Heat is the amount of thermal energy transferred from one body to another because of the temperature differences between those bodies [Sears and Zemansky, 1970]. In the environmental context, about half the solar energy reaching the earth's surface is absorbed by oceans and land. Of the energy absorbed by the ocean, most part of it is released locally to the atmosphere by evaporation and infrared radiation [Stewart, 2005]. And heat lost by the tropical oceans is the major source of energy needed for driving the atmospheric circulation. It is also known that the solar energy retained in the ocean from summer to winter influences the earth's climate to a large extent [Laing and Evans, 2010]. For these reasons it is imperative to understand the heat budget even at a regional level, in order to understand local climatic features and its variability on different space and time scales.

Upper layers of the ocean are in constant interaction with the atmosphere above, and undergoes incoherent input-output of heat energy at its surface and over a time, this imbalance changes the energy stored in those upper layers; the transfer of heat through the surface is named heat flux. The total flux of energy into and out of the ocean must be balanced (zero), otherwise the ocean as a whole would get heated up or cool down. This sum of heat fluxes into and out of the ocean is termed as the heat budget of the ocean. The flux data used in this study was obtained from Woods Hole Oceanographic Institute under the International
Satellite Cloud Climatology project (ISCCP). The gridded temperature profile data was obtained from CORIOLIS IFREMER. The following are the important parameters that influence the oceanic heat budget [Stewart, 2005]:

1. Incoming solar Radiation ($Q_{SWR}$): the flux of sunlight into the sea,
2. Outgoing long wave radiation ($Q_{OLR}$): net flux of infrared radiation from sea,
3. Sensible heat flux ($Q_{SHF}$): flux of heat out of the sea due to conduction,
4. Latent heat flux ($Q_{LHF}$): flux of heat carried by evaporated water,
5. Advection ($Q_{ADV}$): heat carried away by the ocean currents.

Therefore the total heat budget sums up, algebraically, as follows with its units as Watts/m$^2$.

$$Net\text{HeatFlux}(Q_T) = Q_{SWR} + Q_{OLR} + Q_{SHF} + Q_{LHF} + Q_{ADV} \quad (6.1)$$

### 6.1.1 Incoming Solar Radiation ($Q_{SWR}$)

Incoming solar radiation is the amount of sunlight received by the earths surface. Insolation (incoming solar radiation) on the earth surface is dependent on latitude, season, time of the day and cloudiness [Ahrens, 2007]. Thus the Polar Regions are heated less than the tropical regions; areas which are prevailed by cool weather during winter are warmer during summer and also cloudy days have less insolation transferred than a warm sunny day. Insolation also depends on the height of the sun above the horizon, length of the day, cross-sectional area of the surface absorbing the sunlight. Attenuation of insolation is due to the presence of clouds in the path, gas molecules which absorb radiation, aerosols which scatter and absorb radiation [Ahrens, 2007]. Reflectivity of the earths surface is dependent on the solar elevation angle and roughness of the sea surface. As observed from the figure (6.1), the net shortwave radiation was always positive in the study region with maximum recorded during February to April and minimum during June and July, along the coast owing to the presence of clouds during the SWM season. During the rest of the year, $Q_{SWR}$ did not show much variation, even
spatially. The climatological net shortwave radiation in the region is observed to be varying between 170 and 280 Wm$^{-2}$.

6.1.2 **Outgoing Longwave Radiation ($Q_{OLR}$)**

Ocean surface radiates as a black body at temperature of 290° K at wavelength near 10µm which are strongly absorbed by cloud and water vapour. The long wave radiation also is influenced by cloud thickness, cloud height, atmospheric water vapour content, water temperature and ice / snow cover [Ahrens, 2007]. Water vapour and clouds determine the net loss through infrared radiation. The hotter the water, more the heat that is radiated at a rate proportional to the fourth power of temperature. Tropical regions loose less heat than cold Polar Regions. If we assume that the temperature difference between poles and the equator is approximately 25° K i.e., between 273°K and 298°K, then the ratio of maximum to minimum is 298/273 = 1.092; when raised to the fourth power it is 1.42. This indicates 42% increase in the emitted radiation from poles to equator and over the same distance; water vapour can change the net emitted radiance by 200% [Stewart, 2005].

From the climatological monthly mean (Figure 6.2) of net long wave radiation that is emitted from the earths surface, it was observed that except during the SWM season, rest of the year has greater OLR. This is because of the presence of clouds during monsoon period would trap long wave radiation from getting released into the upper atmosphere. The climatological range of OLR over SEAS was between 15 and 100 Wm$^{-2}$. From insolation and OLR data, it was observed that the months of June and July have less insolation as well as less OLR. Also, February, March and April are the months when the region receives maximum insolation and also releases maximum OLR.

6.1.3 **Sensible Heat Flux ($Q_{SHF}$)**

Sensible heat flux is the process where heat energy is transferred from the earths surface to the atmosphere by conduction and convection. Sensible heat flux can be expressed by the amount of heat transmitted per unit of area per unit of time (http://disc.nasa.gov/hydrology). In the oceans, it is primarily influenced
Figure 6.1: Monthly variability of Net Shortwave radiation over SEAS
Figure 6.2: Monthly variability of net long wave radiation over SEAS
by wind speed and air-sea temperature difference. Heavy winds and large temperature gradients often cause high outward fluxes. Compared to the other three terms of the surface fluxes, sensible heat flux is very low and non variant over the region all through the year (6.3). Only during June and July months, considerable amount of heat was released into the atmosphere and again in December. This is because of the high SWM winds prevalent during June and July and the NEM winds during December. Also, during strong upwelling periods, the sensible heat flux is directed downwards and this will stabilize the air, trapping the evaporated water vapour in the lowest layer of the atmosphere and thereby reducing the evaporative power of the air with increasing humidity [Hareesh Kumar and Mathew, 1997]. For rest of the year, the region was prevailed by less intense winds and also low air-sea temperature difference. The climatological mean values ranged from $-1.0$ to $15.5 \text{Wm}^{-2}$ in SEAS.

### 6.1.4 Latent Heat Flux ($Q_{\text{LHF}}$)

Latent heat flux is the flux of heat from the earths surface towards the atmosphere which is associated with the evaporation of water at the surface and subsequent condensation of water vapour in the troposphere. It is influenced by wind speed and relative humidity. High winds and dry air evaporate much more water than weak winds with relative humidity near 100% [Stewart, 2005].

Compared to sensible heat flux, the contribution of latent heat flux is greater towards the heat budget of the region. From figure 6.4, it is observed that the latent heat flux was high during May, June and July and for the rest of the year, it was relatively less. This indicates that the ocean looses heat during the above period much more rapidly. The monthly mean values ranged from 60 to $180 \text{Wm}^{-2}$.

### 6.1.5 Net Heat Flux ($Q_{\text{net}}$)

The net heat flux over SEAS is shown in figure 6.5. From the figure, it is observed that the region gains heat for most part of the year except during June, November, December and up to some extent during January. Pre-monsoon heating is observed from February and the maximum gain in the heat at the surface
Figure 6.3: Monthly variability of sensible heat flux over SEAS
Figure 6.4: Monthly variability of Latent heat flux over SEAS
Figure 6.5: Monthly variability of Net heat flux over SEAS
was during March and April. This intense heat during the pre-summer monsoon season was due to the prevailing weak winds and clear skies. With the onset of SWM winds over the region, there is a net loss in the energy (∼ −20 W m⁻²) during that period. Reduced insolation and large latent heat flux are the reasons for this loss of heat during June. In July, the coastal region continues to receive heat though in lesser magnitude. During the season of coastal upwelling, the evaporation drops to less than 3mm/day [Hastenrath and Lamb, 1979]. Also, the flux of sensible heat is directed downward. This will stabilize the air, trapping evaporated water vapour in the lowest layer and reducing the evaporative power of the air with increasing relative humidity [Hareesh Kumar and Mathew, 1997]. Surprisingly, even during the monsoon months of August and September when the region receives heavy rainfall and runoff, the ocean is seen to be gaining more heat than compared to some of the relatively dry months like that of February and November. The reason for such unconventional behaviour could be due to the processes that are intrinsic to the ocean and not much dependent on the atmosphere. This is explored in the following sections.

6.2 Evaluating the Heat Budget Terms

Southeastern Arabian Sea has been little studied especially beyond the monsoon related parameters, Arabian Sea mini warm pool, to some extent seasonal upwelling and the circulation pattern in the region. It is in this context that a detailed heat budget study has been carried out for SEAS in order to gauge the mechanisms in heat budget terms over a year, especially the role of coastal upwelling in modulating the heat budget terms with available high quality data sets. A preliminary heat budget over the Arabian Sea for the first time was computed by Duing and Leetma [1980] and they concluded that the heat loss owing to advection across the equator and due to upwelling exceeding the heat gained from atmosphere during summer monsoon. Coastal upwelling was observed to be having a profound influence on the heat budget during SWM. Later many studies based on both observed and simulated, on the heat budget were carried out by Hareesh Kumar and Mathew [1997], Loschnigg and Webster [2000], Shenoi et al.
6.2 Evaluating the Heat Budget Terms

[2002], de Boyer Montegut et al. [2007] and Wilson-Diaz et al. [2009] over the Arabian Sea. However, as all these studies were executed taking into consideration the Arabian Sea as a whole, wherein the domination of Somalia coastal upwelling and convective cooling in the Northern Arabian Sea were the dominant factors that influenced the heat budget. Also, it is a region where there are heavy rainfall events, over the ocean and river runoff during the SWM leading to stratification. This brings about positive feedback that result in deep convection in the atmosphere. For this reason and as rightly remarked by Shenoi et al. [2002], eastern Arabian Sea requires a detailed heat budget of its own. Therefore, the aim of this chapter is to develop an understanding on how the air-sea fluxes behave and influence the ocean and up to what extent, especially during the upwelling season.

Thermodynamics of the upper ocean are influenced by the cooler subsurface waters entraining towards the surface during upwelling [by changing the stratification]. In order to estimate and understand the processes that influence the upper ocean heat budget of SEAS, heat budget has been computed between $7^\circ$ - $15^\circ$N and $70^\circ$ - $78^\circ$E up to a depth of 50m as followed by Shenoi et al. [2002]. The depth of the volume is considered as 50m because the mixed layer in this region for all seasons was shallower than 50m. Area average of the flux data is computed and averaged for a climatological year to generalize the phenomena taking place in this region. The period covers the years 2000 to 2008 as it is the common period over which all data was available.

The rate of change of heat in the region is balanced by the flux of heat through its boundaries due to advective and non-advective processes. Assuming that the fluid is incompressible and following Gauss theorem to express the conservation of heat as follows:

$$\frac{1}{A} \frac{\partial}{\partial x} \int (\rho_w C_p T) \, dv = -\frac{1}{A} \int (\rho_w C_p T) u \cdot n \, ds - \frac{1}{A} \int F \cdot n \, ds \quad (6.2)$$

Where $T$ is the temperature, $u$ is the velocity; $F$ is non advective fluxes, $n$ is the unit vector normal to the surfaces bounding the control volume (directed outwards). $\rho_w$ is the density of sea water 1026 kgm$^{-3}$, $C_p$ is the specific heat capacity of sea water at constant temperature and pressure 4000 Jkg$^{-1}K^{-1}$. The temperature $T$ is utilized from the temperature profiles of the region obtained from CORIOLIS gridded product which is of high spatial ($0.25^\circ x 0.25^\circ$) and
6.2 Evaluating the Heat Budget Terms

temporal (weekly) resolution than the available climatological datasets. The non-
advective term $F$ consists of the surface fluxes and diffusion through the bottom of
the control volume; diffusion through the lateral boundaries is negligible as there
is an opening on both northern and southern boundaries of the region where one
can assume owing to the direction of currents in the region, the energy entering
the control volume is balanced by the energy leaving that control volume. All
the fluxes are divided by the area at surface $A$ so that all the surface fluxes
are represented in $W m^{-2}$ to bring uniformity in the notation. The terms are
classified as the surface fluxes which are resultant of the atmospheric processes
and the advective terms which are treated separately as they are intrinsic to the
ocean.

6.2.1 Surface Fluxes ($Q_{SF}$)

The net heat flux through the surface is:

$$Q_{SF} = Q_{SWR} + Q_{OLR} + Q_{LHF} + Q_{SHF} \quad (6.3)$$

Where $Q_{SWR}$ is the net shortwave radiation, $Q_{OLR}$ is net long wave radiation,
$Q_{LHF}$ is latent heat flux and $Q_{SHF}$ is sensible heat flux through the surface
respectively. Climatological monthly mean surface fluxes are shown in figure
6.6. Except $Q_{SHR}$ and $Q_{OLR}$, rest of the fluxes show bi-modal distribution.
$Q_{SWR}$, increased from January to April and later, decreased later till June. From
June, it again showed increase, up to September followed by a drop to November;
this was followed by a slight raise towards December to match January values.
Overall, the $Q_{SWR}$ was observed to be greater than 200 $W m^{-2}$ all throughout
the year except during June and July when it was hampered by the presence of
thick clouds during to the SWM. $Q_{OLR}$ too increased from January to March
and later showed a decreasing trend up to July. $Q_{OLR}$ increased from August till
December. $Q_{OLR}$ was greater than 50 $W m^{-2}$ during all the year except during
the SWM season. Unlike $Q_{SWR}$ and $Q_{OLR}, Q_{LHF}$ showed decreasing trend from
January to April and later as the summer season approached the region, the
latent heat flux increased till June with heightened evaporation. Later from July,
the $Q_{LHF}$ term showed a decline till October and again increased towards the end
6.2 Evaluating the Heat Budget Terms

Figure 6.6: Climatological monthly mean fluxes at the surface of the Ocean: Net surface Flux ($Q_{SF}$), Net shortwave radiation ($Q_{SWR}$), Net Longwave radiation ($Q_{OLR}$), Net latent heat flux ($Q_{LHF}$) and Net sensible heat flux ($Q_{SHF}$)

of the year. The contribution of sensible heat flux is very feeble over the region as it did not show any variation and its magnitude remained near to zero indicating very minute contribution towards the heat budget of this region. As evidenced from the above figure, the net surface flux peaks during NEM season. A fact worth mentioning here are that all the fluxes showed positive values indicating gaining of heat throughout the year from the surface. Therefore this gain in heat from the surface has to be compensated by oceanic processes.

6.2.2 Oceanic Processes

Diffusion through the bottom and advection of heat together can be clubbed as the oceanic processes. The advective processes are classified into two kinds depending on the direction of flux. Firstly, flux on both north and southern boundaries. Here, in this region, depending on the direction of the currents over a year, it is assumed that the amount of flux entering from one end is leaving from the other end thereby making it sufficient to compute the flux at one boundary to know the total flux through the lateral boundaries. Second part of the advective fluxes is the cross shore flow at the eastern boundary resulting in upwelling /
downwelling along the eastern boundary. This vertical mass flux through the bottom of the coastal region is balanced by a vertical mass flux through the bottom of the rest of the control volume which is away from the coastal region. Such a process is called the coastal pumping [Shenoi et al., 2002]. This causes over turning and thereby removes heat from the control volume. Vertical mass flux at the western, northern and southern boundaries are ignored in comparison to the eastern boundary as it the region of coastal upwelling. Thus the net flux due to the oceanic processes is:

\[ Q_{OP} = Q_{mo} + Q_{cp} + Q_d \]  \hspace{1cm} (6.4)

where \( Q_{op} \), \( Q_{cp} \) and \( Q_d \) are the heat fluxes due to meridional over turning, coastal pumping and diffusion respectively.

**Meridional Overturning**  The transport along the lateral boundaries in the north-south direction is balanced by a mass flux through the bottom of the control volume. The difference between the vertical average of temperature at the bottom of the control volume results in a net flux of heat due to this process. This heat flux is estimated as:

\[ Q_{mo} = \rho_w C_p \frac{1}{A} \int v \Delta T_{mo} \, dx \]  \hspace{1cm} (6.5)

Where \( v \) is the transport per unit distance along the southern boundary (here only southern boundary is considered to know the flux that is entering the control volume from southern end and leaving from the northern boundary),

\[ \Delta T_{mo} = T_{sb} - < T > \]  \hspace{1cm} (6.6)

The integral is evaluated along the southern boundary. \( T_{sb} \) is the vertically averaged temperature (over the depth of the control volume) at the southern boundary, and \( \langle T \rangle \) is the basin wide average of the temperature at the bottom of the control volume. Both Ekman and geostrophic flows contribute to \( v \); where \( Q_{mo} = Q_{moe} + Q_{mog} \) and \( v = v_e + v_g \) and

\[ v_e = \frac{\tau_x}{\rho_w f} ; v_g = \frac{g D \partial \eta}{f \partial x} \]  \hspace{1cm} (6.7)
6.2 Evaluating the Heat Budget Terms

The subscripts ‘e’ and ‘g’ refers to Ekman and geostrophic flows respectively. The geostrophic current is assumed to constant over the depth of the control volume, is the zonal wind stress and is the sea surface height, is the Coriolis parameter, \( g = 9.81 \, m/s^2 \) is the acceleration due to gravity and \( D \) is the depth of the control volume considered to be 50m. Wind stress climatology computed from the QuikScat measured winds was used to estimate the zonal wind stress and the absolute dynamic topography obtained from AVISO was used to compute climatological sea level of the region.

From the figure (6.7), it is observed that none of the fluxes show any variation till the onset of SWM which is coincident with coastal upwelling in the region along the eastern boundary of the ocean. An exception is observed with geostrophic component of meridional overturning \( (Q_{mog}) \), showing heat gain (\( \sim 10W/m^2 \)) up to June, which can be attributed to oceanic flow into and out of the control volume. The heat loss due to meridional overturning owing to the geostrophic component is compensated by the heat gained through the Ekman component during the SWM season. \( Q_{mog} \) contributed to bulk of the heat gained in the region during October and November which started to decrease by December. From the above, it is be stated that Ekman is a major contributor to
6.2 Evaluating the Heat Budget Terms

heat (∼ 40Wm$^{-2}$) gain during the SWM season while geostrophy is the major contributor (∼ 90Wm$^{-2}$) during post SWM period. In short, heat gained due to Ekman component is being compensated by the heat lost due to geostrophy during the SWM. Thus it can be firmly stated that meridional overturning is not a significant contributor to heat lost during the upwelling season.

Coastal Pumping  Cross shore flux at the eastern boundary has to be compensated by the vertical mass flux through the bottom of the control volume over rest of the region. The difference in average temperature between the coastal region and the rest of the control volume at the bottom results in a flux of heat. This process is known as coastal pumping [Shenoi et al., 2002] and is derived as follows:

$$Q_{cp} = -\rho_w C_p \Delta T_{cp} \frac{1}{A} \int l u.n \, dl$$  (6.8)

Where $u$ is the cross-shore transport per unit distance along the coast, $n$ is the unit vector normal to the eastern boundary (positive out of the coast), $l$ is the coordinate along the coast and the integral is evaluated along the coast. $\Delta T$ is the average difference in temperature at the bottom of the control volume between the coast and the rest of the basin, arrived at as:

$$\Delta T_{cp} = < T_c > - < T_I >$$  (6.9)

Where $< T_c >$ is the average temperature along the coast and $< T_I >$ is the average temperature at the coast of the control volume. Like meridional overturning, coastal pumping is also a combination of Ekman pumping and geostrophic components, i.e., $Q_{cp} = Q_{cpe} + Q_{cpg}$ and $u = u_e + u_g$, where

$$u_e = \frac{K x n \tau}{\rho_w f}; u_g = \frac{g D}{f} \Delta \eta$$  (6.10)

where $\tau$ is the wind stress vector, $k$ is the unit vector normal to the earths surface, D is the depth of the control volume, and is the sea level. The dynamic topography obtained from merged altimetry data has been used for computing geostrophic component ($u_g$), of coastal pumping.

As suggested by Duing and Leetma [1980] for the entire Arabian Sea basin, herein, the SEAS also, coastal upwelling forced by Ekman pumping along the
6.2 Evaluating the Heat Budget Terms

Figure 6.8: Heat Fluxes due to Oceanic Processes Meridional Overturning, Coastal Pumping, Diffusion

eastern boundary of the ocean is the dominant term in the heat budget of the region during SWM. The upwelling triggered by the monsoon winds along the coast brings up the cooler subsurface waters towards the surface and thereby cooling the region by ∼ 2.5°C during the SWM (as observed in chapter 5). This results in heat loss due to coastal Ekman pumping (60 W m$^{-2}$) during the SWM.

For rest of the year, coastal Ekman pumping is non-existent as seen from figure (6.7). Geostrophic component of coastal pumping resulted in the heat gained over the region up to 30 W m$^{-2}$ during the summer monsoon. The region began to lose heat from August and reached its peak by November due to geostrophic component of coastal pumping.

**Diffusion** The flux of heat through bottom of the control volume due to diffusion process is computed as:

\[
Q_d = -\rho_w C_p k \frac{1}{A} \int_A \frac{dT}{dz} dS \tag{6.11}
\]

where \( k = 2 \times 10^{-4} m^2 s^{-1} \) is the eddy diffusivity coefficient, \( \frac{dT}{dz} \) is evaluated at 50m depth, and the integral is over the surface area of the basins. The choice of \( k \) has been considered to be constant over the region as followed by Shenoi.
et al. [2002]. Diffusion results in loss of heat from the control volume as seen in figure 6.8. The loss is minimum during February - March ($\sim 15 W m^{-2}$) and is maximum during September - October ($\sim 60 W m^{-2}$).

In SEAS, the dominant oceanic processes are coastal Ekman Pumping, Meridional overturning due to Ekman flow and diffusion during the summer monsoon. Among these, the Ekman components are directly influenced by winds, while the coastal pumping removes the heat, meridional over turning brings in heat to the control volume which is a new finding of this study. The same is the case with geostrophic components of coastal pumping and meridional overturning, when the geostrophic component ($Q_{mog}$) removes heat, $Q_{cpg}$ introduces heat during the summer monsoon. Hence, we attribute this as the reason for overall weak upwelling in the region when compared to the eastern boundary upwelling regions elsewhere. Diffusion also results in loss of heat from the control volume throughout the year and is the only component that is not influenced by winds. Here, among the overall oceanic processes, meridional overturning results in gain of heat flux into the region, while the coastal pumping and diffusion results in loss of heat.

6.3 Rate of Change of Heat

The resultant heat flux owing to surface fluxes and oceanic processes result in change of heat content in the control volume. The rate of change of heat is obtained using the left hand side of the equation 6.1. Both $Q$ and $Q_t$ show bi-modal distribution in the region (Figure 6.9) with minima during the SWM and the coincident upwelling season of JJAS. From net heat flux curve, it can be understood that all through the year, SEAS gain heat though with varying intensities. Thus for this region, it can be concluded that the heat flux from the surface is far greater than that can be compensated by oceanic process. The sources of heating (primarily the fluxes with considerable contributions from Ekman and geostrophic components of the flow) are found to be compensated by cooling from vertical diffusion and coastal upwelling. However, it is evident that upwelling is the major factor in regulating the heat budget of the region during the SWM. For
6.3 Rate of Change of Heat

The net heat flux \( Q = (Q_{sf} + Q_{op}) \) into and out of the control volume and the rate of change of heat \( Q_t \) for the rest of the year, only the surface fluxes contribute towards the heat budget while the oceanic processes are feeble.