See also

Climate Variability: Seasonal to Interannual Variability. **Coriolis Force**. **Cyclogenesis**. **Dynamic Meteorology:** Overview; Waves. **Stationary Waves (Orographic and Thermally Forced)**. **Stratosphere–Troposphere Exchange:** Global Aspects.

Further Reading

James IN (1994) *Introduction to Circulating Atmospheres*. Cambridge: Cambridge University Press.

Wallace JM and Gutzler DS (1981) Teleconnections in the geopotential height field during the Northern Hemispere winter. *Monthly Weather Review* 109: 784–812.

THERMAL LOW

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Introduction

A thermal low (sometimes referred to as a heat low) is a low-pressure area resulting from high temperatures in the lower troposphere caused by a localized area of intense heating at the Earth's surface. Thermal lows occur typically during the summer over subtropical continental areas and are most intense in the desert regions of the world. These regions are characterized by clear skies and a lack of vegetation. Consequently, there is a large diurnal cycle of surface heating, which in turn creates a pronounced diurnal cycle in the intensity of thermal lows, with a maximum intensity (i.e., minimum surface pressure) during the afternoon. Because thermal lows are linked directly to surface properties, they are nonmigratory in nature. Moreover, they do not exhibit any frontal characteristics, nor are clouds or precipitation associated with them.

Since thermal lows arise from surface heating, the maximum amplitudes of temperature and circulation anomalies associated with them are confined to the lower troposphere (i.e., below 5 km or 500 hPa). Most of the desert areas of the world that exhibit strong thermal lows are surrounded by bodies of water. As a result, horizontal heating gradients that develop generate sea and land breezes that influence the thermal-low circulations. In regions away from the equator, the low-level inflow associated with the daytime sea breeze produces, through the action of the Earth's rotation (the Coriolis force), a cyclonic vorticity anomaly in the lower troposphere. The converging air at low levels in the thermal low also produces upward motion in the lower troposphere, but sinking motion occurs aloft. The cyclonic circulation often persists into the nighttime hours above the surface following the development of a shallow, nocturnal temperature inversion. By the morning, subsidence extends all the way down to the surface.

Because of the clear-sky conditions and high surface reflectance (albedo) of the desertlike areas where thermal lows persist, these regions have often been regarded as large-scale radiative energy sinks. However, in some thermal low regions dust is raised from the surface by the intense heating which results in short-wave absorption that makes at least a portion of the thermal low region a radiative energy source.

Geographical Distribution

The strongest thermal lows are located over the great deserts of the world, e.g., the Sahara, Arabian, Kalahari, Australian Great Western Desert, and Mojave/Sonoran Deserts. In these regions the thermal lows are so strong that they appear as closed lows or troughs in the mean-sea-level pressure maps of the summer hemispheres (**Figure 1**). For example, note the low-pressure troughs over the African Sahara, the Indian subcontinent, and the Southwest US in July and over northern Australia in January. The trough over southern Asia near India in July is closely associated with cross-equatorial flow over the western Indian Ocean and the Asian summer monsoon.

The locations of thermal lows also correspond to regions of intense surface heating, as can be seen from a map of the global distribution of surface sensible heat flux (**Figure 2**). The lack of cloudiness, vegetation, and surface moisture in these regions accounts for the large values of surface sensible heat flux. Maximum values upwards of 60–70 $\mathrm{W\,m^{-2}}$ can be seen in the regions of the major deserts. The close relationship between the thermal lows and surface sensible heat flux patterns indicates a close physical linkage between these two features. Specifically, intense, localized heating warms the lower atmosphere and hydrostatistically reduces the surface pressure in that region.

The wind vectors illustrated in **Figure 1** show generally confluent surface flow into the thermal low-pressure areas. This flow arises from strong horizontal gradients in the surface sensible heat flux between the desert areas and the surrounding oceans

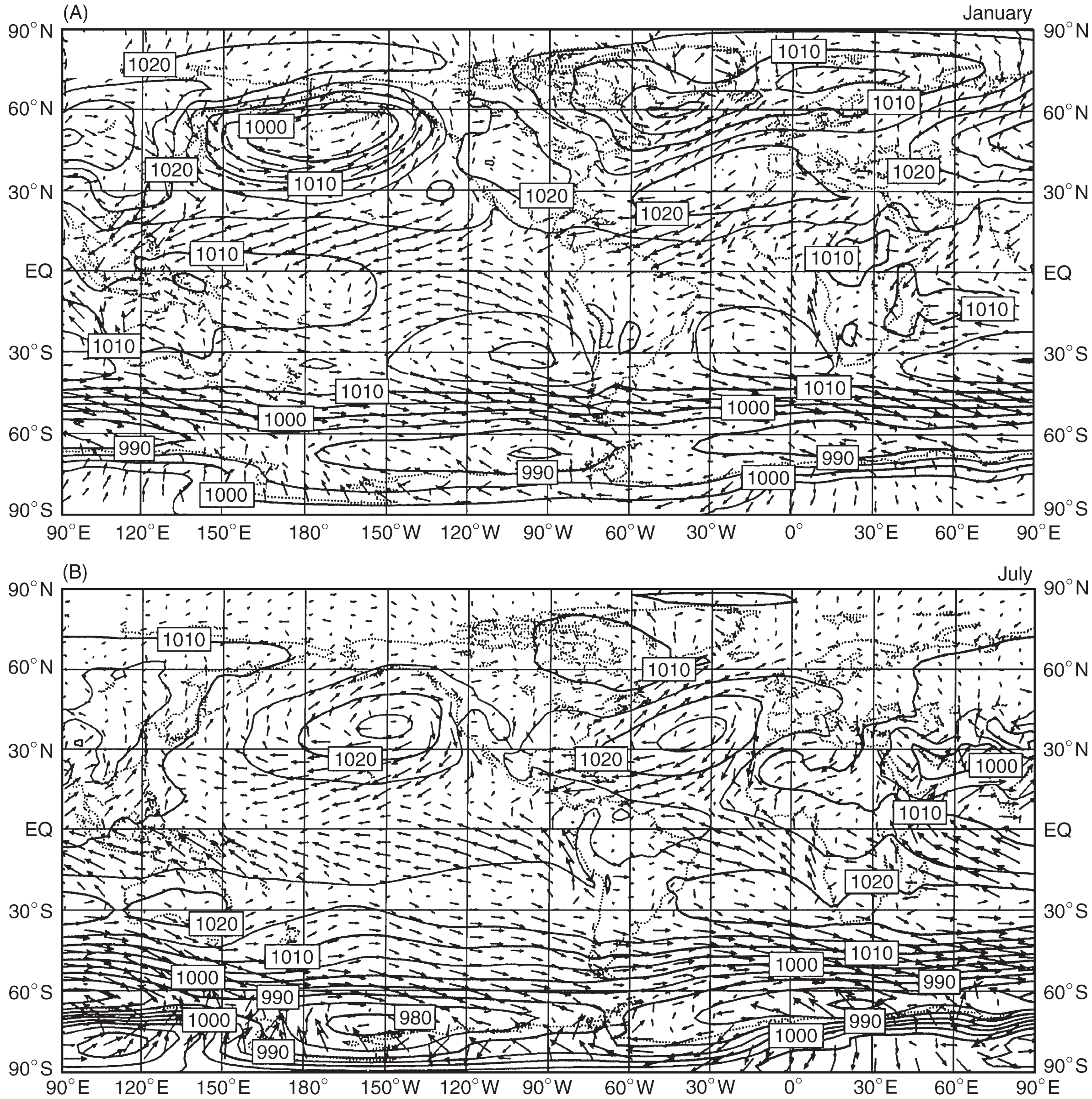

Figure 1 Maps of mean sea-level pressure for January and July. Wind vectors for the 1000 hPa level are superimposed. Data are 1980–87 analyses from a forecast model. The contour interval is 5 hPa and the largest vector represents a wind speed of 12 m s^{-1}. (Reproduced with permission from Hartmann DL (1994) *Global Physical Climatology*. San Diego, CA: Academic Press.)

(sensible heat fluxes over the ocean are typically 10 W m^{-2} or less), which drives sea breeze circulations during the daytime hours. At night, a land breeze or offshore surface flow develops, but the intense heating causes the sea breeze to dominate over the land breeze circulation.

Vertical Structure and Energetics

Thermal lows generally develop over arid lands or deserts in the summertime. **Figure 3** depicts the thermodynamic structure of the troposphere within a thermal low. The case shown represents a synthesis of data from research aircraft flights over the Arabian desert, although it is typical of other desert regions. To illustrate mixing processes in the atmosphere, vertical profiles of two conserved thermodynamic quantities for dry atmospheric motions, the specific humidity q and the potential temperature θ, are plotted in **Figure 3**. From the surface to about 5 km or 550 hPa, θ is approximately constant and q decreases with height. This structure is characteristic of deep, continental atmospheric boundary layers containing vigorous, dry

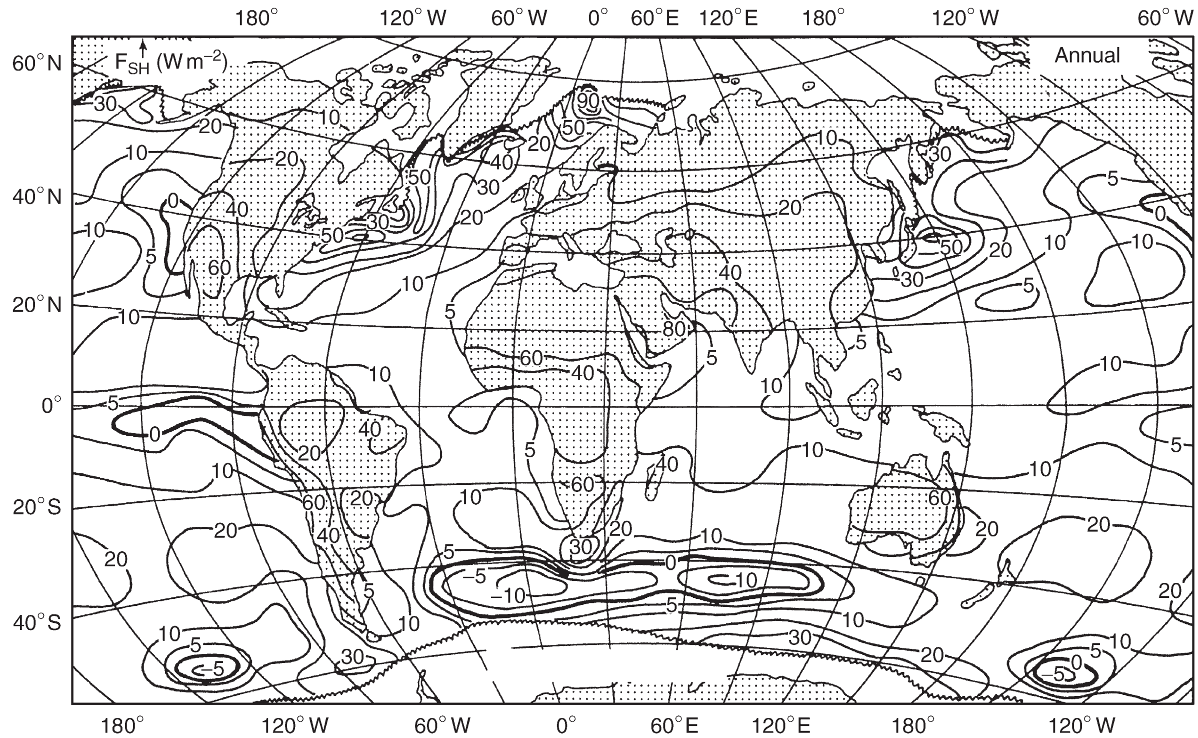

Figure 2 Global distribution of the sensible heat flux from the earth's surface into the atmosphere in $W m^{-2}$ for annual-mean conditions. (Reproduced with permission from Peixoto JP and Oort AH (1992) *Physics of Climate*. New York: American Institute of Physics.)

convective plumes or thermals. The sharp vertical gradients of q and θ near 5 km represent a transition zone between turbulent air in the atmospheric mixed layer below and laminar flow in the free atmosphere above. The vertical gradients in q and θ in the upper part of the mixed layer are a result of entrainment of drier (lower q) and potentially warmer (higher-θ) air from above into the mixed layer. At night, radiational cooling acts to decrease θ in the lowest kilometer, creating a nocturnal inversion.

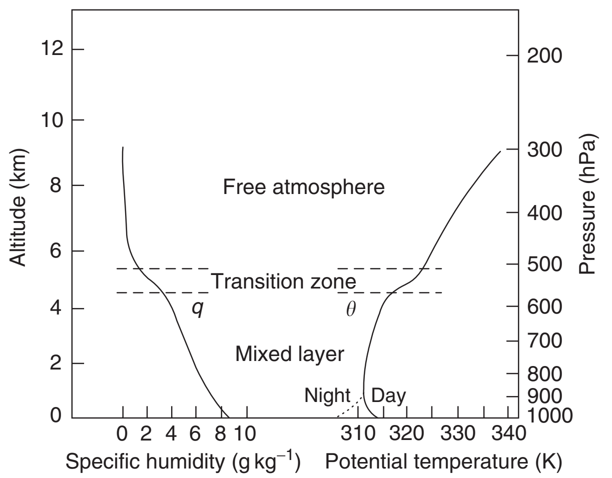

Figure 3 Typical vertical profiles of specific humidity q and potential temperature θ within a thermal low extending from the surface through the mixed layer, the transition zone, and into the free atmosphere. (Adapted with permission from Smith EA (1986) The structure of the Arabian heat low part II: Bulk tropospheric heat budget and implications. *Monthly Weather Review* 114: 1084–1102.)

Thermal lows typically exist in regions of large-scale subsidence. However, owing to the development of sea breezes during the daytime, the vertical motion in the lowest levels can become upward during the day. This behavior can be seen in **Figure 4**, which contains profiles of vertical motion at day and night over the Arabian desert. Above 800 hPa there is sinking motion during both day and nighttime hours, although it is weaker during the day owing to short-wave radiative heating of the surface and atmosphere.

The vertical profiles of vertical motion in **Figure 4** imply horizontal divergence in the lower to middle troposphere. This divergence is consistent with heating in the lower troposphere that lifts isobaric surfaces above the heat source and produces a midlevel high and coincident outflow of air. This outflow in turn reduces the surface pressure and assists in the inflow of air into the surface low. Part of the outflowing air aloft during the daytime represents the return flow of the sea breeze circulation.

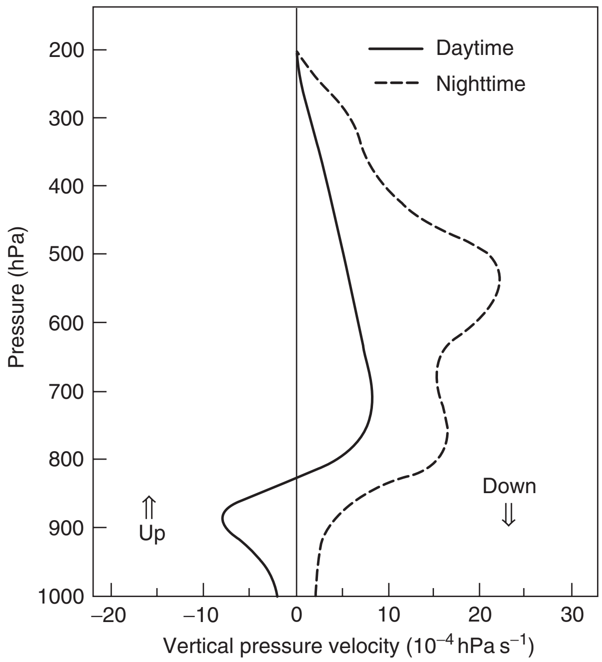

Figure 4 Daytime and nighttime vertical motion profiles (vertical *p*-velocity) over the Arabian heat low. The positive values indicate downward motion. (Adapted with permission from Smith EA (1986) The structure of the Arabian heat low, part II: Bulk tropospheric heat budget and implications. *Monthly Weather Review* 114: 1084–1102.)

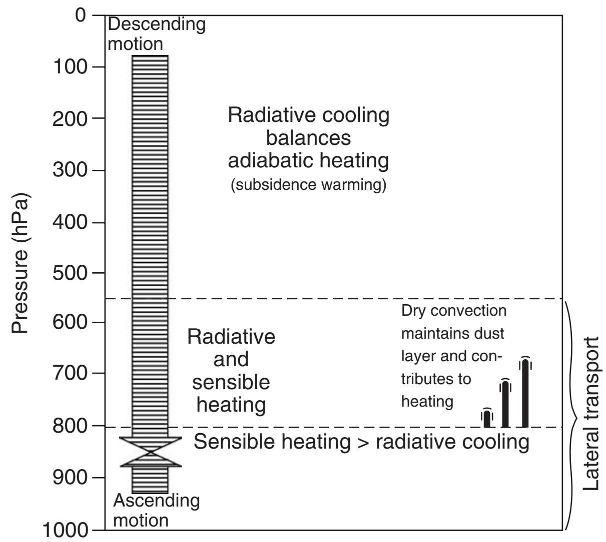

Figure 5 Conceptual three-layer structure of the daytime Arabian heat low. (Adapted with permission from Smith EA (1986) The structure of the Arabian heat low, part II: Bulk tropospheric heat budget and implications. *Monthly Weather Review* 114: 1084–1102.)

The vertical structure of thermal lows can also be considered from an energy budget perspective. Field studies of thermal lows show that the raising of dust by daytime heating makes their energetics rather complex. To illustrate this complexity, the vertical structure of the energy balance of the Arabian heat low is shown in **Figure 5**. This heat low can be characterized as a three-layer system. The lower atmosphere forms a deep mixed layer from the surface to 550 hPa. An upper layer from 550 hPa to the upper boundary is a region where radiative cooling is approximately balanced by adiabatic heating (subsidence warming). A middle layer from 550 to 800 hPa undergoes both sensible and radiative heating. The sensible heating arises from the convergence of eddy heat flux owing to mixed-layer turbulence. The radiative heating results from enhanced daytime shortwave absorption due to a substantial aerosol or dust loading. The dust is generated locally by daytime vigorous boundary layer thermals and associated gusty surface winds. In the lower layer (surface to 850 hPa) the convergence of eddy heat fluxes dominates radiative cooling, resulting in a net lower-tropospheric warming.

The processes illustrated in **Figure 5** indicate that the middle and lower troposphere undergoes a net energy gain due to diabatic processes. The only mechanism to balance this energy gain is a lateral transport of energy out of the region. For the Arabian heat low, the main lateral transport is over the Arabian Sea, where the warm air serves to maintain the West Arabian Sea inversion. This inversion acts to trap water vapor in the lower troposphere as it is carried toward India in the strong south-west summer monsoon flow. This is one mechanism by which thermal lows can have an important influence on surrounding weather patterns.

Dynamics and the Diurnal Cycle

Direct observations indicate that a low-level cyclonic circulation develops typically within thermal lows with a maximum amplitude during the daytime hours. Observations are rather sparse, however, and details of the diurnal cycle of this circulation have not been well documented. Therefore, numerical modeling studies have been used to provide further insight into the dynamics of thermal lows.

Numerical simulations of thermal lows over land surrounded by an ocean show a pronounced diurnal cycle in the circulation patterns. In fact, the circulation cannot be understood without consideration of the diurnal cycle. Simulations show that while the thermal low has a surface pressure minimum in the late afternoon following strong solar heating of the land, the relative vorticity is strongest in the early morning hours as a result of a prolonged period of low-level

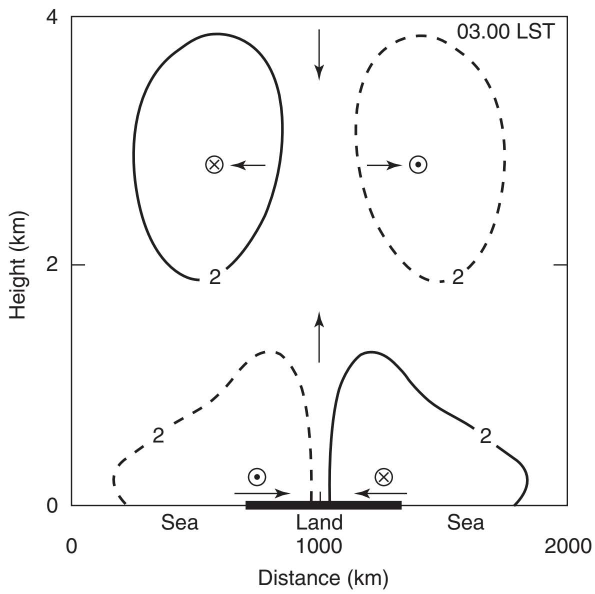

Figure 6 Vertical west–east cross-section at 03.00 LST through a thermal low that has formed over a 600×600 km square land area surrounded by ocean in the Northern Hemisphere. Contours indicate meridional (N–S) wind (m s^{-1}). $\otimes$ indicates the center of the maximum southerly flow, $\odot$ the center of the maximum northerly flow. The horizontal arrows denote the centers of maximum zonal (E–W) flow and the vertical arrows the centers of maximum vertical motion. The heavy line indicates the land area. (Adapted with permission from Rácz Z and Smith RK (1999) The dynamics of heat lows. *Quarterly Journal of the Royal Meteorological Society* 125: 225–252.)

convergence. Thus the thermal low is not approximately in quasi-geostrophic balance.

A depiction of the zonal, meridional, and vertical velocity components at 03.00 LST for an idealized simulation of a thermal low over a 600 km $\times$ 600 km land area surrounded by ocean at 20° N is shown in **Figure 6**. Low-level convergence, initially developed in connection with a daytime sea breeze, is seen in the early morning hours to evolve into a nocturnal low-level jet. This jet develops as a result of strong nocturnal surface cooling over land. The earth's rotation acting on this circulation generates a cyclonic circulation at low levels, evident in the meridional wind field. The sea breeze return flow aloft is horizontally divergent and it generates an anticyclonic circulation at upper levels, again evident in the meridional wind field. Rising motion occurs at low levels and sinking motion aloft, consistent with the daytime pattern illustrated in **Figure 4**. The vertical motion over the land eventually reverts to sinking at all levels by 06.00 LST, as in the nighttime curve in **Figure 4**.

The dynamics of thermal lows can also be considered from a potential vorticity perspective. Observations of the summertime thermal low over the Iberian Peninsula (Spain) show that as an unstable lapse rate forms in the afternoon, a negative potential vorticity anomaly develops over the peninsula. This negative anomaly exists within a large-scale environment of positive potential vorticity. At night this negative anomaly disappears as stable air develops near the surface due to nocturnal cooling. It has been postulated that the dome of negative potential vorticity associated with the thermal low over the Iberian Peninsula can act to inhibit the development of nearby Algerian lows by increasing the effective interaction distance between upper-level and low-level potential vorticity anomalies.

See also

Aerosols: Role in Radiative Transfer. **Boundary Layers:** Surface Layer. **Coriolis Force**. **Deserts and Desertification**. **Diurnal Cycle**. **Dust**. **Energy Balance Model, Surface**. **Land–Sea Breeze**. **Reflectance and Albedo, Surface**.

Further Reading

Griffiths JF and Soliman KH (1972) The northern desert. In: Griffiths JF (ed.) *World Survey of Climatology, vol. 10. Climates of Africa*, pp. 75–111. New York: Elsevier.

Hartmann DL (1994) *Global Physical Climatology*. San Diego, CA: Academic Press.

Peixoto JP and Oort AH (1992) *Physics of Climate*. New York: American Institute of Physics.

Rácz Z and Smith RK (1999) The dynamics of heat lows. *Quarterly Journal of the Royal Meteorological Society* 125: 225–252.

Ramage CS (1971) *Monsoon Meteorology*. New York: Academic Press.

Smith EA (1986) The structure of the Arabian heat low, pt II. Bulk tropospheric heat budget and implications. *Monthly Weather Review* 114: 1084–1102.