Role of the Indian Ocean in Regional Climate Variability

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Abstract

The role of the Indian Ocean in the regional climate variability has been studied for a long time. Whether the Indian Ocean plays an active role or simply responds passively to the wind and heat flux variability generated elsewhere remains somewhat of an open question to this day. Here we attempt a fairly comprehensive review of the literature relating to the Indian Ocean variability at all time-scales and the current understanding of the role this tropical ocean plays in the coupled climate system. Despite an investigative history of more than a century, it is fair to say that the role of the Indian Ocean in regional monsoon variability, the most important climate process in this sector, remains to be fully understood and remains an active area of diagnostic and modeling research. This translates into limited success of the state of the art monsoon forecast systems, statistical and dynamical. Much attention in the last few years has been focused on the east-west mode of variability, referred to as the dipole or the zonal mode. While there is incontrovertible evidence that the Indian Ocean plays an active role in some of these dipole/zonal mode events, it is still a matter of debate as far as the self-sustainability of this mode and its impact on regional and global climate is concerned. The most significant impediment to improving the predictive understanding of the region is the basic fact that sea surface temperature variability of the Indian Ocean is most often in the range of observational errors at all time-scales. The need for coordinated and sustained observations and diagnostic studies along with continued investigations with a hierarchy of models has been well-recognized along with the fact that monsoons form an integral part of the global climate system and the ENSO cycles. It is thus hoped that the role of the Indian Ocean in the coupled climate system will be much better quantified in the coming years, leading to significant improvements in regional coupled climate forecasts.

1. Introduction

1.1 Background

Anomalous events in the Indian Ocean (IO) during 1994 and 1997 have generated much renewed interest in all aspects of the coupled climate system over this sector. The purpose of this study is to review the literature on IO dynamics and thermodynamics and place these studies in the context of the role of the IO in regional climate variability. The IO basin can be considered to consist of the northern and southern basins with their own distinct features since the Northern IO is subject to the seasonally reversing monsoon circulation. The equatorial band that connects the northern and southern basins undergoes some unique air-sea interactions of its own. In this section we provide a rather extensive, albeit not comprehensive, literature survey to indicate that despite the overwhelming number of high quality studies of the IO, they are not always focused on the role of the IO in climate variability.

Most earlier studies of the IO were typically motivated by the unique aspect of the region, namely, the seasonally reversing monsoon forcing (Lighthill 1969, Leetmaa 1972, Anderson and Rowlands 1976, see Schott and McCreary 2001 for a review). Numerous studies reported so far address the impact of monsoonal forcing on the physical, biological, and biogeochemical variability of the IO basin. Majority of them relate to climatological features as opposed to the Pacific basin where the number of studies addressing the interannual to decadal variability far outnumber the climatological studies (see TOGA Special Issue of J. Geophys. Res., 1998).

Figure 1a shows the standard deviation (a measure of interannual variability) of sea surface temperature (SST). It is evident that unlike the tropical Pacific (e.g., McPhaden 1999), the SST variance in the IO is relatively small. This may account for the lack of motivation to address the heat budgets that control SST variability, the main variable of interest for atmospheric forcing.

Although small in magnitude, the SST variability, particularly that over the Southern IO is tightly coupled to thermocline variability (Figure 1b), and therefore expected to influence the regional climate variability. Yet, many of the studies investigating the heat budgets for SST variability in the IO did not seek thermocline/SST feedbacks as a main motivation (Shetye 1986, Molinari et al. 1986, McCreary and Kundu 1989, McCreary et al. 1993, Anderson and Carrington 1993, Behera et al. 2000). Lately, such a feedback is being investigated to find the role of IO in regional monsoons or the IO response to ENSO forcing (Hastenrath and Greischar 1993, Murtugudde and Busalacchi 1999, Schiller et al. 2000, Loschnigg and Webster 2000, Rao and Sivakumar 2000, Xie et al. 2002, Meehl et al. 2003).

Some atlases for the IO have been constructed which provide an excellent background for any IO studies (e.g., Hastenrath and Greischar 1989, Rao et al. 1989). Recently, Schott and McCreary (2001) have provided an excellent review of the physical oceanography of the IO. Unlike the numerous volumes of literature, review articles and books on El Niño-Southern Oscillation (ENSO) and its global impacts (e.g., Philander, 1990, TOGA Special issue of J. Geophys. Res., 1998), a comprehensive review on the role of IO on regional climate variability is unavailable. In the present review, our intention is not to provide any new results but to place the past studies in context and highlight some outstanding issues that must be addressed in the coming years. After the anomalous IO events of 1994 and 1997, there is a surge of interest to understand the interannual variability of the IO dipole/zonal mode (Saji et al. 1999, Webster et al. 1999, Anderson 1999, Murtugudde et al. 2000, Yu and Rienecker 1999). In light of this recent interest in the IO, such a review is appropriate and timely. In the following subsections, articles related to dynamics and thermodynamics and their collective impact on SST in various regions of the tropical IO are reviewed. This will lay the foundation for the role of these regional SST anomalies on the regional climate variability, particularly on the Asian-African-Australian monsoon systems.

1.2 Northern Indian Ocean

1.2.1 Arabian Sea: The IO variability is fundamentally affected by the northern land boundary that consists of the Asian continent and the Indian subcontinent, which bisects the Northern IO into the Arabian Sea and the Bay of Bengal. The massive land in the north sets up the required land-sea thermal contrast that is instrumental for the development of the monsoon. The response to the reversing wind field is most dramatic in the Somali upwelling region in the western boundary of the Arabian Sea with the Somali Current directed equatorward during the Northeast (winter) Monsoon (NEM) and poleward during the Southwest (summer) Monsoon (SWM). Related circulation changes are seen over the entire Northern IO in terms of surface current reversals (Hastenrath and Greischar 1989, McCreary and Kundu 1993, Schott et al. 1994, Shankar et al. 1996, Shankar et al. 2002), coastal upwelling (Shetye et al. 1990, Vinaychandran and Yamagata 1998, Murtugudde et al. 1999), sea level variability (Clarke and Liu 1993, Clarke and Liu 1994, Eigenheer and Quadfasel 2000, Brandt et al. 2002), and associated biological responses (Bhanse and McClain 1986, Brock et al. 1991, Murtugudde et al. 1999, Prasanna Kumar et al. 2002). The variability of the subsurface currents over the Arabian Sea has also been studied (Luyten and Swallow 1976, Leetmaa and Stommel 1980, Jensen 1991, Flagg and Kim 1998, McCreary et al. 1996, Vinaychandran et al. 1999, Beal et al. 2000, Dengler and Quadfasel 2002).

The Arabian Sea receives high salinity waters from the Red Sea and the Persian Gulf that tend to stay below the surface layer (Maillard and Soliman 1986, Cember 1988, Prasanna Kumar and Prasad 1999, Prasad et al. 2001, Prasad and Ikeda 2001). A pertinent issue that remains unknown is the effect of these physical and dynamical changes on the SST variability. It is known from the above studies that most of the SST variability in the Arabian Sea can be attributed to changes in surface heat fluxes (McCreary and Kundu, 1989, Rao et al. 1989, Murtugudde and Busalacchi, 1999, Rao and Sivakumar 2000). There is evidence that SST anomalies over the Arabian Sea do influence the interannual variability of the monsoon (Section 2.2).

1.2.2 Bay of Bengal: The Bay of Bengal receives freshwater input from several major rivers such as Irrawaddy, Ganga, and Brahmaputra in addition to significant amounts of rainfall during the SWM season (Shetye et al. 1996, Vinayachandran et al. 2002). While synoptic observational estimates show the formation of a freshwater induced barrier layer (Lukas and Lindstrom 1991, Sprintall and Tomczak 1992, Vinayachandran et al. 2002), buoy observations and modeling studies indicate that the impact of freshwater input on regional SST is not as obvious (Sengupta and Ravichandran 2001, Han et al. 2001, Howden and Murtugudde 2001). Apart from the differences from the Arabian Sea in terms of local forcing, the Bay of Bengal is also characterized by remote forcing from the equatorial region (Yu et al. 1991, Potemra et al. 1991). The wind forcing and the freshwater forcing also result in temperature inversions in the Bay (Shetye at al. 1996, Howden and Murtugudde 2001) and lead to higher sea levels in the Bay of Bengal than in the Arabian Sea (Shankar and Shetye 2001). The circulation around the southern tip of India and Sri Lanka connects the Bay of Bengal to the Arabian Sea (Schott and Fischer 1994, Reppin et al. 1999, Vinayachandran et al. 1998). The extent of the intrusion of the Bay of Bengal waters into the Arabian Sea has not been observed sufficiently (Shetye 1993, Jensen 2001). In a recent study, Jensen (2003) shows that relatively fresh Bay of Bengal water is transported southward across the equator throughout the year east of 90^{o}E, but during the SWM as far west as 60^{o}E. The surface wind speed variability over the Bay of Bengal is less than that over the Arabian Sea which results in differences in biological and physical responses (Shenoi et al. 2002, Prasanna Kumar et al. 2002). This along with the freshwater input results in less vertical mixing in the upper layers of the ocean leading to high mean SST (Rao 1987, Shenoi et al. 2002, Bhat et al. 2001, Webster et al. 2002). The higher mean SST in the head Bay (Figure 2) is instrumental for the maintenance of the higher mean precipitation during the monsoon season (Figure 3b, Shenoi et al. 2002). In summary, the SST variability in the Bay of Bengal is small except at intraseasonal time scales (Section 2.6).

1.3 Equatorial Region

Unlike the other two oceans, the IO has no significant upwelling in the eastern equatorial region. The upwelling is, however, stronger in the western equatorial region and the seasonal cycle of the SST propagates eastward unlike the equatorial Atlantic and Pacific Oceans where it propagates westward. The IO is the only tropical ocean where the annual-mean winds on the equator are westerly, resulting in a deeper thermocline in the eastern equatorial region (Reverdin et al. 1986). The lack of equatorial upwelling is related to the lack of trade wind-like structure. This leads to an equatorial current structure that is characterized by annual mean eastward surface current and an absence of a semi-permanent, basin-wide, equatorial under current (Swallow 1967, Taft 1967, Leetmaa and Stommel 1980, McPhaden 1982, Murtugudde and Busalacchi 1999).

The most prominent feature of the equatorial IO is the semi-annual equatorial Kelvin waves and the associated eastward jets (Wyrtki 1973, O'Brien and Hurlburt 1974, Yamagata et al. 1996, Sprintall et al. 2000). The eastward Wyrtki Jets affect the eastern equatorial IO warm pool variability directly (Han et al. 1999, Murtugudde and Busalacchi 1999). All year around, the SST in the eastern equatorial Indian Ocean in the vicinity of (90^{o}E-110^{o}E, 10^{o}S-0) is warmer than 27.5^{o}C (Figure 2), the threshold required for deep convection in the tropics (e.g., Gadgil et al. 1984). Unlike the Northern IO, the eastern equatorial IO experiences heavy rainfall throughout the year (Figure 3) resulting in a barrier layer (Sprintall and Tomczak 1992, Masson et al. 2002), and the barrier layer variability there may play a role in the interannual variability of the tropical IO (Murtugudde and Busalacchi 1999, Murtugudde et al. 2000, Annamalai et al. 2003a, Section 3.1.3). Recent research interest lies in understanding the SST and thermocline variability in the near-equatorial region (Section 1.5). The coupled process in the equatorial IO has considerable impact on the Asian-African-Australian monsoon systems (Section 2).

1.4 Southern Indian Ocean

1.4.1 Southwest Indian Ocean (SWIO): The Southern IO is influenced by predominant southeasterlies to the south of about 10^{o}S and seasonally reversing southwest and northeast monsoons to the north of 10^{o}S. Thus the deep tropics between the equator and about 10^{o}S is subject to seasonally reversing Ekman pumping which leads to a doming thermocline structure, especially in the west (Reverdin 1987, Murtugudde and Busalacchi 1999, Xie et al. 2002). This band of shallower thermocline results in a biological signature in response to wind-forced surface entrainment (Murtugudde et al. 1999) that is not seen as clearly in SST variability (Schott and McCreary 2001). As suggested by Figure 1b, there is a strong linkage between SST and thermocline variations over the SWIO in the vicinity of 8^{o}S, 60^{o}E. Xie et al. (2002) showed that the SWIO thermocline variability is influenced by Rossby waves, whose signal may extend into the equatorial region (Tsai et al. 1992, Hastenrath et al. 1993, Murtugudde et al. 2000, Vinayachandran et al. 2003). The importance of SWIO SST variability on the South African rainfall variability will be addressed in Section 2.

1.4.2 Southeast Indian Ocean and Indonesian Throughflow: One major factor in the variability of the Southern IO is the introduction of Pacific waters through the Indonesian Seas via the Indonesian throughflow (ITF) which has been studied extensively (see Godfrey 1996 for a review). To first order, the effect of upwelling on SST variability off Java is reduced due to the spreading of the thermocline by the ITF (Murtugudde et al. 1998). The basin scale impacts of the ITF have typically been studied in model simulations with the ITF closed (Hirst and Godfrey 1993, Schneider and Barnett 1997, Murtugudde et al. 1998, Wajsowicz and Schneider 2001). The role of the heat carried in via the ITF into the IO on regional climate variability remains to be fully understood (Reason et al. 2000, Schneider and Barnett 1997, Banks 2000, Wajsowicz and Schneider 2001).

In contrast to the Pacific and the Atlantic Oceans, the Southeast IO is devoid of significant upwelling. The seasonal upwelling off Java/Sumatra is fairly weak compared to similar latitudes in the eastern Pacific and Atlantic Oceans. The west coast of Australia has persistent alongshore winds which are upwelling favorable. However, the steric gradient set up by the ITF drives a coastal current poleward into the prevailing winds leading to the Leeuwin current complex (Godfrey and Ridgeway 1985, Godfrey and Weaver 1991, Holloway 1995, Morrow and Birol 1998). Despite these constraints, occasionally, there is significant upwelling of cold thermocline waters off the coasts of Java and Sumatra (as indicated by thermocline feedback in Fig. 1b), leading to a cold tongue for a few months. In summary, most of the observed SST variance in the Northern IO is due to heat fluxes (e.g., Klein et al. 1999, Rao and Sivakumar 2000) while ocean dynamics contribution is substantial over the Southern IO, particularly in the neighborhood of the equatorial zone (e.g., Murtugudde et al. 2000; Xie et al. 2002). The SST variability over the subtropical Southeast IO (poleward of 20^{o}S) is largely due to heat fluxes. Xie et al. (2002)

suggested that the SST variability off Sumatra excites atmospheric waves that pass over this region and generate SST variability there.

1.5 ENSO, and non-ENSO Variability in the Indian Ocean

1.5.1 ENSO signature: The IO response to ENSO is well known to be a basin scale surface warming/cooling with a lag of about a season (Nigam and Shen 1993, Tourre and White 1995, Nicholson 1997, Klein et al. 1999, Enfield and Mesats-Nunez 1999, Venzke et al. 2000), whereas the subsurface response appears to be simultaneous, forced by ENSO related wind anomalies (Reverdin et al. 1986, Hastenrath et al. 1993, Shen and Kimoto 1999, Murtugudde and Busalacchi 1999, Chambers et al. 1999, Webster et al. 1999, Schiller et al. 2000, Hastenrath 2002, Krishnamurthy and Kirtman 2003, Masson et al. 2003, Allan et al. 2003). Some recent studies argue that subsurface temperature variability in the IO is predominantly quasi-biennial and is independent of ENSO variability (Rao et al. 2002). In the basin-wide structure the maximum SST variance is observed over SWIO (e.g., Nigam and Shen, 1993; Figure 1a). Analysis of in situ measurements and a model-assimilated dataset reveals a strong influence of subsurface thermocline variability on SST over the SWIO, and ENSO is found to be the dominant forcing for the SWIO thermocline variability (Xie et al. 2002).

As is well known, the Pacific undergoes extended El Niño or La Niña like conditions (Trenberth and Hoar 1996) and the African, Asian, and Australian monsoons are affected differently by these conditions than the traditional ENSO effects as noted by Allan and D'Arrigo (1999) and Allan et al. (2003). The associated IO response and likely feedbacks to ENSO or monsoons have not yet been investigated in detail. The ISM-monsoon relations also display interdecadal variability (Krishna Kumar et al. 1999, Krishna Kumar et al. 2000, Krishnamurthy and Goswami 2000), the causal links for which are not well understood.

1.5.2 Indian Ocean Zonal/Dipole Mode: A new wave of interest has emerged in understanding the air-sea interactions in the equatorial Indian Ocean, known as Indian Ocean Zonal or Dipole mode (Reverdin et al. 1986, Kapla et al. 1994, Hastenrath et al. 1993; Murtugudde et al. 1998, Murtugudde and Busalacchi, 1999, Saji et al.1999, Webster et al. 1999, Yu and Reickner 1999, 2000). Of particular importance is the SST variability off Java/Sumatra and its relationship to thermocline (Figure 1). Some of the interpretations of the surface and subsurface variability associated with the dipole/zonal mode have been reported to be independent of ENSO (Saji et al. 1999, Iizuka et al. 2000, Rao et al. 2002) but the debate continues (Nicholls and Drosdowsky 2001, Baquero-Bernal et al. 2002, Hastenrath 2002, Huang and Kinter 2002, Krishnamurthy and Kirtman, 2003, Li et al. 2003, Murtugudde et al. 2003, Yamagata et al. 2003, Philander 2003). Wang et al. (2003) proposed that the atmospheric Rossby wave-SST dipole feedback is involved in maintaining the air-sea interaction in the southern IO during the developing phase of ENSO. On the other hand, from a simple correlation analysis Behera and Yamagata (2003) suggest that IO dipole/zonal mode can influence the pressure anomalies over Darwin and hence ENSO itself. The ENSO related IO response is also seen to produce a cooling off Java/Sumatra during the developing phase of El Niño (Wolter and Hastenrath 1989, Yu and Reinecker 1999, Krishnamurthy and Kirtman 2003, Hendon 2003) that appears to be surface heat flux driven. The anomalous cooling during certain years such as 1994 and 1997, however, appear to involve a coupled mode with a Bjerknes feedback between the thermocline movement, SSTs, precipitation and winds (Bjerknes 1969, Murtugudde et al. 1998, Murtugudde and Busalacchi 1999, Saji et al. 1999, Ueda and Masumoto 2001, Annamalai et al. 2003a). We return to these issues in Section 3.

1.6 Features common to the subregions of Indian Ocean

Despite the differences in the subregions of the IO, there are many features that are common to the IO basin. One of the ubiquitous features is the presence of Rossby waves at seasonal to interannual time-scales (Periguad and Delecluse 1992, Perigaud and Delecluse 1993, Masumoto and Meyers 1998, Brandt et al. 2002, Birol and Morrow 2001, Subramnyam et al. 2001). The monsoonal winds over the Indian Ocean are also known for their intraseasonal variability which induces multiple recirculation zones, eddies, and gyres as noted in many of the observational and modeling studies (Cox 1979, Bruce et al. 1981, Kindle and Thompson 1989, Woodberry et al. 1989, Bruce et al. 1994, Fischer et al. 1996, Garternicht and Schott 1997, Sengupta et al. 2001, Shetye et al. 1993, Tsai et al. 1992, Vinayachandran and Yamagata 1998, Morrow et al. 2003).

The annual mean heat fluxes in the Indian Ocean show a net heat gain to the north of about 10^{o}S (Josey et al. 1999). Considering that the IO is bounded to the north by the Asian landmass, the heat gained in the IO has to be transported out across the equator to the south. The issues of this meridional circulation and the details of how water masses can cross the equator are addressed in a number of studies (Wacogne and Pacanowski 1996, Lee and Marotzke 1998, Miyama et al. 2003).

The review article is organized as follows. In Section 2, we provide a comprehensive review of the various studies that explore the role of the IO in the monsoon variability and Tropospheric Biennial Oscillation (TBO). Recent interest in the IO region has been spiked by the anomalous events of 1994 and 1997 with some published works hypothesizing an internal IO mode that is independent of ENSO. The role of the IO dynamics in this mode, and the role of this mode on the regional climate variability are discussed in Section 3. Section 4 addresses some outstanding issues in the IO variability. We summarize our review in Section 5.

2 Role of the Indian Ocean in Monsoon Variability

The tropical IO and the land mass around its rim experiences one of the most energetic components of the Earth's climate system, the Asian-Australian-African monsoon system. In fact, deep convection/intense precipitation occurs in all the seasons around the IO rim (Figure 3). The Asian summer monsoon peaks during June-September (Fig. 3b), the African monsoon establishes in October-November (short-rains, Fig. 3c), and the Australian monsoon occurs during December-February (Fig. 3d). During boreal spring (March-May) and winter the SWIO, extending into Southern Africa experiences deep convection (Figs. 3a,d).

The variability of the monsoons, at both intraseasonal and interannual time scales exert considerable influence on the socio-economic aspects in many regions such as south Asia, Africa, and Australia (e.g., Webster et al. 1998). There is increasing evidence that the diabatic heating associated with the monsoon system influences the global climate (e.g., Rodwell and Hoskins 1996). Owing to both scientific and social importance, a large volume of literature that describes the annual cycle and variability of the monsoons is available in the form of review articles (e.g., Shukla 1987, Webster et al. 1998, Hastenrath 1988, Holland 1986, McBridge 1987) and books (e.g., Lighthill and Pearce 1981, Fein and Stephens 1987, Chang and Krishamurti 1987, Hastenrath 1988, Pant and Rupakumar 1997). Despite these sustained efforts the monsoons still remain one of the toughest challenges for the modeling community (Sperber and Palmer 1996, Gadgil and Sajani 1998).

In this section, first we present the arguments for why the SST anomalies are important for the interannual variability of the monsoon (Section 2.1), then we review the observational and modeling studies that focused on the role of IO SST on the Indian summer monsoon (ISM, Section 2.2), the African monsoon (AFRM, Section 2.3), and the Australian monsoon (AUSM, Section 2.4). Based on the above linkages, we review the studies that implicate the role of IO on the TBO (Section 2.5). Recent studies suggest that the statistics of intraseasonal variability can influence the interannual climate variability, particularly over the monsoon regions. Therefore, we conclude the section with a review of the effects of IO SST on the summer and winter season intraseasonal variability (Section 2.6).

2.1 Effect of Lower Boundary Forcing on the Monsoon Variability

The land-sea thermal contrast between the Asian land mass and tropical IO plays an important role during the onset and developing stages of the ISM (e.g., Shukla 1987, Webster et al. 1998). The north-northwestward migration of convection associated with the annual cycle is tightly linked to the poleward migration of maximum SST over the Asian-Australian monsoon region (Meehl 1987, Webster et al. 1998, Figures 2, 3). Joseph (1990) hypothesized that the warm pool of the northern Indian Ocean in May-June forces the onset of the ISM. From a modeling study the instrumental role of the seasonal cycle of IO SST on the monsoon evolution was demonstrated by Shukla and Fennessy (1994).

A successful prediction of the seasonal mean rainfall and circulation is based on the premise that the monsoon is dynamically a stable system and its seasonal mean rainfall and circulation and their interannual variability are largely governed by slowly varying boundary conditions such as SST, snow cover, soil moisture, etc. (Charney and Shukla 1981, Shukla 1981). Since the seminal works of Walker and Bliss (1932), many observational (e.g., Sikka 1980, Rasmusson and Carpenter 1982, Webster and Yang 1992, Slingo and Annamalai 2000, Miyakoda et al. 2002) and modeling studies (e.g., Keshvamurthy 1982, Palmer et al. 1992, Ju and Slingo 1995, Yang and Lau 1998, Soman and Slingo 1997, Nigam 1998, Lau and Nath 2001) have confirmed the strong influence of the contemporaneous SST anomalies in the central-eastern Pacific on the

ISM. Similarly, the variations of the AUSM (e.g., Nicholls 1984) and AFRM (e.g., Farmer 1988, Hastenrath et al. 1993) are linked to ENSO.

At interannual time scales, despite the above evidence that ENSO dominates the monsoon variations, the SST variability in the IO can also influence the monsoons, particularly in years when the local SST anomalies are strong and/or years when the SST anomalies in the tropical Pacific are small and insignificant. Such a premise is possible since much of the central-eastern tropical IO lies in a warm pool region (SST > 28^{o}C, Figure 2b), and due to the non-linearity in the Clausius-Clapeyron equation a small variation in SST (~ 0.5^{o}C) could have larger impact on moisture availability, and therefore on the tropical convection (Soman and Slingo 1997, Zhang 1999). From a diagnostic study, Lau and Wu (1999) note that about 19% of the Asian-Australian monsoon variability is due to local coupled processes.

2.2 Indian Ocean and the Indian Summer Monsoon

2.2.1 Observational Studies: During the pre-monsoon season (March-May) warmest SST in the global oceans is observed over the Northern IO (e.g., Joseph 1990, Sengupta et al. 2002, Figure 2a). Therefore, a moderate fluctuation in SST over the Northern IO during spring and early summer is expected to have a larger impact on the ensuing ISM primarily through changes in evaporation and moisture convergence. With the availability of reliable SST observations/analyses for the last 4-5 decades, many observational studies indicated a causal relationship between Arabian Sea SST anomalies and ISM. Ellis (1952) carried out the first ever case study and noted that during a strong (weak) ISM year the SST anomalies were above (below) normal over the Arabian Sea. This case study, probably motivated the subsequent observational studies since the physics involved was rather appealing: warm SST can produce more evaporation and therefore stronger ISM rainfall (Shukla 1987). Using long observations along particular ship tracks, many studies examined the correlations between ISM and SST over the Arabian Sea (e.g, Shukla and Misra 1977, Weare 1979). Rao and Goswami (1988) find that SST anomalies during March-April in the region of 5^oN-10^oN, 60^oE-75^oE to be significantly correlated with the ISM. Clark et al. (2000) propose predictive relationships between Arabian Sea SSTs over autumn and winter to ISM in the following year while Terray (1995) found quasi-biennial and ENSO time-scale SST modes in the IO that were related to ISM. Lau et al. (2000) showed that boreal spring warming in the northern Arabian Sea has a significant impact on the ISM. Li et al. (2001) conclude that IO SSTs do influence ISM on the TBO time-scale through local moisture convergence.

The first-ever observational study to examine the importance of Arabian Sea water vapor flux on the ISM was done by Pisharoty (1976). He reported that the in situ flux of water vapor over the Arabian Sea was much larger than that obtained from the Southern IO and underscores the importance of local SST anomalies. The research that followed these findings, however, obtained contradictory results. The calculations of Saha and Bavadekar (1973), Saha (1974), Cadet and Reverdin (1981), and Cadet and Diehl (1984) suggested that the Southern IO is the main moisture source region for the ISM. Yet, there is no demonstrative relationship between ISM and SST over the Southern IO except for the correlation between the SST gradient, Somali Jet, and ISM reported by Murtugudde et al. (1998) and Murtugudde and Busalacchi (1999). The latter appears to be driven by the monsoonal heating rather than being the cause for it.

Nicholls (1983, 1995) proposed that SSTs over northwest Australia (5^oS-10^oS, 120^oE-160^oE) has a predictive value for ISM since warm SSTs during April there were correlated with stronger ISM in the following summer. In a related study, Sadhuram and Wells (1999) find a significant correlation between ISM and SSTs over 0^oN-5^oN, 80^oE-85^oE during November of the previous year. In summary, a synthesis of all the correlation studies is that the significant area of correlation between ISM and IO SST is observed only over small regions and depends very much on the selective filtering applied to the SSTs.

It is now well accepted that the ISM is a large-scale phenomena. The questions of how the SSTs over such small regions would persist to influence seasonal rainfall over such a large region have not been fully explained. It has been suggested that part of the controversy may be related to the lack of sufficient accuracy in the observed SSTs (Shukla 1987, Rao and Goswami 1988, Terray 1994). Even though subjective filtering of observations seems to strengthen the correlations (Rao and Goswami 1988, Sadhuram and Wells 1999, Li et al. 2001) concrete mechanistic connections are still lacking. Two constraints, small magnitude in SST variance (Figure 1a), and small areas of significant correlations diminish the value of any IO SST parameter in predicting the ISM.

The IO zonal/dipole mode has been shown to influence the ISM (Behera et al. 1999, Ashok et al. 2001, Annamalai et al. 2003a). Figure 4 taken from Annamalai et al. (2003a), shows the composites of anomalous precipitation and surface winds during strong zonal/dipole mode events. The negative precipitation anomalies in the eastern equatorial IO are surrounded by positive precipitation anomalies in the equatorial western-central IO and along the entire monsoon trough, from the Indian sub-continent extending eastwards into the tropical northwest Pacific. The surface winds indicate that the convergence over the Bay of Bengal is due not only to the westerly anomalies but also due to southerly flow originating off Sumatra. The inference from the Figure 4b is that the north-south heating gradient (precipitation) over the eastern IO favors a local meridional circulation (Gill 1980, Slingo and Annamali 2000), which is crucial for transporting moisture towards the monsoon trough.

Another possible dynamical interpretation of how the zonal/dipole mode influence the ISM can be obtained from Figure 4c, which shows the composites of the anomalous rotational part of the winds at 1000hPa. The twin anticyclones straddling the eastern equatorial IO are Rossby-wave response to the heat sink there. The northern component enhances the mean monsoon flow over the entire Northern IO, and thereby influences the ISM (Annamalai and Liu, 2003).

2.2.2 Modeling Studies: Motivated by the observational studies mentioned in Section 2.2.1, Shukla (1975) carried out the first ever modeling study to understand the impact of Arabian Sea SST on the ISM. His atmospheric general circulation model (AGCM) results corroborated the observational findings in that when cold SST anomalies over the Arabian Sea were imposed, the model simulated rainfall over India and adjoining oceanic regions was significantly reduced. Results from a different AGCM with similar SST forcing produced conflicting response over the ISM region (Washington et al. 1977). The differences in the response primarily depend on the individual AGCM's capability in simulating the mean monsoon (Shukla 1984). Consistent with the results of Shukla (1975, 1984), in a recent modeling study, Arpe et al. (1998) re-emphasized the importance of Northern IO SST anomalies on the ISM.

Recently, some modeling studies focused on the impact of equatorial central-eastern IO SST anomalies on the ISM. In an idealized case, Chandrasekar and Kitoh (1999) imposed warm (cold) SST anomalies over the equatorial IO and their model simulated suppressed (enhanced) rainfall over the Indian sub-continent. In the absence of an El Niño during the summer of 2000, Krishnan et al. (2003) attributed the deficient rainfall over India in that year to the abnormal warm SST anomalies in the IO. From model simulations they suggested that the persistent warm SST anomalies favored anomalous convection over the equatorial IO that subsequently, resulted in an extended monsoon break over India. Both these studies indicated that SST anomalies in the equatorial IO modulate the local Hadley circulation, and thereby the monsoon precipitation over India.

Lau and Nath (2000) carried out a series of model experiments to assess the role of remote versus local SST on the ISM. In the run called TOGA (Tropical Ocean-Global Atmosphere), the AGCM was forced by ENSO SST variability while the run term TOGA-ML, the ENSO forcing from the Pacific is included with air-sea coupling in the India Ocean. In another run, GOGA (Global Ocean-Global Atmosphere), the SST variability over the global oceans is prescribed. The simulated precipitation and winds between TOGA-ML minus TOGA would provide information about air-sea coupling outside the tropical Pacific. In a similar way, the difference between GOGA and TOGA runs would indicate the role of SST outside the tropical Pacific. The difference charts in all the four panels of Figure 5 suggest enhanced southwesterly flow and precipitation over India. The inference is that the basin-wide air-sea interactions over the tropical Indian Ocean opposes the remote ENSO forcing on the ISM.

Schubert and Wu (2001) examined the predictability of the 1997 and 1998 ISM low-level winds. Using an AGCM, they made a 10-member ensemble simulation with prescribed SST anomalies over the tropical Indo-Pacific regions. Their model simulations suggested that the 1998 monsoon is considerably more predictable than the 1997 monsoon. They found that during May and June of 1998 the predictability of the low-level wind anomalies is largely associated with a local response to anomalously warm IO SSTs. If not in a statistically significant sense, case studies for selective years imply the importance of IO SST anomalies on the ISM.

Annamalai and Liu (2003) carried out a series of AGCM experiments to isolate the role of Pacific and Indian Ocean SST anomalies on the ISM. Their results are broadly consistent with Lau and Nath (2000) in that the remote forcing from the Pacific dominates the ISM variability while the local forcing from the Indian Ocean adds to the details in the ISM response. The authors note that prescribing SST anomalies associated with the Indian Ocean zonal/dipole mode events alone do not significantly influence the precipitation over the Indian subcontinent.

A direct hypothesis for the regulation of monsoon involves the Ekman transport of heat across the equator into and out of the Northern IO (Webster et al. 2001) that would put the IO in a regulating role for regional climate variability with potential feedbacks to ENSO via monsoon variability. Building on the bienniel SST regulation mechanism for the IO proposed by Loschnigg and Webster (2000), Webster et al. (2003) argue that the monsoon is a self-regulated system where stronger ISM leads to enhanced draining of the heat content from the northern IO across the equator leading to a cooling of SST, hence a reduced meridional SST gradient and a weakening of the ISM in the following year. The reverse chain would work for a weaker ISM.

2.3 Indian Ocean and the African Monsoon

2.3.1 Observational Studies: The African monsoon consists of two subcomponents, the East African monsoon and South African monsoon. There is increasing evidence that different regional SST anomalies within the tropical IO influence each of the subcomponents.

2.3.1.1 Indian Ocean and East Africa: The seasonally reversing Asian monsoon circulation (NEM and SWM) also affects large parts of East Africa producing two seasons of rainfall during the transition from SWM to NEM and NEM to SWM, namely, during boreal spring and fall seasons (Ogallo 1988, Farmer 1988, Hutchinson 1992, Nicholson and Kim 1997). The rainfall during spring (April-May) is abundant, while it is more variable during fall (October-November, (Hastenrath et al. 1993, Figure 7a). As for the Asian monsoons, majority of the studies exploring the variability of the East African rainfall have looked to ENSO as a remote driver (Farmer 1988, Ogallo 1988, Hutchinson 1992, Hastenrath et al. 1993, Nicholson and Kim 1997). But

Hastenrath et al. (1993) pointed out that the mechanisms for the precipitation anomalies in the two rainy seasons are distinct with only the October-November rainfall anomalies showing a strong correlation with the Southern Oscillation. They also pointed to the role of the SST anomalies over western IO in suppressing convection during fall, particularly during the high phase of the Southern Oscillation (La Niña) and affecting East African rainfall (Figure 6). The SST anomalies themselves, however, appear to be driven by wind changes associated with ENSO that is consistent with Nicholson (1997), Nicholson and Kim (1997), and the recent findings of Allan et al. (2003) and Annamalai et al. (2003a).

Birkett et al. (1999) suggested that the abnormal East African rainfall during fall 1997 is due to IO SST. Black et al. (2003) examined in detail the role of IO SST on the AFRM system during boreal fall. The standard deviation of rainfall over the equatorial region is twice as strong compared to the region slightly south of it (Figure 7). The SST anomalies associated with heavy rainfall for both regions are shown in Figure 8 (from Black et al. 2003). It is clear that high rainfall is associated with a warming of the western Indian Ocean near the East African coast, and another robust signal is the cooling off the Sumatran and Australian coast. In addition, there is a pronounced warming in the equatorial Pacific. Of all the features, the SST anomalies off Sumatra show high statistical significance (right panels in Figure 8). Yet, the rainfall during 1994, one of the strongest IO dipole/zonal mode events, is not anomalous (Figure 7). Despite the lack of robustness, it should be noted that the average SST anomaly over the Pacific is strongly positive, which supports the hypothesis that El Nino conditions are associated with high rainfall in equatorial Africa. In a similar study, Clark et al. (2003) also note that correlations of East African rainfall with IO SSTs are not only significant over the western IO but also off Sumatra/Java in the eastern IO consistent with the Indian Ocean zonal or the dipole mode. Like in Hastenrath et al. (1993), Clarke et al. (2003) also note high simultaneous correlation between boreal fall East African rainfall and ENSO indices. However, the SST-rainfall relationship reversed during 1983-1993 for some unexplained reasons. Although not reviewed here in detail, the interannual variability in the water discharge of the Senegal River, and the rainfall in the West African Sahel during July-August are correlated with SST anomalies in the Western tropical IO (Bhatt 1989). In addition, Bhatt (1989) found that the discharge of rivers in the Nile basin is highly correlated with the ISM.

2.3.1.2 Indian Ocean and Southern Africa: There is observational evidence that the South African rainfall variability is forced remotely by SST anomalies in the western tropical IO (Mason and Jury 1995, Jury 1996, Goddard and Graham 1999). The variability of summer rainfall over Zimbabwe is related to tropical IO SSTs (Makarau and Jury, 1997). Landman and Mason (1999) argue that the relation between IO SSTs and summer rainfall over South Africa has changed since the late 1970s, coincident with hypothesized changes in the ENSO characteristics (Trenberth and Hoar 1996, Balaji et al. 1999). Warmer SSTs in the western equatorial IO were associated with drier conditions over Southern Africa prior to the 1970s whereas they are associated with wetter conditions since the late 1970s even though the ENSO influence appear to remain the same (Landman and Mason 1999).

The seasonal mean rainfall can be significantly affected by the number tropical cyclone days. Jury et al. (1999) show that SSTs over central and southern IO have an impact on the tropical cyclone days in the SWIO and have a predictive value with a lead of a few months. Xie et al. (2001) showed a similar predictive value based on the Rossby waves associated with ENSO and also with the IO zonal/dipole mode and the associated SST anomalies.

Van Heerden et al. (1988) note that South African summer rainfall is impacted by ENSO with warm (cold) ENSO corresponding to a dry (wet) spell and the correlations have decadal and interdecadal variabilities. Southern IO also undergoes interannual to interdecadal variability (Allan et al. 1995, Reason et al. 1996a, 1996b, Reason 1999) but the influence of this SST variability on Southern African climate variability has not yet been investigated in detail (Jury et al. 1993, Mason 1995). Recently, Behera and Yamagata (2001) identified a dipole-like variability in SST over the subtropical Southern IO which is correlated with rainfall over Southern Africa. Hydrographic sections show that surface freshwater and heat fluxes in the Antarctic region are manifested as decadal changes at depths in the southern IO (Bindoff and McDougal 2000). How these subsurface changes may affect the surface conditions is also not fully understood and may in fact explain some of the model/data discrepancies in Reason et al. (1996a).

2.3.2 Modeling studies: Since the number of modeling studies on the African monsoon variability is meager we will combine the AGCM studies that address the East and South African monsoon subcomponents together in this subsection. Latif et al. (1999) used an AGCM to show that East African rainfall anomalies that led to severe flooding during December-January of 1998 were forced largely by the western IO warming. This conclusion was arrived at when El Nino-related SST anomalies in the equatorial Pacific did not directly drive the changes in the climate over Eastern Africa. Figure 9 taken from Latif et al. (1999) indicates the direct role of IO SST anomalies on the rainfall over South Africa. Although encouraging, to what degree the IO SST anomalies in 1997-98 were driven by ENSO vs. local air-sea interactions remains an open issue. Similar AGCM experiments by Goddard and Graham (1999) show that Southern African rainfall anomalies were also forced by the western IO SST anomaly. In a recent modeling study, the subtropical SST dipole identified by Behera and Yamagata (2001) has been found to influence the rainfall over Southern Africa.

2.4 Indian Ocean and the Australian monsoon

2.4.1 Observational studies: Australian monsoon is dominated by seasonal rainfall during austral winter with a well-known ENSO driven interannual variability (Quinn et al. 1978, McBride and Nicholls 1983, Nicholls 1988, Webster et al. 1998). While much of Australia tends to be drier during warm ENSO events, the correlation is not perfect similar to the case of the ISM (Nicholls 1989). Indonesian Seas are subject to the reversing seasonal monsoons albeit a much weaker seasonal contrast than the Indian subcontinent (Nicholls 1981, Hackert and Hastenrath 1986). The most dominant interannual signal over Indonesia is of course the ENSO related drought (Hendon 2003) that not only leads to severe agricultural devastation but also forest fires and haze (Nichol 1997, 1998).

Braak (1919) first suggested that high local atmospheric pressure over Indonesia in the first half of the year was an indicator of a drier second half of the year with a late onset of the wet season and vice verse. Nicholls (1981) found evidence to support this hypothesis based on a positive feedback in the air-sea interactions causing the initial SST anomalies to persist from austral winter into spring. Nicholls (1978) also pointed out that the air-sea interactions in the Indonesian Seas and Northern Australia are dependent on the seasonally reversing relation between SST and pressure anomalies because the background winds themselves change sign giving rise to the quasibiennial nature to regional anomalies.

Hackert and Hastenrath (1986) analyzed ship and land observations in the IO region to investigate the mechanisms of rainfall anomalies over Indonesia and arrived at the conclusion that the interannual anomalies are largely a modulation of the annual cycle. Wet (dry) years are thus associated with stronger (weaker) Northeast monsoon with coincident cloudiness and SST anomalies although the ocean-atmosphere interactions are distinct in the austral spring and fall seasons, which was noted to be crucial for reversing the SST anomalies between the two seasons. The analyses of Hackert and Hastenrath (1986) clearly captured the correlation between Java rainfall and eastern and central IO SSTs.

Nicholls (1989) investigated the role of Indo-Pacific SSTs on the Australian winter monsoon variability. From an EOF analysis, he identified two different spatial patterns of rainfall variability. The correlations between the PCs and Indo-Pacific SSTs revealed some interesting results. The dominant spatial pattern (EOF1) was significantly correlated with the SST difference between south-central IO (10^oS-20^oS, 80^o-90^oE) and Indonesian Sea (0-10^oS, 120^o-130^oE) while EOF2 was correlated with ENSO-related SST anomalies in the equatorial Pacific. He applied partial correlation analysis that removes the effects of ENSO. Again, EOF1 was highly correlated with SST in the south-central IO and Indonesian Seas (Figure 10). Based on this statistical analysis Nicholls (1989) concluded that the SST anomalies in the south-central IO may be independent of ENSO, and has a strong predictive skill for the winter monsoon over Australia. Even though causality was not evident, it was clear that the IO SSTs were important for Australian winter rainfall.

The central IO and Indonesian SST gradient could also be inferred from the analyses of Reverdin et al. (1986) in the context of anomalous events in the IO, especially 1961. Hendon (2003) relates the SST zonal gradient in the IO to the Indonesian rainfall variability during the dry season centered on August. He concludes that the anomalous gradient in the IO is largely controlled by seasonally varying air-sea interactions in the eastern IO and is strongly related to ENSO. The role of ocean dynamics and air-sea interactions in SST variability of the southeastern

IO has since been reported in several studies (Murtugudde and Busalacchi 1999, Behera et al. 1999, Vinayachandra et al. 2000, Murtugudde et al. 2000, Ansell et al. 2000). Probably for the first time, Murtugudde et al. 1998, and Murtugudde and Busalacchi (1999) reported that this was one of the regions in the IO where Bjerknes (1969) feedbacks occur due to interactions between the thermocline and SSTs. We will return to this issue in Section 3.

2.4.2 Modeling studies: Motivated by the correlation statistics, many AGCM studies have investigated the impact of Pacific and IO SST anomalies on the Australian rainfall variability (e.g., Voice and Hunt 1984, Simmonds and Smith 1986, Simmonds and Trigg 1988, Simmonds et al. 1989, Simmonds 1990, Simmonds and Rocha 1991, Frederiksen and Balgovind 1994, Frederiksen and Frederiksen 1996, Frederiksen et al. 1999). Most of these studies confirm the role of the SST gradient between the Indonesian archipelago and the south-central IO on the winter rainfall variability.

2.5 Indian Ocean SST, Monsoons and the TBO

The processes within the tropical IO have been proposed to play an "active" role in the life cycle of the TBO, one of the major climate signals in the tropics (e.g., Meehl 1987, 1993, Li et al. 2001, Meehl et al. 2003). A dry ISM is followed by a spatially large warm SST anomaly over the tropical IO that is observed to persist up to the following monsoon (Joseph and Pillai 1984). Such a large warm SST anomaly is found to result in a wet monsoon in the following year producing a biennial type oscillation (Joseph 1981, Joseph and Pillai 1986). In the context of the Asian-Australian monsoon, the TBO is defined as the tendency for a relatively strong monsoon to be followed by a relatively weak one, and vice-versa (Meehl and Arblaster 2002). Therefore, it has been suggested that the TBO is not so much an oscillation, but a tendency for the system to flip-flop or transition from year to year. There is compelling observational evidence that a strong

ISM is followed by a strong AUSM, and the strong AUSM is followed by a weak ISM (e.g., Meehl 1987, Chang and Li 2000, Yu et al. 2003). An analysis of observed data shows that the TBO encompasses most ENSO years (with their well-known biennial tendency) as well as additional years that contribute to biennial transitions (Meehl and Arblaster 2002). Thus it has been hypothesized that the TBO is a fundamental feature of the coupled climate system over the entire Indian-Pacific regions (see Li et al. 2001 for a review on TBO theories).

Figure 11 taken from Yu et al. (2003) shows the lagged correlation coefficients calculated between monthly ISM and AUSM rainfall anomalies. Two large correlation coefficients, positive one (in-phase) with ISM leading the AUSM by about 2 seasons, and negative one (out-of-phase) with the AUSM leading the ISM by about 2 seasons, are readily apparent (Fig.11). Yu et al. (2003) examined the role of Indo-Pacific oceans for the TBO from a series of coupled model integrations. The coupled model run, when both the Indian and Pacific Oceans are interactive produces both the transitions (Figure 11b). The run, only with the Pacific (Indian) produces the in-phase (out-of-phase, Fig. 11d) transition (Fig. 11c). These model simulations, in agreement with observations (e.g., Meehl and Arblaster 2002) suggest that IO plays an "active" role in the life cycle of the TBO. Meehl and Arblaster (2002), and Loschnigg et al. (2003) argue that the IO zonal/dipole mode events are large amplitude excursions of the TBO in the tropical IO. If the hypothesis for TBO proposed by Meehl et al. (2003) and related works are accurate, then it is not surprising that the Indian Ocean affects the African and Australian monsoons through its impact on the Walker Cells.

2.6 Indian Ocean on Intraseasonal Variability

Both during summer and winter seasons, convection associated with the intraseasonal oscillations (ISOs), typically with a time scale of 30-50 days, originate over the equatorial Indian

Ocean (e.g., Madden and Julian 1971, 1972, Yasunari 1979, Sikka and Gadgil 1980, Lau and Chan 1986, Annamalai and Slingo 2001). Although the ISOs are inherent to the atmosphere, both observational and coupled modeling studies clearly demonstrate the crucial role played by IO SST in the organization, intensification, and propagation of the convection and circulation associated with the ISOs (e.g., Krishnamurti et al. 1988, Li and Wang 1994, Hendon and Glick 1997, Flatau et al. 1997, Shinoda et al. 1998, Waliser et al. 1999, Jones et al. 1999, Woolnough et al. 2000, Sengupta and Ravichandran 2001, Sengupta et al. 2002, Kemball-Cook et al. 2002, Fu et al. 2003, Waliser et al. 2003a,b).

SST observations from moored buoys in the North Bay of Bengal during July-August of 1998 indicate that the amplitude variation of SST at ISO time scale could be as large as 2^oC (Sengupta et al. 2001, Bhat et al. 2001). From satellite derived SST data, Vecchi and Harisson (2002) also noted similar magnitudes in SST fluctuations at ISO time scales. BOBMEX, the Bay of Bengal Monsoon Experiment was carried out under the auspices of the Indian Climate Research Program during the boreal summers of 1999-2001 (Bhat et al. 2001). The SST observations from BOBMEX also indicate the large amplitude variations in SST at ISO time scales over the North Bay of Bengal, and Bhat et al. (2001) suggest that the SST variations have direct impact on the genesis of monsoon depressions.

The phase of the ISOs is linked to the onset, and active/break phases of the ISM (e.g., Joseph et al. 1994, Sikka and Gadgil 1980), and the AUSM (Hendon and Liebmann 1990a, b). The large spatial scale in convection and circulation associated with the ISOs modulate the mean monsoon circulation (e.g., Krishnamurthy and Shukla 2000, Sperber et al. 2000, Goswami and Ajayamohan 2001) and also the genesis of synoptic systems (e.g., Liebmann et al. 1994). Therefore, it has been hypothesized that the statistical properties of the ISOs can modulate the interannual variability of the monsoons (e.g., Palmer 1994; Sperber et al. 2000, Krishnamurthy and Shukla 2000, Goswami and Ajayamohan 2001). Therefore, a better representation of the ISOs in the models may be pre-requisite for addressing the interannual variability.

The AGCMs have limitations in the simulation of ISOs (e.g., Slingo et al. 1996). Motivated by the observational evidence, some recent studies indicate that an AGCM is coupled to either a simple or a comprehensive ocean model has a better representation of the ISOs (e.g., Waliser et al. 1999, Woolnough et al. 2001, Kemball-Cook et al. 2003). Figure 12 taken from Fu et al. (2003) suggests that at ISO time scales, SST from the coupled model leads the convection by about 10-days. This lead-lag relationship between SST and convection at ISO time scale appears to be crucial for the organization, and propagation of ISOs (e.g., Fu et al. 2003).

3. Role of the Indian Ocean in the Dipole/Zonal Mode Events

Even though the IO zonal/dipole mode events with significant climatic impacts occurred only during 1961, 1994, and 1997, they produce the largest SST anomalies in the IO, and clearly involve an active role by the ocean. We focus on some details of the IO zonal/dipole mode events in this section to highlight the renewed interest in the IO. Much research has been devoted to different aspects of the IO sector climate variability including ocean dynamics and thermodynamics in the aftermath of the anomalous IO dipole/zonal mode (Behera et al. 1999, Chambers et al. 1999, Saji et al. 1999, Webster et al. 1999, Yu and Reinecker 1999, 2000, Vinayachandran et al. 2000, Murtugudde et al. 2000). A number of recent studies have also addressed the ocean variability associated with the IO dipole/zonal mode (Murtugudde et al. 2000, Rao et al. 2002, Huang and Kinter 2002, Krishnamurthy and Kirtman 2003, Vinayachandra et al. 2003, Feng and Meyers 2003). Several ocean-atmosphere coupled general circulation models have, with a varying degree of success, simulated the IO dipole/zonal mode (e.g., Gualdi et al. 2003, Lau and Nath, 2003, Cai et al. 2003, Loschnigg et al. 2003, H. Spencer 2003, personal communication). The diagnoses of the IO mode have led to extensive debate about its relation or lack thereof to ENSO phenomenon in the Pacific (Iizuka et al. 2000, Allan et al. 2001, Baquero-Bernal et al. 2002, Hastenrath 2002, Yamagata et al. 2003, Annamalai et al. 2003a, Saji and Yamagata 2003). It is evident that this is a debate that will continue to generate new hypotheses for the processes involved in the IO climate variability and its impact on regional and global climate. Since the focus of this review is to synthesize the role of the IO in regional climate variability, we will not delve into the ongoing debate here.

Rao et al. (2002) and Feng and Meyers (2003) argue based on model simulations that the heat content and sea level variability in tropical IO is predominantly quasi-biennial with a delayed oscillator like process driven by equatorial waves providing the phase change mechanism. Huang and Kinter (2002) and Krishnamurthy and Kirtman (2003) on the other hand argue, also based on model simulations, that heat content variability in the tropical IO has a significant correlation with ENSO variability. Saji et al. (1999) and Saji and Yamagata (2003) find that some IO dipole/zonal mode events do occur when the Pacific is not undergoing a warm ENSO event with some events even occurring during a cold ENSO event. Here, we base our analyses on years of significant cooling in the eastern equatorial IO as defined by Annamalai et al. (2003a) which is also consistent with other studies such as Hendon (2003). In addition, Black et al. (2003) note that the cooling off Sumatra is a robust signal in the relationship between East African rainfall and IO SST. The results are not greatly altered by using the zonal gradient (Murtugudde et al. 1998, Saji et al. 1999). We simply seek oceanic conditions that accompany the anomalous events in the IO that point to the role of the IO in the evolution of these events.

As discussed in the Introduction, climatological conditions in the equatorial IO are quite unlike the Pacific and Atlantic Oceans with no significant equatorial upwelling except the seasonal upwelling in the west. The eastern equatorial IO is somewhat similar to the western Pacific warm pool in terms of annual mean SSTs that remain at or above 28.5^{o}C (Fig. 2) with a barrier layer that decouples the SSTs from thermocline variability. Seasonal upwelling off the coast of Java extends northwestward along the coast of Sumatra in response to the atmospheric Rossby wave associated with the heat source of the ITCZ (Annamalai et al. 2003a, Wang et al. 2003). However, during certain years, the alongshore winds and the coastal upwelling achieve a trade wind-cold tongue like structure as noted in the literature cited above. We use a model simulation to exemplify the various stages of evolution of the oceanic conditions.

Annamalai et al. (2003a) used an index of SST anomalies in the eastern IO, which cools significantly during every dipole/zonal mode year and categorized the interannual events into strong, weak, and aborted zonal mode years. Figure 13, reproduced from Annamalai et al. (2003a) shows that during strong zonal mode years, the cooling in the eastern IO is initiated in the early part of the year and is greater than one standard deviation during the boreal fall months whereas during weak zonal mode years the eastern IO cools during boreal spring months and the cooling is significantly weaker albeit persistent through boreal fall months. There are years when anomalous upwelling occurs off Java in terms of stronger than seasonal upwelling during boreal spring months but the SSTs return to normal or even warmer than normal by boreal summer/fall seasons. These years can get categorized as dipole mode years based on the gradient even though climatically they may have very little impact. Here we categorize the dipole/zonal mode years as consisting of several phases of evolution, namely, preconditioning (Jan-Mar), onset (Apr-May), growth (Jun-Aug), mature (Sep-Nov), and decay (Dec-Feb) phases with the decay phase extending into the year following the dipole/zonal mode year. We present the composite anomalies for each phase from the strong years and analyze the weak and aborted years to delineate the differences that result in the events being weak or aborted.

3.1 Phases of evolution of Dipole/Zonal Mode Events

Wyrtki (1985) was the first to argue that a build up of the western Pacific warm pool occurs prior to the onset of El Niño events. Such a preconditioning is now observed with TAO data and also in the barrier layer thicknesses (McPhaden 1999, Maes et al. 2002). Our model results appear to indicate the thermocline and the barrier layers tend to be positive in the eastern IO during the beginning of dipole/zonal mode years. Altimeter data from TOPEX captured the variability during the 1994 and 1997 events which also shows positive sea level anomalies in the east during the beginning of both years (Fig. 14). As of this writing, the eastern IO also has undergone substantial cooling off-Java during the summer of 2003 and the TOPEX data shows that the sea levels were also positive during early 2003 although the coast of Java shows negative sea level anomalies and the Southern IO is dominated by a downwelling Rossby wave signature. It is thus likely that similar to the ENSO events, the IO zonal/dipole mode events may also be preconditioned prior to the onset of these events. However, like in the Pacific, not every preconditioned state in the early part of the year leads to a zonal/dipole mode. Annamalai et al. (2003a) demonstrated that the onset of the zonal/dipole modes occurs during boreal spring months off of Java/Sumatra. We thus consider the evolution of the zonal/dipole modes to consist of a preconditioning, onset, growth, mature, and decay phases.

3.1.1 Wind stress anomalies

Figure 15 shows the composite wind anomalies from NCEP reanalyses for strong zonal/dipole mode years which show that during the early part of the calendar year, the wind anomalies are actually downwelling favorable off Sumatra and Java but the wind stress curl offshore generates weak Ekman pumping and initiates the uplifting of the thermocline (Fig. 17) and erosion of the barrier layer (BL, Fig. 19). By the boreal spring, the onset phase is well established with easterly anomalies off Java and alongshore-upwelling favorable winds off Sumatra. The easterly wind anomalies along the equator in the eastern IO favor a weaker eastward Wyrtki Jet despite the semi-annual downwelling Kelvin wave generated in the central western IO. As noted by Yamagata et al. (1996) and Sprintall et al. (2000), the downwelling Kelvin wave deepens the thermocline and generates a reduction in ITF that actually assists the subsequent rebound of the thermocline and the susceptibility of the SSTs to entrainment cooling (Murtugudde et al. 1998, Annamalai et al. 2003a). The trade wind like structure is fully in place by the growth phase and strengthens significantly during the mature phase before weakening rapidly as the decay phase sets in by the end of the calendar year. The negative wind-stress curl in the central-western part of the Southern IO is clearly seen in nearly all phases which leads to the Rossby wave mechanism that extends the warming in the west (Murtugudde et al. 2000, Xie et al. 2002, Vinayachandra et al. 2003).

The weak zonal/dipole mode years are characterized by considerably weaker upwelling favorable winds in the coastal regions during the onset phase (Fig. 16). In fact, the growth phase has stronger alongshore winds than the mature phase leading to much weaker coupled Bjerknes feedbacks during the summer and fall months. The aborted years (Fig. 16) have weak alongshore winds off Java during the onset phase and the winds reverse completely during the growth phase. This deepens the thermocline off Sumatra and Java and the reponse to weak upwelling alongshore winds during the mature phase are simply unable to generate any cooling let alone

SST-thermocline feedbacks. There is no well-organized wind-stress curl in the Southern IO for weak and aborted events.

3.1.2 Thermocline anomalies

The thermocline anomalies for strong zonal/dipole mode years clearly show the upwelling Rossby wave signatures in the southern IO in the preconditioning phase (Fig. 17) as was noted by Vinayachandran et al. (2003) for late 1993 and 1996 prior to the 1994 and 1997 events. The thermocline anomalies near Java are slightly positive corresponding to the wind stress curl anomalies in Fig. 15. The onset phase is characterized by negative thermocline anomalies along the entire coast of Java/Sumatra that also extend into the Bay of Bengal as Kelvin wave signatures (Potemra et al. 1991, Yu et al. 1991). The growth phase has significantly negative thermocline anomalies off Java that spread northwestward as upwelling favorable winds along the coasts strengthen. Note that the mature phase has reduced thermocline anomalies in the far eastern equatorial IO that become positive as the decay phase commences. The weak and aborted events have no significant shallowing of the thermocline during the onset and growth phases (Fig. 18). The weak events have some negative thermocline anomalies during the mature phase and they are not very strong on the equator indicating that coastal upwelling and offshore advection are relatively weak. As noted earlier, the strong downwelling Rossby wave in the southern IO seen during strong events is hypothesized to play a dominant role in the western IO warming (Webster et al. 1999, Murtugudde et al. 2000, Vinayachandra et al. 2003) and also hypothesized to provide predictability for some aspects of the East African climate variability (Xie et al. 2002). Note that Rossby wave signature is barely present during the mature phase of the weak and aborted years (Fig. 18). The thermocline depth anomalies are in fact positive off the coast of Java during the mature phase of an aborted event.

3.1.3 Barrier layer anomalies

The BL anomalies during the preconditioning phase of the strong years are slightly positive in the eastern IO (Fig. 19). This is akin to the deeper BL and sea levels in the far western Pacific prior to the onset of warm ENSO events in the Pacific (McPhaden 1998, Maes 2001). By the onset phase, positive BL anomalies begin to disappear in the eastern IO as negative precipitation anomalies begin to set in over Java/Sumatra. By the growth phase, the BL anomalies are significantly negative and they decrease further into the mature phase and do not disappear even as the decay phase sets in. Weak zonal mode years have some negative BL anomalies in the mature phase whereas aborted events have slightly positive BL anomalies (Fig. 20).

3.1.4 Sub-surface temperature anomalies

The role of the IO in the evolution of the zonal/dipole mode events is made clearer by examining the subsurface temperature anomalies. The first point to note is that the largest subsurface temperature anomalies occur south of the equator unlike ENSO events where the largest subsurface temperature anomalies tend to be centered on the equator (McPhaden 1998, Picaut et al. 2002). The temperature anomalies along 5^{o}S for composite strong events shown in Fig. 21 correspond to the positive thermocline depth anomalies during the preconditioning phase in the far-east. The effects of the upwelling favorable curl generates mild near surface cooling away from the coast. By the onset phase, the subsurface warm anomalies have nearly dissipated away with cold subsurface anomalies appearing at the eastern coast. The cold anomalies in the east begin to increase and spread to the surface and downward by the growth phase with positive subsurface anomalies first occurring at 60^{o}E where the thermocline doming provides the largest interannual variability in the IO during non-zonal/dipole mode years. This is the region that is most sensitive to any changes in wind-stress curl in the Southern IO. By the mature phase, entire central-western IO along 5^{o}S has warm anomalies at the thermocline level with the negative anomalies in the east at their coldest. The eastward spreading of the warm anomalies during the decay phase is similar to the eastward propagation of subsurface anomalies during warm ENSO events in the Pacific (Picaut et al. 2002).

The subsurface conditions during weak and aborted events are dramatically different with no systematic east-west contrast during any time of the year (Fig. 22). The subsurface cooling for weak events is only modest during the mature phase with no indication of a Bjerknes feedback in terms of thermocline-SST interactions. The aborted events are in fact dominated by warm subsurface anomalies in the east during the entire year pointing to eastward spreading of western upwelling during the preconditioning phase with no notable shallowing of the thermocline near the coast at all. These are the events where any surface temperature anomalies are mostly driven by surface fluxes rather than ocean dynamics (Latif et al. 1994, 2000). The role of the ocean dynamics and thermodynamics is thus clearly evident during strong zonal/dipole mode years whereas no significant role for the IO is evident during weak or aborted years.

3.2 Decadal Phasing of the Zonal Mode Events

It has been noted by Slingo et al. (1999) and Webster et al. (1999) that the IO SST has trends and decadal variability (also see Saji and Yamagata 2003). The monsoon-ENSO correlations themselves undergo decadal and longer-term variations (Kumar et al. 1999a, b). Annamalai et al. (2003b) note that there is a decadal clustering of the zonal mode events that they propose is related to decadal phasing of thermocline depth in the eastern IO forced by Pacific Decadal Variability through an atmospheric bridge. We visit the decadal variability issue here briefly to highlight some tantalizing results and discuss the potential impacts and issues to be understood in the future.

The low-pass filtered variance of the ITF from the model results used in Murtugudde et al. (1998) and Annamalai et al. (2003b) is shown in Fig. 12 that clearly displays decadal and longer-term variability. The 1950s are marked by relatively higher variance with a notable shift to lower variance during 1958-59. An increase in variance is initiated in the mid-1970s which continues into the 1990s. Note that the warming trend in the IO SSTs reported by Slingo et al. (1999) and Webster et al. (1999) over this period has a similar character. The role of the ITF in the long-term heat balance of the IO needs to be investigated further. The potential impact of the long-term variability of the ITF can be expected to manifest itself in the heat content and thermocline variability also. Figure 24 shows that the first EOF of the thermocline anomalies in the IO display a low-frequency variability whereas the time-series associated with the first EOF of thermocline anomalies is predominantly interannual when the ITF is closed. It is the second EOF of the thermocline anomalies with the ITF that captures the interannual variability as seen in Fig. 25. The variance of ITF shown in Fig. 23 and the first EOF of the thermocline with the ITF open is remarkably similar to the decadal pattern observed in low-pass filtered Pacific basin precipitation and its regression against the 200-hPa stream function field shown by Higgins et al. (2000). It is thus likely that the decadal and longer-term variability in the Pacific is transmitted to the IO through the oceanic bridge provided by the ITF. It is also likely that the IO heat content variability and associated SST signatures can feedback to the Pacific decadal variability as is evident in some of the correlations seen between IO and the North Pacific (Deser et al. 2003). The interdecadal changes in the Southern and Northern Pacific correspond to decadal and longer time-scale changes in the gyres (Latif and Barnett 1994, Wajsowicz 1994, Geise et al. 2002, Zhang et al. 2003) and also correspond to the changes in the relative contributions from the North and South Pacific to the ITF. Thus the decadal variability in the IO may be intimately tied to the Pacific decadal variability through the ITF with a potential to play a role in the secular trends of regional SSTs. Since the IO gains heat in the annual mean at the surface and there is an annual mean heat input into the IO by the ITF (Godfrey 1996), the meridional flushing of this heat may occur at lower frequencies. We raise these issues here simply to stress on the need for further data analyses and model synthesis activities to understand the role of IO in regional climate variability at decadal and longer time-scales.

4. Outstanding Issues

We have attempted to provide a review of published literature on the role of the Indian Ocean in the regional climate variability and some model simulations on oceanic conditions associated with the different phases of the strong, weak, and aborted IO dipole/zonal mode events. Numerous excellent studies have been reported on all aspects of the IO variability from diurnal to decadal time-scales though not necessarily seeking to understand the role of the IO in climate variability. Some of the earliest oceanographic expeditions took place in the 1970s (Wyrtki 1971) that were followed by coupled ocean atmosphere experiments such as MONSOON-77 and MONEX-79 (Krishnamurti 1985, Rao 1987). Much of the 1980s were dominated by extensive investments in the tropical Pacific to understand and forecast the ENSO phenomenon (for e.g., McPhaden et al. 1998). One of the achievements under the auspices of TOGA includes the successful ENSO forecasts with simple coupled models, which led Cane (1991) to claim that ENSO was a simple, robust and large-scale phenomenon. However, no such claim can be made about any aspect of the IO sector climate variability, especially the monsoons. Webster and Yang (1992) argued strongly that it is fallacious to assume that ENSO and monsoon are two separate climate processes. The two-way interactions between ENSO and monsoon and their net impact on IO processes are yet to be fully understood. The impact on IO SST on the Pacific climate needs to be studied in detail. Although not reviewed here, recent studies point to the possible influence of IO SST on west Pacific convection (Watanabe and Jin 2002, 2003) and on northern hemisphere extra-tropics (Farra et al. 2000, Deser et al. 2003). More observational and modeling efforts are needed to extend and validate these results.

Webster and Yang (1992) and Webster et al. (2002) also make a strong case that it is absolutely essential to simulate the annual cycle accurately in order to be able to capture the interannual variability of large-scale systems such as ENSO and monsoons. Based on Slingo et al. (1996), Ferranti et al. (1997), Sperber et al. (2000), and Krishnamurthy and Shukla (2000), it is further noted by Webster et al. (2002) that the intraseasonal variability in the monsoon regions need to be faithfully represented before the annual cycle can be accurately simulated. As alluded to earlier with regards to the monsoon correlations to IO SSTs, the greatest challenge in the IO sector is to be able to capture the anomalous variability of SSTs which are at present in the range of observational errors at all time-scales. This can be placed in the context of recent experiments in the IO region and what has been learned from them.

The summer monsoon rainfall over the Indian subcontinent consists of westward and northward propagating synoptic disturbances. It was learned from MONTBLEX-90 that warm SSTs and high heat contents in the Bay of Bengal provided regions of genesis for the synoptic-scale convective events (Rao 1987). The SSTs were cooled by 0.2^oC to 3^oC, depending on the strength of these events. The SSTs recovered within about a week after the systems moved away. The need for new data could easily be justified leading to more recent observational studies such as BOBMEX and JASMINE (Bhat et al. 2001, Webster et al. 2002). The new experiment called ARMEX (Arabian Sea Monsoon Experiment) will shed insight into the air-sea processes that control the precipitation maximum off the west coast of India. Understanding of the coupled system at intraseasonal time-scales gained from such meticulous observational efforts has been invaluable and as pointed out by Webster et al. (2002) the knowledge will likely be transferred to improve monsoon forecasts. Contributions from other multinational (for e.g. WOCE, see Ffield 1997) and national efforts (for e.g., Indian/German current meters, Japanese Triton buoys, see http://www.incois.gov.in/Incois/iogoos/home.jsp) and volunteer observing ships have systematically enhanced hydrographic and dynamic database of the IO region. While the growing in situ and remotely sensed observations of the IO continue to shed more light on the role of the IO in the coupled climate system, the need for a more coordinated and sustained observing system in the IO can not be overemphasized (Godfrey et al. 1995, Meyers et al. 2000). Such efforts are underway and should lead to vast improvements in coordinated and sustained data gathering efforts in the coming years (for e.g., Indian Ocean-GOOS (Global Ocean Observing System); http://www.clivar.org).

Continued modeling and synthesis activities such as CLIVAR-Monsoon Intercomparison Project (Kang et al. 2002) are needed if the potential role of the IO in monsoon forecasting at intraseasonal to interannual time-scales are to be understood. Such inter-comparison studies should involve all components of the monsoons including African and Australian monsoons. Ocean modelers lag far behind the atmospheric modelers in intercomparison studies that have proven to be instrumental in focusing community efforts in improving AGCMs (e.g., Slingo et al. 1996, Sperber and Palmer 1996, Gadgil and Sajani 1999). The need for multiple surface fluxes to force ocean models is a great inhibitor of ocean model intercomparison studies but time has come to address such issues head on if real progress is to be made in understanding the role of the IO in regional climate variability. As discussed in Section 2, numerous observational and modeling evidences exist pointing to role of heat content variability in the IO to regional climate variability including the TBO which appears to be a fundamental mode of interannual variability in the Indo-Pacific domain. However, no systematic studies exist to consistently connect heat content anomalies to SST anomalies so that the causal linkages can be made to quantify the role of IO in the coupled climate system. As pointed out by Webster et al. (2002) and Waliser et al. 2003, heat transports and currents display significant intraseasonal variability and yet no direct impact on SSTs has been quantified other than inferential connections.

Traditionally, the ocean's role is invoked as a memory of the coupled climate system at interannual and lower frequencies. If the IO plays a significant role in the intraseasonal variability as evidence seems to suggest, then this clearly adds great difficulty to the task of intraseasonal prediction and the observational need for predictive understanding. The ocean modeling requirements in terms of horizontal and vertical resolutions have not been investigated in a climate forecasting context. Such an effort also has to consider the issues of coupling an appropriate resolution ocean model to the AGCMs and related data assimilation issues for climate forecasting. It is likely that assimilating data to improve model simulations of intraseasonal variability will be fraught with problems similar to the assimilation of tropical instability wave like phenomenon in the Pacific Ocean. Even if the modeling exercises are split into an engineering approach to forecasting (optimum initial conditions through data assimilation and short term integrations for forecasting) and process oriented model improvement activities, each task will demand enormous computational capabilities.

The Arabian Sea is noted for the diurnal variability of the mixed layer which determines the seasonal variability of biomass as shown by model studies (McCreary et al. 2001, Wiggert et al. 2002). And yet, no such studies have been carried out to understand the impact of the diurnal cycle on SST variability in the Arabian Sea. The dynamics and thermodynamics of the Bay of

Bengal is only beginning to be understood at intraseasonal time-scales. It is clear that this region is remarkably different in terms of surface fluxes and oceanic structure and response from the western Pacific warm pool region despite many similarities. It is not obvious if what was learned during TOGA-COARE can be applied to the Bay of Bengal in terms of surface fluxes and convective parameterizations. The role of the annual cycle of wind stress curl and thermocline doming in the Southern tropical IO has been noted repeatedly as discussed in the previous sections. Despite the largest interannual variability of SSTs in this region and the variability of the ITCZ in the Southern IO, detailed observational and modeling studies are lacking to understand the SST, and heat content variability at intraseasonal time-scales in this band. The modulation of the variability in this band is most likely related to the ITF at annual, interannual, and longer time-scales that also have not been studied in detail.

The recent attention to the IO based on dipole/zonal mode variability has led to numerous modeling studies. However, even when the coupled models generate a dipole/zonal mode like variability, they are not always accompanied by realistic monsoon and/or ENSO variability, especially the nonstationary monsoon/ENSO interactions. This casts doubts on the ability of the present models to capture the details of the coupled ocean-atmosphere processes realistically in the IO region. It behooves the modeling and observational community to understand the relation between the TBO, zonal/dipole mode events, ENSO, and monsoons apart from the role of the intraseasonal variability in each mode of variability and their interactions. Such efforts are already underway with simple models and large scale coupled GCMs. This is indeed a monumental task and will probably require far more modeling exercises with simple process oriented models to regional models to coupled ocean-atmosphere models focused on capturing the annual cycle of the monsoons accurately including their intraseasonal variability. While designing coordinated observational experiments and sustained observational systems is important, transferring the knowledge from such data gathering efforts to improving processes in the hierarchy of models is absolutely essential if ultimate improvements in forecasting at intraseasonal to interannual time scale are to be achieved. Such modeling/synthesis activities should in fact be initiated immediately based on data gathered during field experiments such as BOBMEX, JASMINE and ARMEX in addition to other mooring data.

Stand-alone ocean models can address several issues such as the role of the ITF in the variability of the IO at all time-scales. The Southern Indian Ocean has a gyre structure that is unique compared to the Atlantic and Pacific southern gyres since the South Equatorial Current in the Indian Ocean is never a part of the equatorial current system which is mainly eastward. The injection of the ITF occurs into this gyre and may modulate the low-frequency variability of the gyres. With the availability of atmospheric reanalyses and multidecadal flux products, it has become routine to carry out long simulations of the IO. At this juncture, it is important to analyze the ocean model simulations carefully to determine if the decadal and longer time-scale variability in the IO is crucially dependent on the Pacific through the ITF and if so, what the best way is to represent the impact of the ITF on the IO. Similar issues may exist as far as any long term trends in the freshwater input in the Bay of Bengal which have not even been thought of so far. Explicit representation of river inputs into large-scale ocean models remains somewhat of an art thus far and needs to be revisited with a focus to understand its impacts on regional dynamics and thermodynamics.

The variability of the Southern IO gyre has not been studied much compared to the gyres in the Pacific and Atlantic Oceans. The mechanisms of transporting waters across the equator are only beginning to be addressed in models but may have a direct bearing on the TBO in the region. Ocean models can also quantify the oceanic mass and heat budgets associated with strong and weak monsoons to explain the feedbacks that occur in the real world and if similar feedbacks occur in the coupled models (systematic biases in continental vs. oceanic ITCZ seen in AMIP simulations). The failure of the models to produce any of the observed monsoon-SST correlations in the CGCMs has to be understood in terms of the ocean's response to errors in surface heat fluxes and wind stresses. Fate of the freshwater input into the Bay of Bengal and its impact on the upper ocean hydrological cycle has to be revisited especially in terms of its impact on air-sea interactions at intraseasonal time-scales.

Note that many of the Asian-Australian monsoon issues and the need for diagnostic studies and data mining efforts in the context of climate variability and predictability are discussed in related CLIVAR documents (http://essic.clivar.org) and a research prospectus for U.S. CLIVAR (http://www.usclivar.org).

5. Concluding Remarks

A fairly comprehensive review of the published literature on the role of the IO in regional climate variability has been attempted. It is clear that despite a lack of coordinated observational and modeling attempts to quantify the role of the IO in the coupled climate variability, the predictive understanding has made considerable progress in the past decade even if as a result of the recognition of monsoons as an integral part of the ENSO cycle. Despite very focused efforts towards ENSO prediction over the last decade or so, many issues remain in the basic understanding and predictive understanding of the ENSO cycle. In contrast, monsoon prediction attempts can be traced back to more than a century and yet the dynamic prediction of the monsoon regions is in various stages of progress with much of the gain having been accomplished as an ENSO teleconnected climate process. It has become more and more evident that the earlier conjectures on the role of the intraseasonal variability in determining the seasonal mean monsoons have been confirmed by more recent observational and modeling activities.

Despite the enormity of the task of accurately modeling and predicting intraseasonal to interannual variability in the IO region as a coupled system, the convergence of ideas on observational and modeling requirements for achieving such a goal provides hope for future success. With observational systems such as Argot and Atlas/Triton buoys achieving their critical mass in the IO, the knowledge of IO variability and its role in regional climate variability will improve greatly. The need for understanding the intrinsic variability as opposed to ENSO response of the IO will lead to significant progress towards monsoon forecasts especially in terms of improving forecasts at intraseasonal time-scales. With the growth of the observational database in the coming years, it is hoped that the IO variability will improve at all spatial and time-scales. We look forward to the future with great expectations as far increased understanding of the IO sector climate variability and the role of the IO in this coupled system.

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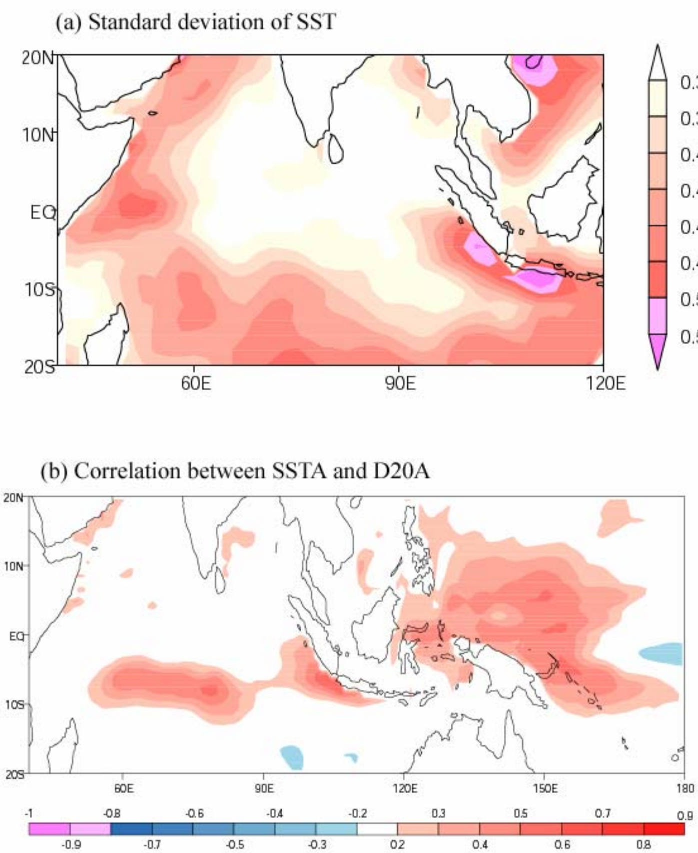


Figure 1: (a) Standard deviation of interannual SST, and (b) simultaneous correlation between anomalous SST and thermocline as measured by the depth of the 20°C isotherm.

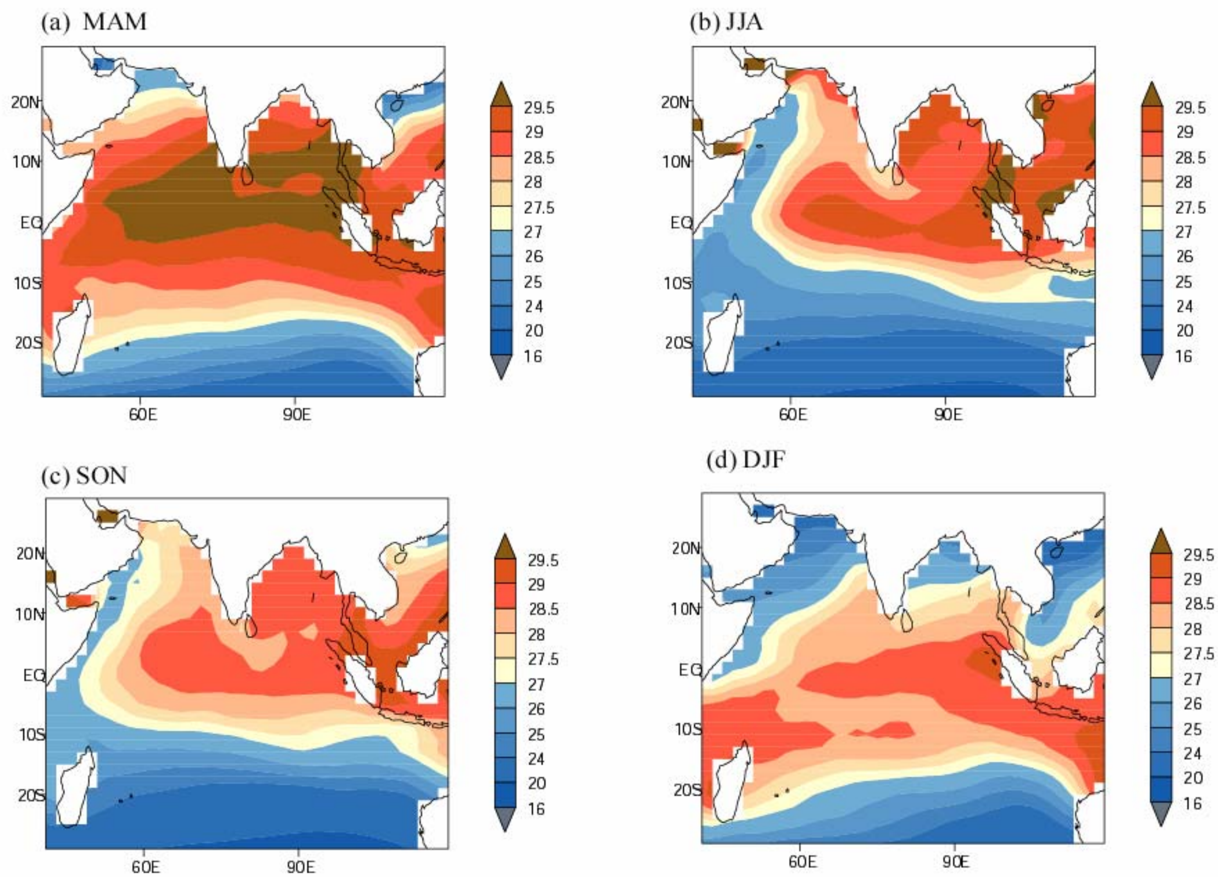


Figure 2: Seasonal SST climatology over the tropical Indian Ocean, (a) March-May, (b) June-August, (c) September-November and (d) December-February. The gure illustrates the seasonal c ycle of SST.

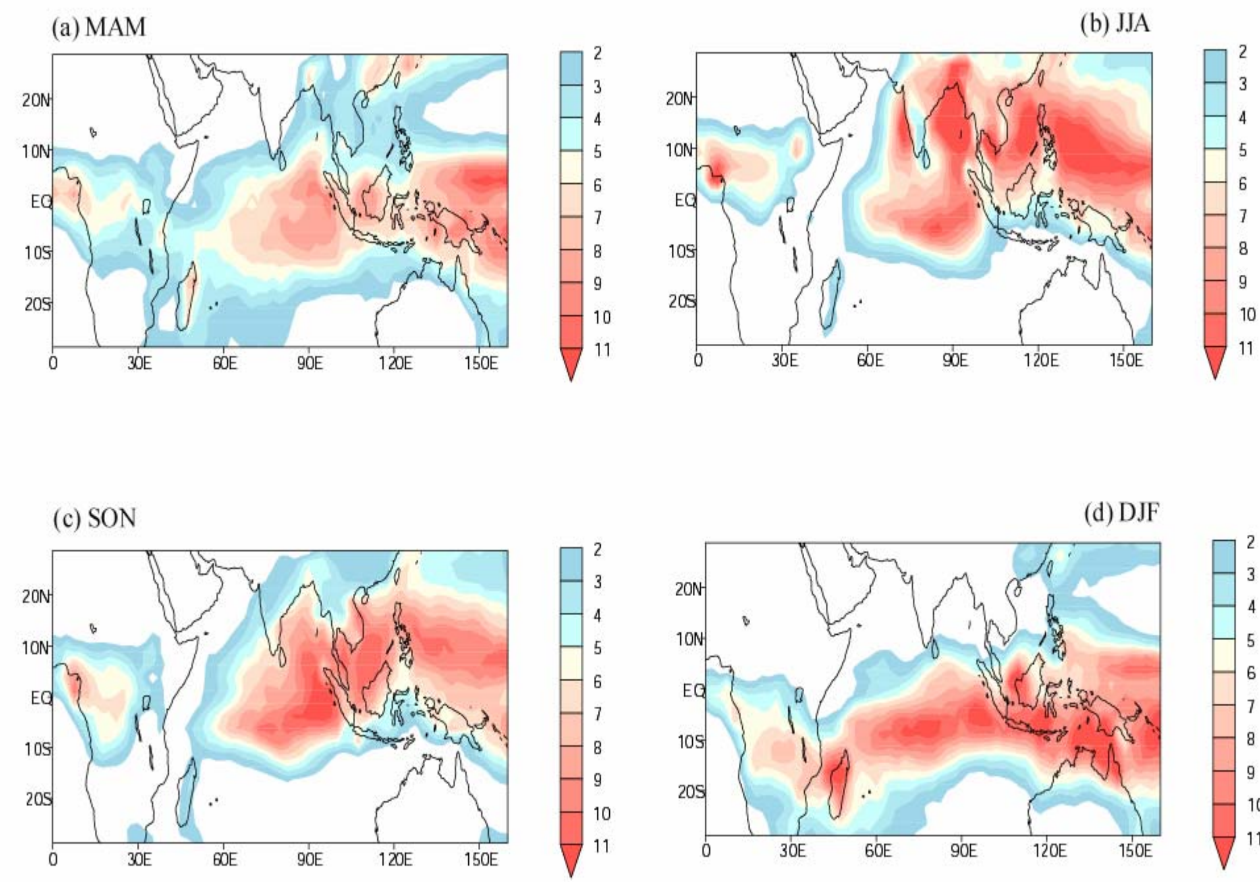
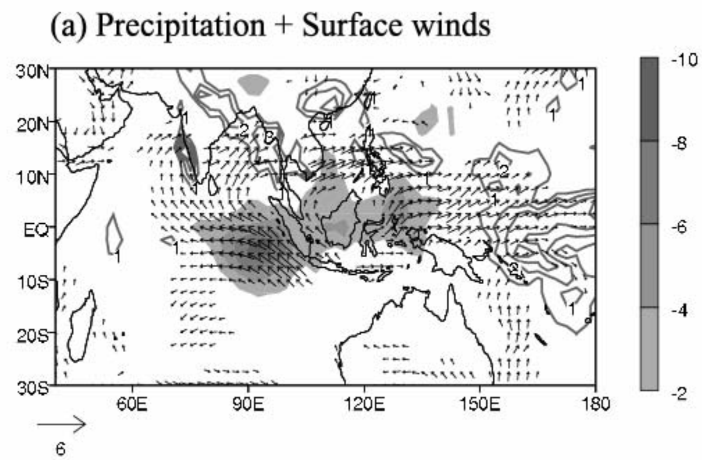


Figure 3: Observed precipitation (mm/day) climatology over the Indian Ocean for the four seasons, (a) March-May, (b) June-August, (c) September-November, and (d) December-February.



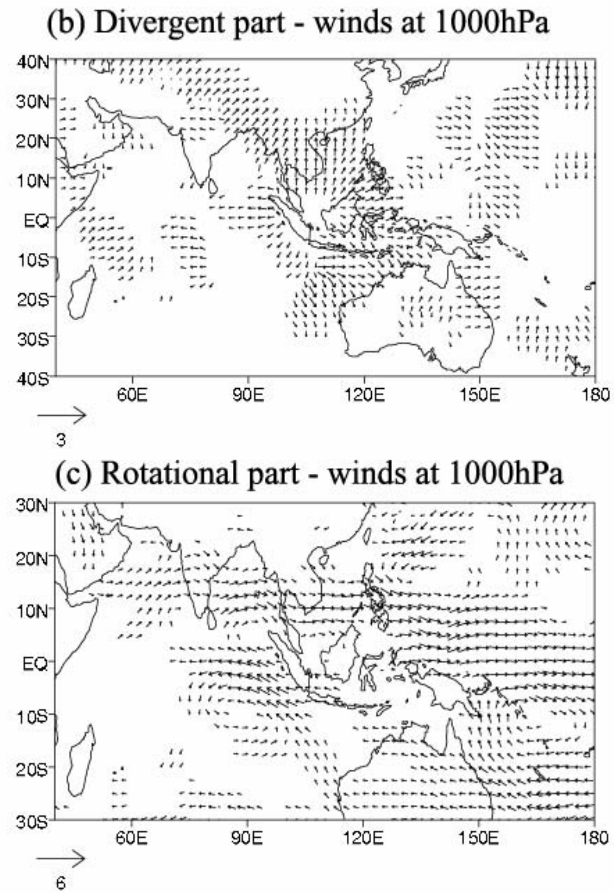


Fig. 4: July-August composites of (a) precipitation and surface winds. Negative (positive) values of precipitation are shaded (contour). The contour interval is 1mm/day while the shading interval is 2mm/day.
(b) divergent winds and (c) rotational part of the winds at 1000hPa.
Reference vector is shown in all the panels. (from Annamalai et al. 2003).

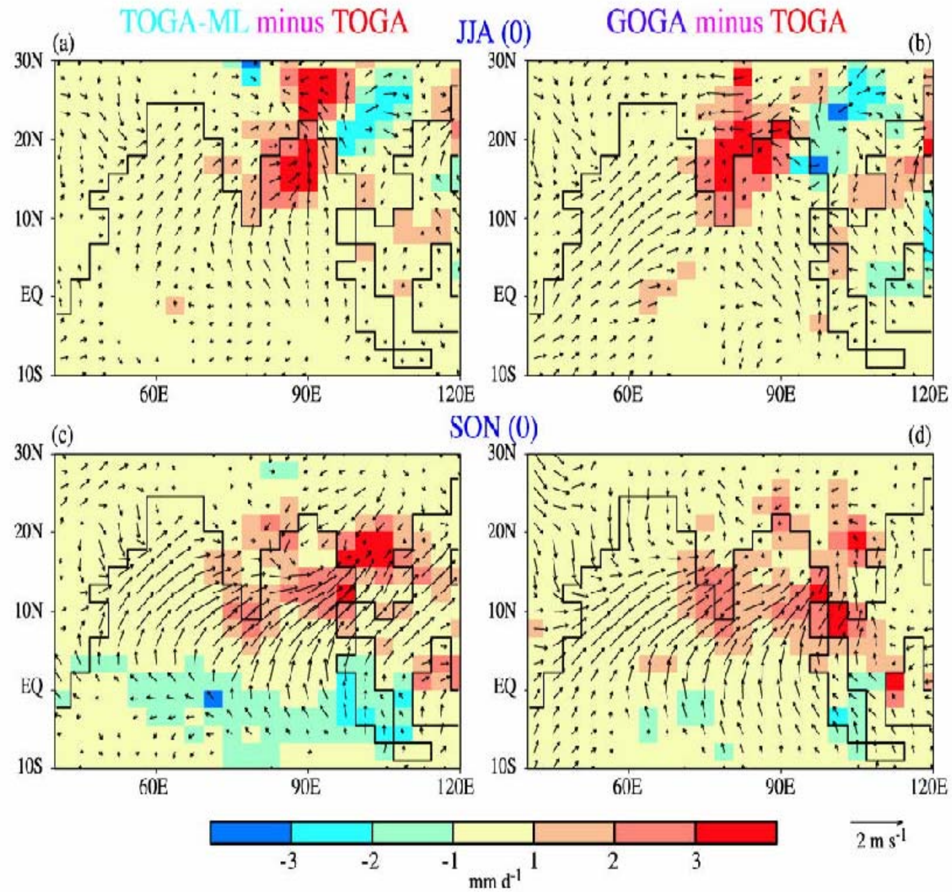
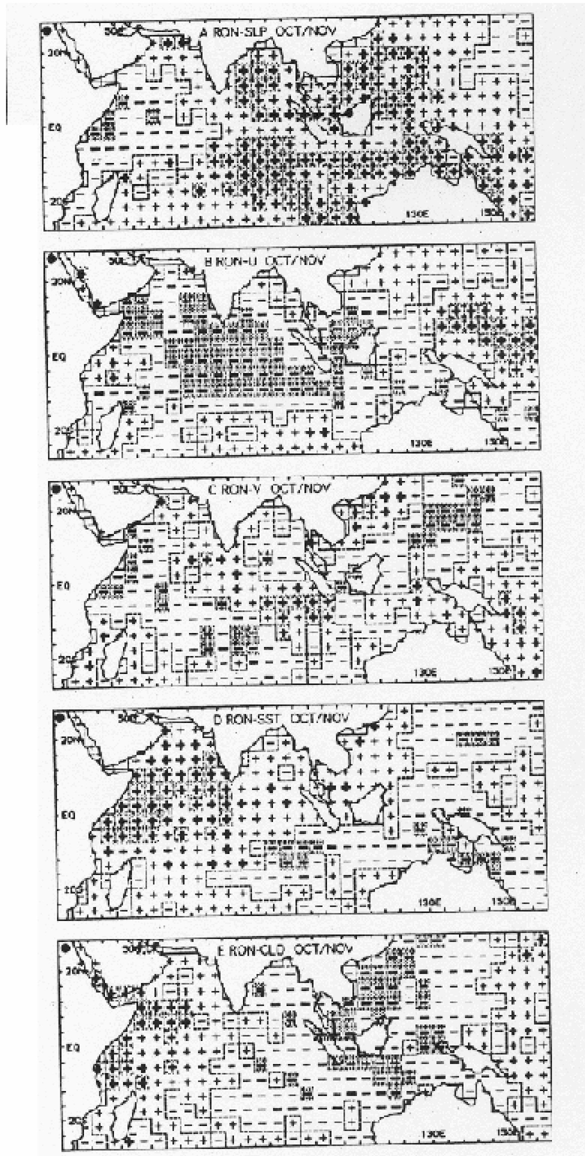


Figure 5: Difference patterns as obtained by subtracting the warm minus cold composites for TOGA from the corresponding composites for TOGA-ML (left panels), and by subtracting the composites for TOGA from those for GOGA (right panels). Results are shown for the surface wind vector (arrows; see scale at bottom right) and precipitation (shading; see scale bar at bottom; units: mm day1) during [(a), (b)] JJA and [(c), (d)] SON (from Lau and Nath, 2000).



(a)

(b)

(c)

(d)

(e)

Figure 6: Patterns of correlations between October-November rainfall over equatorial Africa and (a) SLP, (b) Zonal wind, (c) Meridional wind, (d) SST and (e) Cloud for October-November. Plus and minus symbols denote the sign of the correlation, and boldface type indicates values with local significance at 5% level. Shading represent absolute values larger than 0.4. Base period is 1948-87. (from Hastenrath et al. 1993).

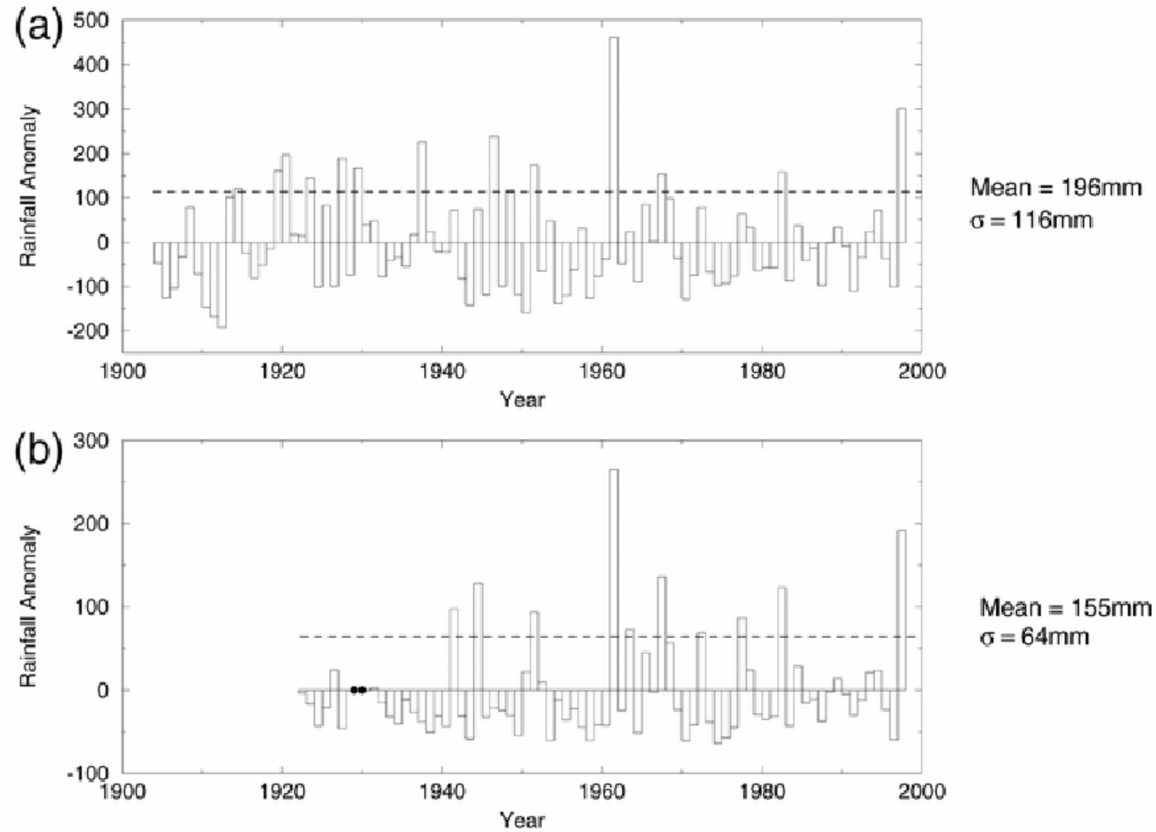


Figure 7: Time series of the seasonal mean rainfall anomaly for SON between 37.5° and 41.25°E for (a) 2.5°S-2.5°N and (b) 10°-12.5°S. The black circles indicate missing data. This gure is tak en from Black et al. (2003).

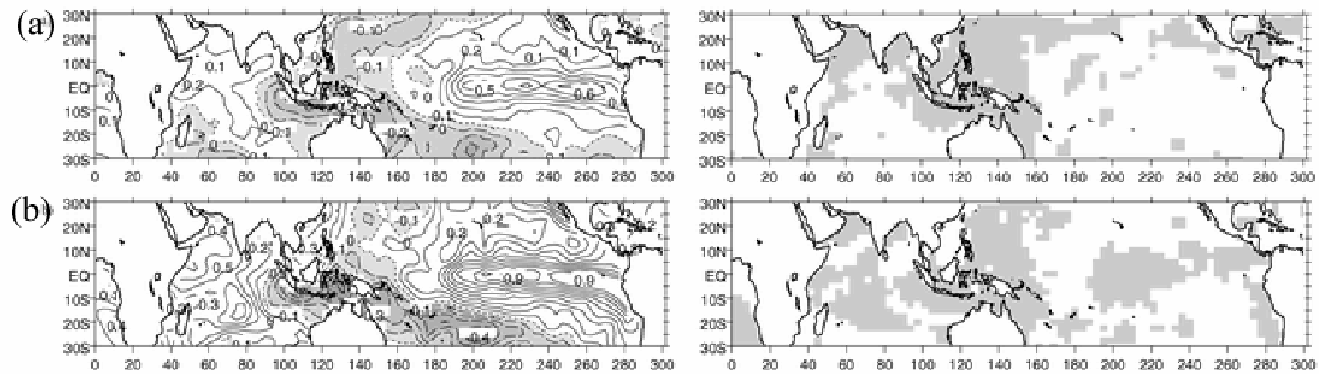


Figure 8: SST anomaly patterns associated with heavy East African short rains based on SON rainfall from Figure 7. Extreme rainfall years are based on anomalies greater than one standard deviation from the mean: (a) equatorial region (2.5°S-2.5°N), and (b) southern region (10°-12.5°S). SST anomalies are calculated relatibe to the 1900-97 climatological mean. The left panels show the SST anomalies; contour interval is 0.1 K, and neagtive anomalies are shaded. Gray shading in the right panels indicates regions where anomalies are signicant at the 95% le vel (adopted from Black et al. 2003).

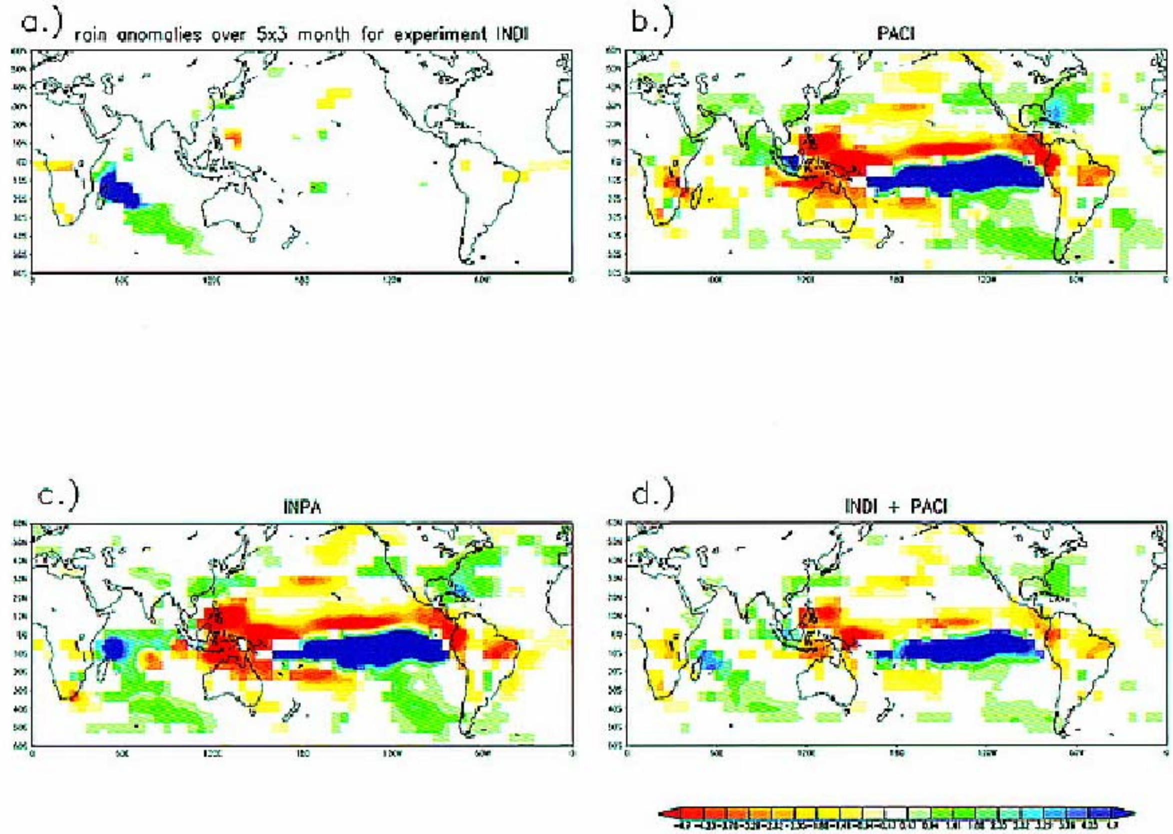


Figure 9 Ensemble mean precipitation responses (mm day1) of the atmosphere model ECHAM3 (T21) to the SST anomaly patterns (a) Response to the Indian Ocean SST anomaly, (b) response to the Pacific Ocean SST anomaly, (c) response to the complete Indo-Pacific SST anomaly, and (d) linear superposition of the response shown in (a) and (b). Shown are the anomalies (relative to a 50-yr control run with climatogical SSTs) that exceed the 95% significance level according to a *t* test (from Larif et al. 1999).

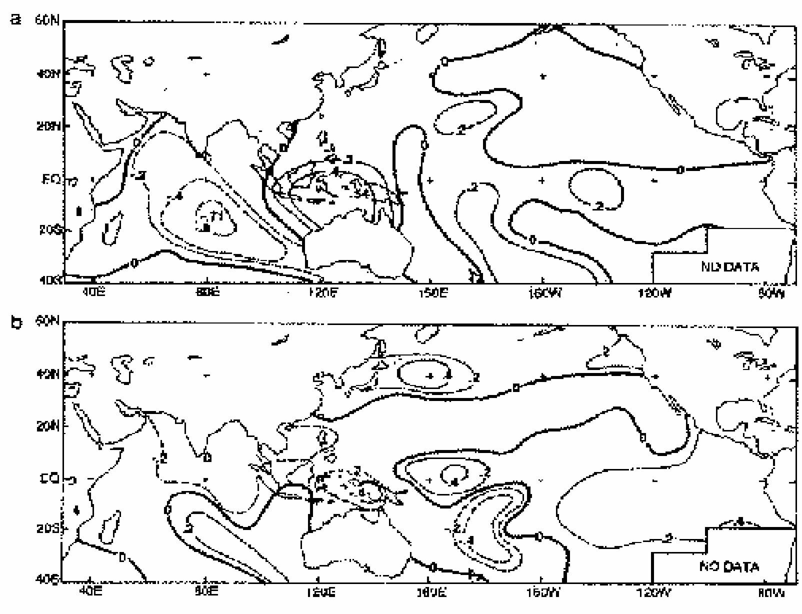


Figure 10: Partial correlations between SSTs and (a) the rst rotated principal component of Australian rainf all, after removal of effects of ENSO, and (b) the second rotated principal component of Australian rainfall, after removal of effects of ENSO. (adopted from Nicholls, 1989).

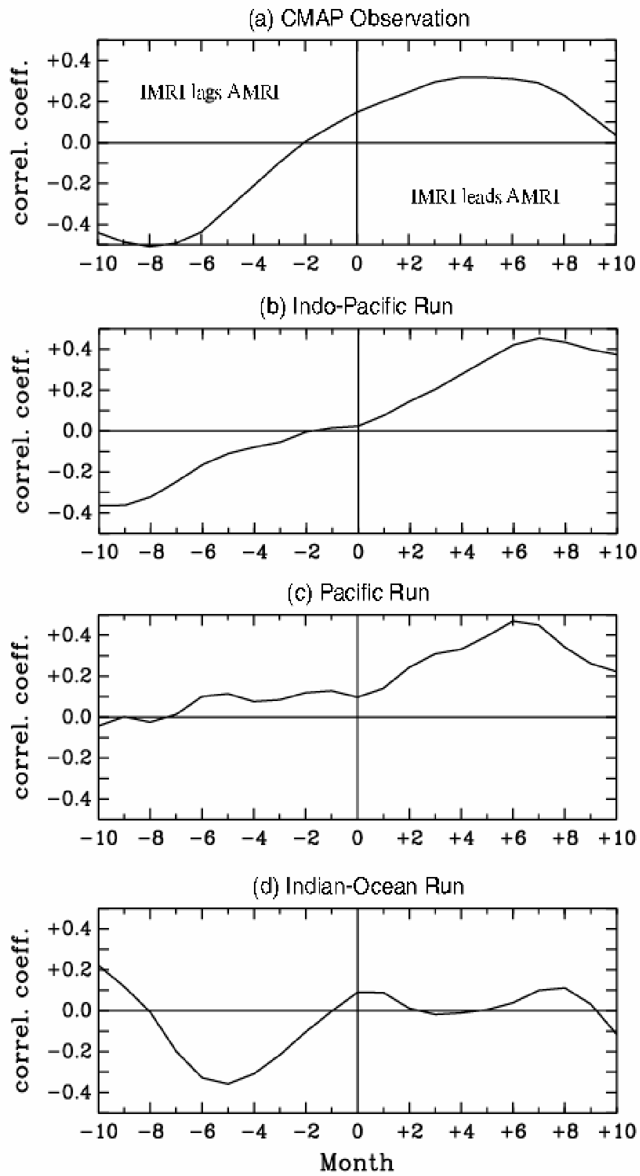


Figure 11: Time-lagged correlation coefcients between monthly Indian Monsoon Rainf all Index (IMRI) and Australian Monsoon Rainfall Index (AMRI) calculated from (a) CMAP observation, (b) Indo-Pacic CGCM run, (c) Pacic CGCM run, and (d) Indian Ocean CGCM run (from Yu et al. 2003).

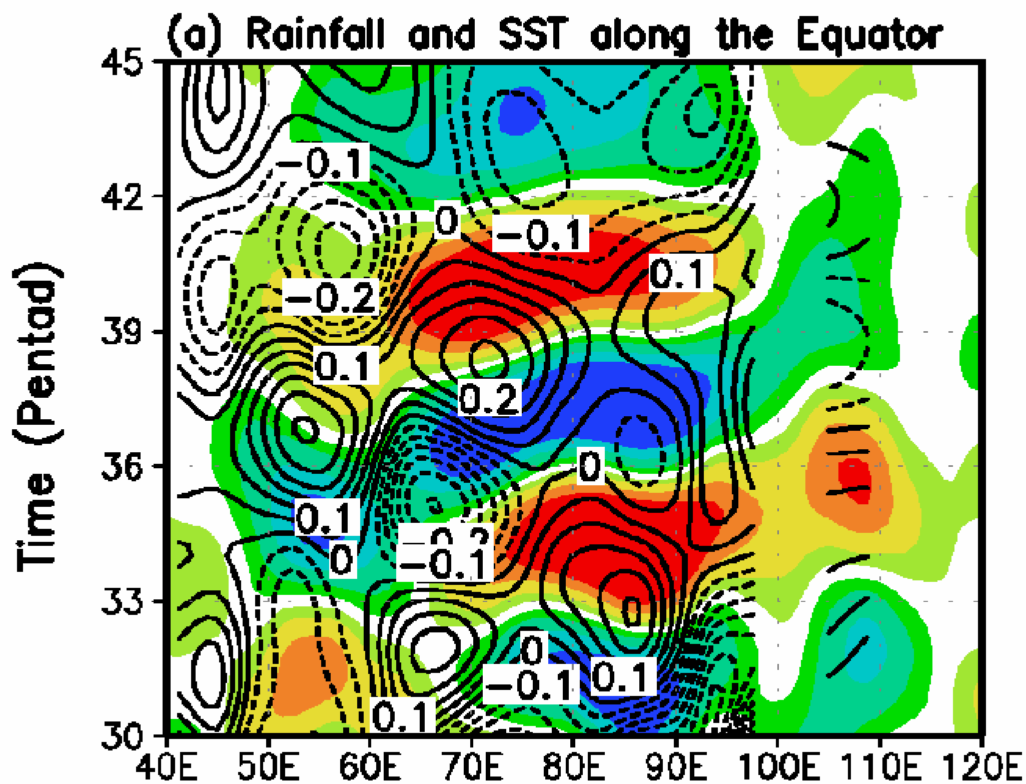


Figure 12: Longitude-time Hovmoeller diagrams of intraseasonal variability of rainfall rate (positive in yellow-red; negative in green-blue; the shading interval is 2 mm/day) and SST (contours; negative values in dotted lines; contour interval is 0.05°C) along the equator. Pentad 30 corresponds to 01 June (from Fu et al. 2003).

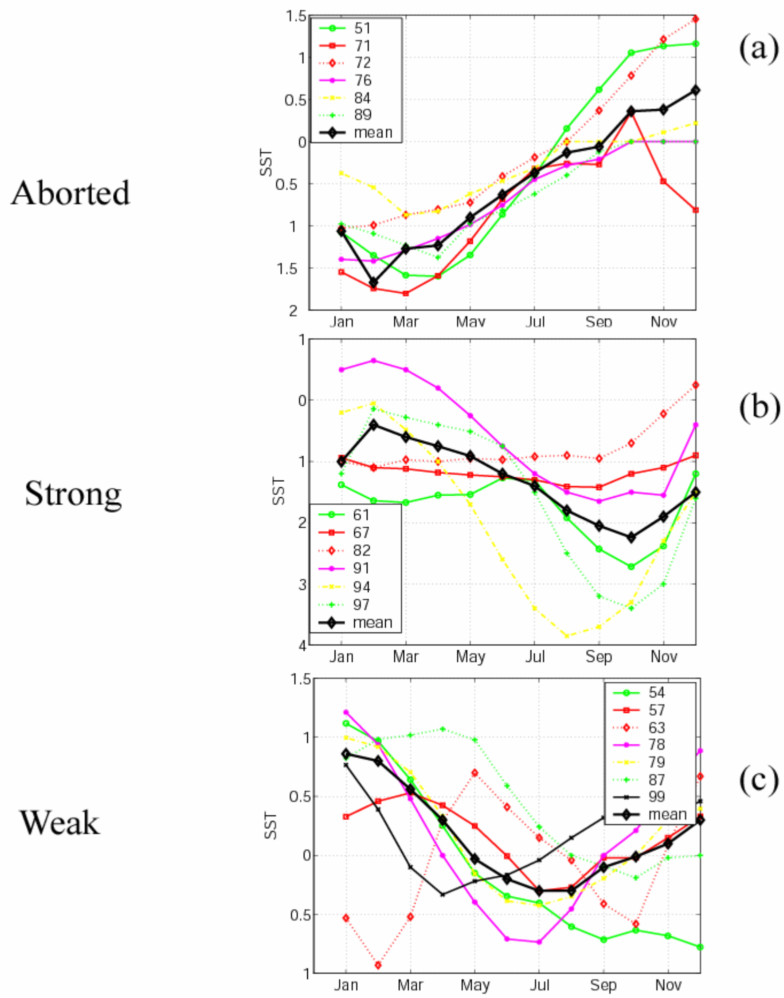


Figure 13. SST anomalies in the eastern equatorial Indian Ocean standardized and averaged over 90°E-110°E, 10°S-Equator to define (a) aborted, (b) strong, and (c) weak zonal mode years. The monthly s.t.d of SST is 0.3°C. The figure is reproduced from Annamalai e al. (2003), *Deep Sea Res.*, **Part II, 50**.

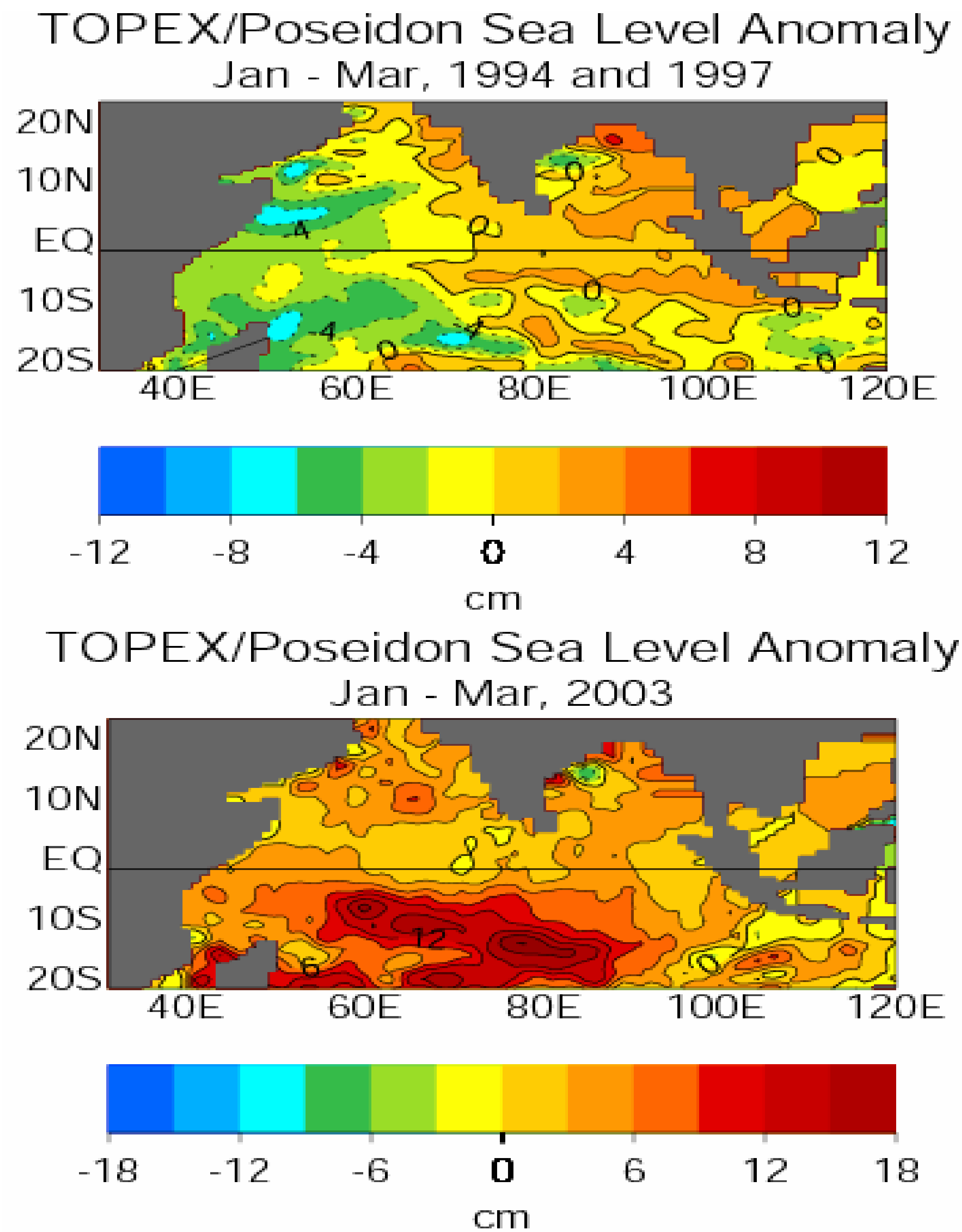


Figure 14. Composite sea level anomalies from TOPEX/POSEIDON for early 1994 and 1997, strong zonal/dipole mode years, show that the eastern Indian Ocean tends to have positive sea level anomalies prior to the onset of zonal/dipole mode years. As of August 2003, coast of Java/Sumatra had anomalous cooling. Sea level data show preconditioning in the east although the downwelling Rossby wave is the strongest signal in the Southern IO.

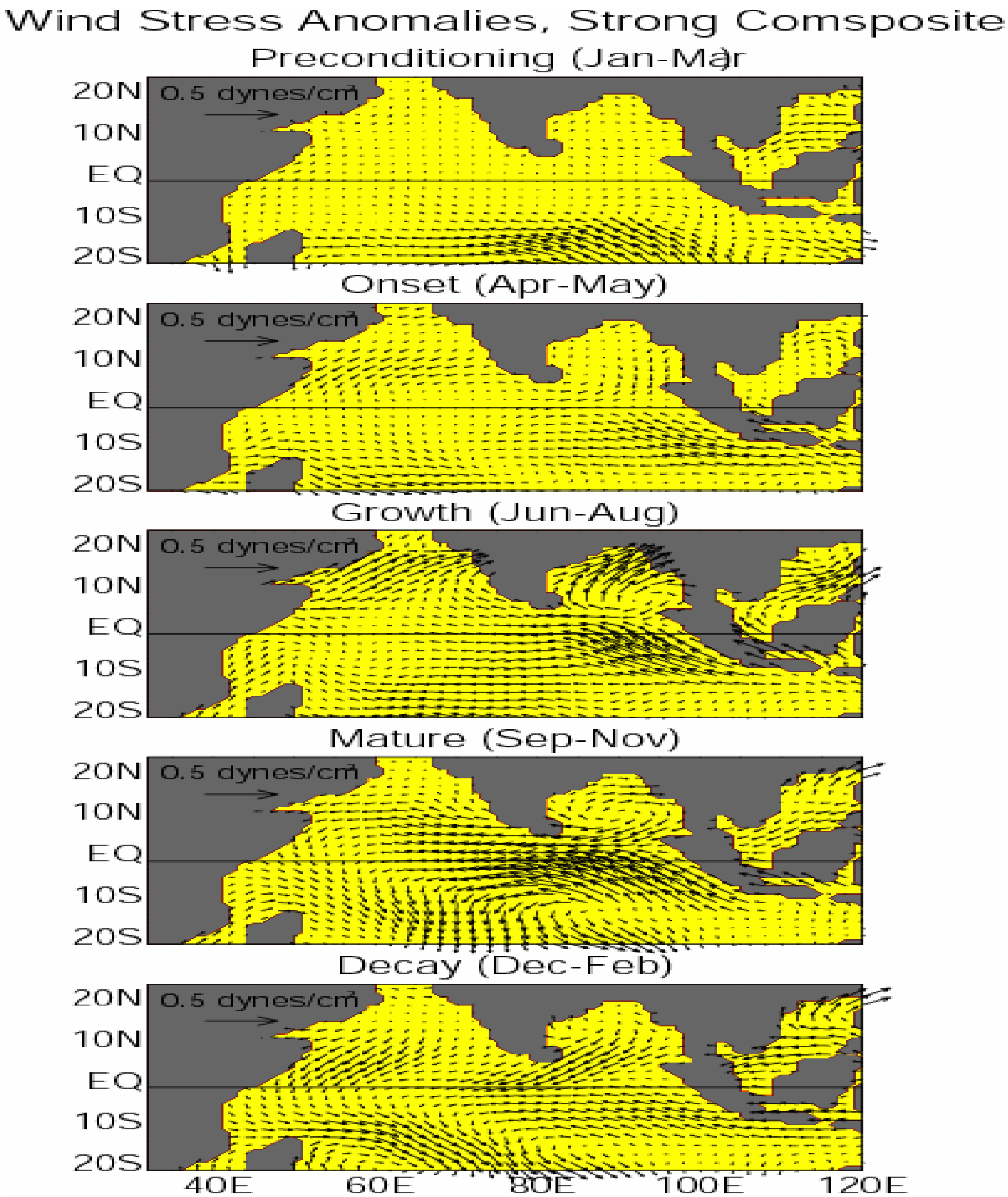


Figure 15. Wind stress composites for strong dipole/zonal mode years. By the boreal spring months, the onset phase has upwelling favorable alongshore winds off Java and Sumatra which strengthen into a tradewind-like structure by the growth phase. The winds diminish rapidly as the decay phase sets in.

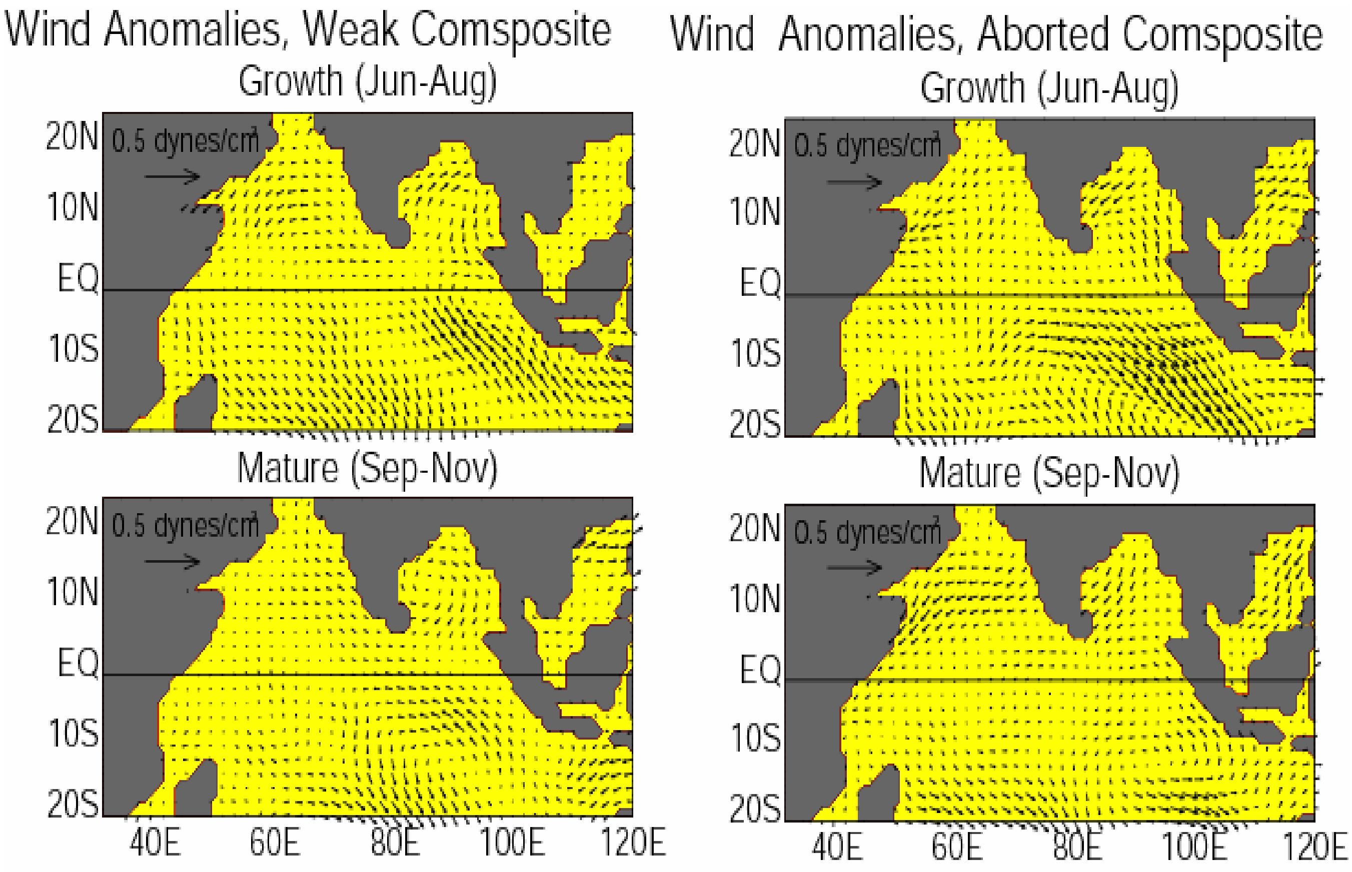


Figure 16. Wind anomaly composites for the growth and mature phases of weak and aborted zonal mode years. For the weak events, growth phase has stronger alongshore winds than the mature phase. For the aborted years, the wind anomalies are downwelling favorable during the growth phase and thus no surface cooling occurs during the mature phase despite weak upwelling favorable winds.

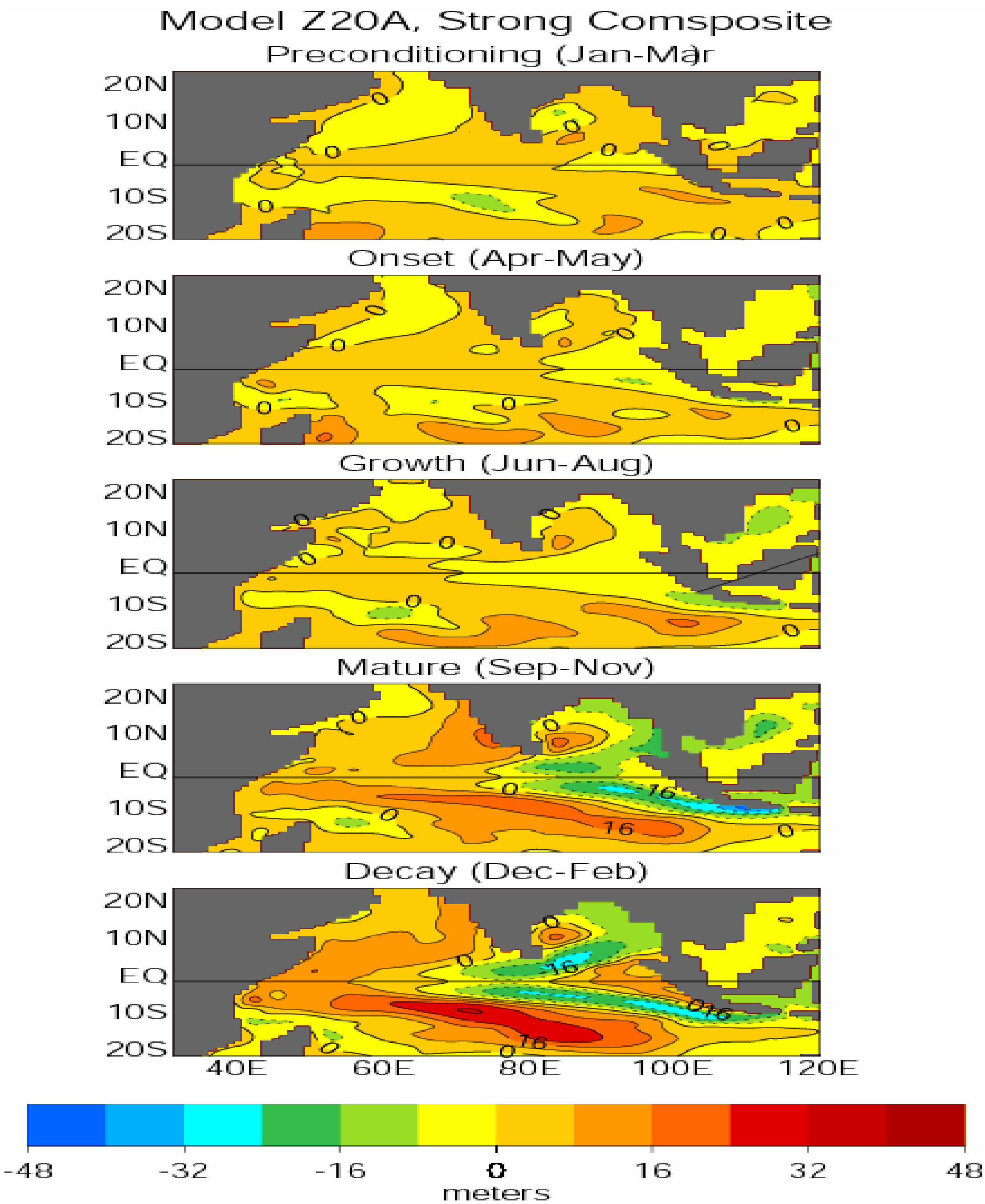


Figure 17. Composite thermocline anomalies for strong zonal mode events shows an upwelling Rossby wave and deeper eastern thermocline in the Southern IO during the preconditioning phase. The thermocline in the east shallows rapidly once the onset occurs and the downwelling Rossby wave becomes dominant in the southwestern IO during the growth and decay phases.

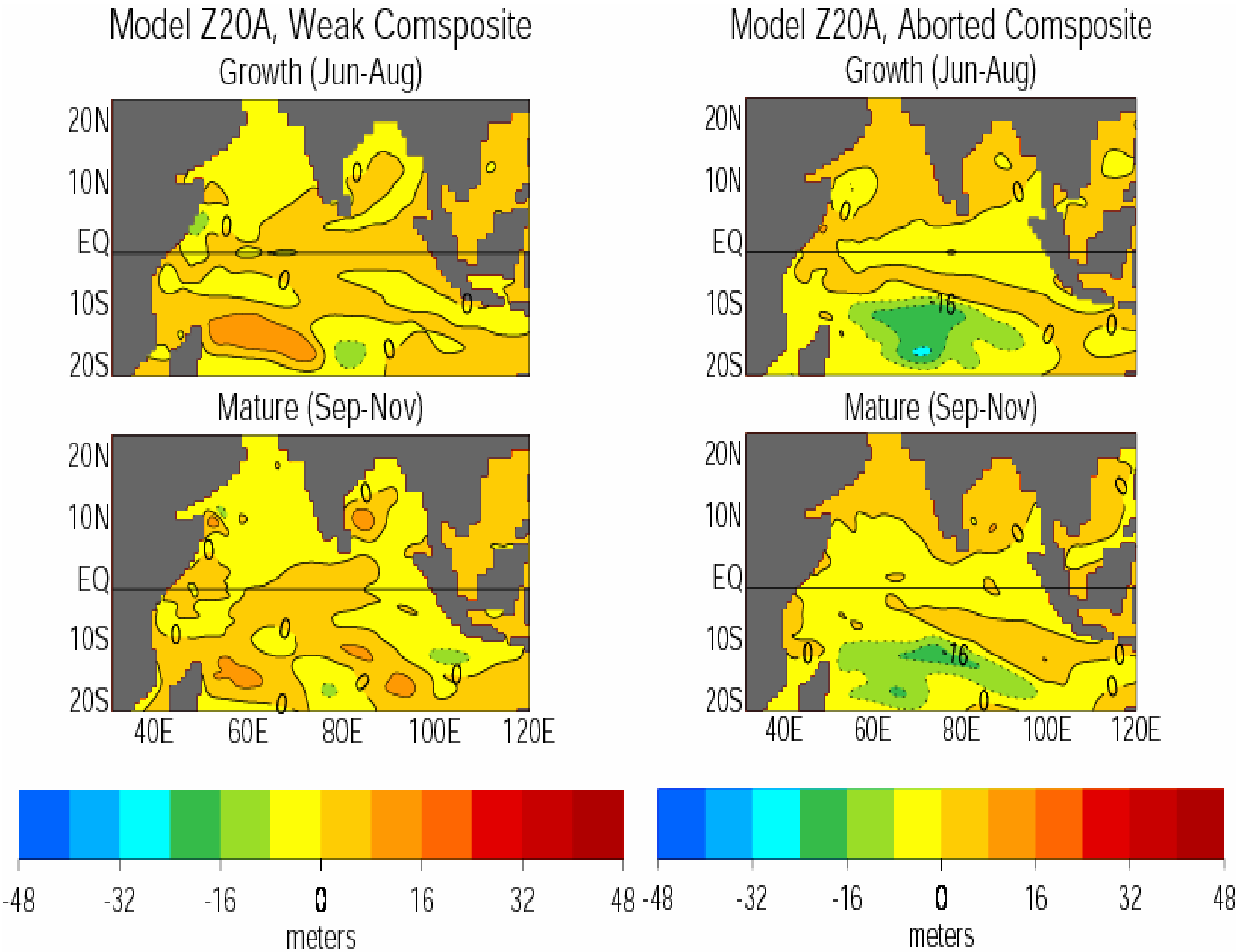


Figure 18. Thermocline anomalies for weak and aborted zonal mode years show no significant shallowing in the eastern IO even during the mature phase. Note also that downwelling Rossby wave in the Southern IO is also only weakly forced.

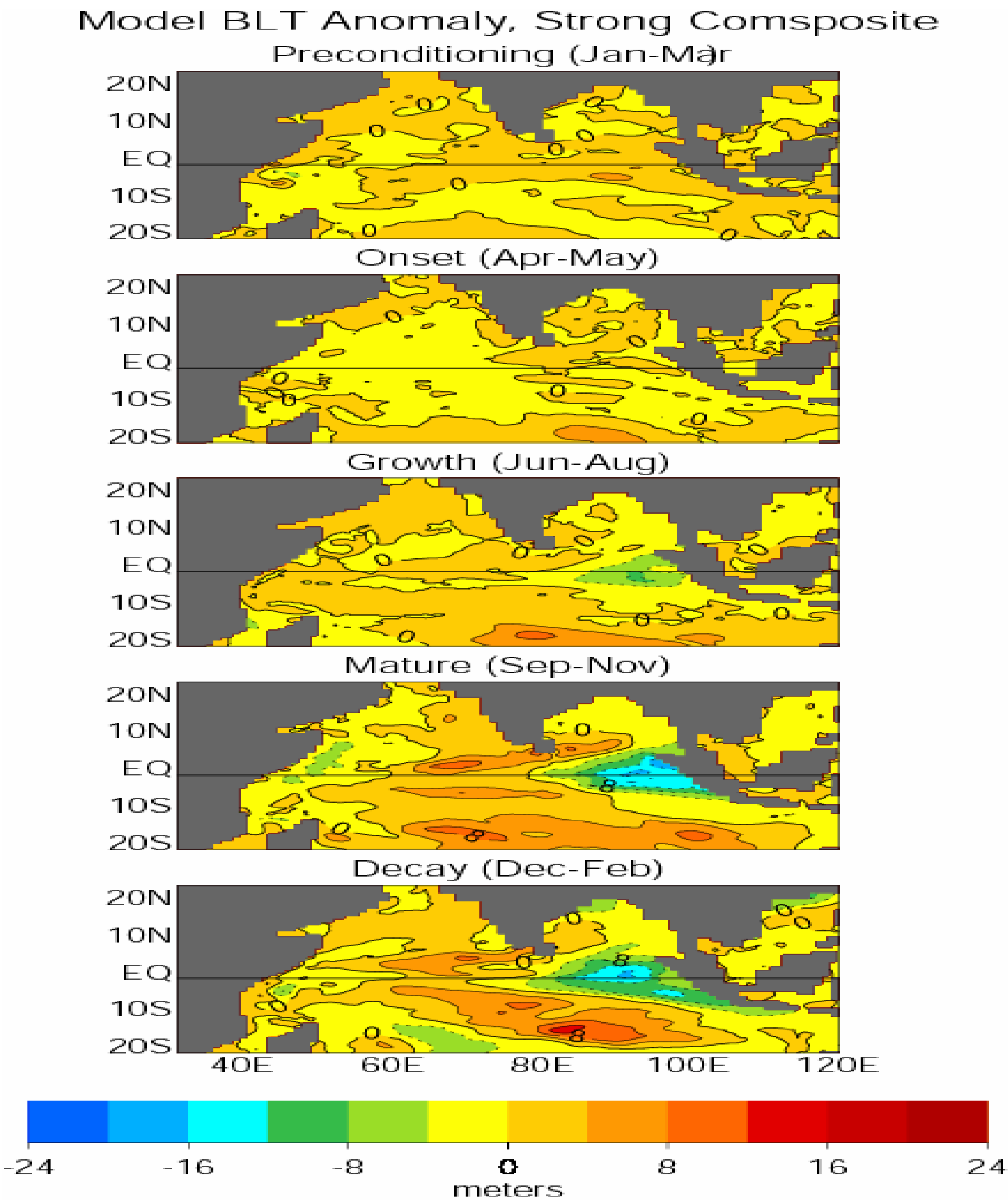


Figure 19. The barrier layer (BL) variability in the eastern IO plays a crucial role in the zonal mode evolution. The BL anomalies are positive during the preconditioning and begin to disappear during the onset. During the growth and mature phases, the BL anomalies are negative and they do not recover completely even during the decay phase.

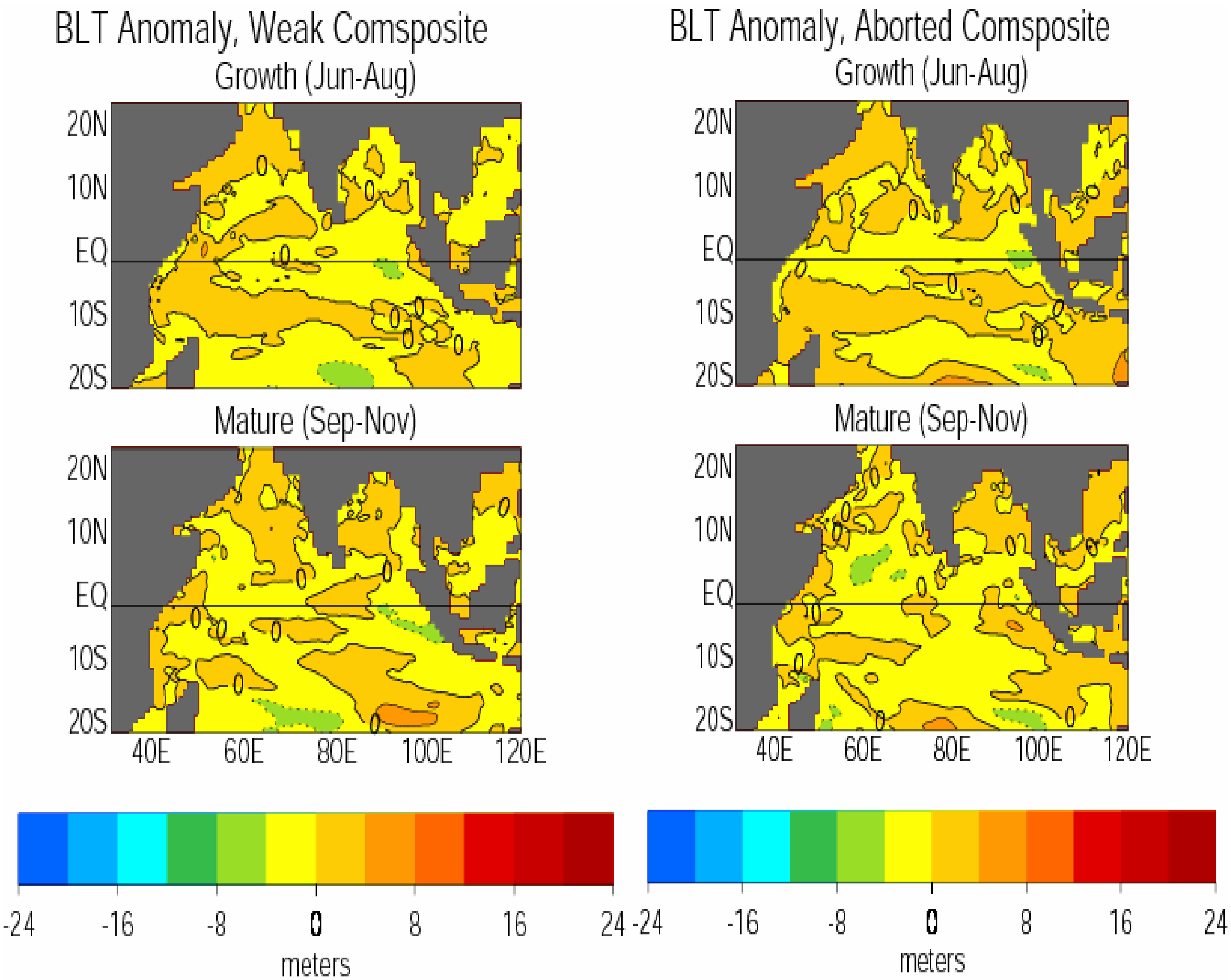


Figure 20. Barrier layer anomalies during the growth and mature phase of the weak and aborted events indicate that they are not reduced greatly, especially for the aborted events. Thus the thermocline-SST feedback may not be fully operative, resulting in either weak or no ocean-atmosphere coupling in the eastern IO.

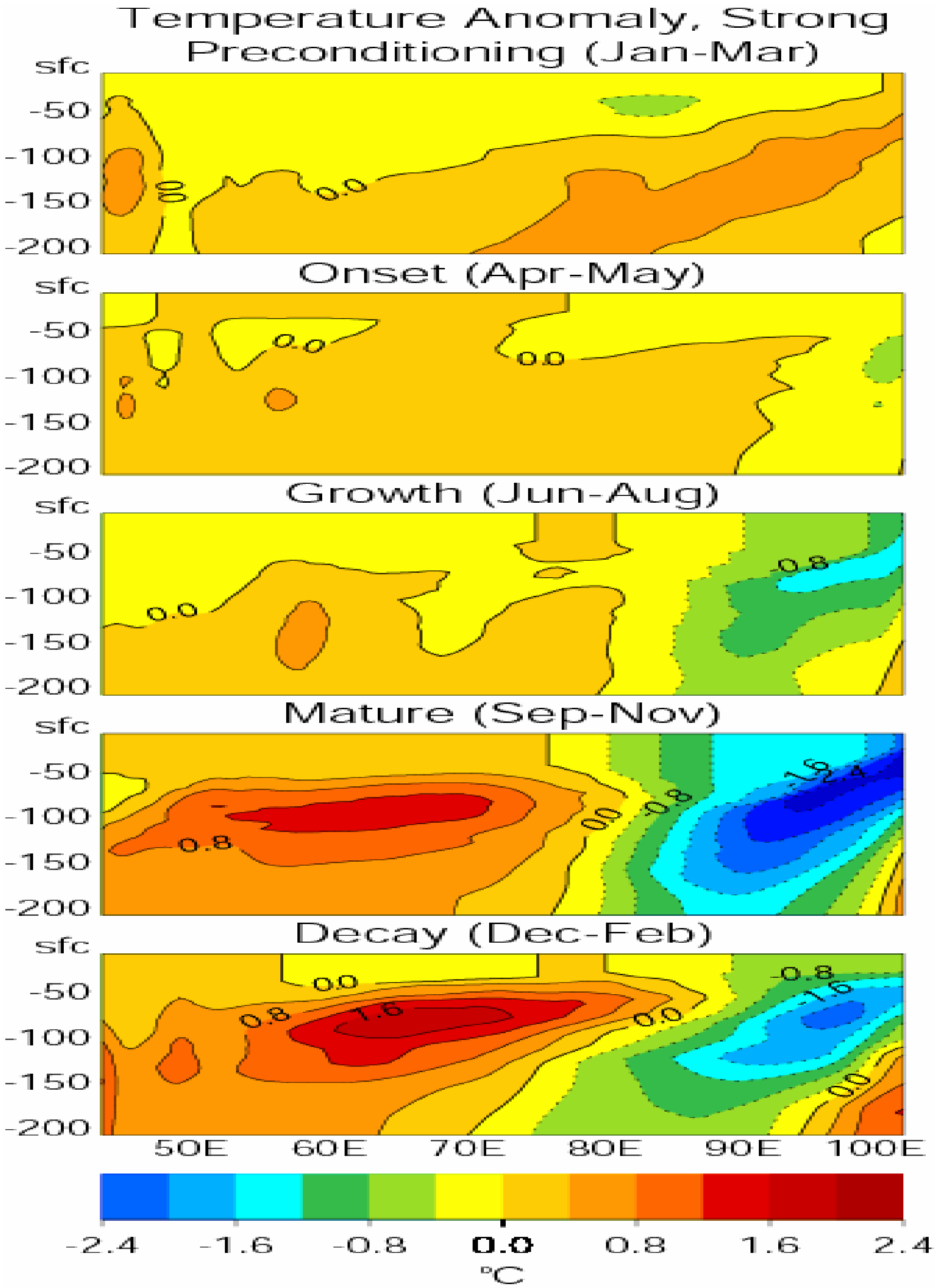


Figure 21. Subsurface temperature anomalies along 5°S for a composite strong event shows that the preconditioning phase in the east corresponds to deeper thermocline in the east and the upwelling favorable windstress curl away form the coast. The shallowing of the thermocline in the east is initiated during the onset phase and continues to shoal through to the mature phase before beginning to deepen. The deepening of the thermocline extends eastward through the year.

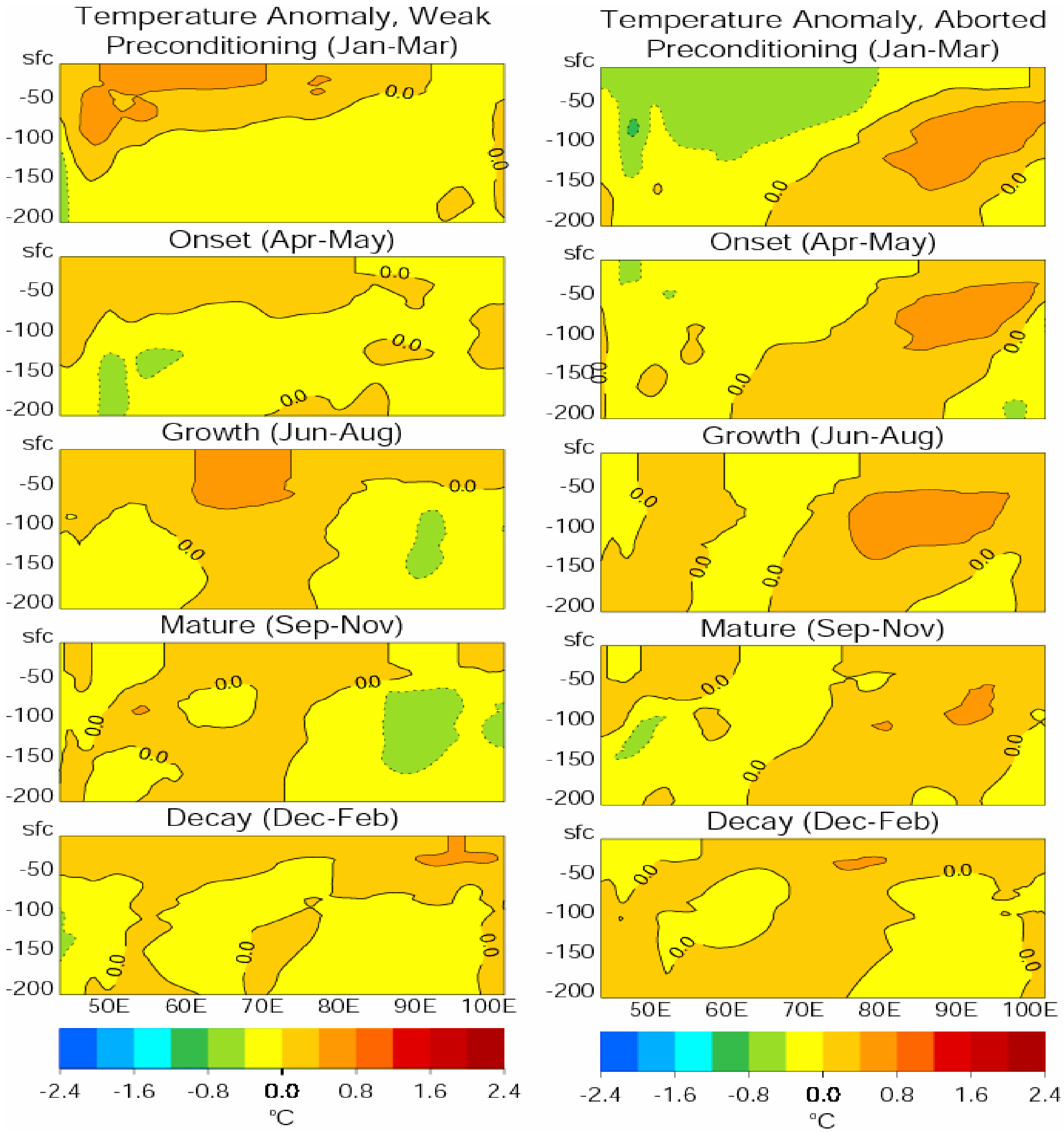


Figure 22. Subsurface temperature anomalies along 5°S for weak and aborted events show that weak events have only minor upwelling during the growth andmature phase with no strong feedbacks between SSTs and the thermocline. Aborted events have no role for ocean dynamics at all during much of the year with only weak cooling associated with the eastward extension of the western upwelling during early part of the year.

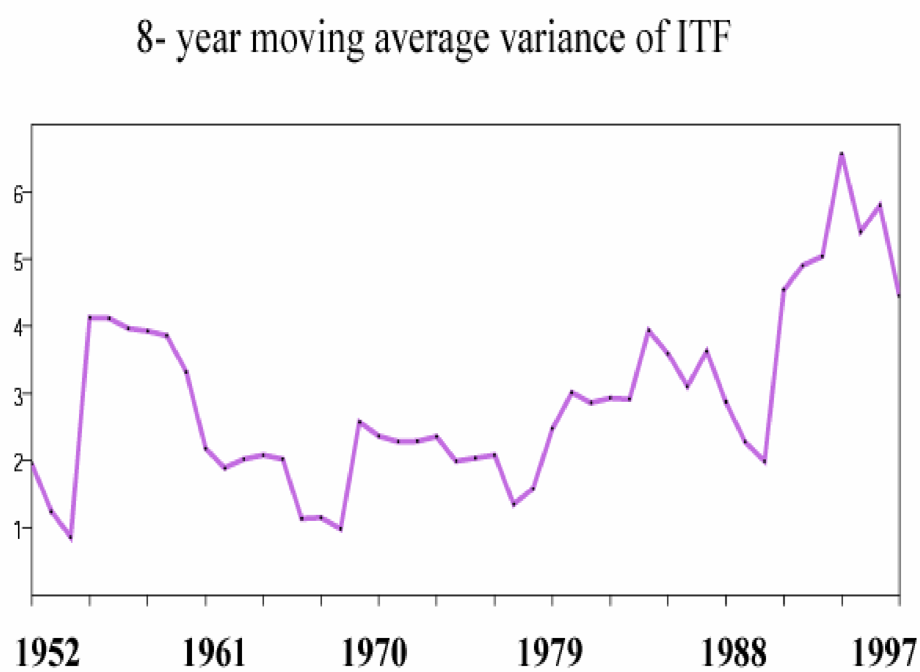


Figure 23. Variance of the low-pass filtered (8-year running mean) Indonesian throughflow displays a interdecadal variability similar to the variance of the IO Zonal Mode index defined by Annamalai et al. (2003) as the standardized SST averaged over 90°E-110°E, 10°S-Equator. The impact of ITF on the IO decadal variability may manifest itself in the heat content and SST variability.

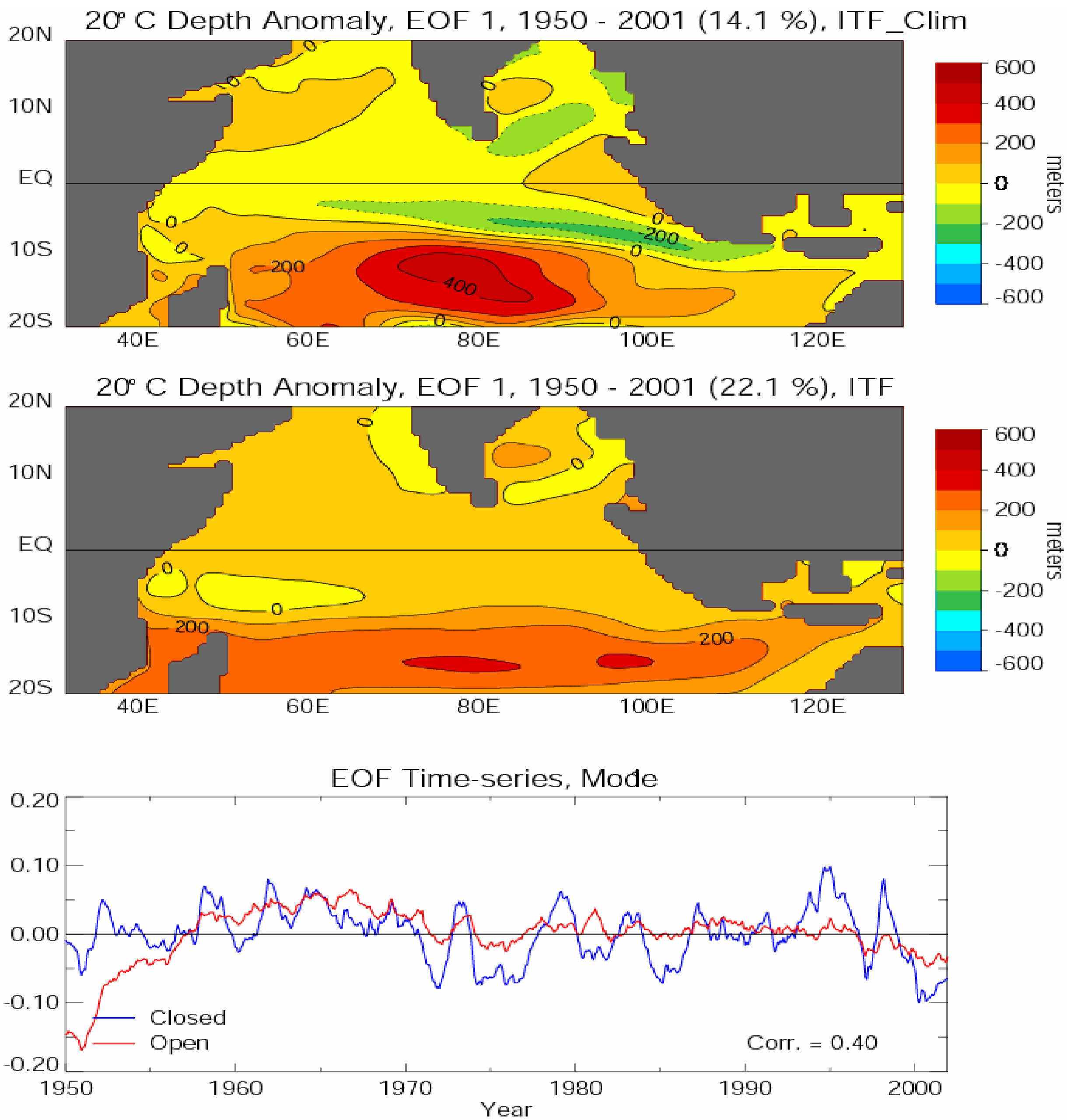


Figure 24. The first EOF of thermocline depth anomalies is dominated by a decadal pattern when the interannual Indonesian throughflow (ITF) is represented in the model whereas interannual variability is predominant when the ITF is fixed to its climatology.

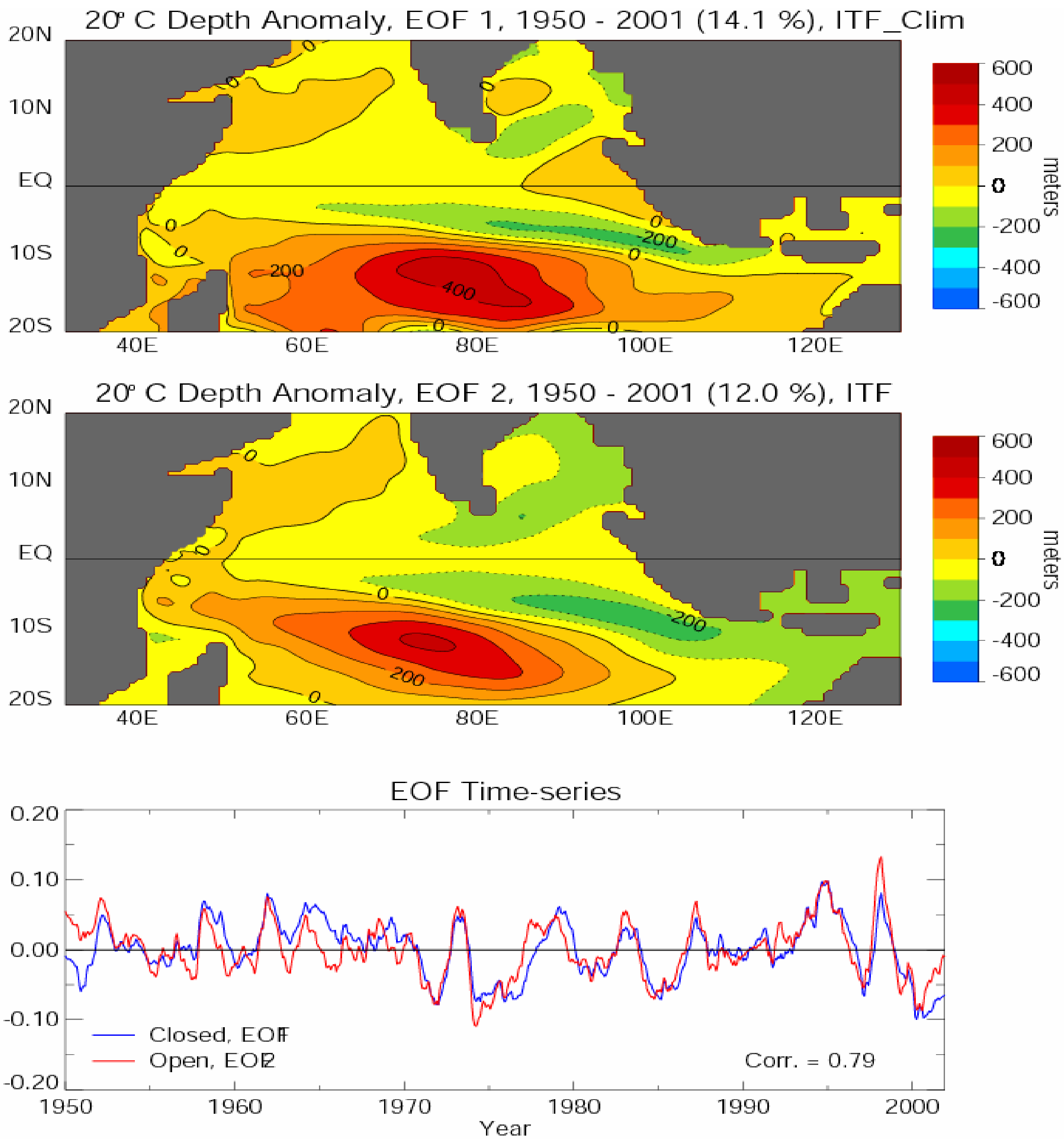


Figure 25: The second EOF of thermocline anomalies with the real ITF corresponds to the first EOF with the climatological ITF since the first EOF is a decadal mode when the ITF is allowed to vary. Note that the interannual mode is dominated by Rossby-Kelvin wave structure whereas the decadal mode is a basin mode with maximum amplitude in the Southern IO where ITF has maximum impact.