Spatial-temporal variation and mechanism associated with mini-cold pool off the southern tip of India during summer and winter monsoon season

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ABSTRACT

Sea surface temperature (SST) from different sources suggests that the occurrence of a mini-cold pool (MCP) off the southern tip of India (STI) is a persistent phenomenon which occurs during both the summer and the winter monsoon seasons. However, the associated mechanism is different in both scenarios and, hence, numerical experiments are conducted to study and ascertain the mechanism. The dynamics that govern the occurrence of MCP during the summer season is mainly due to upwelling, caused by the divergence in the near-surface circulation off STI, advection of the cold upwelled water from the western Arabian Sea and the southwest coast of India. In contrast, during the winter monsoon, the model studies suggest that circulations driven by positive Ekman dynamics and outgoing heat flux are mainly responsible for the formation of MCP off STI during December–February. The cold water intrusion in both seasons occurs in accordance with the monsoon and coastal currents, which underlines the importance of advection. The position and extent of cooling differs during both seasons because wind stress varies significantly.

Key words | coastal upwelling, mini-cold pool, monsoon, salinity, wind-driven circulation

INTRODUCTION

The southwest and the northeast monsoons serve as lifelines to the economy of India. The onset and active phases of the southwest monsoon are of utmost importance to the Asian countries. The Bay of Bengal (BoB) and Arabian Sea (AS) are the two important basins of the North Indian Ocean which exhibit seasonal circulation, forced by the southwest and the northeast monsoons. During the early 1990s, McCreary et al. (1993) showed that wind over the equator and winds over the BoB contribute to the circulation in the Bay. There is ample evidence (Shetye et al. 1990, 1991; McCreary et al. 1993) to show that no part of the BoB or AS can be studied in isolation without taking into account the rest of the North Indian Ocean basin, in particular, the equatorial Indian Ocean. It should also be noted that wind-forcing along the eastern boundary of the BoB has significant impact on the circulation along the western boundary. Similarly, the strong southwest summer and the winter monsoon currents affect the upper ocean dynamics at the southern tip of India (STI) and also the south central BoB and AS.

The AS is first to experience the effect of the summer monsoon in terms of circulation and thermo-haline distributions compared with the BoB. During the summer monsoon (June–September), an anticyclonic circulation evolves in the AS. The equator-ward component of the eastern boundary of this anticyclonic circulation, along the Indian coast, is known as the West India Coastal Current (WICC). The WICC is northward in winter (November–January), which favours the sinking processes along the eastern AS (Stramma et al. 1996; Gopalakrishna et al. 2008, 2010). Hence, the equator-ward WICC during the summer monsoon acts as a communication channel from AS to BoB. Similarly, the cyclonic circulation in the BoB which constitutes the southward East India Coastal Current (EICC) in
winter gives rise to water mass transport from BoB to AS (Shankar & Shetye 1997; Shetye & Gouveia 1998). A cyclonic eddy, Lakshadweep Low (LL) forms during the early summer monsoon centred at 75° E and 10° N and migrates westward across the southern AS subsequently after its genesis. The time of formation of WICC and the southward phase in early summer, coincide with that of LL (Bruce et al. 1994; Shankar & Shetye 1997; Rao et al. 2008a). McCreary et al. (1995) suggested that westward migration of LL and Lakshadweep High (LH) is a result of Rossby waves propagating in the AS. The generation of these waves by the wind and their propagation was earlier noted by Jensen (1991).

The in situ data collected along the west coast of India show that the temperature inversion (TI) in this area is a stable seasonal feature (Thadathil & Gosh 1992). The downwelled AS high-saline sub-surface waters are warm and lead to TI in the mixed layer (Hareesh Kumar et al. 2009). Durand et al. (2004) explained that TI is known to occur in the near surface ocean regime where salinity stratification is large enough to influence the density field. Recent studies by Kurian & Vinayachandran (2006) suggest that in the southern part of the South Eastern Arabian Sea (SEAS), advection of cooler low-salinity BoB water over warm salty SEAS water leads to the formation of TI, which occur at a greater depth (~80 m) because of the downwelling within the Lakshadweep High. In the northern part, TI occur at a much shallower depth (~20 m) and are caused by shortwave radiation penetrating below the mixed layer of about 12 m. During the summer monsoon, the southerly WICC favours the upwelling along the eastern AS. Stramma et al. (1996) observed a typical eastern boundary current along the southwest coast of India during August 1993. A recent field experiment conducted by Sanilkumar et al. (2004) during July 2003 off the southwest coast of India indicated intense upwelling within the upper 60 m water column all along the coast but the off-shore width of upwelling reduced significantly from 10 N (>200 km) to 15° N (~50 km). All these features seem to affect the circulation and thermo-haline characteristics and, hence, any or all the processes of the AS are to be studied.

The observed mini-cold pool (MCP) off the STI and its intrusion into the south central BoB and AS during the summer and the winter monsoon seasons play an important role in understanding the evolution of sea surface temperature (SST). The MCP has potential implications for the active and break cycles of the summer and the winter monsoon (Joseph et al. 2005; Shankar et al. 2007). This phenomenon also shows pronounced intra-seasonal fluctuations that vary from year to year (Schott et al. 1994). Luis & Kawamura (2002a) and Rao et al. (2006) studied the probable governing mechanism of MCP of waters off the STI and its intrusion into the south central BoB and AS during the summer and the winter monsoon seasons (Luis & Kawamura 2002b; Rao et al. 2008b) and they concluded that local upwelling is responsible for this process. The upwelling process starts in late June and peaks in August and subsides gradually in September; and the intrusion takes place through the Summer Monsoon Current (SMC). However, their studies were limited to a few observations (Rao et al. 2006; Rao et al. 2008b) and were episodic in nature (Luis & Kawamura 2002a, 2002b). These authors also stressed that modelling efforts are needed to resolve the relative importance of the processes associated with the evolution of SST in the BoB and AS, and that knowledge of this phenomenon might eventually lead to improved understanding of the intra-seasonal variability of the summer and the winter monsoons. Observations of SST from different sources (e.g. AVHRR5 (Advanced Very High Resolution Radiometer), TMI (Tropical Rainfall Measuring Mission Microwave Imager) and OSTIA (Operational Sea Surface Temperature and Sea Ice Analysis)) indicate that the formation of the MCP off the STI and its intrusion takes place right through the month of September during the summer season.

In this study, the nature of the cooling episodes observed off the STI and in the south-central BoB and AS on an intra-seasonal time scale and its year to year variability is examined. In addition, the co-evolution of the divergence in the near-surface circulation resulting in upwelling and cooling episodes on an intra-seasonal time scale is also explored. To investigate the occurrence of the MCP off the STI during the summer and the winter monsoons and a mechanism for its sustenance during the summer monsoon season, a three-dimensional Princeton Ocean Model (POM) has been configured for the North Indian Ocean. The three-dimensional ocean models, POM, Regional Ocean Modelling Systems
(ROMS), Hybrid Co-ordinate Ocean Model (HYCOM) and others, have been used extensively in several parts of the world’s oceans and in different water bodies (e.g. Lake Ontario in Canada (Huang et al. 2010), Philippine Archipelago region (Han et al. 2009)) and for different thermohaline circulation, upwelling and downwelling processes. These models are also used for resolving cyclonic and anticyclonic eddies. The POM model has been used especially for tracking pollutant transports (Chen et al. 2011) in the Grand Banks of Canada region and several places in the Great Lake regions of Canada.

THE MODEL AND NUMERICAL EXPERIMENTS

POM, as described by Blumberg & Mellor (1987) and Mellor (1992), is a free surface ocean model with a turbulence closure scheme of Mellor & Yamada (1982). The momentum equations are nonlinear and incorporate β-plane approximation. The model uses an orthogonal curvilinear grid and a terrain following sigma coordinates in the vertical. The boundary conditions in the vertical are no-slip condition at the bottom and no through flow at the surface (i.e. $\omega = 0$, $\omega$ is the vertical velocity component normal to the sigma surfaces) for the continuity equation. For the momentum equations, the boundary conditions are suitably specified using the surface and bottom turbulence momentum fluxes. The boundary conditions for velocities include radiation conditions at the lateral open boundaries. For a detailed description of the POM, one can refer to the user manual and related materials at http://www.aos.princeton.edu/WWWPUBLIC/htdocs.pom/publications.htm.

The analysis area is the entire North Indian Ocean covering 30° S to 30° N and 30° E to 120° E. The model has 150 x 150 grid points and 26 vertical ($\sigma$) levels and uses curvilinear grids with varying horizontal resolution from 20 km to 50 km. The model incorporates a higher resolution near the coast, and in the vertical a terrain-following sigma coordinate is used, with finer resolution near the surface and bottom and relatively coarse resolution in the middle. The open (deep) ocean has a relatively coarser resolution of about 50 km. The model time steps are based on the Courant-Friedrichs-Lewy (CFL) condition, in which the two-dimensional external mode has a short time step and the internal mode a longer time step, depending on the associated external and internal wave speeds, respectively. The model uses staggered Arakawa ‘C’ grid with external and internal time steps of 15 and 600 s, respectively. The model utilized the bottom topography derived from the ETOPO5 (Earth Topography Five-Minute) from the NGDC (National Geophysical Data Center) database as shown in Figure 1. Topographic gradients are smoothed to avoid spurious along-slope currents in a sigma coordinate model (Haney 1991). A mathematical tool, SeaGrid (http://woodshole.er.usgs.gov/operations/modeling/seagrid/seagrid.html) for Matlab, is used to convert the rectangular grid to an
orthogonal curvilinear grid, which also enhances model resolution at the selected regions. However, since we are dealing with a basin-scale model, it is important that the model should attain the so-called quasi-steady-state before carrying out any experiments and drawing conclusions thereby. To achieve this, we proceed through three broad stages which can be classified as: (i) diagnostic runs of the model with annual fields of input; (ii) prognostic model runs for spin-up; and (iii) prognostic runs for 10 years with near real-time wind forcing. These are described in detail in the following sections.

The diagnostic run

To start with, the initial density field is set, which is defined by annual climatological data of temperature and salinity, obtained from the World Ocean Atlas 2005 (WOA05). To avoid any noise in model integration, the model ocean is initialized with the present ocean state until the circulation, sea level and velocity are consistent with the prescribed water mass structure. In other words, the model ocean is adjusted geotropically to its initial state. To accomplish this, the model has been spun up in diagnostic mode with constant forcing, keeping the temperature and salinity constant until a steady state is reached. The experiments in the diagnostic mode attain steady-state conditions after about 90–120 days. The diagnostic approach not only provides a powerful tool for deducing circulation but also permits a consistent way of initializing a prognostic mode. Numerical experiments involve the use of both diagnostic and prognostic modes.

The prognostic run (30 years) for spin-up

In the prognostic mode integration that follows the diagnostic run, the model equations are integrated as an initial value problem. In the present case, the model is forced with monthly Quikscat climatological wind stress, annual values of temperature and salinity from WOA05 and monthly Southampton Oceanography Centre (SOC) fluxes (http://iridl.ldeo.columbia.edu/SOURCES/SOC/GASC97/) and integrated for 30 years until it reached quasi-steady-state. This is indicated by the seasonal cycle oscillations, typical of the North Indian Ocean region, established as seen by stabilization of the model kinetic energy (not shown here). During the 30-year spin-up, the model temperature and salinity are relaxed to the climatology of the month concerned through a weak relaxation coefficient of 30 days which otherwise would have caused excess heat accumulation in a shallow water region.

The prognostic model runs (10 years) for experiment

The model initialized by the spin-up results is further integrated for 10 years (from 1 August 1999 to 1 August 2008) with real time daily Quikscat wind. The wind stress data are obtained from http://dss.ucar.edu/datasets/ds744.4, which is referred to as QSCAT/NCEP blended ocean winds with a spatial resolution of $0.5 \times 0.5$ degrees, and are available for a 6-hourly basis from July 1999 to 31 July 2009. However, monthly climatological SOC fluxes continue to be utilized in this component of prognostic mode integration. From the daily model output of 10 years, multi-year monthly averaged values are analysed for the study area as shown in (inset) Figure 1, where the rectangular box shows the study area for both the monsoon seasons. The lines 1, 2 and 3 in the figure refer to the latitudinal transects for which model simulated results of depth-wise temperature distribution are discussed. The inset picture on the right shows the grids overlaid for the southern tip of India region. However, the open ocean has further coarser resolutions compared with the coastal region.

In order to assess the cooling episodes and contribution from the air–sea interaction through surface fluxes, results of surface fluxes from the site http://nomad1.ncep.noaa.gov/ncep_data/index.html were taken into consideration. This is done in order to validate our model data qualitatively with other model data and also TMI SST as AVHRRS SST data have their own limitations especially during the monsoon season. The fluxes taken for consideration are turbulent heat fluxes (sum of latent and sensible heat fluxes), averaged outgoing heat fluxes and net shortwave fluxes. The upper panel in Figure 2 depicts the summer (2008) and the winter monsoon (2008–2009) cases for these fluxes, averaged over a box of $2 \times 2$ from 77.5° to 79.5° E and 6° to 8° N. The strong cooling episode is indicated by the time encompassed by vertical lines where marked changes (heat loss with respect to ocean) in the
fluxes are seen (Luis & Kawamura 2002a, 2002b). However, there are secondary cooling episodes which are much more pronounced in the case of the summer monsoon than in the winter monsoon as indicated by Figure 2. Nevertheless, these secondary heat loss episodes contributed to the prolonged cold pool during the summer monsoon. As only peak monsoon months (December–January, July–August) are chosen, the variability, increase or decrease is not significantly marked here but the peaks are clearly visible. Correspondingly, the lower panel in Figure 2 depicts the model multi-year averaged daily SST and TMI SST imagery (from http://www.remss.com/tmi) for the above-mentioned period over the same box. The cooling trend, as indicated by the figure is clearly visible in both the monsoon cases whereas the summer cooling is much stronger compared with the winter monsoon. The change in the air–sea interaction as reflected in the heat fluxes is clearly correlated to the cooling episodes, both in the model and the TMI SST. The reason the model SST is flat compared with that of TMI SST is because the model was initialized with the monthly climatological temperature and salinity of WOA05. Further, the heat fluxes in the model do not comprise any short-term fluctuations, for example diurnal variations. Nevertheless, the figure qualitatively depicts the role of air–sea interaction on the cooling episodes, observed during the monsoon season off STI.

The heat budget analysis assesses the cause of cooling and warming episodes. Therefore, the heat budget analysis of the mixed layer is presented (Santoso et al. 2010) briefly below:

\[
\theta_t = Q_{\text{net}} - u \cdot \nabla \theta - \frac{w_{\text{ent}}}{h}(\theta - \theta_d) + \text{Res}
\]

(e.g. Qiu 2000; Qu 2005; Du et al. 2005), where \( \theta \) is the potential temperature averaged over the mixed layer, the subscript \( t \) denotes time differential operator, and \( h \) is the mixed layer depth (MLD). Here, \( Q_{\text{net}} \) represents the effective net surface heat flux retained within the mixed layer, specified by

\[
Q_{\text{net}} = \frac{Q}{\rho_0 c_p h}
\]
where \( \rho_0 \) is a reference density (1,026 kg m\(^{-3}\)), \( C_p \) is the specific heat capacity of seawater (3,986 J kg\(^{-1}\) K\(^{-1}\)), and \( Q \) is the net air–sea heat flux absorbed within the mixed layer. Here, \( Q \) can be further decomposed into its components: \( Q = SW + LW + LH + SH - q_d \), where \( SW \), \( LW \), \( LH \), and \( SH \) are shortwave, net longwave, latent heat and sensible heat fluxes at the ocean surface, respectively, and \( q_d \) is the portion of the shortwave radiation that escapes through the base of the mixed layer. Following Paulson & Simpson (1977), \( q_d \) is estimated as

\[
q_d = SW \left[ Re^{(-h/\gamma_1)} + (1 - R) e^{(-h/\gamma_2)} \right]
\]  

(3)

where \( R \), \( \gamma_1 \), and \( \gamma_2 \) are coefficients that depend on water turbidity as classified by Jerlov (1968). For example, the Southern Ocean approximately falls into a water type with \( R = 0.67, \gamma_1 = 1, \) and \( \gamma_2 = 17 \) (i.e. Jerlov’s water type IB; see also Dong et al. 2007). However, for the Indian Ocean a water type 1A is often used (Rasmussen 1999) with \( R = 0.62, \gamma_1 = 0.6, \) and \( \gamma_2 = 20 \).

The second term on the right of Equation (1) expresses the oceanic advection of heat. The horizontal current velocity \( u \) is an average over the MLD and can be decomposed into the Ekman \( u_e = (\rho_0 h)^{-1} (\tau_y, -\tau_x) \) and non-Ekman components, where \( f \) is the Coriolis parameter and \( (\tau_x, \tau_y) \) are the zonal and meridional components of sea surface wind stress.

The third term on the right of Equation (1) expresses the heat flux due to entrainment. Ignoring diffusion, the entrainment rate \( (\dot{w}_{ent}) \) can be diagnosed as

\[
\dot{w}_{ent} = h_t - \nabla \cdot (hu), \quad \dot{w}_{ent} > 0
\]

\[
\dot{w}_{ent} = 0, \quad \dot{w}_{ent} \leq 0
\]

(4)

after Qiu & Kelly (1993), where \( h_t \) is the rate of change of the MLD, and \( \nabla \cdot (hu) \) represents the divergence of mass in the mixed layer, which includes vertical advection at the base of the mixed layer. Here, \( \theta_d \) is the temperature of entrained fluid immediately below the mixed layer (taken at 5 m below MLD; Du et al. 2005).

The residual term (Res) represents all remaining unresolved processes, such as lateral diffusion which is taken as zero here. The results of the heat budget of the mixed layer are discussed in the heat budget analysis of the mixed layer.

**RESULTS AND DISCUSSION**

**Intra-seasonal variability of the MCP**

**MCP off the southern tip of India: AVHRR5 SST data**

The multi-year monthly averaged AVHRR5 SST data, obtained from the PODAAC site (http://poet.jpl.nasa.gov) of the Jet Propulsion Laboratory (JPL) for May–October show the observed spatial variability around the southern peninsula of India (Figure 3(a)) with a distinct MCP (~26 °C) off the STI. In addition, this highlights the advection of cold waters from the western AS and the southwest coast of India in this region. The AVHRR5 SST data for all the 9 years (from 1 August 1999 to 1 August 2007) show that the formation and genesis of MCP off STI starts by June and peaks in August and the SST attains as low a value as 25.5 °C, essentially due to upwelling. The AVHRR5 SST data also indicate that the upwelling process enhances with time. It starts off in the Somalia region during the southwest monsoon season and the cold upwelled water is advected as far as 18° N by a strong southwest monsoon current which reaches the west coast and STI. This process continues until September as is evident by AVHRR5 SST data when the upwelling off the STI is believed to be on the wane with the recession of the southwest monsoon. To supplement our understanding of the cause of cooling off the STI, the model results such as multi-year averaged SST, vertical temperature profile off the STI, sea surface currents and sea surface salinity along with AVHRR5 SST data, divergence of near surface circulations and Ekman transports are analysed. Details of Ekman transport are discussed in the section on ‘Sea surface current’.

Figure 4(a) depicts the AVHRR5 SST data for December to February. It shows the formation and dissipation of the cold pool and its intrusion into the AS. Qualitatively, it can be seen that the cold pool position is different for both the monsoon seasons. The winter monsoon is short compared with the summer monsoon, which can be
attributed to the relative strengths and duration of the monsoonal winds in both the cases.

**Sea surface temperature**

Figure 3(a) depicts multi-year (1999–2008) averaged AVHRR5 SST from May to October which clearly shows the development and dissipation of MCP during the summer monsoon season. The cold water flows south of Sri Lanka and intrudes into the south-central BoB as a cold tongue with an anticlockwise curvature. The figure suggests that the upwelling process starts as early as June and increases gradually with the advancement of the southwest monsoon. The minimum SST for September and October are 26.7 and 27.3 °C, respectively, highlighting the occurrence of MCP in September as well. This qualitatively agrees with the model SST data as shown in Figure 3(b). Comparison of the climatology with model-simulated SST gives the intensity, i.e. the temperature gradient, of MCP off the STI as 0.5–0.7 °C for May–June, 1.3–1.5 °C for July–September and 0.7 °C for October. The intrusion of this MCP into the south-central BoB was also captured by the model. The intrusion of MCP has an anticlockwise curvature as indicated by the AVHRR5 and the model SST data, and it also peaks in August as indicated by 27.9 °C contours which were replaced by a 28.1 °C isoline for both July and September. This supports our argument about the advection of cold waters in September. In October, the intrusion ceases to exist as indicated by the 28.7 °C isotherm which limits itself to just the southeast region off Sri Lanka. The spread of warm water from around 70°E in the equatorial region into the eastern AS (or west coast of India) from
August to October is also confirmed by the AVHRR5 SST data.

Figures 3(a) and 3(b) suggest that there are three contributing factors for the formation of the MCP off the STI: (i) wind-driven local upwelling off the STI; (ii) advection of cold upwelled water from the Somalia region (western AS); and (iii) advection of upwelled water from the southwest coast of India which was at its peak in July and August. Intense blooming of chlorophyll $a$ in this region lends additional support to the upwelling process (Yapa 2000; Vinayachandran et al. 2004, 2005; Rao et al. 2006). However, the chlorophyll $a$ bloom off the STI may have contributed to the local upwelling associated with the west coast of India (Wiggert et al. 2006; McCreary et al. 2009). The strong monsoonal current decreases the chlorophyll $a$ bloom off the STI, which may not necessarily rule out upwelling in that region. As depicted in Figures 4(a) and 4(b) during the northeast monsoon, in contrast to the summer monsoon season, the low saline waters from the south-central BoB intrude into the AS guided by the winter monsoon current (WMC), hence, gradually dissipating the cold pool. The cold pool intensifying in the month of January ($\sim 0.9$ °C) and subsiding in February is clearly seen. In this case, the chlorophyll $a$ bloom compared with the summer monsoon is not seen, although upwelling is there and this may be confirmed by numerical model results as discussed subsequently (McCreary et al. 2009).

### Sea surface current

Horizontal advection characterized by the SMC is an important process in the redistribution of temperature in areas where the surface currents and horizontal gradients in SST are relatively strong, which is responsible for the genesis of the MCP off the STI and its sustenance till September. Figure 5(a) shows the multi-year monthly averaged model sea surface current for May to October. It shows the strengthening and the progress of the SMC during the southwest summer monsoon season. In May, the current is basically in the southwest direction and the strength is about 40–50 cm s$^{-1}$ off the Somali region. It is zonal in the central AS region and becomes southeast in direction when it reaches the southwest coast of India. Finally, it skirts the STI where the current speed is about 30–35 cm s$^{-1}$.
to enter the south-central BoB. Although the characteristic feature remains the same, the strength of the SMC increases from June to July and remains strong in August as ascertained by historic ship drift vectors (Cutler & Swallow 1984; Mariano et al. 1995) and direct current measurements from moored stations and shipboard profiling (Schott et al. 1994). The model sea surface current shows a strong seasonal cycle with eastward flowing SMC. The major part of the SMC curves around Sri Lanka; this is about 300 km wide and extends more than 100 m deep (sub-surface model currents are not shown here) in agreement with Vinayachandran et al. (1999) and Shankar et al. (2002).

The gradual progress in the SMC showed the sustenance of the MCP off the STI in September due to the advection of cold waters from the western AS. The current in October, being feeble and southward around the west coast of India, is different from the previous months and a sharp gradient is observed off the STI which delineates it from the rest of the AS, and after crossing Sri Lanka the current is in the south-east direction which breaks the intrusion of the MCP into the south-central BoB. The southward current along the east coast of Sri Lanka ensures the anticlockwise curvature of the MCP intruding into the BoB.

In contrast to the summer monsoon case, as depicted in Figure 5(b) the wind-driven sea surface current pattern is completely different in the winter monsoon case. The current direction is in accordance with the northeast monsoon wind and the magnitude is more at the Gulf of Mannar compared with an adjoining area. In the Gulf region, the magnitude of the southwestward sea surface current increases from about 25–30 cm s⁻¹ in December to 35–45 cm s⁻¹ in January and reduces to 20–25 cm s⁻¹ in February. This is favourable for upwelling in the region but more towards the Sri Lanka coast than the Indian tip, a fact that is supported by Figure 6 which shows the wind-driven Ekman transport (kg m⁻¹ s⁻¹) in this region. The classical Ekman dynamics is one of the conventional methods which is considered here for the winter monsoon case. The Ekman transport, $M_e$, can be computed as $M_e = \tau/f$. Here, $M_e$ is the wind generated mass transport per unit width integrated over the depth of the Ekman layer (kg m⁻³ s⁻¹); $\tau$ is wind stress, and $f$ is the Coriolis parameter. For details about
Ekman transport in the Indian Ocean one can refer to Vimal Kumar et al. (2008) and Jayaram et al. (2010). There is an increased Ekman transport indicated to the eastern side of the Gulf which showed strong upwelling in the month of January and may be responsible for the intensification of the cold pool. The negative chlorophyll bloom may be attributed to the depletion of essential nutrient minerals which can be confirmed by three-dimensional biological model (Wiggert et al. 2006). The shape of the WMC may be responsible for the limited intrusion of low saline water from BoB to AS since it turns northward after crossing the Sri Lanka coast and entering the eastern AS. This will be evident when we consider the surface salinity structure.

**Model ocean surface salinity**

Figure 7(a) shows the multi-year monthly averaged model simulated sea surface salinity from May to October. The figure suggests that there is a gradual development of a front from July which is pronged towards the equator and grows into a well developed front from August to September in the zonal direction and is suppressed in October. This feature suggests that during the summer monsoon the highly saline AS waters show their intrusion into the southern BoB (Wyrtki 1971; Conkright et al. 2002; Rao & Sivakumar 2003). The salt budget analysis of the surface mixed layer has revealed the importance of horizontal advection of salinity into the BoB during the summer monsoon season. The model sea surface current also supports the advection of highly saline water into the BoB. The prominence of the front in the months of August and September is supported by Jensen (2001) who used passive tracers in a model as a tool to map the pathways and noted that inflow of highly saline waters from the AS into the BoB is significant and occurs after the mature phase of the summer monsoon.

The multi-year averaged sea surface salinity from December to January shown in Figure 7(b) shows the opposite trend to that of the summer monsoon and indicates a possible intrusion of low saline waters into the eastern AS guided by WMC and WICC. The effect of this intrusion is shown beyond 65° E and, hence, an intrusion of more than 1,500 km is indicated. The figure indicates a salinity drop of about 0.8 PSU around 67° W and 5° N, which plays a significant role in altering the upper ocean characteristics of the AS in winter. The decrease in salinity in and around the Gulf region may inhibit chlorophyll a bloom near the surface during the winter monsoon season. The zonal front which is seen in all of the 3 month simulations is associated with the south equatorial counter current (SECC), which is not shown in the sea surface current since our focus region is the STI (Schott et al. 1994; Donguy & Meyers 1995; Shankar et al. 2002). A similar but opposite front can be seen in Figure 7(a) for the summer monsoon case which lies between the equator and 10° S.

**Transport due to advection**

The cooling episodes during the summer monsoon are much more intense than that of the winter monsoon. This underlines the importance of the role of advection, which on one hand enhances the cooling over a prolonged period during...
summer and on the other hand limits the cooling period during the winter monsoon. To substantiate this fact, the advection due to model transport is estimated for both the monsoon periods for the upper 100 m of the ocean. In the present study, we consider only coastal transport as far as advection is concerned. The pictorial depiction of the volume transport is not shown in the present study and only quantitative discussion is made. For details of the computation of volume transport, refer to Stammer et al. (2003).

During the summer monsoon period, the strong Somali current drives the cool upwelled waters in a poleward direction at a rate of $35 \times 10^6 \text{ m}^3 \text{s}^{-1}$, of which $8-10 \times 10^6 \text{ m}^3 \text{s}^{-1}$ reaches as high as 20° N and is mixed with the eastern Arabian Sea waters. In turn, about $4-5 \times 10^6 \text{ m}^3 \text{s}^{-1}$ is advected to the STI region along with the locally upwelled water off the west coast of India. Along with the SMC and the southward WICC, a transport of $9-11 \times 10^6 \text{ m}^3 \text{s}^{-1}$ is witnessed around the STI. This strong transport helps in establishing the cold pool tongue to the central BoB during the summer monsoon.

In contrast, during the winter monsoon, the transport is much stronger at about $14 \times 10^6 \text{ m}^3 \text{s}^{-1}$ around the STI which, in turn, enters the AS by strong poleward WICC as clearly seen by the AVHRR5 and the model SST. Owing to the WICC, a strong transport of the order of $7-8 \times 10^6 \text{ m}^3 \text{s}^{-1}$ occurs during the winter monsoon along the west coast of India. This strong transport and its direction play an important role in confining the cold pool off the STI in that region only. During February–March, there is change in transport direction along the west coast and around the STI region which causes the cold pool to cease to exist. Here, it should be noted that we consider the upper 100 m of the ocean since the mixed layer depth is very shallow in the AS region. The model-predicted volume transport is in agreement with the analysis and results of several workers (Schott et al. 1994; Shankar 2000; Haugen et al. 2002).
Heat budget analysis of the mixed layer

Figure 8 represents the heat budget analysis in the mixed layer based on Equation (1) for the southern tip of India. The different terms shown are averages over the mixed layer of rate of change with respect to time for potential temperature ($\theta_t$), effective net surface heat flux retained within the mixed layer ($Q_{net}$), advection of cold (or warm) water and entrainment of cold (or hot) water ($\omega_{ent}$) for $69^\circ$E to $85^\circ$E and $4^\circ$N to $14^\circ$N. The subsequent analysis will unequivocally establish the mechanism of formation of the cold pool at the southern tip of India during both monsoon seasons. The results for the aforesaid parameters are shown for December–January and July–September.

The formation of the cold pool and its persistence is evident from the rate of change of potential temperature ($\theta_t$) in December–January and July–September. However, the contribution due to net heat flux ($Q_{net}$) is more and advection of cold water is less in December and January compared with entrainment. Hence, we can conclude that the local upwelling and net heat flux are responsible for the formation and sustenance of the cold pool during the winter monsoon. During the summer monsoon period, the patches of cold pool near the southwest coast of India and contribution by entrainment confirms local upwelling in this region. The same fact holds true for the southern tip of India. Since the Somalia upwelling is an established fact, we concentrate only on the southern tip of India region along with the southwest coast of India. The advection of upwelled water from the Somalia region along with upwelled water from the southwest coast of India and southern tip of India contributes to the genesis of the cold pool at the southern tip of India. The net heat flux during the summer monsoon season shows how its role is smaller.
compared with that in the winter monsoon. Therefore, our argument that heat flux plays its role in contributing only during the winter monsoon holds true as far as the formation of the cold pool is concerned. However, during the summer monsoon season, the local upwelling and contribution of remote upwelled water due to advection supplements the formation and sustenance of the cold pool near the southern tip of India.

**Inter-annual variability of the MCP**

**MCP off the STI: Divergence of near surface circulation**

The divergence of the near surface circulation, henceforth referred to as divergence, is an indicator of upwelling of sub-surface waters giving rise to the formation of the cold pool. Figure 9(a) depicts inter-annual variability of the divergence of near surface circulation for every alternate year of 2000–2008. The figures for individual years show a wide range of variability from year to year and also from month to month (here only August and September are shown). However, our focus is mainly on the southern tip of India and its adjoining area. Every year there are two strong divergence zones, one around the Somalia region and another above the Gulf of Aden, which results in strong upwelling in those regions which is a well-known feature and hence not shown in the figure. Compared with all the years, in 2000, there is less divergence for both August and September but it is clearly evident that in September the divergence is less than that of August off the STI. For all other years the same feature is there confirming the fact that upwelling is reduced in September compared with August and, therefore, there must be some other source which sustains the cooling in September, which may be advection of cold (or upwelled) water from other places. The position of divergence occurring near the STI is

![Figure 9](https://iwaponline.com/wqrj/article-pdf/47/3-4/333/163541/333.pdf)
important for the intensity of cooling of the MCP. For instance, in 2001 (not shown) and 2002 the position of divergence patches is such that the intensity of MCP is greater than that of 2004 and 2006 off the STI. The AVHRR5 SST data for the corresponding years support this finding. Another important aspect which is revealed by Figure 9(a) is that there are distinct patches of maximum divergence of different shapes and sizes at the south-central BoB which seem to be a permanent, resident feature for all the years. These give rise to local wind-induced upwelling of the sub-surface waters to the surface which contributes to the MCP off south-central BoB. The intensity and position of the divergence may determine the shape of the tongue of the MCP that intrudes into south-central BoB.

In order to assess the inter-annual variability of the surface circulation for the winter monsoon we consider Ekman transport as an indicator of upwelling. Figure 9(b) denotes the contours of Ekman transport for all the alternate years of 2000–2008 for December and January. The figures as in the case of the summer monsoon verify that there is a large inter-annual variability in the Ekman transport and hence the degree of upwelling. For instance, except for 2002, the Ekman transport increases from December to January and so does the radius of the MCP. In 2002, it decreases and the MCP also showed marked decrease in its intensity. It is observed that if the position of the Ekman transport is towards the eastern side of the Gulf of Mannar, that is, the west coast of Sri Lanka, and its intensity is increasing from December to January then it happens to be a normal monsoon as in the case of 2003 (not shown) and 2008. But, if the position of large Ekman transport is towards the Adams Bridge then that year is a deficient monsoon case as in 2002 and 2004. However, this needs to be verified with other factors involving monsoonal rainfall.

**MCP off the STI: Vertical temperature along 7.5°N**

Upwelling of sub-surface waters driven by divergence of near surface circulation is the primary cause of formation of the MCP off the STI. To substantiate this, Figure 10(a) shows model-computed vertical temperature along 7.5°N and 50–80°E (line 2 in Figure 1) for the same period as depicted in Figure 9. The left side of the figure shows a region off the Somalia coast and the right side shows that off the STI. In all figures for August there is upwelling as indicated by the upwarping of the isotherms off the STI region. In the Somalia region also there is an indication of strong upwelling for all the years. As indicated by the years 2000 and 2006 there is an intense upwelling even greater in September than that of August driven by the strong southwest monsoon wind. These cold upwelled waters are ultimately advected to the southwest coast of India contributing to the MCP off the STI. The large inter-annual variability in the divergence of the near surface circulation leads to varied episodes of upwelling giving rise to different strength of advection of the cold waters in different years. The depth-wise distribution of temperature is similar for all the years which may be a result of the limitations in the model input fields of temperature and salinity. In Figure 10(a) for September, there are cold waters at the surface and upwelling is suppressed for almost all the years, which supports the fact that these cold waters reach the STI by advection. The weekly or pentad analysis of the model results may help to quantify the extent of upwelling and advection of the cold waters, which is beyond the scope of this paper.

Figure 10(b) represents the vertical profile of temperatures for the winter monsoon of December and January for alternate years from 2000 to 2008. Here also, increased upwelling is indicated from December to January by upwarping of isotherms. As discussed above, in some years there is more intense upwelling than in other years. But compared with summer, there is decreased activity in the winter monsoon as far as upwelling is concerned.

Figure 11 depicts the model-computed vertical temperature cross-section off 10°N and 50–75°E (line 1 of Figure 1, upper panel of Figure 11); and off 5°N and 65–85°E (line 3 of Figure 1, lower panel of Figure 11). The upper panel shows the upwelling phenomena off the southwest coast of India during the summer monsoon period of 2000 and 2002 chosen arbitrarily. Similarly, the lower panel shows the local upwelling that takes place in the south-central Bay of Bengal. Figures 10 and 11 together support the reasoning that the MCP off the STI is formed by cold upwelled waters of the Somalia region as well as from the southwest coast of India driven by the strong summer monsoon current. Subsequently, the MCP intrudes into the south-central BoB and is mixed with the local
upwelled water and hence its sustenance continued until September.

**MCP off the STI: Sea surface current and sea surface temperature**

The most important factor that contributes to the formation, sustenance and dissipation of the MCP off the STI is the SMC during the summer monsoon season. The inter-annual variability of the wind as described earlier is reflected on the model ocean current. Figure 12(a) shows the model ocean current (upper panel) and model ocean SST (lower panel) for alternate years of 2000–2008. The vectors are drawn at every third grid point for the sake of clarity. The variability is less marked in Figure 12(a), possibly because it shows a monthly averaged picture. Here, the current strength, direction and their vertical extent plays an important role as far as the MCP off the STI is concerned, although the role is only to support advection of cold water to the STI and then to south-central BoB. This feature is highlighted in the figures for 2006 and 2008 where the current near the STI region is almost in the zonal direction obstructed by Sri Lanka, indicating hampering of cold water advection. After intruding into the south-central BoB again the current direction helps the spread of the cold tongue more compared with other months. The lag in advection and local upwelling off the STI probably lend a hand in the prolonged existence of the MCP through

![Figure 12(a)](https://iwaponline.com/wqrj/article-pdf/47/3-4/333/163541/333.pdf)
September. The distribution of SST also looks similar as far as inter-annual variability is concerned, but the extent of cooling differs for the MCP off the STI and highlights the fact that the variability is intense on a daily or weekly basis.

The WMC plays an important role in driving the low salinity BoB waters into the AS and dissipation of the MCP in the winter monsoon. To observe the inter-annual variability Figure 12(b) depicts the sea surface currents (upper panel) and sea surface temperature (lower panel) for alternate years of 2000–2008. The figure shows large variability in both direction and magnitude of the currents thereby also changing the evolution of SST. The SST contours clearly show the pathways of WMC and its interaction with the MCP to dissipate it. The currents are mostly westward in direction but differ in some years such as in 2000 and 2008 which show some equator-ward movement near Sri Lanka and the Indian tip. The branching out of WMC to form WICC varies from year to year and hence the extent of intrusion of low saline waters from the BoB to the AS also differs.

**SUMMARY AND CONCLUSIONS**

The MCP off the STI during the summer and the winter monsoon season is simulated using the Princeton Ocean Model. The simulation suggests that the occurrence of the MCP off the STI and its persistence is mainly the result of three factors during the summer monsoon season. The first is the local wind-driven upwelling off the STI, the second is the advection of the Somali or western AS upwelled waters, and the third is the advection of cold water from the Somali region.

Although the upwelling phenomena are inhibited gradually by September, the SMC advected cold water from the western AS still reach the STI, which is responsible for the prolonged presence of the MCP off the STI. The simulation also shows that the
upwelled waters are brought from a depth greater than 100 m. Although the process starts with wind-driven upwelling in May and is intensified by August, its sustenance in September is primarily by the advection of cold upwelled waters by SMC. Intense chlorophyll $a$ blooming in this region confirms upwelling during the summer monsoon season. The model-simulated sea surface current mimics the SMC well compared with ship-drift vectors and shows an anticlockwise curvature making its entry into the south-central BoB. Thus, the summer monsoon cooling in the BoB as indicated by model studies and SST data from different sources is determined by the local upwelling around the STI and advection of cooler waters by the SMC from the western AS and the southwest coast of India. The intra-seasonal and inter-annual variations in the divergence of near surface circulation are primarily responsible for the variability in the intensity and spatial distribution of the MCP off the STI. Weekly analysis of results might throw some light on the relation between the MCP at the south-central Bay of Bengal and active and break phases of the summer monsoon.

During the winter monsoon season, the WMC, which is the main driving force, acts as a deterrent to the formation of the MCP, in contrast to the summer monsoon season. The formation of the MCP is primarily due to local upwelling as supported by increased Ekman transport. The spread of the MCP is trapped by the WMC. It is dissipated by the WMC and ceased to exist in February. The intensity of the MCP both spatially and temporally during the winter is much lower compared with the summer monsoon; the major contributory factor is the difference in monsoonal wind strength. The wind being different in both cases
initiates upwelling of varied strength and at different places. During the summer it is towards the southern tip of India but during winter it is more towards the eastern side of the Gulf or the west coast of Sri Lanka. The heat budget analysis of the mixed layer confirms the above facts. More detailed study of the MCP should be done with reference to the Indian rainfall to assess the correlation between the active and break phase and normal or drought season.

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