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"*Workshop on El Niño, Southern Oscillation and Monsoon*"
15 - 26 July 1996

"Indian Ocean & Monsoon"

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INDIAN OCEAN AND MONSOON

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INDIA

Workshop on

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Introduction

In the present lecture a review of India Ocean features and its relationship with the summer Monsoon has been given. There has been considerable activity over the past few decades, directed towards establishing the relationship between some of the tropical Indian Ocean parameters with the summer monsoon activity over Indian subcontinent, however, no definite conclusions have yet been drawn. The most studied parameter in the India Ocean is the Sea Surface Temperature (SST). Attemptes have been made to study the effect of seasonal and the interannual variability of Indian Ocean SST on monsoon and its variability. Since SST are primarily controlled by the behaviour of the upper ocean, emphasis is also placed by many workers on the interannual variability of other surface variables, eg., surface meteorological data, surface currents, mixed layer depth, etc. Attempts have also been made to correlate the Indian monsoon activities with one of the most important currents in the western sector of the India Ocean, i.e., the Somali Current. Another phenomena that has been studied extensively is the heat flux in the Indian Ocean. Although our awareness and understanding of the role of the tropical Indian Ocean and its relationship with the monsoon has increased considerably as a result of the TOGA programme, a clear picture is still unavailable.

1. Monsoon Winds

Monsoon are identified by the reversal of winds during the year. Amongst other features, Ramage (1971) suggested principal features of monsoon as (i) a change in the prevailing wind direction by 120° between January and July, and (ii) the average frequency of prevailing wind directions in January and July should exceed 40 percent. Figure 1.1 depicts the areas with large change in wind direction during the year and the constancy of the wind in these areas (Flöhn, 1960).

It may be seen from the figure that over the Indian Ocean, north of 10°S, the change in the direction of winds and their constancy is consistent with the features outlined by Ramage (1971). Monsoon winds are most pronounced in the summer season of either hemisphere, that is during the period June to mid-September in the northern hemisphere and in January and February in the southern hemisphere. During northern summer, the trade winds from the southern hemisphere penetrate deep into the northern hemisphere towards India and the wide stretch of southeast Asia, and to a lesser extent towards Africa. On the other hand, in the northern winter the northeast trades move southwards moving around an anticyclone over Siberia

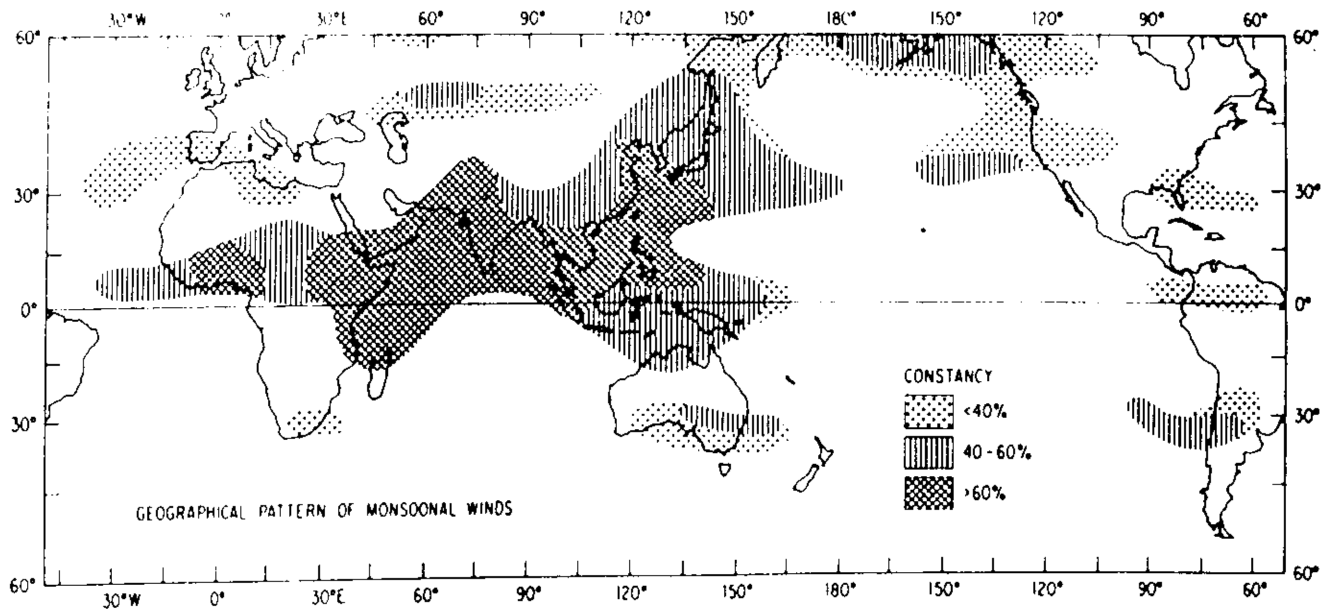
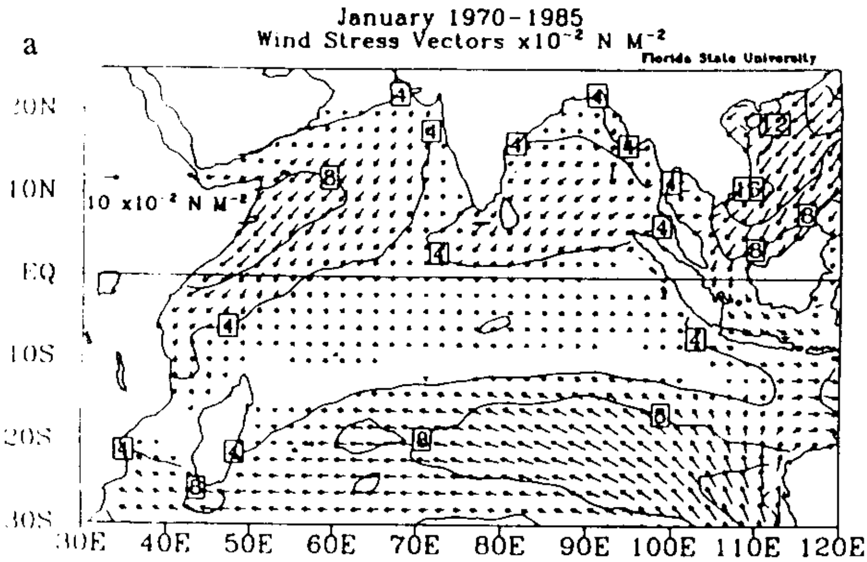


Figure 1.1 Regions where the resultant wind changes by more than 120° during the year are shown by shading. Constancy of the wind in these regions is indicated by differences in shading. The chart is from Flőhn (1960)

a



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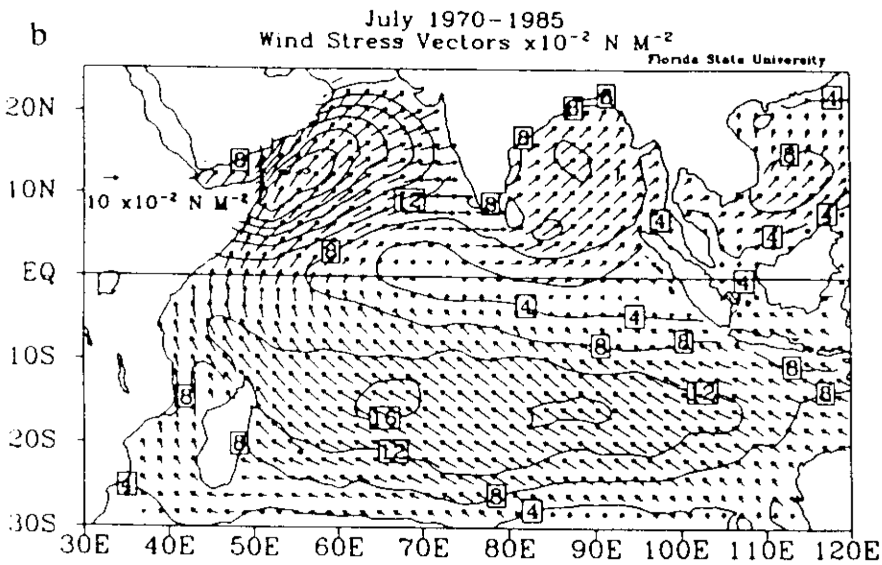


Fig. 1.2 Monthly mean wind-stress vectors over the Indian Ocean (a) January, and (b) July (Stricherz et al., 1993).

sweeps into south America, east Africa and northeast Australia. During this period, a branch of the northeast trades moving around an anticyclone over Siberia sweeps across Indonesia, Malaysia and the southern half of the Indian peninsula.

Over the Indian Ocean, the strongest winds of the northeast monsoon are in January and the strongest winds of the southwest monsoon are in July, each monsoon persists for about five months. Figures 1.2 (a) and 1.2 (b) show the monthly mean wind stress vectors over the Indian Ocean for the months of January and July respectively (Stricherz et al. 1993).

In January the northeast monsoon extends across the equator and the Intertropical Convergence Zone (ITCZ), where the northeast monsoon and southeast trades meet to the south of the equator. In July the southeast trades blow across the equator and the wind vector north of the equator rotates clockwise. Over the Arabian Sea and the Bay of Bengal the winds are thus southwesterly. Wind speeds during the southwest monsoon are considerably stronger than during the northeast monsoon.

2. Surface Circulation of the Indian Ocean

The near surface circulation of the North Indian Ocean is primarily wind-driven. On acount of the seasonal reversal of winds in the northern part of the Indian Ocean the general surface circulation differes from winter to summer. Most striking feature of the circulation system is the seasonally changing monsoon gyre. Surface current data for the Indian Ocean has been compiled in several excellent atlases (Duing, 1970; Wyrtki, 1971; Hastenrath and Lamb, 1979; Cutler and Swallow, 1984).

Figures 2.1 (a) and 2.1 (b) show the typical large-scale circulation pattern associated with the winter and summer monsoons as compiled by Wyrtki (1973) from the analysis of the ship drift data. Figures although provide excellent information about the large-scale dynamics of the tropical Indian Ocean, information on the smaller scales, particularly about the circulation in near coastal regions are somewhat limited.

In the following sections we will discuss some of the main features of the circulation in the North Indian Ocean and South Indian Ocean separately. Although figures given are for January & July only, we will discuss the currents for the three periods, viz. November to January, February to April and May to September (Pickard and Emery, 1982).

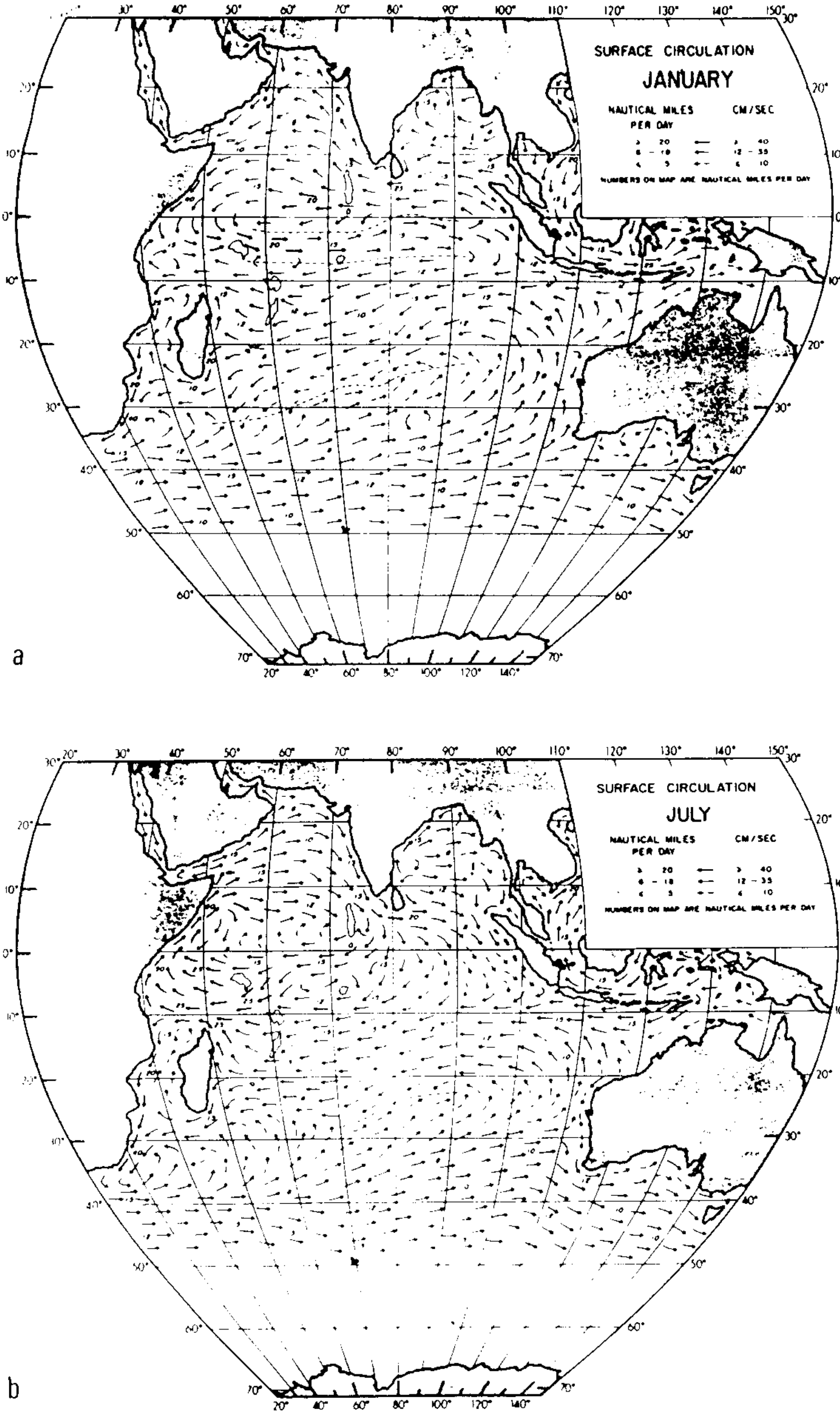


Figure 2.1 Observed large scale surface circulation during (a) NE and (b) SW Monsoons from analysis of ship drift data (Wyrtki, 1973).

2.1 North Indian Ocean

- The currents in the greater part of the North Indian Ocean, including the Arabian Sea and the Bay of Bengal, are reversed in direction seasonally by the monsoons.

- The NE-monsoon current circulation occurs during the height of that monsoon from November to Januaryy. During the later part of this monsoon, February to April, the circulation changes.

- The SW-monsoon circulation prevails from May to September. October is a transitional month.

2.1.1 NE-Monsoon Circulation (November - January)

- In the open waters of the Arabian Sea and the Bay of Bengal the currents sets in a westerly direction. These westerly sets extend southwards beyond the equator.

- Near the coasts of the Arabian Sea there is a week circulation in a counter - clockwise sense following the coasts.

- In the northern part of the Bay of Bengal there is a clock-wise circulation.

- A stronger current sets southwards down the east coast of Africa from Ras Hafun to about latitude 2°S. This, the *Somali Current*, turns eastwards between about latitudes 2°S and 5°S, to form the begining of the Equatorial Counter Current.

- Between Ras Hafun and Ras Asir (C. Guardafui) the current is northerly.

2.1.2 Later NE-Monsoon Period (February - April)

- The flow in the open waters of the Arabian Sea and the Bay of Bengal remains westerly, though the currents are more variable than from November to January.

- Towards the equator, westerlies which are well marked in February become less so in March, and in April the flow changes to easterly.

- The coastal circulation of the Arabian Sea, however, is reversed to a clockwise direction.

- In February the Somali Current flows south-westwards along the African coast to about latitude 3°S. During March the flow becomes variable and by April it reverses to north-easterly.

2.1.3 SW-Monsoon Circulation (May - September)

- In the open water the drift is easterly.

- The coastal circulation of the Arabian Sea and Bay of Bengal remains clockwise and is strengthened.

- The Somali current continues to flow northwards, from Cabo Delgado to Ras Asir (Cape Guardafui), and is greatly strengthened.

- It devides in about latitude 7°N, part continues along the coast to Ras Asir (Cape Guardafui), but the bulk turns eastwards and passes south of Socotra into the general easterly current.

- The current south of Socotra in July-September is the strongest known in the world in the open ocean and speed upto 7kn have been recorded.

2.1.4 Circulation in Red Sea and Gulf of Aden

The currents in the Red Sea and Gulf of Aden conforms to the monsoon.

- During the NE-Monsoon it sets eastwards in the Gulf of Aden and passes through the Straits of Bab-al Mandab to flow up the axis of the Red Sea.

- During SW-Monsoon the current in the Gulf of Aden sets eastwards; in the Red Sea the water flows down the axis of the sea and into the Gulf of Aden.

2.2 South Indian Ocean

2.2.1 Main Features

- The main circulation of the South Indian Ocean is counter-clockwise.

- The northern flank of the circulation is formed by the west-going South Equatorial Current. The northern boundary of this South Equatorial Current is usually between latitudes 6°S to 10°S but it varies according to longitude and season.

- The South Equatorial Current, after passing the northern extremity of Madagascar, meets the African coast near Cabo Delgado.

- At Cabo Delgado, this current divides and some of the water flows northwards along the coast. The remainder flows southwards to form a strong coastal current which, from Cabo Delgado to Lourenzo Marques, is known as the *Mozambique Current*.

- Southward continuation of the Mozambique Current is the *Agulhas Current*. This is reinforced by water from the South Equatorial Current setting past the southern extremity of Madagascar.

- Some of the water of the Agulhas Current recurves to southeastwards between about 25°E and 35°E and enters the northern aprt of the *Southern Ocean Current*.

- The remainder of the Agulhas Current continues along the coast line, and passing over the Agulhas bank, enters the South Atlantic Ocean, where it joins the Benguela current.

- The southern side of the main circulation is formed by the cold water of the Southern Ocean Current, setting in a generally easterly direction in latitudes south of about 35°S.

- The east side of the circulation is not well marked. In the northern winter months the Southern Ocean Current turns northwards as it approaches Cape Leeuwin and form a north-going current parallel to the west coast of Australia, though there is a narrow belt of south-going current close inshore.

- In the northern summer the Southern Ocean Current off south-western Australia sets easterly and turns southerly towards the coast in latitudes south of latitude 26°S. Between about latitudes 20°S and 26°S the current near the coast runs southwards from about March to August but is northerly in other months.

2.2.2 Equatorial Counter Current

- In the Indian Ocean, the *Equatorial Counter Current* is readily distinguished during the NE monsoon when it forms a belt of easterly currents lying between about latitudes 2°S and 8°S.

- In the SW monsoon, however, the criterian for determining the northern boundary of this belt disappears. Since easterlies are then more or less continuous north of about latitude 8°S. This easterly flow tends to be lighter and more variable in the south and to become more pronounced further north.

- In July and August the currents are light and rather variable between about latitudes 8°S and 2°N with a more-decided easterly flow further north.

- In June and September the region of marked easterly flow extends south wards as far as latitudes 2°S-4°S.

2.2.3 Extreme Eastern Part of the Ocean

- The currents here, including those of the Arafura Sea, are not well known, owing to the scarcity of observations.

- Eastward of Christmas Island, between the parallels of about 10°S and 12°S, there is a pre-dominance of westerly sets during most of the year and they form the most easterly part of the Equatorial Current.

3. The Role of the Indian Ocean

The strength and stability of the Indian Summer Monsoon is one of the important feature that is thought to involve intimate interaction between the Indian Ocean in general and Arabian

Sea in particular and the monsoon circulation. This is a two way interaction (Keshavamurty and Sankar Rao, 1992) :

- Monsoon current picks up copious moisture by evaporation from the Arabian Sea and the Indian Ocean. This moisture is essentially not only for the monsoon rainfall but is also important as a driving force for the monsoon - as the latent heat release in the monsoon trough strengthens the monsoon circulation.

- The top layers of the Arabian Sea cool by interaction with the monsoon current possibly by :

    - loss of heat due to evaporation,
    - upwelling off the Somalia coast and spread of the cool waters eastwards,
    - convection mixing and deepening of the mixed layer.

4. Sea Surface Temperature (SST)

The thermal structure of the upper part of the tropical Indian Ocean undergoes a marked annual cycle which is dominated by the seasonal reversal of winds. Inorder to have the better understanding of the oceanic processes which alter SST on the space and time scale and ultimately influence monsoon climate, one should have a good knowledge of its mean distribution and variability.

4.1 Mean Distribution and seasonal variability

From the Climatic Atlas of Hastenrath and Greischar (1989) and Conkright et al (1994), the following features are seen:

- In "quiet" ocean locations from land, SST is usually warmest at the end of local summer, and coldest at the end of winter.

- During the northern hemisphere winter (January-March, Fig 4.1(a)), the surface water are warmest in the equatorial belt, from where the temperature decreases northward into the Arabian Sea and the Bay of Bengal, and southward into the Southern Indian Ocean.

Conkright et al., 1984

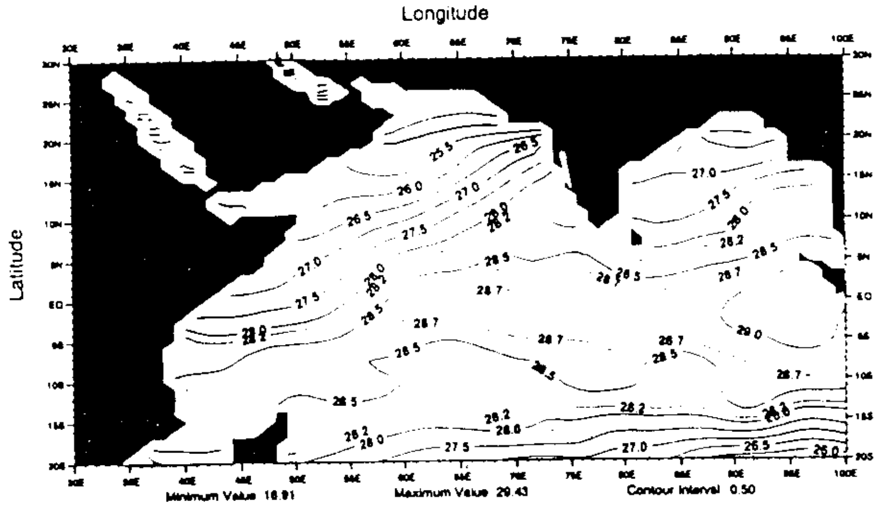


Fig.4.1 (a) Winter (Jan.Mar.) mean temperature (°C) at the surface

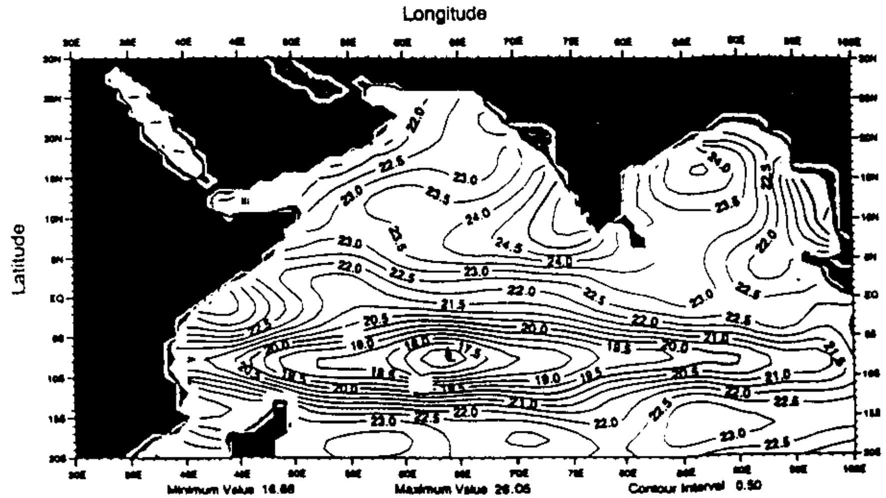


Fig.4.1 (b) Winter (Jan.Mar.) mean temperature (°C) at 100 m depth

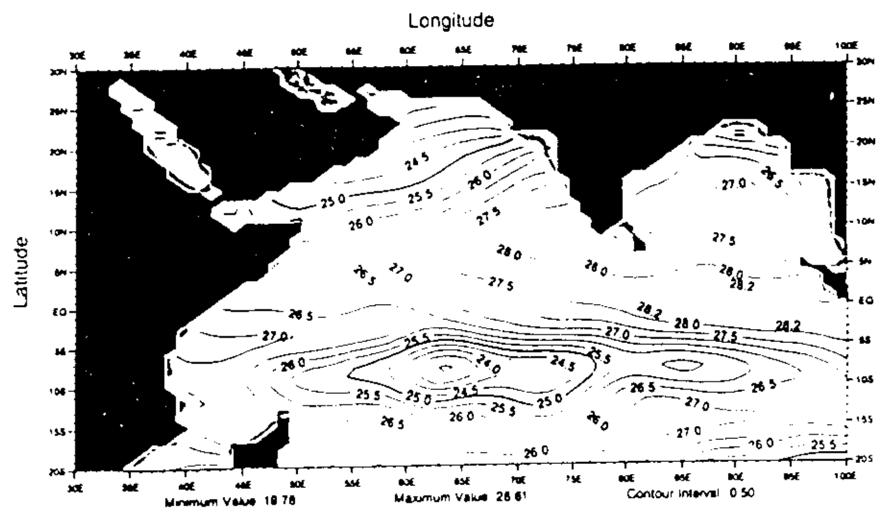


Fig.4.1 (c) Winter (Jan.Mar.) mean temperature (°C) at 50 m depth

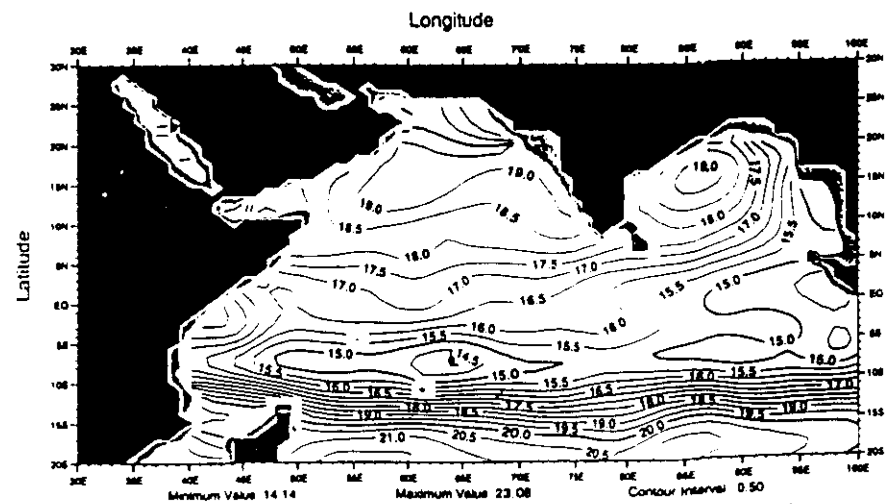


Fig.4.1(d) Winter (Jan.Mar.) mean temperature (°C) at 150 m depth

- The surface isotherms during winter in the low latitudes assume a zonal orientation. Meridional gradients grow particularly strong in the southern portion of the Red Sea.

- Horizontal temperature pattern change remarkably with depth. Overall horizontal gradients are particularly steep and patterns diverse at 50,100 and 150 m levels which are broadly the regions of the thermocline (Figs. 4.1 (b), (c), (d)).

- The surface temperature pattern characteristic of the Indian Summer Monsoon evolves gradually from March to June (Fig 4.2(a)), accompanying the establishment of the southwest monsoon winds.

- The zone of maximum SST shifts north of the equator during March to May. Highest SST is found in a broad southwest to northeast oriented band extending from the equatorial region of the Arabian Sea to the Bay of Bengal.

- With the development of strong cross-equatorial southwest monsoon winds in May, relatively cool surface waters appear along the coasts of Somalia and Arabia.

- During northern hemisphere summer (July-Sept), the cooling off the coasts of Somalia and Arabia gets accentuated due to upwelling associated with strong cross-equatorial winds. Relatively low sea surface temperature also occur off the south and south-east coast of India (Fig 4.3(a)).

- Some other distinctive features of the summer SST pattern include cool surface waters in the western Bay of Bengal; high SST in the Red Sea and Persian Gulf, a zone of warmest surface waters near the equator in the Central and eastern Indian Ocean and strengthening of the meridional SST gradient across the southern tropical Indian Ocean.

- With the weakening of the cross equatorial airflow during fall months (Oct-December), the cold surface waters off the coast of Somalia, Arabia and Southern Indian Ocean disappear (Fig 4.4(a)).

- The transition of the winter monsoon is further marked by cooling of surface waters in the Arabian Sea, Bay of Bengal and the appearance of broad zone of warmest surface waters in equatorial belt.

Conkright et al., 1984

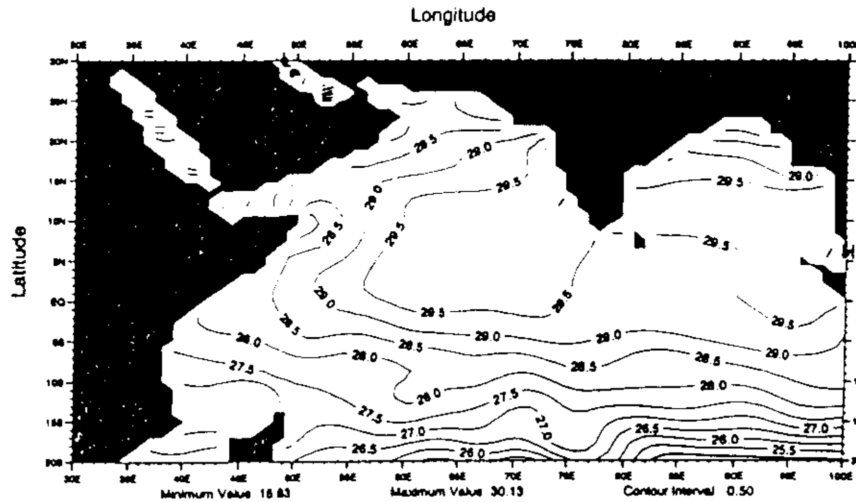


Fig.4.2 (a) Spring (Apr.-Jun.) mean temperature (°C) at the surface

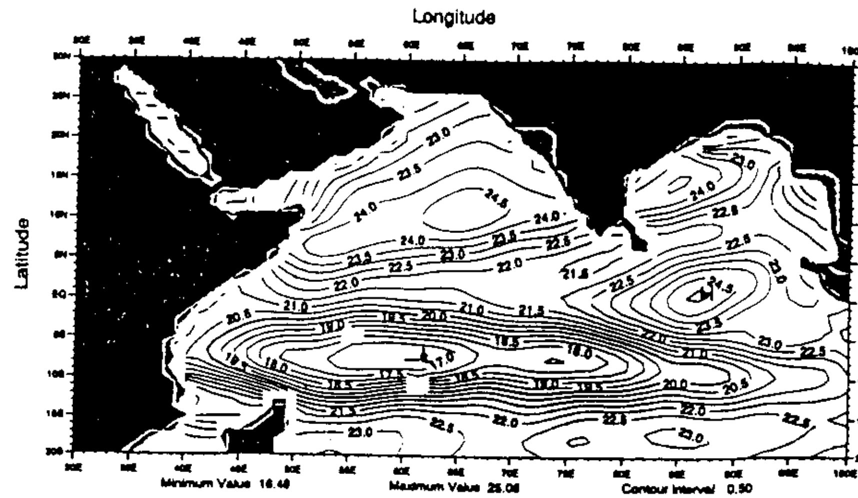


Fig.4.2 (b) Spring (Apr.-Jun.) mean temperature (°C) at 100 m depth

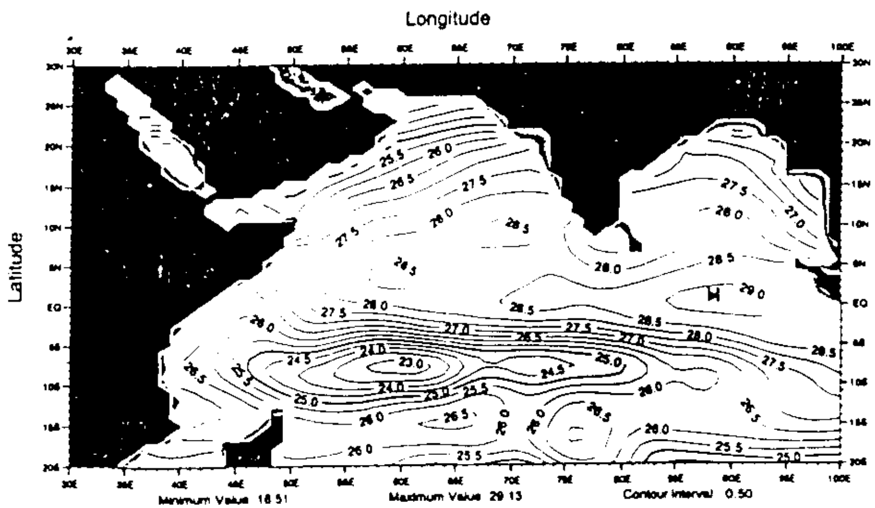


Fig.4.2 (c) Spring (Apr.-Jun.) mean temperature (°C) at 50 m depth

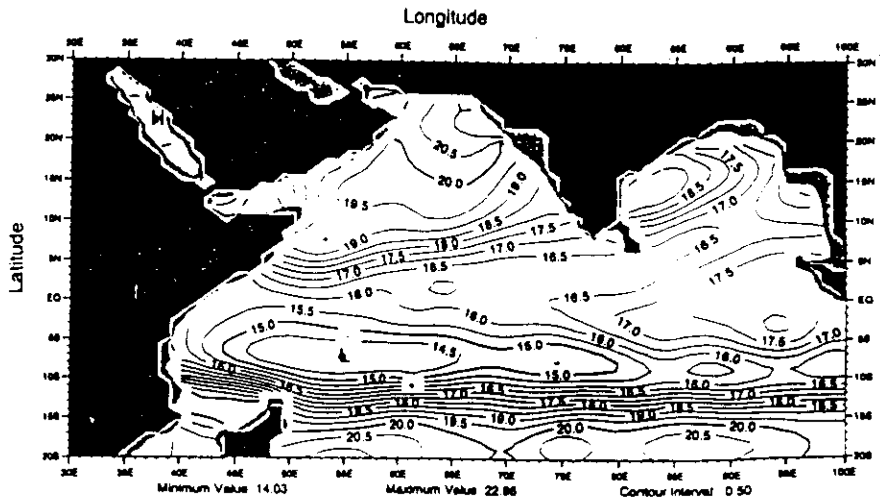


Fig.4.2 (d) Spring (Apr.-Jun.) mean temperature (°C) at 150 m depth

Conkright et al., 1984

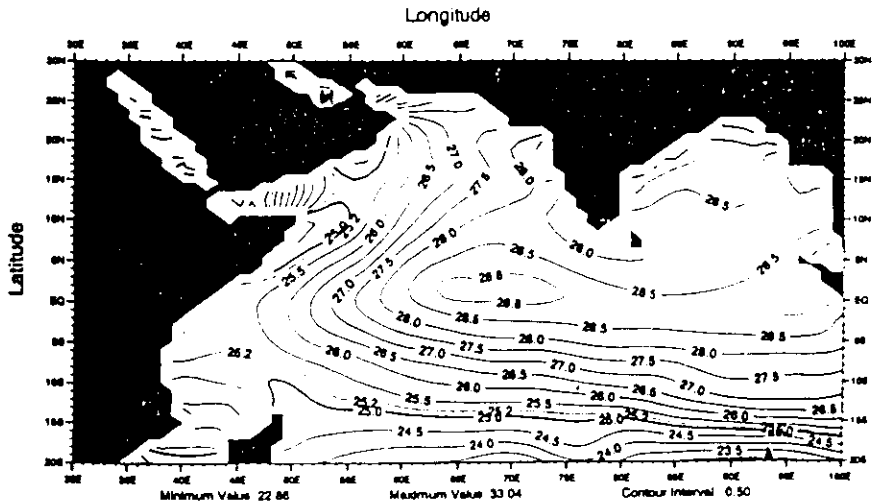


Fig.4.3 (a) Summer (Jul.-Sep.) mean temperature (°C) at the surface

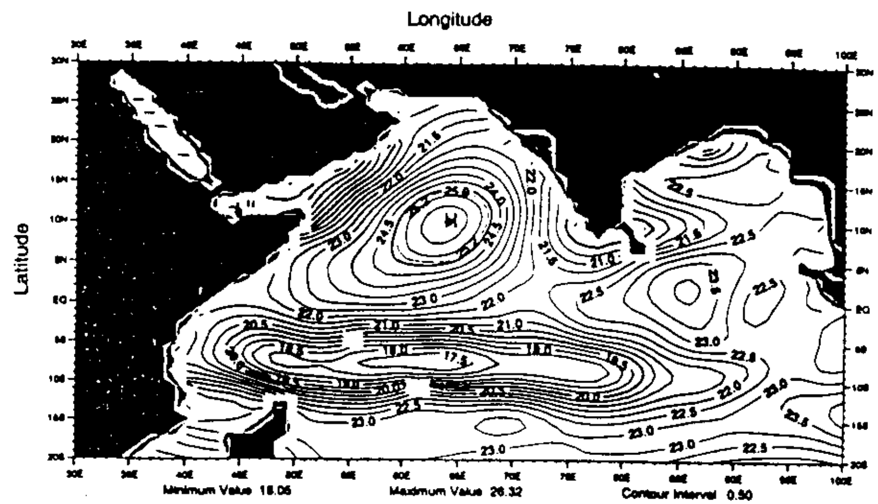


Fig.4.3 (b) Summer (Jul.-Sep.) mean temperature (°C) at 100 m depth

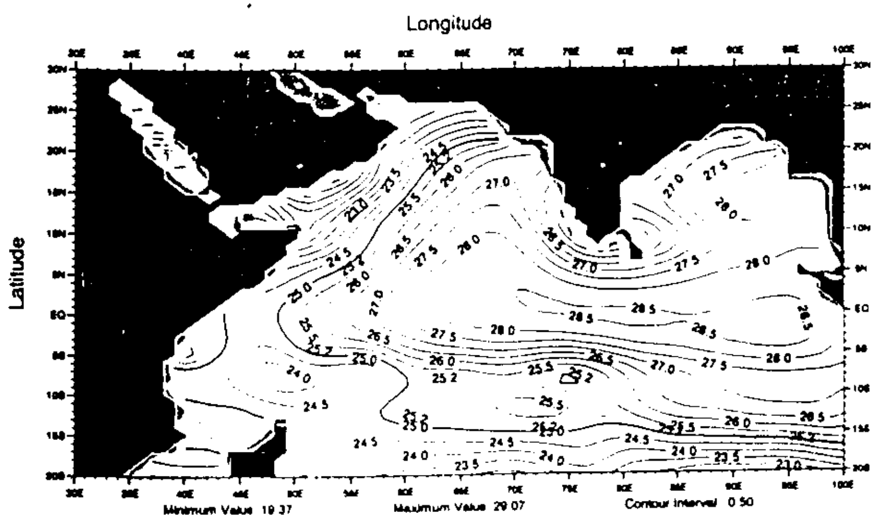


Fig.4.3 (c) Summer (Jul.-Sep.) mean temperature (°C) at 50 m depth

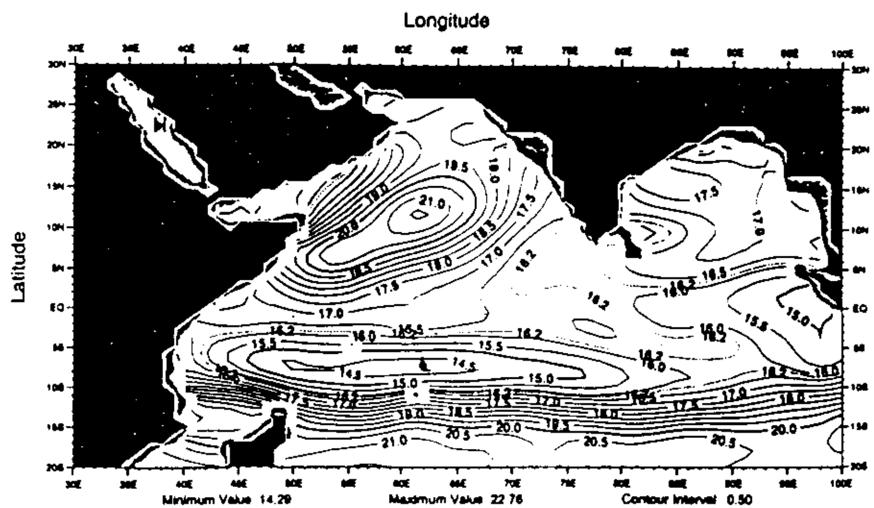


Fig.4.3 (d) Summer (Jul.-Sep.) mean temperature (°C) at 150 m depth

Conkright et al. (1994)

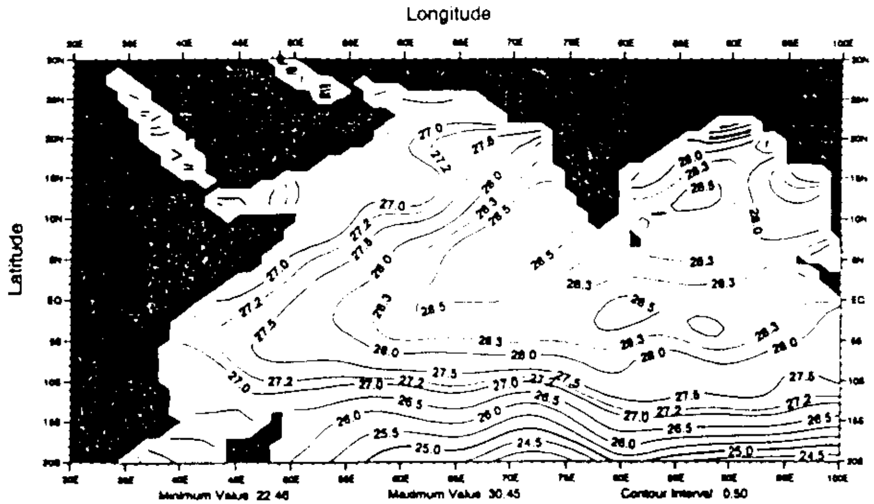


Fig. 4.4 (a) Fall (Oct.-Dec.) mean temperature (°C) at the surface

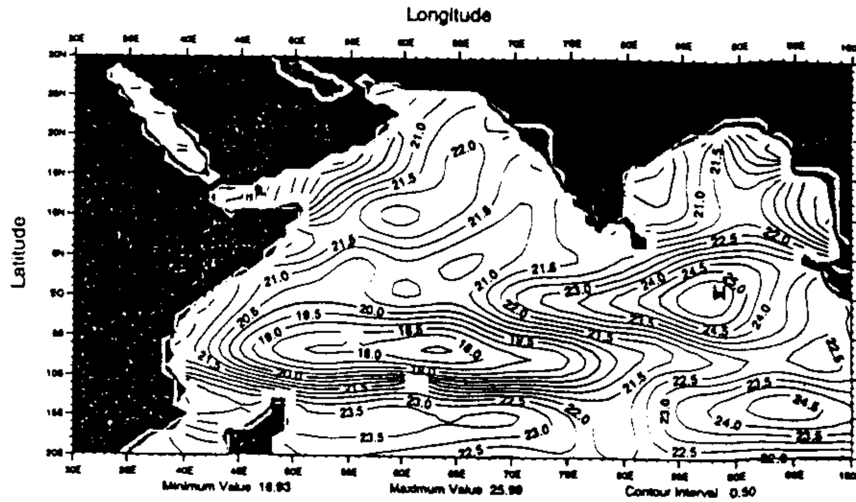


Fig. 4.4 (b) Fall (Oct.-Dec.) mean temperature (°C) at 100 m depth

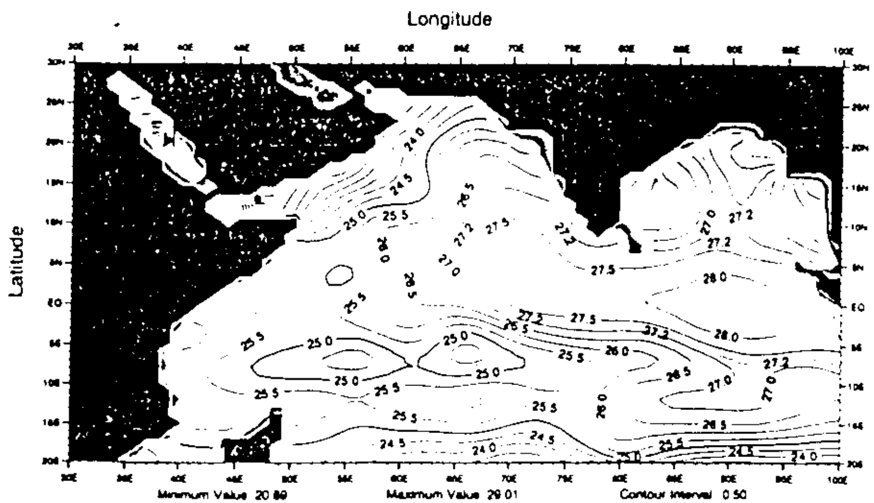


Fig. 4.4 (c) Fall (Oct.-Dec.) mean temperature (°C) at 50 m depth

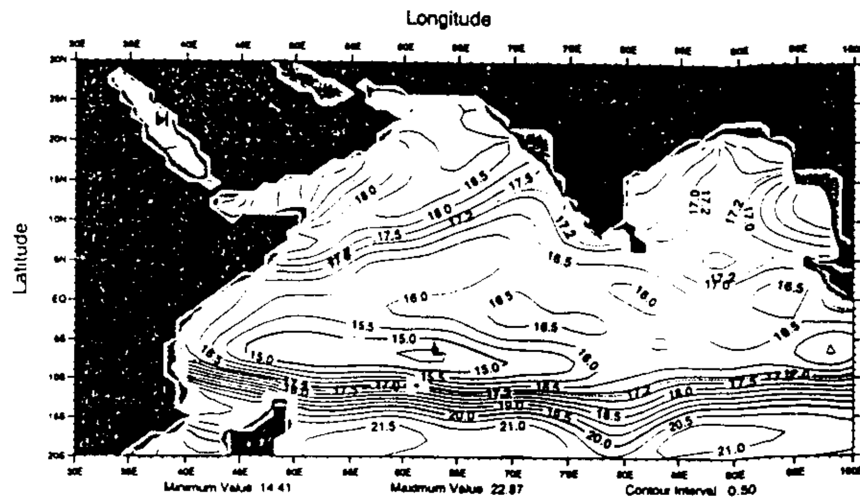


Fig. 4.4 (d) Fall (Oct.-Dec.) mean temperature (°C) at 150 m depth

4.2 Main Features of subsurface thermal structure

(a) Dramatic changes from the northern hemisphere winter (January to March) to the summer monsoon (July-September) half-year take place, however, to the north of the equator, especially in the northwestern part of the tropical Indian Ocean. A broad tongue of maximum temperature somewhat to the north of the equator and extending from the eastern into the Central Indian Ocean evolves from spring to summer, and decay again in fall. This feature broadly located around the middle of a large gyre in the upper oceanic circulation is more pronounced at 50m (Figs. 4.1(b), 4.2(b), 4.3(b), 4.4(b) than at the surface. It largely fades out by the 100m level (Figs 4.1(c), 4.2(c), 4.3(c), 4.4(c)).

- Another prominent feature is the shallowing of thermocline base along the coast of Somalia and Arabia and the deepening in the South Central Arabian Sea. These two distinct nuclie are reflected in the temperature patterns (Figs 4.3(c), 4.3(d)) and mixed layer depth (Fig 4.5).

- Most dramatic is the pattern evolution of the northern hemisphere summer in the northwest sector of the Indian Ocean which is characterised by a shallowing of the thermocline base along the coasts of Somalia and Arabia and a deepening in the South Central Arabian Sea (Figs. 4.6(a) to 4.6(f)), which is reflected in the development of a band of coldest waters along the shores and a core of maximum temperature in the open ocean particularly apparent in the maps for the 50, 100 and 150m levels (Figs 4.2(b), 4.2(c), 4.2(d), 4.3(b), 4.3(c), 4.3(d), 4.4(b), 4.4.(c), 4.4(d)). The gradual waning of these features of surface thermal structure from the summer to the winter can be traced in both the maps of thermocline (Figs 4.6(e), 4.6(f), 4.7(a), 4.7(b)) and the temperature at standard depths (Figs 4.4(b), 4.4.(c), 4.4(d), 4.1(b), 4.1(c), 4.1(d)).

4.3 Seasonal Variability

Southern Indian Ocean, cools by 3°-5°C from March to September, south of 15° south (Figs 4.1(a), 4.3(a)). This is mainly because of the fact that SST is usually warmest at the end of local summer and coldest at the end of winter. However, it may be seen that in an average year SST also cools by 2°-3°C from May to September throughout most of the Indian Ocean (Figs 4.2(a), 4.3(a)). Cause of this general cooling may be traced to enhanced loss of surface heat away from the western and northern boundaries and the effects of upwelling near them.

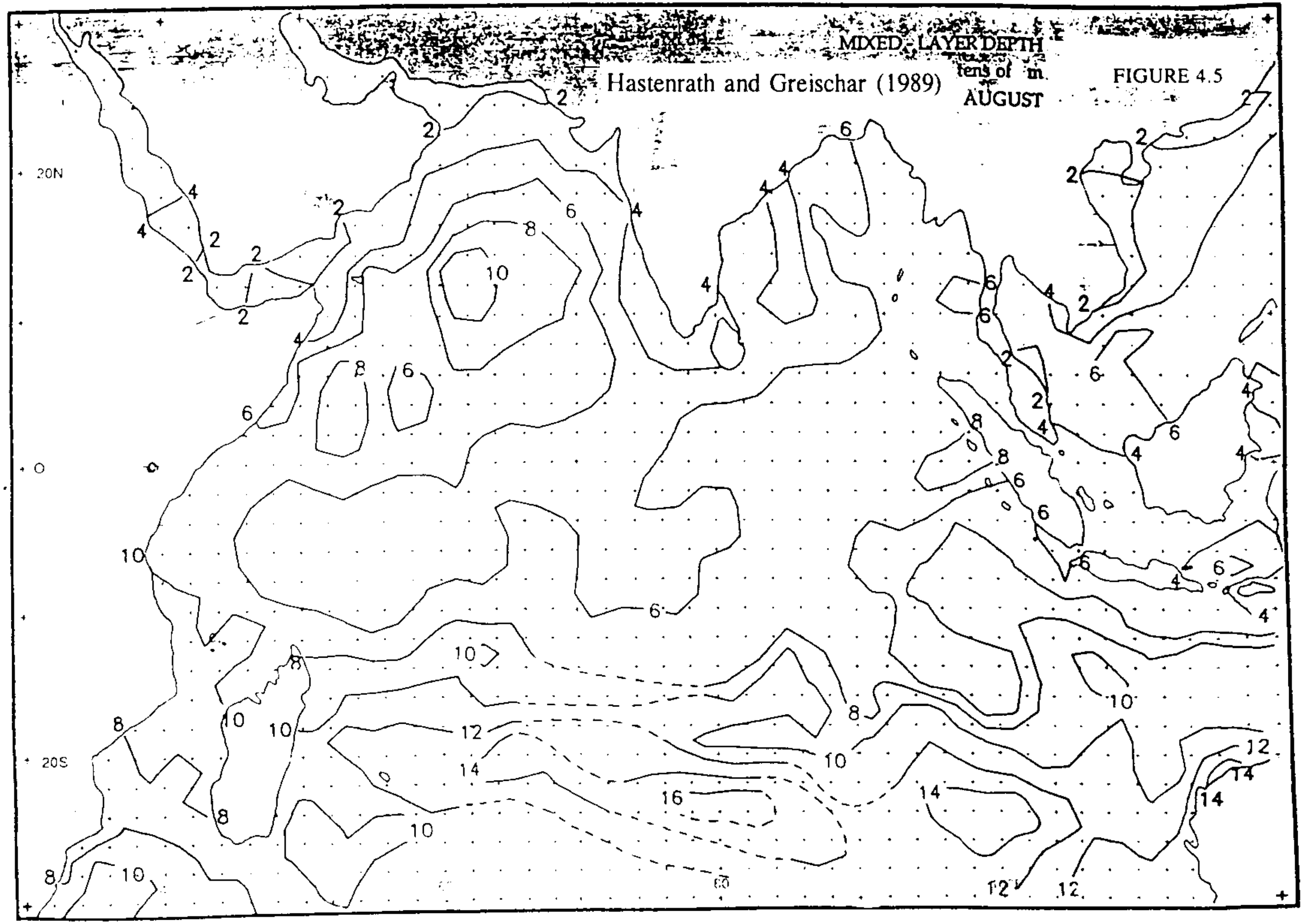

MIXED LAYER DEPTH
tens of m
AUGUST
Hastenrath and Greischar (1989)
20N
0
20S

FIGURE 4.5

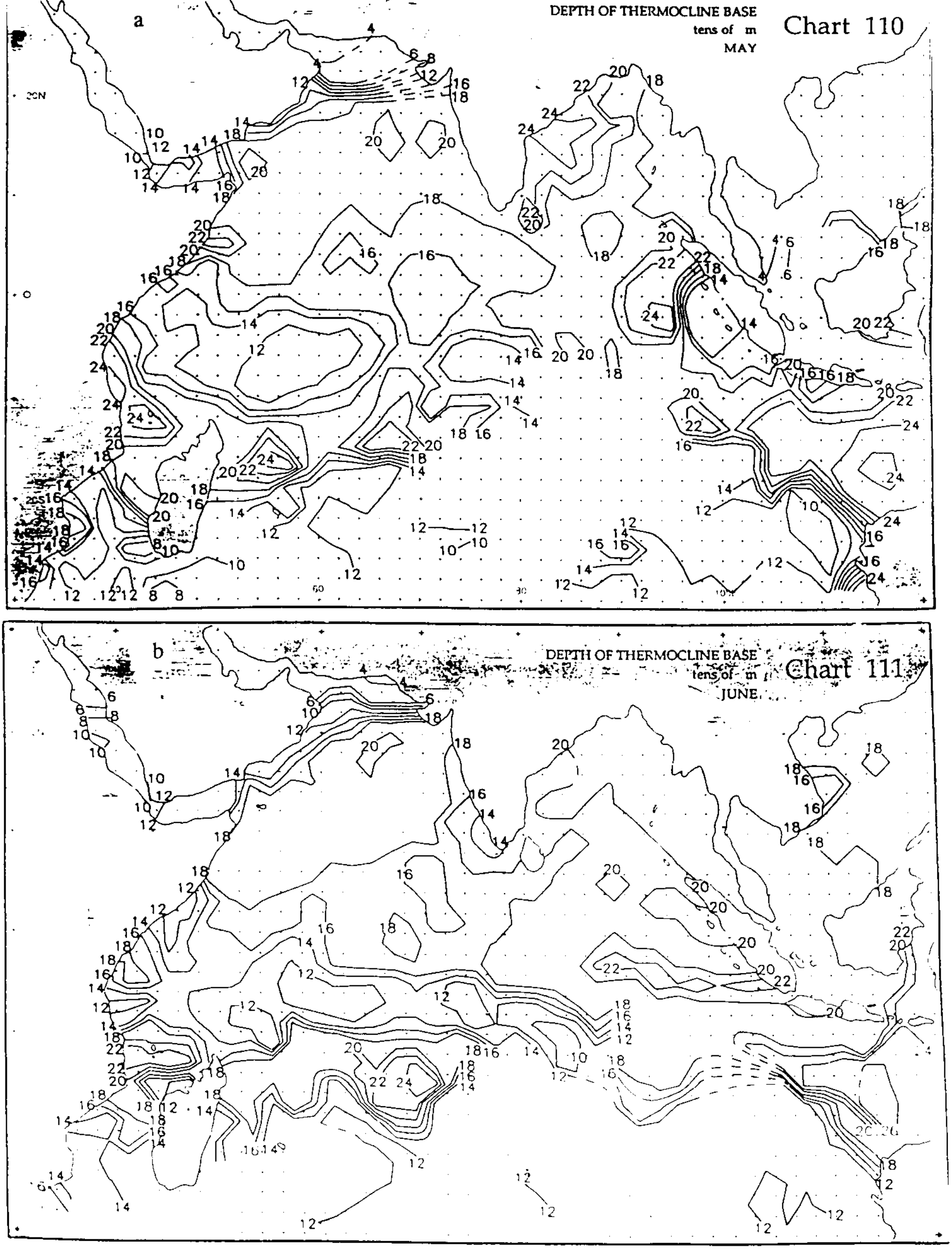
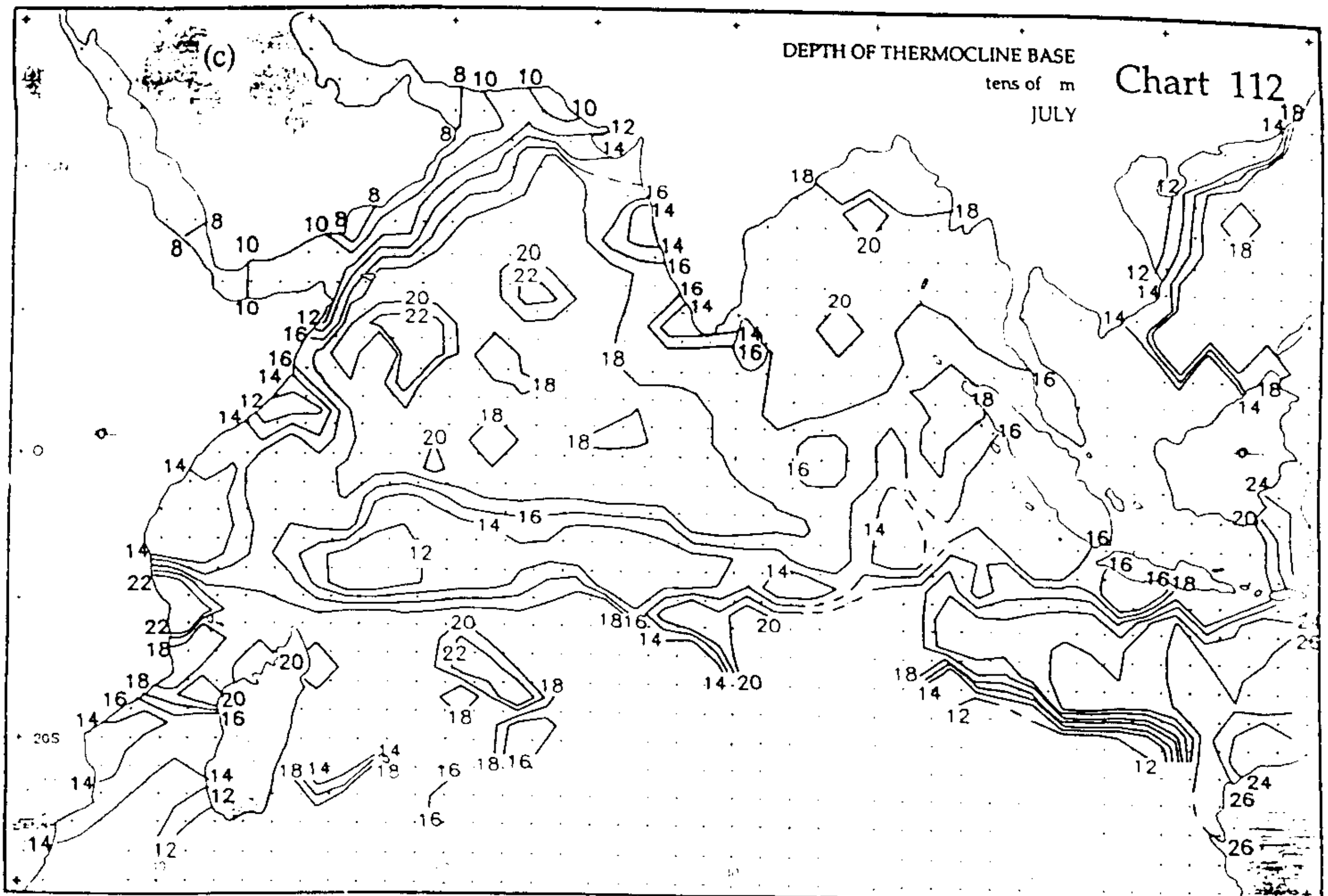


Fig.4.6 Depth of Thermocline Base (a) May, (b) June, (c) July, (d) August, (e) September, and (f) October (Hastenrath and Greischar, 1989).



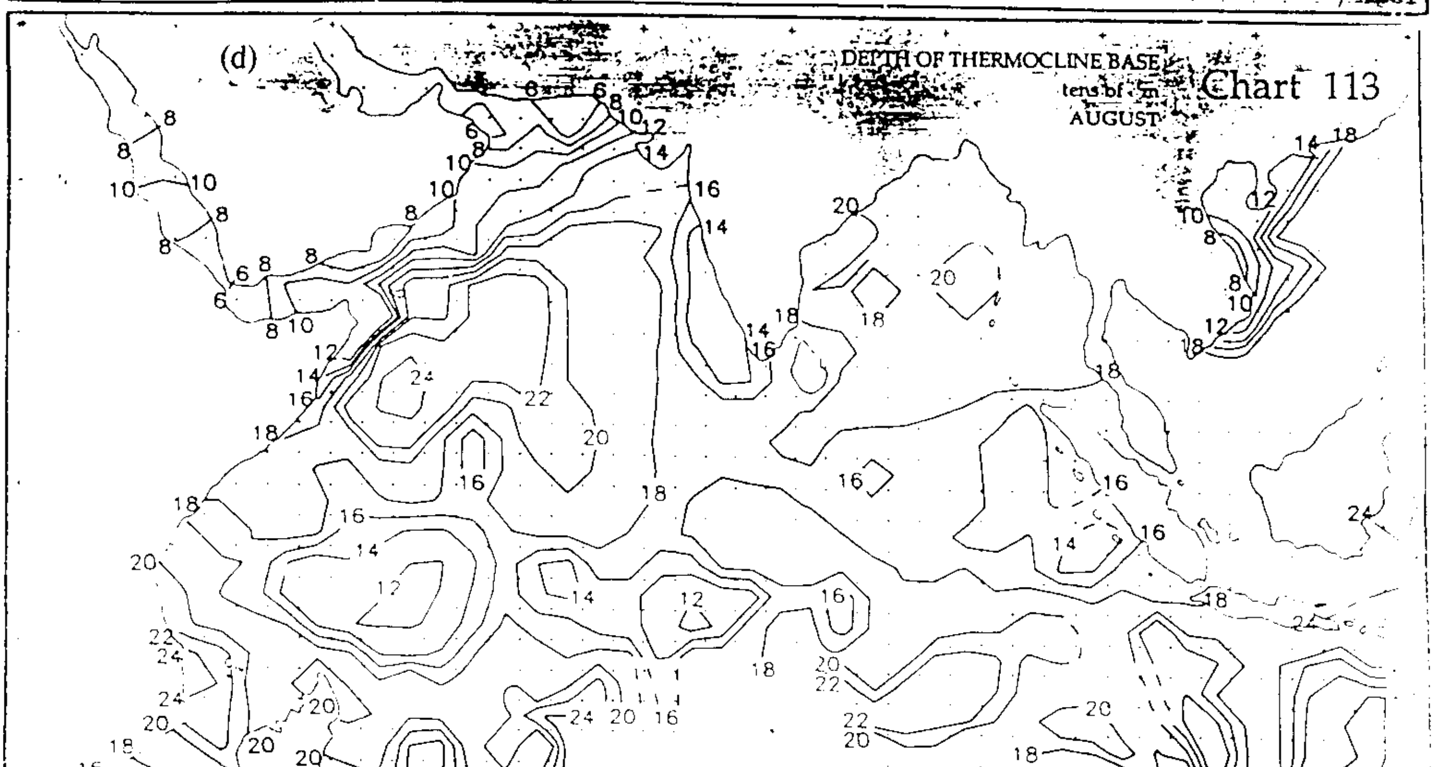
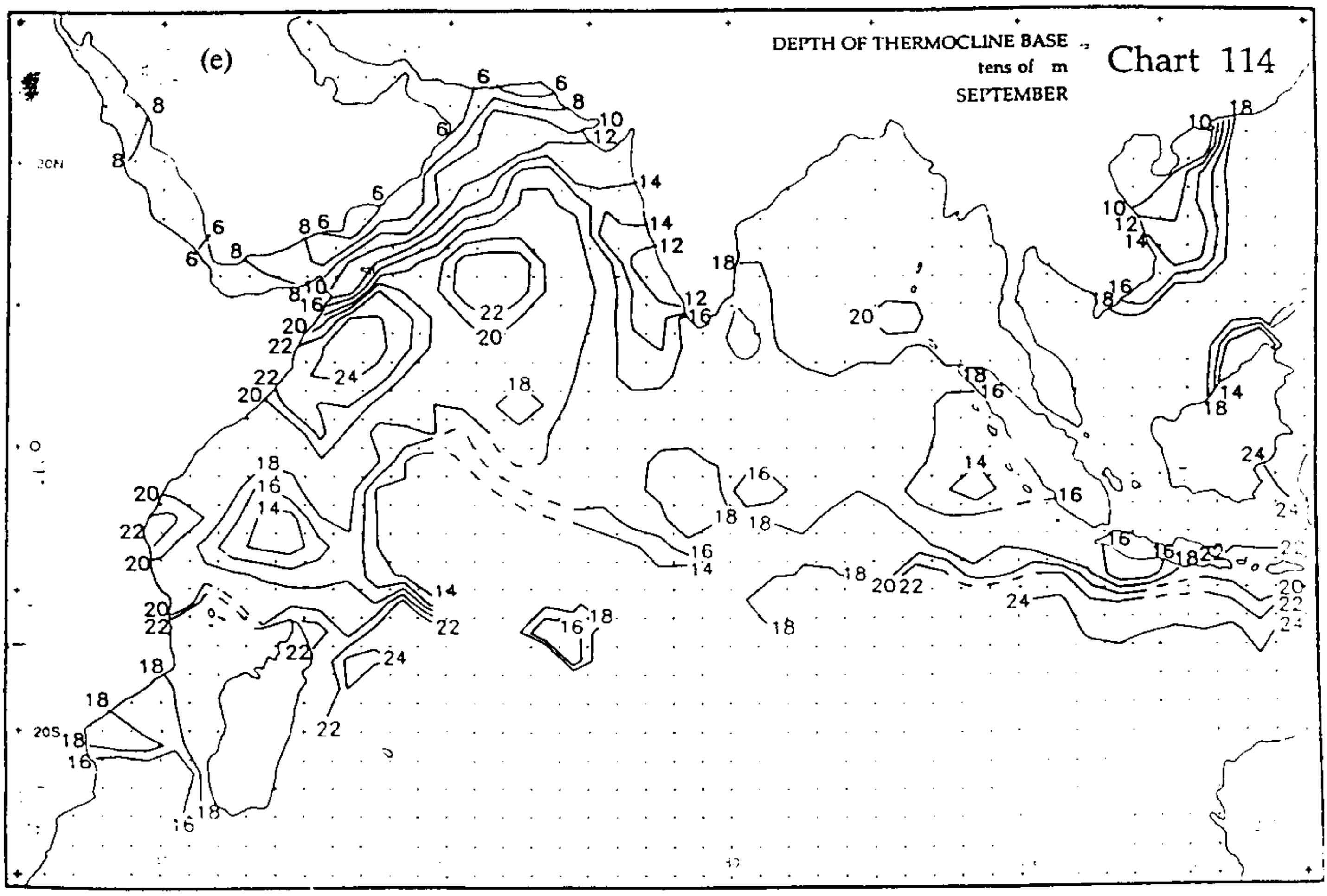


Fig. 4.6 Continued



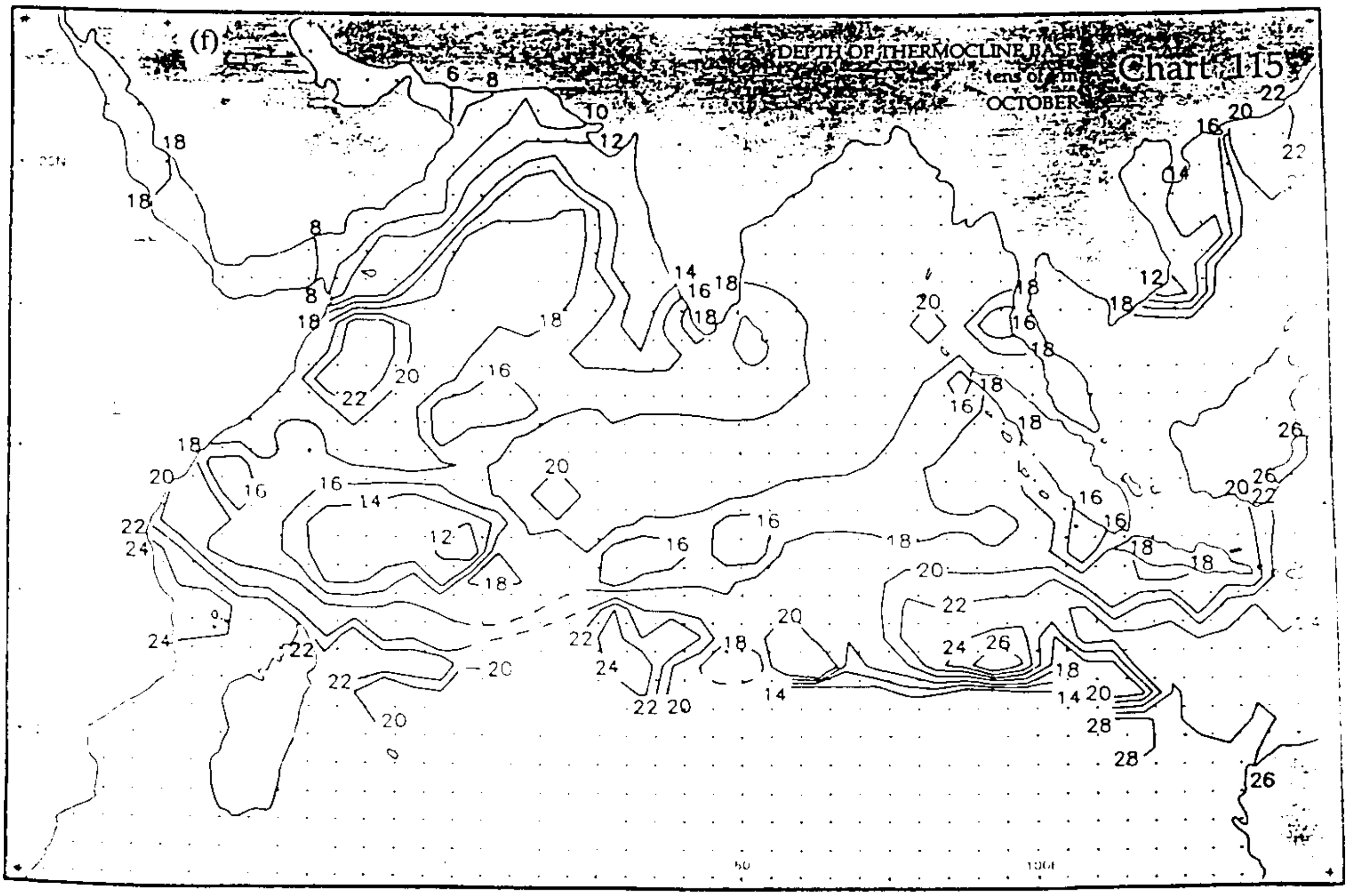
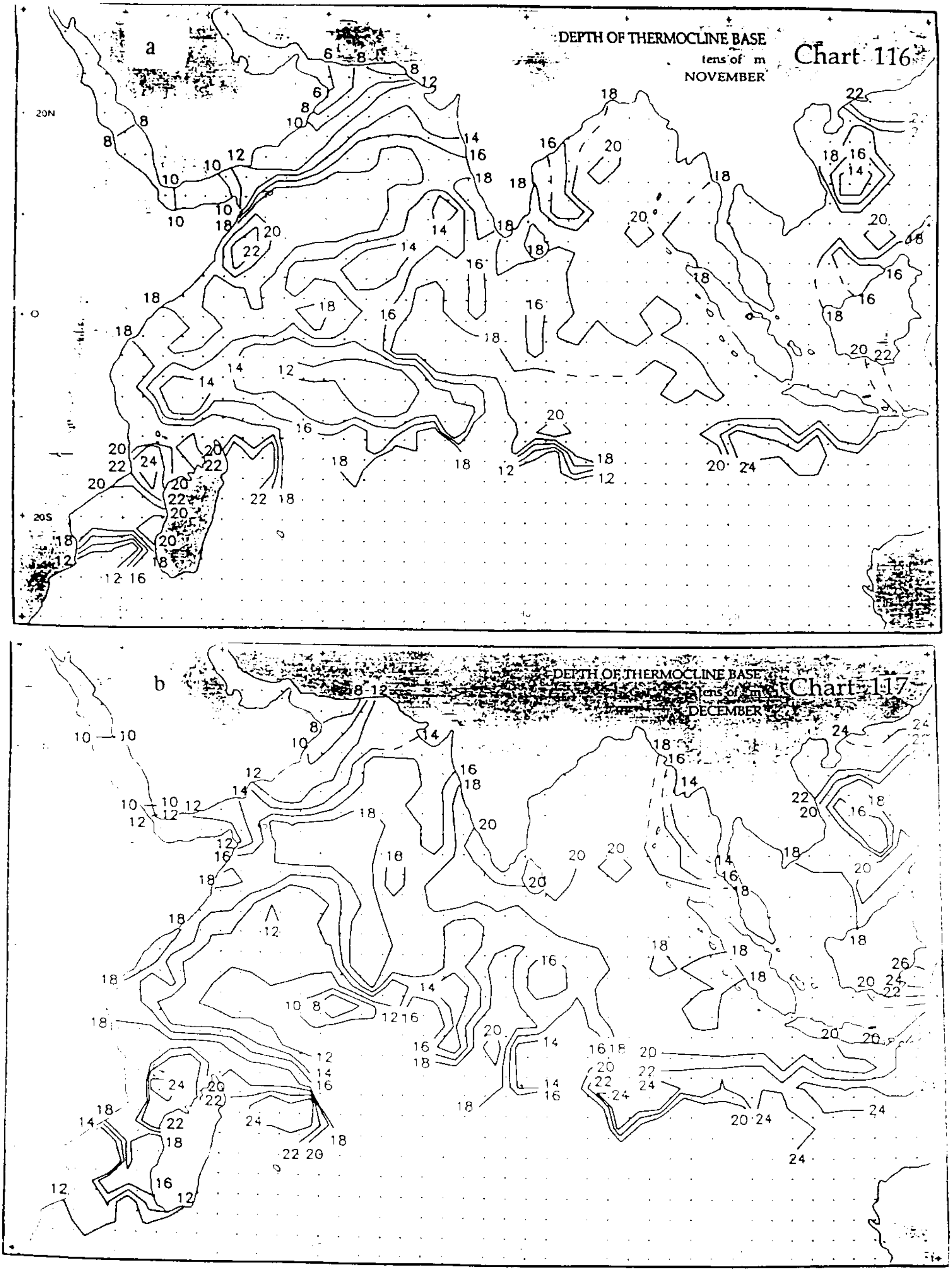


Fig. 4.6 Continued

Fig.4.7 Depth of Thermocline Base (a) November, and (b) December (Hastenrath and Greischar, 1989).



Summer upwelling is also seen at a number of places other than the coast of Africa and Arabia. Apart from the east coast of India (Shetye 1991; Johns et al, 1991, 1992, 1993; Dube et al, 1992, 1993, 1995; Rao et al, 1993, 1995), it must be particularly marked off the southern tip of Srilanka and India where strong winds blow parallel to the coast, although this region is not well studied. There are no regions with shallow thermocline under strong easterly winds along the Indian Ocean equator that might give rise to SST cooling due to equatorial upwelling. A consequence is that east-west SST gradients are small everywhere on the equator except in the very west, where there is coastal upwelling in the summer months.

McCreary and Kundu (1989) studied the SST variability in the Arabian Sea with the help of a numerical hydrothermodynamic model. Their results showed that in response to SW monsoon, SST along the coast of Somalia and Arabia decreases owing to coastal upwelling, and some of this cool water advects offshore (Fig 4.8(a)). The decrease of SST in the Central Arabian Sea is mostly due to evaporative cooling. A sharp temperature front is generated across the Gulf of Aden, due to the low winds and the lack of advection within the Gulf. During the NE Monsoon, the surface isotherms take a NE-SW orientation due to evaporative cooling by the monsoon winds (Fig 4.8(b)). There is additional cooling of SST at this time near the northern boundary of the basin because of entrainment generated by surface heat loss. There is no coastal upwelling, since the Ekman drift is directed offshore in this season. Another factor contributing to the SST decrease is wind mixing away from the coastal upwelling regions. This fact is reflected in the deepening of the mixed layer from May to September (Fig 4.9, Rao et al 1989).

4.4 *Interannual Variability (Godfrey et al, 1995)*

The largest know pattern of SST in the Indian Ocean is associated with the largest known interannual climate signal, namely ENSO. It was Walker (1923) who first hinted that variations of the monsoon may be related to events in the equatorial Pacific Ocean. Later, several workers explored the possible linkage between the Indian Summer Monsoon and the ENSO events. Cadet (1985), Wright et al. (1985) and Villwock and Latif (1994) have provided detailed analysis of this phenomena. Rasmusson and Carpenter (1983) showed that there is an apparent connection between ENSO and the Asian summer monsoon. Drought years over India are associated with the warm phase of ENSO, and the flood years with the cold phase (Shukla, 1987, 1991). Without going into greater details of this topic, I will illustrate here some of the conclusions of Wright et al. (1985) as also reported by Godfrey et al. (1995). These conclusions are based on analysis of COADS data sets. There analysis show that:

(a)

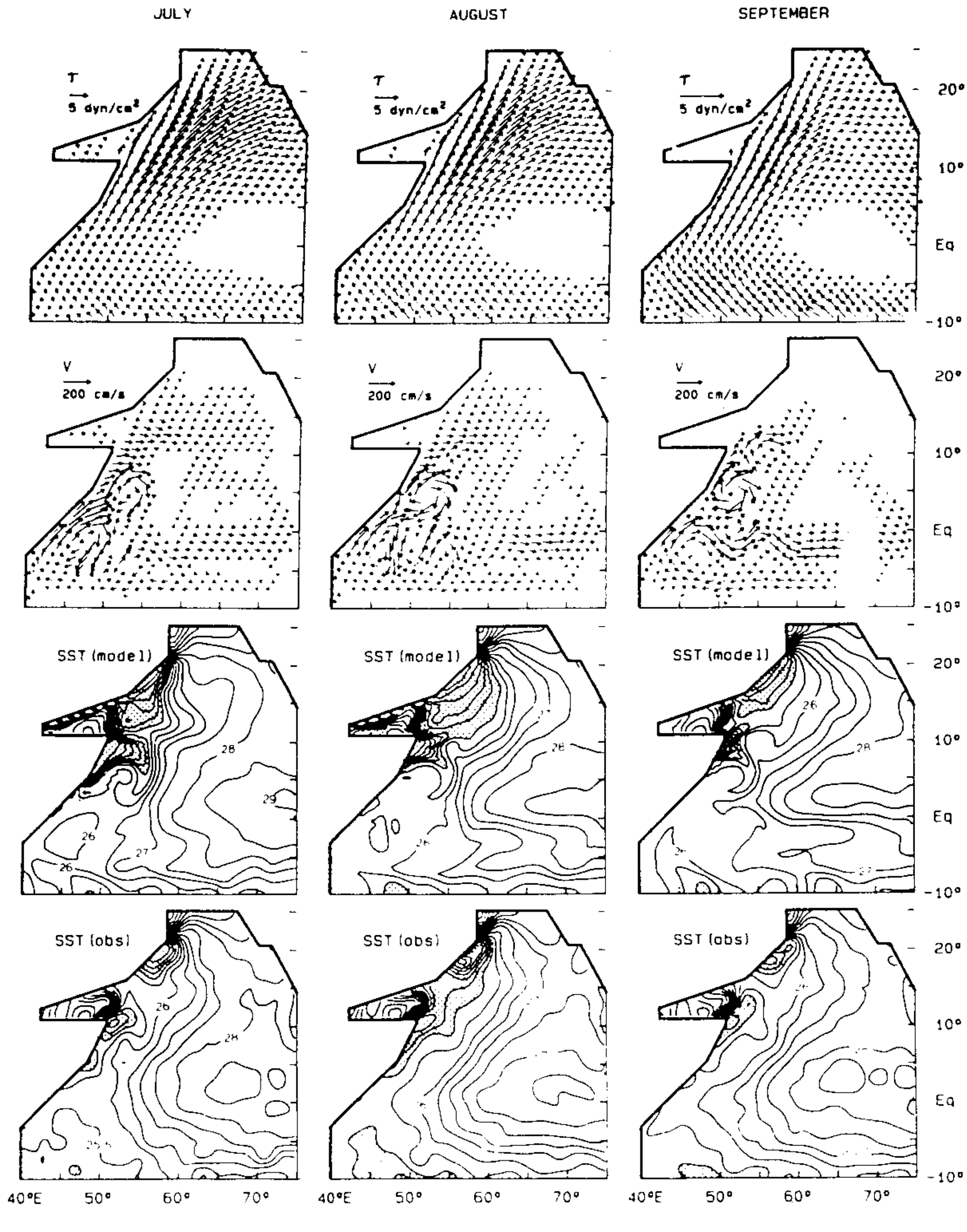


Fig.4.8 Monthly distribution of wind stress, model current, model SST, and climatological SST. The SST has a contour interval of 0.5°C, and regions less than 25°C are shaded. (a) July-September, (b) January-March (McCreary & Kundu, 1989).

(b)

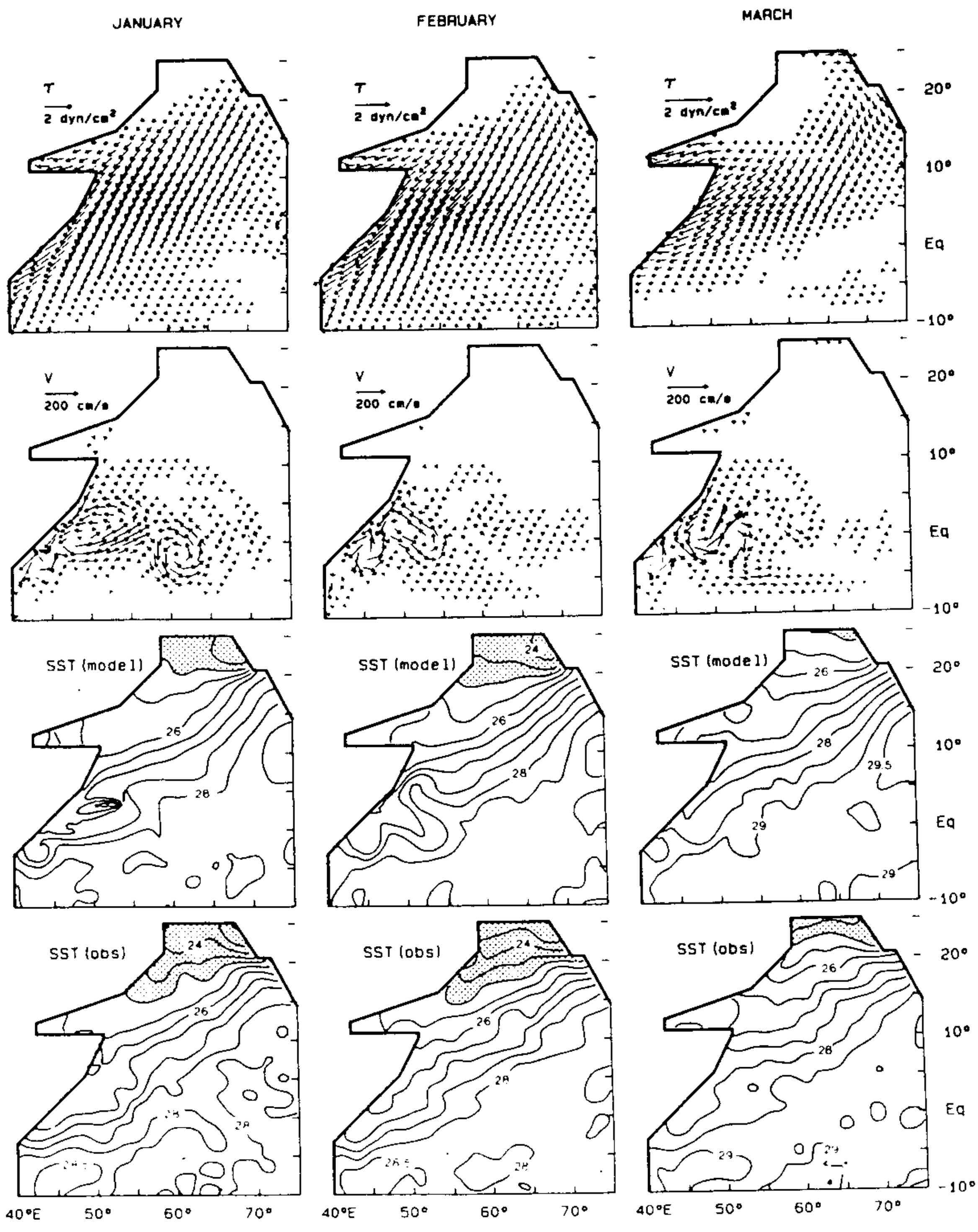


Figure 4.8 continued

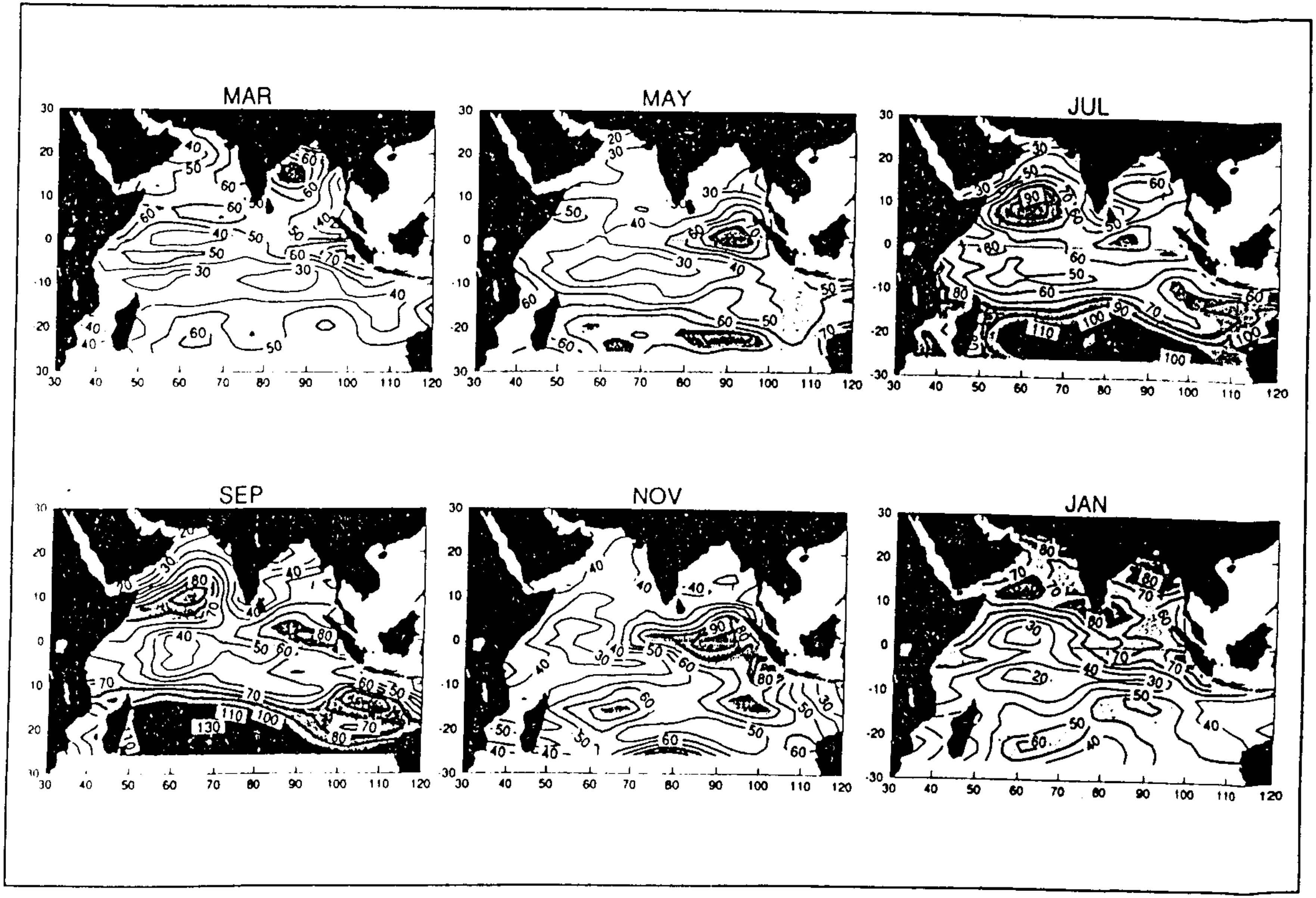


Fig 4.9 . Seasonal mean mixed layer depth (defined by the depth at which temperature is 1°C less than at the surface) at two month intervals. Contour interval is 10 m (from Rao et al., 1989).

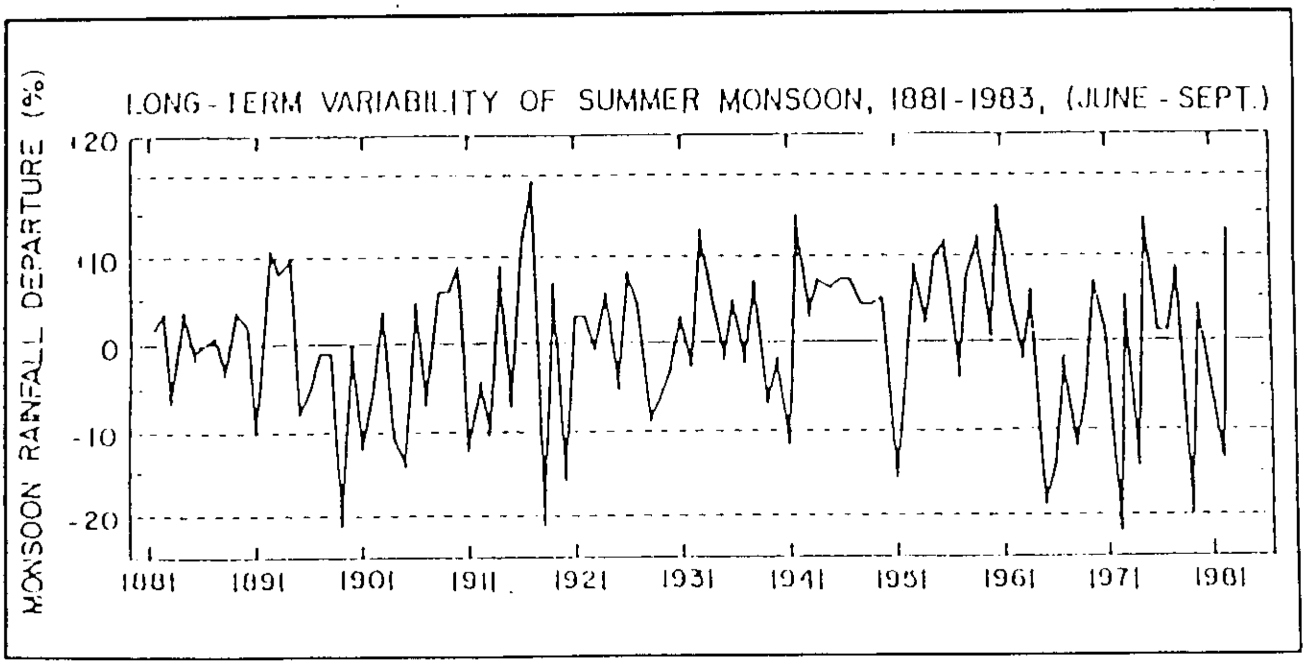


Fig.4.10 Indian monsoon rainfall 1881-1983 expressed as a percentage departure from the long-term mean. Values averaged over India for June-September are updated by Verma et al., 1984 (from Meehl, 1987).

- There is no noticeable change in the southwest Monsoon Winds along the Somali Coast in June-August during ENSO events contrary to the fact that monsoon rainfall is definitely weaker over India during ENSO events on average, and winds and rainfall are usually closely correlated.

- Decrease in cloudiness over most part of the tropical India Ocean during March-May at the start of "Darwin Pressure" year.

These patterns in winds and cloudiness have been used to explain the variations in SST in the Indian Ocean through an ENSO event. SST anomalies start slightly cold in December-February preceding the "Darwin Pressure year". They are slightly positive in the northern Indian Ocean in March-May preceding the monsoon, but this part of the ocean warms steadily over the following six months. The eastern tropical Indian Ocean remains cool till December-February at the end of the "Darwin Pressure Year".

Since the winds over the North Indian Ocean do not change significantly during an ENSO event, therefore rise in the SST may not probably be due to changes in the upwelling or evaporation. It may probably be attributed to increased insolation because of decreased cloudiness (Wright et al, 1985). However, the same argument does not hold good for SST changes in the eastern Indian Ocean where SSTs remain low till December-February, despite reduced cloudiness. Reduced SST in this region may probably be associated with the reduction of the sea level (thermocline depth) during ENSO events, bringing colder water closer to the surface (Wyrtki, 1975; Bye & Gordon, 1982). As the Indonesian region is marked by vigorous tidal mixing (Ffield and Gordon, 1992) it may produce lower SST.

4.5 Non-ENSO Interannual Variations (QBO) (Godrey et al, 1995)

There are several attempts to study the possible relationship between the SST of the Indian Ocean and summer monsoon and its variability. Most of these studies mainly attempted to correlate the SST anomalies in the Arabian Sea and the monsoon rainfall over India. The results of these studies differ even in sign of correlation obtained and, therefore, the present picture of the relationship between Indian Ocean SST and the Indian Summer Monsoon is somewhat contradictory. It is now realised that besides Indian Ocean SST there are other factors, such as, SST in the Pacific Ocean and Eurasian Snow Cover preceding the monsoon which have also strong influence on the variability of the Indian Monsoon rainfall. The problem of finding cause-and-effect relationship is further complicated due to the fact that each of these are also correlated with each other. Mention may be made to some of the significant studies carried in this regard (Saha, 1970, 1974; Shukla, 1975; Sikka and Raghvan, 1976; Raghvan et al, 1978; Joseph 1981; Joseph and Pillai, 1984, 1986; Shukla and Misra, 1977; Weare, 1979; Washington et al, 1977; Druyan, 1982 (a,b); Shukla, 1987; Cadet and Diehll, 1984; Barnett, 1984; etc.). It may not be possible to describe the results of all these studies and therefore, I am leaving it to the readers. However, I would briefly describe some of the recent works which have attracted attention.

Shukla (1987) analysed the Indian Ocean SST observed along the shipping route from the Red Sea to Singapore and found that this SST does not significantly contribute to Indian Monsoon Rainfall. In other studies Meehl (1987, 1993) and Yasunari (1990) associated the biennial oscillations (Fig 4.10) of the Indian Monsoon rainfall with a large-scale air-sea interaction phenomenon over Indian and Pacific Oceans. Palmer et al (1992) used the runs of atmospheric GCM for the ENSO (1987) and non-ENSO (1988) years to suggest that large SST anomalies in the east pacific are probably more influential in causing the Asian and Australian monsoon to be strong than the Indian Ocean SST anomalies. Shukla and Fennesy (1994) conducted sensitivity experiments using COLA GCM. Their results indicate that the annual cycle of SST is as important as the annual cycle of solar forcing in the establishment of the Asian Summer Monsoon circulation and rainfall over India.

4.6 Intraseasonal Oscillations

Interaseasonal variability in the monsoon rain is very common leading to periods of heavy rain for a few-weeks followed by weak monsoon and then again periods of active monsoon. This problem of variability during the monsoon season has still not been understood. There have been several attempts to explain these oscillations by correlating them with various features affecting SST of the India Ocean.

Sikka and Gadgil (1980) and Gadgil and Asha (1992) suggested that these intraseasonal variations in the Indian monsoon rainfall are associated with variability in the northwards migration, decay and reappearance of new rainfall bands on the equator. These rainfall bands form over the eastern equatorial Indian Ocean and move northward to northern India.

Tsai et al (1992) observed a 26-day oscillation in SST over Equatorial western Indian Ocean (52°E to 60°E). The SST data used in the study was obtained from NOAA9 satellite. This fluctuation of SST at a period near 26-days is found to be antisymmetric about the equator and is trapped within the equatorial waveguide (equator $\pm$ 6°). Using linear wave theory, authors conclude that the equatorially trapped mixed Rossby-gravity (or Yanai) wave is responsible for the 26-day fluctuation observed in the SST data in this region.

The amplitude of this 26-day oscillation in the SST data is found during the summer monsoon season (July to September) with a maximum amplitude of 0.4°C in August of 1987 and 0.8°C in August of 1988. This observation is consistent with the temporal variability of the Yanai wave as seen in the modelling studies of Kindle and Thompson (1989) and Woodberry et al (1989). It is unlikely that this oscillation is a result of direct atmospheric forcing, since the dominant periods of atmospheric oscillation in the equatorial region is of 40 to 60 days (Mertz and Mysak, 1984; Madden and Julian, 1972). Marked SST variations on 40-60 day timescale have also been observed in the Bay of Bengal and the eastern equatorial Indian Ocean (Krishnamurti et al, 1988).

5. Surface Fluxes

Surface fluxes of heat, mositure and momentum provide the link in the interaction between the atmosphere and Ocean (Fig.5.1). For understanding the weather and climate in the tropics this relationship is of great significance. However, the major difficulty in the estimation of these fluxes arises due to inadequacy in the measurement of meteorological and oceanographic variables on which these fluxes are critically dependent.

There have been large number of studies pertaining to the estimation of climatologies of surface fluxes for the global oceans as well as the regional ocean basins. Hastenrath and Lamb (1979) published an atlas of the oceanic heat budget in the Indian Ocean. Similarly Oberhuber (1988) and Fu et al (1990) created atlases of the net heat over the global oceans. Rao et al (1991) compiled the historic surface marine meteorological measurements and prepared the atlas of the tropical Indian Ocean.

The atlases of the Hastenrath and Lamb (1979) and Oberhuber (1988) discuss the annual cycle of the various surface heat flux components. In the north and equatorial Indian Oceans, the two largest components are the incoming solar radiation and the latent heat. The net surface flux into the ocean is then the difference between these two terms (with contributions from the net longwave radiation and sensible heat). Solar radiation falls by 40-80 W m^{-2} from April to July in most part of the Northern Indian Ocean, whereas latent heat loss from the ocean increases from 80 W m^{-2} to 120-160 W m^{-2} (Hastenrath and Lamb, 1979). This is because with

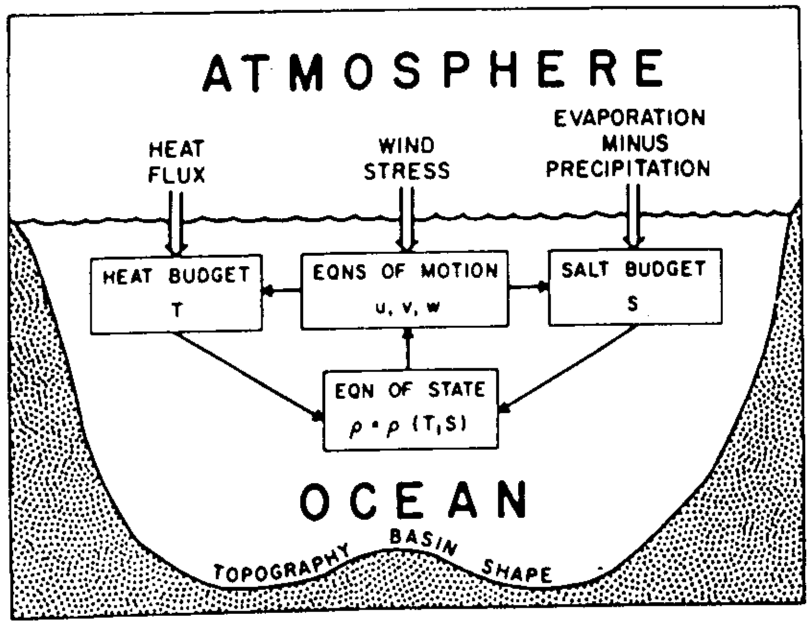
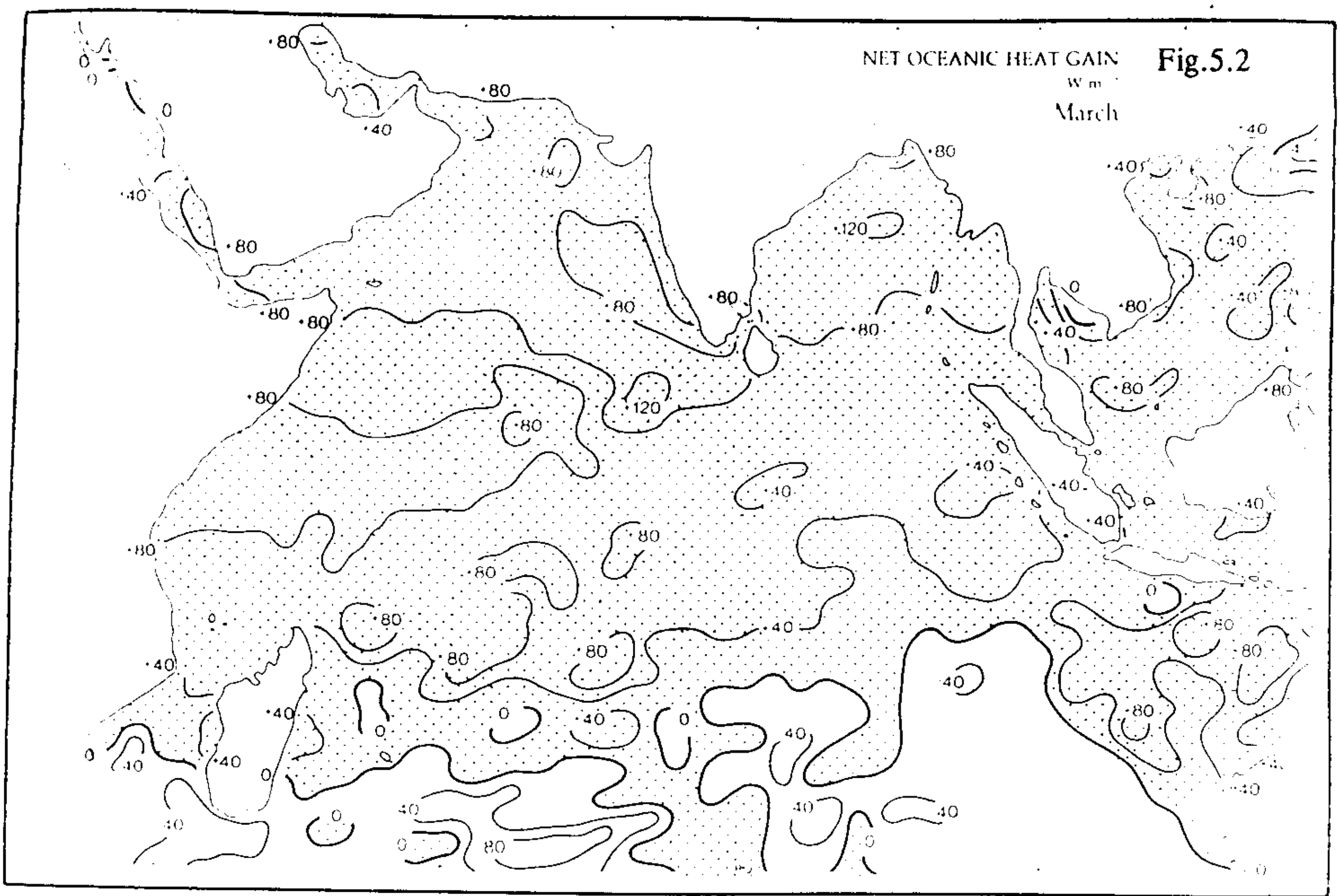
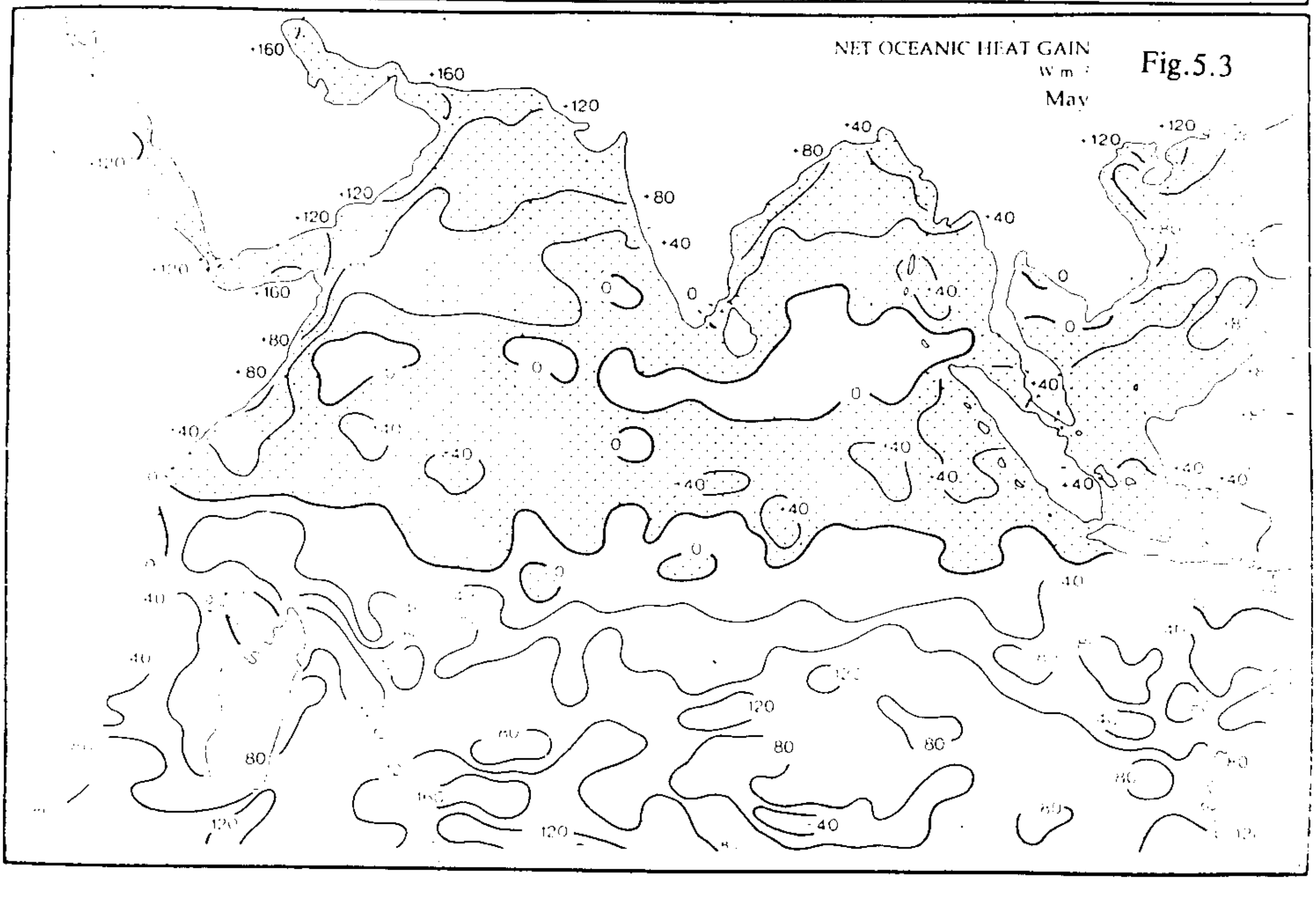


Fig.5.1 Air-Sea Exchange and Coupling

(Hastenrath and Lamb, 1979).





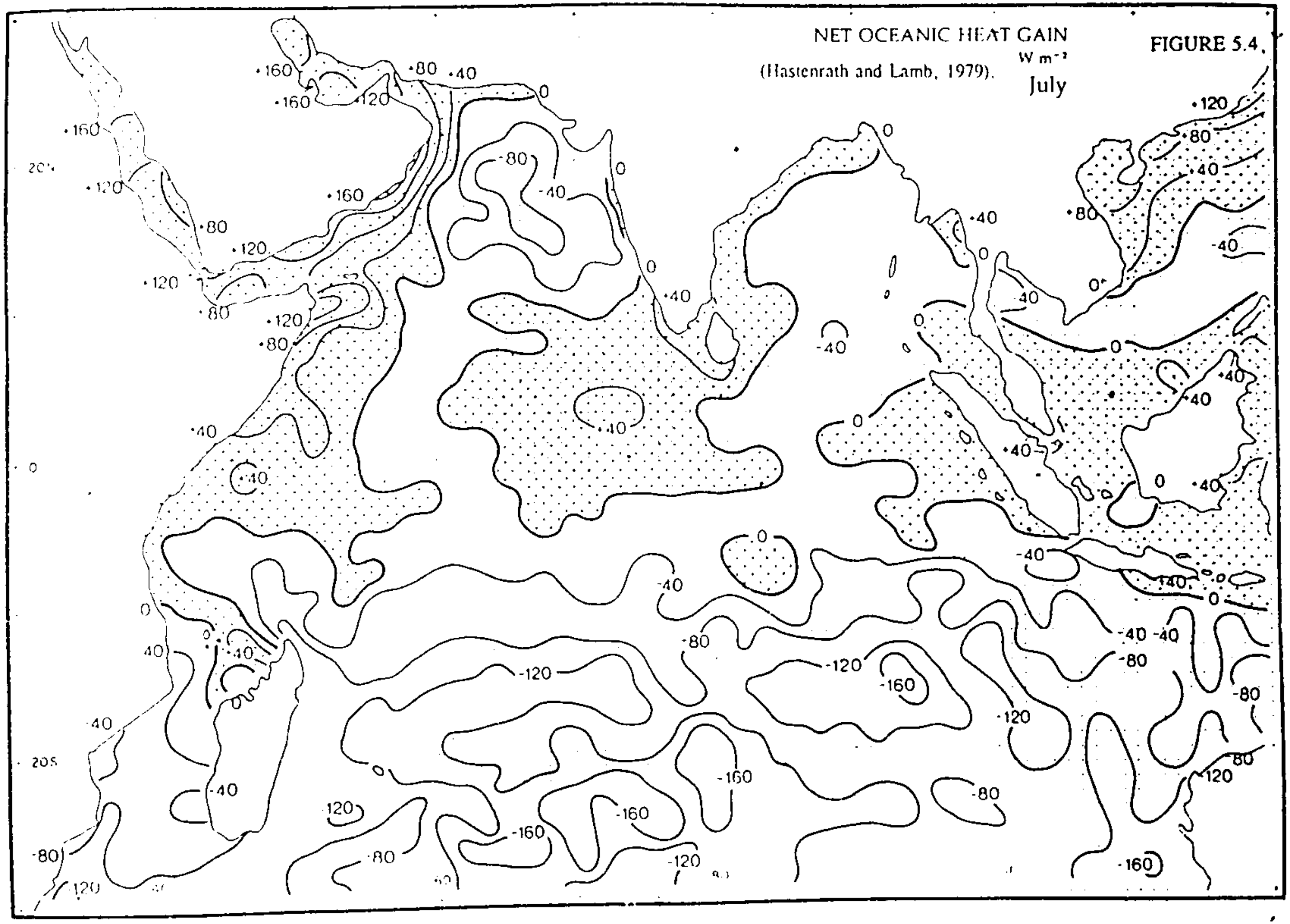


FIGURE 5.4.

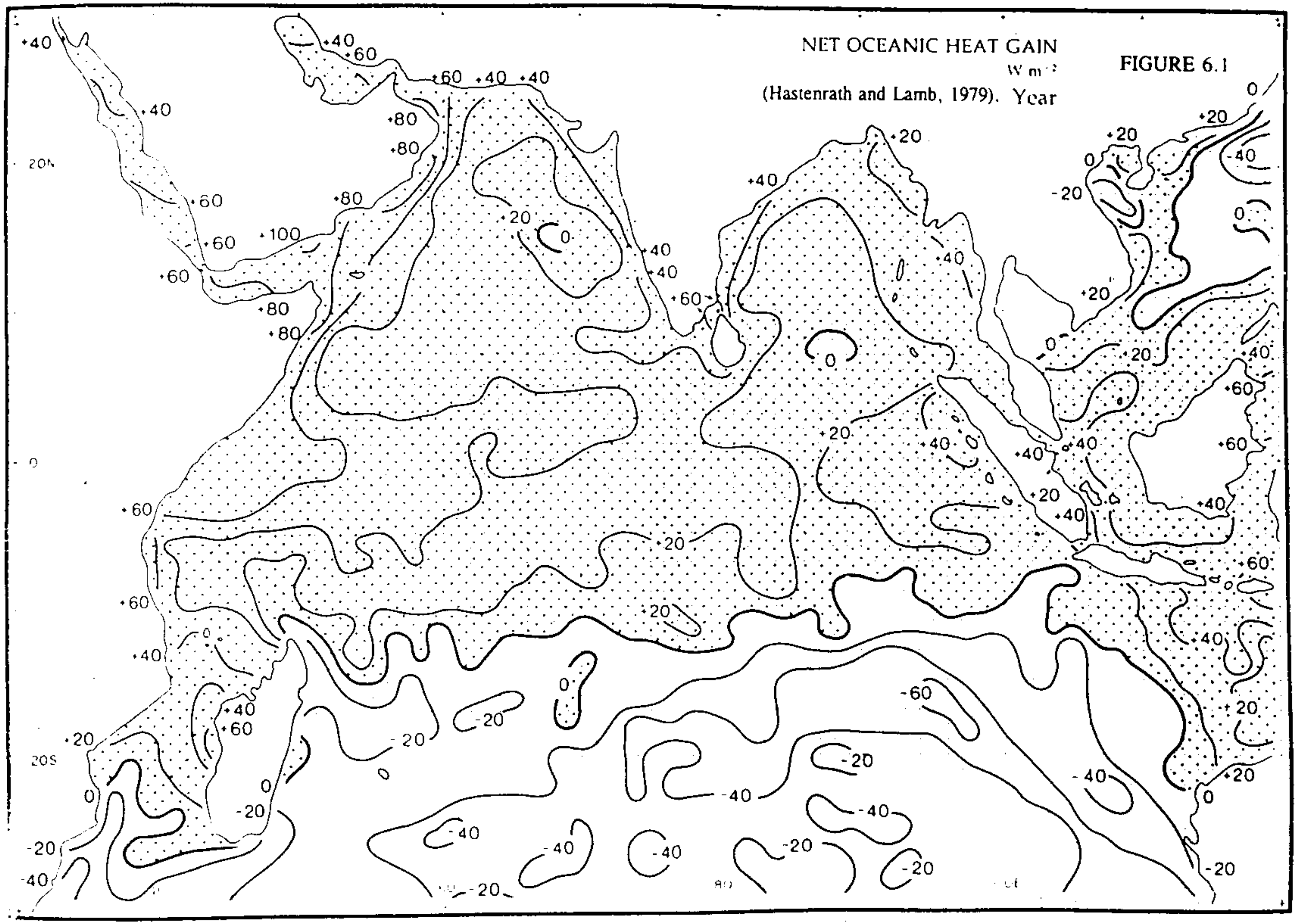
the onset of the summer monsoon, the entire region becomes windy and cloudy. A remarkable exception is the western Arabian Sea, where intense upwelling cools the sea surface as a result, the latent heat flux stays close to 80 W m^{-2}, further, the region remains cloud free and solar radiation remains nearly constant.

The highest net heat flux gain by the Indian Ocean occurs around March when the equatorial region gains 60-80 W m^{-2} and the northern Indian Ocean gains 80-120 W m^{-2} (Fig.5.2). Infact, along the coasts of Somalia and Arabia during northern hemisphere summer monsoon, the net oceanic heat gain is particularly large, as a result of the large net all wave radiation, small latent heat loss, and downward directed sensible heat flux (Hastenrath and Lamb, 1979). By may the monsoon winds have begun, and the resulting upwelling seems to be responsible for maxima in net heat flux along the western boundaries of the ocean - off Somalia, Arabia and the east coast of India (Fig.5.3). The net gain in the Bay of Bengal at about 40 W m^{-2}, is about half that in the Arabian sea where the skies are clear. At the peak of the summer monsoon, in July (Fig.5.4), there is intense evaporation in the Arabian Sea and in the Bay of Bengal away from the western boundary of the Ocean. Coupled with high cloudiness, this makes the net heat flux negative in these regions, i.e., the oceanic heat loss. From September untill March, almost the entire north Indian Ocean gains heat with exception of the northern Arabian Sea and Bay of Bengal, where cold, dry winter winds blowing off land cause strong latent heat loss.

6. Transport of Heat (Godfrey et al., 1995)

The transport of heat from one region and its release in another region for away and many years later is one of the important aspects of the role of the ocean in climate. This process is controlled by the horizontal advection of heat. We will briefly discuss the advection of heat, and the related heat fluxes through the Indian Ocean Surface. All the climatologies agree that the Indian Ocean north of 15°S gain heat in annual mean (Fig.6.1), although they do not agree with each other regarding the magnitude of the net heat flux. The integral of the net heat flux into the Indian Ocean over the area north of 15°S ranges between 0.5 - 1.0 x 10^{15} Watts, depending on the choice of climatologies.

If the temperature structure in the ocean is to have a steady annual cycle, therefore, this region of the Indian ocean has to expel the excess heat southward, since there are land boundaries to the north. The only way the ocean can export heat is through southward transport



NET OCEANIC HEAT GAIN
W m⁻²
(Hastenrath and Lamb, 1979). Year

FIGURE 6.1

of warm water along with northward transport of an almost equal volume of cooler water.

There is very little known about the way the Indian Ocean transports heat. The hydrographic measurements of Toole and Warren (1993) at 32°S show a deep overturning circulation with inflow of about 25 x 10^6 m^3 s^{-1} of cold water below 2000 m and a southward return flow of this volume plus 7 x 10 ^{6} m^3s^{-1} of Indonesian throughflow, mostly above 1000 m (Fig. 6.2). The heat transport associated with this circulation is adequate to account for the total heat received through the surface north of 15°S, which is about 0.5 to $1x10^{15}$ Watts. Recently Hastenrath and Greischar (1993) obtained a pattern of oceanic heat transport in the upper layer (Fig. 6.3).

The net northward heat transport into the upper layers of the Indian Ocean is highly seasonal. It is strongly southward (negative) in the northern summer months and northward in winter, giving net southward transport in the annual mean. This result can be explained with the help of Ekman transport theory.

Ekman transports in the Indian Ocean north of 15°S are everywhere to the south in the northern hemisphere summer and winter. Since the summer monsoon winds are stronger, the annual mean circulation has southward transport in the upper few tens of meters, and a shallow return flow. This shallow meridional circulation (Godfrey et al., 1995) can act to transport heat southward. However, this explanation is not fully complete as the intermediate and deep ocean may certainly be active in the advective transport of heat.

One seasonal time scales, the view that Ekman transports and their subsurface replacements play a major role in controlling heat flows in the northern Indian Ocean is strongly supported by two recent numerical model studies of the heat budget of this region (McCreary et al., 1993, McCreary, 1994, Wacongne, 1994 and Wacogne and Pacanowski, 1995). Neither of these models have produced the large northward flow of deep water into the Indian Ocean observed by Toole and Warren (1993). These models produce spatial patterns of surface heat flux which agree reasonably well with observed climatologies. These modelling studies have contributed greately to the understanding of dynamic and the thermodynamic processes in the Indian Ocean.

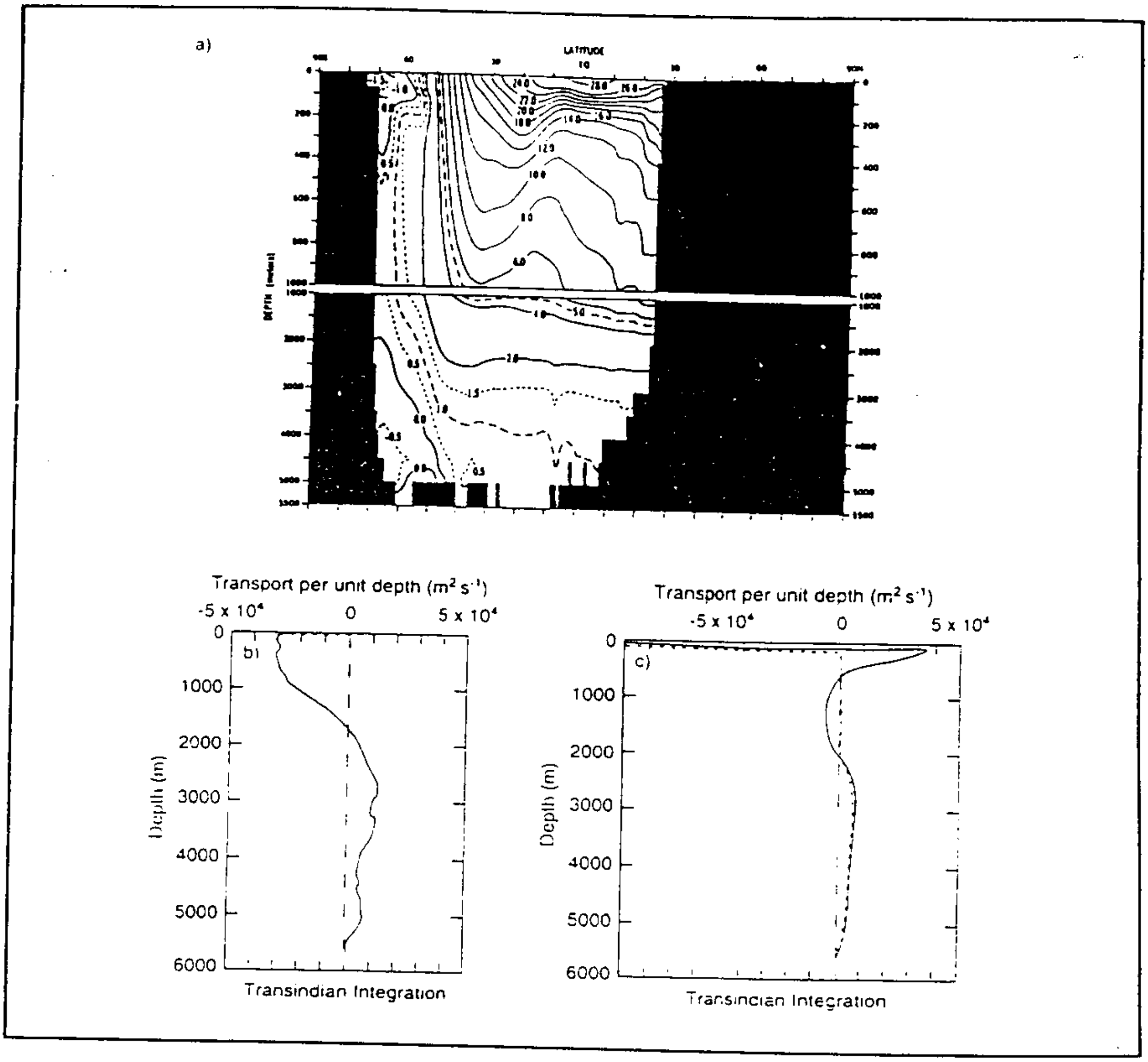


Fig.6.2 (a) Zonally-averaged temperature in the Indian Ocean, from Levitus (1982). (b) Meridional mass transport across 32°S as a function of depth (from Toole and Warren, 1993). Solid line: geostrophic flow: dashed line: total (Ekman plus geostrophic) flow, with Ekman flow distributed over the top 50 m. (c) Qualitative sketch of two possible forms for the net meridional mass transport across 10°S, as a function of depth. In both, 10 Sv of Ekman flow are distributed over top 100 m. For the solid line, the deep inflow is assumed all to leave at the depth of the oxygen minimum (about 800-1000 m). The Ekman transport is all replaced by flow within the thermocline, i.e., above 800 m. For the dashed line, the deep inflow rises to the surface to become part of the Ekman outflow. The real flow pattern is probably more like the solid line than the dashed line. Because of the very strong near-surface tropical stratification (a), the upper cell (above 600 m, in the solid line) transports more heat southward than the deep cell.

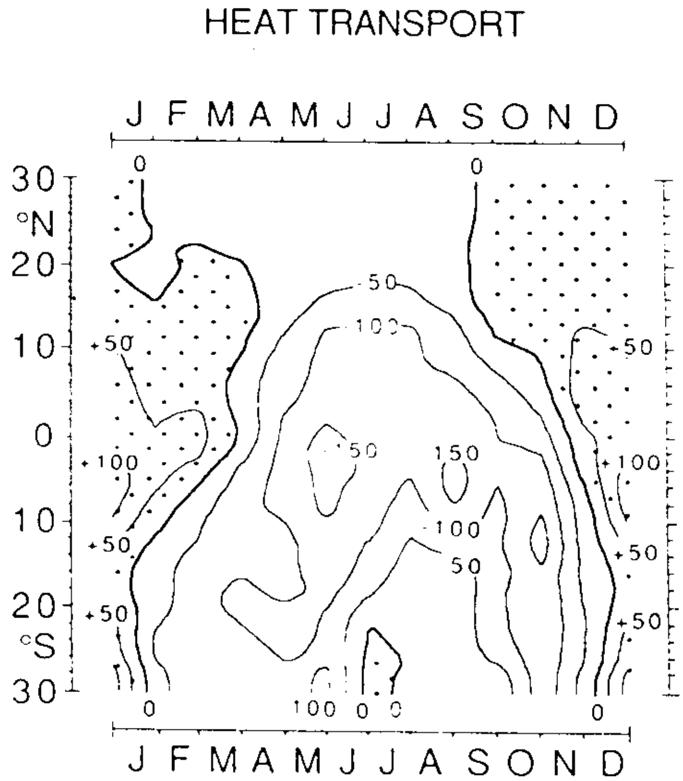


Fig.6.3 Oceanic heat transport in the upper layers of the Indian Ocean. Units are 10^3W. Stippled region corresponds to northward heat transport; southward transport is negative (Hastenrath and Greischar, 1993).

7. Variability of Fluxes and Relationships to Monsoon

Seasonal and spatial variations of fluxes of heat and freshwater across the sea surface are very large in the intertropical Indian Ocean, especially in its northern half. These variations are integral characteristics of the monsoon climate of the region.

Latitudinal and annual variation of residually determined net oceanic heat gain in the Indian ocean has been presented by Hastenrath and Lamb (1980). They found that the area mean SST in the southern hemisphere follows a 'traditional' seasonal cycle, with highest SST in southern autumn (March) and lowest in southern spring (September). However, the northern hemisphere revealed a much more complex behaviour, due to the monsoons. For southern hemisphere they found that at any time, the net heat flux into the ocean must be balanced by the sum of the rate of change of heat storage (Q_t) and the meridional advection of heat (Q_v), and Q_t must be zero on annual average, so Q_v must be small. On the other hand, their analysis for the northern Indian Ocean showed that Q_v and Q_t must be of similar magnitude, and that both must have strong and complex seasonal variations.

Later on Hsiung et al., (1989) and Hastenrath and Greischar (1993) preformed detailed analysis of Leviuts' (1982) global data and extensive hydrographic data set respectively to come to the conclusion that in the northern Indian ocean the difference between Q_t and the area integrated net surface heat flux are well defined and large. Their difference must be the net meridional heat transport Q_v in the Northern Indian Ocean.

There have been large number of studies on the estimation of heat budget in the tropical Indian Ocean and its possible relationship with summer monsoon activities. It would not be possible to describe the results of all of these, however the mention may be made to the earlier works of Düing and Leetma (1980); Golovastov (1980) and Nuzhdin (1982) and recent study by Gautier and Peterson (1994).

Latent and sensible heat fluxes can have major impacts on the large scale sea surface temperature anomaly fields in the mid-latitudes. A few studies have tested the relationship between the fluxes and the SST field over broad scales. These include observational studies (Frankignoul and Reynolds, 1983; Cayan, 1992 a,b,c,) and an ocean general circulation model (Haney, 1985). These studies indicated that anomalous heating by the surface fluxes were a major component in producing monthly thermal anomalies in the Northern Hemisphere oceans. Majority of these studies focussed on the climatological and/or mean annual estimates of the surface fluxes. In one of the recent studies Jones et al., (1995) objectively determined series of monthly fields of surface fluxes and surface parameters in the Indian Ocean basin for a 30 year period (January, 1960 - December, 1989) based on ship reports (COADS data). Importance of this study is that first time fields of surface fluxes and the surface variable which determine these fluxes have been jointly analyzed. It would be of some interest to see the results of this analysis.

Jones et al (1995) indicate significant annual, interannual and decadal variability in surface fluxes over the Arabian Sea, Bay of Bengal and the southern trade wind belt. Climatology of the latent heat flux created from 30 years of monthly objective analysis results exhibit two maxima (70 E, 20 S and 100 E, 15 S) in southern trade winds during July and August. Another maxima is located in the Arabian sea 65E, 10N (Fig.7.1, bottom). Fluxes in both areas are sustained by the strong summer monsoon winds. During the northeast monsoon, latent heat flux exhibits maxima again in the Arabian sea(55E, 10N and 70E, 20N) and in the southern trade wind just west of Australia (Fig.7.1, top)

Sensible heat flux varies little in comparison to latent heat flux. Values are generally positive except where coastal upwelling occurs in the Arabian Sea 60E, 10N (Fig. 7.2, bottom). Authors indicate that on a monthly time scale latent heat flux was always positive, i.e., the ocean cooled the atmosphere warmed. Latent heat flux variability is more correlated to the wind variability, however SST exerts a secondary influence which was seen in regions of coastal upwelling especially near Saudi Arabia in June, July and August. Areas of high wind speed such as the southern trade winds and the Arabian Sea are characterised by larger values of latent heat flux. Latent heat in these two regions were somewhat correlated.

EOF analysis of the wind stress fields of this analysis shows strong biennial variability, but little interannual variability. In contrast a similar EOF analysis of the latent heat flux shows very little annual or biennial variability, and instead indicates significant interannual variability. Authors also found the evidence of decadal-scale variability in the latent heat and wind fields for many locations, but especially in the Arabian Sea and southern trade wind regions.

Mcreary and Kundu (1989) studied the seasonal variability of heat flux in the Arabian sea by using a numerical model. His results show that in July, the heat flux is into the ocean near the western ocean and it closely resembles the structure of SST field (Fig.7.3). Net annual heat gain computed by the model is in close agreement with the observations (Fig.7.4). Authors

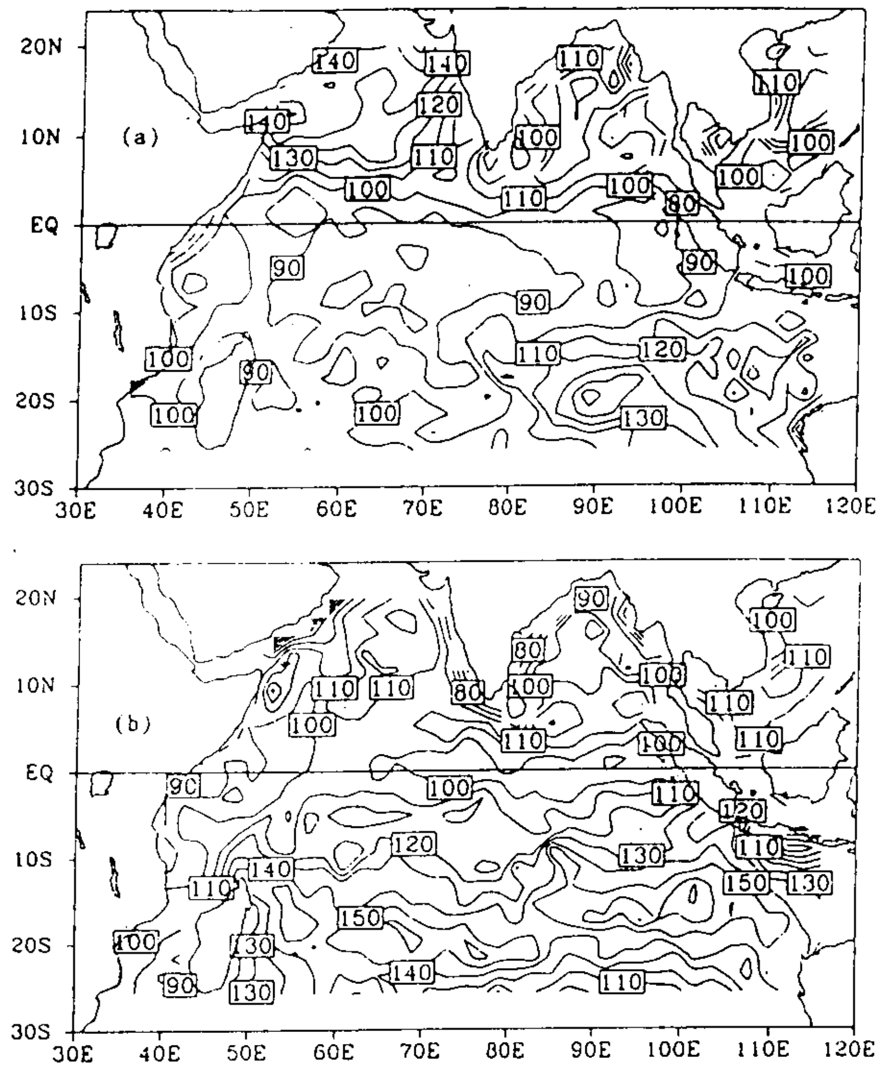


Fig.7.1

Climatology of the latent heat flux created from 30 years of monthly objective analysis results. (a) January and (b) August are shown as representative of the winter and summer monsoon periods. Units are in Wm^{-2} and contour intervals are 10 Wm^{-2} (Jones et al., 1995).

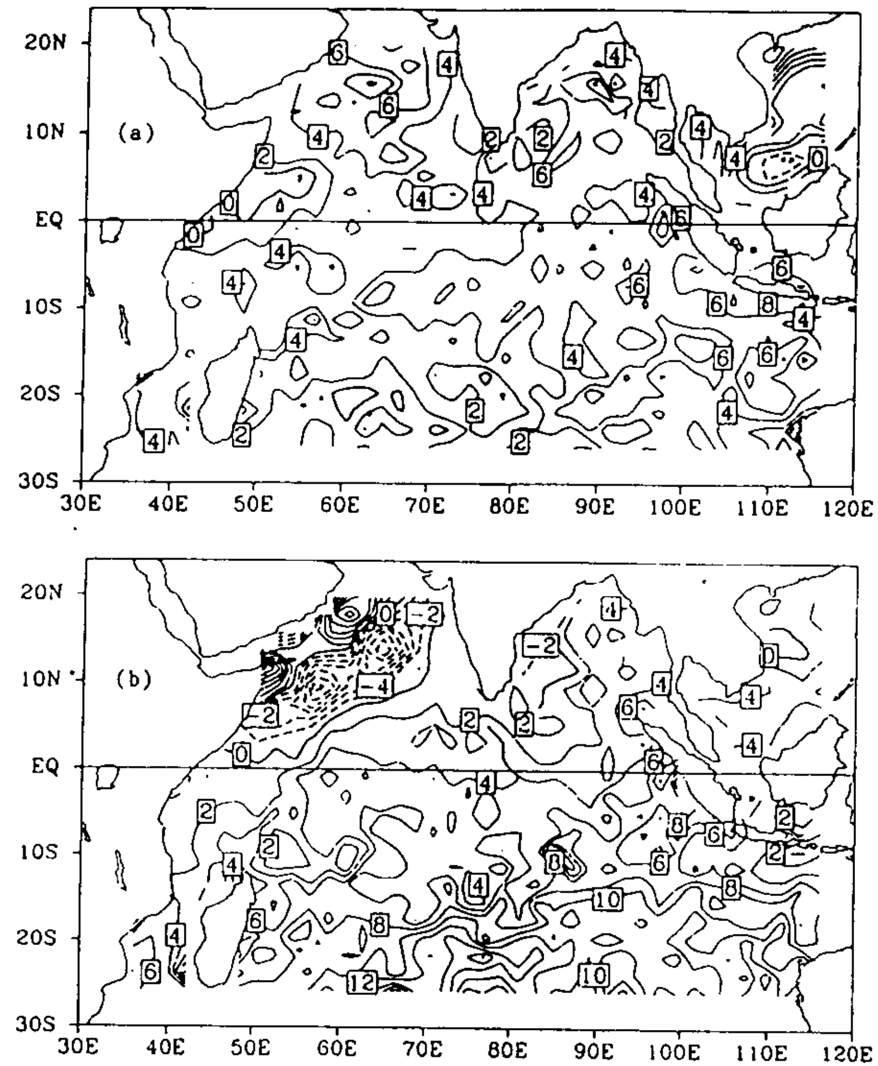


Fig.7.2

Climatology of the sensible heat flux (a) January and (b) Augus[t] representative of the winter and summer monsoon periods. Units are in [Wm^{-2}] and contour intervals are 2 Wm^{-2}. The thicker line is the zero contour (Jo[nes et] al., 1995).

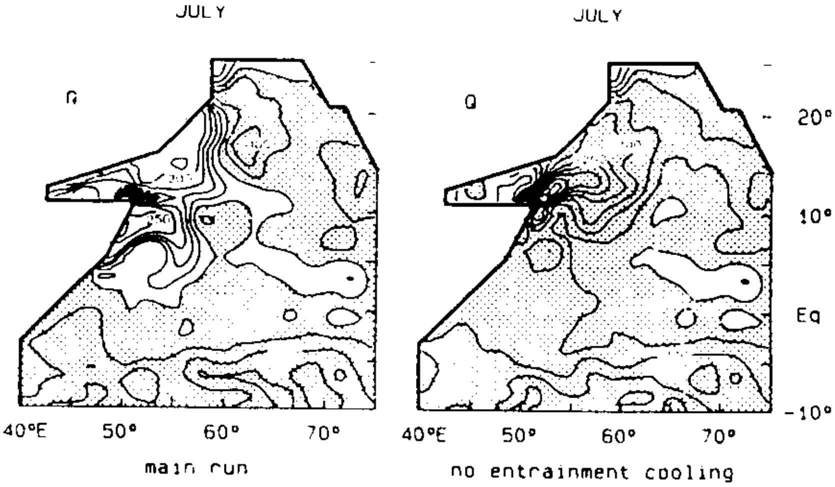


Fig.7.3 Model surface heat flux into the ocean during July (left) for the main run and (right) when there is no cooling by entrainment of subsurface water. The contour interval is 50 W/m^2, and negative regions are shaded. In the main run, Q is positive near the western boundary and closely resembles the SST field. In the right panel the heat flux pattern near the western boundary is completely different, with large negative values there (McCreary & Kundu, 1989).

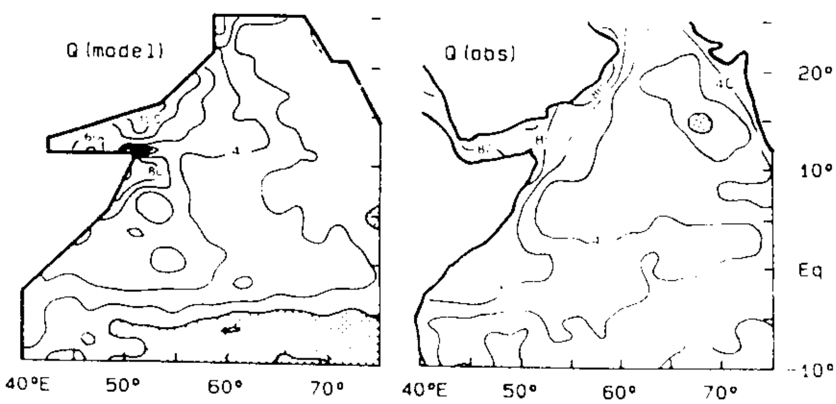


Fig.7.4 Annual average surface heat flux into the ocean (left) for the main run and (right) as estimated from climatological data by Hastenrath and Lamb [1979]. The contour interval is 20 W/m^2, and negative regions are shaded. The heat flux is almost everywhere into the ocean (positive) and increases markedly toward the western boundary (McCreary and Kundu, 1989).

demonstrated that the model heat gain exists only because entrainment, primarily along the Somali and Arabia coasts, provides a source of cold water to the upper layer ; the resulting cooler SST limits the magnitude of the latent heat flux out of the ocean by lowering the saturation specific humidity. The net surface heat input in the central and equatorial regions of the basin is due to cooler water being advected toward the south in the central Arabian Sea and toward the east along the equator.

8. Mixed Layer Depths

Knowledge of the climatological state of the ocean is required to address adequately the possible role of SST anomalies on atmospheric circulation. Of particular importance is the mean state of the upper thermal structure, i.e., the depth of the mixed layer. Molinari et al (1986) derived the mean monthly charts of the mixed layer depth (defined as the first depth at which the temperature is 0.5°C less than the SST) and thermocline depth (defined as the depth of the 20°C isotherm) in the western Indian Ocean from the data collected between 1948-1981. Monthly maps of mixed-layer depth for the Indian Ocean have been prepared by Rao et al (1989) (Fig.4.9) and Hastenrath and Greischar (1989) by utilising a larger data sets. It is seen from these maps that throughout the year, the mixed layer tends to be relatively shallow in equational region, and deep in the Southern Indian Ocean. Seasonal variations are particularly pronounced in the Northern Indian Ocean.

The observed seasonal pattern of mixed layer depth (MLD) (Fig. 4.9) shows a band of minimum MLD along 10°S. MLD is shallowest throughout the Northern Indian Ocean during the month of May, except for a region of deep MLD in the eastern equatorial Indian Ocean. The latter reflects the accumulation of water in this region by the strong eastward surface jets along the equator, driven by the equatorial westerly winds of May and November (Wyrtki, 1973). The May minimum in MLD throughout the Northern Indian Ocean is to be expected, since it follows a period of strong net heat flux into the ocean and weak winds. By July, after two months of strong monsoon winds, the MLD pattern changes drastically. Deep maxima in MLD occur in the Central Arabian Sea and Bay of Bengal. Coastal MLDs do not deepen nearly so much, from the Somali coast to Bangladesh. The maximum MLDs in July occur near (10°N, 60°E), to the southeast of the strongest SW monsoon winds, where Ekman surface fluxes converge. Very deep mixed layers have developed south of 15°S, under the influence of strong net surface heat lose (Fig 5.4), and strong wind mixing and Ekman convergence. Rao (1986) and Rao and Matheu (1990) have given the details of these processes in the Arabian Sea.

These conditions persist through September. By November the SSTs have warmed somewhat under the light winds of October, except in the northern Bay of Bengal and Arabian Sea where cold, dry NE Monsoon winds start to cool them, and MLD shallows in most part of the Northern Indian Ocean. By January the NE Monsoon winds have apparently caused substantial cooling and MLD deepening in the northern most Indian Ocean. South of 15°S net heat flux into the Ocean and reduced wind speeds have resulted in a shallowing of the mixed layer and surface warming.

9. Seasonal and Interannual Variability of other parameters of Indian Ocean

Although the knowledge of the interannual variability of SST and associated fluxes of heat and MLD are important to understand how ocean affects the overlying atmosphere, in order to understand the mechanism of SST variability one also needs information on the variations of other parameters, such as, winds, thermal structure of subsurface waters, salinity and currents. We will briefly discuss the variability of some of these parameters based on various studies.

9.1 Studies of Meteorological Fields

The interannual variability is well resolved by the analysed fields of the 23 years (1954-1976) prepared, by Cadet and Diehl (1984), from ship reports which include surface winds, SST, and cloudiness. Authors have also studied the relationship of these fields with the activity of the southwest and northeast monsoons. The analysis indicates that the end of the 1960's and the begining of the 1970's were characterized by below normal trade winds in the southern hemisphere during all the months, weaker cross-equational flow, particularly significantly along the coast of East Africa during the summer, and weaker winds over the western Arabian Sea. Associated with this weakening of the wind was an increase of SST and air temperature due to reduced cooling and turbulent mixing of the ocean by evaporation. This large time scale fluctuation can also be associated with a similar long-term tendency of summer rainfall over India. Fields of correlation coefficients between summer rainfall over India and anomaly of fields show that weaker trade winds during all months, weaker Somali Jet during the summer and higher SST over the Ocean were associated with weaker rainfall over India.

Reverdin et al (1986) investigated the interannual variability of surface observations in the equatorial Indian Ocean from 23 years of ship reports (1954-1976). In October and November, during certain years, the monthly analyses show strong wind anomalies in the eastern and central equatorial Indian Ocean. Simultaneously, cloud cover anomalies of opposite signs are observed in the eastern Indian Ocean and in the Central Indian Ocean. A linear analysis of the heat induced circulation carried out by the authors suggests that the surface wind anomalies are forced by anomalies of rainfall. During the years with anomalously little cloudiness in the eastern Indian Ocean, SST lower than normal are found in the eastern Indian Ocean. Authors suggest that coupled air-sea dynamics must be taken into account to understand the interannual variability.

Miles et al (1993) used a variational method to produce five-day averaged wind fields from SEASAT scattermeter and ship data with $1^{\circ} \times 1^{\circ}$ resolution for the 1978 summer monsoon period over the Indian Ocean. This method is used first for 30 day results over the Indian Ocean. Ten five-day results are then calculated, each based on the previously determined longer term results. The results reveal five-day time scale information about the intramonsoonal variability of the region in six areas of interest including the monsoon and trade wind regions.

Breidenbach (1990) carried out EOF analysis of FSU wind field (Legler et al., 1989) to show that interannual variations of Indian Ocean winds are not very large. Recently Jones et al (1995) used a newly developed variational technique to objectively analyse Indian Ocean monthly mean, 2° resolution surface fields of winds, temperature, humidity, as well as wind stress, sensible and latent heat fluxes, for the period 1960-1989. EOF analyses of the wind stress fields from this analysis shows strong biennial variability, but little interannual variability.

9.2 Variability in the Somali Current

The most spectacular event in the annual march of oceanic response to monsoon winds is the reversal of the Somali current along the east coast of Kenya and Somalia. Somali Current is the strongest time varying current in any of the oceans and the speed of the surface current may reach as high as 3.7 m S^{-1} (Düing et al., 1980). Observational data to investigate the interannual variability in the Somali Current are very sparse. The major source of information about the structure and variability of the Somali Current have come from 1960s International Indian Ocean Expedition (IIOE), INDEX (Indian Ocean Experiment) which was the oceanographic component of MONEX (Monsoon Experiment), carried out during the summer monsoon period of 1979. A large number of observational studies have appeared in the August 1980 issue of 'Science' and December 1982 issue of Jour. Phys. Oceanography. An extensive review of these papers has been given in Knox (1987).

Beginning with the pioneering efforts of Cox (1970,1976,1979), numerous modelling efforts have sought to explain the observed flows in the tropical Indian Ocean, with particular attention given to the semi-annual reversals in the Somali Current along the east coast of Africa (e.g. Hurlburt and Thompson, 1976; Lin and Hurlburt, 1981; Luther and O'Brien, 1985; Luther et al., 1985; McCreary and Kundu, 1988; see Luther (1987) or Knox (1987) for a review). While the Somali Current is similar to mid-latitude western boundary currents in some respects, it is unique in many others. The most striking feature of this current is its reversal with the changing monsoon winds (Schott, 1983; Knox, 1987). During boreal summer, the boundary current flows toward the north and northeast from the coast of Mozambique (11°S) to the island of Socotra at 12°N, driven by the southwest monsoon winds. A "two gyre" system is often observed in this current (Swallow and Fieux, 1982), with a southern gyre that straddles the equator, flowing offshore at 3-4°N, and a northern gyre, called the great whirl, between 5°N and 9°N. Wedge shaped areas of cold, upwelled water are found along the coast to the north of these gyres. Late in the summer, the southern gyre and its cold wedge migrate rapidly northward and coalesce with the great whirl (Brown et al., 1980; Evans and Brown, 1981; Swallow et al., 1983). This appears to be the case in most years; however, there is evidence of some years when the two gyre system does not form (Swallow and Fieux, 1982). The summer Somali Current to the north of 2-3°S gradually beaks up during the fall transition period, and is replaced by the south westward winter Somali Current with the onset of the northeast monsoon in December. The boundary current to the south of 3°S flows the north throughout the year and is called the East African Coastal Current (EACC). During the summer, the EACC feeds the northward Somali Current; during winter, it meets the southwestward Somali Current and both flow offshore into the South Equatorial Counter Current (SECC). Leetmaa (1972, 1973) concludes that the early reversal of the Somali Current south of the equator with the onset of the southwest monsoon is due to both local wind forcing and an intertidal overshoot (or switching action) in the EACC. Woodberry et al. (1989) show that this region of the EACC is a tropical analog to a mid-latitude western boundary current recirculation region that is strongly modified by the presence of the equatorial wave guide and by the seasonal reversals in the wind. It closes the circulation in a tropical Sverdrup-like gyre in the southern hemisphere, consisting of the eastward SECC that meanders between the equator and 8°S and the westward South Equatorial Current (SEC) between 10 and 20°S.

As mentioned earlier, the observational data to investigate the interannual variability in the Somali Current are very sparse. Swallow and Fieux (1982) analyse historical ship drift data and SST to provide the only long term observational data set on this interannual variability. They find that the two gyre system is present in most years where sufficient data are available, although there are years where the southern gyre appears to be absent (Fig. 9.1 (a) and (b)). These data are too sparse to delineate changes from year to year in the complicated evolution of the Somali Current during the southwest monsoon.

Luther and O'Brien (1985) developed a nonlinear reduced gravity model for the Indian Ocean that is driven by observed winds. Integrations using various versions of the model have used a monthly mean climatology of ship's winds (Luther and O'Brien, 1985), a monthly mean of the FGGE 1000 mbar winds for 1979 (Luther et al., 1985), objectively analyzed ships' winds for 1985 (Simmons et al., 1988), and the Hellerman and Rosenstein (1983) monthly mean climatological winds (Woodberry et al., 1989) as forcing fields.

Luther and O'Brien (1989) contrast the variability inherent in the model physics to interannual variability that is driven by variability in the winds. From 10 years of model integration driven by the climatological monthly mean of the Cadet and Diehl (1984) winds, Luther and O'Brien (1989) compute the mean and standard deviation of the model upper layer thickness (H) field for the 16th of each month of the year. The standard deviation fields are thus a measure of the inherent variability of the model physics, since the wind cycle is exactly repeating from year to year. Over most of the basin, on a particular day of the year, H varies by less than 1m from one year to the next; for instance, the H field for August 16, year 7, is within 1m of the value it had on August 16, year 8, over most of the basin, indicating that the model solution is a nonlinear periodic response to the seasonal winds. Exceptions to this occur only in limited regions such as in the intense shear zone around the great whirl during the summer monsoon, but this is confined to very small scale motions around the periphery of the great whirl; the center of the great whirl is found in the same position with the same intensity from one year to the next. The highest values of H standard deviation in these regions are in the range of 12-15m, indicating a more chaotic nature in the model response. In the region around the great whirl, the chaotic nature of the flow stems from horizontal shearing (barotropic) instabilities; still, the solution over much of this region is repeating to a high degree.

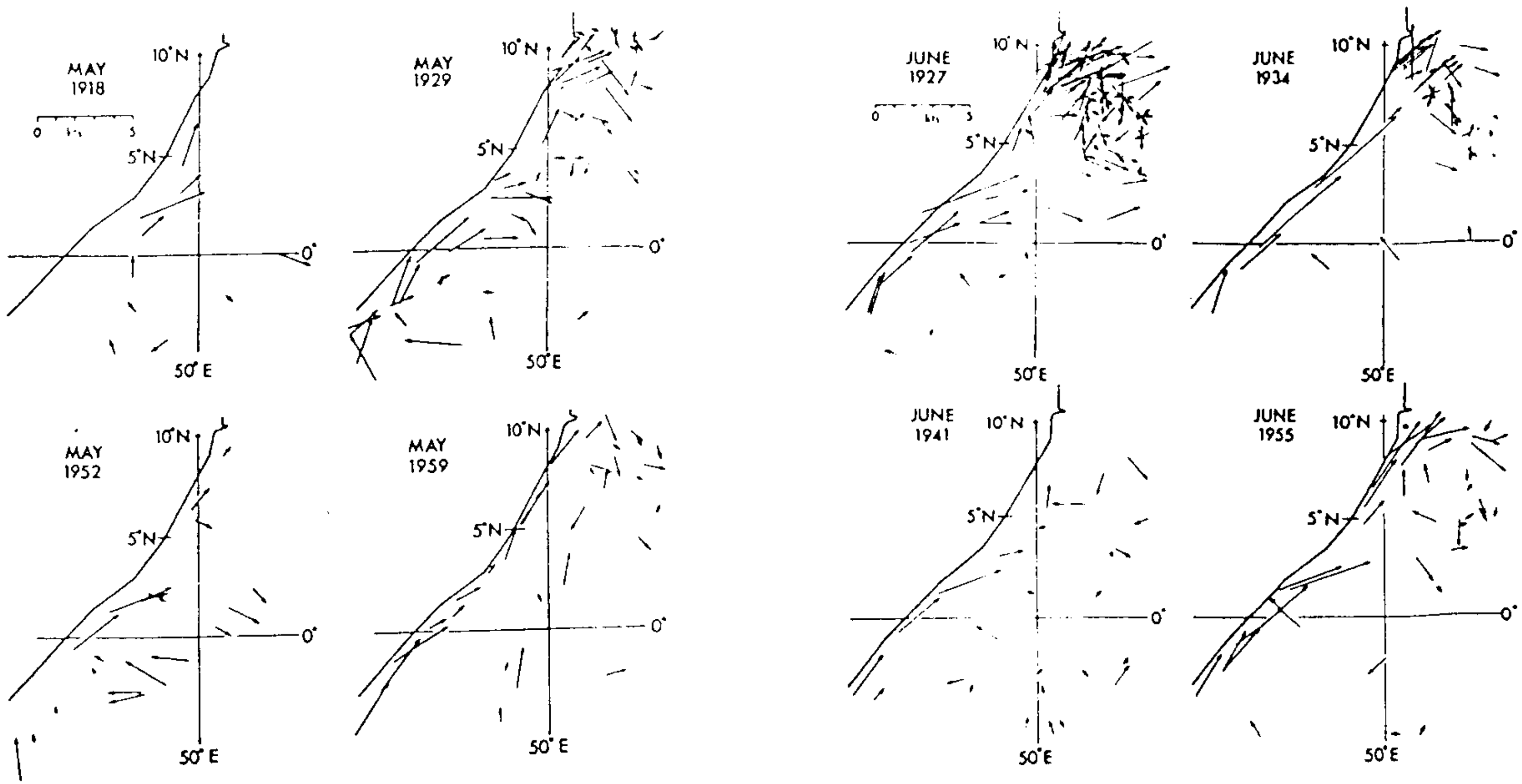


Fig.9.1 May and June Surface Currents for selected years (Swallow and Fieux, 1982).

The situation is much different when the interannual winds are used to drive the model. From the same standard deviation calculation from the model driven by 23 years of interannually varying winds, it is shown that the H standard deviation is everywhere an order of magnitude larger than for the climatological case, except in the highly nonlinear shear zones around the great whirl, indicating that interannual variability in the model response solely due to variability in the winds, rather than to inherent variability contained in the model physics.

Dube et al. (1990) investigate the relationship between the interannual variability in the model fields from this 23 year model integration and variability in Indian monsoon rainfall. They find that the period 1954 to 1967 is a period during which the southern gyre of the Somali Current is generally stronger, the cross-equatorial winds are stronger and the upper layer is thinner (the thermocline is shallower) in the central Arabian Sea, implying that sea surface temperatures are lower (Figs. 9.2(a), 9.2(b), 9.3(a), and 9.4). This period is identified by Cadet and Diehl (1984) as one of generally higher Indian monsoon rainfall, stronger summer cross equatorial air flow, and lower Arabian Sea sea surface temperature (SST). In contrast, the period 1968 to 1974 is one of generally lower Indian monsoon rainfall, weaker cross equatorial air flow, and higher Arabian Sea SST, and also is a period during which the model southern gyre is weaker and the model upper layer is thicker (deeper thermocline) across the central Arabian Sea implying higher SST (Figs. 9.2(c), 9.2(d), 9.3(b) and 9.4). Cadet and Diehl (1984) find that the summers of 1956 and 1972 represent two extremes of this interannual variability, with 1956 being a wet year and 1972 being an extremely dry year. A pronounced quasi-biennial oscillation (QBO) is found by Dube et al. (1990) in the model fields and in the wind stress curl fields during the period of higher rainfall but was absent during the period of lower rainfall. The significance of the presence of absence or this QBO signal is not yet clear, but is certainly suggestive.

Detailed investigation on the interannual variability in the two gyre system of the Somali Current during the southwest monsoon is done by Luther (1990) using a numerical model forced by observed winds for the period 1954-1976. He found that great whirl is always present during the southwest monsoon but the southern gyre exhibit large year to year varibility. Author suggest that the interannual variability in two gyre system of Somali Current can be categorised into 4 dominant patterns. The mechanism responsible for each of these pattern has been discussed. Another important finding is that this model integration shows the evidence that the remote forcing by equatorial Rossby waves play a role in the collapse of the two gyre system.

Szekielda (1988) attempted to estimate the response of the Somali upwelling to the Southwest Monsoon with the help of data sets of GOSST COMP/MCSST, HRIR and selected VHRR data during July 1966. This study may also be of interest to the readers.

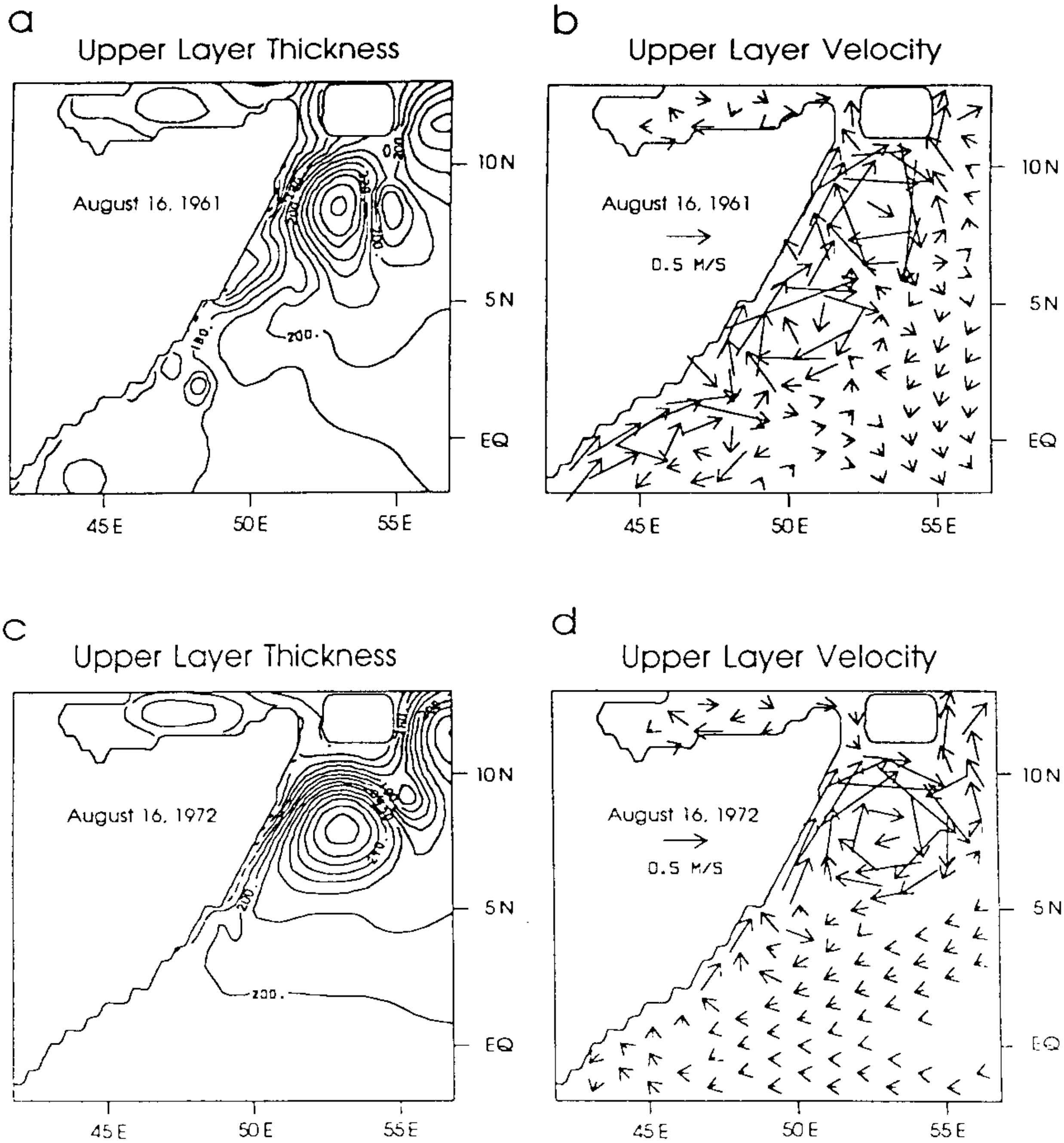


Fig.9.2 (a) Upper layer thickness and (b) circulation pattern in model domain during mid-August of 1961. Arrows indicate upper layer velocity. Contour interval for ULT is 10m. (c-d) Same as (a-b) for mid-August 1972 (Dube et al.,1990)

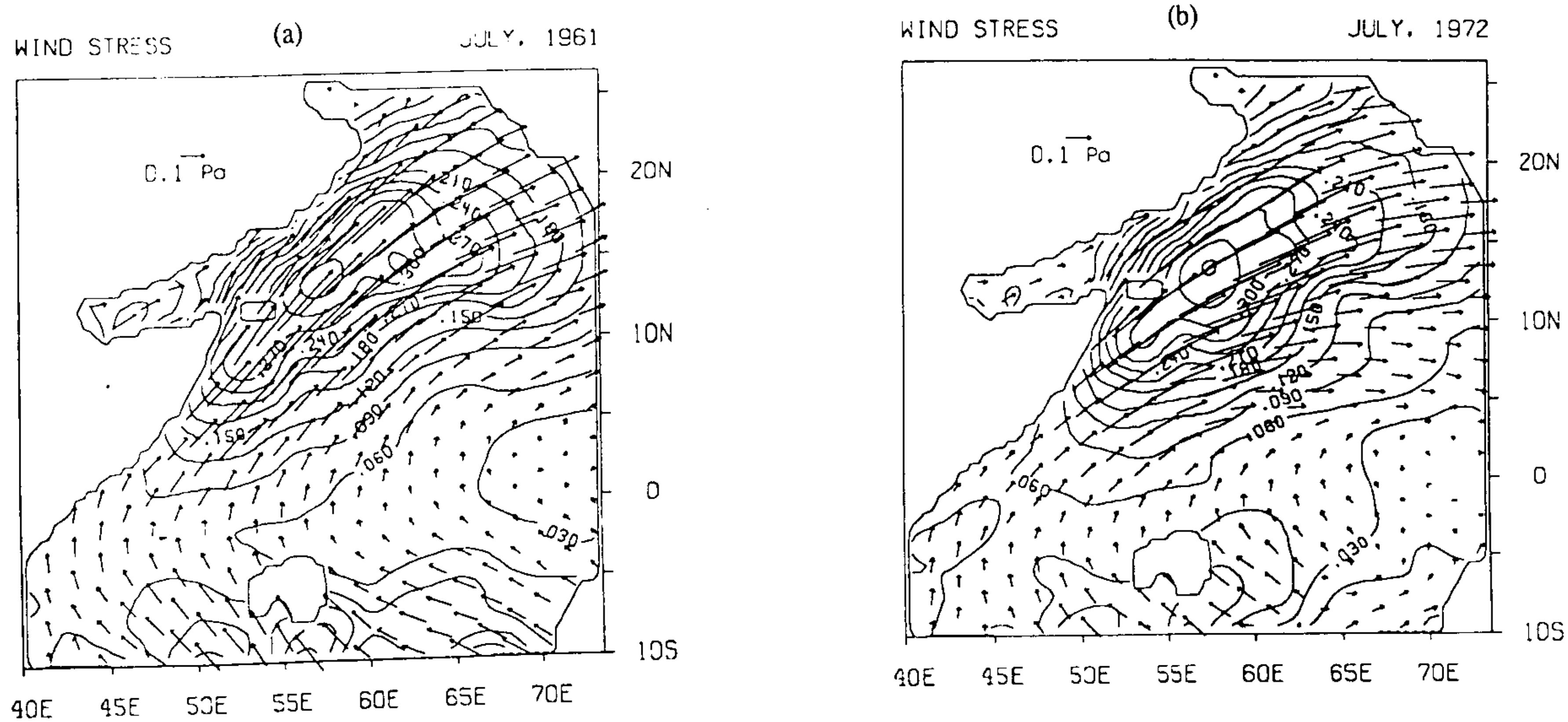


Fig.9.3 Wind stress from Cadet and Diehl (1984) used force the model. (a) July 1961 (a typical wet year) (b) July 1972 (a typical dry year). Arrows indicate wind stress while contours indicate magnitude. Contour interval is 0.03 Pa (Dube et al., 1990).

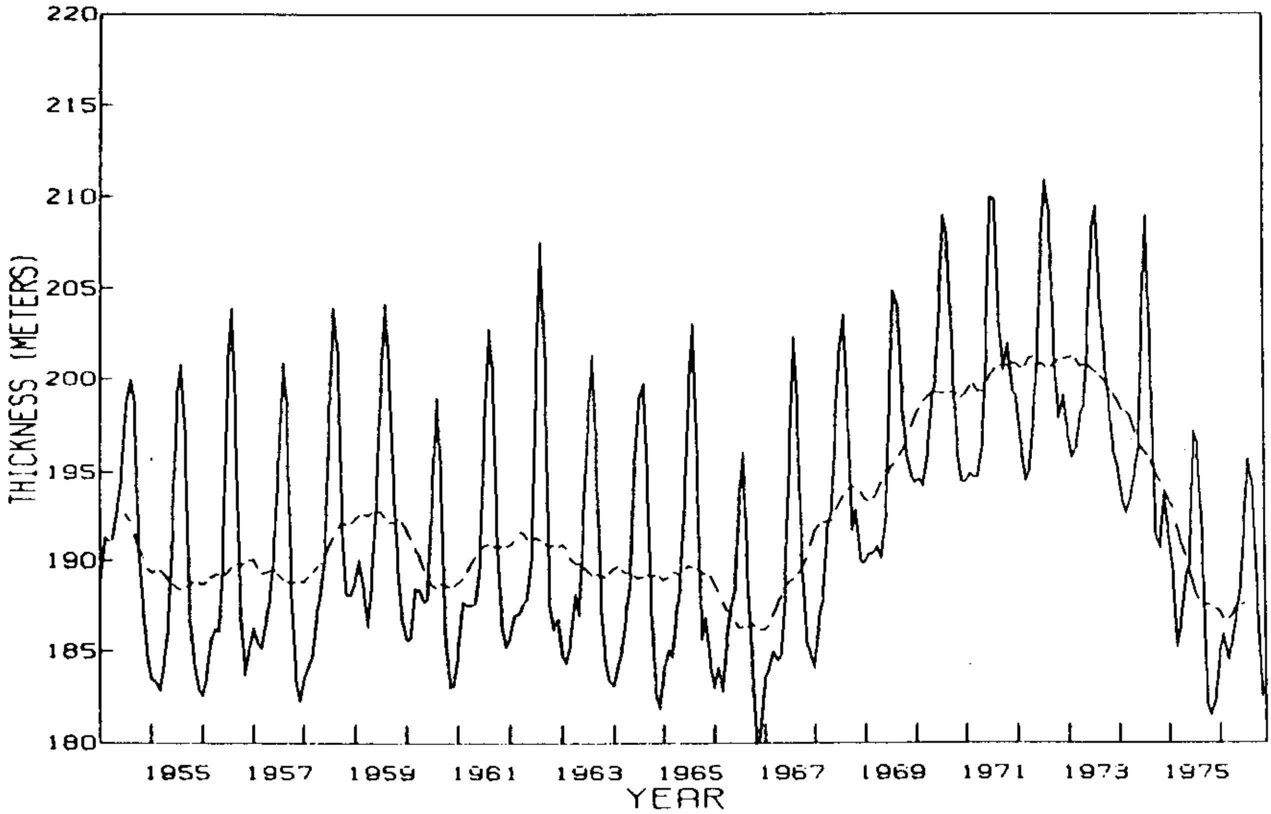


Fig.9.4 Monthly mean upper layer thickness in the central Arabian Sea. Superimposed is the 12-month running mean (dashed lines) (Dube et al., 1990)

In one of the very interesting modelling study, Jensen (1991) described the seasonal variability of the undercurrents in the Somali Current system. He found very good agreement between the observed under currents and the simulations in the model. Some of the interesting suggestions made by Jensen (1991) are :

- barotropic instability is likely to cause the generation of the Great Whirl in early June,
- equatorial onshore flow below the thermocline in June is associated with the disappearance of the undercurrent below the Somali Current,
- the early return of the southward under current in the fall is caused by baroclinic instability of the Great Whirl followed by a reversal to eastward undercurrents along the equator,
- remote winds probably control the intermediate depth flow along the Somali coast acting through equatorial waves and Rossby waves generated along the west coast of India.

10. Equatorial Variability

Although the most remarkable response of the Indian Ocean due to seasonal monsoon winds is the annual reversal of Somali current, strong seasonal signal is also found in the vicnity of the equator. There is considerable energy in the equatorial oceans at periods 2 weeks to 2 months, especially in the meridional velocity component. All these frequencies, the available waves involving meridional motion on the equator are Yanai waves (mixed Rossby waves) and a few Rossby waves, and the observed variability is often interpreted in terms of these linear waves. Recent observations from the western Indian Ocean provide a well documented example of such variability (Tsai et al, 1992).

Indeed there is a significant variation of the equatorial wind fields with strong annual and semiannual periodicities (Knox, 1976).

10.1 Observational Studies

The equatorial transient response to wind change was first observed in the upper Indian Ocean by Wyrtki (1973). He found an intense eastward equatorial jet during May and October, when the westerly winds intensified. A semiannual period in the zonal wind stress component and surface currents at Gan was reported by Knox (1976). He presented a 2-year time series of the current above the main thermocline at 0.5°S, 73°E and found that strong eastward currents were present from April to June and from mid october through December, with reversals occuring in January, March-April, June and September. The fall jet was the more intense of the two. Based on measurements from 21 drifting buoys, Reverdin et al. (1983) found the jet to be most intense in May and November with eastward monthly mean velocities of 55 cm/s and 80 cm/s, respectively. Rao et al. (1989) analysed current data from Cutler and Swallow (1984), and found a very strong semi-annual signal in zonal curents along the equator. They found a maximum in the semiannnual harmonic near 70°E, accounting for more than 75% of the total variance, and the annual harmonic accounted for less than 25% of the variance east of 55°E. Recently, Molinari et al. (1990) presented detailed maps of the seasonal tropical currents based on 142 surface buoy tranjectories, finding that the eastward equatorial jet was most intense in the central Indian Ocean during May-June and during November.

Luyten and Swallow (1976) found an eastward upper layer flow and a westward flow underneath during May and June 1976. Below the thermocline a strong semiannual current was also observed in the western Indian Ocean by Luyten and Roemmich (1982). From moored instruments at depths of 200, 500, and 750 m they found a strong eastward jet during February and March and during August and September, whereas a westward jet was observed during May and June and during November and December. This is nearly 180° out of phse with the observed surface jets. They also found westward and upward propagation of phase and attributed the currents below the thermocline mainly to first-horizontal-mode, vertically propagating Rossby waves with some influence from Kelvin waves.

10.2 Modeling Studies

The equatorial jet has been investigated theoretically by several authors, (e.g., Yoshida, 1959; O'Brien and Hurlburt, 1974; Gill, 1975; Cane, 1980). Cane and Moore (1981) demonstrated theoretically the existence of a standing equatorial mode composed of long Kelvin and Rossby waves. They pointed out the possibility of resonance, when the ocean is forced by wind stress with a period equal to 4 times the time it takes an equatorial wave to cross the basin.

Gent et al., (1983) presented an analytical linear model, which when forced by the large semiannual zonal component of the Hellerman and Rosenstein (1983) wind stress, produced semiannual reversals of the equatorial currents. It was demosntrated that reflections at lateral walls were essential for a realistic result. They questioned whether realistic coast lines and the presence of the Somali Current would alter the reflections and perhaps decrease the strong semiannual response.

Reduced-gravity models with a single active layer and realistic geometry have been used to model various regions of the Indian Ocean, in particular the Arabian Sea. Luther and O'Brien (1985) and Luther et al (1985) used the NOAA Global Marine Sums climatological wind stress to successfully model the seasonal cycle of the Somali Current. The study by Kindle and Thompson (1989) focussed on the monthly and bimonthly oscillations in the western equatorial region, which were also found in the model solutions of Woodberry et al. (1989). Kindle and Thompson (1989) forced their 1.5-layer model with the original Hellerman and Rosenstein wind stress and found periods in the range 20-30 days and wavelengths from 800 km to 1400 km, which they identified as mixed planetary-gravity waves, often referred to as Yanai waves. They suggested that the source of Yanai waves is instability of the Southern Gyre associated with the Somali Current, while instability waves with periods 40-60 days were attributed to barotropic instability of the currents south of the equator. In recent studies of the Somali current, McCreary and Kundu (1988, 1989) used a 2.5-layer reduced-gravity model of the Arabian Sea, with the addition of a mixed layer, which includes a prognostic temperature equation. However, due to the limited region, the equatorial current system is not fully developed. For instance, the strong eastward equatorial jet observed in the spring is missing in their model. Moore and McCreary (1990) investigated the vertical structure of the Yanai waves and short Rossby waves forced by monthly and bimonthly winds. An important result of their study was that energy is radiated downward, with upward phase propagation. They also demosntrated that the waves may be generated at the western boundary by broad-scale wind fields.

Jensen (1993) used a 3.5-layer isoppycnal ocean model for the Indian Ocean (north of 25°S) to study the equatorial variability and resonance. I will briefly describe some of the interesting results of this study. Model simulates a well defined westward North Equatorial Current during the boreal winter and a seasonal cycle of the transport in the South Equatorial Current (Wyrtki, 1971; Schott, 1983). The semiannual equatorial jets with reversing flow during the same months are also simulated. This feature is in close agreement with earlier reported observations (Molinari et al., 1990). Model also gives good results for the deep flow. For example the equatorial surface jets in the model are associated with intense flow in the opposite direction below the thermocline. This is in aggrement with the work of Luyten and Roemmich (1982), who have observed that the deep semiannual currents change direction during the year.

The highest variability in the model solution is found at the equator near the western boundary. Yanai waves with periods of 20-30 days are responsible for high variability in the meridional velocity field, as in the simulations by Kindle and Thompson (1989) and Woodberry et al., (1989). The Yanai waves dominate eastward and downward propagation of energy along the equator in the western Indian Ocean during the late summer and during the NE monsoon, and seem to be the main contributors to internal variability.

Valenti et al (1995) used a numerical model of the North Indian Ocean to show a propagating pattern connecting the equatorial wave guide with higher latitudes in both the hemispheres.

11. Remote Forcing of the Circulation In the Indian Ocean

The major role that remote forcing plays has stimulated a number of studies in the Pacific and Atlantic Oceans, studies such as the interannual occurrence of warming SST in the eastern Pacific (Wlyrtki, 1975) and the seasonal upwelling in the Gulf of Guinea (O'Brien et al, 1978). In the Pacific during ENSO downwelling Kelvin waves radiate from the western Pacific, deepen the pycnocline on the eastern Pacific, and hence weaken the upwelling of cold water. A similar process may also occur in the Indian Ocean affecting the eastern Indian Ocean circulation.

Lighthill (1969) pointed out that low-latitude oceans respond to atmospheric forcing much faster than do oceans in mid-latitudes and proposed that the onset of the Somali current is the remote response to the changes of the monsoon wind in the interior of the Arabian Sea. Although Leetma (1972, 1973) presented evidence that the onset of the Somali current is closely coupled with the local wind stress, remote forcing is still considered to be an important mechanism of intensifying the Somali Current (Knox, 1987). Anderson and Moore (1979) and Anderson (1981) suggests that Somali Current is remotely forced by the Southern hemispheric trade winds.

Like Somali current, the currents in the Bay of Bengal reverse their directions semiannually (Legeckis, 1987; Molinari et al., 1990). On the basis of the data obtained from the polar orbiting NOAA Satellite, Legeckis (1987) showed that the circulation in the Bay of Bengal during the winter is anticyclonic, with colder water to the northeast and warmer water to the southwest. By using a multiple layer ocean model, driven by climatological monthly mean winds, Potemra et al., (1991) have reproduced circulation patterns similar to those presented by

Legeckis (1987). Their results also display the cyclonic circulation during the summer and two gyre transitional circulation during the spring and autumn. It is clear from their results that the reversal of the currents is closely related to the monsoon system.

As a result of wind forcing, several types of equatorial trapped waves are generated and propagate along the equator away from the forcing region. These trapped waves act as signal carriers, transferring the input wind-forcing energy to unforced area. The capability to transmit energy rapidly away from the forcing region by the equatorial trapped waves and the nondispersive characters of the two most important waves, Kelvin waves and long Rossby waves, are the keys to understanding the physics of the remote forcing.

For a large-scale and low-frequency forcing associated with monsoon winds, only long Rossby waves and Kelvin waves are generated. These long Rossby waves propagate westward and finally strike the eastern African coast. Kelvin waves travel eastward with a faster speed. Upon reflection from the eastern boundary, some of the Kelvin wave's energy will be reflected back to the equatorial waveguide as long Rossby waves, while the remaining energy forms two coastally trapped Kelvin waves traveling poleward (Moore, 1968). As a coastally trapped Kelvin wave propagates poleward, it excites long Rossby waves with the same frequency.

Yu et al., (1991) pointed out that remotely excited Kelvin waves in the equatorial waveguide of tropical Indian Ocean play an important role in determining the variability of the circulation in the Bay of Bengal. Using a simple reduced gravity oceanic model, they tested the hypothesis that the seasonal variability in the Bay of Bengal is caused by the long Rossby waves which are radiated westward from the coastally trapped Kelvin waves with an annual frequency, the turning latitude of the Rossby waves is further north than the northern limit of the Bay of Bengal (about 25 N); therefore Rossby wave radiation occurs all along the eastern boundary of the Bay of Bengal.

In one of the recent studies Dube et al. (1995) proposed a dynamical explaination of the semiannual reversal of the circulation in the Bay of Bengal. They used a model based on the conservation of absolute vorticity to a one and half layer (reduced gravity) ocean to illustrate how variability in the circulation pattern of the Bay of Bengal can be induced by Rossby waves excited by the remotely forced Kelvin waves. Their computations indicate that remote forcing contributes significantly to the local variability of the Bay of Bengal circulation.

Dube et al. (1995) on the basis of theoretical analysis suggest that the local rate of change of circulation is generated by the sum of the terms representing local curl of the wind stress and a southern boundary term representing the northward propagation of Kelvin waves along the eastern meridional boundary formed by Sumatra, the Andaman - Nicobar Islands, and Malaysia. A comparison of the magnitude of southern boundary term and local wind forcing term show that the former is significantly larger than the later during most of the monsoon and postmonsoon period (Fig.11.1).

A number of example of remotely forced circulations are revealed by the numerical investigation of the dynamical processes in the Indian Ocean (McCreary et al. 1993). Their model runs identity following instances of remotely forced circulations:

- During the spring a northeastward counter current flows against the prevailing winds along the Somali coast north of 4°N, and from October through February a southwestward Somali Undercurrent is present from the tip of Somalia to 3°N; both of these flows result in part from forcing during the previous SW monsoon.

- From March through May there is another southwestward Somali Undercurrent south of 7°N, generated primarily by the propagation of a Rossby wave from the west coast of India.

- The Currents along the west coast of India are either strongly influenced or dominated by remote forcing from the Bay of Bengal throughout the year.

- A northeastward flow is well established along the east coast of India in March, long before the onset of SW Monsoon; it is remotely forces either by upwelling - favourable, alongshore winds elsewhere within the Bay of Bengal or by negative wind curl in the western Bay, Remote forcing from the equator, a process emphasised by Potemra et al. (1991), Yu et al. (1992) and Dube et al (1995), does influence the flow field of the McCreary et al. (1993), primarily along the eastern boundary of the Bay of Bengal during the spring.

- The Agulhas Current is strengthend considerably in a solution that includes throughflow from the Pacific Ocean.

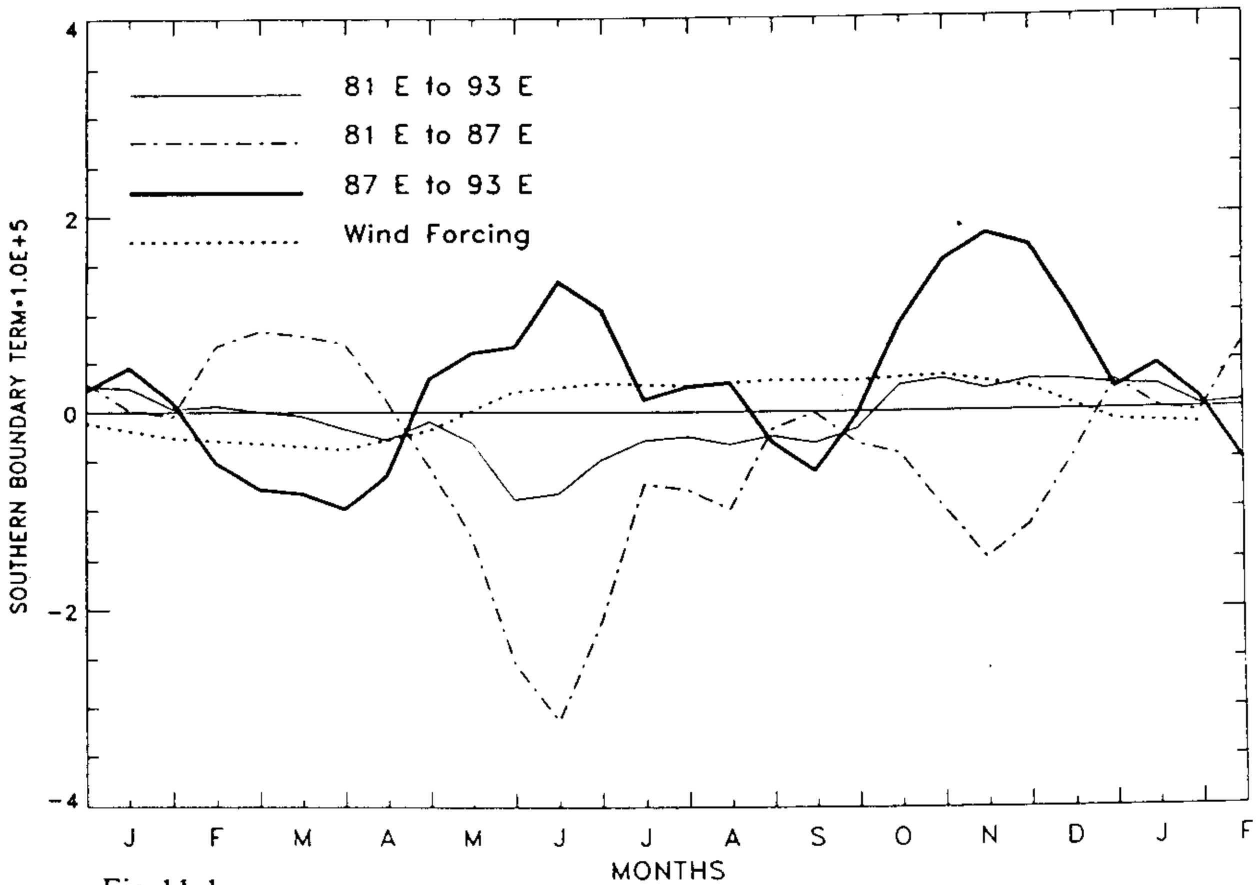


Fig.11.1 Time Integrated Boundary Term at 8 N & Wind Forcing

12. Indo-Pacific Throughflow

The material contained in this section has basically been taken from the works of Verschell et al (1995) and Godfrey et al (1995). Several other studies which have been consulted and the references of these are made below in this section.

The connection between the Pacific and Indian Ocean in the Indonesian Archipelago (PACIO region) is composed of three major passages (Figure 12.1). These passages are Lombok Strait to the south of Java, Savu Strait to the North of Timor Island, and Timor Strait between Timor Island and Australia. The minor passage are both narrow and shallow, and no significant transport is assumed to pass through them. There have been few long term observational studies in the region and most determinations of net transport and regional dynamics have been made from indirect observations, theoretical studies, and compueter modelling (Table 12.1).

Godfrey and Golding (1981) performed an early study that investigated the dynamic importance of the PACIO region. Their calculations of depth-integrated steric height in the Indian Ocean determined that net transport from the Pacific to the Indian Ocean through the PACIO region could have a large effect on the Sverdrup circulation pattern in the southern Indian Ocean, which would be otherwise be more like that of other tropical oceans. They hypothesized that throughflow of Pacific water is also an important factor in the Leeuwin current, the east Australian flow, and the South Equatorial Current, and that the throughflow may be responsible for the observed lack of upwelling (or lack of cold, nutrient rich water) along the Western Australian coast. Godfrey (1989) calculated annual mean depth-integrated steric height and stream function and determined closing off the PACIO region could cause a $6^{o}C$ drop in the upper 500m over the entire South Indian Ocean, with smaller changes elsewhere.

Other studies have also looked at the effect of the transport through the PACIO region on Indian Ocean circulation. Kundu and Mcreary (1986) hypothesized that southward bending of the throughflow may contribute to the Leeuwinn current, although it would not be dominant mechanism. However, Godfrey and Weaver (1991) determined that warm west Pacific water can flow through the Indonesian Archipelago creating basin-scale buoyancy-driven circulation in the Indian Ocean, and that Pacific heating and winds may be the main driving force behind the Leeuwin current. Hirst and Gidfrey (1993) examined the effects of the throughflow on surface

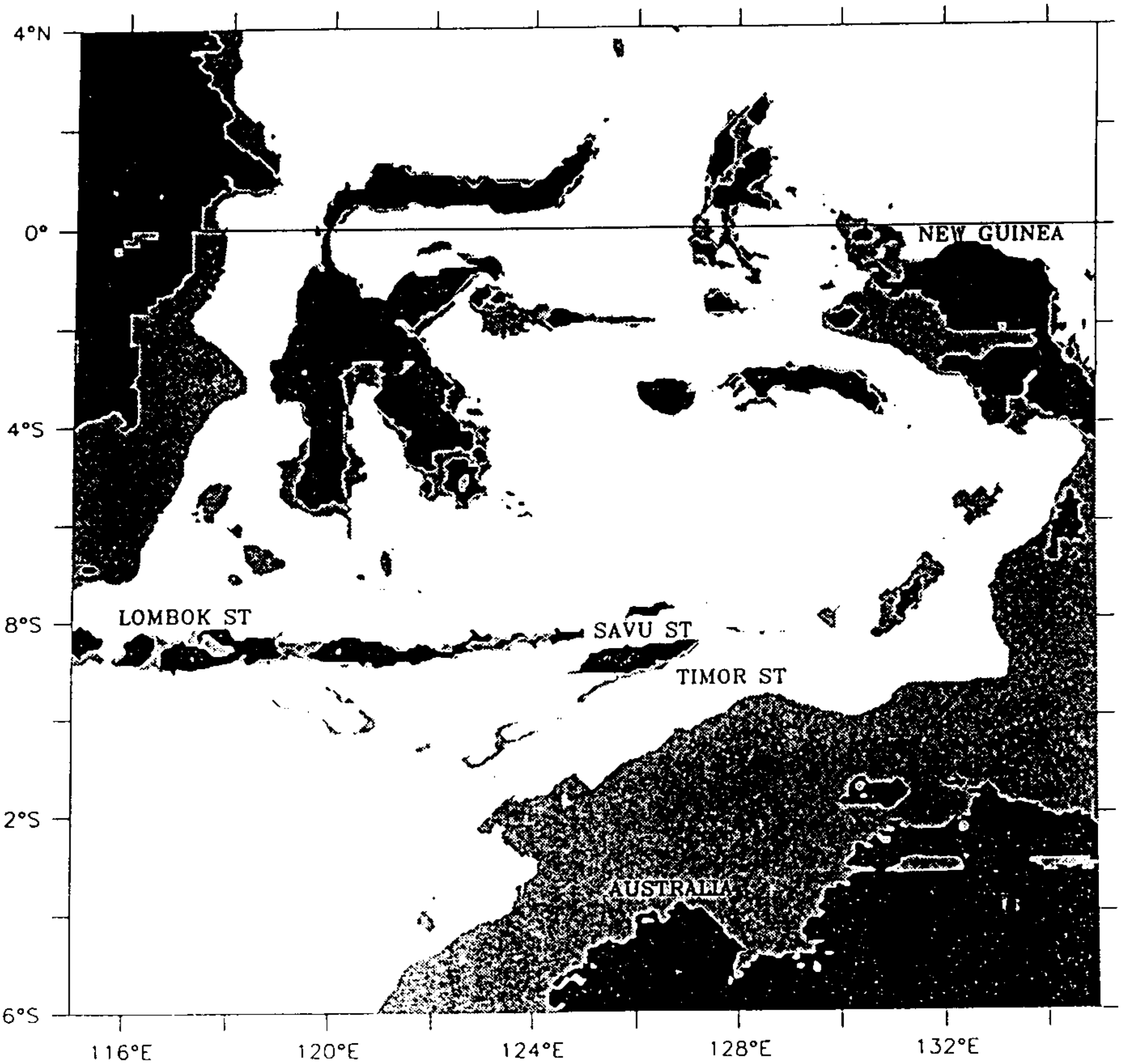


Fig.12.1 PACIO Region and the Three Major Passages (Lombok, Savu and Timor Straits). Light Grey area is ocean depth < 200 m. Overlayed is the ten year average current pattern as determined by the numerical model (Verschell et al., 1995).

TABLE 12.1

PACIO Throughflow

Study	Estimate in Sv	Notes
Godfrey and Godling [1981]	10.0	From depth-integrated storic height
Piola and Gordon [1984]	14.0	Freshwater box model
Fine [1985]	5.0	Tritium box model
Kindel et al. [1987]	4.5	Global reduced-gravity numerical model
Murray and Arief [1988]	1.7	Current meters in Lombook St. only
Godfrey [1988]	16.0	Global Sverdrup model
Clarke [1991]	5.0	Residence times from radiocarbon data
Broecker [1991]	6.1	Geostrophic estimate from sea level data
Inoue and Welsh [1993]	9.8	Two-basin reduced-gravity numerical model
Wajsowicz [1993a]	14.0	Friction reduces godfrey [1989] estimate
Cresswell et al. [1993]	7.0	ADCP and Dropsonde in Oct'87: Mar'88
Fieux et al. [1994]	18.6	Geostrophic estimate from Aug'89 cruisc
Molcard et al. [1994]	4.5	Current meters in Savu and Timor only
Meyers et al. [1994.	5.2	Geostrophic estimate from XBT data

Sv ≡ $10^6 m^3 s^{-1}$

heat flux and sea surface temperature using a coarse resolution GCM. They found that, in general, throughflow warms the Indian and cools the Pacific, but any large changes in surface temperatures or heat flux were confined to certain "well-defined" regions in these basins. Wajsowicz (1994a) described numerical model results that suggest PACIO transport is an element for a good simulation of the eastern Indian Ocean.

The justification behind several additional studies of the PACIO throughflow were attempts to balance the mass, heat and salt fluxes between the Indian and Pacific Oceans. Wunsch et al., (1983) concluded as a result of inverse calculations on sections across the South Pacifc that there was negligible net transport through the PACIO region, implying that PACIO throughflow was unimportant in mass, heat and salt flux balances. Other researchers have concluded that the PACIO region, is important in balancing these fluxes. Piola and Gordon (1984); Toole (1987); Kindle et al., (1989) each maintain that an accurate determination of the mass, heat and salt fluxes in the Pacific and Indian Ocean requires adequate knowledge of the magnitude and variability of the PACIO throughflow.

The PACIO region has been suggested as the major route for return flow of thermocline water from the Pacific to the Atlantic Ocean (Gordon and Piola, 1983; Gordon, 1985, 1986). Broecker (1991) believes that the PACIO region accounts for only about one quarter of this return flow, but he thinks that this flow is still an important component of what he calls "The Great Ocean Conveyor", the major pathway of global circulation.

Attention has also been turned to this region because of it's possible connection to the ElNino-Southern Oscillation (ENSO). Bye and Gordon (1982) claim that basin-wide exchange between the Pacific and Indian Oceans through the Indonesian Archipelago is an important factor in the oceans' role in the Southern oscillation. Nicholls (1984) determined that anomalously warm sea surface temperature in the Indonesian Sea precede ENSO warm events by a few months, and White et al., (1985) found a large increase in the heat content of the warm pool in the tropical Western Pacific before the 1982-83 ENSO that might be important to the onset of this event. Kindle et al., (1987, 1989) found anomalously high Pacific to Indian Ocean transports during a modeled ENSO warm event. They believe that the extent to which the throughflow modifies the regional circulation may be an important factor in the maintenance and variability of the warm pool of the tropical Western Pacific, and therefore may have a significant effect on the interannual variability of the coupled ocean-atmosphere system in that region. Wajsowicz (1994b), using Godfrey's Island Rule, showed that depth-integrated PACIO transport decreased by 4Sv (Sv $\equiv 10^6 m^3 s^{-1}$) near the onset of ENSO warm event, and increased a similar amount near the onset of a cold event. Clarke and Liu (1994) also made a similar connection to PACIO transport and ENSO. However, Wyrtki (1987), using sea level data to calculate transport from dynamic height differences, found no connection between PACIO throughflow and ENSO.

McCalpin (1987); Clarke (1991); du Pensoat and Cane (1991) examined the reflection of Rossby wave energy by the irregular western boundary of the Pacific Ocean and the possible connection to PACIO throughflow, Clarke and du Penhoat and Cane concluded that neither the highly irregular geometry of the coastline or the passages from the Pacific into the Indian will strongly affect the reflection of lower mode Rossby waves as compared to a meridional wall. McCalpin predicted that, unlike a meridional wall, as the period of the wave decreased past the minimum reflection point near 65 days reflection would increase, du Penhoat and Cane (1991) also concluded that there is no anomalous transport through the PACIO region associated with either Rossby wave reflection or ENSO.

Recent indications that the PACIO region may be important in general ocean circulation, and in particular large-scale phenomena such as ENSO, make this an important area to study.

Verschell et al (1995) examined the dynamics of the PACIO region and its influence on the circulation of the Pacific and Indian Ocean with the help of a reduced gravity model. Their model is forced by either monthly 1980-1989 ECMWF wind stresses, or by an imposed Rossby wave of 40 day period. Their numerical experimnt yields a mean westward transport through PACIO region at $7.5 \pm 4.75 \times 10^6 m^3 s^{-1}$. Authors suggest that episodes of large eastward transport are determined to be linked to passage of interannual Rossby wave energy from the Pacific to the Indian Ocean during the northern winter following an ENSO warm event. There appears to no appreciable effect of PACIO transport on interannual variability in the easter tropical Pacific. The average upper layer thickness of the Indian Ocean is much deeper and more variable because of the influence of PACIO transport, while the effect in the Pacific Ocean is much smaller (Fig 12.2).

Ten year time series of transports from the model run through Lombok, Savu and Timor Straits and combined transport through the region were constructed (Fig 12.3(a-(d)). A strong annual signal is apparent. The transport is mostly westward throughout the model run. An annual climatology constructed from the model output (Fig 12.4(a)-(d)) show that transport values are smallest during northern winter (October-January) and largest during northern summer (June-August).

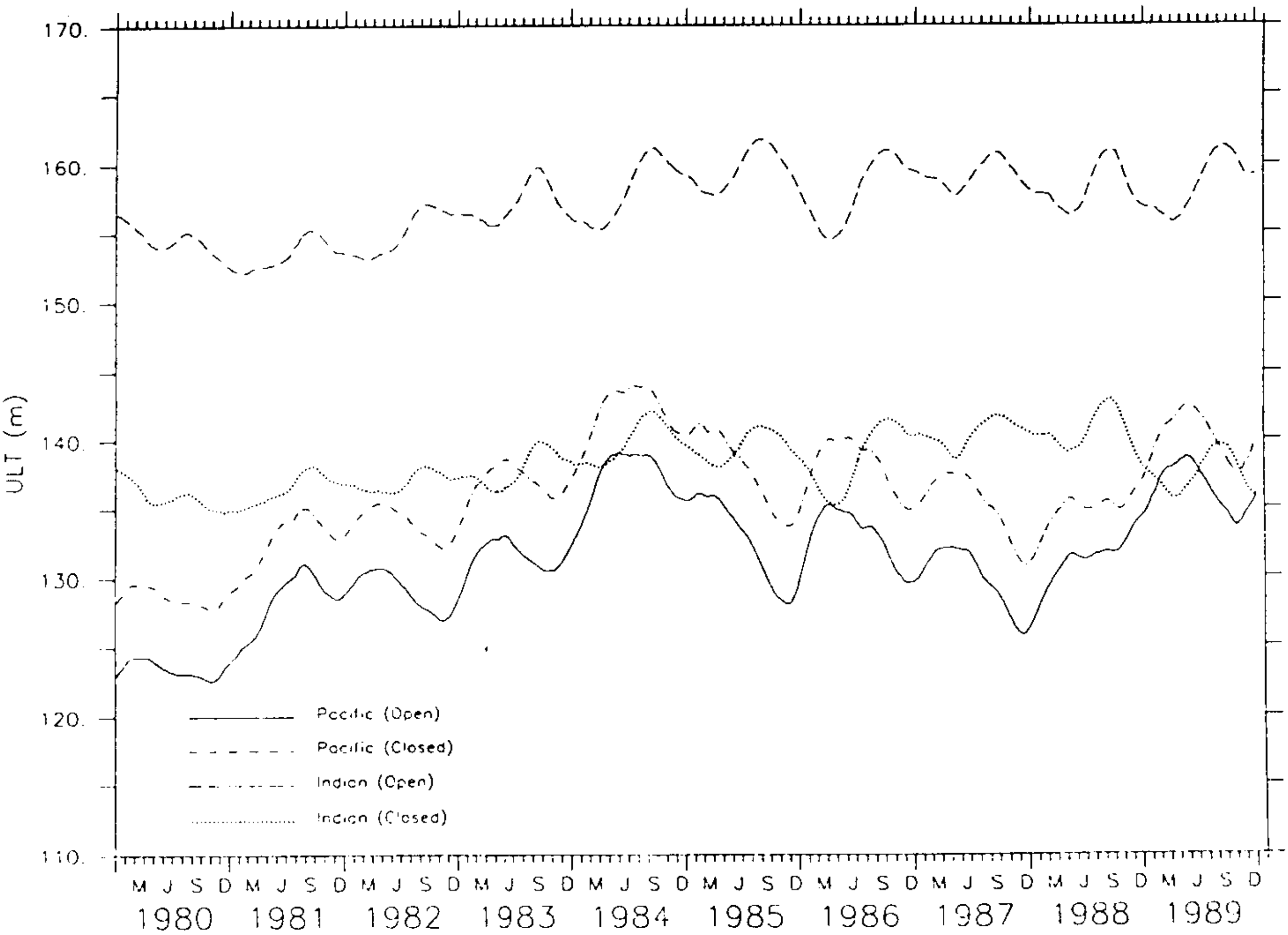


Fig. 12.2 Mean upper layer thickness (in meters) in the tropical Pacific and Indian Oceans for Open and Closed experiments (Verschell et al., 1995).

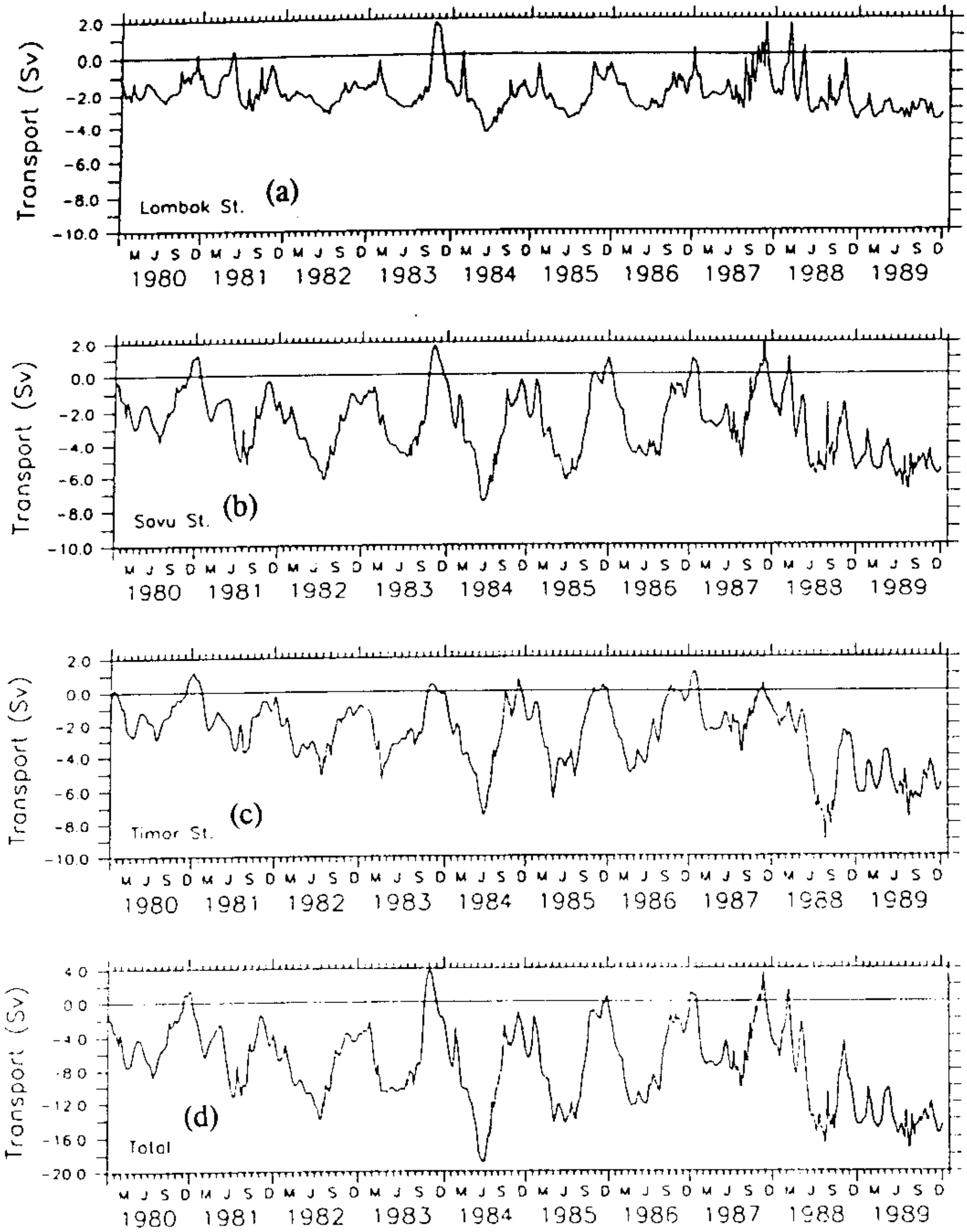


Fig.12.3 Transport through PACIO Region for Open case, in Sv. (a) Transport through Lombok Strait. (b) Transport through Timor Strait. (d) Total PACIO transport. Positive values show Indian to Pacific transport (Verschell et al., 1995).

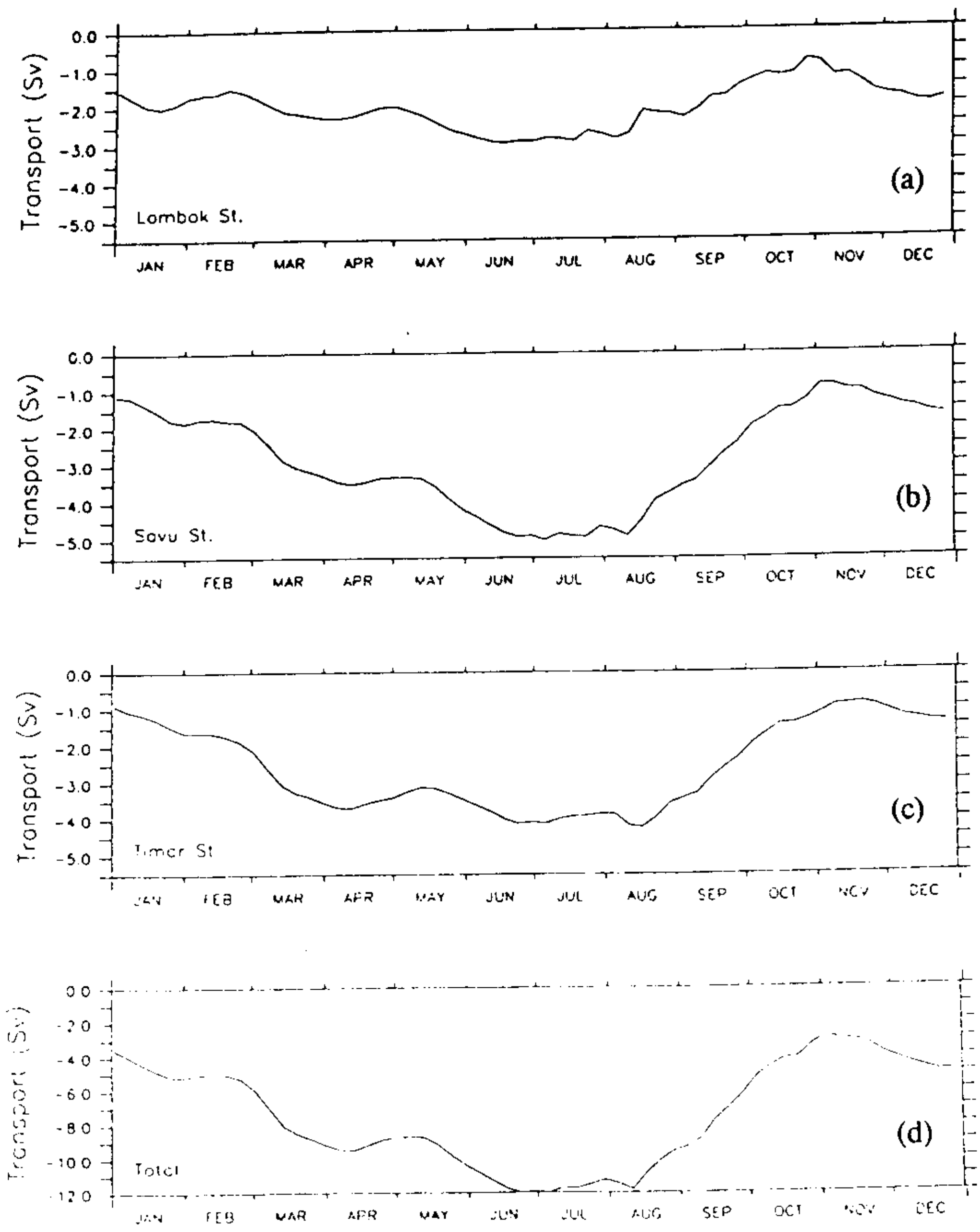


Fig. 12.4 Annual climatology of transport through PACIO Region for Open case, in Sv. (a) Transport through Lombok Strait. (b) Transport through Savu Strait. (c) Transport through Timor Strait. (d) Total PACIO transport. Positive values show Indian to Pacific transport (Verschell, et al., 1995).

13. Summary

In this lecture an attempt has been made to give the feeling that how the Indian Ocean works as an important component of Indian monsoon system. Large number of papers have been cited in where attempt have been made to study the variability of Indian Ocean parameters and processes, and their relationships with Indian Summer Monsoon and its variability. It is definite now that there is fairly good understanding of the influence of oceans on the global climate system and weather and of the tropical ocean circulation in the maintenance of the global circulation of the atmosphere and its variability. Annual and interannual variations in the strength of surface winds over the equatorial oceans cause changes in the ocean circulation and vice versa.

Contradictory results have been brought out by large number of studies carried out to correlate the SST over the North Indian Ocean and the Indian Summer Monsoon rainfall. However, there is an enhanced understanding that SST anomalies cause changes in the patterns of cloudiness, precipitation and heating of the atmosphere leading to changes in the mixed layer dynamic and surface fluxes. In order to have better understanding of the atmospheric and ocean interannual variabilities in the Indian Ocean region we may have to make further studies to enhance our knowledge about the complex process of the atmospheric response to large scale anomalies of SST and the oceanic response to the atmospheric forcing.

There have been several studies correlating the heat content of the mixed layer of Indian Ocean and variability of the summer monsoon. Modelling of the thermodynamics of the upper ocean layer have also been undertaken. These modelling studies have produced interesting results about the variability of the SST, heat flux, mixed layer depth and the depth of the thermocline and their linkage with the monsoon variability. It is however, clear that in order to fully understand and predict the variability of the monsoon climate one will have to use the results of realistic ocean-atmosphere coupled models.

Only a brief account of the interannual variability in the Indian Ocean associated with ENSO events has been given. However, I am sure that more detailed account of ENSO events and the Indian Summer Monsoon activity has been given by various other experts in the workshop.

Some interesting results pertaining to the variability of the currents in the equatorial Indian Ocean and in the western sector of the Arabian Sea (Somali Current) together with the response of monsoon circulation activities are reported. Studies on the Role of the remote forcing in the seasonal variability of the circulation in the equatorial region and in the Bay of Bengal are significant, however probably further studies are needed to confirm the hypothesis proposed by these workers.

Model results also suggest that variability in the Indonesian throughflow is reflected into the variability of the upper layer thickness which may alter the surface heat flux and SST pattern of a number of locations in the Indian Ocean.

14. Outstanding Problems

For recommending the outstanding problems, I have relied mainly on the recently prepared Science Plan report of Indian Climate Research Programme (ICRP, 1995), the Report of Group of Experts on 'Ocean Observing Systems and Services' (Department of Ocean Development, Govt. of India, 1996) and CLIVAR, Science Plan (1995).

14.1 Ocean dynamics

Low frequency coastal Kelvin waves and Rossby waves are thought to be of vital importance in the circulation of the entire north Indian Ocean. Yet there are no observations available of these waves. Time series of direct current observations are required both in the coastal region and offshore in order to document the role of low frequency waves. These and other observations (such as ocean colour or turbidity) will also help to elucidate the influence of the riever flow on the circulation, which is a basic unsolved problem of great interest.

14.2 Heat and freshwater transports

How the north Indian Ocean transports heat southward is unknown. Similarly, the way in which the evaporation-dominated Arabian Sea and and the rain and river-dominated Bay of Bengal maintain their salinities is not clear. It is known that the excess heat received by the two basins through the surface must be expelled southward. It is also likely that there is oceanic exchange of salt between the Arabian Sea and the Bay.

Repeat hydrography sections and direct current measurements will provide important ideas about these transports. They will also help to initialise and vadiate OGCM's with which such issues can be addressed.

14.3 Air-Sea fluxes

The transfer of heat and moisture from the ocean to the atmosphere depends on the inequilibrium between the ocean and the air immediately overlying it, and on the wind speed. Air-sea fluxes over the north Indian Ocean are poorly known away from shipping lanes. Direct measuremnet of fluxes of heat and moisture by fast response sensors and radiation sensors is a major requrement in this region.

14.4 Atmospheric structure and synoptic disturbances

Compared to other oceanic regions in the tropics, the frequency of occurrence of synoptic systems in the Bay during the active monsoon period is very high. Normally, the atmosphere becomes stable to moist convection following a synoptic disturbance. The surface layer has to regain its moisture and temeperature before the atmosphere can support another disturbance. The mechanism of the recovery of the surface layer and response of the tropospheric thermal structure need to be elucidated. The ocean's role in these processes is unclear as yet.

14.5 Thermohaline structure of the upper ocean and air-sea coupling

The coupling of the Bay of Bengal to the atmosphere during active and weak phases of the monsoon, and the transition from one to the other are not understood. One outstanding general question in this context is the role of low surface salinity and shallow haloclines in determining SST and in air-sea coupling.

14.6 Evolution of Atmospheric Mixed Layer over the Bay of Bengal

A large area in the Bay has SSTs over 28 C from April onwards. However, the monsoon lags SST by about two months. This may be related to the structure of the atmospheric mixed layer and the strength of the overlying inversion due to subsidence. The processes that trigger the transition from a subsidence-dominated (cloud free or shallow cloud) conditions to a convection-dominated (deep cloud) phase needs to be understood.

14.7 Propagation of synoptic disturbances

Cloud bands associated with the TCZ over the Bay exhibit northward propagations at intervals of 2-6 weeks throughout the summer. Observational experiments must be carried out to elucidate the mechanism underlying northward propagations.

14.8 Deep circulation of the north Indian ocean

The pattern and magnitude of the meridional circulation in the Indian ocean are not known. Yet circulation determines the heat balance and controls the distribution of tracers in the ocean. Present day models cannot capture the deep circulation correctly, and more observations are required. In conjunction with models, these observations of deep hydrography and currents will give a true picture of the deep flows.

Efforts be made to model the three-dimensional circulation and thermohaline structure of the North Indian Ocean with particular emphasis on the contrast between Arabian Sea and Bay of Bengal, arising from differences in thermohaline forcing.

14.9 The coupled system dynamics

- Understanding the variability and prediction of organised convection (from intra-seasonal to interannual time scales) over the warm Indian Seas in relation to the variability of the ocean beneath and impact of other oceans.

- Impact of the North Indian Ocean on the interannual variability of the monsoons.

- Modelling variability of the ocean and the coupled system on decadal and longer times scales including changes generated by gases, desertification etc.

14.10 Ocean State Forecasting

Recent studies have indicated that significant variations in SST may take place over the Arabian Sea and the Bay of Bengal even within a week's period and thus affecting the prediction of monsoon activity even on medium range scales (5-10 days). Therefore, there is need of not only the development of realistic tropical Indian Ocean Atmosphere coupled models to investigate aspects of ocean circulation variability in the context of monsoon climate variability, but also the development of operational ocean state forecasting system.

Following are some of the specific recommendations in this regard :

- Operational analysis of the state of the tropical Indian Ocean by applying data assimilation techniques to the data set derived from various observing systems.

- Seat-state bulletins, on the lines of weather bulletins, will be very useful for agencies involved in weather predictions and other related activities. The bulletin should aim at providing 3 to 5 days forecast of important oceanic features such as sea surface winds, surface currents, eddies, gyres, SST etc.

- Carrying out predictability studies and conducting experimental prediction on time scales of months to seasons based on coupled models designed to simulate the interaction between tropical Indian Ocean and the overlying atmosphere.

In the end I would like to suggest for the benefits participants of this workshop that they should refer to the excellent scientific questions given in Godfrey et al., (1995), to enhance our understanding of the role of the upper Indian Ocean in the global climate system in general and monsoon climate in particular.

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