Variability of the Turkana Low-Level Jet in Reanalysis and Models: Implications for Rainfall

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Key Points:

- The Turkana Low-Level Jet may be slowing down on decadal timescales. However, reanalysis simulations of the trend and annual cycles differ.
- The jet slowdown is a plausible consequence of changes to zonal surface pressure gradients, linking the jet to the Walker Circulation.
- Coupled Model Intercomparison Project 5 (CMIP5) representations of the jet are resolution-dependent.
Abstract

The complex topography of East Africa poses challenges for accurate modelling of regional climate. The Turkana Channel in northwestern Kenya is an important feature because of a persistent low-level jet (LLJ) which blows through it, which has complex interactions with local and regional rainfall. We establish the annual cycle and interannual variability of the LLJ in the ERA5, MERRA-2, and JRA-55 reanalyses. The jet is strongest during wet seasons in the surrounding region. Results suggest a statistically significant weakening of the LLJ over the last 30-40 years in two out of the three reanalyses. We propose an explanation based on the jet’s relationship with regional warming patterns and zonal surface pressure gradients, which link the jet to larger-scale climate dynamics including the Walker Circulation. If these changes continue in future, there may be significant implications for rainfall including increases in northwest Kenya and decreases further inland. However, the global models used to produce climate projections vary in their simulations of the LLJ in part because of their representations of topography. Consequently, it is not possible to assess how future LLJ changes will affect regional climate using CMIP5 models alone. Differences between the reanalyses preclude their direct use for model evaluation. Improving the processes by which topographical observations are mapped onto model grids could lead to improvements in simulation of East African climate. A field campaign to measure the LLJ directly could resolve uncertainties in the literature, help constrain reanalyses, and determine which models have the most realistic LLJ representation.

Plain Language Summary

The Turkana Low-Level Jet is an important physical feature of the East African climate. It is present throughout the year, and is both a factor in the aridity of the channel through which it blows, and a source of moisture for rainfall in the wider region. We investigate how the jet varies throughout the year and across longer timescales, and suggest that it may have been getting weaker over the last 30-40 years. We suggest that this is due to connections between the jet and large-scale climate dynamics, such as the Walker Circulation. However, because the jet is represented differently in different datasets, and climate models struggle to simulate it, we can't be sure of the impact these changes might have in future. One potential way forward is to measure the jet directly in a field campaign.

1 Introduction

The Turkana Channel in northern Kenya is one of the most prominent topographic features in East Africa (Figure 1). It separates the East African highlands to the south and the Ethiopian highlands to the north. The channel is oriented on a SE-NW axis, and is around 700km long. Its width varies from 700km at its widest point to around 150km at its narrowest. The channel floor is between 300-600m above sea level, with hills forming the sides of the channel rising to ≥1000m. There are a number of hills within the channel itself, notably Mount Marsabit (1531m), Mount Kulal (2293m) and the Huri Hills (1252m). The area experiences an arid hot desert climate; the regional capital, Lodwar, has a mean annual temperature of ~29°C and a mean annual rainfall of ~186mm.
Figure 1 – Topography of the Greater Horn of Africa region from the GTOPO30 dataset at 3 arc minute resolution. The purple, blue, orange, and red boxes indicate the jet core, Western Indian Ocean, Turkana Jet entrance, and Turkana jet exit regions respectively. The black line indicates the cross-section used for vertical profiles of the Turkana Jet core.

Aircraft pilots working in Kenya have long been aware of the existence of strong low-to-mid-tropospheric winds in this region (Kinuthia and Asnani 1982). A southeasterly low-level jet (LLJ) is present in the Turkana Channel throughout the year at an altitude of 600m, and was first described by Kinuthia and Asnani (1982) based on pilot balloon measurements. They reported mean wind speeds of around 25 m s$^{-1}$ and peak speeds of over 50 ms$^{-1}$ at Marsabit. More extensive pilot balloon analysis at multiple locations in the channel was conducted by Kinuthia (1992). He found that the jet was characterised by strongly divergent flow in the morning, with higher jet speeds in the narrower parts of the channel, suggesting Bernoulli forcing of the flow. The shape of the channel results in a southeasterly jet regardless of whether the prevailing low-level winds to the southeast of the channel, dominated in this region by the seasonal reversal of the Somali Jet, are northerly or southerly (Kinuthia 1992). He reported jet maxima in the morning and at midday, although jet occurrence was possible at any time of day, and found the jet to split vertically into two distinct cores, one at 300m-2000m and another at 2000m-3100m.
(except at Marsabit owing to the influence of mountainous terrain there). Kinuthia’s (1992) observations also indicated very high peak jet speeds of >50 m s\(^{-1}\). It should be noted, however, that single theodolite pilot balloon observations, which assume a constant balloon ascent rate, may be biased towards high wind speeds as a result of turbulence.

Nicholson (2016) presented the first climatology of the jet using reanalysis data. She found jet maxima of ~18 m s\(^{-1}\), much lower than Kinuthia and Asnani (1982) and Kinuthia (1992). Additionally, the reanalysis climatology suggested the jet was mainly nocturnal, with maxima at midnight or 6am in agreement with analyses of other LLJs (Allen and Washington 2013, Rife et al. 2010), but not the observations of Kinuthia (1992) who reported 9am or 12am maxima. Nicholson (2016) found the jet to be strongest in September-October-November and December-January-February, and weakest from May to August.

Indeje et al. (2001) used a regional climate model (RCM) to investigate the drivers of the Turkana LLJ. Their findings supported the Kinuthia (1992) hypothesis that the main forcing mechanism for the jet is the Bernoulli effect, with the channel dimensions controlling its velocity profile at different locations. Additionally, the magnitude of the jet core wind speed was found to be affected by the large-scale background flow associated with the Somali Jet, but equal importance for jet formation and maintenance was given to local-scale thermal and frictional effects. Hartman (2018) argued - based on reanalysis data - that thermal forcing is of primary importance in jet formation, maintenance, and variability. He suggested that the jet strength is controlled by the temperature gradient along the Turkana Channel, finding it to be enhanced during the April to October period by the advection of cooler air from the Indian Ocean into southeastern Kenya by the Somali Jet, and reduced between November and March when the Somali Jet is reversed; in this period, the Somali Jet is less energetic (Boos and Emanuel 2009) and is advecting air from a different (continental) air mass. This implies that the jet maximum and minimum are coincident with those of the Somali Jet, in contrast to the reanalysis climatology obtained by Nicholson (2016); neither of the two published observational studies ran for long enough to produce an annual cycle. On shorter timescales, Vizy and Cook (2019) propose a connection between high pressure ridging off the Kenyan coast and strong jet events.

The influence of the jet on rainfall within the Turkana Channel itself is evident from the arid nature of the local climate. Kinuthia (1992) noted that the desert regions of northern Kenya have the same size and shape as the LLJ. This aridity results from the highly divergent flow aloft in the channel (Kinuthia 1992), which leads to dominant downward motion (positive omega) and the consequent inhibition of convective rainfall (Indeje et al. 2001, Hartman 2018). Nicholson (2016) also ascribed the aridity of the area surrounding the Turkana Channel to the jet’s divergent flow, as well as suggesting a link between the jet’s diurnal cycle and rainfall (with rainfall typically occurring when the jet is weakest), although it is difficult to distinguish these effects from the diurnal cycle of convection. Vizy and Cook (2019) extend this analysis, finding strong negative correlations between reanalysis jet strength and rainfall in the channel and surrounding regions. The jet has a complex relationship to rainfall in the exit region of South Sudan, with a strong jet likely to trigger low-level convergence and convective rainfall in the early morning in the west of the country but not the east. This is due to the increased convergence in the strong jet case being offset by reduced atmospheric moisture content.

The LLJ may also have an impact on rainfall further afield. Despite the aridity of the Turkana Channel, the jet itself is a key vector for moisture flux into East Africa. Viste and Sorteberg (2013) showed that 32% of the moisture falling as rain over the Ethiopian Highlands
passes through the Turkana Channel. Indeed, LLJs in Eastern Africa and southern Africa are key conduits of water vapour from the Indian Ocean to the African interior. Kinuthia (1992) suggested that the Turkana LLJ could have an impact on the African Easterly Waves (AEWs), which are responsible for much of the convective rainfall in West Africa, by imparting vorticity to the African Easterly Jet (AEJ) at the jet exit. This hypothesis is supported by Lin (2009), who demonstrated that AEWs and their associated mesoscale convective systems (MCSs) can be generated by vortex shedding associated with horizontal wind shear at the Turkana Jet exit. Furthermore, Lin (2009) suggested that this vorticity contribution from the Turkana Jet and the Ethiopian Highlands could be a precursor to tropical cyclones and hurricanes in the Atlantic Ocean. This is comparable to the effect reported by Holbach and Bourassa (2015) in the Pacific, where topographically forced winds from central America contribute vorticity to cyclone formation. The jet also contributes to the local economy through the generation of electricity by the Lake Turkana Wind Power Project, one of Africa’s largest wind farms.

Rainfall change and variability have substantial impacts on East African societies and economies. It is therefore important to be able to make reliable projections of how anthropogenic climate change will affect rainfall (Washington et al. 2006, Giannini et al. 2018). The ability of climate models to represent East African climate is a caveat on studies using models to understand the drivers of variability and change, and on future projections (Rowell et al. 2015, James et al. 2018, Hirpa et al. 2019). General circulation models (GCMs) participating in Coupled Model Intercomparison Project 5 (CMIP5; Taylor et al. 2012) tend to project a wetter future for East Africa (Ongoma et al. 2017, Han et al. 2019) – however, this is in opposition to the observed drying trend, a problem known as the East African climate paradox (Rowell et al. 2015). Models are also known to have substantial rainfall biases in East Africa, underestimating the magnitude of the long rains and overestimating the short rains (Tierney et al. 2015). A number of possible reasons for models’ limited reliability in the region are under investigation, including biases towards a positive Indian Ocean Dipole (IOD) state (Cai and Cowan 2013), erroneous representation of zonal winds (Hirons and Turner 2018, Walker et al. 2019) and the Walker Circulation (King et al. 2020), and the impact of aerosols (Scannell et al. 2019). Differences in the background state of the circulation can determine the trajectory of model projections of East African rainfall (Cook et al. 2020). One significant source of uncertainty arises from East Africa’s highly complex topography, which has a major impact on rainfall (Hession and Moore 2011, Ogwang et al. 2014), but which CMIP5 models are unable to capture in detail owing to their resolution (James et al. 2018). Therefore, a key task for improving our understanding of East African rainfall change is to investigate the effects of topographic features (such as the Turkana LLJ) on the observed climate over the last few decades.

Given its links to rainfall, and importance for wind power, investigation of any multidecadal trends in the Turkana LLJ strength is important, with implications for future climate change impacts in East Africa. Additionally, the nature of the jet as a regional-scale topographically-forced feature with wider-scale climate impacts means it is a challenge to represent in climate models (Collins et al. 2018, James et al. 2018). The Turkana LLJ specifically has been suggested as a reason for the limited ability of seasonal forecasting models to represent interannual variability in East African rainfall (MacLeod 2019). This paper therefore presents the first analysis (to our knowledge) of the long-term behaviour of the jet, as well as its representation in CMIP5 models. We aim to address the following research questions:
1. What is the climatology of the Turkana LLJ in MERRA-2, ERA5, and JRA55, and how does it interact with regional rainfall?

2. How has the Turkana LLJ changed over recent decades (1981-2016) and what might account for these changes?

3. How is the Turkana LLJ and its links to rainfall represented in CMIP5 models, and how does model topography affect this representation?

The remainder of the paper is organised as follows. Section 2 introduces the datasets used and methods of analysis. Section 3 describes the climatology of the Turkana LLJ and its links to regional rainfall. Section 4 evaluates changes in the Turkana LLJ strength over the period 1981-2016, and suggests potential mechanisms for observed trends; Section 5 assesses the ability of CMIP5 models to represent the jet, its annual cycle, and its correlation with regional rainfall. Section 6 discusses and summarises the findings, and Section 7 concludes the paper.

2 Data and Methods

Rainfall data used in this study were obtained from version 2 of the Climate Hazards Group Infrared Precipitation with Stations dataset (CHIRPSv2.0; Funk et al. 2015). This is a high-resolution (0.05°) gridded dataset which combines observed rainfall from raingauges with remotely sensed infrared measurements of cold cloud duration. It has performed well in assessments of precipitation datasets over East Africa relative to station data (Kimani et al. 2017, Gebrechorkos et al. 2018) especially for the monthly means used in this study (Dinku et al. 2018).

Given the lack of available climatological observations of low-to-mid tropospheric wind speed in the Turkana Channel, reanalysis data are used to assess the decadal behaviour of the Turkana LLJ. Reanalysis products differ in their representations of the tropical African troposphere (Washington et al. 2013, Maidment et al. 2015); consequently, three reanalysis data sets were used: version 2 of the Modern Era Retrospective Analysis for Research and Applications (MERRA-2; Gelaro et al. 2017); the European Centre for Medium-Range Weather Forecasts Reanalysis 5 (ERA5; Hersbach et al. 2020); and the Japanese 55-year Reanalysis Project (JRA-55; Kobayashi et al. 2015). The JRA55 data used here are on the dataset’s native hybrid pressure levels rather than isobaric levels; conversions to hPa are as per Japan Meteorological Agency (2014), whereby hybrid level 12 equates to 846.96hPa. MERRA-5 and ERA5 are on isobaric levels. Details of the reanalyses used are presented in Table 1.
Table 1 - Details of the reanalysis products used in this study.

| Product | Institution        | Spatial Resolution | Temporal Resolution | Time Period     | Assimilation Scheme |
|---------|--------------------|--------------------|---------------------|-----------------|--------------------|
| MERRA-2 | NASA (USA)         | 0.625° x 0.5°      | 3 hrs               | 1980-present    | 3D-VAR             |
| ERA5    | ECMWF              | 0.2815° x 0.2815°  | 1 hr                | 1950-present    | 4D-VAR ensemble    |
| JRA-55  | JMA (Japan)        | 0.5625° x 0.5625°  | 3 hrs               | 1958-present    | 4D-VAR             |

MERRA-2 has been suggested to be the most realistic reanalysis of those currently available at representing tropospheric wind fields over equatorial Africa when compared to quality-controlled radiosonde observations (Hua et al. 2019), whereas ERA5 performed best in a global-scale assessment of near-surface winds (Ramon et al. 2019). However, the latter study did not assess reanalysis performance against measurements in tropical Africa, and found little difference in seasonal mean wind performance between datasets (Ramon et al. 2019). Torralba et al. (2017) found good agreement between MERRA-2, JRA55, and ERA-Interim (the predecessor of ERA5) for both 10m and 850 hPa wind speed trends, although both reduced wind speeds over land and stronger wind speed trends were found in JRA55 relative to the other datasets. 10m surface wind and 2m air temperature observations were obtained from the UK Met Office Integrated Data Archive System (MIDAS; Met Office (2012)) for three stations located along the channel axis: Wajir (1.75°N, 40.067°E), Marsabit (2.3°N, 37.9°E), and Lodwar (3.117°N, 35.617°E). Topography data are obtained from the US Geological Survey GTOPO30 dataset at 3 arc minute resolution (Gesch et al. 1999).

The climate model data used are from CMIP5 (Taylor et al. 2012). A subset of 25 models was selected based on wind data availability from historical coupled runs, and availability of wind data under forcing experiments using the RCP8.5 scenario for use in future work (Table 2.). RCP8.5 is the closest CMIP5 scenario to observed increases in atmospheric CO2 concentrations (Sanford et al. 2014). The first ensemble member for each model was used over the period 1975-2005.

3 Results

3.1 Turkana LLJ Climatology

3.1.1 Long-Term Means

We first plot the annual cycle of the Turkana LLJ using the MERRA-2, ERA5, and JRA55 datasets. Figure 2 indicates that in the MERRA-2 reanalysis, the jet is present in the long-term mean windspeed field. The jet core is between 10 and 14 m s⁻¹, with the core lying between 900 and 800 hPa. The ERA5 LLJ is also present at the same height range but with less extension. In JRA55 the jet has a weaker core speed on average than MERRA-2 or ERA5 (not shown). The JRA55 LLJ is also more vertically extensive than in MERRA-2 or ERA5, extending from hybrid level 5 (978hPa) to hybrid level 13 (818hPa) (not shown).
Figure 2 – Long-term mean wind speed at the cross-section through the jet core shown in Figure 1, for MERRA-2 (a) and ERA5 (b). Data averaged from 1981-2016.

Plotting wind vectors at 850hPa/hybrid level 12 (Figure 3) indicates the presence of a strong easterly flow in the Turkana Channel in the monthly averages throughout the year. It is noteworthy that the jet persists as a southeasterly flow when the prevailing wind at this height, the Somali Jet, undergoes a seasonal reversal from northerly cross-equatorial flow in boreal winter to southerly cross-equatorial flow in boreal summer associated with the Asian Monsoon (Boos and Emanuel 2009).

The long-term means of 850hPa windspeed (Figure 3), however, indicate differences between the datasets. The annual JRA55 jet is weaker than the annual means of MERRA-2 and ERA5, with an annual mean of ~8 m s\(^{-1}\) in the former dataset compared to ~12 m s\(^{-1}\) in the latter datasets. Additionally, the directionality of the wind is slightly different for each reanalysis. ERA5 shows winds entering the Turkana channel from the east/north-east whereas MERRA-2 and JRA55 show winds entering from the south/south-east.

Figure 3 – as Figure 2, but for wind speed at 850hPa/hybrid level 12
3.1.2 Annual Cycle

The jet is primarily a zonal wind feature, with a core centred at around 850 hPa in the reanalysis datasets. The annual cycle of zonal wind at the jet core area (850hPa, 3°N-5°N, 35°E-39°E, pink box in Figure 1) (Figure 4) for the reanalysis datasets is broadly in agreement with that obtained by Nicholson (2016) from the ERA-Interim reanalysis. The reanalysis LLJs exhibit a bimodal climatology with peak wind speeds in boreal spring and boreal autumn/winter. Minimum wind speeds are experienced in July (boreal summer). While the seasonality of the jet is the same in all three reanalyses, there is variation in the jet’s magnitude. MERRA-2 consistently has the strongest jet throughout the year, with average peak zonal wind speeds 2 m s\(^{-1}\) faster than ERA5 between December and March. ERA5 and MERRA-2 do have similar jet strength in May-September when the jet is weaker; therefore, the annual cycle of jet strength is more pronounced in MERRA-2 than ERA5. JRA55 has similar peak speeds to ERA5 from November to March, but much lower speeds between May and September as the jet almost disappears (~1 m s\(^{-1}\)) in July. JRA55 has the most pronounced annual cycle of all three reanalyses.

![Figure 4](image)

**Figure 4** – seasonal cycle of jet core zonal wind at 850hPa from reanalysis datasets plus the ensemble mean of the 25 CMIP5 models listed in Table 2. The blue shaded area indicates ±1 standard deviation of the CMIP5 ensemble mean.

The ensemble mean CMIP5 model annual cycle is in good agreement with the reanalyses in terms of seasonality, though with a faster zonal wind in February-March (7.8 m s\(^{-1}\)) compared to November-January (6.8 m s\(^{-1}\)) - the reanalyses do not have such a seasonal difference between zonal wind speeds. Similar results were obtained using scalar wind speed integrated in the jet core region between 800hPa and 900hPa (not shown).

Vizy and Cook (2019) also note the stronger LLJ in MERRA-2 when compared to ERA5. They ascribe this to Bernoulli forcing, noting that the underlying topography in MERRA-2 features a narrower Turkana Channel, as well as either lower or absent topography in the channel.
which is present in ERA5. Our analysis concurs with these findings. However, JRA55, which also has a narrower Turkana Channel than ERA5 and even less representation of within-channel topography (Figure 5), has a jet which is comparable in strength to ERA5 during peak months and much weaker during boreal summer. Therefore, differences between reanalysis datasets cannot be attributed to topography alone.

Figure 5. Comparison of underlying topography of the Turkana Channel from GTOPO30, ERA5, MERRA-2, and JRA55.

We next consider the annual cycles of dynamic fields which may play a role in forcing the jet. Following Hartman (2018), we plot annual cycles of the differences between the jet exit and the jet entrance, and between the jet exit and the western equatorial Indian Ocean (WEIO) for three key fields: near-surface temperature, 900hPa geopotential height, and 850hPa vertical velocity (omega). The areas used for averaging are shown as boxes in Figure 1. This indicates how the strength of the gradients along the jet axis vary seasonally and between the different reanalyses.
Figure 6. Differences in near-surface temperature for each month in reanalysis between the jet entrance box and the jet exit box (A) and between the Western Indian Ocean box and the jet exit box (B; for 900hPa geopotential height (C and D); and for 850hPa vertical velocity (omega; E and F).

For the jet entrance, exit, and WEIO areas, the annual cycle of temperature gradients in reanalysis are in broad agreement. In all areas, temperature peaks in boreal spring and experiences a minimum in boreal summer. For the entrance zone, MERRA-2 is the hottest reanalysis apart from in May, July, and January when ERA5 is hotter by ~0.1-0.3°C; the temperatures of the 2 datasets are in general very similar. JRA55 is consistently around 1°C cooler at the entrance. At the jet exit, the annual cycle of temperature has greater amplitude, with a more pronounced peak in March for all datasets. MERRA-2 is again narrowly the hottest and JRA-55 cooler by up to 1°C, but there is closer agreement between the datasets than the exit. Peak jet speeds coincide with higher temperatures in the channel, but boreal spring is markedly warmer than boreal autumn despite similar jet strength. The temperature gradient between the entrance and exit and between the WEIO and exit peaks in boreal spring when the jet is strong, and is lowest in boreal summer when the jet is weak (Figure 6 A and B). This does imply a link between the temperature gradient and the jet strength – however it does not entirely account for the annual cycle because the temperature gradients in boreal autumn are weaker than spring.
despite the jet having a similar magnitude. Furthermore, in all months apart from May and June, the reanalysis with the strongest along-channel temperature gradient is JRA55, which has the weakest jet of the three datasets for most of the year. The entrance minus exit gradient for JRA55 is significantly different ($p > 0.05$) from MERRA-2 and ERA5.

For 900hPa geopotential height ($Z_g$; Figure 6, C and D), the gradients between the entrance/WEIO and exit also vary seasonally. For the entrance minus exit differences (Figure 6 C), MERRA-2 and ERA5 have a similar annual cycle with the steepest gradient in April and the shallowest in January; MERRA-2 consistently has a smaller $Z_g$ difference than ERA5, and this difference is statistically significant at $p > 0.05$. JRA55 has a stronger gradient peaking in July, which is effectively in antiphase with the annual cycle of jet zonal wind strength. For the $Z_g$ difference between the WEIO and the jet exit (Figure 6 D), the datasets have boreal spring and summer peaks that correspond more closely to the jet annual cycle – implying a role for ocean-land pressure gradients in the jet’s climatology, although the stronger gradients in JRA55 do not translate straightforwardly into a faster jet. As the jet lies almost on the equator, with the Coriolis parameter tending to zero, the jet cannot be described in terms of geostrophic wind; however, neither is it straightforwardly dependent on the pressure gradient force. The $Z_g$ over the WEIO is influenced by the strength of the Somali Jet, suggesting its influence on the Turkana LLJ may be more complex than the temperature advection mechanism proposed by Hartman (2018).

The omega gradients (Figure 6, E and F) indicates stronger gradients in vertical motion when the jet is strong. The weaker gradients during boreal summer could be attributable to the influence of the Somali Jet, which results in suppression of the pressure gradient due to prevailing downward motion across Kenya. It is notable that in JRA55, the vertical motion averaged across the entrance region is downward in boreal summer whereas it is upward in the other 2 reanalyses. In spring and summer, JRA55 has weaker ascent in this region. This results in an omega gradient from the entrance to the exit of almost zero in boreal summer, during which the jet strength in JRA55 drops almost to zero. The differences between all three reanalyses for the WEIO minus exit difference in omega are statistically significant ($p > 0.05$). In all three variables, the reanalyses are more similar for the exit region than the entrance regions. This may reflect the greater influence of parameterisations relating to moist processes over the WEIO and/or the differences in the relative strength of the Somali Jet between the reanalyses.
Figure 7 – seasonal cycles of near-surface temperature from reanalysis data at Wajir (a), Marsabit (b), and Lodwar (c) stations, compared to station observations.

Figure 7 compares surface temperature for the stations at Wajir, Marsabit, and Lodwar with the nearest reanalysis grid box. All three stations lie within the entrance or core regions of the channel. Marsabit station is located on a hill and is the highest above sea level of the stations (1345m), whereas Lodwar (515m) and Wajir (244m) are located on the channel floor. Figure 7 indicates that the reanalysis annual cycle of temperature has a slight cool bias at Wajir and Lodwar. The strong warm bias for Marsabit can be attributed to area-averaged elevation in reanalyses which results in an overall more flattened topography. There are statistically significant differences ($p < 0.05$) between the mean temperatures at all stations and the reanalysis temperatures interpolated to the station locations, as well as between the individual reanalyses at Lodwar (all reanalyses different from one another), Wajir (significant differences between JRA55 and the other reanalyses) and Marsabit (significant differences between MERRA-2 and the other reanalyses). Quality control issues with the station data in this area mean these observations should be interpreted with some caution; nevertheless, the reanalyses are consistent in being cooler than observations at Wajir and Lodwar, and warmer than observations at Marsabit.
Vizy and Cook (2019) observe that on synoptic timescales, a strong Turkana LLJ is associated with high pressure ridging off the coast of East Africa which drives an anomalous southeasterly flow. For the purposes of our investigation into the jet’s climatology, we plotted the correlation between jet core zonal wind speed and mean sea level pressure (MSLP) for each of the three reanalyses during the months of the two main East African rainy seasons (Figure 8). The results demonstrate that the mechanism suggested by Vizy and Cook (2019) has an analogue
on climatological timescales, with jet zonal wind speed exhibiting significant correlations with MSLP along the East African Coast in all rainy season months for both MERRA-2 and ERA5. This correlation is however not evident in JRA-55 other than for November, and significant correlations of the opposite sign for the Kenya coast are obtained for April and October. Given the physical plausibility of Vizy and Cook’s (2019) mechanism, this may cast doubt upon the ability of JRA-55 to represent the jet’s variability.

3.1.2 Wind-Rainfall Covariability

This section investigates the link between the Turkana LLJ and regional rainfall. We plot the Pearson product moment correlation coefficient between the 850hPa zonal wind speed in the jet core for each reanalysis dataset and CHIRPSv2.0 rainfall for each grid point in each bimodal rainy season month over the period 1981-2016 (Figure 9). There are strong and significant correlations between the jet and rainfall in the surrounding area, with similar relationships obtained for each of the three reanalyses. The jet strength is negatively correlated with precipitation across Tanzania, Kenya, and southern Ethiopia, as well as in the Turkana Channel itself, reflecting the divergent flow of the jet (Kinuthia 1992, Vizy and Cook 2019). Meanwhile, jet strength is positively correlated with rainfall in the African interior, especially across the northern part of the Congo Basin. This may be linked to the jet’s suggested influence on MCS propagation via vortex shedding effects on the African Easterly Jet (Lin 2009), or on downstream convergence. In January-September, when the jet is at its weakest, higher jet strength is linked with lower rainfall in the Turkana channel and across South Sudan, central and northern Ethiopia, and eastern Sudan and Chad (not shown).
Figure 9 – Pearson product moment correlation coefficients between reanalysis jet core zonal wind timeseries and CHIRPSv2.0 rainfall data from 1981-2016 for MERRA-2 (a), ERA5 (b), and JRA55 (c). Shading indicates significance at the 95% confidence level.

4. Trends

4.1 LLJ Strength

To our knowledge, no investigation has been published into the strength of the Turkana LLJ on multidecadal timescales. Here, we present such an analysis using reanalysis data (Table 2). Trends were computed with linear regressions of both zonal wind and wind speed based on monthly average values at the jet core for the period 1981-2016. A Mann-Kendall test with Theil-Sen slope estimation was also performed to provide an independent metric for estimating trends; the two tests were in agreement in the majority of cases for both variables, with the exceptions of April (MERRA-2 zonal wind and ERA5 windspeed trends were significant at p<0.05 for Mann-Kendall but not linear regression), May (MERRA-2 wind speed trend was significant for linear regression but Mann-Kendall was not), and September (MERRA-2 wind speed trend was significant for linear regression but Mann-Kendall was not). The results indicate a strong possibility that the Turkana LLJ has been weakening since 1981. For the MERRA-2 reanalysis, there are negative trends in jet core zonal wind which are statistically significant (p<0.05) for all months except for February and June (although there is a significant zonal wind trend in June using the Mann-Kendall test at p<0.1), and wind speed trends for all months excluding February, April, and June. For JRA55, there are significant (p<0.05) negative trends for January, March, May, July, and December in jet core zonal wind, and January, March, May, July, September, and December for wind speed. A significant negative trend at p<0.1 is obtained for February using the Mann-Kendall test. By contrast, the only trend obtained in ERA5 was for April, where there is a significant increase in jet core zonal wind speed at the p<0.05 confidence level using the Mann-Kendall test. This is opposite to the trend in MERRA-2 for the same month. Although the jet is primarily a zonal wind feature, we also found significant decreases in the strength of the meridional wind component at the jet core in 7 months of the year (MERRA-2) and 8 months of the year (JRA55), with ERA5 displaying contrasting increases in meridional wind in February and April (not shown).
Table 2. Turkana jet core wind trends and significance values for MERRA-2, ERA5, and JRA55. Trends are computed by taking the monthly average zonal wind (U) and wind speed (WS) at the jet core at 850 hPa defined in Figure 1. Trends are denoted by the slope of the linear regression equation. Diacritical marks denote p < 0.05 as follows: * = significant trend in linear regression. † = significant trend with Mann-Kendall test. Note that u wind is negative, so a positive slope means a decrease in u wind. Units are m s\(^{-1}\) decade\(^{-1}\).

|       | Jan MERRA-2 | Jan ERA5 | Jan JRA55 | Feb MERRA-2 | Feb ERA5 | Feb JRA55 | Mar MERRA-2 | Mar ERA5 | Mar JRA55 | Apr MERRA-2 | Apr ERA5 | Apr JRA55 | May MERRA-2 | May ERA5 | May JRA55 | Jun MERRA-2 | Jun ERA5 | Jun JRA55 |
|-------|-------------|----------|-----------|-------------|----------|-----------|-------------|----------|-----------|-------------|----------|-----------|-------------|----------|-----------|-------------|----------|-----------|
| U     | 0.81†       | 0.05     | 0.58†     | 0.15        | -0.29    | 0.33      | 0.48†       | -0.1    | 0.53†     | 0.38†       | -0.21    | 0.34      | 0.41†       | -0.01    | 0.42†     | 0.34       | 0.07     | 0.11      |
| W     | -0.87†      | -0.03    | -0.59†    | -0.21       | 0.38     | -0.35     | -0.59†      | 0.17    | -0.58†    | -0.37       | 0.30†   | -0.39     | -0.31†      | 0.10     | -0.49†    | -0.28      | -0.03    | -0.31      |
| U     | 0.54†       | 0.12     | 0.26      | 0.46†       | 0.17     | 0.08      | 0.44†       | 0.26    | 0.16      | 0.61†       | 0.17    | 0.13      | 0.68†       | 0.04     | 0.24      | 0.87†      | 0.09     | 0.55†     |
| W     | -0.51†      | -0.09    | -0.58†    | -0.41†      | -0.14   | -0.26     | -0.44†      | -0.20   | -0.45†    | -0.65†      | -0.15   | -0.26     | -0.78†      | -0.04    | -0.30     | -0.94†      | -0.07    | 0.55†     |

4.2 Mechanisms

To better understand the mechanisms behind the weakening jet strength in reanalysis, as well as the differences between the datasets, we analysed changes in 900 hPa Zg. A composite difference approach was taken; for each month in reanalysis, the average conditions for 1981-1987 were subtracted from the most recent 7-year period available for each field (2010-2016).
For MERRA-2, wind trends are observed for 9 months of the year at $p<0.05$ and 11 months at $p<0.1$. In the 900 hPa $Zg$ composite differences (Figure 10), there is a general increase
in geopotential height inland in MAM and OND, with some decreases along the GHA coast and over the Indian Ocean in boreal autumn/winter. Zg increases to a greater extent in the jet exit than the jet entrance, apart from in February, when there is no significant trend in LLJ zonal wind speed (not shown). The changing gradient in geopotential height along the channel axis from the jet entrance to the jet exit, with higher Zg in the jet exit region compared to the jet entrance, is consistent with the weakening of the LLJ shown by MERRA-2 in OND. This is because an increase in the geopotential heights at the jet exit relative to the entrance is indicative of a reduction in the pressure gradient force acting along the jet axis. This effect is less noticeable in MAM for MERRA-2, potentially due to reduced influence from larger-scale tropical circulations in this season; however, increases in Zg offshore are reduced compared with inland.

In JRA55, there are significant decreases in Turkana LLJ zonal wind in 4 months of the year at p<0.05 and 6 months at p<0.1. JRA55 shows a strong increase in geopotential height at the jet exit in MAM and a reduced increase in OND. The increasing inland geopotential height is evident in several months in which JRA55 shows a decrease in jet strength (March, May, and December).

The ERA5 dataset does not have any significant changes in LLJ zonal wind at p<0.05, and only a single month (September) has a decreasing trend at the p<0.1 level. The Zg in ERA5 shows weaker increases than MERRA-2 in MAM, and the sign of change is reversed in November and December with decreases in geopotential height inland of the Turkana Channel. In ERA5, inland decreases in Zg result in little change in the along-channel pressure gradient. By contrast, MERRA-2 and JRA55 have Zg changes that could explain the jet slowing trends as a function of a changing geopotential height gradient along the channel.
Figure 1 shows the linear trend in mean sea level pressure (MSLP) for March-April and October-December for each of the three reanalysis datasets. While there is variation in the locations of the signals between the reanalyses, some general patterns emerge. For March, April, and May, there are statistically significant increases in MSLP over land in East Africa, on the order of 0.05 hPa month$^{-1}$ year$^{-1}$. These are confined to the south of the Equator in ERA5, and are most intense in JRA-55, whereas the trend in MERRA-2 extends in a band extending further north into East Africa. Increasing MSLP trends inland are also seen in MERRA-2 and JRA55 in
October, November, and December, though not in ERA5 which has trends of the opposite sign. Additionally, large-scale statistically significant decreases in MSLP of up to -0.1 hPa month$^{-1}$ year$^{-1}$ are seen in the Indian Ocean during OND in MERRA-2 and ERA5, and in ON in JRA55. While we note the limitations of using pressure reduced to sea level for areas of high relief such as much of East Africa, we nevertheless consider the broad-scale nature of these changes, taken together with geopotential height, to be suggestive of potential mechanisms for the slowing trend in the reanalysis jet.

5. CMIP5 Models

The previous sections demonstrated that Turkana LLJ is a significant feature for the regional climate, and is associated with rainfall variability in the Turkana Channel and downstream in the Congo Basin. Here we assess the abilities of CMIP5 models to represent the LLJ. We note that, because of the close relationship between the jet and regional topography, modelling studies focussing on the jet specifically should use regional-scale climate models (RCMs) (e.g. Indeje et al. 2001). However, CMIP5 models are used in many estimates of future regional climate change, including in East Africa. Assessing how well they reproduce a key control on regional rainfall is therefore important.

Table 3. CMIP5 models used in this study. Expansions of acronyms are available online at http://www.ametsoc.org/PubsAcronymList. Models highlighted in bold are the focus of this section.

| Model Name | Modelling Group | Native Resolution of Atmosphere (lat x lon, °) |
|------------|-----------------|-----------------------------------------------|
| ACCESS1-3  | Commonwealth Scientific and Industrial Research Organisation and Bureau of Meteorology (Australia) | 1.25 x 1.875 |
| BCC-CSM1-1-M | Beijing Climate Centre, China Meteorological Administration | 2.7906 x 2.8125 |
| BNU-ESM   | College of Global Change and Earth System Science, Beijing Normal University | 2.7906 x 2.8125 |
| CanESM2   | Canadian Centre for Modeling and Analysis | 2.7906 x 2.8125 |
| CCSM4     | National Center for Atmospheric Research (USA) | 0.9424 x 1.25 |
| Model         | Institution                                                                 | Ratio          |
|--------------|------------------------------------------------------------------------------|----------------|
| CESM1-CAM5   | National Science Foundation, Department of Energy, National Center for Atmospheric Research (USA) | 0.9424 x 1.25  |
| CMCC-CM      | Centro Euro-Mediterraneo per i Cambiamenti Climatici (Italy)                  | 0.7484 x 0.75  |
| CNRM-CM5     | Centre National de Recherches Météorologiques (France)                        | 1.4008 x 1.40625 |
| CSIRO-Mk3-6-0| Commonwealth Scientific and Industrial Research Organisation in collaboration with the Queensland Climate Change Centre of Excellence (Australia) | 1.8653 x 1.875 |
| EC-EARTH     | EC-EARTH consortium (Europe)                                                  | 1.1215 x 1.125 |
| FGOALS-g2    | LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences; and CESS, Tsinghua University | 2.7906 x 2.8125 |
| FIO-ESM      | The First Institute of Oceanography, SOA (China)                              | 2.7906 x 2.8125 |
| GFDL-CM3     | Geophysical Fluid Dynamics Laboratory (USA)                                   | 2 x 2.5        |
| GISS-E2-R    | Goddard Institute for Space Studies (USA)                                     | 2 x 2.5        |
| HadGEM2-AO   | Met Office Hadley Centre (UK)                                                 | 1.25 x 1.875   |
| INMCM4       | Institute for Numerical Mathematics (Russia)                                  | 1.5 x 2         |
### 5.1 Topography

As discussed in section 3.1, good representation of regional topography is important for accurate modelling of the jet (Vizy and Cook 2019). This poses problems for GCMs in regions of complex topography, especially in Africa where inaccurate representation of topographic features can lead to large-scale model errors and biases (Munday and Washington 2018). In this

| Model Name           | Institute and Location                                                                 | Domain Size |
|----------------------|----------------------------------------------------------------------------------------|-------------|
| IPSL-CM5A-LR         | L’Institut Pierre-Simon Laplace (France)                                               | 1.8947 x 3.75 |
| IPSL-CM5A-MR         | L’Institut Pierre-Simon Laplace (France)                                               | 1.2676 x 2.5  |
| IPSL-CM5B-LR         | L’Institut Pierre-Simon Laplace (France)                                               | 1.8947 x 3.75 |
| MIROC5               | National Institute for Environmental Studies (Japan), Atmosphere and Ocean Research Institute (University of Tokyo), Japan Agency for Marine-Earth Science and Technology | 1.4008 x 1.40625 |
| MIROC-ESM-CHEM       | National Institute for Environmental Studies (Japan), Atmosphere and Ocean Research Institute (University of Tokyo), Japan Agency for Marine-Earth Science and Technology | 2.7906 x 2.8125 |
| MPI-ESM-LR           | Max Planck Institute for Meteorology (Germany)                                         | 1.8653 x 1.875 |
| MPI-ESM-MR           | Max Planck Institute for Meteorology (Germany)                                         | 1.8653 x 1.875 |
| MRI-CGCM3            | Meteorological Research Institute (Japan)                                              | 1.12148 x 1.125 |
| NorESM1-M            | Norwegian Climate Centre                                                               | 1.8947 x 2.5  |
This section, we therefore investigate the representation of the jet in two high-resolution models (CMCC-CM and CCSM4) and two low-resolution models (BNU-ESM and FGOALS-g2). Figure 12 shows the substantial differences in topography between these models. CMCC-CM and CCSM4 have a channel of width ~250km at the core which is comparable to the observed topography in Figure 1. BNU-ESM has a much more simplified channel as a result of its lower resolution, with reduced channel depth relative to the sides. Representation of the channel is not, however, purely a function of resolution; FGOALS-g2 has the same spatial resolution as BNU-ESM but features flatter topography across the domain. The channel floor in this model is approximately 600m higher than in GTOPO30. Given the close relationship between the jet and regional topography (Indeje et al., 2001), models which do not accurately represent the topography are likely to struggle in capturing the salient features of the jet. We investigate this below.

Figure 12 – model topography over the East African domain for CMCC-CM (A), CCSM4 (B), BNU-ESM (C), and FGOALS-g2 (D).

5.2 Annual Cycle

We plot the annual cycle of CMIP5 850 hPa zonal wind (Figure 13). The spread of individual model zonal winds is greater than the annual variation in the jet itself. The model annual cycle is more pronounced than the reanalyses – five models (CCSM4, EC-EARTH, BNU-ESM, MRI-CGCM3, and FIO-ESM) have stronger zonal winds than any reanalysis in March and seven in April (the above plus NorESM1-M and ACCESS1-3), but 21 models have weaker zonal winds than MERRA-2 and ERA5 in July. Only one model, MIROC-ESM-CHEM, exceeds MERRA-2 jet speeds in November. The amplified Turkana LLJ annual cycle may result from the effects of simplified topography on the regional circulation – models with simplified
representation of the East African highlands are likely to feature Somali Jets extending further inland than in observations, as these highlands are important for constraining the Somali Jet (Slingo et al. 2005). This jet peaks in boreal summer, when its flow at the latitude of the Turkana Channel is primarily meridional. Therefore, if the Somali Jet is not as well constrained by the East African Highlands in the models, it may extend inland as far as the Turkana Jet core region and produce an exaggerated reduction of the zonal wind component during boreal summer.

Topography also has an important impact on constraining the Walker Circulation in which the zonal winds across the IO are embedded (Naiman et al. 2017) – given that these winds have been found to be erroneous in some CMIP5 models (Hirons and Turner 2018), models with simplified topography may also have biases in the prevailing winds into the Turkana Channel. Some outlying models in Figure 13 exhibit unrealistic annual cycles. For example, MRI-CGCM3 has stronger winds in boreal summer than in boreal spring or autumn/winter. In INMCM4, the sign of the zonal wind reverses in boreal summer, indicating that the wind at the jet core becomes westerly – this is not seen in reanalysis where the zonal wind is always easterly.

![Figure 13](image)

*Figure 13 – annual cycle of zonal wind at jet core for CMIP5 models listed in Table 2.*

The reanalyses are characterised by similar peak zonal wind speeds in MAM and OND, with slightly stronger winds in OND;). By contrast, 13/25 models have stronger winds in MAM.

### 5.3 Sensitivity to Topography

To investigate the effects of topography on CMIP5 model simulations of the LLJ, we compare the behaviour of the models in Figure 12.
Figure 14 indicates substantial differences between the high and low-resolution models. In CMCC-CM and CCSM4, the jet is well-resolved within the Turkana Channel, with a clear acceleration and narrowing of the mean flow. The highest resolution model, CMCC-CM, also has the fastest mean jet speed (~11 m s$^{-1}$). By contrast, the jet is not evident in the low-resolution models. While BNU-ESM and FGOALS-G2 do feature stronger southeasterly winds in the vicinity of the jet entrance, the wind vectors show little evidence that a low-level jet is being formed as a result of the channeling of the mean flow by the topography. Instead, a larger-scale area of increased zonal wind is found over Kenya in the long-term average. This may reflect inland incursion of the offshore Somali Jet as a result of the simplified representation of topography in these models, as has been suggested by sensitivity experiments (e.g. Slingo et al. 2005). The wind speeds are also lower in these models compared to CMCC-CM and CCSM4. This implies a role for resolution in determining the ability of CMIP5 models to simulate the Turkana LLJ.
Figure 15 – as Figure 14 but for a cross-section through the Turkana Channel as per Figure 1.

Figure 15 demonstrates the role of topography representation in CMIP5 models’ simulations of the Turkana LLJ. A cross-section through the wind speed field at the location of the reanalysis jet shows clear differences between models in terms of topography. The Turkana Channel is represented well in CMCC-CM and to a reasonable degree in CCSM4, where the simplified topography nevertheless features a channel of comparable magnitude to that in reality. The vertical profile of the jet in CMCC-CM compares well to that in reanalyses in terms of height and location. The jet core is less clearly defined in CCSM4. In BNU-ESM and FGOALS-g2, the topography of the Turkana Channel is absent, and as a result the low-level jet is not simulated adequately. This both demonstrates the role of topography in forcing the jet, and the importance of considering topography as a factor in process-based climate model evaluation over East Africa.

5.4 Wind-Rainfall Covariability

We next show the association between CMIP5 model 850 hPa zonal wind at the jet core, and model rainfall across the East African domain (Figure 16). The results show substantial variations in models’ correlations between the zonal winds in the channel and regional rainfall, which vary depending on the resolution. The higher-resolution models exhibit negative correlations between jet strength and rainfall in the area surrounding the jet, in agreement with reanalysis and theory, but also have much broader positive correlations extending across the south of the domain. By contrast, the lack of a clear within-channel negative correlation in the lower-resolution models emphasizes their lack of a clearly defined jet. South of the Turkana Channel is a sharp dividing line between 0°-10°S, which separates regions of negative and positive correlations. Compared to reanalysis, the models tend to extend the region of negative
correlation much further west. The large spatial extent of the correlation fields in the lower-resolution models, and the sharp meridional dividing line between positive and negative correlations, suggests that a larger-scale feature than the Turkana Jet (potentially the Somali Jet) is controlling the wind at the location of the Turkana Channel. It is evident that lower-resolution CMIP5 models do not have adequate representation of the Turkana LLJ, which is important in light of the need to identify reasons for poor CMIP5 performance over East Africa (Rowell et al. 2015). By contrast, higher-resolution models within in the ensemble are able to simulate the jet’s structure and aspects of its correlations with rainfall.

Figure 16 – As Figure 9 but for CMCC-CM (A), CCSM4 (B), BNU-ESM (C), and FGOALS-g2 (D) over 1975-2005.

6. Discussion

6.1 Climatology of the Turkana LLJ

We examined the climatology of the Turkana LLJ across three reanalysis datasets with varying resolution and underlying topography. Consistent with earlier studies (e.g. Nicholson
2016) we find that the jet is a persistent feature throughout the year. The annual cycle of jet core zonal wind was similar in all the datasets, with jet maxima in boreal spring and autumn/winter and minima in summer. MERRA-2 consistently had faster jet speeds than ERA5 or JRA55; we concur with Vizy and Cook (2019) that the narrower topography of the channel plays a role due to Bernoulli forcing, but there is also a likely contribution from differences in surface heating and land/atmosphere coupling. The jet has significant correlations with rainfall in the region throughout the year, which are reasonably consistent between the reanalyses. A stronger jet - and attendant low-level divergence - is associated with lower rainfall in the Turkana Channel and in the jet entrance region. Positive correlations with rainfall further inland are likely to reflect some combination of the jet’s impact on triggering convection (Vizy and Cook 2019), moisture transport from the Indian Ocean (Viste and Sorteberg 2013), and contribution of vorticity to the AEJ which favours the development of MCSs (Lin 2009).

6.2 CMIP5 Performance and Implications for Projections

Given the impacts of the Turkana LLJ on the regional climate, we briefly assessed the ability of CMIP5 models to represent it. We found that the ensemble mean performed well compared to reanalysis at representing the annual cycle of jet core zonal wind, but that this was obscuring a wide range of models, some of which had incorrect annual cycles or 100% biases in wind direction. We were able to attribute some of this variation to resolution and topography, with higher-resolution models having a much better representation of the Turkana Channel than those with lower resolutions. Model topography impacts on both the representation of the Turkana LLJ, and the extent to which the Somali LLJ moves inland (Slingo et al. 2005). We therefore suggest that measurements of 850hPa zonal wind at the real-world location of the Turkana LLJ in some models may in fact be reflective of the simulated Somali LLJ. This latter suggestion is consistent with the interannual correlation field between jet strength and rainfall, which is continental in scale in some models. We contend that improving model topography over Africa should be a major priority in light of this and other studies (e.g. Naiman et al. 2017, Munday and Washington 2018). However, other model features, such as parameterisations of land-atmosphere interactions, should also be taken into account when identifying model improvement foci.

The deficiencies in the simulation of the Turkana LLJ, as well as the relationship between jet strength and rainfall, erode our confidence in model-derived projections of mean change and changes to rainfall variability in East Africa - and especially Kenya. This is a particular problem when using the ensemble mean model to estimate future change, which includes contributions from models with fundamental errors in the present-day representation of the jet, and its relationship with regional climate. More positively, improving the simulation of the jet - for example by increased model resolution or representation of topography, could help to reduce the uncertainty in future projections of rainfall in East Africa.

6.3 Long-Term Trends in the Turkana LLJ

We undertook the first analysis of the jet’s behaviour on decadal timescales. A statistically significant decrease in jet core zonal wind speed was found in MERRA-2 for most of the year and for JRA55 in around half the year. The reality of the jet slowdown is tentative because the most recently developed reanalysis used, ERA5, shows no trend for most of the year and increases in April. However, it is noteworthy that MERRA-2 performed best in an
assessment of reanalysis data over tropical Africa based on quality-controlled radiosonde observations of winds in the lower troposphere (Hua et al. 2019).

We investigated dynamical fields in the reanalysis datasets to better understand what might be causing the putative decrease in the jet strength. We note that reductions in surface wind speeds have been suggested to occur worldwide (e.g. McVicar et al. 2012) as a result of the poleward expansion of the Hadley Circulation under anthropogenic climate change (Lu et al. 2007) and the consequent decrease in surface pressure gradients between the tropics and the midlatitudes. Taking a composite approach, we found that in MERRA-2 the region around the jet exit (in the continental interior of Africa) had a change in geopotential height gradient from the entrance to the exit whereby the height increases at the exit and decreases at the entrance. This is consistent with Gebrechorkos et al. (2019) who showed that the interior of East Africa has experienced greater warming than the coastal jet entrance region over the period 1979-2012. This change in the along-axis temperature gradient is reflected in This is a hypothesis for the jet slowdown which is thermodynamically consistent with expected spatial patterns of global warming, whereby inland areas experience faster warming than coastal areas. Variations on this pattern of regional geopotential height change are present in ERA5 and JRA55 but do not always translate into a slowing of the jet; while this slowdown is present in JRA55, it is not in ERA5 except for the month of September. It is suggested that this may reflect a weaker land-atmosphere coupling strength in the region for ERA5 and JRA55 compared to MERRA-2. We note that MERRA-2 replaces modelled precipitation with observed values before it reaches the surface in order to simulate land surface variables (Draper et al. 2018) - which ERA5 does not (Albergel et al. 2018) – which results in slight overestimations of latent heat flux (Draper et al. 2018). Differences in physical parameterisations, for example of convective mass flux over the WEIO, could also contribute to the differences.

Long-term windspeed plots (Figure 4) indicate marked differences between the reanalyses in terms of the jet’s connection to prevailing winds in the region. The jet tends to enter the Turkana Channel from the north-east in ERA5, but from the south-east in MERRA-2 and JRA55. Given that the latter datasets indicate a decreasing jet strength, and the former does not, an explanation for these differences is needed. The reliance on reanalysis datasets for the study of mid-troposphere winds in this region means we cannot determine the correct direction for entry from this analysis alone. Figure 4 suggests the differences between the datasets may be related to the ways in which they simulate the connection between the Somali and Turkana Jets. The southeasterly orientation of the jet entry in MERRA-2 and JRA55 implies a closer connection to the southwesterly phase of the Somali Jet, which has shown slowing trends at 850 hPa (Rai and Dimri 2017). This may account for the slowing trends in these datasets which are not found in ERA5. However, the Turkana LLJ is present in both MERRA-2 and JRA55 when the Somali Jet is not active, or is blowing in the opposite direction as part of its seasonal reversal. Resolving the relationship between the two LLJs is an important future research direction.

Following Vizy and Cook (2019), we find a significant positive correlation on climatological timescales between the jet core zonal wind speed and MSLP off the coast of Kenya, as well as significant negative correlations inland. This implies a role for pressure gradients in the jet’s climatology. Linear regression analysis shows that MERRA-2 and JRA-55 exhibit increasing pressure inland of the Turkana Jet and decreasing pressure offshore. The dipole in pressure trends could relate to changes to regional overturning circulations (Hartman 2018), or to broader scale intensification to the East African Walker circulation (Mutai et al., mutants, etc.).
2012, L’Heureux et al. 2013, McGregor et al. 2014, King et al., 2020). This, again, suggests a physically plausible mechanism for the jet slowdown in these datasets that warrants further investigation. Future work should also examine why this pressure trend is not present in ERA5; the opposing sign of the MSLP gradient over continental East African in this reanalysis may go some way towards explaining the lack of a jet slowdown, but the more fundamental differences in SLP trends between the reanalysis datasets require explanation. A decreasing wind speed trend in the Turkana Channel at 850 hPa was indicated by Torralba et al. (2017) as part of a global study demonstrating agreement between reanalysis trends over the period 1980-2015. They argued that the agreement between 10m and 850hPa trends indicates that interannual variability in reanalysis wind is controlled by changes in the large-scale circulation, a view supported by our findings.

Basic meteorology suggests that the stronger surface heating in the Turkana Channel and further inland should result in decreasing surface pressure, through the formation of heat lows, but reanalysis suggests the surface pressure has increased over the last 40 years in several months. How should we reconcile these differing trends? We suggest that a combination of more intense small-scale overturning circulations in the channel (as discussed by Hartman (2018)) and the wider-scale Indian Ocean Walker Circulation which results in climatological descent over equatorial East Africa (Mutai et al. 2012, King et al. 2020) may be responsible for the increasing pressure by acting to cap rising air driven by strong surface heating. Although our understanding of the nature of the Walker Circulation is still incomplete, it has been suggested to have been intensifying in recent decades (McGregor et al. 2014, Chung et al. 2019) and to be increasingly important as an influence in future East African climate (Giannini et al. 2018, King et al. 2020).

Building on the links between the Turkana Jet and regional heating and pressure patterns found by Hartman (2018) and Vizy and Cook (2019), we therefore postulate a link between the strength of the Turkana Jet on climatological timescales and regional atmospheric dynamics including the overturning Walker Circulation in the Indian Ocean, expressed through the relationship between the large-scale zonal pressure gradient between the jet exit region and the equatorial Indian Ocean, and the strength of the jet itself. Future work will investigate this relationship in GCM projections.

The strengthening of the Pacific Walker Circulation has been linked to variability associated with the Pacific Decadal Oscillation (PDO; McGregor et al. 2014) and the Atlantic Multidecadal Oscillation (AMO; Chung et al. 2019). Pacific SST variability affects East Africa via perturbations to the Walker Circulation (e.g. Lyon and DeWitt 2012, Hoell et al. 2014, Vigaud et al. 2017). The Atlantic Walker Circulation also affects East Africa (Zhao and Cook 2021) and on decadal timescales is affected by the AMO. The AMO and PDO may therefore exert a high-order control on the Turkana LLJ through their impact on zonal pressure gradients. Investigating this hypothesis is beyond the scope of this paper, though we would encourage further research in this direction, potentially using model experiments with idealised SST. However, the AMO has recently been suggested to be externally forced (Mann et al. 2021) and the role of internal climate variability in 20th century trends is limited (Haustein et al. 2019). Therefore, its invocation as an explanation for climatic trends should be treated with caution.

The issue of trends in East African climate is important because of the uncertainty surrounding projected rainfall changes highlighted by Rowell et al. (2015). We argue that the identification of plausible trends in the Turkana LLJ strength contributes to our understanding of this uncertainty, because we demonstrate that many CMIP5 models are not able to accurately
simulate the Turkana LLJ. A better understanding of the mechanisms controlling the long-term variability of the Turkana LLJ will therefore contribute towards resolving the ‘East African Climate Paradox’ (Rowell et al. 2015).

Further work is also needed to account for the mechanisms of the annual cycle of the Turkana LLJ. This should enable understand deeper understanding of the reliability of the slowing trend. Analysis of the hotspot in low-level inertial instability associated with the Turkana Jet may prove useful (Cook 2015), given its development in the tropics stems from an imbalance between the Coriolis parameter (which increases along the jet axis) and the pressure gradients generated by strong temperature differences between coastal regions and continental interiors (e.g. Hagos and Cook 2007). Recent advances in high-resolution regional climate modelling of the African climate (Stratton et al. 2018) provide an opportunity to investigate jet drivers in additional detail.

Our understanding of the Turkana LLJ ultimately rests on observations taken using pilot balloons in the 1980s. The differences between the reanalyses reflect not only the limited observational network in the region, but also the limitations in our understanding of jet dynamics. A targeted field campaign to measure the Turkana LLJ directly using more up-to-date equipment (such as LIDAR) would be of great benefit and could help resolve outstanding questions about what drives the jet. This approach has been of value in investigating low-level winds in arid zones elsewhere in Africa (Allen and Washington 2013).

7. Conclusion

It is increasingly clear that the Turkana LLJ plays an important role in East African climate, but it remains under-investigated. We analyse its annual cycle and variability in three reanalysis datasets, demonstrating that the jet is present throughout the year with peak speeds between November and March in all three datasets. The jet strength is found to have statistically significant relationships with observed rainfall, with negative correlations in the Turkana Channel and positive correlations in the East African interior. This indicates its role in the aridity of the Turkana region as well as moisture transport and convection triggering in the East African continental interior. An analysis of the interannual variability of the jet strength suggests that it may be slowing down, with consequent implications for East African hydroclimate; potentially increased rainfall in northern Kenya, but drier conditions in South Sudan and southwestern Ethiopia. There are also implications for wind power projects in the Turkana region. However, this trend is not found in the ERA5 reanalysis. We suggest that a slowing jet is a physically plausible consequence of observed patterns of climate change in East Africa, including coast-to-continent temperature gradients and/or the intensification of the Walker Circulation, but further research is needed to determine the robustness of this trend and to explain the differing behaviour of reanalysis datasets.

Given the jet’s relationship with rainfall and the potential slowing trend, it is important to be able to say with confidence how it will change in future. We assessed the ability of CMIP5 models to represent the jet. We found that, while some higher-resolution models were able to simulate the jet, lower-resolution models did not resolve topography at a sufficiently high level of detail. Consequently, there is a large degree of variation within the CMIP5 ensemble in terms of jet representation and links between the jet and rainfall. We argue that models that do not simulate the jet will not be able to make reliable projections of future East African hydroclimate. Future projections should make use of high-resolution regional climate models where possible,
and any assessment of the performance of CMIP6 models in East Africa should include the Turkana LLJ. Finally, there is a need for field measurements of the jet with modern remote sensing equipment such as LIDAR. A targeted field campaign could resolve outstanding questions about the jet’s diurnal and annual cycles, as well as the physical mechanisms controlling its variability. This would enable better constraints on the reanalysis datasets, and a better understanding of the impacts of climate change on East Africa.

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