THERMODYNAMIC STRUCTURE OF THE ATMOSPHERIC BOUNDARY LAYER OVER THE ARABIAN SEA AS REVEALED BY MONSOON-77 DATA

SURENDRA S. PARASNIS and SAVITA B. MORWAL

Indian Institute of Tropical Meteorology, Dr Homi Bhabha Marg, Pune 411 008, India

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Abstract. The thermodynamic structure of the Atmospheric Boundary Layer (ABL) over the Arabian sea region has been studied with the help of 135 aerological observations obtained during MONSOON-77 in the region (10–14° N, 64–68° E) by USSR research vessels. Low-level inversions were observed over the western Arabian sea region (west of 66° E) in association with suppressed convection. The different sublayers of the ABL, viz. the mixed layer, the cloud layer and the inversion/isothermal/stable layer were identified. The low-level stability analysis indicated that in the region east of 66° E, conditions were favourable for deep convection. The thermodynamic transformation of the boundary layer after precipitation was documented.

1. Introduction

Knowledge of the thermodynamic structure of the Atmospheric Boundary Layer (ABL) over the Arabian sea especially during the summer monsoon would be valuable for understanding of the monsoonal flow. The structure of the marine boundary layer over the equatorial Pacific regions and Atlantic oceanic region have been studied extensively (Augstein *et al.*, 1973, 1974; Brummer, 1976, 1978; Firestone and Albrecht, 1986; Betts and Albrecht, 1987; Kloesel and Albrecht, 1989). For the first time, radiosondes from ships and dropsondes from aircraft taken during the International Indian Ocean Expedition (IIOE, 1963–65) provided some insight into the temperature and moisture stratification present in the lower troposphere in the Arabian sea particularly in the region north of 10° N. However, the IIOE data were not adequate to delineate the different sublayers of the ABL in this region and bring out their characteristic features. A picture of the marine boundary layer over the Arabian sea has begun to emerge in recent years when the observations from ISMEX-73, MONSOON-77, MONEX-79 were made available (Pant, 1977, 1978, 1982; Ghosh *et al.*, 1978; Mohanty *et al.*, 1983). The mean structure of the marine boundary layer has been studied using limited observations from aircraft and ships during MONEX-79 (Holt and Sethuraman, 1985, 1987).

The lower tropospheric circulation in this region of the globe differs significantly from that observed in other tropical oceanic regions during the northern summer. With the onset of the summer monsoon over India and neighbourhood, towards the end of May, a cool and moist air mass from the south Indian ocean establishes itself in the lower troposphere above the warm waters of the Arabian sea. A relatively warm and dry airmass of continental origin prevails above the moist

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monsoon current. While the pressure gradients and winds are weak in other tropical regions at this latitude during this period, strongest winds occur over this region in the lower as well as in the upper troposphere. Occurrence of a low-level jet (wind speed exceeding 25 m s^{-1}) off Somalia and its further extension northeastward across the Arabian sea in the layer 1.0–2.0 km with associated strong vertical shear is another feature which has no parallel elsewhere (Pant, 1978). Earlier studies (Holt and Sethuraman, 1987) have also revealed that the structure of the monsoon boundary layer is different from a trade-wind boundary layer in several aspects, one of which is its generation. A number of not yet fully understood processes of different scales interact in the monsoon boundary layer.

In this paper, a detailed analysis is presented of the radiosondes from the USSR research vessels in the region (10–14° N, 64–68° E) during the two periods 14–20 June and 13–16 July during MONSOON-77. The purpose of the paper is to describe the characteristic features of the marine boundary layer over the Arabian sea as far as the thermodynamical and dynamical aspects are concerned. Although some of the results conform with those reported in earlier studies using ISMEX-73 and partly using MONSOON-77 (Ghosh *et al.*, 1978; Pant, 1978) data sets, aspects such as low-level stability, and the thermodynamic transformation of the boundary layer, have not been documented earlier.

2. Location of Observations and Meteorological Conditions

Aerological observations were carried out during MONSOON-77 by the USSR research vessels that were stationary in the region (10–14° N, 64–68° E) of the Arabian sea (Figure 1). The data utilized in this paper pertain to two periods viz. 14–20 June and 13–16 July. As far as synoptic conditions during 14–20 June 1977 are concerned, the trough off the west coast was present. For the period in July 1977, for which the surface pressure distributions are shown in Figure 2, the monsoon trough was aligned with the Gangetic plains (on 15th and 16th July) and protruded over the Head of the Bay of Bengal. The position of the trough line has an influence on monsoon activity. It is seen from Figure 2 that synoptic conditions were favourable for an active monsoon during this period. An active monsoon is usually associated with widespread rainfall over the Indian region (see e.g., Parasnis and Goyal, 1990). The trough off the west coast of India persisted during the period 13–16 July, 1977 with a well marked embedded vortex near the coast (15° N).

3. Method of Analysis

The data consisted of temperature (T, °C), dew point temperature (T_d, °C), humidity mixing ratio (q, gm kg^{-1}) and wind (speed and direction) at intervals at about 25 hPa. The interval is not fixed because significant changes occur at different

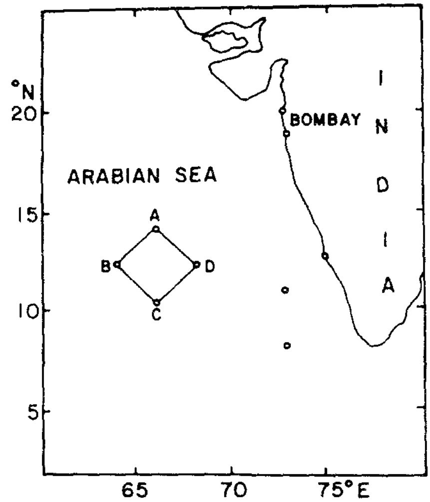

Fig. 1. Locations of ships during periods of observations. A: Okean, B: Priboy, C: Priliv, and D: Shirshov.

levels. These data were used to compute thermodynamic parameters such as virtual potential temperature (θ_v, °K), equivalent potential temperature (θ_e, °K), saturated equivalent potential temeprature (θ_{es}, °K), zonal (U, m s^{-1}) and meridional (V, m s^{-1}) components of the wind.

There were 75 observations during 14–20 June 1977 from three ships and 60 observations during 13–16 July 1977 from four ships (Figure 1). Earlier studies of the Convective Boundary Layer (CBL) during the summer monsoon (Parasnis and Morwal, 1991; Parasnis, 1991), suggested a height of the CBL between 700–650 hPa (during weak monsoon activity) and 500 hPa and above (during active monsoon conditions). The CBL is thermodynamically coupled to the surface heat and moisture fluxes. This includes the subcloud, cloud and inversion/stable layers. The extent of the CBL, to higher levels, characterises the extent of the monsoon current. Therefore observations up to 500 hPa were used to study the thermodynamic structure of the ABL in this paper.

4. Characteristics of the ABL over the Arabian Sea

4.1. Details of inversion/isothermal layers

During the IIOE (1963–65), boundary-layer measurements were made by Bunker (1965). He found that the sensible heat transport was downward above 1.5 km over the west Arabian sea and upward over the east Arabian sea. The IIOE data also revealed for the first time the presence of a sharp temperature inversion in the layer 900–750 hPa over the western and central regions of the Arabian sea

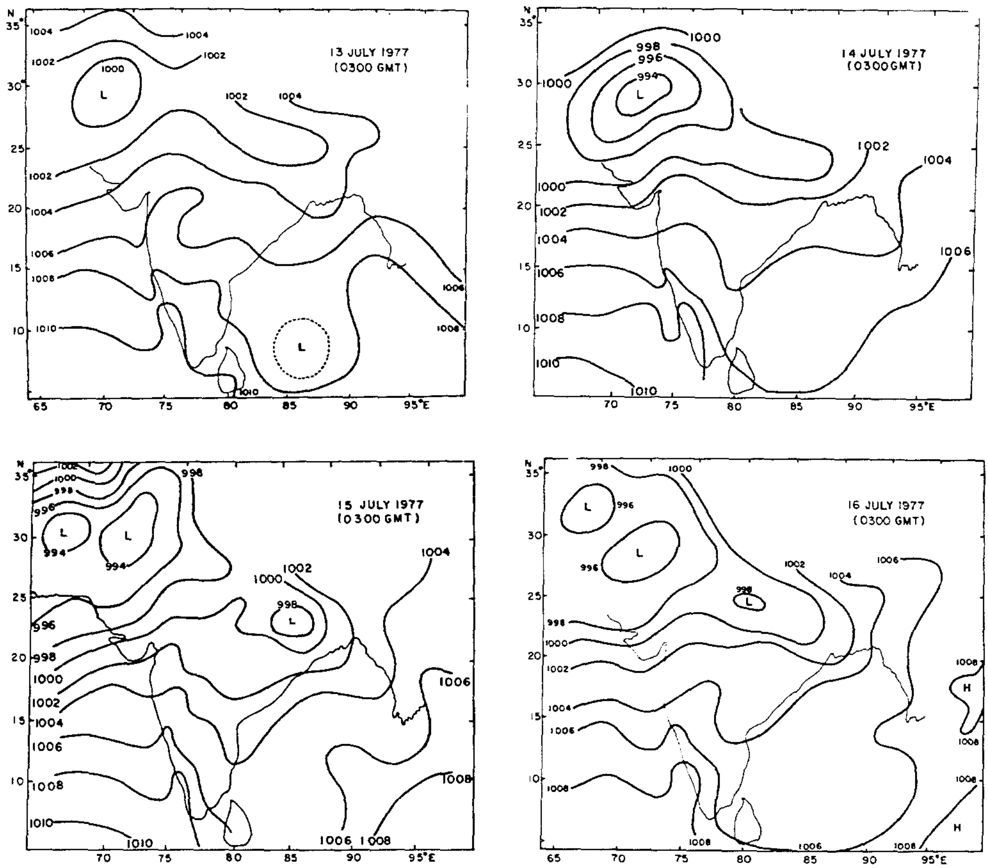

Fig. 2. Surface pressure distributions (hPa) 13–16 July 1977.

with a dry air mass prevailing above (Colon, 1964; Ramage, 1966; Desai, 1968, 1970). There have been different opinions as to the causes for the inversions, whether they are 'subsidence type' or 'air-mass type' (Sikka *et al.*, 1981). The analysis of aerological observations over the Arabian sea during ISMEX-73 showed that the vertical structure of the Planetary Boundary Layer was different in the region north of about 10° N from that observed to the south of it (Pant, 1978). Also, the temperature profile indicated a strong upward slope of the inversion layer from the western Arabian sea (50° E) to the eastern Arabian sea (68° E). A very different picture is observed in the extreme eastern region of the Arabian sea, where strong convective mixing weakens the layered structure and the inversion is generally absent. In this region, the top of the boundary layer cannot easily be defined based on temperature and humidity profiles. During 14–20 June, the inversion layers were observed in 20% of the observations. During June, the belt 10–15° N is usually free of inversions. With advancement of the monsoon, convection on a large scale begins in this belt. During the 13–16 July period, in

80% of the cases, inversion layers were observed over Priboy (12° N, 64° E) and Okean (14° N, 66° E). At Priliv (10° N, 66° E), isothermal layers were observed in 40% of the observations. It is worth noting that at Shirshov (12° N, 68° E) there was only one case of inversion. Thus, the conditions at Shirshov were different from those observed over the other three ships. During June the average inversion strength was 5 °C per 100 hPa which was based on 15 inversions cases. Considerable variation in inversion strength was observed (range 2–13°). During July, the average inversion strength was 3 °C per 100 hPa at Okean (11 cases of inversions) and 4 °C per 100 hPa at Priboy (13 cases). Inversion strength during July varied between 1–8°.

Inversion layers were observed at 875 and 825 hPa levels at Priboy and Okean respectively. This latitudinal variation in the inversion base is in agreement with the results obtained by Ghosh *et al.* (1978) for ISMEX-73 data.

4.2. Height of the ABL

The top of the ABL has been taken as the base of the inversion/isothermal layer. In the absence of an inversion/isothermal layer, the base of the stable layer has been taken as the height of the ABL. θ_{es} is a very useful parameter to identify the base of the stable layer. In earlier studies of the CBL over the Pacific Oceanic region, it was found that minimum and maximum values of θ_{es} coincided with the base and top of the inversion respectively (Betts and Albrecht, 1987). The ABL heights are shown in Figures 3a, b (solid lines) during both periods. From the figures it is seen that the ABL height varied between 900–580 hPa. The variation in ABL height is more during June. During July, except at Shirshov, the ABL height did not show much variation. The mean ABL height was between 800–750 hPa during June. During July in the eastern region (i.e., at Shirshov), it was higher and because of the presence of low-level inversions, it was lower (between 875–825 hPa) at Priboy, Okean and Priliv as seen in the previous section.

4.3. Mixed layer

The mixed-layer heights have been obtained by examining the gradients of θ_v. In the moist atmosphere, θ_v is the more suitable measure than θ for the mixed layer. The constancy of θ_v has been used to find out the extent of mixed layers. The heights of the mixed layer have been determined in such a way that the gradient of θ_v should not exceed 2 °K km^{-1} in the mixed layer. The heights of the mixed layer thus obtained are shown in Figure 3a, b (dot dash lines). From Figures 3a, b, it is seen that the mixed-layer heights showed diurnal characteristics, although these were not pronounced at Shirshov. The heights of the lifting condensation level (LCL) are also shown in Figures 3a, b by broken lines. In most cases, it is seen that the mixed-layer heights are higher than the LCL heights, implying saturation in the upper part of the mixed layer.

Thermal stability has been found by examining the difference in θ_v in the surface

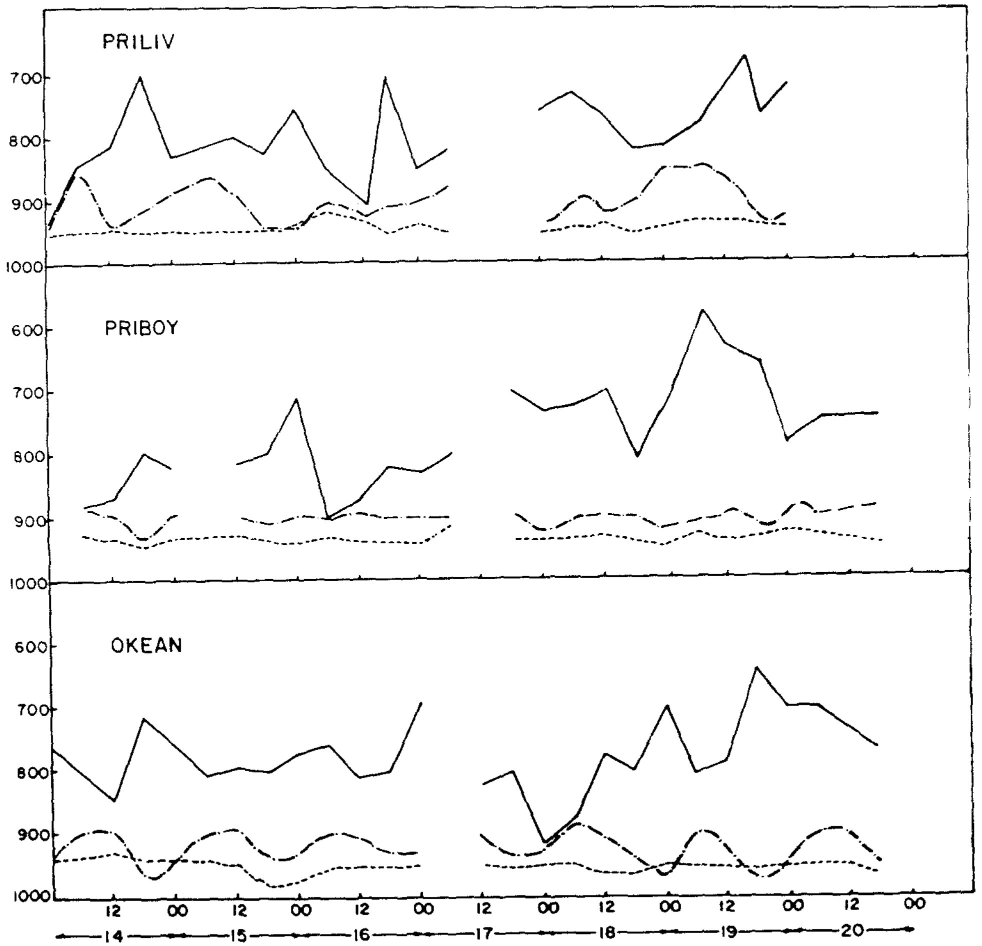

Fig. 3a. Variations in height of boundary layer (solid), top of the mixed layer (dash-dotted) and lifting condensation level (dashed) during 14–20 June 1977.

and 1000 hPa layer (i.e., in the 60–70 m layer). The neutral, unstable and stable cases have been categorized as

$$\begin{array}{lccccc} \text{Unstable} & & & & & > \\ \text{Stable} & \text{if} & \begin{array}{c}\theta_v \\ \text{(Surface)}\end{array} & - & \begin{array}{c}\theta_v \\ \text{(1000 hPa)}\end{array} & < 0. \\ \text{Neutral} & & & & & = \end{array}$$

This stability pertains to the subcloud part of the ABL. In the majority of the cases in June and July, the subcloud layer was observed to be unstable.

4.4. Cloud layer

The LCL heights (Figure 3a, b) have been considered as the cloud bases. During both periods, the cloud base did not show much variation. The average cloud base

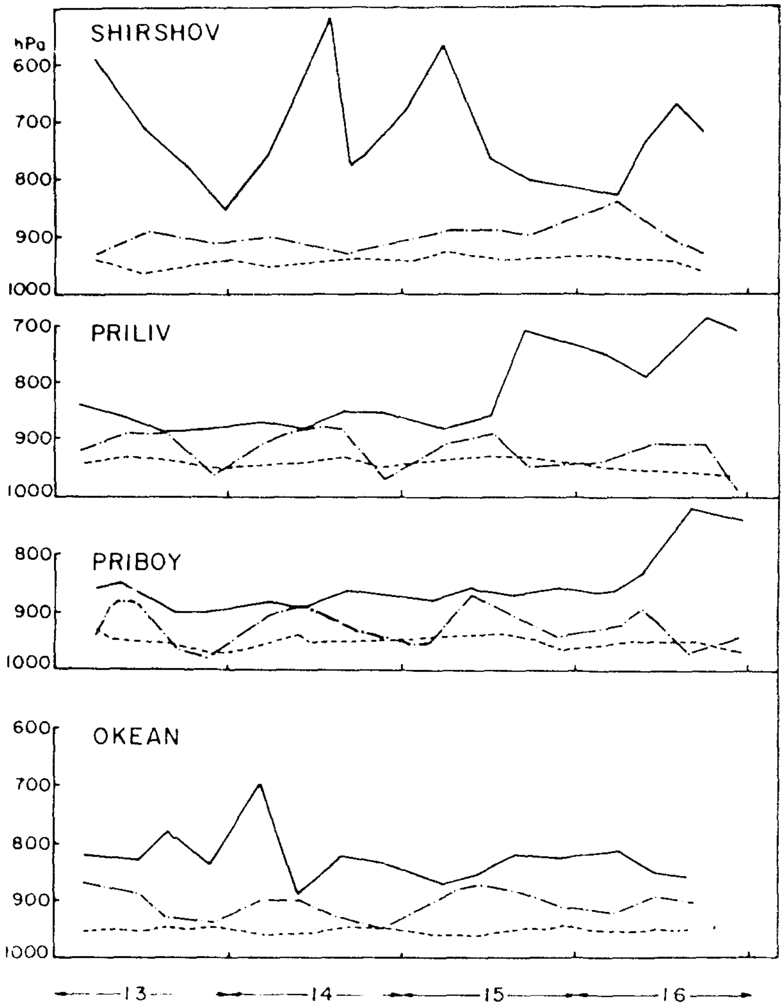

Fig. 3b. Same as Figure 3a, except for the 13–16 July 1977 period.

was ~450 m. The cloud layer has been taken as from the LCL to the base of the inversion/isothermal layer. Marked variation in the depth of the cloud layer has been noticed. During June, in the absence of an inversion, the cloud layer depth varied between 1.0–2.0 km. But during July at the three ships (Priboy, Okean and Priliv), there were clouds of shallow depth (0.7–1.0 km) while at Shirshov the depth of cloud layer was ~2.5 km. It can be seen that the western (west of 66° E) and northern (north of 10° N) regions were free from convective clouds whereas the eastern region (east of 66° E) had deep convective clouds.

4.5. Assessment of low-level stability

To assess the low-level stability, the mean θ_e and θ_{es} soundings in the two periods have been examined. Firestone and Albrecht (1986) used the θ_e and θ_{es} profiles

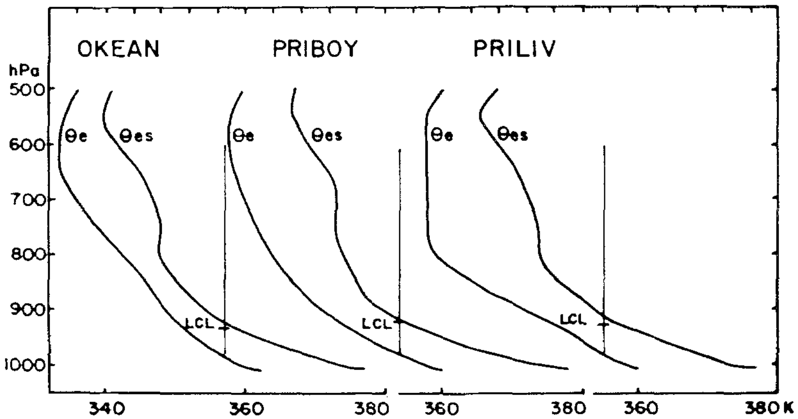

Fig. 4a. Mean θ_e and θ_{es} profiles for June 1977. The straight line from 980 hPa upwards represents the ascent of an air parcel.

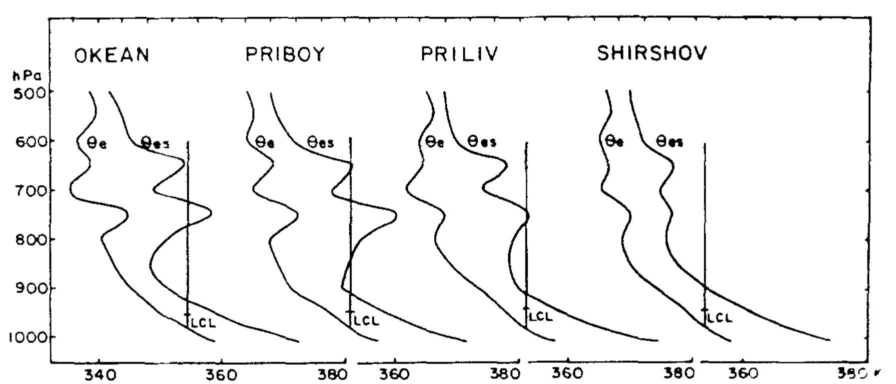

Fig. 4b. Same as Figure 4a, except for the July 1977 period.

for characterising the differences in the thermodynamic structure of the boundary layer associated with deep convection, shallow convection, and conditions where both deep and shallow convection are suppressed. Also, Kloesel and Albrecht (1989) used the same methodology for the classification of observations obtained from FGGE. The low-level stability, in a case study, has been assessed similarly for the observations carried out in the vicinity of a monsoon trough region (Parasnis *et al.*, 1991).

The mean θ_e, θ_{es} profiles for the two periods are shown in Figures 4a, b. A straight line represents the θ_e path for a non-entraining parcel originating from 980 hPa. Above the lifting condensation level, the difference between this path and the θ_{es} profile will be proportional to the temperature difference between a parcel originating from 980 hPa and the temperature of the environment. Mean θ_e and θ_{es} profiles during 14–20 June shown in Figure 4a indicate the conditions associated with deep convection. The θ_e path cuts the θ_{es} profile more or less at the LCL; upwards, the value of θ_e is higher than the θ_{es} profile, which represents the ambient temperature. In an earlier study (Sikka *et al.*, 1981) it was observed that during this period, the satellite picture showed a cloud cluster covering the three locations. No stable layers were observed at low level during this period. Important features of these profiles during the July period are the stable layers shown by the mean profiles of θ_{es} (Figure 4b). At Priboy, the stable layer at 900 hPa clearly inhibits the convection there. For Priboy the buoyancy is seen to be sufficient to raise the parcel between 910–840 hPa. Similarly at Okean and Priliv, the parcel originating from 980 hPa would be positively buoyant between 925–775 hPa and 905–770 hPa, respectively. The ascent of an non-entraining parcel at the three ships would clearly be inhibited by inversion/stable layers between 900–750 hPa. Another stable layer of 50 hPa depth is seen between 700–650 hPa. At Shirshov, the θ_e path of a non-entraining parcel indicates that there seems to be sufficient buoyancy to support convection to at least 500 hPa. Thus at Shirshov, conditions were favourable for deep convection. At the other three ships, inversion/isothermal layers inhibited the convection.

The stable layers between 900–750 hPa are clearly of sufficient strength to inhibit deep convection. It can be seen from Figure 4b that they are maintained by subsidence, induced either by nearby deep convection or by large-scale subsidence. These low-level stable layers in a way play a significant role in maintaining the organization of convective systems by suppressing deep convection in the surrounding regions.

4.6. Assessment of Vertical Velocity

For the purpose of computation of vertical velocity, the observational area (Figure 1) has been divided into four triangles, viz., western (west of 66° E), eastern (east of 66° E), northern (north of 12° N) and southern (south of 12° N). The vertical velocity values were obtained at the centroid of these triangles (Ryan *et al.*, 1989). Plots of vertical velocities are shown in Figure 5. The values of divergence have been estimated from observations of U and V. In calculating the vertical velocities, the divergences were adjusted using the method of O'Brien (1970).

From Figure 5 it is seen that in the eastern region there was evidence of strong ascent up to 640 hPa and descent in the 640–500 hPa layer. In contrast, in the western region there was strong descending motion in the region 900–640 hPa and ascending motion in the 640–500 hPa layer. This clearly reflects the differences in the prevailing weather conditions. In the northern region, descending motion from 900 hPa up to 650 hPa was noticed. The southern region shows ascending motion up to 800 hPa and in the layer 700–650 hPa. This indicates the existence of multi-layer cloud structure in the southern region. The ascending and descending mo-

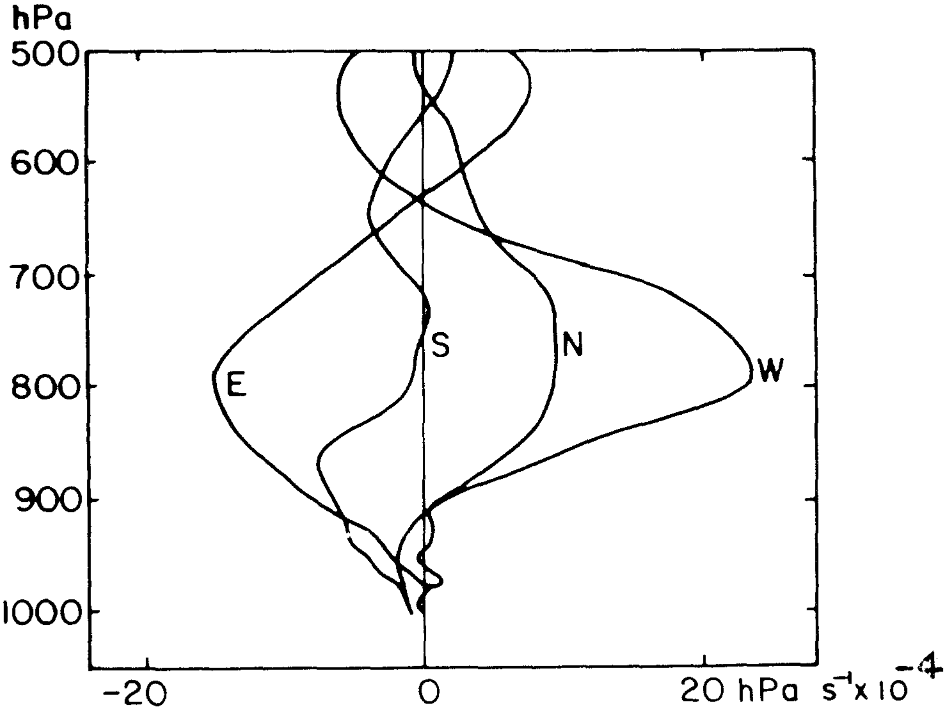

Fig. 5. Mean vertical velocities in the regions eastern (E), western (W), northern (N) and southern (S) during 13–16 July 1977.

tions shown in Figure 5 for different regions are consistent with the prevailing weather conditions in those regions.

4.7. Wind profiles

Average profiles of the zonal and meridional wind components are shown in Figure 6. The existence of a low-level jet is clearly seen in the zonal wind. The wind maxima are located between 900–800 hPa in the average profiles. These profiles show that during June and July (i.e., during the two periods considered here) the region was dominated by westerly flow. The westerly wind increases from the surface upwards, reaching a maximum in the 900–800 hPa layer and then weakening with height (Figure 6).

From the profiles of the meridional component of wind, it is seen that it is southerly in the lower layers and tends to weaken with height to a minimum at the height ranging from 900–650 hPa. It appears that the flow was more westerly in the month of June and gradually became southwesterly in July. In the western region (i.e., at Okean, Priboy and Priliv) of the Arabian sea, the zonal wind maximum was found in the cloud layer. Also, it is seen that this maximum nearly coincided with the top of the mixed layer in the eastern region (at Shirshov).

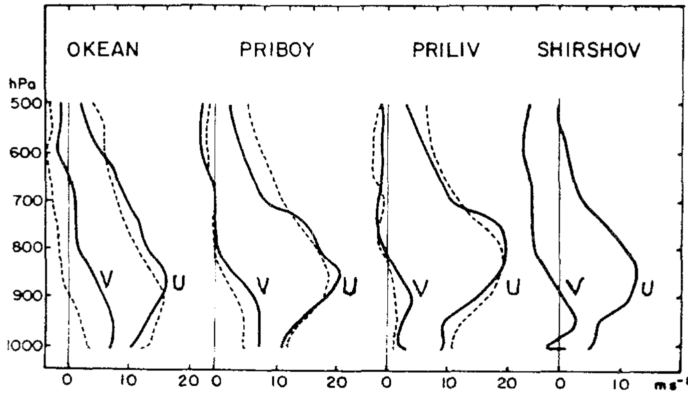

Fig. 6. Mean U and V profiles for June (dashed) and July (solid) 1977.

4.8. Thermodynamic transformation of the boundary layer due to precipitation

It is interesting to see the transformation of the boundary-layer structure and the properties of thermodynamic parameters as a result of precipitation. From the hourly surface observations, two cases were considered.

At Priboy (12° N, 64° E) on the 15th of July 1977 just before the time of the radiosonde flight (22 UTC), there was a light rain shower. The profiles of θ, q, θ_e obtained from an earlier flight (at 16 UTC) have been compared with the profiles obtained from the 22 UTC flight. Also, at Priliv (10.3° N, 66° E) on the 14th of July 1977, there was rain before the radiosonde flight at 22 UTC. The profiles of θ, q, θ_e from the 16 UTC flight have been compared with the profiles obtained from the 22 UTC flight. Similarly the U and V components for the four flights were compared.

The θ, q, θ_e profiles for Priboy and Priliv for the two flights are shown in Figures 7 and 8, respectively. The U and V profiles for the two ships are shown in Figures 9 and 10, respectively. The surface temperature, dew point temperature, q and other observations such as total cloud amount, low cloud amount, LCL height at the two ships are given in Table I. Conditions before the precipitation will be referred as "undisturbed" and after precipitation as "disturbed" in the following discussion.

It is seen from Table I that there was an increase in q by nearly 1 gm, an increase in the amount of low clouds, and a lowering of cloud base heights (LCL) by nearly 150 m during the transition of weather conditions from undisturbed to disturbed. The well-mixed subcloud layer structure during undisturbed conditions is transformed into a more stable structure after precipitation (Deardorff, 1972). In earlier studies of the transformation of the tropical subcloud layer by precipitation and downdraft (Betts, 1976), it was shown that with the onset of precipitation, there are two processes which increase the coupling between the subcloud

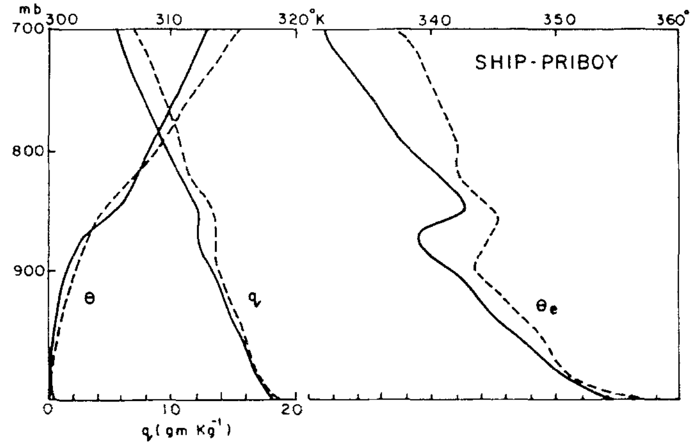

Fig. 7. θ, q, θ_e profiles at ship Priboy before (solid) and after (dashed) precipitation.

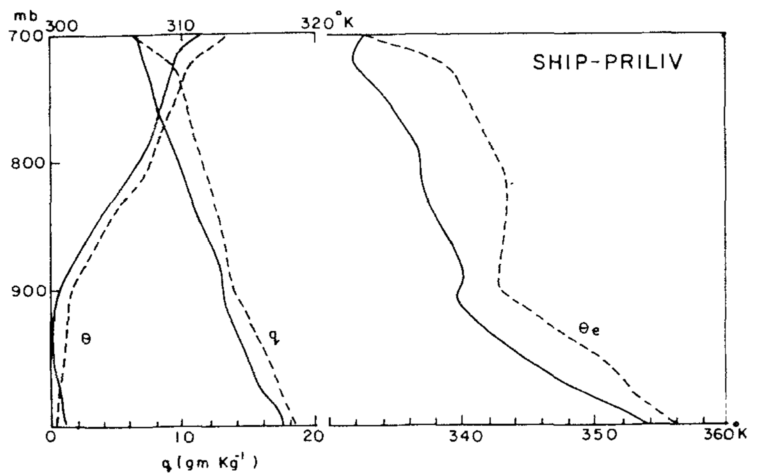

Fig. 8. Same as Figure 7 for ship Priliv.

and the cloud layers and transform the structure of the mixed layer. The effect of evaporation of falling droplets into the unsaturated subcloud layer is to cool and moisten the layer, thereby bringing the layer closer to saturation. In addition, evaporative cooling can drive convective-scale downdrafts with mass transport comparable to the convective-scale downdrafts. These downdrafts bring down potentially warmer and drier air into the subcloud layer. These two processes oppose each other in the sense that evaporation cools and moistens while downdrafts warm and dry. Therefore, any combination of these processes can occur. Comparison of θ profiles during undisturbed and disturbed conditions indicates that the stability of the boundary layer increased during disturbed conditions. The well-mixed subcloud layer structure before precipitation (Figure 8) has been transformed into a stable structure after the precipitation. The inversion layer

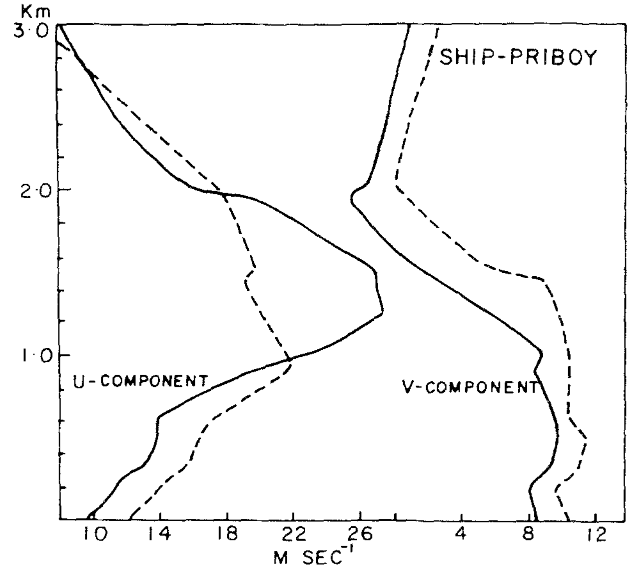

Fig. 9. U and V profiles at ship Priboy before (solid) and after (dashed) precipitation.

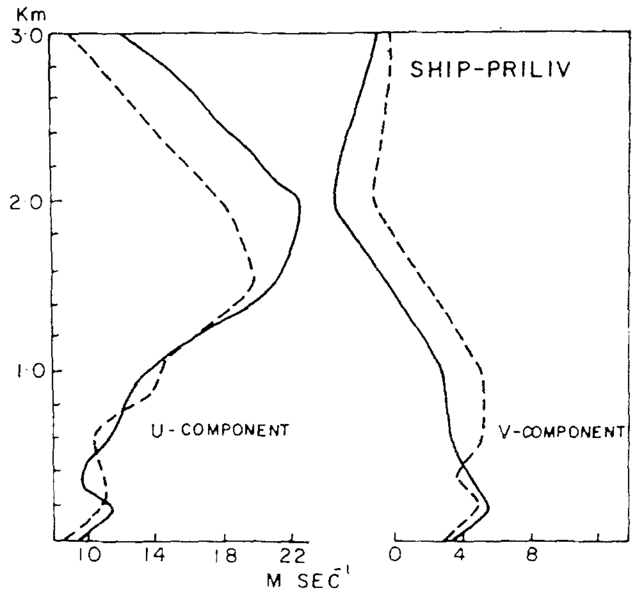

Fig. 10. Same as Figure 9 for ship Priliv.

between 900–800 hPa at Priboy (16 UTC) was transformed to an isothermal layer after precipitation. The striking feature of the q profile is that it shows an increase at all levels up to 700 hPa. θ_e values also show an increase at all levels. The slope of the θ_e profile shows a decrease during disturbed conditions. In the cloud layer, a neutral or stable θ_e profile is observed (Figures 7 and 8). In the earlier study (Betts, 1976), it was observed that the subcloud layer became cooler and drier

TABLE I
Changes in surface temperatures and mixing ratios, total cloud amount, low cloud amount and LCL heights before and after precipitation at the ships Priboy and Priliv.

Time (UTC)	Surface temperature dry-bulb (°C)	Surface temperature dew-point (°C)	Surface mixing ratio (gm kg^{-1})	Total cloud (okta)	Low cloud (okta)	LCL height (m)
Ship – Priboy (12° N, 64°3′ E)						
15.8	27.6	23.6	18.6	6	2	500
21.8	27.6	24.8	20.0	7	3	350
Ship – Priliv (10°3′ N, 66° E)						
15.9	28.4	23.3	18.2	7	3	640
21.9	28.1	24.1	19.0	7	6	500

after precipitation and downdrafts. From the above analysis it is seen that the subcloud layer has become moist and warm after the precipitation in both cases. The evaporative cooling observed in the earlier study (Betts, 1976) has not been observed here perhaps because the effect of evaporative cooling may be of subtle nature; the atmosphere remains nearly saturated during the summer monsoon season.

The wind components profiles at the two ships before and after precipitation are shown in Figures 9 and 10.

5. Summary and Conclusions

The aerological observations obtained from USSR research vessels during MONSOON-77 were used to explore the thermodynamic structure of the atmospheric boundary layer over the Arabian sea.

Inversion layers were observed in the layer 900–800 hPa during July in the western region (west of 66°) in 80% of the observations. Isothermal layers were observed in 40% of the cases. During June only 20% of the observations were associated with inversion layers. The soundings with low-level inversions were generally obtained in the regions of suppressed convection. The eastern Arabian sea (east of 66°) was associated with deep convection. Despite the large-scale upward motion in the eastern Arabian sea (east of 66° E), compensating subsidence in regions surrounding deep convection are probably responsible for the low-level stable layers. These stable layers, in turn, inhibit convective activity. Over this region of the Arabian sea, the temperature and moisture in the surface layer did not show any significant differences in the eastern and western regions; hence low-level inversions are important for regulating deep convection.

The height of the atmospheric boundary layer over the Arabian sea as determined by the height of the inversion base and the minimum of θ_{es}, varied from 900–680 hPa. Low-level stability assessed by θ_e and θ_{es} profiles showed that during

June, conditions were favourable for deep convection. During July the presence of a low-level stable layer inhibited the convective activity in the region west of 66° E. The stable layers might have resulted from convectively-induced subsidence.

The thermodynamic transformation of the marine boundary layer associated with mesoscale disturbances like convective rain has been examined diagnostically. The characteristic well mixed, dry adiabatic subcloud layer preceding the rain system has been transformed by updrafts and precipitation. This new structure of the marine boundary layer consisted of a warmer, more stable subcloud layer, and a more well-mixed cloud layer. Zonal wind maxima have been observed at lower levels following the precipitation.

Acknowledgements

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