Climatology of the heat low and the intertropical discontinuity in the Arabian Peninsula

Ricardo Fonseca1 | Diana Francis1 | Narendra Nelli1 | Mohan Thota2

1 Environmental and Geophysical Sciences (ENGEOS) Lab, Khalifa University of Science and Technology, Abu Dhabi, United Arab Emirates
2 National Centre for Medium Range Weather Forecasting, Ministry of Earth Sciences, Noida, India

Correspondence
Diana Francis, Khalifa University of Science and Technology, P.O. Box 127788, Abu Dhabi, United Arab Emirates.
Email: diana.francis@ku.ac.ae

Abstract
In this article, the climatological state and the seasonal variability of the Arabian heat low (AHL) and the intertropical discontinuity (ITD) are investigated over the Arabian Peninsula using the 1979–2019 ERA-5 reanalysis data. The AHL is a summertime feature, mostly at 15°–35°N and 40°–60°E, exhibiting a clear strengthening over the last four decades in line with the observed increase in surface temperature. However, no clear shift in its position is detected. The AHL, driven by both thermodynamic and dynamic forcing, is broader and stronger during daytime and exhibits considerable variability on day-to-day timescales, likely due to the convection associated with the Asian summer monsoon. The ITD is the boundary between the hot and dry desert air and the cooler and more moist air from the Arabian Sea. It lies along the Arabian Peninsula’s southern coastline in the cold season but reaches up to 28°N between 50° and 60°E in the summer months. While in the former it has a rather small diurnal variability; in the latter it shows daily fluctuations of up to 10° in latitude. The presence of the Sarawat Mountains over southwestern Saudi Arabia precludes a northward migration of the ITD in this area. The ITD exhibited a weak northward migration in the 41-year period, likely due to the increased sea surface temperatures in the Arabian Sea. On inter-annual timescales, the El Niño-Southern Oscillation, the Indian Ocean Dipole, and solar-geomagnetic effects play an important role in the AHL’s and ITD’s variability.

KEYWORDS
arid regions, convergence zone, heat Low, intertropical front, land–sea interactions, monsoon system

1 | INTRODUCTION

Thermal heat lows and convergence zones between moist and dry air masses are ubiquitous features of tropical and subtropical regions (i.e. Flamant et al., 2007; Dumka et al., 2019). They play a crucial role in modulating the mesoscale meteorological features in these regions, such as the mesoscale convection, dust storms and monsoon surge and associated rain (e.g. Bou Karam et al., 2014). Like the other deserts regions, the Arabian Peninsula witnesses the development of a thermal heat low during the summer season and consequently the inland advance of
the intertropical front or the intertropical discontinuity (ITD), located over the Arabian Sea during the winter season. In this study, we aim at characterizing these two features over the Arabian Peninsula and establishing the knowledge on their seasonal and inter-annual variability.

The Arabian Desert, a vast arid region located in the Middle East that extends over 2.33 x 10^6 km², is one of the driest places on Earth (Cosnafroy et al., 1996). The majority of the meagre and irregular precipitation falls in the cold season, in association with mid-latitude baroclinic systems (e.g. Niranjana Kumar and Ouarda, 2014; Al Senafi and Anis, 2015; Wehbe et al., 2018). Summer-time rainfall is mostly confined to the southern part of the Arabian Peninsula in association with the Asian summer monsoon (Babu et al., 2016), where localized convective events form during summer (Steinhoff et al., 2018; Branch et al., 2020; Wehbe et al., 2020). As a result of the strong heating of the surface by the Sun, a thermal low, hereafter denoted as the Arabian heat low (AHL), develops, with a well-mixed layer from the surface up to 650 hPa during mid-afternoon hours (e.g. Blake et al., 1983). While ascent prevails in the lowest 1 km above ground level (AGL), at all levels above 1 km there is a descending motion, associated with the downward branch of the Hadley circulation and Asian summer monsoon (Steinhoff et al., 2018). The AHL is a shallow, cyclonic, warm-core low-pressure system in the lowest kilometre of the atmosphere (e.g. Racz and Smith, 1999).

A more comprehensive investigation of the structure of the AHL is given in Smith (1986a, 1986b), using satellite, aircraft and surface data from field experiments conducted in the warm season of 1979–1982. A swing of about 50°C in the 2-cm deep temperature at a weather station in the Arabian Desert between day and night is observed, with the surface energy budget, in a daily averaged sense, dominated by a balance between the sensible heat flux and the net radiation flux. The latter has also been reported by Nelli et al. (2020a), in an analysis of eddy-covariance measurements at Al Ain in the United Arab Emirates (UAE). The interaction between the AHL and the southwest Asian monsoon is found to be two-way: the monsoon helps to maintain the heat low through large-scale descent and adiabatic warming, while the thermal advection into the western Arabian Sea confines the low-level moisture within the monsoon low-level jet, aiding in its sustenance (Ackerman and Cox, 1982). On intra-seasonal timescales, the two systems are also intertwined. As noted by Steinhoff et al. (2018), active phases of the Asian summer monsoon are associated with enhanced upward motion and convection over the Arabian Sea, with the resulting anomalous subsidence over the Arabian Peninsula leading to a stronger AHL. The AHL also plays an important role in modulating the occurrence of convection. An intense AHL over land during the day drags maritime air masses inland and enhances low-level convergence, which is essential for cloud formation in this region (Schwitalla et al., 2020; Francis et al., 2021).

Convergence lines are of the utmost importance in meteorology, given their role in convection initiation and subsequent occurrence of precipitation (e.g. Weller et al., 2019; Branch et al., 2020). A well-known convergence line in the tropics and subtropics is the ITD, where the hot and dry desert air converges with the cooler and moist tropical maritime air somewhere in the 0°–30° latitude band (Bou Karam et al., 2008; Pospichal et al., 2010). This interface marks the leading edge of the monsoon flow, and is often mixed with the intertropical convergence zone (ITCZ), even though the two features are clearly distinct, as noted by Williams (2008). The ITCZ generally lies some 500 km south of the ITD (Hamilton and Archibald, 1945; Lélé and Lamb, 2007), and represents the core region of tropical convection. The ITD separates the moist monsoon layer to the south from the dry boundary layer to the north (Williams, 2008).

Several studies have been conducted over Africa on the variability and dynamical role of the ITD. For example, Bou Karam et al. (2008) characterized the ITD using airborne measurements and highlighted the importance of the ITD in lifting and subsequent transport of dust over the Sahara. Furthermore, the combination of the ITD and the presence of orography can lead to dry cyclogenesis and subsequent dust emissions over the Sahel (Bou Karam et al., 2009b). The role of the ITD on dust activities has also been noted by, for example, Knippertz (2008) and Lyngsie et al. (2013), and more recently by Francis et al. (2020b) on its contribution to the historical dust storm of June 2020 that extended into the tropical Atlantic Ocean and North America.

Odekunle (2010) and Odekunle and Adejuwon (2017) reported on the effects of the ITD on the variability of the precipitation regimes in Nigeria. While on seasonal scales it follows the annual march of the Sun, over a diurnal cycle the ITD position can fluctuate by 1°–2°, northwards during the night and southwards during the day, in line with the diurnal cycle of the planetary boundary layer (PBL; Lothon et al., 2008), and the expansion of the thermal low during the day (Sultan et al., 2007). Elsewhere, there are very few studies on the dynamics of the ITD. Rashki et al. (2019) found that it plays an important role in the dust variability over the Arabian Sea, with the convergence of the northwesterly (Shamal) winds, southwesterly (monsoon) winds, and northerly (Levar) winds leading to the accumulation of large amounts of dust aerosols over northern and central parts of the Arabian Sea in the summer. Dumka
et al. (2019) characterized the ITD over northern India and stressed the role it played in dust lifting and vertical transport. Both studies focused on an individual period during a summer month, and do not explore how the ITD fluctuates in the region throughout the year, nor on its inter-annual variability.

**FIGURE 1** (a) Orography (m) and spatial extent of the WRF model’s 12 km domain. (b)–(d) Comparison of the observed (black), ERA-5 (red) and WRF-predicted (blue) hourly air temperature (°C) and sea-level pressure (hPa) at the location of the three stations highlighted by a star in (a). The time is given in UTC. [Colour figure can be viewed at wileyonlinelibrary.com]
Despite being key elements of the regional climate and weather patterns, the characteristics of the thermal low and the ITD, as well as their spatio-temporal variability over the Arabian Peninsula, have not been established yet. The aim of this work is to investigate the mean state and variability of the AHL and ITD on different timescales, an important step given their wide-range implications for weather and climate processes. While a similar analysis has been conducted in the neighbouring Sahara Desert (e.g., Lothon et al., 2008; Lavaysse et al., 2009; Pospichal et al., 2010), to the best of the authors’ knowledge, it has not been performed over the Arabian Peninsula and adjacent regions.

This article is structured as follows. In Section 2, the methodologies used to detect the thermal low and the ITD, as well as the set-up of the numerical model used to test some of the hypotheses put forward in the analysis of the reanalysis data, are described. The discussion of the spatio-temporal variability of the AHL and ITD is then presented in Sections 3 and 4, respectively, while in Section 5, the main findings are summarized.

2 | METHODOLOGY

In this section, the most commonly used measures to identify a heat low and the ITD are outlined. They have been employed in the Arabian Peninsula using ERA-5 reanalysis data (Hersbach et al., 2020), extending from 1979 to 2019. ERA-5 was selected due to its higher spatial (0.25° or ~27 km) and temporal (hourly) resolution when compared to other commonly used reanalysis datasets such as ERA-Interim and the Climate Forecast System Reanalysis (e.g., see table 2 of Francis et al., 2021), and its overall good performance over this region. The latter can be seen in Figure 1b–d, where the ERA-5 predictions at the closest grid-point to the location of three airports in the region are compared with the observed values for a 1-week period in July 2018. The slight offset in the mean sea-level pressure at station #1 and, in particular, at station #3, is due to differences in the terrain, which at ~27 km resolution is not fully captured by ERA-5.

2.1 | Thermal low detection

To detect the AHL, we consider the same metrics implemented by previous studies to identify and characterize the Sahara heat low (SHL), the AHL’s counterpart over the Sahara Desert, for which several measures have been employed in published works. For example, Lavaysse et al. (2009) used the low-level atmospheric thickness (LLAT) criterion to detect the SHL. The LLAT is the 700–925 hPa geopotential height thickness, with the SHL corresponding to the region with the 10% highest LLAT in the domain 20°W–30°E and 0°–40°N. A similar approach was followed by Engelstaedter et al. (2015) and Wang et al. (2015). The idea is that the presence of the SHL and associated low-level temperature-increase leads to an expansion of the lower atmosphere and therefore higher LLATs. Chauvin et al. (2010) used the maximum in the 850 hPa potential temperature, $\theta_{850 \text{ hPa}}$, a level located at roughly 3,000 m AGL, to characterize the SHL. This field is correlated with the LLAT (Roehrig et al., 2011). Flamant et al. (2007) identified the SHL using the 1,006 hPa threshold of the mean sea-level pressure. Here we use all the three metrics described above over the domain 10°–35°N and 40°–60°E for 0300 UTC. As explained in Lavaysse et al. (2009), the detection of the SHL is conducted just before local sunrise, at 0600 UTC in West Africa, as at this time the low-level temperature field is least affected by the presence of clouds and the complex surface albedo pattern. As West Africa is roughly three time zones behind the Arabian Peninsula, the choice of 0300 UTC is justified. In addition to the referred metrics, the relative vorticity at 850 hPa, $\zeta_{850 \text{ hPa}}$, used by Steinhoff et al. (2018) in their study of a convective event over western UAE in late August 2011, is also considered.

In order to assess the performance of the different metrics, Figure 2 shows them on a single day (11 July 2018) at 0000, 0600, 1200 and 1800 UTC (similar conclusions are reached for other days, not shown). The first panel gives the LLAT in shading, while the stipple shows the AHL using the LLAT metric proposed by Lavaysse et al. (2009). In order to construct this plot, and for a given timestamp, the cumulative distribution of the LLAT for the domain 10°–35°N and 40°–60°E, which encompasses the Arabian Peninsula where the AHL develops, is first generated. As the geopotential height is extrapolated below orography in the reanalysis dataset, all regions for which the 925 hPa pressure level is below the surface are excluded from the analysis. In addition, the water bodies are also masked out, as the thermal low is, by definition, a land-based feature. The top 10% of the values are then plotted. The second to fourth panels show $\theta_{850 \text{ hPa}}$, sea level pressure, and $\zeta_{850 \text{ hPa}}$, with the AHL region also stippled. In all panels, the 2-m temperature equal to 40°C isotherm is drawn as a solid contour, so as to highlight the warmest regions at a given time.

The AHL extends over a broad region around the Arabian Gulf. Spatially, it is not only more intense during daytime, but also broader, as a result of the strong surface heating. At night, and in particular around local sunrise (0600 UTC) when it is defined following Lavaysse et al. (2009), the AHL largely coincides with the region...
where $\theta_{850\,\text{hPa}}$ is the highest and the sea-level pressure is the lowest, in line with theoretical arguments (e.g. Blake et al., 1983; Smith, 1986a, 1986b; Rácz and Smith, 1999). However, during daytime and evening hours, the heat low in the pressure field is centred over southeastern Saudi Arabia, where the air temperature is higher and a cyclonic circulation at 850 hPa is present. The AHL defined using the LLAT metric only includes the northern part of the core of the referred low pressure, whose signature is also not present in the $\theta_{850\,\text{hPa}}$ plot. This is a drawback of using the LLAT to identify the thermal low during daytime, even though it is important to note that, as stated before, the definition used here was designed for nighttime hours and just before sunrise.

The $\zeta_{850\,\text{hPa}}$ field also appears to be a good metric to diagnose the thermal low but only on a daily basis, as no signature of the heat low using this metric is detected in the climatological analysis as it will be shown later.
Figure 2 shows a region of low-level cyclonic vorticity over southeastern Saudi Arabia to the south of the UAE at all times, with a higher amplitude of about $10^{-4}$ s$^{-1}$ at 0600 UTC in line with Rácz and Smith (1999), and only partially captured by the AHL using the LLAT definition. The low-level convergence on 11 July 2018 is broadly in the same area (although the peak region is further north) to the one on 30 August 2011 shown in Steinhoff et al. (2018), their figure 2. The vortex is slightly displaced with respect to the minimum in the sea-level pressure, which is indicative of a baroclinic signature in line with expectations (e.g. Smith, 1986a, 1986b), and is of a smaller spatial extent. What is more, no clear signal is present in that region in the LLAT field, with a relative minimum shown in the $\theta_{850\text{hPa}}$ plot. An inspection of the vertical velocity at 850 hPa indicates stronger ascent in the area where $\theta_{850\text{hPa}}$ is lower (not shown), suggesting that the near-surface potential temperature here is reduced when compared to neighbouring regions, likely due to stronger radiative cooling at this inland site.

In summary, the AHL defined using the LLAT field is able to diagnose the thermal low mostly at night and just before sunrise, with the defined AHL largely coinciding with the areas where $\theta_{850\text{hPa}}$ is the highest and the sea-level pressure is the lowest, in line with Smith (1986a, 1986b). During daytime, however, when it is broader and has a larger magnitude, the LLAT is not as good of a metric to identify it, with the sea-level pressure or $\xi_{850\text{hPa}}$ preferred. As the AHL is defined around 0000–0600 UTC, the LLAT definition will be used in the subsequent discussion.

### 2.2 Intertropical discontinuity detection

As the boundary between the hot and dry desert air and the cooler and moist marine air, the ITD has been widely defined using the 2-m dew-point temperature. For example, Buckle (1996), Bou Karam et al. (2009a), Flamant et al. (2009) and Chaboureau et al. (2016) used a value of 14°C for this purpose, while Grams et al. (2010) and Lafore et al. (2017) considered 15°C. Lothon et al. (2008) and Kalapureddy et al. (2010) identified the ITD as the line of maximum horizontal gradient of the dew-point temperature. Bou Karam et al. (2009b) defined the ITD as the region of maximum meridional gradient of the integrated water vapour content, focusing on the moisture contrast across the interface. In addition to moisture-based diagnostics, some authors have considered the wind reversal associated with the ITD to detect it. For example, Flamant et al. (2007) and Pospichal et al. (2010) defined the ITD as the latitude of the reversal of the 925 hPa wind field from southerly to the northerly along the interface, a similar approach to that followed by Bou Karam et al. (2008). Lavaysse et al. (2009) also looked into the wind reversal but by finding the minimum in the 925 hPa geopotential height between the equator and 28°N. Odekonle (2010) used a combination of the two criteria above (i.e. surface wind convergence and a dew-point temperature threshold of 15°C), together with the discontinuities in dry bulb temperature and surface pressure. The ITD is usually defined at night, when it is most stable and not affected by the presence of the daytime thermal convection (Lothon et al., 2008). In a study over the Arabian Sea, Rashki et al. (2019) identify the ITD as the region of zero meridional wind at 850 hPa, while Dumka et al. (2019) over India defines it as the area of weak winds of less than 4 m s$^{-1}$ at 925 hPa, where two opposing wind regions converge. For the 925 hPa pressure level, considered in the vast majority of the studies above to detect the ITD, using the reversal of the wind direction to identify this feature gives similar results to employing the dew-point temperature criterion with a threshold of 15°C at 0000 UTC (not shown). Given this, the dew-point criterion, with a threshold of 15°C applied over the domain 13°–38°N and 43°–70°E at 0000 UTC, is considered due to its added simplicity.

### 2.3 Numerical simulations

In order to better understand the role of the orography on the ITD, simulations with the Weather Research and Forecasting (WRF; Skamarock et al., 2008) model are also conducted. Two runs are performed: a real-case, where

| Parameterization scheme | Option |
|-------------------------|--------|
| Cloud microphysics      | Thompson aerosol-aware scheme (Thompson and Eidhammer, 2014) |
| Planetary boundary layer | Mellor-Yamada Nakanishi Niino (MYNN) Level 2.5 (Nakanishi and Niino, 2006, 2009) |
| Radiation               | Rapid Radiative Transfer Model for Global Circulation Models (Iacono et al., 2008) |
| Cumulus                 | Kain-Fritsch (Kain, 2004), with subgrid-scale cloud feedbacks to radiation (Alapaty et al., 2012) |
| Land surface model      | Noah LSM with MultiParameterization options, Noah-MP (Niu et al., 2011; Yang et al., 2011) |
FIGURE 3  Legend on next page.
WRF is run as per normal, and a semi-idealized run, where the orography is removed. The period targeted for the modelling work is 11–18 July 2018. The run is initialized on 10 July, with the first day regarded as spin-up and its output discarded. WRF is run with a single nest at 12 km spatial resolution, with the domain comprising the whole of the Middle East and surrounding region, as shown in Figure 1. The model configuration, listed in Table 1, is as in Francis et al. (2021), which has been found to work well for warm season simulations in this region.

A comparison between the WRF predictions and the observed measurements at the three sites, given in Figure 1b–d, shows a clear cold bias, more pronounced in the evening and nighttime hours. This is a well-known model bias, reported, for example, in Weston et al. (2018), Nelli et al. (2020b), Sch witalla et al. (2020) and Temimi et al. (2020), that is also seen in other arid regions such as the Sahara Desert (e.g. Fekih and Mohamed, 2019). It has been attributed to deficiencies in the physics schemes, in particular in the land surface model (LSM) and radiation scheme, and/or an incorrect representation of the surface properties and concentration of atmospheric gases and dust. Using a different PBL parameterization scheme (Chaouch et al., 2017), tweaking tunable parameters inside the LSM (Weston et al., 2018) and surface layer (Nelli et al., 2020b) schemes, or employing a more realistic representation of the soil texture and land use land cover (Temimi et al., 2020), does not seem to alleviate this problem. A correction of the WRF cold bias is beyond the scope of this work. Despite these biases, however, the WRF model gives a good representation of the AHL and atmospheric fields such as sea-level pressure, and low-level potential temperature and relative vorticity. This can be seen by comparing Figure S1 with Figure 2. The cold bias can be seen as the 2-m temperature contour covers a much-reduced area in Figure S1 when compared to Figure 2, where ERA-5 data, which is more accurate as seen in Figure 1b–d, are used. However, the WRF-predicted AHL is not that different from that predicted by ERA-5, with the minimum in sea-level pressure and the cyclonic vortex in $\zeta_{850\, \text{hPa}}$ being roughly in the same place, and having a comparable magnitude, to the correspondent ones in the reanalysis plot. This highlights the potential use of the WRF model to investigate further this feature, at least on a day-to-day basis. Having said that, in this study the model will only be employed to investigate the role of the topography on the position of the ITD by comparing the output of a real-case to that of a semi-idealized simulation. The bulk of the analysis conducted here only makes use of the reanalysis dataset.

3 | THE ARABIAN HEAT LOW

As summarized in Section 2.1, LLAT, $\theta_{850\, \text{hPa}}$, sea-level pressure and $\zeta_{850\, \text{hPa}}$ diagnostics have been used to detect heat lows. Here, the analysis will be extended to monthly mean fields averaged over the 41-year (1979–2019) ERA-5 data, with the results given in Figure 3 at 0300 UTC (0700 LT), just before or around local sunrise.

3.1 | Spatial extent and annual cycle

The AHL is found to be exclusively a summer feature, only present from June to September (JJAS), and hence the fields are only shown in the warmer months. The AHL achieves its highest amplitude in July, when the surface/air temperatures in the region peak (e.g. Al Senafi and Anis, 2015; Branch et al., 2020; Nelli et al., 2020a). It covers a broad region, extending from the Rub’ Al Khali desert in central Saudi Arabia to the shores of the Arabian Gulf and Sea of Oman. It also comprises some of the valleys in Iran. Indeed, the AHL is part of the Asian summer monsoon trough, which has a rather broad spatial extent (e.g. Yu et al., 2016). The solid blue line corresponds to the 2-m temperature equal to $40^\circ\text{C}$ contour, and it largely overlaps with the AHL, as expected from theoretical considerations (e.g. Smith, 1986a, 1986b). On top of the thermodynamic forcing, the AHL is also partially dynamically driven. Figure 4a gives the vertical velocity and horizontal winds at 850 hPa for the same months and time of the day. It shows a broad region of low-level convergence and ascent over central Saudi Arabia, where the northwesterly turning northeastlies (Shamal) winds slow down, and also converge with the southwesterly (monsoon) winds. This region of upward motion roughly matches the area where
the highest $\theta_{850\text{ hPa}}$ are seen, Figure 3b, and is also part of the AHL obtained with the LLAT criterion.

The similarity between the LLAT and $\theta_{850\text{ hPa}}$ has been noted for example, by Roehrig et al. (2011), even though for the Arabian Peninsula in the summer months, the latter is dominated by the dynamic forcing. Conversely, the sea-level pressure field, Figure 3c, picks up more of the thermodynamic forcing, with the AHL encompassing the regions of lowest sea-level pressures. The fact that the near-surface low lies in regions of low topography is consistent with the findings of Lavaysse et al. (2009, 2013), whom reported that low-elevation regions are preferred sites for thermal lows. As opposed to the LLAT, $\theta_{850\text{ hPa}}$ and sea-level pressure fields, and to the conclusions reached in the analysis of the 1-day event in Figure 2, however, the vorticity field, $\zeta_{850\text{ hPa}}$ given in

![Image](https://example.com/image.png)

**Figure 4** (a) June to September monthly mean 850 hPa omega (shading; Pa s$^{-1}$) and horizontal wind vectors (arrows; m s$^{-1}$). (b) Hourly AHL index, defined as the spatial average of the 700–925 hPa thickness (m) corresponding to the AHL. The white shading in (a) denotes regions for which the 850 hPa pressure surface is below orography. Both plots are generated with ERA-5 1979–2019 data. (c) and (d) are as (b) but showing the 850 hPa potential temperature (K) and sea-level pressure (hPa) averaged over the AHL region [Colour figure can be viewed at wileyonlinelibrary.com]
Figure 3d, does not seem to properly capture the signature of the AHL. The areas of low-level convergence are rather weak and small in size, with this field being dominated by negative values that arise from the presence of the subtropical anticyclone (Spinks et al., 2015). This is expected, as ζ_{850 hPa} is a noisy field, with a great deal of cancellation taking place when doing long-term averages. Low-level convergence is primarily confined to regions of
complex topography, such as southern and western parts of Yemen, and the high-terrain over Zagros Mountains and the Hindu Kush. A comparison with the annual cycle of the SHL, given in figure 4 of Lavaysse et al. (2009), reveals that the AHL has a more intricate structure, owing to the more complex topography and land-sea patterns of the Arabian Peninsula.

3.2 Diurnal variability

In order to investigate the diurnal variability of the AHL, the spatially-averaged LLAT over the AHL region, regarded as the AHL index, is plotted for the June to September months in Figure 4b. The lowest LLAT occurs around 0500–0600 UTC (0900–1000 LT) and the highest around 1400 UTC (1800 LT), roughly 2 hr after the surface and air temperatures in the region reach their extrema (e.g. Nelli et al., 2020a). This lag arises from the fact that it takes time for the sensible heat fluxes to warm up the lower levels of the atmosphere. The phase of the diurnal cycle is robust throughout the warm season, with the higher thicknesses in July consistent with the observed temperature annual cycle (e.g. Al Senafi and Anis, 2015; Branch et al., 2020; Nelli et al., 2020a). Figure 4c, d are as Figure 4b but for $θ_{500 \text{ hPa}}$ and sea-level pressure, respectively. They show a similar monthly variability and phase of the diurnal cycle as the LLAT. The double minima, at roughly 0600 UTC (1000 LT) and 1900 UTC (0300 LT), seen in the pressure field also arises from the atmospheric tides (Nelli et al., 2020a).

3.3 Intra-seasonal variability

On top of the diurnal variability, the AHL also exhibits considerable intra-seasonal variability. This can be seen in Figure 5a, which shows the hourly AHL index from May to September 2018. In order to obtain this index, and for a given timestamp, we first compute the AHL based on the proposed LLAT metric discussed in Section 2.1. Then, we take the averaged LLAT over the AHL region, a measure of its strength. The averaged latitude and longitude of the AHL region, which give information regarding its spatial variability, are plotted in Figure 5b.

Intra-seasonal variability in the strength and spatial extent of a thermal low has been reported in arid regions such as the Sahara Desert. For example, Wang et al. (2017) noted a 10-day cycle in the Saharan heat low, where it first built up over the core of the desert, migrated westwards over the following 3–5 days, and finally collapsed around the West African coastline. The westward migration may be related to (a) warm and dry air advection from the north to the west of system in association with the cyclonic circulation; (b) cold and moist air along its eastern edge due to convective systems; (c) propagation of mid-latitude Rossby waves along the North Atlantic–North African waveguide. The collapse of the heat low around the coast is due to cold and humid air advection, in association with a cold oceanic current that flows near the coastline (e.g. Parker and Diop-Kane, 2017). Day-to-day variability has also been reported for the AHL, such as in Ackerman and Cox (1982) and Steinhoff et al. (2018), and is found to be linked with the Asian summer monsoon.

An inspection of Figure 5a reveals a variability of the AHL on timescales of about 5–15 days. A comparison with Figure 5b suggests that a north-westward migration of the AHL, that is, towards the core of the Arabian Desert, is typically accompanied by its intensification, while a movement to the southeast, that is, closer to the Arabian Sea, sees its weakening, a behaviour consistent with that of the SHL described in Wang et al. (2017). In order to illustrate the two extreme states, Figure 5c, d show the LLAT, AHL, and 10-m horizontal winds at 0300 UTC on 7 and 16 June 2018, respectively, roughly when the index reaches its highest and lowest value for the summer of 2018. The two snapshots reveal a remarkable contrast in the spatial extent and magnitude of the thermal low: while on 7 June, the LLAT exceeds 2,450 m over the bulk of the AHL region, on 16 June it dropped up to 50 m in that region, with the AHL found further south, over Oman and northern Yemen. In addition, the northwesterly (Shamal) winds are stronger over the Gulf on 16 June, and advect the lower thicknesses over Iraq/Kuwait southeastwards. The higher magnitude wind speeds are consistent with a deeper and broader monsoon trough on 16 June (not shown), which gives rise to a steeper pressure gradient with respect to the subtropical high (Yu et al., 2016). As the monsoon trough is closely related to the Indian Summer Monsoon, this suggests that summertime weather conditions in the Arabian Peninsula are linked with the intra-seasonal variability of the Indian Summer Monsoon. This connection is well known and has been explored, for example, by Steinhoff et al. (2018) and Attada et al. (2019). In a nutshell, enhanced convective activity, and therefore ascent, over the Arabian Sea and Indian subcontinent leads to stronger descent over the Arabian Peninsula, and subsequently a deeper thermal low, which in turn strengthens the lower-level winds. As a result of increased moisture advection from the surrounding seas, summertime convection over western UAE is more frequent during the decay phase of the AHL, as noted by Steinhoff et al. (2018).
FIGURE 5  Hourly AHL (a) index, defined as the averaged LLAT over the AHL region, and (b) position, given by the AHL’s averaged longitude (blue; °; left axis) and latitude (red; °; right axis), from May to September 2018 from ERA-5 data. (c) LLAT (m; shading), 10-m horizontal wind vectors (m s⁻¹; arrows), and AHL (stippled region) at 0300 UTC on 7 July 2018, the time around which the highest value of the AHL index for May–October 2018 is reached. The white shading denotes water bodies and areas for which the 925 hPa pressure surface is below orography. (d) is as (c) but on 16 July 2018 at 0300 UTC, roughly when the lowest value of the AHL index for May–October 2018 is reached [Colour figure can be viewed at wileyonlinelibrary.com]
In addition to the referred intra-seasonal variability, it is interesting to note in Figure 5a the sudden build-up of the AHL from late May to early June, accompanied by a north-eastward migration, Figure 5b, and its rapid decline in September, associated with a south-westward movement. This is reminiscent of what has been reported over
the Sahara Desert (e.g. Redelsperger et al., 2002; Sultan and Janicot, 2003). Besides the annual march of the Sun, this may arise from the fact that the majority of the precipitation in the region occurs in winter and early spring (e.g. Nelli et al., 2020a). Once the soil (and the atmosphere) is bone dry, the heat low can develop very quickly.

**FIGURE 6** (a) Yearly JJAS AHL index at 0300 UTC for the period 1979–2019. The blue line is the best linear fit to the data, with the equation given on the top left. (b) is as (a) but for the surface temperature (K) averaged over the AHL region, while (c) is as (b) but with the surface temperature averaged over the AHL domain (10°–35°N and 40°–60°E) outside the AHL region. In (d), the yearly JJAS sea surface temperatures over the Arabian Sea (red line; 5°–15°N, 55°–70°E) and Arabian Gulf (blue line; 20°–30°N, 30°–55°E) are plotted. All trends are statistically significant at the 95% confidence interval, except the one for the Arabian Sea SSTs (red line in panel (c)) [Colour figure can be viewed at wileyonlinelibrary.com]
The AHL exhibits a clear positive trend, which is statistically significant at the 95% confidence level, with the LLAT increasing at a rate of about 0.2 m·year$^{-1}$, giving a relative increase of roughly 0.3% over a 40-year period. This is not surprising, as several studies have highlighted the increase in surface/air temperature in the region over the last few decades (e.g. Almazroui et al., 2014), which is also seen in the ERA-5 data (Figure 6b) and is also statistically significant at the 95% confidence level. It is also interesting
to note that, while the surface temperature in the AHL domain (10°–35°N, 40°–60°N) but outside the AHL region has been increasing over the last four decades (Figure 6c), the rate of increase is roughly 35% lower and the temperatures about 2 K less than in the AHL region. In other words, while the warming tendency is not confined to the AHL, it is more pronounced in the heat low region. Further analysis of the surface temperature

FIGURE 7 (a) Wavelet power spectrum, using the Morlet wavelet, of the JJAS AHL time-series (4 months for 41 years giving a total of 164 data points). (b) as (a) but for the JJAS ITD time-series. The purple line denotes the cone-of-influence, which gives the maximum period of useful information at a particular time [Colour figure can be viewed at wileyonlinelibrary.com]
variability in the region, including the land–sea temperature gradient, is outside the scope of this study.

There is considerable inter-annual variability in the AHL index, with anomalously high values in 1998 and 2017, and low values in 1992, 2004 and 2013. As the summers of 1998 and 2017 featured a La Nina and those of 1992 and 2004 an El Nino (2013 was a neutral summer), it appears that the El Nino-Southern Oscillation (ENSO; Huang et al., 2017) may play an important role in the variability of the AHL. This is confirmed in a wavelet analysis, Figure 7a, which shows higher power for timescales of 2–4 years, the prevailing timescales of ENSO variability (e.g. Berner et al., 2020). ENSO is known to have a significant impact on the Indian summer monsoon: the
FIGURE 9  (a) Monthly mean 2-m dew-point temperature (shading; °C), 10-m horizontal wind vector (arrows; m s⁻¹), and position of the ITD (solid green line), defined as the latitude at which the dew-point temperature at 0000 UTC, decreasing northwards, is equal to 15°C, from ERA-5 (1979–2019) data. (b) Six-hourly diurnal cycle of the ITD for each month. The ITD position at 0000 UTC is shown in red, at 0600 UTC in green, at 1200 UTC in blue, and at 1800 UTC in orange [Colour figure can be viewed at wileyonlinelibrary.com]
enhanced convection over the Maritime Continent in La Nina events, leads to anomalous westerlies over the tropical Indian Ocean, which act to strengthen the monsoon’s circulation, a relationship modulated by the Indian Ocean SSTs (e.g. Hrudya *et al.*, 2020; Srivastava *et al.*, 2020). At the same time, and as highlighted before, enhanced convection over the Indian subcontinent and Arabian Sea leads to anomalous descent over the Arabian Peninsula, and hence to an intensification of the AHL. In addition to ENSO, the Indian Ocean Dipole (IOD; Saji *et al.*, 1999) also has a peak in that region of the spectrum (Wang *et al.*, 2019), and may impact the AHL through changes in the convection in the Arabian Sea. In particular, positive IOD events are accompanied by higher than average SSTs over the western Indian Ocean, which will increase the likelihood of convective activity over the Arabian Sea, and subsequently descent and adiabatic warming over the Arabian Peninsula. The higher power around a timescale of 2 years may be linked to the biennial oscillation in the Asian-Australian monsoon system, connected to both ENSO and IOD (e.g. Konda *et al.*, 2018). In addition to the referred modes of variability, increased solar activity and the resulting stronger surface heating are also likely to strengthen the AHL (e.g. Misios *et al.*, 2016).

In order to better visualize the change in the AHL’s magnitude and spatial extent with time, Figure 8a,b shows Hovmoller plots averaged over 40°–60°E and 15°–25°N, respectively, where the bulk of the thermal low is found (Figure 3). Here, the LLAT anomalies with respect
to the 1979–2019 monthly climatology normalized by its standard deviation are plotted. A positive trend is seen, with negative values prevailing before 1997–1998, and positive values since then. The transition, which takes place around 1997–1998, is likely linked with a change in the dominant phase of ENSO and associated teleconnections. As reported by Niranjan Kumar and Ouarda (2014) and Aldababseh and Temimi (2017), El…
Nino events were more frequent from the late 1970s to the mid-1990s, while La Nina events have occurred more often from mid-1990s. Figure 8a,b does not show any significant shift in the position of the AHL over the 41-year period, with the largest anomalies seen around 15°–25°N and 40°–50°E, in the western half of the AHL.

4 | THE INTERTROPICAL DISCONTINUITY

4.1 | Spatial extent and annual cycle

Being the interface between hot and dry northerly winds and the cooler and moist southerly winds, both moisture-based and wind-based diagnostics have been employed to detect the ITD (e.g. Pospichal et al., 2010; Lafore et al., 2017). Figure 9a shows the monthly-mean 2-m dew-point temperature at 0000 UTC with the 10-m horizontal wind vector superimposed. The near-surface atmospheric circulation in the Arabian Sea features northeasterly winds in the cold season, and southwesterly winds in the warm season, in association with the Asian monsoon (Schott et al., 2009). The background flow over the Arabian Gulf, on the other hand, is from the northwest, stronger in the warm season in response to a more vigorous pressure gradient between the subtropical high over Africa and the western Arabian Peninsula and the Asian summer monsoon trough (Al Senafi and Anis, 2015; Bou Karam Francis et al., 2017). The dew-point temperature is higher in the warm season, reaching or even surpassing 30°C in the Arabian Gulf and Sea of Oman, as a result of enhanced surface evaporation (Xue and Eltahir, 2015).

The ITD, shown by a green colour contour, is along the Saudi Arabian and western Indian coastlines in the cold season, penetrating inland from April to October, up to 28°N just to the west of 60°E. In its furthest inland position (July and August), it reaches the southern Iranian coast and the Arabian Gulf, extending into Pakistan. In other words, while the ITD may be of little dynamical interest in the south-western Arabian Peninsula, on its eastern side it is likely to play an important role for example, in the development of convective events, as a convergence line favours ascent and cloud development (e.g. Francis et al., 2021), and dust episodes (e.g. Rashki et al., 2019). Figure 9b gives the 6-hr diurnal cycle of the ITD. Over the eastern Arabian Peninsula, there is very little diurnal variability in the cold season, but in the warmer months a clear southward shift during daytime and northward shift at night, by as much as 10°, is seen. These daily fluctuations are roughly 5 to 10 times larger than those seen over Africa, and are most probably due to the location of the AHL, which is closer to the nearby seas than the SHL over Africa. The link between the diurnal cycle of the ITD and the daytime expansion of the thermal low has been noted, for example, by Lothon et al. (2008) and Lavaysse et al. (2010) over Africa. The intensification of the AHL leads to increased moisture advection inland and a northward displacement of the ITD. As the cooler and moist air is advected northward with the ITD, the heat low weakens, and the ITD moves southwards again.

In order to investigate the role of the topography in the position of the ITD, two simulations with the WRF model were performed: a real-case simulation, and a semi-idealized run where all the orography in the model’s 12 km domain, Figure 1, was removed. The WRF predictions on 11 July 2018 at 0000 and 1200 UTC for both simulations are given in Figure 10. As in the climatological plots in Figure 9, the ITD is mostly along the southern Arabian coastline, although it migrates northwards in the eastern side of the peninsula at 0000 UTC. In the run without orography, however, the ITD is always inland over the Arabian Peninsula, continuing to exhibit a larger diurnal cycle on the eastern side due to the presence of the Arabian Gulf and subsequent moisture advection by the sea-breeze circulation. The near-surface circulation is also very different in the semi-idealized simulation, with northeasterly winds over the Gulf, in response to a more zonally elongated monsoon low that extends from the UAE and Oman into northern India (not shown), as it is not constrained by the topography. The results of the semi-idealized run highlight the role of the orography in shaping the ITD, and confirm that the coastal position of the ITD over the southwestern Arabian Peninsula is due to the presence of steep topography (Sarawat Mountains) over the site. In particular, the high-terrain over Yemen and parts of Saudi Arabia, where the mountains reach up to 3,666 m above sea-level, blocks the inland progression of the ITD, which is a low-level feature extending up to 2 km in altitude.

4.2 | Inter-annual variability and climatological trends

Figure 11a shows the variability of the summertime ITD position over 40°–60°E. As opposed to the AHL, no marked trend is seen in the data, with a slightly northerly position from 1997–1998 to 2008–2009, even though the overall fluctuation of the median does not exceed 2°. This is confirmed in Figure 11b, which shows a linear fit to the yearly mean JJAS ITD latitude over the 41-year period. The slope of the line is 0.0164°·year⁻¹, which corresponds to an overall change in position of roughly 0.7°

...
in 40 years. This weak (but positive and statistically significant at the 95% confidence level) trend likely reflects the increase of the SSTs (and hence surface evaporation) in the Arabian Sea (Kumar et al., 2009), also present in the ERA-5 data (Figure 6d). In addition, the recent warming of the Arabian Gulf (Figure 6c) and Sea of Oman (Noori et al., 2019), and subsequent inland moisture advection by the sea-breeze circulation (Eager et al., 2008), may also promote a northerly shift of the ITD position. The wavelet spectrum for the JJAS ITD time-series, presented in Figure 7b, shows the highest power in the 2–4 and 6–14 year bands. The former is also present in the AHL spectrum, Figure 7a, and has been attributed to the influence of ENSO and IOD on the atmospheric circulation over the Arabian Peninsula. ENSO is also likely to play a role in the ITD, as the weather conditions in the southern Arabian Peninsula are more humid in La Nina years (Babu et al., 2016), such as in 1983, 1986, 1998 and 2017 (Huang et al., 2017). In addition, the above-average SSTs in the western Indian Ocean in positive IOD events will enhance the amount of moisture in the atmosphere through evaporation that is subsequently advected inland (Anil et al., 2016). This may explain the northern migration of the ITD in such episodes. The peak in the 6–14 year band is likely due to solar-geomagnetic effects (e.g. Sunkara and Tiwari, 2016). Figure 11 also highlights that, despite a weak positive trend in the ITD latitude, there has been a slight southward displacement since 2010. A possible explanation is the shift to more negative IOD events, co-occurring with the more frequent La Nina episodes (e.g. Niranjan Kumar and Ouarda, 2014; Lestari and Koh, 2016).

5 | CONCLUSIONS

In this article, the climatologically mean state and variability of the AHL and ITD are investigated over the Arabian Peninsula with the 1979–2019 ERA-5 reanalysis data. The usage of ERA-5 data, which has high spatial (0.25° or ~27 km) and temporal (hourly) resolution compared to other reanalysis datasets, is justified, as a comparison with hourly data at individual stations in the region revealed a good agreement with the observed air temperature and sea-level pressure. The AHL is a deep thermal low that develops in response to the strong surface heating, and is therefore mostly a summertime feature. The ITD denotes the boundary between the hot and dry air from the Arabian Desert and the cooler and moist marine air from the Arabian Sea. Both features have been extensively investigated over Africa, but not so much over southwestern Asia, where only the ITD was discussed and for individual case studies.

The AHL is typically detected with a LLAT, θ850 hPa, sea-level pressure or ζ850 hPa diagnostic. Here, the LLAT-based methodology proposed by Lavaysse et al. (2009) is employed over the region 10°–35°N and 40°–60°E at 0300 UTC (before local sunrise). The AHL is found to exhibit variability on intra-seasonal timescales, in association with the active and break periods of the Indian Summer Monsoon (e.g. Attada et al., 2019): increased convective activity and ascent over the Arabian Sea and Indian subcontinent lead to descent and adiabatic warming over the Arabian Peninsula, and hence to a stronger AHL. This feature also shows a pronounced diurnal variability, with a maximum at 1400 UTC (1800 LT) and a minimum at 0300–0500 UTC (0900–1100 LT). The phase is largely invariant from June to September, with the highest amplitudes seen in July. Over the 41-year period, the AHL shows a strengthening that is statistically significant at the 95% confidence interval, in line with the observed surface warming in the region (Almazroui et al., 2014). However, no clear shift in its position is noted. On inter-annual timescales, ENSO and IOD play an important role in the AHL’s variability, as they modulate the convective activity in the Indian Ocean.

The ITD is identified using moisture-based and/or wind-based diagnostics. Here, a 15°C threshold in the dew-point temperature is used to detect the ITD at 0000 UTC. In the cold season, the ITD is located along the Arabian Peninsula coastline, but it migrates northward starting in April, and reaches up to 28°N just to the west of 60°E in particular in July and August. In the western Arabian Peninsula, it shows a rather small diurnal variability, being located mostly along the southern Saudi Arabian coastline, while in the eastern side, and in the warm season, its position can fluctuate by as much as 10°, reaching the Arabian Gulf and southern Iranian coastline at night. A comparison between a real-case and a semi-idealized numerical simulation where the orography is removed for a 7-day period in July 2018, revealed that the coastal position of the ITD is due to the presence of topography, as without it the ITD is placed well inland over the full southern Saudi Arabia. Only a weak positive trend, but statistically significant at the 95% confidence level, in the location of the ITD is seen in the data, likely driven by the increased SSTs in the Arabian Sea and Arabian Gulf. While ENSO and IOD seem to account for a significant fraction of the inter-annual variability of the ITD, like the AHL, the signal from solar-geomagnetic effects (e.g. Misios et al., 2016; Sunkara and Tiwari, 2016) is also clearly seen in a wavelet analysis. Positive IOD events promote a northward migration of the ITD, as La Nina events also do.
It is important to note that the AHL and ITD are interrelated, as the strengthening of the AHL will lead to the advance of the opposing flows towards it and a northward displacement of the ITD. Both features play a crucial role in the weather conditions in the Arabian Peninsula by modulating the atmospheric circulation at different altitudes. For instance, the ITD helps in triggering dust storms and convective events (e.g. Francis et al., 2020a, 2021) as convergence zones promote ascent and favour cloud development while increasing near-surface turbulence which favours dust uplift. An extension of this work would include investigating how processes such as dust storms, convection initiation, and sea-land-breeze circulations are modulated by the AHL and ITD.

ACKNOWLEDGEMENTS
We would like to acknowledge the Copernicus Programme for making the ERA-5 reanalysis data freely available through the Climate Change Service portal (https://climate.copernicus.eu/climate-reenalysis). We wish to acknowledge the contribution of Khalifa University’s high-performance computing and research computing facilities to the results of this research. We would also like to thank two anonymous reviewers for their several detailed and insightful comments that helped to significantly improve the quality of the article.

CONFLICT OF INTEREST
The authors declare that they have no conflict of interest.

AUTHOR CONTRIBUTIONS
Ricardo Fonseca: Data curation; formal analysis; investigation; writing-original draft; writing-review & editing.
Diana Francis: Conceptualization; project administration; resources; supervision; writing-review & editing.
Narendra Nelli: Data curation; formal analysis; methodology; validation; visualization. Mohan Thota: Methodology; validation.

ORCID
Ricardo Fonseca https://orcid.org/0000-0002-8562-7368
Diana Francis https://orcid.org/0000-0002-7587-0006
Narendra Nelli https://orcid.org/0000-0003-0066-6040

REFERENCES
Ackerman, S.A. and Cox, S.K. (1982) The Saudi Arabian heat low: aerosol distributions and thermodynamic structure. Journal of Geophysical Research, 87, 8991–9002.
Al Senafi, F. and Anis, A. (2015) Shamals and climate variability in the Northern Arabian/Persian Gulf from 1973 to 2012. International Journal of Climatology, 35, 4509–4528.
Alapaty, K., Herwehe, J., Otte, T.L., Nolte, C.G., Bullock, O.R., Mallard, M.S., Kain, J.S. and Dudhia, J. (2012) Introducing subgrid-scale cloud feedbacks to radiation for regional meteorological and climate modeling. Geophysical Research Letters, 39, 24809. https://doi.org/10.1029/2012GL054031.
Aldababseh, A. and Temimi, M. (2017) Analysis of the long-term variability of poor visibility events in the UAE and the link with climate dynamics. Atmosphere, 8, 242. https://doi.org/10.3390/atmos8120242.
Almazroui, M., Islam, M.N., Dambul, R. and Jones, P.D. (2014) Trends of temperature extremes in Saudi Arabia. International Journal of Climatology, 34, 808–826.
Anil, N., Kumar, M.R.R., Sajeev, R. and Saji, P.K. (2016) Role of distinct flavours of IOD events on Indian summer monsoon. Natural Hazards, 82, 1317–1326.
Attada, R., Dasari, H.P., Parekh, A., Chowdary, J.S., Langodan, S., Knio, O. and Hoteit, I. (2019) The role of the Indian summer monsoon variability on Arabian Peninsula summer climate. Climate Dynamics, 52, 3389–3404.
Babu, C.A., Jayakrishnan, P.R. and Varikoden, H. (2016) Characteristics of precipitation pattern in the Arabian Peninsula and its variability associated with ENSO. Arabian Journal of Geosciences, 9, 186. https://doi.org/10.1007/s12517-015-2265-x.
Berner, J., Christensen, H.M. and Sardeshmukh, P.D. (2020) Does ENSO regularity increase in a warming climate? Journal of Climate, 33, 1247–1259.
Blake, D.W., Krishnamurti, T.N., Low-Nam, S.V. and Fein, J.S. (1983) Heat low over the Saudi Arabian desert during May 1979 (summer MONEX). Monthly Weather Review, 9, 1759–1775.
Bou Karam, D., Flamant, C., Knippertz, P., Reitebuch, O., Pelon, J., Chong, M. and Dabas, A. (2008) Dust emissions over the Sahel associated with the West African monsoon intertropical discontinuity region: a representative case-study. Quarterly Journal of the Royal Meteorological Society, 134, 621–634.
Bou Karam, D., Flamant, C., Tulet, P., Chaboureau, J.-P., Dabas, A. and Todd, M.C. (2009a) Estimate of Sahelian dust emissions in the inter-tropical discontinuity region of the West African monsoon. Journal of Geophysical Research, 114, D13106. https://doi.org/10.1029/2008JD011444.
Bou Karam, D., Flamant, C., Tulet, P., Todd, M.C., Pelon, J. and Williams, E. (2009b) Dry cyclogenesis and dust mobilization in the intertropical discontinuity of the West Africa monsoon: a case study. Journal of Geophysical Research, 114, D015115. https://doi.org/10.1029/2008JD010952.
Bou Karam, D., Williams, E., Janiga, M., Flamant, C., MacGraw-Herdeg, M., Cuesta, J., Auby, A. and Thornicroft, C. (2014) Synoptic-scale dust emissions over the Sahara Desert initiated by a moist convective cold pool in early August 2006. Quarterly Journal of the Royal Meteorological Society, 140, 2591–2607. https://doi.org/10.1002/qj2326.
Bou Karam Francis, D., Flamant, C., Chaboureau, J. P., Banks, J., Cuesta, J., Brindley, H. and Oolman, L. (2017) Dust emission and transport over Iraq associated with the summer Shamal winds. Aeolian Research, 24, 15–31. https://doi.org/10.1016/j.aeolia.2016.11.001.
Branch, O., Behrendt, A., Gong, Z., Schwitalla, T. and Wulfmeyer, V. (2020) Convection initiation over the East Arabian Peninsula. Meteorologische Zeitschrift, 29, 67–77.
Buckle, C. (1996) Weather and Climate in Africa. Harlow: Addison-Wesley Longman Ltd.

Chaboureau, J.-P., Flamant, C., Dauhut, T., Kocha, C., Lafore, J.-P., Lavaysse, C., Marnas, F., Mokhtari, M., Pelon, J., Martinez, I. R., Scheppanski, K. and Tulet, P. (2016) Fennec dust forecast intercomparison over the Sahara in June 2011. Atmospheric Chemistry and Physics, 16, 6977–6995.

Chaouch, N., Temimi, M., Weston, M. and Ghedira, H. (2017) Sensitivity of the meteorological model WRF-ARW to planetary boundary layer schemes during fog conditions in a coastal arid region. Atmospheric Research, 187, 106–127.

Chauvin, F., Roehrig, R. and Lafore, J. (2010) Intraseasonal variability of the Saharan heat low and its link with midlatitudes. Journal of Climate, 23, 2544–2561.

Cosneauy, H., Leroy, M. and Briottet, X. (1996) Selection and characterization of Saharan and Arabian desert sites for the calibration of optical satellite sensors. Remote Sensing of the Environment, 58, 101–114.

Dumka, U.C., Kaskaoutis, D.G., Francis, D., Chaboureau, J.-P., Rashki, A., Tiwari, S., Singh, S., Liakokou, E. and Mihalopoulos, N. (2019) The role of the intertropical discontinuity region and the heat low in dust emission and transport over the Thar desert, India: a premonsoon case study. Journal of Geophysical Research: Atmospheres, 124, 13197–13219.

Eager, R.E., Raman, S., Wootten, A., Westphal, D.L., Reid, J.S. and Al Mandoos, A. (2008) A climatological study of the land and sea breezes in the Arabian Gulf region. Journal of Geophysical Research, 113, D15106. https://doi.org/10.1029/2007JD009710.

Engelstaedter, S., Washington, R., Flamant, C., Parker, D.J., Allen, C.J.T. and Todd, M.C. (2015) The Saharan heat low and moisture transport pathways in the Central Sahara – multi-aircraft observations and Africa-LAM evaluation. Journal of Geophysical Research: Atmospheres, 120, 4417–4442.

Fekih, A. and Mohamed, A. (2019) Evaluation of the WRF model on simulating the vertical structure and diurnal cycle of the atmospheric boundary layer over Bordj Badji Mokhtar (southwestern Algeria). Journal of King Saud University – Science, 31, 602–611.

Flamant, C., Chaboureau, J.-P., Parker, D.J., Taylor, C.M., Cammas, J.-P., Bock, O., Timouk, F. and Pelon, J. (2007) Airborne observations of the impact of a convective system on the planetary boundary layer thermodynamics and aerosol distribution in the inter-tropical discontinuity region of the West African monsoon. Quarterly Journal of the Royal Meteorological Society, 133, 1175–1189.

Flamant, C., Knippertz, P., Parker, D.J., Chaboureau, J.-P., Lavaysse, C., Agusti-Panareda, A. and Kergoat, L. (2009) The impact of a mesoscale convective system cold pool on the northward propagation of the intertropical discontinuity over West Africa. Quarterly Journal of the Royal Meteorological Society, 135, 139–159.

Francis, D., Chaboureau, J.-P., Nelli, N., Cuesta, J., Alshamsi, N., Temimi, M., Pauluis, O. and Xue, L. (2020a) Summertime dust storms over the Arabian Peninsula and impacts on radiation, circulation, cloud development and rain. Atmospheric Research, 2020, 105364. https://doi.org/10.1016/j.atmosres.2020.105364.

Francis, D., Fonseca, R., Nelli, N., Cuesta, J., Weston, M., Evan, A. and Temimi, M. (2020b) The atmospheric drivers of the major Saharan dust storm in June 2020. Geophysical Research Letters, 47, e2020GL090102. https://doi.org/10.1029/2020GL090102.

Francis, D., Temimi, M., Fonseca, R., Nelli, N.R., Abida, R., Weston, M. and Wehbe, Y. (2021) On the analysis of a summertime convective event in a hyperarid environment. Quarterly Journal of the Royal Meteorological Society, 147, 501–525.

Grams, C.M., Jones, S.C., Marshall, J.H., Parker, D.J., Haywood, J. M. and Heuveline, V. (2010) The Atlantic inflow to the Saharan heat low: observations and modelling. Quarterly Journal of the Royal Meteorological Society, 136, 125–140.

Hamilton, R.A. and Archibald, J.W. (1945) Meteorology of Nigeria and adjacent territory. Quarterly Journal of the Royal Meteorological Society, 71, 231–265.

Hersbach, H., Bell, B., Berrisford, P., Hirahara, S., Horanyi, A., Munoz-Sabater, J., Nicolas, J., Peubey, C., Radu, R., Schepers, D., Simmons, A., Soci, C., Abdalla, S., Abellan, X., Balsamo, G., Bachtold, P., Biavati, G., Bidlot, J., Bonavita, M., De Chiara, G., Dahlgren, P., Dee, D., Diamantakis, M., Dragani, R., Fleming, J., Forbes, R., Fuentes, M., Geer, A., Haimberger, L., Healy, S., Hogan, R.I., Holm, E., Janiskova, M., Keeley, S., Laloyaux, P., Lopez, P., Lupu, C., Radnoti, G., de Rosnay, P., Rozum, I., Vamborg, F., Villaume, S. and Thepaut, J.-N. (2020) The ERA5 global reanalysis. Quarterly Journal of the Royal Meteorological Society, 146, 1999–2049.

Hruby, P.H., Varikoden, H. and Vishnu, R. (2020) A review on the Indian summer monsoon rainfall variability and its association with ENSO and IOD. Meteorology and Atmospheric Physics, 133, 1–14. https://doi.org/10.1007/s00703-020-00734-5.

Huang, B., Thorne, P.W., Banzon, V.F., Boyer, T., Chepurin, G., Lawrimore, J.H., Menne, M.J., Smith, T.M., Vose, R.S. and Zhang, H.-M. (2017) Extended reconstructed sea surface temperature, version 5 (ERSSTv5): updates, validations, and intercomparisons. Journal of Climate, 30, 8179–8205.

Iacono, M.J., Delamere, J.S., Mlawer, E.J., Shephard, M.W., Clough, S.A. and Collins, W.D. (2008) Radiative forcing by long-lived greenhouse gases: calculations with the AER radiative transfer models. Journal of Geophysical Research, 113, D13103. https://doi.org/10.1029/2008JD009944.

Kain, J.S. (2004) The Kain-Fritsch convective parameterization: an update. Journal of Applied Meteorology, 43, 170–181.

Kalapureddy, M.C.R., Lothion, M., Campistron, B., Lohou, F. and Said, F. (2010) Wind profiler analysis of the African easterly jet in relation with the boundary layer and the Saharan heat low. Quarterly Journal of the Royal Meteorological Society, 136, 77–91.

Knippertz, P. (2008) Dust emissions in the West African heat trough – the role of the diurnal cycle and of extratropical disturbances. Meteorologische Zeitschrift, 17, 553–563.

Konda, G., Chowdary, J.S., Srinivas, G., Gnanaseelan, C., Parekh, A., Attada, R. and Krishna, R. (2018) Tropospheric biennial oscillation and south Asian summer monsoon rainfall in a coupled model. Journal of Earth System Sciences, 127, 46. https://doi.org/10.1007/s12040-018-0948-x.

Kumar, S.P., Roshim, R.P., Narvekar, J., Kuumr, D. and Vivekanandan, E. (2009) Response of the Arabian Sea to global warming and associated regional climate shift. Marine Environmental Research, 68, 217–222.

Lafore, J.-P., Chapelon, N., Beucher, F., Diop-Kane, M., Gaymard, A., Kasimou, A., Lepape, S., Mumbai, Z., Orji, B., Osika, D.,
Parker, D.J., Poan, E., Razafindrakoto, L.G., Vincendon, J.C. and ACMAD and the West Africa Forecasting Community. (2017) West African synthetic analysis and forecast. In: Parker, D.J. and Diop-Kane, M. (Eds.) Meteorology of Tropical West Africa. Hoboken, New Jersey, USA: Wiley Blackwell, https://doi.org/10.1002/9781118391297.ch11.

Niranjan Kumar, K. and Ouarda, T.B.M.J. (2014) Precipitation variability over UAE and global SST teleconnections. Journal of Geophysical Research: Atmospheres, 119, 10313–10322. https://doi.org/10.1002/2014JD021724.

Niu, G.-Y., Yang, Z.L., Mitchell, K.E., Chen, F., Ek, M.B., Barlage, M., Kumar, A., Manning, K., Niyogi, D., Rosero, E., Tewari, M. and Xia, Y. (2011) The community Noah land surface model with multiparameterization options (Noah-MP): 1. Model description and evaluation with local-scale measurements. Journal of Geophysical Research, 116, D12109. https://doi.org/10.1029/2010JD015139.

Noori, R., Tian, F., Berndtsson, R., Abbasi, M.R., Naseh, M.V., Modabberi, A., Soltani, A. and Klove, B. (2019) Recent and future trends in sea surface temperature across the Persian Gulf and Gulf of Oman. PLoS One, 14, e0212790. https://doi.org/10.1371/journal.pone.0212790.

Odekonu, T.O. (2010) An assessment of the influence of the intertropical discontinuity on inter-annual rainfall characteristics in Nigeria. Geophysical Research, 48, 314–326.

Odekonu, T.O. and Adejuwon, S.A. (2017) Assessing changes in the rainfall regime in Nigeria between 1961 and 2004. Geojournal, 70, 145–159.

Parker, D.J. and Diop-Kane, M. (2017) Meteorology of Tropical West Africa: The Forecasters’ Handbook. Hoboken, New Jersey, USA: Wiley Blackwell 468 pp.

Poseichal, B., Bou Karam, D., Crewell, S., Flamant, C., Hunerbein, A., Bock, O. and Said, F. (2010) Diurnal cycle of the intertropical discontinuity over West Africa analysed by remote sensing and mesoscale modelling. Quarterly Journal of the Royal Meteorological Society, 136, 92–106.

Racz, Z. and Smith, R.K. (1999) The dynamics of heat lows. Quarterly Journal of the Royal Meteorological Society, 125, 225–252. https://doi.org/10.1002/qj.49712555313.

Rashki, A., Kaskaoutis, D.G., Mofidi, A., Minvielle, F., Chiapello, I., Legrand, M., Dumka, U.C. and Francois, P. (2019) Effects of monsoon, Shamal and Levar winds on dust accumulation over the Arabian Sea during summer – the July 2016 case. Aeolian Research, 36, 27–44.

Redelsperger, J.-L., Diongue, A., Diehiou, A., Ceron, J.-P., Diop, M., Gueremy, J.-F. and Lefore, J.-P. (2002) Multi-scale description of a Sahelian synoptic weather system representative of the West African monsoon. Quarterly Journal of the Royal Meteorological Society, 128, 1229–1257.

Roehrig, R., Chauvin, F. and Lefore, J. (2011) 10-25 day intraseasonal variability of convection over the Sahel: a role of the Saharan heat low and midlatitudes. Journal of Climate, 24, 5863–5878.

Saji, N.H., Goswami, B.N., Vinayachandran, P.N. and Yamagata, T. (1999) A dipole mode in the tropical Indian Ocean. Nature, 401, 360–363.

Schott, F.A., Xie, S.-P. and McCreary, J.P., Jr. (2009) Indian Ocean circulation and climate variability. Reviews of Geophysics, 47, RG1002. https://doi.org/10.1029/2007RG000245.

Schwitalia, T., Branch, O. and Wulfmeyer, V. (2020) Sensitivity study of the planetary boundary layer microphysical schemes to the initialization of convection over the Arabian Peninsula. Quarterly Journal of the Royal Meteorological Society, 146, 846–869.

Skamarock, W.C., Klemp, J.B., Dudhia, J., Gill, D.O., Barker, D.M., Duda, M.G., Huang, X.-Y., Wang, W. and Powers, J.G. (2008) A
description of the Advanced Research WRF version 3, 113 pp. https://doi.org/10.5065/D68S4MVH.

Smith, E.A. (1986a) The structure of the Arabian heat low. Part I: surface energy budget. *Monthly Weather Review*, 114, 1067–1083.

Smith, E.A. (1986b) The structure of the Arabian heat low. Part II: bulk tropospheric heat budget and implications. *Monthly Weather Review*, 114, 1084–1102.

Spinks, J., Lin, Y.-L. and Mekonnen, A. (2015) Effects of the subtropical anticyclone over North Africa and Arabian Peninsula on the Africa easterly jet. *International Journal of Climatology*, 35, 733–745.

Srivastava, G., Chakraborty, A. and Nanjundiah, R.S. (2020) Multidecadal variations in ENSO-Indian summer monsoon relationship at sub-seasonal timescales. *Theoretical and Applied Climatology*, 140, 1299–1314.

Steinhoff, D.F., Bruintjes, R., Hacker, J., Keller, T., Williams, C., Jensen, T., Al Mandous, A. and Al Yazeedi, O.A. (2018) Influences of the monsoon trough and Arabian heat low on summer rainfall over the United Arab Emirates. *Monthly Weather Review*, 146, 1383–1403.

Sultan, B. and Janicot, S. (2003) The West African monsoon dynamics. Part II: the “preonset” and “onset” of the summer monsoon. *Journal of Climate*, 16, 3407–3427.

Sultan, B., Janicot, S. and Drobinski, P. (2007) Characterization of the diurnal cycle of the West African monsoon around the monsoon onset. *Journal of Climate*, 20, 4014–4032.

Sunkara, S.L. and Tiwari, R.K. (2016) Wavelet analysis of the singular spectral reconstructed time series to study the impacts of solar-ENSO-geomagnetic activity on Indian climate. *Nonlinear Processes in Geophysics*, 23, 361–374.

Temimi, M., Fonseca, R., Nelli, N., Weston, M., Thota, M., Valappil, V., Branch, O., Wizemann, H.-D., Kondapalli, N.K., Wehbe, Y., Al Hosary, T., Shalaby, A., Al Shamsi, N. and Al Naqb, H. (2020) Assessing the impact of changes in land surface conditions on WRf predictions in arid regions. *Journal of Hydro meteorology*, 21, 1–60. https://doi.org/10.1175/JHM-D-20-0083.1.

Thompson, G. and Eidhammer, T. (2014) A study of aerosol impacts on clouds and precipitation development in a large winter cyclone. *Journal of the Atmospheric Sciences*, 71, 3636–3658.

Wang, H., Kumar, A., Murtugudde, R., Narapsetty, B. and Seip, K. L. (2019) Covariations between the Indian Ocean dipole and ENSO: a modeling study. *Climate Dynamics*, 53, 5743–5761.

Wang, W., Evan, A.T., Flamant, C. and Lavaysse, C. (2015) On the decadal scale correlation between African dust and Sahel rainfall: the role of Saharan heat low-forced winds. *Science Advances*, 9, e1500646. https://doi.org/10.1126/sciadv.1500646.

Wang, W., Evan, A.T., Lavaysse, C. and Flamant, C. (2017) The role of the Saharan heat low plays in dust emission and transport during summertime in North Africa. *Aeolian Research*, 28, 1–12.

Wehbe, Y., Temimi, M. and Adler, R.F. (2020) Enhancing precipitation estimates through the fusion of weather radar, satellite retrievals, and surface parameters. *Remote Sensing*, 12, 1342. https://doi.org/10.3390/rs12081342.

Wehbe, Y., Temimi, M., Ghebreyesus, D.T., Milewski, A., Norouzi, H. and Ibrahim, E. (2018) Consistency of precipitation products over the Arabian Peninsula and interactions with soil moisture and water storage. *Hydrological Sciences Journal*, 63, 408–425.

Weller, E., Jakob, C. and Reeder, M.J. (2019) Understanding the dynamic contribution to future changes in tropical precipitation from low-level convergence lines. *Geophysical Research Letters*, 46, 2196–2203.

Weston, M., Chaouch, N., Valappil, V., Temimi, M., Ek, M. and Zheng, W. (2018) Assessment of the sensitivity to the thermal roughness length in Noah and Noah-MP land surface model using WRF in an arid region. *Pure and Applied Geophysics*, 175, 1–17.

Williams, E.R. (2008) Comment on “Atmospheric controls on the annual cycle of North African dust” by S. Engelstaedter and R. Washington. *Journal of Geophysical Research*, 113, D23109. https://doi.org/10.1029/2008JD009930.

Xue, P. and Eltahir, E.A.B. (2015) Estimation of the heat and water budgets of the Persian (Arabian) Gulf using a regional climate model. *Journal of Climate*, 28, 5041–5062.

Yang, Z.-L., Mitchell, K.E., Chen, F., Ek, M.B., Barlage, M., Kumar, A., Manning, K., Niyogi, D., Rosero, E., Tewari, M. and Xia, Y. (2011) The community Noah land surface model with multiparameterization options (Noah-MP): 2. Evaluation over global river basins. *Journal of Geophysical Research*, 116, D12110. https://doi.org/10.1029/2010JD015140.

Yu, Y., Notaro, M., Kalashnikova, O.V. and Garay, M.J. (2016) Climatology of summer Shamal wind in the Middle East. *Journal of Geophysical Research: Atmospheres*, 121, 289–305.

**SUPPORTING INFORMATION**

Additional supporting information may be found online in the Supporting Information section at the end of this article.

**How to cite this article:** Fonseca, R., Francis, D., Nelli, N. & Thota, M. (2022). Climatology of the heat low and the intertropical discontinuity in the Arabian Peninsula. *International Journal of Climatology*, *42*(2), 1092–1117. https://doi.org/10.1002/joc.7291