Response of subsurface waters in the eastern Arabian Sea to tropical cyclones

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1. Introduction

Passage of a cyclonic storm is an epitome of strong interaction between the atmosphere and the ocean. A moving storm, even though relatively of short time-scale, is a means of intense localized surface wind stress. It generates several different modes of oceanic response which can be divided into two stages, namely, the forced and the relaxation stage. The forced stage baroclinic response includes the mixed layer currents (Sanford et al., 1987) and cooling of surface layers due to vertical mixing (Black, 1983). The barotropic response consists of a geostrophic current and an instantaneously set up associated trough in sea surface height. The typical time-scale for the forced stage response is of the order of half-a-day. The relaxation stage response following the storm passage is an inherently non-local baroclinic response to the wind stress. The energy of the mixed layer currents is dispersed in the spreading wake of near inertial internal waves (Gill, 1984) that penetrate into the thermocline (Shay and Elsberry, 1987; Dickey et al., 1998) leaving behind a baroclinic geostrophic current along the storm track. As the cyclone passes, large inertial currents are generated in the upper ocean and large-amplitude temperature oscillations are set up in the thermocline below the mixed layer. These are basically internal gravity waves with inertial period. The restoring force for these gravity waves is proportional to the product of gravity and the density difference between the adjacent layers. At internal interfaces this difference is much smaller than the density difference between air and water. Internal waves can therefore attain much larger amplitudes than surface waves. It also takes longer time for the weak restoring force to return particles to their mean position. At the surface, the inertial oscillations may take less time (about a week) to reach normal state as the restoring force is large. While, the subsurface temperature oscillations take longer period (about two weeks) as the restoring force is small. Hence, the time-scale of relaxation stage response is of about two weeks. The most important response of the tropical storm and the open ocean interaction is the marked cooling of the upper ocean. The SST anomalies forced by the tropical cyclone have been observed to vary from 1 to 6 °C (Black, 1983; Sadhuram, 2004). The SST cooling also depends upon the translation speed of the cyclone. The slow, medium and fast moving cyclone results in a surface ocean cooling of high, medium
and low respectively (Bender and Ginis, 2000; Rao et al., 2007). The recent study of Oey et al. (2006, 2007) describes hurricane-induced motions and its interaction with loop current warming by Hurricane Wilma.

The cyclones that generally form in the Arabian Sea (AS) in May and October originate in the southeast while; the ones in June ensue over the central/east central region. A few cyclones of the Bay of Bengal (BOB) travel across the peninsula and emerge into the AS and may again intensify into cyclonic storms. There are few records of severe cyclonic storms in the northeastern AS covering Gujarat and Maharashtra coasts. The modeling studies related to response of the ocean due to passage of cyclone over this region are limited. Murthy and Hareesh Kumar (1991) studied the response of coastal waters off south of Maharashtra to a deep depression in 1986 using one-dimensional model. Chittibabu et al. (2000) and Jain et al. (2006) used barotropic models to simulate surface circulation and associated storm surges along the eastern AS. Deo et al. (2002) studied the upper ocean response to tropical cyclones in the Indian Ocean using 1½-layer wind-driven reduced gravity ocean model.

Ly and Kantha (1993) and Ly (1994) used Princeton Ocean Model (POM) to study the nonlinear interaction of Hurricane Camille and Hurricane Frederic with the Gulf of Mexico Loop Current. In their study the model considers both barotropic and baroclinic effects. Keen and Glenn (1999) studied shallow water currents during Hurricane Andrew. It was found that the presence of downwelling favorable flow regime on the right and upwelling favorable flow on the left of the storm track as the eye of the cyclone made the landfall. Modeling studies made by Mahapatra et al. (2007) suggests the maximum cooling shifts to the left of the cyclone track just before the cyclone crosses the coast.

The turbulent mixing and entrainment of subsurface cold water is responsible for sea surface cooling in deep water (Price, 1981). Shen and Ginis (2001, 2003) found that the hurricane-induced sea surface cooling is insensitive to the ocean depth if it is much deeper than the mixed layer and the hurricane-induced sea surface fluxes dominate the surface cooling in the shallow regions. Zedler et al. (2002) studied the upper ocean response to hurricane Felix. It was observed that during the passage of cyclone there is cooling in the upper 30 m and warming at depths of 30–70 m. In their study they tried to resolve the temperature evolution using one-dimensional mixed layer model. The observational data in the eastern AS at a deep buoy location also suggests similar subsurface warming during the passage of a cyclone. Hence, two historical cyclones of June 1998 and May 1999 are chosen for this study that made landfall in the western Gujarat coast in order to demonstrate the
evolution of ocean response and possible mechanism for subsurface warming using the POM.

2. Data

The available cyclone tracks of 1998 and 1999 as shown in Fig. 1 along with its intensity information are obtained from the India Meteorological Department (IMD) reports. Real-time measurements of marine meteorological and oceanographic parameters are acquired at DS1 (Deep Sea) buoy location (15.5 N, 69.2 E) in the AS (Fig. 1). The water depth at the buoy location is about 3500 m. The numbers on the track give the dates of the cyclone position at 0600 UTC. These cyclones have almost the same intensity of about 40 m s\(^{-1}\) at the time of crossing the buoy. However, the translation speed of these cyclones is highly variable and increases its speed as the cyclone approaches towards the coast. On an average, the translation speed is about 9.5 km/h and 10 km/h for 1998 cyclone and 1999 cyclones respectively. However, the speed of the 1998 storm is slightly higher to that of 1999 near the DS1 location. It is to be noted that 1998 cyclone track is to the left of the buoy whilst; the 1999 cyclone passes through the buoy location.

The meteorological parameters such as air pressure and winds are measured at 3 m above the sea surface while; surface currents and SST are observed at 3 m below the sea surface. The detailed description of the buoy data measured is reported in Premkumar et al. (2000). The UCM 60 ultrasonic current meter is used to measure current speed and its direction. Ultrasonic pulses are simultaneously propagated in opposite directions through a water column with defined path length, and the arrival time difference of these pulses is measured with high precision. The transit time for these pulses is dependent on the fluid velocity component along the same path as the pulse is propagating, and the difference in transit time for the two pulses simultaneously sent in opposite directions will consequently be a direct expression for the current speed resolved along the same axis. Since current speed and its direction are measured at 3 m below the sea surface with ultrasonic technique, there is no interference with wave. Apart from these parameters, one of the critical parameters i.e. water temperature at different depths (3, 8, 13, 8, 23, 28, 33, 38, 43, 46, 53, 58, 73, 98, 123 m) is measured using thermistor chain (Aanderaa make, model: 3678B, 2000 \(\Omega\) Fenwall thermistor) fitted along with the mooring rope. The surface data are transmitted in every 3 h interval to shore for real-time application, whereas subsurface data from thermistor string are stored onboard system. While servicing the buoy, the onboard data are acquired and processed to obtain the temperature at different depths for every 3 h. The temperature at different depths is from the averaged values from 10 min data obtained once in 3 h interval. Though the buoy data is available for every 3-h interval, the comparison in the present study is made at every 12-h interval.

These observations provide an opportunity to understand the variation of temperature and current due to passage of the cyclone. Fig. 2 depicts the observed time series of temperature at different depths for the 1998 cyclone at DS1 location. The analysis of the data suggests that cooling in the surface layers is found up to 30 m depth and warming beneath. Strong inertial oscillations generated by the
cyclone in the subsurface layers are observed with a periodicity of 2 days.

3. Numerical model

The POM model (described in details by Blumberg and Mellor, 1987) is implemented for the eastern AS. It uses the 2.5 level turbulence closure scheme described by Mellor (2001). The model is fitted with realistic bottom topography from ETOPO5 of NGDC. As shown in the Fig. 1, the analysis area (bounded by dashed line) extends approximately from 5°N to 25°N along the eastern AS which is almost parallel to the coast with a maximum offshore extent of ~800 km on the southern open boundary and reduces towards north. A bilinear smoother is used in order to reduce sharp bottom slopes and associated pressure gradient errors in sigma coordinated models (Haney, 1991; Mellor et al., 1994, 1998). There are about 175 × 250 grid points in the horizontal and 26 computational levels in the vertical. The curvilinear coordinates are positive in the eastward and northward directions. Resolution in the zonal direction varies from 6 to 9 km with finer resolution near the coast while; it is 7–11 km in the meridional direction. In the vertical, a terrain following sigma coordinate is used with fine resolution of about 0.5 m and 15 m near the surface and the bottom respectively at a typical local depth of 500 m while; a relatively coarse grid of about 30 m is used in the mid-depths. The horizontal time differencing is explicit whereas the vertical time differencing is implicit. It uses mode splitting technique with a barotropic time step of 11.25 s and a baroclinic mode time step of 450 s. Horizontal boundary conditions over the land are implemented by a land mask, which ensures that the velocity over the land and the normal velocity along the coastline are zero. Radiation conditions along the open boundaries are used which allow disturbances generated in the interior to travel outwards in the form of progressive waves when they reach the boundaries. For temperature and salinity, upstream advection scheme is applied.

4. Numerical experiments

The study is confined to the simulations of temperature and currents during and after the passage of the cyclone to understand the mechanism of surface cooling and the subsurface warming on either side of the cyclone track. Initial data fields of temperature and salinity are obtained from the World Ocean Atlas 2001. Thermal forcing is used from Lisan and Weller (2007) heat flux data at 1° × 1° resolution in terms of daily incoming solar radiation and the net heat flux. The wind stress at the model surface is derived from ECMWF winds (2.5° × 2.5° gridded data). With these initial data fields, the model integration is started about 15 days before the cyclone forms over the ocean. It is to be noted that neither ECMWF nor high resolution scatterometer winds are able to capture the high intensity of the cyclone. Hence, the cyclonic wind field is generated using the asymmetric wind module of Jelesnianski and Taylor (1973). The formulation is based on using pressure drop (Δp) and radius of maximum winds (R_{\text{max}}). Six hourly
surface observations are available from the India Meteorological Department (http://www.imd.gov.in/section/nhac/dynamic/bestrack.htm) web site and IMD Report (1999, 2000). The surface parameters are linearly interpolated at model time step for obtaining the wind distribution.

Comparison of the model wind speed with that of observed at DS1 location is depicted in Fig. 3 for both the cyclonic events. In the case of 1998 cyclone, maximum wind speed observed is of \( w = 25 \text{ m s}^{-1} \) (Fig. 3a) on 7 June when the cyclone is near the buoy position. The computed cyclonic winds at the DS1 location are comparable with buoy observations. Since, the wind speed becomes normal after 9 June, the model is again forced with the ECMWF daily winds. The model net heat flux during this period is also depicted in the figure along the secondary axis in which positive values indicate the gain and negative corresponds to the loss with reference to the ocean. The net heat flux is negative during the cyclone period 6–9 June and then positive till 18 June. Later, the flux decreases again as onset of the monsoon winds strengthen. Similarly, Fig. 3b depicts winds and fluxes for May 1999 cyclone. The heat flux reduces significantly during cyclone period compared to that of 1998. The net heat flux is noticed negative even after one week of the cyclone landfall. This may be attributed to the onset of vigorous southwest monsoon over this region.

The daily SST of the model and buoy for the 1998 and 1999 cyclones are depicted in Fig. 4(a,b) respectively at DS1. There is a sharp fall of surface temperature of about 2.5 °C within 24 h after crossing the cyclone at DS1 on 7 June as the net heat flux reduces significantly. The maximum cooling due to the passage of the cyclone is in agreement with the buoy value. After 10 June, the SST tends to regain normalcy. However, it could not return to the pre-storm condition due to the influence of evaporation from the overlapping southwest monsoon that assumed its strength after 19th. As a result, the sea surface is cooled, as evident from the fluxes that are negative during this period (Fig. 3a). The simulations for the 1999 cyclone also suggest similar pattern. However, the associated cooling is only about 1.5 °C which is about 1 °C less compared to the 1998 cyclone.

In order to understand the thermal response across the depth at DS1, the computed temperature is plotted at different depths in Fig. 5a,b for both the cyclones. In case of June 1998 cyclone (Fig. 5a), the temperature at 5 m depth shows a sharp decrease of more than 2.5 °C after the cyclone passed the location. It is important to mention here that a sudden warming is noticed in the waters below 60 m at the time of cyclone crossing. A maximum temperature increase of about 1.7 °C is simulated at 60 m. The increase of temperature is underestimated in comparison to the observations (shown in Fig. 2). However, the model is still able to simulate...
The temperature decreases at 60 m depth after 15 June and this can be related to the surface cooling as the monsoon winds become stronger. The inertial oscillations generated by the cyclone continued for about two weeks which are consistent with the observations (Fig. 2). The cooling of only 1 °C at 5 m depth is simulated for the 1999 cyclone (Fig. 5b), because the buoy location is coincided with the centre of the cyclone. However, in the subsurface depths of 60–200 m, a sudden rise of temperature is computed as the cyclone crosses the location. The associated subsurface warming of about 2 °C is noticed below the depths of 60 m.

Comparison is also made for current speed and direction between the model and buoy data at DS1 for both the cyclones as depicted in Fig. 6. It suggests that there is a rapid increase in magnitude of computed current speed from 28 cm s\(^{-1}\) on 5 June to 135 cm s\(^{-1}\) on 7 June as the cyclone crosses the buoy.

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**Fig. 6.** Comparison of (a) current speed (b) current direction with buoy data for 1998 cyclone events.
The current shows some wavy disturbance even after the cyclone crosses the coast on 9 June and slowly stabilizes its magnitude to 25 cm s\(^{-1}\) by 21 June. It is noted that the simulated currents are underestimated as the model winds during this period are weak compared to that of buoy winds (Fig. 3a). The simulated currents at DS1 (Fig. 6b) and the buoy data are in phase. It also suggests that observed inertial oscillations, generated by the passage of the cyclone are well simulated by the model. The oscillations continued till 21 June and then stabilize to a southward flow which is consistent with the observations. This shows the response of the ocean due to the passage of the cyclone is limited to about 2 weeks. Though the magnitudes of the current and its direction are not exactly matching with the buoy observations, the model is able to simulate well the phase of the current and normalcy of the ocean state after the cyclone crosses the coast.

In order to investigate further the temperature increase in the subsurface waters, vertical velocities across the latitude of DS1 are plotted in Fig. 7 for different days during 1998 cyclone. When the cyclone crosses the DS1 location on 7 June, there is a strong downwelling at the centre of the cyclone with an upward velocity on either side. Vertical velocities of about \(7 \times 10^{-2} \) cm \(s^{-1}\) are simulated in the deep waters. It is observed on the subsequent days, alternate convergence and divergence bands in the vertical column. The downward velocity causes increasing temperature while it is opposite for the upward velocity as a result of Ekman pumping processes. Analysis for the vertical velocity suggests that the warming in the subsurface waters is as a result of strong surface convergence just beneath the cyclone. As the cyclone advances ahead, the surface waters are cooled due to upwarping of the subsurface waters. Hence, the maximum surface cooling is taking place after the occurrence of maximum rise of temperature in the lower layers with about a day lag. This can be clearly seen in the time series data presented in Figs. 2 and 5. Further, enhancement of inertial oscillations is also noticed with depth. Similar vertical velocity pattern is simulated for 1999 cyclone (not shown here) with a maximum strength of about \(8 \times 10^{-2} \) cm \(s^{-1}\).

Fig. 8 shows the corresponding vertical thermal structure across the latitude of DS1 location. Beneath the centre of the cyclone on 7 June, the isotherms are pushed down as a result of strong downwelling velocity causing an increase of temperature.
in the subsurface. After the cyclone crosses the buoy location, the surface waters are cooled on 8 June due to cyclone induced upwelling. The passage of cyclone over the ocean leaves strong inertial oscillations and that leads to formation of convergence and divergence zones at all levels. It is evident from the figure that these zones excite strong oscillations even in the thermal field. Fig. 9a,b gives the SST anomaly on 7 June when the cyclone crosses the DS1 location with reference to the pre-storm conditions with and without using fluxes in the model. The numbers on the cyclone track give the dates of the cyclone position at 0600 UTC. Clearly, the cooling associated with fluxes is higher by only about 0.5 °C to right of the track indicating that the turbulent mixing and entrainment are dominant compared to the role of surface heat fluxes in the deep waters. This is consistent with the study of Shen and Ginis (2003).

Further to understand the change of temperature in the vertical water column, vertical stratification at DS1 is depicted in Fig. 10a before and after the 1998 cyclone crosses the location. As the thermocline is considered as the region of vertical temperature gradient with more than 0.5 °C/10 m, hence it is noticed that the thermocline ranges from 20 m to 200 m depth at DS1 on 1 June. On 7 June, when the cyclone crossed the DS1, the mixed layer depth (MLD) deepened upto 50 m and the thermocline is pushed beneath which in turn rises underlying waters temperature. Analysis of the observations during Hurricane Felix (Zedler et al., 2002) also suggests that the subsurface warming occurs between the depths 40–70 m which coincidence with the thermocline region. Hence, the vertical extent of subsurface warm zone is variable and depends on the local thermocline. The warming at 100 m depth is noticed about 1.5 °C on 7 June which is in good agreement with the observations. The surface cooling of about 2.5 °C is also very well comparable with the observations as shown in Fig. 3.

The centre of the 1999 cyclone coincides with the DS1 location. Hence, it may be interesting to see how the surface cooling and subsurface warming on either side of the location. Fig. 10b provides the computed thermal profiles at selected locations through depth before (on 15 May) and during the cyclone (on 17 May) at DS1. Two additional profiles are extracted at locations selected about one degree left and right to the buoy location on 17 May. These profiles are compared with the initial temperature profile at DS1 on 15 May. The warming of subsurface waters occurs only to the left of the track as a result of downwelling in the region while, the cooling is seen to the right of the track due to favorable upwelling processes. In this case, the simulated warming and cooling is about 1 °C at 100 m depth to the left and right of the track respectively when compared to the pre-storm conditions of 15 May. The surface cooling is more than 1 °C to the right of the track compared to that of left due to strong upwelling in addition to evaporative cooling on either side.
Fig. 9. SST anomaly on 7 June 1998 with reference to pre-storm conditions (a) without fluxes (b) with fluxes.

Fig. 10. Vertical temperature profile (a) beneath the 1998 cyclone and (b) at locations to the left, right and centre of the 1999 cyclone.
5. Conclusions

The computed maximum surface temperature cooling for the 1998 and 1999 cyclones at DS1 agrees qualitatively well with the buoy and thermister chain data. In this study, an attempt is made to account the subsurface warming observed during the 1998 cyclone. The model simulates warming of subsurface waters below 60 m and cooling in the surface layers. The mechanism for the fall and rise of temperature at the surface and subsurface respectively is explained through the convergence and divergence in the vertical water column. The subsurface warming precedes the surface cooling with a lag of a day as the cyclone crossed the location. The local temperature stratification is important for the cooling of the upper ocean and warming of subsurface waters. The maximum surface cooling occurs to right of the track due to combined effect of upwelling and evaporation. The evolution of surface currents produced by the passage of cyclone is also studied. The simulated currents are slightly underestimated whilst; the current direction is in good agreement with the buoy observations. The model simulations suggest that the inertial oscillations generated due to the passage of cyclone continued for about two weeks even after its landfall and then stabilize with the normal southward flow which is consistent with the buoy observations.

References