January for the period 1982 to 1994. In the regions with precipitation above 12 mm/day, the GMS is confined to the values in the range of 8 to 12 kJ/kg. In view of the strong coupling between GMS and precipitation, it is appropriate to ask whether low GMS is a necessary condition for high precipitation or does high precipitation lead to moistening of the lower troposphere and hence a reduction of GMS. The inspection of the daily variation of GMS and precipitation in the NCEP reanalyses indicates that in most cases the decrease in GMS leads the increase in precipitation and hence it looks as though a low GMS is a pre-requisite for the incidence of high rainfall. This issue needs to examined in greater detail.

Conclusions

We have shown that there are two conditions for the existence of DCC in continental and oceanic regions. The first condition is that the net energy convergence in the troposphere should be positive. The second condition is that the mean moist static energy of the lower troposphere should be above 335 kJ/kg. We have shown that the absence of DCC in the Sahara desert during Boreal summer could be on account of high albedo as well as high

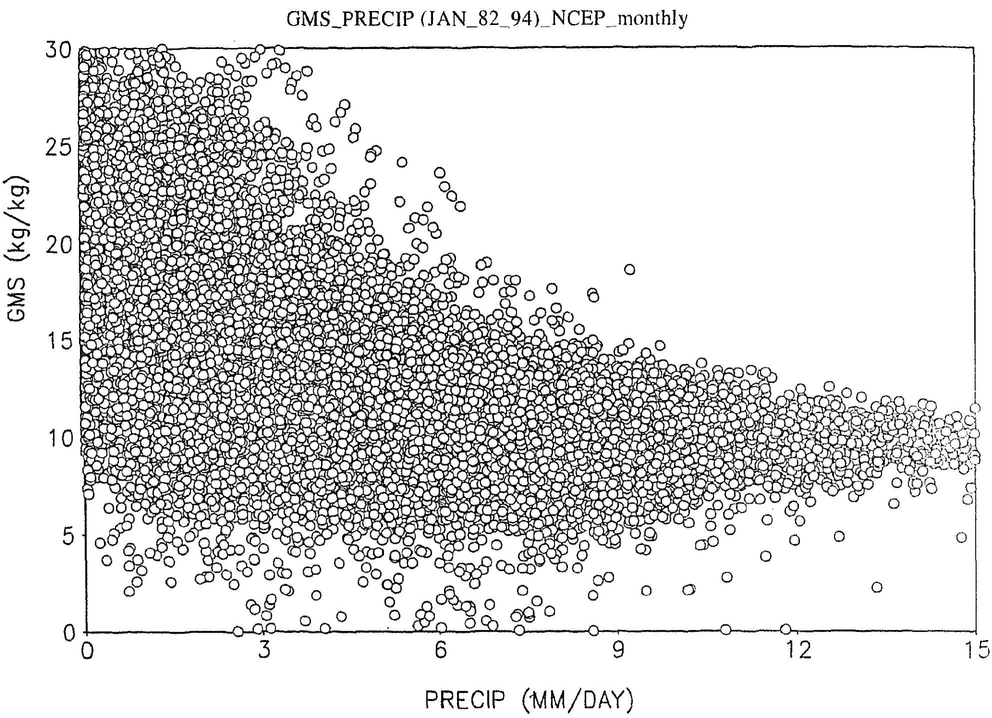


FIG. 16. The gross moist stability (GMS) versus precipitation based on monthly mean fields from 1982 to 1994 from NCEP renalyses.

OLR. We have thus shown that there can be more than one route to desertification when vegetation is removed from a continental region.

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