Seasonal-to-Interannual Forecasting of Tropical Indian Ocean Sea Surface Temperature Anomalies: Potential Predictability and Barriers

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ABSTRACT

Whether seasonally phased-locked persistence and predictability barriers, similar to the boreal spring barriers found for El Niño–Southern Oscillation (ENSO), exist for the tropical Indian Ocean sector climate is investigated using observations and hindcasts from two coupled ocean–atmosphere dynamical ensemble forecast systems: the National Centers for Environmental Prediction (NCEP) Coupled Forecast System (CFS) for 1990–2003, and the NASA Seasonal-to-Interannual Prediction Project (NSIPP) system for 1993–2002. The potential predictability of the climate is also assessed under the “perfect model/ensemble” assumption.

Lagged correlations of the indices calculated over the east and west poles of the Indian Ocean dipole mode (IDM) index show weak sea surface temperature anomaly (SSTA) persistence barriers in boreal spring at both poles, but the major decline in correlation at the east pole occurs in boreal midwinter for all start months with an almost immediate recovery, albeit negative correlations, until summer approaches. Processes responsible for the change in sign of SSTAs associated with a major IDM event effect a similar change on much weaker SSTAs. At the west pole, a major decline occurs at the end of boreal summer for fall and winter starts when the thermocline deepens with the seasonal cycle and coupling between the ocean and atmosphere is weak.

A decline in skillful prediction of SSTA at the east pole over boreal winter is also found in the hindcasts, but the relatively large thermocline depth anomalies are skillfully predicted through this time and skill in SSTA prediction returns. A predictability barrier at the onset of the boreal summer monsoon is found at both IDM poles with some return of skill in late fall. Potential predictability calculations suggest that this barrier may be overcome at the west pole with improvements to the forecast systems, but not at the east pole for forecasts initiated in boreal winter unless the ocean is initialized with a memory of fall conditions.

1. Introduction

To date, investigations on the prediction of climate variability have tended to focus on El Niño–Southern Oscillation (ENSO), as discussed by Latif et al. (1994), and the North Atlantic (see e.g., Griffies and Bryan 1997). Skillful prediction of interannual variability in the Asian–Australian monsoons, while recognized as an important goal, remains elusive due to the complexity of the system (Webster et al. 1998). However, prospects for the recently identified Indian Ocean dipole mode (IDM) (Saji et al. 1999) in which tropical sea surface temperature (SST) over the Indian Ocean seesaws zonally, resulting in occasions of severe drought over Indonesia and severe flooding over East Africa, appear more hopeful (Wajsowicz 2004). In the present study, the potential predictability of SST variability over the tropical Indian Ocean, and the existence of predictability barriers, are investigated using observations and results of multiyear ensemble hindcasts from the National Centers for Environmental Prediction (NCEP) Coupled Forecast System (CFS) and the National Aeronautics and Space Administration (NASA) Seasonal-to-Interannual Prediction Project (NSIPP) system.

There are two types of climate prediction problem (Lorenz 1975). In the boundary value problem, the task is to assess the change in climate due to some change in external forcing, for example, anthropogenic changes. 
In the initial value problem considered here, compare with ENSO forecasting, the task is to predict the evolution of the climate system, given some estimate of its present state. The evolution is sensitive to the initial conditions specified, and trajectories initially close together diverge, so rendering the forecasts useless after some time. “Perfect model,” or “perfect ensemble” experiments, in which time integrations of the climate system model are conducted from initial conditions that differ by only a small perturbation, address this error and enable the “potential predictability” of the climate system to be assessed. Model estimates of potential predictability of sea surface temperature (SST) variations over the equatorial Pacific Ocean are about three seasons, and typically estimates for surface temperature over the oceans are greater than over land (Collins 2002).

The upper limits to predictability of the climate system given by the perfect model experiments may, or may not, be robust depending on whether they are due to inherent instabilities in the coupled ocean–atmosphere–land system, ensemble bias, or model inadequacies. Forecasting ENSO more than a season or two in advance is apparently limited by a boreal “spring predictability barrier.” The rapid decline in observation–prediction correlations in boreal spring occurs irrespective of the initiation month of the forecast and in a wide range of coupled ocean–atmosphere models (see, e.g., Moore and Kleeman 1996; Chen et al. 1997). A similar springtime decline is found in the lagged correlations (equivalent to forecasting simple persistence of the initial anomaly) of the Southern Oscillation index (Webster and Yang 1992). A persistence/predictability barrier may be expected in boreal spring, as the Walker circulation and zonal equatorial gradient in sea surface temperature are at their weakest, so the coupled ocean–atmosphere system over the Pacific is least stable and most susceptible to external influences and noise.

Lagged correlations of observed SSTAs over the tropical Indian Ocean suggest that there may be barriers to seasonal prediction in boreal summer and winter. However, unlike for ENSO, correlations increase again in the east, albeit of the opposite sign, emphasizing the importance of the quasi-biennial nature of the Indian Ocean sector climate. These results are presented in section 2 along with a discussion of mechanisms and feedbacks, which modulate the signal seasonally. The coupled ocean–atmosphere–land forecast systems are introduced in section 3 with a discussion of their drifting climate, which may affect the evolution of anomalies. Hindcasts of SSTAs over the eastern poles of the IDM are presented at one and two season lead times. Seasonal variations in forecast skill measured by correlation of predicted SSTAs and thermocline depth with observations are described in section 4. They suggest that predictability barriers exist in the present generation of forecast systems. Their robustness is investigated under the perfect model/ensemble assumption in section 5. A discussion and summary are given in section 6.

2. Observed SST variability and persistence barriers

a. The dataset and indices

The SST data described in this section are from version 2 of the NCEP analyses (Reynolds et al. 2002) from 1982 to 2004 inclusive. Anomalies are calculated by linear detrending of the time series and subtracting the monthly mean of the series. Over the tropical Indian Ocean, interannual variability in SST has two dominant structures, namely, a basinwide mode in which the eastern and western halves vary in unison (Venzke et al. 2000) and a dipole mode in which the halves vary oppositely (Saji et al. 1999). Whether these modes are just seasonal expressions of a mode forced by ENSO has been the subject of much debate (Baquero-Bernal et al. 2002; Saji and Yamagata 2003), and will be discussed further in the next subsection, as it relates to the return of skill in the forecasts. In the analysis presented herein, the SST variability over the tropical Indian Ocean is gauged by averaging SSTAs over the eastern (10°S–0°, 90°–110°E) and western (5°S–5°N, 50°–70°E) poles of the IDM index, as defined by Saji et al. (1999).

The signal-to-noise ratio in the tropical Indian Ocean is much worse than over the eastern tropical Pacific Ocean, as given by the power spectra in Fig. 1. The Niño-3 index power spectrum has peaks in the 2-5 yr range corresponding to ENSO; the Niño-3 index is given by averaging SST over 5°S–5°N, 150°W–90°W. At the west pole of the IDM index the major peaks also lie between 2 and 5 yr. The spectrum of the east pole of the IDM index has major peaks around 18 months and 3 yr. The quasi-biennial nature of the IDM index was noted by Saji et al. (1999). The differing nature of the quasi-biennial signal at the two poles is explored further below.

b. Seasonal modulation and quasi-biennial nature

Interannual variability in SST over the eastern equatorial Pacific Ocean has a peak in December with a sharp decline in spring coincident with the “spring pre-
dictability/persistence barrier” (Fig. 2a). Variability in the thermocline depth (measured by the depth of the 20°C isotherm) peaks a month or so earlier (Fig. 2a). There is a close relationship between SSTAs and thermocline depth anomalies, as indicated by the high correlation in Fig. 2b with the 20°C isotherm depth anomalies (D20A) leading SSTAs over the eastern equatorial Pacific by a couple of months. The D20As are derived from the revised version of NCEP’s Global Ocean Data Assimilation System (GODAS) for 1982–2004 and are described in more detail in section 3a [D. Behringer et al. (2006, unpublished manuscript); also see information online at http://www.cpc.ncep.noaa.gov/products/GODAS/].

Features of the seasonal cycle, which explain the seasonal modulation of the interannual SSTAs and the relationship between SSTAs and D20As, are shown in Fig. 3. SSTAs are largest in the east Pacific, as SSTs are too cool there for the SST–cloud–incident shortwave radiation negative feedback to be effective; low OLR values, signifying deep convection, are confined to the western Pacific (Fig. 3a). Also, the SST–wind–evaporation feedback is positive over the warm (cool) SSTAs and associated westerly (easterly) anomaly, as the zonal surface winds are easterly throughout the year (Fig. 3b). The end of year peak in SSTA, when the equatorial Pacific is most stable, that is, coupling strength between the ocean and atmosphere is minimal rather than during spring and early summer when the coupling strength is maximal, has been explained using linear dynamics and a simple delayed oscillator mechanism by Tziperman et al. (1998) and is illustrated in Fig. 3c.

Assuming coupling strength is proportional to thermocline depth, then westerly anomalies generated in boreal spring when the ENSO event is weak generate weak upwelling westward propagating long Rossby waves (LRWs) and weak downwelling eastward propagating equatorial Kelvin waves (EKWs). The LRWs amplify as they propagate across the Pacific over boreal summer as coupling is strong (Fig. 3c). The large amplitude upwelling EKWs, generated by reflection at the western boundary, then balance the large amplitude downwelling EKWs generated by the now large westerly anomalies, which are only weakly amplified in the east Pacific during boreal winter (Fig. 3c). Hence, the ENSO event peaks. A similar argument can be used to show that an ENSO event cannot peak when coupling is strongest in boreal spring and summer.

As discussed by Balmaseda et al. (1995), the close-
ness of the relationship between SSTAs and D20As given by the above seasonally modulated delayed oscillator mechanism may enable forecast skill to return when the coupled system is disturbed. Suppose the disturbance does not destroy the memory of the coupled system in both the surface and subsurface layers simultaneously. Then, if skill in predicting thermocline depth anomalies declines, but that for SSTAs remains good, the predicted SSTAs could drive wind anomalies that force thermocline depth changes, thus permitting recovery of the thermocline depth skill. Alternatively, if skill in predicting SSTAs drops, then the predicted thermocline depth changes could result in upwelling changes that permit recovery of the SST skill.

In contrast with the seasonal modulation of the ENSO signal, peak SST variability over the east IDM pole occurs in September (about half that of the ENSO peak) with a secondary peak in March (Fig. 2a). The minimum in between, at the onset of the boreal summer monsoon, is secondary to that in December. D20A variability is a maximum over boreal winter when SST variability is a minimum, and a minimum in May (Fig. 2a). As in the Niño-3 region, the SSTAs and D20As are related (Fig. 2b), but here D20As lag SSTAs by a month or so.

These results may be explained by considering IDM-type mechanisms. The SSTAs are plotted as time series for each year, including the year before and the year after, in Fig. 4, in sets according to the strength of the IDM index that year. Features of the seasonal cycle, which help determine the seasonal modulation, are plotted in Fig. 5. Peak SSTA variability in boreal fall is expected and is due to the extreme cooling associated with the 1994 and 1997 positive IDM events and warming associated with the 1998 negative IDM event, as well as lesser warming in 1996, 1992, and 2001 (Fig. 4). The rapid growth to the west of Sumatra–Java of an isolated cool SSTA, or basinwide dipole SSTA, occurs in late summer and fall, as the associated surface southeasterly wind anomaly on the eastern flank (cf. Gill 1980) combines with the climatological southeasterlies to give a positive SST–wind–evaporation feedback (Li et al. 2003), similarly for a warm SSTA and the associated northwesterly wind anomaly. In addition, the associated seasonal reduction in thermocline depth promotes air–sea coupling, and the shift in the region of deep convection reduces the effectiveness of the SST–cloud–incident shortwave radiation negative feedback (Fig. 5). The northward (southward) alongshore wind anomaly results in a shoaling (deepening) of the thermocline locally, thus accounting for the good correlation between SSTAs and D20As with D20As lagging by a month or so, as seen in Fig. 2b.

The relative weakness of the SSTAs over the eastern tropical Indian Ocean compared with the Pacific may be attributed in part to the relative narrowness of the former basin. It results in a relatively rapid baroclinic adjustment to an equatorial zonal wind stress anomaly. Transit times permit little amplification of equatorial baroclinic waves as they propagate across the basin and only 2–3 months for local coupling to amplify an east IDM pole SSTA before EKWs reflected from LRWs at the western boundary help cancel the opposite-signed EKWs propagating directly to the east; see the schematic paths in Fig. 5c.
Reasons for the secondary peak in SST variability in March, seen in Fig. 2a, are twofold: First, there is a tendency for a SSTA in boreal fall, associated with an IDM event, to decrease in magnitude and emerge as an anomaly of opposite sign the following boreal spring (Saji et al. 1999; Rao et al. 2005); zero magnitude occurs around December (Fig. 4). This tendency accounts for the weak anticorrelation of SSTAs with thermocline
depth anomalies leading by 6 months, as shown in Fig. 2b; correlations stratified by season (not shown) give a peak correlation of boreal spring SSTAs with D20As about 4–5 months earlier. The demise in the boreal fall SSTAs is due to the seasonal reversal of the alongshore winds, which turn the positive SST–wind–evaporation feedback into a negative one and deepen the thermocline, so weakening the coupling between the ocean and atmosphere (Li et al. 2003); see Fig. 5. Also, the intertropical convergence zone (ITCZ) moves southwestward in boreal winter to lie over the east pole of the IDM, so enhancing the negative SST–cloud–shortwave radiation feedback. The sign of the SSTA reverses as the oppositely signed (downwelling for a cool SSTAs) EKWs arrive (Wajsowicz 2004; Vinayachandran et al. 2002). These waves are generated by reflection at the western boundary of LRWs generated by the zonal equatorial wind stress anomaly associated with the east IDM SSTA. As the seasonal cycle progresses into boreal spring and the ITCZ moves northward again, SSTAs to the southwest of Sumatra–Java are less readily damped. Ocean–atmosphere coupling increases
again, as the thermocline averaged over the east IDM pole shoals in response to the seasonal reversal of the equatorial zonal surface wind component; see Fig. 5.

The D20As remain large during boreal winter when the SSTAs change sign; following Balmaseda et al. (1995), if there is a loss of skill in forecasting the weak SSTAs over boreal winter, then the large D20As may enable skill to return.

A second reason for the March peak in variability is due to the other major mode of variability of the tropical Indian Ocean in which SSTs in both halves of the basin vary in unison. Figure 4 shows sizeable SSTAs in spring at the east IDM pole when the IDM index is weak, for example, 1988 and 2000, and which do not follow a major IDM event. These may be attributed to ENSO, which produces tropical tropospheric warming around the globe (Horel and Wallace 1981). Over the tropical Indian Ocean, deep convection is reduced, resulting in basinwide warming of SST about 4 months later, see, for example, Chiang and Sobel (2002); an explanation framed in terms of changes to the Walker circulation, and its impact on the components of surface heat flux, is given by Venzke et al. (2000). These anomalies decay with ENSO, accounting for the dip in variability over boreal summer in Fig. 2a.

2) WEST IDM POLE

In contrast with the above, there is little seasonal modulation of the low frequency SST variability at the west IDM pole (Fig. 2a); there is a slight peak at the onset of the southwest monsoon. In addition, the variability is only about half that at the east IDM pole in boreal fall. The thermocline depth anomalies are largest in boreal fall/winter and a minimum in summer (Fig. 2a). The D20As are weakly correlated with SSTAs for about a season before and a season after (Fig. 2b).

In different years, peak SSTAs occur at different times of the year (Fig. 6) with different mechanisms responsible for their initiation and evolution. Warm (cool) SSTAs develop over boreal summer at the western IDM pole in association with weaker (stronger) than normal southwest monsoons yielding reduced (increased) evaporation and decreased (increased) coastal upwelling along Somalia. They develop in boreal spring as a result of warming aloft due to ENSO, as described for the east IDM pole, and in boreal fall and winter in association with an opposite-signed anomaly at the east IDM pole. In this latter case, the zonal equatorial wind stress anomaly associated with the SST dipole generates westward propagating Rossby waves, which promote the growth of the SSTAs at the western IDM pole (Webster et al. 1999); its demise is essentially due to the demise of the SSTA at the east pole and decay of the equatorial zonal wind anomalies.

The key difference with the east IDM pole region is that the reversing monsoon winds do not modulate the sign of the SST–wind–evaporation feedback or the depth of the thermocline, and hence the strength of air–sea coupling, as effectively over an interior western ocean box as over a coastal eastern box. There is seasonal modulation in the SST–cloud–shortwave radiation negative feedback. It is least effective over the west IDM pole in boreal spring, when the ITCZ is farthest south in the west, and most effective in boreal summer, when deep convection expands northwestward with the evolution of the monsoon; see Fig. 5. This could also help account for the slight peak and dip seen in Fig. 2a.

Figure 6 also suggests that interannual SST variability at the west IDM pole is quasi-biennial. However, the structure is different from that at the eastern IDM pole. SSTAs associated with IDM events in year (0) continue into the next year, year (+1). They decrease and emerge again with the opposite sign continuing into the year after that, that is, year (+2). The period of about 3 yr is suggestive of ENSO forcing and, indeed, through much of the year SST variability at the west IDM pole is well correlated with SSTAs in the eastern equatorial Pacific the previous winter and anticorrelated with them the winter before that (not shown).

c. Lagged correlations (forecasting persistence)

Examination of the time scale over which forecasting persistence of an SSTA remains valid, and when and how it breaks down, helps identify key mechanisms in the anomaly evolution. Lagged correlations of the SSTAs and D20As indices, calculated from observations, for the IDM poles and the Niño-3 region are shown in Fig. 7. As described by Webster and Yang (1992) for the Southern Oscillation index, persistence makes a reasonably good forecast (correlation coefficient 0.6) for the Niño-3 SSTAs for any start month up until the following spring (Fig. 7c). Therefore, a forecast of persistence in July of year (0) is quite good through to March–April of year (1), that is, for more than two seasons. However, a forecast of persistence in March of year (0) is barely good for the next couple of months, and beyond the spring persistence barrier there is no significant return of skill. The tendency for a La Niña to follow an El Niño, and vice versa, accounts for the peak in negative correlations.

The pattern for persistence of the thermocline depth anomalies (D20A) is similar in terms of a spring persistence barrier, but now there is some return of skill signified by correlations of the same sign for starts in
boreal spring and of the opposite sign for fall starts. The return of skill is consistent with the notion that whatever processes destroy the SSTAs in boreal spring do not completely destroy the baroclinic oceanic upwelling (downwelling) LRWs generated directly by the ENSO westerly (easterly) wind anomalies. The increased correlation after the spring decline for early year starts, and the negative correlation for later in the year starts, is due to the continuing arrival of upwelling (downwelling) EKWs generated by reflection of the LRWs at

Fig. 6. Quasi-biennial nature of the SSTA at the west IDM pole from observational data. Each year [denoted year (0)] is plotted with the two following years [year (+1), year (+2)]; years grouped by IDM index magnitude as in Fig. 4. SST anomalies are multiplied by sign of the maximum IDM index in year (0); dark (light) gray denotes positive (negative) IDM index; thick black line denotes the mean of the plotted group. Data detrended and filtered as described in Fig. 2 caption.
Fig. 7. Contour maps of correlation coefficient between initial observed anomaly assumed to persist with time and actual observed anomaly, as a function of start month and target month, for (left) SSTA and (right) D20A for the (a) east IDM pole, (b) west IDM pole, and (c) Niño-3. Absolute correlations greater than 0.9 (between 0.9 and 0.6) shaded dark (light) gray. Contour interval is 0.1. Solid (dashed) lines denote positive (negative) values, and dotted line denotes zero; contour value of 0.6 highlighted by the dot–dash line. Data detrended and filtered as described in Fig. 2 caption.
the western boundary, which led to the demise of the ENSO event (Schopf and Suarez 1988).

1) East IDM Pole

For each start month, persistence of the SSTAs is a reasonable forecast for a season or so unless December is encountered, in which case the anomaly must be assumed to switch sign subsequently (Fig. 7a). Therefore, December presents a possible predictability barrier, but the quasi-biennial nature described in section 2b(1) suggests a return of skill for forecasts initiated in fall and, to some extent, those initiated in spring. Although forecasting persistence of the SSTAs is reasonable until the end of the year for May through July starts, there is a relative lack of a return of skill the following year.

The corresponding correlation plot for persistence of the D20As suggests a spring barrier (Fig. 7a). The subsequent sign reversal of the anomaly correlation coefficient is due to the arrival of oppositely signed EKWs generated by reflection of LRWs at the western boundary during IDM events (Wajsowicz 2004, Fig. 10); compare this with the Niño-3 region and the processes phase locked to the seasonal cycle responsible for the sign reversal of the SSTAs described in section 2b(1). The increased positive correlation in the second fall following the start indicates that the quasi-biennial nature of IDM events extends into the subsurface ocean, which may help in their long lead time prediction. The uniform basinwide STA following ENSO in boreal spring does not generate surface wind stress anomalies and thus does not have a significant subsurface signature; its successful prediction depends on accurately forecasting ENSO and its teleconnection with the Indian Ocean sector climate.

2) West IDM Pole

Forecasting persistence is reasonable for a season or so for initial SSTAs throughout the year. However, correlations then fall off rapidly in late boreal summer/early fall for initial boreal fall and winter SSTAs (Fig. 7b) coinciding with the seasonal deepening of the thermocline there (Fig. 5c). For June–November initial anomalies, skill returns from April through July of year (2) by assuming the anomaly is of the opposite sign from the original, as noted in section 2b(2). For persisted D20As, correlations fall off rapidly in boreal spring for initial anomalies in the first half of the year. In contrast, correlations for boreal fall persisted anomalies remain fairly good through boreal spring, consistent with the continuing arrival of Rossby waves generated by zonal equatorial wind stress anomalies associated with an IDM-type event in the fall. The return of negative skill for persisted summer/fall SSTAs in boreal spring in year (2) is not associated with any return of skill in D20A then. There is a fair return of negative skill for persisted winter D20As in the spring of year (2), consistent with the interpretation that, if the persisted December D20A anomalies were due to an IDM event in the fall of year (0), then there would be D20A anomalies in the spring of year (2) due to the arrival of opposite-signed Rossby waves from an oppositely-signed IDM event in the fall of year(1).

3. The dynamical forecast systems

The NCEP CFS (Saha et al. 2006) is briefly described in section 3a, and differences from the NSIPP system (Wajsowicz 2004) highlighted. Biases in the summer and fall climatologies of the NCEP CFS tropical Indo-Pacific sector, as it drifts from observed initial conditions, are described in section 3b; biases in the annual-mean Indian Ocean sector climate were shown in Wajsowicz (2005b) and biases in the boreal summer and winter climate over the globe are given in Saha et al. (2006). Some aspects are similar to the systematic biases found in the free-running NSIPP coupled system described in Wajsowicz (2004, 2005a), and these are discussed. Retrospective forecasts of SSTAs at the east and west IDM poles from the NCEP CFS for initiations from January 1990 to December 2003 are presented along with similarities with and differences from hindcasts for the NSIPP system (Wajsowicz 2004) for initiations from January 1993 to December 2002 in section 3c.

a. The NCEP coupled forecast system

The NCEP CFS is described in detail in Saha et al. (2006). It is similar to that developed by NSIPP, described in detail in Wajsowicz (2004), in that both consist of coupled 3D numerical ocean and atmosphere models, but the dynamical cores and physics, and numerical formulations, are different, as is the method of ensemble generation; a side by side summary is given in Table 1. The NCEP CFS consists of coupled ocean–atmosphere GCMs and a land model. The atmosphere GCM is spectral in the horizontal with triangular truncation at 62 waves (approximately equivalent to a 200-km grid), and finite differencing in the vertical with 64 sigma layers. The ocean model is the Geophysical Fluid Dynamics Laboratory Modular Ocean Model, version 3 (MOM3), which is cast on a staggered Arakawa B grid in the horizontal with a resolution of 1° zonally and $1/8^\circ$ within 10° of the equator decreasing to $1^\circ$ poleward of 30°. The vertical coordinate is $z$ with 40 levels in the
vertical (27 levels in the upper 400 m). It extends from 74°S to 64°N. The land model is a two-layer surface hydrology model. The ocean and atmosphere models are coupled once a day and exchange daily averaged quantities.

Initial conditions for the NCEP atmosphere are from the NCEP–Department of Energy Atmosphere Model Intercomparison Project II reanalysis (NCEP R-2) data (Kanamitsu et al. 2002); the NSIPP atmosphere was initialized from their AMIP-style runs, which prior to 2003 did not include data assimilation. Initial conditions for the ocean model are obtained from running the ocean model in stand-alone mode forced by weekly fluxes from the NCEP R-2 data with subsurface temperature data assimilated using a 3D variational technique; subsurface salinity is adjusted using a climatological temperature–salinity relationship. The temperature data assimilated includes data from expendable bathythermographs, the Tropical Atmosphere Ocean (TAO) Array of moored buoys, and the Argo network of floats. Sea surface temperature is relaxed on a 90-day time scale to the observed SST [NCEP optimum interpolation (OI) data: Reynolds and Smith (1994)]. The whole ocean system comprises the Global Ocean Data Assimilating System (GODAS); the NSIPP ocean model was also initialized using the same observational

### Table 1. Summary of dynamical forecast systems.

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<th>NCEP CFS</th>
<th>NSIPP (GMAO/CGCM version I)</th>
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<tr>
<td>AGCM</td>
<td>Spectral ’T62–2° grid, 64 sigma levels</td>
<td>NSIPP – 12° × 2.5°; 34 sigma levels (Bacmeister and Suarez 2002)</td>
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<tr>
<td>OGCN</td>
<td>GFDL MOM3 1° × 1/6° within 10° of equator and 1° poleward of 30°; 40 levels (27 in upper 400 m)</td>
<td>Poseidon (reduced gravity) 1/6° × 1/6°; 27 layers (Schopf and Loughe 1995)</td>
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<tr>
<td>Land model</td>
<td>Two-layer surface hydrology model (Mahrt and Pan 1984)</td>
<td>MOSAIC model after SiB model (Koster and Suarez 1996).</td>
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<tr>
<td>Ocean initialization</td>
<td>Real-time subsurface ocean temperature data (GODAS; D. Behringer et al. 2006, unpublished manuscript) assimilated into ocean model forced by weekly surface fluxes from NCEP R-2 data (Kanamitsu et al. 2002) with damping of SST to NCEP’s OI analysis of weekly values and SSS to monthly climatology; subsurface salinity corrected using climatological T–S profiles</td>
<td>Real-time subsurface ocean temperature data (GODAS with substitutions) assimilated using a univariate OI scheme into ocean model forced by SSM/I- and QuikSCAT-derived wind stresses; monthly climatological surface heat fluxes from COADS with damping of SST to NCEP’s OI analysis of weekly values and SSS to monthly climatology</td>
</tr>
<tr>
<td>Atmosphere initialization</td>
<td>NCEP R-2 data</td>
<td>None; AMIP-style run (operational version: NCEP operational forecast)</td>
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<tr>
<td>Coupling</td>
<td>Once per day, daily averaged values exchanged; no flux correction</td>
<td>Once a day using the Goddard Earth Modeling System (GEMS) coupler; land–ocean mask for coupled model defined on ocean’s lat–lon grid; each grid box is either all ocean or all land; atmosphere to ocean coupler: bilinear interpolation from atmospheric grid to mass point of underlying ocean boxes; ocean to atmosphere coupler: underlying ocean grid-box values averaged; land to atmosphere coupler: conservative aggregation of fluxes computed over each land tile. No flux correction</td>
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<tr>
<td>Ensembles</td>
<td>15 ensemble members created by initiating runs from 15 days each month</td>
<td>Six members (operational version: 19 members)</td>
</tr>
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<td></td>
<td>• Daily atmospheric initial states from 9th through 13th day of month with 5-day-averaged ocean initial state for 11th day of month</td>
<td>• Three perturbed ocean states generated from differences between analyzed states chosen randomly within 15 days of forecast initialization time</td>
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<td></td>
<td>• 19th through 23d day of month with 5-day-averaged 21st day of month</td>
<td>• Two perturbed atmosphere states generated by choosing a random pair of ensemble members from an existing up-to-date AMIP run and adding fraction to base state</td>
</tr>
<tr>
<td></td>
<td>• Last 3 days of month and first 2 days of next month with 5-day-averaged 1st day of next month</td>
<td>• Unperturbed</td>
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<tr>
<td>Runs</td>
<td>9 months long, initiated from January 1981 to December 2003; only subset January 1990–December 2005 used</td>
<td>12 months long, initiated on first day of every month from January 1993 to December 2002</td>
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systematic biases that affect the forecasts presented in the next subsection.

1) BOREAL SUMMER

The coupled system is drifting toward an El Niño state in the tropical Pacific (Fig. 8a). There is warming of the SSTA and deepening of the thermocline in the east. The ITCZ, indicated by outgoing longwave radiation (OLR) values less than 240 W m$^{-2}$, remains north of the equator but is shifted eastward, and convection on the equator is increased. The equatorial easterlies are reduced to the west of the convective anomaly, and the surface latent heat flux (into the atmosphere) reduced. To the east, the surface latent heat flux is increased. Over the tropical Indian Ocean, the coupled system is drifting toward a positive IDM state. SSTs are cooling to the south and west of Sumatra–Java as the thermocline shoals. Associated with the cooling is a decrease in the surface latent heat flux and convection. The surface wind changes are consistent with the decrease in convection.

In the western equatorial basin, SSTs are warming as the thermocline deepens. There is increased surface latent heat flux and convection. The “Warm Pool” no longer extends westward from the Maritime Continent. There is a central pool around 60°E. Similar biases were found in the climatology of the century-long run of the free-running NSIPP coupled system reported in Wajsowicz (2004, 2005a). Over the Arabian Sea and Bay of Bengal, SSTs are cooling and the thermocline deepening. There is another center of suppressed deep convection and much-reduced surface latent heat flux to the southwest of India, consistent with the reduction in the southwesterly monsoon flow. Over the southern tropical Indian Ocean (around 20°S) the southeasterlies, and surface latent heat flux, are reduced.

2) BOREAL FALL

The coupled system is drifting toward a transitioning La Niña state in the equatorial Pacific (Fig. 8b). SSTs are cooling over the western and central Pacific as the thermocline shoals, and there is an associated decrease...
in the surface latent heat flux. There is decreased convection over the west Pacific warm pool. The change in surface wind anomalies is consistent with the change in convection. Over the Indian Ocean, the climate is drifting to a weaker positive IDM state than is found in summer. However, the cooling of SSTs over the east IDM pole is sufficient for the warmest SSTs over the Indian Ocean in the 8–9-month lead climatology to be found in the central basin. As reported in Wajdowicz (2004, 2005a), this results in the SST, convective, and wind anomalies generated at the east IDM pole during an IDM event moving slowly westward, reminiscent of the coupled Rossby mode described by Gill (1985) and Hirst (1986).

In general, it is difficult to simulate the observed strength in anomalies associated with a tropical phenomenon in a coupled system in which the background state is biased toward the phenomenon. For example, consider the east IDM pole: if SSTs are too cool for deep convection to occur, it is difficult to trigger a positive IDM event that depends on the cooling of SST, a decrease in deep convection, and an associated surface wind anomaly, which adds to the mean wind to increase evaporation, thus causing the SSTs to cool further. Similarly, it could be difficult to trigger a negative IDM event that requires the warming of SSTs and an increase in deep convection unless the system was close to the threshold for deep convection. The biases due to the drifting climate could be interpreted as the coupled system spinning down after initiation. However, Wajdowicz (2004, 2005a) found similar biases in the climatology deduced from the century-long run of the NSIPP coupled system, running freely, in which ENSOs and IDM events with reasonably realistic frequency and magnitude occurred. An overly shallow thermocline at the east IDM pole led to overly strong ocean–atmosphere coupling, and a lack of deep convection prevented an SSTA from being damped by the SST–cloud–incident shortwave radiation negative feedback.

c. The retrospective forecasts

Forecasts of the SSTAs averaged over the east and west IDM poles at 2–3- and 5–6-month leads are displayed in Figs. 9a,b and 10a,b, respectively, as time series from 1990 to 2004; compare with Wajdowicz (2004) for the NSIPP system from 1993 to 2003. Figs. 9c,d and 10c,d. Major positive IDM events, as defined by Saji and Yamagata (2003), when the eastern tropical basin cools and the west warms, occurred in 1994 and 1997. Major negative IDM events, when the eastern tropical basin warms and the west cools, occurred in 1992, 1993, and 1996. For reference, forecasts of the Niño-3 index are plotted in reduced size for each model; El Niños occurred in 1991–92 and 1997–98 and La Niñas occurred in 1998–99 and 1999–2000. The difference in spread given by the ensemble members between the modeling systems is expected from the greater number of members for the NCEP CFS for the time period considered. In both models, the spread is much greater for the Indian Ocean indices than for Niño-3; few data were assimilated into the Indian Ocean prior to 2003 to generate the initial conditions.

1) East IDM pole

At 2–3-month lead in the NCEP CFS, the cold SSTA associated with the positive IDM event in 1994 is larger than observed (Fig. 9a). There is also a slight delay in forecasting the onset of major coolings, which may be related to the boreal summer biases discussed above. Looking at only the east pole index, false alarms of a positive IDM event would have been forecast in 1991, 2000, and 2004. The NSIPP system (Fig. 9c) captured the timing and magnitude of the major coolings in 1994 and 1997 reasonably well.

At 5–6-month lead in the NCEP CFS (Fig. 9b), the magnitude and timing are only a little worse. However, the 1990s are plagued with false alarms of a similar character to those in the NSIPP system (Fig. 9d), namely, extreme coolings prior to the 1997–98 events in boreal fall, and extreme warmings following the 1997–98 events. Investigation of the ensemble mean fields for forecasts initiated in May (not shown) indicate that for the cold alarms in the early 1990s, prevailing El Niño-like conditions are responsible for cool SSTAs over the western Pacific and Maritime Continent. In observations, they never encompass the Maritime Continent south of Java and have all but decayed by July. However, in the forecast system, the cool SSTAs are more extensive over the Maritime Continent, and importantly south of Java and under the South Pacific convergence zone (SPCZ), and persist into July. Meanwhile in the western tropical basin, warm SSTAs are both observed and forecast until June. They are no longer present in observations in July, but in the forecast system they have started to grow. By July with the east and west IDM poles anomalously cool and warm, respectively, and with associated convective and surface wind anomalies, the system is ready to spawn a positive IDM event in the fall.

For the late 1990s, La Niña-like conditions prevail over the tropical Pacific in May with warm SSTAs over the western equatorial Pacific and Maritime Continent. SSTs are anomalously cool over much of the tropical Indian Ocean. These Indian Ocean anomalies decay by July in observations, but persist in the forecasts initiated in May. Once again, there are convective and wind
anomalies associated with the SSTAs such that the system, initiated in May, by July is set to forecast a negative IDM event in the fall.

2) West IDM Pole

With the exception of the cooling associated with the negative IDM event in 1993, major SSTAs at the west IDM pole are reasonably well forecast at both 2–3- and 5–6-month leads (Figs. 10a and 10b); there is a suggestion at 5–6-month lead that false warmings are beginning to emerge in the forecast in the early 1990s and coolings following the 1997–98 event, as discussed above. There is a slight delay in forecasting the onset of the warming associated with the 1997–98 positive IDM event, which could be related to an increasing delay in forecasting the 1997–98 El Niño, but a similar problem was found in the NSIPP system (Figs. 10c and 10d) even though the timing of the El Niño was well forecast. Therefore, it is more likely due to the systematic bias in the boreal summer climatology discussed above.

4. Seasonal forecast skill

Skill in the NCEP CFS is measured by calculating the correlation between the ensemble mean anomaly and the observed anomaly for each of the 9 months of the retrospective forecast as a function of the start month; for the NSIPP system, this is done for each of the 12 months of the hindcasts. Historically, forecasts yielding correlations above 0.6 are considered skillful (Hollingsworth et al. 1980).

For the Niño-3 index over the period 1990–2003, the NCEP CFS yields correlations above 0.6 for start months from February to August for each of the following 9 months (Fig. 11). For September through January starts, correlations fall below 0.6 for forecasts beyond April irrespective of start month. This is the “spring predictability barrier,” which corresponds to the rapid decrease in lagged correlations (Fig. 7c). Encouragingly, the dynamical forecasts beat the assumption of persistence for every month. The decline in the forecast skill of the thermocline depth anomalies, mea-

![Fig. 9. Retrospective forecasts of the SSTA at the east IDM pole from NCEP CFS at (a) 2–3-month and (b) 5–6-month lead for starts initiated from January 1990 to December 2003, and from NSIPP system hindcasts at (c) 2–3-month and (d) 5–6-month lead for starts from January 1993 to December 2002; corresponding plots for the Niño-3 index given beneath. Ensemble mean denoted by the thick black line and 95% confidence interval (1.96 std dev) provided by ensemble members shaded light gray. Verification (thick midgray line) from GODAS for NCEP CFS and NCEP OI v.2 (Reynolds et al. 2002) for the NSIPP system; two std dev (dotted line). Positive (thin solid) and negative (thin dot–dash) IDM events, as defined by Saji and Yamagata (2003), marked through October of the given year. No detrending or filtering performed.](image-url)
sured by the anomaly in the depth of the 20°C isotherm, lags that of the SSTA for forecasts initiated in boreal winter, so enabling the coupled system to recover slightly, as described by Balmaseda et al. (1995). For boreal summer and fall starts, skill declines in forecasting thermocline depth anomalies over the Niño-3 region in late boreal winter ahead of SSTAs in spring, thus contributing to the spring predictability barrier; compare with Fig. 2b.

a. East IDM pole

Correlation between retrospectively forecast and observed anomalies is above 0.6 for a season or so for most start months (Fig. 11). It is as long as 5 months for a May or June start, but only 2 months for an April start; for January and February starts skill is lost almost immediately, but returns briefly. There is a late boreal spring/early summer predictability barrier for SSTA forecasts initiated in boreal fall through early spring, which is much more pronounced and affects a wider range of start months than the SSTA persistence barrier (Fig. 7a); the similarity in part could be due to the “observed” D20A in Fig. 7a being generated by GODAS. The ocean–atmosphere coupling strength decreases with deepening of the thermocline during the monsoon transition season (Fig. 5), so the coupled system cannot recover fully. There is return of weak skill for boreal spring starts as the coupling strengthens again in late summer.

As an illustration of the seasonal variation in forecast skill, the predicted anomalies for July 1997, the onset of a major ENSO and a major positive IDM event, are shown in Fig. 12a, for forecasts initiated in May 1997 and November 1996. As expected from Fig. 11 (see also Fig. 9a), warming over the eastern equatorial Pacific Ocean and cooling to the west and south of Sumatra–Java are well captured by the forecasts initiated in May (2–3-month lead). However, the forecast surface wind anomalies over the eastern tropical Indian Ocean are weaker than observed. Inspection of the forecast OLR and surface latent heat flux anomaly fields indicates that poor simulation of the warming and increased latent heating over the Bay of Bengal are associated with the too weak southeasterly wind anomaly to the west of Sumatra. Over the equatorial Pacific, problems with the position and magnitude of the latent heating are emerging. Also, extensive cooling over the Maritime Conti-
A dynamical forecast system simulates the quasibiennial nature of the SST variability well, as it captures the reduction in magnitude of the SSTAs in December and their reemergence the following year with opposite sign, as discussed in section 2b(1). For forecasts initiated in boreal summer, there is a sharp decline in skill over boreal winter, but skill returns by spring. Thermocline depth anomalies are well forecast through boreal winter (Fig. 11) across the tropical Indian Ocean, which suggests that the ocean’s memory helps the coupled system recover in late winter, as the ocean–atmosphere coupling strength increases with the seasonal shoaling of the thermocline (Fig. 5c).

Calculations for the NSIPP dynamical forecast system for hindcasts initiated from January 1993 to December 2002 start months (Fig. 13) also have a problem in predicting fall SSTAs unless the hindcasts are initiated after May. However, as the hindcasts are 12 months long, the return of skill in forecasting the No-
November and December anomalies from forecasts initiated in March is evident; there is also a hint of a return of weak skill for hindcasts initiated from November through February. In section 5 on potential predictability, under the perfect model/ensemble assumption, the return of skill correlations exceed 0.6, and possible reasons for the collapse of the predictability barrier are discussed.

The NSIPP system is relatively poor at forecasting anomalies around January irrespective of start month (cf. Figs. 11 and 13), which corresponds to the sharp reduction in lagged correlations of the observed anomalies (Fig. 7a). However, once again, the dynamical forecast system simulates the quasi-biennial behavior well and skill returns in late winter through spring.

The ability of the NSIPP system to capture the onset of the major coolings at the east IDM pole at 3-month lead (Fig. 9c) is reflected in the high skill scores (> 0.9) for the prediction of fall SSTAs from July onward in Fig. 13. The NCEP CFS shows little seasonal phase locking of skill in this regard; skill just decreases with increasing lead time for the first season.

b. West IDM pole

Forecasting SSTAs at the west IDM pole is more robust than at the east IDM pole, as expected from Fig. 10. Correlations are above 0.6 until mid-June is encountered (Fig. 11). Ocean–atmosphere coupling weakens then due to the deepening thermocline (Fig. 5c); however, skill in prediction of thermocline depth anomalies has already declined in April. These predictability barriers correspond to the persistence barriers in Fig. 7b, but occur a month or so earlier and are more obvious for the boreal summer start months.

From the Fig. 12a conditions in July 1997 forecast from May 1997, the observed widespread warming over the western tropical Indian Ocean is not predicted, although the southeasterly surface wind anomaly, albeit weaker, is predicted; see also the delay in predicting the onset of the warming in Fig. 10a. The observed warming is thought to be due to the arrival of downwelling equatorial Rossby waves generated by the easterly equatorial wind stress anomaly and anomalous surface heat fluxes due to a weaker than usual southwest monsoon (Yu and Rienecker 2000; Murtugudde et al. 2000). The D20As are not well predicted (Fig. 11), and a broad band of increased surface latent heat fluxes to the south of the equator is predicted in contrast to the observed major reduction to the north and east of Madagascar. As found for the east IDM pole, there is no obvious skill in the prediction of any of the anomaly fields for forecasts initiated 8–9 months earlier in November 1996.

In contrast, the continuing warming over the western tropical basin in December 1997 is well forecast from October 1997 (Fig. 12b). The decreased surface latent heat flux and increased convection are also captured, though the surface wind anomalies are too weak and zonal, which is consistent with the too equatorial east-
ern anomalies. For forecasts initiated earlier, in April 1997, only a very weak warming over the western basin is predicted in line with weaker eastern Indian and Pacific anomalies. Conditions to the east of Madagascar—warming of SSTs, surface northwesterlies, reduced surface latent heat flux, and increased convection—continue to be predicted.

An early boreal summer predictability barrier is also evident in the NSIPP system (Fig. 13) with the seasonal phase locking even more apparent. The barrier is reduced to a temporary weak loss of skill around October in the NSIPP potential predictability calculations of the next section suggesting that the problem is related to systematic biases in the simulation of the boreal summer climatology. NSIPP’s 12-month forecasts confirm the return of some skill in midfall.

The NSIPP system also indicates that a dynamical forecast system is capable of simulating the biennial nature of the variability over the western basin to the extent that, when the lagged correlations of observed anomalies are negative (Fig. 7b), the forecast correlations are positive; there is even skill for December anomalies forecast 12 months earlier.

5. Potential predictability

Even with a perfect coupled model, exact initial conditions, and the optimum choice of ensemble members, a climate phenomenon still may not be predicted with confidence due to the system’s chaotic nature. An assessment of its potential predictability can be made by assuming that the forecasting system is perfect, then taking each ensemble member in turn as truth, and comparing it with the ensemble mean of the remaining members. High skill scores indicate that the phenomenon is predictable, or that the ensemble is biased. Low scores suggest a lack of predictability. A significant increase in the skill scores calculated by comparing the forecasts with real observations indicates the scope for improvement in the forecast system.

From the NCEP CFS, the potential predictability of Niño-3 SSTAs over the last 14 years is good at lead times up to three seasons (Fig. 14) with scores dropping to around 0.8 only for boreal summer SSTAs forecast from winter starts (as also reported in Saha et al. 2006). The decline in potential predictability of SSTAs is preceded by a decline in that of the thermocline depth (as...
given by D20A). The spring predictability barrier is evident in the D20A map with skill dropping rapidly after April for boreal fall starts.

a. East IDM pole

Over the eastern tropical Indian Ocean, prediction of boreal fall SSTAs from the start of the year appears limited by a late boreal spring predictability barrier (Fig. 14); compare with the observed persistence barrier in Fig. 7a. There is a corresponding, but even sharper, decline in D20A predictability. The boreal winter decline in SSTA predictability, corresponding to the winter persistence barrier, with a return of skill by spring is evident for forecasts initiated in the second half of the year. The thermocline depth remains predictable through this time, which supports the notion that the return of predictability of SSTAs is due to the ocean’s memory. There is only a temporary loss of skill in SSTA prediction as the ocean–atmosphere coupling weakens over boreal winter due to the deepening seasonal thermocline (Fig. 5c).

The winter (January) persistence barrier, so evident in the hindcast skill map (Fig. 13) for the NSIPP system, is present again in the potential predictability map (Fig. 15), but is shifted to February and is much weaker. Evidence of a late spring predictability barrier is also found. However, the NSIPP results for 12-month hindcasts suggest that the problem may just be limited to boreal winter starts and that skillful forecasting of fall anomalies could be achieved from as early as March in the same year and from the previous boreal fall. Figure 15 shows a return of skill in the following boreal summer for fall starts, as noted for the hindcast–observation anomaly correlations plotted in Fig. 13. These results suggest that the observed quasi-biennial nature of SST variability at the eastern IDM pole (Fig. 4) is robust to the processes causing the loss of skill in late spring.

The quasi-biennial nature of SST variability over the tropical Indian Ocean is attributed to the crisscrossing of equatorial baroclinic ocean waves of opposite sign generated by the zonal equatorial wind stress anomaly associated with the original SST anomalies (see, e.g., Shinoda et al. 2004; Rao et al. 2005). The return of skill for fall, but not winter, starts suggests that the tropical Indian Ocean is better initialized in fall when SST and wind stress anomalies are strong. They generate strong thermocline depth anomalies due to the shallow seasonal thermocline (Fig. 5c). In winter, the SST and wind stress anomalies are typically weaker, as the SSTA at the east IDM pole is changing sign (Fig. 4). Only weak thermocline depth anomalies are generated, as the seasonal thermocline is deep; thus the ocean–atmosphere coupling is weak. A caveat to this argument is that predictability of the thermocline depth also declines in late spring in the east and a month or so later in the west (Fig. 14). A mechanism needs to be invoked, similar to that proposed by Balmaseda et al. (1995), whereby the staggering of the decline in predictability in SSTA and D20A at each IDM pole and between poles enables the coupled system to recover when it has been well initialized.

b. West IDM pole

The potential predictability of SSTAs in the western tropical Indian Ocean is high (Fig. 14) except for the prediction of late summer and boreal fall anomalies from start months in boreal winter and early spring. A decline in the potential predictability of thermocline depth anomalies precedes that in SSTAs by a month or so and is typically more severe. However, there is a weak return of skill, which could be attributed to the coupled dynamics as noted above for the east IDM pole.

The NSIPP system results (Fig. 15) suggest that SSTAs at the west IDM pole could be skillfully forecast
at lead times up to 12 months for all start months. There is some decline around October associated with the seasonal deepening of the thermocline (Fig. 5c), but the coupled system recovers; lagged correlations of the observed anomalies declined at the end of the summer monsoon (Fig. 7b).

The contrast in potential predictability for the two systems could be due to ensemble bias with the fewer ensemble members in the NSIPP system (6 versus 15 in the NCEP CFS) giving overly optimistic results. Comparison of both systems with the corresponding prediction–observation skill maps suggest that the problem of predicting anomalies in boreal summer and early fall may be overcome with model improvement, for example, reducing the biases in the annual mean and seasonal climatologies shown in Wajsowicz (2005b) and Fig. 8 for the NCEP CFS and in Wajsowicz (2004, 2005a) for the NSIPP system, respectively. However, reducing biases in coupled systems with such large numbers of degrees of freedom is difficult, as attempting to reduce one problem can emphasize or introduce another.

6. Summary and discussion

Seasonal modulation of the interannual SST variability over the eastern tropical Pacific Ocean results in peak variability in December and a minimum in boreal spring (Fig. 2a). This minimum coincides with a “predictability barrier” in many dynamical forecast systems, whereby irrespective of the start month, the forecast skill rapidly declines in boreal spring (Fig. 11); there may be some return of skill in fall. A similar decline occurs in lagged correlations of observed SSTAs (Fig. 7c), which is equivalent to predicting persistence; moderate negative correlations found 18–24 months later simply reflect that a La Niña typically follows an El Niño on that time scale. The co-occurrence and seasonal phase locking of these barriers is thought to be due to the Walker circulation over the Pacific being at its weakest in boreal spring and the coupled ocean–atmosphere system being most unstable, so the system is most susceptible to noise and external interference.

To investigate the existence of persistence and predictability barriers over the tropical Indian Ocean, SSTAs indices are constructed (cf. the Niño-3 index), which capture the two major modes of interannual variability. The east and west poles of the IDM index, as defined from observations by Saji et al. (1999), are chosen following Wajsowicz (2004). In section 2, the peak in observed SST variability over the east IDM pole is found to occur in September with a secondary maximum in March (Fig. 2a); there is a sharp decline in December. Over the west IDM pole, the observed SST variability is found to be almost independent of season (Fig. 2a). The east pole results are attributed to IDM-like variability, which is strongly seasonally phase locked with SSTAs of opposite sign occurring in boreal spring following the fall anomaly peak; the sign switching occurs in midwinter. An additional contribution to the boreal spring variability is given by the basinwide response to ENSO’s effect on tropospheric temperatures around the globe. The west pole results are due to the IDM-like variability, giving SSTAs lasting from fall through spring. ENSO teleconnections as for the east pole, and over boreal summer due to variability associated with the southwest monsoon. As for the Pacific, there is a close link between SST and thermocline depth anomalies (Fig. 2b), which is essential for forecasting SSTAs beyond a couple of weeks.

As expected, lagged correlations of observed SSTAs (Figs. 7a and 7b) show declines over the monsoon transition seasons, but only weakly at both poles in late spring for initial boreal winter SSTAs, and only at the west pole in early fall for initial boreal fall and winter SSTAs. There is a sharp decline in correlation in midwinter at the east pole, corresponding to the sign switching of the SSTAs associated with IDM-like variability. Persisted thermocline depth anomalies (Fig. 7a) remain well correlated with observed values through winter, suggesting that this persistence barrier will not yield a strong predictability barrier.

Before presenting results on predictability barriers given by the dynamical forecast systems (see section 3), the NCEP CFS dynamical forecast system is described briefly and differences from the NSIPP system, described in Wajsowicz (2004) highlighted (summarized in Table 1). The biases in the NCEP CFS seasonal climatologies (boreal summer and fall; see Fig. 8), which develop as the forecast proceeds past a season or so, are presented. In summer, the Indian Ocean sector drift resembles that of a positive IDM event and that in the Pacific sector, an El Niño. For the same lead time, the drift is weaker in other seasons. Time series of the predicted SSTAs at one and two seasons lead time (Figs. 9 and 10) show that the major warmings and coolings at the poles are reasonably well predicted at 2–3-month lead with no major anomaly predicted when none was observed. However, at 5–6-month lead time, false alarms are raised, notably at the east pole, through the 1990s, with a similar character to those reported in the NSIPP system (Wajsowicz 2004). The positive IDM event alarms are associated with biases in simulating the prevailing El Niño conditions prior to 1997, and the negative IDM event alarms with biases in simulating the prevailing La Niña conditions post-1998.
Forecast skill as a function of start month and target month for the two systems is described in section 4. Both systems struggle to beat persistence with their SSTA forecasts at the IDM poles over the first couple of months (Figs. 11 and 13). However, as the forecasts proceed, they start to do better than persistence, most notably through boreal winter and beyond at the east IDM pole due to the skillful forecast of thermocline depth anomalies. Both systems display a late boreal spring/early summer predictability barrier at the IDM poles, with that at the west pole occurring about a month later. Staggering in loss of skill between the poles and between the surface and subsurface is important for the coupled system, as it may enable some skill to be recovered; compare with ENSO (Balmaseda et al. 1995). Any return of skill in the spring or fall could be attributed to ENSO, when SSTA s over the eastern equatorial Pacific and tropical Indian Oceans are well correlated (Baquero-Bernal et al. 2002). However, the return of skill in the NSIPP system in the fall at the east pole for fall starts in the previous year, and not winter starts, suggests that good initialization of the subsurface Indian Ocean is needed with information about fall conditions. Similarly, the return of skill at the east pole in spring can be linked to the continuing skillful prediction of the thermocline depth anomalies. However, ENSO’s influence is not ruled out, especially at the west pole where climate drift leads to a much closer correlation between SST variability there after a couple of seasons than is found in practice over boreal summer (not shown); the correlation between ENSO and SST variability at the east pole remains similar to observations, as does the correlation between the poles. To illustrate the seasonal variation in skill, anomaly fields for the target months of July 1997, the onset of a major El Niño and a major positive IDM event, and December 1997, when the both have matured, are shown in Fig. 12 for different lead times.

Finally, in section 5, results on the potential predictability of the tropical Indian Ocean climate are presented assuming a “perfect model/ensemble” (Figs. 14 and 15). Care is necessary so as not to give an overly optimistic interpretation, as high skill scores could arise from a biased ensemble. However, the calculations suggest that skillful forecasting of SSTs over the western tropical basin is achievable for all months of the year from forecasts initiated in any month of the year from good initial conditions; the early boreal summer predictability barrier in Figs. 11 and 13 may be caused in part by the climate drift shown in Fig. 8a. Over the eastern tropical basin, the calculations suggest that the problem in forecasting boreal fall anomalies from boreal winter starts may be insurmountable even with a well-initialized ocean. However, they may be able to be skillfully forecast from the previous fall.

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