See also

Monsoon: Dynamical Theory; ENSO–Monsoon Interactions; Prediction. **Tropical Meteorology:** Tropical Climates.

Further Reading

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Dynamical Theory

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Elements of a Monsoon Circulation

A monsoon is a circulation system with certain well-defined characteristics. During summer, lower tropospheric winds flow toward heated continents away from the colder oceanic regions of the winter hemisphere. In the upper troposphere the flow is reversed, with flow from the summer to the winter hemisphere. Precipitation generally occurs during summer, centered in time on either side of the summer solstice and located over the heated continents and the adjacent oceans and seas in the vicinity of a trough of low pressure referred to as the 'monsoon trough'. Most summer rainfall is associated with synoptic disturbances that propagate through the region. However, these disturbances are grouped in periods lasting from 10 to 30 days. Such envelopes of disturbed weather and heavy rainfall are referred to as 'active periods of the monsoon'. The intervening periods of mini-drought are referred to as 'monsoon breaks'. The location of the monsoon trough and axis of heavy monsoon precipitation is generally well poleward of the position of the oceanic intertropical convergence zone (ITCZ), within which the majority of tropical oceanic precipitation occurs. For example, the rainfall associated with the South Asian monsoon falls at the same latitudes as the great deserts of the planet.

Monsoon systems are associated with colocated pairs of continents such as Asia and Australia, or continents straddling the Equator such as north-west and south-west Africa, and North and South America defining, respectively, the Asian–Australian monsoon system, the West African monsoon, and the American monsoon. Each system is different in terms of intensity and circulation characteristics. For example, the northern arm of the American monsoon is a relatively weak counterpart of the other major monsoon systems and there does not appear to be a discernible cross-equatorial component during the summer. In that sense, the North and South American monsoons may be thought of as almost separate entities. Rainfall that occurs over the continents that span the Equator (e.g., equatorial Africa and South America, and Indonesia) is not strictly monsoonal and possesses double rainfall maxima occurring with the equinoxes. Monsoon climates, on the other hand, possess a single solstitial rainfall maximum, while solstices demark the dry seasons for equatorial climates.

Basic Driving Mechanisms of the Monsoon

It is helpful to consider first a simple prototype geography that will allow us to identify the basic elements of a monsoon system. The geographical model we adopt is an oceanic planet with a continental cap extending from the subtropics to the pole in one hemisphere. After establishing the important processes that drive the monsoon for this simple geography, we will return to the consideration of local influences.

Monsoons arise from the development of cross-equatorial pressure gradients produced or modified by the following physical properties of, or processes associated with, the land–ocean–atmosphere system: differential heating of land and ocean produced by the different heat capacity of land and water; the different manner in which heat is transferred vertically and stored in the ocean and the land; modification of differential heating by moist processes; the generation of meridional pressure-gradient forces resulting from the differential heating; and the meridional transport of heat in the ocean by dynamical processes. Each of these processes and properties has to be considered relative to the rotation of the planet, and the influence of local effects such as the geography of the ocean and the land masses, and regional topography.

Differential Heating

Heat capacity differences There is roughly a factor of 4 difference between the specific heat of water ($4218\,\mathrm{J\,kg^{-1}\,K^{-1}}$) and dry land (roughly $1300\,\mathrm{J\,kg^{-1}\,K^{-1}}$). Wet soil may have a heat capacity 30% higher than dry soil. For some net heating rate, the temperature of a mass of dry land the increment in temperature will be nearly four times greater than that of a similar mass of water. In the late seventeenth century, Halley (of Halley's comet) was the first to suggest that monsoon circulations were driven by heating gradients produced by the heat capacity differences between the land and the ocean and used his theory to explain aspects of the West African and South Asian surface monsoon winds that had been reported by explorers and traders. He also understood the role of the annual cycle of solar heating that produced the strong seasonality of the monsoon and the reversal of the circulation during the winter.

Halley had defined a basic factor that determines the existence of monsoons. However, to understand how different heat capacities produce motion and why the Halley's theory has to be expanded, it is necessary to delve deeper into the physics of the atmosphere and the ocean. If the heat flux into the surface layer is F ($\mathrm{W\,m^{-2}}$) and if there is no heat flux out of the bottom of the layer at some depth $z = z_1$ (m), the heating rate of the layer will be determined by the flux divergence in the layer (eqn [1]).

$$\frac{\mathrm{d}T}{\mathrm{d}t} = -\frac{1}{\rho C_p}\frac{\mathrm{d}F}{\mathrm{d}z} = \frac{1}{\rho C_p}\frac{F_{z=0}}{\Delta z} \qquad [1]$$

In eqn [1], $F_{z=0}$ is the net flux at the surface and Δz is the thickness of the layer. The surface energy balance is given by eqn [2], where I_{net} is the net radiation at the surface given by the sum of the net solar radiation, the upwelling infrared radiation, and the re-radiation from the atmosphere (the greenhouse effect), respectively.

$$F_{z=0} = I_{\mathrm{net}} - H_{\mathrm{s}} - H_{\mathrm{e}}, \qquad [2]$$

I_{net} is given by eqn [3], where S is the solar flux at the surface, a is the system albedo, ε is the emissivity of the atmosphere, T_{g} and T_{a} are the surface and atmospheric temperatures; H_{s} and H_{e} are the sensible and latent turbulent heat transports away for the surface as described in **Figure 1**.

$$I_{\mathrm{net}} = S(1-a) + \varepsilon\sigma T_{\mathrm{g}}^4 - \sigma T_{\mathrm{a}}^4 \qquad [3]$$

From eqn [1] it is apparent that the heating rate of a slab will depend on the heat capacity the layer, its thickness, and the net energy flux into or out of the layer at the surface. The temperature of a motionless oceanic slab (i.e., no vertical mixing or horizontal advection) is determined by the net heating at the surface. South of the Equator, in the winter hemisphere, the slab ocean would cool by a combination of evaporative cooling and negative net radiational heating. To the north of the Equator, the ocean would heat if net radiational heating exceeded the evaporative cooling. During summer the land heats more rapidly than the adjacent ocean because of its smaller specific heat and shallow Δz. These factors easily compensate for the fact that dry land has a larger albedo than the ocean (20–40% versus about 10%). In the winter the land surface will cool much more quickly than the ocean simply because there is little available heat in the subsurface that can be made available to heat the surface on seasonal time scales because of the slowness of the diffusive processes. Given that the sensible heat exchange between the land surface and the atmosphere depends to a large degree on their temperature difference, the atmospheric column over the land will be warmer than over the ocean.

In the simple model described above, the differences between the heating rates of land and the ocean reside in their different heat capacities and densities, and the depth Δz of the slab that is defined as the depth over which the heating is spread. Over land, the Δz is very small because of the opacity of the soil to radiation, and the depth of the 'active' layer in which there is a discernible signal of the annual cycle is only a meter or so. This depth is constrained by the inefficiency of conductive heat transfer. If the ocean is assumed to be immobile, then its effective depth over which heating is spread may be defined as the *e*-folding depth of solar radiation. Here radiative transfer and the opacity of the ocean determine the effective depth. Observations suggest that the solar radiation *e*-folding depth is about 10 m. The ocean temperature variation will also lag the surface heating. This may be seen by setting the surface flux, $F(z=0)$ in eqn [1], proportional to $\sin(\omega t)$. In this case the temperature variation will be proportional to $-\cos(\omega t)$, therefore lagging the forcing by a quarter period.

In summary, when the ocean is assumed to be immobile, an annual cycle of ocean–land temperature difference and meridional pressure-gradient force is achieved. Substitution of numbers into eqns [1]–[3] shows that the variability of the ocean temperature is much larger than observed as long as a depth of about 10 m is used. More realistic amplitude can be achieved by 'tuning' the depth of the active ocean layer to be considerably greater. It turns out there are good physical reasons why we may expect a deeper active layer.

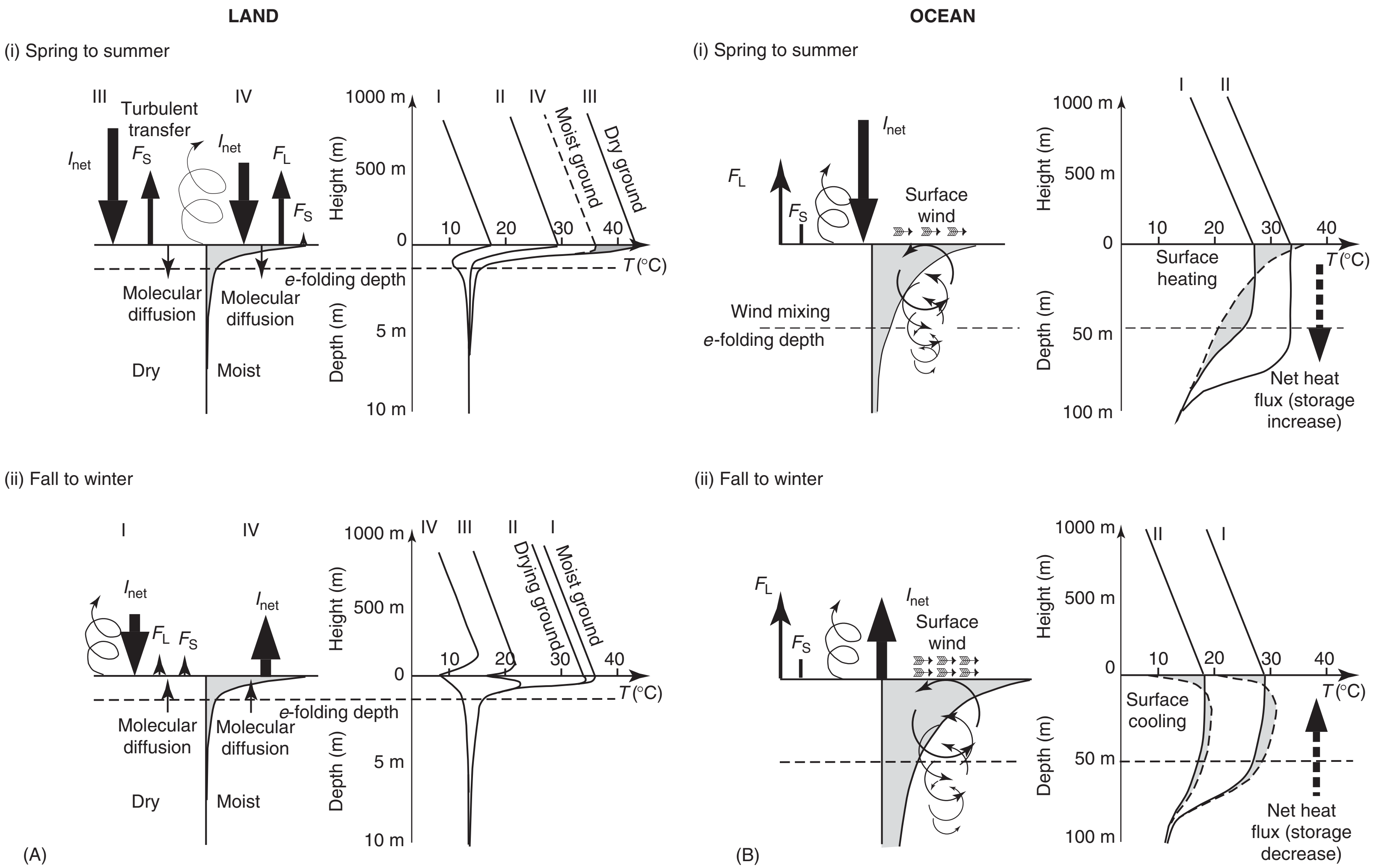

Figure 1 Schematic diagram of the dominant physical processes determining the surface temperature over (A) land, and (B) the ocean for the transitions from (i) spring to summer and (ii) fall to winter. The radiative and turbulent fluxes are labeled relative to eqns [1]–[3]. Relative *e*-folding depths of diffusion, solar radiation penetration, and turbulent mixing are shown as horizontal dashed lines in each section. The net incoming radiation is attenuated in the first few millimeters of soil, from where it is transferred by molecular diffusion for a few meters. Wind stress at the ocean surface causes substantial transfers of heat (upward in winter, downward in summer) between the surface layers and the subsurface ocean by inducing turbulent mixing. Successive lower tropospheric temperature profiles are also shown (marked I, II, and so on). The shading in (A(i)) shows the change in temperature if the soil is moist. The shading in (B) shows the changes in the ocean temperature profiles when heated surface water moves downward in the summer or when cooled surface water is replaced by warmer subsurface water.

Mixing and storage of heat So far, the fluid nature of the ocean has been ignored. In a fluid, wind forcing and gravitational instabilities formed by the cooling of the surface layer may induce turbulence and mixing of the surface and subsurface water. Wind stress also can move a body of water horizontally, producing ocean currents that can advect heat and mass from one place in an ocean basin to another. The impact of lateral transports will be considered later. Stable layers near the surface can be produced by the freshening effect of precipitation. These fresh layers may reduce the impact of wind stirring. With these factors in mind, we can return to the consideration of the heat balances of the ocean and the land regions and the atmospheres above. These processes are shown schematically in **Figure 1**.

During the summer, when the net heating of the ocean surface is positive, wind-induced turbulence mixes the warm surface water downward. As long as the net surface flux of energy into the ocean is positive, wind mixing will increase the heat content (or heat storage) of an ocean column. The mixing is very effective and observations show that in the tropical ocean a constant-temperature mixed layer may extend down below the surface to depths of 50–100 m. During winter, when the net surface heating is negative, the colder surface water (formed by the negative heat balance at the surface) is mixed downward to be replaced by warmer subsurface water that had been mixed down into the ocean column during the previous summer. As long as the surface energy balance is negative, wind-induced turbulence will decrease the total heat content (i.e., reduce heat storage) in the ocean column. As turbulent mixing occurs over a much deeper layer than the *e*-folding depth of solar radiation, the heat absorbed in the surface layer of the ocean is spread through a depth greater than the *e*-folding penetration depth of solar radiation. That is, Δz is larger and the overall sea surface temperature (SST) changes are smaller in the presence of turbulence than if the ocean were immobile.

The impact of changes in heat storage on the ocean temperature is twofold. First, it moderates the SST, which in turn modulates the temperature and moisture content of the air adjacent to the ocean surface. Atmospheric turbulent mixing produced either mechanically by wind stress or by buoyancy effects extends the imprint of the SST into the troposphere. Second, the mixing processes in the ocean column produces the observed lags between the ocean temperature and the solar cycle. Land surface temperature tends to follow the solstices, although, because of moist processes, the maximum land temperature occurs before the onset of the summer rains.

The Generation of Monsoonal Pressure Gradient Forces

To account for the observed reversal of the monsoon circulation with height we require a basic driving force that changes in magnitude or reverses with height. The only force available is the horizontal pressure-gradient force. It is relatively simple to show that the horizontal pressure-gradient force between the summer and winter hemispheres may change with height and, under certain circumstance, even reverse. Eliminating density between the equation of state and the hydrostatic equation gives eqn [4].

$$\frac{\partial p}{\partial z} = -\frac{g}{R}\frac{p}{\overline{T}} \qquad [4]$$

In eqn [4], $p(z)$ is the atmospheric pressure, g is the acceleration due to gravity, R is the gas constant, and $\overline{T}$ is the mean temperature of the atmospheric column. Equation [4] states that the change of pressure with height is inversely proportional to the mean temperature of the column. Therefore, over the warm summer continent, the pressure will decrease with height at a lesser rate than over the cold ocean, as shown in **Figure 2**. The relationship can be explored by integration in the vertical through the thickness of a slab of atmosphere between heights $z = 0$ and $z = z_1$. The difference in pressure $\Delta \ln p(z)$ at height $z = z_1$ between the warm and cold columns of **Figure 2** can be expressed as eqn [5], where $\overline{T}_c$ and $\overline{T}_w$ are the mean temperatures of the atmospheric columns over the heated land and the cooler ocean.

$$\Delta \ln p(z_1) = \frac{g}{R} z_1 \left(\frac{1}{\overline{T}_c} - \frac{1}{\overline{T}_w} \right) + \Delta \ln p(0) \qquad [5]$$

Here, Δ refers to the difference of a quantity between the warm and cold columns along a constant height surface. From eqn [5], the condition for $\Delta \ln p(z_1) > 0$, assuming that the surface pressure difference between the warm and cold columns is zero (i.e., $\Delta \ln p(0) = 0$), is that $\overline{T}_w > \overline{T}_c$. In this case air above the surface will be forced to flow from the summer to the winter hemisphere, with mass continuity providing a lower tropospheric return flow from the winter to the summer hemisphere.

However, in general, the surface pressure over the winter subtropics is higher than the surface pressure over the heated continent, perhaps by as much as 20 hPa. In fact, as the solar heating increases over the continent, the surface pressure is observed to fall so that $\Delta \ln p(0)$ becomes increasingly negative (**Figure 3**). Thus the criterion $\overline{T}_w > \overline{T}_c$ is not sufficient to ensure that there will be a reversed upper tropospheric pressure gradient and a return flow to the winter

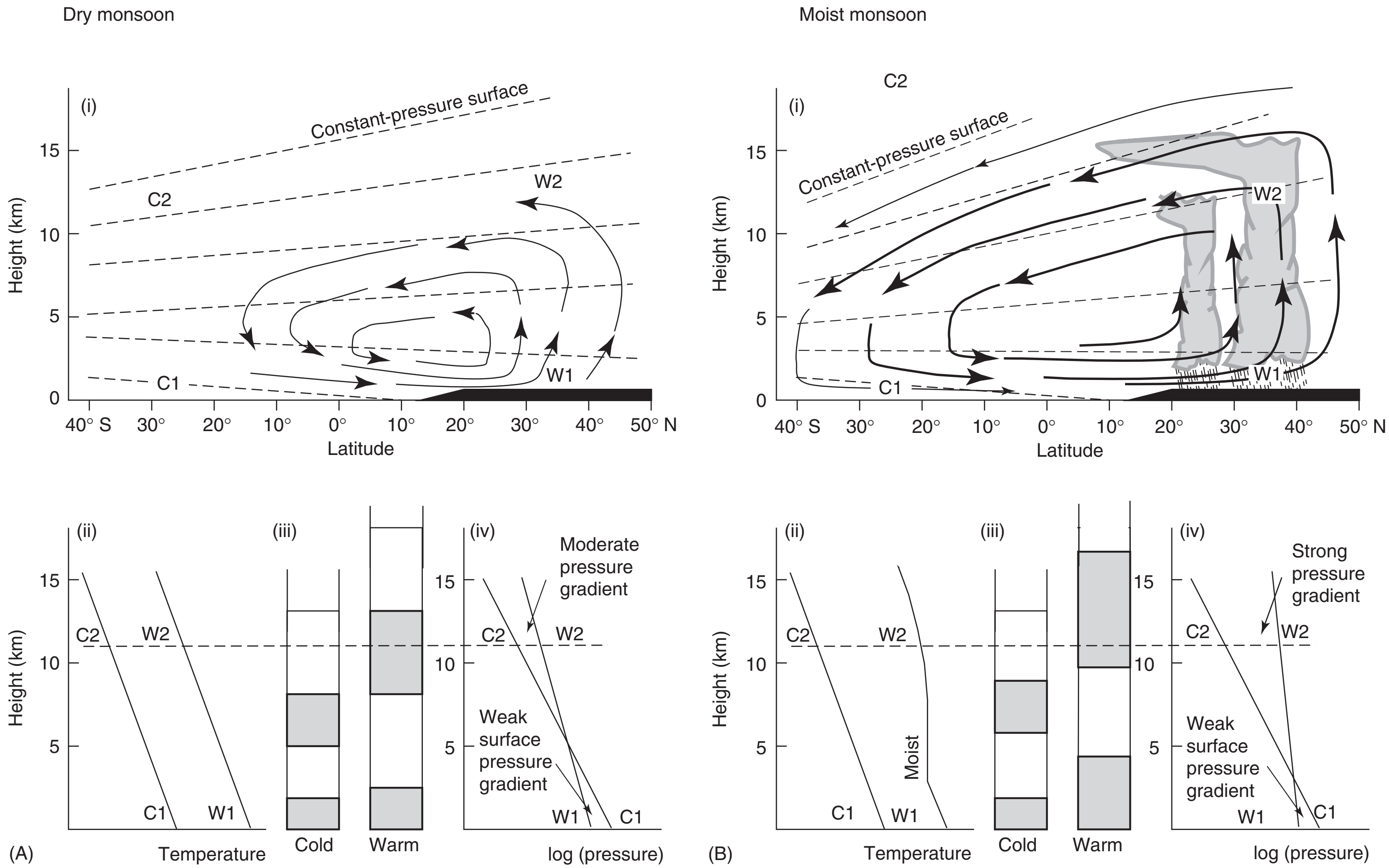

Figure 2 A mechanistic view of the development of the meridional monsoon circulation (A) when moist processes are ignored and (B) when moist processes are taken in to account. The panels show (i) the resultant circulation, (ii) the temperature profiles, (iii) the distribution of mass in the vertical columns, and (iv) the change of pressure with height. Dashed lines in panel (i) show constant-pressure surfaces. Dashed lines in panels (ii)–(iv) denote a constant height. In both examples it is assumed that the difference in temperature of the warm and cold columns is sufficient to generate a reversing pressure gradient with height in the presence of the surface pressure gradient as described in eqn [6]. The figure is discussed extensively in the text.

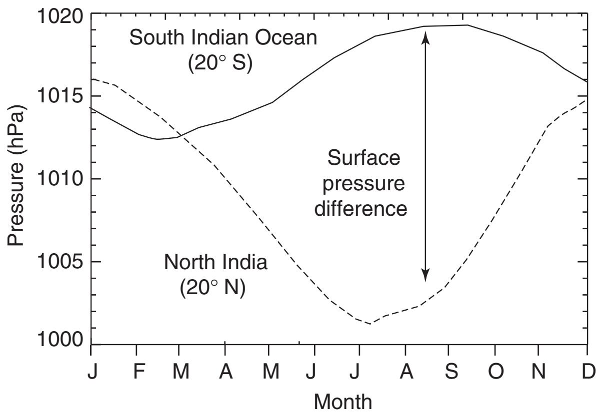

Figure 3 The mean annual cycle of sea-level pressure at 20° N and 20° S along 80° E, representing the monsoon trough and South Indian Ocean, respectively.

hemisphere. From eqn [5], a general condition for the temperature difference for $\Delta \ln p(z_1) > 0$ in the presence of a surface pressure gradient can be found (eqn [6]).

$$\overline{T}_{\mathrm{w}} > \frac{gz_1\overline{T}_{\mathrm{c}}}{gz_1 + R\,\Delta\,\ln\,p(0)\overline{T}_{\mathrm{c}}} \qquad [6]$$

These simple arguments suggest that there may be a threshold in temperature difference between the winter hemisphere and the summer hemisphere in the presence of a surface pressure gradient, before a reverse pressure gradient in the upper troposphere is established. At that stage, pressure gradients throughout the troposphere are conducive for the maintaintence of a direct thermal circulation. This threshold in mean tropospheric temperature gradient may be the reason for the observed sudden onset of the monsoon over South Asia in late May or early June, at which time deep convection and heavy precipitation occur.

The post-monsoon onset circulation, in the absence of moist processes, is shown in **Figure 2A**. If it assumed that the monsoon is in steady state, the amount of heat gained by surface heating must be balanced by heat lost to space by radiative processes. For a given stratification, the vertical extent of the dry monsoon is determined by the input of heat at the lower boundary. The longitudinal extent of the circulation is determined by the time it takes for a parcel to radiate away excess heat gained at the continental surface. If radiative processes were very efficient, the longitudinal scale of the monsoon would be very small. However, radiative processes are slow, with *e*-folding dissipative time scales of about 20 days. Thus, the parcel takes a considerable time to cool and a parcel in the upper troposphere travels a considerable distance while cooling.

Moist Processes and the Monsoon Solar Collector

So far, moist processes have been ignored except for their implicit inclusion in surface evaporation. Moist processes change the character of the monsoon by moistening the land surface and being the agent of strong mid-tropospheric heating through the release of latent heat over the summer continent or adjacent marginal seas.

The source for summer monsoon rainfall is water evaporated from the ocean as air flows toward the heated continent under the action of the pressure-gradient forces discussed above. **Figure 4** shows the source regions of moisture for the monsoons. The figure plots the vertically averaged moisture transport, B_{q}, defined as in eqn [7], where $q(z)$ and $\tilde{\mathbf{V}}(z)$ are the specific humidity and the horizontal velocity vector, respectively.

$$B_{\mathrm{q}} = \int_0^{\infty} q\tilde{\mathbf{V}}dz \qquad [7]$$

As moisture tends to decrease exponentially above the surface, the greatest contributions to B_{q} come from the surface boundary layer. Moisture accumulation zones can be seen to the south of Asia extending well into the winter hemisphere during the boreal summer (**Figure 4A**) and, to a lesser extent, to the north of Australia during the boreal winter (**Figure 4B**). Even though evaporation cools the surface of the ocean, the boundary layer air flows across a gradient of increasing SST due to the net positive radiation budget at low latitudes. Consequently, the boundary layer air becomes warmer along its trajectory and, as the surface saturated vapor pressure increases, the moisture content of the boundary is elevated.

One might imagine that the dry monsoon model shown in **Figure 2A** is applicable to the monsoons during spring. Between the spring equinox and the summer solstice, the temperature of the land increases, producing a low-level pressure gradient, causing a steady advection of moist air toward the continent. Eventually, sufficient water vapor will be imported over the land so that rising motion will result in the release of latent heat and an increase in temperature of the continental atmospheric column, eventually producing a reversed pressure gradient at higher levels. The strengthening of the monsoon occurs with the rapid development of the upper-tropospheric meridional temperature gradient. The increase in columnar temperature necessary to produce the reversal (see the previous Section) is directly attributable to the release of latent heat. At this point, the acceleration of the monsoon is substantial. Surface winds that were relatively weak prior to the onset of the monsoon exceed $10\,\mathrm{m\,s^{-1}}$ at the surface when the monsoon is

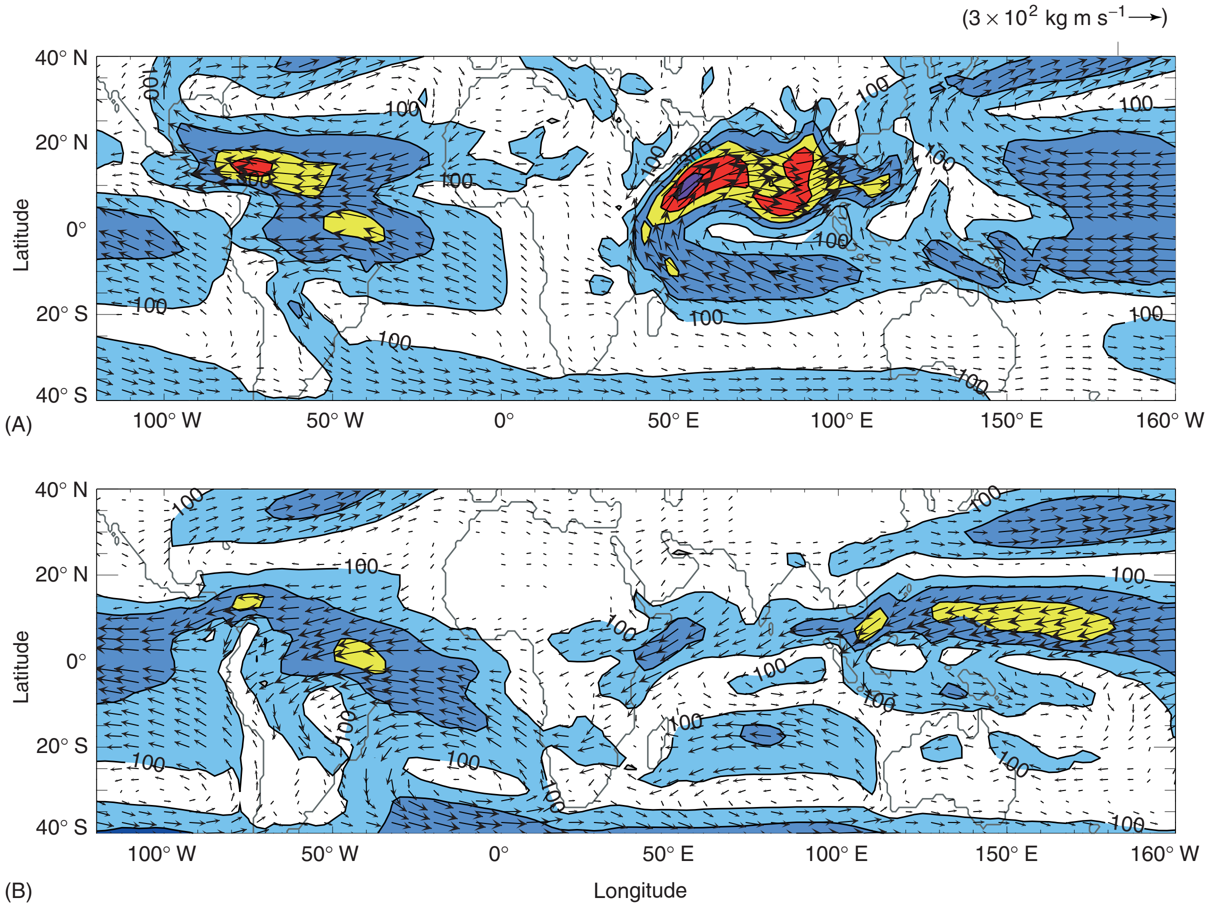

Figure 4 Distribution of mean vertically integrated moisture transport from eqn [7] for the period (A) June–September and (B) December–February. Viewed in the context of moisture transport, the Asian–Australian monsoon system appears in both (A) the boreal summer and (B) the boreal winter as strong interhemispheric systems with moisture sources clearly defined in the winter hemisphere. Both the African summer and winter monsoons are less clearly defined. Weak moisture fluxes into north-west Africa are evident, for example, but the region is dominated by strong westward moisture fluxes associated with the Trade Wind across the Atlantic. Furthermore, the moisture fluxes associated with the North and South American monsoons appear restricted to their respective hemispheres. Only the Asian–Australian monsoon possesses a truly interhemispheric solar collector.

established. Evaporation and increased ocean mixing accompany the strengthening winds and the SST of the North Indian Ocean drops rapidly by 1–2°C.

At the time of the year when large-scale precipitation occurs, two very important transitions occur in the monsoon. First, the dry land area becomes moist, sometimes so wet that it adopts many of the characteristics of a warm shallow ocean or lake (**Figure 1**). Accompanying the surface moistening is a substantial decrease of the surface land temperature. For example, the daily maximum surface temperature of New Delhi, India often exceeds 45°C prior to the onset of the monsoon, while during active rainy periods following the monsoon onset the surface temperature maxima are between 30°C and 32°C. At the same time, the intensity of the monsoon increases substantially, with both the surface and upper tropospheric winds strengthening considerably.

It would seem that the simultaneous decrease in surface temperature of the land and the cooling of the adjacent ocean with the strengthening of the monsoon would act as a negative feedback and that the monsoon strength should decrease. However, the stronger winds also cause a very large increase in the amount of water vapor imported over the continental regions. The enhanced convergence of water vapor causes the release of latent heat to increase substantially. Thus, once the monsoon strengthens, the importance of the surface temperature gradient decreases and the overall driving of the monsoon is taken over by the heating of the troposphere over the continents through the release of latent heat. Also, prior to precipitation, the depth of the dry circulation is relatively shallow (**Figure 2A**). However, in the moist monsoon, the circulation occupies the entire troposphere as a result of the buoyant moist parcels releasing their latent energy as they ascend to great heights (**Figure 2B**). Furthermore, the vertical lapse rate decreases, moving closer to moist adiabatic. Thus, even though the surface temperature over the land decreases, the average tropospheric temperature increases.

The monsoon may be thought of as a great solar collector. The latent heat of condensation that is released over the heated continents, and which drives the established monsoon, is the summation of the evaporation occurring at the surface of the ocean between the winter hemisphere and the heated continents (**Figure 4**). As the evaporation results from a net positive radiation excess at the ocean surface or from the winds driven by the SST gradients, evaporation may be thought of as the integration of the solar heating across the ocean. Thus, the latent heating of the atmosphere over the continents is the accumulation of a large fraction of the solar energy incident over the vast ocean fetch of the monsoon winds. However, the effect of this solar heating is concentrated into a relatively small region of summer hemisphere continents and the adjacent seas where precipitation occurs.

The vertical and longitudinal scales of the moist monsoon are much larger than those of the dry monsoon as may be seen by comparing **Figures 2A** and **2B**. With moist processes, there is a very large amount of heat added to the system in the vigorous moist ascent over the heated continents. This is heat accumulated during the long trajectories over the oceans: the solar collector effect. The input of energy is far larger than in the dry monsoon and, as a result, the amount of time required for the upper tropospheric flow to radiate excess heat away is much longer. Thus, the upper tropospheric pressure gradient is maintained over a much larger scale and the moist monsoon acquires a correspondingly large longitudinal scale.

The Impact of Rotation

Halley was the first to understand the fundamental role of land–sea differences in producing onshore monsoon winds. Some 50 years, later in 1735, Hadley published a paper that considered the effects of the rotation of the planet and produced a general theory for the Trade Wind regime and the monsoons.

The monsoon model we have developed so far defines a circulation system in which air parcels move about under the action of body forces whose direction and magnitude are determined by the distribution of heating. In essence, we have added to Halley's monsoon model the physics of moist processes and a more realistic view of the ocean heat capacity. However, the planet is rotating and the effects of rotation are extremely important in the equatorial regions as the Coriolis force ($f = 2\Omega \sin \phi$) changes sign and the gradient of the Coriolis force ($\beta = 2\Omega(\cos \phi)/a$) is a maximum. As monsoon motion is cross-equatorial, the dynamics of low-latitude phenomena become very important. Here we restrict ourselves to the discussion of two immediate effects of rotation on the monsoon system: the general form of the flow under the action of a constant cross-equatorial pressure gradient on a rotating planet, and the forcing of boundary layer mass and heat transports in the upper ocean.

Impacts on the atmosphere The horizontal equation of motion on a rotating platform may be written as eqn [8], where α (s^{-1}) is a dissipation coefficient. In the boundary layer, α^{-1} is large and has an e-folding scale of about 2 days.

$$\frac{d\mathbf{V}}{dt} = -\frac{1}{\rho}\nabla p - f\,\mathbf{k} \times \mathbf{V} - \alpha\mathbf{V} \qquad [8]$$

Consider now a constant pressure gradient in the lower troposphere extending from high pressure in the wintertime subtropics to the low pressure over the heated continent in the summer hemisphere. Without rotation, steady flow will occur when the frictional force (proportional to wind speed) balances the pressure gradient force. From eqn [8] this balance may be written as eqn [9], where v is the meridional component of the velocity vector.

$$\alpha v = \frac{1}{\rho}\frac{\partial p}{\partial y} \qquad [9]$$

The resultant flow, which is purely meridional, is depicted schematically in **Figure 5A(i)**. With rotation, steady flow comes about from a balance between the Coriolis force ($f\,\mathbf{k} \times \mathbf{V}$ in eqn [8]), and the pressure gradient force ($\partial p/\rho\partial y$). From eqn [8], the balance may be expressed as in eqn [10]

$$f\,\mathbf{k} \times \mathbf{V} + \alpha\mathbf{V} = -\frac{1}{\rho}\frac{\partial p}{\partial y} \qquad [10]$$

Between the high-pressure region and the Equator the flow must cross the isobars, as shown in **Figure 5A(ii)** because of the frictional force eventually adopting the form given by eqn [9] at the Equator. At all latitudes in this simple example, the flow is across the pressure gradient, flowing from low to high pressure. The resultant surface flow diverges out of the winter hemisphere surface anticyclone, flows across the Equator, and finally converges into the continental low-pressure region over the heated continent. In the

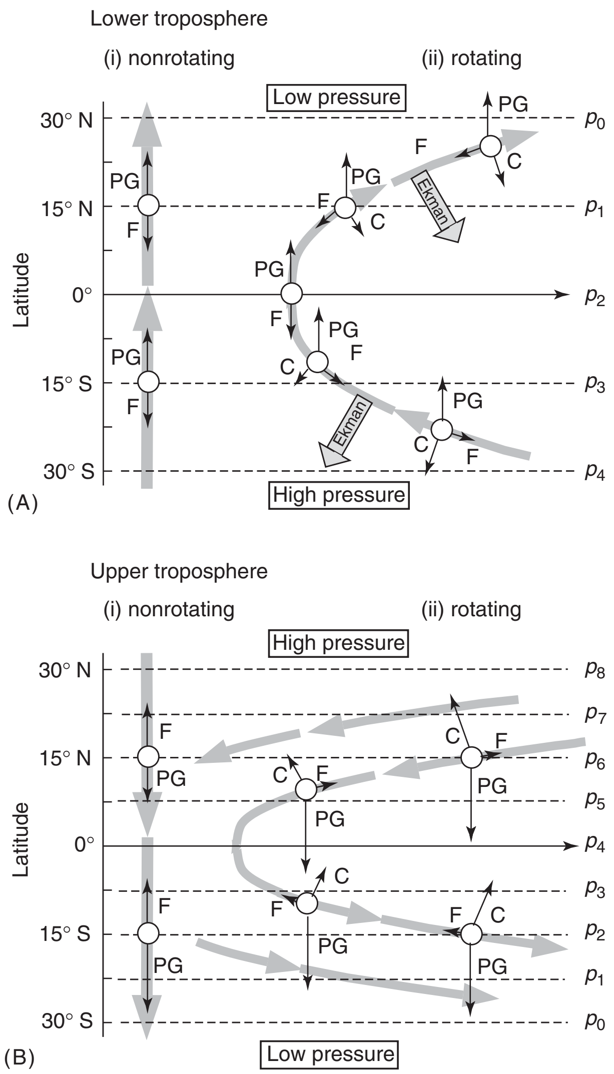

Figure 5 The circulation across the Equator in (A) the lower troposphere and (B) the upper troposphere on (i) a nonrotating planet, and (B) a rotating planet. Circulation is forced by a pressure-gradient force (PG) that is constant in latitude and subject to Coriolis effects (C) and frictional dissipation (F) that is assumed to be proportional to the parcel speed. The broad arrows denote the integrated Ekman mass ocean transport forced by the surface winds.

winter, when the polarity of the pressure systems is reversed, the flow is a mirror image to the flow shown in **Figure 5A**.

As discussed earlier, the flow in the upper troposphere is in the opposite meridional direction through the action of a reversed pressure gradient force that increases with height. Furthermore, compared to the surface boundary layer, the dissipation coefficient (α) is an order of magnitude smaller. The resulting flow is closer to geostrophic than the surface boundary layer flow but still cross-gradient as it spirals out of the upper anticyclone and moves westward. The trajectory that an upper-tropospheric air parcel takes is such that it will move a much greater distance in the longitudinal direction (compared to the surface flow) before it crosses the Equator and eventually descends into the winter hemisphere. The strong upper tropospheric easterlies are referred to as the Easterly Jet that extends from South Asia across East and Central Africa and attains speeds of 40–50 m s^{-1}. The extent of the Easterly Jet can be seen in **Figure 6A**, which shows the mean northern hemisphere 200 hPa flow.

The vertical variation of the geostrophic flow is given by the thermal wind equation, which can be obtained by differentiating eqn [10] in the vertical, using the equation of state and neglecting frictional effects. This gives eqn [11], where $\overline{T}$ is the mean temperature of a layer.

$$\frac{\partial \mathbf{V}_g}{\partial z} = -\frac{g}{f\overline{T}}\nabla\overline{T}\times\tilde{\mathbf{k}} \qquad [11]$$

Thus if the mean temperature of a layer decreases toward the poles, then the vertical shear will be positive (e.g., lower tropospheric easterlies and upper tropospheric westerlies) in either hemisphere. However, in the monsoon regions, the shear is negative (lower level south-westerlies and upper level easterlies) so that temperature must increase towards the pole. **Figure 7** plots the mean global upper tropospheric (500–200 hPa) temperature for the boreal summer and winter. During the boreal summer, the upper troposphere of South Asia/Tibetan Plateau region is the warmest region on the planet. Importantly, the mean temperature is >5°C warmer than the Equator at the same longitude. During the boreal winter, a weak mean upper tropospheric temperature maximum exists in the vicinity of North Australia. The anomaly is about 1°C warmer than at the Equator. This reversed temperature gradient is sufficient to drive a weak upper tropospheric easterly jet stream over North Australia. Thermal ridges may be seen over other parts of the world (e.g., North America), but nowhere else does the temperature gradient reverse between the Equator and the pole except over South Asia and North Australia.

The Easterly Jet produced by the reversed temperature gradient has a pronounced effect on the North African monsoon. **Figure 6A** shows maximum jet speeds are attained over the central North Indian Ocean. Over central and western Africa, the flow decelerates in the exit region of the jet. The deceleration of the jet stream is important as it produces a secondary circulation over West Africa that is in the opposite sense to the local monsoon direct circulation. The circulation is produced in the following manner. Consider a parcel flowing through the Easterly Jet in geostrophic balance. From **Figure 6A** it is evident the

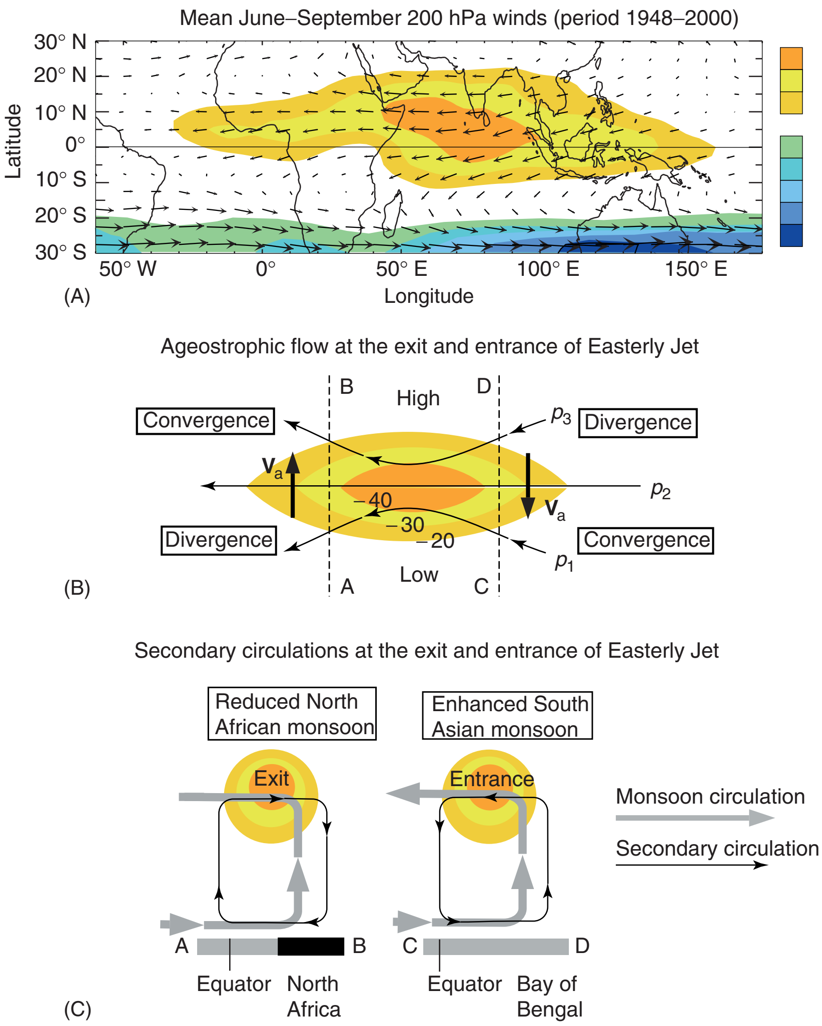

Figure 6 Impacts of the large monsoon heating over South Asia. (A) The upper tropospheric circulation. Note the extent of the Easterly Jet that commences over the north-eastern Indian Ocean and extends out across the Atlantic Ocean. (B) The secondary circulation associated with the entrance and exit regions of an easterly jet stream. Note that the ageostrophic flow produces a northward deflection of the winds that is consistent with subsidence on the northward side of the jet and rising surface pressure. (C) Schematic diagram showing the secondary circulation in opposition to the monsoon meridional cell.

parcel moves through the jet stream it will experience first an increase in easterly wind speed (i.e., $du_g/dt < 0$) in the entrance region, where u_g is the zonal component of the geostrophic wind. In the exit region of the jet, the easterly wind decreases in intensity (i.e., $du_g/dt > 0$). The zonal momentum equation can be written as eqn [12], where we have decomposed the meridional velocity component into a geostrophic and ageostrophic component (i.e., $v = v_g + v_a$).

$$\frac{du_g}{dt} = fv - \frac{1}{\rho}\frac{\partial p}{\partial x} = f(v_g + v_a) - \frac{1}{\rho}\frac{\partial p}{\partial x} = fv_a \qquad [12]$$

Thus, in the entrance region of the jet there must be a southward displacement of the parcel. On leaving the jet, the parcel must move northward. This ageostrophic flow is shown in **Figure 6B**. Over Africa, the result is the production of a secondary circulation made up of the ageostrophic northerly flow with descending air on the poleward side of the jet stream and ascending air on the equatorward side. This secondary circulation is in opposition to the monsoon meridional circulation that would have rising air over North Africa (**Figure 6C**). As a result, the secondary circulation restricts the northward extension of monsoon rainfall over the continent and may help explain why the West African monsoon is restricted to the coastal belt and why rainfall decreases so rapidly over the Sahel and the Sahara. On the north side of the entrance region of the Easterly Jet, the South Asian monsoon circulation is enhanced by the secondary circulation. This enhancement occurs in the Bay of Bengal region and perhaps explains the existence of the most intense rainfall in the Asian monsoon.

It is paradoxical that the limitations on the poleward extent of the African rainfall are the result of the South Asian monsoon rainfall falling so far to the north. Reasons for the anomalous latitudinal location of the South Asian monsoon rainfall will be discussed in the next section. These are the same reasons that determine the longitudinal extent of the jet stream and hence the limit of its effect on African rainfall.

Impacts on the ocean Rotational effects are very important in the ocean and play a key role in determining both the overall character of the annual cycle of the monsoon and its interannual variability. The heat balance of the ocean mixed layer may be written as eqn [13].

$$C_{\mathrm{w}}\frac{\mathrm{d}hT}{\mathrm{d}t} = F_{z=0} + F_{\mathrm{adv}} \qquad [13]$$

Here h and T are the thickness and temperature of the mixed layer, and $F_{z=0}$ is defined in eqn [2]. F_{adv} refers to the advection of heat in and out of the column. Equation [9] allows for the horizontal advection of heat by ocean currents. The depth of the mixed layer is determined, to a large degree, by wind-forced turbulent mixing as discussed earlier.

Observations in the North Indian Ocean suggest that the advective term may be important. In the period between March and May, the average net surface flux into the North Indian Ocean is $>100\,\mathrm{W\,m^{-2}}$. However, the SST increases by only about 2°C during this period compared to about 9°C that would be expected if $F_{\mathrm{adv}} = 0$. It is important to understand the reason for this SST modulation, as the monsoon would be very different if the temperature of the North Indian Ocean were much warmer. Modulation might occur by either heat storage increases (by deepening of the mixed layer) or by the horizontal advection of heat away from the summer hemisphere. As it turns out, both processes are important.

The instantaneous northward flux of heat across the Indian Ocean between two meridians x_1 and x_2 and the change in heat storage in the ocean column are given by eqns [14] and [15].

$$F_{\mathrm{adv}} = \rho_{\mathrm{w}} C_{\mathrm{w}} \int_{x_2}^{x_1}\int_{z}^{0} (HvT)\,\mathrm{d}x\,\mathrm{d}y\,\mathrm{d}z \qquad [14]$$

$$S_{\mathrm{Q}} = \rho_{\mathrm{w}} C_{\mathrm{w}} \int_{x_1}^{x_2}\int_{y_1}^{y_2}\int_{z}^{0} (HT)\,\mathrm{d}x\,\mathrm{d}y\,\mathrm{d}z \qquad [15]$$

Here ρ_{w} and C_{w} represent the density and specific heat of sea water, and the coordinates x, y, and z represent longitude, latitude, and depth, respectively. **Figure 8** shows the total cross-equatorial flux of heat flux by oceanic processes from eqn [14], the change in oceanic storage of heat of the entire Indian Ocean north of the Equator from eqn [15], and the net surface heat flux into the North Indian Ocean from [2]. In addition, the atmospheric flux of latent heat across the Equator is also plotted. Compared to the magnitude of the net surface heat flux into the ocean (about 0.5 PW), the changes in storage and the heat flux across the Equator are much larger, with amplitude swings of ± 2 PW through the annual cycle. Clearly, the surface heat flux is not driving either meridional heat transport or changes in heat storage. A large part of the changes in storage occur through cross-equatorial transports of heat. **Figure 8** shows that transport has an extremely large annual cycle, with a maximum northward

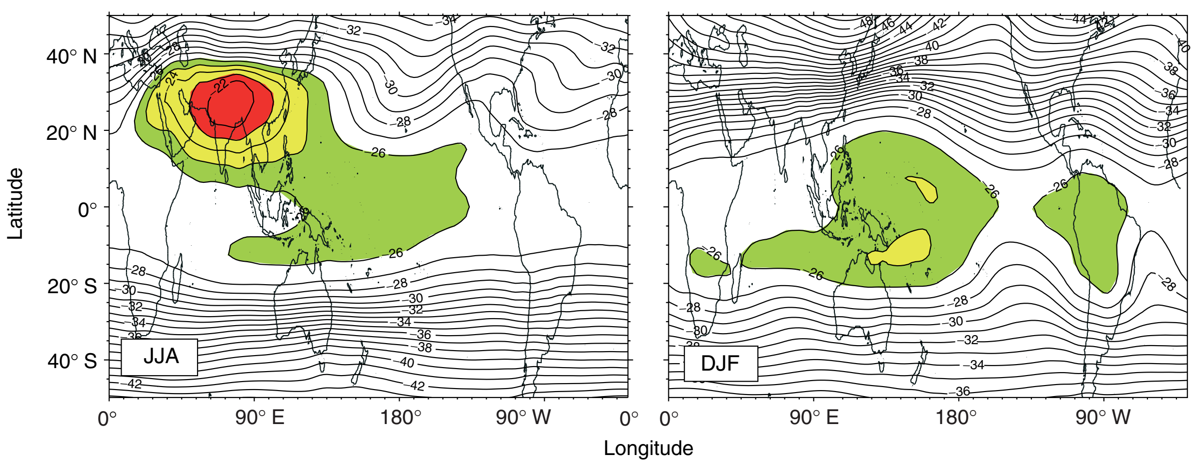

Figure 7 Distribution of the mean upper tropospheric temperature averaged between 200 and 500 hPa for the boreal summer (JJA) and winter (DJF). Note the two locations where the mean temperature is warmer than the equatorial temperature: over the Tibetan Plateau during the boreal summer and to the north-east of Australia in the austral summer.

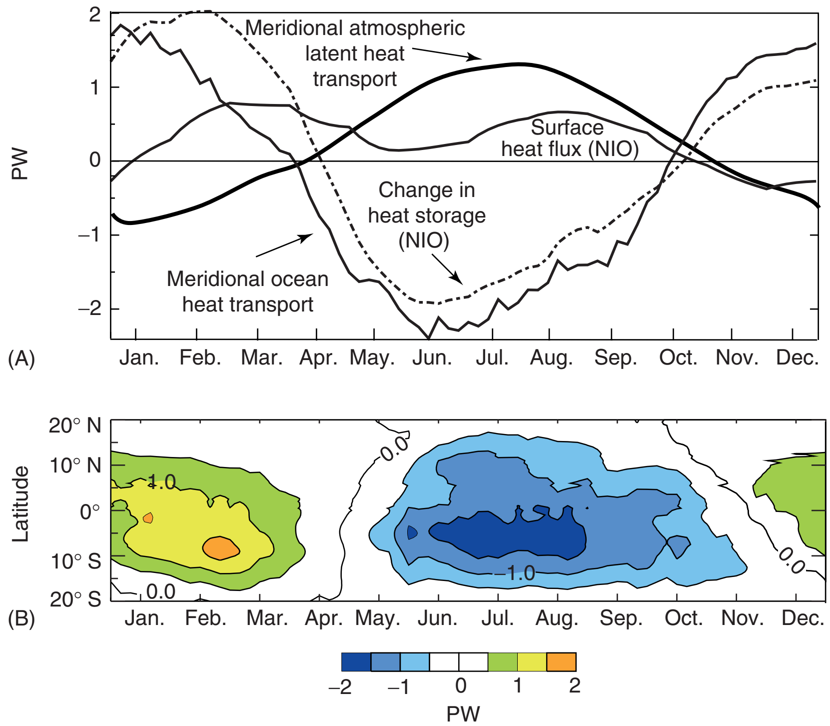

Figure 8 (A) Annual cycle of the meridional ocean heat transport across the equator, changes in the surface heat flux integrated over the North Indian Ocean, the change in total heat storage in the North Indian Ocean (NIO), and the atmospheric meridional transport of latent heat across the Equator. All quantities are averaged across the Indian Ocean. (B) Time–latitude plot of the cross-equatorial ocean flux of heat. All units are in PW (10^{15} W).

transport occurring in winter and early spring and a reverse transport occurring in the summer. Finally, it should be noted that the atmospheric latent heat flux is almost exactly out of phase with, and roughly equal in magnitude to, the lateral oceanic heat transport across the Equator. Clearly, the ocean and the atmosphere are working in tandem in some manner to determine the overall heat budget of the monsoon.

If the atmospheric and oceanic components of the monsoon are coupled, what are the processes responsible for the coupling? The fundamental clue to understanding the coupling comes from noting that the majority of mass and heat transport in the ocean is wind-driven. Transport is a combination of geostrophic and Ekman transport such that $F_{adv} = F_{geos} + F_{Ek}$. The meridional Ekman transport is especially interesting and may be written as in eqn [16], where τ_x is the zonal wind stress and T' is the degree to which the temperature of the upper ocean is warmer than the lower ocean.

$$F_{Ek} = C_w \int \frac{-\tau_x}{f} T' \, dx \qquad [16]$$

In the Northern Hemisphere ($f > 0$), the transport will be southward if the winds are westerly ($\tau_x > 0$) and northward for easterly winds ($\tau_x < 0$). As $f < 0$ in the Southern Hemisphere, the reverse associations are true. More generally, vertically averaged oceanic Ekman transports are to the right of the surface wind in the Northern Hemisphere and the left of the wind in the Southern. Thus, in a monsoon system, the heat transport will be southward in both hemispheres in the boreal summer and northward in the winter hemisphere, as shown in **Figure 5A**. This means that the overall oceanic heat transport has the opposite sign to the atmospheric heat transport that is in the direction of the lower tropospheric divergent wind. The overall effect of the wind-driven oceanic heat transports is to cool the SST in the summer hemisphere and warm the SST in the winter hemisphere, thus reducing the cross-equatorial SST gradient.

Finally, it should be noted that Ekman transports are not really applicable very close to the Equator as the ‘spin-up’ time for an Ekman transport to begin after winds goes as the inverse of the Coriolis parameter. At the Equator the spin-up time would be infinite. However, the approximation is valid a few degrees from the Equator and ocean models show a smooth transition of heat transport across the Equator that matches the Ekman transport to the north and south.

Regional Effects

In reality, the geography of the monsoon regions is far more complicated than the idealized model discussed so far. Each monsoon region possesses different land–sea distributions and topography. It might therefore be expected that the monsoon flows will have strong regional character. Nowhere is topography more important than around the Indian Ocean basin where the East African Highlands and the Tibetan Plateau complex extend along the eastern boundary of the basin and over South Asia, respectively.

The meridionally oriented East African Highlands act as a mechanical barrier to the South Indian Ocean south-east trades and concentrate the low-level flow intercepting the barrier crossing the Equator as an intense low-level flow called the Somalia Jet. Wind speeds in the core of the jet exceed 20–25 m s^{-1}. The impact of the East African Highlands is shown schematically in **Figure 9**. The jet stream drives a strong cross-equatorial Somalia Current and farther up the coast produces intense coastal upwelling that is associated with the 3–4°C drop in SST in the eastern Arabian Sea compared to 1–2°C over the entire North Indian Ocean.

During the spring and summer, the role of the Tibetan Plateau complex thought to be thermal, acting as an elevated heat source rather than as a barrier. To understand the concept of an elevated heat source, one needs to consider the heat balance on the surface of the plateau and the heat balance of the free atmosphere at the same elevation but away form the plateau. During spring and summer, the interception of radiation on the plateau causes a transfer of sensible heat and latent heat to the atmosphere of about 100 W m^{-2} compared to a much smaller value in a layer of atmosphere over the plains to the south. The heating of the plateau produces a localized heating anomaly. **Figure 10** shows the heating that occurs throughout the troposphere at 10° N, 20° N, 30° N, and 40° N at 80° E after May 1. During the boreal winter, there is no evidence of heating over the plateau. Instead, the topography acts as a mechanical barrier to the strong boreal winter westerlies.

To a large degree, the Tibetan Plateau is major factor determining the character of the South Asian summer monsoon. The location of the elevated heat source to the north of 25° N is sufficiently strong to produce rainfall well into the subtropics, which, on a global perspective, are usually regions of low rainfall and deserts. The location of the upper tropospheric anticyclone is determined by Tibetan Plateau, which, in turn, sets the position of the upper tropospheric Easterly Jet and the region influenced by the secondary circulations in the exit region (**Figure 6**).

Regulation of the Monsoon System

One of the interesting features of the South Asian monsoon is its relative constancy from year to year. For example, the mean precipitation over the Indian subcontinent is 852 mm with a standard deviation of 84 mm. These statistics represent a high-rainfall

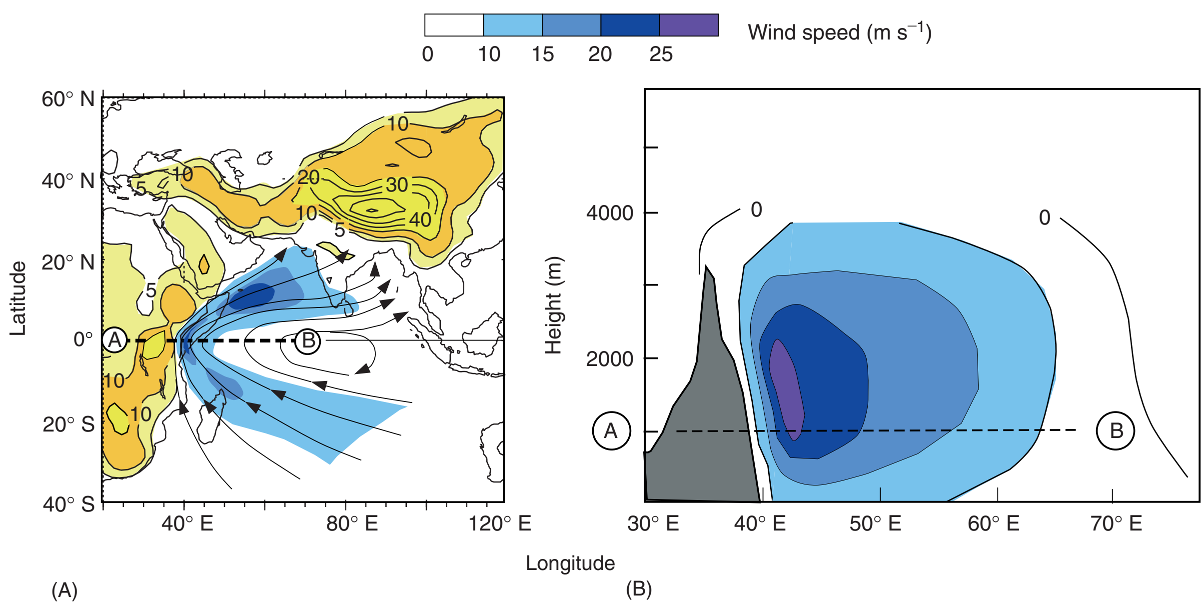

Figure 9 (A) Horizontal section of the Somalia Current at 1 km showing its lateral extent. Region north of the equator is a region of strong ocean upwelling forced by the wind. (B) Latitude–height section of the Somalia Jet along the equator. Units are m s^{-1}.

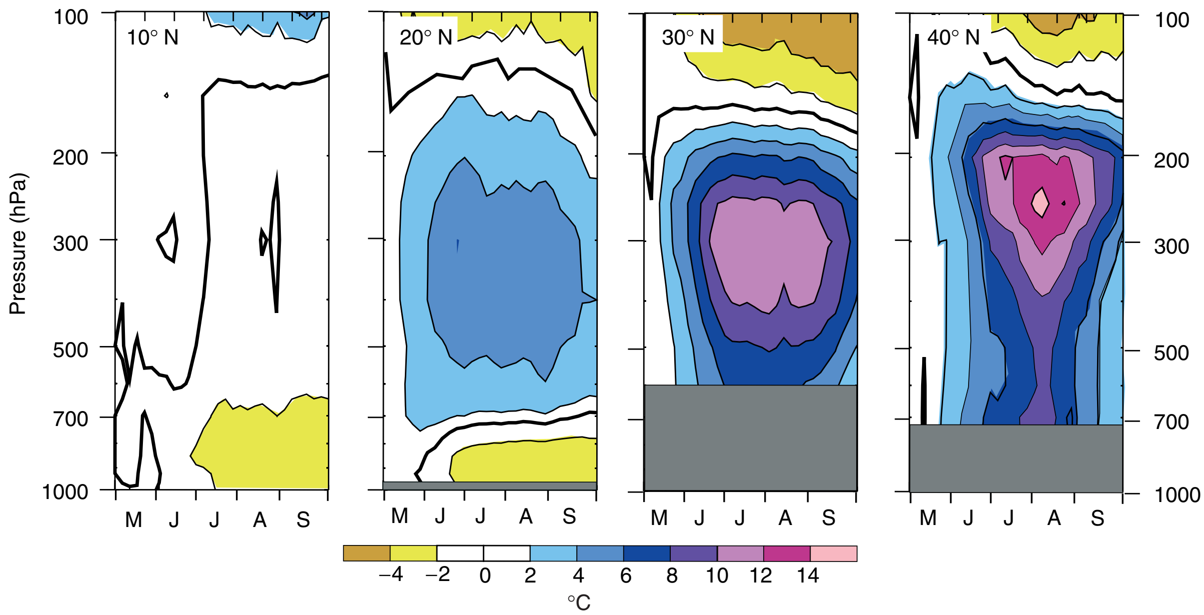

Figure 10 Time section of the change in atmospheric temperature over the Tibetan Plateau based on temperatures at 1 May.

region with very small variability about the mean. Furthermore, prolonged droughts or floods (i.e., mean summer rainfall outside the range of 852 ± 84 mm) are very rare. Between 1870 and 1998 there were 17 flood years and 22 drought years. Equally interesting is the lack of prolonged anomalies. Excessive rainfall occurred in successive years only twice (in 1892–94, and 1916–17), while deficient rainfall has occurred in successive years three times (1904–05, 1965–66, and 1985–87). The El Niño Southern Oscillation (ENSO) phenomenon explains about 40% of the variance, with El Niño often being associated with drought and La Niña with flood. Also, it has been noted that there are periods where there is strong biennial trend in the rainfall record, with strong and weak monsoon years occurring successively. Overall, it would seem that the monsoon is regulated in some fashion to fall within rather narrow limits of variability. In the following paragraphs, it will be suggested that regulation of monsoon intensity occurs through the interaction of the ocean and the atmosphere in the manner introduced described under 'The Impact of Rotation'.

Regulation of the Monsoon Annual Cycle

Figure 11 shows a schematic of the atmospheric circulation and corresponding ocean heat transport for the boreal summer and winter. The impact of the wind-driven ocean heat transport is to homogenize the SST gradient across the Equator, lowering it in the summer hemisphere and warming it in the winter hemisphere. The reduction of the SST gradient will have two major impacts. First, the cross-equatorial pressure gradient will be reduced, which will have the effect of reducing the intensity of the monsoon winds. Second, the reduced SST in the summer hemisphere reduces the saturated vapor pressure of an air parcel approaching the continental region. As the SST is very warm, small changes in the SST will invoke large changes in the saturated vapor pressure, as determined by the nonlinear Clausius–Clapeyron equation. The reduced monsoon winds will lower the convergence of moisture into the continental regions. In tandem with the reduced saturated vapor pressure this reduces the release of latent heat in the precipitating regions. The overall impact of the two processes is to reduce the amplitude of the monsoon annual cycle. Without ocean heat transports, the SST extremes would produce a monsoon that would be very different from that which is observed today and would quite likely be much wetter.

It is important to note that the annual cycle of oceanic heat transports, as described above, is a phenomenon that is restricted to regions that have strong cross-equatorial pressure gradients and a surface flow from the winter to the summer hemisphere. For example, the Pacific Ocean is characterized by a Trade Wind regime that converges into the western Pacific from both hemispheres. The Ekman transports of mass and heat are away from the tropics in both hemispheres: to the right of the north-easterly Trades and the left of the south-easterly Trades. The ocean heat transport produced by the Pacific Trade Winds is probably important in regulating the surface temper-

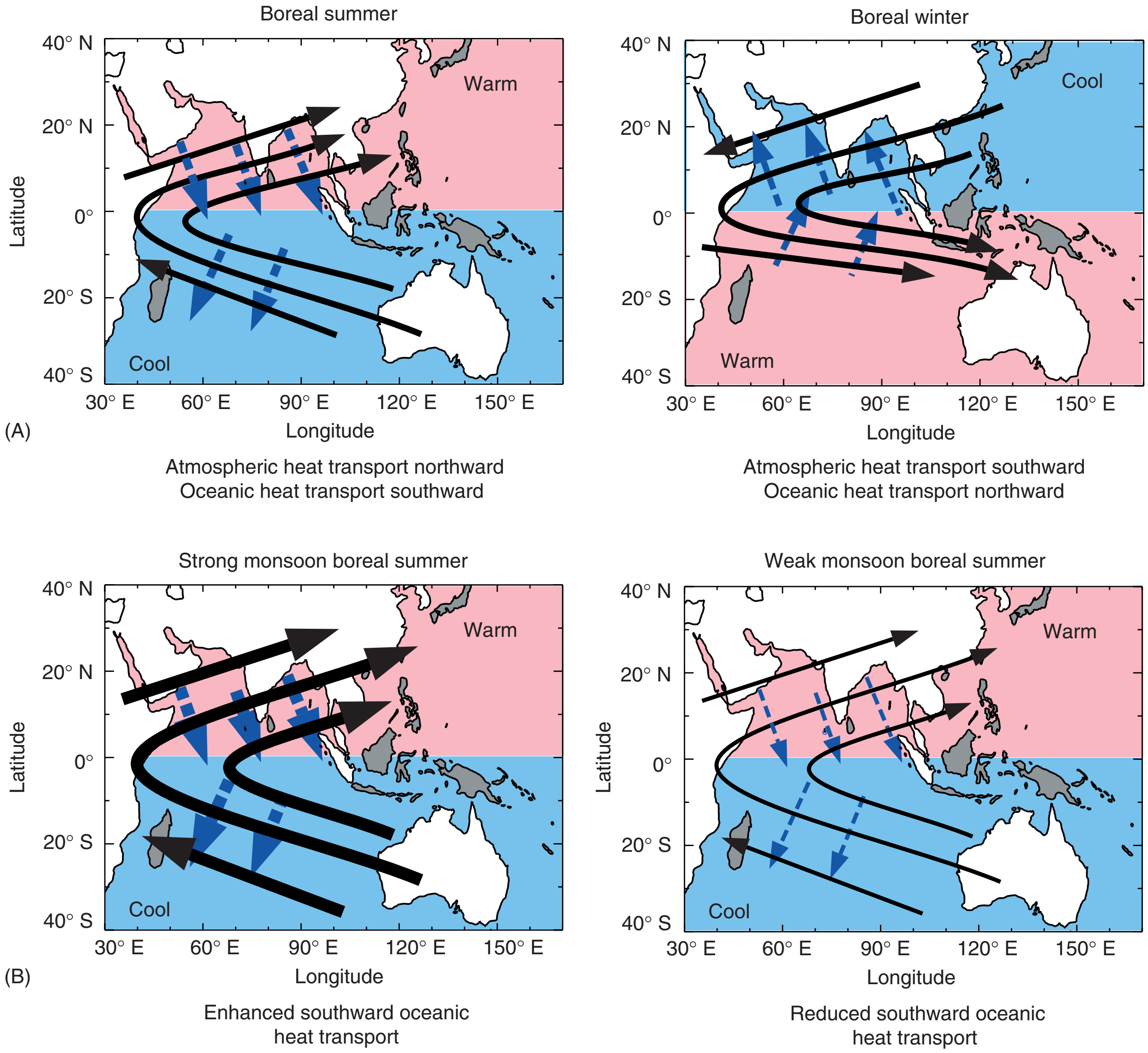

Figure 11 (A) A regulatory model of the monsoon for the annual cycle. Climatological lower tropospheric winds (solid black lines) are shown for summer (left panel) and winter (right panel). Blue dashed lines show the direction of Ekman mass transports which are to the right of the wind in the northern hemisphere and to the left in the southern hemisphere. Given the general sea-surface temperature gradient, heat is transported from the summer hemisphere to the winter hemisphere in both seasons. From **Figure 8** it can be seen that these heat transports can be substantial enough to change the temperature of the entire North Indian ocean by 2–3°C. (B) Summer representations of the surface wind in strong and weak monsoon years. In a strong year, the Ekman transport will be large while in a weak year it will be smaller. Thus during a strong year the northern Indian Ocean will cool at a greater rate than during a weak year. Given that a strong monsoon is usually preceded by warmer than average sea-surface temperatures in the Indian ocean, a strong monsoon year is often followed by a weak monsoon and vice versa.

ature of the Pacific warm pool. However, the strong seasonality and change in sign of the heat transport is restricted to the Indian Ocean monsoon regime.

Regulation of Monsoon Interannual Variability

Imagine that, for some reason, that the SST in the northern Indian Ocean during spring was higher than normal. Such alterations to the climatology could be imposed by external factors such as ENSO, anomalous winter or spring snowfall over Eurasia, or the result of stochastic processes. Warmer than average SSTs would lead to a stronger than average monsoon winds. In fact, there is observational evidence for the relationship. Indian Ocean SSTs during the previous winter and early spring correlate positively with the strength of the ensuing monsoon. Overall, the same processes that regulate the annual cycle of the monsoon ensure that the anomalous imposed conditions (e.g., high SST, weak winds) are eradicated.

For the case of higher than average springtime North Indian Ocean SSTs, the meridional pressure gradient would tend to be stronger and the near-surface saturated vapor pressure higher, causing a stronger than average monsoon to develop. A strong monsoon with greater than average winds would (1) cause deep mixing and increase the ocean storage of heat, and (2) drive stronger than average southward wind-driven oceanic heat transports. The impact of the strong monsoon would be to cool the North Indian Ocean so that the SSTs during the following winter would be lower than average. In turn, these lower SSTs would produce a weaker than average monsoon in the following spring and summer. Associated with the weaker than average monsoon winds would be a lowering of the southward heat transport.

Figure 12 shows an example of the monsoon interannual regulation process at work. In 1987 a strong El Niño occurred and was followed in 1988 by an equally strong La Niña. In 1987 the South Asian monsoon was weak, with Indian summer rainfalls 17% below normal. In 1988 the monsoon was strong, with rainfall over India 13% above normal. The figure shows a time section of the annually averaged northward oceanic heat transport for these two years calculated using a stand-alone ocean model driven by observed winds. **Figure 12** plots annually averaged values. The long-term average cross-equatorial heat transport is about −0.2 PW which matches the climatological net surface heating over the North Indian Ocean. However, during the weak monsoon year of 1987, the net surface flux was slightly higher and, associated with the weaker monsoon winds, the cross-equatorial ocean heat transport reversed from the climatological value of −0.2 PW to +0.2 PW. During the following year, the net surface flux into the North Indian Ocean was reduced while, responding to the much stronger monsoon winds, the oceanic heat transport was twice the long-term mean value of −0.2 PW at −0.45 PW. The two years 1987 and 1988 illustrate the manner in which the monsoon system responds to forcing anomalies and returns the system to equilibrium.

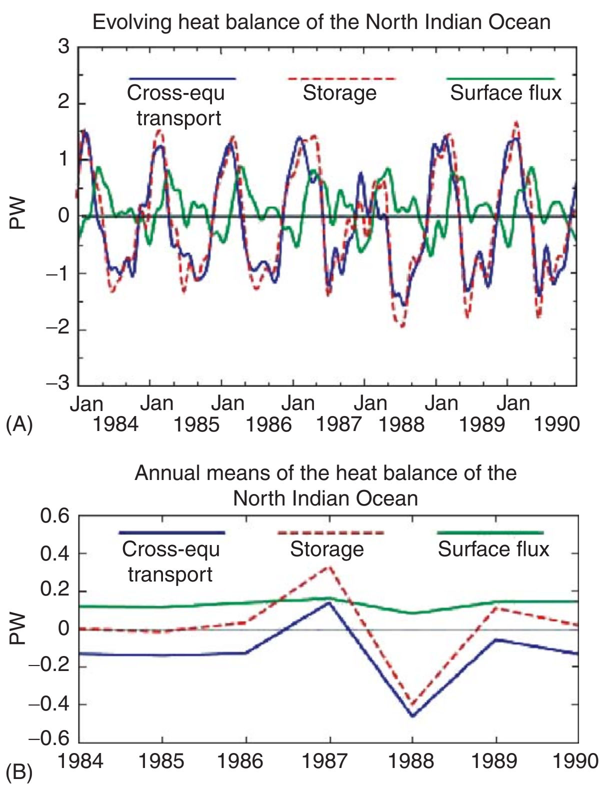

Figure 12 Time sections of the oceanic cross-equatorial heat flux, the net surface heat flux into the North Indian Ocean, and the change in ocean heat storage in the North Indian Ocean for the years 1983–89 as computed from an intermediate ocean model driven by observed winds and fluxes: (A) 5-day averages, (B) annual averages. 1987 and 1988 were weak and strong monsoon years associated, respectively, with El Nino and La Nina.

Summary

A rather different picture of the monsoon has been presented. Rather than describing the monsoon as a gigantic sea breeze driven by the differential heating between the ocean and land, a self-regulating system has been introduced in which dynamic aspects of the ocean play critical roles in modulating the strength of the monsoon. On this view, it is clear that if the variability of the monsoon is to be predicted, it will be necessary to consider the system as a fully coupled interactive system. As such, the numerical models that will be used to simulate and predict monsoon behavior will have to have interactive oceans.

See also

Climate Prediction (Empirical and Numerical). **Coupled Ocean–Atmosphere Models**. **Hurricanes**. **Monsoon:** ENSO–Monsoon Interactions; Overview; Prediction. **Ocean Circulation:** General Processes. **Tropical Meteorology:** Equatorial Waves; Inter Tropical Convergence Zones (ITCZ); Overview and Theory; Tropical Climates. **Weather Prediction:** Regional Prediction Models.

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ENSO–Monsoon Interactions

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Introduction

The word ‘monsoon’ is derived from the Arabic word *mausam*, which means season. In monsoon regions, the change of season is accompanied by a reversal of the prevailing wind direction and abrupt changes in rainfall patterns. The most pronounced monsoon climate is found in the Asian–Australian region. Other regions that exhibit monsoonal characteristics include the maritime continent, western Africa, central America, south-western North America, and South America. These regions are often adversely affected by severe droughts and floods caused by the strong year-to-year-variability in monsoon rainfall. During the summer of 1997, southern China experienced severe flood while northern China was gripped by one of the driest season on record. In the summer of 1998, a monsoon depression in Bangladesh devastated the country, causing major floods in the Ganges and the Brahmaputra river, displacing over 30 million people, and resulting in property and agricultural loss of over $US 3.4 billion. In the same summer, a flood of biblical proportion ravaged the Yangtze River basin and northeastern China, displacing over 220 million people, inflicting a huge economic loss of over $US 12 billion. Understanding and predicting monsoon variability is therefore vitally important for benefit of society.

Scientists have long suspected that there is a connection between monsoon rainfall variability and components of the global circulation system. G. T. Walker in the 1920s found that the Indian monsoon rainfall anomalies may be foreshadowed by seasonal surface pressure variations over several ‘strategic’ points remote from India. Of the many planetary-scale patterns found by Walker, the most important is the Southern Oscillation (SO), which was described as ‘a swaying of pressure on a big scale backwards and forwards between the Pacific Ocean and the Indian Ocean’. However, his effort to translate the monsoon–SO relationship into seasonal and interannual prediction of Indian monsoon rainfall was largely unsuccessful. Several decades later, Bjerknes first suggested that the SO as found by Walker is closely linked to El Niño (*see* **El Niño and the Southern Oscillation:** Observation; Theory), Bjerknes’ work and subsequent research have led to the development of various empirical monsoon forecast schemes based on the El Niño–Southern Oscillation (ENSO). However, results are mixed, because even though ENSO may be important in influencing monsoon rainfall variability, there are a large number of factors that may confound or limit monsoon predictability. With the advances in our understanding and prediction of ENSO, there is now renewed interest in reexamining the monsoon–ENSO relationship and its possible utilization for monsoon prediction. The discussion here will focus mainly on the most pronounced monsoon system, the Asian–Australian monsoon (AAM), unless specified otherwise.

The Basic Relationship

Figure 1 shows the time-series of all-Indian summer rainfall (June–September) from 1871 to 1998. Large interannual variability can be seen. There also appears to be interdecadal modulation of the amplitude of the anomalies. Relatively small anomalies are found in the 1880s–90s, the 1920s–40s, and in the 1990s. Large anomalies are found in the 1900s–20s, 1960s–80s. As indicated by the color shading, the Indian monsoon is generally below normal preceding the peak of a warm sea surface temperature (SST) event (El Niño) and above normal preceding a cold event (La Niña). However, there are warm or cold events that produce very weak signals or even anomalies of the opposite sign. Most important, a large number of major rainfall anomalies are not related to either El Niño or La Niña. Similarly relationship can be found for the austral monsoon (not shown). Hence, not all monsoon floods and droughts can be explained by ENSO.