A Wall-Like Sharp Downward Branch of the Walker Circulation Above the Western Indian Ocean

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Abstract In the zonal direction, the downward branch of the Walker circulation above the Indian Ocean is only 20° wide, whereas the Pacific counterpart is 90° wide. This zonal sharpness is notable because atmospheric disturbances smaller than the planetary scale can interact with the planetary-scale Walker circulation through this branch. This zonal sharpness is also imprinted on a unique zonal discontinuity of the tropical rain belt above Northeast Africa. Therefore, we refer to this narrow downward branch as the “Wall,” investigate its climatology and interannual variability, and aim at determining its reason for existence. The strongest season of the lower tropospheric Wall in boreal summer is sustained by horizontal cold advection associated with the Asian Summer Monsoon, whereas another peak in boreal spring is explained by the seasonal migration of zonal winds. Two weak phases of the Wall correspond to two rainy seasons at the Eastern Horn of Africa, which are not reproduced well by the state-of-the-art global climate models. Experiments using a convection-permitting atmospheric model show that vertical mixing forced by mountain waves in East Africa is necessary for sustaining the Wall. After flattening the East African topography, zonal discontinuity of the tropical rain belt disappears. As for interannual variability, one standard deviation variability of vertical motions at the Wall is associated with one degree of sea surface temperature in the tropical Pacific, and the relationship is strongest in boreal autumn. Nevertheless, total variance is explained more by sea surface temperature in the tropical Indian Ocean.

Plain Language Summary The Walker circulation is the largest tropical east-west circulation on a global scale. In this Walker circulation, there is a region with particularly strong downward flow at the western edge of the Indian Ocean. This study shows that, contrary to common knowledge of large-scale tropical circulations, this strong downward flow is supported by cooling effects of horizontal flows and mountain waves generated by African mountains, in addition to radiative cooling. Numerical simulations based on a high-resolution atmospheric model with flattened African mountains confirm that East African mountains allow this strong downward flow to exist. This result implies that the East African mountains are responsible for the semi-arid climate and the two wet seasons in the Eastern Horn of Africa.

1. Introduction

The Walker circulation is the most prominent planetary-scale tropical atmospheric circulation in the zonal direction (e.g., Bjerknes, 1969; Walker, 1923, 1924). It has been understood that, to first order, the vertical motion associated with the Walker circulation consists of upward branches over relatively warm surfaces (e.g., the warm pool in the western Pacific) and downward branches over relatively cool surface (e.g., the cold tongue in the eastern Pacific; e.g., Lau & Yang, 2003). In the context of climate variability, trends and interannual variability of the Walker circulation have long been investigated in association with climate modes and the greenhouse gas forcing (e.g., Tanaka et al., 2004). In particular, the Pacific branches of the Walker circulation have received much attention, because its interannual fluctuation serves as the atmospheric component of the El Niño Southern Oscillation (ENSO), the most dominant interannual climate mode on Earth (e.g., Bjerknes, 1969).

As a mean state, however, a downward branch of the Walker circulation above the western Indian Ocean also exhibits a strong subsidence, which is at least comparable to the Pacific downward branch. Figure 1a shows the annual-mean equatorial vertical motion calculated by taking the meridional mean over the equatorial
The strong and sharp downward branch stands at the western edge of the Indian Ocean \((10^\circ E - 40^\circ E)\), whereas the weak and wide downward branch lies over the eastern Pacific \((90^\circ W - 150^\circ W)\). Considering the size of the two oceanic basins, one might find this interbasin contrast counterintuitive. In fact, the interbasin difference in apparent strength of the downward motion emanates from latitudinal dependence. An important caveat of this meridional-mean view shown in Figure 1a is that the strength of the downward motion depends upon latitudinal choices of the equatorial belt, as confirmed by Schwendike et al. (2014) who defined the local Walker circulation by shifting the equatorial belt southward \((35^\circ S - 10^\circ N)\) when taking this meridional mean. Figure 2a shows the annual-mean meridional-mean equatorial vertical motion over the equatorial belt of \(10^\circ S - 10^\circ N\), \(5^\circ S - 5^\circ N\), and \(2^\circ S - 2^\circ N\). While the downward branch above the Indian Ocean relatively keeps its strength notwithstanding, the latitudinal choices that of the eastern Pacific becomes stronger when narrower equatorial belts are chosen. A key to understand this difference is the zonal-mean cross-section (Figure 2b). The outstanding difference between the Indian and the Pacific downward branches is the degree of meridional asymmetry. The apparent weak downward motion above the Pacific, shown in Figure 1a, originates from an offset of a strong upward branch over the northern off-equatorial region \((4^\circ N - 10^\circ N)\) against a strong equatorial downward motion over \(10^\circ S - 2^\circ N\). On the other hand, the downward branch above the Indian Ocean shown in Figure 1a consists of a strong downward branch over \(4^\circ S - 10^\circ N\) with a hint of weak upward branch located to the south. Though it could be misleading in some context, we will carefully keep using the equatorial belt of \(10^\circ S - 10^\circ N\) in this study, because we are interested in the meridionally broad equatorial downward branch above the Indian Ocean.

Figure 1. (a) Observed annual-mean vertical motion averaged meridionally over the equatorial region \((10^\circ S - 10^\circ N)\) based on the ERA-Interim Reanalysis data. (b) As in (a), but the horizontal map at 500 hPa. (c) Observed annual-mean outgoing longwave radiation (OLR) based on the NOAA interpolated OLR data. (d) Observed annual-mean precipitation based on the GPCP data.
A more fundamental difference between the two downward branches lies in the longitudinal width. That is, the downward branch above the Indian Ocean is only 20° thick in the zonal direction, whereas the Pacific counterpart is about 90° wide. As we shall see in later sections, this zonal sharpness of the Indian Ocean branch is remarkable in the sense that phenomena smaller than the planetary scale, such as the Asian Summer Monsoon, the Turkana low-level jet (King et al., 2021), and possibly mountain waves, can easily interact with the planetary-scale Walker circulation via this branch. In this regard, Takasuka et al. (2021) recently demonstrated that this zonally sharp downward branch plays a fundamental role for the Madden-Julian Oscillation convective initiation triggered by amplification of upper-tropospheric mixed Rossby-gravity waves. This zonal sharpness is also notable in the lower tropospheric layer, where the downward branch subsides only onto the coastal western Indian Ocean and not onto the African continent (Figure 2c), which is consistent with Yang et al. (2015). In addition, from the viewpoint of moist circulations, the narrow downward branch is imprinted on a unique zonal discontinuity of the tropical rain belt. Figures 1b–1d show the annual-mean vertical motion at 500 hPa, outgoing longwave radiation (OLR), and precipitation, respectively, over the tropics. The tropical rain belt is typically characterized by the narrow convective band that circles the Earth along the equatorial region (e.g., Nicholson, 2018), but if we carefully look at the tropical rain belt, a discontinuity is found at the western edge of the Indian Ocean.

One of the major implications of this tropical rain belt discontinuity, and thereby, of the narrow downward branch, is the relatively dry climate at the so-called “Eastern Horn of Africa”, whose mean state, annual cycle, variability, and change have long been investigated in many previous studies (e.g., Camberlin, 1995; Liebmann et al., 2014, 2017; Lyon, 2014; Schreck & Semazzi, 2004; Tierney et al., 2015). In this regard, Zhao and Cook (2021) recently examined the influence of the narrow downward branch on East African rainfall. Another recent work by King et al. (2020) also demonstrated, based on the model ensemble that participated in the Coupled Model Intercomparison Project Phase 5 (CMIP5), that the poor representation of the narrow downward branch explains rainfall biases over Kenya during the main East African rainy seasons. In addition, the sensitivity of the climate in this region to topography in East Africa has long been examined from multiple perspectives (e.g., Naiman et al., 2017; Munday et al., 2021; Slingo et al., 2005). In line with these previous studies, by exploring the physics of the narrow downward branch in a comprehensive way, we expect a better understanding of the climatology of the Eastern Horn of Africa, whose annual cycle of precipitation is poorly reproduced by state-of-the-art, air-sea coupled global climate models (Tierney et al., 2015; Yang et al., 2014).

Figure 2. (a) As in Figure 1a, but over the equatorial belt of 10°S–10°N (top), 5°S–5°N (middle), and 2°S–2°N (bottom). (b) Observed annual-mean vertical motion averaged zonally over the Western Indian Ocean (45°E–60°E) and the Eastern Pacific Ocean (160°W–90°W). (c) Observed annual-mean vertical motion averaged vertically over the lower troposphere (800–1,000 hPa). All panels are produced based on the ERA-Interim Reanalysis data.
Therefore, in this study, we refer to this meridionally wide, zonally narrow sharp downward branch above the Indian Ocean as the “Wall” of the Walker Circulation, and will investigate its climatology and interannual variability in the hope that its reason for existence will be determined. Data and methods are described in the next section. In Section 3, we describe the seasonality of the Wall, and highlight a role of horizontal cold advection to support its strongest phase. We then perform model experiments to identify the East African topography as a necessary condition for the existence of the Wall, and discuss implications for the climate at the Eastern Horn of Africa. In Section 4, we define the Wall index to describe the interannual variability of the Wall and point out that both remote and local SST explain the interannual variance. Conclusions are presented in Section 5.

2. Data and Model

2.1. Data

Observed vertical motion, wind, and temperature data are from the European Center for Medium range Weather Forecasting (ECMWF) ERA-Interim reanalysis data (Dee et al., 2011). Observed OLR data is from the National Oceanic and Atmospheric Administration (NOAA) interpolated OLR (Liebmann & Smith, 1996). Observed precipitation data is from the Global Precipitation Climatology Project (GPCP; Adler et al., 2003). The time span of the atmospheric data used in this study is from 1979 through 2017. The horizontal resolutions used in this study are $3^\circ$ for vertical motion, wind, and temperature, and $2.5^\circ$ for OLR and precipitation. Observed SST data is from the National Oceanic and Atmospheric Administration (NOAA) Optimum Interpolation SST (OISST; Reynolds et al., 2007). Only when the relationship between atmospheric variables and SST is analyzed, the data from 1982 through 2017 is used, due to the availability of the satellite-based SST dataset. The resolution of the SST data is $1^\circ$ in both longitudes and latitudes.

2.2. Atmospheric General Circulation Model (AGCM) Experiments

We use the Nonhydrostatic Icosahedral Atmospheric Model (NICAM; Satoh et al., 2008, 2014; Tomita & Satoh, 2004), the version of which used for our experiments is the latest stable version, NICAM16-S (Kodama et al., 2020). The condensation processes are explicitly calculated using the single moment water 6 microphysics scheme (Tomita, 2008). Sub-grid scale turbulence is calculated by a modified version of the Mellor-Yamada scheme (Mellor & Yamada, 1982; Nakanishi & Niino, 2004; Noda et al., 2010). The radiation model with two stream radiative transfer scheme employs a correlated k-distribution method (msrnx; Sekiguchi & Nakajima, 2008). Surface fluxes are calculated with a modified version of the Louis scheme (Louis, 1979; Uno et al., 1995). For the land processes, the minimal advanced treatments of surface interaction and runoff (MATSIRO) land model (Takata et al., 2003) is used. Orographic gravity wave drag is considered to be sufficiently resolved in our simulations to obviate the need for parameterization of this process at subgrid scale.

The horizontal resolution is approximately 14 km on an icosahedral hexagonal-pentagonal mesh (Tomita et al., 2002). A terrain-following vertical grid coordinate is employed with the model top of approximately 40 km and 38 vertical layers, whose thickness increases with height. The model time step is 60 s. Our simulations are initialized on 00 UTC 28 June 2013 (ENSO neutral) and 2016 (La Niña), and are integrated for 93 days for each year. Initial conditions of the atmosphere and the ocean are derived from the National Centers for Environmental Prediction (NCEP) Final Operational Model Global Tropospheric Analysis (NCEP-FNL; NCEP, 2015). Time evolution of the sea surface temperature is prescribed externally from the interpolation of the NCEP-FNL data at 00 UTC on each day. To mitigate the effect of the model bias over land, the initial conditions of the land surface are taken from the monthly climatology derived from the last 5 years of a 10-year simulation of NICAM at 220 km horizontal resolution following Kodama et al. (2015, 2020).

Because it takes approximately 45 days for the values of vertical motions to converge to realistic climatological values, the first 63 days of the integrations are taken as a spin-up period, and the last 30 days of the integrations starting from September 1 are analyzed in this study. For comparison, we have also performed the same simulation but initialized on 00 UTC April 28, 2016 to capture the strongest month of the Wall, that is, July. Climatology of the Walker circulation, however, is not reproduced well, presumably because
the observed downward branches are not established until the end of May, during which the integration cannot be used as a spin-up period to capture the target circulation. Related difficulty in this season is also discussed by King et al. (2020) in association with dry biases (i.e., too strong subsidence) in AMIP experiments by the CMIP5 models. Though the AGCM used in this study realistically simulates the mean field over long time periods, the reproducibility of quick variations within relatively short time scales, such as a transition of seasons, is insufficient, to which future improvement is needed.

In addition to control runs, we have performed an experiment named "Flat East Africa (FEA)," in which the elevations are set to be 1 m over the entire East African region (30°S–30°N, 30°E–50°E; see Figure 6a) for 2013 (ENSO neutral) and 2016 (La Niña). We have also performed the same experimental sets but for 2015, an El Niño year, but the control run does not reproduce the observed features of the Walker circulation. This poor reproducibility of the Wall during an El Niño year appears to be because the observed Wall is weaker than those of other years (see Figure 9b). In our model, the water vapor supply from the anomalously warm eastern equatorial Pacific is biased to be excessive. The Walker Circulation over the tropics is distorted by this bias, to which the zonally-thin downward branch is sensitive. Therefore, in this study, only AGCM experiments in ENSO neutral and La Niña years will be discussed.

3. Climatology of the Wall

In this section, we first overview the seasonality of the Wall and the consistency with the local rainy seasons. Next, from the energetic viewpoint, we show that the strongest phase of the Wall is supported by horizontal cold advection associated with the Asian Summer Monsoon. Then, we perform model experiments to demonstrate the role of topography as a necessary condition of the Wall.

3.1. Bimodal Seasonality of the Wall

The Wall exhibits bimodal seasonal variability in its strength of the subsidence. The left panels of Figure 3 shows the monthly climatological-mean equatorial vertical motion averaged over the base period of 1979–2017. The Wall exhibits moderate subsidence from January through March, almost disappears from April
through May, reaches its strongest phase from June through September, and becomes weak from October through December.

The phase of this bimodal seasonality corresponds well to the annual precipitation cycle of the Eastern Horn of Africa, where two rainy seasons are known to exist. In this region, the term “Long Rains” denotes the longest and wettest rainy season that lasts from March through May, and the term “Short Rains” denotes the shorter and drier rainy season that peaks in October. As King et al. (2020) recently showed based on analyses of the CMIP5 model ensemble, the insufficient representation of this bimodality by state-of-the-art global climate models (Tierney et al., 2015; i.e., the models produce excessive “Short Rains” and insufficient “Long Rains”) is inseparable from the reproducibility of the seasonal variability of the Wall.

### 3.2. MSE Framework to Understand the Energy Balance for Sustaining the Wall

Hereafter, under the moist static energy (MSE) framework (e.g., Neelin and Held, 1987), the heat budget of the Wall will be discussed. The definition of MSE is as follows:

$$\text{MSE} = C_p T + L q + g Z$$

where $C_p$ is the specific heat capacity, $T$ is air temperature, $L$ is the latent heat of condensation, $q$ is specific humidity, $g$ is the gravitational acceleration, and $Z$ is geopotential height.

When this framework is applied to the budget at a constant height in the Wall region, the $L q$ term is of second order importance assuming that the Wall region is dry enough:

$$\frac{\partial (\text{MSE})}{\partial t} \bigg|_{Z=\text{const.}} = C_p \frac{\partial T}{\partial t} \bigg|_{Z=\text{const.}} + L \frac{\partial q}{\partial t} \bigg|_{Z=\text{const.}} \simeq C_p \frac{\partial T}{\partial t} \bigg|_{Z=\text{const.}}$$

Hence, the temperature tendency is of greatest interest, which is also employed by Veiga et al. (2011).

To understand climatological vertical air motion, the MSE tendency, and thereby temperature tendency, is assumed to be negligible so that the following steady-state balance is sustained:

$$\frac{\partial T}{\partial t} \bigg|_{Z=\text{const.}} = -\overline{\vec{v}} \cdot \nabla_h T - \overline{\vec{w}} \left( \frac{\partial T}{\partial z} + \Gamma_d \right) + \text{(eddy transport)} + \text{(diabatic heating)} \equiv 0$$

where the overlines denote temporal mean values, the $-\overline{\vec{v}} \cdot \nabla_h T$ term denotes horizontal temperature advection, $\overline{\vec{w}}$ denotes vertical motion, and $\Gamma_d$ denotes the dry adiabatic lapse rate. This balance yields the following formula:

$$\text{(subsidence)} \propto \text{(horizontal cold advection)} + \text{(eddy cooling)} + \text{(diabatic cooling)}$$

In this study, one of our goals is to determine which terms in Equation 4 are of first-order importance at each vertical level through the tropospheric Wall. In general, within the tropics, diabatic heating of large-scale downward motion is balanced with diabatic cooling (i.e., radiative cooling), and this energy budget is mostly true for the Walker circulation as well (Veiga et al., 2011). In the Wall region, however, we will show that horizontal cold advection and eddy cooling are of first-order importance to sustain the lower and upper tropospheric downward motion, respectively.

### 3.3. Role of Horizontal Cold Advection in the Strongest Phase of the Wall

One of the essential features of the strongest phase of the Wall is that the subsidence reaches the surface, which is not the case in weaker phases (Figure 3). To sustain the Wall, horizontal temperature advection plays a key role for lower tropospheric atmospheric subsidence to extend to the surface. The right panels of Figure 3 shows the mean horizontal temperature advection, which is defined as the inner product of climatological horizontal wind and the horizontal gradient of climatological temperature. Our definition of the mean horizontal advection does not take eddy heat transport into account.

The strongest subsidence observed in the lower troposphere from June through September is supported by the mean horizontal advection, and the horizontal cold advection is tightly connected to the Asian Summer
Monsoon. Figure 4 shows vertical-mean views of the horizontal temperature advection decomposed into zonal and meridional components. Here, we look at vertical-mean view, despite the baroclinicity of the tropical troposphere, because we are interested in the net energy budget realized by both upper and lower atmospheric horizontal advection. The horizontal advection cools the Wall region in boreal summer, when the Wall reaches its maximum phase (Figure 4, top).

This horizontal advection is realized by the meridional advection (Figure 4, bottom right), rather than the zonal component (Figure 4, bottom left). In Figure 5, these components are further decomposed into zonal wind, zonal temperature gradient, meridional wind, and meridional temperature gradient. Based on these panels, the maximum horizontal advection in boreal summer originates from the southerly winds associated with Asian Summer Monsoon, which blow toward the upgradient direction of the temperature field in this season. Based on Figure 4, the second strongest subsidence observed from January through March...
also appears to be supported by the horizontal advection. Nevertheless, because Figure 5 shows that the horizontal advection in this season originates from the seasonal migration of zonal wind signals, we cannot explain this second strongest peak by invoking the Asian Summer Monsoon.

Presumably, the aforementioned cooling effect drags down the lower tropospheric Walker Circulation to the surface, which is capable of strengthening the downward flow of the Wall further. This notion is consistent with the disappearance of the Wall from April through May, because this season is the period when the Asian summer monsoon is weakened to switch its direction before the onset of the strong Somali jet in early June (e.g., Findlater, 1969). From an energetic viewpoint, the relevance of the Somali jet is also consistent with Heaviside and Czaja (2013), who showed that the Somali jet dominantly accomplishes the atmospheric cross-equatorial heat transport.

Figure 5. As in Figure 4, but for zonal wind (top left), zonal temperature gradient (top right), meridional wind (bottom left), and meridional temperature gradient (bottom right). All panels are produced based on the ERA-Interim Reanalysis data.
3.4. Role of the East African Topography for Sustaining the Wall

Though we have concluded in the previous subsection that strong subsidence in the lower troposphere is associated with horizontal temperature advection, it remains unexplored what makes the subsidence in the Wall so strong that the Wall penetrates the entire troposphere in the vertical direction. In particular, cooling mechanisms of the upper troposphere have been largely unexplored. Therefore, in this subsection, we perform model experiments to highlight the role of topography for sustaining the Wall. Some implications for the climate of the Eastern Horn of Africa are also discussed.

3.4.1. Background

Our experiments are inspired by Naiman et al. (2017), who showed, in an interesting way, that topography can play major roles in determining the tropical circulation. Using the Geophysical Fluid Dynamics Laboratory (GFDL)-Earth System Model (ESM) 2M, they performed an experiment called “Pancake,” in which they removed all the topography on Earth and simulated the air-sea coupled system with flat lands. Because the Wall disappears in their “Pancake” run, we have hypothesized that, by flattening topography regionally rather than globally, it is possible to pinpoint the location of mountains that directly contribute to the realization of the Wall.

In this regard, at the end of last century, Rodwell and Hoskins (1995) already pointed out the importance of East African Highlands for the existence of the Somali jet by flattening the topography in this region as well as disabling the land-sea contrast in surface friction in their model. Liang et al. (2005) also hinted that the land-sea distribution is an important determinant of the tropical rain belt and the Wall. Furthermore, Ogwang et al. (2014) also investigated a precipitation response to regionally flattened African topography and demonstrated that the mean rainfall significantly reduces to the west of the Wall region. Considering their results, by eliminating the topography in East Africa, the strength and the hydrology at the center of the Wall may also be modulated.

3.4.2. Model Experiments With Flat East African Topography

In this study, in addition to a control run, the FEA experiment is arranged where the East African topography is flattened in the region shown in Figure 6a (see also Section 2). The top and middle panels of Figure 6b shows the monthly-mean equatorial vertical motion in September 2013 (ENSO neutral) and 2016 (La Niña) based on observations and the control, and the Flat East Africa (FEA) experiment in this order from the top panel. In the FEA experiment, the topography in the East African region, shown as the black box in panel (a), is flattened and set to be 1 m.
The FEA experiment reveals that, without the East African topography, the Wall disappears almost entirely through the troposphere (Figure 6b, bottom). By comparing the control and FEA runs, at least in both September 2013 and 2016, the East African topography is a necessary condition for existence of the Wall. Because the spring peak of the Wall is weaker than the summer-autumn peak, it is more challenging to reproduce the Wall in the control run. Therefore, to keep the robustness of the experimental result, we have restricted our conclusion to the summer-autumn peak. Determining whether the same mechanism works for the spring peak remains to be our future work.

A promising hypothesis is that the lack of turbulence generated by mountain waves suppresses vertical mixings. Because the lower troposphere generally has lower potential temperature than the upper troposphere, the reduction of vertical heat exchange weakens the subsidence of upper tropospheric air. Figure 7a shows the difference of equatorial vertical motions between the control and FEA runs, which should be interpreted as the downward motion forced by the East African topography. In this model, the downward branch is shifted westward compared to observations, so 30°E–45°E is drawn. Also shown in the left panel of Figure 7b is the energetic tendency contributions by the sum of eddy heat transport (EHT) and longwave radiation. Here, the EHT contribution is calculated as follows.

\[ \text{EHT} = \frac{1}{2} \left( \frac{\partial u' T'}{\partial x} - \frac{\partial u' T'}{\partial y} - \frac{\partial u' T'}{\partial z} \right) \]  

where \( x, y, \) and \( z \) denotes zonal, meridional, and vertical coordinates, respectively; \( u, v, \) and \( w \) denotes zonal wind, meridional wind, and vertical motion, respectively; \( T \) denotes temperature. The overlines denote the mean over September simulated in the model, and the primes denote deviations from the mean.

EHT and longwave radiation explain how East African topography works for realizing the vertical motion. In particular, the downward motion at higher levels than 10 km is predominantly explained by the eddy heat transport (Figure 7b, middle). The phase of heat and momentum transport is shifted, which suggests a hint of stationary gravity waves forced by mountains. Because the mountain waves suppress the upper tropospheric cloudiness, radiative cooling is enhanced in the middle tropospheric layer (Figure 7b, right), which in turn strengthens the downward motion further. Note that, because here we focus on the upper
and middle tropospheric energy budget, sensible heat flux from the surface cannot directly contribute to the downward motion. It is expected that the surface sensible heat flux can only modulate the lowest layer of the vertical level. Another possible form of diabatic heating is latent heating, but it is reasonably assumed to be negligible because the Wall region is sufficiently dry.

A caveat of this heat budget analysis is that the cooling effects of eddy transport and longwave radiation are quantitatively insufficient to explain the total downward motion. By assuming the dry adiabatic lapse rate, the vertical motion of 500 m/day requires approximately 5°C/day of cooling, but the aforementioned effects only explains 1°C/day of cooling. Though a more dominant effect may exist, it is still hard to close perfectly the heat budget based on an analysis of the 6-hourly snapshots available in our AGCM. Nevertheless, by using the convection permitting model without artificial gravity wave drags, our experiments at least confirm that the upper tropospheric downward motion is sustained by eddy heat transport forced by East African topography, rather than radiative cooling. Radiative cooling, enhanced by the clearer condition, is only capable of contributing the downward motion in lower altitudes where specific humidity is higher.

This vertical mixing effect serves as a good example where interscale interaction plays a fundamental role in downward branches, in addition to convective upward branches, to realize the large-scale atmospheric circulation in the current tropical climate. Specifically, the narrowly localized downward branch above the western Indian Ocean is realized as a result of interactions between large-scale motions and disturbances in smaller horizontal scales than the weak temperature gradient approximation (Sobel et al., 2001).

3.4.3. Implications for the Climate of the Eastern Horn of Africa

Without the East African topography, the relatively dry climate at the Eastern Horn of Africa would become wetter than in the real world. Figure 8 shows the monthly mean OLR and precipitation near the Wall in 2013 and 2016. The control run of the high-resolution convection-permitting model reproduces both OLR and precipitation well, particularly the discontinuity of the tropical rain belt. In the FEA run, as the East African topography is flattened, the discontinuity of the tropical rain belt disappears, which results in extension of the tropical rain belt over East Africa. Our result is consistent with a model experiment without topography performed by Chou and Neelin (2003), which does not exhibit the discontinuity of the tropical rain belt.

Both local processes and remote forcings can contribute to the “closing” of the tropical rain belt discontinuity in the FEA run. Locally, the removal of the Wall enhances convection above the Eastern Horn of Africa. This enhancement is due to the reduction of large-scale atmospheric subsidence as demonstrated by the simulated vertical motion. In addition to this local instability effect, clouds and moist air, which are
advected remotely by zonal winds, are also allowed to enter the Eastern Horn of Africa from the interior of the African continent, because topographic obstacles do not exist.

4. Interannual Variability of the Wall

In this section, we define the Wall index to point out that both remote and local sea surface temperature (SST) variability explains the interannual variations of the Wall.

4.1. Definition of the Wall Index

To highlight the interannual variability of the Wall, we define the Wall index as the average over vertical motions at 250, 550, and 850 hPa. This index should be physically interpreted as the mass-weighted average of vertical motions for the upper (100–400 hPa), middle (400–700 hPa), and lower (700–1,000 hPa) tropospheric layers. Figure 9a shows that, to the first order, the downward motion for the interannual scale is vertically constant, which justifies the definition of the Wall index for the purpose of representing the downward motion throughout the troposphere.

4.2. Relationship With ENSO, IOD, and Local SST Variability

The interannual variability of the Wall is explained by ENSO, which reminds us of the notion that the Wall is a part of the Walker circulation. The top panel of Figure 9b shows the 5-month running-mean time series of the Wall index. Also shown is the 5-month running-mean Niño 3.4 index, which is defined as SST anomalies averaged over the Niño 3.4 region (5°S–5°N, 170°W–120°W) standardized by its own standard deviation and the sign is reversed (red). These two indices exhibit a correlation of 0.54 (1982–2017), which is significant at the 95% confidence level. During an El Niño event, the Wall region exhibits a weakly downward motion, and vice versa for a La Niña event. In particular, the strong negative peaks of the Wall index in 1982, 1997, and 2015 are well-explained by strong...
El Niño events, whereas the strong positive peaks of the Wall index in 1998, 2010, and 2016 are well-explained by big La Niña events.

In addition to ENSO, the Indian Ocean Dipole (IOD; Saji et al., 1999) also explains the Wall index well. The bottom panel of Figure 9b is the same as top but for the 5-month running-meaned Dipole Mode index, which is defined as the SSTA difference calculated in the manner of the western equatorial Indian Ocean (50°E−70°E and 10°S−10°N) minus the south eastern equatorial Indian Ocean (90°E−110°E and 10°S−0°N). The correlation between the Wall index and the Dipole Mode index is 0.63 (1982–2017), which is significant at the 95% confidence level. During a negative IOD event, the Wall region exhibits a similar response to a La Niña event, though some of which could be explained by the covariance between ENSO and IOD.

The association with ENSO and IOD is also verified by SST spatial patterns. Figure 9c shows the regression map of SST anomalies on the monthly mean Wall index. This map clearly shows that the Wall variability is projected onto the equatorial Pacific SST variability. Because interannual variance of the tropical SST variability in the Pacific is dominated by ENSO, it is virtually certain that ENSO is one of the key factors to understand the Wall climate variability. That being said, the clear regression pattern in the Pacific does not necessarily mean that the Wall variance is predominantly explained by ENSO; it reflects on the large interannual variance of the eastern equatorial Pacific SST. Figure 9d shows the same map as Figure 9c but for correlation coefficients. Based on this correlation map, though ENSO still retains its importance, local SST variability, particularly an IOD-like zonal SST gradient, explains more variance of the Wall. Also notable is the high positive correlations over the maritime continent, presumably because the strength of the upward motions allowed in this region is also inseparable from the amount of the downward motions realized in the whole tropics, following the continuity equation.

The association with SST patterns is seasonally dependent. Figure 10 shows that correlations between the Wall index and the other climate indices reach their maxima during boreal autumn, which corresponds to the season of the “Short Rains,” while the connection to the “Long Rains” season is weaker. This seasonal dependence is also confirmed by the SST regression and correlation patterns shown in Figure 11. Our result
is consistent with previous studies (e.g., Camberlin & Philippon, 2002; Ogallo, 1988) that showed that, while the “Short Rains” are closely related to ENSO, the “Long Rains” do not exhibit as strong linkage to external climate forcings, possibly because the ENSO is more active in boreal autumn (e.g., Hastenrath et al., 2002).

That being said, our result also exhibits weak but significant correlations in MAM, which is consistent with previous studies (Dunning et al., 2016; King et al., 2020; Williams & Funk, 2011).

5. Summary and Discussions

We have reconsidered the climatology and the interannual variability of the Walker circulation by focusing on its sharp downward branch, which we refer to as the Wall, observed at the western edge of the Indian Ocean (Figures 1 and 2). A distinctive feature of the Wall is the two-peak seasonality (Figure 3). The two weak phases of the Wall, one in boreal spring and the other in boreal autumn, correspond well to the two rainy seasons at the Eastern Horn of Africa, which is not reproduced well by state-of-the-art GCMs. Another distinctive feature is that the subsidence of the Wall in its strongest phases reaches the surface (Figures 1 and 3). This “subsidence extension” in the strongest season (June–September) appears to be sustained by horizontal cold advection associated with the Asian Summer Monsoon, whereas the second strongest subsidence (January–March) appears to be caused more by the seasonal migration of zonal wind signals (Figures 3–5).

AGCM experiments show that the East African topography determines the strength of the Wall (Figure 6). In the FEA experiment, where the East African mountains are broadly flattened, the Wall almost vanishes throughout the entire tropospheric layer. This result leads to a conclusion that the East African topography is necessary for the existence of the Wall. We hypothesize that the role of topography is to generate mountain waves in response to large-scale circulation. The stationary vertical mixing enhances vertical heat exchange to cool the upper troposphere, which makes the Wall rigid (Figure 7). Assuming this mechanism, climate variability of the Wall could also be understood based on interscale interactions between global-scale circulation and local-scale mountain waves. In addition, we could also hypothesize that the modulation of the Somali jet (Chakraborty et al., 2009) and the Turkana low-level jet (Hartman, 2018; Nicholson, 2016; Vizy & Cook, 2019) by the African topography could control the lower tropospheric downward motion. Our simulation, however, does not reproduce the lower tropospheric downward motion realistically enough to make a case for the role of horizontal advection. In general, reproducing detailed features of the Asian Summer Monsoon in the state-of-the-art GCMs is still a challenging task (e.g., Johnson et al., 2016). Further process studies are needed to improve the robustness of these physical processes.
An implication of our simulation is that the dry and clear climate at the Eastern Horn of Africa is sustained by the East African topography (Figure 8). As a local effect, the large-scale subsidence associated with the Wall suppresses the local convection by drying the environment and by suppressing upward motion. At the same time, the high mountains in East Africa serve as obstacles that prevent clouds and moist air from being conveyed from the interior of the African continent. Because both of these local and remote processes are inseparable from the existence of the East African topography, it remains to be an open question which process serves as the dominant cause of the tropical rain belt discontinuity.

The interannual variability of the Wall is in no doubt associated with ENSO, but more variance is explained by SSTs in western equatorial Indian Ocean and over the maritime continent (Figure 9). This kind of association between the Wall and the tropical SST is strongest in boreal autumn (Figures 10 and 11), as suggested by Hastenrath et al. (2002) and many others. Nevertheless, our result also exhibits weak but significant correlations in boreal spring, which is consistent with some previous studies that focused on the Indian Ocean component of the Walker circulation in this season (King et al., 2020; Williams & Funk, 2011). Because these observational pieces of evidence in our present work are based on a single reanalysis data set, further verifications using multiple datasets are needed. In particular, a more up-to-date version of the ECMWF analysis, ERA5, could be of interest as a potential candidate for this purpose.

Data Availability Statement
This study is based on the ERA-Interim data set available online at https://apps.ecmwf.int/datasets/data/interim-full-moda/levtype=pl/, the NOAA interpolated OLR data set available online at https://psl.noaa.gov/data/gridded/data.interp_OLR.html, and the GPCP data set available online at https://psl.noaa.gov/data/gridded/data.gpcp.html. The OISST data are available online at https://www.esrl.noaa.gov/psd/data/gridded/data.noaa.oisst.v2.highres.html.

Acknowledgments
T. Kohyama is supported by the Japan Society for the Promotion of Science (JSPS)-Kakenhi Grant Numbers 19K23460 and 20K14554. T. Sueumas is supported by JSPS-Kakenhi Grant Numbers 16H07769, 20H05730, and 21K13991. H. Miura is supported by JSPS-Kakenhi Grant Number 21H04048 and 20H05729. D. Takasuka is supported by JSPS-Kakenhi Grant Numbers 16J07769, 20H05730, and 21H04048. The numerical computations using NICAM are provided by the University of Tokyo through the HPCI System Research Project hp210085.

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KOHYAMA ET AL. 15 of 17
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