Detection and Analysis of an Amplified Warming of the Sahara Desert

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ABSTRACT

Evaluation of three reanalyses (ERA-Interim, NCEP-2, and MERRA) and two observational datasets [CRU and Global Historical Climatology Network (GHCN)] for 1979–2012 demonstrates that the surface temperature of the Sahara Desert has increased at a rate that is 2–4 times greater than that of the tropical-mean temperature over the 34-yr time period. While the response to enhanced greenhouse gas forcing over most of the globe involves the full depth of the atmosphere, with increases in longwave back radiation increasing latent heat fluxes, the dryness of the Sahara surface precludes this response. Changes in the surface heat balance over the Sahara during the analysis period are primarily in the upward and downward longwave fluxes. As a result, the warming is concentrated near the surface, and a desert amplification of the warming occurs. The desert amplification is analogous to the polar amplification of the global warming signal, which is concentrated at the surface, in part, because of the vertical stability of the Arctic atmosphere. Accompanying the amplified surface warming of the Sahara is a strengthening of both the summertime heat low and the African easterly jet and a weakening of the wintertime anticyclone and the low-level Harmattan winds. Potential implications of the desert amplification include decreases in mineral dust aerosols globally, decreases in wintertime cold air surge activity, and increases in Sahel rainfall.

1. Introduction

It is clear from model simulations and observations that the ongoing global warming signal is not, and will not be, distributed evenly. For example, surface warming at high Northern Hemisphere latitudes is amplified because of the ice albedo–temperature feedback and, close to the pole, the vertical stability of the atmosphere (e.g., Manabe and Stouffer 1980; Washington and Meehl 1996; Serreze and Barry 2011).

In the tropics, enhanced longwave back radiation due to increased levels of greenhouse gases increases the latent heat flux from the surface, reducing the surface temperature response and heating the middle and upper troposphere. However, if the surface is relatively dry, enhanced latent cooling of the surface will be inhibited. In this case, which would apply most generally to the subtropics, surface warming can be stronger (Sutton et al. 2007).

The purpose of this paper is to examine observational and reanalysis data to identify and understand the potentially most extreme case of this amplification of the subtropical warming signal: namely, in the Sahara Desert, the largest nonpolar desert on the planet. This investigation is complicated by two factors. One is that other processes may be important to, or even dominate, the desert’s response to greenhouse gas forcing. Possibilities include changes in solar forcing associated with changes in clouds and/or aerosols, changes in precipitation distributions, a modified sensible heat flux, and changes in circulation (e.g., the downbranch of the Hadley circulation and locations of storm tracks). Also of primary importance may be changes in the regional climate near the desert.

The other complicating factor is observational limitations, including uncertainty in the identification of trends driven by greenhouse gas forcing resulting from short record lengths and/or high interannual variability. This uncertainty is especially acute in regions with sparse ground-based observations, such as deserts. To ameliorate this concern, we use multiple datasets, including three reanalyses and two sets of observations, and investigate the physical processes of trends using the constraints of the surface heat budget.

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Background on the Sahara climate, subtropical warming, and the use of reanalysis products for trend detection is presented in the next section, followed by a discussion of methodology in section 3. The evidence for an amplification of surface warming over the Sahara is presented in section 4a, and, in section 4b, the surface heat balance is examined to understand the physical processes of the surface temperature trend. Implications of the amplified Sahara warming are discussed in section 4c. A summary and conclusions are provided in section 5.

2. Background

The Sahara Desert is the largest nonpolar desert on the planet. With an area of approximately $9.2 \times 10^6 \text{ km}^2$, it is more than $10^5 \text{ km}^2$ larger than the contiguous United States and covers about 6\% of the land surface area of the earth. Annual-mean temperatures exceed 302 K across much of the African Sahel and southern Sahara (about 10$^\circ$–25$^\circ$N), extending into the Arabian Peninsula and Somalia. No other extensive region on the planet has such high annual temperatures. These extreme temperatures combine with exceptionally low precipitation rates; north of about 15$^\circ$N, Africa receives less than 100 mm of water each year.

During the boreal winter months, the Sahara is 10–12 K cooler than tropical Africa (Fig. 1a). High geopotential heights extend eastward from the North Atlantic subtropical high, establishing the strong meridional gradients that characterize the Sahel region to the south. The northeasterly low-level flow, known as the Harmattan winds, constitutes the largest source of aerosol dust on the planet.

During the summer months, temperatures exceed 312 K and low geopotential heights cover northern Africa. This thermal low extends across the continent centered near 20$^\circ$N, and it is characterized by the convergence of low-level northerly flow across the Mediterranean Sea and the southwesterly African

![Fig. 1. Climatological ERAI December–February mean (a) surface temperature (K) and (b) 950-hPa geopotential heights (m; shading) and winds (m s$^{-1}$; vector scale shown below). (c),(d) As in (a),(b), but for the June–August mean.](image-url)
monsoon flow. The thermal low is shallow (not shown) and overlain by the Saharan high at 600–700 hPa.

To our knowledge, there have been no previous studies of the contemporary climate change signal in the Sahara Desert, but there are a number of studies that evaluate land–sea warming contrast under increasing atmospheric greenhouse gas levels (e.g., Joshi et al. 2008; Boer 2011; Byrne and O’Gorman 2013). Sutton et al. (2007) evaluate the latitudinal dependence of the ratio of land–sea warming in 20 coupled GCMs forced with greenhouse gas increases. They find that the global land surface warms more than the ocean surface in these models and that the ratio of land–sea warming is greatest in the subtropics.

Joshi and Gregory (2008) note that water vapor–temperature longwave radiative feedbacks over unsaturated land surfaces will be limited by restricted evaporation, and because the source of atmospheric moisture over dry land is horizontal advection in the atmosphere and, therefore, at temperatures cooler than the surface.

Sutton et al. (2007) also find that the enhanced warming over land is simply not a matter of heat capacity differences between land and ocean surfaces; it is not only a transient difference in the models, but also occurs in equilibrium simulations. Many climate feedback processes, such as the ice albedo–temperature feedback, are important and even dominant on regional space scales, and these are generally independent of surface heat capacity. Further evidence supporting the idea that the distribution of heat capacity does not dominate the pattern of regional warming comes from Wu et al. (2012). They examine ocean reanalyses and find that SSTs in the subtropical western boundary currents are increasing at an accelerated rate, 2–3 times that of the global mean, possibly in conjunction with a poleward shift of the currents and an intensification of the anticyclonic wind stress curl.

Three reanalysis products are used here as sources of information about surface temperature trends in the Sahara. Reanalyses are also used to investigate the physical basis of the surface temperature trends through an examination of the surface heat balance. Caution is needed in this approach because it is not clear that reanalyses are accurate and reliable datasets for evaluating decadal-scale trends. One concern is that changes in the observing systems that constrain reanalyses are sometimes discernible by unrealistic shifts in values. Another concern is their application in data-sparse regions when few stations are available, such as the Sahara Desert. Despite these concerns and the need for caution, reanalyses have proved to be useful and generally consistent in providing information about trends in atmospheric temperature (e.g., Simmons et al. 2014), winds and surface wind stress (e.g., Swart and Fyfe 2012), ocean heat storage (Loeb et al. 2012), and ocean circulation (Tillinger and Gordon 2009; Johnston and Gabric 2010); comparisons with direct surface temperature measurements in data-rich regions are encouraging (e.g., Vose et al. 2012).

There are a number of in-depth investigations of the accuracy of the individual surface heat budget terms in various reanalyses. Brunke et al. (2011) evaluated sensible and latent heat fluxes over the oceans in six reanalyses, including ERA-Interim (hereinafter ERAI; Dee et al. 2011), the NASA Modern-Era Retrospective Analysis for Research and Applications (MERRA; Rienecker et al. 2011), and four satellite-derived products. Each dataset relies on the use of algorithms or parameterizations to calculate the turbulent fluxes. Inputs to these algorithms, which are based on the bulk aerodynamic equations, are called the “bulk variables.” Brunke et al. (2011) find that MERRA and the satellite-derived observations produce relatively low biases in the daily mean latent heat flux (all under 2.6 W m$^{-2}$), primarily due to errors in the bulk variables, while errors in ERAI are larger at 15–20 W m$^{-2}$. For the sensible heat flux, MERRA has biases under 1 W m$^{-2}$.

Roberts et al. (2012) evaluate the annual-mean climatology, seasonal variations, and ability to represent extremes for the turbulent heat fluxes in MERRA. They find that the sensible and latent heat fluxes are relatively accurate for typical conditions, but they are too low in regions with strong temperature and moisture gradients. Bosilovich et al. (2011) find that the MERRA solar flux at the surface is about 5% stronger than the satellite-based values compiled by Trenberth et al. (2009), and they show that the averaging period used may affect the global averages as a result of changes in the observing system. There is also a downward trend punctuated by abrupt changes that are related to the analysis increment, which is a correction to the calculated fields that includes influences from both assimilated observations and inadequacies in the model physics. Surface longwave fluxes in MERRA, both the back radiation and the upward flux, are within about 3 W m$^{-2}$ of the satellite-based estimates.

The downward surface flux of shortwave radiation is about 4 W m$^{-2}$ greater in the ERAI climatology for the 1989–2008 period than in the satellite-based estimates of Trenberth et al. (2009), according to the analysis by Berrisford et al. (2011), although it is improved significantly over the previous generation of reanalysis. The longwave back radiation is also larger by about 8 W m$^{-2}$ in the global mean, with most of that error over the oceans. The upward longwave flux is within 2 W m$^{-2}$ of the observed global mean, but this is an average of errors.
on the order of 10 W m$^{-2}$ with different signs over land and ocean. The sensible heat flux from the surface is about 1 W m$^{-2}$ greater than in the satellite estimates over both land and ocean, and the latent flux is also about 8%–10% larger than in the observations. Given uncertainties in the estimates of the turbulent fluxes from direct observations, Balmaseda et al. (2008) found that using ERAI surface fluxes improves predictive skill for tropical SST.

When reanalyses disagree on the magnitudes of variables, it is still possible that they are producing realistic representations of trends. For example, Park et al. (2013) compared the atmospheric water balance over the ocean from 60°S to 60°N in MERRA with satellite-observed and merged observational datasets. While there were large differences in absolute values among all of the datasets, and the water budget was not closed, there was a good correspondence in how the components of the water balance varied on monthly time scales.

In summary, reanalyses provide complete, internally consistent datasets that allow one to analyze physical processes, requiring the use of both assimilated and calculated variables. Here we examine and compare three reanalyses. Confidence is bolstered by agreement between the three reanalyses, a comparison of surface temperature trends with more direct observations, and by a diagnosis that makes physical sense.

3. Methodology

a. Observations and reanalyses

We analyze the 1979–2012 time period because of the numerous contributions from satellite data to the reanalyses that begin in 1979. The reanalyses used are as follows:

- ERAI (Dee et al. 2011) is a global 6-hourly reanalysis that is available at 1.5° resolution from the European Centre for Medium-Range Weather Forecasts (ECMWF).
- NCEP–DOE AMIP-II reanalysis (NCEP-2; Kanamitsu et al. 2002) provides 6-hourly output at 2.5° resolution, and at approximately 1.9° resolution for surface and 2D diagnostic fields. This lower-resolution product has existed for a longer time than ERAI and has, therefore, undergone longer scrutiny. It is used to constrain ocean reanalyses and to provide variables for satellite retrieval algorithms.
- NASA MERRA (Rienecker et al. 2011) provides upper-air fields at 6-hourly intervals, with 2D diagnostic fields that include precipitation and surface fluxes at 1-hourly increments on a 0.5° × 0.66° resolution grid; 3-hourly assimilated and diagnostic fields are available at 1.25°.

We considered the use of a fourth reanalysis; namely, ERA-Interim/Land (Balsamo et al. 2013). This is a global reanalysis for 1979–2010 that is generated using the ERAI and GPCP precipitation to drive a state-of-the-art land surface model. It emphasizes the accuracy and closure of the water balance on the surface. We find that there are large shifts in the surface temperature, perhaps indicating influences from changes in observing systems. For this reason, we did not adopt this dataset for the Sahara analysis.

Surface temperatures (skin temperatures) from the three reanalyses are compared to two observational datasets:

- Climatic Research Unit Time Series, version 3.21 (CRU TS3.21; Mitchell and Jones 2005), includes 0.5°-resolution monthly surface temperature estimates gridded over land from 1901–2012 using more than 4000 land-based weather stations. Antarctica is not included.
- Global Historical Climate Network merged land–ocean analysis (GHHCN; Peterson and Vose 1997; Jones and Moberg 2003) monthly surface temperature (version 3.2.2) is blended with Extended Reconstructed SST, version 3b (ERSST.v3b), over the ocean to create this dataset. Station data are interpolated onto a 5° × 5° grid, and the coverage is approximately 70°S–70°N. Anomalies relative to the 1980–2010 mean are provided.

Station data from GHHCN are also consulted for the few stations with long records in the Sahara Desert. Note that these gridded observations provide 2-m surface air temperature, while, for the reanalysis, skin temperature is used.

Trends in reanalysis and observed values are estimated using a linear least squares fit. Standard deviations and the significance of the results are calculated using the ANOVA F test and lag-1 autocorrelation is accounted for by calculating an effective sample size (e.g., Frankignoul and Hasselmann 1977; Wilks 1995; Santer et al. 2000). It is important to note, however, that a statistically significant result from a reanalysis does not necessarily mean that a trend is confidently identified. A statistically significant but incorrect trend may emerge in reanalysis values when variabiity is unrealistically small or when a large, spurious trend occurs, or by some combination of these effects. Similarly, unrealistically large variability in the reanalysis can obscure the significance of a trend. For this reason, the analysis also relies on an examination of physical processes to evaluate confidence in the results.

In this study, we refer to expert evaluations of the reanalyses, both in the literature and currently ongoing.
(see section 2). Many of these evaluations are conducted by the teams that develop the reanalyses and satellite retrieval methodologies, and it is outside of the scope of this project to repeat this work. Here, the analysis focuses on the identification of physical processes and understanding them in the context of governing equations. Methods used for that analysis are discussed below.

b. Diagnostic methods

The first step in the analysis is to examine trends in the annual- and seasonal-mean surface temperature $T_S$, comparing area averages for the Sahara Desert with tropical and global means and comparing among the three reanalyses listed above.

To better understand surface temperature trends and their reliability in the observations and reanalyses, terms in the annual-mean surface heat balance are calculated from the reanalyses. These terms are not assimilated in the reanalyses, and studies of their reliability are reviewed in section 2.

The heating rate of the surface is given by

$$C \frac{dT_S}{dt} = S_{ABS} + F_{BACK} - F_{UP} - H_S - H_L - D,$$  \hspace{1cm} (1)

where $C$ is the heat capacity of the surface. According to Eq. (1), the earth’s surface is warmed by the absorption of solar radiation $S_{ABS}$ and longwave back radiation $F_{BACK}$ emitted by atmospheric greenhouse gases. It is cooled by the emission of longwave radiation from the surface $F_{UP}$, which is related to $T_S$ by the Stefan–Boltzmann equation:

$$F_{UP} = \varepsilon \sigma T_S^4,$$  \hspace{1cm} (2)

where $\varepsilon$ is the longwave emissivity of the surface, and $\sigma$ is the Stefan–Boltzmann constant. Over the Sahara, the surface broadband (8–13.5 $\mu$m) emissivity is approximately 0.9 (Ogawa and Schmugge 2004). The sensible heat flux from the surface to the atmosphere $H_S$ and evaporation $H_L$ (i.e., the latent heat flux from the surface to the atmosphere) also cool the surface, and these terms are also functions of $T_S$. Heat is redistributed within the surface primarily by downward vertical diffusion and ocean currents $D$.

Each term of Eq. (1) depends on $T_S$, either directly or indirectly or both; each term in the equation has adjusted to changes in the other terms. For this reason, one cannot identify cause and effect from the heat balance equation alone without additional information and/or constraints.

Taking an equilibrium approach to the surface heat balance analysis, Eq. (1) is written

$$[R] = [S_{ABS}] + [F_{BACK}] - [F_{UP}] - [H_S] - [H_L],$$  \hspace{1cm} (3)

where the square brackets indicate the annual mean, and

$$[R] = \left[ C \frac{dT_S}{dt} \right] + [D].$$  \hspace{1cm} (4)

The $[R]$ cannot be calculated from atmospheric reanalysis values, and it is evaluated here as a residual. Note that $[R]$ will also include error and terms, such as the assimilation term used in MERRA (Cullather and Bosilovich 2012).

4. Results

a. Detection of enhanced warming over the Sahara

Annual-mean surface temperatures anomalies over the Sahara Desert (20°–30°N, 10°W–30°E) in ERAI from 1979 through 2012 are plotted in red in Fig. 2a. Also shown are annually averaged global (black), tropical (30°N–30°S; blue), and tropical land (30°N–30°S; green) temperature anomalies. All anomalies are calculated as differences from the 1979–2012 mean. Figures 2b and 2c are surface temperature anomalies from NCEP-2 and MERRA, respectively. Figure 2d displays values from the CRU observations, which are based on station data and, therefore, report values only over land (section 3). Figure 2e shows values from the merged GHCN data, which extends to about 70° in both hemispheres. In each dataset, surface warming is amplified over the Sahara Desert compared with global and tropical averages. Table 1 shows correlations between the different data sources for annual-mean Sahara and tropical averages. In most cases, the correlations exceed 0.85, indicating consistency in the surface temperature trends among the data sources.

For added perspective, we examine the station data for the four long-term reporting stations in and near the Sahara (Fig. 3). For the 1979–2013 averaging period, each station exhibits a smooth, regular warming trend with no sudden shifts in temperature. Prior to 1979, two of the stations (Bilma, Niger, in Fig. 3a and In Amenas, Algeria, in Fig. 3b) have no surface temperature trends, while the other two stations (Tamanrasset, Algeria, in Fig. 3c and Agadez, Niger, in Fig. 3d) exhibit warming throughout their records.

Figures 4a–c show surface temperature time series for the three reanalyses and two observational datasets for the tropical (blue), tropical land (green), and Sahara (red) averaging regions, as in Fig. 2 but smoothed using a 5-yr running mean. Least squares linear fit trends are indicated by the dashed lines. These trends are listed in

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Table 2 for each of the averaging regions shown in Fig. 2 and extratropical regions for comparison. The table includes standard deviations in parentheses, and significance at the 90%, 95%, and 98% levels is indicated.

Surface temperature trends over the Sahara range from 0.84 to 1.71 K (34 yr)\(^{-1}\), with the larger trend in NCEP-2 not emerging as statistically significant; trend magnitudes are similar [1.36 and 1.39 K (34 yr)\(^{-1}\)] and significant at the 90% and 95% level in the CRU and GHCN observations. The warming trend in the Sahara exceeds the global warming trend by a factor of 2 in ERAI and by a factor of 2.5 in MERRA. In NCEP-2, the Sahara warming is greater than the global warming by a factor of 2.9, but neither warming trend emerges as statistically significant. When compared with the tropical warming trend instead of the global-mean trend, the amplification of the warming over the Sahara is greater by factors of 3.2 in ERAI, 4.8 in NCEP-2, and 4 in MERRA. For the GHCN observations, the Sahara
warming trend is 3.3 times the tropical-mean trend. In the reanalyses, the warming trend in the Sahara is comparable to the warming trend over extratropical land where the ice albedo–temperature feedback operates. While all five datasets show an amplified warming trend in the Sahara, there are differences in the distribution of this warming in the Sahara and throughout the African continent. The spatial distribution of the linear trend in surface temperature from 1979 to 2012 is shown in Fig. 5 for the three reanalyses and the two observational datasets. Here, the trend at each grid point (in kelvin per year) is calculated and multiplied by 34 yr. Stippling indicates the statistical significance of the trends using an effective sample size that takes into account lag-1 autocorrelation. In ERAI (Fig. 2a), warming exceeds 1 K over almost all of the Sahara, with largest values and significance in the central Sahara. In NCEP-2 (Fig. 2b), the warming is centered near 20°N, south of the maximum warming in ERAI. It exceeds 3 K in two regions: over northern Niger and northern Sudan. Despite the large magnitude of the NCEP-2 trend, it is not significant at even the 90% level. Ranges and significance in the warming signal across the Sahara are similar to ERAI in MERRA (Fig. 2c) and CRU data (Fig. 2d), with stronger warming tending to be located in the central desert (near 10°E) and to be somewhat more uniform in coverage. The low-resolution GHCN data place a fairly uniform warming trend of 1–1.5 K throughout the Sahara. As an aside, we note that the strong warming trend in the Congo basin and the cooling trend in East Africa in MERRA are associated with the inclusion of NOAA-15 Advanced TIROS Operational Vertical Sounder (ATOVS) observations beginning in 1998 (Rienecker et al. 2011). The inclusion of ATOVS is associated with a reduction
in precipitation over the Congo, enhanced rainfall farther east over the Horn of Africa, and subsequent changes in cloud cover and solar heating of the surface. Because of this inhomogeneity, the equatorial Africa trends in MERRA (Fig. 5c) are not reliable.

Examining the ability of the three reanalyses to represent the present-day observed surface temperature climatology can provide support for confidence in their representation of trends, although an extensive comparison is not the purpose of this paper. Figures 6a–c show climatological (1979–2012) surface temperatures for each reanalysis, and the CRU observations are shown in Fig. 6d. ERAI, NCEP-2, and MERRA (Figs. 6a–c, respectively) are similar to the CRU data in placing the maximum annual-mean temperature between 10° and 20°N. ERAI and MERRA are quite similar to each other over the domain, and about 0.5–1 K warmer overall than the CRU observations. NCEP-2

![Fig. 4. The 5-yr running means of annual-mean surface temperature anomalies (K) averaged over the Sahara Desert (red), the entire tropics (blue), and tropical land (green) for (a) ERAI, (b) NCEP-2, (c) MERRA, and the (d) CRU TS3.21, and (e) GHCN, version 3.2.1 (V3.2.1), observations. The dashed lines represent linear trends (least squares fit) calculated from unsmoothed values.](image-url)
surface temperatures are cooler than the CRU data by 2–5 K over the Sahara, so confidence in the strong Sahara warming in this reanalysis (Figs. 2b and 4b) may be reduced.

The vertical dependence of the warming trend for the three reanalyses is displayed in Fig. 7, averaged over the Sahara (red), the entire tropics (blue), and tropical land (green). Atmospheric temperatures, which are assimilated variables in reanalyses, are warming throughout the lower troposphere. Below 500 hPa, warming rates over the Sahara are significantly larger than warming rates averaged over tropical land and the entire tropics. Above 500 hPa, warming rates are similar over the three averaging regions.

The seasonality of the Sahara warming signal is evaluated and compared with the seasonality of the tropical warming signal at the surface in Figs. 8a and 8b, respectively, which shows trends in surface temperature for each month in the three reanalyses and two observations for the Sahara and tropical averaging regions. ERAI and MERRA produce reduced warming trends in May, June, and November that are not highly significant, and the CRU observations and NCEP-2 do not produce an identifiable seasonal signal over the Sahara. Over the tropical averaging region (Fig. 8b), there is a tendency for the warming trend to be larger during June, July, and August in the reanalyses but not in the GHCN observations.

In summary, analysis of surface and atmospheric temperature trends in three reanalyses and two observational datasets show that the warming trend over the Sahara Desert in the 1979–2012 period is 2–4 times greater than the tropical-mean trend. All five datasets are consistent in the sense that they indicate an amplification of warming over the Sahara. To deepen our understanding of and confidence in the amplified Sahara warming signal, it is essential to explore its physics.

b. Analysis of the enhanced Sahara warming

An examination of trends in the annual-mean, equilibrium surface heat balance [Eq. (3)] is used to better understand the amplified warming signal over the Sahara Desert and add confidence to its detection. Comparisons are made between the Sahara and tropical mean rather than the global mean, because the ice albedo–temperature feedback amplifies the extratropical signal, and this feedback is irrelevant over the Sahara.

Table 3 shows changes [linear trends in W m⁻² (34 yr⁻¹)] in the surface heat budget terms for ERAI, NCEP-2, and MERRA averaged over the entire tropics (30°S–30°N) and over the Sahara (20°N–30°N, 10°W–30°E). Standard deviations are in parentheses, and statistical significance is indicated, with autocorrelation accounted for by using an effective sample size in its assessment. Statistical significance testing is useful here to identify trends in the reanalyses relative to the interannual variability of a particular reanalysis, but this significance does not mean the change is “correct,” because the degree of accuracy of the surface heat budget terms and their variability is not known exactly. As discussed earlier, we also rely on a physical understanding to evaluate confidence.

The reanalyses are in disagreement about the signs and magnitudes of trends in the solar radiation absorbed at the surface, and there is no statistical significance. NCEP-2 produces a small positive trend in the tropical average and a negative trend over the Sahara. In MERRA, there is a large negative trend in the tropical mean with essentially no change over the Sahara. This disagreement occurs because many of the climate features that influence solar fluxes, such as clouds (cloud type, albedo, and altitude), aerosols, and surface albedos, are not assimilated variables in the reanalyses. The shortwave component of the surface heat balance is not directly involved in the hypothesis about the desert amplification, but it has the potential to disrupt and/or compete with the effect.

The longwave back radiation from the atmosphere to the surface $F_{\text{BACK}}$ increases in each reanalysis in both the tropical and Sahara mean, with some statistical significance emerging. Values range from about 1 to 6 W m⁻² in the tropical mean, and from 4 to 11 W m⁻² for the Sahara mean. [See Wang and Dickinson (2013) for a review of observed and reanalysis downward longwave fluxes.] In ERAI and NCEP-2, the increase in $F_{\text{BACK}}$ in the Sahara mean is greater by a factor of 4 or 5 than the tropical mean. 

Table 2. Linear trends [K (34 yr⁻¹)] in annual-mean surface temperature for various regions for 1979–2012. Standard deviations are in parentheses. Statistically significant values at the 90%, 95%, and 98% confidence interval are italicized, boldfaced, and italicized and boldfaced, respectively, after taking autocorrelation into account by calculating an effective sample size.

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<th>Dataset</th>
<th>Sahara</th>
<th>Global</th>
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<th>Tropical Land</th>
<th>Extratropics</th>
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<td>0.43 (0.21)</td>
<td>0.77 (0.21)</td>
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<td>0.43 (0.22)</td>
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<td>—</td>
<td>0.66 (0.17)</td>
<td>—</td>
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<tr>
<td>GHCN</td>
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<td>—</td>
<td>0.42 (0.11)</td>
<td>—</td>
<td>0.39 (0.13)</td>
<td>—</td>
</tr>
</tbody>
</table>

*a Antarctica is not included in the CRU dataset.

*b The GHCN datasets extend to 70° latitude in both hemispheres.
Fig. 5. Trends in annual surface temperature \(\text{[K (34 yr)}^{-1}\) for (a) ERAI, (b) NCEP-2, (c) MERRA, and the (d) CRU and (e) GHCN observational datasets. Values greater than 3 K (34 yr)\(^{-1}\) are contoured at a 1 K (34 yr)\(^{-1}\) interval. White, green, and purple stippling denotes significance at the 90%, 95%, and 99% levels of confidence, respectively, when applying an effective sample size that takes into account lag-1 autocorrelation.
mean, but $F_{\text{BACK}}$ increases are similar over the two averaging regions in MERRA.

Recognizing that these changes in the annual-mean $F_{\text{BACK}}$ in the reanalyses are related to and equilibrated with other components of the heat balance, additional information must be brought to bear to understand the reasons for these changes. One potential explanation for the increase in the longwave back radiation is an increase in the atmospheric greenhouse gas concentration due to human activity. CO$_2$ is the primary anthropogenic greenhouse gas. Prata (2008) used an atmospheric radiative transfer model to estimate that the increase in longwave back radiation at the surface is 0.3 W m$^{-2}$ decade$^{-1}$. For the 34-yr period of this analysis, during which CO$_2$ levels increased from 339 to 393 ppmv, that amounts to an increase of about 1 W m$^{-2}$. This value is smaller than any of the changes in $F_{\text{BACK}}$ listed in Table 3, suggesting that there are additional processes working to increase the longwave back radiation.

An additional possible contributor to the trend in $F_{\text{BACK}}$ is changes in atmospheric water vapor. Water vapor is a greenhouse gas, and, when the surface warms, evaporation increases and loads more water vapor into the atmosphere, at least over wet surfaces. Although it is not possible to generate a simple estimate of the change in $F_{\text{BACK}}$ because of changes in atmospheric water vapor, we examine trends in total column water vapor $W$ over the tropical and desert averaging regions in the three reanalyses (Fig. 9). This value is calculated as the mass-weighted vertical integral of the specific humidity $q$, using

$$W = \frac{1}{\rho_w g} \int q \, dp,$$

where $\rho_w$ is the density of water, $g$ is the acceleration due to gravity, and the vertical integral is over the full depth of the atmosphere in pressure.
In ERAI (Fig. 9a), there is a vanishingly small negative trend in the tropical averaging region and a small negative trend in the Sahara averaging region [-0.39 mm (34 yr)^{-1}]; neither is statistically significant. Both NCEP-2 and MERRA indicate positive trends over both averaging regions. In NCEP-2 (Fig. 9b), which has a large surface warming response over the Sahara (Figs. 5b), the tropical trend of 0.27 mm (34 yr)^{-1} is not statistically significant, while the positive Sahara trend of 1.23 mm (34 yr)^{-1} is significant at the 80% level. Since the low climatological surface temperatures over the Sahara in NCEP-2 (Fig. 6b) are accompanied by relatively high soil moisture values (not shown), the trend in column water vapor in Fig. 9b may not be accurate. MERRA (Fig. 9c) produces a larger water vapor trend over the tropics than over the Sahara: 2.64 and 0.53 mm (34 yr)^{-1}, respectively.

The investigation of column water vapor trends in the three reanalyses is inconclusive. The three datasets are not in close agreement, even in the tropical mean. Evaluations of reanalysis column water vapor values have been conducted over the ocean. Schroeder et al. (2013) find that ERAI is consistent with more direct satellite-based measurements for ocean points. Other studies (e.g., Trenberth et al. 2005) use different satellite-derived datasets and find a global-mean trend in column water vapor over the oceans of approximately 0.44 mm decade^{-1} [1.5 mm (34 yr)^{-1}]. Trenberth et al. (2005) assess the reliability of the different reanalyses and find that NCEP-2 tends to not be very realistic over the ocean in terms of the mean, variability, and trends of water vapor, in part because this product does not assimilate water vapor information from satellites. NCEP-2 is more realistic over land where radiosonde water vapor values are assimilated, but this information provides only weak constraint in remote areas, including the Sahara.

A third possible reason for the increase in the longwave back radiation is an increase in the upward longwave radiation. The lower troposphere and the surface are tightly coupled by the exchange of longwave energy, so an increase in surface temperature and the upward longwave flux will warm the lower atmosphere and lead to an increase in the back radiation and, potentially, an amplified surface warming. Increases in the upward longwave radiative flux, which are proportional to the fourth power of the surface temperature according to the Stefan–Boltzmann relation, are positive in each reanalysis and in both averaging regions (Table 3). Because greenhouse gas absorption lines are not saturated, increases in the upward longwave flux from the surface are associated with increases in the downward flux of longwave radiation.

As an example, Figs. 10a and 10b show annual-mean values of the longwave back radiation from the atmosphere to the surface from ERAI for the tropical and Sahara averaging regions, respectively. Figures 10c and 10d are annual values for the upward longwave, and even a visual inspection shows that the upward flux is highly correlated with the downward flux. As a result of the strong exchange of longwave energy between the surface and atmosphere, trends in the net upward longwave radiation produced in the reanalyses are small in ERAI and NCEP-2. In ERAI, the increase in upward longwave radiation is greater than the increase in the downward longwave radiation in both the tropical and Sahara mean, so there are net longwave cooling trends of 0.45 and 1.32 W m^{-2} (34 yr)^{-1}, respectively (Table 3). The opposite occurs in NCEP-2,
but these trends are small and not significant (in the context of the variability of the reanalyses).

In contrast to ERAI and NCEP-2, the trend in the longwave back radiation in MERRA is much larger than the increase in the upward longwave flux in the tropical mean (Table 3). This is consistent with the large decrease in the solar radiation absorbed in the tropical mean, since large increases in cloudiness would account

TABLE 3. Surface heat budget component trends [W m\(^{-2}\) (34 yr)\(^{-1}\)] and standard deviations (in parentheses) of the annual-mean surface heat balance averaged over the tropics (30°S–30°N) and the Sahara Desert (20°N–30°N, 10°W–30°E). Rows 2, 3, 4, 7, and 8 appear on the rhs of Eq. (3). Statistically significant values at the 90%, 95%, and 98% confidence interval are italicized, boldfaced, and italicized and boldfaced, respectively.

<table>
<thead>
<tr>
<th>Term</th>
<th>Tropics ERAI</th>
<th>Tropics NCEP-2</th>
<th>Tropics MERRA</th>
<th>Sahara ERAI</th>
<th>Sahara NCEP-2</th>
<th>Sahara MERRA</th>
</tr>
</thead>
<tbody>
<tr>
<td>Solar absorbed</td>
<td>−0.47 (1.14)</td>
<td>1.44 (1.01)</td>
<td>−9.51 (1.27)</td>
<td>0.12 (0.80)</td>
<td>−4.24 (1.29)</td>
<td>0.27 (0.76)</td>
</tr>
<tr>
<td>Longwave back</td>
<td>1.13 (1.08)</td>
<td><strong>2.07</strong> (1.12)</td>
<td>6.18 (1.18)</td>
<td>4.26 (2.54)</td>
<td>11.35 (3.39)</td>
<td>5.85 (2.61)</td>
</tr>
<tr>
<td>Upward longwave</td>
<td><strong>1.58</strong> (0.73)</td>
<td><strong>1.70</strong> (0.80)</td>
<td><strong>1.63</strong> (0.83)</td>
<td><strong>5.58</strong> (2.16)</td>
<td>9.81 (2.55)</td>
<td><strong>6.59</strong> (2.07)</td>
</tr>
<tr>
<td>Net upward longwave</td>
<td>0.45 (0.52)</td>
<td>−0.36 (0.54)</td>
<td>−4.55 (0.65)</td>
<td>1.32 (1.17)</td>
<td>−1.54 (1.79)</td>
<td>0.75 (1.24)</td>
</tr>
<tr>
<td>Net radiative heating</td>
<td>−0.92 (0.79)</td>
<td>1.81 (0.68)</td>
<td>−4.96 (0.94)</td>
<td>−1.19 (0.44)</td>
<td>−2.70 (0.77)</td>
<td>−0.48 (0.55)</td>
</tr>
<tr>
<td>Sensible heat</td>
<td>0.11 (0.25)</td>
<td>1.06 (0.64)</td>
<td>−0.84 (0.57)</td>
<td>0.01 (0.45)</td>
<td>−2.68 (0.69)</td>
<td>−0.20 (0.43)</td>
</tr>
<tr>
<td>Latent heat</td>
<td>8.01 (1.28)</td>
<td>11.29 (1.87)</td>
<td>4.22 (1.52)</td>
<td>−0.35 (0.23)</td>
<td>−0.08 (0.70)</td>
<td>−0.37 (0.85)</td>
</tr>
<tr>
<td>Residual</td>
<td>−9.05 (1.85)</td>
<td>−10.54 (2.05)</td>
<td>−8.34 (2.34)</td>
<td><strong>−0.86</strong> (0.23)</td>
<td>0.06 (0.15)</td>
<td>0.09 (0.15)</td>
</tr>
</tbody>
</table>

FIG. 8. The 1979–2012 monthly trends in surface temperature [K (34 yr)\(^{-1}\)] for the (a) Sahara and (b) tropical averaging regions for various datasets. Squares denote values statistically significant at a 90% level of confidence when taking into account autocorrelation using an effective sample size, while diamonds (triangles) denote values statistically significant at the 95% (99%) level of confidence.
for both responses. Association of the change in the net tropical upward flux in MERRA with changes in cloudiness decreases confidence in that trend. Over the Sahara, the increase in the upward longwave flux exceeds the increase in the longwave back radiation in MERRA, leading to a small positive (cooling) trend in the net longwave flux.

Net radiative heating trends \( (S_{\text{ABS}} + F_{\text{BACK}} - F_{\text{UP}}) \) are shown in Table 3. The only positive (surface warming) trend in the net radiative flux occurs in NCEP-2 in the tropical mean. Despite the positive trend in surface temperatures, all three reanalyses produce net radiative cooling trends over the Sahara. The change in the net radiative heating over the Sahara in ERAI is only a little larger than the tropical-mean value at 1.2 W m\(^{-2}\), and it occurs because the trend in the upward longwave flux is greater than the trend in the downward longwave flux. In NCEP-2, the net radiative cooling of the Sahara occurs because of a large decrease in the solar radiation absorbed, which overwhelms the warming trend in the net upward longwave flux. In MERRA, the net radiative cooling trend occurs because the positive upward longwave trend exceeds the positive longwave back radiation trend.

The net radiative heating trends are balanced by changes in the turbulent heat fluxes and the residual \( (H_S + H_L + R) \). For annual-mean values, the surface temperature tendency is small, and the residual \( R \) is primarily reflecting horizontal transport of heat by ocean currents and the vertical diffusion of heat. Over the Sahara in all three reanalyses, there is a negative (cooling) trend in the net radiative heating of the surface. In ERAI, the small cooling trend is balanced by small decreases in the residual and the latent heat flux. In NCEP-2, decreases in the sensible heat flux balance the radiative cooling trend, and, in MERRA, the small negative net radiative trend is balanced by small negative trends in the turbulent heat fluxes.

Despite some differences between the reanalyses in the details of how the surface heat balance over the Sahara has changed over the last 34 years, they agree that the primary adjustment has occurred in the longwave radiation components of the surface heat balance. In contrast, for the tropical mean, changes in the latent heat flux and the residual (heat flux by ocean currents) play a major role (Table 3). In ERAI and NCEP-2, the largest trends are in the latent heat flux term and the residual, representing increases in evaporation and a weakening of heat fluxes by ocean currents and/or vertical diffusion. In MERRA, there are relatively large trends in all components of the surface heat balance except the sensible heat flux. Similar to ERAI and NCEP-2, however, MERRA produces a small trend in

![Figure 9](image-url)
the sensible heat flux, a larger positive trend in the latent heat flux, and a large negative trend in the residual term.

Further diagnosis of these changes in the tropical mean is outside the scope of the current study. They are presented here to show the contrast with the trends in the heat balance over the Sahara to explain the amplified desert warming.

c. Consequences of enhanced warming over the Sahara

The implications of the observed amplification of warming over the Sahara Desert are direct for the estimated three million people living in this region in terms of heat stress and decreased water availability through increased evaporation from oases and reservoirs. Changes in the large-scale circulation in the lower troposphere can bring added impacts. Amplified warming of the Sahara in boreal summer is related to a deepening of the thermal low (see Fig. 1a). Following Lavaysse et al. (2009), the strength of the Saharan heat low is expressed as the 700–925-hPa thickness. Figures 11a–c display the time series of this thickness over the Sahara averaging region for the July–September (JAS) mean in ERAI, NCEP-2, and MERRA, respectively. Consistent with the surface and lower-tropospheric warming, the thickness is increasing at an average rate of 0.31 gpm yr\(^{-1}\) in ERAI, 0.36 gpm yr\(^{-1}\) in NCEP-2, and 0.34 gpm yr\(^{-1}\) in MERRA, with the ERAI and NCEP-2 trends significant at the 90% confidence level and the MERRA trend significant at the 95% confidence level. For the tropical mean (not shown), the rate of increase of the thickness is much smaller: for example, at 0.6 gpm yr\(^{-1}\) in ERAI.

During winter, when relatively high geopotential heights extending eastward from the North Atlantic subtropical high are located over the Sahara (Fig. 1b), lower-tropospheric warming and increased thickness represents a weakening of the high. Figures 11d–f show thickness trends over the Sahara averaging region for the December–February (DJF) mean in ERAI, NCEP-2, and MERRA, respectively. The thickness is increasing at an average rate of 0.24 gpm yr\(^{-1}\) in ERAI, 0.36 gpm yr\(^{-1}\) in NCEP-2, and 0.44 gpm yr\(^{-1}\) in MERRA, with the NCEP-2 and MERRA trends significant at the 90% confidence level and ERAI trend significant at the 80% level.

Amplified warming over the Sahara in summer strengthens the meridional temperature gradient and,
thereby, the African easterly jet (Cook 1999). This relationship is observed on interannual time scales (e.g., Lavaysse et al. 2010). The time series of the JAS zonal wind at 600 hPa in ERAI is shown in Fig. 12a, and in NCEP-2 and MERRA in Figs. 12b and 12c, respectively. The least squares linear fit indicates that the jet is strengthening by 0.01 m s$^{-1}$ each year, which amounts to about 0.4 m s$^{-1}$ over the 34-yr period, in ERAI. In NCEP-2 the jet strengthens by 0.85 m s$^{-1}$ over the 34-yr period and in MERRA by 0.32 m s$^{-1}$. This rate of strengthening is certainly at the edge of accuracy in our observing systems, and it is only statistically significant in NCEP-2 (at the 90% confidence level), but their consistency and physical association with changes in the surface temperature adds confidence and has implications as global warming progresses.

Weakening the wintertime high over the Sahara (Figs. 11d–f) has the potential to modify mineral dust loading into the atmosphere. The Sahara and Sahel are primary sources of mineral dust for the global atmosphere when surface material is suspended by the northeasterly surface winds in winter, often in association with nocturnal
low-level jet formation (Fiedler et al. 2013), that are part of the anticyclonic circulation around the surface high (Fig. 1b). Amplified Sahara warming in winter would weaken this high and the northeasterly flow and reduce the suspension of mineral dust. As seen in Figs. 12d–f, each reanalysis captures a weakening of the surface easterly flow during the 34-yr analysis period. Such a reduction in the African mineral dust source would diminish its fertilizing effects in the Amazon basin, the Caribbean Sea, the equatorial Atlantic Ocean (e.g., Bristow et al. 2010), Europe (e.g., Helmert et al. 2007), and the Mediterranean Sea (e.g., Guieu et al. 2002). There may also be some effects on the global warming signal due to changes in the radiative effects of these aerosols and in reducing the uptake of CO₂ in the Amazon if the reduction in fertilizing micronutrients leads to a reduction in net primary
production. A reduction in this important source of iron for the world’s oceans may also weaken the ocean biological pump and its oceanic uptake of atmospheric CO$_2$ (e.g., Archer and Johnson 2000). In addition, Washington et al. (2009) show that the Bodélé dust source is very sensitive to even minor changes in the atmospheric circulation, and they suggest that this sensitivity could generate abrupt changes in climate.

Weakening of the wintertime anticyclone and northeasterly low-level flow over northern Africa associated with amplified Sahara warming may weaken the frequency and/or intensity of midlatitude cold air surges that propagate equatorward over northern Africa (Vizy and Cook 2009, 2013). These surges have a known drying effect over northern and tropical Africa. Hence, a weakening of the low-level anticyclonic circulation due to amplified warming may result in a reduction in rainfall breaks along the Guinean coast and northern Congo basin during the winter. Vizy and Cook (2013) show a significant positive relationship exists between eastern Saharan low-level temperature variability and rainfall anomalies over tropical Africa during the winter [e.g., see Fig. 10 in Vizy and Cook (2013)].

5. Conclusions

Evaluation of three reanalyses (ERAI, NCEP-2, and MERRA) and two observational datasets (CRU and GHCN) for 1979–2012 demonstrates that the surface temperature of the Sahara Desert has increased at a rate that is 2–4 times greater than that of the tropical-mean temperature over the 34-yr time period. The amplified warming is confined to the lower troposphere, below about 500 hPa, and it is not strongly dependent on season.

An evaluation of the surface heat balance in the three reanalyses is used to better understand the difference in surface temperature trends over the Sahara compared with the tropical mean. While there are significant differences between the reanalyses, some patterns emerge. Averaged over the tropics, the primary trend in the surface heat balance components is in the latent heat flux from the surface and in a residual term. The residual includes heat storage within the surface, assumed to be small, and the effects of redistribution of heat within the surface, primarily as a result of ocean currents. Increases in the latent heat flux cool the surface, and this cooling is balanced by a reduction in the residual term. Such a strong response in the latent heat field distributes heating throughout the vertical column.

Over the Sahara, the perturbation of the surface heat balance is different because the surface is dry and redistribution/storage of heat within the surface is very small. In other words, both mechanisms that contribute to the adjustment of the surface heat balance in the tropical mean are unavailable over the Sahara. Instead, the primary changes in the surface heat balance occur in the longwave fluxes. In each reanalysis, large increases in both the longwave back radiation and the upward longwave radiation occur. Small increases in the longwave back radiation occur because of increased atmospheric CO$_2$. Increases in column water vapor also occur in two of the three reanalyses. But the large increases in the back radiation are primarily associated with increases in the upward longwave radiation in the time-mean (adjusted) surface heat balance. Over the Sahara, the strong longwave radiative coupling between the surface and the atmosphere amplifies the surface warming. With no mechanism for distributing the added back radiation vertically (i.e., latent cooling/evaporation) the desert amplification is analogous to the amplification of the greenhouse gas–induced warming in the vertically stable polar atmosphere.

In winter, preferential warming over the Sahara weakens high surface pressures that are essentially an eastward extension of the North Atlantic subtropical high, and the reanalyses indicate a weakening of the low-level easterly flow. Potential implications include a weakening of the North African source of mineral dust and diminished dry periods over tropical Africa because of a decrease in the frequency and/or intensity of midlatitude cold surges.

In summer, amplified warming over the Sahara strengthens the thermal low in all three reanalyses. The African easterly jet, which is directly related to surface meridional temperature gradients, also strengthens in all three reanalyses.

Changes in Sahara surface temperatures and the heat low are related to rainfall variations over the Sahel on all time scales. A relationship is also found in the context of global warming by Vizy et al. (2013), who analyze 30- and 90-km-resolution regional model simulations for the mid- and late twenty-first century as well as five CMIP5 CGCMs under greenhouse gas increases from the IPCC representative concentration pathway 8.5 (RCP8.5) forcing scenario. The average surface temperature increase in the seven simulations for midcentury over the Sahara averaging region is 2.5 K, representing 50 years of CO$_2$-induced warming. This is 0.05 K yr$^{-1}$, or 1.7 K over a 34-yr period. This value is similar in magnitude to the observed Sahara warming values listed in Table 2 but somewhat higher than the average warming of 1.3 K, as would be expected under future exponential increases in atmospheric CO$_2$.

Figure 13 shows results from Vizy et al. (2013) for their 30-km regional model simulations and from the Community Climate System Model (CCSM) for the...
mid-twenty-first century. A deepening of the thermal low of 3–4 gpm in the regional simulation and 8–9 gpm in the CCSM CGCM are associated with a stronger meridional geopotential height gradient across the central Sahel, enhanced southwesterly flow (and moisture transport), and precipitation increases of about 3 mm day$^{-1}$ across the Sahel that are significant at the 95% confidence level.

This comparison suggests that the well-known recovery in Sahel rainfall since the later 1990s may be associated, in part, with greenhouse gas warming through the desert amplification mechanism, especially in the central Sahel. We imagine that this is one of a suite of influences on the Sahel rainfall recovery; Sahel rainfall is known to be sensitive to global SSTs, and the recent recovery has also been related to such factors as Indian Ocean warming (e.g., Hagos and Cook 2008), the Atlantic multidecadal oscillation (e.g., Mohino et al. 2011), and aerosol forcing (e.g., Ackerley et al. 2011).

The analysis of trends in the adjusted surface temperatures and surface heat balance presented here cannot definitely attribute the amplified Sahara warming to greenhouse gas forcing. In addition, there is no way to quantitatively associate these changes with changes in precipitation: for example, with the increases in West African precipitation in the recent decade. However, the agreement between the RCM simulations, the GCM simulations, and this observational analysis suggests that the amplified warming over the Sahara is a response to increasing atmospheric CO$_2$ levels and that it may be playing a role (among other factors) in the apparent recovery of Sahel rainfall.

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REFERENCES


