Arabian Sea mini warm pool and the monsoon onset vortex

P. N. Vinayachandran[1,]*, D. Shankar[2], J. Kurian[1], F. Durand[2,3] and S. S. C. Shenoi[2]

[1]Centre for Atmospheric and Oceanic Sciences, Indian Institute of Science, Bangalore 560 012, India
[2]National Institute of Oceanography, Dona Paula, Goa 403 004, India
[3]IRD, LEGOS, UMR5566 CNRS-CNES-IRD-UPS, 14 Edouard Belin, 31400, Toulouse, France

Warm pools – regions with sea surface temperature (SST) in excess of 28°C – in the ocean occupy a special place in tropical climate owing to their impact on tropical convection. The southeastern Arabian Sea presents SST in excess of 30°C for two to three months preceding the onset of the summer (southwest) monsoon over India. Recently, several attempts have been made to understand the processes that cause such high SSTs in the region. The literature is also strewn with observations and modelling of a 'monsoon onset vortex' forming over the warm pool. This paper is a review of our current understanding of the Arabian Sea mini warm pool and its possible role in the formation of the monsoon onset vortex.

Keywords: Arabian Sea, mini warm pool, monsoon onset vortex.

TROPICAL atmospheric convection is highly sensitive to the sea surface temperature (SST) of the underlying ocean. Deep atmospheric convection in the tropics is found to increase sharply when SST exceeds a threshold of about 27.5°C[1,2]. The convection is also sensitive to fluctuation in SST above 28°C[3]. Although SST above 28°C is not sufficient to produce organized deep convection in the tropical atmosphere, it is a necessary condition[4–6]. Therefore, oceanic regions with SST above 28°C, known commonly as 'warm pools', occupy a significant place in tropical climate.

A huge warm pool occupies the western Pacific and the Indian Oceans[7,8], where the SST is higher than anywhere else in the world oceans (Figure 1).

During the months preceding the summer monsoon, the SST within this warm pool generally exceeds 29°C (Figure 2). In the Indian Ocean, the warmest SSTs are found in three preferred locations: in the western equatorial Indian Ocean, in the southeastern Arabian Sea (SEAS), and in the eastern Bay of Bengal. The region in the SEAS where temperature exceeds 29°C during February–May has been called the Arabian Sea Mini Warm Pool (ASMWP, Figure 2); it is this ASMWP, which is identifiable even within the huge Indo-Pacific warm pool that forms the focus of this review.

The warming in the SEAS takes place in February–March, well before the surrounding seas[9,10], suggesting that the processes determining the SST response in the SEAS during this season are distinctly different from the rest of the North Indian Ocean, where the rise in SST follows the seasonal march of the sun. There is also speculation that the warm SSTs in the SEAS affect monsoon onset over Kerala and lead to the formation of a monsoon onset vortex[7–11].

Recently, a major field experiment, called the Arabian Sea Monsoon Experiment (ARMEX), was conducted under the Indian Climate Research Program (ICRP) to investigate the role of high SST in the SEAS in the onset process of the Indian summer monsoon. The field experiment, as well as the accompanying modelling studies, has generated new ideas regarding the warm SSTs in the SEAS. A synthesis of these studies is in order, to provide a consolidated account of what has been known and what should be the future direction. This need is the motivation for this review.

Recent reviews have synthesized our current knowledge of the basin-scale monsoon circulation of the Indian Ocean[12] and of the immediate results from the oceanic component of the ARMEX field programme[10] and the ICRP field programmes like BOBMEX (Bay of Bengal Monsoon Experiment) and ARMEX[13]. Notwithstanding these reviews, the exponential increase in our understanding of the oceanography of the SEAS and of the underlying processes since the beginning of ARMEX, from both observational and modelling efforts, needs to be synthesized. In this review, we synthesize all available observations and modelling studies that address different aspects of the ASMWP and attempt to arrive at a comprehensive picture of our present understanding and suggest some directions for future research.

The Arabian Sea mini warm pool

The existence of the ASMWP was first observed in data from the Monsoon Experiment (MONEX) of 1979; these data showed that a mini warm pool (SST > 30.5°C)

*For correspondence. (e-mail: vinay@caos.iisc.ernet.in)

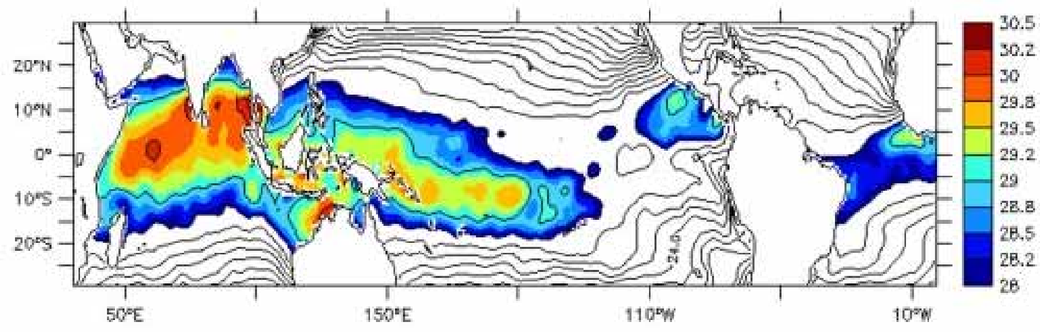

Figure 1. Climatological monthly mean SST during April, from World Ocean Atlas 2005 (WOA05) climatology[84]. Contour interval is 1°C. Temperature higher than 28°C is shown in colour shading.

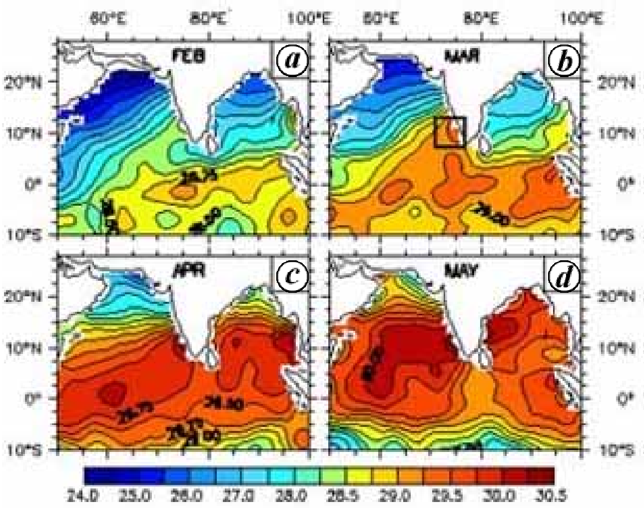

Figure 2. Climatological monthly mean SST (°C) from WOA05 data set[84] during February–May. Contour and shading interval is 0.5°C for values below 28°C and 0.25°C for values above. The box marked in panel (***b***) shows the SEAS.

existed in the SEAS east of 65°E, a week before the genesis of the onset vortex of the Indian summer monsoon[11]. The ASMWP can be identified using the 29°C contour during February–April (Figure 2). This fails in May because the temperature rises to values above 29°C in the southern Arabian Sea. The SST in the SEAS still stands out as the highest in the region[9]. Interannual variability in SST also makes it difficult to define a threshold for delineating the ASMWP. In May, prior to the onset of the summer monsoon, the SEAS become the warmest region of the world ocean[7]. Interest in this warm pool was revived by the investigations of Shenoi *et al.*[9] and Rao and Sivakumar[14] who examined the processes that lead to the formation of the warm pool and implications of its existence.

Observations

The SEAS showcases several features in its hydrography and circulation, the combination of some being unique. These features include:

- The presence of a well-defined warm pool during February–May.
- Large amplitude in the annual cycle of sea surface salinity and the presence of high salinity water at subsurface depths throughout the year and low salinity water at the surface during November–April.
- Formation of the Lakshadweep High, its subsequent westward propagation, and the upwelling and downwelling circulations associated with gyres and eddies.
- Presence of subsurface temperature inversions, particularly during November–February.
- Presence of a barrier layer (the layer between the base of the mixed layer and the top of the thermocline) during November–May.

The oceanographic setting dictated by the above features is crucial for the formation of the ASMWP and we review them below.

Temperature: Early hydrographic surveys[15,16] had shown that the temperature in the SEAS during March–April is higher than 29°C. The depth of the isothermal layer varied between 30 m[17] and 100 m, depending on the location and month of observation. The thermocline, which was located below 75–100 m during November, was observed to shift to shallower depths during this period. Most of these surveys also showed that the temperature is lower to the north of the SEAS, off the Indian coast, consistent with the recent identification of the ASMWP. With the onset of the summer monsoon, the SST decreases considerably, and as the season progresses, the SST near the coast decreases further owing to upwelling[15,16].

Recent hydrographic surveys in the SEAS have made it possible to chronicle the evolution of the warming. After the withdrawal of the summer monsoon, during October–November, the SEAS warms slightly and the SST reaches 28.5°C[18,19]. There are indications that during December–January, SST dips to 28°C[17,18,20–22], although patches of higher SSTs (29°C) are seen in the Lakshadweep Sea during this time[19]. During December–January, the region to the south of the tip of the Indian peninsula cools by more than 1°C, SST dropping to around 27°C[23], whereas SST in the SEAS remains higher than 28°C. During January, SSTs in the range 28–29°C have been observed in various parts of the SEAS[24–26]. Under the regime of strong, dry, northeasterly winds associated with the winter (northeast) monsoon, regions to the south of India cool extensively during January[23], which sets up large SST gradients near 6°N. The SEAS warms considerably from mid-February to March (Figure 3). SSTs in the range 29–30°C have been observed in the SEAS during March[19,27,28]. During the ARMEX observations in March 2003, SST was higher than 30°C in the eastern and southern part of the SEAS, while it was above 29°C to the north and west[10]. These observations illustrate that the SST in the SEAS reaches its peak well before the onset of the summer monsoon.

Conductivity, temperature, depth (CTD) observations during ARMEX[29] have shown a warming of about 1°C from March to May in 2003. SST up to 32°C and above has been observed in localized areas within the SEAS during mid-April to mid–May, with the mean SST remaining above 30°C[18,19,25,28,30–33]. During late May and early June, SST falls below 29°C close to the coast[34,35], but remains above 30°C in the offshore areas[25,32,34,35]. The isothermal layer in the SEAS during the winter monsoon is generally shallow (Figure 4) and there is no signature of its deepening from October till the end of spring. The thermocline, on the other hand, is pushed downward by the prevailing anticyclonic circulation and begins to rise during March–April.

Temperature inversions: The SEAS hosts temperature inversions during October–April. Observations show that inversions are a stable seasonal feature and occur frequently in the SEAS[15,21,36]. During December, the magnitude of inversion is about 0.5°C, and during its peak, the magnitude varies from 0.5°C to 1.2°C[21,37]. The thickness of the inversion layer varies in the range 20–80 m. Inversions typically occur below 20–30 m depth[10,24,38]. Inversions form at shallow depths (~20 m) in the northern part of the SEAS, but occur at deeper depths (80 m) in the southern parts[10,39]. The frequency of occurrence of temperature inversions is low during October–November, increases dramatically during December, peaks during January–February, declines in March, and disappears subsequently[37,38]. XBT observations, in consonance with theory, also show an apparent westward propagation of temperature inversions[38].

The formation of temperature inversions has been attributed primarily to the intrusion of cooler low-salinity water from south[21,37–39] and loss of penetrative radiation to the barrier layer[38,39]. Surface heat loss can also contribute to its formation during the initial stages[21] and on smaller timescales[39]. The shallow inversions observed during April have been attributed to night-time cooling[38]. Inversions in the SEAS are locally formed[39] and near-surface haline stratification is necessary to sustain the temperature inversions on seasonal timescales[39,40].

Sea level and currents: Hydrographic observations and altimeter sea-level data show that an anticyclonic eddy, known as the Lakshadweep High (LH), which has a diameter of 500–800 km, forms in the SEAS during the winter monsoon[41]. The LH has been shown to be a part of a seasonal cycle, which features a Lakshadweep Low (LL) during the summer monsoon[42]. Monthly maps of sea-level anomalies from TOPEX/Poseidon (Figure 5) elucidate this seasonal cycle clearly; the low during the summer monsoon is replaced by the high during winter monsoon, with the transition taking place in the intervening months. Averaged over the SEAS sea level begins to rise during October and reaches its peak during December–January, and then drops continuously till August.

The ocean circulation within the SEAS during the winter monsoon is in a stage of continuous evolution, a characteristic feature of the monsoon currents in the north Indian Ocean[43]. The westward flowing Winter Monsoon Current (WMC) flows westward to the south of India and

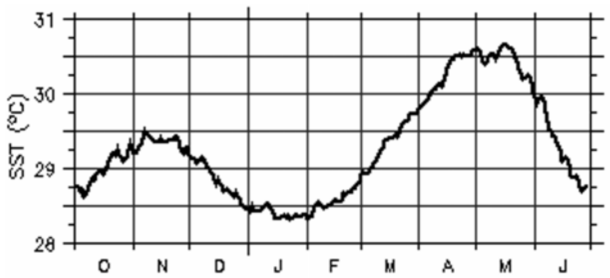

Figure 3. Evolution of SST (°C) over the SEAS (68°–77°E and 6°–15°N) from TMI climatology for the period 2000–2006. The three-day mean TMI SST version 'bmaps_v04' has been obtained from SSMI (http://www.ssmi.com/tmi/tmi_browse.html).

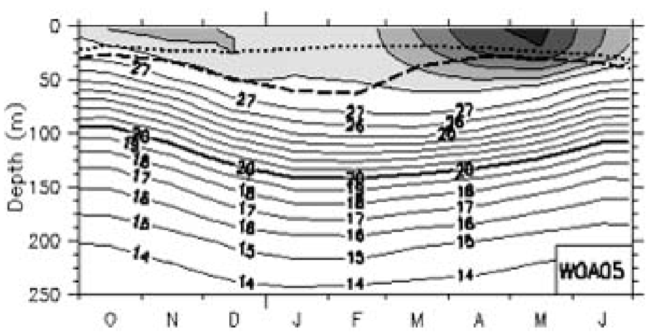

Figure 4. Time–depth section of temperature over the SEAS (68°–77°E and 6°–15°N) from WOA05[84], during October–June. Contour interval is 1° below 28°C and 0.5° above that. Temperature above 28°C is shaded in addition to contouring. The thick continuous line show the 20°C isotherm and thick dashed line is the isothermal layer depth (defined as in the depth at which temperature decreases 0.5°C from SST) and the thick dotted line is mixed layer depth (defined as the depth at which density increases from that at surface by an amount equal to decrease in temperature by 0.5°C from SST).

Sri Lanka. South of Sri Lanka, the WMC is fed by the East India Coastal Current (EICC), which carries low-salinity water from the northern Bay of Bengal. In the SEAS, the WMC splits into two branches; one branch continues westward and the other flows around the LH, in geostrophic balance, feeding into the West India Coastal Current (WICC)[43,44]. The anticyclonic circulation suggests that a major part of the water involved is likely to recirculate within the SEAS. The anticyclonic circulation associated with the LH pushes down the isotherms and isohalines, and thus actively modifies the SEAS hydrography[45]. This downwelling is clearly seen in model simulation[39,46].

Salinity: The connection of the open-ocean monsoon current, the WMC, with the coastal current, the EICC, enables the transport of low-salinity water from the Bay of Bengal into the SEAS. The supply of low-salinity water occurs during November–February (WOA05), and, as a result, the salinity in the SEAS drops by about 1.5 psu from October to February (Figure 6). The low-salinity water trapped within the anticyclonic circulation creates a fresh pool in the SEAS, causing strong stratification in the near-surface layers; it is believed to be one of the reasons for the warming of the SEAS.

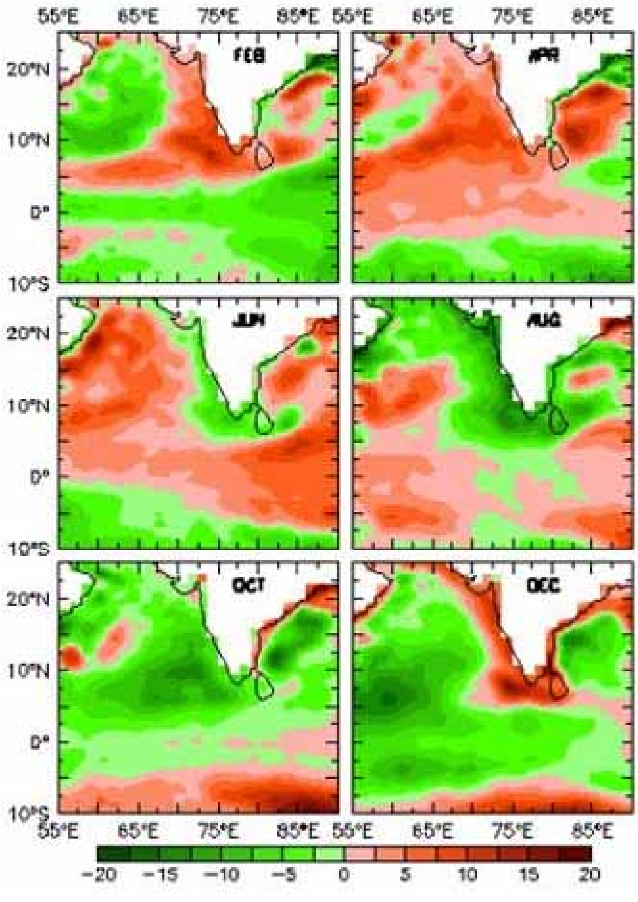

Figure 5. Climatology of the annual cycle of sea level anomalies (cm) from AVISO merged product of satellite altimeters.

During October 1957, Darbyshire[15] observed that salinity decreased with depth and to the south off the west coast of India between 8.45°N and the southern tip of India in the upper 100 m. Salinity was lower close to the coast, which was attributed to the discharge of freshwater from the backwaters of Kerala during and after the summer monsoon. During January 1959, Banse[16] observed salinity below 33 psu off the continental slope around 10°N. Observations during December 1987 also have shown low-salinity waters near the coast, a salinity maximum below 75 m, and negligible salinity stratification below 150 m[20]. During December 1985 to January 1986, salinity along the Indian coast was less than 33 ppt in the south and well above 35.5 ppt in the north[47]. While the low-salinity water lay over high-salinity water in the south, the water column was well-mixed in the north. Lack of systematically collected observations precludes the delineation of the pathways of this low-salinity water at the surface. Nevertheless, available observations indicate that the low-salinity water advects along the west coast of India as well as westward along with the WMC[20,37,38]. There seems to be considerable interannual variation in the sea surface salinity in the SEAS, particularly between good and bad monsoon years[37]. All available observations of salinity and currents suggest that the source region of the low-salinity water in the SEAS is the Bay of Bengal, the low-salinity water being carried equatorward along the Indian east coast by the EICC and then into the SEAS by the WMC and WICC[15,20,47]. The high-salinity water seen below the fresher surface layer originates from the northern Arabian Sea. When the WICC flows southward along the Indian coast during summer, it carries the Arabian Sea High Salinity Water into the SEAS. This high-salinity water appears as a salinity maximum during November–February.

Barrier layer: The capping of the upper layer by low-salinity water creates strong stratification in the upper layer. The mixed layer is much thinner than the isothermal layer, resulting in the formation of a barrier layer. ARMEX observations showed the existence of a 20 m thick barrier

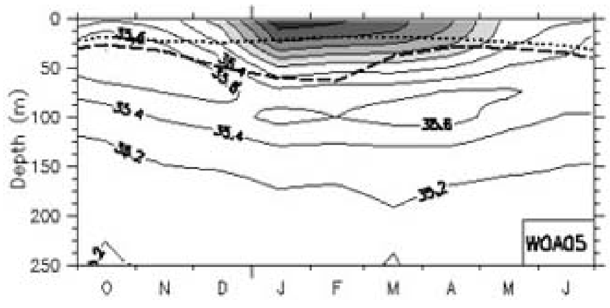

Figure 6. As in Figure 4, but for salinity from WOA05[85]. Contour interval is 0.2 psu and salinity less than 35 psu is shaded in addition to contouring.

layer in the SEAS during March–April 2003. The barrier layer had formed earlier during December–January owing to a surface layer of low-salinity water advected from the Bay of Bengal[48]. Based on the analysis of *in situ* and satellite data, Shenoi *et al.*[48] argued that the process of annihilation of barrier layer starts in early April and is caused first by remotely forced upwelling and then in May by the advection of high-salinity surface water from the north into the SEAS.

The thermodynamic impact of the barrier layer on the SST is not simple. The mechanisms by which a barrier layer can induce a SST rise are twofold. First, the thinner mixed layer, in the presence of a barrier layer, is more reactive to the incoming atmospheric heat fluxes. Second, in situations where the barrier layer also supports a temperature inversion with warmer water than the surface layer, the SST can increase by mixing of the surface water with the underlying inversion layer[45–49]. The barrier layer, however, has also the potential to induce a drop in SST. This is typically the case when the ocean–atmosphere net heat flux is close to zero, and when the mixed layer is fairly shallow; under such conditions, a significant part of the incoming shortwave radiation can escape through the bottom of the mixed layer, thereby inducing a warming of the barrier layer water rather than of the surface layer[39,49]. All these effects are not exclusive. Hence, the overall impact of the barrier layer on the SST is difficult to assess. Quantifying its net effect from the available observations is not feasible.

Modelling

Most of the earlier modelling efforts in the Indian Ocean were aimed at simulating the seasonal cycle of temperature and currents and little attention was paid to salinity. Recent modelling studies, however, have had considerable success in simulating the evolution of the ASMWP, in which the role of salinity has been emphasized by the observational studies, and the associated circulation; these modelling studies have also been successful in analysing the processes that are involved in the formation of the ASMWP. Most of these studies came in the wake of the ARMEX observations and these results have not been synthesized yet.

The first comprehensive analysis of the processes involved in the upper–ocean thermodynamics of the Indian Ocean was the seminal work of McCreary *et al.*[50], who used a two-and-half-layer reduced-gravity model that incorporated mixed-layer physics (but did not include salinity). They did not focus on the ASMWP, but their simulations showed the formation of the LH (the feature was yet to be named thus) and a mini warm pool, with SST greater than 29°C, during the winter monsoon in the SEAS. The following year came the formal report of the LH (then called 'Laccadive High') in altimetry and historical hydrography by Bruce *et al.*[41], who also used a three-layer reduced-gravity model (which did not, however, include a mixed layer) to show that the LH could be simulated using climatological winds and was due to both local and remote forcing. Tracer experiments showed that the source region of low-salinity water in the SEAS is the northern Bay of Bengal; the bifurcation of the WMC east of the Maldives led to the intrusion of this low-salinity water into the LH. Simulations with the OPA OGCM[51] (Ocean General Circulation Model) with a resolution of 0.5° confirmed this finding a decade later[45]. Lagrangian particles experiments showed that the low-salinity water reaches the SEAS via two different routes, one via the EICC and the other via the WMC south-east of Sri Lanka. Before entering the SEAS, these two branches merge with a third branch of higher salinity water that recirculates within the SEAS[45].

An analysis of the dynamics of the LH and LL with a reduced-gravity model and an analytic model showed that the LH and LL, which form part of the annual cycle of circulation in the SEAS, are a consequence of linear wave dynamics[42]. The analysis showed that these features are due to westward propagating Rossby waves that are radiated by the coastal Kelvin waves propagating poleward along the west coast of India. The LH was forced primarily by the winds blowing along the east coast of India (remote forcing), but the LL owed its existence to these winds (remote forcing) as well as the winds blowing along the southwest coast (local forcing)[42,43].

Most numerical models have had rather limited success in bringing low-salinity water from the Bay of Bengal into the SEAS at the correct time. The surface salinity simulation in most of the models[41,52–56] shows a weakness in the northward advection of low-salinity water along the west coast of India. Han and McCreary[57] noted that the arrival of the low-salinity water in the SEAS was delayed by about a month if the Palk Strait was closed in their four-and-half-layer model, a result confirmed by OGCM simulations[45]. The OGCM simulations of Durand *et al.*[45], which included a realistic representation of the climatological river discharge, also showed that inclusion of the runoff from the minor rivers scattered along the Indian coasts, which discharge freshwater to the Arabian Sea during the monsoonal and post-monsoonal months, but are generally not accounted for in the standard forcing strategy of numerical models, curb to some extent the salty bias of the SEAS simulations.

The problem of simulating the arrival of low-salinity water into the SEAS is not trivial. The EICC reverses to flow poleward first in the north, where the source of freshwater lies, and then in the south off India. Off Sri Lanka, the EICC flows equatorward over most of the year[58,59]. Hence, for the low-salinity water to make its way into the SEAS, it is necessary that it reaches the east coast of Sri Lanka before the EICC reverses off India in December–January. It appears that three conditions have to be satisfied by numerical models in order to simulate correctly the salinity in the SEAS and further north along the Indian west coast. First, the model wind forcing should have high temporal resolution. The strong interannual variability seen in the salinity in the SEAS (with respect to both timing and values)[37] suggests that it is difficult to capture this intrusion with climatological wind forcing because climatological winds have a poor temporal resolution. Second, the need to get right speed of the EICC implies the need for a reasonably high model resolution to resolve the coastally trapped current; else, the model EICC will be weaker, resulting in less freshwater making its way to the SEAS before the EICC reverses. Third, as shown by Durand *et al*.[45], inclusion of the discharge from minor rivers is important. Most of the model studies satisfy one of the three conditions. It is therefore necessary to carry out a complete set of sensitivity experiments to decide what determines the salinity in the SEAS. A recent model simulation by Kurian and Vinayachandran[46], which met most of these conditions, showed considerable improvement (see below) in the simulation of arrival time of low salinity water into the SEAS.

Shenoi *et al*.[9] and Rao and Sivakumar[14] examined the processes that lead to the formation of the ASMWP using climatological data sets. The feature that distinguishes the SEAS most, from the surroundings, is the arrival of low-salinity water and its trapping within the LH by the anticyclonic circulation. The ensuing stratification causes the mixed layer to shoal due to the formation of barrier layer. Considering that the heat flux falling on the SEAS is not significantly different from the rest of the North Indian Ocean, they suggested that the difference in trapping of the radiation must be necessary for the higher warming of the SEAS. Stratification caused by the low-salinity cap was central to this hypothesis.

The availability of OGCMs with high enough resolution in the horizontal, and realistic enough vertical physics, has allowed this hypothesis to be tested. Durand *et al*.[40] used a seasonal simulation of OPA OGCM at 0.5° resolution[51]. They found that the SST build-up in the SEAS during the pre-monsoonal season is primarily driven by subsurface heating from the inversion layer embedded in the barrier layer.

Recently, Kurian and Vinayachandran[46] (hereafter KV07) used an Indian Ocean configuration of MOM4 (Modular Ocean Model Version 4)[60] forced by (daily) climatological forcing to examine the dynamics and thermodynamics of the SEAS. The model had a high resolution (0.25°) in the horizontal and 40 levels in the vertical and included climatological river discharge. This model simulation offered excellent comparison with the climatological distribution of salinity in the SEAS (see their figure 4). This model reproduced the arrival of low-salinity water into the bay in a manner similar to that in climatology (Figure 7). In order to test the hypothesis that salinity plays an important role in the formation of the ASMWP, KV07 carried out an experiment by prescribing and maintaining a uniform salinity of 35 psu over the entire model domain. Surprisingly, in this experiment, the ASMWP formed just as in the control run (Figure 8), which included salinity variations, suggesting that factors other than salinity are responsible for the formation of the ASMWP. KV07 also evaluated the temperature equation in order to determine the role of subsurface warming on the SST. They found that the contribution from temperature inversions to the warming of the SEAS is negligibly small. KV07 re-examined the role of surface fluxes and found that a low in latent-heat flux over the SEAS during the winter monsoon has implications for the warming of the SEAS during February–March. The winds over the SEAS are blocked by the Western Ghats (Figure 9), and consequently the latent-heat loss over the SEAS is much lower than over its surroundings. In their model, this process leads to the warming during February–March and the later warming in April–May was due to local air sea fluxes.

To our knowledge, the only modelling study to specifically address the issue of the interannual variability of the ASMWP SST is that of de Boyer Montégut *et al*.[61]. They used a model very similar to that of Durand *et al*.[40]. They did not focus explicitly on the SEAS, but rather analysed the whole eastern Arabian Sea. The seasonal picture of their interannual simulation for 1992–2001 is consistent

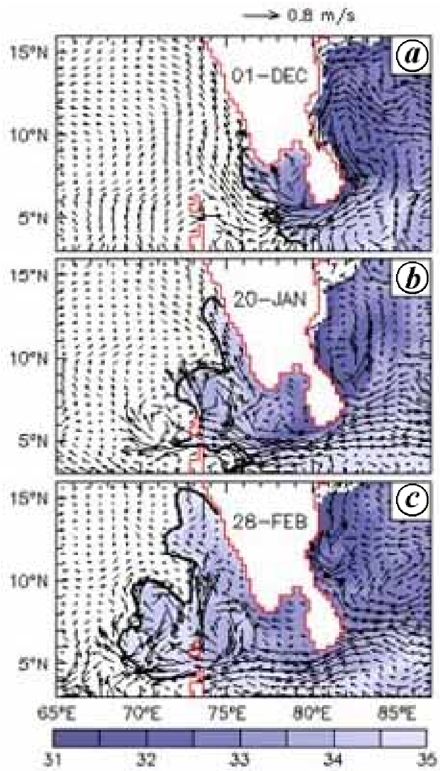

Figure 7. Currents (vectors) and salinity (shading and contour) in the SEAS during December–February from climatological model simulation[46]. Surface salinity less than 35 psu is shaded. The contour shows 35 psu isohaline.

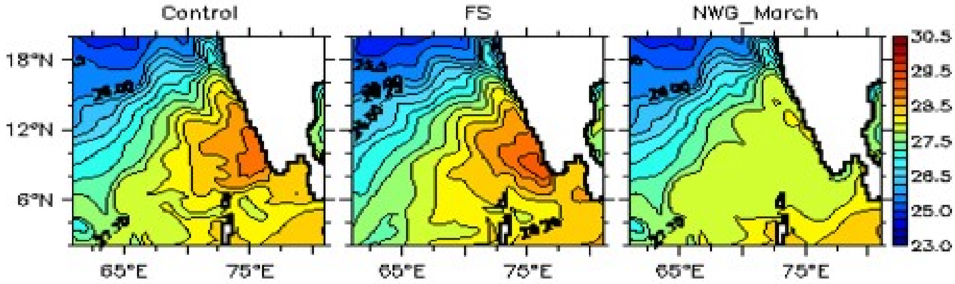

Figure 8. SST from different experiments described in KV07[46]. Control is the standard experiment, without any modification to the forcing fields and it has realistic freshwater forcing. In FS (fixed salinity) experiment, the salinity has been held constant at 35 psu everywhere in the model domain and the freshwater forcing has been switched off. NWG_March (no Western Ghats during March) is similar to the control run, except that the wind speed, specific humidity, air temperature, and wind stress over the SEAS have been replaced by average values over the region west of the SEAS during March in the sixth of the control run. For details, see KV07.

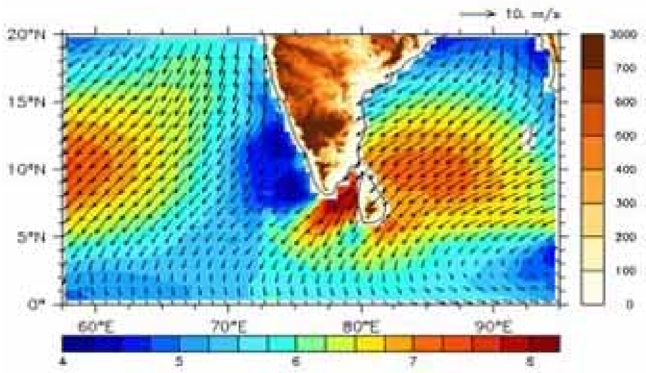

Figure 9. November–February mean winds (vectors, m/s) from QuikSCAT climatology for the period 2000–2006. Colour shading over ocean represents wind speed with the key shown at the bottom and colour shading over land represents topography (m) with the key shown to the right of the plot.

with that of Durand *et al.*[40], with the ASMWP being caused by warming from the temperature inversions in the barrier layer. At interannual timescales, on the contrary, the picture is basically in line with the mechanism revealed by KV07 in the SEAS at seasonal timescales: the year-to-year variability of SST over the eastern Arabian Sea is mostly driven by the year-to-year variability of ocean–atmosphere heat fluxes. Only over specific periods did the vertical oceanic physics act to modulate the atmospheric effect; one such example was the anomalously cold winter of 1998.

Clearly, the state-of-the-art oceanic general circulation models have proved to be useful tools for exploring the processes responsible for the ASMWP thermal structure and variability. Nevertheless, those two models of the same class, that of Durand *et al.*[40] and KV07, that yielded completely different results, open several issues in numerical modelling. Although they used comparable forcing strategies, the resolution of KV07, in both horizontal and vertical directions, is significantly higher than that in Durand *et al.*[40]. In order to settle these issues as well as to determine issues limiting the realism of the OGCM, carefully designed sensitivity experiments should be undertaken. For example, the impact of the parameterizations that is inherent to any OGCM need to be ascertained. This can be done by running various models using the same resolution and the same forcing strategy. We must also assess the impact of the sub-grid scale physics neglected in the previous studies. The method of incorporating air–sea heat flux in the model also requires careful evaluation. The circulation and thermohaline structure simulated by an OGCM changes a lot when the resolution is increased from mesoscale up to the kilometric scale (e.g. Lévy *et al.*[62]). How our understanding of the ASMWP behaviour will change once we are able to simulate explicitly the whole oceanic spectrum (including the sub-mesoscale turbulence) remains unknown. Given the explosion in computing capability, this challenge could be taken up soon. Finally, a major source of uncertainty in the realism of the models of the ASMWP lies in the air–sea boundary conditions that are specified at the ocean surface. At present, we have to rely on remotely sensed datasets to prescribe these conditions. The possibility of validation of the satellite products (typically heat fluxes and atmospheric variables) by independent observations is severely limited by the scarcity of *in situ* data in the ASMWP.

Monsoon onset vortex

The Indian summer monsoon is a complex phenomenon that involves processes coupling the atmosphere, the ocean and the land. Though almost all aspects of the monsoon are fascinating (and present a formidable research challenge), perhaps the most dramatic feature of the monsoon is its spectacular onset. (See Frater[63] for a popular description of the drama involved in a monsoon onset.)

The monsoon sets in over the Indian subcontinent over the southwestern state of Kerala. The normal date of monsoon onset over Kerala is 1 June (mean date is 30 May, 1

June being the mode and median date), the standard deviation being eight days[64]. Associated with the onset is a sharp increase in the sea-level-pressure difference between Thiruvananthapuram and Mumbai[64]. This pressure difference fluctuates around 0.5 mbar in the fortnight before the onset, but rises sharply to about 3 mbar around two days before the onset. It peaks at about 5–10 mbar 4–8 days after onset. Ananthakrishnan *et al.*[64] speculated that this increase in pressure gradient was associated with the low-pressure system or onset vortex that forms at around 10°N in the SEAS and moves northward as the monsoon advances along the Indian west coast[65,66]. The vortex, an intense cyclonic circulation, accelerates the northward advance of the monsoon over India[67,68]. It often deepens into a cyclonic storm, resulting in establishment of a region of convergence over the SEAS, which is followed by the strengthening of monsoon westerlies. In turn, the westerlies lead to sustained rainfall over Kerala, heralding the onset of the summer monsoon[67]. An exhaustive description of the occurrence of the vortex since 1901 is available in Ananthakrishnan *et al.*[65].

A typical example of the vortex is shown in Figure 10 for 1998. Warm SST is seen in the SEAS prior to the monsoon onset over Kerala (the onset date over Kerala was 2 June in 1998). The OLR (outgoing longwave radiation) map clearly shows low values, indicating deep convection over the SEAS and winds indicate cyclonic circulation.

One such vortex was observed over the SEAS during MONEX-79 after the low-level westerly flow had intensified. This prompted Krishnamurti *et al.*[69] to propose that barotropic instability due to horizontal shear leads to the formation of the vortex. Mak and Kao[70] suggested, however, that vertical shear and stratification in the atmosphere are responsible for the genesis of the vortex. Though there have been several studies using atmospheric models (see Krishnamurti[71] for a review of MONEX), successful prediction of the monsoon during its onset phase proved elusive (see, for example, Krishnamurti *et al.*[72]).

In an attempt to predict the onset date of the monsoon over Kerala, Kung and Sharif[73] found it necessary to include the SST in a 10° × 10° box in the southern Arabian Sea; excluding this SST from the multi-parameter regression led to a poorer prediction. This box was much larger than what we now call the mini warm pool, but SST data were sparser and less accurate in those days. Using MONEX data, Seetaramayya and Master[11] showed that a mini warm pool (SST > 30.5°C) existed in the SEAS east of 65°E, a week before the genesis of the onset vortex. Kershaw[74,75] used their SST analysis to demonstrate that predictability of the formation of the depression (vortex) on 11 June 1979, coinciding with monsoon onset over the Indian subcontinent, improved when he used SST anomalies rather than the SST climatology. This study not only highlighted the relation between the SST in the SEAS and the onset of monsoon, but also pointed to the need for more accurate SSTs in the SEAS. Since interannual variability in this region's SST is small, accurate SSTs and special temporal filters are needed to bring out a relationship between Indian–Ocean SST and the monsoon[76].

Joseph[7], using better SST data than was available to Kung and Sharif[73], refined their 10° × 10° SST box to focus on the SEAS. He noted that the onset date and SST in the SEAS showed decadal variability, later onsets during the 1900s and 1970s and earlier onsets during the 1950s coinciding with lower and higher SSTs[77], respectively. Joseph and Pillai[78] had already suggested a link between the warm SEAS and the pre-monsoon rains in Kerala during mid-April.

Following these studies, which primarily came in the wake of MONEX, interest in monsoon onset seemed to wane. Joseph[7] had pointed to the need for studying the warm pool and its possible role in the Indian monsoon, but this did not happen for a decade, by when there was

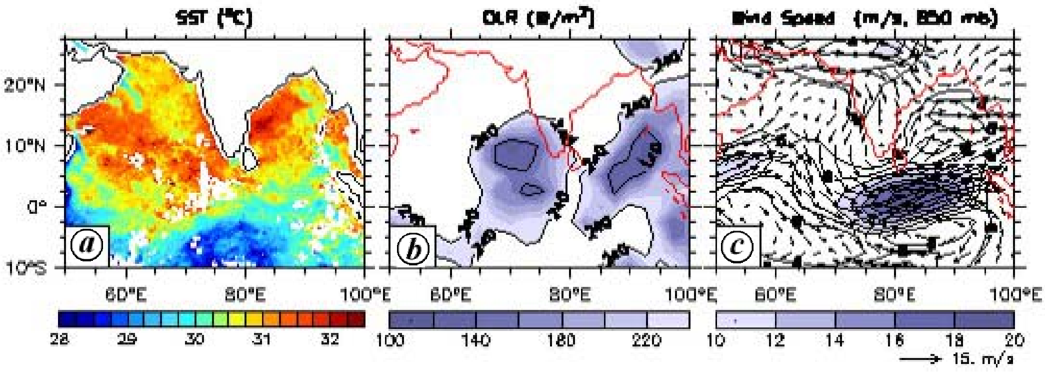

Figure 10. (*a*) SST for 25 May 1998, one week prior to monsoon onset over Kerala from TMI dataset, (*b*) NOAA interpolated Outgoing Longwave Radiation (OLR, W/m^2) for 1 June 1998 and (*c*) 850 mbar winds (m/s) from NCEP reanalysis[82] for 1 June 1998. In (*b*), contours correspond to 240 and 140 W/m^2 and values less than 240 are shaded. In (*c*), wind speed above 6 m/s are contoured at an interval of 2 m/s and values above 10 m/s are highlighted by colour shading. The onset date for summer monsoon over Kerala in 1998 was 2 June.

much more data from ship surveys and data from satellites that had created a revolution in meteorology and oceanography (see, for example, the review by Srinivasan and Joshi, this issue). Analysing the available historical data for 120 years (1877–1996), Rao and Sivakumar[14] concluded that the onset vortex forms either in the SEAS during May or in the eastern Arabian Sea during June (see their figure 9). They found a distinct correspondence between the maximum in the upper ocean heat content (with respect to 28°C isotherm) in May and June and the location of genesis of the vortex. They also found that the vortex mostly forms during late May in the SEAS, where the seasonal build up of thermal energy is largest. They opined that a great deal of thermal energy is then extracted from the SEAS during the genesis of the vortex, the vortex subsequently shifting its position in June to the north, where the available thermal energy is larger following the cooling of the SEAS (see Shenoi *et al.*[48]).

The other major development during the 1990s, coinciding with and due to the massive increase in available data and computer technology, was a significant improvement in our understanding of the dynamics of the north Indian Ocean on seasonal timescales. Numerical models of ocean circulation had improved to the point that they could be used to study processes. Application of the new knowledge to problems of air–sea interaction associated with the monsoon, including its onset, was a natural corollary. Following the demonstration of the importance of remote forcing from the Bay of Bengal for the circulation in the SEAS[42,50], Shenoi *et al.*[9] argued that the downwelling associated with the LH played a role in keeping the SST higher in the SEAS as compared to the surrounding Arabian Sea. This core of high SST, evident as early as March, retained its identity within the larger Arabian Sea warm pool till the onset of the summer monsoon. Assuming that the co-location of the LH, the SST high, and the onset vortex in the SEAS could not be a mere coincidence, Shenoi *et al.*[9] argued that the warm SST in the SEAS satisfies a condition necessary (but not sufficient) for the genesis of the onset vortex, and hypothesized a classical coupled ocean–atmosphere system work in the region. According to them: (1) the collapse of the summer monsoon winds in October and the onset of the winter monsoon over the north Indian Ocean generates downwelling Kelvin waves at the equator and along the east coast of India; (2) the Kelvin waves from the bay enter the Arabian Sea along the southwest coast of India, leading to downwelling and therefore to a stable surface layer in the SEAS, permitting warm SSTs there; (3) the ocean provides necessary/favourable conditions to trigger an event, the monsoon onset vortex, in the atmosphere (see the schematic in figure 7 in Shenoi *et al.*[9]).

The papers by Rao and Sivakumar[14] and Shenoi *et al.*[9] formed the basis for the second phase of ARMEX (see Shenoi *et al.*[10]), the results of which have added considerably to our understanding of the SEAS.

Following the observational and modelling studies based on ARMEX data, Masson *et al.*[79] used a coupled general circulation model to examine the potential impact of the barrier layer in the SEAS on monsoon onset. Their 100-year control simulation reproduced the observed climatic features of the region reasonably well. Then, in a sensitivity experiment, they suppressed, but only in the SEAS, the impact of salinity stratification on the vertical mixing (as done by Vialard and Delecluse[49]). This localized perturbation of the model physics had a significant, but local, impact: the SEAS was cooler during April–May in the experimental run, the rainfall over the SEAS decreased by 1–3 mm/day, and the onset of the monsoon, inferred from the burst in rainfall, was delayed by 10–15 days. Thus, though the perturbation of the model physics in the SEAS had no impact on the rest of India, the impact on the monsoon onset in the region was significant. The resolution of the ocean model in this study was very low (2°); hence, in the light of earlier studies, the results of this coupled experiment need to be verified at higher resolutions.

Evidence from satellite data (Reynolds[80] SST) and model reanalyses[81] (850 hPa winds from NCEP/NCAR) also suggests a possible influence of the warm pool on the onset vortex[82]. This analysis, though restricted to six years, showed that the mini warm pool was important for the formation of the onset vortex, the onset vortex forming when the mini warm pool and low-level jet[83] are present along the west coast and the cyclonic shear at 850 hPa along the northern flank of the jet overlies the mini warm pool. This study suggests a possible connection between the mini warm pool and the onset of the summer monsoon, but more data and model studies are necessary to test the hypothesis and refine it.

The definition of the mini warm pool used by Deepa *et al.*[82] was different; the mini warm pool of Deepa *et al.* was the patch of highest SST in the Arabian Sea and it was not necessarily restricted to the SEAS. This study underscores the need to discriminate between the high SSTs in the SEAS and patches of high SST elsewhere in the Arabian Sea. Any patch of high SST can justifiably be called a mini warm pool, but what distinguishes the mini warm pool in the SEAS is its repeatability. As shown by Shenoi *et al.*[9], this SST high in the SEAS can be discerned even in a climatology of Reynolds SST[81] and is seen to appear, with some difference in timing, almost every year. This repeatability has to be clearly demonstrated for other patches of high SST for them to be analysed similarly. Else, we are left with a confusing picture that becomes even more blurred when air–sea interaction and the onset vortex are part of it.

Concluding remarks

Oceanic observations illustrate that the SEAS is the warmest region in the North Indian Ocean prior to the summer monsoon and this warming takes place earlier than over the surrounding seas. A significant enhancement in our knowledge of the ASMWP has taken place recently owing to ARMEX programme, one of the major observational programmes conducted under ICRP[10,13]. Most of this new information is derived from rather 'patchy' data collected during individual cruises. A systematically collected time history of the evolution of the ASMWP is still lacking. An important point here concerns the arrival of low-salinity water into the SEAS. In particular, the ARMEX observations could not delineate the pathways of intrusion of low-salinity water into the SEAS.

Numerical experiments aimed at unravelling the processes that lead to the formation of the ASMWP have not given a conclusive answer on whether oceanic process dominate air–sea fluxes in warm pool formation. While KV07 suggest that the air–sea fluxes are the key factor, Durand *et al.*[40] argue in favour of oceanic processes. An intercomparison of their models would be useful in order to shed light on the model dependency of these results. A pre-requisite for model studies to unravel the processes responsible for the formation of the ASMWP is their capability to reproduce the observed features of the SEAS and this demands a systematic set of observations. At present, it appears that our modelling ability is superior to the quality of available data. High-quality time-series measurements are needed to make unambiguous progress.

There is also considerable ambiguity as far as the onset vortex is concerned. While the ASMWP is a regular feature of the seasonal cycle in the SEAS, the onset vortex displays no such regularity. It is obvious that the mere existence of the ASMWP is not a sufficient condition to guarantee the existence of an onset vortex. More reliable data and better descriptions of the process of monsoon onset are needed. It is necessary, however, to ensure that such descriptions are based on atmospheric and oceanic data as well as data from land-based meteorological observatories, i.e. they look at the process of monsoon onset as a coupled air–sea–land process. In modelling such coupled phenomena, however, our modelling ability is still poor. The model resolutions are still too low to capture several of the processes that atmospheric and oceanic models suggest are important.

1. Gadgil, S., Joseph, P. V. and Joshi, N. V., Ocean–atmosphere coupling over monsoon regions. *Nature*, 1984, **312**, 141–143.
2. Graham, N. E. and Barnett, T. P., Sea surface temperature, surface wind divergence, and convection over tropical oceans. *Science*, 1987, **238**, 657–659.
3. Waliser, D. E. and Graham, N. E., Convective cloud systems and warm–pool sea surface temperatures: Coupled interactions and self-regulation. *J. Geophys. Res.*, 1993, **98**, 12881–12894.
4. Bony, S., Lau, K. M. and Sud, Y. C., Sea surface temperature and large scale circulation influences on tropical greenhouse effect and cloud radiative forcing. *J. Climate*, 1997, **10**, 2055–2077.
5. Lau, K. M., Wu, H. T. and Bony, S., The role of large-scale atmospheric circulation in the relationship between tropical convection and sea surface temperature. *J. Climate*, 1997, **10**, 381–392.
6. Gadgil, S., The Indian monsoon and its variability. *Annu. Rev. Earth Planet. Sci.*, 2003, **31**, 429–467.
7. Joseph, P. V., Warm pool over the Indian Ocean and monsoon onset. *Trop. Ocean–Atmos. Newslett.*, 1990, Winter, 1–5.
8. Vinayachandran, P. N. and Shetye, S. R., The warm pool in the Indian Ocean. *Proc. Indian Acad. Sci.* (*Earth Planet. Sci.*), 1991, **100**, 165–175.
9. Shenoi, S. S. C., Shankar, D. and Shetye, S. R., On the sea surface temperature high in the Lakshadweep Sea before the onset of southwest monsoon. *J. Geophys. Res.*, 1999, **104**, 15703–15712.
10. Shenoi, S. S. C., Shankar, D., Gopalakrishna, V. V. and Durand, F., Role of ocean in the genesis and annihilation of the core of the warm pool in the southeastern Arabian Sea. *Mausam*, 2005, **56**, 147–160.
11. Seetaramayya, P. and Master, A., Observed air–sea interface conditions and a monsoon depression during MONEX–79. *Arch. Meteorol. Geophys. Biocl.*, 1984, **A33**, 61–67.
12. Schott, F. A. and McCreary, J. P., The monsoon circulation of the Indian Ocean. *Prog. Oceanogr.*, 2001, **51**, 1–123.
13. Sanjeeva Rao, P. and Sikka, D. R., Intraseasonal variability of the summer monsoon over the north Indian Ocean as revealed by the BOBMEX and ARMEX field programs. *Pure Appl. Geophys.*, 2005, **162**, 1481–1510.
14. Rao, R. R. and Sivakumar, R., On the possible mechanisms of the evolution of a mini-warm pool during the pre-summer monsoon season and the onset vortex in the southeastern Arabian Sea. *Q. J. R. Meteorol. Soc.*, 1999, **125**, 787–809.
15. Darbyshire, M., The surface waters off Kerala, south-west India. *Deep-Sea Res.*, 1967, **14**, 295–320.
16. Banse, K., Hydrography of the Arabian Sea Shelf of India and Pakistan and effects on demersal fishes. *Deep-Sea Res. Oceanogr. Abstr.*, 1968, **15**, 45–48.
17. Shetye, S. R., Seasonal variability of the temperature field off the southwest coast of India. *Proc. Indian Acad. Sci.* (*Earth Planet. Sci.*), 1984, **93**, 399–411.
18. Srinivas, K. and Dinesh Kumar, P. K., Atmospheric forcing on the seasonal variability of sea level at Cochin, southwest coast of India. *Cont. Shelf Res.*, 2006, **26**, 1113–1133.
19. Luis, J. A. and Kawamura, H., Seasonal SST patterns along the west India shelf inferred from AVHRR. *Remote Sensing Environ.*, 2003, **86**, 206–215.
20. Shetye, S. R., Gouveia, A. D., Shenoi, S. S. C., Michael, G. S., Sundar, D., Almeida, A. M. and Santanam, K., The coastal current of western India during the northeast monsoon. *Deep–Sea Res. Part A*, 1991, **38**, 1517–1529.
21. Thadathil, P. and Ghosh, A. K., Surface layer temperature inversion in the Arabian Sea during winter, *J. Oceanogr.*, 1992, **48**, 293–304.
22. Antony, M. K., Narayanaswamy, G. and Somayajulu, Y. K., Offshore limit of coastal ocean variability identified from hydrography and altimeter data in the eastern Arabian Sea. *Cont. Shelf Res.*, 2002, **22**, 2525–2536.
23. Luis, A. J. and Kawamura, H., Wintertime wind forcing and sea surface cooling near the south India tip observed using NSCAT and AVHRR. *Remote Sensing Environ.*, 2000, **73**, 55–64.
24. Prasanna Kumar, S. *et al.*, Intrusion of the Bay of Bengal water into the Arabian Sea during winter monsoon and associated chemical and biological response. *Geophys. Res. Lett.*, 2004, **31**, doi:10.1029/2004GL020247.
25. Sanilkumar, K. V., Hareesh Kumar, P. V., Jossia Joseph and Panigrahi, J. K., Arabian Sea mini warm pool during May 2000. *Curr. Sci.*, 2004, **86**, 180–184.
26. Kumar, S. and Ramesh, R., 15N enrichment in the surface particulate organic nitrogen of the north-eastern Arabian Sea from the middle to the waning phase of the winter monsoon: possible causes. *Ocean Sci. Discuss.*, 2007, **4**, 245–264.

27. Ramesh Babu, V., Varkey, M. J., Kesava Das, V. and Gouveia, A. D., Water masses and general hydrography along the west coast of India during early March. *Indian J. Mar. Sci.*, 1980, **9**, 82–89.
28. Jossia Joseph, K., Harikrishnan, M., Rajesh, G. and Premkumar, K., Moored buoy observations in Arabian Sea warm pool. *Mausam*, 2005, **56**, 161–168.
29. Shenoi, S. S. C. *et al*., Hydrography and water masses in the southeastern Arabian Sea during March–June 2003. *J. Earth Syst. Sci.*, 2005, **114**, 475–491.
30. Hareesh Kumar, P. V. and Mohan Kumar, N., On the flow and thermohaline structure off Cochin during pre-monsoon season. *Cont. Shelf Res.*, 1996, **16**, 457–468.
31. Sengupta, D., Ray, P. K. and Bhat, G. S., Spring warming of the Arabian Sea and Bay of Bengal from buoy data. *Geophys. Res. Lett.*, 2002, **29**, doi:10.1029/2002GL015340.
32. Hareesh Kumar, P. V., Madhusoodanan, P., Ajai Kumar, M. P. and Raghunadha Rao, A., Characteristics of Arabian Sea mini warm pool during May 2003. *Mausam*, 2005, **56**, 169–174.
33. Rajesh, G., Jacob, K. J., Harikrishnan, M. and Premkumar, K., Observations on extreme meteorological and oceanographic parameters in the Indian seas. *Curr. Sci.*, 2005, **88**, 1279–1282.
34. Ramesh Babu, V., Sastry, J. S., Gopalakrishna, V. V. and Rama Raju, D. V., Premonsoonal water characteristics and circulation in the east central Arabian Sea. *Proc. Indian Acad. Sci.* (*Earth Planet. Sci.*), 1991, **100**, 55–68.
35. Singh, O. P. and Hatwar, H. R., Response of sea state to the monsoon onset. *Mausam*, 2005, **56**, 59–64.
36. Thompson, B., Gnanaseelan, C. and Salvekar, P. S., Seasonal evolution of temperature inversions in the north Indian Ocean. *Curr. Sci.*, 2006, **90**, 697–704.
37. Gopalakrishna, V. V., Johnson, Z., Salgaonkar, G., Nisha, K., Rajan, C. K. and Rao, R. R., Observed variability of sea surface salinity and thermal inversions in the Lakshadweep Sea during contrast monsoons. *Geophys. Res. Lett.*, 2005, **32**, doi: 10.1029/2005GL023280.
38. Shankar, D. *et al*., Observational evidence for westward propagation of temperature inversions in the southeastern Arabian Sea. *Geophys. Res. Lett.*, 2004, **31**, doi:10.1029/2004GL019652
39. Kurian, J. and Vinayachandran, P. N., Formation mechanisms of temperature inversions in the southeastern Arabian Sea. *Geophys. Res. Lett.*, 2006, **33**, doi:10.1029/2006GL027280.
40. Durand, F., Shetye, S. R., Vialard, J., Shankar, D., Shenoi, S. S. C., Ethe, C. and Madec, G., Impact of temperature inversions on SST evolution in the South-Eastern Arabian Sea during pre-summer monsoon season. *Geophys. Res. Lett.*, 2004, **31**, doi:10.1029/2003GL018906.
41. Bruce, J. G., Johnson, D. R. and Kindle, J. C., Evidence for eddy formation in the eastern Arabian Sea during the northeast monsoon. *J. Geophys. Res.*, 1994, **99**, 7651–7664.
42. Shankar, D. and Shetye, S. R., On the dynamics of the Lakshadweep high and low in the southeastern Arabian Sea. *J. Geophys. Res.*, 1997, **102**, 12551–12562.
43. Shankar, D., Vinayachandran, P. N. and Unnikrishnan, A. S., The monsoon currents in the north Indian Ocean. *Prog. Oceanogr.*, 2002, **52**, 63–120.
44. Shenoi, S. S. C., Saji, P. K. and Almeida, A. M., Near-surface circulation and kinetic energy in the tropical Indian Ocean derived from Lagrangian drifters. *J. Mar. Res.*, 1999, **57**, 885–907.
45. Durand, F., Shankar, D., de Boyer Mont´egut, C., Shenoi, S. S. C., Blanke, B. and Madec, G., Modelling the barrier-layer formation in the South-Eastern Arabian Sea. *J. Climate*, 2007, In Press.
46. Kurian, J. and Vinayachandran, P. N., Mechanisms of formation of the Arabian Sea mini warm pool in a high-resolution Ocean General Circulation Model. *J. Geophys. Res.*, 2007, **112**, doi: 10.1029/2006JC003631.
47. Pankajakshan, T. and Rama Raju, D. V., Intrusion of Bay of Bengal water into Arabian Sea along the west coast of India during north east monsoon. In *Contributions in Marine Sciences* (eds Rao, T. S. S., Natarajan, R., Desai, B. N., Swami, G. N. and Bhat, S. R.), Dr S. Z. Qasim Sastyabdapurti felicitation volume, National Institute of Oceanography, Goa, India, 1987, pp. 237–244.
48. Shenoi, S. S. C., Shankar, D. and Shetye, S. R., Remote forcing annihilates barrier layer in southeastern Arabian Sea. *Geophys. Res. Lett.*, 2004, **31**, doi:10.1029/2003GL019270.
49. Vialard, J. and Delecluse, P., An OGCM study for the TOGA decade. Part I: Role of salinity in the physics of western Pacific fresh pool. *J. Phys. Oceanogr.*, 1998, **28**, 1071–1088.
50. McCreary, J. P., Kundu, P. K. and Molinari, R. L., A numerical investigation of dynamics, thermodynamics and mixed-layer processes in the Indian Ocean. *Prog. Oceanogr.*, 1993, **31**, 181–244.
51. Madec, G., Delecluse, P., Imbard, M. and Levy, C., Ocean General Circulation Model Reference Manual, Technical Report XX, Institut Pierresimon Laplace (IPSL). Internal Report, France, 1999.
52. Howden, S. D. and Murtugudde, R., Effects of river inputs into the Bay of Bengal. *J. Geophys. Res.*, 2001, **106**, 19825–19843.
53. Jensen, T. G., Arabian Sea and Bay of Bengal exchange of salt and tracers in an ocean model. *Geophys. Res. Lett.*, 2001, **28**, 3967–3970.
54. Jensen, T. G., Cross-equatorial pathways of salt and tracers from the northern Indian Ocean: Modelling results. *Deep-Sea Res. Part I*, 2003, **50**, 2111–2127.
55. Masson, S., Delecluse, P., Boulanger, J. -P. and Menkes, C., A model study of the seasonal variability and formation mechanisms of the barrier layer in the eastern equatorial Indian Ocean. *J. Geophys. Res.*, 2002, **107**, doi:10.1029/2001JC000832.
56. Esenkov, O. E., Olson, D. B. and Bleck, R., A study of the circulation and salinity budget of the Arabian Sea with an isopycnic coordinate ocean model. *Deep-Sea Res. II*, 2003, **50**, 2091–2110.
57. Han, W. and McCreary, J. P., Modelling salinity distributions in the Indian Ocean. *J. Geophys. Res.*, 2001, **106**, 859–877.
58. Shankar, D., McCreary, J. P., Han, W. and Shetye, S. R., Dynamics of the East India coastal current. 1. Analytic solutions forced by interior Ekman pumping and local alongshore winds. *J. Geophys. Res.*, 2004, **101**, 13975–13991.
59. McCreary, J. P., Han, W., Shankar, D. and Shetye, S. R., Dynamics of the East India coastal current. 2. Numerical solutions. *J. Geophys. Res.*, 1996, **101**, 12993–14010.
60. Griffies, S. M., Harrison, M. J., Pacanowski, R. C. and Rosati, A., A Technical Guide to MOM4. *GFDL Ocean Group Technical Report No. 5*, 2004, Princeton, NJ, available online at www.gfdl.noaa.gov.
61. de Boyer Mont´egut, C., Vialard, J., Shenoi, S. S. C., Shankar, D., Durand, F., Eth´e, C. and Madec, G., Simulated seasonal and interannual variability of mixed layer heat budget in the northern Indian Ocean. *J. Climate*, 2007, In Press.
62. L´evy, M., Klein, P. and Treguier, A.-M., Impact of sub-mesoscale physics on production and subduction of phytoplankton in an oligotrophic regime. *J. Mar. Res.*, 2001, **59**, 535–565.
63. Frater, A., *Chasing the Monsoon*, Penguin Books, India, 1991.
64. Ananthakrishnan, R., Pathan, J. M. and Aralikatti, S. S., The onset phase of the southwest monsoon. *Curr. Sci.*, 1983, **52**, 755–764.
65. Ananthakrishnan, R., Srinivasan, V., Ramakrishnan, A. R. and Jambunathan, R., Synoptic features associated with onset of southwest monsoon over Kerala, Forcasting Manual. FMU Technical Rep. IV-18.2, India Meteorological Department, Poona, India, 1968.
66. Rao, Y. P., Southwest monsoon. Meteorological Monograph, India Meteorological Department, New Delhi, India, 1976.
67. Ananthakrishnan, R., Tracks of storms and depressions in the Bay of Bengal and Arabian Sea. Technical report, India Meteorological Department, New Delhi, India, 1964.
68. De, U. S. and Joshi, K. S., Genesis of cyclonic disturbances over the north Indian Ocean, 1981–1990. Pre-published Scientific

Report 1995/3, India Meteorological Department, Poona, India, 1995.
69. Krishnamurti, T. N., Ardanuy, P., Ramanathan, Y. and Pasch, R., On the onset vortex of the summer monsoons. *Mon. Weather Rev.*, 1981, **109**, 344–363.
70. Mak, M. and Kao, J. C. Y., An instability study of the onset vortex of southwest monsoon, 1979. *Tellus*, 1982, **34**, 358–368.
71. Krishnamurti, T. N., Summer monsoon experiment – a review. *Mon. Weather Rev.*, 1985, **113**, 1590–1360.
72. Krishnamurti, T. N., Inglas, K., Cocke, S., Pasch, R. and Kitade, T., Details of low latitude medium range numerical weather prediction using a global spectral model. Part II: Effect of orography and physical initialization. *J. Meteorol. Soc. Jpn.*, 1984, **62**, 613–649.
73. Kung, E. C. and Sharif, T. A., Long-range forcasing of the Indian summer monsoon onset and rainfall with upper air parameters and sea surface temperature. *J. Meteorol. Soc. Jpn.*, 1982, **60**, 672–681.
74. Kershaw, R., Onset of the southwest monsoon and sea surface temperature anomalies in the Arabian Sea. *Nature*, 1985, **315**, 561–563.
75. Kershaw, R., The effect of sea surface temperature anomalies in a prediction of the onset of southwest monsoon over India. *Q. J. R. Meteorol. Soc.*, 1988, **144**, 325–345.
76. Rao, K. G. and Goswami, B. N., Interannual variations of sea surface temperature over the Arabian Sea and the Indian monsoon: A new perspective. *Mon. Weather Rev.*, 1988, **116**, 558–568.
77. Shukla, J., Inter-annual variability of monsoons In *Monsoons* (eds Fein, J. S. and Stephens, P. L.), John Wiley and Sons, New York, 2001, pp. 399–463.
78. Joseph, P. V. and Pillai, P. V., 40-day mode of equatorial trough for long range forcasting of Indian summer monsoon onset. *Curr. Sci.*, 1988, **57**, 951–954.
79. Masson, S. *et al.*, Impact of barrier layer on winter spring variability of the southeastern Arabian Sea. *Geophys. Res. Lett.*, 2005, **32**, doi:10.1029/2004GL021980.
80. Reynolds, R. W. and Smith, T. M., Improved global sea surface temperature analysis using optimum interpolation. *J. Climate*, 1994, **7**, 929–948.
81. Kalnay, E. *et al.*, The NCEP-NCAR 40-year reanalysis project. *Bull. Am. Meteorol. Soc.*, 1996, **77**, 437–471.
82. Deepa, R., Seetaramayya, P., Nagar, S. G. and Gnanaseelan, C., On the plausible reasons for the formation of onset vortex in the presence of the Arabian Sea mini warm pool. *Curr. Sci.*, 2007, **92**, 794–800.
83. Joseph, P. V. and Sijikumar, S., Intraseasonal variability of the low-level jet stream of the Asian summer monsoon. *J. Climate*, 2004, **17**, 1449–1458.
84. Locarnini, R. A., Mishonov, A. V., Antonov, J. I., Boyer, T. P. and Garcia, H. E., *World Ocean Atlas 2005, Volume 1: Temperature* (ed. Levitus, S.), NOAA Atlas NESDIS 61 Technical report, NODC, U.S. Government Printing Office, Washington, D.C., 2006, p. 182, Available online at http://www.nodc. noaa. gov/OC5/WOA05/pr woa05.html.
85. Antonov, J. I., Locarnini, R. A., Boyer, T. P., Mishonov, A. V. and Garcia, H. E., *World Ocean Atlas 2005, Volume 2: Salinity* (ed. Levitus, S.), NOAA Atlas NESDIS 62 Technical report, NODC, U.S. Government Printing Office, Washington, D.C., 2006, 182 pp, Available online at http://www.nodc.noaa. gov/OC5/WOA05/pr woa05.html.

ACKNOWLEDGEMENTS. Financial support from the ARMEX programme of DST and the INDOMOD programme of INCOIS, MoESS, Govt. of India is greatly acknowledged. F.D. acknowledges the support from IRD and IFPREWAC (Indo-French Programme of Research on Weather and Climate). The altimeter products were produced by SSALTO/DUACS and distributed by AVISO with support from CNES. This is NIO contribution 4259.