

Projected Changes in East African Rainy Seasons

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ABSTRACT

A regional climate model with 90-km horizontal resolution on a large domain is used to predict and analyze precipitation changes over East Africa caused by greenhouse gas increases. A pair of six-member ensembles is used: one representing the late twentieth century and another the mid-twenty-first century under a midline emissions scenario. The twentieth-century simulation uses boundary conditions from reanalysis climatology, and these are modified for the mid-twenty-first-century simulation using output from coupled GCMs. The twentieth-century simulation reproduces the observed climate well. In eastern Ethiopia and Somalia, the boreal spring rains that begin in May are cut short in the mid-twenty-first-century simulation. The cause is an anomalous dry, anticyclonic flow that develops over the Arabian Peninsula and the northern Arabian Sea as mass shifts eastward near 20°N in response to strong warming over the Sahara. In Tanzania and southern Kenya, the boreal spring's long rains are reduced throughout the season in the future simulation. This is a secondary response to precipitation enhancement in the Congo basin. The boreal fall "short rains" season is lengthened in the twenty-first-century simulation in the southern Kenya and Tanzania region in association with a north-eastward shift of the South Indian convergence zone.

1. Introduction

Precipitation distributions over tropical East Africa exhibit pronounced regional variations, and the seasonal cycle is complicated. In most regions there are two peak rainfall seasons that are nominally associated with solar-heating maxima in the equinox seasons, but topography, SST forcing, and teleconnections to the West African and Indian monsoon systems are among the other important factors influencing the timing and intensity of seasonal rainfall.

The tropical East African population is dependent on rain-fed agriculture and is, therefore, vulnerable to climate variation and change. Figure 1a shows the annual number of growing-season days in tropical East Africa from the Food and Agriculture Organization (FAO) of the United Nations Natural Resources Climate, Energy, and Tenure Division (FAO 1978). According to the FAO (1978) method, any day on which precipitation exceeds one-half of the potential evapotranspiration rate counts as a growing-season day. There are additional growing-season days

added at the end of the season to take surface water storage into account. The number of growing-season days ranges from less than 30 days in northern Somalia to about 200 days in some regions of Tanzania. Smaller areas in the Ethiopian highlands, around Lake Victoria, and in the Congo basin have 300 growing-season days or more.

Figure 1b is adapted from Cook and Vizy (2012), showing that their 90-km resolution regional model simulations of the late twentieth century are able to reproduce the region's climate with sufficient accuracy to capture the observed distribution of growing-season days over East Africa. The current paper is motivated by projections of large decreases in the number of growing-season days in East Africa for the mid-twenty-first century (2041–60) discussed in Cook and Vizy (2012). As seen in Fig. 1c, large reductions in growing-season days are projected for eastern Ethiopia, Somalia, southern Kenya, eastern Uganda, and much of Tanzania in these simulations.

The purpose of this paper is to better understand the precipitation changes that lead to these simulated reductions in growing-season days over East Africa, since a physical understanding of the change is useful for evaluating confidence in the projections. Background on East African rainy seasons is provided in the next section,

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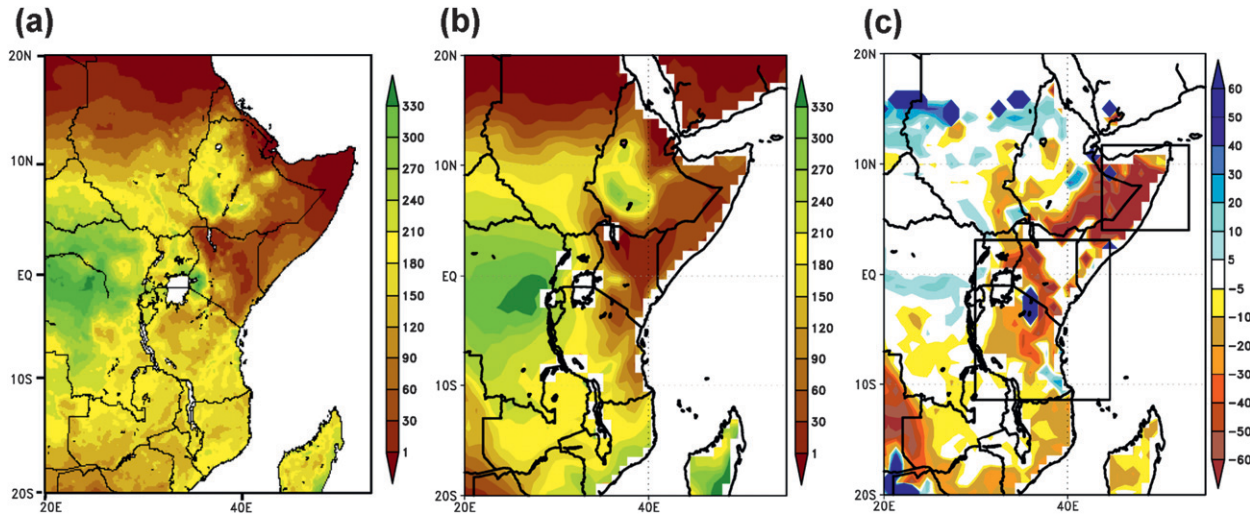


FIG. 1. Number of annual growing-season days in tropical East Africa from (a) FAO (http://www.fao.org/nr/climpag/cropfor/lgp_en.asp) and (b) the regional model simulations of Cook and Vizzy (2012) representing 1981–2000. (c) Percentage differences in the number of growing-season days from regional model simulations between the late twentieth century (1981–2000 mean) and the mid-twenty-first century (2041–60). Averaging regions for Ethiopia–Somalia and Kenya–Tanzania are indicated in (c).

along with an overview of projected changes from increased atmospheric greenhouse gas concentrations. The regional model simulations are described and their realism is evaluated in section 3. Results are presented in section 4, and conclusions are summarized in section 5.

2. Background

Rainfall in tropical East Africa, within about 15° of the equator, is often delivered during two seasons. One rainy period occurs during boreal spring, known as the “long rains” season in Kenya, “Belg” in Ethiopia, and “Gu” in southern Somalia. A second rainy period occurs in the boreal fall over much of the region, although central and northern Ethiopia has only one summer rainy season. In Kenya, this second season is termed the “short rains”. In southern Ethiopia the second rainy season is “Keremt”, and in southern Somalia it is “Deyr”. Here, for simplicity, we adopt the terminology common in Kenya and refer to the boreal spring and fall rain seasons as the long and short rains, respectively.

Herrmann and Mohr (2011) construct a detailed classification of the seasonality of precipitation over Africa based on the high-resolution Tropical Rainfall Measuring Mission (TRMM) precipitation data (Huffman et al. 2007), and display the pronounced regionality and complexity of the rainy seasons in East Africa. In some regions, such as eastern Kenya and southern Ethiopia, the occurrence of two wet seasons is stable and the same pattern is likely to recur each year. However, in northern Tanzania and farther inland in western Kenya, Uganda, and central

Ethiopia, the occurrence of two wet seasons manifests in the climatology but does not recur most years.

Patterns of interannual variability are similarly complex and display pronounced regional variations. Diro et al. (2011) divided Ethiopia into 6 regions to identify teleconnections with sea surface temperature anomalies (SSTAs) and found that each region displayed different teleconnections. The annual cycle of rainfall is the dominant mode of variation, but influences on interannual time scales are significant with, for example, summer drying in central and northern Ethiopia during warm ENSO events (Beltrando and Camberlin 1993; Segele and Lamb 2005; Segele et al. 2009). During boreal fall, rainfall is enhanced in the greater Horn of Africa region during warm events (Mutai and Ward 2000; Bowden and Semazzi 2007).

A positive correlation between rainfall over Ethiopia and India has been identified, including in the early work of Walker (1910) and, more recently, Whetton and Rutherford (1996) and Camberlin (1995, 1997). The modeling study of Vizzy and Cook (2003) reveal pronounced regional variations in the teleconnection. Their process-study simulations show that the monsoon trough that extends from East Africa into southwestern Asia mediates the covariability of Indian and Ethiopian precipitation. When this trough is weak, northern Ethiopia is dry in association with dry advection from the north but the rest of Ethiopia experiences positive precipitation anomalies.

Farther south, over Kenya and northern Tanzania, variations in the short rains have been related to the

Indian Ocean dipole model of SST anomalies. Kabanda and Jury (1999) find that east–west SSTA gradients in the tropical Indian Ocean are associated with an anomalous zonal circulation that enhances convection over the western ocean basin and northern Tanzania. Zorita and Tilya (2002) find that variations in the long rains in March and April are also accompanied by zonal wind anomalies but, in May, they are associated with meridional surface temperature structures that suggest a correlation with the Indian monsoon.

As reported in the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC; Solomon et al. 2007), coupled general circulation models (CGCMs) from the World Climate Research Programme's phase 3 of the Coupled Model Intercomparison Project (CMIP3) multimodel dataset (Meehl et al. 2007) suggest that increasing atmospheric greenhouse gas concentrations will lead to increases in East African precipitation during boreal winter. Shongwe et al. (2011) analyzed a multimodel ensemble of CMIP3 CGCMs and reported that these models project overall upward trends in precipitation rates and intense rainfall events in the twenty-first century, with a decreased propensity for drought. Although they approximate the observed sensitivity of East African precipitation during October–December to Indian Ocean SSTAs, the CGCMs do not generally produce an accurate representation of the East African climate. The very important role of topography in determining East African rainfall distributions (Hession and Moore 2011) and the observed regionality of the rainy seasons (Herrmann and Mohr 2011) demonstrate that a regional approach with higher-resolution simulations is needed.

Other studies project drying over East Africa during this century from greenhouse gas increases. One component of this drying is an increase in evaporation associated with surface warming (Marshall et al. 2012). Patricola and Cook (2010, 2011) simulated large precipitation reductions at the end of the twenty-first century in parts of East Africa for August and September in a regional climate model, but they did not ascribe strong confidence to the projection because the rainfall reductions are coupled with changes in the Indian monsoon, which were not captured well in the model's simulation of the present-day climate.

Williams and Funk (2011) perform a principal component analysis (PCA) of observed Indian Ocean SSTs and East African rainfall and find that a consequence of the observed Indian Ocean warming trend is enhanced subsidence near southern Somalia, eastern Ethiopia, and northern Kenya as well as regional suppression of convection over central Ethiopia and eastern Sudan. Lyon and DeWitt (2012) relate the observed decline in

East African long rains to large-scale SSTA forcing from the tropical Pacific that recurs during boreal spring.

Cook and Vizzy (2012) simulate severe reductions in growing-season days in East Africa during growing seasons of both the long and short rains. Increases in evaporation throughout the continent from increasing surface temperatures play a role in decreasing the number of growing-season days. However, when decreases in precipitation accompany evaporation increases, the severe reductions in growing-season days over East Africa shown in Fig. 1c result. The purpose of this paper is to focus on the months and regions with large reductions in growing-season days as simulated by Cook and Vizzy (2012), and to understand the physical processes of the precipitation changes that are responsible for those reductions.

3. Model simulations and evaluation

The National Oceanic and Atmospheric Administration (NOAA)/National Center for Atmospheric Research (NCAR) Weather Research and Forecasting (WRF; Skamarock et al. 2005) regional model, version 3.1.1, is used with 32 vertical levels, 90-km horizontal resolution, and a time step of 3 min. The top of the atmosphere is set at 20 hPa. Figure 2 shows the full model domain with topography as resolved. A large domain is used to minimize the effects of lateral boundary constraints over East Africa and to allow the development of subtropical anticyclones over the oceans. Over the Ethiopian plateau, the model topography reaches about 2500 m. Mt. Kilimanjaro, near the Kenya–Tanzania border and separated from the Ethiopia topography by the Turkana Channel, extends above 1500 m at this resolution.

The model domain, resolution, and physical parameterizations are chosen based on testing and include the Yonsei University planetary boundary layer (Hong et al. 2006), Monin–Obukhov surface layer, new Kain–Fritsch cumulus convection (Kain 2004), Purdue–Lin microphysics (Chen and Sun 2002), Rapid Radiative Transfer Model (RRTM) longwave radiation (Mlawer et al. 1997), the Dudhia shortwave radiation (Dudhia 1989), and the unified Noah land surface model (LSM; Chen and Dudhia 2001).

An ensemble simulation approach is used. Two ensembles, each with six members, are generated. The first represents the average late twentieth-century conditions for 1981–2000 and is referred to as 20C. Initial, lateral, and surface boundary conditions for each ensemble member are derived from the 1981–2000 monthly climatology in the National Centers for Environmental Prediction reanalysis 2 (NCEP2; Kanamitsu et al. 2002). Lateral boundary conditions for horizontal winds, temperature, relative humidity, and geopotential heights are

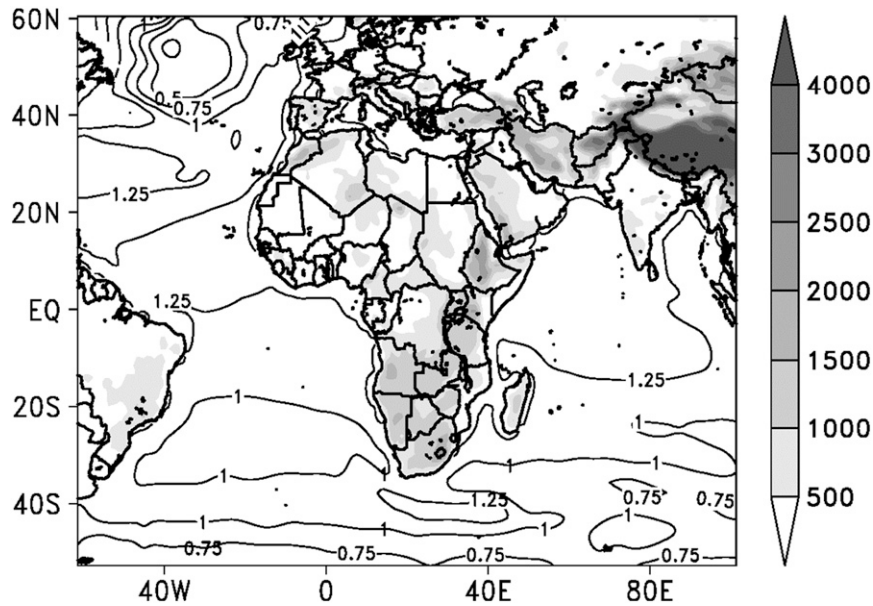


FIG. 2. Regional climate model domain with topography (shaded; m) as resolved at 90-km resolution and SSTAs [contour interval (CI): 0.25 K] as prescribed for the 21C simulation.

updated every 6 h using NCEP2 climatological values. The monthly means are assumed to represent the middle of the month, and boundary conditions every 6 h are generated using linear interpolation. These boundary conditions include seasonality, but synoptic time scales are filtered out. This climate-mode methodology has proved to be useful for tropical and subtropical regional modeling studies over Africa (Vizy and Cook 2002; Hagos and Cook 2007; Patricola and Cook 2010, 2011), although it fails in midlatitude applications where transients must be included on the lateral boundary. Simulations on this domain with synoptically varying boundary conditions confirm the usefulness of the climate-mode approach.

The six, year-long 20C simulations have different initial conditions, as detailed in Cook and Vizy (2012). A 9-month spinup period is used to allow the soil moisture, which is initialized with reanalysis values, to equilibrate with the model climate. Three-hourly output from the six runs is averaged to produce the 20C ensemble mean.

The second ensemble, 21C, represents the average mid-twenty-first century conditions for 2041–60 under the IPCC AR4 A1B emissions scenario, which is a mid-line scenario that is very similar through the twenty-first century to the representative concentration pathway (RCP) 8.5 scenario devised for the fifth assessment report. The atmospheric CO_2 concentration is increased to 536 ppmv, the 2041–60 average in the A1B scenario. (This is 10 ppmv lower than the RCP8.5 estimate for midcentury.) Effects of other greenhouse gases and aerosols are not included.

Boundary conditions for the 21C simulations are derived from CMIP3 CGCM simulations and applied as anomalies to the reanalysis boundary conditions. Cook and Vizy (2012) provide the details and validation of the specification of future boundary conditions. Output from nine CGCMs is used to generate future anomalies by taking differences between the simulated climate mean for 2041–60 under A1B forcing and the climate simulated for 1981–2000. The models used are provided in Table 1. The CGCM anomalies are then averaged to reduce the dependence of the projections on any one GCM. The averaged anomalies are added to the reanalysis boundary conditions for 1981–2000 to form the boundary forcing for 2041–60. While these boundary condition anomalies are derived from GCM simulation, this methodology is not producing regionally downscaled simulations because the imposed lateral and surface boundary conditions do not occur in any one model. This approach, combined with the use of a very large domain, is aimed at producing regional projections that are as independent as possible from GCM projections.

Contours in Fig. 2 show the annual mean of the imposed SSTAs derived from the CGCM simulations. The specified SSTs vary seasonally, but seasonal variations in the anomalies are small. Maximum open-ocean warming within the domain is less than 2 K. A very low-amplitude Indian Ocean dipole pattern is imposed, and the northern tropical Atlantic Ocean is slightly warmer (by about 0.25 K) than the southern tropical Atlantic. Increases of SST in the high-latitude North Atlantic are

TABLE 1. Model research institutions, acronyms, and expansions.

Research Institution	Model Acronym	Model Expansion
Canadian Centre for Climate Modelling and Analysis (CCCma)	CGCM3.1 (T47)	Coupled Global Climate Model, version 3.1 (T47)
Météo-France/ Centre National de Recherches Météorologiques (CNRM)	CNRM-CM3	CNRM Coupled Global Climate Model, version 3
Max Planck Institute (MPI)	MPI-OM	MPI Ocean Model ECHAM5
NOAA/Geophysical Fluid Dynamics Laboratory (GFDL)	GFDL CM2.0	GFDL Climate Model, version 2.0
Center for Climate Research	MIROC3.2 (medres)	Model for Interdisciplinary Research on Climate, version 3.2 (medium resolution)
Meteorological Research Institute (MRI)	MRI-CGCM2.3.2	MRI Coupled Atmosphere–Ocean General Circulation Model, version 2.3.2
National Center for Atmospheric Research (NCAR)	CCSM3.0	Community Climate System Model, version 3.0
NCAR	PCM	Parallel Climate Model
Met Office Hadley Centre	HadCM3	Third climate configuration of the Met Office Unified Model

small because of the slowdown in the Atlantic meridional overturning circulation simulated in the CMIP3 CGCMs (e.g., Gregory et al. 2005).

This approach to simulating future climate with regional models produces projections that are as independent as possible from the CGCM simulations. This is not a traditional downscaling of GCM simulations since all intraseasonal and synoptic variations are generated within the large domain by the regional model dynamics and physics. This approach avoids propagating errors in the twentieth-century simulation from CGCMs into the regional model domain.

Extensive evaluation of the model simulations for the entire continent is presented in Cook and Vizy (2012) and Vizy and Cook (2012). Here, we concentrate on the representation of the East African climate. We also focus the analysis on the ensemble mean, rather than examining each ensemble member. Cook and Vizy (2012) examine individual ensemble members for these simulations and find that there are no strong outliers, and that area-averaged precipitation for each 21C ensemble member falls outside of the envelope defined by the area-averaged precipitation from the 20C ensemble members.

Figure 3 shows precipitation from the 14-yr National Aeronautics and Space Administration (NASA) 3B42V6 TRMM climatology (1998–2011; Huffman et al. 2007) and the 20C simulation averaged over the traditional seasons. (These averaging months are not necessarily optimal for examining rainfall in all regions of East Africa, but they are used here to be concise.) During boreal winter (Fig. 3a), the precipitation distribution is dominated by the South Indian convergence zone (SICZ) that stretches to the southeast from the continental interior (Cook 2000). Precipitation is amplified over the 1000-m topography of Madagascar (Fig. 2). The model simulation captures the structure of the observed SICZ very well

(Fig. 3b), but with rainfall rates that are about double the observed values.

Precipitation rates weaken and rainfall maxima move to the equator in boreal spring, centered in the vicinity of Lake Victoria (Fig. 3c). A similar shift and weakening occurs in the 20C simulation (Fig. 3d). In both the simulation and observations, a rainfall maximum develops over southern Ethiopia, and rainfall rates above 2 mm day^{-1} are widespread over East Africa. Somalia receives about 1 mm day^{-1} during these months.

In the boreal summer months, rainfall is mostly confined to the Northern Hemisphere in both the observations and model simulation (Figs. 3e,f). Again, the simulation produces an accurate distribution of rainfall, but with excessive magnitudes. The rainfall maximum over Ethiopia is farther north than in boreal spring, concentrated on the western slopes of the Ethiopian Highlands (see Fig. 2). Precipitation over much of Kenya and Tanzania is minimal during these months.

The rainfall maximum returns to the equator during the boreal fall season in both the observations (Fig. 3g) and the 20C simulation (Fig. 3h). A secondary maximum lingers to the north over southern Ethiopia and Somalia.

The low-level flow and surface temperature distribution in the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis Interim (ERA-Interim) (Dee et al. 2011) and the 20C simulation are compared in Fig. 4. During the boreal winter, the flow is generally from the northeast over East Africa (Fig. 4a). Over the Horn of Africa, this inflow extends from the Indian winter monsoon and forms the western side of the South Indian Ocean subtropical high. Farther inland, winds consist of northerly flow from the Mediterranean region. The 20C simulation reproduces this flow field well (Fig. 4b). The major discrepancy is that Southern Hemisphere wind speeds in the model are greater than in the

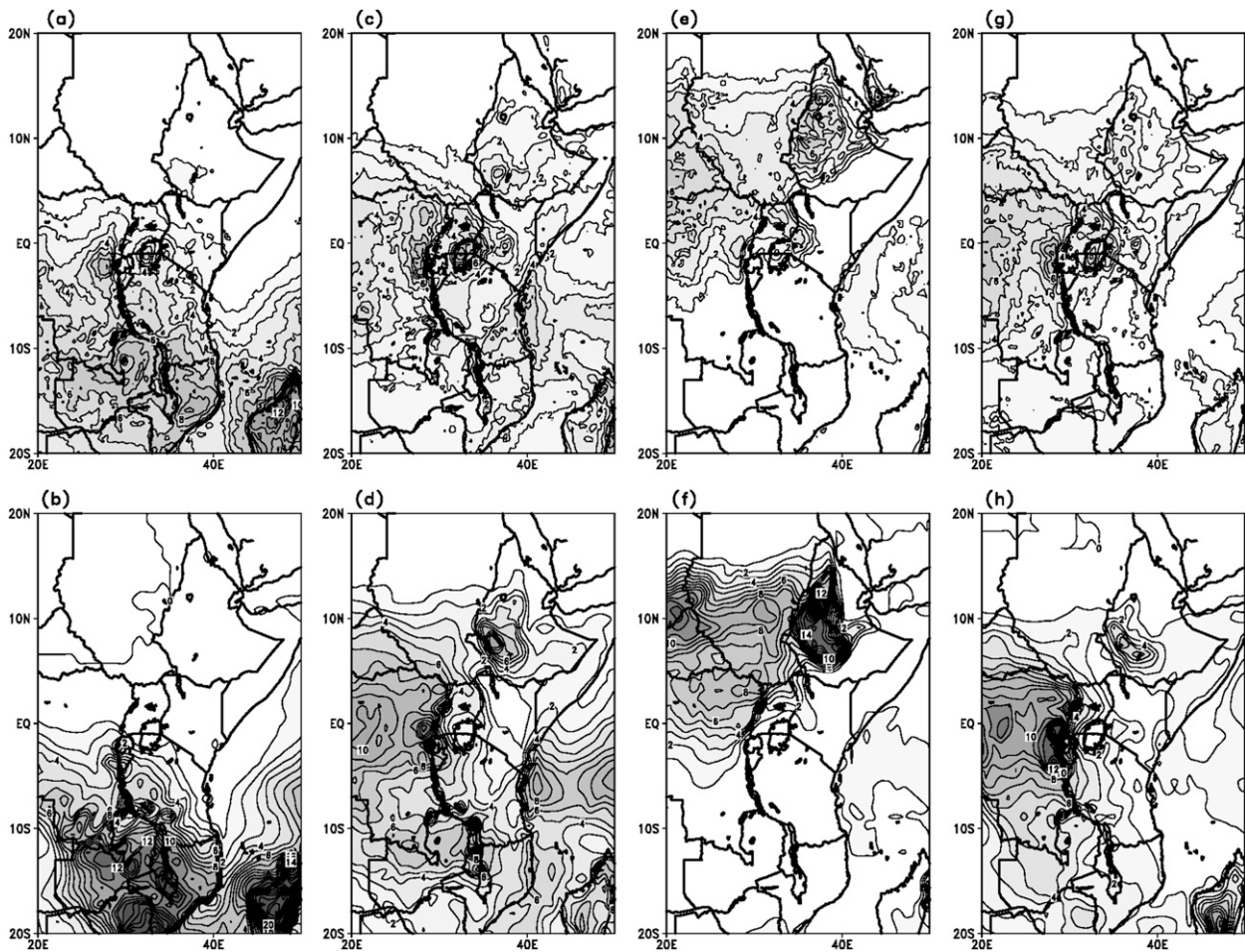


FIG. 3. December–February (DJF) precipitation climatology from (a) TRMM and (b) 20C. March–May (MAM) precipitation climatology from (c) TRMM and (d) 20C. June–August (JJA) precipitation climatology from (e) TRMM and (f) 20C. September–November (SON) precipitation climatology from (g) TRMM and (h) 20C. Shading interval is 1 mm day^{-1} . The CI is 1 mm day^{-1} for precipitation rates between 1 and 4 mm day^{-1} and every 2 mm day^{-1} for precipitation rates above 4 mm day^{-1} .

ERA-Interim, but this could be partially caused by the fact that the reanalysis is on a coarser grid. The surface temperature field is also accurately reproduced in the 20C simulation.

During boreal spring, the flow is directed onto the continent from the Indian Ocean. In the reanalysis (Fig. 4c), the onshore flow is primarily in the Southern Hemisphere and has a southeasterly orientation. In the simulation (Fig. 4d), easterly flow is directed onto the continent along the entire coast, although its penetration inland occurs only over Kenya and Tanzania similar to the reanalysis.

The direction of the low-level flow in boreal summer (Fig. 4e) is opposite to that in boreal winter (Fig. 4a). The Somali jet is fully formed in both the reanalysis and the 20C simulation (Fig. 4f), placing southeasterly flow over East Africa in the Southern Hemisphere and southwesterly flow over the Horn of Africa. Winds extend into the continental interior over Kenya and Tanzania, and also

flow through the Turkana Channel (see Fig. 2). Flow converges from the west and the north over western Ethiopia and Sudan.

Low-level winds in boreal fall (Figs. 4g,h) are similar to those in boreal spring (Figs. 4c,d). The model underestimates the cross-equatorial flow, and overestimates the northerly flow over Sudan.

In summary, the 20C simulation captures an accurate, though not perfect, representation of the East African climate and its seasonal variations. Rainfall rates are excessive, as is typical, but the distribution of rainfall and the circulation features are simulated well.

4. Results

As discussed above, the purpose of this paper is to understand the physical processes of the climate change that lead to the simulated decrease in growing-season days over

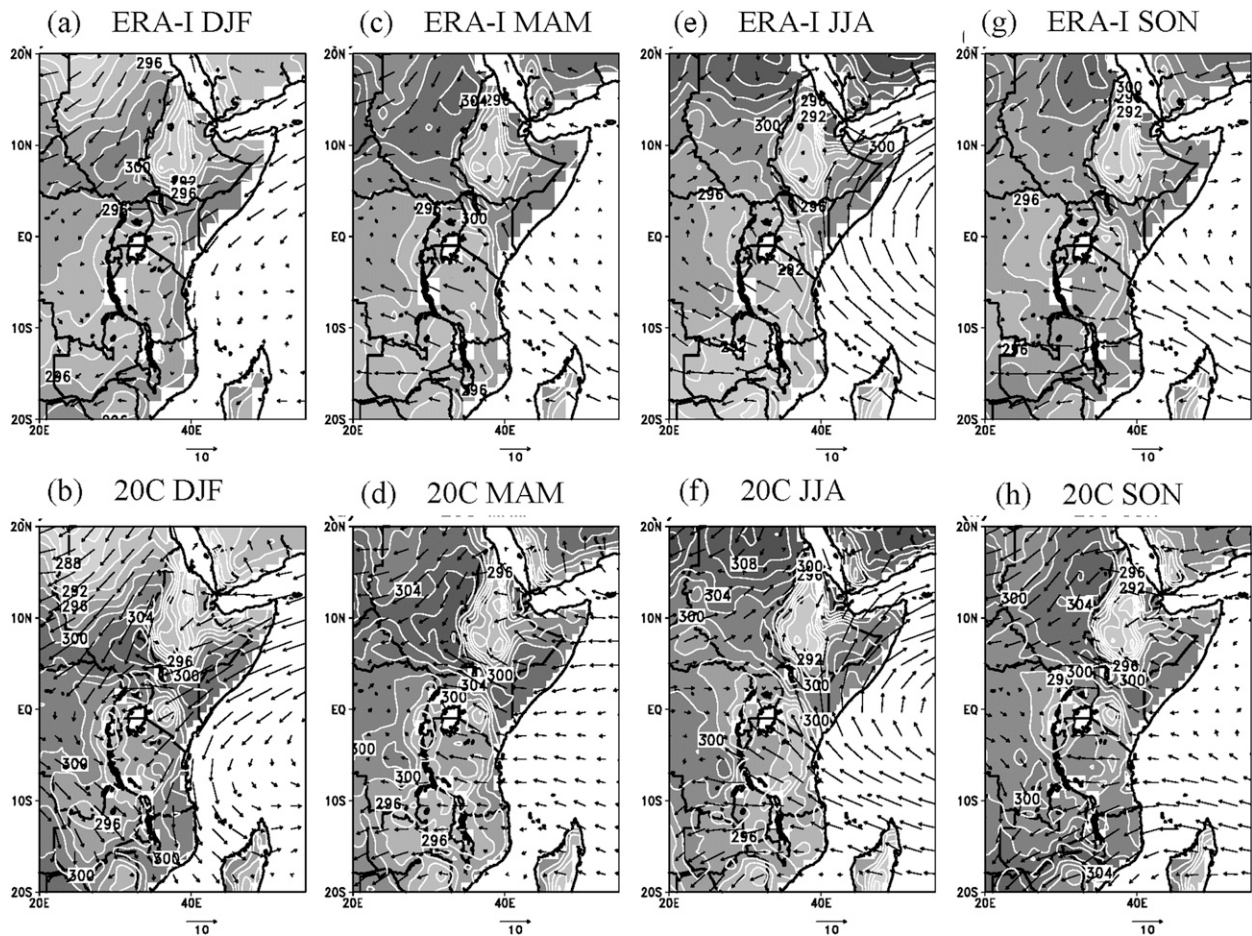


FIG. 4. (a)–(h) Surface air temperature (K) over land and 850-hPa winds (vectors; m s^{-1}) for the traditional seasons from the 1981–2000 ERA-Interim climatology (top) and 20C simulation ensemble mean (bottom). The vector scale is indicated below the panels, and the CI for temperature is 2 K.

East Africa reported by Cook and Vizy (2012; Fig. 1c). Two regions, drawn in Fig. 1c, are defined for the analysis based on the loss in growing-season days. One covers eastern Ethiopia and Somalia (4.0° – 11.7°N , 43.7° – 53.2°E) and the other covers southern Kenya and Tanzania (11.4°S – 3.1°N , 29.9° – 44.6°E). Only the grid points over land within these regions are included in the averaging.

Days on which precipitation exceeds one-half of the potential evapotranspiration rate are counted as growing-season days, with an extension of the season added as water stored in the surface is consumed (FAO 1978). As surface temperatures rise, potential evapotranspiration rates increase and the number of growing-season days decreases. However, changes in precipitation are the dominant cause of the loss in growing-season days in the averaging regions (Fig. 1c), and these are the focus of the current study.

Figure 5a shows the time series of the ensemble-mean precipitation averaged over the eastern Ethiopia–Somali

region for the 20C (black line) and 21C (gray line) simulations. The long rains begin in early May in both the 20C and 21C simulations, but after 10 May the long rains collapse in the 21C simulation. The short rains season (October–November) is similar in the 20C and 21C simulations.

The time series of the ensemble-mean precipitation averaged over the southern Kenya–Tanzania region for the 20C and 21C simulations is plotted in Fig. 5b. The long rains (March–May) are reduced throughout the season in 21C compared with 20C. In contrast, the short rains period is extended in the 21C simulation. In 20C, rainfall rates are greatest in November, exceeding 4 mm day^{-1} for most of the month. In December, relatively high rainfall rates above 3 mm day^{-1} are maintained. In the 21C simulation, the high rainfall rates of November remain, and rainfall rates increase to greater than 5 mm day^{-1} in December.

To understand these results, including their relationship to the climate forcing and their reasonableness, the

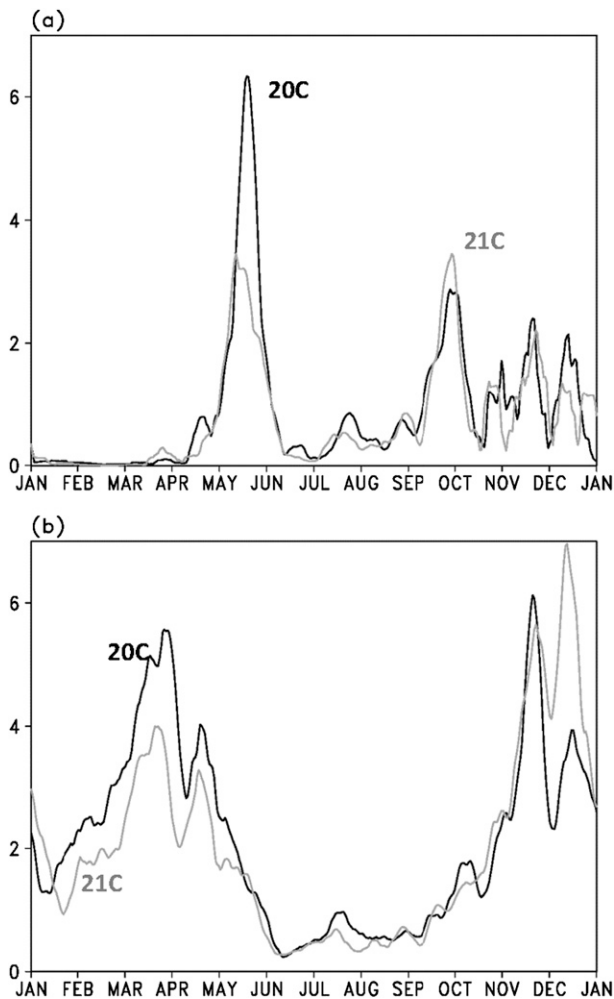


FIG. 5. (a) Daily precipitation (mm day^{-1}) smoothed using an 11-day running mean averaged over the Ethiopia–Somalia region in the 20C (black) and 21C (gray) simulations. (b) As in (a), but averaged over the Kenya–Tanzania region.

climate dynamics of each of these three responses is investigated.

a. Ethiopia–Somalia long rains

Figure 6a shows 800-hPa moisture transport vectors and precipitation from the 20C simulation over East Africa averaged over 11–31 May, the period during which the spring rains fail in the 21C simulation (Fig. 5a) in the eastern Ethiopia–Somalia region. At this time, the meridional branch of the Somali jet has formed (Riddle and Cook 2008) and there is strong, southeasterly, cross-equatorial moisture transport into the continental interior south of about 10°N . The flow splits over northern Kenya and southern Ethiopia (near 5°N , 40°E). One branch continues northwestward through the Turkana Channel, and the other curves northward and converges

over the Horn of Africa as it decelerates. This eastern branch of the meridional Somali jet is the primary source of moisture for the long rains in the Ethiopia–Somalia averaging region, and it supports the 9 mm day^{-1} precipitation maximum over the Horn of Africa.

In the 21C simulation, precipitation rates are less than half their value in the 20C simulation (Fig. 6b). Moisture transport in the western branch of the meridional Somali jet is similar to that in the 20C simulation, but the eastern branch is weaker and diverges over eastern Ethiopia and Somalia.

To understand the association between this regional rainfall response and large-scale greenhouse gas climate forcing, consider the structure of 800-hPa geopotential heights and winds in the 20C simulation shown in Fig. 7a. Subtropical highs are centered near 20°N over Africa, with local maxima over the Arabian Peninsula and the eastern Sahara. Both of these highs extend into the upper troposphere (not shown). The anticyclonic flow associated with high geopotential heights over the Arabian Peninsula places easterly flow to the north of Somalia. Below 900 hPa (not shown), a shallow layer with low geopotential heights is associated with the warm surface temperatures of the Sahara and the Arabian Desert. A shallow vertical circulation system characterizes these dry regions, with high surface temperatures associated with a thermal low and wind convergence below 900 hPa and subtropical highs and wind divergence aloft.

In the warmer 21C simulation, shown in Fig. 7b, geopotential heights rise throughout the domain, but the strengthening of the high over the Arabian Peninsula is especially pronounced. The anticyclonic circulation associated with this high extends southward over the northern Arabian Sea. As a result, the easterly flow that was located to the north of Somalia in the 20C simulation (Fig. 7a) moves to the south, and easterly flow impinges on the Somali coast.

Examining the anomalous fields shows this difference more clearly. Figure 7c shows differences in the 800-hPa geopotential heights (contours) and winds (vectors) across northern Africa for 21C minus 20C. Here, geopotential height differences are normalized by subtracting the average difference in geopotential heights over the region shown in the figure (10.13 gpm) to focus on differences in the gradients since these are related to differences in the flow. The fact that the high over the Arabian Peninsula strengthens more than the Saharan high is evident in these normalized fields since the normalized difference is negative over the Sahara. This structure generates a southerly flow anomaly over the eastern Sahara and Sahel. Over the Horn of Africa, the expansion of the Arabian high generates easterly anomalies north

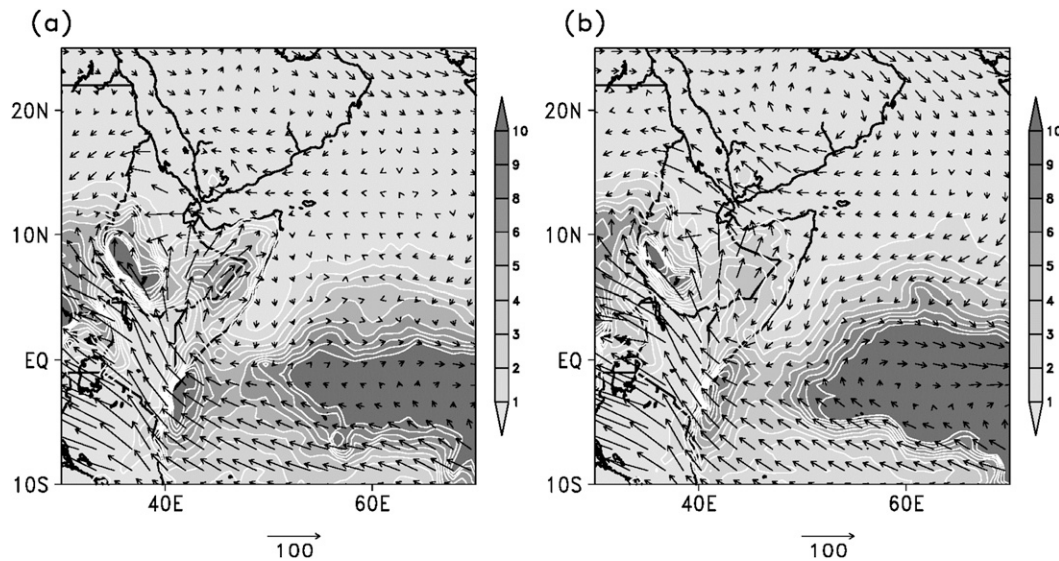


FIG. 6. Ensemble-mean moisture transport (vectors; $10^{-3} \text{ kg m kg}^{-1} \text{ s}^{-1}$) at 800 hPa with precipitation (shaded and white contours; CI is 2 mm day^{-1}) averaged from 11 to 31 May for the (a) 20C and (b) 21C simulations. Precipitation rates greater than 10 mm day^{-1} have uniform shading. The vector scale is indicated below the panels.

of 5°N that suppress the formation of the split jet, cutting off the moisture supply for the spring rains.

The simulated differences in low-level geopotential heights (Fig. 7c) amount to an eastward shift of mass from the Sahara to the Arabian Peninsula. Surface temperature increases over the Sahara in the 21C simulation are larger than those over the Arabian Peninsula (not shown). Warming prescribed over the Indian Ocean is about 1 K, and this relatively modest warming tempers the warming over the adjacent land surface. Over the eastern Sahara, surface temperatures increase by nearly 5 K, but over the Arabian Peninsula the land surface warms by about 3 K or less.

An examination of the atmospheric moisture budget shows that, while the anomalous onshore flow shown in Fig. 7c advects moisture onto the continent, the net result is not a regional precipitation increase. Figure 8a displays 800-hPa mixing ratio and wind differences for 21C minus 20C. The onshore flow anomaly is directed down the anomalous atmospheric moisture gradient over the Indian Ocean until it reaches the coast, suggesting enhanced moisture transport into Somalia and Ethiopia. Since the precipitation anomaly is negative over the Horn of Africa, this positive moisture advection anomaly from the Indian Ocean is clearly not increasing precipitation over the land. The reason is that the flow is divergent, being associated with an anticyclonic circulation system. As seen in Fig. 8b, which shows 800-hPa moisture convergence representing the lower atmosphere, the region over the Indian Ocean with a strong anticyclonic flow anomaly exhibits large-scale divergence.

b. Long rains in Kenya–Tanzania region

The reduction in the long rains in the Kenya–Tanzania region occurs throughout the season, with decreases of 20%–30% persisting from mid-January to mid-May (Fig. 5).

Figure 9a shows 950-hPa moisture transport vectors and precipitation contours from the 20C simulation averaged from 15 January through 15 May. Close to the surface (below 900 hPa), moisture transport is directed onto the continent along the east coast from 18°S to 10°N , and on the west coast into the Congo basin north of 10°S and into West Africa from the Gulf of Guinea. (Nearly all of the southerly moisture transport along the Guinean coast occurs in April and May.) There are two prominent precipitation maxima; one in the Congo basin (near 2°S , 20°E) and the other west of Lake Malawi centered over the Katanga Plateau (near 12°S , 28°E). There is a pronounced precipitation gradient across Tanzania with rainfall rates increasing to the southwest.

Differences for 21C – 20C are displayed in Fig. 9b. Low-level moisture transport into the Congo basin, the Guinean coast, and along the west coast north of 10°S is enhanced by about 10% in the future climate simulation. Over the western Indian Ocean, the moisture transport anomaly is directed toward the northwest, which, when added to the northeasterly flow in the 20C simulation (Fig. 9a), brings the moisture more directly onto the west coast.

Despite this increase in moisture transport along the east coast of Africa, precipitation rates in southern Kenya

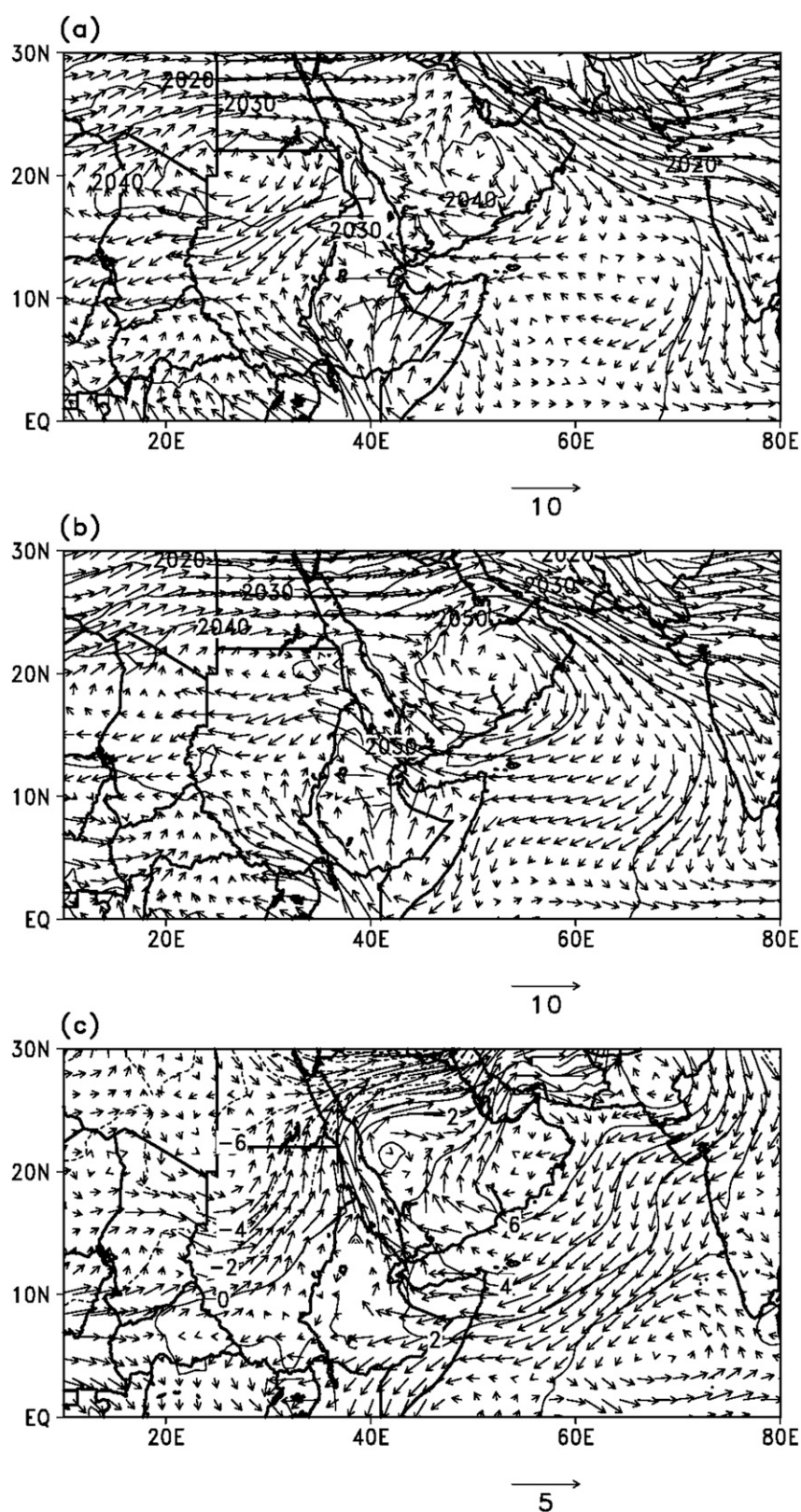


FIG. 7. (a) Ensemble-mean winds (vectors; m s^{-1}) and geopotential heights (contours; CI is 2 gpm) at 800 hPa averaged from 11 to 31 May for the (a) 20C and (b) 21C simulations. (c) Differences (21C - 20C) in the ensemble-mean winds (vectors; m s^{-1}) and normalized geopotential heights (black contours; CI is 2 gpm) at 800 hPa averaged from 10 to 31 May. The vector scale is indicated below the panels.

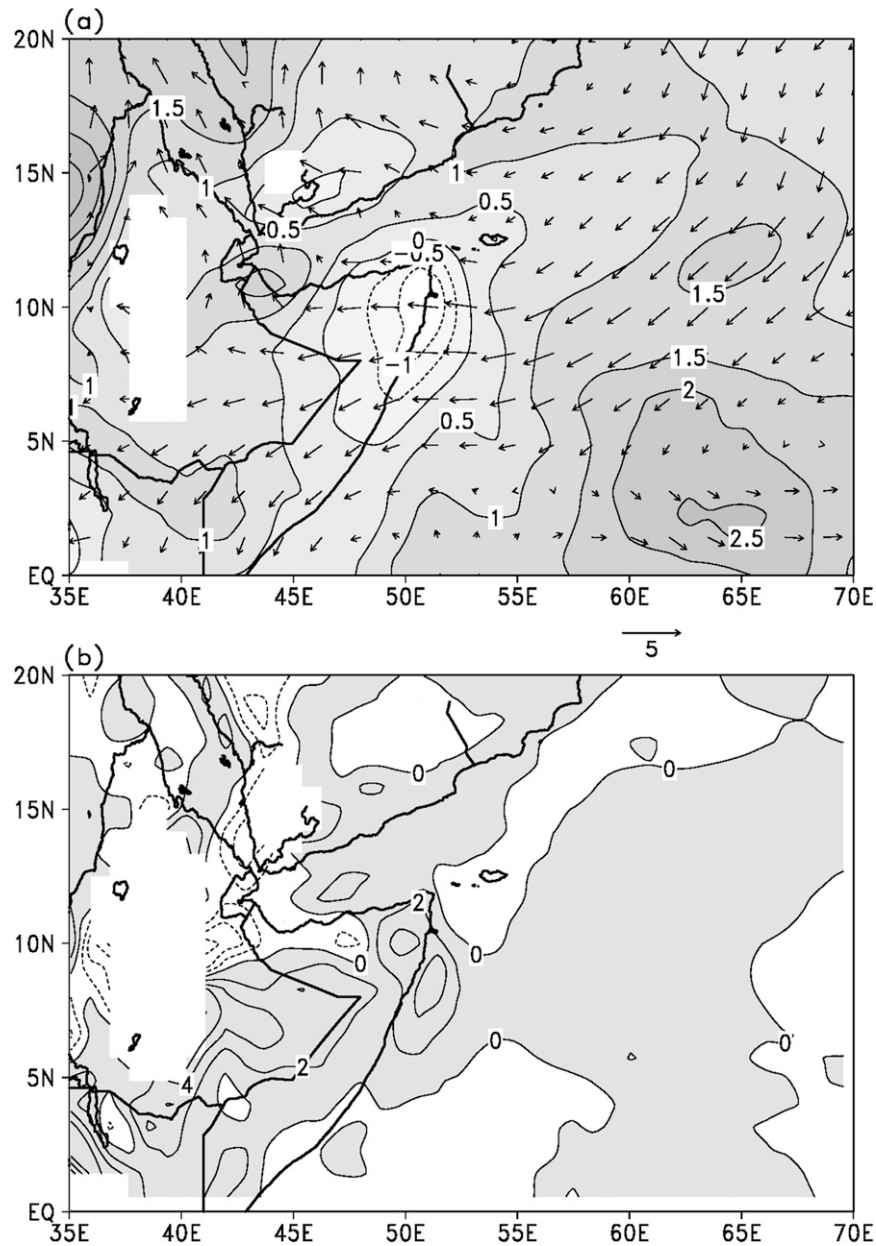


FIG. 8. Differences (21C – 20C) in the 800 hPa ensemble-mean (a) winds (vectors; m s^{-1}) and mixing ratio (contours; CI is 2 kg kg^{-1}) and (b) moisture convergence (CI is $1 \times 10^{-5} \text{ kg s}^{-1} \text{ kg}^{-1}$). All values are averaged from 11 to 31 May. The vector scale is indicated below (a).

and Tanzania are $1\text{--}2 \text{ mm day}^{-1}$ smaller in the 21C simulation than in the 20C simulation, as seen in Fig. 10. In contrast, precipitation rates in the Congo basin are $1\text{--}2 \text{ mm day}^{-1}$ greater in the 21C simulation. The decrease in precipitation over Tanzania and southern Kenya in the 21C simulation is associated with anomalous moisture transport and divergence in the overlying layer that carries moisture away from this region and into the Congo basin. The vectors in Fig. 10 are the 800-hPa moisture

advection differences. A westward moisture flux between 850 and 600 hPa develops in the 21C simulation that moves moisture from Kenya and Tanzania westward into the Congo basin.

While it is not possible to identify cause-and-effect from climatology, one can build consistent hypotheses to test with future analysis. One hypothesis is that the warming imposed in the Gulf of Guinea (about 1 K for the January–April mean; see Fig. 11a) enhances

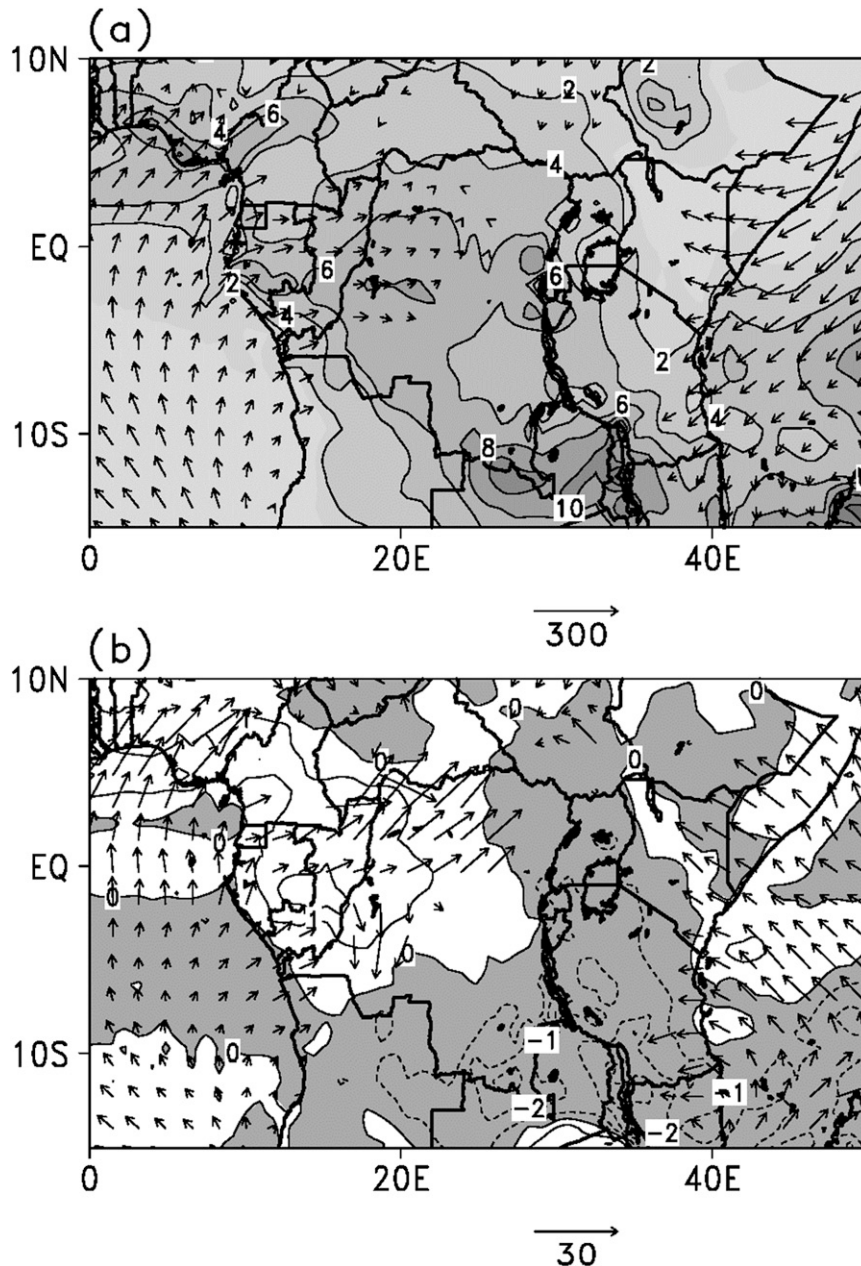


FIG. 9. 15 January–15 May ensemble-mean 950-hPa moisture transport (vectors; $10^{-3} \text{ kg m kg}^{-1} \text{ s}^{-1}$) and precipitation (CI is 1 mm day^{-1} ; negative values shaded) for the (a) 20C and (b) 21C – 20C simulations. The vector scale for each panel is indicated below.

evaporation from the ocean and increases the moisture content of the low-level flow into the Congo basin (Fig. 9). This enhanced moisture flux into the Congo basin triggers a precipitation intensification that is further enhanced and deepened through positive evapotranspiration feedbacks and strong surface water recycling typical of rain forest climates, similar to the intensification of MJO-induced precipitation in this region noted by Tazalika and Jury (2008). While surface temperature increases in the Congo

basin are comparable to those in the Gulf of Guinea, the increase in evaporation is 2–4 times greater (Fig. 11a). This intensification of the Congo basin precipitation results in negative geopotential height anomalies between 850 and 600 hPa (Fig. 11b). As a result, additional moisture converges down the geopotential height gradient in this layer, largely unimpeded by Coriolis accelerations at these low latitudes. The result is moisture divergence and drying over Tanzania and southern Kenya.

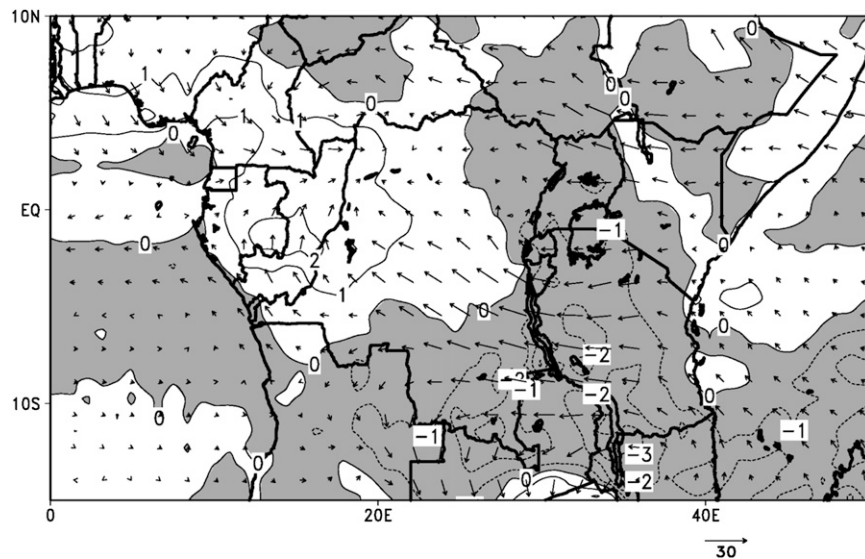


FIG. 10. Differences in the 15 January–15 May ensemble-mean 800-hPa moisture transport (vectors; $10^{-3} \text{ kg m kg}^{-1} \text{ s}^{-1}$) and precipitation (CI is 1 mm day^{-1} ; negative values shaded) for the 21C – 20C simulations. The vector scale is indicated below.

c. Short rains in the Kenya–Tanzania region

The short rains season in southern Kenya and Tanzania in the 21C simulation begins in October and develops through the middle of November as in the 20C simulation (Fig. 5). Rainfall rates decrease in the second half of November in the 20C simulation, but in the 21C simulation they are maintained above 3 mm day^{-1} until early January.

In the 20C simulation, SICZ precipitation extends from the center of the continent to the southeast (Fig. 12a), similar in structure but stronger in magnitude than in the observations (see Fig. 3). This places a pronounced precipitation gradient across the Kenya–Tanzania averaging region (Fig. 1c), with dryer conditions in the northeast and up to 9 mm day^{-1} rainfall rates in the southwest. In the 21C simulation, rainfall rates in the averaging region increase by $2\text{--}4 \text{ mm day}^{-1}$ (Fig. 12b). This regional precipitation increase is associated with a continental-scale pattern of enhanced rainfall along a diagonal to the northeast of the position of the SICZ in the 20C simulation (Fig. 12a), and drying to the southwest. This represents a northeastward shift of the SICZ.

Convergence zones such as the SICZ, the South Pacific convergence zone (SPCZ), and the South Atlantic convergence zone (SACZ) occur where the cyclonic flow about a thermal low converges with the anticyclonic flow about a subtropical high (Cook 2000; Ninomiya 2008). Figure 13a displays 850-hPa geopotential heights and wind vectors over southern Africa for the 20 November–31 December period from the 20C simulation. The

continental thermal low, known as the Angola low, during this period is centered near 20°S , 20°E with the Muchinga Mountains and Rift Valley topography to its east. Cyclonic flow associated with this low is northwesterly from 5° to 15°S over the continental interior. To the east, the warm-season South Indian Ocean subtropical high is centered near 10°S , with the strongest positive zonal geopotential height gradients concentrated along the east coast. The northeasterly flow associated with the subtropical high converges with the northwesterly flow about the thermal low to support the SICZ (Fig. 12a). On synoptic time scales, this region supports numerous tropical-temperate troughs (TTTs) that are associated with intense rainfall events (Todd and Washington 1999; Todd et al. 2004; Hart et al. 2010).

In the 21C simulation, the continental thermal low is strengthened relative to the subtropical high (Fig. 13b), and this repositions the SICZ to the northeast. The result is enhanced rainfall over Tanzania and southern Kenya. This response by the SICZ is similar to the observed northeastward shift of the SICZ and TTTs associated with warming in the Indian Ocean and ENSO warm events (Cook 2001; Reason and Jagadheesha 2005; Manhique et al. 2011), and it is consistent with the increase in intense events in this simulation analyzed by Vizzy and Cook (2012).

5. Summary and conclusions

Regional climate model simulations by Cook and Vizzy (2012) project severe decreases in the number of

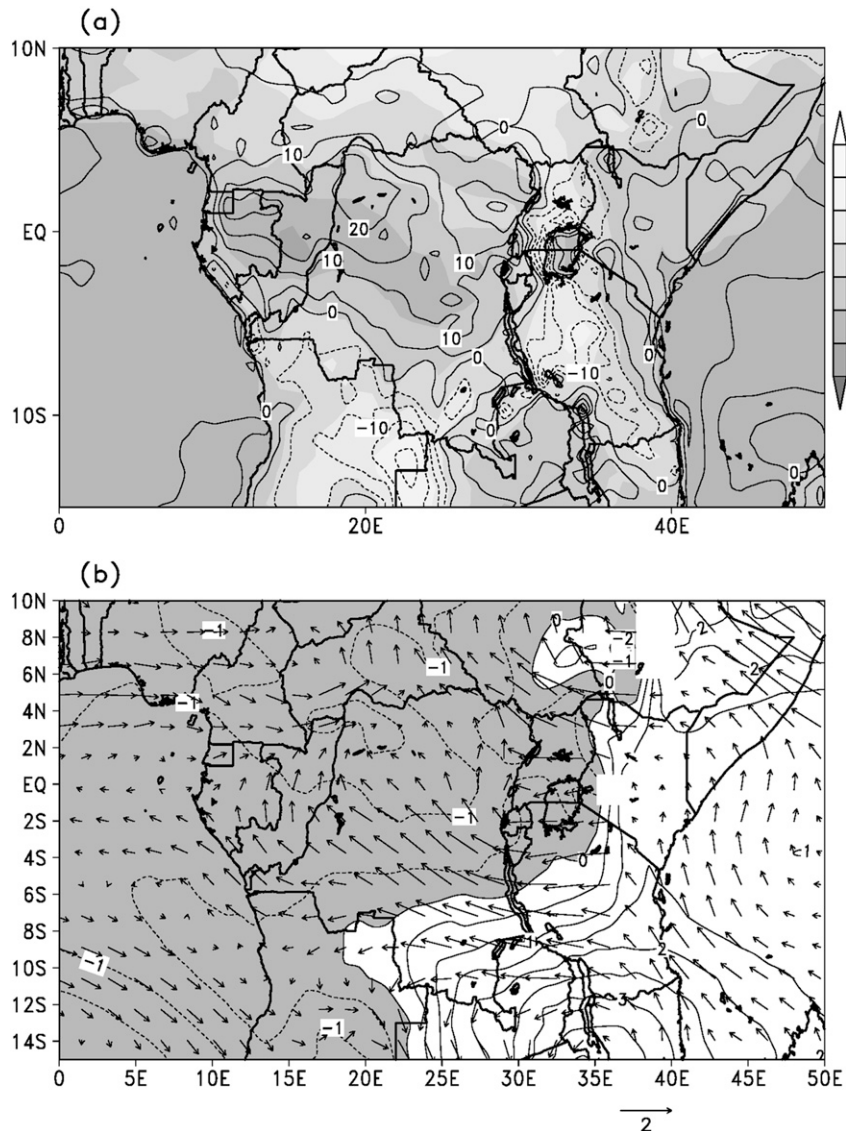


FIG. 11. Differences (21C – 20C) in the 15 January–15 May ensemble-mean (a) surface temperature (shaded; CI is 0.5 K) and the surface latent heat flux (CI is 5 W m^{-2}) and (b) 800-hPa winds (vectors; m s^{-1}) and geopotential heights (CI is 0.5 gpm; negative values shaded). The vector scale is indicated below.

growing-season days in East Africa by the mid-twenty-first century because of increasing greenhouse gas concentrations. The associated reductions in rainfall during the growing season are analyzed here to better understand the physical processes of the change.

A regional climate model with 32 vertical levels, 90-km horizontal resolution, and a time step of 3 min is used over a large domain—including all of Africa and the adjacent oceans—to minimize the effects of lateral boundary constraints and to allow for the development of subtropical anticyclones within the model. An ensemble simulation approach is used, and two ensembles

with six members each are generated. Each ensemble member is a 1-yr integration. One ensemble represents the average late twentieth-century conditions (1981–2000) and the other the average mid-twenty-first-century conditions (2041–60) under the IPCC AR4 Special Report on Emissions Scenarios (SRES) A1B emissions scenario. The twentieth-century simulation uses lateral and surface boundary conditions updated every 6 h from reanalysis climatology. For the twenty-first-century simulation, boundary conditions are modified using anomalies derived from coupled GCM simulations of the 2041–60 period with A1B forcing. This approach provides a

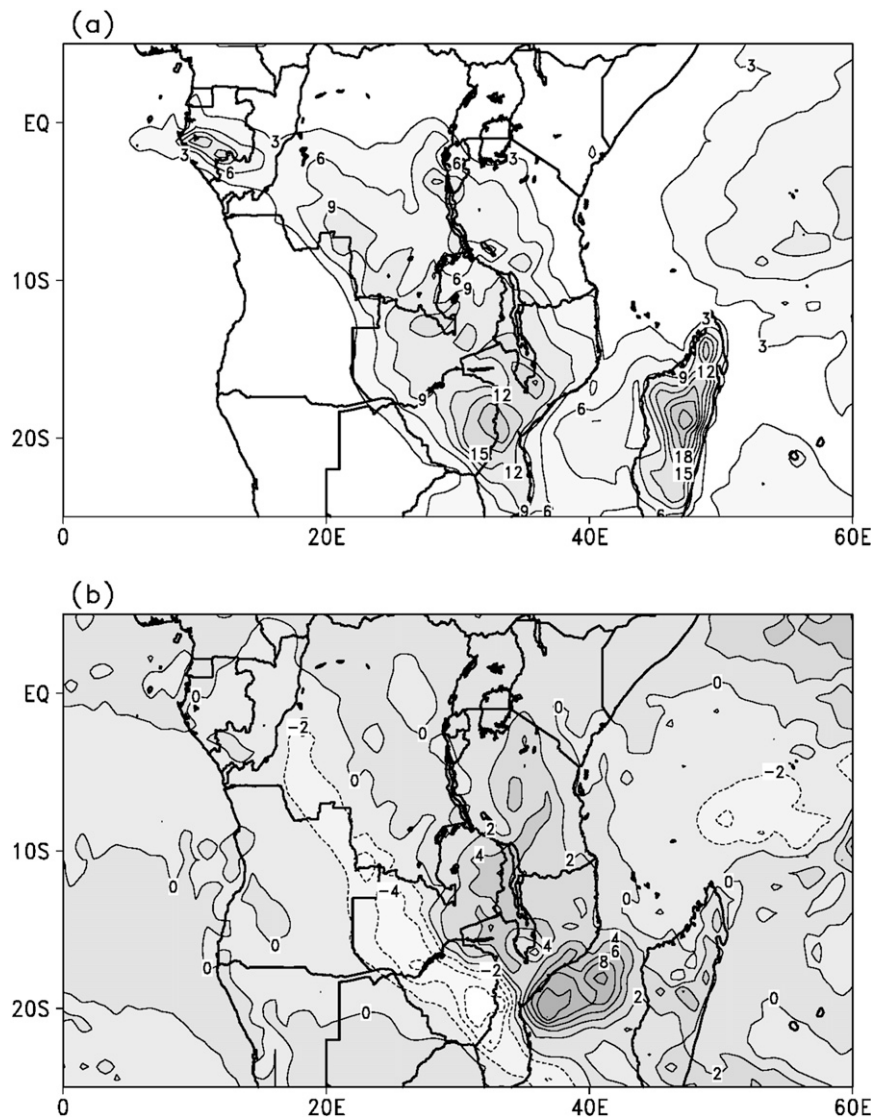


FIG. 12. Ensemble-mean precipitation averaged from 20 November to 31 December for the (a) 20C simulation (CI is 3 mm day^{-1}) and for (b) the difference $21\text{C} - 20\text{C}$ (CI is 2 mm day^{-1}).

properly functioning model climate and avoids propagating errors from coupled GCMs into the regional model domain.

Two regions are defined for the analysis based on the simulated reduction in growing-season days in the twenty-first-century simulation. One covers eastern Ethiopia and Somalia (4.0° – 11.7°N , 43.7° – 53.2°E) and the other covers southern Kenya and Tanzania (11.4°S – 3.1°N , 29.9° – 44.6°E).

In the eastern Ethiopia–Somali region, the boreal spring's long rains begin in early May in both the twentieth- and twenty-first-century simulations, but after 10 May precipitation rates fall dramatically in the twenty-first-century simulation. At this time of year,

the meridional branch of the Somali jet has formed (Riddle and Cook 2008) and southeasterly, cross-equatorial moisture transport flows into the region. The flow splits over northern Kenya and southern Ethiopia (near 5°N , 40°E), and one branch curves northward and converges moisture over the Horn of Africa to support the boreal spring rains in the Ethiopia–Somali averaging region. In the twenty-first-century simulation after 10 May, this northward moisture transport is weakened considerably and the rainy season ends prematurely. The cause is anomalous anticyclonic flow that develops over the Arabian Peninsula and the northern Arabian Sea in the twenty-first-century simulation. The surface temperature over the Sahara warms more than the surface

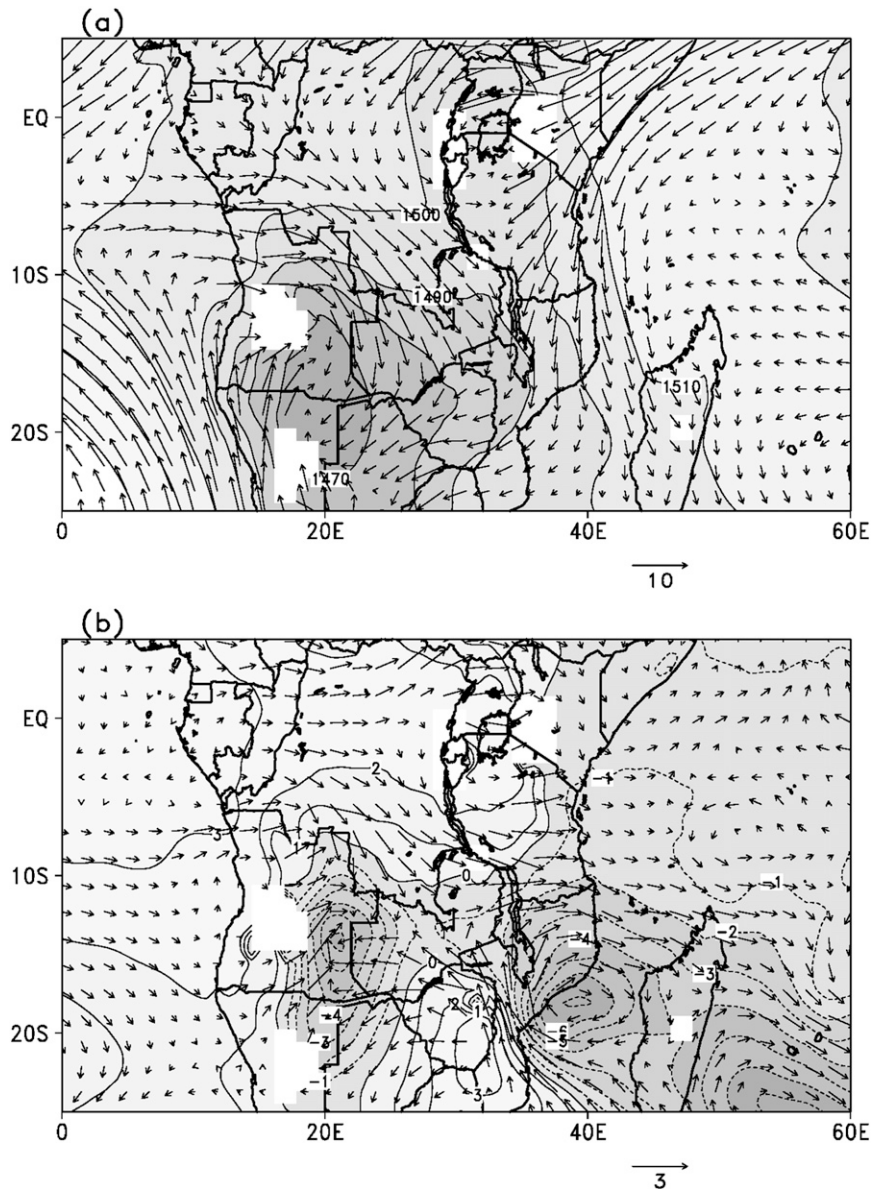


FIG. 13. Ensemble-mean 850-hPa winds (vectors; ms^{-1}) and normalized geopotential height (contours; gpm) averaged from 20 November to 31 December for the (a) 20C simulation (CI is 10 gpm) and (b) the difference 21C - 20C (CI is 1 gpm). The vector scale is indicated below the panels.

temperature over the Arabian Peninsula during this time period, resulting in an eastward shift of mass and higher relative geopotential heights over the Arabian Peninsula. The strengthening and expansion of the high over the Arabian Peninsula generates easterly flow north of 5°N along the East African coast. This flow advects additional moisture onto the continent, but the added moisture is not realized as precipitation because the flow is divergent.

In the southern Kenya–Tanzania region, the long rains (March–May) are reduced throughout the season in the twenty-first-century simulation. This reduction is not

associated with differences in the low-level moisture transport. Rather, drying in the Kenya–Tanzania region is a secondary response to changes in the Congo basin. Low-level moisture transport into the Congo basin, the Guinean coast, and along the west coast north of 10°S is enhanced by about 10% in the future climate simulation. This strengthens and deepens the rain forest precipitation system in the Congo basin. The moisture flux anomaly associated with drying over Kenya and Tanzania is a northwestward moisture transport between 850 and 600 hPa out of the region and into the Congo basin.

The boreal fall, short rains period is lengthened by about two months in the twenty-first-century simulation in the southern Kenya and Tanzania region. This regional precipitation increase is associated with a continental-scale pattern with enhanced rainfall along a diagonal to the northeast of the climatological position of the SICZ, and drying to the southwest. This represents a northeastward shift of the SICZ in response to a strengthening of the thermal low (Angola low). This response by the SICZ is similar to that observed in association with warming in the Indian Ocean and/or ENSO warm events.

This diagnosis of greenhouse gas-induced differences in rainfall over East Africa clearly demonstrate that climate change in this region will be regional and seasonal. Just as the region's climate variability is associated with various forcing factors, climate change will also be the result of different forcings and teleconnections that operate differently in different regions and times of year. Improved prediction requires higher-resolution simulations, in-depth physical analysis, and a regional approach.

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