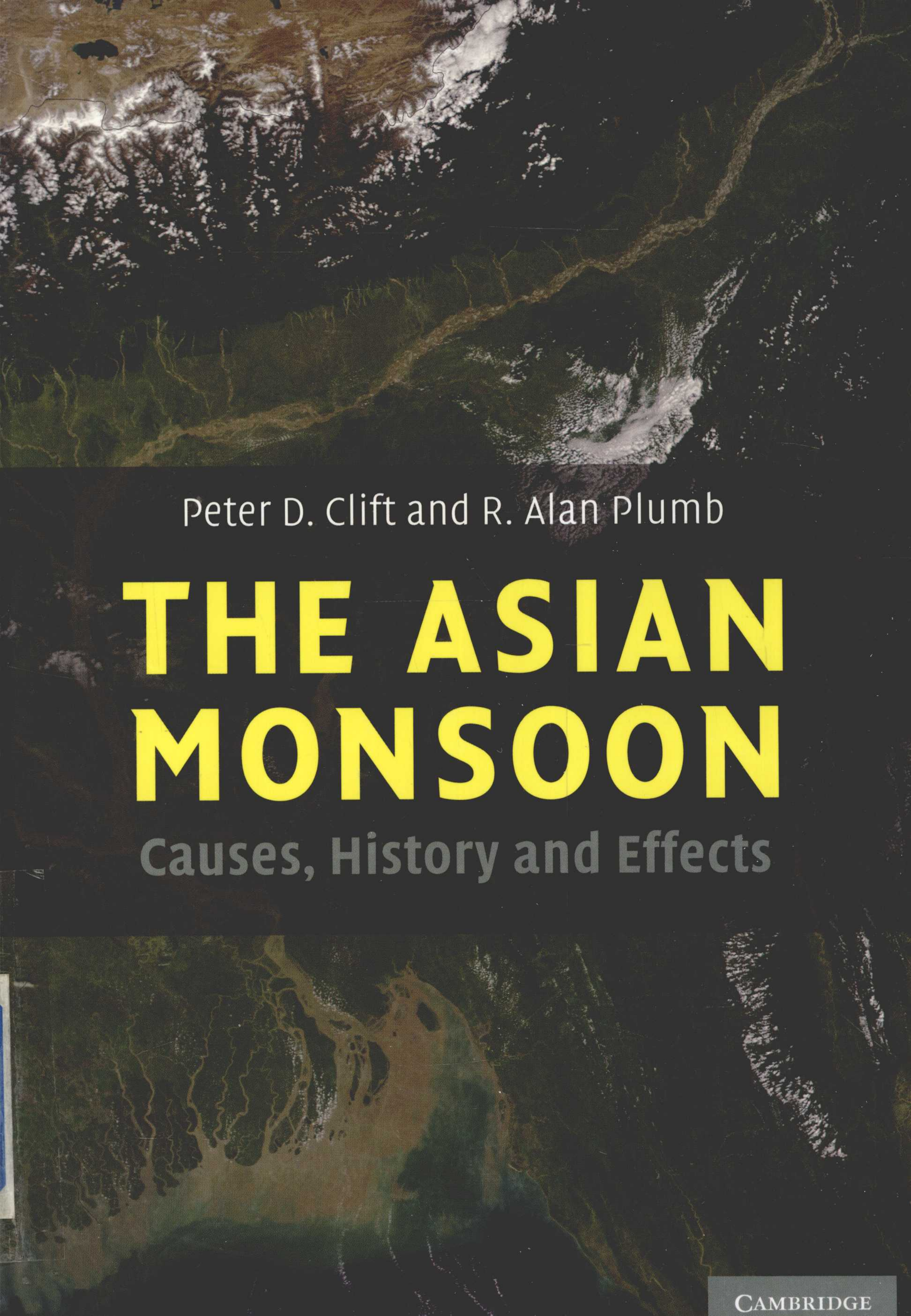

Peter D. Clift and R. Alan Plumb

THE ASIAN MONSOON

Causes, History and Effects

CAMBRIDGE

The Asian Monsoon

Causes, History and Effects

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The Asian Monsoon: Causes, History and Effects

The Asian monsoon is one of the most dramatic climatic phenomena on Earth today, with far reaching environmental and societal effects. But why does the monsoon exist? What are its driving factors? How does it influence the climate and geology of Asia? How has it evolved over long periods of geologic time?

Almost two-thirds of humanity lives within regions influenced by the monsoon. 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. Monsoon strength and variability have been and will continue to be crucial to the past and future prosperity of the region.

The Asian Monsoon describes the evolution of the monsoon on short and long timescales, presenting and evaluating models that propose a connection between the tectonic evolution of the solid Earth and monsoon intensity. The authors explain how the monsoon has been linked to orbital processes and thus to other parts of the global climate system, especially Northern Hemispheric Glaciation. Finally, they summarize what is known of the monsoon evolution since the last ice age and note how this has impacted human societies, as well as commenting on the potential impact of future climate change.

This book presents a multi-disciplinary overview of the monsoon for advanced students and researchers in atmospheric science, climatology, oceanography, geophysics and geomorphology.

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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.

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Contents

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.

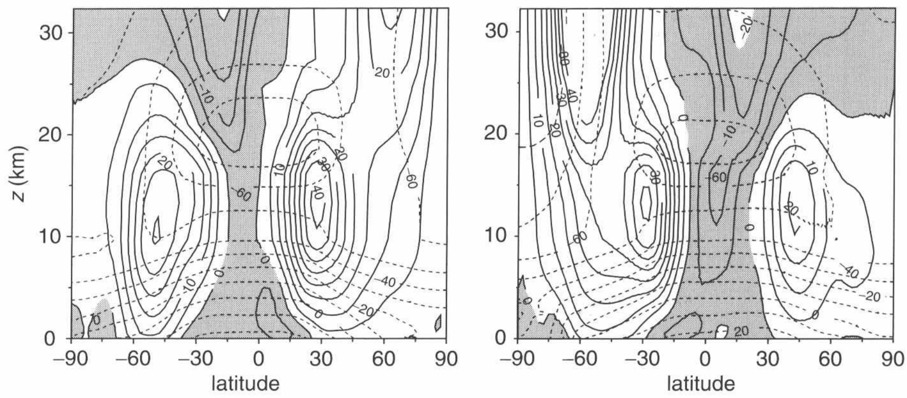

Figure 1.1 Climatological zonal mean zonal wind (solid; ms^{-1}) and temperature (dashed; °C) for (left) December–February and (right) June–August. Contour intervals are $5\,\mathrm{ms}^{-1}$ and 10°, respectively; easterly winds are shaded. The data are averaged on pressure surfaces; the height scale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

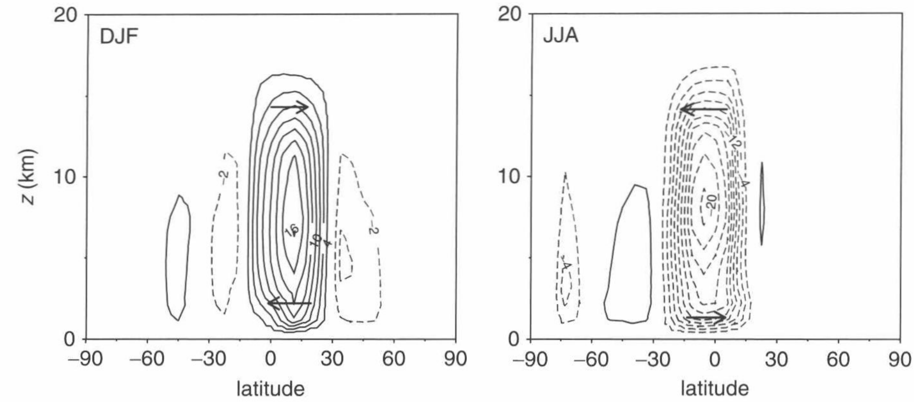

Figure 1.2 Climatological zonal mean overturning streamfunction (10 kg s^{-1} for December–February (left) and June–August (right). Solid contours denote positive values, dashed contours are negative; the zero contour is not plotted. The meridional flow is directed along the streamfunction contours,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

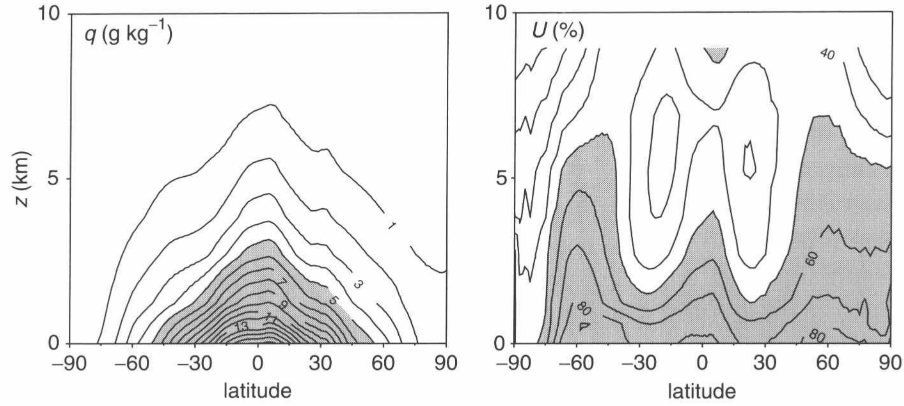

Figure 1.3 Climatological annual- and zonal-mean specific humidity (left, $g\,kg^{-1}$; values greater than $5\,g\,kg^{-1}$ are shaded) and relative humidity (right, %; values greater than 50% are shaded). Data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, through their website 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: $58\,\mathrm{ms}^{-1}$ at $20°$, $134\,\mathrm{ms}^{-1}$ at $30°$, $328\,\mathrm{ms}^{-1}$ 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;