# Heat budget of the Gulf of Aden: surface, advective and upwelling heat fluxes



Heat fluxes
Surface
Upwelling
Advective
Gulf of Aden

Flux thermique
Surface
Upwelling
Advection
Golfe d'Aden

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ABSTRACT

Surface heat fluxes for the Gulf of Aden have been estimated using bulk formulae. The annual averages of sensible, latent and net back radiation fluxes are $-16$, 43 and 52 W $m^{-2}$, respectively. The annual mean of observed solar radiation is 192 W $m^{-2}$, thus the annual heat surplus amounts to 113 W $m^{-2}$. This is equivalent to a total energy of $2.5 \times 10^{13}$ W, considering the surface area of the Gulf to be $220 \times 10^{3}$ $km^{2}$. Two processes seem to balance the heat gain: upwelling and water exchange with the Red Sea and with the Arabian Sea. Upwelling appears to be a major diffuser of the heat accumulated by the surface processes. About 48% ($1.2 \times 10^{13}$ W) of the heat gain is lost by upwelling. The net annual advective heat flow into the Red Sea due to water exchange at Bab-al-Mandab disperses about 34% ($0.84 \times 10^{13}$ W) of the heat input. The remaining 18% ($0.46 \times 10^{13}$ W) may be carried away into the Arabian Sea.

RÉSUMÉ

Bilan thermique du golfe d'Aden: flux superficiel, advection et upwelling.

Les flux superficiels de chaleur dans le golfe d'Aden ont été estimés à l'aide de formules globales. Les valeurs moyennes annuelles des pertes de chaleur sensible, latente et radiative sont respectivementt $-16$ W $m^{-2}$, 43 W $m^{-2}$ et 52 W $m^{-2}$. L'apport solaire annuel est 192 W.$m^{-2}$ , d'où un gain annuel de 113 W $m^{-2}$, soit au total $2{,}5 \times 10^{13}$ W pour le golfe ($220 \times 10^{3}$ $km^{2}$).
Deux phénomènes équilibrent ce gain de chaleur: l'upwelling et les échanges d'eau avec la mer Rouge et avec la mer d'Arabie. L'upwelling diffuse environ 48 % ($1{,}2 \times 10^{13}$ W) du gain thermique. L'advection transfère environ 34 % ($0{,}84 \times 10^{13}$ W) de la chaleur dans la mer Rouge par le détroit de Bab-el-Mandab. Le reliquat, 18 % ($0{,}46 \times 10^{13}$ W), passe dans la mer d'Arabie.

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## INTRODUCTION

The Gulf of Aden is an area of particular oceanographic interest in that it constitutes a transition zone between the Red Sea and the Arabian Sea. It extends eastward from the narrow Strait of Bab-el-Mandab to the line between Ras Baghashwa on the southern coast of Arabia and Ras Asir on the northern coast of the Somali peninsula (Fig. 1). The approximate surface area of the Gulf is $220 \times 10^{3}$ $km^{2}$ ($\approx$ length 785 km; $\approx$ width 280 km); its average depth, excluding the relatively narrow (40-50 km) continental shelf, is about 1800 m. Towards the south-eastern parts, the depth exceeds 3000 m (Wyrtki, 1971).

The climatology of the area is characterized by two distinct seasons: NE Monsoon (December to March), and SW Monsoon (May/June to September/October), with two transition periods. The SW Monsoon winds are strong and steady (Beaufort 4 to 5), while the NE Monsoon winds are weak. During the transition periods, the winds are weak and variable. Tropical cyclones only on rare occasions reach the Gulf of the Aden (Murty and El-Sabh, 1984). The surface circulation is mainly dependent on the changing monsoon winds, with average surface currents directed along the Arabian Coast into the Gulf of Aden and the Red Sea during winter and in the opposite direction during summer.

The hydrographic structure is also controlled by the reversal of the wind regime and water exchange with the Red Sea through the Strait of Bab-al-Mandab and the Indian Ocean. In the upper 1000 m, there are three water masses, namely Surface Water (SW), Subsurface Water (SSW) and Red Sea Water (RSW) (Piechura and Sobaih, 1986). The surface water (SW), formed locally, is characterized by high temperature and salinity and is well-oxygenated. The physicochemical properties of this water mass undergo pronounced spatial and temporal changes. In summer, this water mass is restricted to the upper 70-80 m layer; in winter it reaches a depth of 200-300 m. The subsurface water (SSW), distinguished by minimal salinity and oxygen, is believed to originate at the subtropical front in the Indian Ocean (Seriy, 1968). Although small, seasonal changes in its properties can still be detected, this water mass is observed with the lowest temperature and highest salinity during the SW Monsoon; the opposite occurs during the NE Monsoon (Piechura and Sobaih, 1986). The Red Sea Water (RSW) is characterized by high temperature, salinity and oxygen content. Seasonal changes in the RSW characteristics have also been observed, due to a variable water exchange through Bab-el-Mandab Strait (Bogdanova, 1966; Piechura and Sobiah, 1986). At greater depths down to 2500 m, the upper Indian Ocean deep water is observed. Close to the bottom (3000-4000 m), low temperature and salinity water probably originating in the Antarctic region is found.

During the NE Monsoon, winds drive the surface water towards Bab-el-Mandab Strait, particularly along the Arabian coast. Along the Somali coast, there is a weak current in the opposite direction. During the SW Monsoon, the surface water generally moves in the opposite direction. The longshore wind stress causes a net offshore Ekman transport and upwelling of subsurface water occurs at different points along the Arabian and Somali coasts. During this season, three upwelling regions with varying intensity and mechanism have been identified in the Arabian Sea (Prell and Streeter, 1982; Swallow, 1984; Savidge *et al.*, 1990). In consequence, the Arabian Sea including the Gulf of Aden is characterized by a pronounced cooling of the upper mixed layer (Wyrtki, 1971; Ramage *et al.*, 1972; Fieux and Stommel, 1976; Hastenrath and Lamb, 1980; Duing and Leetmaa, 1980). The extent of cooling depends on the intensity and duration of upwelling, which vary from one year to another. Swallow and Bruce (1966) showed that the lowest temperature in 1964 was 13.2 °C. In 1979, the lowest temperature observed was 17 °C (Swallow *et al.*, 1983). In the Gulf of Aden, the lowest temperature, based on 50 years of data, was 24 °C (Sharaf el-Din and Mohammed, 1984). The distribution of surface temperature in August 1984 shows three partially-separated sources of coastal upwelling in the Gulf, with the lowest temperature below 17 °C in the Mukalla region (Piechura and Sobaih, 1986). In the present study, the lowest temperature observed in July 1986 was 22.6 °C.

A striking feature of the upwelling along the coast of Yemen is its extent over a wide area, up to nearly 150 miles (Bottero, 1969; Currie *et al.*, 1973; Swallow, 1984). As a consequence, the upwelled waters are supplied from a much greater depth than is usual in other coastal upwelling regions (Currie *et al.*, 1973).

Undoubtedly, upwelling plays an important role in maintaining the heat budget of the Arabian Sea (Hastenrath and Lamb, 1980). Duing and Leetmaa (1980) considered the total heat loss during the SW Monsoon cooling to be essentially balanced by three processes: positive heat gain from the atmosphere; negative northward heat flux across the equator; and heat loss due to upwelling. These authors concluded that the rate of cooling due to upwelling was 2-3 times greater than that from the advection of cool surface water across the equator, and more than sufficient to offset the gain of heat from the atmosphere.

The heat budget of the Gulf of Aden seems to be governed by three processes: surface heat fluxes; heat exchange with the Red Sea and the Arabian Sea; and upwelling. The present study is based on monthly means of sea-surface temperature and meteorological parameters obtained during the year 1986. The objective of the investigation is to calculate the surface heat fluxes and to assess the relative importance of the various processes contributing to the heat budget of the Gulf of Aden. It must be admitted that the estimates of the individual terms are necessarily crude, and possess large margins of error. However, these estimates can shed some light on the relative importance of the various processes.

## MATERIAL AND METHODS

### Data source

The present study is based on meteorological data obtained from Riyan airport and sea-surface temperatures at two stations, Ash-Shihir and Mukalla, on either side of the airport, during 1986. Riyan airport is not far from the coast and the winds are mainly is an onshore direction; therefore the data will be representative of a coastal stations. Figure 1 shows the study area and the station positions. At Riyan airport, the meteorological parameters were measured three-hourly. Monthly means of the parameters were obtained from the daily means. Water temperatures were measured on a weekly basis at the two stations, and monthly means were obtained from the weekly measurements (Ormond and Banaimoon, 1994). The sources of error include the variability of the upwelled water temperature and the estimates of the meteorological data at the coastal station. Previous studies indicate that the upwelled temperature varies from 13 °C to 24 °C. The dominant wind direction in the area is NE during winter and SW during summer (Currie *et al.*, 1973; Hastenrath and

Lamb, 1980). Therefore, the prevailing winds are mainly from the sea and the meteorological data at the coastal station can be considered to represent the sea conditions.

### Surface heat fluxes

The sensible heat flux $Q_h$ and the evaporative flux $Q_e$ are estimated using the Bulk Aerodynamic method (Pond, 1975). The equations in their usual form are:

$$Q_h = \rho_a C_p C_T (T_s - T_a) W \quad (1)$$

$$Q_e = \rho_a L C_E (q_s - q_a) W \quad (2)$$

where $\rho_a$ is the density of air (1.2 kg m$^{-3}$), $C_p$ is the specific heat of air at constant pressure ($\approx$1050 J kg$^{-1}$ °C$^{-1}$), $C_T$ is the heat transfer coefficient, $T_s$ is the sea-surface temperature, $T_a$ is the air temperature, $W$ is the wind speed, $L$ is the latent heat of evaporation (2.45 $\times$ 10$^6$ J kg$^{-1}$), $C_E$ is the latent heat flux coefficient, $q_s$ is the saturated specific humidity at sea-surface temperature and $q_a$ is the specific humidity in the air. The saturated specific humidity at sea-surface temperature and the specific humidity at the air temperature are related to saturated vapour pressure at the sea-surface temperature and the vapour pressure at the air temperature as :

$$q_s = 0.622\, e_s/p$$

$$q_a = 0.622\, e_a/p$$

The evaporative and sensible heat fluxes depend on the choice of the transfer coefficients $C_E$ and $C_T$. In constructing the "Atlas of the heat balance of the earth" Budyko (1963) used a constant value of 2.1 $\times$ 10$^{-3}$ for calculating evaporation from the world ocean. This value is favoured by Banker *et al.* (1982) and by Ahmad and Sultan (1987) in their calculations of the heat balance of the Red Sea. In Robinson's review (Robinson, 1966), a value of (2.4 ± 0.4) $\times$ 10$^{-6}$ g cm$^{-3}$ for $\rho_a C_E$ was given. Smith (1980) and Large and Pond (1982) found the sensible heat flux coefficient to be larger in unstable than in stable conditions. Masagutov (1981) proposed similar coefficients for heat and moisture fluxes for wind speed from 2 to 21 m s$^{-1}$ and water-air temperature differences from – 4 to 4 °C. Blanc (1985) summarized ten selected bulk transfer schemes, in which the coefficients $C_E$ and $C_T$ vary with wind speed and water-air temperature differences. These coefficients are intended for use with hourly means of vector winds, humidity and water-air temperature difference. However, Marsden and Pond (1983) concluded that these coefficients are also valid for monthly means if scalar-average wind speeds are used. Based on the monthly averages of water-air temperature differences and wind speed in the Gulf of Aden, the monthly values of transfer coefficients were calculated from the table given by Masagutov (1981). From the monthly values the averages of the coefficients are:

$$C_T = C_E = 1.02 \times 10^{-3}$$

substituting the numerical values for the various parameters given above the bulk aerodynamic equations for $Q_h$ and $Q_e$ (both in W m$^{-2}$) reduce to

$$Q_h = 1.3 (T_s - T_a) W \quad (3)$$

$$Q_e = 1.87 (e_s - e_a) W \quad (4)$$

where $e_s$ and $e_a$ are the saturated vapour pressure at the sea-surface temperature and actual vapour pressure at the air temperature respectively, these values being expressed in millibars. Equation (4) for the calculation of evaporative flux is applied when $e_s$ is greater than $e_a$. But when $e_s$ is less than $e_a$ the following equation is used (Ramage, 1971).

$$Q_e = 3.7 \times 10^{-5} (444 - 0.56\, T_s) (0.98 e_s - e_a) \quad (5)$$

During the three-month period from July to September, $e_s$ is less then $e_a$. Therefore, during this period Eq. (5) was used to estimate the evaporative heat flux. During the remaining months, when $e_s$ was greater than $e_a$, equation (4) was used to compute the evaporative flux.

The effective back radiation depends on the sea-surface temperature, humidity and the amount of cloud cover. The formula given by Budyko (1974) is applied to estimate the monthly mean net long-wave radiation.

$$Q_b = \varepsilon \sigma T_s^4 (0.39 - 0.05 \sqrt{e_a}) (1 - c \cdot n^2) \quad (6)$$

where $\varepsilon$ is the coefficient of emissivity with an average value of 0.985 (Krauss, 1972), $\sigma$ is the Stefan-Boltzman constant equal to 5.74 $\times$ 10$^{-8}$ J m$^{-2}$ s$^{-1}$ °C$^{-4}$, $T_s$ is the absolute sea-surface temperature, $e_a$ is the water vapour pressure in millibars, $c$ is a tabulated function of latitude which runs from 0.5 at the equator to 0.82 at 75 °C latitude and $n$ is the cloud cover. Equation (6) reduces to:

$$Q_b = 5.7 \times 10^{-8}\, T_s^4 (0.39 - 0.05 \sqrt{e_a}) (1 - 0.54\, n^2) \quad (7)$$

In the absence of recorded solar radiation in the study area, solar radiation figures at two stations, Sabya and Sirr Lason (Fig. 1) (Ministry of Agriculture and Water Resources, 1991), on the border between Yemen and Saudi Arabia, are used.

## RESULTS AND DISCUSSION

The monthly means of sea-surface temperature, air temperature, wind speed, relative humidity and cloud cover for the year 1986 are reported in Table 1 and shown in Figure 2.

The effect of the SW Monsoon and the associated upwelling on the sea-surface and air temperatures is clearly discernible. The sea-surface temperature increases steadily from February (24.9 °C) to reach maximum in June (33.1 °C) and then drops sharply to 22.6 °C in July. In fact, the surface temperature in July is the lowest during

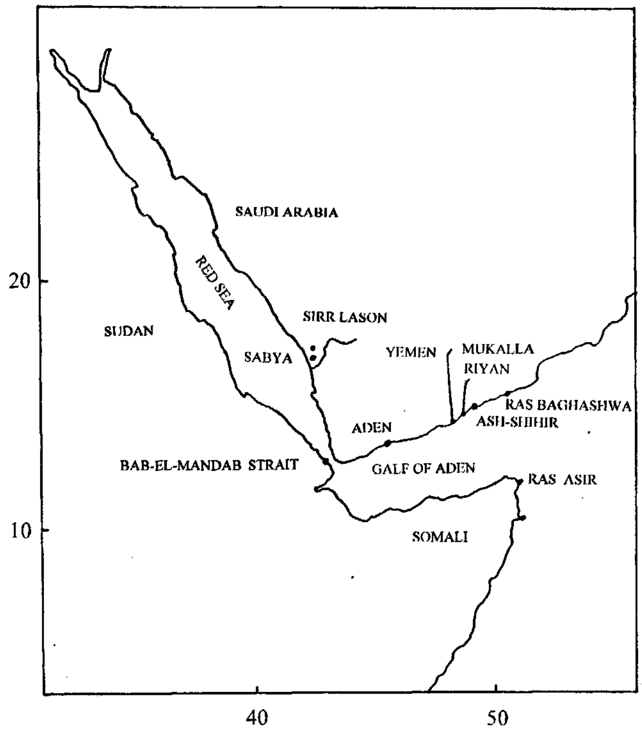


Figure 1

*Map of the study area showing the station positions.*

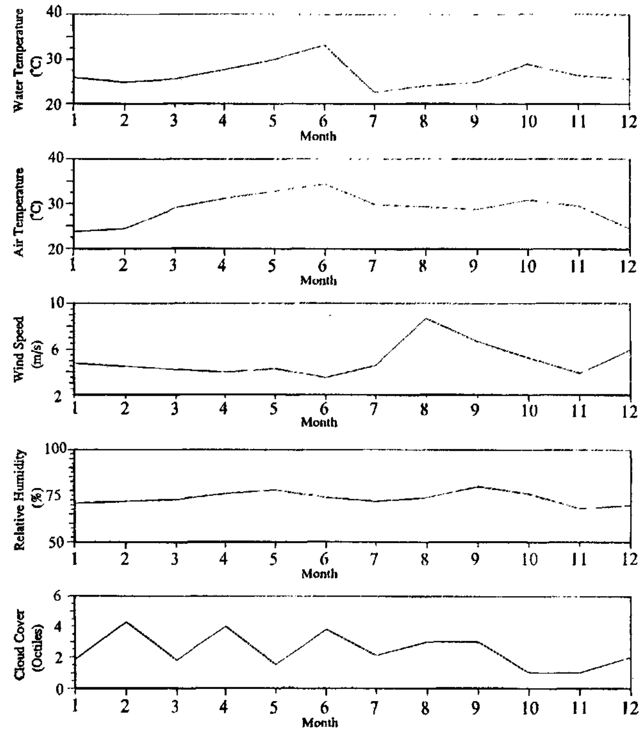


Figure 2

*Monthly means of sea-surface temperature, air temperature, wind speed, relative humidity and cloud cover for the Gulf of Aden during 1986.*

the whole year. Towards the end of the SW Monsoon, the sea surface temperature begins to increase, until October (transition month), when it reaches 28.9 °C and then decreases with the onset of winter. The air temperature displays a similar pattern, increasing from winter and attaining a maximum value in June (34.4 °C). During the SW Monsoon (July to September), it decreases to reach a low value in September (28.7 °C). However, the drop in the air temperature is not as abrupt as that of the sea temperature. Once again, air temperature increases in October and then decreases towards winter. Unlike that of the water temperature, minimum air temperature is observed in January (23.7 °C). It would appear that during summer the decrease in air temperature is a reaction to decreasing sea-surface temperature, and that in winter, the decrease in sea-surface temperature is a reaction to decreasing air temperature and the increasing loss of latent heat due to the reduced relative humidity which could be due to the advection of dry air from the north. The relative humidity is also affected by the SW Monsoon, but to a lesser degree. Figure 2 shows that the relative humidity is slightly reduced during this period. This decrease cannot be attributed to a change in air temperature, as any decrease in the latter will increase the relative humidity.

The sensible heat flux is calculated using Eq. (3). The latent heat flux is computed by either Eq. (4) or Eq. (5) depending on whether $e_s$ is greater or less then $e_a$. The effective back radiation is estimated using Eq. (7). The monthly means of the various heat fluxes, together with the solar radiation, are reported in Table 2.

Table 1

*Montly means of sea-surface temperatures ($T_s$), air temperature ($T_a$), wind speed (W), relative humidity (RH) and cloud cover (n) for the Gulf of Aden.*

| Month | $T_s$ (°C) | $T_a$ (°C) | W (m s$^{-1}$) | RH (%) | n (octiles) |
|---|---|---|---|---|---|
| January | 25.9 | 23.7 | 4.8 | 71 | 1.8 |
| February | 24.9 | 24.4 | 4.5 | 72 | 4.3 |
| March | 25.6 | 29.4 | 4.2 | 73 | 1.8 |
| April | 27.7 | 31.2 | 4.0 | 76 | 4.0 |
| May | 29.9 | 32.8 | 4.3 | 78 | 1.5 |
| June | 33.1 | 34.4 | 3.5 | 74 | 3.8 |
| July | 22.6 | 29.8 | 4.6 | 72 | 2.1 |
| August | 24.0 | 29.3 | 8.7 | 74 | 3.0 |
| September | 24.9 | 28.7 | 6.7 | 80 | 3.0 |
| October | 28.9 | 30.8 | 5.3 | 76 | 1.0 |
| November | 26.4 | 29.5 | 3.9 | 68 | 1.0 |
| December | 25.5 | 24.4 | 6.0 | 70 | 2.0 |

Figure 3 displays the monthly mean changes of the various surface heat fluxes and the net heat gain/loss. Computation of the heat fluxes shows that the Gulf loses heat by conduction during the peak of the NE Monsoon (December to February). During the remainder of the year, the Gulf gains heat, with maximum heat being received during the peak of the SW Monsoon (July to September). The reduction of sea-surface and air temperatures induced by the upwelling of cold water results in sensible and evaporative heat fluxes that are characterized by two cycles during the year (Fig. 3). The annual average of sensible heat amounts to –16 W m$^{-2}$.

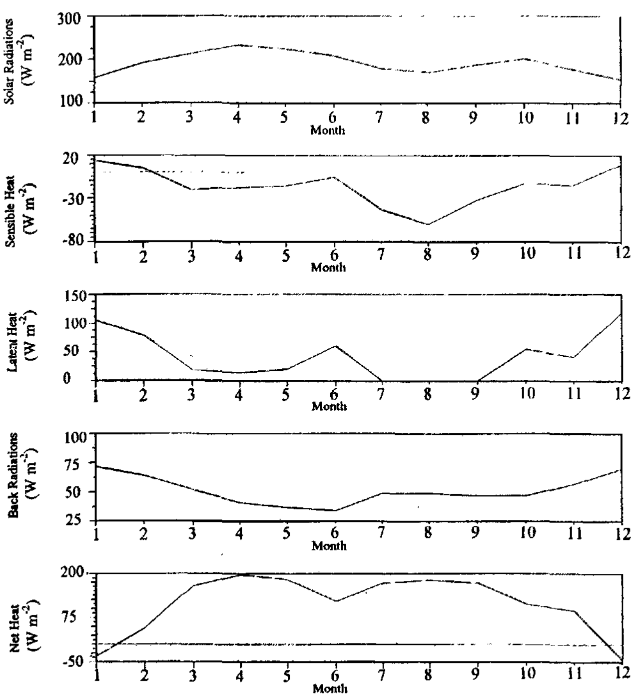


Figure 3

*Monthly means of solar radiation, sensible heat, latent heat, back radiation and the net heat for the Gulf of Aden during 1986.*

The latent heat flux varies greatly in the course of the year. The highest value occurs during the NE Monsoon period (December to February) and then decreases sharply to reach 14 W m$^{-2}$ in April. Just before the onset of the SW Monsoon (June), the loss of heat by evaporation increases to 60 W m$^{-2}$. This figure is reduced to zero during July to September, and then increases towards winter. This is to be expected, as the drop in the sea-surface temperature reduces the air temperature at the boundary layer and stabilizes the air over the sea. The annual average of evaporative flux is 43 W m$^{-2}$, significantly lower than the averages for the Red Sea (169 W m$^{-2}$) and the Arabian Gulf (168 W m$^{-2}$) (Ahmad and Sultan, 1987, 1989, 1991), and also less than the evaporative heat flux in the Gulf of Oman (100 W m$^{-2}$) (Sultan and Ahmad, 1993). The large difference in the evaporative heat flux between the Gulf of Aden and the adjacent waters may be attributed to the influence of upwelling.

The effective back radiation is also higher during the NE Monsoon period, and relatively lower during the SW Monsoon. The back radiation displays slight seasonal changes and appears to be less vulnerable to the semi-annual variation of water temperature induced by upwelling (Fig. 3). The annual average of the net back radiation amounts to 52 W m$^{-2}$.

The annual average of recorded solar radiation obtained from nearby stations is 192 W m$^{-2}$. Thus, an annual heat surplus of 113 W m$^{-2}$ is gained through the surface. This value is slightly higher than that of Hastenrath and Lamb (1980), who give an average value of 100 W m$^{-2}$ in their calculation of the heat budget for the Indian Ocean. However, the latter represents the long-term annual mean (1911-1970). The difference between the present estimate and that given by Hastenrath and Lamb (1980) can be attributed to the lower rate of evaporation in the Gulf of Aden in comparison with the adjacent waters. The evaporation depends on the choice of exchange coefficient $C_E$, water-air temperature difference and wind speed. In the present study, a small value ($C_E = 1.02$) for the exchange coefficient has been used according to the monthly water-air temperature difference and wind speed. The decrease in surface temperature caused by upwelling of deep cooler water appears to contribute significantly to the heat budget

Table 2

*Monthly averages (in W m$^{-2}$) of absorbed solar radiation $Q_s$, computed values of sensible heat flux $Q_h$, latent heat flux $Q_e$, effective back radiation flux $Q_b$ and net heat flux at the air-sea interface $Q_t$.*

| Month | $Q_s$ | $Q_h$ | $Q_e$ | $Q_b$ | $Q_t$ |
|---|---|---|---|---|---|
| January | 158 | 14 | 106 | 72 | - 34 |
| February | 192 | 5 | 78 | 64 | 45 |
| March | 213 | - 21 | 18 | 52 | 164 |
| April | 234 | - 18 | 14 | 41 | 197 |
| May | 224 | - 16 | 20 | 37 | 183 |
| June | 208 | - 6 | 60 | 34 | 120 |
| July | 179 | - 43 | 0 | 49 | 173 |
| August | 170 | - 60 | 0 | 49 | 181 |
| September | 187 | - 33 | 0 | 47 | 173 |
| October | 201 | - 13 | 54 | 47 | 113 |
| November | 177 | - 15 | 41 | 57 | 94 |
| December | 154 | 9 | 119 | 70 | - 44 |
| Annual mean | 192 | - 16 | 43 | 52 | 113 |

of the Gulf. Such a decrease would increase the heat gain by conduction and reduce the evaporative flux, thus increasing the net heat gain. The surface temperature during the SW Monsoon in 1986 is less than the long-term average. On the other hand, Duing and Leetmaa (1980) found that the average annual heating rate above and beyond the seasonal fluctuations in the Arabian Sea equals 24 W m$^{-2}$. These authors also showed that upwelling plays a major role in the heat balance of the Arabian Sea. Further east in the Gulf of Oman, the annual average of the surface heat flux is estimated to be 55 W m$^{-2}$ (Sultan and Ahmad, 1993).

The net surface heat gain or loss (Fig. 3) displays a pattern similar to that of sea-surface and air temperatures, except for the month of June. In winter (December and January), the sea loses heat, whereas it gains heat during the remainder of the year. The heat gain increases to reach a maximum in April (transition month), decreases in June and then increases during the SW Monsoon (July to September). Finally it decreases gradually from October, and by December it becomes negative. A semi-annual variation in the net heat gain/loss is evident in Figure 3.

Now, taking the average area of the Gulf as 220 × 10$^3$ km$^2$, the total heat received annually is 2.5 × 10$^{13}$ W. The heat accumulated by the surface processes must be balanced by other processes. Water exchange with the Red Sea and the Arabian Sea is one mechanism that can disperse some of the accumulated heat. Upwelling of cold water appears to be another mechanism that can diffuse heat efficiently. The estimation of the heat carried away by these processes is discussed below.

## Effect of advection

On the basis of fifty years of hydrographic data, Sharaf el-Din and Mohammed (1984) have studied the seasonal vertical temperature distribution in the southern Red Sea and the Gulf of Aden. Ahmad and Sultan (1989) used these data and computed an advective oceanic heat flow of 19 W m$^{-2}$ into the Red Sea at Bab-el-Mandab by taking into account the average temperature difference between the upper and lower layers and the variations of the in- and out-flows. Taking the surface area of the Red Sea as 0.44 × 10$^6$ km$^2$, this is equivalent to an annual heat advection of 0.84 × 10$^{13}$ W into the Red Sea.

## Effect of upwelling

During the SW Monsoon, the winds blow almost parallel to the southern coast of Yemen. The Ekman transport associated with the wind-driven currents is directed to the right of the wind stress. Hence the surface layer is skimmed away from the coast in the offshore direction. The continuity of mass requires upwelling of colder subsurface water. The rate of upwelling along the coast is given by $\tau L\, f^{-1}$, where $\tau$ is the mean wind stress during the SW Monsoon, $L$ is the length of the coastline and $f$ is the Coriolis parameter. The heat loss due to upwelling can be estimated by the equation below.

$$Q_{upw} = \tau\, L\, f^{-1}\, \Delta T\, C_p \tag{8}$$

In Eq. (8), $\Delta T$ is the temperature difference between the coastal and offshore waters and $C_p$ is the specific heat of sea water at constant pressure. The temperaure $\Delta T$ is taken at the position where the horizontal temperature gradient is minimum. The calculated value of the wind stress ($\tau = \rho_a\, C_D\, w^2$ where $C_D$ depends on the wind speed and has an average value of 1.4 × 10$^{-3}$ in the present case) during this period is 0.5 dynes cm$^{-2}$. According to the above, the length of the coastline where pronounced upwelling and intensive cooling takes place is approximately 300 km and the temperature difference between coastal and offshore waters is 6 °C (Sharaf el-Din and Mohammed, 1984). The Coriolis parameter is evaluated at 12° N latitude. Substituting the values for the various parameters in Eq. (8) results in a total heat loss of 1.2 × 10$^{13}$ W.

On an annual basis, the difference between the surface heat gain and the loss due to advection into the Red Sea and upwelling is therefore 0.46 × 10$^{13}$ W. This heat must be carried away by water exchange with the Arabian Sea. Unfortunately, the present data do not permit direct estimation of the heat advected from the Gulf of Aden into the Arabian Sea. However, they do demonstrate the importance of the heat transported by currents in maintaining the heat budget of the Gulf, and also show that upwelling constitutes a major diffuser of the accumulated heat: annually, about 48 % of the heat input by the surface processes is diffused by upwelling. Upwelling has been thought to be the dominant factor in maintaining the heat balance of the Arabian Sea (Duing and Leetmaa, 1980; Hastenrath and Lamb, 1980; Swallow, 1984).

One important aspect of upwelling is the subsequent fate of the upwelled waters, and the associated question of how the cold upwelled water is replenished. There is some uncertainty concerning the fate of this water. For the Arabian Sea as a whole, Swallow (1984) indicates that any upwelled water entrained by the boundary current is carried offshore and resubmerged below less dense surface water. Duing and Leetmaa (1980) suggested several important processes at work in balancing the upwelled water. These authors believe winter cooling of the surface water at the northwestern end of the Arabian Sea and its subsequent sinking and spreading southward to be a possible mechanism.

For the Gulf of Aden, water exchange with the Red Sea does not appear to compensate the upwelled water. In fact, the annual volume of water advected to the Red Sea in the surface layer exceeds that advected into the Gulf in the subsurface layer by about 10 % (Ahmad and Sultan, 1989). Additionally the Red Sea water can be traced in the Arabian Sea, which signifies a further loss of water from the Gulf. The replacement of the upwelled subsurface water could be accomplished by two processes: submerging of the upwelled water offshore, and water exchange with the Arabian Sea. The relative contribution of these processes cannot be assessed on the basis of the present data. However, the present study does elucidate the importance of water exchange with the Arabian Sea in maintaining the heat budget of the Gulf. About 18 % of the heat gain by the surface processes seems to be taken away by advection to the Arabian Sea. On the other hand, the presence of different water masses – the subsurface water mass (SSW) originating at the subtropical front of the Indian Ocean (Seriy, 1968), the upper Indian deep water and bottom water from the Antarctic region – provides clear evidence of the importance of water exchange in replacing upwelled water.

## Error analysis

The present computations are crude, and may possess large uncertainties that preclude an exhaustive description of the heat budget of the Gulf. However, despite these disadvantages, the results of this study have been instructive, particularly in showing that upwelling contributes significantly to the cooling of the Gulf. An estimate of the errors are given below.

Uncertainties in the estimates of surface heat fluxes depend on the choice of coefficients in the bulk formulae and on the errors inherent in such heterogeneous oceanographic and meteorological data. The standard deviations of the fluxes are calculated using the equation of the form (Weare *et al.*, 1981)

$$S = \left[ \left(\frac{\partial \nu}{\partial x}\right)^2 S_x^2 + \left(\frac{\partial \nu}{\partial y}\right)^2 S_y^2 + \left(\frac{\partial \nu}{\partial z}\right)^2 S_z^2 \right]^{1/2}$$

where $S$ is the standard deviation, $v$ is a dependant variable (sensible heat or evaporation or back radiation) and $x, y, z$ are independent variables such as $(T_s - T_a)$, $(e_s - e_a)$, wind speed and coefficients etc. upon which the sensible heat, evaporation or back radiation depend.

In the absence of actual standard deviations in the monthly oceanographic and meteorological data sets, it is assumed that the standard deviations are 10 % of the annual mean oceanographic and meteorological data. Therefore the values are:

$(T_s - T_a) = -2.4 \pm 0.2$ K

$(e_s - e_a) = 5.7 \pm 0.6$ mb

$w = 5.0 \pm 0.5$ m.s$^{-1}$

$C_E = C_T = (1.02 \pm 0.1) \times 10^{-3}$

$T_s = 297.2 \pm 2.4$ K

$e_s = 30.0 \pm 3.0$ mb

$\tau = 0.5 \pm 0.05$ dynes.cm$^{-2}$

$L = 300 \pm 30$ km

$\Delta T = 6 \pm 0.6$ K

Based on the above standard deviations, the error in the annual averages flux estimates are: ± 3, ± 6 and ± 4 W m$^{-2}$ respectively for sensible heat, evaporation and back radiations.

For $Q_{upw}$ the error is $\pm 0.2 \times 10^{13}$ W.

These values will not affect the results significantly.

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