The Southern African Heat Low: Structure, Seasonal and Diurnal Variability, and Climatological Trends

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ABSTRACT

Heat lows are key features of subtropical climates and monsoon systems. In southern Africa, they are pivotal to understanding divergent climate change projections, in particular the veracity of future rainfall decline. Compared to other heat lows, including in West Africa and Australia, the southern African heat low remains poorly documented. Here, we analyze the diurnal cycle, seasonal variability and trends of the heat low in reanalysis data. In ERA5, 462 strong heat low days are detected between September and March from 1990-2019, equating to 7.3% of days sampled. These events feature ascent (exceeding -0.2 Pa/s) at low levels (strongest between 800-600 hPa) and subsidence aloft, generating low-level cyclonic flow with anticyclonic flow above. This flow exhibits strong diurnal variability, with peak windspeeds between 0600-0900 UTC and maximum ascent at ~2300 UTC. Heat lows form preferentially over Angola in September (~14ºS) and October (15-20ºS), and in Namibia from November to March (~20-26ºS). Strongest ascent occurs over areas of high elevation. Finally, we show a rapidly increasing frequency of strong heat low days, with a 175% increase between 1960-1989 and 1990-2019. The greatest increase (459%) has occurred in the early summer months of September and October, consistent with projections of delayed rainfall onset. Strikingly, more strong heat lows are detected in the most recent 5 years of analysis (2014-2019) than in the 30-year period from 1960-1989. These results suggest the heat low is an important feature in determining drying trends over southern Africa and is a vital indicator of climate model accuracy.

SIGNIFICANCE STATEMENT

This work documents the heat low that forms in southern Africa in the lowest levels of the atmosphere. The feature is present during austral summer (from September to March) and is associated with below average rainfall across much of the subcontinent. The frequency of strong heat lows has rapidly increased in line with regional amplified warming trends. The heat low is identified as an important control on circulation and precipitation patterns and changes in the frequency or intensity of the feature in the future are likely to influence the strength of declining rainfall trends across southern Africa.

1. Introduction

Southern Africa is a region with a particularly unique and complex climate owing to a combination of tropical and mid-latitude influences, contrasting sea-surface temperatures on
the east and west coasts and a high interior plateau (Tyson and Preston-Whyte, 2000). Considerable gaps remain in our knowledge of the southern African climate system, hindering prediction relating to interannual variability and trends associated with anthropogenic climate change. As an arid region with high exposure to climate across many economic sectors, southern Africa is particularly vulnerable to climate change (Conway et al., 2015). Annual precipitation rates are projected to decrease across the region throughout the 21st Century owing to a delayed onset of the summer rainy season (Dunning et al., 2018; Ranasinghe et al., 2021). Despite intermodel agreement in the future drying trend, there are considerable differences in the projected magnitude of change between models (Munday et al., 2019). In addition, many models overestimate current rainfall – some by as much as 300% – casting doubt on their ability to simulate future precipitation accurately (Munday and Washington, 2018). Improving our understanding of contemporary climate dynamics at a regional scale is therefore essential, not only to improve forecasting of inter-annual and inter-seasonal variability but also to better constrain models and evaluate future projections (James et al., 2018).

A key feature of southern African climate that remains understudied is the heat low that forms during austral summer due to intense solar heating, initially as the Angola (Heat) Low and subsequently as the Kalahari Heat Low, as it shifts southward to the Kalahari Desert around 25ºS (Vizy and Cook, 2016; Munday and Washington, 2017; Howard and Washington, 2018). The heat low in southern Africa exerts a strong influence on regional rainfall and the accuracy of model rainfall estimates (Munday and Washington, 2017). Furthermore, the Angola and Kalahari heat lows are expected to strengthen due to anthropogenic climate change, which will affect millions of people, particularly via effects on agriculture (Vizy et al., 2015; Vizy and Cook, 2016).

Considering the importance of the regional heat low for the current and future climate of southern Africa, it is essential that the feature and its seasonal evolution are understood, yet no paper to date has provided a full climatology of the feature. In this study, we address the following questions:

1. How does the location and structure of the heat low vary through the annual cycle?
2. How does the heat low evolve through the diurnal cycle?
3. Are there trends in the strength of heat low events?
Section 2 summarizes the understanding of heat lows gained from theoretical work and observed data. Section 3 details the data and metric used to detect heat lows and Section 4 describes how this was used to detect strong heat low days and identifies the mean structure of the feature during these cases. Section 5 deals with the seasonal variability of the heat low, and Section 6 discusses the diurnal cycle of strong heat lows. Section 7 documents trends in the strongest heat lows. The final section provides a summary and conclusions.

2. Heat Lows and Heat Low Circulation

a. Heat Lows

Heat lows occur in the lower troposphere of arid regions, particularly during summer months, where intense solar heating and low soil moisture levels initiate dry convection (Spengler and Smith, 2008), creating an area of low surface pressure that drives a cyclonic circulation near the surface. Ascent is capped by subsiding air above at approximately 600 hPa, where divergence generates an anticyclone. The rhythm of this circulation is set by the diurnal cycle of solar radiation; greatest ascent occurs after the afternoon surface heating maximum (Rácz and Smith, 1999). This diurnal cycle can dictate low-level winds, rainfall, heatwave occurrence and monsoonal inflow (Hoinka and Castro, 2003).

Numerical models can reproduce the circulation patterns associated with heat lows and have demonstrated the importance of topography, sea breezes and low-level jet inflows in their formation (Reichmann and Smith, 2003). Model simulations exhibit a nocturnal peak in low-level winds and a corresponding overnight maximum in cyclonic vorticity (Rácz and Smith, 1999). Daytime turbulent mixing in the boundary layer weakens the cyclonic flow. The absence of this inflow allows surface pressure to decrease throughout the day until surface heating reduces and the flow towards the low-pressure center of the heat low can resume, reducing the pressure gradient (Zängl and Chico, 2006; Spengler and Smith, 2008). Model experiments have also demonstrated that heat lows often occur in areas of high orography where heating is concentrated over a shallower column of air (Zängl and Chico, 2006; Smith and Spengler, 2011).

Model experiments have been supported by studies of heat lows using both reanalysis and observational datasets, such as for the Iberian Peninsula (e.g., Hoinka and Castro, 2003), the Arabian Peninsula (e.g., Blake et al., 1983; Smith, 1986), Australia (e.g., Lavender, 2017), North America (e.g., Rowson and Colluci, 1992) and West Africa (e.g., Parker et al., 2005; Accepted for publication in Journal of Climate. DOI 10.1175/JCLI-D-23-0522.1.)

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Comparatively few studies have focused on southern Africa, where the heat low has, however, been identified as an important feature (e.g., Vizy and Cook, 2016; Munday and Washington, 2017).

**b. Heat Lows in Southern Africa**

The southern African heat low initially forms in the plateau region of southern Angola as the Angola Low (Vizy and Cook, 2016). The Angola Low is only a heat low in the early summer season (September to November); from December the low shifts from dry to moist convection (Howard and Washington, 2018), becoming a tropical low as the heat low migrates southward to become the Kalahari Heat Low in the core summer months (December to February) (Munday and Washington, 2017). The tropical low and heat low forms of the Angola Low can be distinguished via their vertical profiles; the deeper, tropical Angola low features negative relative vorticity from the surface to 400 hPa, whereas negative relative vorticity values associated with the heat low do not extend above 600 hPa (Fig. 3 in Howard and Washington, 2018; Crétat et al., 2019; Pascale et al., 2019). At low levels, the heat low (tropical low) features lower (higher) specific humidity and a lower (higher) potential temperature lapse rate (Howard and Washington, 2018). The heat low is embedded in the regional atmospheric circulation; the pressure gradient between the South Atlantic and Indian Ocean anticyclones and the heat low drives moisture flux and low-level winds across the subcontinent (Cook et al., 2004; Munday and Washington, 2019) particularly via low-level jet inflow from the east through the Limpopo and Zambezi valleys (Spavins-Hicks et al., 2021).

Howard and Washington (2019) show that the southern African heat low is linked to the presence of a dryline known as the Congo Air Boundary (CAB). The CAB marks the meeting point of moist northerly winds and dry southeasterlies in a convergence line feature which is a key determinant of rainfall onset over southern Africa (Howard and Washington, 2019). The CAB lies to the north of the heat low and sets up the cloud-free skies to the south required for strong solar heating and heat low formation. As the season progresses the dryline shifts to the Kalahari Desert, forming to the east of the heat low (Howard and Washington, 2019; Van Schalkwyk et al. 2022). The drylines are themselves maintained by the strong temperature and moisture gradients generated by the heat low circulation (Howard and Washington, 2019).
The formation of the heat low is also linked to the African Easterly Jet South, which is driven by the anticyclone (the Botswana High) that forms between heat-low-induced ascent and upper-level subsidence from 600 - 500 hPa (Adebiyi and Zuidema, 2016; Reason, 2016; Vizy and Cook, 2016; Kuete et al., 2020). Subsidence associated with the Botswana High generates the clear skies required for intense surface heating, thus strengthening the heat low, yet the heat low simultaneously strengthens the Botswana High by providing inflow via ascent in the lower troposphere (Kuete et al., 2023). The heat low is thus an important control on interannual rainfall and temperature variability and the onset of the summer rainy season over southern Africa (Mulenga, 1999; Vizy et al., 2015).

Studies have indicated a strengthening of the heat low in recent decades due to amplified surface warming trends over the Kalahari region (Vizy et al., 2015; Vizy and Cook, 2016). Cook and Vizy (2013) and Engelbrecht et al. (2009) show that the heat low circulation strengthens in response to increased greenhouse gas concentrations in two separate regional climate model simulations. Potential changes in the heat low have been posed as a mechanism of late rainfall onset over southern Africa in future projections (Dunning et al., 2018). The capacity of climate models to accurately resolve the heat low is linked to model ability to simulate current climate over the region. The majority of coupled climate models simulate a large positive precipitation bias (Munday and Washington, 2017). The tropical low form of the Angola Low in late summer has been shown to influence model precipitation biases (Reason and Jagadheesha, 2005; Dieppois et al., 2015; Munday and Washington, 2017). As a stronger southern African heat low has been linked to suppressed rainfall across the subcontinent (Vizy et al., 2015), it is possible that models with anomalously weak or absent heat lows exhibit the greatest positive precipitation biases. Assessing the structure, diurnal cycle and seasonal evolution of the regional heat low are hence important steps from which to progress understanding of both current and future climate over southern Africa.

3. Data

This study uses hourly ECMWF Reanalysis v5 (ERA-5) data (with a horizontal resolution of 0.28125 degrees) to investigate the structure and variability of the heat low over southern Africa, focusing primarily on 1990 to 2019 and extending to 1960 to assess observed trends in Section 7. ERA-5 has a high spatial resolution (a native resolution of approximately 31 x 31 km) across 137 vertical levels (Hersbach et al., 2020). The dataset performs well over Africa, with greatly reduced biases in temperature and precipitation compared to its...
predecessor, ERA-interim (Gleixner et al., 2020; Steinkopf and Engelbrecht, 2022; Gbode et al., 2023). Despite this, the accuracy of ERA-5 over areas of southern Africa where station data is limited is less well known and tropospheric values in particular are hard to evaluate due to a lack of upper air sounding data (Hersbach et al., 2020; Roffe and van der Walt, 2023). Additionally, care must be taken when analyzing trends in reanalysis datasets due to potential inhomogeneities introduced by changes in the observational network and data assimilation. In acknowledgement of these limitations, we test the robustness of key results by comparison with Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2) reanalysis data at a horizontal resolution of 0.5° × 0.625°, which is available from 1980 to present at three-hourly timesteps (Gelaro et al., 2017).

Following previous studies, primary heat low detection is performed using Low-Level Atmospheric Thickness (LLAT) – a metric used to determine the dilation of the lower levels of the troposphere (Lavaysse et al., 2009). LLAT is taken here as the difference in geopotential heights at 700 hPa and 850 hPa. A lower bound of 950 hPa falls below the surface level across much of the southern African plateau.

Trends in heat low frequency are assessed in two ways: 1) the Wilcoxon Signed Rank test for difference between the periods 1960-1990 and 1990-2020 and 2) Poisson regression for the timeseries of strong heat low days.

4. Heat Low Detection

A 30-year time-series of austral summer (September to March) LLAT was generated across southwest Africa (10-30ºS, 12-24ºE) at 1500 Coordinated Universal Time (UTC). Based on our analysis of the diurnal cycle, the 1500 UTC timestep was chosen to coincide with the afternoon LLAT maximum (demonstrated in Section 5). Days where the spatial mean 1500 UTC LLAT was above the 90th percentile of 1990-2019 values were added to an initial dataset of strong heat low days. This absolute rather than relative monthly detection threshold was applied to enable a comparison of the strength of the heat low throughout the season. As there is no absolute threshold for determining the presence of a heat low, this stringent threshold enables an analysis of the structure and variability of particularly strong and well-defined heat low events. Similar thresholds have been employed for other heat lows, such as in West Africa and Australia (Lavaysse et al., 2009; Lavender et al., 2017). Events exceeding the 90th percentile of 1500 UTC LLAT were included in the final sample if the
grid point of maximum LLAT met the additional criteria of featuring subsidence in the daily mean at 300 hPa to remove events associated with deep moist convection. Events were also removed if the LLAT maximum occurred outside the core study area, where the Angola and Kalahari heat lows are known to occur (10-30°S, 12-24°E). Monthly composite vertical profiles of potential temperature and vertical velocity were assessed to ensure the characteristics were not contingent on the chosen detection scheme.

These criteria yielded a total of 462 strong heat low days (7.3% of all days sampled), with the highest number detected in October (114) followed by January (89), November (75), February (57), December (55), September (37) and with fewest detected in March (35). The spatial distribution of events demonstrates the transition from the Angola (Heat) Low in the early season (September and October) to the Kalahari Heat Low, following the seasonal shift of the solar radiation maximum (Fig. 1). However, the LLAT maxima are clustered with the heat low favoring two broad positions, rather than undergoing a gradual southward progression. LLAT maxima are clustered in areas of high elevation, such as the Huila and Bié plateaus in Angola and Khomas Highland in Namibia, indicating the role of topography, which is discussed further in Section 5. Due to the high elevation of the Namibian plateaus, when defined by maximum thickness, most strong heat lows that form in the late season (as the Kalahari Heat Low) are centered to the northwest of the Kalahari Desert. The heat low detection scheme was also applied to MERRA-2, which yields a similar spatial distribution of events, although slight differences in the location of the LLAT maxima are inevitable in part due to the higher resolution of ERA-5 (Supplementary Fig. 1).
Fig. 1. The location of detected strong heat lows in southwest Africa in ERA-5 from September to March 1990-2019, defined by the grid point of maximum low-level atmospheric thickness (LLAT). The size of the marker corresponds to the number of heat lows at that grid point.

Latitude-height composite profiles of potential temperature, vertical velocity and zonal and meridional winds point to a coherent structure of the lower atmosphere for strong heat low days across the annual cycle (Fig. 2). High surface potential temperature values indicate clear skies and high surface heating in the heat low region. Little change in potential temperature with height at the center of the heat low indicates the central region of instability where the atmosphere is well mixed due to dry convection; at 1200 UTC the 317 K isoline extends from 650 hPa to the surface (Fig. 2). This corresponds to negative vertical velocity values of -0.2 Pa s\(^{-1}\) from 700-600 hPa at the center of the heat low, denoting ascent at low levels. Weaker values extend up to 500 hPa, above which the heat low is capped by subsidence. Zonal winds highlight the mid-level African Easterly Jet South with mean windspeeds exceeding 7 m/s. Westerly flow associated with the westerly subtropical jet dominates towards the south. Meridional winds show cyclonic converging flow below 700 hPa and divergent, anticyclonic flow from 700-500 hPa (~2-3 m/s). The composites capture a classic heat low structure.
Fig. 2. Composite daily mean latitude-height vertical profiles of strong heat low days in ERA-5 from September to March for potential temperature (K), vertical velocity (Pa s$^{-1}$), zonal wind (m s$^{-1}$) and meridional wind (m s$^{-1}$) during detected heat low days. Profiles are averaged over 5º and centered over the low-level atmospheric thickness (LLAT) maximum to account for the seasonal movement of the heat low. Negative (positive) latitude values represent degrees north (south) from the heat low center. Negative (positive) longitude values represent degrees west (east) from the heat low center.
5. The Seasonal Cycle of Low-Level Atmospheric Thickness and Heat Low Presence

This section considers the annual cycle of the heat low. At 1500 UTC, LLAT over southwest Africa reaches its maximum (~1,660 meters) between late September and early March, declining to a climatological minimum of 1,620 meters in austral winter (Fig. 3). Both ERA-5 and MERRA-2 demonstrate similar annual cycles, although ERA-5 exhibits a slightly lower LLAT during austral summer. The climatological LLAT is nonetheless similar between the two reanalyses at 1656.6 meters and 1658.9 meters in ERA-5 and MERRA-2 respectively. Average LLAT remains somewhat constant within the study period of September to March, although the LLAT maximum (defined as values exceeding the 99th percentile) moves southwards by 10º latitude (Fig. 4). In September, the LLAT maximum is situated above the Angolan highlands (at approximately 14ºS), indicating the dominance of the Angola (Heat) Low in the early season. October marks the point of transition when the LLAT maximum is found between Angola and Namibia. By November, the LLAT maximum is situated across western-central Namibia, where it remains until the end of the summer season in April (22-24ºS).

Fig. 3. The annual cycle of low-level atmospheric thickness (LLAT) at 1500 UTC for southwest Africa (10-30ºS, 12-24ºE) in ERA-5 and MERRA-2. The red and blue lines denote the multiyear mean from 1990-2019 of ERA-5 and MERRA-2 respectively and individual lines show each year for ERA-5 only.
Fig. 4. The annual cycle of low-level atmospheric thickness (LLAT) in ERA-5 at 1500 UTC averaged from 1990-2019. Dashed contours denote the 95th percentile and solid contours denote the 99th percentile of LLAT relative to each month.
The relationship between topography and heat low location was tested by comparing ERA-5 surface elevation for all land grid points across detected heat low days (n = 3,338) and a subset of grid points with strong ascent in the seasonal mean (September to March). Grid points of strong ascent are defined by integrated 800-600 hPa vertical velocity values below -0.15 Pa s\(^{-1}\) (n = 205). A left-tailed z-test reveals a statistically significant difference in surface elevation at grid points with strong ascent, demonstrating that high elevation is associated with strong overlying ascent (z-score = 19.36, p-value < 0.0001). Strong ascent in the seasonal mean occurs above land that is an average of 267 meters higher in elevation than the average for southwest Africa (1110 m). Dry convection is enhanced above areas of high elevation, where surface heating induces strong heating of the comparatively shallow atmospheric column (Smith and Spengler, 2011).

Fig. 5 shows the seasonal shift in vertical velocity during strong heat low days. In September and October, low-level ascent (below 600 hPa) is favored equatorward of 16°S, above which there is subsiding air. As the season progresses, low-level ascent favors more southerly latitudes, with a northern limit of 14°S in November, shifting to 17°S by February. This southward movement is accompanied by a reduction in the latitudinal extent of the heat low; a region of deep ascent encroaches from the north as the summer progresses, visible throughout the vertical extent of the troposphere and reaching a southern limit of ~16°S in January and February. This deep layer of ascent is associated with increasing specific humidity – the 5.0 g kg\(^{-1}\) specific humidity contour at 600 hPa moves by over 10° latitude throughout the season – indicating tropical moist convection associated with the southward movement and subsequent breakdown of the CAB. Potential temperature contours further support this conclusion; in the heat low region, weak vertical gradients of potential temperature indicate low static stability (Fig. 5). In addition, high surface potential temperatures point to strong surface heating due to clear skies and thus dry, rather than moist, convection. The area with the weakest vertical potential temperature gradient tracks southward through the season and, at lower latitudes, gradients of potential temperature increase with growing deep ascent, indicating the release of latent heat associated with moist convection. Figure 5 thus captures the seasonal cycle of the heat low whereby dry convection shifts from the Angola (Heat) Low to the Kalahari Heat Low as tropical moist convection becomes dominant at low latitudes from late October.
Fig. 5. Average daily mean latitude-height composites of vertical velocity (Pa s$^{-1}$), specific humidity (g kg$^{-1}$; solid black lines) and potential temperature (K; dotted blue lines) during strong heat low days ($n = 462$) in ERA-5, averaged from 12-24°E. Red shades indicate negative (upward) vertical velocity. Black shading indicates averaged topography.

In alignment with this latitudinal shift, the zone of greatest convergence moves southward and the low-level winds providing the inflow shift from being dominantly associated with the Zambezi low-level jet (September to November) to the Limpopo low-level jet (December to
March) (not shown). Portela and Castro (1996) similarly demonstrate that surface winds flowing into the Iberian heat low tend to follow major river valleys and the importance of the heat low in setting up the subcontinental pressure gradient for low-level jet flow has been noted for southern Africa (Spavins-Hicks et al., 2021; Van Schalkwyk et al. 2022). These studies highlight the importance of the heat low for low-level winds across the entire subcontinent.

A striking feature of strong heat low events is that, across all months, they are associated with below average precipitation across much of southern Africa, particularly over southern Angola, Namibia, Botswana, South Africa and southern Zambia (Fig. 6). The average daily negative precipitation anomaly increases each month as the mean climatological rainfall increases. From December, the northern extent of negative precipitation anomalies follows the southwest-northeast alignment of the CAB (Howard and Washington, 2019). In January and February, positive rainfall anomalies are visible north of the expected boundary, with anomalously high precipitation on heat low days in northern Angola (January) and the southern Congo basin (February). This dipole pattern suggests that heat lows are linked to the CAB, whereby dry convection south of approximately 14ºS associated with the heat low prevents equatorial moisture from moving south (Kuete et al., 2020). In the late rainy season months of January and February, the dipole anomaly thus reflects an anomalously late-season CAB structure. The earlier summer months (September-November) do not exhibit this dipole anomaly as the CAB is present on most days during this period (Howard and Washington, 2019).
Fig. 6. ERA-5 daily average precipitation anomalies (mm day$^{-1}$) during heat low days from September to March, relative to the 1990-2019 climatology. Stippling denotes areas where the precipitation anomaly exceeds the 0.05 significance level using a Mann Whitney U test and where the monthly average precipitation exceeds 30 mm.

6. The Diurnal Cycle of Strong Heat Lows

Average LLAT across southwest Africa undergoes a diurnal oscillation of approximately 12 meters (Fig. 7). Maximum LLAT occurs in the afternoon (1500 UTC; 1700 local time) in response to daytime surface heating and the minimum is reached in the early morning (0700...
UTC; 0900 local time). Diurnal characteristics are consistent across all months studied, albeit with varying magnitudes of the daytime maximum in agreement with the seasonal cycle outlined in Section 5. Composites of strong heat lows show that winds in the low-level cyclone peak in the morning between 0500-0900 UTC, whereas upper-level winds exhibit less variability (Fig. 8). Low-level winds weaken during the day following increased boundary layer turbulence (Parker et al., 2005). Low-level moisture flux also peaks overnight, demonstrating the role of the nocturnal low-level jets in bringing moisture from the southwest Indian Ocean (not shown) (Spavins-Hicks et al., 2021).

![Figure 7: Average hourly ERA-5 low-level atmospheric thickness (LLAT) across southwest Africa (10-30°S, 12-24°E) from 1990-2019.](image)

The afternoon peak in LLAT coincides with the steepest potential temperature gradient where values of 317 K extend below 850 hPa, indicating strong vertical mixing associated with daytime turbulence (Fig. 8). Vertical velocity also exhibits a strong diurnal cycle at low levels; subsidence occurs throughout the troposphere in the morning hours and is replaced by ascent up to 500 hPa during the afternoon, which reaches its maximum strength overnight at

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~2300 UTC (0100 local time), in agreement with Howard and Washington (2019). The diurnal cycle of ascent is uniform across all months of analysis and is consistent between ERA5 and MERRA-2 (Supplementary Fig. 2). Streamlines show that low-level ascent begins in the early afternoon, with upward motion extending up to 400 hPa at 1500 UTC (1700 local time) but persisting at lower levels until the early morning (Fig. 9). Comparable strong evening ascent in the heat low region of northern Australia was documented by Hart (1990), and similarly Huaman et al. (2023) found that the ascent maximum of the West African Heat Low occurs around 0000 UTC in ERA-5 reanalysis. It is not immediately clear what drives the strong nocturnal ascent. One hypothesis is that nocturnal ascent occurs due to the convergence of inflow from the Zambezi and Limpopo low-level jets (for example, as documented by Spavins-Hicks et al. (2021)). However, since the maximum in convergence in the early morning occurs when subsidence dominates the lower-troposphere (Fig. 8), other processes may be important.

The lack of strong daytime ascent is surprising. Typically, model studies and observations of heat lows indicate that strong ascent should occur throughout the day (e.g., Blake et al., 1983; Gaertner et al., 1993; Peyrillé and Lafore, 2007). As noted by Howard and Washington (2019), the diurnal cycle of ascent within the heat low occurs out of phase with the diurnal cycle of potential temperature. Little change in potential temperature with height in the afternoon – as seen at 1200 UTC when only 2 K separates potential temperature values at 600 hPa and the surface (Fig. 8) – indicates the atmosphere is at its least statically stable due to turbulent mixing, and thus strong ascent is expected at this time (Rácz and Smith, 1999). Howard and Washington (2019) suggest that the absence of this strong daytime ascent may result from turbulence parameterization in reanalysis; at a horizontal resolution of 30 km, ERA-5 is unable to explicitly represent sub-grid scale turbulent fluxes which account for daytime ascent in reality (Peyrillé and Lafore, 2007). Likewise, MERRA-2 has a spatial resolution of approximately 50km. Following this argument, ascent in reanalyses is simulated at night when vertical velocity is a product of the large-scale flow but the ascent that occurs as the sum of daytime turbulent fluxes may not be captured. Observational data would be invaluable in assessing the accuracy of the diurnal cycle in reanalysis and establishing whether strong daytime ascent is also absent in reality.
Fig. 8. Cross sections of ERA-5 vertical velocity (Pa s\(^{-1}\)) and contours of potential temperature (K) (left), meridional winds (m s\(^{-1}\)) (center) and zonal winds (m s\(^{-1}\)) (right) during detected heat low events from September to March, at 6-hourly (UTC) timesteps. Data is centered over the low-level atmospheric thickness (LLAT) maximum of each event, averaged over 5º longitude/latitude to account for the seasonal movement of the heat low. Green/blue shades of meridional wind (center) indicate positive values (northward flow). Green shades of zonal wind (right) indicate positive (westerly) flow. Negative (positive) latitude values represent degrees north (south) from the heat low center. Negative (positive) longitude values represent degrees west (east) from the heat low center.
Fig. 9. Average streamlines \((v; w \text{ (Pa s}^{-1}) \times 10^3)\) across all heat low days in ERA-5 from September to March, averaged over 5º latitude centered over the low-level atmospheric thickness (LLAT) maximum of each event to account for the seasonal movement of the heat low. Negative (positive) longitude values represent degrees west (east) from the heat low center.


Section 4 defined 462 strong heat low days from 1990-2019. Here, we analyze trends in the frequency of strong heat low days in the extended period of 1960-2019 using hourly
ERA-5 data. Using the same thickness threshold, we find only 168 strong heat low days between 1960 and 1989. This represents a significant 175% increase between the 1960-1989 and 1990-2019 study periods (p < 0.0001). Fig. 10 depicts the increasing frequency of the strongest heat lows and the corresponding temperature and geopotential height trends. Fig. 10b shows a high increase in heat low frequency in the most recent decade; more heat low days were detected in the most recent 5-year period (2015-2019) than in the 30-year period from 1960-1989. A Poisson regression on the annual frequency of strong heat lows from 1960-2019 indicates that approximately 42% of the variability in strong heat low frequency can be explained by the temporal trend.

To confirm the robustness of this result, we analyze trends in the MERRA2 data, which is available from 1980. We find a similarly strong, and significant trend (Supplementary Fig. 3). In MERRA2 the increase in the frequency of strong heat lows from 1980-1999 compared to 2000-2019 is 73.3%. The increase in ERA-5 between the same time periods is 74.9%. The similarity in results between the two reanalysis datasets supports the reliability of the trend irrespective of model structure and changes in the observations assimilated into both products. The increasing frequency of strong heat lows in both datasets supports the findings of Vizy et al. (2015) who document a strengthening heat low in southern Africa from 1982 to 2013 in multiple reanalysis products. Lavaysse et al. (2016) and Fonseca et al. (2022) similarly detect increases in the Saharan Heat Low and Arabian Heat Low, respectively, in reanalysis data.
Fig. 10. ERA-5 trends in (a) annual average September to March 850-hPa temperature (K) and 400-hPa geopotential height (m) from 1960-2019 across southwest Africa (10-30°S, 12-24°E) and (b) the detected number of strong heat lows from 1960-2019, using the initial detection criteria (detailed in Section 4), for each month of the study period.

Daily time-series of temperature at 850 hPa and geopotential height at 400 hPa in ERA-5 (in addition to 300 hPa and 500 hPa geopotential height, not shown) for the same period depict a similar trend (Fig. 10a). Both of these trends are determined to be statistically significant at the 0.05 significance level via using a Mann Kendall test (in both cases, $p < 1 \times 10^{-7}$). The positive lower tropospheric temperature trend of approximately 2°C from 1960 to 2019 implies a strengthening heat low circulation. The trend in 400 hPa geopotential height...
(an increase of approximately 40 meters since 1960) indicates tropospheric expansion in line with this increased heating. Correspondingly, the average LLAT at 1500 UTC across southwest Africa has increased in all months between 1960-1989 and 1990-2019 (Fig. 11a). Despite increases in LLAT, no shift in the location of monthly maximum LLAT was detected, further demonstrating the controlling influence of topography on constraining heat low location.

The greatest trend in LLAT is attributable to the early season (September and October), where the mean increase is approaching the same magnitude as the diurnal cycle (Fig. 11b). This corresponds to an increase in the relative contribution of detected September and October heat low days to the seasonal total – by 7.4% and 8.9% respectively from 1960-1989 to 1990-2019 – and an absolute increase in the number of strong heat low days during these two months of over a factor of 5 (Fig. 10). Evidence of early season strengthening of the heat low is supported by the results of Moses et al. (2023) who identify a warming and drying trend due to a strengthening of the Botswana High in October and November in ERA-5. LLAT increases are also noteworthy in February and March, for which the number of strong heat low days detected has increased by almost a factor of 3, suggesting an overall lengthening of the season in which strong heat lows occur (Fig. 10; Fig. 11). As heat low events are associated with largely negative precipitation anomalies (Fig. 6), these results support projections of shortening rainy season length over southern Africa and indicate that the heat low circulation may be an important factor in the dynamics of this trend (Engelbrecht et al., 2009; Cook and Vizy, 2013; Dunning et al., 2018).

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Fig. 11. The difference in 1960-1989 and 1990-2019 ERA-5 low-level atmospheric thickness (LLAT) at 1500 UTC for (a) the spatial mean across southwest Africa (10-30°S, 12-24°E) throughout the annual cycle and (b) the monthly mean spatial change.
8. Summary and Conclusions

This study provides the first detailed account of the characteristics, seasonal variation and diurnal variability of the southern African heat low. Using ERA-5 and MERRA-2 reanalyses, the heat low is shown to be an important feature of the summertime southern African climate, present in the climatology between September and March. The heat low initially forms over Angola, shifting southwards to Namibia in the core summer months as moist convection in the rainbelt moves south from the Congo Basin (Fig. 5). Analyzing the strongest heat lows yields a daily mean vertical structure common to theoretical and observation-based studies of heat lows: ascent occurs at low levels, capped by subsiding air above (Fig. 2). As this ascent is associated with dry convection, clear skies and intense surface heating, strong heat lows are associated with negative precipitation anomalies across much of southern and central Africa (Fig. 6). Ascent is driven by converging low-level winds that generate a cyclonic circulation below 700 hPa. Above the layer of ascent, outflow generates an anticyclonic circulation. At its diurnal maximum, ascent occurs throughout much of the troposphere (above 500 hPa), refuting the classification of heat lows as only shallow features confined to levels below 600 hPa. This reflects both the high elevation of the southern African plateau and the strength of the circulation during strong heat low days.

Strongest ascent occurs over areas of high elevation, demonstrating the role of orography in constraining the heat low location, as has been noted in model-based studies (Zängl and Chico, 2006; Smith and Spengler, 2011). Poor resolution of topography may thus explain why low-resolution climate models often fail to reproduce the heat low circulation and generate positive rainfall biases across southern Africa (Munday and Washington, 2018). Our results support the conclusion that the ability of models to simulate the location and structure of the heat low may be indicative of their ability to accurately reproduce the rainfall climatology over the region, due to the influence of the heat low on regional precipitation (Fig. 6).

Composite plots of strong heat low days demonstrate a marked diurnal cycle of low-level atmospheric thickness which peaks in the late afternoon following the diurnal cycle of solar radiation (Fig. 7). Accordingly, vertical potential temperature gradients and instability are strongest in the afternoon (Fig. 8). Ascent, however, is weak at this time – out of phase with the afternoon surface potential temperature maximum – in agreement with the results of Howard and Washington (2019). Clarifying whether the lack of daytime ascent is due to

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turbulence parameterization in reanalysis is an important avenue of future research. If such a signal is present due to the parameterization of sub-grid scale processes, this has implications for the ability of the heat low to be accurately simulated not only in reanalysis datasets but also climate models, which typically operate at coarser resolutions. Considering the lack of observed data with which to constrain models and reanalysis data over Africa, observational analysis of the diurnal cycle would provide invaluable insight into this key feature of the southern African climate. This is the subject of future research.

Analysis of strong heat lows from 1960-2019 shows a significantly increasing trend: the number of heat lows detected from 1990-2019 has increased by 175% with respect to 1960-1989. Over 40% of the variability in strong heat low frequency is attributable to the temporal trend. This is in agreement with surface temperature and mid-level geopotential height trends (Fig. 10). Dunning et al. (2018) suggest that changes to the relative strength of the Saharan and Angolan heat lows may be responsible for future drying trends over southern Africa owing to their influence on the pressure gradient associated with the movement of the tropical rainbelt. Comparing daily LLAT between 1960-1989 and 1990-2019 shows greatest increases during early and (to a lesser extent) late austral summer, indicating a lengthening of season in which strong heat lows occur. A continuation of this trend would align with the projected shortening of the rainy season over Africa, considering the occurrence of largely negative precipitation anomalies during strong heat low days. Late rainy season onset is expected to occur due to the delayed breakdown of the Congo Air Boundary (Howard and Washington, 2020). The results presented in this study support the conclusion that the circulation associated with heat lows helps to maintain the Congo Air Boundary (and vice versa) and indicate that this relationship may be important for future rainfall trends (Howard and Washington, 2019).

A strengthening of the southern African heat low is likely to play a role in future amplified warming the region, where rates of temperature change are expected to exceed the global rate (Engelbrecht et al., 2015; Vizy and Cook, 2016; Landman et al., 2018; Fan et al., 2021). Alongside the link between heat lows and precipitation, this demonstrates that understanding the regional heat low is essential to constrain model projections of future change across southern Africa in addition to informing forecasting and understanding interannual variability. Such efforts provide vital opportunities to minimize risks to health, infrastructure and agriculture (e.g., Lazenby et al., 2014; Landman et al., 2018). The
increasing frequency of strong heat lows is symptomatic of the rapidly changing nature of the southern African climate system which requires careful observation to inform responses to changes in future seasonality and extremes.

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Data Availability Statement.

All ERA-5 reanalysis data used in this study are available from the Copernicus Climate Data Store at https://doi.org/10.24381/cds.143582cf as documented by Hersbach et al. (2020).

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