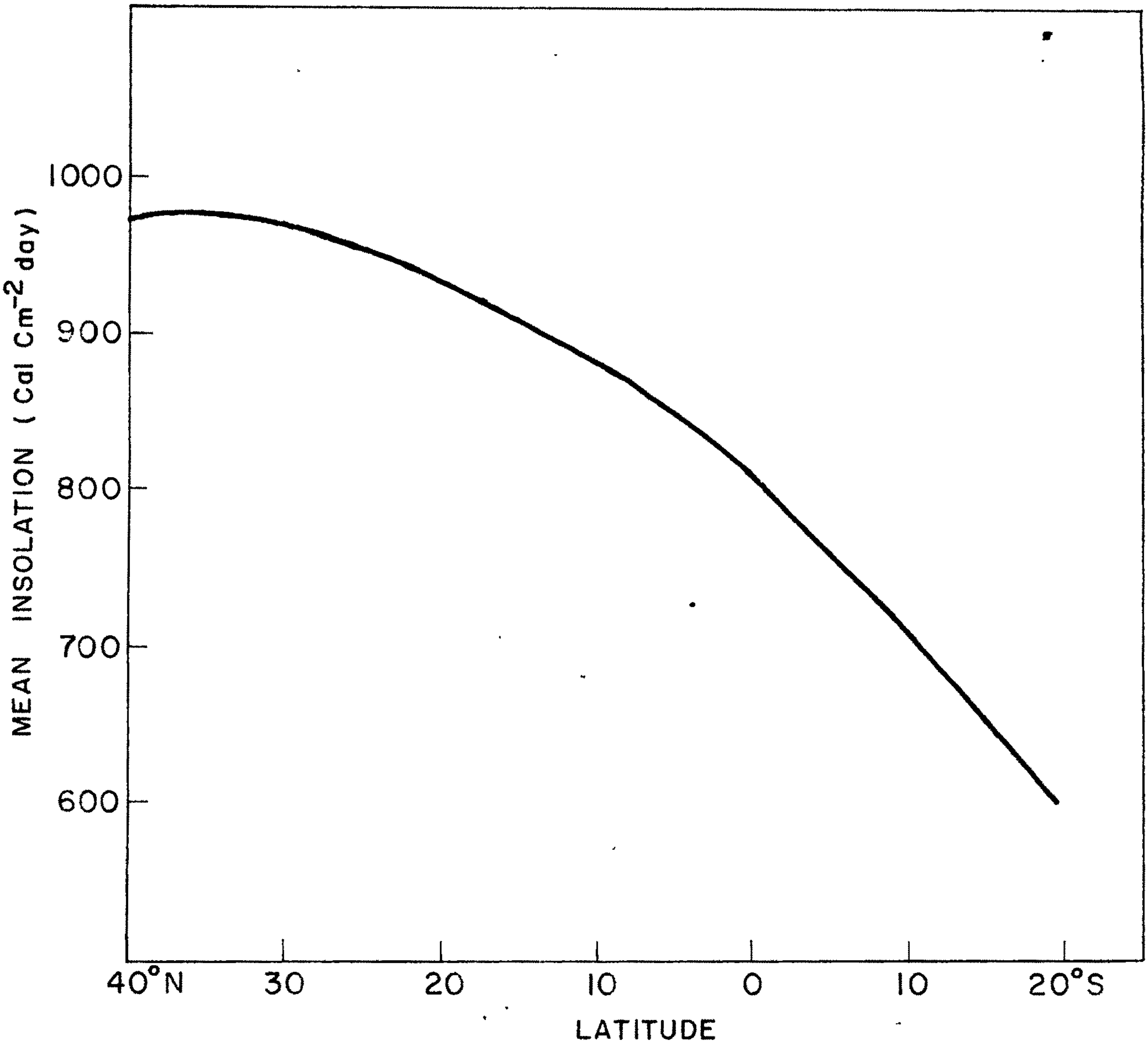


Figure 3.1.1 Variation of solar insolation (cal cm$^{-2}$ day$^{-1}$) with latitude at the top of the atmosphere.

-ble heat transfer across the earth-atmosphere interface. The major components that contribute to the diabatic heating and cooling, thus consist of radiative heat exchange between the earth and the free atmosphere and eddy exchange of heat and moisture in the surface boundary layer.

The importance of diabatic heating has been stressed by Petterssen (1956) in the theory of cyclone development. The short-term influence of heat and moisture on individual weather systems have been investigated by Pettersen et al (1962). The effect of radiation on such systems has been examined by Godbole (1963). These investigations have clearly shown that the distribution of diabatic heat sources and sinks excercise profound influence upon the weather systems.

Since the monsoon system is essentially of thermal origin, it is instructive to examine as to what extent diabatic heating is responsible for its develop--ment and maintenance. With this end in view, an attempt is made in the following to analyse the fluxes of solar and infra red radiations at the surface for the entire region under consideration and the fluxes of sensible and latent heat for the oceanic part of the region at the ocean-atmosphere interface.

## 3.1 Solar Radiation

Figure 3.1.1 shows the distribution of solar insolation as taken from the Smithsonian Tables (Table

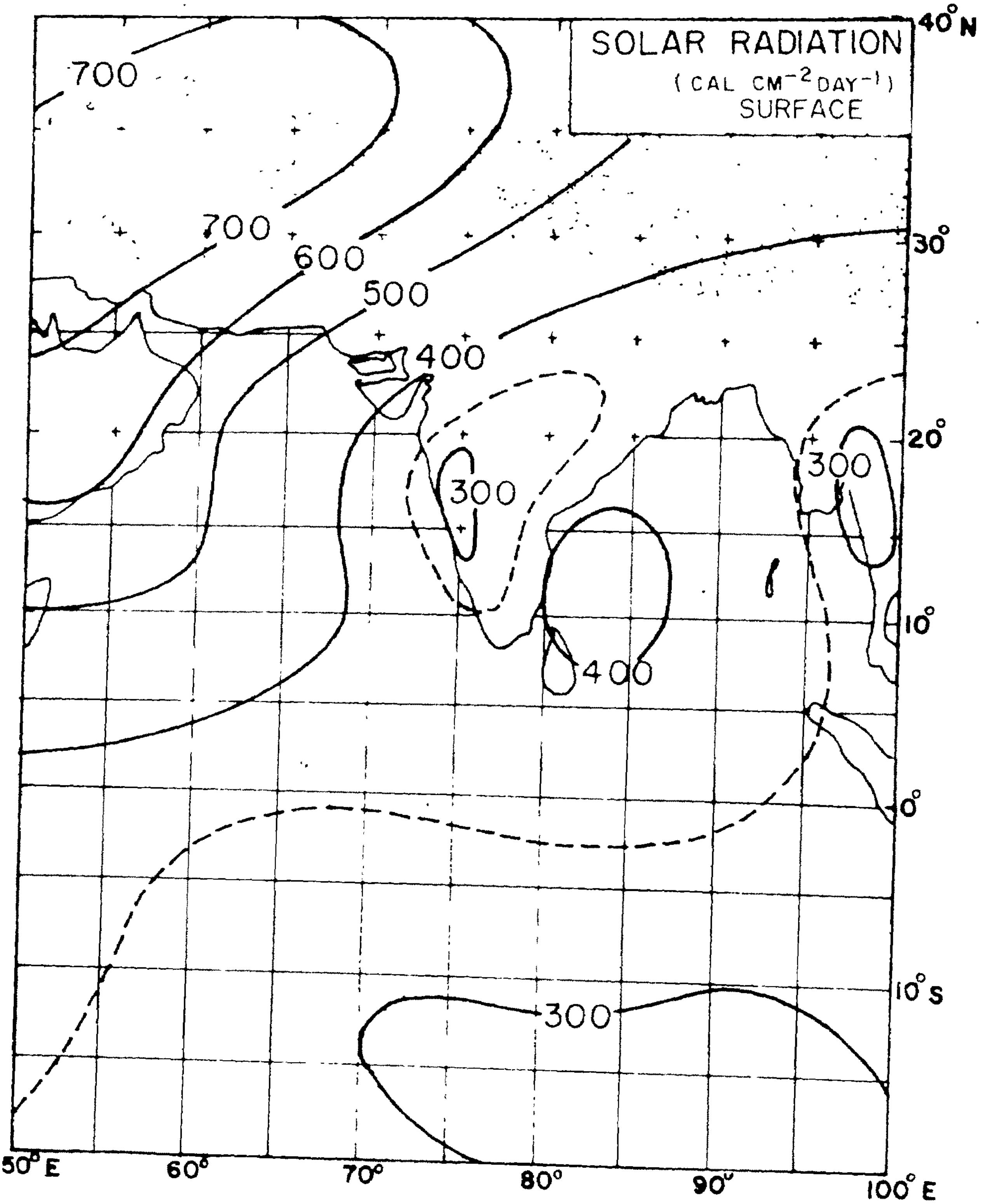


Figure 3.1.2 Distribution of global solar radiation (cal cm$^{-2}$ day$^{-1}$) at the surface for July.

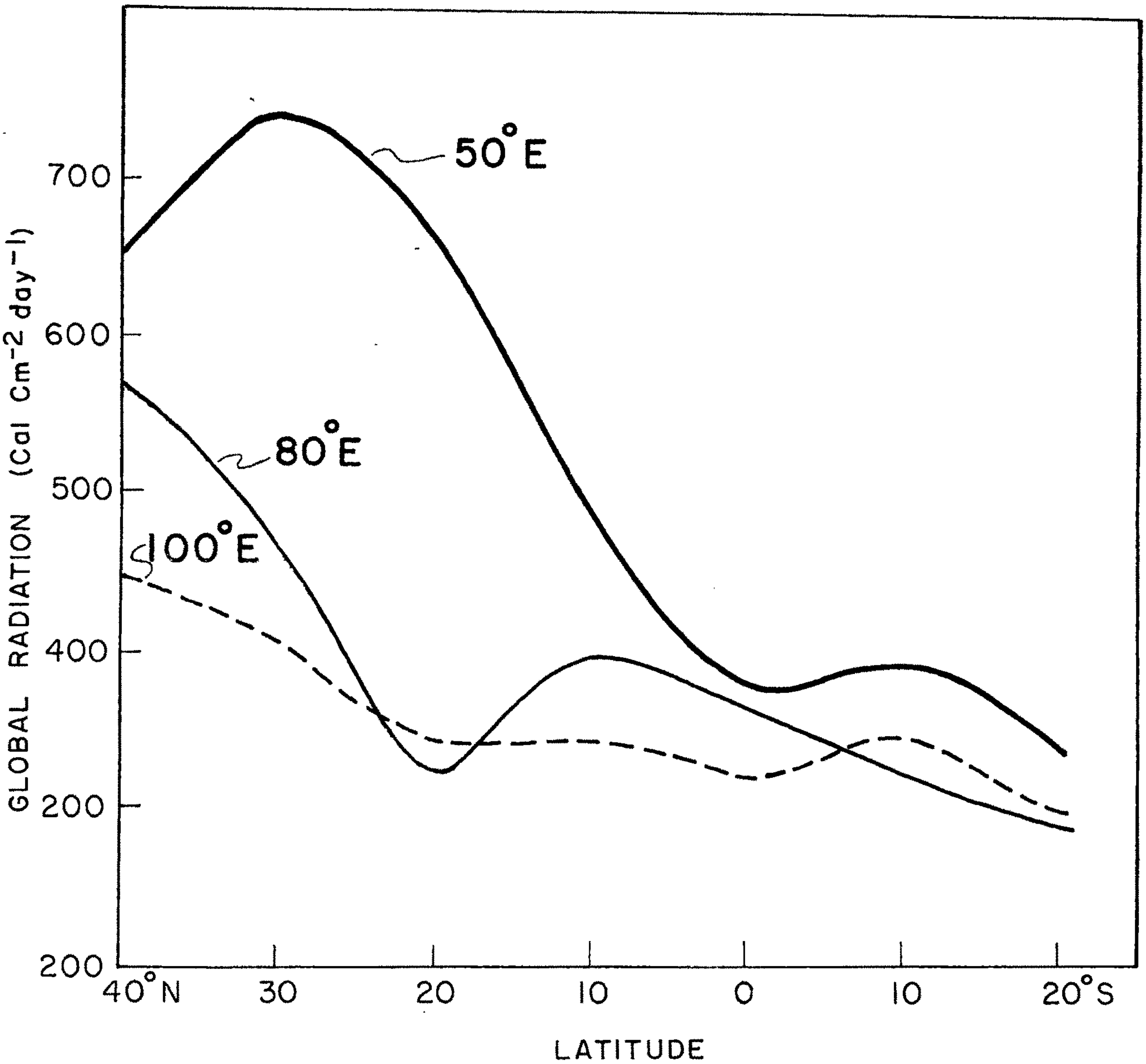


Figure 3.1.3 Variation of global solar radiation (cal cm$^{-2}$ day$^{-1}$) at the surface with latitude for 50°E, 80°W, and 100°E for July.

No. 132), at the outer boundary of the atmosphere for the month of July. The maximum insolation is found at 35° N because, at this latitude, the product of the duration of sunlight and the zenith distance of the sun is maximum for July.

The solar radiation, in course of its passage through the atmosphere, undergoes absorption by the gases of the atmosphere and reflection from cloud tops. The scattering due to air molecules and dust particles gives rise to an additional component of radiation called the sky radiation. The downward flux of solar ratiation or the global solar radiation, therefore, consists of direct radiation from the sun and the sky radiation. The spatial distribution of global solar radiation at the earth's surface for July, as reproduced from Mani et al (1967), is shown in Figure 3.1.2. It is seen that the distribution of land and sea distorts the zonal pattern which is implicit in Figure 3.1.1. Minimum radiation ( $< 300$ cal $cm^{-2}$ $day^{-1}$ ) is received over the west coasts of India and Burma which are characterised by heavy clouding (see Figure 3.2.1), and the largest amount of precipitable water (Godbole and Kelkar, 1969 a). Minimum values are also observed over the south Indian ocean but this is due to the low solar trajectory. Maximum radiation ( $> 700$ cal $cm^{-2}$ $day^{-1}$ ) is found over the north-western part of the region (i.e. Arabia and Afganistan ) where the clouding and moisture content are the least.

Figure 3.1.3 has been prepared from Figure 3.1.2 to show more clearly the variation of global radiation with latitude and longitude. Comparision with Figure 3.1.1 reveals that the distribution of global radiation, unlike insolation at the top of the atmosphere, is dependent upon longitude also.

The amount of solar radiation which is absorbed by the earth's surface depends upon the surface albedo. The value of the surface albedo not only determines the surface temperature but also affects the air temperatures in the vertical, more so at the lower levels (Godbole et al, 1969 b).

## 3.2 Long Wave Radiation

As a result of the absorption of solar radiation the earth and the atmosphere are heated and, in turn,emit radiation in the infra red or long wave range. Unlike solar radiation, the long wave radiation undergoes repeated absorption and emission by the gases of the atmosphere throughout the vertical column. It can be shown from the first principles that the result of the radiative heat exchange between the earth and the atmosphere is that there is a net radiation in the upward direction at every level and that the amount of net upward radiation increases with height. Long wave radiation, therefore, unlike solar radiation which heats up the earth-atmosphere system, contributes to its cooling.

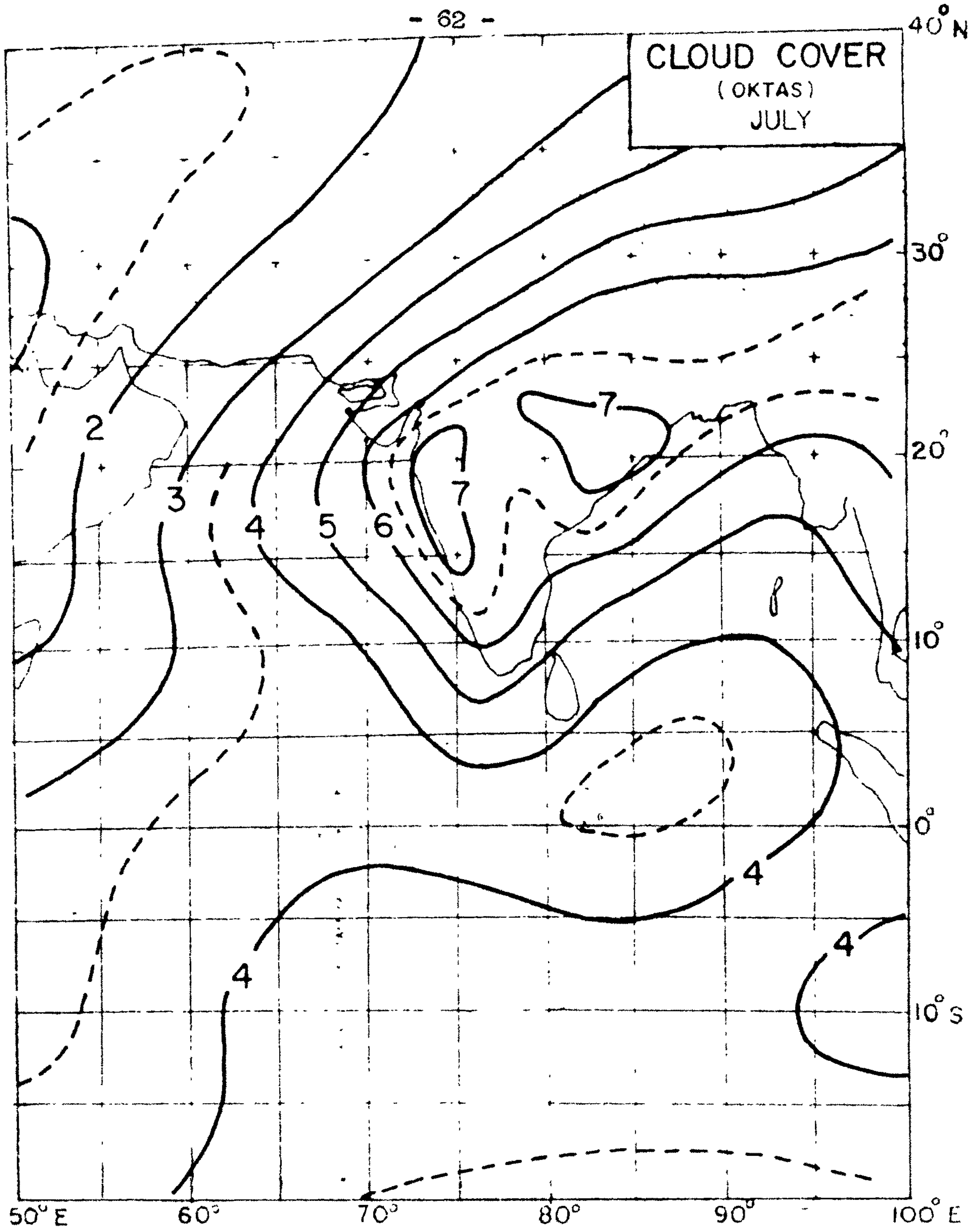


Figure 3.2.1 Distribution of cloud cover ( oktas ) for July.

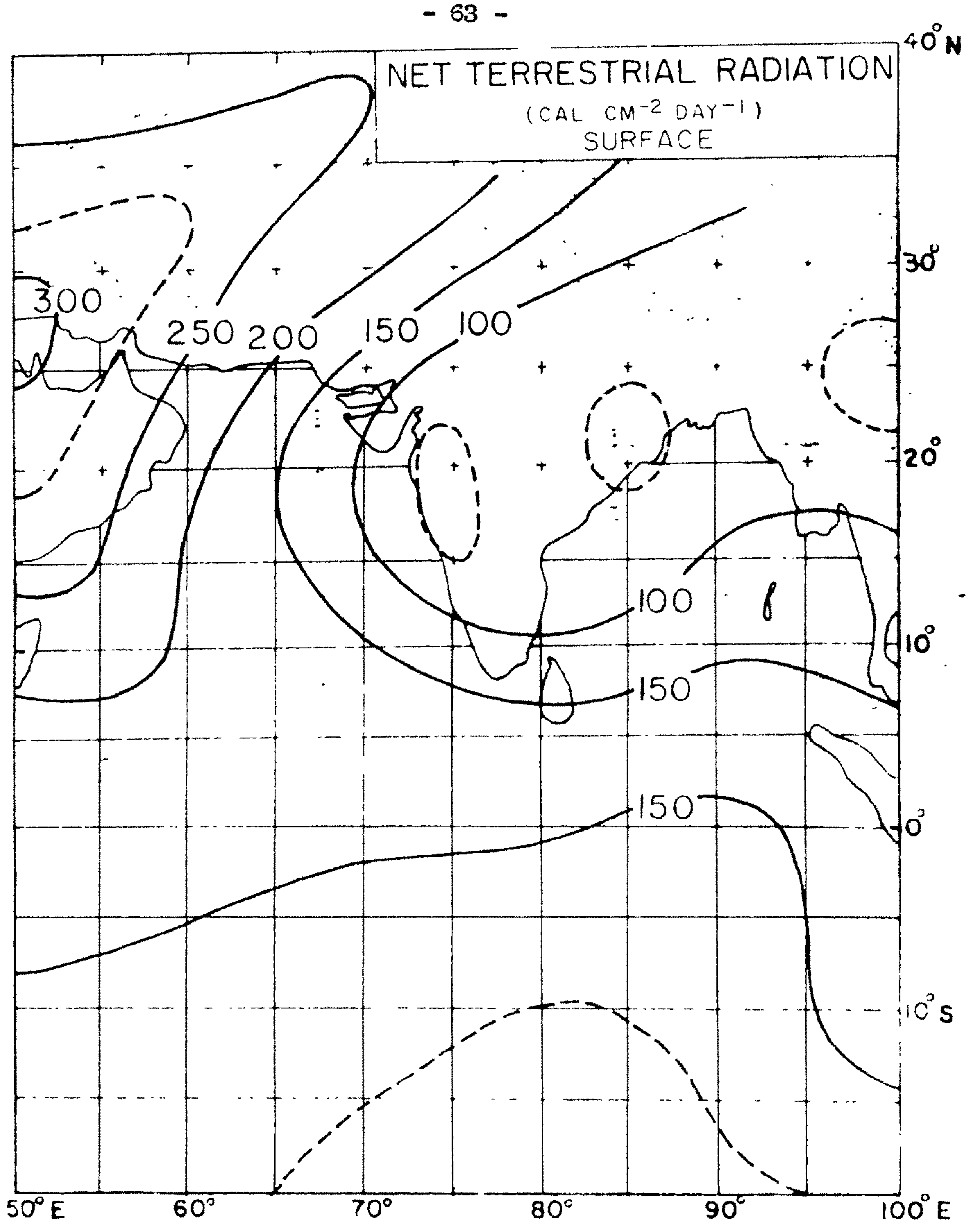


Figure 3.2.2 Distribution of net terrestrial radiation (cal cm$^{-2}$ day$^{-1}$) at the surface for July.

Long wave radiation at any level is a function of air temperature at that level and of the amounts of absorbing gases above and below that level. Water vapour and ozone together with carbon dioxide play important roles in the long wave radiative heat exchange processes respectively in the troposphere and the stratosphere (Godbole et al, 1971). The most dominant factor influencing the distribution of long wave radiation at any level is the cloud cover (Kelkar et al, 1970) because cloud is considered to absorb and emit long wave radiation like a black body.

The spatial distribution of the net long wave radiation at the surface has been calculated following the earlier procedure ( Godbole et al, 1969 a ). The distribution of cloud cover for the month of July as seen by the meteorological satellite (Bunker et al,1969; Sadler, 1969 ) and supplemented by surface observations particularly over India as shown in Figure 3.2.1, has been made use of in the calculation. The results obtained are shown in Figure 3.2.2. The radiative heat loss at the surface is minimum over the Western Ghats and northeast India on account of the maximum clouding over those regions ( Figure 3.2.1). The radiative heat loss is maximum over the desert regions in the northwest which is due to the combined effect of high temperature and low amounts of water vapour and clouding.

## 3.3 Sensible and Latent Heat

The boundary layer next to the earth's surface (sea and land) is of crucial importance because it is here that the basic processes of energy exchange take place which ultimately influence the general circulation of the atmosphere (Rolls,1965). As the information about the meteorological elements required for computing the transfer rates of sensible and latent heat are available for the oceanic region only, the analysis presented in the following is limited to the oceanic region. Such analysis has been reported earlier (Saha, 1970) but it takes into account only single year's data, namely of 1964. In order to filter out the effect of year-to-year variation of the monsoon activity, climatological data which are available for a period of more than 15 years (Deutshes Hydrographisches Institute, 1960, and London Met. Office, 1971) have been made use of in the present study. The procedure adopted in the analysis is as follows.

The eddy fluxes of sensible and latent heat are computed by the bulk aerodynamic method (Jacobs, 1951; Sverdrup, 1951). The flux of sensible heat ( $F_s$ ) is expressed as

$$F_s = k_1 V_a (T_s - T_a) \quad \text{cal cm}^{-2}\text{day}^{-1} \qquad (3.3.1)$$

where $k_1$ is called the exchange coefficient, $V_a$ is

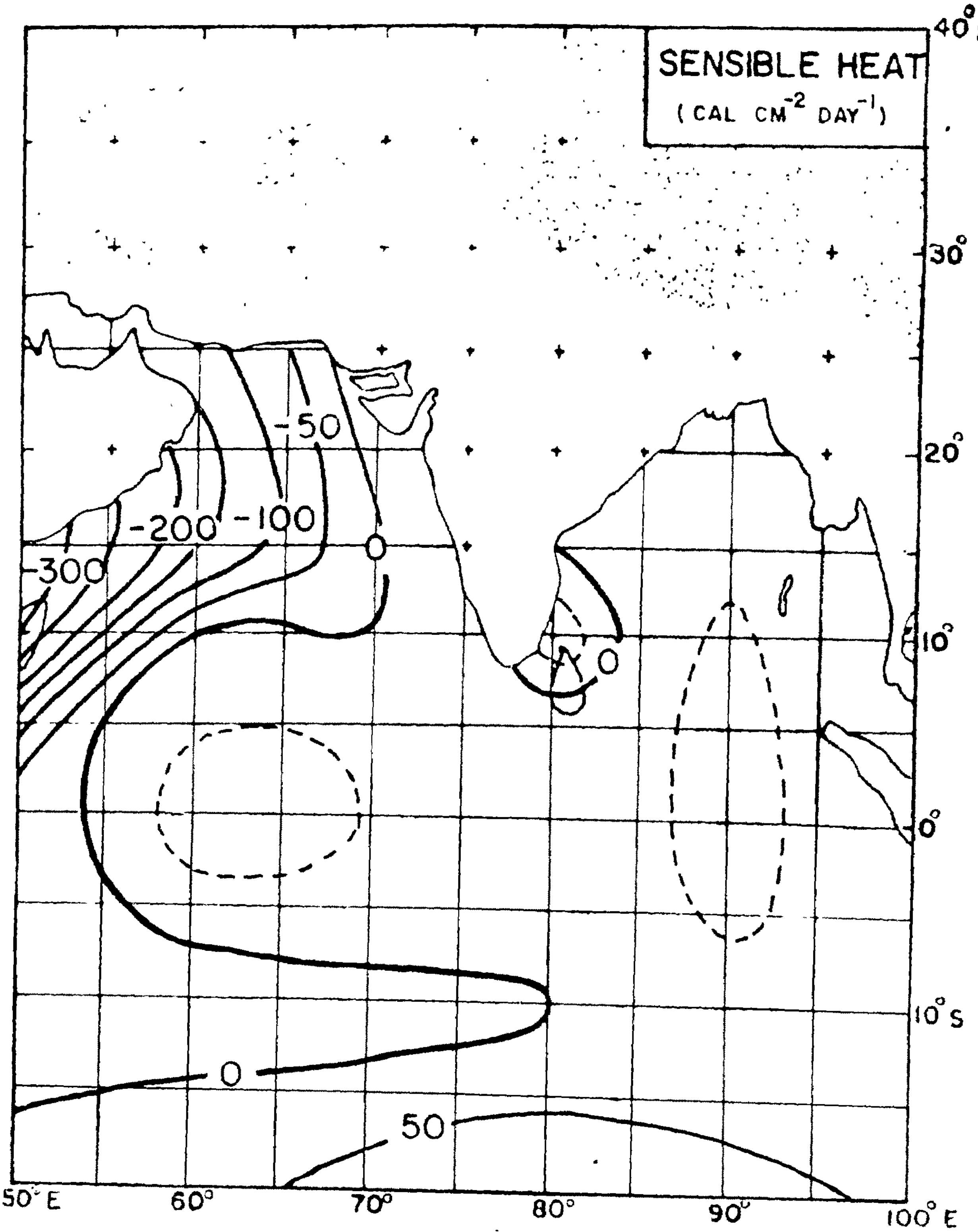


Figure 3.3.1 Distribution of sensible heat (cal cm$^{-2}$ day$^{-1}$) at the surface for July.

the wind speed in knot and $T_s$ and $T_a$ are respectively the sea surface temperature and air temperature in °C. The flux of latent heat ( $F_L$ ) is similarly expressed as

$$F_L = k_2 V_a (e_s - e_a) \quad cal\, cm^{-2} day^{-1} \qquad (3.3.2)$$

where $k_2$ is the exchange coefficient, $e_s$ is the saturation vapour pressure corresponding to $T_s$ and $e_a$ is the vapour pressure of air. In the above, the subscript 'a' refers to the observations at ship's deck level.

A considerable amount of variation is found in the literature as regards the values used for the exchange coefficients $k_1$ and $k_2$ (Budyko, 1956; Riehl et al, 1958; Sheppard, 1958; Roden, 1959, Priestley, 1959; Swinbank, 1959). In the following, the values used by Petterssen et al, (1962) which are based on the work of Sheppared (1958) and Swinbank (1959) have been adopted.

3.3.1 Sensible heat flux

The spatial distribution of the flux of sensible heat obtained is shown in figure 3.3.1. Over the west Arabian Sea, the sensible heat flux is negative which means that, in that region the atmosphere loses heat to the ocean. This feature which is due to upwelling of cold waters from below is intense near the coast of Somalia and Arabia ( $\approx$ - 300 cal $cm^{-2}$ day$^{-1}$). The effect of

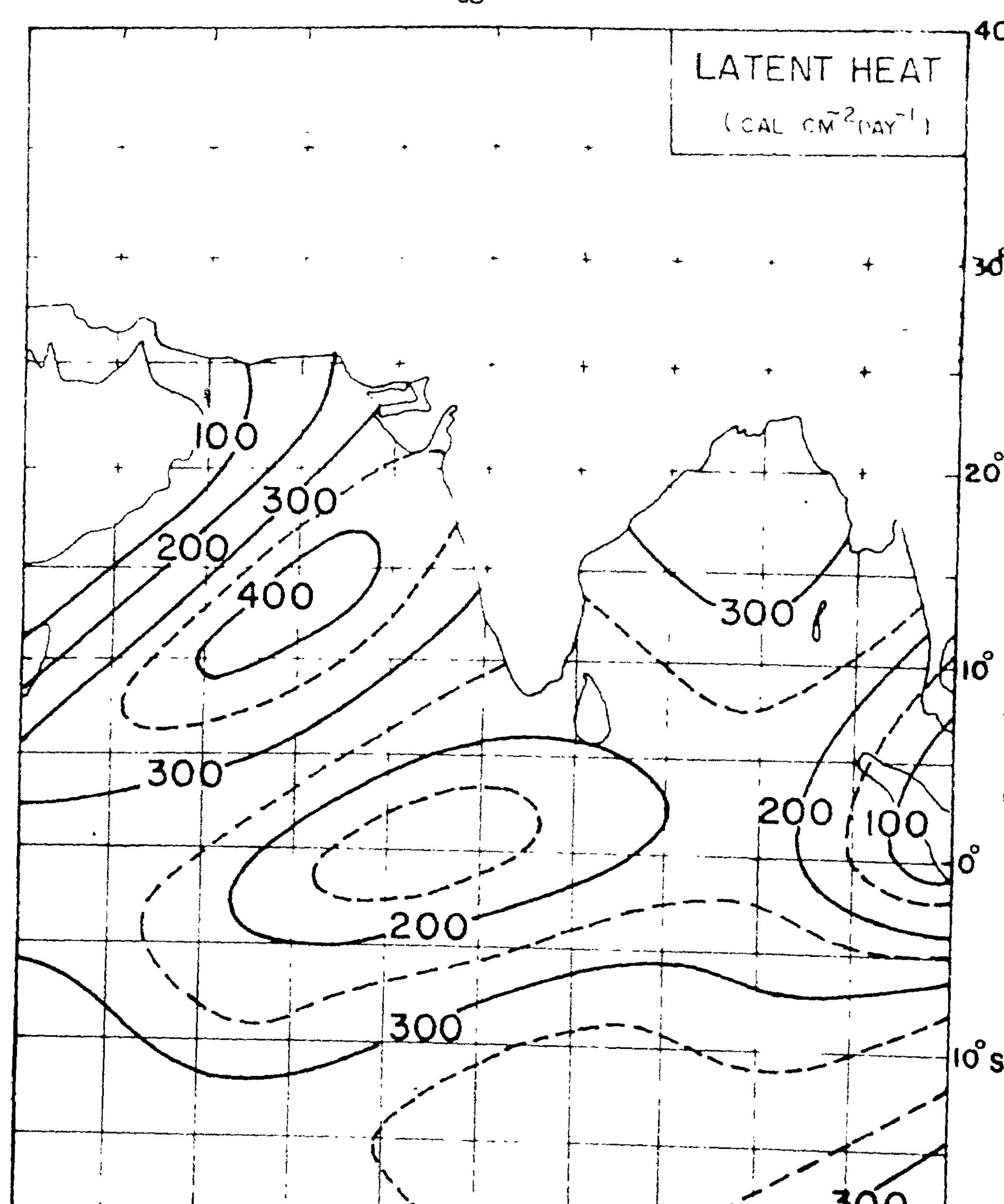


Figure 3.3.2 Distribution of latent heat ( cal cm$^{-2}$ day$^{-1}$) at the sea surface for July.

slight upwelling is discernible near the east coast of peninsular India too, where there is a region of downward sensible heat flux. Over the remaining part of the region, the sensible heat flux is relatively weak and directed from the ocean to the atmosphere.

### 3.3.2 Latent heat flux

The distribution of latent heat flux, through evaporation from the ocean, is shown in Figure 3.3.2. Unlike sensible heat flux, the latent heat flux is everywhere positive meaning thereby that the atmosphere gains heat from the ocean. The maximum transfer of latent heat noticed is 400 cal $cm^{-2}$ $day^{-1}$ and takes place over the central Arabian Sea. This region has been examined and characterised for the flux divergence of water vapour (Pisharoty, 1965). The transfer of latent heat is also large over the north Bay of Bengal and over the south Indian ocean between 10° and 20° S. Over the equatorial belt, as well as near the coast of Somalia and Arabia the transfer is small.

The values of the sensible heat flux obtained in the present study are about four times higher than the values reported by Saha (1970). As this discrepancy is mainly due to the widely different values adopted for the exchange coefficient $k_1$ in the studies ( whereas the numerical value used by Saha is 0.83 for $T_s < T_a$ that used by the present author is 4.1 ), and as there is no experimental evidence in support of the values used for $k_1$ in either of the studies, the situation calls for the urgent need to take up the necessary measurement program of work with reference to the Indian Ocean region.

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