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Department of Marine, Earth and Atmospheric Sciences, North Carolina State University, Raleigh, North Carolina, USA

Simulation of the sensitivity of Lake Victoria basin climate to lake surface temperatures

R. O. Anyah and **F. H. M. Semazzi**

With 7 Figures

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Summary

The response of Lake Victoria basin climate to changes in the lake surface temperatures (LST) has been examined using NCAR-Regional climate model (RegCM2). In the control run uniform lake surface temperature of 24 °C was prescribed and the model integrated for four months, starting at the beginning of September, 1988. In the anomaly experiments the LST was perturbed by ±1.5 °C, and kept constant during the entire period of the integrations.

Simulation results show significant relationship between basin-wide spatial distribution of rainfall and changes in LST. In general during the short rains at warmer/cooler LSTs, significant increase/decrease in the simulated rainfall occurs over the lake surface and surrounding areas. Rainfall exceeding the amount in the control run by more than 50%, particularly over the western, south/southwestern and central parts of the lake is simulated in the run in which the LST is 1.5 °C warmer than the control. It is also evident from our results that different parts of the lake basin respond differently to LST changes which is in contrast to the common characterization of the lake basin as a single homogeneous climate regime in many previous studies.

In general the results show that regions with largest response to LST anomalies during the short rains are collocated with the ITCZ. In October when the ITCZ is directly located over the lake, the largest response (maximum rainfall) is also located over the same region. As the season progresses and the ITCZ shifts out of the lake into northern Tanzania, the regions of rainfall maxima also shift with it. This appears to explain the unexpected reduction in over-lake rainfall in December in spite of the LST being warmer than control by 1.5 °C. We believe this is a direct consequence of the enhanced convection to the south of the lake (over ITCZ) and the tendency of the system to conserve local moisture budget over the lake.

1. Introduction

The major sources of seasonal and inter-annual climate variability over the East African region include El Nino/Southern Oscillation (ENSO), Quasi-bennial Oscillation (QBO), Inter-Tropical Covergence Zone (ITCZ) and large scale monsoonal winds (Nicholson, 1996; Ogallo, 1988; Indeje and Semazzi, 2000). However, meso-scale systems generated by orography, large inland lakes, and other surface contrasts have also been shown to contribute significantly to the diurnal, seasonal and even inter-annual climate variability over the region (Datta, 1981; Mukabana and Pielke, 1996; Sun et al., 1999). Localized circulations associated with the presence of complex orography and inland lakes may significantly impact regional climate anomalies initially triggered by large scale systems. In this study we investigate the relationship between the thermal characteristics of Lake Victoria of eastern Africa and the seasonal variability of the lake basin climate. We particularly investigate, through numerical simulations, the sensitivity of the lake basin climate to lake surface temperatures (LSTs).

The role large lakes play in modifying regional climates in different parts of the globe has been the subject of a number of investigations (Bates et al., 1993, 1995; Small et al., 1999, 2001; Sun et al., 1999). Large inland lakes affect atmospheric circulations in the surrounding regions mainly through frictional and thermal contrasts between the lake surfaces and the adjoining land areas (Mukabana and Pielke, 1996). Furthermore, lakes are sources of significant moisture and latent heat to the ambient environment (Bates et al., 1993). The contribution of Lake/Land Breeze effects on the diurnal cycle of local climates is also now well understood (Fraedrichs, 1972; Datta, 1981; Okeyo, 1987; Mukabana and Pielke, 1996; Bates et al., 1993, 1995). Mesoscale circulations triggered by large inland lakes also interact with large scale circulations and significantly contribute to seasonal and inter-annual climate variability. Latent heat is a primary energy source driving tropical climate (Asnani, 1993) and therefore, accurate simulation of tropical climates requires proper understanding and treatment of the interplay among cumulus convection, mesoscale and large scale climate systems (Sun et al., 1999). Large inland lakes are an important meso-scale "factor" in the modulation of regional climate since they are associated with large amounts of latent heat release.

Bates et al. (1993, 1995) have studied the influence of the Great Lakes in USA using the RegCM2 coupled to 1D thermal lake model. Their results demonstrated that the model captured the basic characteristics of the lake-induced precipitation of the region reasonably well. In particular, their RegCM2-ID Lake modeling system reproduced more realistic lower surface pressure and surface layer wind patterns over the entire Great Lakes, compared to the RegCM2-only simulations. Using the 1-D lake model of Hostetler et al. (1990) their study showed that when the lake model was coupled to RegCM2, there were remarkable improvements in the simulations of the lake-induced precipitation over most parts of the Great Lakes region. In this study we apply a slightly different approach, whereby we use prescribed LST in our simulations to investigate the effects of Lake Victoria of eastern Africa on the regional climate. We adopted a simple technique of prescribing LST over the lake surface to examine how the local climate responds to the LST anomalies. This approach is widely applied in Atmospheric Global Circulation models (AGCMs), where prescribed sea surface temperatures (SSTs) are normally used to evaluate the sensitivity of global climate to SST variations over different Ocean basins.

Compared to other parts of the world (tropics), Eastern Africa region is disproportionately covered by large inland lakes and thus provides a suitable geographical setting for the interaction between large scale climate systems and lake-induced meso-scale flows. Lake Victoria is the largest among the family of the East African Great Rift-valley fresh water lakes. It is also the second largest in the world (Asnani, 1993). The lake's location at an elevation of more than 1200 m (Fig. 1), the diverse topographical features in its vicinity, together with the intense equatorial insolation provide a unique setting for strong mesoscale circulations over the lake and surrounding regions.

Lake Victoria, together with its rich agricultural hinterland is home to and directly supports over 30 million people with average economic

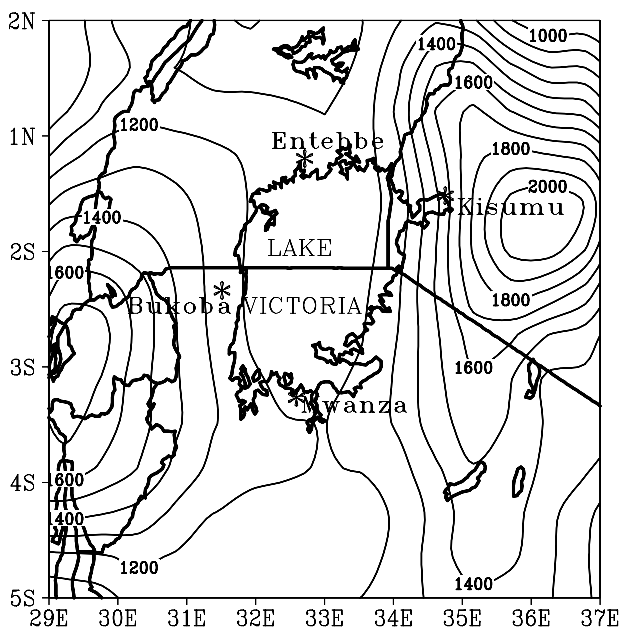

Fig. 1. The topography of the Lake Victoria region (contour heights in meters showing regions which are 1200 m or higher AMSL) based on 30 minute topographical dataset

productivity estimated at between \$3 and \$4 billion per year (FAO, 1999). This makes the lake basin climate a primary factor in the regional economies especially in terms of food security, water and hydro-electric power supply. Since Lake Victoria has only one major outlet (river Nile) and several inlets hydrologically, the lake behaves like a closed system (Spigel and Coulter, 1996). The hydrological balance of the lake is dominated by rainfall, evaporation and runoff, with river inflow and outflow making negligible contributions (Ba and Nicholson, 1998). The large proportion of rainfall and evaporation in the lake's water budgets make it particularly sensitive to climate change (Spigel and Coulter, 1996). These physical attributes and the lake's dynamical characteristics make it a significant contributor to diurnal and seasonal variability of the regional climate.

The available observational records and historical information show that variations in the level of Lake Victoria and other East African lakes is closely related to previous trends in rainfall pattern over the region. For, example, Nicholson (1996) has used a long time series of observations, dating back to the late 1880s, derived from both historical information and modern instruments to show that historical changes in the Lake Victoria levels has been changing in response to changes in rainfall over the region. The above average rainfall experienced over the region during the 1960s led to a record lake level rise of 2 m, accompanied by heavy flooding over most parts of the lake basin (Anyamba, 1983). In contrast, during the early periods of the 19th century when the region experienced prolonged drought, the lake level fell significantly (Nicholson, 1996; Ba and Nicholson, 1998). However, lack of adequate observations has hampered objective and comprehensive diagnostic or numerical characterization of the climate anomalies over the Lake Victoria region. In terms of lake surface temperature variations, limited available observations indicate that the lake's temperature vary by only a few degrees from month to month. For, example, measurements carried out on the Kenyan side of the lake by Ochumba (1996) show the LST to vary between 24.3 °C and 28 °C most of the year, except during the months of May through July, when temperatures are about 2–3 °C cooler than the average which is about 24 °C. The temperature profile measurements on the Ugandan side of the lake (Bugenyi and Magumba, 1996) indicate that Lake Victoria surface temperatures were warmer by more than 0.5 °C during the 1990s compared to the 1960s.

The main objective of this case study is to investigate the relationship between the Lake Basin climate and changes in the lake surface temperatures. We examine how the anomalous warming or cooling of the Lake Victoria surface may influence the ambient atmosphere and interact with the prevailing large scale circulation to modulate Lake Basin rainfall variability.

A brief overview of the study area and general climatology of Lake Victoria basin is given in Section 2. Section 3 gives a brief description of the model and design of numerical experiments. The results of the numerical experiments are discussed in Section 4 and conclusions presented in Section 5.

2. Study area and general climatology

The study area covers Lake Victoria basin of East Africa. Lake Victoria is located approximately at the center of East Africa comprising the countries of Kenya, Uganda and Tanzania. It is part of a depression within the East African Great Rift Valley, between two mountain ranges, generally running north–south, on both sides (Fig. 1). The lake is bounded by latitudes 0.5° N and 2.5° S and longitudes 32° E and 34° E. The approximately rectangular-shaped lake is 280 km wide, 400 km long; with a total surface area of about 69,000 km^2 (IDEAL, 1993; Ochumba, 1996). The lake's mean depth is about 40 m, with maximum depth of 92 m. The volume of the lake is about 2760 km^3 (IDEAL, 1993) with the mean annual evaporation and rainfall almost in perfect balance (Ba and Nicholson, 1998).

The large scale winds over the lake basin are mainly easterly monsoons most of the year. Superimposed on this basic current are the south easterly (SE) or north-easterly (NE) components depending on the migration and location of the Inter-tropical Convergence Zone (ITCZ) and the position of the nearby sub-tropical high pressure systems. The seasonal distribution of rainfall during the year is bimodal, with the long rains occurring in March to May, and short rains in October through December. Due to the interaction between lake-induced mesoscale flows and the prevailing large scale flows, the rainfall distribution is asymmetric over the lake basin. Thus, the general climate of the lake basin ranges from a modified equatorial type with substantial rainfall occurring throughout the year, particularly over the lake surface, to semi-arid conditions characterized by high frequency of droughts, over some areas. The climate of East Africa in general and Lake Victoria basin in particular are described in detail in Asnani (1993) and Nicholson (1988).

3. Model description and numerical experiments

The NCAR-RegCM2 regional climate model used in this study is described in Giorgi et al. (1993a, b) and it has been customized for eastern Africa domain (Sun et al., 1999a, b). The horizontal grid spacing adopted was 60 km on a 94×85 grid-mesh. The meteorological initial and lateral boundary conditions used to drive the atmospheric component were interpolated from the 4-D ECMWF reanalysis data, constructed from four times daily observations. The lower boundary conditions were derived from the monthly mean sea surface temperatures (SSTs) on $1^\circ \times 1^\circ$ horizontal resolution (Shea et al., 1992).

The parameterization of radiation is based on the NCAR CCM3 package, while the land-surface physics are estimated using the Biosphere-Atmosphere Transfer Scheme (BATS 1e; Dickinson et al., 1993). The use of BATS requires an explicit planetary boundary layer (PBL) to represent the lowest atmospheric level; hence we apply the non-local boundary layer scheme developed by Holtslag et al. (1990). The PBL scheme was implemented with 6 of the 16 model vertical levels located in the PBL and it also includes a 'counter-gradient' transport term for representing non-local transport due to dry, deep convection and turbulent eddy transport. The cumulus parameterization scheme by Grell (1993) was applied, while the split-explicit time integration scheme by Madala (1989) which solves for the shortest gravity wave modes using shorter time-steps than those used for the rest of the model integration period was used. A more detailed description of this model can be found in Giorgi et al. (1993a, b) and Sun et al. (1999a, b).

The simulations started at the beginning of September, through the end of December 1988. This period covers the full short rains season over East Africa. We focus on the year 1988 since it exhibited near normal climate conditions over the region (Ininda, 1998). In the control run we prescribe constant LST of 24 °C at all the 33 lake points within the domain and the length for each numerical integration is four months. To investigate the response of the lake-basin climate to changes in LSTs, a suite of anomaly simulations are performed. In these experiments the LST prescribed in the control run was perturbed by adding or subtracting constant temperature anomaly ranging within ± 1.5 °C across the entire lake surface. These anomaly values are consistent with the observations by Ochumba (1996), Bugenyi and Magumba (1996) over different points on the lake. Lehman (1997) also found that Lake Victoria was more than 0.5 °C warmer in the 1990s than it was in the 1960s which is also within the range of our model sensitivity simulations.

4. Results and discussions

4.1 Validation of the model control simulations over eastern Africa domain

The model control simulations are first evaluated against NCEP reanalysis (Kalnay et al., 1996) monthly wind data, monthly CPC Merged Analysis of Precipitation (CMAP; Xie and Arkin, 1996a) and gauge observations for eastern Africa domain. CMAP data is derived from a combination of gauge observations, satellite estimates and numerical model predictions.

4.2 Comparison of control and NCEP reanalysis 850 hPa wind patterns

In Fig. 2a–f we compare the control simulations and NCEP reanalysis monthly mean flow fields at 850 hPa level for October, November and December. In October, the simulated monthly mean flow pattern (Fig. 2a) and NCEP reanalysis (Fig. 2b) at 850 hPa reasonably agree over most of the domain. There are however, significant differences off the coast of East Africa where the simulated flow south of the equator is mainly southeasterly, westerly along the equator and

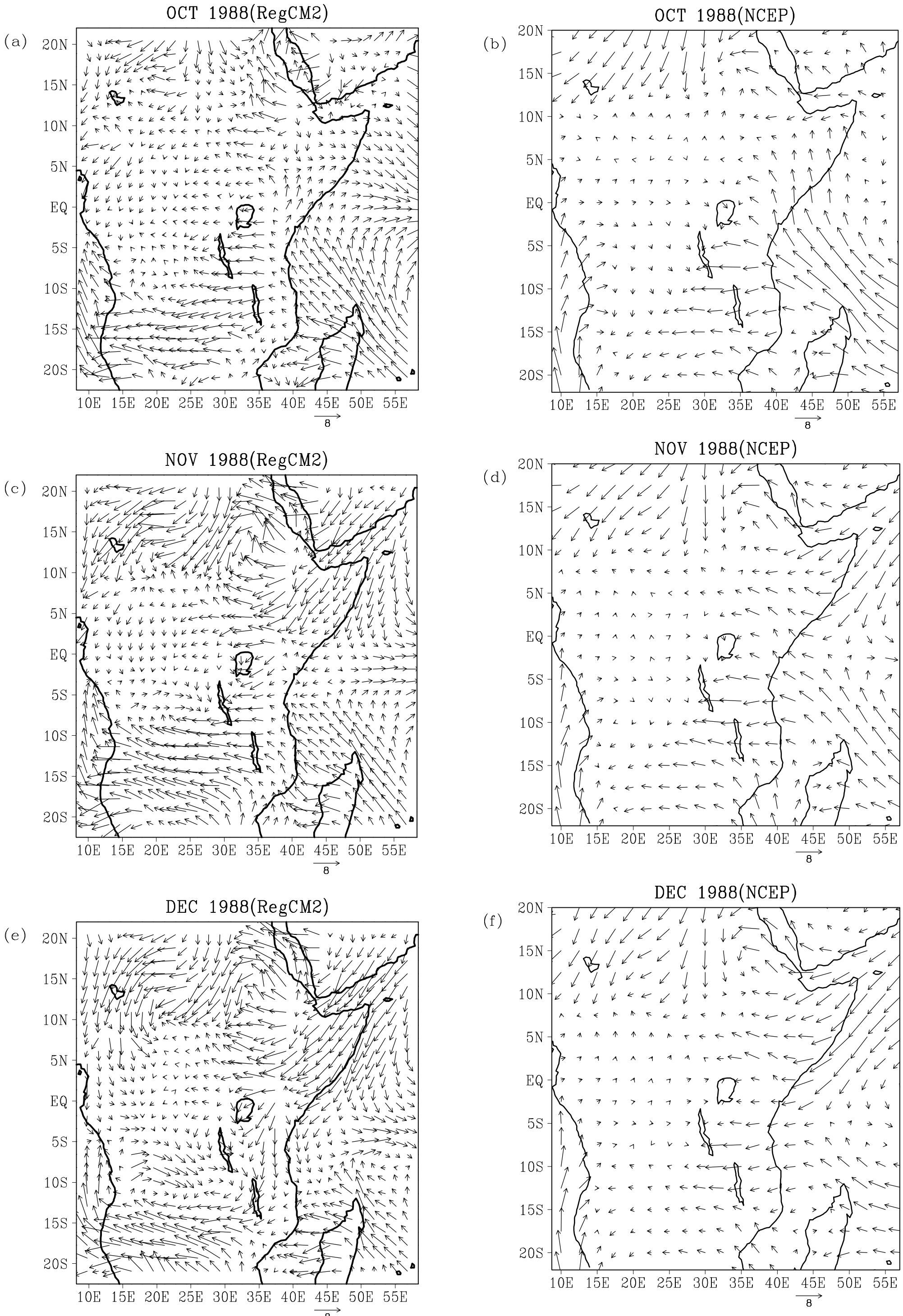

Fig. 2. Comparison of 850 hPa winds between RegCM2 simulations and NCEP reanalysis

turns to southwesterly on crossing the equator. On the other hand, in the NCEP reanalysis the flow along the East African coast is southeasterly south of the equator, but becomes predominantly southerly on crossing the equator. The simulated flow in November (Fig. 2c) indicate the transition of the monsoon flow over the Arabian sea and northern Indian Ocean from southwesterly (southerly) in October (Fig. 2a) to northeasterly which is in agreement with reanalysis wind pattern (Fig. 2d). Significant differences are observed over the Ethiopian highlands, located approximately between 5N–10N and 35E–37E. The simulated flow here bifurcates into southeasterly and northeasterly and flow around the mountain is also evident (Fig. 2c). These features are not present in the reanalysis which shows that RegCM2 resolves some of the sub-grid scale topographical features within our domain that are not represented in the coarse resolution of the NCEP reanalysis. There is also good agreement in the flow patterns between model and reanalysis in December, except the simulated easterly wind flow (Fig. 2e) over the southern parts of the domain is more intense than in the reanalysis (Fig. 2f).

4.3 Simulated monthly mean rainfall compared to CMAP data

Due to the sparse observational data available over most regions in our study domain, we preferred to evaluate our control simulated monthly rainfall against CMAP data. Schreck and Semazzi (2004) have shown, using EOF analysis, that CMAP represents both the spatial and temporal variability of eastern Africa rainfall quite well and is thus a suitable proxy to observations. The simulated rainfall in the control run is compared with CMAP data over the interior part of the domain, covering most regions of eastern and central Africa.

The comparison between the simulated and CMAP monthly mean rainfall for October (Fig. 3a and b, respectively) show significant agreements in the rainfall amounts and distribution over eastern and central Africa. In particular, the regions of rainfall maximum ($\sim$300 mm) over the Democratic Republic of Congo (DR Congo) in the control simulations are consistent with CMAP data. Similarly, dry conditions simulated over northern Kenya, Ethiopia and Tanzania with rainfall amounts less than 50 mm also agree with CMAP data. However, the East African coast is shown to be much drier in CMAP compared to the simulation. In November (Fig. 3c) the shift in regions of rainfall maxima southward in the simulation is relatively coherent with the rainfall amount and distribution pattern in CMAP (Fig. 3d). However, relatively more rainfall is simulated over the East African coast. In addition, rainfall in excess of 500 mm is simulated over eastern DR Congo which may be associated with the effects of the Congo (tropical) rainforest and high topography between the border of DR Congo and Rwanda/Burundi. Over the central and southern Tanzania it is a little drier in the model simulations compared to CMAP data. However, the simulated rainfall and CMAP distribution and amounts are the same over most of the domain in December (Fig. 3e and f). The shift of regions of rainfall maximum farther to the south of DR Congo, southern Tanzania and coast of Tanzania, compared to October (Fig. 3a) and November (Fig. 3c), in the simulations is in agreement with CMAP data.

We conclude that the evolution of the simulated rainfall in the control runs for all the three months indicate that the RegCM2 model reasonably reproduces the climatological pattern as well as the month-to-month (intra-seasonal) rainfall variability over our domain. In general the model captured the migration of the rainfall southward over eastern and central Africa between October and December which is consistent with the north–south movement of the ITCZ over the region during the short rains season. In particular, the gradual increase in rainfall from October, reaching maximum over most areas in November and finally beginning to subside in December over most parts of eastern Africa is clearly evident in the model simulations.

4.4 The simulated mean circulation over the lake basin

The mean circulation fields in the control and anomaly runs at 850 hP are presented in Fig. 4a–i. In the control (CTRL) experiment the mean circulation in the lower troposphere (850 hPa) in October (Fig. 4a) is dominated by easterly flow to the east and south of the lake. To the west of the lake the flow is very weak or almost stagnant. In November (Fig. 4b) the circulation mainly comprise of northeasterly and northwesterly/westerly winds to the east and west of the lake, respectively. The flow is also relatively stronger (by about 2 m/s) and more confluent compared to October. The transition of the large scale wind pattern over the lake basin from predominantly easterly flow to northeasterly flow between October and November is in agreement with

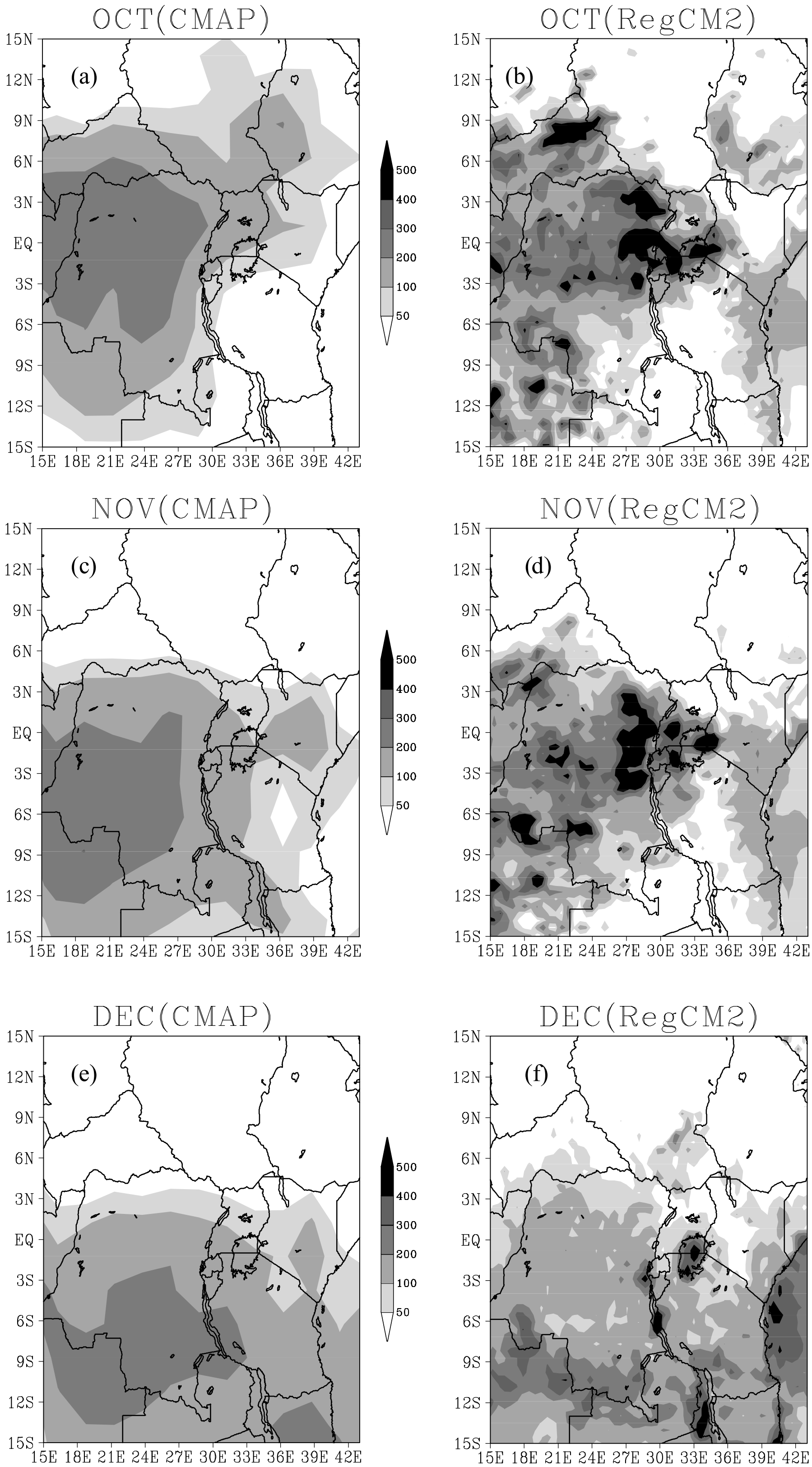

Fig. 3. Comparison of simulated and CMAP monthly mean rainfall over eastern and central Africa for the short rains season, 1988

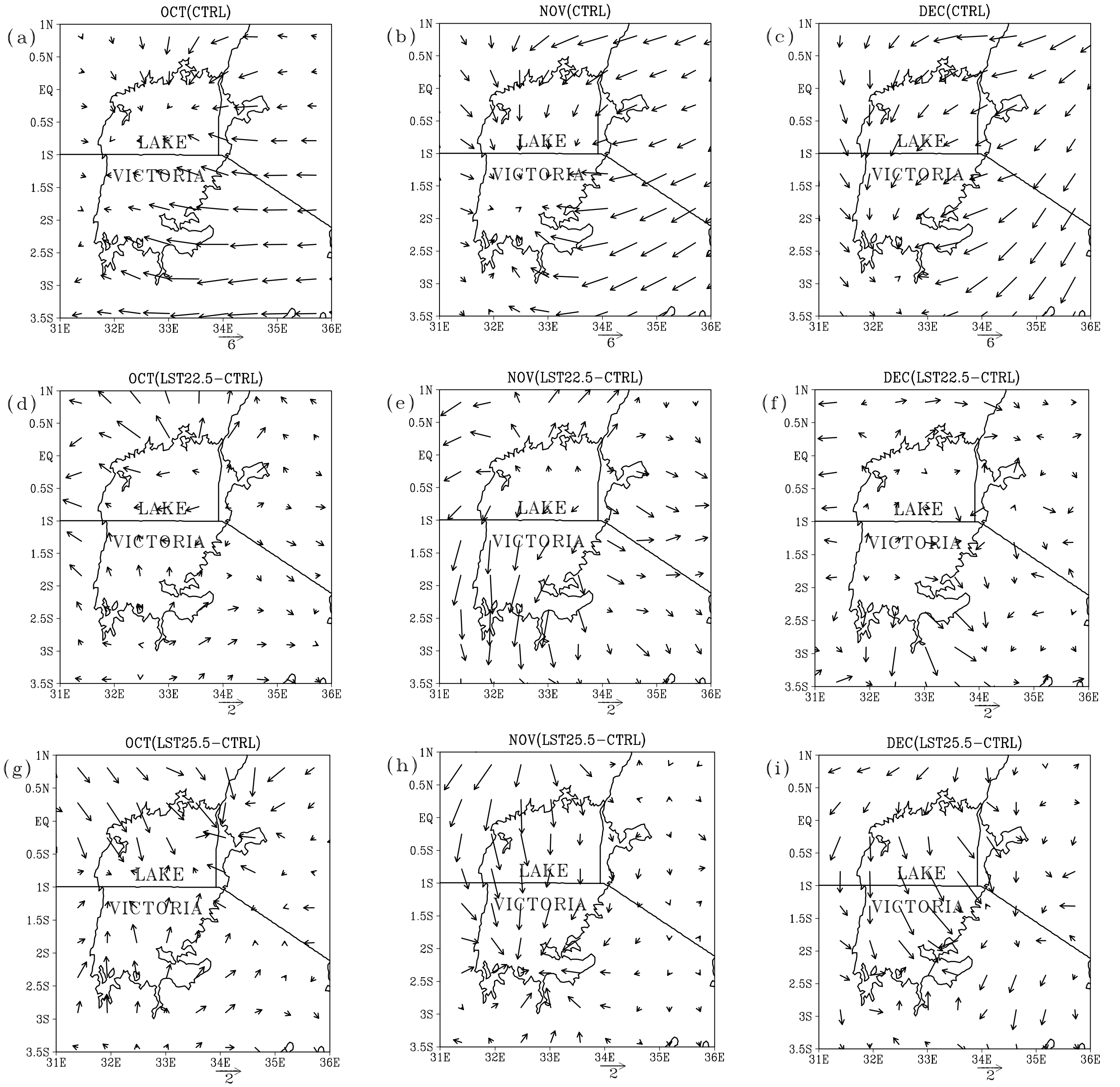

Fig. 4. Simulated monthly mean wind fields at 850 hPa (**a**) $LST_{-1.5}$ expts (top panels) (**b**) $LST_{-1.5}$ minus Control (middle panels) (**c**) $LST_{+1.5}$ minus Control (bottom panels)

the NCEP reanalysis flow pattern during the same months (Fig. 3a–f). In December (Fig. 4c), we find that in the control run the primary region of convergence is shifted farther south of the lake, compared to October and November runs which is consistent with the migration of the ITCZ. Since at this time of the year the ITCZ is approximately located over the lake, the interaction between the prevailing monsoonal flow and the lake-induced meso-scale circulations may enhance convergence and convective development over the lake region.

In Fig. 4d–i, we show the differences between anomaly and CTRL simulated flow patterns. In the cooler LST ($LST_{-1.5}$) minus control for October (Fig. 4d) the 850 hPa flow pattern is predominantly divergent over the northern parts of the Lake. We hypothesize that despite the location of ITCZ over the lake region during this time, cooler LST weaken the ITCZ locally resulting in a more divergent flow pattern over the lake surface as shown in our simulations. The interaction between the ITCZ and the LST is also evident in November (Fig. 4e) and December (Fig. 4f). For, example, there is relatively stronger southerly flow directed away from the lake in $LST_{-1.5}$ minus control simulations in these two months compared to October (Fig. 4d). Thus the monthly simulated flow pattern is consistent with the expected ITCZ southward migration pattern across the lake region.

In the warmer LST ($LST_{+1.5}$) minus control case, the flow pattern show the lake surface to be the most active convergence area during all the three months. The simulated difference for October (Fig. 4g) show that when the lake surface is warmer, the large scale easterly flow and the westerly flow from opposite sides of the lake tend to converge over the northern part of lake. In November (Fig. 4h) and December (Fig. 4i), convergence over the lake still persists in the simulations except the locations have shifted southward. A warmer lake seems to enhance the low pressure center over the lake which in turn interacts with the ITCZ to intensify the axis of flow (moisture) convergence over the lake surface. The warmer lake surface also appears to accelerate the southward advance of ITCZ over the region (Fig. 4i).

4.5 Simulated rainfall over the lake basin

In Fig. 5a–i we show the simulated rainfall over the lake basin in the control and anomaly runs. In the October control simulation (Fig. 5a) maximum rainfall regions are located over the western, central and northeastern sections. The distribution of the simulated rainfall is generally consistent with the flow pattern shown earlier in Fig. 4a which indicate strong easterly flow to the east of the lake and convergence over the northern/central parts of the lake. The effect of the easterly prevailing currents is to shift the region of (moisture) convergence to the west. In November (Fig. 5b) the simulated rainfall is more widely distributed over the lake basin and the rainfall amount is also relatively higher compared to October. There is an NE/SW axis of rainfall maximum across the lake with the central and southwestern parts having an increase of about 150 mm (over 50%) compared to October (Fig. 5a). The better organized and more symmetric distribution of precipitation in November (Fig. 5b) and to some extent in December (Fig. 5c) may be explained by the location of the ITCZ. The simulated rainfall pattern and amounts generally agree well with the flow pattern (Fig. 4b) which shows a well established center of convergence over the lake surface. We suggest that the even distribution and high rainfall amounts in November and December (Fig. 5b and c) demonstrates the interactions among the ITCZ, prevailing monsoon flow and the thermodynamic characteristics of the Lake. Since during November the zonal branch of the ITCZ is located almost across the center of the lake, it helps to induce convergence over the lake leading to widespread rainfall over the basin. This is also consistent with the general observed intra-seasonal climatological rainfall distribution over the lake basin. The seasonal rains normally have a peak in November when the ITCZ is located over the lake as it migrates southward. The simulated rainfall in the control run for December (Fig. 5c) on the other hand indicates significant reduction in rainfall amount over most areas within the catchment, except over the lake and coastal areas to the south. The climatological rainfall pattern over East Africa shows that rainfall normally shifts southwards, into parts of central and southern Tanzania during December following the migration of the ITCZ (Asnani, 1993).

4.6 Rainfall anomaly simulations

We show the differences between simulated rainfall in the control and anomaly runs in Fig. 5d–i. The October $LST_{-1.5}$ minus control runs (Fig. 5d) indicate that there is a negative rainfall anomaly on both sides of the lake, except over the central part of the lake where about 40 mm more rainfall is simulated than in the control. In November (Fig. 5e) less rainfall is simulated over most of the lake basin in the $LST_{-1.5}$ run where the lake surface is much cooler than in the control. Thus in the $LST_{-1.5}$ minus control, negative rainfall anomaly over the entire lake basin indicate that when the lake surface is cooler by 1.5 °C the simulated rainfall amount is less than the control by about 100 mm (~50%). This decrease is consistent with the wind flow

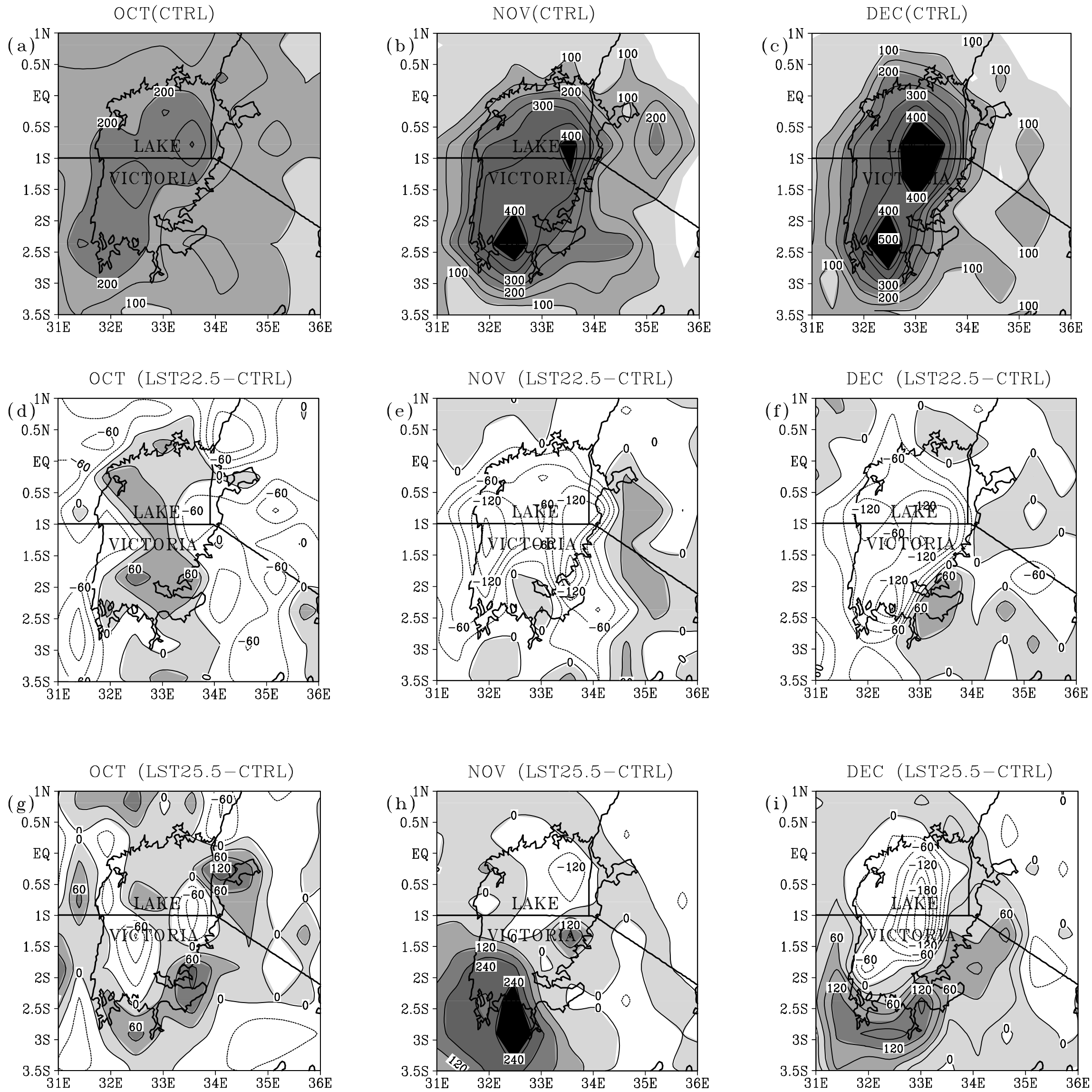

Fig. 5. Simulated monthly rainfall totals (**a**) $LST_{-1.5}$ expts (left panels) (**b**) Control (middle panels) (**c**) $LST_{-1.5}$ expts. (Right panels)

anomaly shown in Fig. 4e where most of the flow is directed away from the lake when the surface is cooler. The flow becomes more divergent and hence does not favor moisture convergence (more rainfall) over the lake basin. Similar rainfall changes are simulated in December (Fig. 5f) which again is consistent with the flow anomaly shown in Fig. 4f.

In contrast, in $LST_{+1.5}$ minus control runs (Fig. 5g–i) we show that the rainfall anomaly is mostly positive, especially in October (Fig. 5g) and November (Fig. 5h). We note dramatic increase in the simulated rainfall when the lake surface is much warmer than the control. In some areas the difference represents more than 100% increase in the $LST_{+1.5}$ runs compared to control. This is also in general agreement with the flow anomaly pattern (Fig. 4g–i) which indicate well organized horizontal wind convergence over the lake when the surface is warmer than the control. For example, in October (Fig. 5g) the simulated rainfall difference between $LST_{+1.5}$ and control runs show significant increase in rainfall amounts especially to the north and northeastern areas and also to the western and eastern regions of the basin. The difference between the two runs indicates positive rainfall bias over the whole basin except for a narrow strip to the west of the lake.

Similarly, in November (Fig. 5h), the simulated rainfall maximum increased by about 300 mm (100%) over the central and southwestern sections of the lake, compared to the control. There is however a negative rainfall bias over the northern parts of the lake which also features in the $LST_{+1.5}$ minus control simulations in December (Fig. 5i). The simulated wind anomaly patterns shown in Fig. 4h and i for November and December, respectively, are consistent with these rainfall anomalies.

Perhaps the most unexpected result is the fact that in December, and to some extent in November, despite the higher LST, the rainfall is significantly reduced over most of the lake surface. We postulate that this is as a result of several factors which combine to contribute to this anomaly. In December the ITCZ is located to the south of the lake as evident in the flow patterns (Fig. 4h and i). We hypothesize that the increased moisture over the lake due to higher evaporation induced by warmer LST is transported by the prevailing flow southward into the ITCZ where it contributes to the dramatic increase in rainfall. We further postulate that the increase in rainfall may help to sustain anomalous latent heat release that drives a local direct circulation over the lake basin. The generation of the local circulation should be expected to promote low-level convergence over the ITCZ and outflow over the lake. We envisage that this could result in a positive feedback process creating a circulation pattern that enhances southward moisture transport in the same sense as the prevailing wind currents.

4.7 Comparison of simulated rainfall over the lake to gauge observations

No direct observational data over the lake were available during this study. Hence, in order to validate our model results we compare model output with observations from a few coastal stations. We also show comparisons for November only since it is normally the peak month for the short rains over the lake basin. The stations are located within less than 50 km from the lake shore (Fig. 1). Stations used are Bukoba (west), Kisumu (northeast), Entebbe (north/northwest) and Mwanza (southcoast). The comparison between the model output and observations is performed by first interpolating the model data to respective station locations. The cubic interpolation scheme is used. The model data are then compared with observations and the climatological long term mean for the 1961–90 base period. The simulated rainfall for Entebbe (Fig. 6a) and Kisumu (Fig. 6b) in both the control

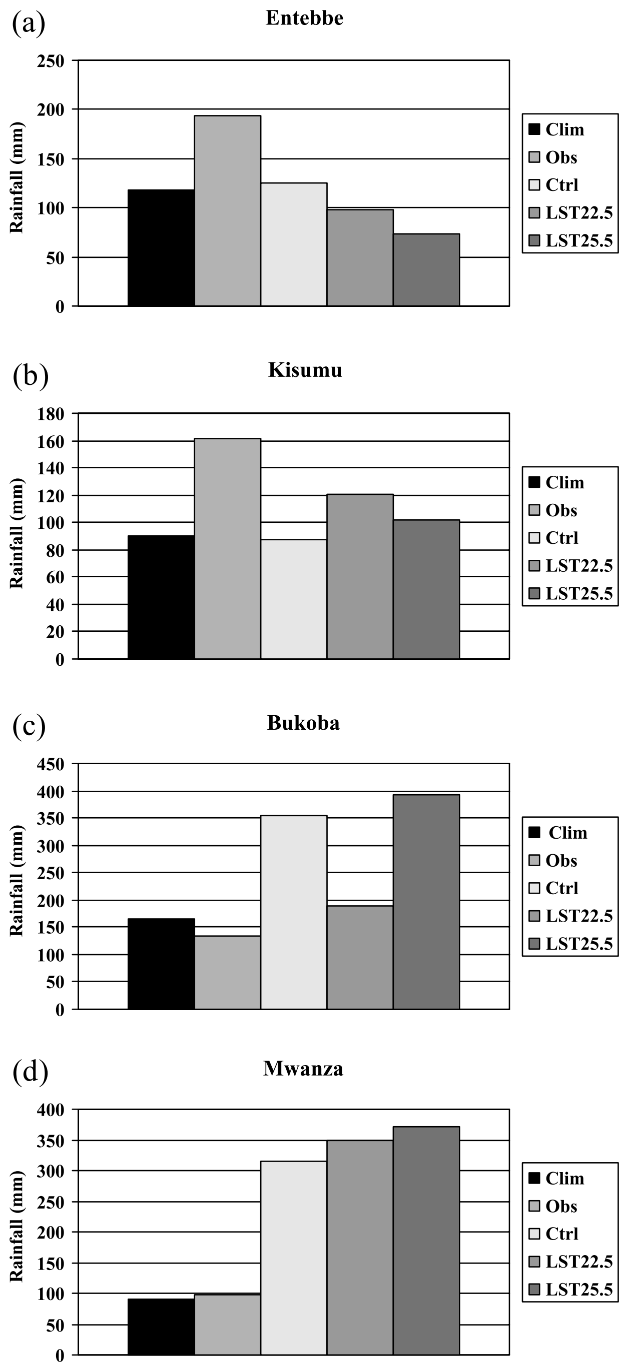

Fig. 6. Comparison of model and observed mean monthly rainfall at four coastal stations. (**a**) Entebbe (**b**) Kisumu (**c**) Bukoba (**d**) Mwanza

and anomaly experiments are less than the observed and the mean climatology. Conversely, the simulated rainfall for Bukoba (Fig. 6c) and Mwanza (Fig. 6d) are much higher than observations. Thus the model underestimates rainfall over the northern/northeastern stations, but overestimates for the stations to the western and southern parts of the lake basin. The simulated rainfall for coastal stations to the north/northeast show relatively high amounts when the lake is cooler than the control than when it is warmer. On the other hand, simulated rainfall for the coastal stations to the west/southwest and southern parts of the lake indicates higher rainfall amounts when the lake is warmer when compared to the control. This asymmetry in the distribution of simulated rainfall is a further manifestation that the lake basin is characterized by complex climate regimes. Hence, the Lake Victoria basin may not be treated as one homogeneous climate zone as has been the case in most of previous studies.

4.8 Evaporation over the lake surface

The over lake evaporation is averaged within longitudes 32E and 34E, and latitudes 2.5S and Equator which covers most of the lake surface. Figure 7a–c shows the averaged evaporation for the control and anomaly runs for the short rains season. For all the control and anomaly simulations evaporation ranges between 80 mm and 147 mm for the coolest and warmest LST, respectively. These values are comparable to results of other previous studies which have used various methods to compute the water budget over the lake. For example, Yin et al. (1998) used infrared satellite data and estimated evaporation over the lake to be about 145 mm in October, 135 m in November and 129 mm in December. However, while our simulated evaporation, especially for the control and $LST_{+1.5}$ cases are within the same range (see Table 1 and Fig. 7), the averaged rainfall over the lake has significant differences with Yin et al. (1998, 2000) estimates. Furthermore, their study estimated rainfall to be in near perfect balance with evaporation but rainfall simulated in both our control and anomaly runs are much higher than evaporation (cf: Figs. 5 and 7). It is difficult to determine from these limited comparisons whether the model simulations or the water budget calculations of evaporation based on limited observations is accurate. The model could have significant positive bias as witnessed in many regional climate models (Giorgi et al., 1993), thus overestimating rainfall and evaporation over the lake. Furthermore, the prescribed LST may have limited more realistic lake-atmosphere interactions to enable simulations of more accurate rainfall and evaporation over the lake basin. On the other hand, the

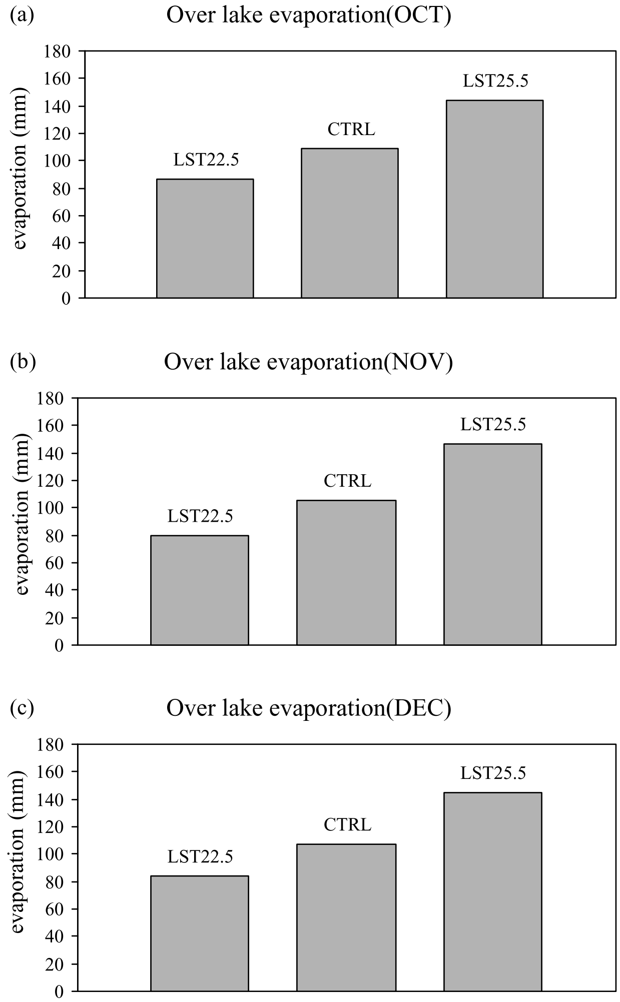

Fig. 7. Evaporation over the lake averaged over (32E–34E, 2.5S-EQ)

Table 1. Evaporation over the lake

Experiment	Oct	Nov	Dec
LST22.5	86.2	79.8	84.0
LST24.0 (CTRL)	108.9	105.0	106.8
LST25.5	144.0	146.7	145.2

observations based on satellite/remote sensing could underestimate lake evaporation because of sub-optimal observational network that may not include calibration (ground truth) observation over lake regions with most intense rainfall.

However, the systematic increase in evaporation over the lake surface as the lake becomes warmer is clearly evident. Another notable result is that when we prescribe LST for each of the control and anomaly runs the rate of evaporation over the lake remained the same within the season. However, the simulated rainfall changed significantly for all the months in the control and anomaly runs. Thus, although there is firm evidence that increase in evaporation over the lake led to increase rainfall, the over all rainfall total is not entirely due to lake evaporation. Also in December (Fig. 7c), and to some extent in November (Fig. 7b), increase in evaporation over the lake does not lead to a proportionate increase in rainfall. In these months more rainfall is simulated in the control than in the $LST_{+1.5}$ cases. This evidence supports our earlier hypothesis that during late November/December the ITCZ position favors moisture transport southward of the lake (into the ITCZ) so that higher evaporation may not necessarily lead to more rainfall over the lake but rather over the ITCZ location.

5. Conclusions

We have demonstrated that there is significant response of Lake Victoria basin climate to changes in the lake surface temperatures using RegCM2 and prescribed LST. However, based on the simulated rainfall anomalies for the short rains the relationship between rainfall anomalies and LST variations is not directly proportional. This may be explained by the complex interactions among the lake-induced meso-scale circulation, topography, prevailing monsoonal flow regimes and the location of ITCZ over the region that may greatly influence the rainfall pattern. Hence, there is need to take into account the full hydrodynamic characteristics of the lake, the strength of the monsoonal flows, topography and other land surface characteristics in order to better understand the contribution of the lake to the regional climate variability.

The results generally show distinct response of the basin-wide climate to changes in the prescribed LST. The intensification/weakening of flow convergence/divergence when the lake is warmer/cooler is clearly evident in the simulations. The model results show realistic month-to-month (intra-seasonal) rainfall variability over the lake basin, with rainfall steadily increasing from October, reaching peak over most areas in November and eventually shifting southwards out of the basin in December. The rainfall-LST response is also attributed to the changes in the locations of the axis of flow convergence/divergence patterns over the lake which is consistent with the expected ITCZ migration pattern over the region.

The simulations show that at warmer LSTs rainfall increased in amount and spatial coverage over most parts of the basin for October and November. On the other hand, as the lake surface gets cooler than in the control the rainfall diminishes in most areas over the lake. Our simulations also revealed surprising response for the month of December when both warmer and cooler (than control) lake results in decreased rainfall over the lake surface. This may be explained qualitatively in terms of the interactions between the prevailing flow, moisture availability over the lake and the location of the ITCZ.

Regions of rainfall maxima in the runs when the prescribed LST is warmer than in control runs by 1.5 °C are located over the western, southwestern and central parts of the lake. In the runs where the prescribed LST is cooler than in control simulations by a similar LST deviation, simulated rainfall reduced over most areas on the basin, except over the center of the lake surface. One notable exception, however, is that in October when more rainfall is simulated for coastal stations to the north and northeastern parts of the lake, represented here by Entebbe and Kisumu, when LST is cooler than the control.

The results from both the anomaly and control runs also display significant differences in the rainfall distribution over the lake basin. In addition, we find that the relationship between the simulated rainfall over the Lake Basin and variations in LST appear to be non-linear although this cannot be conclusively demonstrated by the limited anomaly experiments performed in the present study. This suggests that the lake basin has a complex climate pattern and thus does not necessarily constitute a single homogeneous climate region as has been commonly assumed in many previous studies. The interactions among various local and large scale physical mechanisms such as, ITCZ, monsoon circulations, topography and the lake itself seem to determine the overall distribution and amount of rainfall over the lake basin. This observation is supported by the southward shift in regions of maximum rainfall anomaly which is consistent with the expected ITCZ movement over the region during the October–November–December short rains season. This means better understanding of the temporal and spatial variability of the Lake Victoria basin rainfall should benefit from adapting the full 3-D dynamics of the atmosphere-lake coupling, the influence of the surrounding topography as well as interactions with the prevailing monsoonal circulations.

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Authors' address: Richard Anyah (e-mail: richard_anyah@ncsu.edu), Fredrick H. M. Semazzi, North Carolina State University, Department of Marine, Earth & Atmospheric Sciences, Box 8208, Raleigh, NC 27695-8208, USA.