East African Rainfall and the Tropical Circulation/Convection on Intraseasonal to Interannual Timescales

CHARLES C. MUTAI

Kenya Meteorological Department, Nairobi, Kenya, and
Cooperative Institute for Mesoscale Meteorological Studies, University of Oklahoma, Norman, Oklahoma

M. NEIL WARD

Cooperative Institute for Mesoscale Meteorological Studies, University of Oklahoma, Norman, Oklahoma

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ABSTRACT

It is known that the East African short rains (October–December, OND) have a positive correlation with El Niño–Southern Oscillation (ENSO). A prediction scheme based on sea surface temperature eigenvectors (SST EOFs) including Pacific, Indian, and Atlantic Ocean variability showed skill higher than one based on ENSO alone. First the authors assess the extent to which the large-scale SST predictors correlate with rainfall averaged over smaller subregions of East Africa. Most regions correlate consistently well, though some pockets of lower correlations suggest that interaction with orographic features may modulate the large-scale ENSO and other coupled ocean-atmosphere signals in the region.

Next, to evaluate the atmospheric teleconnections giving rise to the SST correlations, the circulation anomalies associated with East African rainfall are investigated using National Centers for Environmental Prediction–National Center for Atmospheric Research reanalysis data and, as a proxy for large-scale tropical convection anomalies, outgoing longwave radiation (OLR). In the seasonal OND mean, a sequence of three horseshoe structures are evident in the OLR anomalies, with anomaly sign in phase over the central Pacific and East Africa, and out of phase over the Maritime Continent. This gives rise to the positive ENSO association with East African rainfall. The horseshoe structures in the Indian Ocean are absent in September and through much of the long rains in March–May, though weakly evident again in May. It is suggested that the presence or absence of this teleconnection structure is related to the state of the background annual cycle. When the ENSO variance is removed (by linear regression) from the datasets, there emerges more strongly a positive correlation between East African rainfall in OND and enhanced convection through equatorial Africa and into the equatorial Atlantic and Amazon region, in turn associated with warm equatorial and tropical South Atlantic SST.

The lead–lag structure of intraseasonal teleconnections with East African rainfall suggests that 5–10 days before the rainfall event, low-level dynamics start to develop in the equatorial Atlantic, and these penetrate across equatorial Africa and into East Africa during the event itself, at which time anomalies in the Pacific Ocean, which were strong 15 days before the event, are now weaker. Five days after the rainfall event, 200-hPa divergence over East Africa pulls off the east into the Indian Ocean and shows structures that resemble the Madden–Julian oscillation. For above-normal rainfall in East Africa, the seasonal mean teleconnection across the Indian Ocean resembles this intraseasonal picture with strong convection particularly just off the East African coast, prompting discussion of the interaction between the intraseasonal and seasonal anomalies.

1. Introduction

Rainfall in East Africa mostly occurs during the boreal spring (long rains, March–May) and autumn (short rains, September/October–December) seasons as the intertropical convergence zone (ITCZ) migrates through the equator from south to north, and vice versa. Like other tropical regions, interannual variability of rainfall in East Africa results from complex interactions of forced and free atmospheric variations. These include interactions between sea surface temperature (SST) forcing, large-scale atmospheric patterns, and synoptic-scale weather disturbances including monsoon and trade winds, persistent mesoscale circulations, tropical cyclones, subtropical anticyclones, easterly/westerly wave perturbations, and extratropical weather systems.

Relative to the long rains, the short rains tend to have stronger interannual variability, stronger spatial coherence of rainfall anomalies across a large part of the region, and a substantial association with El Niño–Southern Oscillation (ENSO) (e.g., Ropelewski and
These studies described that warm events of ENSO tend to be associated with above-average rainfall and cold events of ENSO are associated with below-average rainfall during the short rains. However, much remains to be understood about the global, regional, and local circulation mechanisms that generate rainfall anomalies in East Africa and how they are connected to ENSO. Hastenrath et al. (1993) showed a particularly close association between near-surface zonal wind anomalies in the equatorial Indian Ocean and East African rainfall, an association stronger than that with ENSO. There is a need to assess if such an enhanced correlation results from a local response to increased East African rainfall, or whether it is part of a large-scale ocean–atmosphere teleconnection structure partially independent of ENSO. Indeed, some severe droughts and floods observed over parts of East Africa appear to have had very little to do with ENSO. Some of the local circulation features associated with droughts and floods have been studied, emphasizing the role of local near-surface westerly wind anomalies in above-normal rainfall seasons (Nakamura 1969; Anyamba 1984; Davies et al. 1985) and intraseasonal wet spells (Camberlin and Wairoto 1997; Camberlin 1997; Okoola 1999). However, how these features relate to structures in the global Tropics on intraseasonal to interannual timescales is not known and is addressed in this paper.

Anomalous SST in the Atlantic and Indian Oceans has also been associated with rainfall fluctuations in eastern Africa (Reverdin et al. 1986; Nicholson and Entekhabi 1987). Predictive schemes based solely on ENSO do show skill in the region (Farmer 1988; Hutchinson 1992), but two of the three predictors used in Mutai et al. (1998) were largely independent of traditional ENSO measures, including aspects of SST variability in the northwest tropical Pacific–Indian Ocean and in the equatorial–tropical South Atlantic. There is a need to diagnose the teleconnection circulation structures associated with these SST patterns as well as ENSO, to gain a better understanding of the SST–rainfall links, and to underpin confidence in prediction schemes based on SST forcing.

This paper is sensitive to the issues of temporal and spatial scales of climate anomalies. Traditionally, climate research in Africa has tended to focus on seasonal mean area-average rainfall totals, guided partly by available data and partly by a desire to diagnose the large-scale relations as a first step. However, it is now clear that for a full understanding and a maximum predictive utility, there is a need to study spatial and temporal variability within a season. Inspection of daily rainfall sequences for Kenya reveals a tendency for a season to be made up of a few isolated rainfall events combined with a few more major rainfall spells (5–10-day duration), punctuated by dry spells. As a first step toward better understanding the nature and role of the major rainfall spells, this paper diagnoses the global Tropics circulation before, during, and after the rainfall spells. The primary aim is to assess how the intraseasonal teleconnection structure relates to the seasonal atmospheric teleconnection structure. A secondary aim is to assess if there is any potential to foresee the likelihood of major rainfall spells a week or two in advance by monitoring circulation in the global Tropics (e.g., Waliser et al. 1999). The study is undertaken against the backdrop of the known strongest signal at the intraseasonal timescale in the Tropics, the Madden–Julian oscillation (MJO; Madden and Julian 1971, 1994), which propagates eastward around the global Tropics with a spectral signature of around 30–60 days, rather shorter and more coherent in atmospheric circulation and rather broader and longer in tropical convection (Salby and Hendon 1994). Thus, this paper generates some unique lead–lag results by focusing on the coherent short rains season over East Africa (centered on October–November) to construct intraseasonal global Tropics signals, at a time of the year when most convection is close to the equator from Africa to the western Pacific.

The East African region contains some of the most varied topography in the world, including large lakes, rift valleys, and snow-capped mountains on the equator. This heterogeneity gives rise to dramatic variations in climatological mean rainfall totals. In addition, seasonal rainfall anomalies can tend to have a coherence that is confined across small subregions. It is postulated that some of this is generated by interaction of the large-scale regional atmospheric forcing with the topography (Sun et al. 1999a, b). It is therefore possible for large-scale atmospheric changes associated with, for example, ENSO to have a stronger signal in some parts of the region. This is very difficult to prove statistically and likely requires a combination of statistics with regional numerical modeling. Nonetheless, we provide in this paper an indication of the extent to which the large-scale forcing is manifest in the observed rainfall anomalies of smaller subregions. For this, we use the eight homogeneous rainfall anomaly subregions defined recently by Indeje et al. (1999) on the basis of empirical orthogonal function (EOF) and point correlation analysis. These results are intended to serve as benchmarks for comparison with numerical model results and provide a practical indication of the extent to which teleconnections with SST are reflected in local regions.

The fundamental goal of this study is to improve existing climate monitoring and forecasting in East Africa. First, the study addresses the representativeness of the predictability of the large-scale area average at smaller spatial scales. Second, the study diagnoses atmospheric circulation structures associated with ENSO and non-ENSO variability, with a view to establishing the physical basis for remote teleconnections with SST and therefore improving reliability and confidence in SST-based prediction schemes. Third, the paper aims to make a contrast with the teleconnection structures as-
associated with the long rains, to provide a basis for why
the ENSO links in that season are weaker. Finally, the
study takes a first look at the role of extended wet spells
in East Africa in the seasonal rainfall anomalies and
associated teleconnection structures. This is intended to
shed light on the potential for anticipating intraseasonal
rainfall events in East Africa.

In section 2, the data sources are given in more detail,
along with an outline of analysis procedures. Section 3
focuses on the monthly/seasonal East African rainfall
and tropical circulation/convection anomalies including
structures that are independent of ENSO. In section 4,
the intraseasonal lead–lag structures of wind fields in
the global Tropics are defined using 5-day East African
rainfall as the host index. Section 5 reevaluates the SST
predictors used in Mutai et al. (1998), by assessing the
atmospheric structures associated with these SST pre-
dictors and comparing them to the results in sections 3
and 4.

2. Data and analysis procedures
a. Rainfall data

The grid-box monthly rainfall database from the Cli-
mate Research Unit (CRU) with a resolution of 2.5°
latitude × 3.75° longitude is utilized in this study.
Hulme (1994) described in detail the dataset and its
quality control. The results are complemented using
monthly station rainfall data for East Africa and daily
data for Kenya kindly made available by the Drought
Monitoring Centre, Nairobi, Kenya. The monthly re-
cords were available for 64 stations across East Africa
(1961–90) as used in Indeje et al. (1999) for subregional
rainfall indices. Daily values for 20 stations in Kenya
(1979–96) were used to calculate the pentad (5 day)
rainfall totals.

b. Outgoing longwave radiation data

The outgoing longwave radiation (OLR) data are
from the Advanced Very High Resolution Radiometer
of the National Oceanic and Atmospheric Administra-
tion (NOAA) operational polar orbiting satellites. The
data are available from June 1974 with a gap in 1978
due to a fault on the NOAA-5 satellite. Resolution is
2.5° latitude × 2.5° longitude. The derivation of the
OLR dataset is well documented (e.g., Gruber and Krue-
ger 1984; Liebmann and Smith 1996). Additional in-
formation on changes in instrumentation, equator cross-
ing times, and other inherent biases that have largely
been corrected for are found in Chelliah and Arkin
(1992) and Waliser and Zhou (1997). The pentad OLR
dataset is only available from 1979.

The sensitivity of OLR at the top of the atmosphere
to small changes in convective cloudiness provides a
good estimate of relative changes in the distribution of
heat sources and sinks in the tropical atmosphere, and
associated vertical motion and precipitation (Gill 1980).
The tropical sparse station rainfall data suffer a host of
problems such as station relocations, instrument chang-
es, changes in time of observation, different lengths of
observation, missing observations, and many others. On
the other hand, the OLR data must be corrected for
different local equatorial satellite crossing times, water
vapor window channels, and instruments. Sometime it
is difficult to isolate regions of active convection from
regions of inactive thick cirrus clouds. We choose here
to use OLR in conjunction with station rainfall and cir-
culation fields in order to identify regions where datasets
agree and are physically consistent, leading us to some
certainty in results.

c. Wind data

The National Centers for Environmental Prediction–
National Center for Atmospheric Research (NCEP-
NCAR) reanalysis data are used to show large-scale
tropical circulation anomalies associated with intrasea-
sonal and interannual rainfall variability in East Africa.
The reanalysis dataset is generated using a state-of-the-
space assimilation scheme and numerical model
(Kalnay et al. 1996). This Climate Data Assimilation
System employs T62 horizontal resolution (about 210
km, interpolated to a 2.5° latitude × 2.5° longitude grid)
and 28 vertical levels. Most results are reported for 850-
and 200-hPa though 700-hPa winds have also been ex-
tensively studied for their consistency with results at
850 hPa. The wind fields were assessed to have a high
confidence rank by the authors of the reanalysis dataset
(Kalnay et al. 1996). Although the reanalysis data are
available for a longer record, we focus on 1974–96, the
period in common with the OLR data (unless otherwise
specifically mentioned).

d. Southern Oscillation and Niño-3 SST indices

The Southern Oscillation index (SOI) measures the
Tahiti MINUS Darwin sea level pressure anomaly
(Troup 1965). Some results are complemented using the
Niño-3 sea surface temperature index as derived from
Rasmussen and Carpenter (1982). Monthly time series
were obtained from the Climate Prediction Center,
Washington, D.C.

e. Analysis procedures

The data are processed in a manner similar to Mutai
et al. (1998) for the creation of monthly as well as
seasonal standardized regional rainfall/OLR indices.
The station rainfall values were used to calculate stan-
dardized indices for the smaller subregions (Fig. 1b)
deﬁned by Indeje et al. (1999), while large-scale indices
were based on grid-box rainfall (CRU) and OLR across
East Africa. The large-scale indices are used in section
3 to calculate the correlation patterns of wind and trop-
Fig. 1. (a) Map showing the location of East Africa. (b) Map showing the regionalization of East Africa defined by Indeje et al. (1999). (c) Standardized Oct–Dec rainfall (solid line, 1961–96) and OLR (dotted line, 1974–96) anomaly index across East Africa.

Phenological convection in relation to East African rainfall. A regression analysis is also applied to remove the variance that is in common with the SOI from each grid-box time series in each dataset. The SOI is in a similar way removed from the East Africa rainfall (OLR) index and a correlation analysis repeated on the (now ENSO independent) residuals.

Point statistical significance for correlations is assessed using the standard $t$ test. For monthly/seasonal interannual series, a full degree of freedom is assumed since rainfall contains no significant lag one serial correlation (Mutai et al. 1998). However, analyses at the intraseasonal timescale are potentially hampered by a reduction in degrees of freedom. The October–November intraseasonal analyses reported here are based on 18 yr, each with 12 pentads in the October–November period, giving rise to time series of length 216. The area-average pentad OLR anomaly for East Africa con-
The study focuses on the October–November months, which contains a lag-1 serial correlation of 0.33, dropping to near 0 at lag 2. We investigated the lag-1 serial correlation of pentad wind anomalies for the October–November season. The $u$ wind close to the equator contained the largest serial correlation, averaging at 0.35. Even in the region of largest lag-1 correlation (central Pacific, around $r = 0.6$), degrees of freedom when correlating with the East African OLR are only reduced from 214 to about 170 (based on the formula of Trenberth 1984), which makes no substantial difference to the absolute correlation that is assessed to be statistically significant. Nonetheless, we have reduced degrees of freedom based on the mean lag-1 wind serial correlation in equatorial (10°N–10°S) and tropical (10°–40°) bands, using the formula in Trenberth (1984).

To better define the associations and lay the basis for physical and dynamic interpretation, statistical significance is mainly offered for guidance and to provide a basis for shading maps for ease of interpretation. Because of the fact that some correlations will always appear statistically significant when calculating many correlations, field significance testing has been applied to the key October–December (OND) convection teleconnection patterns. The field significance is determined through the 500 Monte Carlo simulation procedures as explained in Livezey and Chen (1983). These tests assess the likelihood that the number of significant grid boxes in a field could have been achieved by calculating correlations with random numbers.

Correlation maps with vector wind are presented by forming a vector from the individual correlations with the zonal ($u$) and meridional ($v$) wind components. This enables concise visual interpretations and allows for easy comparison with climatological fields. In addition to monthly/seasonal correlation maps, we present the intraseasonal teleconnection structures before, during, and after (T – 15 days, T – 10 days, T – 5 days, T – 0 days, T + 5 days, T + 10 days, and T + 15 days) prolonged 5-day rainfall events in East Africa, inferred from 5-day OLR over East Africa. The intraseasonal study focuses on the October–November months, which are characterized by a stronger spatial coherence of rainfall anomalies across the region. Composites for the highest and lowest values of the host index (East African rainfall) have also been analyzed to check for any obvious nonlinearity in the relationships discussed.

### 3. Fluctuations of the monthly/seasonal time means

#### a. Variability of the short rains

In Mutai et al. (1998), results were presented using a large-scale standardized area-average rainfall index for the whole of East Africa (Fig. 1a). This region was defined based on rainfall teleconnection studies (point correlation and principal component analysis) that showed a tendency for OND rainfall anomalies to have the same sign across this region, with the southernmost latitudes marking a transition to a tendency for opposite sign rainfall anomalies south of 15°S (e.g., Janowiak 1988; Hastenrath et al. 1993; Mutai et al. 1998). Results with the large-scale rainfall index are useful to diagnose the interaction of East African climate with the global Tropics, but they leave a question about the representativeness of the results at smaller spatial scales over East Africa. This is particularly relevant given the variety of landscapes and associated climates in the region. This section therefore provides information on the representativeness of the large-scale index and associated teleconnections.

In a recent study (Indeje et al. 1999), the region was subdivided into smaller homogeneous rainfall zones as shown in Fig. 1b. These regional groupings represent a simplification of the ones constructed by Ogallo (1989). We explore the relationships between the OND averaged rainfall anomaly indices in these zones with the large-scale index (Table 1, column 3). The results clearly suggest that the large-scale area-average index is very representative of most of the smaller subregions. The mean correlation is 0.76 and over half of the smaller subregions correlate at or above this value. However, there is a suggestion of some independent variability in some western areas (e.g., regions 7 and 8, $r = 0.51$ and 0.49, respectively). For these correlations, we have used

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<table>
<thead>
<tr>
<th>Subregion</th>
<th>No. of stations</th>
<th>CRU OND</th>
<th>Niño-3 SST OND</th>
<th>SOI OND</th>
<th>SST REOF4 Sep</th>
<th>SST REOF5 JAS</th>
<th>SST REOF2 Sep</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>8</td>
<td>0.92**</td>
<td>0.24 (0.37*)</td>
<td>−0.39∗ (−0.59**)</td>
<td>0.30 (0.33)</td>
<td>0.63**</td>
<td>0.16</td>
</tr>
<tr>
<td>2</td>
<td>12</td>
<td>0.85**</td>
<td>0.14 (0.31)</td>
<td>−0.26 (−0.55**)</td>
<td>0.21 (0.25)</td>
<td>0.56**</td>
<td>0.08</td>
</tr>
<tr>
<td>3</td>
<td>6</td>
<td>0.84**</td>
<td>0.15 (0.34)</td>
<td>−0.24 (−0.54**)</td>
<td>0.28 (0.37)</td>
<td></td>
<td>0.51**</td>
</tr>
<tr>
<td>4</td>
<td>8</td>
<td>0.72**</td>
<td>0.30 (0.40*)</td>
<td>−0.37∗ (−0.51**)</td>
<td>0.29 (0.30)</td>
<td>0.47**</td>
<td>0.12</td>
</tr>
<tr>
<td>5</td>
<td>10</td>
<td>0.89**</td>
<td>0.28 (0.40*)</td>
<td>−0.40∗ (−0.58**)</td>
<td>0.32 (0.35)</td>
<td>0.58**</td>
<td>0.29</td>
</tr>
<tr>
<td>6</td>
<td>12</td>
<td>0.87**</td>
<td>0.21 (0.42*)</td>
<td>−0.30 (−0.61**)</td>
<td>0.23 (0.28)</td>
<td>0.60**</td>
<td>0.03</td>
</tr>
<tr>
<td>7</td>
<td>3</td>
<td>0.51**</td>
<td>0.28 (0.31)</td>
<td>−0.38∗ (−0.44*)</td>
<td>0.38∗ (0.38*)</td>
<td>0.21</td>
<td>0.27</td>
</tr>
<tr>
<td>8</td>
<td>5</td>
<td>0.49**</td>
<td>0.11 (0.13)</td>
<td>−0.27 (−0.29)</td>
<td>−0.15 (−0.16)</td>
<td>0.67**</td>
<td>0.07</td>
</tr>
</tbody>
</table>
the period 1961–90, to be consistent with the series length of Indeje et al. (1999).

Next, the coherence of the ENSO signal across the region is evaluated. For this, each rainfall index is correlated with OND values of Niño-3 SST and SOI (Table 1, columns 4 and 5), and the ENSO-based predictor used in Mutai et al. (1998) (September values of global SST rotated eigenvector 4, SST REOF4: column 6). The sign of correlation is always consistent across all regions except SST REOF4 versus region 8, indicating that at least some of the ENSO signal (warm phase corresponding to wet conditions) is manifest in each region. As would be expected simply from the removal of spatial noise, the large-scale area-averaged rainfall shows somewhat higher correlation (0.31, 0.46, 0.35 for 1961–90 and 0.47, 0.70, 0.40 without 1961) with the Niño-3 SST, SOI, and SST REOF4, compared to the mean correlation for the smaller regions (0.21, 0.33, 0.23 and 0.34, 0.51, 0.26, respectively). In fact, 1961–90 is one of the 30-yr periods with weakest ENSO–East African association during this century. This is partly attributable to the extreme year 1961, and its removal led to substantially higher correlations (Table 1 in brackets). Further inspection of Table 1 shows some variation of the ENSO signal across East Africa. For example, region 8 (west of Lake Victoria) fails to achieve significance in any of the correlations and the contrast with other regions is even clearer when 1961 is removed. To assess whether these spatial variations in ENSO signal are more than would be expected due to sampling noise is extremely difficult given the standard errors on the correlations (which are also influenced by the number of stations contributing to each respective index, column 2) and multiplicity considerations (Katz and Brown 1991), which in turn are complicated by spatial correlation among the indices. In terms of the variations in correlation strength, the results can be viewed as laying the basis for evaluation of regional modeling studies (e.g., Sun et al. 1999a,b). For the purposes of the research in this paper, the conclusion is that over much of the region, the large-scale rainfall and associated teleconnection signals in OND are representative of the rainfall anomalies observed over most subregions in East Africa, but with some caution raised for predictability in certain subregions. The final two columns in Table 1 (correlations with July–September and September values of global rotated SST eigenvectors 5 and 2: SST REOF5 and REOF2 used in Mutai et al. (1998)) further support this conclusion. More discussion of precursor SST eigenvectors is returned to in section 5.

b. OLR as a proxy for East African rainfall

Nyakwada et al. (1995) evaluated the rainfall–OLR relationship for all specific seasons in East Africa, mainly focusing on Kenya. They found good relationships in their short rains season of September–November and rather weaker relationships in the long rains of March–May. The utility of OLR has been further confirmed here by comparing standardized area-average rainfall and OLR anomaly indices for the East Africa region as a whole (Figs. 1a and 1c) and three latitude subzones (6.25°N–3.75°S, 6.25°N–6.25°S, and 6.25°–13.75°S), shown in Table 2. Monthly rainfall indicates substantial strong relationships with OLR in eastern Africa except for the early September short rains. On average, the associations are high in November, December, and April and only moderate in March. The September–November season relationships in Nyakwada et al. (1995) may be attributed mainly to the October and November months. The October–December season correlation matrix indicates correlation magnitudes greater than 0.80 for each of the rainfall indices with collocated OLR, confirming OLR to be a good proxy for rainfall in this season. This has been attributed to the strong role of large-scale organized synoptic systems with deep convection at this time of the year (Ogallo 1989).

October–December teleconnection results in subsequent sections are presented using the 6.25°N–13.75°S index while the 6.25°N–3.75°S index is used for March–May and September, since these months do not show such good rainfall anomaly coherence between the northern region and areas south of about 3.75°S. In fact, the large-scale teleconnection structures were not sensitive to the choice of index, though some changes in emphasis were found for the index south of 3.75°S, representing an avenue for further study.
TABLE 3. Pattern correlations to show the similarity of OLR teleconnection structures over 40°N–40°S, 30°E–180°E. The host teleconnection pattern is for Oct 1974–96. For example, the “Oct East African rainfall versus OLR teleconnection pattern” has a pattern correlation of +0.59 with the pattern for Nov.

<table>
<thead>
<tr>
<th>Month</th>
<th>SOI vs OLR pattern</th>
<th>East African rainfall vs OLR pattern</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sep</td>
<td>0.31</td>
<td>−0.06</td>
</tr>
<tr>
<td>Oct</td>
<td>1.00</td>
<td>1.00</td>
</tr>
<tr>
<td>Nov</td>
<td>0.72</td>
<td>0.59</td>
</tr>
<tr>
<td>Dec</td>
<td>0.56</td>
<td>0.56</td>
</tr>
</tbody>
</table>

c. Tropical convection and circulation teleconnection

1) MONTHLY OLR TELECONNECTION

The East African rainfall index and SOI have been used as host indices for correlation maps with tropical OLR. Maps were calculated separately for each of the months September, October, November, and December. Maps for October, November, and (to a lesser extent) December with SOI or East African rainfall were all broadly similar, while those for September were very different. Indeed, the correlation between SOI and September rainfall is actually positive ($r = 0.63, 1974–96$) while SOI and rainfall in October, November, and December are all negative ($r = −0.57, −0.56,$ and $−0.26$) consistent with previous work on the short rains. The similarity of the November and December correlation maps with those of October across the Indian–central Pacific domain is quantified in Table 3. The contrast with September is also quantified in Table 3, most clearly seen with the East African rainfall as the host index.

The results in Table 3 suggest that the early season September rains, and likely also the more widespread precipitation through July–September (JAS) in the western highlands of East Africa [see Camberlin (1997) and Fig. 12b in Ward (1998)], have an opposite sign relationship with ENSO relative to the October–December months. Such a switch of sign is physically plausible given the dramatic change in the background atmospheric circulation from September to October (Figs. 2a,b). September 850-hPa winds have strong southeast to southwesterly cross-equatorial flow near the East African coast, part of the low-level jet associated with the southwest Indian monsoon (Findlater 1974). Westerly monsoon winds penetrate across the Indian subcontinent.
and into the China Sea. The two convection centers over Africa and Southeast Asia/Maritime Continent are distinctly north of the equator. By October, the southwest monsoon is absent, replaced by the first indications of a northeasterly monsoon flow in the northwestern Indian Ocean. The convection centers in both the African and Southeast Asia/Maritime Continent sectors are more symmetric about the equator. Thus, relative to September, equatorial dynamics (Gill 1980) can be expected to play a stronger role in teleconnection structures across the Indian Ocean in October, continuing through November (Fig. 2c) and December (Fig. 2d). The relatively similar teleconnection structures in the OND months (Table 3) are used to justify pooling these months together to reduce noise and produce the clear global teleconnection patterns that are discussed in the next section.

2) Seasonal October–December OLR and Wind Fields

Three equatorial poles stand out clearly in the OLR teleconnection features of the OND season for both the SOI and East African rainfall host indices (Figs. 3a–c), and demonstrates that ENSO is the main large-scale signal in East African rainfall variability. These three horseshoes are present in the individual October, November, and December monthly maps (not shown). For warm ENSO phase, there is reduced convection over the Maritime Continent, with enhanced convection to the east over the central Pacific and to the west over the western Indian Ocean and East Africa. Each equatorial pole forms a point of curvature for horseshoe structures whose outer rims trail northeast and southeast into the subtropics. Previous studies have noted a horseshoe structure associated with ENSO in observed precipitation [station data at least indicating something of the northern rim extending from East Africa; see Wright (1985); Kiladis and van Loon (1988), their Fig. 3d] and in general circulation model vertical velocity (e.g., Miyakoda et al. 1999). These studies, however, have not targeted convection anomalies in the OND season when the third shoe in convection becomes most clearly established over the western Indian Ocean. These structures are of the same essence, but much clearer, than those found in the highly reflective cloud by Hastenrath (1990).

In the rainfall–wind associations (Figs. 4a,b), the equatorial feature in the Indian Ocean basin at 850 hPa has been seen in Hastenrath et al. (1993) and others, whereas reverse features occur at 200 hPa. An apparent Kelvin wave response to anomalous heating (Gill 1980; Reverdin et al. 1986) gives the northeast trades a stronger easterly component over the western Indian Ocean with anomalies converging close to the East Africa coast but not so much over East Africa (this point is returned to in section 4). South of the equator in the Indian Ocean, anomalous winds converge into the first part of the trailing horseshoe rim (about 10°–30°S, 50°–80°E), which can be expected to contribute to destabilizing the southeast trade winds blowing into East Africa from the Mascarene high. Anomalous cyclonic Rossby gyres in the upper troposphere are clear north and south of the Maritime Continent, consistent with the reduced convection in the Maritime Continent. Weak anticyclonic gyres occur near 150°W (Fig. 4b). There is also a weak anticyclonic indication at 20°N and 20°S, 30°–50°E, and this feature is more marked for composites of the wettest five seasons in East Africa (not shown).

If one considers the situation of anomalous ascent over the Maritime Continent, the Gill (1980) model predicts compensating descent particularly centered on the western side of the anticyclonic gyres. This descent could be part of the reduced tropical convection in the rims that connect back to the equatorial pole of anomalous descent that would be located over East Africa in this phase of anomaly. A role for interaction with mid-latitude upper-level jet streams (Meehl et al. 1996) and, specifically, subtropical Rossby waves (Hsu et al. 1990; Hoskins et al. 1983, 1999) is also likely and is supported by the teleconnections north of 40°N in the circulation features on Figs. 4a,b. In the tropical zones of enhanced convection in the horseshoe rims, there are generally weaker and less anticyclonic near-surface 850-hPa trade winds (Fig. 4a), with opposite changes in the zones of reduced convection.

The teleconnection with the anomalous near surface northeasterly over central Africa (Fig. 4a) is largely absent after 1979, and doubt is cast as to its reality. Otherwise, calculating the teleconnections before and after the arrival of satellite data in 1979 produces no changes of any substance to the teleconnection structures. Some changes in teleconnection strength are noted though, with generally stronger teleconnections over the Indian Ocean prior to the introduction of satellite data in 1979.

3) Teleconnection Fields with ENSO Removed

It is worth remembering that the residual variance after ENSO is removed represents about 75%–80% of the interannual variance of OND East African rainfall. Absent now are strong features in the equatorial Pacific and Indian Oceans westward to about 70°E (Fig. 3d). In the Indian Ocean the dominant feature is a structure from East Africa southeastward into the Tropics, though not trailing as far as in the maps with ENSO (Figs. 3a–c). This supports the idea that the southern subtropical horseshoe rim east of 60°E was indeed forced by anomalous convection over the Maritime Continent, in turn triggered by the ENSO SST pattern. However, the result also suggests that the trailing rim west of 60°E may be made up of processes independent of Maritime Continent convection anomalies. Circulation anomalies are also still present west of 60°E, with the anomalous southeasterlies (Fig. 4c) undiminished from the map that includes the ENSO influence (Fig. 4a).
Fig. 3. Correlation for 1974–96 between OND global Tropics OLR field and OND values of (a) SOI, (b) East African rainfall, (c) East African OLR index, and (d) same as (c) but with ENSO variance removed from both datasets. The contour interval is 0.2 with zero contour omitted. Dark (light) shading indicates statistical significance of positive (negative) values at the 5% level. OLR is multiplied by $-1$ in (b) and field significance ($fs$) is shown for each panel.
Fig. 4. Correlation between OND East African OLR multiplied by −1 and OND wind anomalies in the NCEP–NCAR reanalysis, 1974–96: (a) 850 hPa, (b) 200 hPa, (c) same as (a) but with ENSO variance removed, (d) same as (b) but with ENSO variance removed. Vectors are formed from the $u$-wind correlation and $v$-wind correlation and shading indicates statistical significance of either the $u$ or $v$ component at the 5% level.
The rim to the north from East Africa has become much weaker (Fig. 3d), and circulation anomalies from the central Pacific to East Africa are also much weaker (Figs. 4c,d), suggesting that many of these variations are part of the global Tropics ENSO signal. The direct SST forcing may be coming from either the tropical Pacific, sympathetic changes in the Indian Ocean, or a combination of both (Goddard and Graham 1999). Indeed, the continued presence of a small ENSO-free equatorial easterly anomaly in the Indian Ocean (Fig. 4c) and return 200-hPa flow (Fig. 4d) indicate some additional ENSO-free Indian Ocean SST forcing. Mid-latitude teleconnections to circulation in the Mediterranean do remain in the ENSO-free maps (Figs. 4c,d).

An important feature of the ENSO-independent OLR map (Fig. 3d) is a zone of enhanced convection extending westward from East Africa through equatorial Africa and into the equatorial Atlantic and Amazon basin. The 850-hPa wind shows somewhat anomalous cyclonic flow in the Atlantic 0°–30°S (Fig. 4c) that, coupled with enhanced convection south of its climatological position off the coast of West Africa, suggests a warmer equatorial and tropical South Atlantic. Anomalous upper-level divergence above East Africa is now mainly formed by a clear easterly 200-hPa anomaly from East Africa toward the equatorial Atlantic. This is returned to in section 5.

4) CONTRASTS TO MARCH–MAY

SOI–OLR structures in March (Fig. 5a) and April (Fig. 5c) contain Maritime Continent convection nodes centered east of 120°E, much farther east than in OND. While opposite sign anomalies are found northwest and southwest of these convection nodes [consistent with the Gill (1980) model], these opposite anomalies do not wrap westward to form a horseshoe with central curvature point in the western Indian Ocean/East Africa. In May (Fig. 5e) however, Maritime Continent convection anomalies extend (on and north of the equator) to about 90°E, and now a horseshoe structure does wrap round with curvature point in equatorial East Africa (correlations > 0.6). Inspection of the background climatology also shows that the zone of Maritime Continent convection shifts westward during March to May (not shown), and this background near-equator convection in the eastern Indian Ocean is a candidate to explain the presence or absence of the western Indian Ocean horseshoe teleconnection. A further factor in the absence of strong teleconnections to the Indian Ocean in March and April is also likely the lower ENSO variance and its transitory nature at this time of the year (e.g., Meehl 1987; Hastenrath et al. 1993), though this does not prevent ENSO teleconnections eastward, for example, to northern Brazil (Fig. 5a) and northeastern Brazil (Fig. 5c).

Consistent with the above findings, the global Tropics East African rainfall teleconnection pattern is weak in March and April (Figs. 5b,d) but reflects the emergence of the horseshoe SOI structure in May (Fig. 5f). Thus, it appears that May does have features in common with OND, including a tendency for enhanced East African rainfall in warm ENSO phase.

4. Intraseasonal teleconnections with October–November East African rainfall

Correlation and composite atmospheric circulation structures have been computed for time lags ranging from −15 to +15 days relative to the East African area-average 5-day proxy for rainfall (OLR) in the 1979–96 period. To check for the reliability of the pentad OLR data, it was correlated with the pentad area-average rainfall anomaly over Kenya, as estimated from daily rainfall data. The correlation of −0.55 between the 216 pairs of October–November pentad values over 1979–96 suggests a good common signal in the two datasets, especially given the relatively small number of rainfall stations confined to a subregion of East Africa. Furthermore, when pentad values of OLR were replaced with pentad values of the Kenya rainfall index, lead–lag circulation patterns were found to be a little weaker but remarkably similar in pattern (not shown). No filtering was applied to any of the series because, as a first step, we wanted to study the full atmospheric circulation anomalies. Thus, the pictures contain aspects of circulation that are standing quasi-permanent anomalies through the season, combined with some aspects that are propagating intraseasonal disturbances, moving through the quasi-permanent seasonal time-mean circulation. Lead–lag teleconnection structures in the Tropics are shown for winds at 850 (Figs. 6a–g) and 200 hPa (Figs. 7a–g). Lead–lag structures for OLR (not shown) were largely consistent with the zones of divergence and convergence in the 200-hPa fields.

At a lead of 15–10 days, there is an intensification of anomalous 850-hPa divergence over the Maritime Continent, with anomalous easterlies strengthening across the equatorial Indian Ocean (Figs. 6a,b). At 5 days’ lead (Fig. 6c), the easterly wind anomalies turn into Arabia and also southward into the off-equator convection anomalies that form the southern rim of the horseshoe structure. At lag 0, low-level circulation anomalies across the Indian Ocean are generally weaker (Fig. 6d).

The precursor signal in the 200-hPa wind field is particularly striking. At 15 days before the East African rainfall event (Fig. 7a), anomalous equatorial westerlies extend from the Greenwich meridian to the western Indian Ocean. These slowly migrate eastward and reach the eastern Indian Ocean at zero lag, when anomalous easterlies extend over equatorial Africa and into the western Atlantic (Figs. 7b–d). Anticyclonic curvature in westerly anomalies is found at 20°N and 20°S at the longitudes of East Africa, suggesting partly formed Rossby anticyclonic gyres in response to convection
anomalies over East Africa. These westerly anomalies turn equatorward over the Maritime Continent and form the equatorial westerly component of compensating anomalous cyclonic Rossby gyres north and south of the Maritime Continent (the interpretation is that the reduced Maritime Continent convection actually leads to reduced anticyclonic gyres, leading to the cyclonic anomalies). At lag $T + 5$, the circulation anomalies migrate farther eastward, with upper-level divergence centered over the western Indian Ocean. This signature is very indicative of a juxtaposed positive (East Africa–western Indian Ocean), negative (Maritime Continent) diabatic heating anomaly forcing in a Gill (1980) dynamical system, consistent with interpretations of MJO structure in this region (Hendon and Salby 1994; Hendon and Glick 1997). Anomalous easterlies at 200 hPa are well established over East Africa at lag $+5$ days (Fig. 7e), similar to the seasonal mean teleconnection pattern over East Africa (Fig. 4b). We propose that wet seasons in East Africa are marked by favorable con-
ditions for convection over East Africa and the western Indian Ocean, and the circulation picture during the intraseasonal phase of enhanced convection over the western Indian Ocean (lag T + 5 in East Africa, Fig. 7e) tends to dominate the seasonal mean picture (Fig. 4b), perhaps simply due to the stronger magnitude of anomalies in this intraseasonal state, or due to the climate system residing in the state of enhanced convection over the western Indian Ocean on more days than that with the enhanced convection over East Africa.

It has been noted before that, during boreal autumn, the leading mode of interannual variability in the region of the Indian Ocean and Maritime Continent has a dynamic structure that is similar to that of the MJO when its ascending branch is located over the Indian Ocean or (reversed sign) the Maritime Continent (Gutzler and Harrison 1987). It is possible that the MJO is unaltered by the background atmospheric and SST anomalies (the dipole across the Indian Ocean), and standing quasi-permanent anomalies of equatorial convection over the western Indian Ocean (one sign) and Maritime Continent (opposite sign) make the complete contribution to the seasonal mean anomalies in the region. However, the pulsing nature of the convection in this region makes this unlikely. Furthermore, there is mounting evidence that the MJO may be a coupled ocean–atmosphere phenomenon (e.g., Krishnamurti et al. 1988; Wang and Li 1994; Hendon and Glick 1997; Lau et al. 1997). These observations, combined with the analyses here, raise some important questions regarding interannual variability of the MJO and East African rainfall. Instead of a dominance of quasi-permanent convection anomalies, a more plausible hypothesis may be that in years favorable to convection over East Africa and the western Indian Ocean, convection bursts associated with the development of MJO tend to be stronger and/or occur more often. However, the unfavorable background anomalies in the Maritime Continent can be expected to reduce the strength of MJO convection anomalies in this region. This is expected to be one reason why the anomalies that propagate across the Indian Ocean in Figs. 7a–g become very weak as they approach the Maritime Continent. Such scenarios could give rise to the observed seasonal mean teleconnection structures purely through a regionally modified MJO structure. Thus, these results suggest that as a next step, the intraseasonal structures should be stratified according to the background October–November circulation and SST anomalies across the Indian Ocean and Maritime Continent.

Next, the evolution of circulation over the equatorial Atlantic and equatorial Africa is focused upon. An association between local near-surface westerly anomalies and rainfall bursts in East Africa is known (e.g., Cambridge and Wairoto 1997; Camberlin 1997; Okoola 1999). However, the extension of the westerly anomaly back to the equatorial Atlantic, and some evidence of it beginning the pentad before the East African rainfall event (Figs. 6c,d) is previously unknown. It is, however, known from satellite images that wet spells in East Africa are often associated with synoptic disturbances that migrate eastward into the region in association with westerly near-surface wind anomalies (E. Mutoni, Tanzania Directorate of Meteorology 1998, personal communication) and so this result frames further investi-
Fig. 6. Lag correlation between the proxy for East African Oct–Nov pentad rainfall (OLR multiplied by $-1$) and pentad 850-hPa wind anomalies in the NCEP–NCAR reanalysis, 1979–96: (a) $T - 15$ days, (b) $T - 10$ days, (c) $T - 5$ days, (d) $T - 0$ days, (e) $T + 5$ days, (f) $T + 10$
Fig. 6. (Continued) days, and (g) T + 15. Vectors are formed from the $u$-wind correlation and $v$-wind correlation and shading indicates statistical significance of either the $u$ or $v$ component at the 5% level.

gation of the interaction of the disturbances with the westerly anomalies. To see in more detail the changes in African circulation that this evolution corresponds to, Figs. 8a–f show the composite evolution of actual and anomalous 850-hPa winds over East Africa. It can be seen that the westerly anomalies at lag 0 correspond to an actual westerly wind of about 2–5 m s$^{-1}$ extending from the Atlantic coast across equatorial Africa at latitudes 0°–5°S. The meridional arm of convergence is clearly enhanced and shifted east (Fig. 8f). Over North and West Africa, the mass flow has a different kinematic structure at $-10$ days compared to lag 0. At $-10$ days, anticyclonic trades extend from the eastern Sahel into equatorial West Africa and the Atlantic even slightly stronger than over Africa. At lag 0, these trades are weaker over the tropical Atlantic and west Africa while they have a stronger northerly component over central Africa (around 0°–10°N, 15°–25°E) that feeds into the westerlies south of the equator. This component of circulation adds to the converging northeast and southeast trades from the Indian Ocean previously emphasized in East African rainfall. Indeed, Indian Ocean anomalies are weak during the rainfall event itself (Fig. 8e). These results also raise the prospect for an Atlantic influence.
Fig. 7. Same as Fig. 6 but for 200-hPa winds.
on East African rainfall. First, the Atlantic SST anomalies could modify the nature of the Congo air mass that converges from west to east. Second, the SST anomalies over the equatorial Atlantic may change the likelihood of spells of westerly anomalies through the October–November season.

Teleconnection structures are also evident in the central Pacific Ocean 850-hPa winds at considerable lead time. From 10 to 15 days (Figs. 6a,b) before the East African rainfall event, westerly anomalies are found in the central Pacific, connected with an anomalous burst of near-surface divergence over the Maritime Continent and its counterpart strengthening of easterly anomalies across the Indian Ocean. At lag 0, these westerly anomalies in the central Pacific have almost completely disappeared (Fig. 6d). A weaker, but perfectly out-of-phase pulse occurs at 200 hPa (Figs. 7a–d) and similar temporal evolution of convection anomalies is also seen in OLR fields (not shown), with enhanced convection accompanying the westerly anomalies at 10–15 days before the East Africa event, decaying to near-zero anomalies during the rainfall event.

5. Associations of SST eigenvector predictors with tropical convection and circulation anomalies

Mutai et al. (1998) developed an empirical prediction scheme for East Africa OND rainfall based on three
Pentad wet composite for 850-hPa winds
precursor global SST rotated EOF patterns (Figs. 9a–c). This section diagnoses the atmospheric circulation and convection anomalies associated with these predictors, and interprets them in the light of the results in sections 3 and 4.

One predictor (SST REOF4) was noted by Mutai et al. (1998) to closely reflect the ENSO signal (Fig. 9b), and teleconnections with winds (850 hPa, Fig. 10a; 200 hPa, Fig. 10b) and OLR (Fig. 10c) confirm this. Horsehoe structures in the OLR are accompanied by the 850- and 200-hPa wind anomalies noted in Figs. 3 and 4, especially in the Maritime Continent region, stronger in Figs. 10a–c in the central and eastern tropical Pacific. This SST predictor is clearly measuring the state of ENSO in September and showing how that correlates with circulation in OND. It can be noted that the circulation and convection teleconnections in the Indian Ocean do not extend so strongly into the East African region, compared to the simultaneous OND structures (Figs. 3a and 4a). This suggests that some development of ENSO-related indices occurs from September to October–December across the Indian Ocean, such that ENSO indicators measured in OND teleconnect even stronger into East Africa.

A further predictor (SST REOF5) contained SST variance in the Pacific–Indian Ocean sectors. The July–August–September (JAS) time coefficients of SST REOF5 correlate very weakly with JAS values of SOI (−0.26) and Niño-3 SST (0.00) over the 1951–96 period. Correlation of JAS SST REOF5 with OND values of SOI and Niño-3 SST is a little higher ($r = -0.34$ and $r = +0.17$), suggesting a weak tendency for JAS SST REOF5 to anticipate ENSO development in OND. The SST REOF5 predictor has a pattern of weights with a relative minimum in the equatorial western Pacific (Fig. 9a). It was speculated that this could excite near-surface divergence out of the western Pacific [visible in correlation maps with near-surface divergence in Mutai et al. (1998), their Fig. 6a], generating conditions that favored above-normal rainfall in East Africa. The association of JAS SST REOF5 with OND 850-hPa circulation is also divergent along the equator out of the western Pacific in the reanalysis data (Fig. 11a), with strong easterly anomalies across the Indian Ocean extending all the way to the East African coast. The upper-tropospheric wind teleconnections (Fig. 11b) mirror the low-level patterns over the Maritime Continent with anomalous convergence. In addition, there are anomalously strengthened northeasterly trades in the northwestern Pacific (Fig. 11a) and suppressed convection (Fig. 11c) that gives rise to a horseshoe rim structure in this region. Indeed, this suppressed convection is surrounded to the west and east by enhanced convection anomalies, giving rise to a sequence of three trailing rims from equatorial latitudes into the subtropics. The equatorial pole over East Africa is particularly strong, consistent with the strong correlations between this EOF and East African rainfall (Table 1, column 7). The OLR and circulation anomalies in the northwestern Pacific and north Indian Ocean appear plausible with the pattern of SST gradients in the SST EOF. Southeastward trailing rims from the East Africa and Maritime Continent convection poles are also apparent, though weaker than their northern counterparts. The extent to which this arrange-

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ment of ocean–atmosphere associations will be repeated in the future is still of some concern, since it is not a well-known structure and it does not have a sound theoretical basis in ocean–atmosphere coupled dynamics at this stage. Therefore, its further monitoring and study is recommended.

The final SST EOF related to East African rainfall measures warming and cooling events in the equatorial and tropical South Atlantic (Fig. 9c). SST REOF2 in JAS is independent of SOI ($r = 0.00$). The association of September SST REOF2 with tropical OND OLR shows that anomalous warm water tends to draw the ITCZ to be anomalously active over the open ocean just south of the West African coastline, with enhanced convection extending westward into Amazonia and eastward into equatorial and eastern Africa (Fig. 12c). This pattern of correlations has been confirmed using the CRU precipitation dataset, including into the Amazon basin (not shown). The picture is almost identical to that of the East Africa rainfall–OLR signature once ENSO was removed from the data (Fig. 3d). It is strong evidence for an influence of the equatorial and tropical
South Atlantic on East African rainfall. The location of the OLR anomalies connecting from the Atlantic through East Africa is centered at about $0^\circ$-$5^\circ$S, the latitude where westerly winds develop near the surface before and during major rainfall events in East Africa. The relationship between the SST REOF2 and near-surface wind anomalies exhibits a cyclonic anomaly in the equatorial and South Atlantic (Fig. 12a), consistent with the convection anomalies. Associated 200-hPa wind anomalies (Fig. 12b) are very clear, with anomalous easterlies extending from equatorial Africa across the equatorial Atlantic.

6. Summary and conclusions

This paper began by establishing the representativeness of an area-average rainfall index for East Africa during the short rains. It was shown that rainfall anomalies in smaller subregions within East Africa strongly covary with the area-average rainfall index. Furthermore, teleconnections with the broad-scale ENSO indices were also reflected in smaller regions, generally with slightly reduced magnitude of teleconnection. In some cases though, especially subregions near lakes and mountains, the teleconnection strength was more sub-
FIG. 12. Same as Fig. 10 but for Sep SST REOF2.

October–December teleconnections with September SST REOF2

(a) Sep SST REOF2 and OND 850 hPa wind

(b) Sep SST REOF2 and OND 200 hPa wind

(c) Sep SST REOF2 and OND OLR (fs = 85.8%)

For warm ENSO phase, September rainfall in East Africa tends to be part of the July–September reduced rainfall in India, the Ethiopian highlands, and through the western Sahel (Camberlin 1995, 1997; Ward 1998). The teleconnection structure is quite different in October–December, when rainfall in East Africa is enhanced in warm ENSO phase. A sequence of three horseshoe structures are evident in convection anomalies, with western tips (i.e., the center point of curvature) on the equator and outer rims trailing to the northeast and southeast into the subtropics centered over East Africa (positive sign), Maritime Continent (negative sign), and central Pacific (positive sign). The equatorial nodes over East Africa and the Maritime Continent are connected by an anomalous Walker circulation across the Indian Ocean, evident in near-surface and upper-tropospheric winds. One interpretation of the anomalies is in terms of equatorial dynamics driven by equatorial poles of upper-level diabatic heating anomalies asso-
cated with the anomalous convection (Gill 1980; Rev-erden et al. 1986). However, forcing of the marine boundary layer by anomalous SST gradients is likely to play an initiating role too (Lindzen and Nigam 1987; Neelin and Held 1987), with, in warm ENSO phase, anomalously cold SST over the Maritime Continent and warm SST over the western Indian Ocean generating anomalous east–west sea level pressure gradients that drive the observed near-surface easterly wind anomalies. This scenario of SST anomalies, which usually sets up in ENSO0 years, may account for much of the ENSO–East Africa teleconnection. Indeed, a strong role for the Indian Ocean SSTs associated with ENSO in generating East Africa short rains variability was recently found in GCM experiments by Goddard and Graham (1999).

ENSO-free teleconnections contained strong structure from the eastern Indian Ocean westward to Amazonia, indicating that some of the variance not explained by ENSO was still part of coherent large-scale climate processes. From East Africa westward through to the Atlantic and Amazonia, sympathetic convection anomalies were found, suggesting a potential role for the tropical Atlantic SST to excite such fluctuations. Sympathetic convection anomalies were also found southeastward from East Africa, like the horseshoe rim in the full teleconnection pattern, but not extending so far. This suggests that the first half of the rim could be attributed to internal atmospheric variability and/or SST forcing not connected to ENSO.

In March and April, ENSO signal was found to be very different from OND in the Indian Ocean–central Pacific region. Suppressed convection in warm phase ENSO over the Maritime Continent was located farther east and the horseshoe equatorial node over the western Indian Ocean was completely absent, seemingly detaching East Africa from ENSO. However, the horse-shoe structure in the western Indian Ocean convection anomalies returns in May and East Africa regains a teleconnection structure with some similarities to the OND pictures.

It was previously known that seasonal and intraseasonal wet conditions were associated with local westerly anomalies near the surface (Anyamba 1984; Davies et al. 1985; Camberlin and Wairoto 1997; Okoola 1999) and it was reassuring to recover this result using the OLR and reanalysis data. In fact, the reanalysis data suggest the westerlies may start to develop during the 5-day period before the rainfall event in the equatorial Atlantic and during the event, extending through equatorial Africa into East Africa. At 0°–5°S, these correspond to actual westerlies of about 2–5 m s\(^{-1}\) in the pentad reanalysis data. The possible interaction of these winds with synoptic-scale disturbances that are known to propagate eastward into East Africa represents a line for further investigation. Before the rainfall event, anomalous near-surface easterlies grow over the Indian Ocean, giving the northeast and southeast trades a stronger onshore component as they approach East Africa, though during the event itself, these wind anomalies weaken. Other aspects of circulation prior to the event included anomalous near-surface westerlies in the central Pacific (which decay during the event and may be connected to the development of westerlies in the Atlantic and over Africa) and an anomalous upper-level cyclonic wind component over Arabia with associated enhanced convection before the rainfall event.

Inspection of previously published intraseasonal oscillation of equatorial convection (e.g., Rui and Wang 1990; Anyamba and Weare 1995) and velocity potential (Slingo et al. 1996; Sperber et al. 1997) show that bursts often pull off the East African coast and intensify over the Indian Ocean, where they can develop into full MJO events. Thus, the precursory features discussed above are of interest not just to anticipating wet events over East Africa, but also perhaps to MJO development. Indeed, MJO-like structures are seen to migrate across the Indian Ocean in the fields immediately after the East African rainfall event, but in the analyses here, the signal weakens as it approaches the Maritime Continent. This raises issues of interactions between the background ocean–atmosphere seasonal anomaly and the propagating intraseasonal anomalies, which themselves can partially modify the SST field (Emanuel 1987; Neelin et al. 1987; Krishnamurti et al. 1988).

The intraseasonal analyses represent a stringent examination of the utility of the OLR and NCEP–NCAR reanalysis data at pentad timescales. It is therefore a positive endorsement of both datasets when intraseasonal signals are identified by correlating OLR with reanalysis winds [see also Sperber et al. (1997) for case studies in selected years]. In particular, it is encouraging to recover previously anticipated associations, like westerly near-surface anomalies during a rainfall event over East Africa. Further work is, however, still needed to check details, like the exact location of convergence zones over East Africa and the magnitudes of wind anomalies. Also, diagnosis of sub-5-day features, like individual synoptic systems, has not been attempted here. It represents a considerable further demand on the accuracy of the reanalysis data over Africa.

In the light of the above diagnostic studies, the atmospheric teleconnections with the three SST predictors used by Mutai et al. (1998) were also explored. Interpretation of the ENSO predictor was clear, though with a suggestion that ENSO indicators in September may not capture so strongly the extension of ENSO activity right across the Indian Ocean during OND into East Africa. The SST predictor measuring gradients of SST from the northwestern Pacific into the Maritime Continent and northern Indian Ocean teleconnected with a train of convection anomalies from the northwestern tropical Pacific across India into East Africa, representing the northern rims of the horseshoe structures. Such a finding could be consistent with the SST gradients in the predictor, but careful monitoring of the performance of this predictor is recommended. The te-
leconnection results with the Atlantic predictor, measuring warming events in the equatorial and tropical South Atlantic, were particularly insightful and now suggest with some confidence an additional role for Atlantic SST forcing of East African rainfall variance that is independent of ENSO (at least at zero lag). The warm phase of this predictor was clearly associated with enhanced convection south of West Africa in the ITCZ, extending westward to Amazonia and eastward through equatorial Africa and into East Africa. The pattern closely resembled the East Africa rainfall teleconnection with enhanced convection once ENSO was removed. Possible interaction between this SST pattern and the westerly bursts across equatorial Africa bringing rainfall to East Africa at the intraseasonal timescale is an important avenue for further study.

This paper has therefore enhanced the basis for monitoring East African rainfall at interannual and (for October–November) at intraseasonal timescales. This monitoring can be rooted in the knowledge of the teleconnection structures and some of the underlying mechanisms for their generation. In addition, it has diagnosed a set of SST predictors in the light of the circulation anomalies associated with the East African rainfall anomalies and given greater confidence in the potential of these prerainfall season SST patterns to be connected to East African rainfall. At the same time, the intraseasonal analyses, when compared to the interannual analyses, have started to raise issues of scale interaction that are critical to a better understanding of interannual variability in the Tropics. In particular, it is now desirable to stratify intraseasonal variability in the Indian Ocean during October–November according to the phase of the seasonal mean Maritime Continent–East Africa convection anomaly dipole.

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