Recent Atmospheric Variability at Kibo Summit, Kilimanjaro, and Its Relation to Climate Mode Activity

EMILY COLLIER, THOMAS MÖLG, AND TOBIAS SAUTER

Climate System Research Group, Institute of Geography, Friedrich-Alexander University Erlangen-Nürnberg (FAU), Erlangen, Germany

(Manuscript received 15 August 2017, in final form 19 December 2017)

ABSTRACT

Accurate knowledge of the impact of internal atmospheric variability is required for the detection and attribution of climate change and for interpreting glacier records. However, current knowledge of such impacts in high-mountain regions is largely based on statistical methods, as the observational data required for process-based assessments are often spatially or temporally deficient. Using a case study of Kilimanjaro, 12 years of convection-permitting atmospheric modeling are combined with an 8-yr observational record to evaluate the impact of climate oscillations on recent high-altitude atmospheric variability during the short rains (the secondary rain season in the region). The focus is on two modes that have a well-established relationship with precipitation during this season, El Niño–Southern Oscillation and the Indian Ocean zonal mode, and demonstrate their strong association with local and mesoscale conditions at Kilimanjaro. Both oscillations correlate positively with humidity fluctuations, but the association is strongest with the Indian Ocean zonal mode in the air layers near and above the glaciers because of changes in zonal circulation and moisture transport, emphasizing the importance of the moisture signal from this basin. However, the most anomalous conditions are found during co-occurring positive events because of the combined effects of the (i) extended positive sea surface temperature anomalies, (ii) enhanced atmospheric moisture capacity from higher tropospheric temperatures, (iii) most pronounced weakening of the subsiding branch of the Indian Ocean Walker circulation over East Africa, and (iv) stronger monsoonal moisture fluxes upstream from Kilimanjaro. This study lays the foundation for unraveling the contribution of climate modes to observed changes in Kilimanjaro's glaciers.

1. Introduction

Climate change detection and the attribution to anthropogenic or internal atmospheric causes constitute fundamental challenges in the atmospheric and cryospheric research communities. Inherent climate oscillations, such as El Niño–Southern Oscillation (ENSO), mask the long-term climate trends and thereby hamper the detection of recent global warming (e.g., [Pierce et al. 2006](#); [Meehl et al. 2011](#)). Unraveling environmental signals is an especially great challenge in high-mountain

regions, as observational data are sparse and these areas are particularly sensitive to climate change (e.g., [Bradley et al. 2004](#); [Ohmura 2012](#)). However, the character and nature of atmospheric and cryospheric changes at high altitudes is of practical interest, given that they are likely to impact human populations (e.g., [Immerzeel et al. 2010](#); [Kaser et al. 2010](#)).

Numerous studies aim to link glacier changes to ongoing climate trends and internal modes using statistical approaches, focusing on, for example, the relationship between North Atlantic and Pacific modes and glaciers in Europe or North America ([Hodge et al. 1998](#); [Nesje et al. 2000](#); [Marzeion and Nesje 2012](#); [Huss et al. 2010](#)), and between tropical climate modes, such as ENSO, and glaciers at low latitudes (e.g., [Vuille et al. 2008](#); [Francou et al. 2004](#); [Maussion et al. 2015](#)) and midlatitudes (e.g., [Hodge et al. 1998](#); [Bitz and Battisti 1999](#)). While

Denotes content that is immediately available upon publication as open access.

Supplemental information related to this paper is available at the Journal Online website: <https://doi.org/10.1175/JCLI-D-17-0551.s1>.

Corresponding author: Emily Collier, emily.collier@fau.de

Publisher's Note: This article was revised on 3 May 2018 to correct an editing error in the second-to-last sentence of the abstract.

DOI: 10.1175/JCLI-D-17-0551.1

© 2018 American Meteorological Society. For information regarding reuse of this content and general copyright information, consult the [AMS Copyright Policy](#) (www.ametsoc.org/PUBSReuseLicenses).

statistical approaches have proven suitable to relating climate oscillations to glacier mass balance, they are unable to reveal the controlling physical mechanisms. Understanding these mechanisms on all time scales requires process studies at a detailed level. In this study, we aim to provide such a multiscale process-based impact assessment of climate modes on recent atmospheric variability at high elevation as a first step toward decomposing glacier-change signals into internal and anthropogenic parts and identifying potential long-term drivers of glacier mass balance fluctuations.

To achieve this aim, we focus on a case study of Kilimanjaro in Tanzania. Glaciers on Kilimanjaro have retreated dramatically over the twentieth century, with an 85% reduction in ice cover between 1912 and 2011 (Fig. 1; Cullen et al. 2013). In addition to exemplifying strong environmental change signals, this site is selected because (i) the glaciers are exposed to the midtropospheric flow, which may minimize local atmospheric effects on glacier mass balance associated with complex terrain (Sauter and Galos 2016); (ii) we can make use of a long-running observational network for model evaluation (data are available from 2005 through 2013; see section 2a); and (iii) the glaciers have been extensively studied and a great deal of process understanding has already been developed (e.g., Mölg and Hardy 2004; Mölg et al. 2006, 2008, 2009a,b; Cullen et al. 2007; Mölg et al. 2012). These studies have shown that moisture and precipitation are key variables for glacier response, driving the interannual variability in glacier surface energy and mass fluxes (e.g., Mölg et al. 2009a). Seasonal variations in air temperature are less important, both for glacier ablation and the precipitation phase, because these glaciers are located above the mean freezing level all year-round (e.g., Mölg and Kaser 2011). Therefore, moisture and precipitation variability is likely to be the key mechanism through which climate oscillation activity is reflected in glacier mass balance fluctuations on Kilimanjaro.

Equatorial East Africa (EA) is characterized by a bimodal precipitation distribution, with the “long rains” occurring in March–May (MAM) and the “short rains” occurring in October–December (OND), corresponding to the migration of the intertropical convergence zone (ITCZ) and the monsoon transition seasons. Although total rainfall is greater during the long rains, interannual variability of the short rains is greater and shows more spatial coherence (i.e., homogeneity; e.g., Camberlin and Philippon 2002; Black et al. 2003; Hastenrath et al. 2011). Furthermore, since incoming shortwave radiation at Kibo summit peaks between December and February, snowfall during the short rains controls the glacier energy and mass fluxes through the surface albedo more strongly than in other seasons (Mölg et al. 2009a).

Therefore, this study focuses on assessing the impact of internal atmospheric variability on local conditions near the glaciers during the short rains.

Precipitation during the short rains has a well-established relationship with both ENSO (e.g., Indeje et al. 2000; Pohl and Camberlin 2011; Mutai et al. 2012; Hoell and Funk 2014; Nicholson 2015) and the seasonally phase-locked Indian Ocean zonal mode (IOZM; Clark et al. 2003; Behera et al. 2005; Ummenhofer et al. 2009; Nicholson 2015). Precipitation also correlates positively with sea surface temperature in the western Indian Ocean (IO; e.g., Latif et al. 1999; Black et al. 2003; Mölg et al. 2006) and negatively with lower-level equatorial zonal wind, associated with the Walker circulation over the basin (e.g., Hastenrath and Polzin 2005; Hastenrath et al. 2004, 2007, 2010). These relationships show decadal variability (e.g., Clark et al. 2003; Nicholson 2015; Ham et al. 2017), and the interdependence of ENSO and the IOZM has been a prominent topic in the climate dynamics community, with the most recent studies indicating that most IOZM events are ENSO-dependent, or stochastically forced (Stuecker et al. 2017). However, the aforementioned studies have all utilized some combination of data from lowland rain gauges and/or coarse-spatial resolution gridded products, reanalyses, or general circulation models, and are thus not necessarily representative of dynamics at the glacierized altitudes. A small number of studies have utilized observational data to investigate atmospheric variability near the summit to elucidate precipitation dynamics and their relation to the large-scale flow (Chan et al. 2008; Mölg et al. 2009b). However, the short length of the instrumental record available at that time (five years or less) limited elucidation of the physical mechanisms through which mode events impact conditions at Kibo summit.

In this study, we combine subkilometer-grid-spacing numerical simulations with the atmospheric Weather Research and Forecasting (WRF) Model (Skamarock and Klemp 2008) and the aforementioned in situ measurements from the summit of Kilimanjaro to provide the first decadal assessment of drivers of interannual atmospheric variability at the glacierized altitudes, with a specific emphasis on anomalies associated with ENSO and the IOZM states. WRF has previously been used to successfully simulate atmospheric variability in East Africa over seasonal to 1-yr periods at grid spacings of approximately $O(10)$ km (Zhang 2007; Pohl et al. 2011); however, the application of convection-permitting simulations to a decadal period is unprecedented. This analysis provides the first steps for unraveling the contribution of different climate oscillations to the strong observed retreat of Kilimanjaro’s glaciers observed over the twentieth century. In section 2, we describe the data

FIG. 1. Map of Kibo summit, Kilimanjaro, adapted from (Cullen et al. 2013). Gray shading delineates glacierized areas in 2011. The location of the AWSs used in this study are denoted by the italic magenta numbers 3 and 4 (see section 2a). Dashed contours are altitudes (m).

sources and methods we employed in our analysis; in section 3, we present and discuss our results; and in section 4, we provide concluding remarks.

2. Methods

a. Study area and observations

As part of the East African rift system, Kilimanjaro is a massive free-standing mountain that peaks at 5895 m above mean sea level (MSL) (Fig. 1). Ongoing research since the year 2000 includes the installation of a small network of automatic weather stations (AWSs), with four stations in different microclimates of the mountain (Fig. 1). Data from AWS 1 are the subject of a manuscript in preparation by D. Hardy (2018, unpublished manuscript), while AWS 2 was installed 1-m away from a vertical cliff face on the northern ice field and therefore captures microscale circulations that are not represented in the atmospheric model (Winkler et al.

2010). In the present study, we therefore consider data from AWS 3 (3.078°S, 37.354°E; 5873 m MSL), which has been the most extensively analyzed so far (e.g., Mölg et al. 2008, 2009a,b, 2012; Mölg and Kaser 2011), and from AWS 4 (3.081°S, 37.352°E; 5603 m; Mölg 2015). Previous studies demonstrated pronounced annual cycles in moisture-related variables, because of the strong hygric seasonality of this climate zone (cf. section 1; Mölg et al. 2009a,b), while other variables show little variability on scales of daily means or longer (e.g., air temperature and pressure) or no obvious annual cycles (e.g., wind speed).

The relevant measurements from AWS 3 comprise global radiation, incoming longwave radiation, air temperature and humidity at screen level, wind speed, air pressure, and precipitation, while at AWS 4 only air temperature and humidity are available for this study. Details on the employed sensors and processing can be found in Mölg et al. (2008, 2009a). Records are available at AWS 3 from February 2005 to December 2013 and at

TABLE 1. Mean meteorological variables measured at AWS 3 between February 2005 and December 2013, and at AWS 4 between October 2009 and September 2013. (Note that the unit mm w.e. is millimeter water equivalent.)

	AWS 3 (Feb 2005–Dec 2013)	AWS 4 (Oct 2009–Sep 2013)
Global radiation (W m^{-2})	343.9	—
Longwave radiation (W m^{-2})	177.9	—
Air temperature ($^{\circ}\text{C}$)	−7.0	−5.0
Specific humidity (g kg^{-1})	2.6	3.1
Wind speed (m s^{-1})	4.5	—
Air pressure (hPa)	501.8	—
Annual solid precipitation (m)	2.0 (570 mm w.e.) ^a	—

^a Assuming a density of 285 kg m^{-3} (Mölg et al. 2009a).

AWS 4 from October 2009 to September 2013. The mean meteorological variables recorded over their available periods are presented in Table 1.

b. WRF Model

The atmospheric simulations were performed with the advanced research version of the WRF Model, version 3.7.1. The model was configured with three one-way nested domains of 20-km, 4-km, and 800-m grid spacing centered over Kilimanjaro (Fig. 2). The choice of model physics was optimized using test simulations of two months, August 2005 and April 2006, for comparison with previously published WRF simulations of Kilimanjaro (Mölg and Kaser 2011). The resulting configuration (Table 2) is based on that of Collier and Immerzeel (2015), who performed WRF simulations at 1-km grid spacing in the Nepalese Himalaya that compared well against an extensive high-altitude observational network, with the following changes:

- (i) The radiation scheme was changed from the Community Atmosphere Model scheme (Collins et al. 2004) to the Rapid Radiative Transfer Model for General Circulation Models (RRTMG) scheme (Iacono et al. 2008) because of a problem encountered in version 3.6.1 with reproducing incoming longwave radiation values after restarts when using the adaptive time step.
- (ii) The coupled glacier model (e.g., Collier et al. 2013) was not used because of the very small glacierized area on Kilimanjaro ($\sim 2 \text{ km}^2$; Cullen et al. 2013) that exerts minimal feedback on the atmosphere (Mölg et al. 2012).

Amongst the parameterization schemes, Pohl et al. (2011) identified the choice for shortwave radiation, land surface processes, and cumulus convection as being important contributors to WRF-Model spread over East Africa. We employ their best-performing cumulus scheme in the outer model domain (D1) and an improved version of their recommended land surface

model in all domains. However, we selected a different shortwave radiation scheme (RRMTG), as precipitation was strongly underestimated using the Dudhia scheme.

Three modifications were made to the land surface parameters in WRF. First, the default terrain input data was replaced with NASA Shuttle Radar Topography Mission (SRTM) data resampled to 1-km and 500-m grids (<http://www.cgiar-csi.org/data/srtm-90m-digital-elevation-database-v4-1>; accessed 15 December 2015). These data produced a summit height of 5827 m MSL, approximately 70 m lower than in reality, and a topographic height in the closest (glacierized) grid cell to the location of the AWS 3 of 5815 m, compared with the real elevation of 5873 m. Second, vegetation and land-use data were updated following Mölg et al. (2012) based on the data of Hemp (2005, 2006). Third, five grid points were manually classified as glacier based on more recent remotely sensed extents from 2003, 2011, and 2012 (Cullen et al. 2013), producing three and two grid cells for the northern and southern glaciers, respectively. The adjustment provides an overestimate of the glacierized area ($\sim 3.2 \text{ km}^2$) but adequate coverage of the glaciers at the peak. The glacier extents are assumed to be static over the simulation period, as the extent of observed areal decreases occurs at a subgrid scale. Small improvements to parameterization of surface albedo and roughness were also made to the glacier routine in the Noah land surface model with multiparameterization options (Noah-MP; a reference describing the original parameterization is provided in Table 2), comprising a decreased lower bound for albedo (an ice value of 0.32 compared with the previously specified minimum of 0.55) and the introduction of different surface roughness values for snow-covered and snow-free areas.

c. Forcing strategy and simulations

The D1 domain (Fig. 2, top) was forced at its lateral boundaries by the Modern-Era Retrospective Analysis for Research and Applications version 2 (MERRA-2) reanalysis dataset (e.g., Molod et al. 2015) at spatial

FIG. 2. Topographic height (km) of (top) the three WRF domains used in this study, which are centered over Kilimanjaro and configured with grid spacings of 20 km (D1), 4 km (D2), and 800 m (D3) (see Table 2) and (bottom) the finest resolution domain D3. Black contours in the bottom panel delineate terrain heights of 2, 3, 4, and 5 km.

and temporal resolutions of $0.625^\circ \times 0.5^\circ$ in longitude/latitude and 6 h, respectively. We selected this forcing dataset because it represented both monthly mean specific humidity and its fluctuations at 500 hPa compared with the AWS 3 data well. In addition, daily mean Optimum Interpolation Sea Surface Temperature (OISST) analysis data (Banzon and Reynolds 2013; Reynolds et al. 2007), version 2, at 0.25° resolution were used as bottom-boundary forcing.

Grid analysis nudging was applied in D1 to the horizontal winds, potential temperature, and water vapor

TABLE 2. A summary of the WRF domain configuration, physics and dynamics options, and forcing data employed in this study.

Domain configuration	
Horizontal grid spacing	20 km, 4 km, and 800 m (domains 1–3)
Grid dimensions	130×130 , 150×150 , and 150×150
Min/max time step	50/200, 10/40, and 2/8 s
Vertical levels	55
Model top pressure	5000 hPa
Model physics	
Radiation	RRTMG (Iacono et al. 2008)
Microphysics	Morrison (Morrison et al. 2009)
Cumulus	Kain–Fritsch (none in D3) (Kain 2004)
Planetary boundary layer	MYNN level 2.5 (Nakanishi and Niino 2006)
Atmospheric surface layer	Monin–Obukhov (revised MM5) (Jiménez et al. 2012)
Land surface	Noah-MP (Niu et al. 2011)
Dynamics	
Top boundary condition	Rayleigh damping
Diffusion	Calculated in physical space
Lateral boundaries	
Forcing	MERRA-2 reanalysis $0.625^\circ \times 0.5^\circ$, 6-hourly (e.g., Molod et al. 2015)

mixing ratio fields above either the planetary boundary layer or the lowest 10 model levels. The default values for the nudging coefficients were used following Mölg and Kaser (2011), except for the humidity value, which was reduced by a factor of 10 following Otte et al. (2012). Using the aforementioned model configuration and forcing approaches, WRF was used to perform monthly simulations from February 2005 to January 2017. Each simulation included 8 days of spinup, which test runs indicated was sufficient to acclimatize soil and surface fields. A comparison with a continuous simulation from 2005 to 2008 showed minimal differences in monthly mean near-surface atmospheric variables at the location of AWS 3.

d. Climate oscillations

Two indices, the oceanic Niño index (ONI; defined by the Climate Prediction Center of the National Oceanic and Atmospheric Administration) and the dipole mode index (DMI; Saji et al. 1999) were used to diagnose ENSO and IOZM events, respectively. Both indices were computed using the OISST dataset from 1981 through 2016. The ONI was computed as the standardized 3-month running mean of sea surface temperature (SST) in the Niño-3.4 region of the tropical Pacific (5°N – 5°S , 170° – 120°W). The criterion used to identify El Niño and La Niña events was the index exceeding $\pm 0.58^\circ\text{C}$ (normal criterion of $\pm 0.5^\circ\text{C}$ for nonstandardized data,

scaled by the standard deviation of 0.85) for at least five consecutive months. The DMI was computed as the standardized 3-month running mean of the difference in SST between the western (10°N – 10°S , 50° – 70°E) and eastern (0° – 10°S , 90° – 110°E) parts of the IO. A criterion of the mean September–November (SON) index value exceeding plus or minus one standard deviation ($\pm 1\sigma$) was used to identify positive and negative IOZM events. This approach identifies three El Niño (2006, 2009, and 2015) and four La Niña (2007, 2010, 2011, and 2016) events during the study period (Fig. 3). During the key season of influence of the IOZM (SON), three negative events (2005, 2010, and 2016) and four positive events (2006, 2011, 2012, and 2015) are identified. There are four co-occurring in-phase events (2006, 2010, 2015, and 2016), with positive ones emphasized by gray shading in Fig. 3.

e. Analyses

For model evaluation, we performed linear correlation analysis between observational records and WRF data taken from the closest grid cell in domain D3. Because of their proximity, AWS 3 and AWS 4 lie in the same grid cell. To evaluate meteorological variability at lower elevation close to AWS 4, we used the only adjacent glacierized grid cell (model elevation of 5655 m), which is located one grid point to the east of the cell containing both weather stations, producing a modeled elevational difference of 160 m compared with 270 m in reality. We note that AWS 4 is actually installed to the west of AWS 3 (cf. Fig. 1). We also performed linear correlation analysis between monthly anomalies of atmospheric water vapor content simulated by WRF and measured by the MODIS *Aqua* dataset (MYD08_M3; Platnick et al. 2015) and, over the ocean, the SSM/I dataset (e.g., Wentz 1997), both available at a $1^{\circ} \times 1^{\circ}$ spatial resolution. Temporal and autocorrelation was accounted for using the National Center for Atmospheric Research (NCAR) Command Language (NCL) function `equiv_sample_size`, while spatial autocorrelation was addressed using the field significance test developed by Benjamini and Hochberg (1995) and emphasized for climate applications by Wilks (2016).

To examine the response of mountain climate to the two oscillations, following Saji and Yamagata (2003), we performed partial correlations analysis between the indices and standardized 3-month running means of WRF anomalies. Results are presented where they were statistically significant at the 1% level, and for lead-lag analysis, at least two consecutive significant data points were present. We focused our discussion on the simultaneous and lagged response of local summit climate to these modes. For interpreting the correlation results, we note that the IOZM and ENSO peak in boreal fall and winter, respectively.

FIG. 3. Time series of (top) ONI and (bottom) DMI (black curves; data are standardized, as explained in section 2d) for the study period of February 2005–January 2017. Positive and negative events are indicated by red and blue curves, respectively. Vertical gray dashed lines indicate the month of October, the start of the short rains, for each year.

To analyze differences in the short rains, we averaged anomalies from October to January to capture any extensions of the season resulting from climate mode activity and to reduce the influence of the Madden–Julian oscillation, a subseasonal mode of variability with a characteristic period of 30–60 days (Madden and Julian 1994).

Finally, to contextualize the modeling results, we computed the circulation metrics of Hastenrath and Polzin (2005) and Hastenrath et al. (2004, 2007, 2010) using MERRA-2 and OISST data. The metrics include the equatorial lower-level (UEQ; averaged over 4°N – 4°S , 60° – 90°E) and upper-tropospheric [at 200 hPa (U200); 2.5°S – 2.5°N , 60° – 90°E] zonal wind, and vertical velocity at 500 hPa averaged over the western IO and EA (W5w; 2.5°N – 2.5°S , 30° – 50°E). Additionally, the zonal pressure (PWE) and sea surface temperature (TWE) gradients are computed for both fields as the difference between values averaged in the western (8°N – 8°S , 40° – 50°E) and in the eastern equatorial IO (8°N – 8°S , 90° – 100°E). All metrics except for TWE were computed using monthly mean MERRA-2 data between 1980 and 2016, while TWE was computed from OISST data between 1982 and 2016. The deseasonalized monthly data all contained significant linear trends at the 1% level (except for TWE, whose trend was significant at the 5% level) and were therefore linearly detrended before standardized 3-month running means were computed.

3. Results

a. Model evaluation

Figure 4 compares anomalies of meteorological fields measured by AWS 3 and simulated at the nearest grid

FIG. 4. Time series of standardized anomalies of incoming SW and LW, PS, wind speed (WS), T , Q , and PR computed assuming a density of 285 kg m^{-3} (after Mölg et al. 2009a). The black and blue curves show AWS and D3 data while the dashed and solid lines show monthly (used in Table 3) and 3-month running mean (representative of fields used for correlation analysis) data, respectively. The top seven panels show data from AWS 3 while the bottom two panels show data from AWS 4.

point in D3, as well as those measured by AWS 4 and simulated in the adjacent glacierized grid cell at lower elevation. The model represents variability in monthly mean incoming longwave radiation (LW), surface pressure (PS), and specific humidity (Q) particularly

TABLE 3. Linear correlation coefficients (R), mean deviation (MD) and mean absolute deviation (MAD) between monthly mean atmospheric variables (except for precipitation, which is computed as a monthly sum) measured by AWS 3 and the analogous WRF data (see section 2e). All correlations are significant at the 1% level.

	R	MD	MAD
Incoming SW (W m^{-2})	0.75	9.8	14.1
Incoming LW (W m^{-2})	0.91	2.4	9.2
Surface pressure (hPa)	0.91	-1.0	1.0
Wind speed (m s^{-1})	0.73	-0.1	0.5
Near-surface air temperature ($^{\circ}\text{C}$)	0.64	0.4	0.5
Near-surface mixing ratio (g kg^{-1})	0.92	-0.2	0.3
Precipitation (cm)	0.51	5.0	9.3

well (Table 3; Fig. 4). Although discrepancies between the other simulated and observed fields are larger, WRF captures much of the month-to-month variability in local conditions at both weather stations over their available periods. The exception is for simulated precipitation (PR) at AWS 3, which has a relatively low correlation coefficient and the strongest disagreement with observations (Table 3). This fact may be due to both the precipitation treatment in the model and the partly high uncertainty in this observation as a result of the measurement principle (e.g., Mölg and Hardy 2004). A strong positive discrepancy is particularly evident between December 2009 and March 2010, during which time observed anomalies only slightly surpassed 1σ . However, this period coincides with strong positive humidity anomalies at both weather stations and the formation of four tropical cyclones in the western IO, according to the historical track data provided by the Joint Typhoon Warning Center.

As atmospheric humidity and its vertical distribution exert a key control on summit precipitation (Mölg et al. 2009b), we also evaluate the simulation of this field. Figures 5a and 5b compare simulated monthly anomalies of precipitable water in D1 with those measured by the MODIS and SSM/I datasets, respectively. The correlation with both satellite products is high and statistically significant over nearly all of the areas covered in D1, consistent with light nudging of this field (cf. section 2c). It is particularly strong over the moisture source region of the western IO with the SSM/I product, which has been found to capture the variability and trends in humidity over the oceans well (Trenberth et al. 2005). Lower correlation values over lake areas are consistent with the findings of Pohl et al. (2011) of an underestimation of evapotranspiration over these areas by the model. WRF domain D3 has a positive deviation in total-column precipitable water compared with MODIS (Fig. 5c), with a stronger contribution from lower levels

FIG. 5. Linear correlations between deseasonalized monthly anomalies of precipitable water simulated in WRF domain D1 and quantified by the (a) MODIS *Aqua* (MYD08_M3) and (b) the SSM/I satellite products. Stippling indicates statistically significant correlations at the 1% level after correcting for temporal and spatial autocorrelation (see section 2e). The black dot shows the location of Kilimanjaro. Time series of (c) monthly mean and (d) anomaly total-column precipitable water (kg m^{-2}) from the closest grid cell to Kilimanjaro in the MYD08_M3 product and the average over WRF domain D3.

(surface–680 hPa) during the dry season and from upper levels (440–10 hPa) during the rainy seasons (not shown). However, the magnitude and timing of humidity fluctuations are well captured (Fig. 5d; also at all levels), with a correlation coefficient for the total-column amount of $R = 0.84$. We therefore have a relatively high confidence in the model simulation of atmospheric humidity for our study region and focus our analysis on humidity fluctuations rather than precipitation.

b. Variability in local and mesoscale atmospheric conditions

The short rains of 2006, 2009, 2011, and 2015 are characterized by anomalously moist conditions, in particular at higher elevations on Kilimanjaro and in the mid-to-upper troposphere (Figs. 6a,c), corresponding to all positive events except for the IOZM event in 2012 (cf. Fig. 3). Similarly, 2005, 2010, and 2016 are characterized by drier conditions, corresponding to all negative mode events except for weak-to-moderate La Niñas in 2007 and 2011. Anomalies in the vertical profile of moist static energy show a similar pattern with regards

to the mode events; however, their relative amplitude is greater in the midtroposphere around the glacierized altitudes (Fig. 6d), where the rate of change of Q with height is greatest.

In general, the two co-occurring positive events in 2006 and 2015 stand out as being the warmest and wettest, and correspondingly precipitation over much of D3 is anomalously strong during these years (Fig. 6b). However, at higher elevations, the 2009 and 2011 short rains are also strongly positively anomalous. An important contribution to the positive anomalies during El Niño years is the persistence of accumulation into the following January [not explicitly shown but captured by the October–January (ONDJ) averaging period]. The simulated anomalies in moisture and precipitation are consistent with reports of drought conditions in EA in 2005 and 2008 (although the severity of the latter year is not very pronounced in our simulations), and flooding and exceptionally strong snow accumulation at Kibo summit in 2006, and near-average precipitation in 2007 (Hastenrath et al. 2007, 2010; Mölg et al. 2009b).

FIG. 6. Elevational profiles of (a) precipitable water (kg m^{-2}) and (b) precipitation (mm w.e.) averaged in ONDJ (black curve and top x axis), as well as the percent departure for the short rains of each year (bottom x axis, see legend). Vertical profiles of (c) specific humidity (g kg^{-1}) and (d) moist static energy (10^4 J kg^{-1}) and their anomalies. The legend shows the color selection for each year, which was guided by the magnitude of the precipitable water anomalies in (a), as this field is well simulated by WRF in our study region (cf. section 3a).

c. Influence of climate oscillations on high-mountain climate

Figure 7 shows the partial lead-lag correlations between the simulated local atmospheric fluctuations at the glacierized altitudes and the climate mode activity. We focus on the model data because of the longer time series ($n = 144$ for simultaneous correlations) and the ability to spatially average the anomalies, thereby reducing noise in the signal. PS and near-surface air temperature (T) show the strongest positive correlations

over the longest significant ($p < 0.01$) lead-lag periods with ONI, peaking at lags of 2 and 4 months, respectively. Significant positive correlations are found with near-surface Q , lagging the index by 2 months, and weakly with PR approximately simultaneously. Conversely, with the DMI, significant correlations are only found with the radiation fields, a positive one with LW that peaks at lags of 1–2 months and a persistently negative lagged one with incoming shortwave radiation (SW). Considering the observation period and comparing with AWS 3 data, WRF tends to slightly

FIG. 7. Lead-lag partial correlations between ONI (solid lines) and DMI (dashed lines) and anomalies simulated in domain D3 averaged above 5000 m (cf. the innermost black contour in Fig. 2, bottom). Lead (lag) times, where the variable leads (lags) the index, are denoted as being positive (negative). The legend provides the color choices for the correlations with T , Q , PS, WS, incoming SW and LW, and PR. Note that the absolute value of the correlation with SW is plotted.

overestimate (underestimate) the correlation with ONI (DMI; not shown), consistent with the overestimate of precipitation in the short rains of 2009 (cf. Fig. 4, panel labeled PR on the left).

Congruent with the surface signal at Kibo summit, atmospheric anomalies of T and pressure (P) above Kilimanjaro show strong column-wide positive correlations, with the ONI exceeding $R = 0.85$ over a long period (Figs. 8a,b). With the exception of the weak El Niño in 2006, ENSO events are associated with T anomalies around the glacierized altitudes exceeding 2σ (~ 0.7 K; not shown). These results are consistent with the well-established tropical tropospheric temperature response to ENSO (e.g., Chiang and Sobel 2002). There is no significant correlation between these fields and the DMI, confirming that ENSO dominates the fluctuations in T and P at both the local-glacier and mountain scale.

During positive events, the lower-tropospheric (800–600 hPa) southeasterly flow impinging on Kilimanjaro has an enhanced easterly component in boreal fall, coincident with the peak in the DMI. These anomalies are congruent with the easterly anomalies (weaker westerlies) in the equatorial lower-level zonal winds that are characteristic of such events (e.g., Saji et al. 1999).

Conversely, in the upper troposphere lagging the DMI, the climatological weak northerly and northwesterly flow in late boreal fall reverses to easterly flow, and the climatological easterly flow in early boreal winter tends to be enhanced. Thus, positive IOZM and co-occurring mode events are associated with an enhancement of, or regime change to, easterly circulation around Kilimanjaro coincident with and slightly lagging the DMI. Both indices also correlate positively with vertical velocity (m s^{-1}) between lags of about 0 and 5 months, at (~ 600 – 500 hPa) and above (~ 400 – 200 hPa) the glacierized altitudes for ONI and DMI, respectively, while the ONI has a weak positive correlation with meridional wind below 500 hPa between lags of 1 and 6 months (not shown).

Specific humidity has a strong positive correlation with ONI over most of the atmospheric column, from a lead of approximately 2 months through to a lag of 5 months (Fig. 8d). However, there is a gap in the correlation strength and significance in the higher air layers above the glaciers, where the correlation is instead stronger with the DMI. The positive association between Q and ONI reflects fluctuations in the liquid hydrometeors (QL; cloud water and rainwater; Fig. 8e), while fluctuations in the solid hydrometeors (QS; snow, graupel, and ice; Fig. 8f) correlate with the DMI. The latter feature is consistent with observations of anomalously large snowfall at the summit during the short rains of 2006 (Mölg et al. 2009b).

A longitudinal cross section along the latitude of AWS 3 (3.078°S) of the correlation between these oscillations and Q reveals more about the potential contributions of ENSO and IOZM to humidity fluctuations (Figs. 9a,b; data are shown for a lag of 2 months for both indices). Humidity fluctuations correlate more strongly with the ONI in the western part of D1 (Fig. 9a). However, over the western IO, anomalies correlate more strongly with the DMI, in particular between about 500 and 200 hPa and between about 800 and 700 hPa. Similar patterns appear in D3 (Fig. 9b) and a latitudinal cross section along the longitude of AWS 3 (37.354°E) shows a similar feature, namely, a stronger correlation with the DMI than the ONI in the mid-to-upper troposphere above and to the south of Kilimanjaro (Fig. S1 in the supplemental material). The strong positive correlation between humidity fluctuations and ONI primarily reflects the impact of the associated tropospheric temperature changes on the atmospheric capacity for water vapor through the Clausius–Clapeyron relation, as significant correlations with moisture fluxes are generally weak and limited in area (cf. Figs. 9c,d for the zonal component QU; there is a small region of positive correlation between ONI and the meridional

FIG. 8. Lead-lag partial correlations between ONI (color shading) and DMI (pattern-filled shading) and atmospheric anomalies averaged above a terrain height of 2000 m on Kilimanjaro (cf. the outermost black contour in Fig. 2, bottom) as a function of altitude and pressure. The correlations are presented with (a) PS, and (b) T, (c) U, (d) Q, and (e) QL (cloud water and rainwater) and (f) QS (snow, graupel, and ice).

component QV at ~ 500 hPa to the east of the study site; see Fig. S2 in the supplemental material). The relation between the DMI and zonal moisture flux anomalies depends on altitude: strong negative correlations occur in the mid-to-upper troposphere, while positive correlations are present at lower levels to the east of the Kilimanjaro. In the upper troposphere, positive IOZM and co-occurring mode events are associated with easterly moisture transport in late boreal fall and enhanced easterly fluxes in early boreal winter. The flux anomalies result from the aforementioned circulation anomalies in combination with a moister upper atmosphere resulting from enhanced evaporation and convection over the western IO and a higher capacity. The major spatial features of the correlation between zonal moisture flux and the DMI are also reflected in the simultaneous correlation between these anomalies and precipitation at Kilimanjaro, congruent with the importance of east-west moisture transport during precipitation events at the summit reported by Chan et al. (2008). The prominence of the correlations over the western IO in D1 emphasizes the importance of the moisture signal from the IO for moisture and precipitation variability at Kibo summit.

d. Underlying physical processes

Consistent with numerous previous studies of EA rainfall (e.g., Hastenrath et al. 1993; Hastenrath 2001; Hastenrath and Polzin 2005; Hastenrath et al. 2010, 2011; Mutai et al. 2012; Nicholson 2015), changes in the

intensity of the Walker circulation in the IO during boreal fall drive the simulated interannual fluctuations in precipitation (Fig. 10). The IO Walker cell is characterized by surface westerlies at the equator (UEQ), ascending air over the Maritime Continent, easterlies in the upper troposphere (U200), and subsidence over the western IO and EA (W5w). The reduced large-scale subsidence in the western branch and the weakening of the lower-level westerlies at the equator during positive mode events leads to enhanced precipitation around Kilimanjaro (Fig. 10), since convection is less suppressed and lower-level moisture export from EA is reduced. Conversely, the opposite relationships are generally present for reduced precipitation. The changes in the Walker circulation metrics are particularly apparent for co-occurring in-phase events (Fig. 10, bottom). In addition, these events exhibit relatively cohesive anomaly patterns in the total-column moisture fluxes (Fig. 11), with positive (negative) events associated with enhanced (reduced) monsoonal transport upstream from Kilimanjaro, corresponding to southeasterly and northeasterly fluxes during October–November (ON) and December–January (DJ), respectively (Fig. 11; anomalies for other events' years are shown in Fig. S3 in the supplemental material).

4. Discussion and conclusions

In this study, we combined convection-permitting atmospheric modeling and an 8-yr observational record to

FIG. 9. A longitudinal cross section at the latitude of AWS 3 of the partial correlation between anomalies of specific humidity in (a) D1 and (b) D3 and the ONI (color shading) and the DMI (pattern-filled shading) at a lag of 2 months, as a function of altitude and pressure. (c) Longitudinal and (d) latitudinal cross sections of the partial correlation between anomalies of the zonal moisture flux in D1 and the ONI (color shading) and the DMI (pattern-filled shading) at a lag of 2 months, as a function of altitude and pressure. (e) Longitudinal and (f) latitudinal cross sections of the simultaneous (lag 0) linear correlation between anomalies of the zonal moisture flux and of precipitation, as a function of altitude and pressure. The correlation with precipitation averaged over all of D3 is indicated with color shading, while the correlation with precipitation averaged above 5000 m is indicated by pattern shading.

investigate climate oscillation impacts on atmospheric variability at Kibo summit, Kilimanjaro, during the short rains from 2005 through 2016. We focused on two modes known to have strong relationships with rainfall in the region, ENSO and IOZM. Using partial correlation analysis, we demonstrated a signal of both modes in local and mesoscale atmospheric conditions at Kilimanjaro. A key result was strong positive associations between humidity fluctuations and both oscillations,

but a stronger correlation with the IOZM event index in the air layers at and above the glacierized altitudes related to changes in zonal circulation and moisture transport. Interannual fluctuations in humidity and therefore precipitation also covary with changes in the intensity of the Walker circulation in the IO during boreal fall, as has been reported in many previous studies. These associations, in combination with enhanced upstream total-column monsoonal moisture

FIG. 10. A comparison between anomalies of (a) average ONDJ precipitation, as measured at AWS 3 (black dots) and as simulated by WRF at the closest model grid point (blue circles) and averaged over all of D3 (purple asterisks), and (b) average ONDJ metrics of the Walker circulation in the IO (from MERRA-2 data; see section 2e). Note that the observed anomalies are based on a shorter period (2005 through 2013) compared with the WRF data (2005 through 2017).

fluxes, mean that co-occurring positive events are associated with the most anomalously wet and warm conditions during the short rains at Kilimanjaro during our study period.

The importance of co-occurring positive events is consistent with previous studies covering longer periods, which showed that during such events, positive SST anomalies in the western IO are stronger, and easterly anomalies in the 850-hPa zonal winds extend farther west, across the entire IO Basin (Hong et al. 2008). Furthermore, the El Niño event extends the persistence of positive SST anomalies in the western IO by inducing a basin-scale warming in boreal winter following the IOZM event (Xie et al. 2002; Chowdary and Gnanaseelan 2007), which likely contributes to the extension of the short rains into January during El Niño years (cf. section 3b). This result is also consistent with a recent multidecadal analysis of low-altitude rain gauge data from the region (Otte et al. 2017) and reports of highly anomalous precipitation during other strong co-occurring positive events (e.g., in 1982 and 1997; Mölg et al. 2009b; Luo et al. 2010). Previous work has shown that the ONI and DMI indices have a significant positive correlation, in particular if the DMI index slightly leads the ONI (e.g., Ham et al. 2017; Stuecker et al. 2017). Co-occurring in-phase events may reflect one form of ENSO-forced (dependent) IOZM events compared with stochastically forced (independent) ones (Stuecker et al.

2017), where the latter can occur in the absence of ENSO or with the opposite phase. In any case, changes in the frequency of co-occurring in-phase events and how they may have contributed to the historical evolution of Kilimanjaro's glaciers needs to be constrained.

The atmospheric variables recorded by AWS 3 and simulated by WRF are also influenced by other modes of interannual as well as intraseasonal atmospheric variability. Two factors in particular deserve further discussion for our study period. First, in December 2006, the intense (category 4) tropical cyclone Bondo formed on 19 December in the western IO, made landfall on Madagascar on 25 December, then dissipated on 28 December (Mauritius Meteorological Services 2008). This period coincides with exceptionally strong simulated accumulation and relatively uniform high humidity at both Kibo summit and in domain D3 (Fig. S4 in the supplemental material). If mean conditions in December for all years are computed excluding the last 10 days of the month, 2006 and 2015 remain the most anomalously moist short rains in our simulation period; however, it is clear that the tropical cyclone Bondo makes an important contribution toward the extremeness of the former year (e.g., Fig. S5 in the supplemental material). Therefore, the impact of tropical cyclones on conditions at Kibo summit represents an important avenue of future research.

Second, periodicities of approximately 2–3 yr in spectral analysis of long-term EA rainfall records have been attributed to the quasi-biennial oscillation (QBO; Rodhe and Virji 1976; Ogallo 1982; Nicholson and Entekhabi 1986), although we note that these studies predate the seminal papers on the IOZM of Saji et al. (1999) and Webster et al. (1999). Indeje and Semazzi (2000) reported significant correlations between the globally averaged equatorial stratospheric 30-hPa zonal wind index and precipitation during the long rains in the western parts of EA. Furthermore, as noted by the authors of the study, there is a potential for interactions between QBO-driven upper-tropospheric zonal-wind anomalies and those generated by the IOZM to the upper easterly branch of the Walker circulation discussed in section 3c. There is a clear QBO signal in anomalies of T and zonal wind (U) around Kilimanjaro above around 100 hPa, and weak but significant correlations are found between atmospheric anomalies at the peak and the 30-hPa zonal wind index for leads exceeding 10 months (not shown), suggesting that the QBO phase could influence the short rain season of the subsequent year. Thus, a detailed investigation of how the QBO influences atmospheric conditions at the summit remains an important future task.

Our analysis focused on the period during which high-altitude observational data were available to constrain

FIG. 11. Total-column moisture flux ($\text{kg m}^{-1} \text{s}^{-1}$; vectors) and precipitable water (kg m^{-2} ; shading) for (left) ON and (right) DJ; (a) climatology and for anomalies during the in-phase co-occurring mode event years of (b) 2006, (c) 2010, (d) 2015, and (e) 2016. Anomalies of precipitable water are presented as percent differences from the 12-yr mean. The black dot shows the location of Kilimanjaro.

the model performance at the glacierized altitudes. However, ultimately a longer time series is needed to fully elucidate the influence of ENSO and IOZM states at Kibo summit and to address the aforementioned avenues of future research. The impact of climate oscillations during the long rains (March–May), where the anthropogenic signal may be greater, should also be examined as a part of future work. However, our decadal analysis of the drivers of interannual variability at the summit lays the foundation for unraveling the contribution of climate modes to glacier changes and to the strong retreat of Kilimanjaro's glaciers observed over the twentieth century.

Acknowledgments. EC, TM, and TS were funded by the German Research Foundation (DFG) Grants MO 2869/1-1, MO 2869/3-1, and SA 2339/4-1, respectively. The authors gratefully acknowledge the computational resources and support provided by the High-Performance Computing Center (HPC) at the University of Erlangen-Nürnberg's Regional Computation Center (RRZE).

REFERENCES

- Banzon, V. F., and R. W. Reynolds, 2013: Use of WindSat to extend a microwave-based daily optimum interpolation sea surface temperature time series. *J. Climate*, **26**, 2557–2562, <https://doi.org/10.1175/JCLI-D-12-00628.1>.
- Behera, S. K., J.-J. Luo, S. Masson, P. Delecluse, S. Gualdi, A. Navarra, and T. Yamagata, 2005: Paramount impact of the Indian Ocean dipole on the East African short rains: A CGCM study. *J. Climate*, **18**, 4514–4530, <https://doi.org/10.1175/JCLI3541.1>; Corrigendum, **19**, 1361, <https://doi.org/10.1175/JCLI9018.1>.
- Benjamini, Y., and Y. Hochberg, 1995: Controlling the false discovery rate: A practical and powerful approach to multiple testing. *J. Roy. Stat. Soc.*, **57B**, 289–300.
- Bitz, C. M., and D. S. Battisti, 1999: Interannual to decadal variability in climate and the glacier mass balance in Washington, western Canada, and Alaska. *J. Climate*, **12**, 3181–3196, [https://doi.org/10.1175/1520-0442\(1999\)012<3181:ITDVIC>2.0.CO;2](https://doi.org/10.1175/1520-0442(1999)012<3181:ITDVIC>2.0.CO;2).
- Black, E., J. Slingo, and K. R. Sperber, 2003: An observational study of the relationship between excessively strong short rains in coastal East Africa and Indian Ocean SST. *Mon. Wea. Rev.*, **131**, 74–94, [https://doi.org/10.1175/1520-0493\(2003\)131<0074:AOSOTR>2.0.CO;2](https://doi.org/10.1175/1520-0493(2003)131<0074:AOSOTR>2.0.CO;2).
- Bradley, R. S., F. T. Keimig, and H. F. Diaz, 2004: Projected temperature changes along the American cordillera and the planned GCOS network. *Geophys. Res. Lett.*, **31**, L16210, <https://doi.org/10.1029/2004GL020229>.
- Camberlin, P., and N. Philippon, 2002: The East African March–May rainy season: Associated atmospheric dynamics and predictability over the 1968–97 period. *J. Climate*, **15**, 1002–1019, [https://doi.org/10.1175/1520-0442\(2002\)015<1002:TEAMMR>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<1002:TEAMMR>2.0.CO;2).
- Chan, R. Y., M. Vuille, D. R. Hardy, and R. S. Bradley, 2008: Intraseasonal precipitation variability on Kilimanjaro and the East African region and its relationship to the large-scale circulation. *Theor. Appl. Climatol.*, **93**, 149–165, <https://doi.org/10.1007/s00704-007-0338-9>.
- Chiang, J. C. H., and A. H. Sobel, 2002: Tropical tropospheric temperature variations caused by ENSO and their influence on the remote tropical climate. *J. Climate*, **15**, 2616–2631, [https://doi.org/10.1175/1520-0442\(2002\)015<2616:TTVCB>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<2616:TTVCB>2.0.CO;2).
- Chowdary, J. S., and C. Gnanaseelan, 2007: Basin-wide warming of the Indian Ocean during El Niño and Indian Ocean dipole years. *Int. J. Climatol.*, **27**, 1421–1438, <https://doi.org/10.1002/joc.1482>.
- Clark, C. O., P. J. Webster, and J. E. Cole, 2003: Interdecadal variability of the relationship between the Indian Ocean zonal mode and East African coastal rainfall anomalies. *J. Climate*, **16**, 548–554, [https://doi.org/10.1175/1520-0442\(2003\)016<0548:IVOTRB>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<0548:IVOTRB>2.0.CO;2).
- Collier, E., and W. W. Immerzeel, 2015: High-resolution modeling of atmospheric dynamics in the Nepalese Himalaya. *J. Geophys. Res. Atmos.*, **120**, 9882–9896, <https://doi.org/10.1002/2015JD023266>.
- , T. Mölg, F. Maussion, D. Scherer, C. Mayer, and A. B. G. Bush, 2013: High-resolution interactive modelling of the mountain glacier–atmosphere interface: An application over the Karakoram. *Cryosphere*, **7**, 779–795, <https://doi.org/10.5194/tc-7-779-2013>.
- Collins, W. D., and Coauthors, 2004: Description of the NCAR Community Atmosphere Model (CAM 3.0). NCAR Tech. Note NCAR/TN-464+STR, 214 pp., <https://doi.org/10.5065/D63N21CH>.
- Cullen, N. J., T. Mölg, G. Kaser, K. Steffen, and D. R. Hardy, 2007: Energy-balance model validation on the top of Kilimanjaro, Tanzania, using eddy covariance data. *Ann. Glaciol.*, **46**, 227–233, <https://doi.org/10.3189/172756407782871224>.
- , P. Sirguey, T. Mölg, G. Kaser, M. Winkler, and S. J. Fitzsimons, 2013: A century of ice retreat on Kilimanjaro: The mapping reloaded. *Cryosphere*, **7**, 419–431, <https://doi.org/10.5194/tc-7-419-2013>.
- Franco, B., M. Vuille, V. Favier, and B. Cáceres, 2004: New evidence for an ENSO impact on low-latitude glaciers: Antizana 15, Andes of Ecuador, 0°28'S. *J. Geophys. Res.*, **109**, D18106, <https://doi.org/10.1029/2003JD004484>.
- Ham, Y.-G., J.-Y. Choi, and J.-S. Kug, 2017: The weakening of the ENSO–Indian Ocean dipole (IOD) coupling strength in recent decades. *Climate Dyn.*, **49**, 249–261, <https://doi.org/10.1007/s00382-016-3339-5>.
- Hastenrath, S., 2001: Variations of East African climate during the past two centuries. *Climatic Change*, **50**, 209–217, <https://doi.org/10.1023/A:1010678111442>.
- , and D. Polzin, 2005: Mechanisms of climate anomalies in the equatorial Indian Ocean. *J. Geophys. Res.*, **110**, D08113, <https://doi.org/10.1029/2004JD004981>.
- , A. Nicklis, and L. Greischar, 1993: Atmospheric-hydrospheric mechanisms of climate anomalies in the western equatorial Indian Ocean. *J. Geophys. Res.*, **98**, 20 219–20 235, <https://doi.org/10.1029/93JC02330>.
- , D. Polzin, and P. Camberlin, 2004: Exploring the predictability of the “short rains” at the coast of East Africa. *Int. J. Climatol.*, **24**, 1333–1343, <https://doi.org/10.1002/joc.1070>.
- , —, and C. Mutai, 2007: Diagnosing the 2005 drought in equatorial East Africa. *J. Climate*, **20**, 4628–4637, <https://doi.org/10.1175/JCLI4238.1>.
- , —, and —, 2010: Diagnosing the droughts and floods in equatorial East Africa during boreal autumn 2005–08. *J. Climate*, **23**, 813–817, <https://doi.org/10.1175/2009JCLI3094.1>.
- , —, and —, 2011: Circulation mechanisms of Kenya rainfall anomalies. *J. Climate*, **24**, 404–412, <https://doi.org/10.1175/2010JCLI3599.1>.

- Hemp, A., 2005: Climate change-driven forest fires marginalize the impact of ice cap wasting on Kilimanjaro. *Global Change Biol.*, **11**, 1013–1023, <https://doi.org/10.1111/j.1365-2486.2005.00968.x>.
- , 2006: Vegetation of Kilimanjaro: Hidden endemics and missing bamboo. *Afr. J. Ecol.*, **44**, 305–328, <https://doi.org/10.1111/j.1365-2028.2006.00679.x>.
- Hodge, S. M., D. C. Trabant, R. M. Krimmel, T. A. Heinrichs, R. S. March, and E. G. Josberger, 1998: Climate variations and changes in mass of three glaciers in western North America. *J. Climate*, **11**, 2161–2179, [https://doi.org/10.1175/1520-0442\(1998\)011<2161:CVACIM>2.0.CO;2](https://doi.org/10.1175/1520-0442(1998)011<2161:CVACIM>2.0.CO;2).
- Hoell, A., and C. Funk, 2014: Indo-Pacific sea surface temperature influences on failed consecutive rainy seasons over eastern Africa. *Climate Dyn.*, **43**, 1645–1660, <https://doi.org/10.1007/s00382-013-1991-6>.
- Hong, C.-C., M.-M. Lu, and M. Kanamitsu, 2008: Temporal and spatial characteristics of positive and negative Indian Ocean dipole with and without ENSO. *J. Geophys. Res.*, **113**, D08107, <https://doi.org/10.1029/2007JD009151>.
- Huss, M., R. Hock, A. Bauder, and M. Funk, 2010: 100-year mass changes in the Swiss Alps linked to the Atlantic multidecadal oscillation. *Geophys. Res. Lett.*, **37**, L10501, <https://doi.org/10.1029/2010GL042616>.
- Iacono, M. J., J. S. Delamere, E. J. Mlawer, M. W. Shephard, S. A. Clough, and W. D. Collins, 2008: Radiative forcing by long-lived greenhouse gases: Calculations with the AER radiative transfer models. *J. Geophys. Res.*, **113**, D13103, <https://doi.org/10.1029/2008JD009944>.
- Immerzeel, W. W., L. P. H. van Beek, and M. F. P. Bierkens, 2010: Climate change will affect the Asian water towers. *Science*, **328**, 1382–1385, <https://doi.org/10.1126/science.1183188>.
- Indeje, M., and F. H. M. Semazzi, 2000: Relationships between QBO in the lower equatorial stratospheric zonal winds and East African seasonal rainfall. *Meteor. Atmos. Phys.*, **73**, 227–244, <https://doi.org/10.1007/s007030050075>.
- , —, and L. J. Ogallo, 2000: ENSO signals in East African rainfall seasons. *Int. J. Climatol.*, **20**, 19–46, [https://doi.org/10.1002/\(SICI\)1097-0088\(200001\)20:1<19::AID-JOC449>3.0.CO;2-0](https://doi.org/10.1002/(SICI)1097-0088(200001)20:1<19::AID-JOC449>3.0.CO;2-0).
- Jiménez, P. A., J. Dudhia, J. F. González-Rouco, J. Navarro, J. P. Montávez, and E. García-Bustamante, 2012: A revised scheme for the WRF surface layer formulation. *Mon. Wea. Rev.*, **140**, 898–918, <https://doi.org/10.1175/MWR-D-11-00056.1>.
- Kain, J. S., 2004: The Kain–Fritsch convective parameterization: An update. *J. Appl. Meteor.*, **43**, 170–181, [https://doi.org/10.1175/1520-0450\(2004\)043<0170:TKCPAU>2.0.CO;2](https://doi.org/10.1175/1520-0450(2004)043<0170:TKCPAU>2.0.CO;2).
- Kaser, G., M. Großhauser, and B. Marzeion, 2010: Contribution potential of glaciers to water availability in different climate regimes. *Proc. Natl. Acad. Sci. USA*, **107**, 20223–20227, <https://doi.org/10.1073/pnas.1008162107>.
- Latif, M., D. Dommengat, M. Dima, and A. Grötzner, 1999: The role of Indian Ocean sea surface temperature in forcing East African rainfall anomalies during December–January 1997/98. *J. Climate*, **12**, 3497–3504, [https://doi.org/10.1175/1520-0442\(1999\)012<3497:TROIOS>2.0.CO;2](https://doi.org/10.1175/1520-0442(1999)012<3497:TROIOS>2.0.CO;2).
- Luo, J.-J., R. Zhang, S. K. Behera, Y. Masumoto, F.-F. Jin, R. Lukas, and T. Yamagata, 2010: Interaction between El Niño and extreme Indian Ocean dipole. *J. Climate*, **23**, 726–742, <https://doi.org/10.1175/2009JCLI1304.1>.
- Madden, R. A., and P. R. Julian, 1994: Observations of the 40–50-day tropical oscillation—A review. *Mon. Wea. Rev.*, **122**, 814–837, [https://doi.org/10.1175/1520-0493\(1994\)122<0814:OOTDIO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1994)122<0814:OOTDIO>2.0.CO;2).
- Marzeion, B., and A. Nesje, 2012: Spatial patterns of North Atlantic Oscillation influence on mass balance variability of European glaciers. *Cryosphere*, **6**, 661–673, <https://doi.org/10.5194/tc-6-661-2012>.
- Mauritius Meteorological Services, 2008: Cyclone season of the southwest Indian Ocean 2006–2007. Mauritius Meteorological Services Tech. Rep. CS 28, 34 pp., <http://metservice.intnet.mu/pdfs/technical-note-on-cyclone-season.pdf>.
- Maussion, F., W. Gurgiser, M. Großhauser, G. Kaser, and B. Marzeion, 2015: ENSO influence on surface energy and mass balance at Shallap Glacier, Cordillera Blanca, Peru. *Cryosphere*, **9**, 1663–1683, <https://doi.org/10.5194/tc-9-1663-2015>.
- Meehl, G. A., J. M. Arblaster, J. T. Fasullo, A. Hu, and K. E. Trenberth, 2011: Model-based evidence of deep-ocean heat uptake during surface-temperature hiatus periods. *Nat. Climate Change*, **1**, 360–364, <https://doi.org/10.1038/nclimate1229>.
- Mölg, T., 2015: Exploring the concept of maximum entropy production for the local atmosphere-glacier system. *J. Adv. Model. Earth Syst.*, **7**, 412–422, <https://doi.org/10.1002/2014MS000404>.
- , and D. R. Hardy, 2004: Ablation and associated energy balance of a horizontal glacier surface on Kilimanjaro. *J. Geophys. Res.*, **109**, D16104, <https://doi.org/10.1029/2003JD004338>.
- , and G. Kaser, 2011: A new approach to resolving climate-cryosphere relations: Downscaling climate dynamics to glacier-scale mass and energy balance without statistical scale linking. *J. Geophys. Res.*, **116**, D16101, <https://doi.org/10.1029/2011JD015669>.
- , M. Renold, M. Vuille, N. J. Cullen, T. F. Stocker, and G. Kaser, 2006: Indian Ocean zonal mode activity in a multi-century integration of a coupled AOGCM consistent with climate proxy data. *Geophys. Res. Lett.*, **33**, L18710, <https://doi.org/10.1029/2006GL026384>.
- , N. J. Cullen, D. R. Hardy, G. Kaser, and L. Klok, 2008: Mass balance of a slope glacier on Kilimanjaro and its sensitivity to climate. *Int. J. Climatol.*, **28**, 881–892, <https://doi.org/10.1002/joc.1589>.
- , —, —, M. Winkler, and G. Kaser, 2009a: Quantifying climate change in the tropical midtroposphere over East Africa from glacier shrinkage on Kilimanjaro. *J. Climate*, **22**, 4162–4181, <https://doi.org/10.1175/2009JCLI2954.1>.
- , J. C. H. Chiang, A. Gohm, and N. J. Cullen, 2009b: Temporal precipitation variability versus altitude on a tropical high mountain: Observations and mesoscale atmospheric modelling. *Quart. J. Roy. Meteor. Soc.*, **135**, 1439–1455, <https://doi.org/10.1002/qj.461>.
- , M. Großhauser, A. Hemp, M. Hofer, and B. Marzeion, 2012: Limited forcing of glacier loss through land-cover change on Kilimanjaro. *Nat. Climate Change*, **2**, 254–258, <https://doi.org/10.1038/nclimate1390>.
- Molod, A., L. Takacs, M. Suarez, and J. Bacmeister, 2015: Development of the GEOS-5 atmospheric general circulation model: Evolution from MERRA to MERRA2. *Geosci. Model Dev.*, **8**, 1339–1356, <https://doi.org/10.5194/gmd-8-1339-2015>.
- Morrison, H., G. Thompson, and V. Tatarskii, 2009: Impact of cloud microphysics on the development of trailing stratiform precipitation in a simulated squall line: Comparison of one- and two-moment schemes. *Mon. Wea. Rev.*, **137**, 991–1007, <https://doi.org/10.1175/2008MWR2556.1>.
- Mutai, C., D. Polzin, and S. Hastenrath, 2012: Diagnosing Kenya rainfall in boreal autumn: Further exploration. *J. Climate*, **25**, 4323–4329, <https://doi.org/10.1175/JCLI-D-11-00414.1>.
- Nakanishi, M., and H. Niino, 2006: An improved Mellor–Yamada level-3 model: Its numerical stability and application to a

- regional prediction of advection fog. *Bound.-Layer Meteor.*, **119**, 397–407, <https://doi.org/10.1007/s10546-005-9030-8>.
- Nesje, A., Ø. Lie, and S. O. Dahl, 2000: Is the North Atlantic Oscillation reflected in Scandinavian glacier mass balance records? *J. Quat. Sci.*, **15**, 587–601, [https://doi.org/10.1002/1099-1417\(200009\)15:6<587::AID-JQS533>3.0.CO;2-2](https://doi.org/10.1002/1099-1417(200009)15:6<587::AID-JQS533>3.0.CO;2-2).
- Nicholson, S. E., 2015: Long-term variability of the East African “short rains” and its links to large-scale factors. *Int. J. Climatol.*, **35**, 3979–3990, <https://doi.org/10.1002/joc.4259>.
- , and D. Entekhabi, 1986: The quasi-periodic behavior of rainfall variability in Africa and its relationship to the Southern Oscillation. *Arch. Meteor. Geophys. Bioclimatol.*, **34A**, 311–348, <https://doi.org/10.1007/BF02257765>.
- Niu, G.-Y., and Coauthors, 2011: The community Noah land surface model with multiparameterization options (Noah-MP): 1. Model description and evaluation with local-scale measurements. *J. Geophys. Res.*, **116**, D12109, <https://doi.org/10.1029/2010JD015139>.
- Ogallo, L. J., 1982: Quasi-periodic patterns in the East African rainfall records. *Kenya J. Sci. Technol.*, **3A**, 43–54.
- Ohmura, A., 2012: Enhanced temperature variability in high-altitude climate change. *Theor. Appl. Climatol.*, **110**, 499–508, <https://doi.org/10.1007/s00704-012-0687-x>.
- Otte, I., F. Detsch, E. Mwangomo, A. Hemp, T. Appelhans, and T. Nauss, 2017: Multidecadal trends and interannual variability of rainfall as observed from five lowland stations at Mt. Kilimanjaro, Tanzania. *J. Hydrometeorol.*, **18**, 349–361, <https://doi.org/10.1175/JHM-D-16-0062.1>.
- Otte, T. L., C. G. Nolte, M. J. Otte, and J. H. Bowden, 2012: Does nudging squelch the extremes in regional climate modeling? *J. Climate*, **25**, 7046–7066, <https://doi.org/10.1175/JCLI-D-12-00048.1>.
- Pierce, D. W., T. P. Barnett, K. M. AchutaRao, P. J. Gleckler, J. M. Gregory, and W. M. Washington, 2006: Anthropogenic warming of the oceans: Observations and model results. *J. Climate*, **19**, 1873–1900, <https://doi.org/10.1175/JCLI3723.1>.
- Platnick, S., and Coauthors, 2015: MODIS atmosphere L3 daily product. NASA Goddard Space Flight Center, accessed 15 November 2017, https://doi.org/10.5067/MODIS/MYD08_D3.006.
- Pohl, B., and P. Camberlin, 2011: Intraseasonal and interannual zonal circulations over the equatorial Indian Ocean. *Theor. Appl. Climatol.*, **104**, 175–191, <https://doi.org/10.1007/s00704-010-0336-1>.
- , J. Crétat, and P. Camberlin, 2011: Testing WRF capability in simulating the atmospheric water cycle over equatorial East Africa. *Climate Dyn.*, **37**, 1357–1379, <https://doi.org/10.1007/s00382-011-1024-2>.
- Reynolds, R. W., T. M. Smith, C. Liu, D. B. Chelton, K. S. Casey, and M. G. Schlax, 2007: Daily high-resolution-blended analyses for sea surface temperature. *J. Climate*, **20**, 5473–5496, <https://doi.org/10.1175/2007JCLI1824.1>.
- Rodhe, H., and H. Virji, 1976: Trends and periodicities in East African rainfall data. *Mon. Wea. Rev.*, **104**, 307–315, [https://doi.org/10.1175/1520-0493\(1976\)104<0307:TAPIEA>2.0.CO;2](https://doi.org/10.1175/1520-0493(1976)104<0307:TAPIEA>2.0.CO;2).
- Saji, N. H., and T. Yamagata, 2003: Possible impacts of Indian Ocean dipole mode events on global climate. *Climate Res.*, **25**, 151–169, <https://doi.org/10.3354/cr025151>.
- , B. N. Goswami, P. N. Vinayachandran, and T. Yamagata, 1999: A dipole mode in the tropical Indian Ocean. *Nature*, **401**, 360–363, <https://doi.org/10.1038/43854>.
- Sauter, T., and S. P. Galos, 2016: Effects of local advection on the spatial sensible heat flux variation on a mountain glacier. *Cryosphere*, **10**, 2887–2905, <https://doi.org/10.5194/tc-10-2887-2016>.
- Skamarock, W. C., and J. B. Klemp, 2008: A time-split non-hydrostatic atmospheric model for weather research and forecasting applications. *J. Comput. Phys.*, **227**, 3465–3485, <https://doi.org/10.1016/j.jcp.2007.01.037>.
- Stuecker, M. F., A. Timmermann, F.-F. Jin, Y. Chikamoto, W. Zhang, A. T. Wittenberg, E. Widiasih, and S. Zhao, 2017: Revisiting ENSO/Indian Ocean dipole phase relationships. *Geophys. Res. Lett.*, **44**, 2481–2492, <https://doi.org/10.1002/2016GL072308>.
- Trenberth, K. E., J. Fasullo, and L. Smith, 2005: Trends and variability in column-integrated atmospheric water vapor. *Climate Dyn.*, **24**, 741–758, <https://doi.org/10.1007/s00382-005-0017-4>.
- Ummenhofer, C. C., A. Sen Gupta, M. H. England, and C. J. C. Reason, 2009: Contributions of Indian Ocean sea surface temperatures to enhanced East African rainfall. *J. Climate*, **22**, 993–1013, <https://doi.org/10.1175/2008JCLI2493.1>.
- Vuille, M., G. Kaser, and I. Juen, 2008: Glacier mass balance variability in the Cordillera Blanca, Peru and its relationship with climate and the large-scale circulation. *Global Planet. Change*, **62**, 14–28, <https://doi.org/10.1016/j.gloplacha.2007.11.003>.
- Webster, P. J., A. M. Moore, J. P. Loschnigg, and R. R. Leben, 1999: Coupled ocean–atmosphere dynamics in the Indian Ocean during 1997–98. *Nature*, **401**, 356–360, <https://doi.org/10.1038/43848>.
- Wentz, F. J., 1997: A well-calibrated ocean algorithm for special sensor microwave/imager. *J. Geophys. Res.*, **102**, 8703–8718, <https://doi.org/10.1029/96JC01751>.
- Wilks, D. S., 2016: “The stippling shows statistically significant grid points”: How research results are routinely overstated and overinterpreted, and what to do about it. *Bull. Amer. Meteor. Soc.*, **97**, 2263–2273, <https://doi.org/10.1175/BAMS-D-15-00267.1>.
- Winkler, M., G. Kaser, N. J. Cullen, T. Mölg, D. R. Hardy, and W. T. Pfeffer, 2010: Land-based marginal ice cliffs: Focus on Kilimanjaro. *Erdkunde*, **64**, 179–193, <https://doi.org/10.3112/erdkunde.2010.02.05>.
- Xie, S.-P., H. Annamalai, F. A. Schott, and J. P. McCreary Jr., 2002: Structure and mechanisms of south Indian Ocean climate variability. *J. Climate*, **15**, 864–878, [https://doi.org/10.1175/1520-0442\(2002\)015<0864:SAMOSI>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<0864:SAMOSI>2.0.CO;2).
- Zhang, X., 2007: Adapting the Weather Research and Forecasting Model for the simulation of regional climate in East Africa. Ph.D. thesis, North Carolina State University, 227 pp.