

Chapter 2

The background

A brief overview of the coupled ocean-atmosphere system is presented at the beginning of this chapter. The focus will then shift to the monsoon climate system of the NIO.

2.1 The coupled ocean— atmosphere system

The ocean plays a significant role in determining the Earth's climate and weather through its interaction with the atmosphere. Since it is the radiation from the Sun heating the Earth that drives motion in both these fluids, an understanding of climate and weather, and how they evolve, necessarily requires an understanding of how energy is transported within, and exchanged between, the ocean and the atmosphere. The following sections provide a description of these topics; much of what follows is based on Peixoto and Oort [5] and Houghton et al. [6].

2.1.1 The components of the Earth's climate system

Weather can be defined as the state of the atmosphere at a given location. It is characterized by short-lasting meteorological events (with a typical time scale of a few days). Climate proceeds from the average weather patterns over at least a score of years (based on higher momentum statistics) modulated by the seasonal cycle. The state of the atmosphere is determined by state parameters such as density, temperature, pressure, moisture, etc. The atmosphere is an open thermodynamical system that exchanges matter and energy with its environment at its boundary. The oceans, the land surface, the biosphere, and the cryosphere (sea- and land-ice) are open thermodynamical systems, and all together constitute the environment of the atmosphere. Complex interactions are at work between these systems. Complex feedback mechanisms within the individual systems respond to internal stability. Any change in the state of any of the systems can induce

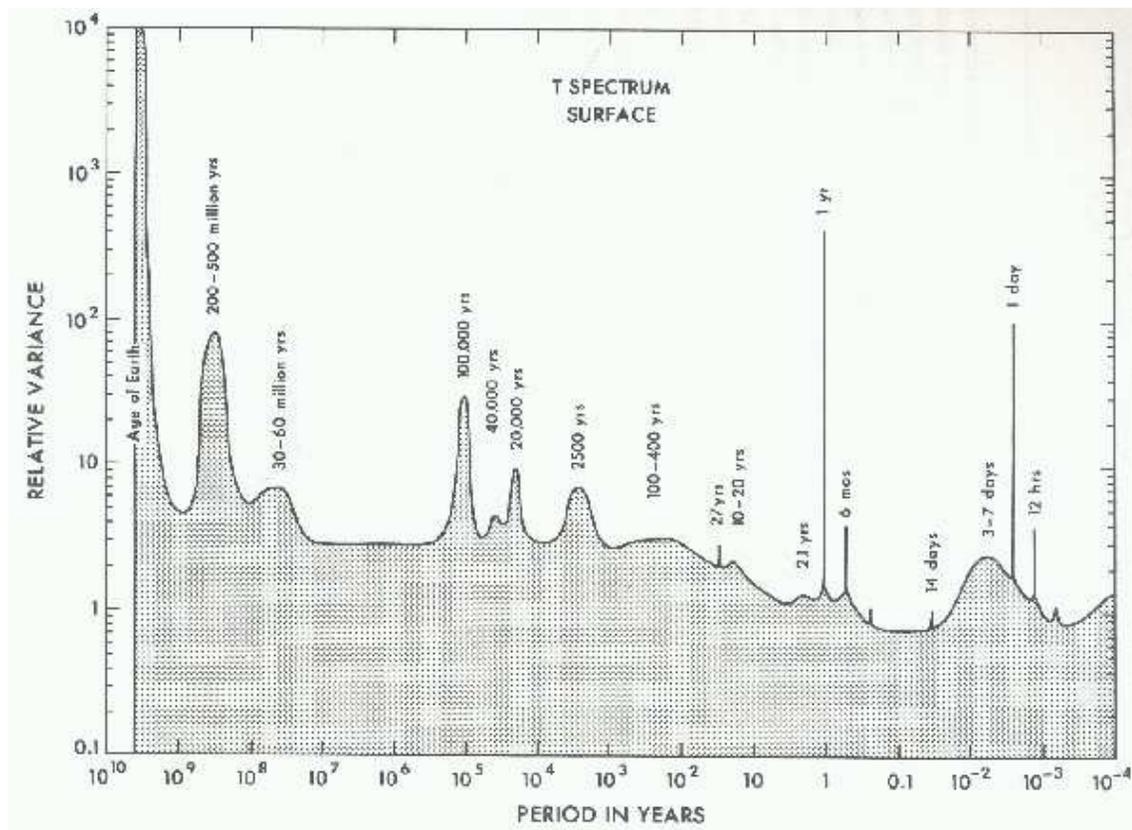


Figure 2.1: Idealized, schematic spectrum of atmospheric temperature between 10^{-4} and 10^{10} year [after Peixoto and Oort [5], their Fig. 2.7]

some change (fluctuation) in the adjacent components. All these interlinked component-systems form the Earth's climate system with the atmosphere as the central component in which we live.

The interaction of the atmosphere with other components of the Earth's climate system can take a number of forms which induce a number of processes at time scales ranging from very small scales (fraction of a second) to very large scales (tens of thousands of years or even more). This is evident from the observed spectrum of the surface air temperature during its past history as evaluated by Mitchell [7] (Fig. 2.1). Analysis of the spectrum shows several spikes and broader peaks. The spikes correspond to astronomically dictated, strictly periodic climate variations, such as the diurnal and annual variations and their harmonics, whereas the broad peaks represent variations that are, according to Mitchell [7], either quasi-periodic or aperiodic, with, however, a preferred time scale of energization. Many of these broad peaks cannot be directly explained by the known external forcing. This indicates strong coupling between the atmosphere and other components of the earth's climate system or internal feedback mechanisms within the atmosphere itself.

2.1.2 The ocean as the flywheel for the Earth's climate system

The major energy source and sink of the climate system are the solar and the terrestrial radiations respectively. The energy from the Sun is received in the form of radiation, with nearly all energy in between 0.2 and 4 μm . Some solar radiation is either absorbed or scattered by the atmospheric gases, water vapour, particles, and clouds, the remaining is either absorbed or reflected by the Earth's surface.

The amount of solar radiation reflected or scattered into space without any change in wavelength is defined as the albedo of the surface. The albedo varies in space and time and also varies for different surface types. The albedo of the ocean is significantly smaller than that of land. The mean global albedo is 30%. The remaining 70% of the incoming solar radiation must be absorbed. Of this remaining flux, about three-quarters (about half of the total incoming flux) penetrates the atmosphere and is absorbed by the Earth's surface. The remainder (i.e. 16% of the total incoming flux) is absorbed directly by the atmosphere. Both the atmosphere and the surface re-radiate this absorbed energy as long-wave radiation.

How the solar radiation is absorbed at the Earth's surface differs profoundly on land and at sea. On land, most of the energy is absorbed close to the surface (a few centimetres). The near surface warms up rapidly, leading to increase of outgoing terrestrial radiation (long wave). At sea, the solar radiation penetrates much deeper, with 20% reaching a depth of 10 m or more. Turbulent diffusion can transport heat from surface to sub-surface layers. So the sea is heated to a much greater depth and the surface warms up much more slowly. The ocean also has a large thermal inertia and it can store heat for a long period, from several months to several years (in contrast to the period of minutes to a few hours exhibited by land). This implies that more energy is stored in the top layer of the ocean and less is lost to space in the form of long-wave radiation. This absorptive capacity of the oceans plays an important part in the dynamics of the Earth's climate system. In effect, ocean act as a huge climatic flywheel, damping down fluctuations in other parts of the climate of the globe.

Because the atmosphere and the oceans transport energy from one place to another, this motion is an integral part of the Earth's radiation balance. On a global scale the most important feature is that, while most solar energy is absorbed at low latitudes (net radiation balance toward the earth), a large amount of energy is radiated toward space at high latitudes (net radiation balance away from the earth). This loss must be balanced by energy transport from lower latitudes. This provides the energy to fuel the engine that drives the global atmospheric and oceanic circulation throughout the year. Peixoto and Oort [5] show that both atmosphere and ocean transport about the same amount

of energy pole-ward. It can be noted that the oceanic component of the meridional heat transport is significantly larger in the tropical region than the atmospheric component, while the opposite occurs in the mid-latitude regions. As regard the oceanic heat transports, the Atlantic carries more heat north-ward than the Pacific. In the southern hemisphere the Pacific transports roughly twice as much energy south-ward than the Indian Ocean, while the Atlantic has a small negative effect, carrying energy north-ward across the Equator.

An interesting feature of the Indian ocean is that the meridional heat transport in the northern basin is negative (toward the equator) in contrast to the other oceans. This is because the Indian Ocean, unlike the other oceans, is blocked by the land masses of the Asian continent to the north around 25° N. The meridional heat transport toward the south pole in the Indian Ocean is smaller than in the other oceans. These unique features of the Indian Ocean have important consequences for the climate of the surrounding regions.

The role of the oceans in the meridional heat transport process cannot be thought as independent of the atmosphere. There is continual exchange of energy in the form of heat, momentum (as winds stir up waves), and moisture (in the form of evaporation and precipitation). This energy flux is particularly large in the vicinity of the major currents, and where cold polar air moves to lower latitudes over warmer water. These regions are the breeding grounds of many of the storm systems that are a feature of the mid-latitude weather.

Maintaining an energy balance by transport from low to high latitudes has another fundamental implication. This relates to the mean temperatures of different latitude belts. Although these temperatures can fluctuate by a certain amount at any time of the year, they will be close to the current climatic normal. Moreover, if one region is experiencing exceptional warmth, adjacent regions at the same latitudes will experience compensating cold conditions. This means that, at short scales, climatic fluctuations are principally a matter of regional patterns. These are often the product of the occurrence of certain extremes (e.g. cold winters, hot summers, droughts, and excessive rainfall) becoming more frequent than usual, but in the long term, it is possible for mean latitudinal temperatures to vary while maintaining the global energy balance. For the most part, however, these overall changes are small compared to the regional fluctuations.

In view of the above description, it is clear that the ocean plays a significant role, in fact a regulating role, in the earth's weather and climate through its interaction with the atmosphere. Understanding this air-sea interaction process at global as well as regional scales is fundamentally important for understanding and predicting the weather and climate of the Earth. In what follows, some specific aspects of these processes will be described.

2.1.3 Air-sea coupling processes and sea surface temperature

The interface between the ocean and the atmosphere is dynamic. Matter and energy are continually being transferred in both directions (see Fig. 2.2). Matter exchange takes place through freshwater flux in the form of evaporation from ocean to atmosphere and of precipitation from atmosphere to ocean. Energy can be exchanged between the two systems in the form of fluxes of momentum, radiation, and heat. The characteristics of these energy fluxes are as follows.

1. Momentum (kinetic energy): The atmosphere supplies energy to the upper ocean by transferring momentum through wind stress. This momentum exchange creates ocean waves and currents, which are mainly responsible for the transport of heat and momentum in the ocean. In contrast, the ocean plays a very sluggish role in momentum transfer to the atmosphere.
2. Radiation: (a) Solar short-wave radiation enters the ocean and constitutes the main energy source for the ocean; (b) long-wave radiation is exchanged between the ocean and the atmosphere in both directions. As far as net long-wave radiation is concerned, it is from the ocean to the atmosphere.
3. Turbulent heat exchanges: There are two ways for energy to be exchanged in the form of turbulent heat fluxes. (a) The sensible heat flux (conduction process) can be exchanged in either of the directions depending on the difference between sea surface temperature (SST) and the temperature of the overlying atmosphere. If the difference is positive, heat flows to the atmosphere. (b) Latent heat flux: In the process of evaporation heat is carried by the water molecules in latent form from the ocean surface to the atmosphere. (This depends on SST, wind speed, and the moisture content of the atmosphere.) The net result is a cooling of the ocean surface.

These flux exchanges are the mediators of the coupling processes, which drive many physical processes (in space and time scales) inside both the systems. These fluxes are directly and indirectly related to SST, which is thus an important indicator of these coupled processes, and hence many investigators put forth their efforts to understand the links between the SST and the various air-sea interaction processes. Some of these studies are reviewed below.

Air-sea interaction is most intense in the tropics (Oceanography Course Team [8]). The intensity of the Hadley circulation and the position of the Inter-Tropical Convergence Zone (ITCZ) are dependent on SST (Bjerknes [9] & [10]). Numerous studies (Gray [11], McBride [12], Briegel and Frank [13], etc.) indicate that SST is crucial for the development and intensification of tropical

cyclones. Tropical cyclones are observed to form only when the SST is greater than 26.5°C, and passage over cooler water causes them to lose intensity. In reverse, cyclones affect the ocean surface/SST, causing strong surface cooling (through an enhanced mixing, owing to strong winds, of cooler sub-surface waters with the surface waters) that can then act to limit the intensity of subsequent cyclones. Lindzen and Nigam [14] suggest that SST gradients are important in determining the location of precipitation over the tropics, including the monsoon region.

On longer time-scales, ocean-atmosphere interactions occur on large spatial scales. The main mode of oscillation on time-scales of a year or more is the El-Niño/Southern Oscillation phenomenon (ENSO). The phase of ENSO (El Niño or La Niña) strongly influences the formation of tropical cyclones in many formation regions. In the Atlantic, during an El Niño (La Niña) event, vertical wind shear is increased (reduced) as a result of stronger (weaker) upper-level wind-shear westerlies over the Caribbean basin and tropical Atlantic (Gray [15], Goldenberg and Shapiro [16]). In the western North Pacific, the relationship with ENSO is weak, but, it still exists (Lander [17], Chan [18]), as for the relationship in the central North Pacific (Chu and Wang [19]). In the south-west Pacific, El Niño causes strong changes in the regions of storm formation, again largely associated with changes in the typical vertical wind shear (Hastings [20], Evans and Allan [21], Basher and Zheng [22]). A similar relationship exists between El Niño and the number of tropical cyclones off the western coast of Australia (Broadbridge and Hanstrum [23]). These changes are fundamentally driven by the large-scale variation in the SST patterns. These cause the zones of cyclone formation to shift eastwards, away from the coast of Australia. The opposite pattern is typically observed in La Niña years. These relationships are used as the basis of a number of statistically-based seasonal forecasting schemes of tropical cyclone formation for various regions of the globe (Gray [24], Elsner et al. [25], Chan et al. [26])

(While SST plays a highly significant role in controlling the climate and weather of the tropical regions, its effect at higher latitudes cannot be ignored. The most frequent mid-latitude (frontal) disturbances are mostly driven by sensible heating (temperature difference between the sea surface and the overlying atmosphere) (Oceanography Course Team [8])).

2.1.4 SST and the concept of mixed layer

Numerous factors affect SST. The local air-sea fluxes have a very direct and strong control. Besides local-air sea fluxes, SST is also controlled by internal processes occurring in the ocean. These processes include horizontal advection, vertical advection (upwelling and downwelling), and vertical diffusion. Horizontal diffusion of heat is much slower in the ocean. How these processes work

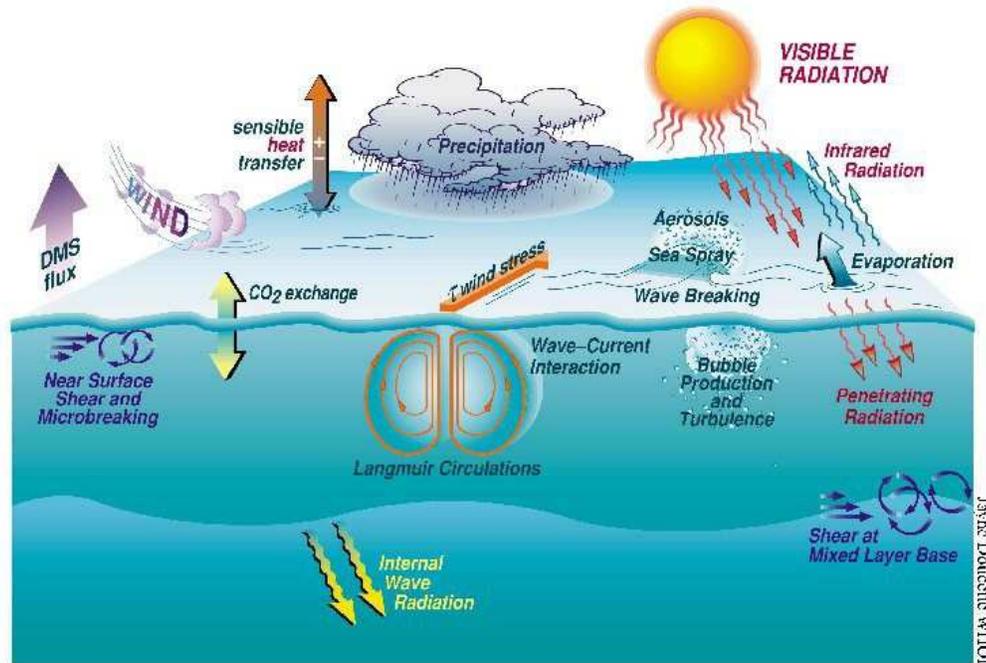


Figure 2.2: A schematic of various processes active in the upper ocean (NOAA [27])

together in the ocean is presented schematically in Fig. 2.2.

The surface mixed layer is a quasi-homogeneous layer (also known as oceanic boundary layer), the site of active air-sea exchange formed in the top layers of the ocean owing to quasi-balance among the various forcing fluxes. Forcing from winds, heating and cooling, and rainfall and evaporation, have a profound influence on the distribution of mass, energy, and momentum in the ocean. Although the effects from this wind and buoyancy forcing are ultimately felt throughout the ocean, the most immediate impact is in the surface mixed layer. The mixed layer is warmed by solar (short-wave) radiation and cooled by (long-wave) radiation emitted from the surface and by latent heat loss due to evaporation. The mixed layer tends to be cooled by sensible heat loss since the surface air is generally cooler than the ocean surface. Evaporation and precipitation change the mixed-layer salinity. These salinity and temperature changes define the surface buoyancy. As the surface loses buoyancy, it becomes denser than the sub-surface waters, causing convective overturning and mixing. Wind forcing can also cause surface overturning and mixing, as well as localized overturning at the base of the mixed layer through shear-flow instability. This wind- and buoyancy-generated turbulence causes the surface water to be well-mixed and leads to vertically uniform temperature, salinity, and density profiles. Furthermore, turbulence can entrain deeper water into the surface mixed layer, causing the surface temperature and salinity to change and the layer of well-mixed, vertically uniform water to thicken. Wind forcing can also set up oceanic

currents and cause changes in the mixed-layer temperature and salinity through horizontal and vertical advection.

Salinity affects the thermal structure by affecting the density field. Density is a strong function of temperature and a weak function of salinity (see Appendix-A). Furthermore, the salinity of the upper ocean usually varies very slowly in space and time. Therefore, in most cases, salinity does not have much control over the thermal field. However, there are some situations where the salinity has strong control over the thermal field. This is due to the so-called “barrier layer” (BL) phenomenon (Vialard and Delecluse [28]). A BL is present when the isohaline layer is shallower than the isothermal layer. In such a case, the mixed-layer (ML) depth is controlled by salinity stratification. The under lying water has a potential temperature close to the SST. A deepening of the ML (e. g., under the effect of a wind burst) will not affect surface temperature via entrainment cooling. This is why the water column between the bottom of the ML and the top of the thermocline has been named barrier layer: it acts as a barriers to cooling owing to entrainment. Using the data from Levitus [29], the presence of the BL in the three tropical oceans has been recognized to be not just a sporadic event but a climatological feature (Sprintall and Tomczack [30]). Moreover, a strong time variability of the BL has been observed and linked to inter-annual variability (Delcroix et al. [31]; Sprintall and Mcphaden [32]). Vialard and Delecluse ([28] & [33]) examined the role of salinity in the formation of the tropical Pacific warm pool. They showed that there are occasions of formation of BL in the equatorial region due to very large bursts of fresh-water flux owing to heavy rain events. The indirect study by Shenoi et al. [34] suggests that salinity stratification has a significant effect on the surface layer heat budget of the Bay of Bengal (eastern NIO), while it has a negligible role in the Arabian Sea (western NIO). The recent model study by Durand et al. [35] has shown that the SST of the eastern Arabian Sea increased by 1.1°C during November–March owing to the presence of a BL below the mixed layer, whereas the atmospheric fluxes cooled the surface by 0.3°C over the same period.

Another important issue related to mixed-layer processes study is the impact of biomass (chlorophyll concentration) on the thermodynamics of the upper ocean through the vertical attenuation of solar irradiance. When biomass concentration increases, the solar irradiance is trapped within a thinner upper layer. This stabilises the ML by reducing the vertical mixing due to convection and wind mixing, and it increases the SST. In reverse, biomass is affected by the local thermodynamics. Such a study was done for the Arabian Sea by Sathyendranath et al. [36] using a coupled bio-optical/physical model. They showed that the rate of biological heating effect is significant (upto 4.0 °C month⁻¹) during the August.

2.2 Coupling between the North Indian Ocean and the Monsoons

The portion of the Indian Ocean that lies north of 10°S is considered here as the North Indian Ocean (NIO) (see Fig. 2.3). The NIO is a small tropical basin, closed by the land masses of south-Asian countries in the north around 25°N . The tropical part of the NIO is divided by the Indian subcontinent into two semi-enclosed basins, the Arabian Sea and the Bay of Bengal. Unlike other world oceans (Atlantic and Pacific), the NIO has no connection to the northern polar seas. This unique size and geographical location has significant consequences for the climate of the surrounding regions. The differential heating between the NIO and the surrounding land masses with the seasonal movement of the Sun causes a north-south pressure gradient. This pressure gradient drives the seasonally reversing (monsoon) winds over the NIO (see Figure 2.3a-b). The monsoon winds generally blow from the south-west during May–September (the SW monsoon or summer monsoon) and from the north-east during November–February (the NE monsoon or winter monsoon), March–April and October being the months of transition with weak winds. The winds are stronger during summer than during winter. Since the SW monsoon winds blow from the NIO to the land, they are loaded with moisture and this brings heavy precipitation to the coastal region. Rainfall during the SW monsoon is the major source of freshwater in the coastal region of the NIO (including India). The NE monsoon winds blow from land to the ocean; hence they are basically dry and do not cause significant rainfall over the Indian sub-continent (except over south-east India). Of the two monsoon seasons, the summer and the winter, the summer (SW monsoon) is particularly important for India. The monsoonal activities over the Indian subcontinent are strongly dependent on the thermodynamical state of the NIO. During the SW monsoon, the SST over the major part of the tropical NIO are warmer than 28°C (a threshold for the atmospheric instability over tropical Indian ocean) (Gadgil et al. [37]). In contrast, during the NE monsoon, the SST over the major part of the tropical NIO remains below the threshold value (see Fig. 2.3c-d).

Numerous attempts were made in the past to understand the various aspects of the Indian summer monsoon (which include genesis, onset, and intensification) and its possible link to the thermodynamical state of the NIO. Reviews of such studies were published by Webster et al. [38], in the Science Plan of the Indian Climate Research Programme [39], and in the monograph by Varadachari [40]. In the following section, some of the important aspects of the Indian summer monsoon (ISM) (air-sea coupling processes) are briefly described .

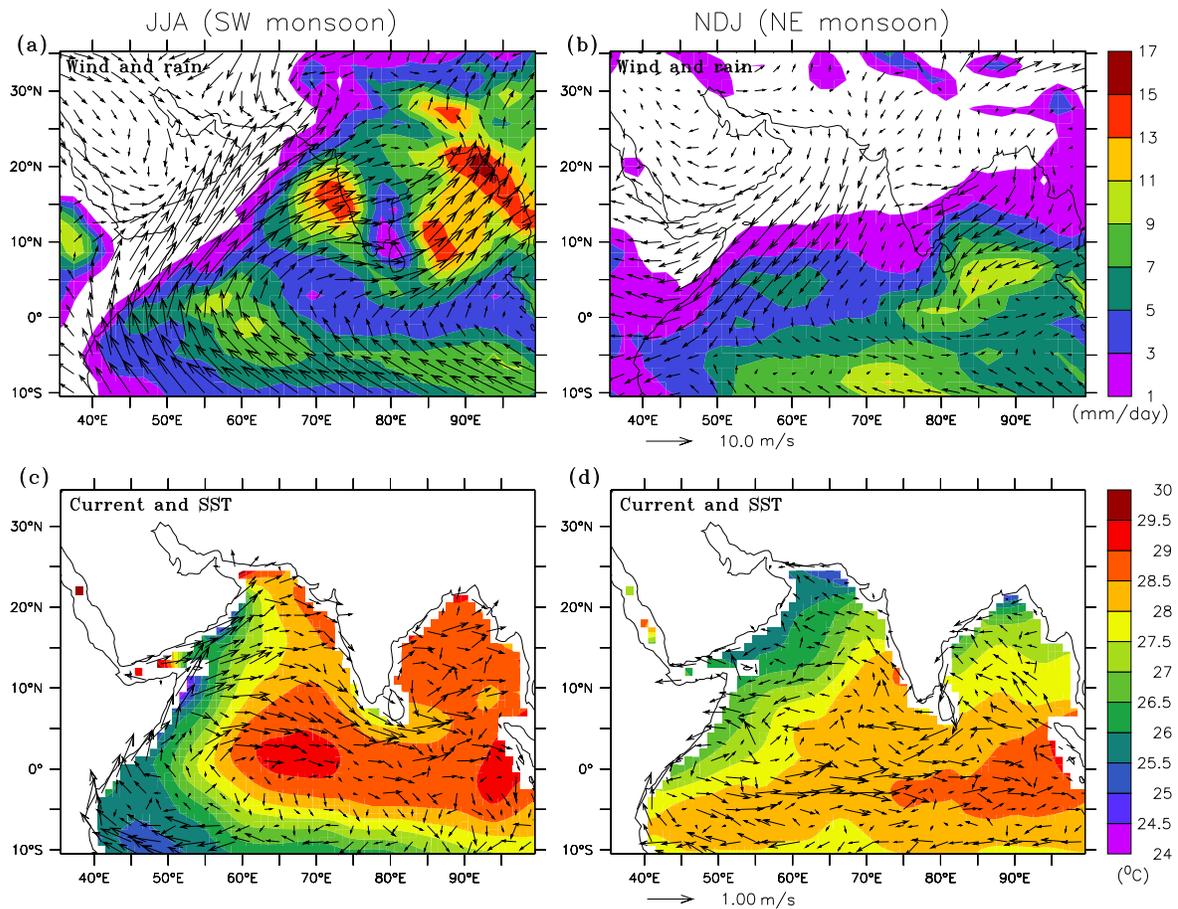


Figure 2.3: (a) Climatologies of wind-vectors and rainfall (color) over the NIO and the surrounding regions during the peak period of the SW monsoon. (b) The same as (a), but for the peak period of the NE monsoon. (c) Climatologies of surface currents (vector) and SST (color) in the NIO during the peak period of the NE monsoon. (d) The same as (c), but for the peak period of the SW monsoon season.

2.2.1 SST and the ISM

SST is an important indicator of the ocean-atmosphere coupling process. There have been numerous attempts to study the possible connections between the SST of the NIO and the ISM variability.

2.2.1.1 Genesis and onset of the summer monsoon

With the northward seasonal advance of the Sun, a progressive shift of the axis of the zone of high SST occurs, and the west-east oriented Inter-Tropical Convergence Zone (ITCZ) can be located along this zone (Choudhury and Karmarkar [41]). Hastenrath [42] showed that some of the precursor characteristics of the ISM are: (i) temperatures over land and the neighboring Arabian Sea

are spatially coherent, as is the surface pressure; and (ii) pressure over land and the wind speed over the Arabian Sea are inversely coupled.

From the climatological atlas of Hastenrath and Lamb [43], the following features can be inferred. The domain of warmest surface water moves northward from February to July with similar changes in the surface pressure patterns. In April, the 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, this being also the domain of low surface pressure. As the domain of the warmest surface waters moves to the central Arabian Sea and Bay of Bengal, a surface pressure pattern with a meridional decrease from the south Indian Ocean to the Indian subcontinent is established. Saha ([44]& [45]) suggested that the existence of a well-defined SST anomaly between the equatorial west and east Indian Ocean could possibly explain phenomena like the equatorial westerlies, the double ITCZ, and the Somali jet (also known as Findlater jet (Findlater [1])).

A number of atmospheric phenomena over the Arabian Sea during summer might arise as feedback effects of the sea, with a complex distribution of SST, on the overlying atmosphere, and the advection of cold water from the coasts of Somalia and Arabia towards the east from June through August could play a dominant role in the interaction. The study by Rao et al. [46] on the thermal structure showed that the SST was higher in May throughout the Arabian Sea (the highest temperature of 31°C being located around 11°N, 65°E) and that there was a decrease of SST ranging from 0.5°C in the northern Arabian Sea to 3°C or more in the western Arabian Sea, and a general increase in the thickness of the mixed layer from May to June–July. The tongue of cold water resulting from the Somali upwelling in summer in the southwestern Arabian Sea extends northward and eastward, and is present throughout the central Arabian Sea between 11°N and 15°N during June–July. Rao et al. [47] studied the time variation of SST and moisture over the Arabian Sea during 25 May to 15 June 1973, covering the pre-onset and onset phases of the ISM. They reported that the onset of ISM was associated with a decrease in the SST especially north of 5°N. This was more marked up to 65°E, probably due to the northeast-ward transport of the cold water from the Somali coast and the strengthening of the low-level Findlater jet.

Studies by Nuzhdin [48] on the energetics of the atmosphere and ocean during MONEX-79, revealed the important role of the Arabian Sea surface-layer heat content in the process of summer-monsoon development. Ocean areas of pronounced heat storage can be viewed as the centres of air-sea interaction (energy-active areas), where evaporation, transfer of sensible heat to the atmosphere, and updraft of warm and moist air masses take place. Such conditions, which promote the onset of monsoon, are characteristic of the central Arabian Sea. Simultaneously,

negative anomalies of the ocean enthalpy are found in the western Arabian Sea during April–May. Therefore, the energy exchange is directed from the atmosphere to the ocean and a downdraft prevails. This contrast between the western Arabian Sea (Somali current region) and the central Arabian Sea energy-active area is one of the factors causing sharp intensification of the south-westerlies and the ITCZ migration northwards to 20°N. This contrast generally becomes critical at the beginning of June (Nuzhdin [48]). In 1979, the onset of monsoon was delayed by 10 days. A strong positive ocean enthalpy in the central Arabian Sea was observed at this time, indicating that the aforementioned contrast would have conditioned the slow-down of the process. It has been observed that lower than normal enthalpy in the central Arabian Sea as compared to the long-term normals favours the SW monsoon to set in.

Gadgil et al. [37] investigated the interaction between the large-scale convective system associated with the continental ITCZ and the ocean over which they were generated, concentrating on the relationship between organized convection and the Indian Ocean SST. They reported that on a monthly basis, the degree of cloudiness correlated well with SST for the relatively cold ocean, but when SST was maintained over 28°C, it ceased to be an important factor in determining the variability of cloudiness. Over the major regions of convection east of 70°E, which are warm year after year, the observed cloudiness could not be correlated with variations in SST.

2.2.1.2 Intensity of the summer monsoon

Jambunathan and Ramamurty [49] showed that during an active monsoon, the sea surface was warmer than the air over the Arabian Sea, and during a weak monsoon the sea surface was colder. They reported that the sea and air temperatures were lowest off the coasts of Somalia, Arabia, and Kerala, and highest over the Gulf of Aden, and north and east of Arabia during both the active and weak monsoon spells. Anjaneyulu [50] concluded that the positive anomalies of SST during May over the west and the adjoining central Arabian Sea, though small, were associated with subsequent good¹ monsoon and vice versa. He also found that in good monsoon years, the difference between the highest SST during the premonsoon and the lowest SST during the monsoon in areas off Arabia and Somalia was large and suggested that the negative anomalies of SST during August, over the western Arabian Sea and the adjoining central Arabian Sea, were associated with good monsoon and vice versa.

Joseph [51] indicated that monsoon failures during the decade 1964–1973 were followed by

¹Monsoon of a region in the Indian subcontinent is good, if it receives 10% more rain fall than its climatological normal.

above normal SST over large areas of the NIO, and these anomalies persisted for one year. These failures in 1965–66, 1968–1969 and 1971–72, were associated with spells of seasonal scale of upper tropospheric westerly intrusion over South Asia. A bad² monsoon warms up the NIO through a decreased upwelling along the Somali and Arabian coasts, weaker surface winds over the Arabian Sea and the Bay of Bengal, reduced mixing and evaporative cooling, and reduced cloud cover (Joseph [52]; Joseph and Pillai [53]). High SST was found to cause the formation of a warm anticyclone in the upper troposphere over the Indian seas during the following winter and the pre-summer monsoon seasons. This was followed by a good monsoon which led to a cooling of the NIO. It was suggested that such a mechanism could be responsible for the three-year oscillation in the upper tropospheric wind over South Asia, the ISM rainfall over India, and the tracks of the post-monsoon cyclonic storms of the Bay of Bengal (Joseph [54]). The SST over the NIO was also found to have a three-year oscillation (Joseph and Pillai [55]). It is interesting to note that Golovastov et al. [56] reported quasi-triennial periodicity of the heat content in the upper layer of the Arabian Sea, an intensification of the Somali Current and the consequent increase in the upwelling. There are reasons to believe that tropospheric wind anomalies could be caused by SST anomalies. Joseph and Pillai [53] studied the relationship between SST and the tropospheric circulation using a Global Circulation Model (GCM) and concluded that SST anomaly over the Pacific (but not over the NIO) can drive the tropospheric circulation as observed over the NIO.

2.2.1.3 Correlation studies

A number of correlation studies were carried out to link the variability of various part of the NIO to variability of ISM. Shukla and Mishra [57] calculated the correlation coefficients between SST anomalies at about 10°N, between 60 and 70°E, and the seasonal mean rainfall over the Indian meteorological sub-divisions, and found that the correlation between SST during July and the rainfall during August over the central and western India was positive and significant. They hypothesized that cold temperatures over the Arabian Sea cause a decrease in evaporation and thus a decrease in the moisture available for downstream precipitation, leading to a rainfall deficit over India. Weare [58] concluded that a warmer Arabian Sea was weakly associated with a decreased rainfall and increased pressure over much of the Indian subcontinent. He explained that periods of higher SST, which might be expected to lead to deeper monsoon lows, were usually accompanied by periods of higher than average pressure over the sub-continent, which would tend to offset the

²Monsoon of a region in the Indian subcontinent is bad, if it receives 10% less rain fall than its climatological normal.

effect of temperature variation. Joseph and Pillai [55] also worked out the correlations between the monsoon rainfall over India and the SST in three selected representative areas in the Arabian Sea and the Bay of Bengal. They found that the SST of the pre-monsoon months is positively correlated with the following monsoon rainfall, but the correlation was low and not statistically significant. However, the monsoon rainfall was found to be significantly negatively correlated with the post-monsoon SST. Joseph and Pillai [53] also concluded that a good monsoon produced a negative SST anomaly in the NIO.

Later, a number of correlation studies (Ramesh Kumar et al. [59] & [60]; Cadet and Diehl [61]) established a positive connection between the SST variability of different parts of the Arabian Sea and the genesis and intensity of the ISM. In general, the correlation between the summer rainfall over India and the anomaly in surface fields (pressure, wind, SST, air temperature) showed that weaker trade winds during the different months of the year, a weaker Somali jet during summer, and higher SST in the Indian Ocean were associated with weaker rainfall over India (Cadet and Diehl [61]). The weaker winds lead to less evaporation, less turbulent mixing, reduced upwelling, and, consequently, less cooling of the ocean surface (Webster [62]).

2.2.1.4 Model studies

The studies described above were mostly based on observations. There were also several model studies carried out to examine possible connection between SST of the NIO and the Monsoon. Washington et al. [63] and Druyan [64] simulated the effect of SST anomalies in the Arabian Sea on the ISM rainfall. They found that the resultant rainfall departures were not statistically significant. Washington et al. [63] showed that global-scale changes in circulation could be caused only by an anomaly of a larger spatial scale and that small spatial anomalies were able to generate only local effects. Washington [65] concluded that the experiments with two Atmospheric General Circulation Models (AGCM) suggested the existence of a local response of SST in the Arabian Sea to precipitation over the Indian sub-continent. The extent to which the Indian Ocean SST anomalies correlate with precipitation over the Indian sub-continent, however, remained to be established.

Shukla [66] pointed out that the results of some of these studies should be viewed with caution. Shukla used about 80 years data to study the correlations of SST anomaly in the NIO with the monsoon rainfall over India. He pointed out that warm (cool) SST anomaly during April, May, and June was not necessarily indicative of an above (below) average monsoon rainfall, though heavy rainfall was followed by negative SST anomalies. He further stated that the predictive value

of this relationship was rather limited because the magnitude of SST anomalies during the pre-monsoon months was within the range of observational error. Shukla also found negative bias in the SST data before the 1940s and positive bias thereafter. Shukla and Fennessy [67] carried out a GCM study in which the solar forcing of the land and ocean were incorporated separately. They showed that the annual cycle of SST in the NIO was crucially important in establishing the monsoon circulation and rainfall. Model simulations also confirmed that the Arabian Sea SST influences subsequent rainfall on time-scales less than 1 month. A cool Arabian Sea SST led to reduced Indian rainfall and vice versa (Shukla [68]; Kershaw [69]; Yang and Lau [70]).

In summary, most of studies described above tried to establish a relationship between SST of the NIO and variability of ISM. Despite contradictions between these studies, most of them agreed that the ISM is dependent on the variability of SST of the NIO, in particular of the Arabian Sea. Thus, examining the SST (ocean state) of the NIO is important for understanding the monsoonal climate over the Indian subcontinent. A number of studies were carried out in the past to understand the upper ocean thermal structure and heat budget of the Arabian Sea and of the NIO. A brief review of such studies follows.

2.2.2 Upper ocean thermal structure and heat budget

2.2.2.1 Basin-scale studies of the NIO

A number of basin-scale studies have been carried out to understand the thermal structure and heat budget of the NIO. A review of such studies is presented here. In the course of the year, the ocean gains a substantial amount of heat from the atmosphere in the northern hemisphere and the equatorial region. In contrast, the southern ocean experiences a net heat loss to the atmosphere. The annual range of the net oceanic heat gain increases with distance from the equatorial belt in both the hemispheres. Hastenrath and Lamb [43] suggested that the ocean operates as an effective heat exporter in the northern hemisphere and the equatorial belt of the Indian Ocean and the adjacent tropical seas, whereas it acts as a heat importer in the higher latitudes of the Southern Indian Ocean. They showed that the oceanic heat budgets in the Arabian Sea and Bay of Bengal exhibit greater annual variations than the other regions in the Indian Ocean because of the reversal of the monsoon systems. The net oceanic heat gain shows large positive values in the Arabian Sea and Bay of Bengal around March to April and September to October, broadly coinciding with the net radiation maxima and latent heat flux minima. The ocean loses heat to the atmosphere in these areas at the start of the summer monsoon and during winter, when enhanced evaporation accompanies reduced net radiation.

Latitudinal and annual variation of residually determined net oceanic heat gain in the Indian Ocean is presented by Hastenrath and Lamb [71]. The equatorial and northern Indian Ocean experience a sizeable heat gain, except for the more northerly region, during the northern winter, while a heat loss prevails in the south tropical Indian Ocean in the northern summer. During the northern winter, the radiative heat input at the top of the atmosphere and the precipitational heating of the atmosphere are very small for the northern Indian Ocean, and are much larger to the south of the equator (and vice versa during the northern summer).

Golovastov [72] computed the heat storage in the surface (0–100 m), sub-surface (100–400 m), and intermediate waters (400–1000 m) in the Indian Ocean using the MONEX-79 data during the SW monsoon. One of the features in the surface thermal field, brought out by the study, was high heat storage north of 2°S and low heat storage south of 2°S. During the development of the SW monsoon, the distribution of heat storage in the different layers was quite unequal. This was closely related to the peculiarities of the water-mass formation and the thickness of the surface layer. The frontal zone between the Arabian Sea and the cold waters of the south equatorial divergence was distinctly outlined in the maximum gradient of heat storage between 3° and 5°S. The ITCZ, located over 3–4°S in April–May, 1979, was associated with it. Analysis of the surface and sub-surface heat storage revealed that the late commencement of the SW monsoon and the ITCZ migration along the Somali coast from the southern to the northern hemisphere occurred mainly due to the weak development of the Somali current and the associated upwelling. It resulted in the formation of positive anomalies in heat storage in the Arabian Sea surface waters and negative anomalies in the sub-surface layer. The large-scale anomalies in turn, have been found to have considerable impact on the development of the atmospheric monsoon circulation (Golovastov [72]).

Golovastov et al. [56] also examined the formation and variability of heat content in different layers in the tropical Indian Ocean using data from 1959 to 1979. Analysis of seasonal variations in the heat content showed that the thermal field varied significantly under the effect of both outer radiative and inner dynamic factors. In several cases, values of heat content variations seem to be higher than the ocean surface heat budget, indicating that dynamic factors prevail in these regions. Analysis of the annual cycle of the heat content also reveals that variations over most of the regions are dependent on the intensity of the ocean general circulation development. A quasi-triennial periodicity in the interannual variation of the heat content is also found. During the north-east monsoon and the spring season, the variations in the surface water heat content of the monsoonal Indian Ocean are mainly dependent upon the heat budget, while during the SW monsoon and the

autumn seasons, it depends on the advective and turbulent heat exchange (Golovastov et al. [56]).

2.2.2.2 Studies focused on the Arabian Sea

The Arabian Sea is an active region of the air-sea interaction process and is recognized as an important driver of the monsoon system over the Indian subcontinent (Nuzhdin [48]). It is characterized by a pronounced summer cooling (Wyrski et al. [74]), whereas water bodies similarly situated geographically (except the Bay of Bengal) show an increase in SST with the advance of the northern summer. The amplitude of the cooling varies from region to region. Off the coast of Somalia, it has been found to be as high as 10°C or more between May and July, while off the west coast of India, it is about $2\text{--}3^{\circ}\text{C}$. There exists a region of minimum cooling around 70°E , separating the regions of cooling in the eastern and western Arabian Sea. Like other tropical oceans, a major portion of the Arabian Sea undergoes winter cooling. Thus, the Arabian Sea exhibits a unique bimodal variation in the annual evolution of surface layer temperature and heat storage (Colborn [75]).

Düing and Leetma [76] computed a preliminary heat budget for the upper ocean layer in the Arabian Sea during the SW monsoon (April–August) in order to assess the importance of possible feedbacks between the ocean and the atmosphere. They suggested that the net radiation gain minus the heat loss through evaporation and sensible heat was positive in the Arabian Sea. The net heat gain at the surface is more than offset by the large south-ward heat flux across the equator and the intense upwelling off Somalia and Arabia. During the rest of the year, the heating more than balances the summer cooling. There must therefore be a mean transport of heat southward across the equator to balance the heat gain in the Arabian Sea on an yearly average (as suggested by Düing and Leetma [76]).

They also suggested various possible mechanisms such as winter cooling in the northern Arabian Sea and southward flow at intermediate depths under the Somali Current for exporting heat out of the Arabian Sea. As indicated earlier, upwelling and advection of cold water appear to be responsible for the surface cooling during summer in the northern Arabian Sea (Düing and Leetma [76]; Sastry and Ramesh Babu [77]), whereas in the eastern Arabian Sea, the entrainment of cold waters into the surface layers and the subsequent turbulent mixing play a dominant role (Sastry and Ramesh Babu [77]). Murty et al. [78] computed the total heat transfer during 15–17 June 1973 over the east-central Arabian Sea and found that the surface heat loss accounted for only about 40% of the total heat loss in the surface layer and the balance heat is lost to the ocean interior owing to mixing across the thermocline. Shenoi et al. [34] examined the role of advection

and of vertical diffusion for the Arabian Sea north of 6°N. They concluded that the net heat gain by the surface layer (upper 50 m) through the air-sea interface during the SW monsoon was offset by southward transport of heat across the equator, by heat losses through coastal pumping (upwelling off Somalia and Arabia), and through diffusion at the bottom of the surface layer.

Recently, the heat budget study by Weller et al. [3] using the high-frequency time-series data of surface meteorology and sub-surface oceanography achieved from the moored-array observations in the central Arabian Sea during October 1994 to October 1995, concluded that the SW monsoon cooling results from horizontal advection, whereas the NE monsoon cooling is due to convective entrainment driven by the net heat loss at the interface. Fischer et al. [4] later estimated the heat fluxes due to horizontal advection using the same time-series data as used by Weller et al. [3]. Their estimate shows strong warming and cooling events during the monsoons. In many cases, these events roughly match the observed temperature trend. They also estimated the vertical advective heat fluxes by balancing the surface layer budget, and then the heat-conserving vertical velocity at 125 m. They found that several strong episodic downwelling events ($\sim 0.02 \text{ cm s}^{-1}$, corresponding to $\sim 18 \text{ m day}^{-1}$) and some smaller upwelling events are necessary to balance the heat budget.

2.2.2.3 Model studies

A number of model studies have also been carried out to examine the upper ocean thermal structure and heat budget of the Arabian Sea and of the NIO. Shetye [79] used the one-dimensional mixed-layer model of Kraus and Turner [80], as modified by Denman [81], to examine the role of various oceanic processes in the annual evolution of surface layer thermal structure of the Arabian Sea. His study was restricted to a 2° latitude zonal strip, centred at 10°N and stretching from the coast of Somalia to the SW coast of India. He concluded that the local surface fluxes alone control the surface layer thermal structure over a year except during the summer (June–August) monsoon. During the summer horizontal and vertical advections play significant roles in removing the heat gained by the mixed layer. Rao [82] used the air-sea fluxes measured during the MONSOON-77 experiment (7–21 of June 1977) in the south-central Arabian Sea at (66°E, 10.5°N) as input to the same mixed-layer model. He showed that part of the cooling was due to wind-driven convective entrainment, but suggested that some additional cooling must be due to horizontal advection (not explicitly examined in his study).

McCreary et al. [83] used a $2\frac{1}{2}$ layer thermodynamical numerical model to study the dynamics, thermodynamics, and mixed-layer physics of the Indian Ocean. They concluded that wind-driven entrainment mixing and horizontal advection are both responsible for the Arabian Sea summer

cooling. Another of their results is that the winter cooling of the northern Arabian Sea is due to the convective entrainment driven by the net heat loss at the air-sea interface.

Wacogne and Pacanowski [84], using a primitive equation model, showed that meridional circulation moves 1×10^{15} W of heat northward across the equator in the Northeast Monsoon (February) and 1.2×10^{15} W of heat southward in the Southwest Monsoon (August). Later, Loschnig and Webster [85] used the same model as used by McCreary et al. [83] and drew similar conclusions as Wacogne and Pacanowski [84]: the meridional transport of heat across the equator is important for the heat budget of the Arabian Sea.

It is to quantify the process involved in the upper ocean thermal structure (heat budget) of the central Arabian Sea and to quantify the meridional transport that the mooring array of Weller et al. [2] was deployed during the US JGOFS programme. We use the data from this mooring array and a numerical model to simulate the upper ocean thermal structure at the mooring location. This allows us to quantify the processes involved and to diagnose their contributions to the heat budget in the region. We begin by describing the data and methods in the next chapter.