Interannual Variability of the Dynamics and Thermodynamics of the Tropical Indian Ocean

RAGU MURTUGUDDE
ESSIC, University of Maryland at College Park, College Park, Maryland, and Laboratory for Hydrospheric Processes, NASA/Goddard Space Flight Center, Greenbelt, Maryland

ANTONIO J. BUSALACCHI
Laboratory for Hydrospheric Processes, NASA/Goddard Space Flight Center, Greenbelt, Maryland

(Manuscript received 5 February 1998, in final form 5 October 1998)

ABSTRACT

Interannual variability of the tropical Indian Ocean is studied with a reduced gravity, primitive equation, ocean general circulation model (OGCM). The OGCM is coupled to an atmospheric mixed layer model for surface heat flux computation. The seasonal simulation of sea surface temperatures (SST), current, and thermocline structures are in good agreement with observations and other models. The seasonal cycle of SST along the equator exhibits an eastward propagation with larger variability in the west. The interannual simulations are carried out over 1980–95 with interannual wind stresses and wind speeds but climatological data for solar radiation and cloudiness. The SST anomalies are smaller than 1°C over most of the basin and the leading EOF shows an ENSO-related warming. However, the correlation between the Southern Oscillation index and the time series of the leading EOF is only $-0.51$ and SST anomalies of similar magnitudes as an El Niño year appear in other years too. ENSO-related equatorial winds determine the SST anomalies along the coast of Sumatra and this anomaly in the eastern southern tropical Indian Ocean (STIO) is typically opposite in sign to the anomaly in the western STIO. The western STIO has some of the largest SSTA because of a shallow thermocline and the entrainment effects associated with wind stress curl anomalies in the region. The quasi-biennial oscillation in the thermocline and the SST gradient in the STIO is correlated with the Somali jet, which in turn is correlated with the Indian summer monsoon. An experiment with climatological wind stresses but interannual wind speeds demonstrates that the wind-driven variations in SST are larger than previously estimated with relaxation type heat fluxes. A parallel experiment with climatological wind speeds but interannual wind stresses shows that heat fluxes contribute significantly to SST variability. Another simulation with interannual data for radiation and cloudiness shows that model simulation is affected significantly in some regions by the use of climatological data for solar radiation and cloudiness. A model experiment with an open eastern boundary provides a simplistic illustration of the effects of the Indonesian Throughflow (ITF). The main influence of the ITF is to warm the Indian Ocean and reduce the effect of upwelling on SST.

1. Introduction

The Tropical Ocean Global Atmosphere decade led to numerous studies of the tropical Pacific region (McPhaden et al. 1997). Only a handful of them included the tropical Indian Ocean (e.g., Meehl 1987; Naggai et al. 1995), although a substantial number of studies focused on understanding the exchange of mass and heat between the Pacific and the Indian Oceans (Godfrey 1996 and references therein). A large part of the literature pertaining to the Indian Ocean has focused on the unique aspect of this particular ocean, namely, the seasonal reversal of the Somali Current (Dueting 1978; Cox 1976; Luther and O’Brien 1989; Jensen 1991). Other seasonal features of the dynamics have been studied by various observational and wind-driven model studies (Rao et al. 1989; Molinari et al. 1990; Luther and O’Brien 1985; Woodberry et al. 1989; Potemra et al. 1991). Among the few models of the seasonal thermodynamics in the Indian Ocean are that of McCreary and Kundu (1989), McCreary et al. (1993), and Murtugudde et al. (1996). Seasonal meridional circulation and heat transport issues were addressed by Wacogne and Pacanowski (1996) and Lee and Marotzke (1997). While some observational and model studies addressed the interannual sea level variability (Perigaud and Delecluse 1993; Clarke and Liu 1994), and wind-driven circulation (Anderson and Carrington 1993; Valenti et al. 1997), to our knowledge, the basin-scale air–sea in-
teraction and the thermodynamics have not been reported thus far. Cadet and Diehl (1984) and Cadet (1985) analyzed available SST and meteorological data for interannual variability and the covariability of these with the Southern Oscillation. Perigaud and Delecluse (1993) concentrated on the Rossby waves in the southern tropical Indian Ocean (STIO). Anderson and Carrington (1993) considered the effects of different wind products on model currents for 1987–90. Valenti et al. (1997, manuscript submitted to J. Climate) ran a high-resolution reduced gravity model for 1979–92 to identify the leading EOFs and propagating features in model currents and layer thicknesses.

Except for the upwelling region off Somalia, SSTs in the tropical Indian Ocean tend to be rather high and there appears to be no coupled unstable mode such as the El Niño–Southern Oscillation (ENSO) in the Indian Ocean. Atmospheric model studies indicate that when the absolute temperatures are high, even small variations in temperatures may produce significant climatic impacts (Palmer and Mansfield 1984). It is thus important to understand the mechanisms responsible for the interannual variability of the the Indian Ocean SSTs. The relationship between the SST variability in the Arabian Sea and the summer monsoon over India has been explored by Shukla and Misra (1977), Weare (1979), and Dube et al. (1990). Dube et al. (1990) used a wind-driven ocean model (without any thermodynamics) simulation for their study. Goswami (1997) carried out coupled model simulations with observed SST and climatological SST and compared the monsoon variability from the two simulations. He concluded that the observed SST anomalies only enhanced the variability but were not necessary to explain most of the Indian monsoon variability. Thus far, no definitive relationship has been established between the air–sea interaction in the tropical Indian Ocean and the summer monsoon over India. Shukla (1987) showed that a warm SST over the Arabian Sea does not necessarily imply an above-average summer monsoon rainfall, but a heavy monsoon rainfall is followed by negative SST anomalies. It is well known that the monsoon over India has a decadal and longer timescale variability (Parthasarathy and Mooley 1978), thus its correlation with the Indian Ocean SSTs may also show such regime shifts. This is confirmed by Goswami et al. (1997), who defined an extended all-India monsoon rainfall index and correlated it with Indo-Pacific SSTs. The correlation patterns show remarkable reversals over the Indian Ocean from 1979–88 to 1989–94 period (see their Fig. 16).

Since we employ a wind-driven ocean general circulation model (OGCM), albeit with a complete upper ocean hydrology and an interactive surface heat flux formulation, we can only conjecture on the possible role of the Indian Ocean SSTs on the Indian monsoon rainfall. Our main interest is to report on the various aspects of the interannual variability of the Indian Ocean dynamics and thermodynamics, and point to possible causal mechanisms through which the SSTs may feed back on to the atmospheric circulation. Our OGCM is coupled to an advective atmospheric mixed layer (AML) in order to allow free evolution of SST with no feedback to observations. In a previous paper (Murtugudde et al. 1996) we showed that significant improvement in the simulation of SST can be attained off the coasts of Africa and Arabia and also northwest Australia when advection and diffusion of humidity is accounted for in surface heat flux parameterizations. We also demonstrated the effects of the Indonesian Throughflow (ITF) on the seasonal-to-interannual dynamics and thermodynamics of the Indian Ocean (Murtugudde et al. 1998a). We resort to simulations of the Indian Ocean only in this study in order to concentrate on its interannual variability. Excluding the ITF obviously affects our results, but through comparisons to available observations and with a crude representation of the ITF in our model, it is shown that the interannual anomalies of the flow fields of interest are not severely degraded. A model description is provided in section 2, seasonal simulations in section 3 and interannual variability in section 4. A discussion and summary of this study are presented in section 5.

2. Model description

The ocean GCM is a reduced gravity, primitive equation, sigma coordinate model (Gent and Cane 1989) with an embedded hybrid mixing scheme (Chen et al. 1994). Surface heat fluxes are computed by coupling the ocean GCM to an advective AML model (Seager et al. 1995). The vertical structure of the model ocean consists of a mixed layer and a specified number of layers below according to a sigma coordinate. The mixed-layer depth and the thickness of the last sigma layer are computed prognostically and the remaining layers are computed diagnostically such that the ratio of each sigma layer to the total depth below the mixed layer is held to its prescribed value. The model domain covers 30°S–26°N, 32°–142°E. The horizontal grid has a longitudinal resolution of ½° and a latitudinal resolution of ½°. In the vertical, there are a total of 15 layers consisting of the variable depth mixed layer and 14 sigma layers with \( \sigma = 0.0286, 0.0286, 0.0286, 0.0286, 0.0429, 0.0429, 0.0429, 0.0429, 0.0714, 0.0714, 0.0714, 0.0714, 0.1429, \) and 0.286. Note that a higher vertical resolution is specified a priori below the mixed layer. The motionless deep layer is at 9°C and 35 practical salinity units (psu). The United Nations Educational, Scientific and Cultural Organization equation of state is used for computing buoyancy from salinity and temperature.

The hybrid vertical mixing scheme (Chen et al. 1994) combines the advantages of the traditional bulk mixed layer model of the Kraus–Turner (1967) type with the dynamic instability model of Price et al. (1986). Freshwater forcing is treated as a natural boundary condition (Huang 1993). The AML model has been described by
Seager et al. (1995) and here we will only provide a summary of its characteristics. The model represents either a dry convective layer or the mixed layer that underlies shallow marine clouds. Within the mixed layer, the air temperature and humidity are determined by a balance between surface fluxes, horizontal advection by imposed winds, entrainment from above the mixed layer, horizontal diffusion, and, for temperature, radiative cooling. Knowing the air temperature and humidity, surface sensible and latent heat fluxes can be calculated in terms of the ocean model SST and the imposed winds. To compute the longwave radiative heat loss from the surface we use a standard bulk formula and observed cloud cover. Solar radiative forcing is taken from the Earth Radiation Budget Experiment (ERBE) satellite data of Li and Leighton (1993). In summary, the complete heat flux can be computed with only the solar radiation, cloud cover, and winds needing to be specified. These are all quantities over which the ocean has only indirect control and can be justifiably specified externally. The quantities over which the ocean does have direct control, air temperature and humidity, are modeled internally in terms of the SST, the winds, and their values at the continental margins.

As stated earlier, the ITF has been neglected in the control run and the eastern boundary between Australia and Java is treated as a solid wall with no flux and no-slip boundary conditions. The southern boundary has a sponge layer between 25° and 30°S where model temperature, salinity, and layer thicknesses are relaxed to Levitus (1994) data. The evaporation (E) is computed by the AML, the climatological precipitation (P) was obtained from Oberhuber (1988) atlas, while the interannual P was generously provided by Xie and Arkin (1995). Interannual simulations are performed over 1980–95 with climatological ERBE solar radiation and International Satellite Cloud Climatology Project (ISCCP) cloudiness data. The sensitivity of model flow fields to cloudiness and radiation is tested by comparing the control run with a simulation using interannual data for cloudiness and solar radiation for the period of their availability, namely, 1984–90. The effect of the ITF is explored by a simplistic representation of the interocean mass and heat exchange, namely, with a simulation including a sponge layer between Indonesia and Australia.

3. Seasonal simulation

The model is spun up from rest with the initial temperature, salinity, and layer thicknesses derived from Levitus (1994) data. The monthly mean winds from Hellerman and Rosenstein (1983) are used to force the model with the surface heat fluxes provided by the AML. The model reaches equilibrium in 20 years of spinup. Monthly mean fields from year 21 are analyzed for model performance.

The seasonal cycle of SST along the equator (averaged over 2°S–2°N) and the standard deviation for the model and Reynolds and Smith (1994) SST (referred to hereafter as Reynolds SST) are shown in Fig. 1. The eastward-propagating feature with maximum SST during boreal spring is similar in both model and Reynolds SST. The seasonal cycle of SST in the Pacific propagates westward (Chen et al. 1994) with a large seasonal variation in the east. Figure 6 shows that the eastern equatorial IO has a barrier layer (an isothermal layer below the mixed layer and above the thermocline), as opposed to the western equatorial Pacific Ocean, which serves to decouple SST variability from thermodynamic variability (Lukas and Lindstrom 1991). Interannual variability of the barrier layer plays a significant role in the SST variability in the equatorial Indian Ocean as discussed later.

The standard deviation of model SST is larger than Reynolds SST in regions with recirculation such as the Somali Current and the northern tip of Madagascar. This is partly due to Reynolds SST being on a coarser grid and smoother. The wind-driven variability in these regions is larger because the effect of upwelling and entrainment on SST is exaggerated due to the missing ITF (Murtugudde et al. 1998a). This is discussed further in the following section. The model SST variability is lower in the northern Bay of Bengal due to the neglect of river discharges.

The annual mean differences between model and Reynolds SST and the annual mean differences in depth of 20°C isotherms (D20) from the model and Levitus (1994) data are shown in Fig. 2. The annual mean SST errors are typically of order 1°C or less. The seasonal errors are larger (not shown), especially in the Somali Current region where Reynolds SST smooths out the features associated with the recirculation. The differences in model and Levitus (1994) SSS, Sea Surface Salinity (not shown) are less than 1 psu over most of the domain, except in the Bay of Bengal where the model errors are as large as 3 psu. Note that the variable depth mixed layer along with the interactive surface fluxes, that is, no relaxation condition on temperature, can produce SSTs that are reasonable. However, the mixed layer depths (MLDs) and the surface fluxes associated with these SSTs must be erroneous in the Bay of Bengal since salinity errors are so large. Due to the lack of observations, it is difficult to quantify the errors in either the MLDs or the surface heat fluxes.

The model D20 is typically deeper than Levitus (1994) except in the northern Arabian Sea and at the exit of the ITF. Large differences are seen off the coast of Madagascar where the resolution of Levitus (1994) data is suspect. The model thermocline is thus slightly weaker over most of the basin. The annual mean model MLDs and their seasonal variations (not shown) are in reasonable agreement with the estimates based on observations by Rao et al. (1989) in other regions. The SSTs and the MLDs show high variability in the Somali Current region extending into the Arabian Sea, around 6°N near the southern tip of Sri Lanka, and between 10°
and $20^\circ$S from $60^\circ$E to $80^\circ$E. The latter region in the STIO also displays the largest seasonal variability in the depth of the thermocline and sea level due to Ekman pumping. Meridional sections of temperature and zonal currents for February and August are shown in Figs. 3 and 4. Reverdin and Fieux (1987) note a pinch in the thermocline north of $8^\circ$S and the model reproduces the corresponding ridge in August. The deepening of the thermocline centered at $8^\circ$N is associated with the anticyclonic Great Whirl. The depth of the $20^\circ$C isotherm ($D_{20}$ hereafter) shows the largest variability north of $5^\circ$N, in qualitative agreement with Fig. 4 of Reverdin and Fieux (1987). The complex seasonal variability of the winds not only affects the surface currents, but also the vertical structure of the currents. An Equatorial Undercurrent with westward surface currents and eastward subsurface currents exists from January through March in the western Indian Ocean (Swallow 1967; Knox 1976). Cane (1980) demonstrated that the seasonal reversal of the undercurrent at Gan is driven by local winds and not by eastern boundary reflections as suggested by Knox (1976). The surface flow reverses by April and the subsurface flow has a semiannual variation (Luyten and Roemmich 1982). Model results reproduce the reversal of the surface and subsurface flows as shown in Fig. 4 with core velocities that are comparable to observations.

Surface currents are shown in Fig. 5 for February and August. The seasonal reversal of the Somali Current from February to August is clearly seen in the model results with peak speeds of over $100 \text{ cm s}^{-1}$ in August. Note that the flow direction in the Arabian Sea and to the south of Sri Lanka have fully reversed in August compared to the circulation in February. As in McCreary et al. (1993), the speeds in the Somali Current are larger than average speeds obtained from ship drift and buoy data (Richardson and McKee 1989; Molinari et al. 1990) but lower than the synoptic cruises (Swallow et al. 1983). The complex structure of the Somali Current with the onshore and offshore currents are reasonably reproduced including the Great Whirl centered around $9^\circ$N. The simulations of McCreary et al. (1993) failed to develop the Great Whirl, although the present version of their model produces the equatorial jet during

![Fig. 1. (a), (b) Seasonal cycle of model and Reynolds SST along the equator, and (c), (d) standard deviation of model and Reynolds SST.](image-url)
both monsoon transitions (J. McCreary 1997, personal communication). The anticyclonic gyre in February over the entire Bay of Bengal is consistent with observations (Cutler and Swallow 1984) and other model results (Potemra et al. 1991). The South Equatorial Counter Current (SECC), which is fed by the confluence of southward Somali Current and the northward East Africa Coastal Current in February, reaches over 50 cm s$^{-1}$ in the west. The structure of the SECC is in close agreement with the historical ship drift data of Richardson and McKee (1989), and the model results of McCreary et al. (1993) and Woodberry et al. (1989). A
pair of upwelling and a pair of downwelling annual Rossby waves are clearly reproduced in the STIO with maximum amplitude at 12°S (Woodberry et al. 1989).

The wind-driven eastward jet near the equator (Wyrtki 1973) is reproduced in the model during transition periods of the monsoons (not shown) and has speeds of over 1 m s⁻¹ in May. The current speeds and the meridional extent and weakening of the jet in August all agree with Richardson and McKee (1989). The seasonal cycle of the model zonal currents at Gan (0°N, 73°E) are similar, except for weaker eastward current in November, to Fig. 4b of Anderson and Carrington (1993) but differ from observations. The model also produces
an equatorial jet (~40 cm s⁻¹) in November, which is clearly much weaker than the spring Wyrtki jet. J. McCreary (1997, personal communication) also obtained similar results in his layer model. Since level models also produce a weaker jet in autumn (D. Sen-gupta 1997, personal communication), wind-stress products used in forcing the models may be responsible for this shortcoming. The possibility that the layer models produce a mixed layer that is deeper in November than in reality due to shear and salinity effects is being in-
vestigated by W. Han and J. McCreary (1997, personal communication).

The barrier layer (BL) is computed according to the definition of Sprintall and Tomczak (1992). The results for May–July (MJJ) and August–October (ASO) are shown in Fig. 6. As noted by Sprintall and Tomczak (1992), the eastern equatorial Indian Ocean the annual mean \( E - P \) is largely negative (Oberhuber 1988) and sustains a shallow isohaline region leading to a BL formation. In the Bay of Bengal in Fig. 6, the BL is notably

**Fig. 6.** Barrier layer thicknesses (contour interval 10 m) for (a) May–Jul and (b) Aug–Sep. Excess precipitation in the eastern equatorial Indian Ocean leads to a shallow isohaline layer.
missing due to missing river discharges during MJJ and is shallower than Sprintall and Tomczak (1992) for ASO. The eastward extensions during ASO into the Arabian Sea and toward Madagascar are remarkably similar to Sprintall and Tomczak (1992). Some of the differences between model results and observational estimates are simply due to differing definitions for mixed layer thicknesses. Sprintall and Tomczak (1992) conjecture that the BL in the zonal belt along 20°S is probably related to the Central Water formation and subduction caused by Ekman pumping (note the deepening of the isotherms at this latitude in Fig. 3).

We are also interested in comparing the air–sea interactions in certain regions of the Indian Ocean at seasonal and interannual timescales. The seasonal cycle of heat budgets normalized with respect to the mixed layer depth for the Arabian Sea (5°–15°N, 60°–70°E), the Bay of Bengal (5°–15°N, 84°–94°E), western STIO (10°–15°S, 55°–65°E), and eastern STIO (10°–15°S, 105°–115°E) are presented in Fig. 7.

Both the Arabian Sea and the Bay of Bengal display a semiannual character in SST and net heat fluxes. The boreal autumn peak in SST is smaller in both regions, mainly due to lower solar insolation. The net downward flux in the Bay of Bengal is significantly lower in October than in March. The entrainment cooling is maximum in both regions in May due to shallow mixed layer during the transition from the winter monsoon to the summer monsoon. The meridional advection brings cold water into the Arabian Sea from the upwelling region off Somalia in August. Despite a net downward flux in August, there is a cooling of SST due to this advection. In the Bay of Bengal, both the zonal and meridional advections contribute to the early summer cooling together with the entrainment cooling. This has to be viewed with caution since including the input from river discharge can form a freshwater cap and trap radiation that will lead to surface warming.

The components of surface heat fluxes and SST from the climatological and interannual simulations at 15.5°N and 61.5°E (Fig. 8) are in reasonable agreement with the analyses for 1994–95 reported in Weller et al. (1997). The discrepancies between model and the analyses are no greater than the differences between various products in Fig. 8 of Weller et al. (1997). The model produces a stronger cooling during summer monsoon with colder SST at this location in August than in January (not shown). The latent heat loss is ~150 W m⁻² during winter due to dry continental winds blowing off Asia and over ~200 W m⁻² in the summer due to strong wind speeds and warmer SST (see Fig. 3 of Weller et al. 1997). As pointed out by Weller et al. (1997), the sensible heat flux is lost from the ocean during winter, but the ocean gains over 10 W m⁻² during the summer.

The net surface heat flux and SST in the STIO has an annual cycle with maximum SST (minimum MLD) in March–April and minimum (maximum MLD) in August–September. We expect entrainment cooling to be important in western STIO since the thermocline is close to the surface (see Fig. 3). While the strengthening South Equatorial Current brings warmer waters from the east, the meridional term mostly cools the region due to northward advection of colder waters. Both zonal advection and entrainment cooling contribute to produce minimum SST during late boreal summer. Seasonal monsoon reversals affect surface currents in the eastern STIO with the meridional advection bringing warm waters southward and causing maximum warming in May–June. But the net outward flux together with zonal advection and entrainment cooling lead to minimum SST in June. Despite the nearly identical seasonal variation of SST and MLD in the eastern and western STIO, the interannual anomalies are typically out of phase in the two regions. This is discussed in further in the following section.

4. Intercalendary simulations

The initial conditions for the interannual simulations are provided by the climatological spinup. The control run was forced with the Florida State University (FSU) winds for the period of 1980–95 with the wind speeds for latent heat loss computed from the wind stresses. A constant drag coefficient of 1.25 × 10⁻³ was used in computing wind stresses from FSU pseudo-stresses. We employed climatological solar radiation (ERBE) and cloudiness (ISCCP) data due to a lack of reliable data for the entire period of simulation. This is our control run. The differences in mean temperatures from the model and from Reynolds SST over 1982–95 are shown in the top panel of Fig. 9. Note that the SST errors are smaller compared to the climatological simulation (Fig. 2). The improvement in the mean SST simulation is achieved by improved simulation of the D20 as seen in the bottom panel of Fig. 9, especially between 10° and 20°S. It is known that Hellerman and Rosenstein (1983) winds are stronger than expected in the Tropics. Since we are computing model anomalies from the interannual run over 1980–95, the errors in the climatological simulation do not affect our analyses except during the first year of spinup of the interannual run.

Since the wind-driven SST in the Indian Ocean are typically small (see Anderson and Carrington 1993), we carried out another simulation for 1984–90 over which ISCCP solar radiation and cloudiness data are available. Since the ISCCP climatology for radiation differs substantially from the ERBE climatology, we also conducted a run for 1980–95 with the ISCCP climatology for solar radiation in place of the ERBE climatology. It is rather difficult to describe the nature of the wind stress anomalies over the tropical Indian Ocean due to their high spatial and temporal variability from weekly to interannual timescales. Hackert and Hastenrath (1986) noted that the characteristic of the zonal wind anomalies over the STIO appears to be a modulation of the annual cycle. An EOF analyses of the winds was carried out.
by Valenti et al. (1997, manuscript submitted to J. Climate). The zonal and meridional anomalies over the four regions in Fig. 7 are generally in the range ~0.2 dyn cm$^{-2}$. The standard deviations of zonal and meridional wind-stress anomalies show maxima in the zonal band between 10$^\circ$ and 25$^\circ$S. The meridional wind stress also has high variance in the Bay of Bengal and along the Somali coast.

\textbf{a. Effects of solar radiation and cloudiness}

In an OGCM with a variable depth mixed layer and an interactive surface heat flux parameterization, the differences in surface fluxes can generate positive feedbacks between SST and MLD (Murtugudde et al. 1996). The ISCCP climatology in the Bay of Bengal is larger than ERBE by as much as 20 W m$^{-2}$ during July–Au-
gust, which leads to SSTs that are warmer by as much as 0.5°C (Fig. 10). In the presence of a freshwater lens due to river discharge, surface warming can be expected to be much larger. Figure 10 also shows the model SST and SST anomalies (SSTA) for the Bay of Bengal for the control run and EXPIS (the simulation with ISCCP radiation and cloudiness product) for 1984–90. The summer cooling is lower and the autumn warming is greater for each year for EXPIS by as much as 0.5°C. The seasonal cycle of SST is shifted upward by nearly 0.8°C in 1987 and even the SSTA is warmer by 0.5°C. It is thus evident that the differences among satellite products for radiation themselves can introduce SST errors that are as large as model errors. The correlations between model and Reynolds SSTA (not shown) for each of the four regions shown in Fig. 7 improves when we used ISCCP interannual data for cloudiness and radiation instead of the climatologies.

b. Effects of heat flux versus wind stresses

We followed the example of Carton et al. (1996) in order to understand the importance of wind stress and heat flux anomalies in determining the interannual variability of the Indian Ocean. In experiment EXPLH, a simulation is carried out with interannual wind stresses and climatological radiation and cloudiness as in the control run except that the wind speeds in computing the latent heat losses are held to their climatological values. This removes the interannual variability of the latent heat loss and allows only the wind-driven variability. Note that since we are using climatological data for cloudiness and radiation, this also eliminates the interannual variability of the net surface heat fluxes. Another experiment, EXPCL, is conducted by holding the wind stresses to their climatological values while the wind speeds for computing latent heat losses are allowed to vary interannually. Thus EXPCL allows us...
to discern the effects of latent heat loss on model variability.

The standard deviation of the SST anomalies and D20 (depth of the 20°C isotherm) anomalies for EXPCL and EXPLH are shown in Fig. 11. The same quantities for the control run are shown in Fig. 16. When the wind stresses are held to their seasonal monthly mean values, the variability of SST anomalies is reduced over the entire basin, especially in the Somali Current region. The D20 variability is reduced in the Somali Current and the STIO. Thus a large part of the SST variability in the model is due to wind stress forcing. The results for EXPLH where interannual variability of wind speeds was eliminated in the latent heating show a reduction in SST variability in the Arabian Sea and the Bay of Bengal in addition to the Somali Current region. The thermocline variability in the STIO remains nearly un-

 changed since it is mainly determined by the wind-forced model dynamics. Thus the SST and thermocline anomalies along the Somali coast, in the Arabian Sea, and in the Bay of Bengal depend both on wind stresses and surface heat fluxes.

c. Interannual variability of dynamics and thermodynamics

The first EOF and the associated time series for SSTA are shown in Figs. 12 and 13. The first EOF of the model and Reynolds SST anomalies explain 22% and
29% of the respective total variability. The two time series show poor correlation (0.34) for the control run (Fig. 13) but the correlation is a significant 0.52 when ISCCP interannual data are used for insolation and cloudiness. The patterns of EOF-1 for both model and Reynolds SST show a maximum at about 10°S, 60°E. Note that the amplitude of model EOF changes sign to the east of 90°E, whereas Reynolds SST has a minimum in the amplitude without changing signs. Part of the blame for poor simulation of SST anomalies may lie with the use of climatological radiation and cloudiness data, but further investigations are under way to identify model deficiencies in simulating SST anomalies. The correlations between the Southern Oscillation index (SOI) and the time series of the first EOF are 0.57 and 0.52, respectively, for model (SOI leads by ~3 months) and Reynolds SST (SOI leads by ~4 months). The peaks for 1983 and 1988 are larger for Reynolds SST than model results. It was noted by Cadet (1985) that the leading EOF of SST is related to ENSO. While the correlations are statistically significant, a modest correlation between SOI and the time series indicate that there are other factors that are also important in determining the SST and this is consistent with the findings of Reverdin et al. (1986) and Anderson and Carrington (1993).

The model and Reynolds SST anomalies for the Arabian Sea, Bay of Bengal, and western and eastern STIO are shown in Fig. 14. Most of the warm and cold SST anomalies in each region are reproduced by the model. The warming in the Arabian Sea and the Bay of Bengal at the end of 1987 are weak in Fig. 14, but are simulated well when interannual ISCCP data is used for radiation and cloudiness (see Fig. 10). Similarly, in the eastern STIO, the warm anomaly in 1988 for the model and Reynolds SST are much closer when ISCCP data are used. Note that if corrective fluxes were used to remove the discrepancies between model and Reynolds SSTs,
they would typically be smaller than the differences between ISCCP and ERBE climatologies for solar insolation. Since SST anomalies for the Indian Ocean are typically of the order 0.5°C, it is evident that accurate forcing fields for radiation and cloudiness are crucial for simulation of interannual variability of the Indian Ocean. The effects of the ITF may also play a role in determining the SST anomalies and the air–sea interaction (Murtugudde et al. 1998a). This is addressed in the following section.

The warm SST anomaly in the Arabian Sea in 1983 is caused mainly by a weaker summer monsoon (negative anomalies in zonal and meridional wind stresses). A shallower mixed layer accompanied by reduced entrainment cooling and the positive anomaly of net downward flux leads to a warm SST anomaly in mid-July. A negative SST anomaly of over 1°C in the western STIO in mid-June of 1984 is in good agreement with Reynolds SST. The mean easterly zonal wind stress is enhanced by a negative anomaly and the southerly meridional component is increased as seen by a positive meridional wind-stress anomaly. The increase in the curl of the wind stress leads to a rapid shallowing of the thermocline and an increase in entrainment cooling. The anom-
FIG. 13. Time series of the first EOF: model and Reynolds SST, model and SOI, and Reynolds SST and SOI. There is poor correlation between the time series of model and Reynolds SST but each is correlated significantly with the SOI.

FIG. 14. SST anomalies for 1980–95 for the Arabian Sea, Bay of Bengal, western STIO, and eastern STIO. The dashed line in each panel is the anomaly for the region from Reynolds SST.

alities in net surface heat flux and meridional advection are positive but the zonal advection is negative, resulting in the large negative SST anomaly. Thus both in the Arabian Sea and in the western STIO, entrainment into the mixed layer seems to be the main cause for SST anomalies. Unfortunately, the simulation is not sufficiently long to analyze the heat budgets for more events with SST anomalies greater than 0.5°C.

Figure 15 shows that the anomalies of precipitation slightly lead the anomalies in BL thickness and SST in the eastern equatorial region (5°–1°S, 95°–102°E) for 1980–95. Sprintall and Tomczack (1992) proposed that the seasonal variation of the BL thickness in this region is driven by rainfall. Our results indicate that the inter-annual variability of the BL thickness may also be mainly determined by local precipitation anomalies. The processes involved in the variability of the BL in other regions is not investigated here. A significant cooling of the surface occurs in the eastern equatorial Indian Ocean during certain years such as 1961, 1994, and 1997. The barrier layer in the east disappears during such events when the equatorial and alongshore wind anomalies drive intense equatorial and coastal upwelling (Murtugudde et al. 1999, manuscript submitted to J. Geophys. Res.; hereafter MMB99). These events have not only a significant influence on the droughts in Indonesia but also on the floods over parts of east Africa. Negative precipitation anomalies in the eastern Indian Ocean favor the removal of barrier layer and large cool-
ing of SST during such events despite the increased radiation associated with reduced precipitation.

Figure 16 shows the standard deviations of Reynolds and model SSTA, and the anomalies of model D20. The correlation between model anomalies of SST and the depths of 20°C isotherm is shown Fig. 17. The standard deviations of seasonal SSTS and the D20 are larger than those of interannual anomalies. The model displays larger variability in the Somali Current region than Reynolds SST. Similar to the seasonal simulation, smoother Reynolds SST on a coarser grid and the lack of ITF result in larger wind-forced variability of the SST in this region. The model SST variability is also slightly larger off the coast of Sumatra and to the south of Sri Lanka. The largest variability of D20 occurs in the western STIO between 8° and 15°S and also in the Somali Current region. A positive correlation between anomalies of D20 and SST in the Pacific and Atlantic Oceans corresponds to regions of ENSO-type air–sea interactions (Carton et al. 1996). The Indian Ocean does not seem to support such a mode as can be inferred from Fig. 17. Significant correlations in Fig. 17 between anomalies of SST and D20 occur in the coastal upwelling regions of Africa and India; in the open ocean regions of STIO, where Ekman pumping leads to thermocline displacements; and along the coast of Sumatra, where equatorial wind variations control the thermocline variability.

The SSTA along the coast of Sumatra (10°–7°S, 102°–106°E) shows a correlation above 0.5 (significant at 95%) with D20 anomalies in Fig. 17. The SSTA in this region has a correlation of 0.7 with equatorial wind-stress anomalies over 2°S–2°N, 82°–92°E. The zonal SST gradient along 8°S also has significant correlation (0.6) with wind-stress anomalies over the same equatorial region. Clarke and Liu (1994) demonstrated that the equatorial winds control the sea level variations along much of the Indian and Indonesian coast (also see Murtugudde et al. 1998a), which is consistent with Fig. 17. We will return to this in the following section.

The longest time series of current data for the Indian Ocean equatorial currents is that at Gan (0.5°S, 73°E) by Knox (1976). The zonal currents and temperatures anomalies for the model averaged over 0.5°S–0.5°N at 73°E are shown in Fig. 18. As noted in the previous section, the model simulation of the Wyrtki Jet in November is weaker than observed. This is also apparent in the interannual simulation. The zonal wind-stress anomalies show a significant correlation of 0.64 with zonal surface current anomalies. Cane (1980) suggested that the interannual variability of the undercurrent depends on the winds during late months of the previous year. For the period of 1980–95, the strength of undercurrent varies considerably (Fig. 18) with the deepest penetration in 1983 and the strongest core speeds in 1992 and 1993. The vertical propagation seen in Fig. 18 was also noted by Reverdin (1987) and Anderson and Carrington (1993).

A more interesting picture emerges when the temperature anomalies are plotted with depth (Fig. 18). While the SSTA are small (~0.5°C), subsurface anomalies can be larger than 2°C. One possible mechanism is that the Ekman pumping associated with the local wind forcing that drives the undercurrent also drives the thermocline movement. The differences in thermocline location appear as subsurface temperature anomalies. The correlation at 73°E between the SOI and zonal wind stress anomalies is greater than 0.6 between 80° and 90°E with SOI leading by one month. The anomalies of SST and the 20°C isotherm depth are correlated with the SOI at slightly greater than −0.51 (significant at 95% level), with SOI leading by 2 months, which explains the subsurface warming associated with the ENSO events.

d. Interannual variability of sea level

The most prominent feature of sea level variability in the Indian Ocean is the ubiquitous Rossby waves (Woodberry et al. 1989; Perigaud and Delecluse 1992, 1993). The theory of westward propagation and their poleward dispersion due to latitude-dependent phase speed has been discussed by White et al. (1990) and by Perigaud and Delecluse (1992, 1993). However, the effect of these Rossby waves on the interannual variability of SSTA is not very apparent. The zonal section of model sea level anomalies (the mean over the period and the annual cycle have been subtracted) along 12°S and a meridional section along 93°E are compared with Geosat and TOPEX/Posidon data in Figs. 19 and 20. The westward phase speeds (~10 cm s⁻¹) during the Geosat period are in qualitative agreement. The phase speeds show significant differences during the TOPEX period between model and TOPEX sea levels, especially along 12°S. The main reason for this discrepancy is the
deficient model simulation of the upper-ocean thermal structure. Note that the sign of the sea level anomaly switches sign abruptly at around 70°E, which is not always reproduced by the model. This feature is prominent in the altimeter data when the annual Rossby waves are computed and the change in the sign of the amplitude is caused by the mid-ocean ridge (L. Wang 1997, personal communication). A reduced gravity model obviously cannot produce the topographic effects on sea level variations. Woodberry et al. (1989) reproduced this feature by including shallow banks as islands. Interannual variability of the wind-stress curl also produces a sudden change in the sign of the wave as seen in early 1994 near 75°E. The meridional sections along 93°E show increased poleward propagation south of about 10°S that is reasonably reproduced by the model.

Clarke and Liu (1994) demonstrated that equatorial wind variations control sea level variability along much of the Indian and Indonesian coast. The anomalies of sea level for coastal stations Vishakapatnam, Madras, Cochin, and Bombay are shown in Fig. 21. The results are consistent with Clarke and Liu (1994). The east coast stations Vishakapatnam and Madras are highly correlated, with Madras leading by about one month while Vishakapatnam leads sea level at Bombay by more than a month. The sea level variation on the northwest coast of Australia is nearly identical to that on the Indonesian coast for the control run since the ITF has been neglected. The effect of the throughflow is to generate a much larger variation along the Australian coast as discussed in the following section.

e. Effects of the Indonesian Throughflow

Previous studies have shown that the most prominent effect of the ITF is to thicken the thermocline and warm
FIG. 17. The map of correlations between the anomalies of SST and depths of 20°C isotherm. Neglecting the Indonesian throughflow leads to shallower thermocline and larger correlations with SST anomalies along the western coast. Correlations larger than 0.5 are shaded (contour interval 0.2).

the upper Indian Ocean (Hirst and Godfrey 1993; Murtugudde et al. 1998a). A simulation, EXPITF, was carried out by including a sponge layer between Indonesia and Australia between 122° and 124°E. Note that in a layer model such as the one used for this study, the layer thicknesses are also adjusted to those based on Levitus (1994) data and thus allow a mass transport across the boundaries.

The westward transport along 112.5°E in the upper 400 m between the coasts of Indonesia and northwest Australia is a measure of the ITF in the model. Figure 22 shows this transport for the control run (dashed line) and for EXPITF. Also shown in Fig. 22 are the differences in annual mean SST and surface currents, and differences in temperature with depth along 90°E between EXPITF and the control run. The seasonality (maximum/minimum during boreal summer/winter) and the magnitude of westward transport (maximum of 15 Sverdrups; Sv = 10^6 m^3 s^-1) for EXPITF are remarkably similar to the results based on data and other models (see Godfrey 1994). The effect of including a sponge layer at the eastern boundary is to warm the SSTs in the STIO and deepen the thermocline over much of the basin. The SSTs are colder with ITF in the Somali Current region and in the eastern boundary sponge layer, which is rather unrealistic. The model seems to reproduce the subsurface warming better than the SST warming due to ITF. A zonal jet across the latitude of ITF exit is seen in Fig. 22, which turns southward at the east coast of Africa and Madagascar. Due to the sponge layer between 25° and 30°S, the ITF returns as a southward zonal jet toward the coast of Australia. The comparison of 20°C isotherm depths from the model and Levitus data (not shown) shows that inclusion of ITF improves model results at the exit of the ITF (the negative errors in the bottom panel of Fig. 2) but slightly degrades the results to the south of 20°S. These results suggest that the sponge layer at the eastern boundary provides an adequate representation of the ITF and allows us to evaluate the effects of ITF on the seasonal and interannual variability of the Indian Ocean.

The ITF reduces the effects of upwelling on SST due to the spreading of the thermocline and this is seen as reduced standard deviation of SST in the Somali Current region (Fig. 1). The correlation between the anomalies of D20 and SST is also reduced along the Somali coast (Fig. 17). The major effect of ITF in our model is the sea level variability along the northwest coast of Australia. Clarke and Liu (1994) argued that the sea level along the northwest coast of Australia is controlled by the equatorial Pacific winds through coastal wave mechanism. This was confirmed in OGCM experiments by Murtugudde et al. (1997). When the ITF is neglected, the sea level anomalies are nearly identical across the channel, while introducing the ITF leads to significantly lower sea level variability along the coast of Indonesia.

5. Discussion and summary

a. Discussion

It is evident that the SST anomalies in the Indian Ocean are smaller than those observed in the Pacific Ocean. However, the absolute SSTs are much warmer. The anomalies of sea level and SST in the STIO mainly vary on a quasi-biennial timescale. It is known that the rainfall over India also displays a biennial variability. A clear understanding of the role of the Indian Ocean air–sea interaction in the variability of the Indian summer monsoon is of great importance for understanding seasonal to interannual climate variability and for predictability on these timescales.

We demonstrated in Fig. 17 that the anomalies of SST and D20 are highly correlated in the eastern Indian Ocean along the coast of Sumatra. As stated earlier, the thermocline is deeper in the eastern STIO than in the western STIO. Most of the significant SST anomalies in the eastern STIO are typically associated with SST anomalies of the opposite sign in the western STIO. The anomalous sea level and D20 are shown in Fig. 23 for eastern and western STIO. Note the quasi-biennial nature of all the quantities and the out-of-phase relation between the eastern and western STIO. Since the STIO supplies a significant fraction of the moisture for the Indian summer monsoon (Reverdin et al. 1986), the interannual variability of the winds over this region must have some bearing on the interannual variability of the Indian summer monsoon. The zonal wind-stress anomaly over 2°S–2°N, 78°–88°E has a correlation of 0.69 with the SOI over 1980–95 (see Fig. 24). These winds generate equatorial Kelvin waves that propagate as coastal Kelvin waves and lead to a region of high correlation between SST and D20 anomalies (Fig. 17; also see MMB99). The gradient defined by difference in SST between 68°–72°E and 98°–102°E over 2°–6°S has a correlation of −0.88 with the zonal wind-stress anom-
Fig. 18. (a) Zonal currents and (b) temperature anomalies on the equator at 73°E. The temperature anomalies at about 100-m depth show ENSO-related warming.

alies above with a lag of about one month. This SST gradient change in the STIO leads the Somali Jet (averaged zonal wind stress over 5°−15°N, 50°−70°E) by two months with a correlation of −0.34. The correlation between the SST gradient and the Somali Jet is rather low over this short period, but it is significant for the Comprehensive Ocean–Atmosphere Data Set data over the period of 1945–93 (Murtugudde et al. 1998b). The Somali Jet is highly correlated with the Indian summer monsoon (Reverdin et al. 1986).
Fig. 19. Zonal sections of sea level along 12°S: (a) Geosat and TOPEX data and (b) model sea level for the Geosat and TOPEX periods. The change in sign of the anomalies related to bottom topography are not reproduced by the reduced gravity model.
Fig. 20. Meridional sections of sea level along 12°S: Geosat and TOPEX data (top) and model sea level for the Geosat and TOPEX periods. Poleward dispersion is clearer in the Southern Hemisphere.
Fig. 21. Sea level anomalies along the coast of India (Vishakapatnam shown in solid line). The stations on the eastern coast in the Bay of Bengal are nearly in phase but the stations on the western coast lag sea level at Vishakapatnam.
Based on these analyses, we propose that the zonal wind-stress anomalies on the equator in the Indian Ocean, which are correlated with the SOI, generate anomalous SST gradient in the STIO. The anomalous SST gradient in the STIO can affect the Somali Jet, which in turn affects the rainfall over India. Such a connection would also explain the previously proposed correlation between summer monsoon rainfall over India and the SST over the Arabian Sea through the air–sea interaction discussed in Figs. 7 and 14. Further analyses based on data and longer simulations of our OGCM forced with National Centers for Environmental Prediction reanalyses winds confirm the correlations (Murtugudde et al. 1998b). However, coupled model experiments will be required to determine if the SSTAs are a passive response to winds generated by other mechanisms such as precipitation forcing or if the ocean plays an active role in the wind variability over the Indian Ocean.

b. Summary

A reduced gravity, primitive equation, sigma-coordinate OGCM coupled to an advective atmospheric
mixed layer model is used to study the seasonal and interannual variability of the tropical Indian Ocean. The model simulation of the seasonal dynamics and thermodynamics are in close agreement with available observations and other model results. The seasonal cycle of SST along the equator shows eastward phase propagation with larger variability in the east, which is opposite of what is observed in the equatorial Pacific. The barrier layer variability is large in the eastern equatorial Indian Ocean where rainfall variability is large and also in the zonal band between 10° and 20°S. Neglecting river discharges in the Bay of Bengal affects model simulation of mixed layer and barrier layer processes. Analyses of seasonal heat budgets for the mixed layer shows that the western STIO has a large seasonal cycle of SST and mixed layer because the thermocline is close to the surface due to Ekman pumping. Interannual sea level anomalies of over 10 cm and SST anomalies of over 1°C are caused by changes in wind-stress curl. The seasonal cycle in the eastern STIO is similar, albeit slightly smaller due to a deeper thermocline. However, the interannual SST anomalies in the western and eastern STIO tend to be of opposite sign.

Our control run for 1980–95 employed interannual wind stresses but climatological fields for solar radiation and cloudiness. Thus only the latent heat losses have interannual variability due to wind-speed variations. A major drawback for the interannual simulations is the lack of solar radiation and cloudiness data. As was noted by Cadet (1985) and Reverdin et al. (1986), interannual anomalies in cloudiness and radiation are significant over the tropical Indian Ocean. The mean SSTs are high, but the interannual SST anomalies are rather small over much of the basin. Anderson and Carrington (1993) concluded that the wind-driven variations in SST are not a good indicator of interannual SST variability in the Indian Ocean. Our sensitivity experiment with climatological wind stresses but interannual wind speeds in the computation of latent heat fluxes show a noticeable reduction in standard deviation of the SST anomalies over much of the basin. A variable depth mixed layer model with interactive surface heat flux formulation leads to a different conclusion than that of Anderson and Carrington (1993). The parallel experiment with climatological wind speeds but interannual wind stresses demonstrates that the variability of SST anomalies in the Arabian Sea, the Somali Current, and the STIO depend on surface heat fluxes as well as wind forcing. Model experiment with the available ISCCP
The exit of ITF. This zonal jet turns southward along the warm waters in a westward zonal jet at the latitude of warming of the tropical Indian Ocean and advection of northern wintertime transport of 5 Sv leads to subsurface observations and other model simulations. A maximum that is remarkably similar to estimates based on observations and other model simulations. A maximum 0.51, suggesting that there are other factors that are equally important in determining SST anomalies over the Indian Ocean. It was noted before (Reverdin et al. 1986) that non-ENSO-related SST anomalies over the Indian Ocean can be as large as those related to ENSO. There are differences in the leading EOF of model and Reynolds SST anomalies that are most likely related to lack of interannual data for radiation and cloudiness. Significant ENSO-related subsurface temperature and zonal current anomalies seen on the equator at 7°E are driven by zonal wind-stress anomalies on the equator. Sea level anomalies along much of the Indian and Indonesian coasts are driven by equatorial wind variations as was suggested by Clarke and Liu (1994). The model Rossby wave speeds and phases are in reasonable agreement with Geosat data but show differences with TOPEX/Poseidon data. A deficient upper-ocean thermal structure in the model simulations during the warm 1990s is largely responsible for the discrepancies with TOPEX data.

The anomalies of SST and depth of 20°C isotherm display positive correlations in the western STIO where the thermocline is close to the surface and also along the coast of Sumatra. The zonal wind stresses along the equator between about 80° and 90°E are correlated to the SOI and these winds drive thermocline variations off Sumatra through coastal wave dynamics. This process was clearly responsible for the anomalous cooling off the coast of Sumatra during the autumn of 1997 (MMB99). The SST anomalies in the eastern STIO driven by the equatorial winds are typically opposite in sign to those in the western STIO. The resultant change in SST gradient across the basin is correlated with the Somali Jet, which in turn is highly correlated with the rainfall over India. It is also known that the evaporation in the STIO supplies a large part of the moisture for the Indian summer monsoon. This connection between the SOI and the Indian monsoon through the STIO is being investigated further.

The effects of the ITF on the interannual variability of the Indian Ocean is examined by introducing a crude representation of the ITF through a sponge layer between Australia and Indonesia (122°–124°E). This produces a seasonal westward transport in the upper 400 m that is remarkably similar to estimates based on observations and other model simulations. A maximum transport of 15 Sv in northern summer with minimum northern wintertime transport of 5 Sv leads to subsurface warming of the tropical Indian Ocean and advection of warm waters in a westward zonal jet at the latitude of the exit of ITF. This zonal jet turns southward along the east African coast and returns eastward to the Australian coast at about 25°S. The warming of SST is seen mainly in the zonal band between 10° and 25°S. Including the ITF results in a reduction of the effects of upwelling on SSTs in the Somali Current region on seasonal and interannual timescales. The sea level anomalies along the coasts of Indonesia and northwest Australia are of similar magnitude and in phase when the ITF is neglected. The variability of sea level on the Australian coast is significantly larger, as in observations, when the ITF is introduced.

One of the Climate Variability and Predictability Program (CLIVAR) goals is to build understanding and predictive capabilities of the interaction of monsoons with the Indian Ocean, ENSO, and land surface processes. A detailed picture of the air–sea interactions and the variability of SST in the Indian Ocean is essential for achieving this goal. While our model study explored this objective with some success, a major limitation seems to be the lack of reliable data for solar radiation and cloudiness. The SST anomalies are typically of order 0.5°C over most of the basin. However, since the absolute temperatures are high, even small variations in temperatures may be climatically important (Palmer and Mansfield 1984). Accurate forcing fields become more important when the signal being simulated is small. It is discouraging that the differences between the climatologies of two satellite products for solar radiation (ERBE and ISCCP) are larger than the corrective fluxes required to remove the differences between model and observed SSTs. However, wind-driven variability is reproduced very well and provides hope for progress toward CLIVAR goals.

Acknowledgments. RM wants to dedicate this paper to his little sister, who lived a short life but suffered too long. Dr. Stephan Howden spent considerable time on helping us with analyses of altimetric data and we are grateful to him. RM was partially supported by NASA RTOP 622-48-02-00. AB would like to acknowledge NASA RTOPS 665-55-04-02 (TOPEX Extended Mission) and 622-48-02-00. We deeply appreciate the assistance of James Beaufort with some of the analyses and figures. We have had unlimited access to the FSU machines and we are indebted to Prof. Jim O’Brien for allowing us access.

REFERENCES


——, and B. Diehl, 1984: Interannual variability of surface fields


Price, J., R. Weller, and R. Pinkel, 1986: Diurnal cycle: Observations...


