Internal Variability of Indian Ocean SST

MARKUS JOCHUM
National Center for Atmospheric Research, Boulder, Colorado

RAGHU MURTUGUDDE
Earth System Science Center, College Park, Maryland

(Manuscript received 6 October 2003, in final form 15 March 2005)

ABSTRACT

A 40-yr integration of an eddy-resolving numerical model of the tropical Indian Ocean is analyzed to quantify the interannual variability that is caused by the internal variability of ocean dynamics. It is found that along the equator in the western Indian Ocean internal variability contributes significantly to the observed interannual variability. This suggests that in this location the predictability of SST is limited to the persistence time of SST anomalies, which is approximately 100 days. Furthermore, a comparison with other sources of variability suggests that internal variability may play an important role in modifying the Indian monsoon or preconditioning the Indian Ocean dipole/zonal mode.

1. Introduction

The countries that rely on the rain provided by the seasonal (Indian) monsoon are home to almost 2 billion people. Their high population density implies that there is only little tolerance for variation in the annual rainfall rate—too little leads to widespread droughts, too much causes flooding. Thus, predicting the strength and pattern of the monsoon is of utmost importance for these societies. Recent studies show that the impact of the monsoon is not restricted to the Indian Ocean (IO), it can also affect El Niño (Wainer and Webster 1996; Kirtman et al. 2004; Annamalai et al. 2005), and the SST in the IO as a whole seems to affect the northern midlatitudes in both the Atlantic and Pacific Oceans (Deser et al. 2004; Hurrel et al. 2004). Understanding and predicting the monsoon is a formidable task and involves issues as diverse as African orography, Himalayan snowfall, ENSO, or ocean dynamics. An excellent review of these topics is provided by Webster et al. (1998); here we focus on ocean dynamics, and specifically their impact on IO SST.

The IO is different from the Atlantic and Pacific in that it is bounded in the north, and that as a result of the Indonesian low the prevailing equatorial winds are westerlies, not easterlies. The main consequence of the westerly winds is the absence of an equatorial upwelling zone, and the main consequence of the surrounding landmasses is a seasonal cycle of winds and currents that is much stronger than in the Atlantic or Pacific. Thus, the annual mean winds and currents are almost negligible north of 10°S, but at their peaks these winds and currents are as strong as any in the tropical oceans (Schott and McCreary 2001 provide an overview over the IO circulation and winds).

The unique boundary conditions make the IO interesting to study in its own right; for climate prediction, however, the attention necessarily focuses on the SST. Much of the success of numerical weather prediction in midlatitudes is a result of the fact that there the ocean responds passively to changes in the atmosphere on daily to seasonal time scales. The predictions are only limited by the nonlinearity of the atmospheric weather systems. Because of the high wave speed it is often argued that in the Tropics the dynamics of both the atmosphere and ocean can be linearized and, therefore, pose no limits on predictability. There, however, nonlinearity can be introduced by positive feedbacks between atmosphere and ocean (Bjerknes 1969), leading to El Niño in the Pacific (McCreary and Anderson 1984) and the zonal modes in the Atlantic (Zebiak 1993) and Indian Oceans (Murtugudde et al. 1998; Murtugudde and Busalacchi 1999; Saji et al. 1999; Webster et al. 1999). Thus, predictability and societal well being in the Tropics relies critically on understanding SST. If SST is entirely determined by surface forcing and

Corresponding author address: Markus Jochum, NCAR, Oceanography Section, Rm. 415, 1850 Table Mesa Dr., Boulder, CO 80307.
E-mail: markus@ucar.edu

boundary conditions, one only has to worry about initial conditions for the prediction model; if, however, there are stochastic or nonlinear processes within the ocean that affect the SST, there will be a limit to predictability.

For the midlatitudes, several authors demonstrated that the observed internal variability of western boundary currents could be understood within the framework of the dynamical systems theory (Simmonet et al. 2003a,b, and references therein). These studies typically use high-resolution shallow-water or quasigeostrophic ocean models set in a rectangular basin with a midlatitude double gyre. For certain ranges of Rossby and Ekman numbers the solutions exhibit chaotic behavior, or limit cycle behavior, which is usually tied to the strength of the inertial recirculation gyres near the western boundary. The not very comforting picture that emerges from these studies is that the nature of the ocean circulation could be sensitive to parameters that are not well known (e.g., friction or boundary conditions). The present authors are not aware of a study that shows the impact of this internal midlatitude variability on large-scale climate, but it can be speculated that it affects the water mass properties and the heat budget of the midlatitude oceans.

For the Tropics it has been suggested that "the tropical ocean response on interannual timescales is reasonably well captured by linear or weakly nonlinear approximations to the ocean dynamics" (Neelin et al. 1998). Main evidence for this is the success of relatively simple coupled ocean–atmosphere models that display El Niño-like behavior in spite of their linear ocean dynamics. However, the present authors recently concluded several studies that show that internal variability in the Tropics is a major source for interannual variability of the cross-hemispheric SST gradient in the Atlantic (Jochum et al. 2004b) and of the zonal SST gradient in the equatorial Pacific (Jochum and Murtugudde 2004). To the extent that tropical climate variability is a self-sustained variability that is made irregular by low-order chaos (Tziperman et al. 1994) or a disturbance of a basically stable state by stochastic noise (Penland and Sardeshmukh 1995; Kessler 2002), this internal variability has a profound impact on the predictability of tropical climate. The present study is a continuation of our work in the Atlantic and Pacific and quantifies internal variability in the tropical IO. It also complements recent studies by Waliser et al. (2003, 2004) who evaluate the impact of intraseasonal atmospheric variability on IO SST.

Internal oceanic variability is caused by instabilities of the wind-driven currents. Two conditions must be met so that the resulting mesoscale eddy field can cause interannual SST variability and introduce atmospheric variability. Firstly, eddies must persist long enough so that they can change the oceanic conditions for the next season where the currents become unstable. The nonlinearities of the instability mechanism can then amplify the small changes in the new initial conditions. Second, because oceanic eddy scales are much smaller than the atmospheric Rossby radius, eddies must modify the ocean–atmosphere heat exchange, thereby affecting the total heat in the mixed layer. Otherwise, they only move heat around horizontally, and there will be no net effect on the atmosphere. The tropical instability waves (TIWs) in the Pacific and Atlantic fulfill both conditions (Jochum et al. 2004a, 2005, hereafter JOC05). Here we will analyze whether the IO eddies meet these conditions. To isolate the effect of eddies on the interannual SST anomalies we will necessarily have to use a numerical model.

The next section describes the model and observations of the IO eddy field, sections 3 and 4 quantify interannual variability in the IO, and the last section provides a discussion on how our results relate to the monsoon and its predictability.

2. Model description

The ocean model that is employed for this study is the reduced gravity, primitive equation, sigma-coordinate model of Gent and Cane (1989). It is coupled to an advective atmospheric mixed layer model (AMLM), which computes surface heat fluxes without any restoring boundary conditions or feedbacks to observations (Seager et al. 1995; Murtugudde et al. 1996). A variable-depth oceanic mixed layer represents the three main processes of oceanic turbulent mixing, namely, the entrainment/detrainment resulting from wind and buoyancy forcing, the gradient Richardson number mixing generated by the shear flow instability, and the convective mixing related to static instabilities in the water column (Chen et al. 1994). The model is initialized with Levitus (1994) temperature and salinity fields, driven by seasonal National Centers for Environmental Prediction (NCEP) winds; thus, the forcing is identical every year. Solar radiative forcing is taken from the Earth Radiation Budget Experiment (Li and Leighton 1993), cloud data are taken from Rossow and Schiffer (1991), and precipitation is based on Xie and Arkin (1998). Evaporation is determined by the AMLM. The ocean general circulation model (OGCM) has a ¼-degree horizontal resolution and 12 layers in the vertical. Along the boundaries at 25°S and 130°E, temperature and salinity are restored to Levitus (1994). The strength of the Indonesian Throughflow (ITF) is determined by these boundary conditions and is ap-

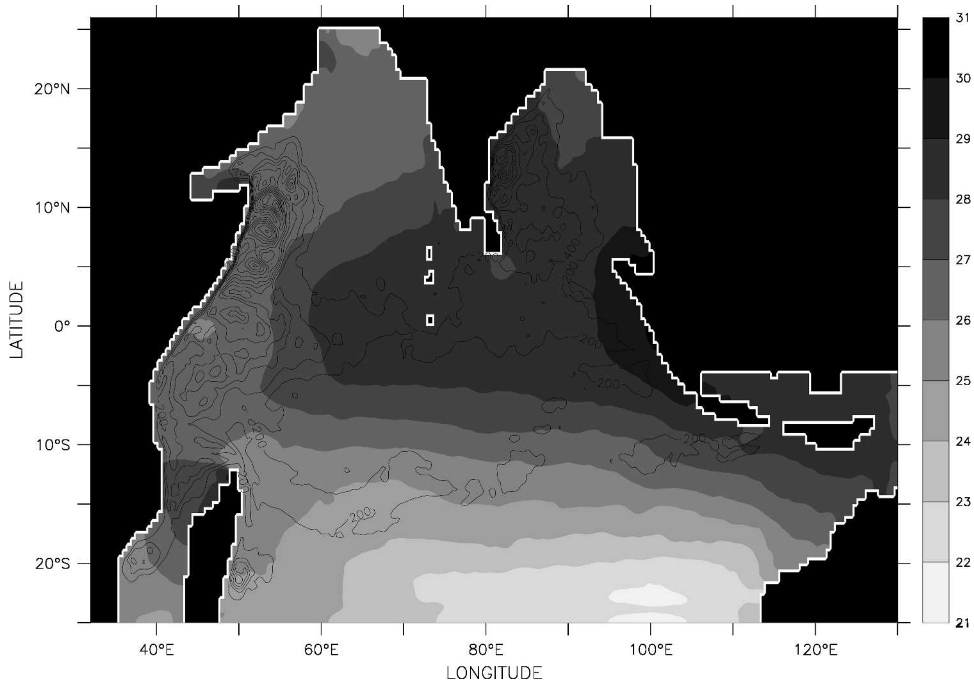

FIG. 1. Eddy kinetic energy (contour line: 200 cm^{-2} s^{-2}, maximum: 2800 cm^{-2} s^{-2}) superimposed on the annual mean SST.

proximately 6 Sv (1 Sv $\equiv 10^6$ m^3 s^{-1}), which is well within the range of observed values (e.g., Godfrey 1996). The model is spun up for 20 yr; to analyze and illustrate the mesoscale eddy field, we then saved every sixth-day snapshots of all the model variables for 5 yr, and to quantify the interannual SST variability we saved monthly mean temperatures for another 40 yr. From these 40 yr we constructed the model climatology; the internal SST anomalies that are discussed here are the deviations there from. Similarily, the eddy kinetic energy is computed as the kinetic energy of the flow after the seasonal cycle has been removed.

It is important to note that with the atmospheric boundary layer model as the upper boundary condition, the model computes its own heat flux and can, therefore, develop its own SST. Neither is the SST artificially damped back to climatology, nor will a positive ocean–atmosphere feedback amplify small perturbations.

Unlike the Atlantic or Pacific Oceans, the IO warm pool is on the eastern side of the domain (Fig. 1), which is a direct result of the westerly winds created by the convection over Indonesia (Gill 1980). The distribution of the eddy kinetic energy (EKE) is no surprise either, and its pattern and strength in the model is largely consistent with observations of the sea surface height variability on subseasonal time scales (see Fig. 10 of Kessler 2004) or in situ observations (cf. Fig. 2, with the observations by Schott et al. 1990). The 15-day waves that have been described by Schott et al. (1994) are not reproduced, because they are directly forced by high-frequency oscillations in the wind field (Sengupta et al. 2001), which, by construction, are not part of the NCEP wind climatology used here.

3. Internal variability of SST

It is obvious that there is internal variability in a turbulent system. The question to be answered here is how large it is in the IO, and if it could affect climate; thus, SST is the variable of choice used to describe the internal variability. The observed SST is the result not only of internal and atmospherically forced variability, but it also reflects positive feedbacks between the ocean and atmosphere. In principle, even an infinitesimally small disturbance could then trigger the rapid growth of anomalies in the coupled ocean–atmosphere system. In that case, every mesoscale eddy would be important. However, the seasonal cycle in the IO is fairly stable and is dominated by the monsoon, which suggests a stable system. The recently much-discussed Indian Ocean dipole/zonal mode (IODZM), whose dynamics shows some similarities to ENSO, seems to be strongly damped and climatically significant events have only been reported 3 times since 1960 (see Annamalai and Murtugudde 2004 for a recent review). This means that the interannual SST variability (Fig. 3) in

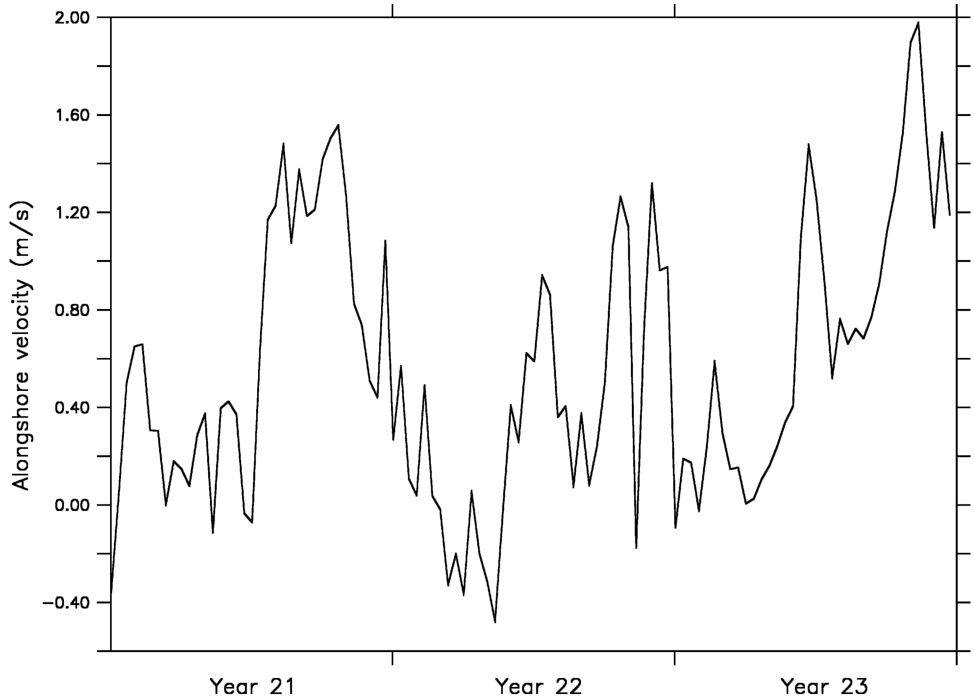

FIG. 2. Alongshore velocity at 0°, 44°E at 100-m depth. Like in the observations the flow shows strong intraseasonal variability and the maximum velocity can reach 2 m s^{-1}. Note that because of nonlinearity of the flow the velocities are different from year to year, although the wind forcing is the same. This makes the interpretation of observations difficult.

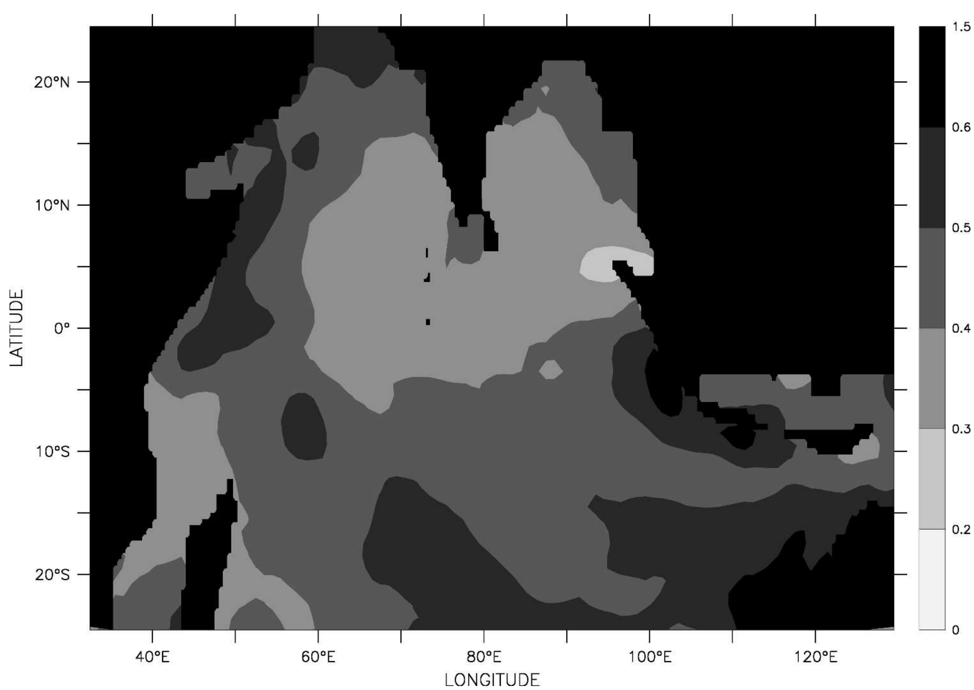

FIG. 3. Standard deviation of the interannual monthly mean SST anomalies (after Reynolds and Smith 1994). Note that the landmass is taken from the model to facilitate comparisons.

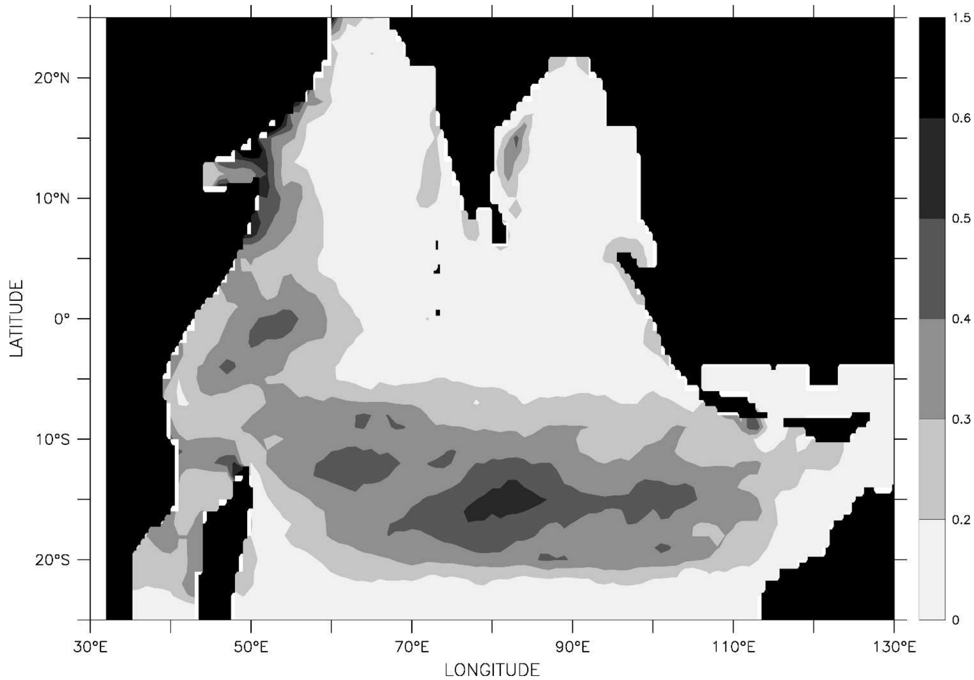

FIG. 4. Standard deviation of the interannual monthly mean SST anomalies (model result). Note that the model is driven by climatological forcing, therefore, the result here is the result of internal variability only.

the IO has to be mostly the result of external atmospheric forcing and internal oceanic variability. In fact, Tourre and White (1995) showed that between one-third and one-half of the observed interannual variability is a result of ENSO. While the debate about the role of ENSO in forcing the IO variability will continue, the focus here is on the magnitude of the internal variability in the IO at interannual time scales.

Waliser et al. (2003, 2004) employ the same OGCM that is used in the present study (albeit with coarser resolution), and find that in the central equatorial IO and the Bay of Bengal the rectification of the Madden–Julian oscillation onto the SST produces interannual SST anomalies of about 0.3 K, which is comparable to the observed interannual variability. This leaves much of the observed interannual variability in the western equatorial IO and south of the equator still unexplained. Interestingly, these are the areas where we find significant internal SST variability (Fig. 4).

The largest internal variability is found in areas of high EKE (cf. Figs. 1 and 4), which have all been discussed already in the literature. Because the present focus will be on SST, only a brief overview of their dynamics will be provided here.

Nof et al. (2002) demonstrate that the Indonesian Throughflow, as it enters the IO through the Lombok Strait and Timor Passage, has to break up into eddies. These eddies are then amplified by baroclinic instability in the South Equatorial Current (Feng and Wijffels 2002). They travel west between 10° and 20°S, and, on encountering the northern tip of Madagascar, they trigger more eddies that travel south in the Mozambique Channel (Schouten et al. 2002).

The mesoscale energy in the Bay of Bengal has been explained by Vinayachandran and Yamagata (1998) as the result of barotropic instability of the Wyrtki Jet and the Southwest Monsoon Current. These instabilities generate vorticies that travel northwestward, and then stall and decay at the eastern coast of India.

The large mesoscale energy along the African coast and the tongue that extends along the equator toward the Maldives is analyzed in detail by Kindle and Thompson (1989). The barotropic instabilities of the seasonally reversing Somali current and the East African coastal current create large gyres that can be seen along the Somali coast, and Yanai waves that carry energy into the eastern IO (Tsai et al. 1992).

The comparison of Figs. 3 and 4 show that in a large part of the IO mesoscale variability explains a substantial part of the observed interannual variability. It is important to notice that this is not just the effect of eddies with long periods, that is, averaging the SST over a season instead of over a month still yields approximately 70% of the SST anomaly shown in Fig. 4 (not

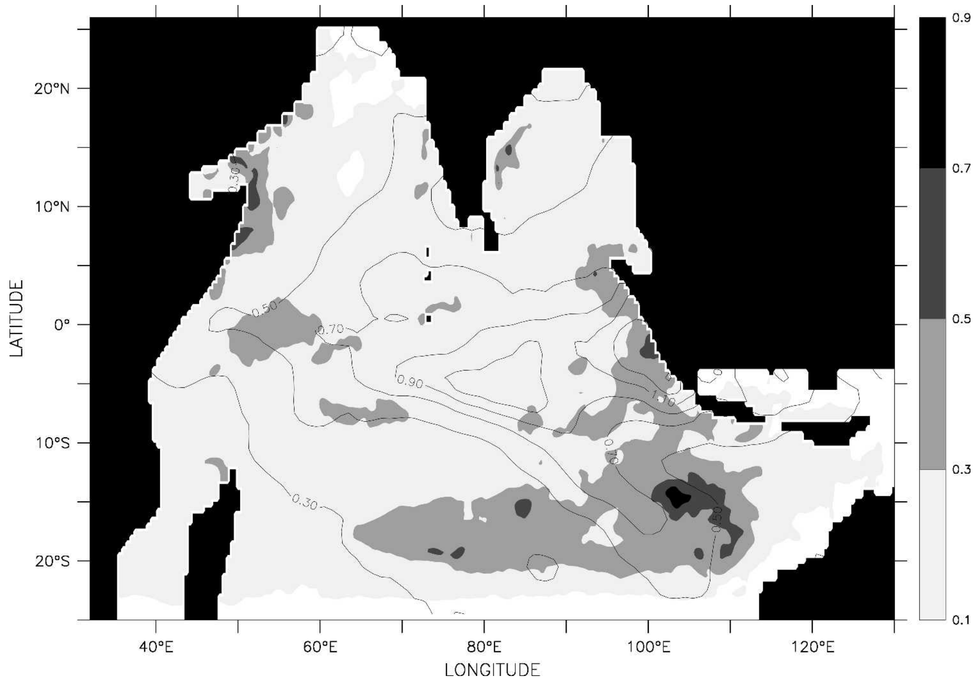

FIG. 5. Standard deviation of the interannual monthly mean SST anomalies divided by standard deviation of the annual cycle. The superimposed contour lines show the same quantity computed from observations.

shown). Thus, not only do the mesoscale eddies stir heat around, they can make a net contribution to the mixed layer heat budget. This is similar to the tropical Atlantic and Pacific, where JOC05 showed that TIWs (the dominant form of mesoscale eddies in the tropical Atlantic and Pacific) make not only a local contribution to the mixed layer heat budget by moving heat from the area north of the SST front to the equatorial cold tongue, but they also make a global contribution because they increase the heat flux from the atmosphere into the equatorial thermocline by increasing the vertical entrainment of heat. Before the details of the eddy heat flux will be discussed in the next section, it is important to clarify where eddies could matter for climate variability.

In the present study the atmosphere is passive and forces the ocean. This enables us to unambigously identify the effect of eddies on the SST, but leaves us to speculate on the effect of eddies on climate variability. It safe, however, to assume that the impact of the SST anomalies depends not only on the magnitude of the anomaly but also on the spatial extent of the anomalous patch. Also, because the IO climate is dominated by the seasonal monsoon, any interannual anomaly of relevance should be at least of the same order of magnitude as the magnitude of the annual cycle. In large areas of the IO, the internal variability is a significant fraction of the annual cycle (Fig. 5); if nothing else, this means that in these areas it takes long observational records to establish confidently the seasonal cycle of SST. Figure 5 also shows the same ratio computed from observations. In the areas of large EKE the values are fairly similar. The biggest difference is in the tongue that stretches from Sumatra toward the western IO, the pattern of which clearly reflects the signal of ENSO and the IODZM (see Annamalai and Murtugudde 2004).

The autocorrelation lengths of the anomalies in the model are approximately 200 km along the coast of Somalia and the two coasts of India, and approximately 400 km along 15°S and the coasts of Sumatra and Java, which is much smaller than the atmospheric Rossby radius of 1000 km. This suggests that there the internal variability of SST that is introduced by the mesoscale eddy field is only of local importance and is unlikely to play an important role in modifying the monsoon or affecting the IODZM. If IO climate can be modified or controlled by mesoscale eddies at all, the western equator around 55°E is the most promising area. First, the autocorrelation length is approximately 1000 km. Second, it is a center of action for the IODZM; SST anomalies there project directly on the IODZM and have the potential to impact the local Walker cell and, thus, the regional coupled variability (Hastenrath et al. 1993). Therefore, the next section will focus on the

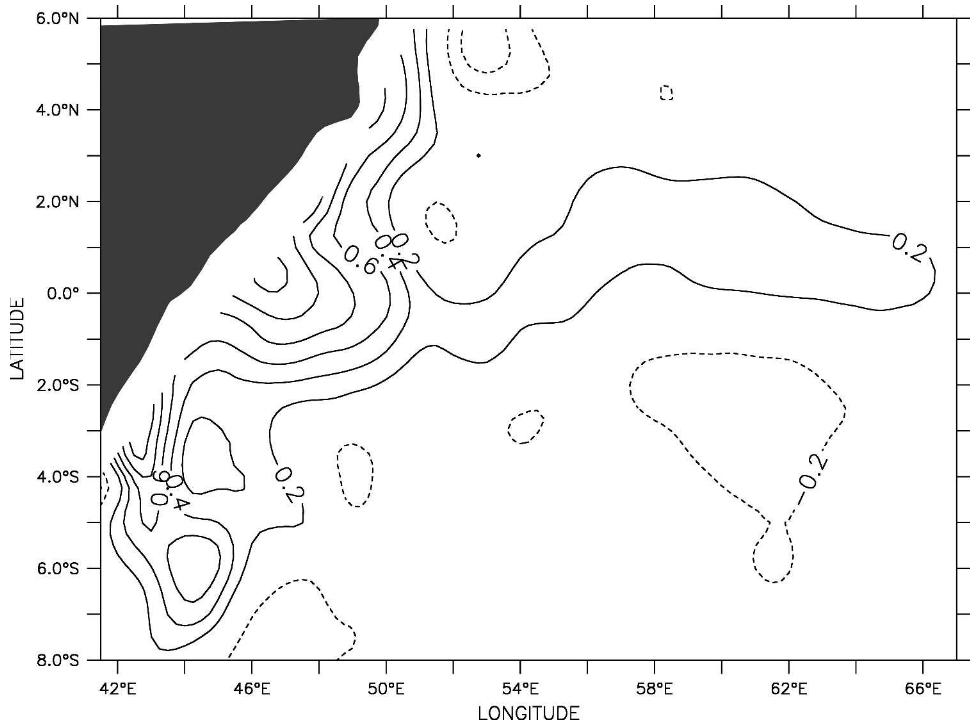

FIG. 6. Mean eddy heat flux convergence in the ML of the WIO (contour lines: 0.2°C month^{-1}).

mesoscale eddy field in the western Indian Ocean (WIO).

4. The western Indian Ocean

The interannual variability of the dynamic fields in the WIO has already been noted in the modeling study of Kindle and Thompson (1989). Wirth et al. (2002) quantified this effect to explain the observed interannual variability in the position of the Great Whirl. In both studies, the interannual variabilities are the result of the nonlinearities of the flow field, and are not due variability in the forcing fields. In the previous section, the effect of the nonlinearities on the internal variability of the SST has been quantified. This section provides a more detailed look into the eddy heat flux and the local heat budget.

If the mesoscale variability varies from year to year, this does not necessarily lead to a different SST. The current mixing length paradigm would suggest that eddies just move heat around, either diabatically near the surface via wave breaking (Kessler et al. 1998) or adiabatically below the mixed layer via thickness fluxes (Gent and McWilliams 1990). Thus, averaging over an eddy time and length scale would leave no net effect on the SST. In the WIO, however, the autocorrelation length of the SST anomalies is approximately 1000 km, and the autocorrelation time is approximately 3 months. This suggests that eddies make a net contribution to the mixed layer (ML) heat budget. A warmer SST is then the result of stronger eddy activity. This process has already been analyzed for the interaction between TIWs and the equatorial ML in the Atlantic and Pacific Oceans (JOC05)—TIWs move surface water back and forth over patches of strong entrainment cooling. Outside of these patches a water parcel is heated by the atmosphere; when the heated water parcel moves across the entrainment area the heat is mixed into the thermocline. Thus, in contrast to the current mixing length paradigm, in the ML eddies can generate a horizontal heat flux with closed particle orbits. For the IO the area of significant net eddy heat flux convergence is shown in Fig. 6. It is restricted to the WIO and is connected to the coastal upwelling off of the Somali coast and the Ekman divergence along the equator during the strong monsoon winds (McCreary et al. 1993; Wacogne and Pacanowski 1996). The analysis of the seasonal cycle further supports the connection between eddy heat flux, ML depth, and EKE (Fig. 7). Whereas EKE has two distinct peaks (July and November), only the second peak is matched by a peak in heat flux convergence. The first peak is accompanied by a deep ML; therefore, the wind stirring fails to further increase the entrainment and cannot mix down the heat that is provided by eddies. Thus, in July, eddies mostly

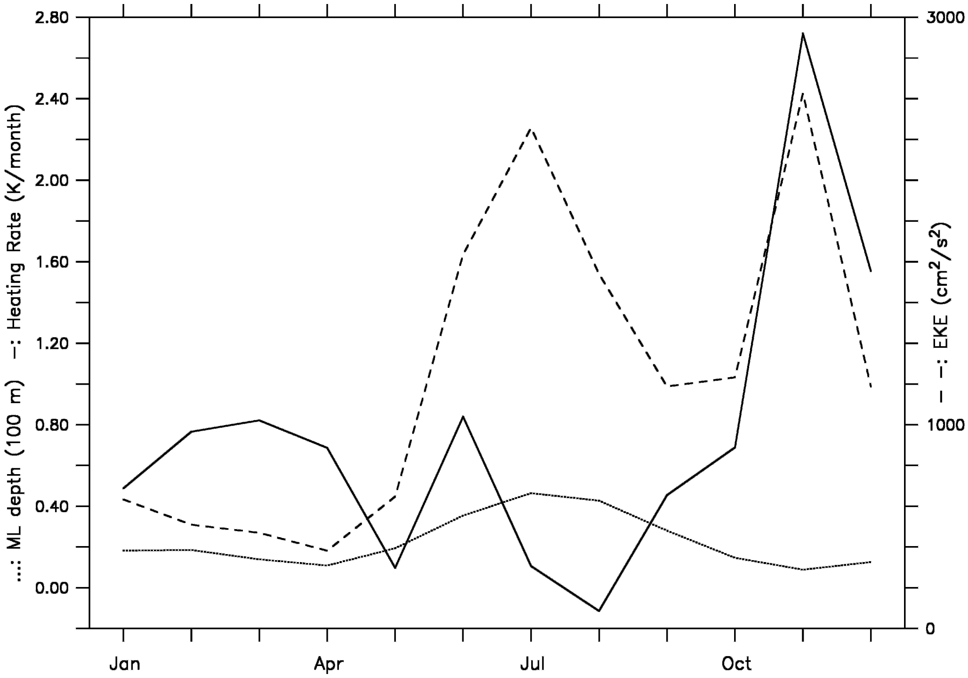

FIG. 7. Seasonal cycle of (left) eddy heat flux convergence (solid line) and ML depth (100 m)$^{-1}$ (dotted line) and (right) the EKE (broken line) in the ML of the WIO (average between 2°S and 2°N and between 46° and 50°W).

just stir heat around adiabatically, but in November they actually change the ML heat budget. Another way to look at this asymmetry is that, in case of a shallow ML, the atmosphere does not provide all of the heat that could possibly be mixed down. Eddies increase the heat supply. In the case of a deep ML, wind stirring does not provide enough energy to mix down the heat that is provided in excess by eddies.

The results so far suggest that there is an analogy between the TIWs in the Atlantic and Pacific and the mesoscale variability in the WIO. In all three domains, eddies generate a significant internal variability of SST and make an important net contribution to the ML heat budget as a result of the interaction between horizontal eddy advection and localized entrainment cooling. In the Pacific and Atlantic, the entrainment is generated by wind stirring of a shallow ML and equatorial Ekman divergence. This is also the case for the WIO, but, in addition, there is also coastal upwelling along the Somali coast. A snapshot from October illustrates how the surface water cools off as it flows toward the coast along 2°S (Fig. 8). In an adiabatic environment the flow would be along isotherms, but here the water flows over an area of shallow ML depth (Fig. 9) where it is exposed to increased entrainment cooling.

The ML heat budget of the IO is largely a local balance between net atmospheric heat flux and vertical entrainment (Rao and Sivkumar 2000; Shenoi et al. 2002), and the WIO is no exception. However, the combination of a thin ML and high background SST can make eddies in the WIO an important source of noise for the atmospheric circulation. Mesoscale activity leads to irregular bursts of warming, which produce the interannual SST anomalies discussed here (Fig. 10). The entrainment cooling is largely determined by the seasonal wind field and, therefore, is more regular, although the mesoscale eddies can occasionally affect the entrainment rate (Fig. 10; early summer of year 22). In this particular location, the net contribution of eddies to the heat budget is small in the years 23–25. In year 22, eddies also increased the entrainment rate so that their heat is mixed into the thermocline; the net oceanic heat flux convergence in the ML does not differ much from the following 3 yr, although it is phase shifted. In year 21 eddies significantly change the ML heat budget, which gives rise to the interannual SST variability discussed here.

The pronounced minimum of the net atmospheric heat flux in the summer is a result of two components: the minimum of solar insolation and the maximum of the latent heat loss during the onset of the summer monsoon. Increased horizontal eddy heat flux is associated with reduced latent heat loss and increased entrainment (years 21 and 22). Thus, eddies mix heat into

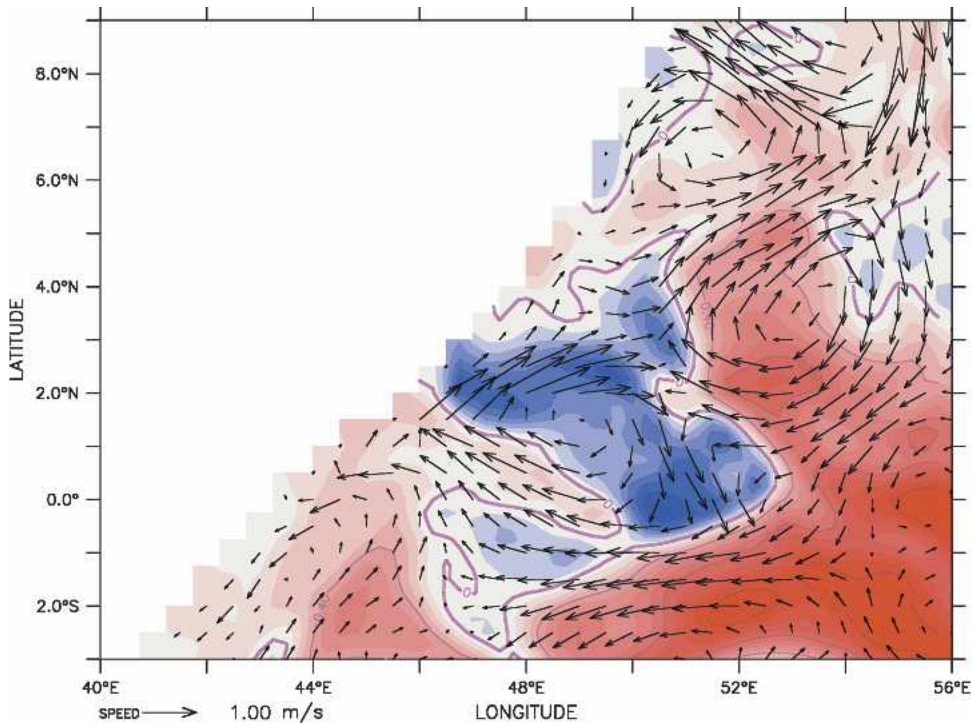

FIG. 8. Snapshot of surface velocity anomalies superimposed on SST anomalies during one particular Oct (contour interval: 0.2°C). Note how at 2°S the water crosses isotherms, giving rise to a positive eddy heat flux convergence.

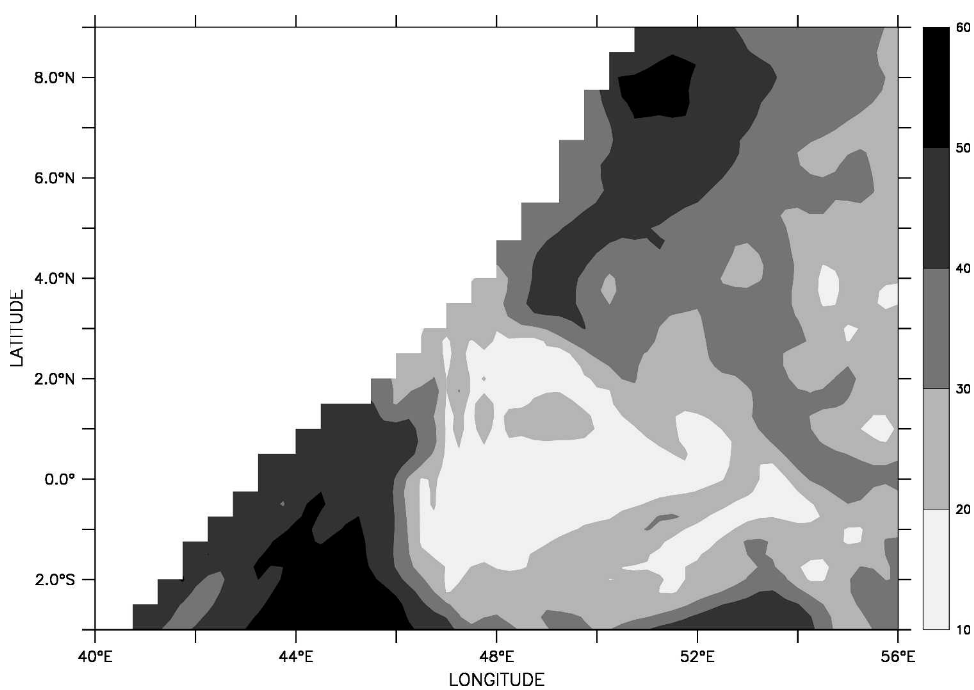

FIG. 9. The ML depth for the same time as in Fig. 8. Comparison with Fig. 8 shows how the strongest cooling of the water occurs in the shallowest part of the ML.

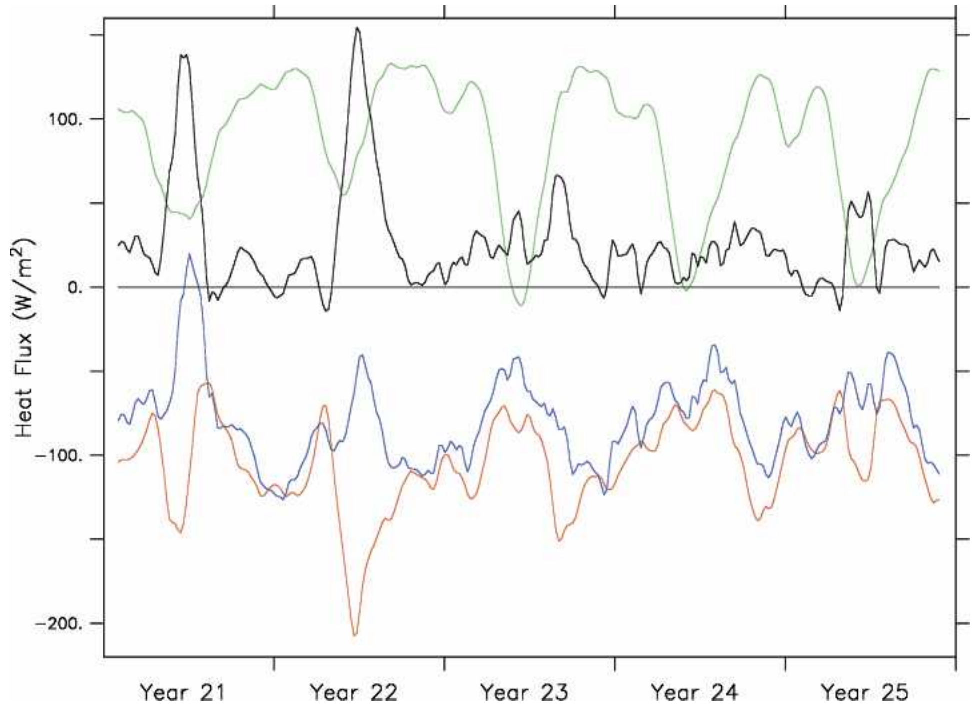

FIG. 10. Five-year time series of oceanic heat flux convergence, averaged from 2°S to 2°N and from 46° to 50°E and smoothed with a 60-day running average: horizontal heat flux convergence (black line), vertical entrainment (red line), total oceanic convergence (blue line), and net atmospheric heat flux (green line). Note that, in contrast, Fig. 7 discusses temperature advection and MLD depth separately, whereas heat flux convergence combines both variables.

the thermocline and prevent it from being lost to the atmosphere. This suggests that during times of strong eddy activity the Findlater jet picks up less moisture, which can have implications for the strength of the monsoon rain over India. The current model configuration, however, is unable to provide insights into the responsible processes. Its atmospheric mixed layer model computes the heat flux based on, among other things, horizontal advection and the diffusion of heat. This depends on the direction of the wind in relation to the SST distribution. The WIO eddies in this model change the SST but not the wind, whereas, in reality, we would expect the wind to change according to the SST distribution. Thus, the model response in latent heat loss resulting from SST anomalies does not necessarily represent all of the processes. The atmospheric mixed layer model was chosen for the current study because it allows for an unambiguous quantification of the internal SST variability; a full AGCM is needed to investigate the effect of this internal variability on the atmosphere.

5. Summary and discussion

A high-resolution primitive equation ocean model with climatological forcing has been used to study the internal variability of the IO. In most of this domain the internal variability appears to be negligible, compared to the seasonal or observed interannual variability. The exception is the WIO, where internal variability explains a significant part of the observed interannual SST variability; it is comparable to interannual SST variability that is introduced by ENSO or the Madden–Julian oscillation. Large internal variability in the WIO has a direct implication for observations: The effect of mesoscale variability in the observational records cannot be removed by simply averaging over the eddy time scale. The eddies make a net contribution to the mixed layer heat budget; because the generating instability processes are nonlinear this contribution will vary and the SST will be different from year to year, even under climatological forcing. Moreover, this uncertainty will affect seasonal climate forecasts because it could reduce the forecast time to the presistence time of SST anomalies, which is, at least for the equatorial Pacific, approximately 100 days (Kessler et al. 1996).

Large internal SST variability is restricted to the WIO, but it projects directly onto the zonal SST gradient and can change its seasonal cycle. This can have implications for the onset and development of the IODZM. It can also affect the amount of moisture that

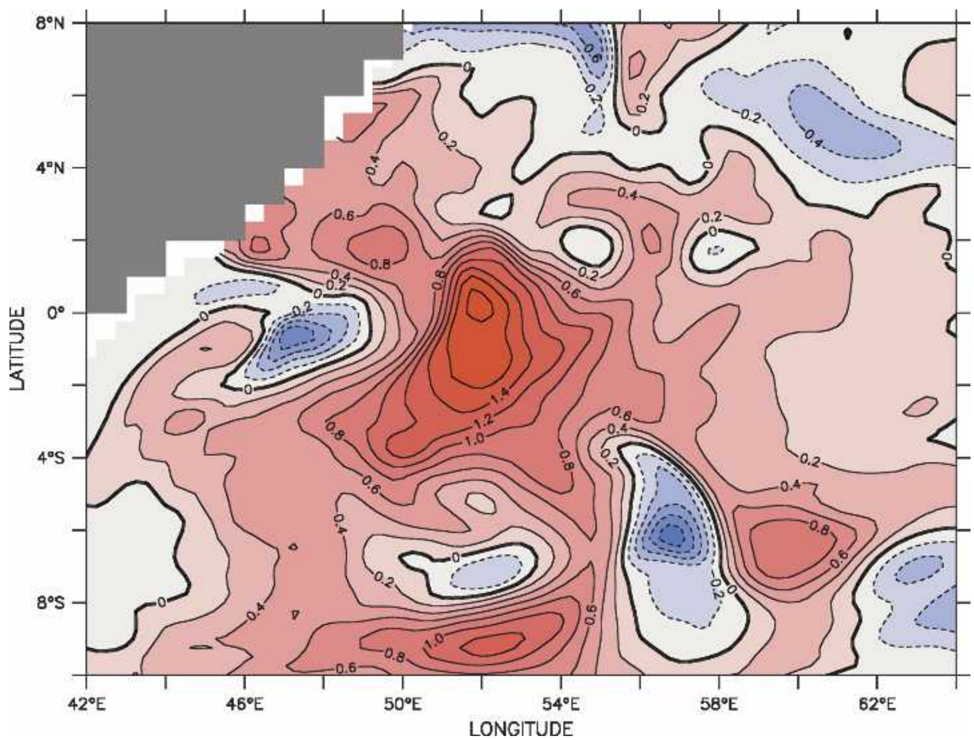

FIG. 11. An eddy-forced extreme event of SST anomalies during one particular Oct (contour intervals: 0.2°C).

is picked up by the Findlater jet and delivered to the Indian subcontinent during the southwest monsoon. Rodwell and Hoskins (1995) demonstrate that the vertical structure of atmospheric heating over the WIO and the Horn of Africa determines the latitude at which the Findlater jet turns east in the Northern Hemisphere; therefore, WIO mesoscale variability may be an important component in the irregularity of Indian monsoon rainfall. It should be pointed out that for societies that are affected by the monsoon rainfall, extreme events are more relevant than the statistical measures that are described in the previous sections. Extreme events from the present OGCM study suggest that large-scale SST anomalies with peaks of up to 2°C are possible (Fig. 11). Of course, the potential effect of internal variability on climate variability as lined out above is speculative and has to be supported by coupled model studies, which will be the focus of the authors' research in the future.

As far as the interaction of eddies with the ML is concerned, the presented results suggest that there is an analogy between the equatorial Pacific and Atlantic and the WIO. In all three domains eddies are created by instabilities and act on a background state of thin MLs, which are caused by the equatorial Ekman divergence. The difference between the WIO and the other two equatorial domains is that in the WIO the upwelling is at the western side of the domain and also has a contribution resulting from coastal upwelling. This complicates not only the analysis of model results, but also the theoretical developments and observational verification. Whereas the TIWs can be analyzed in a two-dimensional framework, the WIO eddies have to be understood in the presence of coastal boundaries and a seasonally reversing western boundary current. For example, much of our knowledge about TIWs comes from surface drifter data, which yield reliable meridional eddy heat fluxes when averaged zonally (e.g., Hansen and Paul 1984; Baturin and Niiler 1997). This, of course, is not practicable in the WIO, and one needs a dense arrays of moorings, similar to the ones used to study the TIWs (e.g., Qiao and Weisberg 1998). In the absence of such observations, the present study relies heavily on trust in the realism of the model, in particular, the vertical mixing and the ML model. However, in the past, the present configuration of the model has been used successfully to study upper ocean and mesoscale physics (see JOC05 and references therein).

Acknowledgments. MJ is grateful for the financial support of Paola Malanotte-Rizzoli, who is funded with NOAA Grant Nr NA16GP1576 and NASA Grant

Nr NAG5-7194 at MIT and ESSIC/UMD. RM gratefully acknowledges the NOAA Monsoon-ENSO grant.

REFERENCES

Annamalai, H., and R. Murtugudde, 2004: The role of the Indian Ocean in regional climate variability. *Ocean-Atmosphere Interaction and Climate Variability, Geophys. Monogr.,* No. 147, Amer. Geophys. Union, 210–250.

——, S. P. Xie, J. P. McCreary, and R. Murtugudde, 2005: Impact of Indian Ocean sea surface temperature on developing El Niño. *J. Climate,* **18,** 302–319.

Baturin, N., and P. Niiler, 1997: Effects of instability waves in the mixed layer of the equatorial Pacific. *J. Geophys. Res.,* **102,** 27 771–27 793.

Bjerknes, J., 1969: Atmospheric teleconnections from the equatorial Pacific. *Mon. Wea. Rev.,* **97,** 163–172.

Chen, D., L. Rothstein, and A. Busalacchi, 1994: A hybrid vertical mixing scheme and its applications to tropical ocean models. *J. Phys. Oceanogr.,* **24,** 2156–2179.

Deser, C., A. S. Phillips, and J. W. Hurrel, 2004: Pacific interdecadal climate variability: Linkages between the Tropics and North Pacific during boreal winter since 1900. *J. Climate,* **17,** 3109–3124.

Feng, M., and S. Wijffels, 2002: Intraseasonal variability in the SEC of the East Indian Ocean. *J. Phys. Oceanogr.,* **32,** 265–277.

Gent, P., and M. Cane, 1989: A reduced gravity, primitive equation model of the upper equatorial ocean. *J. Comput. Phys.,* **81,** 444–480.

——, and J. McWilliams, 1990: Isopycnal mixing in ocean circulation models. *J. Phys. Oceanogr.,* **20,** 150–155.

Gill, A., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.,* **106,** 447–462.

Godfrey, S., 1996: The effect of Indian Ocean throughflow on ocean circulation and heat exchange with the atmosphere. *J. Geophys. Res.,* **101,** 12 217–12 237.

Hansen, D., and C. Paul, 1984: Genesis and effects of long waves in the equatorial Pacific. *J. Geophys. Res.,* **89,** 10 431–10 440.

Hastenrath, S., A. Nicklis, and L. Greishar, 1993: Atmospheric-hydrospheric mechanisms of climate anomalies in the western equatorial Indian Ocean. *J. Geophys. Res.,* **98,** 20 219–20 235.

Hurrel, J. W., M. P. Hoerling, A. Phillips, and T. Xu, 2004: Twentieth century North Atlantic climate change. Part I: Assessing determinism. *Climate Dyn.,* **23,** 371–389.

Jochum, M., and R. Murtugudde, 2004: Internal variability in the tropical Pacific ocean. *Geophys. Res. Lett.,* **31**, L14309, doi:10.1029/2004GL020488.

——, P. Malanotte-Rizzoli, and A. Busalacchi, 2004a: Tropical instability waves in the Atlantic Ocean. *Ocean Modell.,* **7,** 145–163.

——, ——, P. Malanotte-Rizzoli, and A. Busalacchi, 2004b: Internal variability in the Atlantic Ocean. *Ocean-Atmosphere Interaction and Climate Variability, Geophys. Monogr.,* No. 147, Amer. Geophys. Union, 181–187.

——, ——, R. Ferrari, and P. Malanotte-Rizoli, 2005: The impact of horizontal resolution on the equatorial mixed layer heat budget in ocean general circulation models. *J. Climate,* **18,** 841–851.

Kessler, W., 2002: Is ENSO a cycle or a series of events? *Geophys. Res. Lett.,* **29,** 2125, doi:10.1029/2002GL015924.

——, 2004: Intraseasonal variability in the ocean. *Tropical Intraseasonal Variability*, W. K. M. Lau and D. E. Waliser, Eds., Praxis Publishing, in press.

——, M. Spillane, M. McPhaden, and D. Harrison, 1996: Scales of variability in the equatorial Pacific inferred from the TAO buoy array. *J. Climate,* **9,** 2999–3024.

——, L. M. Rothstein, and D. Chen, 1998: The annual cycle of SST in the eastern tropical Pacific as diagnosed in an OGCM. *J. Climate,* **11,** 777–799.

Kindle, J. C., and J. D. Thompson, 1989: The 26 day and 50 day oscillations in the Indian Ocean. Model results. *J. Geophys. Res.,* **94,** 4721–4736.

Kirtman, B. P., K. Pegion, and S. Kinter, 2005: Internal atmospheric dynamics and tropical Indo-Pacific climate variability. *J. Atmos. Sci.,* **62,** 2220–2233.

Levitus, S., 1994: *Climatological Atlas of the World Ocean*. NOAA Prof. Paper 13, 173 pp.

Li, Z., and H. Leighton, 1993: Global climatologies of the solar radiation budgets form 5 years of ERBE data. *J. Geophys. Res.,* **98,** 4919–4930.

McCreary, J. P., and D. Anderson, 1984: A simple model of El Niño. *Mon. Wea. Rev.,* **112,** 934–946.

——, P. K. Kundu, and R. L. Molinari, 1993: A numerical investigation of dynamics, thermodynamics and mixed-layer processes in the Indian Ocean. *Progress in Oceanography*, Vol. 31, Pergamon, 181–244.

Murtugudde, R., and A. J. Busalacchi, 1999: Interannual variability of the dynamics and thermodynamics of the tropical Indian ocean. *J. Climate,* **12,** 2300–2326.

——, R. Seager, and A. Busalacchi, 1996: Simulation of the tropical oceans with and ocean GCM coupled to an atmospheric mixed layer model. *J. Climate,* **9,** 1796–1815.

——, B. N. Goswami, and A. J. Busalacchi, 1998: Air-sea interaction in the southern Indian Ocean. *Proc. Int. Monsoon Conf. and Hydrologic Cycle*, Kyong Ju, Korea, CLIVAR-Monsoon, 184–188.

Neelin, J., D. Battisti, A. Hirst, F. Jin, Y. Wakata, T. Yamagata, and S. Zebiak, 1998: ENSO theory. *J. Geophys. Res.,* **103,** 14 261–14 290.

Nof, D., T. Pichevin, and J. Sprintall, 2002: "Teddies" and the origin of the Leeuwin Current. *J. Phys. Oceanogr.,* **32,** 2571–2588.

Penland, C., and D. Sardeshmukh, 1995: The optimal growth of tropical sea surface temperature anomalies. *J. Climate,* **8,** 1999–2024.

Qiao, L., and R. Weisberg, 1998: Tropical instability wave energetics: Observations from the tropical instability wave experiment. *J. Phys. Oceanogr.,* **28,** 345–360.

Rao, R. R., and R. Sivakumar, 2000: Seasonal variability of near-surface thermal structure and heat budget of the mixed layer of the Indian Ocean from a new global ocean climatology. *J. Geophys. Res.,* **105,** 995–1015.

Reynolds, R., and T. Smith, 1994: Improved global SST analyses using optimal interpolation. *J. Climate,* **7,** 929–948.

Rodwell, M. J., and B. J. Hoskins, 1995: A model of the Asian summer monsoon. Part II: Cross-equatorial flow and PV behavior. *J. Atmos. Sci.,* **52,** 1341–1356.

Rossow, W. B., and R. A. Schiffer, 1991: ISCCP cloud data products. *Bull. Amer. Meteor. Soc.,* **72,** 2–20.

Saji, N. H., B. N. Goswami, P. N. Vinayachandran, and T. Yamagata, 1999: A dipole mode in the tropical Indian ocean. *Nature,* **401,** 360–363.

Schott, F. A., and J. P. McCreary, 2001: The monsoon circulation of the Indian Ocean. *Progress in Oceanography*, Vol. 51, Pergamon, 1–123.

Schott, F., J. C. Swallow, and M. Fieux, 1990: The Somali Current at the equator: Annual cycle and transports. *Deep-Sea Res.,* **37A,** 1825–1848.

——, J. Reppin, J. Fischer, and D. Quadfasel, 1994: Currents and transports of the monsoon current south of Sri Lanka. *J. Geophys. Res.,* **99,** 25 127–25 141.

Schouten, M. W., W. P. M. de Ruijter, and P. J. van Leeuwen, 2002: Upstream control of Agulhas ring shedding. *J. Geophys. Res.,* **107,** 3109, doi:10.1029/2001JC000804.

Seager, R., M. B. Blumenthal, and Y. Kushnir, 1995: An advective atmospheric mixed layer model for ocean modeling purposes: Global simulation of surface heat fluxes. *J. Climate,* **8,** 1951–1964.

Sengupta, D., R. Senan, and B. N. Goswami, 2001: Origin of intraseasonal variability of circulation in the tropical central Indian Ocean. *Geophys. Res. Lett.,* **28,** 1267–1270.

Shenoi, S. S. C., D. Shankar, and S. R. Sheteye, 2002: Differences in the heat budgets of the near-surface Arabian Sea and Bay of Bengal: Implications for the summer monsoon. *J. Geophys. Res.,* **107,** 3052, doi:10.1029/2001JC000679.

Simmonet, E., M. Ghil, K. Ide, R. Temam, and S. Wang, 2003a: Low-frequency variability in shallow-water models of the wind-driven ocean circulation. Part I: Steady-state solution. *J. Phys. Oceanogr.,* **33,** 712–728.

——, ——, ——, ——, and ——, 2003b: Low-frequency variability in shallow-water models of the wind-driven ocean circulation. Part II: Time-dependent solutions. *J. Phys. Oceanogr.,* **33,** 729–752.

Tourre, Y. M., and W. B. White, 1995: ENSO signals in global upper-ocean temperature. *J. Phys. Oceanogr.,* **25,** 1317–1332.

Tsai, P. T., J. J. O'Brian, and M. E. Luther, 1992: The 26-day oscillations observed in satellite SST measurements in the equatorial western Indian Ocean. *J. Geophys. Res.,* **97,** 9605–9618.

Tziperman, E., L. Stone, M. Cane, and H. Jarosh, 1994: El Niño chaos: Overlapping of resonances between the seasonal cycle and Pacific ocean-atmosphere oscillator. *Science,* **264,** 72–74.

Vinayachandra, P. N., and T. Yamagata, 1998: Monsoon response of the sea around Sri Lanka: Generation of thermal domes and anticyclones. *J. Phys. Oceanogr.,* **28,** 1946–1960.

Wacogne, S., and R. Pacanowski, 1996: Seasonal heat transport in a primitive equations model of the Indian Ocean. *J. Phys. Oceanogr.,* **26,** 2666–2699.

Wainer, I., and P. J. Webster, 1996: Monsoon-ENSO interaction using a simple coupled ocean-atmosphere model. *J. Geophys. Res.,* **101,** 25 599–25 614.

Waliser, D., R. Murtugudde, and L. Lucas, 2003: Indo-Pacific ocean response to atmospheric intraseasonal variability: 1. Austral summer and the Madden-Julian Oscillation. *J. Geophys. Res.,* **108,** 3160, doi:10.1029/2002JC001620.

——, ——, and ——, 2004: Indo-Pacific ocean response to atmospheric intraseasonal variability: 2. Boreal summer and the intraseasonal oscillation. *J. Geophys. Res.,* **109,** C03030, doi:10.1029/2003JC002002.

Webster, P. J., V. Magana, T. Palmer, J. Shukla, R. Tomas, M. Yanai, and T. Yasunari, 1998: Monsoon: Processes, predictability and the prospects for prediction. *J. Geophys. Res.,* **103,** 14 451–14 510.

——, A. M. Moore, J. P. Loschnigg, and R. R. Leben, 1999: Coupled ocean-atmosphere dynamics in the Indian Ocean during 1997–98. *Nature,* **401,** 356–360.

Wirth, A., J. Willebrand, and F. Schott, 2002: Variability of the Great Whirl from observations and models. *Deep-Sea Res.,* **49B,** 1279–1295.

Xie, P. P., and P. A. Arkin, 1998: Global monthly precipitation estimates from satellite-observed outgoing longwave radiation. *J. Climate,* **11,** 137–164.

Zebiak, S., 1993: Air-sea interaction in the tropical Atlantic. *J. Climate,* **6,** 1567–1586.