The Tropical Easterly Jet over Africa and its Representation in Six Reanalysis Products

S. E. Nicholson and D. Klotter
Earth, Ocean and Atmospheric Science
Florida State University
Corresponding Author
Sharon E. Nicholson
Earth, Ocean and Atmospheric Science
Florida State University
snicholson@fsu.edu
1 850 644 4022 (office)
1 850 644 9642 (fax)

KEYWORDS

Tropical Easterly Jet, Sahel rainfall, Interannual variability, Teleconnections, Africa

The project was supported by the National Science Foundation, grant numbers NSF AGS-1445605 and 1854511.

Abstract

This paper compares the characteristics of the Tropical Easterly Jet(TEJ) and upper-level winds in six reanalysis products, compares them with soundings over West Africa, and examines the relationship between Sahel rainfall and the TEJ. The reanalysis products utilized are NCEP 1, ERA 5, JRA 55, MERRA 2, CFSR, and 20th Century Reanalysis. For observational winds, pilot balloon and radiosonde data are evaluated at seven locations in West Africa: Ndjamena, Chad; Niamey, Niger; Ouagadougou, Burkina Faso; Bamako, Mali; and Dakar, Senegal, in the Sahel, and Abidjan, Ivory Coast; and Douala, Cameroon in near equator latitudes. The jet characteristics assessed by MERRA2, NCEP 1, JRA 55, and ERA 5 are similar, but CFSR and 20th Century Reanalysis are outliers in nearly every analysis. They tend to overestimate jet wind speeds, being as much as 25% to 40% higher than for the other reanalyses. NCEP 1 is utilized for establishing a long-term climatology of the Tropical Easterly Jet and for comparisons with Sahel rainfall. Over the period 1948 to 2014, the correlation between rainfall and TEJ magnitude is .72. The results clearly demonstrate a strong relationship between the strength of the TEJ and Sahel rainfall and it appears to be a causal one: the strong jet leading to high rainfall. The factors that appear to control the strength of the jet include the sea-surface temperature (SST) contrast between the central equatorial Pacific and central equatorial Indian Ocean (correlation of -.64), SST contrast between the central equatorial and the southern subtropical Indian Ocean (correlation of -.39), the latitude of the shift between upper-tropospheric easterlies and westerlies in the Southern Hemisphere (correlation of -.84 at 150 hPa), the intensity of the Southern Hemisphere westerlies (correlation of +.52 at 200 hPa). This suggests considerable control on the TEJ by extra-tropical circulation in the Southern Hemisphere.

1. Introduction

One of the most prominent features of the general atmospheric circulation in the tropics is the Tropical Easterly Jet (TEJ). Its existence has been known since the 1950s (Krishna Rao 1952, Koteswaram 1958). This jet lies in the upper troposphere and extends from Asia to West Africa, reaching core speeds in excess of 35 ms^{-1}. It results from the convective heating over the North Indian Ocean and Himalayan-Tibetan plateau and, as a large-scale phenomenon, the TEJ can be explained by simple linear arguments related to the heating (e.g., Gill 1980). It is generally considered to be a feature of the boreal summer (e.g., Hastenrath 1988), but evidence exists of a comparable jet in the Southern Hemisphere during the austral summer (Nicholson and Grist 2003).

The TEJ is an important component of the Indian monsoon. It has been well-studied in that region (e.g., Mishra and Tandon 1983, Chen and van Loon 1987, Mishra 1987, 1993) and several authors have examined its interannual variability in the Asian sector (e.g., Rao and Srinivasan 2016, Nithya et al. 2017, Rao et al. 2015). In contrast, studies of the jet over Africa are sorely lacking (Chen and Yen 1993), despite numerous studies showing that a consistent relationship exists between the strength of the TEJ and rainfall over Africa. The factors are governing its magnitude in the African sector may not be same as those operative in the Asian sector. A case in point is that the interannual variability of the TEJ over Asia is only on the order of 5 ms-1 (Roja Raman et al. 2009, Tanaka 1982), while the magnitude over its changes over Africa are two or three times greater (Grist and Nicholson 2001).

The core of the TEJ lies over the Indian Ocean, just to the south of the Indian subcontinent. The intensity of the TEJ appears to dictate the intensity of the Indian monsoon, with a stronger/weaker jet being associated with anomalously high/low monsoon rainfall (Chen and Yen 1993, Pattanaik and Satyan 2000). The association is apparent not only during the monsoon season but also in May, in advance of the monsoon season. Likewise, the TEJ in the Asian sector tends to be stronger during active monsoon phases than during breaks (Raman et al. 2009).

Davies and Sansom (1952) may have been the first to demonstrate that the jet extends over parts of Africa. Some of the earliest suggestions that the TEJ influences African rainfall were the works of Flohn (1964) and Osman and Hastenrath (1969). Flohn applied jet-streak theory to the TEJ, concluding that patterns of upper-level convergence and divergence in its entrance and exit regions could explain the aridity over North Africa during the boreal summer and the summer rainy seasons over the Sahel and India. Similar considerations were applied by Besler (1984) and Pachur and Kröpelin (1987) to explain the role of the TEJ in the paleoclimate of the Sahara.

With the occurrence of drought in the Sahel in the late 1960s and early 1970s, the relationship between the TEJ over Africa and African rainfall received further attention. Kidson (1974) and Newell and Kidson (1984) documented a relation between a weakened TEJ over Africa and the Sahel drought. Similarly, Kanamitsu and Krishnamurti (1978) produced a diagnostic analysis of 1972, a year of Sahel drought, and 1967, a year of normal rainfall in the Sahel. They showed a strong contrast in the intensity of the TEJ in the two years, with the jet being notably weaker in the drought year. Hulme and Tosdevin (1989) reviewed relevant prior work and concluded that the TEJ probably influences rainfall in the Sudan, but they did not directly evaluate the link.

Later work confirmed that one of the strongest contrasts between wet and dry years in the Sahel is the intensity of the TEJ (Nicholson and Grist 2001, Grist and Nicholson 2001, Nicholson 2009a). Its speed over West Africa was nearly twice as great in Sahel wet years as in Sahel dry years, e.g., 20 to 30 ms^{-1} in 1955 versus 15 ms^{-1} in 1983. This contrast was evident in each month from June through September. An intensification of the TEJ was also apparent in wet years over parts of western equatorial Africa (Dezfuli and Nicholson 2013, Nicholson and Dezfuli 2013) and Ethiopia (Segele et al. 2009).

In each of these cases, the changes in the TEJ were part of major changes in the basic state that included concurrent changes in the mid-level African Easterly Jet (AEJ) and the low-level westerly flow. Hence, the question arises as to what extent the TEJ is independently linked to rainfall variability, as opposed to the TEJ in conjunction with other circulation features, such as the latitude of the AEJ. The question is particularly important because some degree of recovery of the rainfall regime has occurred in the Sahel (Nicholson et al. 2018), but the TEJ appeared to have remained relatively weak (Nicholson and Grist 2001).

In view of this, three questions are posed and evaluated in this article. 1) Is the previously documented relationship between the TEJ and Sahel rainfall evident in recent decades? 2) Is the TEJ in the African sector anomalously strong in all wet years in the Sahel? 3) Conversely, is the Sahel anomalously wet/dry in all years in which the TEJ is anomalously strong/weak in the African sector? In answering these questions, the study will also develop a long-term climatology of the TEJ over West Africa, document its year-to-year variability over West Africa, examine factors in the interannual/decadal variability of the TEJ, and evaluate the long-term relationship between the TEJ and Sahel rainfall.

While the questions posed are relatively simple, obtaining answers to them is not. Ideally, one would examine the upper-tropospheric winds at a handful of upper-air stations over the Sahel. Unfortunately, few continuous records exist and coverage is sporadic at most upper-air stations in the region. Moreover, most observations are from pilot balloon (pibal) ascents, which seldom reach the upper-troposphere. In lieu of sounding data, several reanalysis data sets that do provide continuous coverage are available, but most do not extend back to the relatively wet period of the 1950s and early 1960s. Furthermore, the lack of observational upper-air data means that the reanalysis products have not been validated over the Sahel.

Thus, none of the available data sets is ideally suited for an analysis of the long-term behavior of the TEJ over West Africa. For that reason, radiosonde observations are used here along with reanalysis data. Originally, nine reanalysis products were evaluated to determine which is best suited for the study of the TEJ. However, initial results were very similar for three ERA-products (ERA 40, ERA-Interim, ERA 5) and for NCEP 1 and NCEP 2. This is demonstrated in Table 1 and in Figs. S1 and S2. To simplify and condense the results, only six products are evaluated here: NCAR-NCEP 1, ERA-5, MERRA2, CSFR, JRA 55, and 20th Century Reanalysis V2C. Likewise, most analyses are performed only for August, which tends to be the wettest month in the Sahel and the month upon which several prior studies of the TEJ/rainfall association (e.g., Nicholson and Grist 2001) were based. It should be noted that observations are assimilated into the various reanalysis products, with the exception of 20th Century Reanalysis, which over land is based only on surface pressure. Nevertheless, because vast amounts of additional data are assimilated into the products and possibility that we have assembled a larger observational data set, the comparison with observations is useful.

The structure of this article is as follows. The observational and reanalysis data are described in section 2. The data sets are evaluated in section 3 and the best performing reanalysis data set is used to create a long-term climatology of the TEJ over West Africa, presented in Section 4. This includes core speed, and latitudinal and vertical location of maximum winds. In Section 5 the relationship between Sahel rainfall and the TEJ is examined. Section 6 presents a brief analysis of factors associated with the interannual and decadal variability of the TEJ. Section 7 summarizes conclusions concerning the data sets and the results of the various analyses.

2. Data

The study utilizes rainfall from an archive assembled by the first author (e.g., Nicholson et al. 2018). It includes roughly 600 rainfall stations in the Sahel, with the majority of records extending through 2014. In this study, only data for August are utilized and a single time series is derived to represent variability in the Sahel. This approach is reasonable because of the exceedingly strong coherence of interannual variability throughout the region (e.g., Nicholson and Grist 2001), but particularly in its central sector. A spatial average is derived for the sector shown in Fig. 1, ranging from 12 ° N to 18 ° N and 10 ° W to 20 ° E. Rainfall is calculated as in past studies (e.g., Nicholson et al. 2018) by expressing the monthly total at each station as a standardized departure (departure from the monthly mean divided by the monthly standard deviation), then averaging these values over all stations in the region shown. The resultant time series is shown in Fig. 2.

Upper-level winds are derived both from upper air soundings and from six reanalysis products. Reanalysis winds at 150 hPa and 200 hPa are evaluated, with the level shown depending on the specific analysis; the TEJ is generally stronger at 150 hPa, but data for 200 hPa are probably more reliable, as a result of better observational coverage. Observational data are from the Integrated Global Radiosonde Archive (IGRA), an assemblage of radiosonde and pibal observations from over 2700 globally distributed stations (Durre et al. 2006). The advantage is that these are direct observations, but there are several disadvantages. The full complement of West African stations in the archive within the latitudinal span of the TEJ (10 ° S to 20 ° N) was evaluated. From these, seven with relatively complete records were deemed useful, Douala, Abidjan, Niamey, Ndjamena, Ouagadougou, Bamako, and Dakar. Their locations and the total number of records in IGRA are shown in Fig. 1.

To facilitate the discussion, the jet speeds are generally indicated in absolute terms. That is, they are not indicated as negative, although the flow is easterly. The exception is illustrations of vertical profiles, where a distinction between easterly wind (negative) and westerly wind (positive) is necessary.

Table 2 gives the temporal and spatial resolution of the six reanalysis products used. All but ERA 5 and JRA 55, which are relatively new products, have frequently been used for studies of circulation over Africa. NCEP 1 (Kalnay et al. 1996), which begins in 1948, is the only one of the six that covers the Sahelian wet period of the 1950s and early 1960s. The disadvantage of NCEP 1 is its low spatial resolution, 2.5 degrees of latitude and longitude. Merra 2 (Gelaro et al. 2017), ERA 5, JRA 55, and CSFR (Saha et al. 2010) all have higher spatial resolution, but except for JRA 55, commence only in 1979 or 1980. 20th Century Reanalysis V2C (Compo et al. 2011) has a spatial resolution of 2 degrees. Version V2c, which extends back to 1851, is used. Most analyses are based on the period 1980 to 2014. Only the monthly products are utilized.

Additional data utilized are sea-surface temperatures (SSTs), the Southern Oscillation Index (SOI), and the Indian Ocean Dipole Index (DMI). SSTs were derived from the NOAA Extended Reconstructed SST (ERSST) V5 data set (Huang et al. 2017). The SOI (Trenberth 1984) is the normalized pressure difference between Tahiti and Darwin, Australia. The DMI represents the anomalous SST gradient between the two Indian Ocean sectors: 50E-70E, 10S-10N and 90E-110E,10S-0N (Saji and Yamagata 2003).

3. Evaluation of upper-air data

3.1 Inter-comparison of reanalysis products

The six reanalysis products are compared with each other via latitudinal transects of the zonal wind from the surface to 150 hPa, the spatial patterns of vector winds at 150 hPa, local vertical wind profiles, and interannual variability. The 150 hPa level is assumed to represent the maximum development of the TEJ. A quantitative comparison is made for select mean statistics for all four rainy season months (June through September).

3.1.1 Mean zonal wind

Fig. 3 shows for the six reanalysis products the meridional cross-sections of the mean zonal wind in August in the three longitudinal sectors. In each case, consistent with previous depictions of the TEJ (e.g., Flohn 1964, Nicholson 2009b), the further east, the stronger the TEJ. At 10 ° W core speeds (i.e., the speed maximum) are on the order of 15 to 20 ms^{-1} while at 30 ° E, core speeds are 25 to 30 ms^{-1} in most analyses. At 10 ° W, the highest core speeds are in 20th century Reanalysis and CFSR and lowest are in MERRA. Further east the highest core speeds are again in 20th Century Reanalysis and CFSR and the lowest are in NCEP V1.

Fig. 4 shows the August mean vector winds over northern Africa. Only the analysis for 150 hPa is shown, but comparable results were obtained for 100 hPa and 200 hPa. Strongest winds over Sudan and South Sudan and further east. The vector winds are similar for all reanalyses except CSFR 2 and 20th Century Reanalysis. Among the other products relatively small differences are apparent in the latitudinal expanse and intensity of the TEJ, with MERRA 2 and ERA 5 showing a jet core that is somewhat more intense than that of the other two. All products indicate speeds exceeding 20 ms^{-1} over the two Sudans and eastward, but MERRA 2 shows speeds greater than this in an area extending further west. In NCEP 1, MERRA 2, ERA 5 and JRA 55 speeds exceeding 30 ms^{-1} did not extend further west than over the eastern tip of the African continent. In contrast, CFSR shows speeds of this magnitude or greater well into South Sudan, some 20 degrees of longitude further west. CFSR also shows markedly higher speeds than the other products nearly everywhere.

The six reanalysis products are inter-correlated with respect to the mean August core speed at 10 ° E and at 0 °, i.e., in the heart of the Sahel (Table 3). Core speed for a given month is defined as the maximum speed within the TEJ core as evident in the mean latitude/height wind profile for that month. In order to get more robust correlations, the period of comparison is extended to 1980 to 2014. For that period, the 1% significance level is +/- is roughly .42 and all correlations reach this level. However, there is a strong contrast between correlations involving CFRS and 20th Century Reanalysis versus the remaining products. Among the NCEP 1, MERRA 2, ERA 5 and JRA 55 products the correlations range from .68 to .97, but they are mostly in the range of .8 to 9. The correlations between ERA 5 and JRA 55 is particularly high, ranging between .89 and .96. In contrast, every case of a correlation of .6 or lower involves CFSR or 20th Century Reanalysis. With those products, the correlations are typically in the range of .5 to .6.

3.1.2 Characteristics of the TEJ

The contrasts among the reanalysis products are highlighted by Fig. 5, which shows the mean maximum speed in the jet core as well as the mean pressure level and latitude of the maximum for each product. For magnitude there is good agreement among all products at 10 ° W, but the agreement decreases markedly at 30 ° E. 20th Century Reanalysis and, to a lesser extent CFSR, tend to be outliers. This is particularly evident in the higher June and July speeds at 30 ° E. 20th Century Reanalysis consistently shows the largest spread of magnitudes between June and September. In nearly all cases the TEJ is strongest in all products in July and August, weakest in June or September. The spread among the products is generally oy a few ms^{-1}.

There is considerably more variation among the products for pressure level and latitude of the maximum speed. However, no product is a consistent outlier nor does any product appear to show consistent bias with respect to the others. At 10 ° W there is some consistency, with the pressure level of maximum wind being between 150 and 50 hPa and tending to be highest in June and lowest in August or September. The exception is the 20th Century Reanalysis, where the pressure of the core ranges from roughly 100 hPa to 200 hPa. Elsewhere there is considerable spread among the products, with 20th Century Reanalysis, and to a lesser extent CFSR, being on the low side. In terms of latitude of the core, it tends to be highest in June and July, lowest in August and September. In terms of this parameter, 20th Century Reanalysis tends to be an outlier, with generally lower values than the other products and a large spread among the four months at 10 ° W.

3.1.2 Interannual variability of the TEJ

Fig. 6 shows the interannual variability of TEJ strength in August averaged between 5 ° N and 15 ° N. The year-to-year fluctuations are remarkably similar. In most cases, the spread of the estimates is small among NCEP1, ERA 5, MERRA and JRA 55. The only case with strong disagreement among those products is at 30 ° E and 150 hPa. In contrast, 20th Century Reanalysis and CFSR are clearly outliers, showing substantially higher speeds than the other products. Note that the contrasts among the products are primarily in magnitude; the year-to year patterns of variability are generally similar, even with the outliers. The disparity is particularly strong at 150 hPa, where in general the spread of the estimates is greater than at 200 hPa. This might relate to the fact that at 150 hPa fewer observations are available for assimilation into the reanalyses.

Table 4 shows the average speed over the period 1980 to 2014. In all cases the mean speed increases from west to east. Speeds tend to be similar for NCEP 1, ERA 40 and to a lesser extent ERA Interim, roughly ms^{-1} in the west and 16 to 17 ms^{-1} in the east at 200 hPa and 12 to 13 ms^{-1} in the west and 20 to 21 ms^{-1} in the east at 150 hPa. Compared to those products, MERRA 2 is anomalously high in the east by roughly 1 ms^{-1} at 200 hPa and 5 ms^{-1} at 150 hPa. 20th century reanalysis and CFSR tend to be 2 to 4 ms^{-1} higher than the other products at 200 hPa and as much as 6 to 8 ms^{-1} higher at 150 hPa.

3.2 Comparison of NCEP 1 with observational data

A handful of studies have compared reanalysis winds and upper-air data over West Africa. Stickler and Brönnimann (2011) examined pibal observations for the years 1948 to 1957. They showed that NCEP 1 exhibits a bias compared to observations that can exceed 5 ms^{-1}. They concluded that NCEP 1 overestimates the low-level monsoon flow and produces a monsoon layer that is thicker than observed. Their results further suggested the AEJ speed is overestimated. That study also compared 20th century reanalysis with observations during that period and found somewhat lower differences. The bias was generally of the opposite sign for 20th century reanalysis; notably the bias of monsoon flow was much weaker in 20th century reanalysis and more confined to the coastal region. Wartenburger et al. (2013) evaluated the observational data sets available to Sticker et al. (2010), determining bias and root-mean-square error (RMSE) as compared with ERA Interim. For nine Sahelian stations in West Africa they found that the bias was generally less than 0.8 ms^{-1} and the random error was generally 2 to 3 ms^{-1} at 5000 m elevation.

Nicholson and Grist (2003) and Grist and Nicholson (2001) compared several observational data sets over West Africa, including radiosondes, pibals, and the GFDL radiosonde analysis (Oort 1983), with wind estimates from NCEP 1. They concluded that the salient features of the NCEP 1 analyses, such as contrasts in TEJ intensity between wet and dry years, could be confirmed by the observations. However, the contrasts were not quantified.

Here NCEP 1 winds are compared with the time series and vertical profiles of wind at five Sahelian stations (in the latitude band 12 ° N to 14 ° N) and two stations nearer the equator. A single reanalysis product is used here for the sake of simplicity, but in view of the high correlations with MERRA 2, ERA 5 and JRA 55, those products would presumably yield similar results. The Sahelian stations include Ndjamena, Chad; Niamey, Niger; Ouagadougou, Burkina Faso; Bamako, Mali; and Dakar, Senegal. The remaining two stations are Abidjan, Ivory Coast (c. 5 ° N) and Douala, Cameroon (c. 4 ° N). These were chosen because each station has a large number of data points at and above 200 hPa, as well as a long sequence of years with available data. Notably, the number of observations available varies over time, as does the use of pibals vs. radiosondes. However, the time series created are probably the most realistic estimates of the supra-surface wind regime at the locations examined.

The upper-air data for August at each location are compared with a wind average from NCEP 1 over a 3° x 5° box surrounding the station. Some differences are to be expected because point and areal values are being compared. However, the impact of this contrast is not anticipated to be large because there is strong spatial coherence of the wind regime on the monthly time scales examined here.

Fig. 7 shows the year-to-year values of mean August wind at 200 hPa and 150 hPa for the Sahelian stations. For four of the five there is good agreement between NCEP 1 and observational winds. At Niamey, the average difference is only 1.4 ms^{-1} at 200 hPa and 2.4 ms^{-1} at 150 hPa. For Dakar, the average difference is 1.5 and 1.4 ms-1 for these two levels, respectively. The differences are somewhat greater for Ouagadougou and Bamako: 2.8 and 2.0 ms^{-1}, respectively at 200 hPa and 3.2 and 2.6 ms^{-1} at 150 hPa. The differences between NCEP 1 and observations are notably larger at Ndjamena: 3.0 at 200 hPa and 4.2 at 150 hPa. In all cases, the bias of NCEP 1 compared to observations is smaller than suggested by Stickler and Brönnimann (2011). It tends to be negative at Ouagadougou, Niamey and Bamako, positive at Dakar. At Ndjamena the bias tends to be positive in the very early years but generally negative since the late 1960s onward.

The situation is very different at the two equatorial stations. NCEP 1 tends to have a positive bias at 150 hPa. At 200 hPa it tends to have a positive bias prior to the 1980s and a negative bias in later years, but not consistently so. Agreement is good at the 200 hPa level, differences being on average 2.4 ms^{-1} for Douala and 2.2 ms^{-1} for Abidjan. However, the differences are more than twice that at 150 hPa: 5.5 ms^{-1} for Douala and 5.3 for Abidjan. The reason for the larger disparity at 150 hPa is not clear; in these particular cases, roughly the same amount of upper-air data was available at both levels.

Fig. 7 also depicts interannual variability of the TEJ. Data for the three stations with records extending back to the 1950s (Niamey, Dakar, and Abidjan) show decreasing TEJ intensity from the 1950s to the early 1980s. and a smaller increase from that time. After that time there is a gradual increase but the high magnitudes of the 1950s and 1960s are not attained at Niamey or Dakar. Ndjamena, which also commences in the 1950s, does not suggest a reduction since the 1950s. However, that record may not be reliable as large differences are apparent between observational and reanalysis winds at that location. The remaining stations, which commence much later, do suggest increasing TEJ intensity since the early 1980s. These results are consistent with the results of Grist and Nicholson (2001) and with trends in the TEJ over India, as noted by several authors (see section 6.1).

Fig. 8 shows the vertical profile of mean winds in August, averaged for those years for which both observational and reanalysis data are available. The agreement for zonal winds is excellent at each location. However, there is some positive bias for NCEP 1 in the lowest levels and some negative bias (indicating stronger speeds) at the TEJ level. In most cases the bias is less than 2 ms^{-1}. There is most disagreement in the case of meridional winds and it is evident not only the wind magnitude but also in the vertical structure of the meridional wind profile. However, the biases are still relatively small in absolute terms, generally less than 2 ms^{-1}.

Several factors can contribute to contrasts between observational and reanalysis data. One is the lower sampling with observational data: over Africa few stations report daily and generally observations are made only one or two times per day. Moreover, there is instrumental and observer error in the case of pibals and radiosondes and error inherent in models that produce the reanalysis data. The agreement shown in Figs. 7 and 8 between observed and reanalysis winds is encouraging and suggests that NCEP winds are reliable over West Africa.

4. Long-term climatology

NCEP 1 is used to derive a long-term climatology of the TEJ for the period 1948 to 2017 for each month of the year. Mean vector winds, vertical profiles as a function of latitude, and mean characteristics such as core speed and location are presented. The jet profile was examined at two longitudes, 30 ° E (far eastern Sahel) and 10 ° E (central Sahel). As the results were similar, only data for the latter are shown. However, when notable contrasts were apparent between the two longitudes these are described in the text.

Fig. 9 shows the mean zonal wind profile in each month of the year. The major circulation features include the subtropical westerlies of both hemispheres, the African Easterly Jet of the mid-troposphere, the Tropical Easterly Jet of the upper-troposphere, and the low-level equatorial westerlies. Since the annual cycle of the other features is very well known, only the TEJ will be discussed here.

It is best developed during January and February, when its axis is at roughly 10 ° S, and during June through September, when its axis is at roughly 5 ° N. It is completely absent in March-through-May. When it reappears in June its mean core speeds attain 12.5 ms^{-1} at 10 ° E (Fig. 9) and more than 17.5 ms^{-1} at 30 ° E (not shown). The TEJ peaks in August, with mean core speeds on the order of 20 ms^{-1} at 10 ° W and 10 ° E and over 25 ms^{-1} at 30 ° E. The jet is still clearly apparent in September but in October it is extremely weak at 10 ° E and disappears at 30 ° E. The TEJ is completely absent in November and December.

Fig. 10 shows the mean vector winds at 150 hPa for August, the month when the TEJ is best developed in the Northern Hemisphere, and for February, when the TEJ is best developed in the Southern Hemisphere. In August the core extends across the continent into the eastern Atlantic. In eastern regions, such as Ethiopia and the eastern Sudan, the jet extends from the equator to roughly 25 ° N, with speeds between 20 and 30 ms^{-1} in most of that region. In February, when the TEJ develops in the Southern Hemisphere, the jet is weaker, with speeds between 20 and 20 ms^{-1}. It extends between roughly the equator and 10 ° S and from the central Indian Ocean, across the continent and into the eastern equatorial Atlantic.

Table 5 summarizes the climatological statistics for the TEJ at 150 hPa during the four rainy season months of the Sahel. In general, from west to east the core speed increases. This pattern is evident in all months, with few exceptions. In June the core speed is 11.5 ms^{-1} in the west, but reaches 17 ms^{-1} in the east. The jet is notably stronger in July and August, with core speeds in the west of 17.5 and 19.2 ms^{-1}, respectively, in those months but 23.7 and 21.7 ms^{-1} in the east. In September speeds at each longitude are on the order of 14 to 15 ms^{-1}. Pressure level of the core changes substantially from month to month but in a less consistent fashion. In June and July it generally increases eastward, but tends to decrease eastward in August and September. The latitude of the core also varies substantially from month to month and from west to east. At 10 ° W it is furthest north in June, at 11.1 degrees of latitude. and further south in September at 7.6

degrees of latitude. At 10 o° E it also tends to decrease in latitude between June and September, but is furthest north in July, as it is also at 30 ° E.

Another important characteristic of the TEJ is the pattern of divergence/convergence associated with it. These patterns were calculated at 150 hPa and 200 hPa. Because the patterns were extremely similar at the two levels, Fig. 11 presents only the results for 200 hPa. A strong and consistent region of divergence corresponds to the jet, spanning roughly 5 ° N to 15 ° N over western regions and extending diagonally from SW to NE in more eastern sectors. In much of the band of divergence the magnitude exceeds 2 x 10^{-6} s^{-1}. Roughly the same pattern is evident from June through September, but the divergence is strongest in August and September.

5. Relationship between Sahel rainfall and the TEJ

Past studies (section 5.1) suggest a causal relationship between the TEJ intensity and rainfall over West Africa. Potential mechanisms of a link are discussed in section 5.2. There is some evidence that the intensity of the TEJ is affected by latent heat release, and hence dependent on the intensity of convection. This latter effect, if significant, could provide some explanation for the rainfall/TEJ relationship. The issue of the impact of convection is considered heuristically in section 5.3; arguments favor the large-scale control of the TEJ, as opposed to feedback from convection governed by other factors.

A caveat here is that these analyses and the conclusions drawn from them are valid on the monthly time scale. On that time scale the character of rainfall over the Sahel is a result of the combined influence of the TEJ and the AEJ. The TEJ appears to govern primarily the intensity of rainfall in the region while the AEJ appears to govern primarily the latitudinal location of the rain belt. The rain belt lies between the cores of the two jets (Nicholson and Grist 2001). As noted earlier, year-to-year changes in the TEJ are associated with a major change in the basic state that includes concurrent changes in the mid-level African Easterly Jet (AEJ). So the question arises as to whether or not the AEJ might be driving some of the changes observed in the TEJ. On monthly time scales, that is unlikely because on that time scale there is little change in the magnitude of the AEJ, despite large changes in the magnitude of the TEJ. Moreover, in some years the distance between the jet cores can be as great as 15 degrees of latitude. This implies that an influence of the AEJ on the TEJ is unlikely on the monthly scales we are considering. This does not rule out a link on synoptic time scales associated, for example, with the instabilities of the AEJ and resultant wave actiivty.

5.1 Comparisons between the TEJ and Sahel rainfall

Prior analyses of the interannual and interdecadal variability of rainfall over West Africa suggested that the strength of the TEJ over Africa is a major factor. Fig. 12 shows the contrast between the TEJ in August, 1955 (one of the wettest years on record in the Sahel) and August, 1983 (one of the driest). In wet years the mean August TEJ speed exceeds 30 ms^{-1} while it barely 16 ms^{-1} in the dry years (Nicholson and Webster 2007). The difference is much greater than the bias over West Africa as estimated by Stickler and Brönnimann (2011) or in Section 3.2.

This relationship holds during multi-year and decadal time scales (Grist and Nicholson 2001, Nicholson and Grist 2001). However, only a handful of individual years have been examined (Nicholson 2008, 2009a). Here we examine the robustness of the relationship as applied to individual years.

Fig. 13 compares Sahel rainfall in August with 200 hPa winds near Niamey. Rainfall is averaged over the area 12 ° N to 18 ° N and 10 ° W to 20 ° E (see Fig. 1). The area around Niamey is selected for the winds because Niamey appears to have the most reliable soundings (and the highest number). NCEP 1 winds are averaged for an area near Niamey shown in Fig. 1. Here the 200 hPa level is considered because of the greater agreement with observations as evident in Fig. 7. Upper-air data for Niamey from the IGRA archive are also utilized. The correlation between August Sahel rainfall and NCEP 1 winds over the 67-year period 1948 to 2014 is .72; with IGRA wind the correlation is .62. When wind anomalies are positive rainfall is generally well above normal or only slightly below the mean. When wind anomalies are negative rainfall is generally below normal or just slightly above the mean. However, there are some clear outliers: for example, rainfall in 1989 and 1994 was well above average but the jet was anomalously weak.

A comparison with 20th century reanalysis suggests that the relationship between the TEJ and Sahel rainfall generally holds on the century time scale. For the July-August-September rainy season the correlation is .31 over the period 1871 to 2014, not strong but significant at well above the .1 % level (r = .21).

5.2 Mechanisms linking Sahel rainfall and the TEJ

The analysis in section 5.1 clearly demonstrates a strong positive relationship between the intensity of the TEJ and rainfall over the Sahel. The physical mechanism of the link appears to be primarily upper-level divergence associated with the jet core (Fig. 11). The convergence/divergence couplet evident in Fig. 11 is consistent with Flohn's (1965) depiction of a jet streak circulation associated with the TEJ, which he theorizes is a major reason for the rainfall in the Sahelian latitudes and the aridity over the Sahara. The divergence is evident in the strong meridional components associated with the TEJ over Africa (Nicholson and Grist 2003). This pattern is particularly pronounced in wet years in the Sahel and it appears to play a role in the jet's impact on rainfall (Nicholson 2009a). This is consistent with the observation that a stronger TEJ is linked not only to more rainfall in the Sahel, but also a more intense rainbelt (e.g., Nicholson 2008, 2009a).

The increased/reduced vertical and horizontal shear associated with a strong/weak TEJ is probably also a factor. This impacts the development of easterly waves (Grist et al. 2002, Nicholson et al. 2008). Enhanced wave activity appears to be associated with increased rainfall in the Sahel (Grist 2002).

An additional mechanism of the TEJ/rainfall link may relate to the dynamic instability of the TEJ itself (Mishra 1993, Mishra and Tandon 1983). Both barotropic and combined barotropic-baroclinic instability have been demonstrated. This permits the development of waves on the TEJ. Such waves have been observed over West Africa (Nicholson et al. 2007) and their characteristics match those predicted by a numerical simulation of African wave activity (Nicholson et al. 2008).

5.3 TEJ as a driver of or response to rainfall variability

A close relationship between the intensity of the TEJ and West African rainfall has been clearly established. Redelsperger *et al*. (2002) suggest that latent heat released by convection can enhance upper-level shear, and therefore enhance the TEJ. This evokes a question as to whether the stronger TEJ in wet years is a factor in or a product of the more intense rainfall (LaFore and

Chapelon 2017, Thorncroft and Blackburn 1999). The fact that the upper-troposphere near the jet is warmer in dry years (Grist and Nicholson 2001) belies that hypothesis. Several other observational points likewise counter the suggestion that rainfall controls the intensity of the TEJ (Nicholson 2009a). The most compelling are the temporal and spatial characteristics of the rainfall field versus the jet.

One example is that the location of the TEJ core over West Africa is relatively stationary within the Sahel rainy season and from year to year, despite strong latitudinal shifts in the convection (Nicholson and Grist 2001). Also, the intensity of convection can be equally strong in years of a strong or weak TEJ, as shown by a comparison of the years 1950-56-58 (wet composite) and 1968-84-85 (dry composite). In August the latitudinal profile of rainfall in the central Sahel was roughly identical in the two composites, but displaced some 5 degrees southward in the dry case (Fig. 12). The TEJ core speed reached 30 ms^{-1} in the wet composite versus 16 ms^{-1} in the dry composite; its latitude did not shift, despite the 5 degree shift in the location of maximum convection (Nicholson and Webster 2007). In contrast, the day to day variations of TEJ speed, which arguably reflect the impact of latent heat, are only roughly 2 to 3 ms^{-1} and bear no relationship to day-to-day variations in convection (see Fig. S1). In this same vein, it is useful to point out that the TEJ intensity is stronger over the eastern Sahel than over the central and western Sahel, where rainfall is considerably higher.

Another case in point is that the contrast of TEJ intensity between wet and dry years is already apparent in June, a month in which there is little or no contrast in rainfall between wet and dry years in the Sahel (Nicholson 2008, 2009a). In the years examined by Grist and Nicholson (2001), the mean TEJ core speed in June was 18 ms^{-1} in the wet composite versus 8 ms^{-1} in the dry composite. June rainfall as a function of latitude was roughly the same in the two composites, indicating that enhanced convection could not have been responsible for the stronger TEJ in the wet composite.

A further relevant argument is that the mid- and upper-tropospheric temperature gradient is greater in the years or composites with a stronger TEJ and that this contrast is imposed by conditions in the higher latitudes (Grist and Nicholson 2001). The temperature contrasts are strongest in the subtropical latitudes, especially those of the Southern Hemisphere, and not in the vicinity of the jet.

A final argument is based on model simulations using only dry dynamics. When the basic state corresponding to wet years/composites, with an anomalously strong TEJ, is perturbed, waves growth at a faster rate, become more intense, and extend throughout the troposphere (Nicholson et al. 2008, Grist et al. 2002). In contrast, in the dry cases with the weak TEJ, the resultant waves are limited to lower-levels and best developed near the surface, grow very slowly and remain relatively weak. These simulated changes are consistent with observations. The upper-tropospheric shear, which is dependent mainly on TEJ intensity, appears to be the key factor in wave development. Both horizontal and vertical shear are greater in wet years than in dry years (Grist et al. 2002). Potential vorticity considerations likewise suggest that a link between the surface and TEJ is a major factor in the interannual variability of Sahel rainfall (Nicholson et al. 2008).

6. Factors associated with the interannual variability of the TEJ

Few studies have examined the interannual variability of the TEJ over Africa. The most comprehensive was Nicholson and Grist (2001), who examined trends in the TEJ between 1958 and 1997 using NCEP Reanalysis. The core speed of the TEJ decreased markedly throughout the 1960s and early 1970s in all four months of the Sahel rainy season. The decrease was particularly strong in August; core speed dropped from ca. 26 ms^{-1} in 1958 to 15 ms^{-1} in 1972. In later years it varied between ca. 13 and 19 ms^{-1}. Sathiyamoorthy (2005) showed a similar reduction in the African and Atlantic sectors between the 1960s and 1990s.

So far, there have been no studies specifically assessing the factors in the interannual variability of TEJ intensity over Africa. However, several papers give hints as to what might play a role. The intensity appears to change in relationship to the intensity of the extra-tropical Southern Hemisphere westerlies (Dezfuli and Nicholson 2013, Nicholson and Dezfuli 2013, Grist and Nicholson 2001). The stronger jet is also commensurate with cooler upper-tropospheric temperatures in the tropics and stronger latitudinal temperature gradients (Nicholson 2009a). On synoptic time scales, acceleration of the TEJ appears to follow the strengthening of the Indian monsoon (LaFore and Chapelon 2017).

Here the relationship between TEJ core speed and several large-scale factors is evaluated. These include global sea-surface temperatures (SSTs), upper-level temperature gradients, the Southern Oscillation and Indian Ocean Zonal Mode, and the intensity and location of the extra-tropical upper-level westerlies. The core speed is represented by the average zonal wind in August at 150 hPa in a box extending from 5 ° N to 15 ° N, latitudes where the TEJ is best developed, and from 0 ° W to 20 ° E (i.e., central areas of West Africa).

Fig. 14 shows the location of this box and the correlation of the mean zonal wind in this box with global SSTs between 40 ° N and 40 ° S. Correlations are based on the years 1980 to 2014, so that the 5% significance level is +/- .34 and the 1% significance level is +/- .44. The latter is exceeded throughout most of the analysis sector. Strongest correlations are in the central equatorial Indian Ocean, where the correlation exceeds .55, and in the central equatorial Pacific, where the correlation exceeds -.55. This indicates that a stronger TEJ is associated with warmer SSTs in the Indian Ocean, but colder SSTs in the Pacific. This opposition suggests that the SST difference between the two oceans might play a role in the interannual variability of the TEJ. SSTs are calculated for the two small boxes shown in Fig. 14 and the difference calculated as the Pacific SSTs minus the Indian Ocean SSTs. The correlation between TEJ zonal wind speed over West Africa and the SST difference between these two regions is -.73, with a greater warming of the Indian Ocean relative to the Pacific being associated with a stronger TEJ over West Africa.

The speed of the TEJ was correlated with several other parameters with the results indicated in Tables 6 and 7. Fig. 14 suggests a possible relationship with the latitudinal SST gradient in the Indian Ocean, so that Table 6 includes a correlation between the TEJ zonal speed and SSTs in equatorial and subtropical sectors, as well as a correlation with the SST difference between those sectors (indicated by the larger boxes in Fig. 14). The correlation with individual sectors is low but the correlation of the difference is -.39, roughly significant at the 5 % level but much lower than the correlation with the Pacific-Indian Ocean SST difference. Table 6 also shows the correlation of the TEJ zonal speed with the Southern Oscillation (SOI) and the Indian Ocean dipole (i.e., the Dipole Mode Index or DMI). A significant correlation is apparent with the SOI (+.44), but the correlation with the DMI (+.16) is not significant. The relatively low correlation with the SOI is somewhat surprising, since the intensity of the TEJ in the Indian Ocean sector is strongly modified by ENSO (Nicholson 2015).

The final parameters examined relate to upper-level circulation (Table 7). One is the latitude at which the prevailing wind shifts from easterly to westerly. This is assessed for both hemispheres and at both 150 and 200 hPa. The second is the intensity of the subtropical westerlies at these same pressure levels. The correlations are highly significant in both hemispheres for the latitude of the switch, but considerably higher for the Southern Hemisphere. There the correlation is -.79 at 200 hPa and -.84 at 150 hPa. The TEJ speed is also significantly correlated with the intensity of the Southern Hemisphere westerlies, +.38 at 150 hPa and +.52 at 200 hPa. These results suggest substantial control of the TEJ over West Africa by the Southern Hemisphere extra-tropical circulation.

7. Summary and Conclusions

The six reanalysis products evaluated show considerable diversity in the representation of the Tropical Easterly Jet. The jet characteristics assessed by MERRA2, NCEP 1, and the two ERA products are fairly similar, but CFSR and 20th Century Reanalysis are outliers in every analysis. They tend to overestimate jet wind speeds, being as much as 25% to 40% higher than in the other reanalyses. Use of those two products to study West Africa is not to be recommended. The most reliable in this region appears to be NCEP 1. It has the highest correlation with the other products and it compares very favorably with upper-air observations at West African stations.

The reanalysis products, which commence only in 1979 or 1980, do not show a clear trend over time. However, the upper-air observations for Niamey, Dakar and Abidjan show a definite decrease in TEJ intensity since the 1950s and a slight increase since the 1980s. That increase is apparent at the other upper-air sites as well. This is consistent with analyses of the TEJ over India that likewise show a decreasing trend since the 1950s (e.g., Rao et al. 2004).

In the introduction, three questions were posed. The first is whether the previously documented relationship between the TEJ and Sahel rainfall is maintained in recent decades. This work shows that it clearly is for August. The correlation with TEJ intensity over West Africa over a period of 67 years is .72 based on NCEP 1 winds and .62 based on upper-air observations at Niamey. While this is only one month of the boreal summer rainy season, it is the month that is most critical in determining the annual rainfall total in the Sahel. For the 144-year period 1871 to 2014 the Sahel rainfall-TEJ intensity correlation is still -.31, despite the uncertainty of the 20th Century Reanalysis data used to assess the TEJ.

The remaining questions are whether the TEJ is anomalously strong in all wet years in the Sahel and, conversely, whether the Sahel is anomalously wet/dry in all years in which the TEJ is anomalously strong/weak. Clearly an anomalously wet year can occur without an anomalously strong TEJ (Fig. 14). Likewise, a strong/weak TEJ does not guarantee wet/dry conditions. Nevertheless, there are relatively few years that deviate strongly from these associations. Overall, however, a strong/weak TEJ is clearly conducive to wet/dry conditions.

This study also considered the question as to what controls the intensity of the TEJ. Arguments are presented to counter the suggestion by some authors (e.g., Thorncroft and Blackburn 1999) that its intensity is strongly influenced by latent heat release associated with convection. This would suggest control of TEJ intensity by rainfall variability, rather than the jet serving as a causal factor in rainfall variability. The strongest arguments against this hypothesis are 1) that the location of the TEJ core over West Africa is relatively stationary within the Sahel rainy season and from year to year, despite strong latitudinal shifts in the convection and 2) that the intensity of convection can be equally strong in years of a strong or weak TEJ.

A further argument is that the intensity of the TEJ is strongly correlated with global SSTs and with atmospheric circulation parameters. The strongest correlation with SSTs, -.73 over a period of 67 years, is with the difference between SSTs in the central equatorial Pacific and the central equatorial Indian Ocean. This might be indicative of control by the upper-level outflow from the Indian Ocean Walker cell. The highest correlation overall is with the latitude of the shift between upper-tropospheric easterlies and westerlies in the southern Hemisphere (correlation of -.84 at 150 hPa). The TEJ intensity is also strongly correlated (r = +.56) with the strength of the Southern Hemisphere westerlies. Correlations with the Southern Oscillation Index, the Indian Ocean dipole, and Northern Hemisphere circulation parameters are much weaker and generally not highly significant.

These results suggest considerable control on the Tropical Easterly Jet by extra-tropical circulation in the Southern Hemisphere. This is consistent with studies that showed a strong relationship between Sahel rainfall and Southern Hemisphere parameters (e.g., Silvestri and Vera 2009, Sun et al. 2010), as well as a link between African Easterly Waves over the Sahel and the Southern Hemisphere extra-tropics (e.g., Cheng and Thorncroft 2019, Yang et al. 2018). Other evidence of a link between the TEJ and the Southern Hemisphere extra-tropics comes from several papers demonstrating that convection in tropical regions can be affected by potential vorticity streamers associated with Rossby waves in the Southern Hemisphere (e.g., Funatsu and Waugh 2008, Rodwell 1997). These streamers reach monsoonal regions of eastern Asia and are advected across the Indian Ocean into Africa (Ortega et al. 2017). While this phenomenon has been demonstrated only on synoptic time scales, this clearly provides a mechanism of interaction between the tropics and the mid-latitudes of the Southern Hemisphere.

ACKNOWLEDGEMENTS

The project was supported by the National Science Foundation, grant numbers NSF AGS-1445605 and 1854511. The authors would also like to thank Chris Funk and colleagues for making the CHIRPS data set available to the scientific community. Valuable suggestions of Prof. Peter Webster are also appreciated.

TABLE 1: Coefficients of linear correlation between winds in the indicated reanalysis products at two longitudes and at 150 hPa and 200 hPa. Calculations are based on the period 1980 to 2014 and winds are averaged at the indicated longitude over a latitude span of 5 ° N to 15 ° N.

SOURCES CORRELATED		150 hPa		200 hPa	
		10 ° E	0 ° E	10 ° E	0 ° E
NCEP V1	NCEP V2	.87	.90	.94	.95
ERA 40	ERA Int	.85	.89	.79	.96
ERA 40	ERA 5	.87	.85	.86	.94
ERA Int	ERA 5	.90	.93	.93	.98

TABLE 2: Characteristics of Reanalysis Products used in this Study

Reanalysis	Spatial Resolution in Degrees	Years
NCEP V1	2.5	1948 – present
ERA 5	0.25	1979 - 2019
JRA 55	1.25	1958-2019
MERRA 2	0.667 lon, 0.5 lat	1980 - present

CFSR V1	0.5	1979-2010
CFSR V2	0.5	2010-present
20th V2c	2.0	1851 - 2014

TABLE 3: Coefficients of linear correlation between August zonal winds in the reanalysis products at two longitudes and at 150 hPa and 200 hPa. Calculations are based on the period 1980 to 2014. Correlations lower than or equal to .6 are highlighted.

SOURCES CORRELATED		150 hPa		200 hPa	
		10 ° E	0 ° E	10 ° E	0 ° E
NCEP V1	MERRA	0.74	0.79	0.84	0.84
NCEP V1	CFSR	0.69	0.58	0.68	0.59
NCEP V1	20th V2c	0.79	0.69	0.60	0.46
NCEP V1	ERA 5	0.84	0.68	0.87	0.88
NCEP V1	JRA 55	0.75	0.84	0.75	0.89
MERRA	CFSR	0.47	0.61	0.60	0.63
MERRA	20th V2c	0.63	0.61	0.54	0.51
MERRA	ERA 5	0.88	0.95	0.95	0.97
MERRA	JRA 55	0.85	0.96	0.85	0.95
CFSR	20th V2c	0.56	0.57	0.56	0.51
CFSR	ERA 5	0.67	0.72	0.60	0.67
CFSR	JRA 55	0.69	0.71	0.63	0.678

20th V2c	ERA 5	0.75	0.67	0.51	0.53
20th V2c	JRA 55	0.76	0.64	0.64	0.58
ERA 5	JRA 55	0.95	0.94	0.89	0.96

TABLE 4: Mean speed of the TEJ (ms^{-1}) in six reanalysis products, at three longitudes and two levels over West Africa.

	200 hPa			150 hPa		
Source	10W	10E	30E	10W	10E	30E
NCEP 1	13.4	13.8	17.3	14.2	15.4	19.9
ERA 5	12.6	14.4	16.5	13.1	16.4	20.7
JRA 55	12.7	15.0	17.5	13.0	17.1	22.9
MERRA	12.2	13.8	17.1	12.9	15.9	23.7
CFSR	14.7	16.2	19.8	15.3	19.9	28.4
20th VC2	16.1	16.8	20.1	18.1	21.3	26.7

TABLE 5: Climatology of the TEJ, as derived from NCEP 1: mean zonal speed (ms^{-1}), pressure level of maximum wind (hPa), and latitude of maximum wind.

	zonal core speed	pressure of core	latitude of core

	10W	10E	30E	10W	10E	30E	10W	10E	30E
June	11.5	12.2	17.0	89	99	140	11.1	10.6	8.9
July	17.5	18.7	23.7	112	100	138	9.4	11.9	12.3
August	19.2	19.0	21.7	140	113	129	7.9	9.6	11.8
September	14.5	13.7	15.0	130	124	117	7.6	8.5	10.6

TABLE 6: Coefficients of linear correlation of NCEP1 150 hPa wind in August with SSTs, the SOI, and the DMI during the period 1980 to 2014.

August Sea Surface Temperatures (SSTs) in equatorial regions			Indian Ocean August SSTs			SOI	DMI
Pacific	Indian	difference	Subtropical	Equatorial	difference	August	August
-.51	+.39	-.64	-.17	.37	-.39	+.44	+.16

TABLE 7: Coefficients of linear correlation between NCEP1 150 hPa wind and select circulation parameters during the period 1980 to 2014. Winds are averaged over the latitudes 5 ° N to 15 ° N.

Latitude of Shift between Easterlies and Westerlies	Intensity of Extra-tropical Westerlies

Southern Hemisphere		Northern Hemisphere		Southern Hemisphere		Northern Hemisphere	
200 hPa	150 hPa	200 hPa	150 hPa	200 hPa	150 hPa	200 hPa	150 hPa
-.79	-.84	-.58	-.49	+.52	+.38	+.22	+.10

FIGURES

1. Sahelian sector over which rainfall data are averaged (green box) and location of seven upper-air stations used in the analyses. The total number of upper-air observations for each station is also indicated on the right. Small box indicates sector over which NCEP 1 winds are averaged for comparison with rainfall in Fig. 13. D = Dakar, B = Bamako, O = Ouagadougou, N = Niamey, Nj = Ndjamena, A = Abidjan, Do = Douala.

2. August rainfall in the Sahel averaged for the sector shown in Fig. 1. Rainfall is expressed as a regionally-averaged standard departure.

3. Meridional profiles of zonal wind in August for 6 reanalysis products and three longitudinal locations: 10 ° W, 10 ° E and 30 ° E.

4. Mean vector winds in August for six reanalysis products at 150 hPa.

5. Mean jet characteristics at three longitudes as depicted by the six reanalysis products: latitude, elevation, and magnitude of the core speed of the jet. Results are shown for the months of June, July, August, and September.

6. Interannual variability of August zonal winds (ms^{-1}) at three longitudes from 1980 to 2014. Data are given for six reanalysis products and for both the 200 hPa and 150 hPa levels. The wind is averaged over the latitudes 5 ° N to 15 ° N.

7. Interannual variability of NCEP 1 reanalysis zonal wind and upper-air observations for seven stations. Data are for August only and cover all years in which both NCEP 1 and upper-air observations are available. Red portions of the lines indicate cases appears when the NCEP V1 value exceeds observations and blue portions of the lines indicate cases when the value of the observations exceeds that of NCEP 1. Data are in ms^{-1}. NCEP 1 data are for the grid box in which the station lies.

8. Vertical profiles in mean zonal and mean meridional wind at seven West African stations. Red line is wind from NCEP 1; black line is observed wind from pibals and radiosondes.

9. Mean zonal wind profiles in each month at 10 ° E and 30 ° E from NCEP1.

10. Mean vector winds at 150 hPa in February and August, from NCEP1.

11. Mean divergence at 150 hPa in June, July, August, and September, based on NCEP 1.

12. Left: Mean zonal winds at 150 hPa in August of 1955 and 1983 (from Nicholson 2009a). Right: mean August rainfall in the Sahel as a function of latitude in a wet composite and a dry composite (from Nicholson and Webster 2007).

13. Scatter diagrams of zonal wind versus August Sahel rainfall, with the latter expressed in units of standardized departures from the mean. On the left is NCEP 1 wind; on the right is observed wind from pibals and radiosondes at Niamey. Rainfall is averaged over the box shown in Fig. 1; NCEP 1 wind is averaged for an area near Niamey, as shown in Fig. 1.

14. Correlation between NCEP 1 wind in August at 150 hPa and tropical SSTs in August. The black box indicates the area over which wind is averaged. Large and small boxes indicate areas for which the SST difference is correlated with wind (see text for details).

S1. Meridional profiles of zonal wind in August at 200 hPa based on the three ERA products and on observations.

S2. Interannual variability of August zonal winds (ms^{-1}) at three longitudes from 1980 to 2014. Data are given for the ERA reanalysis products and for both the 200 hPa and 150 hPa levels. The wind is averaged over the latitudes 5 ° N to 15 ° N.

S3. Daily values of NCEP 1 wind (ms-1) in the vicinity of Niamey and Niamey rainfall from CHIRPS (Funk et al. 2015) for July and August of the years 1984 and 2007. The standard deviation of wind over the two-month intervals are also indicated. Red line is the mean.

APPENDIX 1

Figs. S1 and S2 demonstrate that the results of the analysis of the three ERA products are very similar and adequately replicate the observed winds. The main contrast among the products is the ERA 40 tends to underestimate wind speed at the two western-most locations, especially at 150 hPa.

APPENDIX 2

Fig. S3 shows the mean daily TEJ speed at 200 hPa for the peak rainy months of July and August during a very wet year (2007) and a very dry year (1984). Here Niamey is used as an example but results were similar for other stations and years. CHIRPS data (Funk et al. 2015) is used for rainfall, as NIC131 does not have daily values. No relationship is evident between rainfall and TEJ speed on daily time scales. The correlation during 2007 is -.11 at 0 lag, -.10 with the TEJ lagged by one day and +.18 with the TEJ lagged by two days. For 1984 the correlation is +.02 at 0 lag and +.04 with the TEJ lagged by one or two days. The standard deviation of TEJ speed is 2.2 ms^{-1} within the dry year and 3.4 ms^{-1} within the wet yet, thus indicating that any possible influence of convection is considerably smaller than the difference in TEJ intensity between wet and dry years.

SUPPLEMENTARY MATERIAL

Supplementary material is included. It consists of a figure (Fig. S1) comparing wind profiles based on ERA products with observations at seven West African stations, a figure (Fig. S2) showing year-to-year variations of the TEJ speed in the ERA products, and a figure showing daily variations in TEJ speed and rainfall during July and August of a wet year and a dry year (Fig. S3), and two appendices explaining the two figures.

REFERENCES

Besler H (1984) The tropical easterly jet as a cause for intensified aridity in the Sahara. In Coetzee, J. A., and van Zinderen Bakker, E. M., (eds), *Palaeoecology of Africa and the Surrounding Islands*, **16**: 163-172.

Chen T-C, van Loon H (1987) Interannual variation of the tropical easterly jet. *Monthly Weather Review* **115**: 1739–1759.

Chen T-C, Yen M-C (1993) The effect of planetary-scale divergent circulation on the interannual variation of the summertime stationary eddies. *Tellus* **45A**: 15–27.

Cheng Y-M, Thorncroft CD, Kiladis GN (2019) Two contrasting African Easterly Wave behaviors. *Journal of the Atmospheric Sciences* **76**: DOI 10.1175/JAS-D-18-0300.1.

Compo GP et al (2011) The twentieth century reanalysis project. *Quarterly Journal of the Royal Meteorological Society* **137**: 1–28.

Davies DA, Sansom HW (1952) Easterly jets above East Africa. *Weather* **7**: 343-344.

Dee DP et al (2011) The ERA-Interim reanalysis: configuration and performance of the data assimilation system. *Quarterly Journal of the Royal Meteorological Society* **137**: 553–597

Dezfuli AK, Nicholson SE (2013) The relationship of interannual variability in western equatorial Africa to the tropical oceans and atmospheric circulation. Part II. The boreal autumn. *Journal of Climate* **26**: 66-84.

Durre I, Vose RS, Wuertz DB (2006) Overview of the Integrated Global Radiosonde Archive. *Journal of Climate* **19**: 53-68.

Flohn H (1964) Investigations on the Tropical Easterly Jet. *Bonner Meteorologische Abhandlungen* **4**: 83 pp.

Funatsu BM, Waugh DW (2008) Connections between potential vorticity intrusions and convection in the eastern tropical Pacific. *Journal of the Atmospheric Sciences* **65**: 987-1002.

Funk C, Peterson P, Landsfeld M, Coauthors (2015) The climate hazards infrared precipitation with stations – a new environmental record for monitoring extremes. *Scientific Data* **2**, No. 150066, doi: 10/1038/sdata.2015.66.

Gelaro R, McCarty W, Suárez MJ et al (2017) The modern-era retrospective analysis for research and applications, version 2 (MERRA-2). *Journal of Climate* **30:** 5419–5454.

Gill AE (1980) Some simple solutions for heat-induced tropical circulation. *Quarterly Journal of the Royal Meteorological Society* **106**: 447-462.

Grist JP (2002) Easterly waves over Africa. Part I: The seasonal cycle and contrasts between wet and dry years. *Monthly Weather Review* **130**: 197-211.

Grist JP, Nicholson SE (2001) A study of the dynamic factors influencing the interannual variability of rainfall in the West African Sahel. *Journal of Climate* **14**: 1337-1359.

Grist JP, Nicholson SE, Barcilon AI (2002) Easterly waves over West Africa II. Observed and modelled contrasts between wet and dry years. *Monthly Weather Review* **130**: 212-225.

Hastenrath S (1988) *Climate and Circulation in the Tropics*. D. Reidel, 455 pp.

Huang B, Thorne PW, Banzon VF, Boyer T, Chepurin G, Lawrimore, JH, Menne MJ, Smith TM, Vose RS, Zhang HM (2017) Extended Reconstructed Sea Surface Temperature version 5 (ERSSTv5), Upgrades, validations, and intercomparisons. *Journal of Climate* **30**: 8179-8205.

Hulme M, Tosdevin T (1989) The Tropical Easterly Jet and Sudan rainfall) A review. *Theoretical and Applied Climatology* **39**: 179-187.

Kalnay E et al (1996) The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American Meteorological Society* **77**: 437–471

Kanamitsu M, Krishnamurti TN (1978) Northern summer tropical circulations during drought and normal rainfall months. *Monthly Weather Review* **106**: 331–347.

Kidson JW (1977) African rainfall and its relation to the upper air circulation. *Quarterly Journal of the Royal Meteorological Society* **103**: 441–456.

Koteswaram P (1958) The easterly jet stream in the tropics. *Tellus* **10:** 43–57.

Krishna Rao PRKri (1952) Probable regions of jet streams in the upper air over India. *Current Science* **21**: 63-64.

Lafore JP, Chapelon N (2017) West African synthetic analysis and forecast: WASA/F. In (Parker DJ and Diop-Kane J, eds) *Meteorology of Tropical West Africa*, Wiley Blackwell, Chichester, United Kingdom, 423-451.

Mishra SK (1993) Non-linear barotropic instability of upper- tropospheric tropical easterly jet on the sphere. *Journal of the Atmospheric Sciences* **50**: 3541–3552.

Mishra SK (1987) Linear barotropic instability of the tropical easterly jet on a sphere. *Journal of the Atmospheric Sciences* **44**: 373–383.

Mishra SK, Tandon MK (1983) A combined barotropic-baroclinic instability study of the upper tropospheric tropical easterly jet. *Journal of the Atmospheric Sciences* **40**: 2708–2723.

Nagarjuna Rao D, Gibies G, Sabeerali CT, Ramu DA, Rao Suryachandra A (2015) Changing relationship between the Tropical Easterly Jet and summer monsoon rainfall: Role of the Indian Ocean warming. Indian Journal of Geo-Marine Science 44: 1678-1683.

Newell RE, Kidson JW (1984) African mean wind changes between Sahelian wet and dry periods. *Journal of Climatology* **4**: 1–7.

Nicholson SE, Grist JP (2001) A simple conceptual model for understanding rainfall variability in the West African Sahel on interannual and interdecadal time scales. *International Journal of Climatology* **21**: 1733-1757.

Nicholson SE, Grist JP (2003) The seasonal evolution of atmospheric circulation over West Africa and Equatorial Africa. *Journal of Climate* **16**: 1013-1030.

Nicholson SE, Barcilon AI, Challa M, Baum J (2007) Wave Activity on the Tropical Easterly Jet. *Journal of the Atmospheric Sciences* **64**: 2756-2763.

Nicholson SE, Webster PJ (2007) A Physical Basis for the Interannual Variability of Rainfall in the Sahel. *Quarterly Journal of the Royal Meteorological Society* **133**: 2065-2084.

Nicholson SE, Barcilon AI, Challa M (2008) An analysis of West African dynamics using a linearized GCM. *Journal of the Atmospheric Sciences* **65**: 1182-1203.

Nicholson SE (2008) The intensity, location and structure of the tropical rainbelt over West Africa as factors in interannual variability. *International Journal of Climatology* **28**: 1775-1785.

Nicholson SE (2009a) On the factors modulating the intensity of the tropical rainbelt over West Africa. *International Journal of Climatology* **29**: 673-689.

Nicholson SE (2009b) A revised picture of the structure of the "monsoon" and land ITCZ over West Africa. *Climate Dynamics* **32**: 1155-1171.

Nicholson SE (2015) Long-term variability of the East African "short rains" and its links to large-scale factors. *International Journal of Climatology* **35**: 3979-3990.

Nicholson SE, Dezfuli AK (2013) The relationship of interannual variability in western equatorial Africa to the tropical oceans and atmospheric circulation. Part I. The boreal spring. *Journal of Climate* **26:** 45-65.

Nicholson SE, Fink A, Funk C (2018) Assessing recovery and regime change in West African rainfall from a 161-year record. *International Journal of Climatology* **38:** 3770-3786.

Nithya N, Manoj MG, Mohankumar K (2017) Effect of El Niño/La Niña on tropical easterly jet stream during Asian summer monsoon season. *International Journal of Climatology* **37**: doi.org/10.1002/joc.5137.

Oort A (1983) *Global atmospheric circulation characteristics: 1958– 73*. NOAA Professional Paper 14, 180 pp (47 microfiches).

Ortega S, Webster PJ, Toma V, Chang H-R (2017) Quasi-biweekly oscillations of the South Asian monsoon and its co-evolution in the upper and lower troposphere. *Climate Dynamics* **49**: 3159-3174.

Osman OE, Hastenrath SL (1969) On the synoptic climatology of summer rainfall over Central Sudan. *Archives für Meteorologie, Bioklimatologie, und Geophysik*, Ser. B **17**: 297-324.

Pachur H-J, Kröpelin S (1987) Wadi Howar: Paleoclimatic evidence from an extinct river system in the southeastern Sahara. *Scienc*e **237**: 298-300.

Pattanaik DR, Satyan V (2000) Fluctuations of Tropical Easterly Jet during contrasting monsoons over India) a GCM study. *Meteorology and Atmospheric Physics* **75:** 51–60.

Raman C, Ramanathan Y (1964) Interaction between lower and upper tropospheres. *Nature* **204**: 31–35.

Rao BRS, Rao DVB, Rao VB (2004) Decreasing trend in the strength of Tropical Easterly Jet during the Asian summer monsoon season and the number of tropical cyclonic systems over Bay of Bengal. *Geophysical Research Letters* **31**: L141031–L141033S.

Rao S, Srinivasan J (2016) The impact of latent heating on the location and strength of the tropical easterly jet. *Meteorology and Atmospheric Physics* **128**: 247-261.

Redelsperger J-L, and Coauthors (2002) Multi-scale description of a Sahelian weather system representative of the West African monsoon. *Quarterly Journal of the Royal Meteorological Society* **128**: 1229–1257.

Rodwell MJ (1997) Breaks in the Asian monsoon: The influence of Southern Hemisphere weather systems. *Journal of the Atmospheric Sciences* **54**: 2597-2611.

Saha S, Moorthi S, Wu XR, Wang J, Nadiga S, Tripp P, Behringer D, Hou YT, Chuang HY, Iredell M, Ek M, Meng J, Yang RQ, Mendez MP, Van den Dool H, Zhang Q, Wang WQ, Chen MY, Becker E (2014) The NCEP Climate Forecast System Version 2. *Journal of Climate* **27**: 2185-2208.

Saji NH, Yamagata T (2003) Possible impacts of Indian Ocean Dipole mode events on global climate. *Climate Research* **25**: 151-169.

Sathiyamoorthy V (2005) Large scale reduction in the size of the Tropical Easterly Jet. *Geophysical Research Letters* **32**: Article ID L14802.

Segele ZT, Lamb PJ, Leslie LA (2009) Large-scale atmospheric circulation and global sea surface temperature associations with Horn of Africa June-September rainfall. *International Journal of Climatology* **29**: 1075–1100.

Silvestri F, Vera C (2009) Nonstationary Impacts of the Southern Annular Mode on Southern Hemisphere Climate. *Journal of Climate* **22**: 6142-6148.

Stickler A, Grant AN, Ewen T and Coauthors (2010) The Comprehensive Historical Upper-Air Network. *Bulletin of the American Meteorological Society* **91**: 741-7511.

Stickler A, Brönnimann S (2011) Significant bias of the NCEP/NCAR and twentieth-century reanalyses relative to pilot balloon observations over the West African Monsoon region (1940-1957). *Quarterly Journal of the Royal Meteorological Society* **137**: 1400-1416.

Sun J, Wang H, Yuan W (2010) Linkage of the boreal spring Antarctic Oscillation to the West African summer monsoon. *Journal of the Meteorological Society of Japan* **88**: 15-28.

Tanaka M (1982) Interannual fluctuations of the Tropical Easterly Jet and the summer monsoon in the Asian region. *Journal of the Meteorological Society of Japan* **60**: 865-875.

Thorncroft CD, Blackburn M (1999) Maintenance of the African easterly jet. *Quarterly Journal of the Royal Meteorological Society* **125**: 763–786.

Trenberth KE (1984) Signal versus noise in the Southern oscillation. *Monthly Weather Review* **112**: 326–332.

Uppala SM, Kallberg PW, Simmons AJ, Andrae U, Bechtold VD, Florino M, Gibson JK, Haseler J, Hernandez A, Kelly GA, Li X, Onogi K, Saarinen S, Sokka N, Allan RP, Andersson E, Arpe K, Balmaseda MA, Beljaars ACM, van de Berg L, Bidlot J, Bormann N, Caires S, Chevallier F, Dethof A, Dragosavac M, Fisher M, Fuentes M, Hagemann S, Holm E, Hoskins BJ, Isaksen L, Janssen PAEM, Jenne R, McNally AP, Mahfouf J-F, Morcrette J-J, Rayner NA, Saunders RW, Simon P, Sterl A, Trenberth KE, Untch A, Vasiljevic D, Viterbo P, Woolen J (2005) The ERA-40 re-analysis. *Quarterly Journal of the Royal Meteorological Society* **131**: 2961-3012.

Wartenburger R, Brönnimann S, Stickler A (2013) Observation errors in early historical upper-air observations. *Journal of Geophysical Research-Atmospheres* **118**: 12012-12028.

Yang G-Y, Methven J, Woolnough S, Hodges K, Hoskins B (1984) Linking African Easterly Wave activity with equatorial waves and the influence of Rossby waves from the Southern Hemisphere. *Journal of the Atmospheric Sciences* **75**: 1783-1809.