ıces at the
ts Institute of
Sciences,
and terrestrial
He has authored
two other books,
eogeography,
soons and

the Program in
te of Technology.
f *Pure and*
viewed papers,
ıdergraduate
ınd Plumb, 2008).

3/2011

The Asian Monsoon

Causes, History and Effects

PETER D. CLIFT
University of Aberdeen, UK
R. ALAN PLUMB
Massachusetts Institute of Technology, USA

CAMBRIDGE
UNIVERSITY PRESS

Preface

The Asian monsoon is one of the most dramatic climatic phenomena on Earth today, with far-reaching environmental and societal effects. Almost two thirds of humanity live within regions influenced by the monsoon. Monsoon strength and variability have been and will continue to be crucial to the past and future prosperity of the region. With the emerging economies of China, Vietnam and India now adding to those of Japan, South Korea and Taiwan the importance of the region to the global economy has never been greater. Continuation of this growth is dependent on the climate and environment. Recent detailed climate reconstructions now show that the development and collapse of civilizations in both South and East Asia have been controlled in large measure by monsoon intensity. Modern technology now allows society to respond more effectively to environmental stresses, yet in the face of the destructive powers of typhoons or long duration droughts there is still little man can do when environmental catastrophe strikes.

As a result, understanding what controls the Asian monsoon and how it has changed in the past is important not only to scientists but also to the general population. In this book we present a multi-disciplinary overview of the monsoon for advanced students and researchers, spanning recent advances in atmospheric sciences, climatology, oceanography and geology. Finally we consider how the evolving monsoon has both helped and hindered the development of human civilizations since the Last Glacial Maximum, 20 000 years ago. The monsoon represents a large-scale seasonal reversal of the normal atmospheric circulation pattern. In this model, low-pressure systems develop in the tropics owing to rising hot air masses that cool and descend in the subtropics, which are thus characteristically arid regions. In contrast, summer heating of the Asian continent, especially around the Tibetan Plateau, generates low-pressure cells and thus summer rains in South and East Asia. In the winter a reversed high-pressure system is established, with dry, cold winds blowing out of Asia.

The links between the Tibetan Plateau and monsoon intensity have formed the basis of a long-running debate because this proposed relationship would appear to be one of the strongest examples of how the solid Earth, which is being continuously deformed and remodeled by plate tectonic forces, may be influencing the global climate system. The intensity of the modern monsoon likely reflects the fact that Tibet is the largest mountain chain seen on Earth for more than 500 million years and has correspondingly made a particularly large impact on the planet's atmospheric systems. Progress has been made in establishing links between the relatively slow growth of the plateau and monsoon strength, yet until the developing altitude of Tibet is better established and a truly long-scale climate history for the monsoon has been reconstructed it will remain impossible to test the linkages definitely. In particular, climatologists need an appropriate, long-duration sedimentary record dating back to the collision of the Indian and Asian plates that generated Tibet in the first place. In practice this means around 50 million years. Such a record exists in the oceans and continental margins around Asia, but has yet to be sampled.

While recognizing that the monsoon has strengthened over periods of millions or tens of millions of years, research focus over the past 10–15 years has demonstrated that not only does monsoon intensity vary dramatically on much shorter timescales, but that these are often linked to other parts of the global climate system. In particular, the detailed climate records now available for the past few million years show coherent, if sometimes lagged, development of the monsoon with the glaciation of the northern hemisphere. Clearly the monsoon cannot be studied in isolation from other systems, especially the oceanic–atmospheric systems of the North Atlantic (Gulf Stream and North Atlantic Deep Water) and the El Niño Southern Oscillation system of the Pacific Ocean. Indeed, it has been suggested not only that these systems control monsoon strength, but also that the monsoon can affect their evolution. A general pattern has emerged of summer monsoons being strong and winter monsoons generally weaker during warm, interglacial periods, and the reverse situation dominating during glacial times. As a result monsoon strength varies on the 21, 40 and 100 thousand year timescales that control periods of glacial advance and retreat. In detail, however, the situation is complicated by lags in the climate system that offset the response of the monsoon to solar forcing. In addition, there continues to be debate regarding how the monsoon differs in South and East Asia over various timescales. Current data suggest a generally coherent development between the two systems over millions of years but differences at the orbital and sub-millennial scale. Determining how and why they differ requires more high-resolution climate reconstructions from across the entire geographic range of the monsoon, involving both the "core area" of monsoon activity, such as the Bay of Bengal, and the "far-field" regions, such as the Sea of Japan and the Gulf of Oman, which may be more sensitive to modest changes in strength. Observations alone are not enough and a deep understanding of how the monsoon evolves and what the key controls are will require better climate models, ground-truthed with both oceanic and continental climate records.

The interactions of monsoon and society are a particularly fertile area of recent and future research. This field has developed as better climate records have been reconstructed over the past 8000 years or so. In particular the resolution permitted by ice cores and some high accumulation rate sediments in the oceans and lakes allows changes in monsoon intensity to be compared with human history. Indeed the ^{14}C dating used to constrain these records is the same method used to date archaeological sites, allowing a robust comparison to be made. Global warming, as a result of human activities, as well as natural processes, would tend to favor a stronger summer monsoon in the long term, yet in detail there is much potential complexity. Melting of the Greenland ice sheet may disrupt the overturn of waters in the North Atlantic and result in a cooling of that region. Comparison with similar natural events in the past suggests that such an event would result in weaker summer monsoons. Not only the strength of the monsoon can be affected by climate change but also its variability. Historical records indicate that the number and intensity of summer typhoons striking the densely populated coast of southern China have increased significantly over the past 200 years. If that trend were to continue, its economic and humanitarian effects could be disastrous.

Whatever part of the Earth we live in, the Asian monsoon is of significance to our lives and understanding of how the planet and society operates. Much work remains to be done in quantifying the monsoon and how it functions at a variety of timescales. Despite this great progress has been made in understanding this system. In this book we have attempted to synthesize what is now known and highlight those areas where significant research remains to be done.

PETER CLIFT
Aberdeen, UK
ALAN PLUMB
Cambridge, Massachusetts, USA

Acknowledgements

Clift would like to thank the following friends and colleagues for their generous help in putting this book together: Mark Altabet, An Zhisheng, David Anderson, Jon Bull, Doug Burbank, Stephen Burns, Kevin Cannariato, Marin Clark, Steve Clemens, Kristy Dahl, Dominik Fleitmann, Christian France-Lanord, Carmala Garzione, Liviu Giosan, Ananda Gunatilaka, Anil K. Gupta, Naomi Harada, Nigel Harris, Ulrike Herzschuh, David Heslop, Ann Holbourn, Yetang Hong, Tomohisa Irino, Hermann Kudrass, Wolfgang Kuhnt, Michinobu Kuwae, Peter Molnar, Delia Oppo, Dave Rea, Stephan Steinke, Ryuji Tada, Federica Tamburini, Ellen Thomas, Ruiliang Wang and Pinxian Wang. The idea for this book came from Clift's involvement in IGCP 476 "*Monsoons and Tectonics*," an international program supported by UNESCO and organized by Ryuji Tada.

Clift would also like to thank the Alexander von Humboldt Foundation for supporting his time in Bremen during a visiting fellowship at the Research Center for Ocean Margins (RCOM) and the Fachbereich Geowissenschaften at the Universität Bremen, when much of the writing was completed. Related research was completed thanks to financial support from the National Science Foundation (USA), the Natural Environment Research Council (UK), the Royal Society and the Carnegie Foundation for the Scottish Universities. Clift wishes to thank his wife Chryseis Fox for putting up with all the lost family time as a result of writing this book and the associated travel and research activities. Without her understanding and support this book would not have been possible.

We also wish to thank Bill Haxby and Suzanne Carbotte at the Lamont-Doherty Earth Observatory for their help with GeoMapApp, which is supported by the National Science Foundation.

We thank all the staff at Cambridge University Press who have helped with the production of this book, especially Matt Lloyd, Susan Francis, Dawn Preston, Denise Cheuk and Annette Cooper.

1 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.

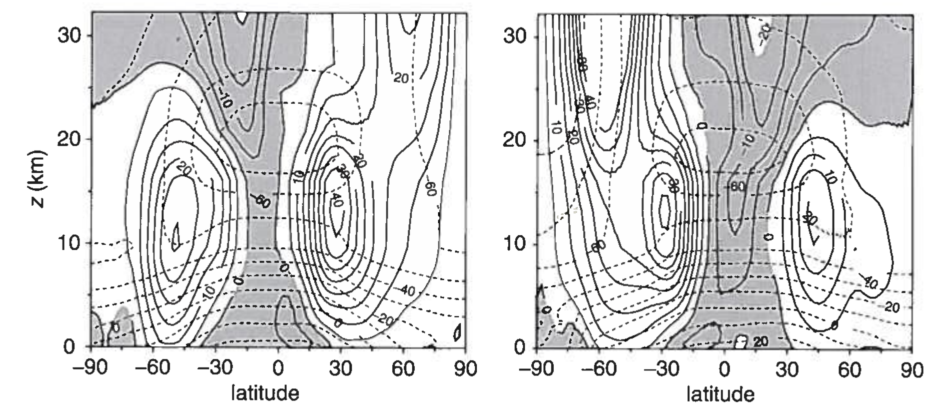

Figure 1.1 Climatological zonal mean zonal wind (solid; ms^{-1}) and temperature (dashed; °C) for (left) December–February and (right) June–August. Contour intervals are 5 ms^{-1} and 10°, respectively; easterly winds are shaded. The data are averaged on pressure surfaces; the height scale shown is representative. 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

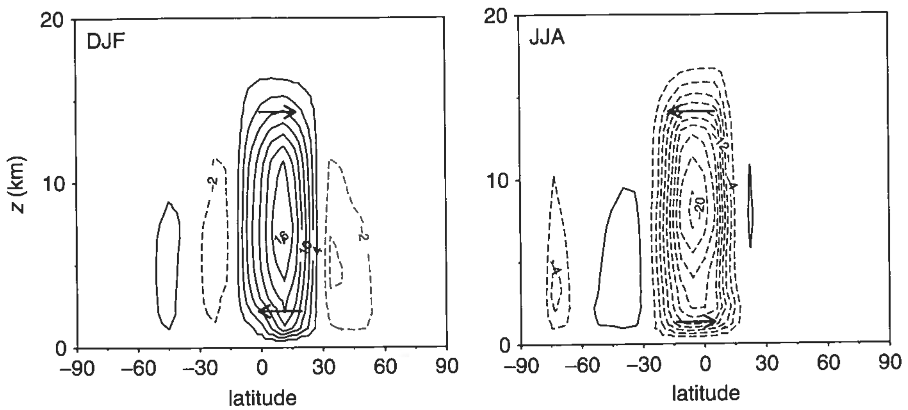

Figure 1.2 Climatological zonal mean overturning streamfunction (10 kg s^{-1} for December–February (left) and June–August (right). Solid contours denote positive values, dashed contours are negative; the zero contour is not plotted. The meridional flow is directed along the streamfunction contours, 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.

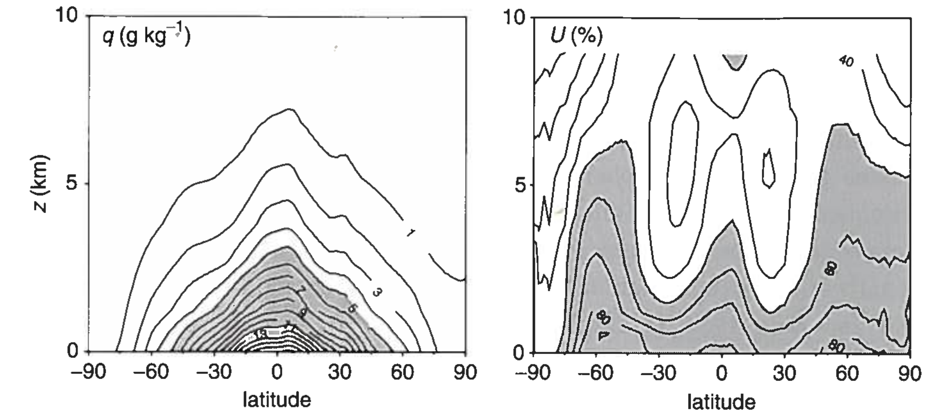

Figure 1.3 Climatological annual- and zonal-mean specific humidity (left, g kg^{-1}; values greater than 5 g kg^{-1} are shaded) and relative humidity (right, %; values greater than 50% are shaded). Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at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

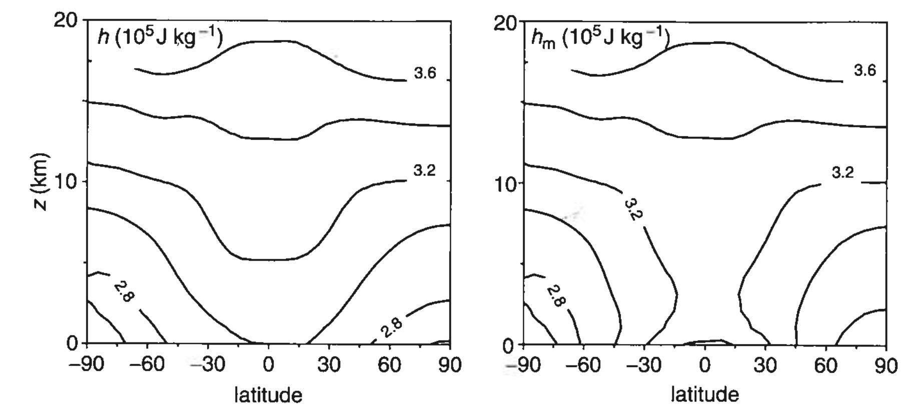

Figure 1.4 Climatological annual- and zonal-mean meridional distribution of (left) dry static energy and (right) moist static energy (J kg^{-1}). Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.

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

$$h = c_p T + gz,$$

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 $-L \times dq/dt$ per unit mass per unit time, where L is the enthalpy (latent heat) of vaporization and q the specific humidity, so that $-dq/dt$ is the rate of condensation per unit mass of air.

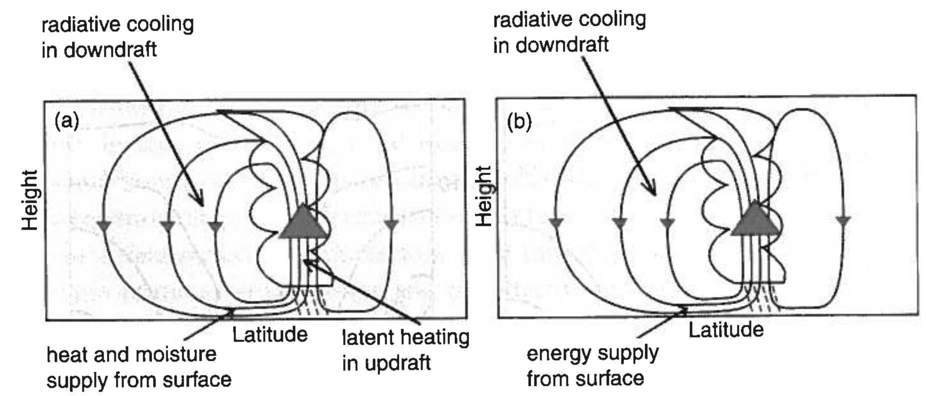

Figure 1.5 Schematic depiction of the energetics of the Hadley circulation. In frame (a), moisture is treated as a source of static energy; in frame (b), it is treated as an integral component of the atmospheric static energy. (See text for discussion.)

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*

$$h_m = c_p T + gz + Lq$$

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.

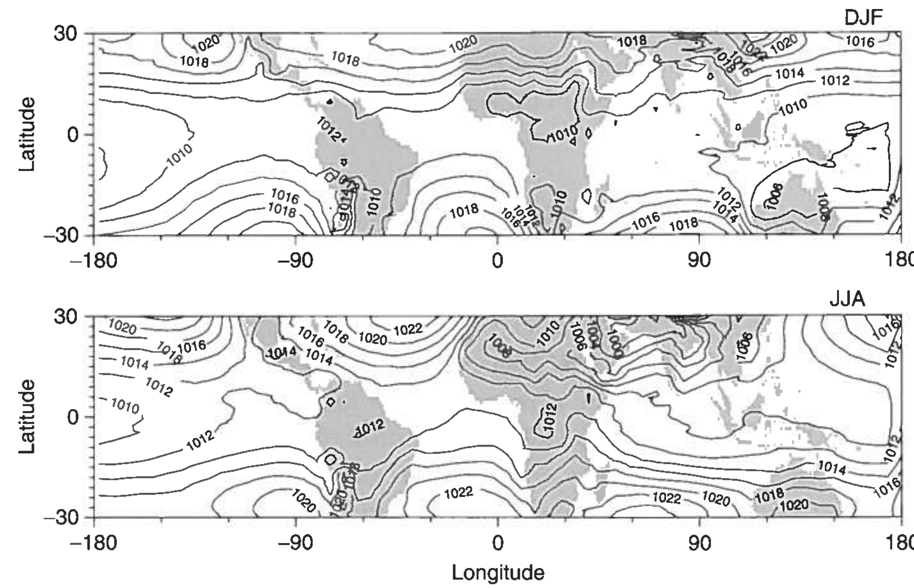

Figure 1.6 Climatological (long-term average) surface pressure (hPa) over the tropics in (top) December–February and (bottom) June–August. Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.

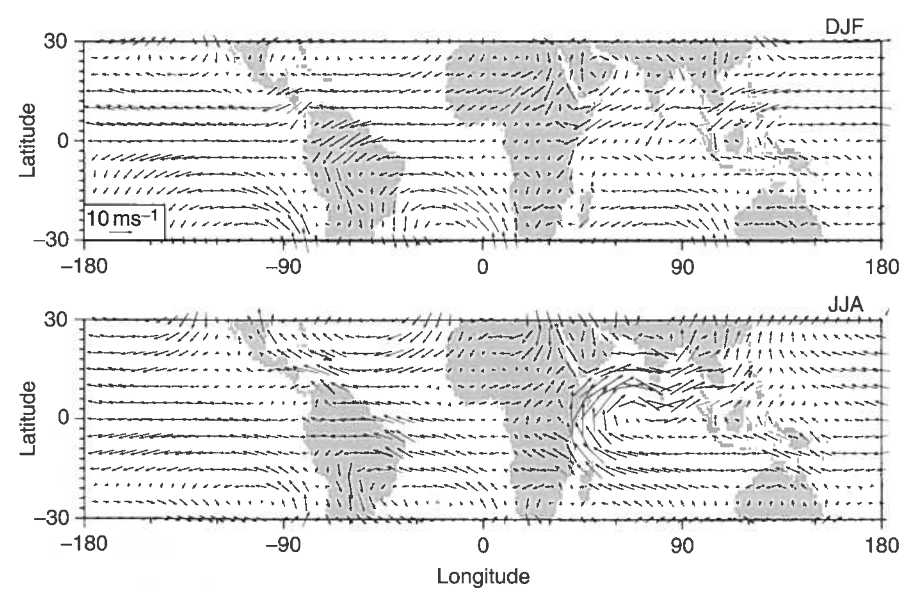

Figure 1.7 Climatological mean winds at 850 hPa (near 1 km altitude) in (top) December–February and (bottom) June–Aug. The scale for the arrows is shown at the lower left of the top plot. Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.

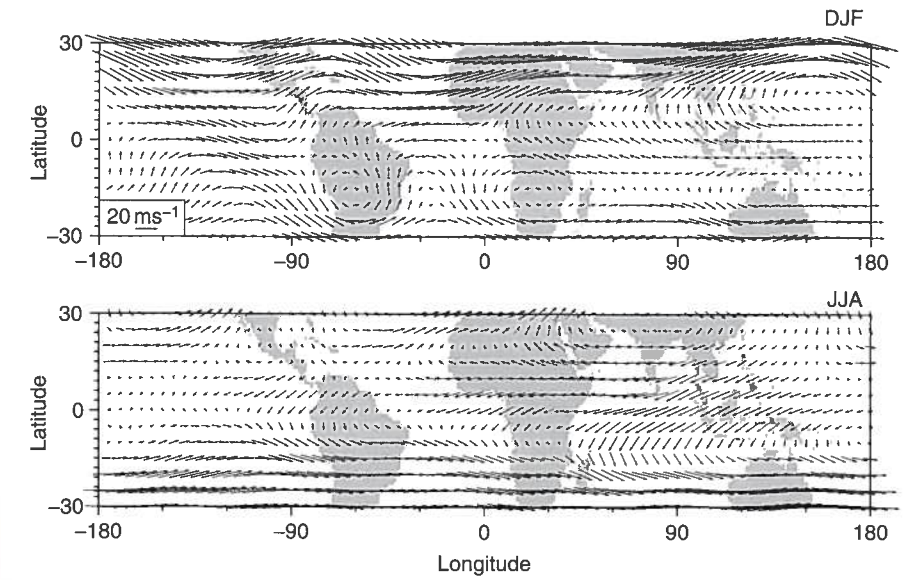

Figure 1.8 As Figure 1.7, but for the 200 hPa pressure level.

The Indian Ocean ITCZ is south of the equator in southern hemisphere summer, but in northern summer there is no clear oceanic convergence zone; rather than converging near the equator, the low-level flow crosses the equator over the western part of the ocean, becoming south-westerly north of the equator. This dramatic feature is the inflow to the South Asian monsoon system, which will be discussed in more detail in Section 1.3. The large surface anticyclones (of high pressure) noted on Figure 1.6 are regions of flow *divergence*; the low-level flow spirals out of these mostly oceanic features to form the Trade winds and ultimately to feed the convergence zones.

The climatological upper-level tropical winds (on the 200 hPa pressure surface, near 11 km altitude) are shown in Figure 1.8. The winds are stronger aloft; they are mostly easterly very close to the equator, and strongly westerly in the subtropics; thus, the zonally averaged picture presented in this section is fairly representative of most longitudes. A few large, strongly divergent anticyclones[2] are evident in the region between the easterlies and westerlies, in fact almost directly over the regions of low-level convergence noted previously, most markedly over the summer continents. There is also strong upper-level outflow from the oceanic ITCZs, but this is somewhat masked on Figure 1.8 by the strong nondivergent part of the flow there. It is here that the large-scale

[2] Anticyclones are evident in the wind plots as clockwise or anticlockwise circulations north or south of the equator, respectively.

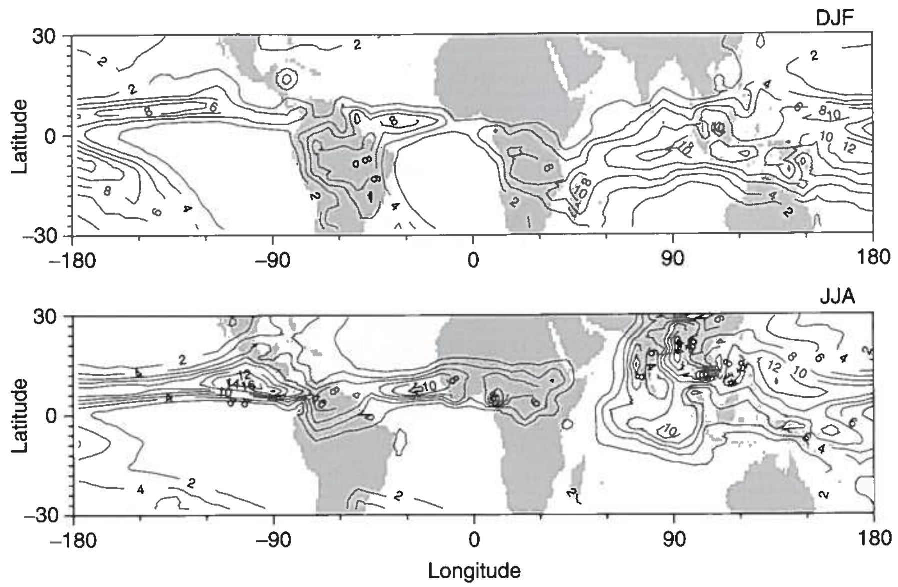

Figure 1.9 Climatological mean rainfall rate (mm day^{-1}) over the tropics in (top) December–February and (bottom) June–August. Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website at www.cdc.noaa.gov/.

tropical ascent is taking place, fed by low-level convergence into surface low pressure and spiralling out in the upper tropospheric anticyclones.

These tropical regions of strong, deep ascent are, of course, characterized by heavy rainfall (Figure 1.9); since the saturation-specific humidity at the outflow level is no more than a small percentage of the specific humidity of the surface air, nearly all the moisture converging at low levels condenses in the updraft to fall as precipitation. In consequence, annual mean rainfall in these regions is as much as several meters.

1.2.4 *Location of the convergence zones*

Why are the convergence regions, with their associated deep convection and intense rainfall, located where they are? Over the oceans, the answer is straightforward: to a large degree, they follow the warmest water. Figure 1.10 shows the climatological distribution of surface temperature in the solstice seasons. The tropical oceans are generally warm, of course; on the large scale, the warmest waters are located consistently in the western tropical Pacific Ocean, and across what is sometimes referred to as the maritime continent, the region lying between the Pacific and Indian Oceans, and occupied by Indonesia, Melanesia, and other island groups. In this region, where sea-surface temperature (SST) can reach 30 °C, the locus of the warmest waters migrates

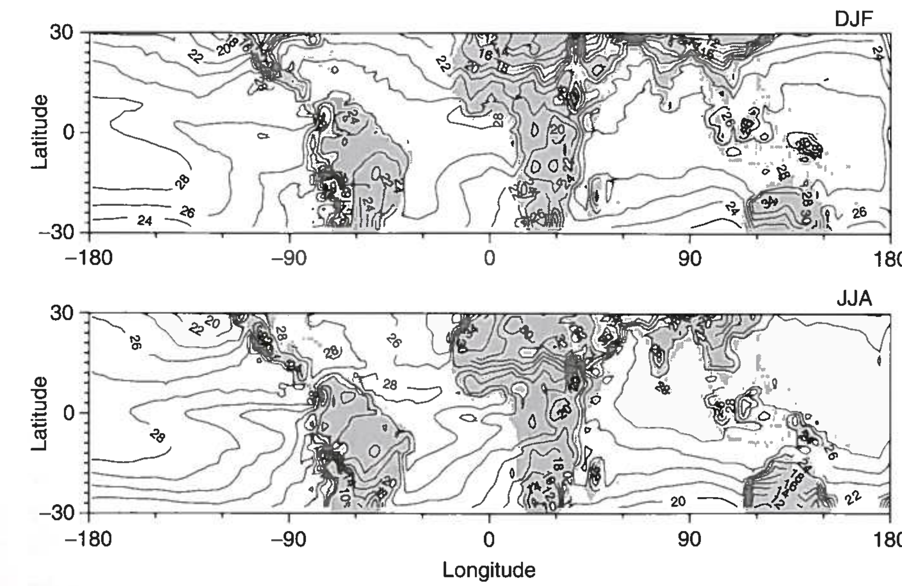

Figure 1.10 Climatological mean surface temperature (°C) in (top) December–February and (bottom) June–August. Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado through their website at www.cdc.noaa.gov/.

a little with the seasons, toward the summer hemisphere. Elsewhere, the warmest waters occur in zonally extensive, but rather narrow, bands which are located mostly north of the equator throughout the year in the Pacific and Atlantic Oceans, but south of the equator in southern summer in the Indian Ocean. Strikingly, in a narrow band a few degrees of latitude wide along the equator, and especially in the eastern Pacific and Atlantic Oceans, the sea surface is much colder, a consequence of the upwelling there of cold water from depth.

Comparing the distributions of low-level winds and rainfall shown in Figures 1.7 and 1.9 with that of surface temperature in Figure 1.10, it is clear that the oceanic high-rainfall convergence zones are collocated with the warmest water. The atmospheric circulation in these regions is thus controlled by the oceans. The ocean surface in these regions is not only warm it is also (to state the obvious) wet and moisture supply to the atmospheric boundary layer is uninhibited. Because of the near-exponential dependence of the saturation specific humidity on temperature, the boundary layer becomes extremely moist, as well as warm, and thus acquires large moist static energy h_m. High boundary layer h_m is the primary condition for atmospheric convection; accordingly, the correspondence between high SST and deep atmospheric convection, intense rain, and low-level flow convergence is straightforward.

One consequence of the strong east-west variations in equatorial SST, especially in the Pacific Ocean, is that the atmospheric circulation comprises not only the Hadley circulation, with its overturning in the latitude–height plane, but also an east–west overturning in much of the tropics. The latter circulation is most evident in northern winter, when it is manifested in Figures 1.7 and 1.8 as low-level equatorial easterly winds overlain by westerlies. The strongest such circulation is in the Pacific basin, where it is known as the Walker circulation, in which air rising over the warm maritime continent spreads westward along the equator, subsides and then returns as a low-level easterly flow. There is, in fact, a strong interplay between the atmospheric Walker circulation and the SST distribution in the equatorial Pacific Ocean: the Walker circulation is controlled by the longitudinal gradient of SST while, in turn, the SST distribution is strongly influenced by the wind stress consequent on the low-level atmospheric flow. Fluctuations in this coupled system manifest themselves as the El Niño–La Niña cycle in the ocean and as the Southern Oscillation in the atmosphere (and, in fact, the coupled phenomenon is now widely known by the concatenated acronym ENSO).

Over the continents, the seasonal shift of the convergence zones is much more marked than over the oceans. There appear to be two types of continental behavior: the region of high rainfall in the interior of Africa migrates relatively smoothly back and forth with the seasons between extremes of about 15° S in southern summer (reaching 20° S in Madagascar) and around 10° N in northern summer. At the equinoxes, the rain band passes the equator; thus, while rainfall peaks once per year (in the wet season of local summer) north and south of the equator, there are two wet seasons per year (at the equinoxes) in the equatorial belt. Similar behavior is also seen in tropical America: it thus appears to be characteristic of continents that span the equator.

In other areas, where tropical land lies poleward of ocean, the seasonal transition from dry to wet to dry is more dramatic. The clearest example of this is the South Asian–Indian Ocean–Australian monsoon system, on which we shall focus in what follows.

1.3 The Indian Ocean monsoon system

The distribution of rainfall over the Indian Ocean region for the four seasons is shown in Figure 1.11; low-level winds over the same region are plotted in Figure 1.12. Through the course of a year, the bulk of the rainfall over the Indian Ocean region migrates systematically from a broad region extending across the maritime continent and the southern equatorial Indian Ocean in northern spring, moving onto the southern and south-eastern Asian continent

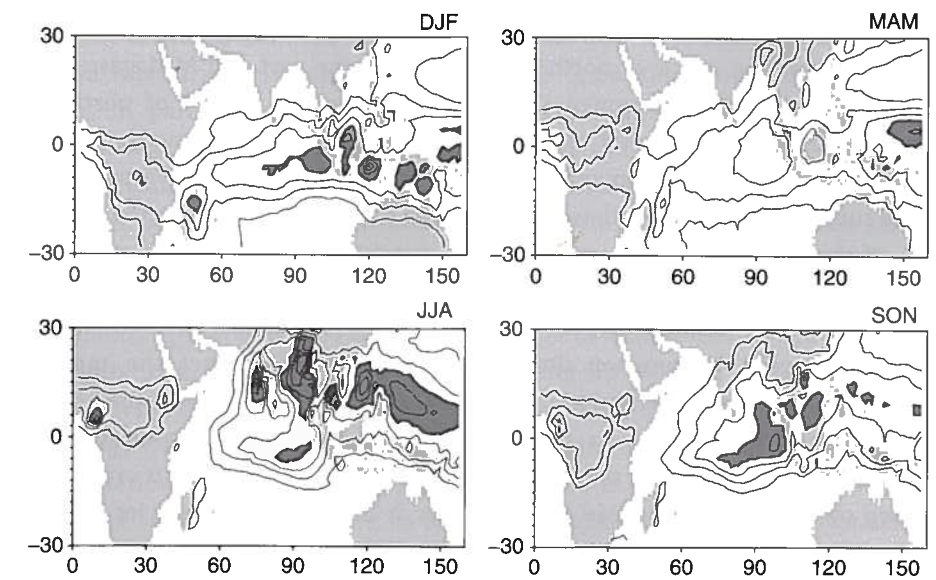

Figure 1.11 Climatological mean rainfall rate for the four seasons. The contour interval is 2.5 mm day^{-1}; heavy shading denotes rainfall greater than 5–10 mm day^{-1}.

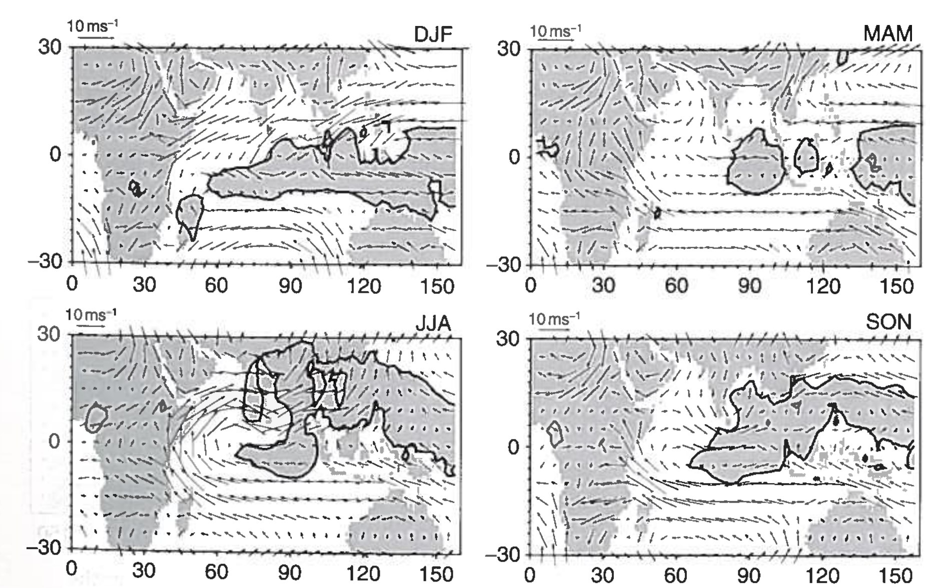

Figure 1.12 Climatological mean low level (850 hPa) winds for the four seasons. The scale for the wind arrows is shown at the top left of each plot. The heavy lines mark regions of seasonal mean rainfall in excess of 7.5 mm day^{-1}.

and nearby oceanic regions in summer, returning southward in northern autumn, reaching as far as northern Australia in the east and Madagascar in the west, in southern summer. During the most active period of northern summer, the intense rainfall maximum near the Asian coast is accompanied by vigorous deep convection and the concomitant net upwelling. Warm, moist air is supplied from the south-west by a strong inflow originating in the easterlies of the southern tropics, crossing the equator in the western ocean and curving north-eastward across the Arabian Sea, and from the south-east across the maritime continent. It is thus clear that this system is not at all local, but a planetary-scale phenomenon. (Indeed, during northern summer, the implied meridional overturning circulation is the dominant contribution to the global Hadley circulation shown in Figure 1.2.) The strong cross-equatorial flow, a feature of most (if not all, see Neelin (2007)) monsoon circulations, is especially strong over the western ocean where in fact it forms a northward jet on the eastern flank of the East African highlands (Findlater, 1969). The cross-equatorial flow over the western ocean is thus concentrated below the peaks of and alongside the mountains. In fact, much of the northward flow crossing the equator in the eastern ocean, in the vicinity of the maritime continent, is also concentrated adjacent to and between the mountains there.

Climatological conditions prior to and following the onset of this large-scale flow are illustrated in Figure 1.13. In May, while some northward cross-equatorial flow is becoming established at the East African coast, the flow is weak and it does not yet extend toward the Indian subcontinent. Flow onto the continent has, however, become established over south-east Asia by this time. The surface waters of the northern Indian Ocean have warmed, especially in the Arabian Sea and the Bay of Bengal. The rain belt (not shown) is beginning to shift north of the equator east of 90° E, but not further west. Hot, dry

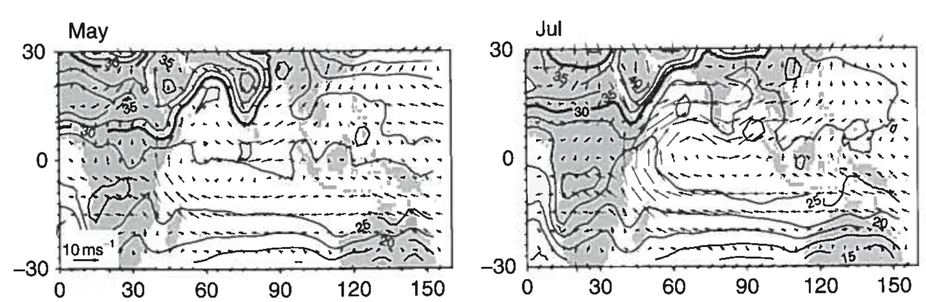

Figure 1.13 Climatological 1000 hPa air temperature and 850 hPa winds over the Indian Ocean region in (left) May and (right) July. Contour interval 2.5 °C; heavy contour is 30 °C. The arrow scale for both frames is shown in the lower left of the left frame.

conditions extend from the Sahara across Arabia and India. In the interior of the Indian subcontinent, temperatures exceed 40 °C, but deep convection does not occur, in part because of the dryness of the low level air, such that the near-surface moist static energy is actually less than that over the somewhat cooler oceans to the south, and because of broad-scale subsidence aloft. However, this state of affairs eventually breaks down, accompanied by the almost simultaneous onset of deep convection and rainfall across the northern Indian Ocean and the continent south of about 20° N, and of the large-scale summertime circulation. The suddenness with which the circulation is established is evident in Figure 1.14, which shows the winds picking up over the Arabian Sea over an interval of 2–3 weeks. With onset, temperatures in the subcontinent drop markedly (Figure 1.13) as cooler but more humid air is advected inland. The precipitation maximum over the Indian Ocean migrates quite rapidly from near and south of the equator to be centered near 10° N, although equatorial precipitation does not cease with monsoon onset. The maximum moves further north over the following month before receding, more gradually than during onset, in October to November. Surface temperatures off the coast also drop, partly in response to enhanced evaporation in the strong winds.

Despite the rather systematic onset of the large-scale flow evident in Figure 1.14, the appearance of monsoon rains is not simultaneous over southern

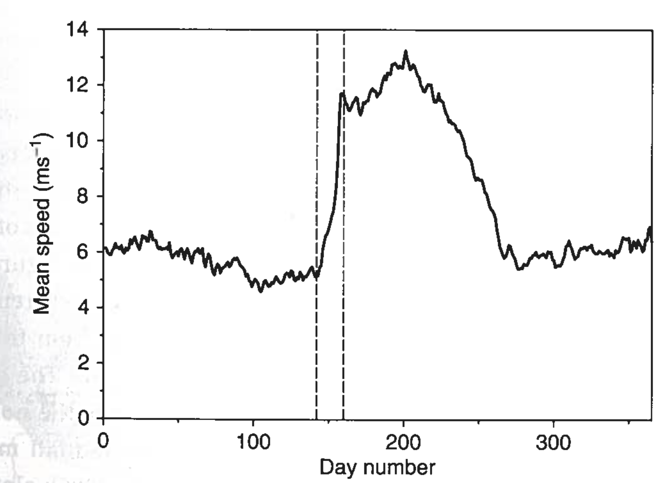

Figure 1.14 Seasonal evolution of an index of mean wind speed over the Arabian Sea. The index is calculated as the square root of specific kinetic energy, averaged over the area 5° S to 20° N, 50° E to 70° E. The day number is a nominal day of year (day 1 is nominally Jan 1), shifted with respect to onset date (the data are composited with respect to monsoon onset). The vertical dashed lines indicate days number 142 and 160. Data courtesy of William Boos.

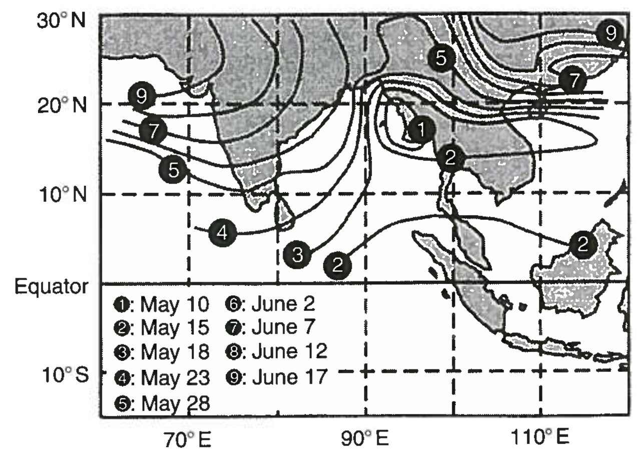

Figure 1.15 Climatological date of the onset of monsoon precipitation, as indicated by the mean location of the 220 Wm^{-2} contour of outgoing long-wave radiation (OLR) observed from space. (Values of OLR as low as this are indicative of cold, high cloud and thus of deep convection and intense rainfall.) (Webster *et al.*, 1998).

and south-eastern Asia. While the actual progression varies somewhat from year to year, on average the rains typically first appear in the eastern Bay of Bengal and Indochina in mid-May, and extend westward and northward from there across the Indian subcontinent and south-eastern Asia over the course of the following month, as illustrated in Figure 1.15. Thus, the actual date of onset, either climatologically or in a given year, is sensitive to the definition used (e.g., Fasullo and Webster, 2003).

Until recently, detailed mapping of monsoon precipitation has been problematic, because of the relative sparseness of high quality rain-gage observations and the highly heterogeneous nature of rainfall. Now, however, high-resolution space-based observations have greatly improved the situation, and maps of mean rainfall rates for June to August from three sources are shown in Figure 1.16. While there are significant differences in intensity between the different data sources (and note that the years making up the averages differ from frame to frame), the benefit of and necessity for high resolution is obvious. The greater part of the precipitation in fact occurs over the ocean, especially the northern and eastern Bay of Bengal and the eastern South China Sea. Rainfall maxima over land are clearly linked to topographic features (Xie *et al.*, 2006): along the Western Ghats, the foothills of the Himalaya, the mountain ranges on the west coast of Burma, Indochina and the west coast of Luzon in the Philippines. These maxima are embedded within the weaker, but still very substantial, region of rainfall extending across the Indian subcontinent and south-eastern Asia evident in both Figures 1.11 and 1.16.

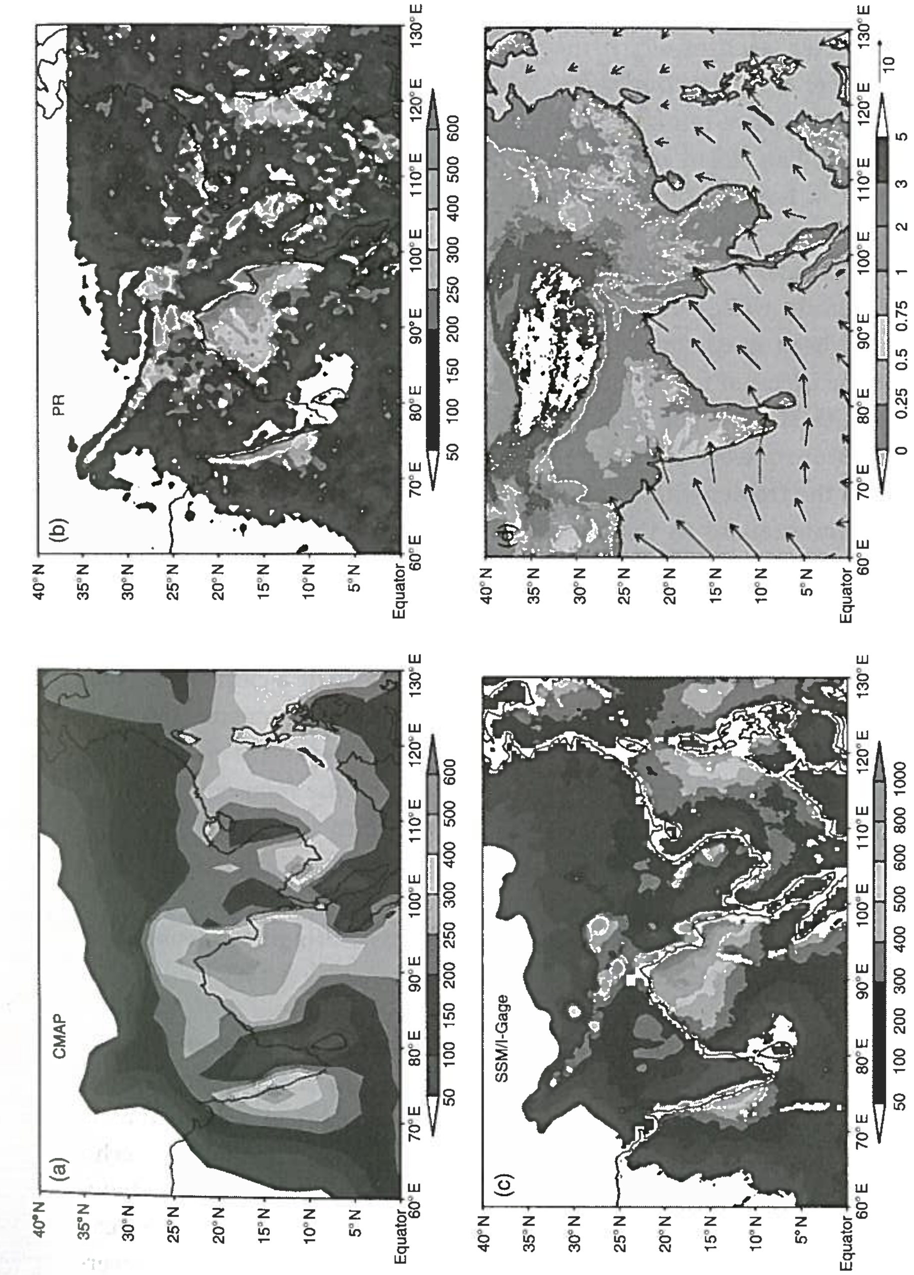

Figure 1.16 Summertime rainfall over the northern Indian Ocean and South Asia. Frames (a)–(c) show June–August mean rainfall rates (mm per month) from different data sources: (a) the Climate Prediction Center Merged Analysis of Precipitation (CMAP; 2.5° resolution, coverage 1979–2003); (b) high resolution data from the space-based Tropical Rainfall Measurement Mission Precipitation Radar (1998–2004) and (c) merged high resolution data from the space-based Special Sensor Microwave Imager and land-based observations (1987–2003). Surface topography (km) is shown in frame (d). (Xie *et al.*, 2006). See color plate section.

1.3.1 *Intraseasonal variability of the monsoon*

Within the wet monsoon season, rainfall intensity fluctuates considerably from day to day and from week to week. There are well defined active and break cycles, each of which may last from a few days to a few weeks (e.g., Webster *et al.*, 1998). During the active cycle, the large-scale monsoonal flow pattern intensifies, rainfall rates increase across much of southern Asia and the northern Indian Ocean and decrease south of the equator. However, just as during onset, a break period does not necessarily imply weak rainfall everywhere. In fact, the rainbands tend to drift northward during the active-break cycle (e.g., Sikka and Gadgil, 1980) such that heavy rains may be falling over the Himalaya foothills while rains subside over much of the rest of the region.

There has been much discussion in the literature about the relationship between these cycles of the Asian monsoon and the Madden–Julian Oscillation (MJO). The MJO (Madden and Julian, 1972, 1994) is a persistent, large-scale and dominant feature of tropical meteorology, which propagates systematically eastward around the tropics with a variable period of around 30–50 days. It has its strongest manifestations in the Indian and Pacific Oceans, but is weakest in northern summer when it might interact with the Asian monsoon. While fluctuations of the Asian monsoon system do have significant variance in the MJO period range (e.g., Yasunari, 1980, 1981) the relationship between the two is not entirely clear (Webster *et al.*, 1998). There does, however, appear to be a clear connection between the MJO and the Australian monsoon (Hendon and Liebmann, 1990).

On shorter timescales, rainfall is modulated by the passage of monsoon depressions. In fact, much of the rain that falls during the monsoon season is associated with these systems. An example is illustrated in Figure 1.17. Typically, four to five of these depressions form in the Bay of Bengal during a monsoon season (Rao, 1976); they propagate generally westward or north-westward over the subcontinent over the course of a few days, bringing intense rain, mostly to the west and south-west of the storm center.

1.3.2 *Interannual variability*

Figure 1.18 illustrates the year-to-year variation of annual mean rainfall for India and for northern Australia. Annual mean rainfall in both regions is dominated by the monsoon rains. Perhaps the first remark to be made, to echo Webster *et al.* (2002), is that interannual variability is, especially for the all-India composite, remarkably weak, in contrast to the more dramatic year-to-year variations that characterize rainfall records across much of the tropics. Nevertheless, year-to-year variations are apparent, the most dominant of which is a two-year cycle in which a wet year is followed by a dry one, and for which the Asian and Australian monsoons appear to be correlated, i.e., a strong South

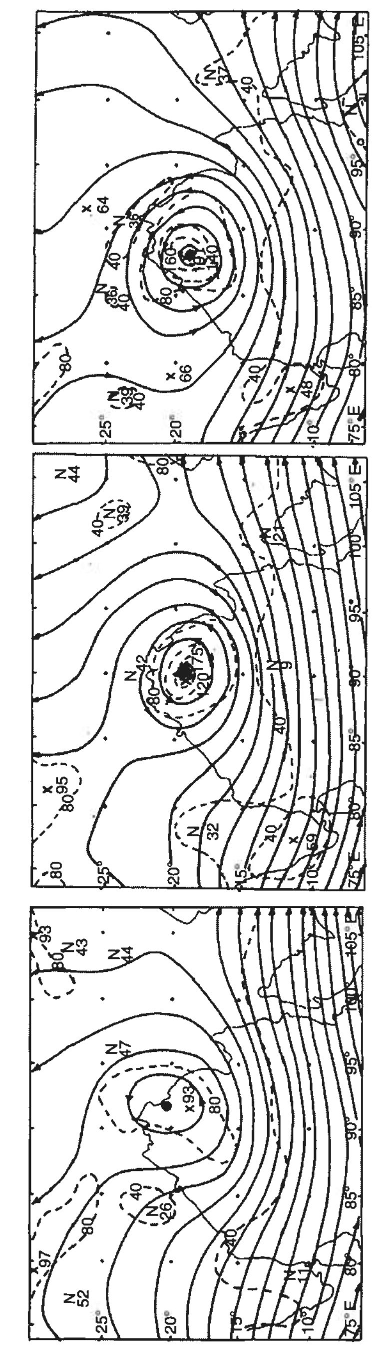

Figure 1.17 An example of a monsoon depression. The solid contours show low-level streamfunction (the dominant, rotational, component of the flow is along these contours, anticlockwise around the storm center). From left to right, the frames are at 24 h intervals beginning 1200 GMT 5 July 1979. (From Sanders (1984)).

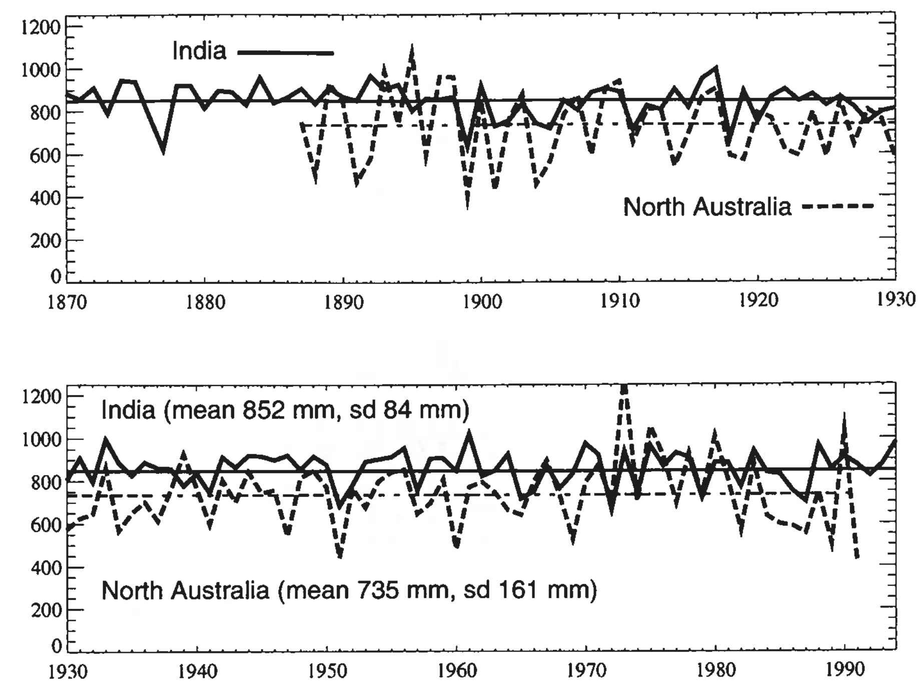

Figure 1.18 Annual-mean rainfall (mm) showing all-India rainfall (solid) and north Australia rainfall (dashed). (Webster *et al.*, 2005).

Asian monsoon tends to be followed, in the ensuing northern winter, by a strong Australian monsoon (Meehl, 1994). Indeed, the two monsoons appear to be intimately linked with Indian Ocean temperatures, with strong monsoon rainfall associated with warm sea-surface temperatures in the preceding northern winter (Harzallah and Sadourny, 1997; Clark *et al.*, 2000).

The dominant signal of interannual variability affecting much of the tropics, especially in the Pacific basin, is that associated with ENSO. El Niño years (identified by anomalously warm surface temperatures in the eastern equatorial Pacific Ocean) are associated with dry conditions in the west Pacific–Indian Ocean region, and this is strongly evident in South Asian monsoon rainfall (Shukla and Paolina, 1983; Torrence and Webster, 1999). Conversely, in the opposite phase of ENSO, La Niña years are usually associated with above-average rainfall.

1.4 Theory of monsoons

1.4.1 *General considerations*

Our understanding of the meteorology of monsoons, and of the factors that control them, has many facets, and is still far from complete. At the very simplest level, the monsoon system is very much like the Hadley circulation

illustration of Figure 1.5: air is supplied with energy at the hot, moist surface, driving an overturning circulation and compensating energy loss in the warm downwelling region. However, real-world monsoons are much more complex than this. For one thing, the underlying surface is heterogeneous: monsoon circulations encompass both oceanic and continental regions. This fact appears to be crucial; the nature, and seasonal variability, of climates and atmospheric circulations in continental and oceanic regions are very different. More to the point, it has long been recognized that it is the land–sea contrast that is at the heart of driving the whole monsoon system. Further, the intense rainfall over land that is characteristic of monsoons, and which is intimately connected to the upward branch of the circulation, cannot be sustained without import of moisture from the ocean, and so there is an inherent two-way feedback between the circulation and the precipitation. Also, monsoon circulations are three-dimensional (i.e., they are longitudinally localized), which makes their dynamics different from that of a simple, zonally symmetric, Hadley circulation.

Just as in Figure 1.5, it is energy input into the atmosphere that drives the circulation, and most of this input is from the surface. Ultimately, of course, all the energy input is from the sun, but the way this energy finds its way into the atmosphere is very different for continental and oceanic regions. Land has a small heat capacity, so that whatever energy the surface receives from the sun (and via thermal radiation from the atmosphere and clouds) is transferred almost immediately to the overlying atmosphere. Not only does the ocean have a much greater heat capacity, but also currents and mixing within the ocean exert a strong influence on SST; accordingly, there is not such a direct, nor immediate, relationship between the input of energy into the ocean and its transfer into the atmosphere. In the subtropics, therefore, as the local downward flux of solar radiation increases with the onset of summer, the surface fluxes over land increase dramatically, thus (other things being equal) increasing h_m, the moist static energy of the boundary layer, while h_m over the ocean is controlled by the more slowly evolving sea-surface temperatures. In early summer, air over the continent is dry; accordingly, almost all of the input of solar energy goes into internal energy: temperature increases, to very high values (cf. Figure 1.10). In the early stages, continental h_m is still less than that over the adjacent ocean, despite its high temperature, because of its low specific humidity. Consequently, convection continues to be located over the ocean, and so there is little change in the large-scale atmospheric circulation. Eventually, however, h_m over the land begins to exceed that over the warmest SSTs; at this stage, convection is favored over the land (it may or may not collapse over the ocean); simultaneously, the associated net upward motion induces inflow, supplying moisture to the continental lower atmosphere, thus feeding moist convection and intense rain. (The input of moisture reduces continental temperatures, since now a significant fraction of moist static energy is in latent form.)

The observed monsoon circulation is, in fact, very much in accord with what theory predicts, given the observed rainfall distribution. Hoskins and Rodwell (1995) investigated the calculated response of the atmosphere, given the longitudinal average of the observed atmosphere as a background state, to heating over South Asia. In both upper and lower troposphere, the induced circulation encompasses about one-third of the globe longitudinally, and reaches across the equator and far into the southern hemisphere. Over the Indian Ocean, the dominant pattern of south-easterly flow south of the equator, crossing the equator over East Africa and becoming the south-westerly flow onto the coast of southern Asia, is well reproduced by the calculation, as is the large, zonally elongated, upper tropospheric anticyclone stretching from the coast of South-east Asia all the way to West Africa. There is a second anticyclone south of the equator, rather like a weak mirror image of that in the northern hemisphere, again in agreement with observations. Thus, most of the observed large-scale features of the tropical atmospheric circulation in northern summer, at least in the region from the Greenwich meridian to the east coast of Asia, can be understood as a response to the diabatic heating over South and South-east Asia. This underscores the far-reaching impact of the Indian Ocean monsoon system, not just to South and South-east Asia, but to the global atmosphere. In fact, Rodwell and Hoskins (1995; 2001) pointed out that one aspect of the calculated response is subsidence over northern subtropical Africa, i.e., the Sahara Desert region; hence, since subsidence is characterized by warm, dry air, there may be a direct link between desertification of the Sahara and the Asian monsoon. As noted above, the global desert belts are products of large-scale subsidence, itself a characteristic of the entire tropical circulation; the point here is that the reason for the size and extreme dryness of the Sahara is probably a result of its geographical location, relative to the Asian monsoon.

The circulation associated with the Asian monsoon, then, is quite well understood once the rainfall pattern is established. Moreover, the calculated low-level flow is such as to transport warm, moist air from over the warm waters of the North-west Indian Ocean onto the continent, thus supplying moisture to sustain the rainfall, and providing air of relatively high moist entropy that will be further heated over the land. Thus, the solution is qualitatively self-consistent. However, since the location of the rainfall is imposed in the calculation, the deeper issue of why the rainfall is located where it is remains unresolved. Why does the monsoon rainfall not extend, e.g., further west or further north, deep into the continent? Simple modeling studies (Chou *et al.*, 2001; Chou and Neelin, 2001; Privé and Plumb, 2007a,b) indicate that the bulk of deep moisf convection, and correspondingly the core of the updraft region, must coincide with maximum boundary layer h_m.

In the Indian Ocean region, the envisaged evolution is therefore as follows. At the start of summer, maximum h_m is invariably found over the ocean, south of the equator; moist convection and rainfall maximize in the ITCZ, more-or-less coincident with the warmest SSTs (Figure 1.11). As summer progresses, the land starts to warm – as do the SSTs in the northern part of the ocean, especially in the Arabian Sea and the Bay of Bengal – and eventually h_m in the vicinity of the south coast of the continent exceeds that over the near-equatorial ocean, at which stage, almost simultaneously, the atmospheric circulation dramatically alters (very rapidly, as was seen in Figure 1.14) and intense rainfall sets in near and over the southern part of the continent. Thus, onset of the monsoon circulation is, as long recognized, associated with a reversal in the land–sea contrast, and that statement can now be made more quantitative: it is the contrast in the geographical distribution of boundary layer moist static energy that is the controlling factor. At first sight, it might appear that this distribution can be explained as a simple and direct consequence of SST over the ocean, and of solar energy input over the land, and that all the essential details are thus understood.

Inevitably, of course, it is not quite so simple. If it were, location of maximum h_m would, once moved onto the land, follow the Sun to 23° latitude whereas, in reality, the core of the monsoon rainfall is never found so far from the equator, especially in South Asia where, as we saw in Figure 1.16, much of the rain remains over the coastal ocean, while the inland penetration of monsoon rainfall into northern Australia is extremely limited, despite the vigorous cross-equatorial flow that supports it (Figure 1.7). To some extent, this may be a reminder that ocean SSTs near the coast themselves become very warm in early summer (although they cool in midsummer, a point to be raised in the following text). However, one must also recognize that the energy budgets over the land are not as simple as the foregoing discussion would imply. Once convection sets in over the land, clouds become a major, and complex, factor in the surface radiation budget. Moreover, radiative heating of the surface is not the only important factor: boundary layer air is also influenced by advection from elsewhere. Just how important this is, and the details of the key factors, depend on geographical details but, in general, air over the midlatitude continents is, even in summer, either dry (in locations uninfluenced by advection from the ocean) or cool (where advection of air from the relatively cool midlatitude ocean is important) and thus, either way, has relatively low h_m. If the monsoon circulation were to penetrate far enough inland to entrain such air, therefore, it would become progressively more difficult to sustain the high h_m required to drive the circulation in the first place. Thus, the low h_m of deep continental air may be the main factor limiting inland penetration (Chou and Neelin, 2001; Neelin, 2007). Moreover, this factor may at the same time limit the intensity of the circulation: a stronger low-level onshore flow would tend to move the convecting region further inland, thus reducing its maximum h_m, and hence reducing its intensity (Privé and Plumb, 2007b).

The potential limiting effects of the onshore flow remind us that the monsoon circulation is fundamentally dependent on surface conditions over both land and ocean. Moreover, sea-surface temperatures in the Indian Ocean are largely controlled by the surface winds, implying a feedback loop involving land, ocean and atmosphere. For one thing, the strong low-level monsoon winds over the Indian Ocean, and especially, over the Arabian Sea in northern summer induce strong sensible and evaporative heat loss to the atmosphere; this is one factor moderating SSTs in summer. In addition, surface temperatures over much of the Indian Ocean are strongly influenced by heat transport within the ocean (e.g., Godfrey *et al.*, 1995), and the ocean circulation that effects this transport is itself wind-driven (McCreary *et al.*, 1993). This has led to suggestions that the biennial oscillation of the Asia–Australia–Indian Ocean monsoon system is a manifestation of land–ocean–atmosphere interactions (Nicholls, 1983; Meehl, 1994, 1997). Meehl (1997) proposed a mechanism whereby the vigorous low-level winds in a strong monsoon year cool the surface waters of the equatorial Indian Ocean, leading to a weak monsoon the following year. Webster *et al.* (1998, 2002) incorporated ocean dynamics into Meehl's picture; the slow timescales of the ocean circulation provide the year-to-year memory that is needed to close the theory.

1.4.2 *Role of orography*

It has long been recognized, and explicitly demonstrated in general circulation models (Hahn and Manabe, 1975), that orographic effects are crucial in making the Asian monsoon the strongest on the planet. Several subsequent studies, using a wide range of such models, have confirmed the importance of Tibetan uplift to the present day monsoon, both in South and East Asia (Prell and Kutzbach, 1992; An *et al.*, 2001; Liu and Yin, 2002; Abe *et al.*, 2003; Kitoh, 2004). While Hahn and Manabe noted that orography can influence the monsoon through both mechanical and thermal effects, much emphasis in the subsequent literature was given to the latter, frequently through vague reference to elevated heating whose significance was unclear until Molnar and Emanuel (1999) demonstrated that, for an atmosphere in radiative–convective equilibrium, boundary layer moist static energy would be greatest, for the same solar input, over the highest elevations. Mechanically, mountains have direct effects, a local impact on precipitation through induced uplift, and a more widespread

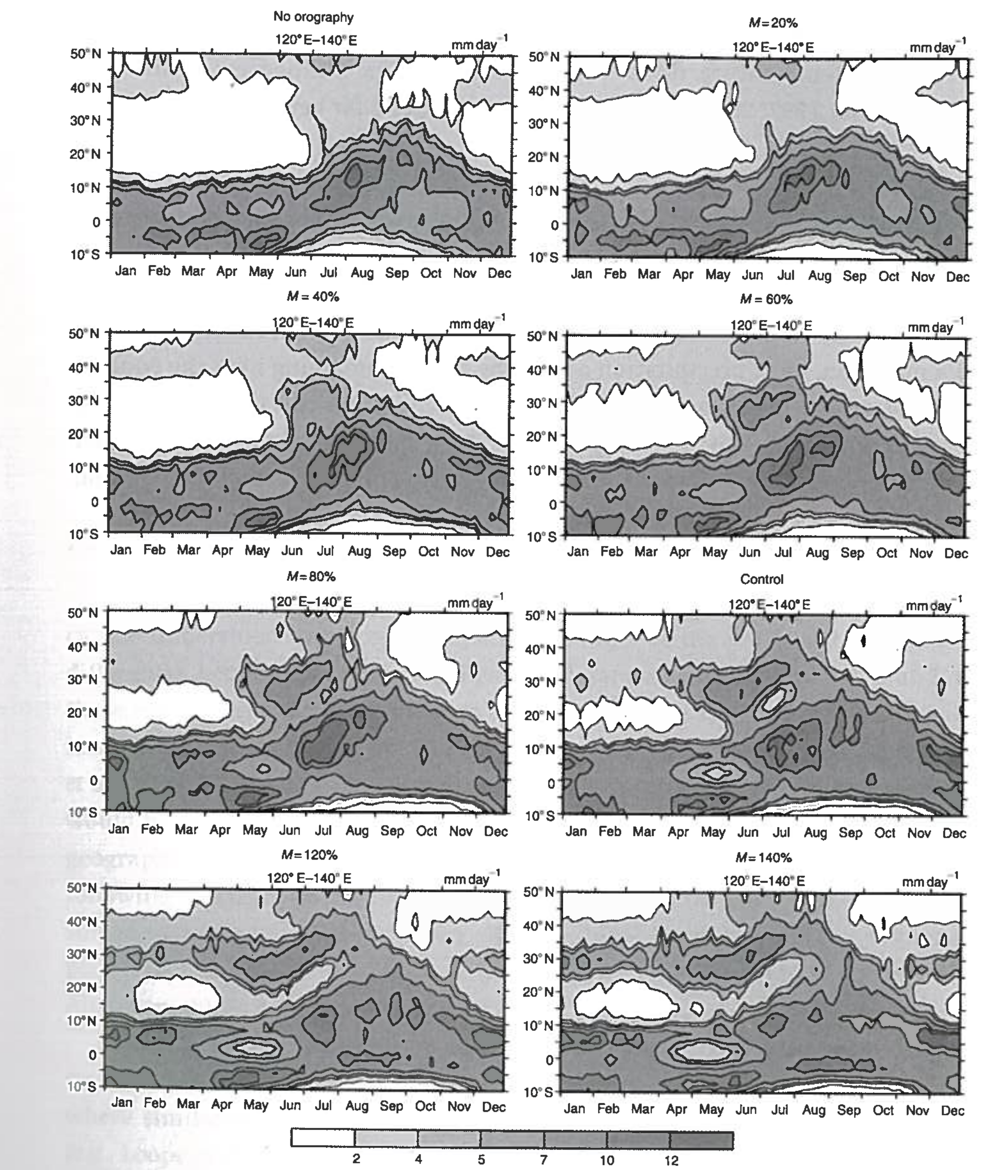

Figure 1.19 Average precipitation rate (mm day^{-1}) between 120° and 140° E from a general circulation model, in which present-day topographic height has been multiplied by a factor *M*, where *M* ranges from 0 through 140%, as indicated. (Kitoh, 2004.)

impact by shielding the monsoon region from the moderating influence of low-level air from higher latitudes with low moist static energy (Neelin, 2007; Privé and Plumb, 2007b). The former effect is visible in the rainfall distribution of Figure 1.16, while the latter may be important in North America as well as in Asia. A further effect, less direct but potentially of great importance, is on sea-surface temperature distributions via orographic influences on low-level winds (Kitoh, 1997; 2004); such influences can extend far beyond the location of the orography.

An example illustrating the dependence of modeled rainfall on orographic heights is shown in Figure 1.19. The results shown are of precipitation averaged over 120–140° E from a series of simulations by Kitoh (2004) using a coupled ocean–atmosphere general circulation model, with all surface elevations equal to present day values multiplied by a factor M. With $M = 0$, there is little evidence of a monsoon, most precipitation across the region remaining near the equator. The characteristic northward migration of precipitation in early summer first appears when $M = 40\%$; it strengthens as M is increased further. Only when orography is close to present values does heavy precipitation extend deep inland.

2

Controls on the Asian monsoon over tectonic timescales

2.1 Introduction

As described in Chapter 1 the intensity of the modern Asian monsoon can be understood as being principally a product of the seasonal temperature differences between the Indian and Pacific Oceans and the Asian continent. It is these differences that drive wind and weather systems in the strongly seasonal fashion that characterizes the climate throughout South and East Asia (Webster *et al.*, 1998; Figure 2.1). As a result, the long-term history of the Asian monsoon would be expected to extend as far back in time as the assembly of these major geographic features themselves, i.e., to the construction of Asia as we know it, following collision of the Indian and Asian continental blocks, generally presumed to be around 45 to 50 Ma (e.g., Rowley, 1996). Even the youngest estimates place the final collision of India and Asia at no younger than 35 Ma (Ali and Aitchison, 2005; Aitchison *et al.*, 2007).

Monsoon weather systems are not unique to Cenozoic Asia. Monsoon climate systems have been recognized in ancient supercontinents, such as Pangaea, where similar major temperature differences existed between land and ocean (e.g., Loope *et al.*, 2001). Indeed, even before the collision of India and Asia there must presumably have been a monsoon system operating between the Tethys Ocean in the south and Eurasia in the north. However, the intensity of that system would have been low because prior to the Indian collision Asia was somewhat smaller and less elevated than it has been for most of the Cenozoic (Wang, 2004). Furthermore, because global sea levels were higher in the early Cenozoic (Haq *et al.*, 1987) wide, shallow seas (Paratethys) covered much of the area of what is now central Asia. These waters had the effect of reducing the temperature contrast further and inhibiting monsoon strength (Ramstein *et al.*, 1997;

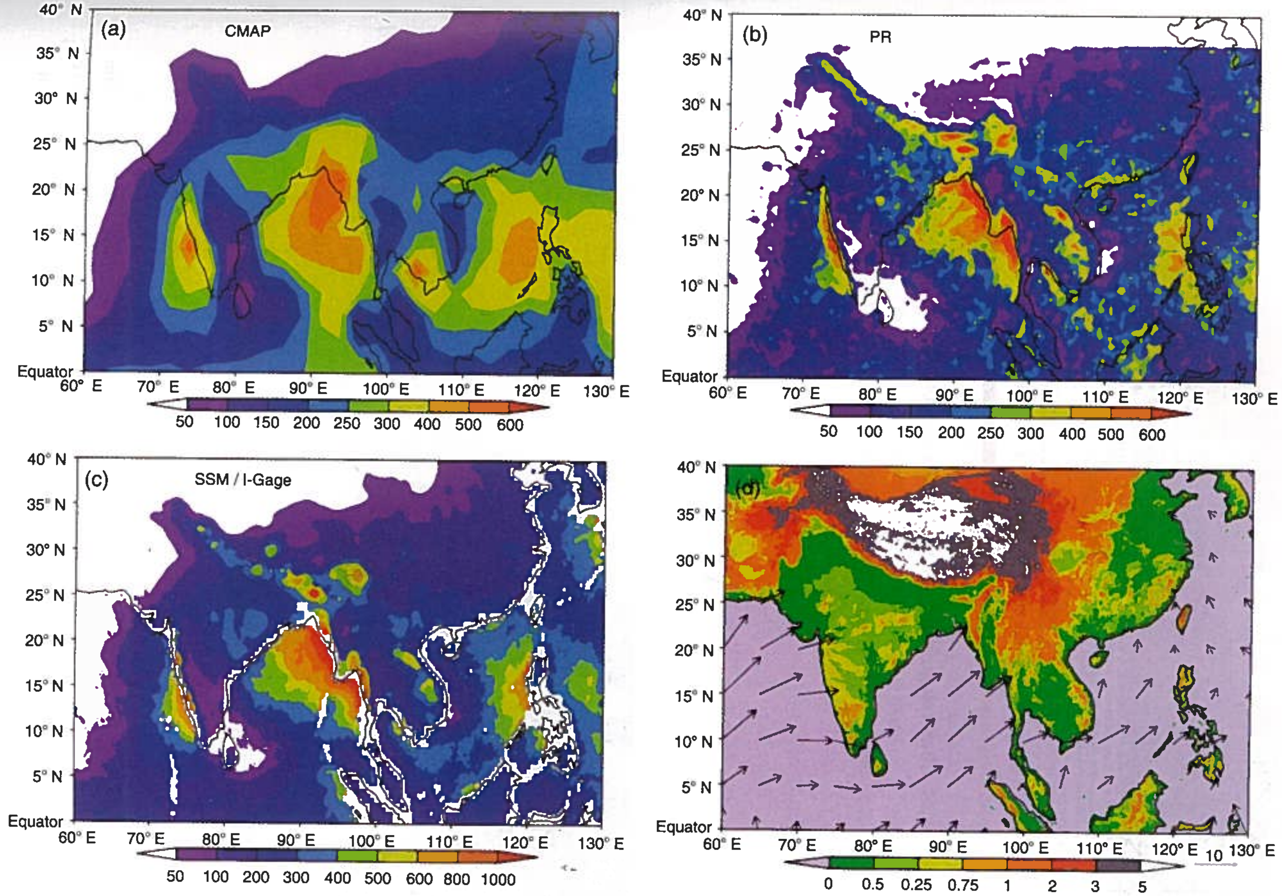

Figure 1.16 Summertime rainfall over the northern Indian Ocean and South Asia. Frames (a)–(c) show June–August mean rainfall rates (mm per month) from different data sources: (a) the Climate Prediction Center Merged Analysis of Precipitation (CMAP; 2.5° resolution, coverage 1979–2003); (b) high resolution data from the space-based Tropical Rainfall Measurement Mission Precipitation Radar (1998–2004) and (c) merged high resolution data from the space-based Special Sensor Microwave Imager and land-based observations (1987–2003). Surface topography (km) is shown in frame (d). (Xie *et al.*, 2006).