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The meteorology of monsoons

1.1 Introduction

Monsoon circulations are major features of the tropical atmosphere, which, primarily through the rainfall associated with them, are of profound importance to a large fraction of the world's population. While there is no universally accepted definition of what constitutes a monsoon, there are some criteria that are widely accepted (see, e.g., the discussions in Ramage (1971), Webster (1987), and Neelin (2007)). Fundamentally, monsoonal climates are found where a tropical continent lies poleward of an equatorial ocean and are characterized by a strong seasonal cycle, with dry winters and very wet summers, and a reversal of wind direction from, in the dry season, the equatorward–easterly flow that is typical of most of the tropics to poleward–westerly flow after monsoon onset. Low-level flow from the ocean imports moisture onto the land to supply the rainfall there (although much of the rainfall within the monsoon system as a whole may actually fall over the neighboring ocean). In fact, in most monsoon systems this inflow includes strong cross-equatorial flow at low levels, from the winter to the summer hemisphere; however, this is not satisfied in all cases (such as the North American monsoon; Neelin (2007)). Indeed, given the differences in detail between different monsoon systems, even though they satisfy the most obvious criteria, it is inevitable that any attempt at definition will be imprecise, and even that classification of some regional meteorological regimes as monsoons may not be universally accepted.

The Asian–Indian Ocean–Australian monsoon system is, by some way, the most dramatic on the planet in terms of its intensity and spatial extent, but there are other regions of the globe, specifically North and Central America, and West Africa, that display similar characteristics and are thus classified as monsoons. It is important to recognize at the outset that, despite these regional classifications, the monsoons form part of the planetary-scale circulation of the tropical atmosphere: they are influenced by, and in turn influence, the global circulation. Accordingly, we shall begin this overview with a brief review of the "big picture" of the tropical circulation, which will lead into a more focussed discussion of the Asian–Indian Ocean–Australian monsoon system.

1.2 Meteorology of the tropics

1.2.1 *Observed zonal mean picture*

A good starting point for understanding the general circulation of the global atmosphere is to look at the zonally (i.e., longitudinally) averaged circulation in the meridional (latitude-height) plane. Since the circulation varies seasonally (an essential fact of monsoon circulations) it is better to look at seasonal, rather than annual, averages. In turn, the atmosphere exhibits interannual variability – it is a matter of basic experience that one year's weather differs from the last, and this is especially true in the tropics – and so, in a general overview such as this, we shall look not at individual years, but at *climatological* averages, i.e., averages over many summers or winters, which show the normal picture for that season.

Figure 1.1 shows the climatological distribution of mean zonal wind and temperature for the two solstice seasons DJF (December through February) and JJA (June through August). The dominant features of the zonal wind distribution are two westerly subtropical jets straddling the equator at altitudes of about 12 km (near 200 hPa pressure). The core of the stronger jet is located at about 30° latitude in the winter hemisphere, while that of the weaker jet is at 40–50° latitude in the summer hemisphere. Within the deep tropics, the zonal wind is easterly, though mostly weak, all the way down to the surface. Outside the tropics, at latitudes greater than about 30°, the mean surface winds are westerly.

Several features of the mean temperature distribution are worthy of note. Temperature generally decreases rapidly through the troposphere up to the tropopause whose mean altitude varies from about 17 km in the tropics down to around 8 km (near 400 hPa) at the poles. Above, temperature increases, or decreases more slowly, with altitude through the stratosphere. As will be seen in Figure 1.3, almost all atmospheric water (along with most dynamical processes relevant to surface weather and climate) is located in the troposphere. Within the troposphere, temperature decreases systematically poleward from a broad maximum centered in the summer tropics. Note, however, the weak temperature gradients between the two subtropical jets, which contrast with the strong gradients in middle latitudes, poleward of the jet cores.

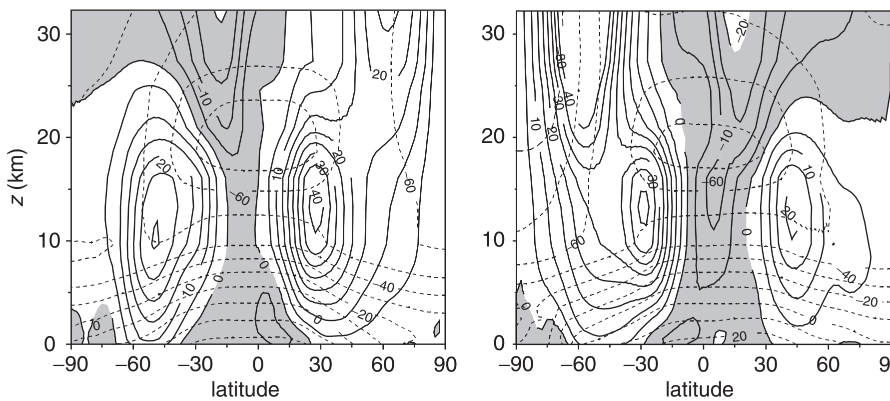

Figure 1.1 Climatological zonal mean zonal wind (solid; ms^{-1}) and temperature (dashed; °C) for (left) December–February and (right) June–August. Contour intervals are 5 ms^{-1} and 10°, respectively; easterly winds are shaded. The data are averaged on pressure surfaces; the height scale shown is representative. Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.

Continuity of mass requires that the zonal mean circulation in the meridional plane be closed, so that northward and vertical motions are directly linked. A convenient way to display the meridional circulation on a single plot is to show the mass streamfunction χ, which is done in Figure 1.2 for the two solstice periods. The mean northward and upward velocities (v,w) are related to the mass streamfunction χ through

$$v = \frac{-1}{2\pi\rho a\cos\varphi}\frac{\partial\chi}{\partial z}; \qquad w = \frac{1}{2\pi\rho a^2\cos\varphi}\frac{\partial\chi}{\partial\varphi},$$

where ρ is the density, a is the Earth's radius and ϕ the latitude. The velocities are thus directed along the χ contours, with mass flux inversely proportional to the contour spacing. In this plane, the mean circulation is almost entirely confined to the tropics. This tropical cell is known as the Hadley circulation, with upwelling over and slightly on the summer side of the equator, summer-to-winter flow in the upper troposphere, downwelling in the winter subtropics, and winter-to-summer flow in the lower troposphere. The latitude of the poleward edge of the cell coincides with that of the winter subtropical jet. There is a much weaker, mirror-image, cell on the summer side of the equator. Around the equinoxes, the structure is more symmetric, with upwelling near the equator and downwelling in the subtropics of both hemispheres.

The distribution of atmospheric moisture is shown in Figure 1.3. Humidity is expressed in two forms: *specific humidity*, the amount of water vapor per unit

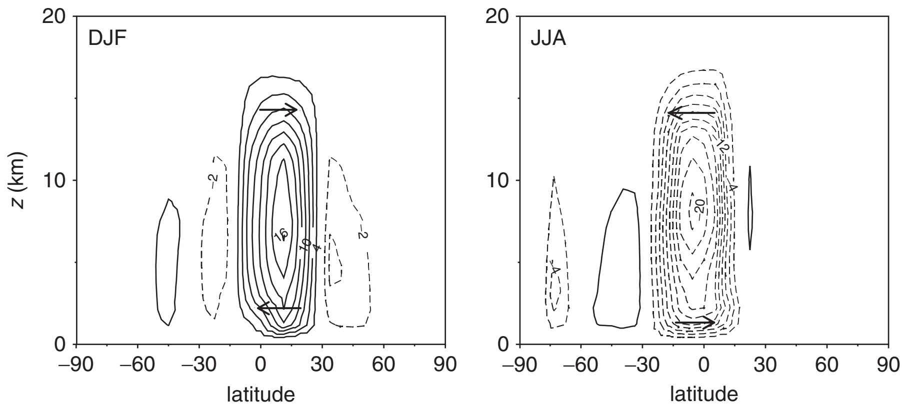

Figure 1.2 Climatological zonal mean overturning streamfunction (10 kg s^{-1} for December–February (left) and June–August (right). Solid contours denote positive values, dashed contours are negative; the zero contour is not plotted. The meridional flow is directed along the streamfunction contours, clockwise around positive cells, anticlockwise around negative cells, as indicated for the dominant cells by the arrows on the plots. The magnitude of the net mass circulation around each cell is equal to the value of the streamfunction extremum in the cell. Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.

mass of air, conventionally expressed as g kg^{-1}, and *relative humidity*, the ratio of specific humidity to its saturation value (the value in equilibrium with liquid water at the ambient temperature and pressure). On this zonally and climatologically averaged view, the near-surface relative humidity varies remarkably little across the globe, being mostly between 65 and 85%. The driest surface regions are near the poles, and in the desert belt of the subtropics. There is a general decrease of relative humidity with height, a consequence of the drying effects of precipitation in updrafts followed by adiabatic descent; the regions of subsidence on the poleward flanks of the Hadley circulation are particularly undersaturated. The zonally averaged specific humidity is as large as 17 g kg^{-1} near the surface just on the summer side of the equator, decaying to less than 1 g kg^{-1} in high latitudes and in the upper troposphere and above. Indeed, the variation of specific humidity is much greater than that of relative humidity, indicating that the former primarily reflects variations of saturation vapor pressure, which has a very strong dependence on temperature (expressed as the Clausius–Clapeyron relationship; see, e.g., Bohren and Albrecht (1998)). Thus, the highest specific humidities are found where the atmosphere is warmest: at low altitudes in the tropics.

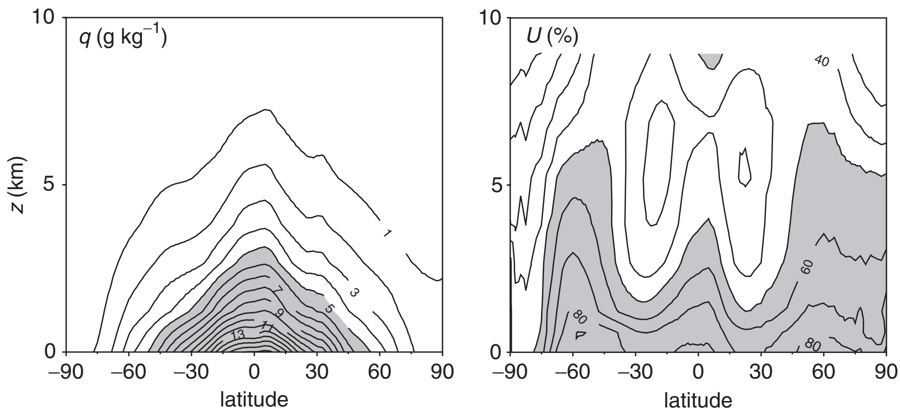

Figure 1.3 Climatological annual- and zonal-mean specific humidity (left, $g\,kg^{-1}$; values greater than $5\,g\,kg^{-1}$ are shaded) and relative humidity (right, %; values greater than 50% are shaded). Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.

1.2.2 *Dynamical and thermodynamical constraints on the circulation*

At first sight, some of the characteristics of the zonally averaged atmosphere may seem puzzling. Ultimately, what drives the atmospheric circulation is the spatial variation of the input of solar energy (per unit surface area) into the atmosphere, which generally decreases monotonically from a maximum in the summer tropics to minima at the poles, yet the meridional circulation is not global in extent. Rather, it terminates at the edge of the tropics where the subtropical jets are located, and there is a distinct contrast between, on the one hand, the tropical region between the jets, characterized by weak horizontal temperature gradients, the strong Hadley circulation, and easterly winds and, on the other hand, the extratropical regions of strong temperature gradients, weak mean meridional flow, and westerly winds poleward of the jets. There is no such sharp distinction in the external forcing.

The most important controlling factor separating the meteorology of the tropics from that of middle and high latitudes is the Earth's rotation. Consider air rising near the equator, and turning toward the winter pole as seen in Figure 1.2. If for the moment we consider zonally symmetric motions, the air aloft (where frictional losses are utterly negligible) will conserve its absolute angular momentum – angular momentum relative to an inertial reference frame, which includes components associated with the planetary rotation as well as with relative motion – as it moves. As air moves away from the equator and thus closer to the rotation axis, the planetary component decreases; consequently, the relative motion must increase. The further poleward the air moves, the more dramatic the effects of rotation become, just because of the geometry of the sphere. Thus, the winds would become increasingly westerly (eastward) with latitude, and dramatically so: 58 ms^{-1} at 20°, 134 ms^{-1} at 30°, 328 ms^{-1} at 45°. In fact, the westerly wind would have to become infinite at the pole. At some point, the atmosphere cannot sustain equilibrium with such winds. Consequently the poleward circulation must terminate at some latitude; exactly where is determined by many factors, most importantly a balance between the strength of the external forcing and the effective local planetary rotation rate (Held and Hou, 1980; Lindzen and Hou, 1988). These termination latitudes mark the poleward boundaries of the Hadley circulation, and the latitude of the subtropical jet. (In reality, the jets are weaker than this argument would imply; processes we have not considered here – most importantly, angular momentum transport by eddies – allow the air to lose angular momentum as it moves poleward.)

Rotational effects are manifested in the balance of forces through the Coriolis acceleration which, for the large-scale atmospheric flow, is more important than the centripetal acceleration. In general, the vector Coriolis acceleration is $2\mathbf{\Omega} \times \mathbf{u}$, where $\mathbf{\Omega}$ is the vector planetary rotation rate and $\mathbf{u}$ the vector velocity. However, the atmosphere is so thin that the vertical component of velocity is necessarily much smaller than the horizontal components and, in consequence, the important components of acceleration can be written as $f\hat{\mathbf{z}} \times \mathbf{u}$, where $f = 2\Omega\sin\varphi$, the Coriolis parameter, is just twice the projection of the rotation rate onto the local upward direction $\hat{\mathbf{z}}$. At low latitudes f, and hence the influence of planetary rotation, is weak, thus permitting the Hadley circulation to exist there. This fact also implies that pressure must approximately be horizontally uniform, just as the surface of a pond must generally be flat (ponds typically being much too small for planetary rotation to matter). Since, in hydrostatic balance, the pressure at any location is just equal to the weight of overlying air per unit horizontal area, and density depends on temperature, the horizontal temperature gradients there must also be weak, as is observed in the tropical atmosphere (Figure 1.1). In fact, the fundamental role of the Hadley circulation is to maintain this state. Thus, the existence of a separation of characteristics between the tropical and extratropical regions of the atmosphere is, in large part, a consequence of planetary rotation.

These, and essentially all other, atmospheric motions derive their energy ultimately from the input of solar energy or, more precisely, from the differential input between low and high latitudes, which creates internal and potential energy within the atmosphere, a portion of which is then converted into the kinetic energy of atmospheric winds. For a compressible atmosphere in hydrostatic balance, internal and potential energy are closely related to each other;

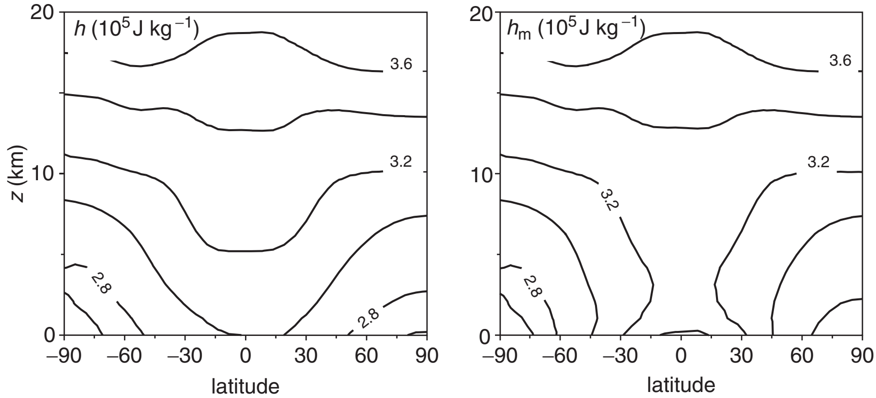

Figure 1.4 Climatological annual- and zonal-mean meridional distribution of (left) dry static energy and (right) moist static energy (J kg^{-1}). Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.

accordingly, it is conventional to combine them into a quantity known as *dry static energy*, which, per unit mass of air, is

$$h = c_\mathrm{p}T + gz,$$

where z and T are altitude and temperature, g is the acceleration due to gravity, and c_p is the specific heat of air at constant pressure. The annual- and zonal-mean distribution of h is shown in the left frame of Figure 1.4.

Just as planetary rotation constrains horizontal motion, so thermodynamic effects and gravity restrict vertical motion. Dry static energy increases with height at all latitudes. Therefore, for near-equatorial air in the upwelling branch of the Hadley circulation to move from the surface up to the upper troposphere (Figure 1.2), its dry static energy must increase. Moreover, in practice what appears in Figure 1.2 as a broadscale, slow, upwelling is in fact the spatial and temporal average of much more rapid motion within narrow convective towers; in such towers, air typically moves from surface to tropopause in an hour or so. Radiation cannot provide the implied diabatic heating: it is much too weak and, besides, radiation is generally a cooling agent in the tropics. However, as was evident in Figure 1.3, tropical surface air is very moist, and the near-equatorial upwelling is thus characterized by saturation, condensation and intense rainfall. Condensation is a major contributor to the thermodynamic balances. Many treatments focus on the thermodynamics of dry air, but add adiabatic heating equal to $-L \times \mathrm{d}q/\mathrm{d}t$ per unit mass per unit time, where L is the enthalpy (latent heat) of vaporization and q the specific humidity, so that $-\mathrm{d}q/\mathrm{d}t$ is the rate of condensation per unit mass of air.

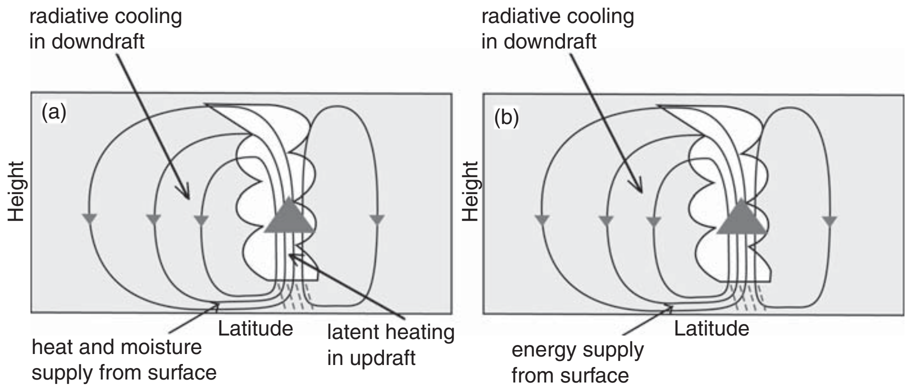

Figure 1.5 Schematic depiction of the energetics of the Hadley circulation. In frame (a), moisture is treated as a source of static energy; in frame (b), it is treated as an integral component of the atmospheric static energy. (See text for discussion.)

The thermodynamic driving of the tropical meridional circulation can thus be described as illustrated in Figure 1.5(a). Air is supplied with heat and moisture in the near-surface boundary, as it flows across the warm tropical ocean (and the stronger the low-level wind, the more turbulent the boundary layer and the greater the evaporation from the ocean surface). Ascent, requiring increasing h, is facilitated by the release of latent heat associated with condensation of water vapor, and consequent precipitation, in the relatively concentrated updraft. The compensating loss of energy occurs in the downwelling region. Descending air tends to warm adiabatically; this warming must result in enhanced emission of thermal radiation. So the picture of Figure 1.5(a) is one of heating in the updrafts, cooling in the downdrafts.

The foregoing description, although commonly found in meteorology texts until quite recently, is imperfect for many reasons. For our purposes, the one important reason is that it misleads us into believing that the underlying driver for such circulations is the latent heat release consequent on precipitation in the updraft, whereas the energy of the circulation is, in fact, supplied ultimately from the warm underlying ocean. More completely, it is the contrast between energy gain in the warm surface boundary layer and the radiative energy loss at lower temperature (in the cooler middle and upper troposphere) in the downwelling. Thus, this kind of circulation is just a classical Carnot heat engine[1] (Emanuel, 1986). To appreciate this fact, we need to recognize that latent heat

[1] This statement is true of many tropical circulation systems. With a modest change of geometry, Figure 1.5 could equally well depict a hurricane or, as we shall see, a monsoon circulation.

release does not constitute an external source of energy. Rather, the process of condensation is internal to an air parcel, and a better way of treating its effects is to include moisture directly in the definition of *moist static energy*

$$h_{\mathrm{m}} = c_{\mathrm{p}}T + gz + Lq$$

per unit mass (e.g., Emanuel, 2000; Holton, 2004). The climatological annual- and zonal-mean distribution of h_{m} is shown in the left frame of Figure 1.4. The greatest difference between h_{m} and h is, not surprisingly, in the tropical lower atmosphere where q is greatest. For our purposes, the most important feature is the elimination of the vertical gradient in the deep tropics: when the contribution of moisture to entropy is properly taken into account, therefore, there is no need to invoke heating in the updraft, since the moist entropy of the air does not change in the updrafts. From this, thermodynamically more consistent, viewpoint there is thus no heating (i.e., no external tendency to increase energy) in the updraft: moisture is lost (to precipitation) but the consequent temperature change is such as to preserve h_{m}. Instead, as depicted in Figure 1.5(b), the energy source driving the circulation is located at the surface, where the crucial role of the supply of both sensible and latent heat from the warm ocean now becomes very explicit. We are thus led to recognize the moist static energy of boundary layer air as a key factor in understanding tropical circulations.

1.2.3 *Longitudinal variations in tropical meteorology*

The distributions of surface pressure over the tropics in the solstice seasons are shown in Figure 1.6. Pressure variations in the tropics are relatively weak, typically a few hPa, as compared with typical variations of 10–20 hPa in extratropical latitudes. A continuous belt of low pressure spans all longitudes, mostly located near the equator over the oceans but displaced into the summer hemisphere over the continents. An almost continuous belt of high surface pressure characterizes the subtropical region around 30° latitude, but in the summer hemisphere the high pressure band is interrupted by continental lows, leaving high pressure centers over the oceans.

The low-level winds (shown for the 850 hPa surface, near 1 km altitude, in Figure 1.7) are dominated by the north-easterly and south-easterly Trade winds in the northern and southern hemisphere, respectively. This general pattern is, however, modulated by features that reflect the pressure distribution. The low-level winds converge into the band of low pressure; in regions, especially oceanic regions, where this band forms longitudinally elongated features, it is known as the Intertropical Convergence Zone, or ITCZ. The Pacific and Atlantic Ocean ITCZs are located north of the equator throughout the year but in this respect, as in many others, the circulation over the Indian Ocean behaves differently.

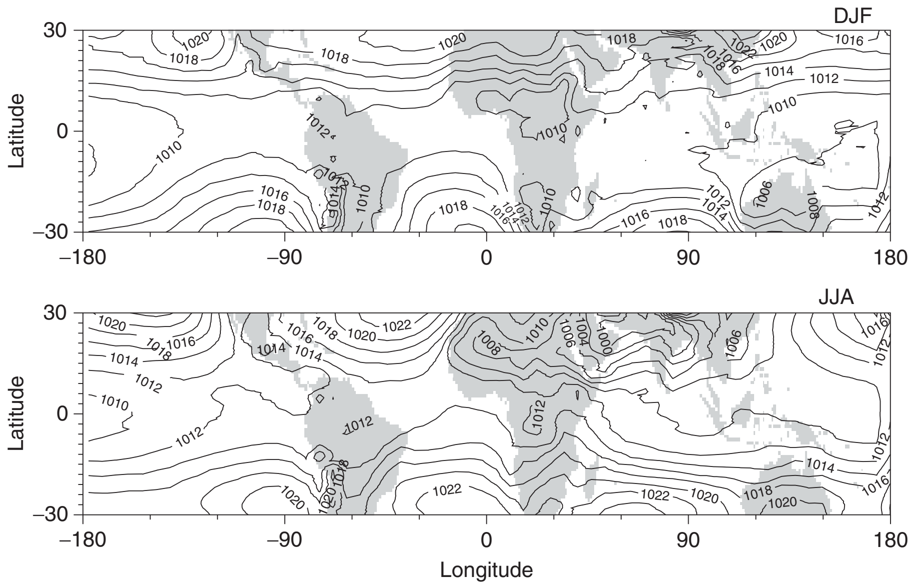

Figure 1.6 Climatological (long-term average) surface pressure (hPa) over the tropics in (top) December–February and (bottom) June–August. Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.

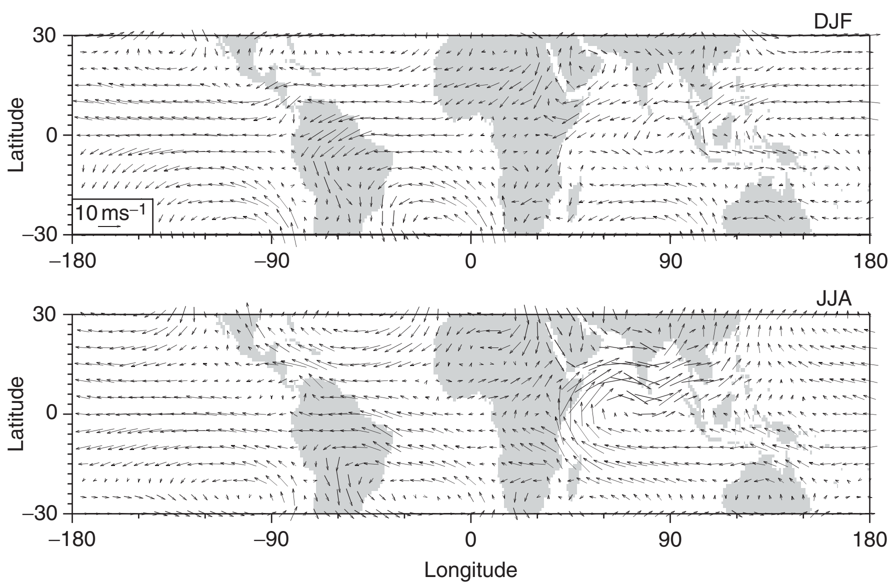

Figure 1.7 Climatological mean winds at 850 hPa (near 1 km altitude) in (top) December–February and (bottom) June–Aug. The scale for the arrows is shown at the lower left of the top plot. Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.