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Preview

This chapter serves as an introduction and preview for the entire book. Topics will be broadly introduced, to be better and more completely explained in the sequel.

1.1 The maritime tropics

It may surprise people living in the midlatitudes that the tropics have such an overwhelming role in the climate of the Earth. Yet it has been shown time and time again that the maritime tropics are the only regions on Earth where changes in the surface-boundary condition, especially sea-surface temperature (SST), have a demonstrable and robust causal correlation with weather effects in midlatitudes. This happens through the ability of warm sea-surface temperature anomalies (deviations of sea-surface temperature from its normal value for that time of year) to organize deep cumulonimbus convection and plentiful rainfall which can then emit large-scale planetary waves which subsequently travel to higher latitudes. The changes of SST, the formation of regions of persistent precipitation, and the resulting forcing of the midlatitude motions by these regions of persistent precipitation, form a set of themes that appear and recur throughout this book.

It is a good rule of thumb (these rules of thumb will be examined in much greater detail in the body of the book), in the tropical Pacific in particular, that regions of persistent precipitation lie over the warmest water, and a good rule of thumb that in the presence of persistent precipitation, the net synoptic motion is upward and the sea-level pressure low. With these rules of thumb, we are in a position to describe the normal conditions over the tropical Pacific, the main region of interest in this book.

1.2 The normal tropical Pacific

The tropical Pacific extends from the coast of South America in the eastern Pacific to the various islands and land masses of Australia and Indonesia that form the

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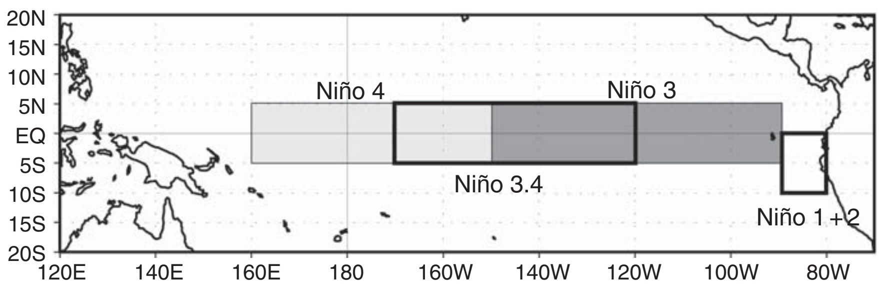

Figure 1.1. The tropical Pacific, including the definition of the four Niño regions. (Courtesy of the NOAA Climate Prediction Center.)

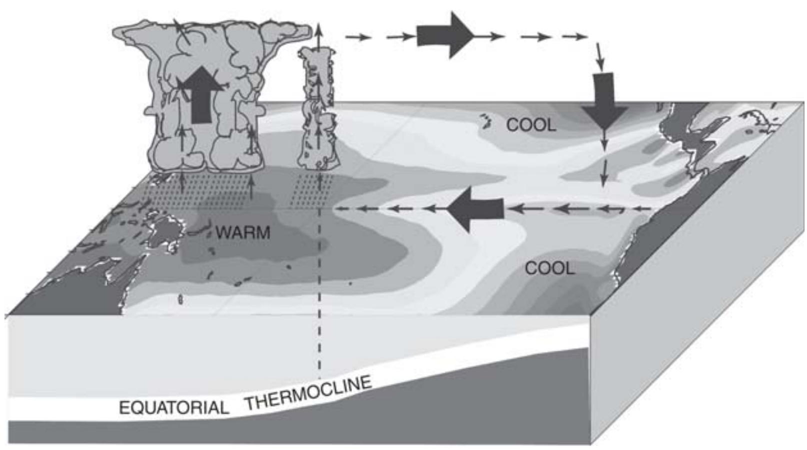

Figure 1.2. Schematic of the normal state of the coupled atmosphere–ocean system in the tropical Pacific during boreal winter. The shading on the surface of the ocean represents sea-surface temperature, warm in the west and cooler to the east and south-west. (Courtesy of the NOAA Climate Prediction Center.)

so-called maritime continent; a somewhat paradoxical idea expressing a collection of land masses without there actually being a land continent present (Figure 1.1). In particular, the equator runs from Ecuador in the east (at 80°W) to Indonesia in the west – the first land the equator crosses in the western Pacific is Halmahera at 129°E and then the more substantial Sulewesi at 120°E. Taking Halmahera as the western boundary of the Pacific gives a total length on the equator of 151 degrees of longitude or 16 778 km, more than one-third of the total distance around the globe.

The climatic state in and over the tropical Pacific is given by a convenient cartoon (Figure 1.2). The surface of the western Pacific is warm and the atmosphere above it is rainy, with the rain coming from deep cumulonimbus clouds. The air rises in the region of the warm water and the rising air is characterized by low pressure at the surface. The winds across the surface of the tropical Pacific blow westward into the region of low pressure, consistent with the westward trade winds. The rising motion in the warm region reaches the tropopause and returns eastward aloft and completes the circuit by descending in the eastern Pacific, leading to higher pressure at the surface.

This tropical Pacific-wide circuit of air proceeding westward at the surface, rising over the (warm) region of persistent precipitation, returning eastward aloft, and descending over the cool eastern Pacific, is called the Walker circulation. Associated with the Walker circulation is the low surface pressure in the western Pacific and the high surface pressure in the east. A measure of the strength of the Walker circulation is the difference in the surface pressure between the east and west – this difference is conventionally called the Southern Oscillation Index (SOI) – we will see below the oscillation to which it refers. When the Walker circulation is strong, the pressure in the west is low and the pressure in the east is high – the SOI is then less negative. When the Walker circulation is weak, the SOI is more negative.

The oceanic part of Figure 1.2 is driven by the westward surface winds; the surface expression of the Walker circulation in the atmosphere. The feature in the ocean called the thermocline is a near-ubiquitous property of the oceans. In the tropics it is a region of such sharp temperature change in the vertical that one may vertically divide the ocean into only two regions, one with warm temperatures and one where the temperatures are cold. The thermocline demarcates the warm-water sphere near the surface from the cold-water sphere below. We will show later that the deeper thermocline in the western Pacific is caused by the westward winds at the surface of the ocean. Thus the stronger the westward surface winds (due to a strong Walker circulation), the deeper the thermocline in the west and the shallower the thermocline in the east. The tilt of the thermocline in the ocean is a measure of the strength of the westward surface winds and, therefore, another measure of the strength of the Walker circulation. The chain of reasoning is continued by noting that the East–West temperature difference, which may be considered to drive the atmospheric motion, is indeed unexpected, since the sun shines equally on the western and eastern Pacific.

The mechanism responsible for the mean East–West sea-surface temperature difference involves both the atmosphere and the ocean. The mean westward surface winds drive ocean motion poleward in both the northern and southern hemispheres within 50 or so meters of the surface very near the equator. Water moving poleward must be replaced by water upwelling on the equator from below. In the eastern Pacific, the thermocline (recall that the thermocline is the demarcation between warm and cold water) is shallower than 50 m, so that cold water is upwelled on the equator, causing the SST to be cold. In the western Pacific, the thermocline lies below 50 m and, while upwelling still occurs, it simply brings up warm water from above the thermocline, allowing the western Pacific SST to remain warm. Heat put into the ocean from the atmosphere counteracts the upwelling influence on SST but it does not win the contest: the eastern Pacific remains cooler than the west.

The cold SST in the eastern Pacific is spread poleward several degrees of latitude by the ocean motions until it encounters another warm region in the northern (but not southern) hemisphere caused by an eastward ocean current. There is again rising motion in the atmosphere above this warm water and a line of deep convection extends pretty much across the entire Pacific at an average latitude of about 6°N. This region of deep convection is called the intertropical convergence zone (ITCZ) and forms the rising tropical branch of a North–South circulation called the Hadley circulation.

Though not mentioned thus far, there is a pronounced seasonal cycle in the tropical Pacific. Unlike midlatitudes, the seasonal extremes are in March–April when the eastern equatorial Pacific is warmest and the ITCZ is closest to the equator, and September–October when the eastern SST is coldest and the ITCZ is furthest north. Since the SSTs in the western Pacific vary only by about 1 °C, the seasonal variations in the East–West gradient co-vary with the eastern Pacific SSTs: weakest in boreal spring, strongest in fall. This annual cycle has a strong influence on the evolution of ENSO phases, which exhibit a marked tendency to be phase-locked to the annual cycle, growing through the (northern) summer and fall to reach a winter peak.

1.3 The phases of ENSO

Superimposed on the normal state of the tropical Pacific is an irregular cycle of warming and cooling of the eastern Pacific with attendant atmospheric and oceanic effects, the panoply of which will be referred to as ENSO. Figure 1.3 shows conditions in and over the tropical Pacific during warm phases of ENSO.

The eastern Pacific warms, and can warm to such an extent that the temperature across the entire tropical Pacific becomes almost uniform. That the temperature reached is that of the western Pacific, rather than that of the eastern Pacific, indicates that the warm phase of ENSO is due to a failure of the eastern Pacific to stay cold. Consistent with this point of view is the relaxation of the westward surface winds (Figure 1.3 shows weaker than normal easterly winds which implies westerly anomalous winds), which produces less upwelling and therefore less cooling. Consistent with weaker westward winds, the thermocline is not as tilted, and any upwelling in the eastern Pacific would bring warmer water to the surface. When the cooling in the eastern Pacific is totally gone and the westward surface winds relaxed to almost zero, the warm phase of ENSO is as strong as it can be, and the temperature over the entire tropical Pacific is uniform and assumes the

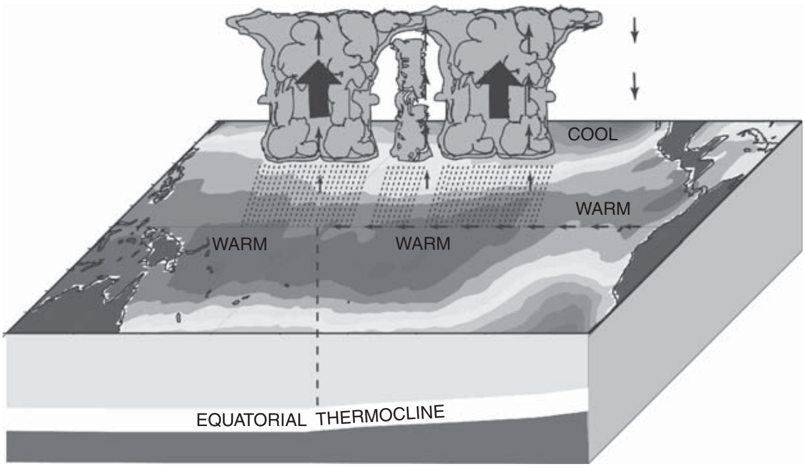

Figure 1.3. Schematic of the coupled atmosphere–ocean in the tropical Pacific during the peak of a warm phase of ENSO during boreal winter. (Courtesy of the NOAA Climate Prediction Center.)

approximate temperature of the western Pacific. This happened in the strong warm phases of ENSO during 1982–3 and 1997–8.

As the eastern Pacific becomes less cold, the region of persistent precipitation that lies over the warmest water expands eastward into the central Pacific. The normally high sea-level pressure (SLP) of the eastern Pacific becomes lower and the sea-level pressure difference between the western and eastern Pacific decreases. Consistent with this decrease is the weakening of the Walker circulation and the relaxation of the normally westward surface winds. As the central and eastern tropical Pacific becomes warm, the ITCZ moves onto the equator and the line of deep convection assumes its southernmost position and the Hadley circulation becomes stronger.

The effect of the warming of the eastern Pacific, and the consequent eastward movement of the region of persistent precipitation, is felt throughout the world (Figure 1.4). In the tropics, the normally rainy western Pacific becomes drier as the region of persistent precipitation moves eastward into the central Pacific. Droughts in Indonesia and in eastern Australia become far more common during the warm phases of ENSO. Rainfall in the normally arid coastal plains of Peru becomes far more likely and warm water spreads north and south along the western coasts of the North and South American continents. The temperature and rainfall in other selected areas of the world (e.g. Zimbabwe, Madagascar) are similarly affected, even though the reasons are either difficult to explain or unknown.

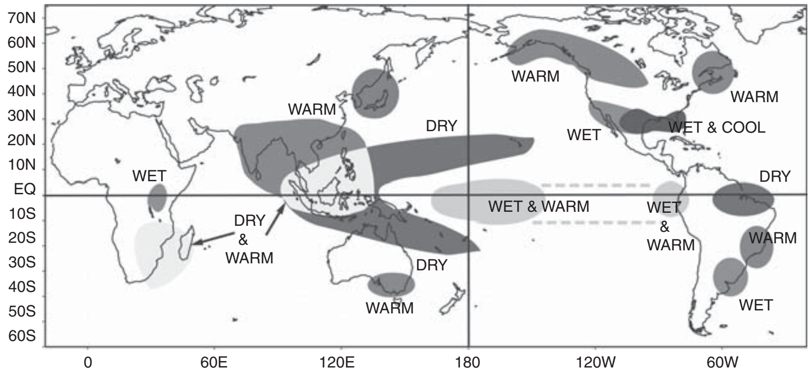

Figure 1.4a. Composite effects of the warm phase of ENSO on global climate during boreal winter. (Courtesy of the NOAA Climate Prediction Center.)

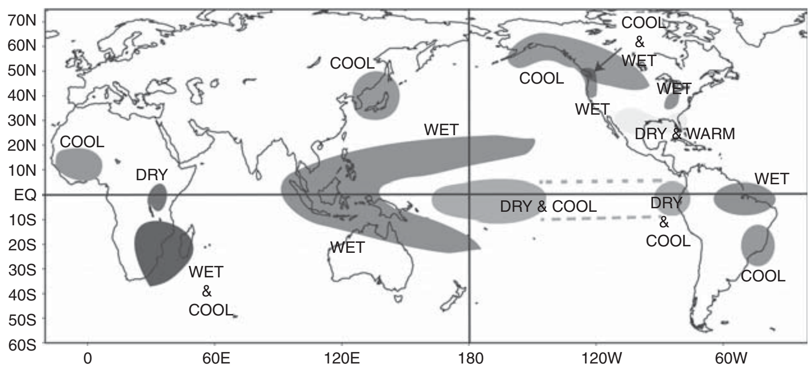

Figure 1.4b. Composite effects of the cold phase of ENSO on global climate during boreal winter. (Courtesy of the NOAA Climate Prediction Center.)

During cold phases of ENSO, the normal cooling of the eastern Pacific becomes even stronger, the surface pressure difference between the eastern Pacific and western Pacific becomes stronger, and the Walker circulation, in general, becomes stronger. Consistent with this, the surface westward winds become stronger, the tilt of the thermocline becomes greater, the stronger westward winds in the eastern Pacific produce even more upwelling and, because the thermocline is closer to the surface, the water upwelled is colder. The regions of warmest water in the western Pacific contract westward under the encroachment of cold water in the east and, with the warm water, the region of persistent precipitation contracts westward onto the maritime continent. Excess rainfall in Indonesia and western Australia becomes far more common during cold phases of ENSO (Figure 1.5).

The SST anomalies (the deviations from the norm) look, in many ways, the obverse of each other (Figure 1.6).

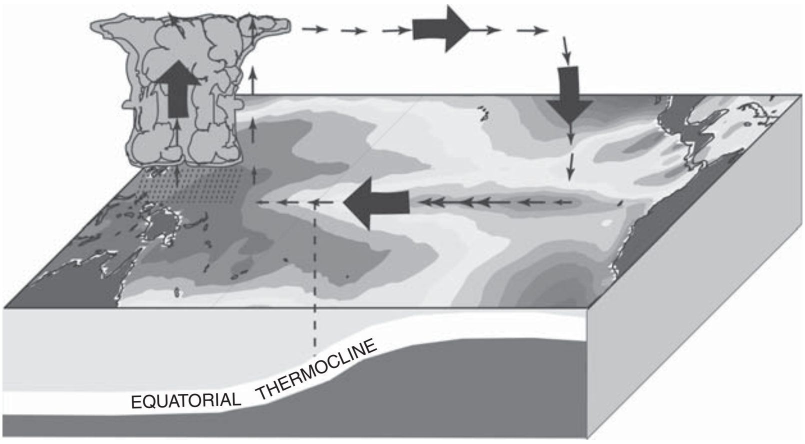

Figure 1.5. Schematic of the coupled atmosphere–ocean in the tropical Pacific during the peak of a cold phase of ENSO during boreal winter. (Courtesy of the NOAA Climate Prediction Center.)

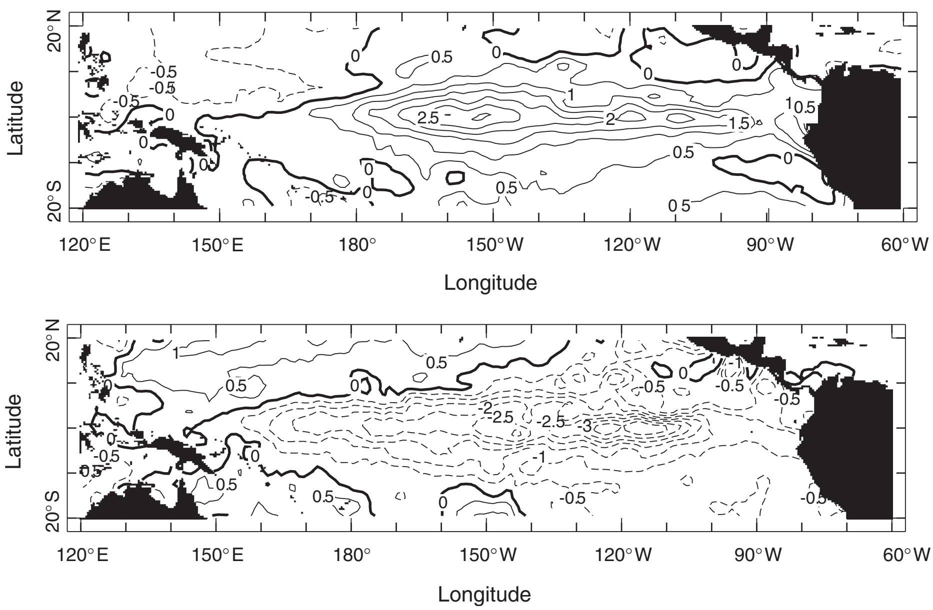

Figure 1.6. Upper panel: SST anomalies for the warm phase of ENSO during December 1991. Lower panel: SST anomalies for the cold phase of ENSO during December 1988. (Downloaded and plotted from www.iridl.ldeo.columbia.edu/ using the Reynolds *et al.*, 2002, updated SST data set.)

It must be kept in mind, however, that in many ways, cold and warm phases are fundamentally different because the quantities that affect the remote atmosphere are not the SST anomalies, but rather the mean location of the regions of persistent precipitation. In the warm phase of ENSO, persistent precipitation extends into the central Pacific, while during the cold phases of ENSO it retreats to the far western Pacific. The SST anomalies can be the inverse of each other but the mean location of the heat source, which drives the response in the low and midlatitudes, is very different. Because the rest of the world is forced by these regions of persistent precipitation and because these regions are in different locations for warm and cold phases of ENSO, there is no expectation that the global effects will be the opposite of each other. Figure 1.4b shows that during cold phases of ENSO, there are some similarities and significant differences in the global response.

1.4 Evolution of phases of ENSO

While we will go into greater detail in later chapters, we simply note here that the phases of ENSO evolve differently each time they appear. The general recurrence time for warm and cold phases is around 4 years, with large variations around this mean. The literature often speaks of an "ENSO band" from 2 to 7 years.

One way of describing ENSO evolution with time is to examine the SST anomalies in various regions of the Pacific defined by Figure 1.1. Figure 1.7 shows the SST anomalies since 1980 in the various regions of the Pacific, and a measure of the strength of the Walker circulation, the SOI. We can infer a number of important properties of the warm and cold phases of ENSO by examining this Figure. First we see that the major phases of ENSO tend to have expression all the

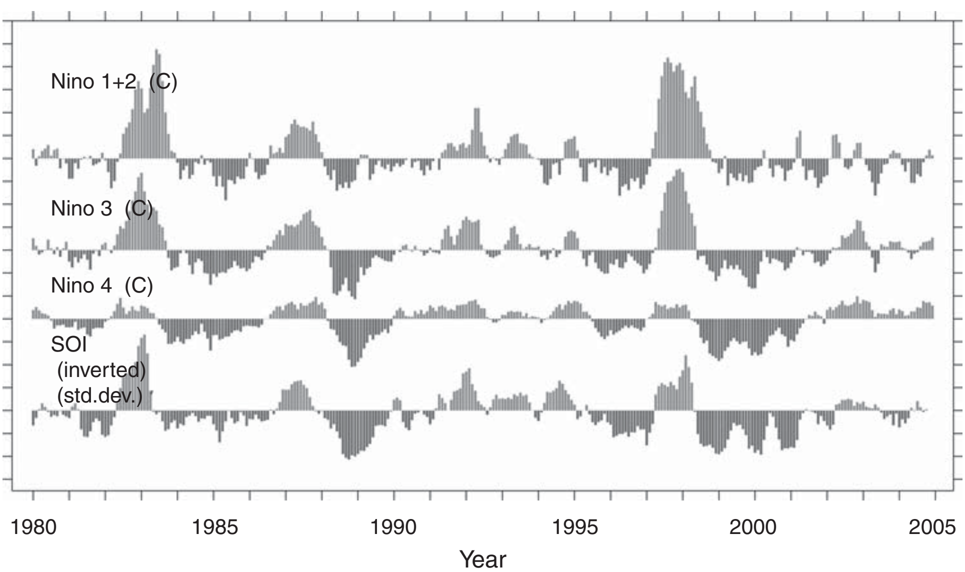

Figure 1.7. Niño region anomalies and Southern Oscillation index with respect to the respective means of 1985–94. (Courtesy of Todd Mitchell. Extended version of Plate 2 of Wallace *et al*., 1998.)

way across the Pacific, from the coast of South America (Nino 1 + 2) to the western Pacific (Nino 4). Second, that the major warm and cold phases tend to set on across the entire Pacific at about the same time. Third, that the larger events seem to start around summer, peak near the end of the year, and end before the next summer, so that the length of warm and cold phases is about a year. Fourth, that there are stretches of time in which not much is happening in the tropical Pacific (the entire 1930s were noted for having no major warm or cold phases – see Figure 1.17) and that these times are punctuated by the appearance of large phases of ENSO. It is worth mentioning that the warm phases in 1982–3 and 1997–8 were the largest of the century.

1.5 Physical ENSO processes

According to what we have seen so far, we need to understand how the SST anomalies characteristic of ENSO are produced, and how the connections of SST with sea-level pressure, precipitation, surface winds, the depth of the thermocline and remote precipitation and temperature are accomplished. Once we have a firm idea of the operation of each of these processes, we will need to know how they fit together to produce ENSO.

1.5.1 The processes that change SST

The temperature of ocean water can change either by directly adding heat (for example, from the sun) or by mixing with water of a different temperature. Because the ocean has no significant internal heat sources, heat can only be added directly at the surface. Heat added at the surface is basically a balance between radiation and evaporation: any net radiation reaching the surface that does not evaporate water is available to cross the ocean surface and heat the ocean water. In general, when water cools, evaporation decreases, and when water warms, evaporation increases. To the extent that the solar radiation reaching the surface is independent of the temperature of the underlying ocean (not entirely true, since the overlying cloudiness can change), warm surface water will have more evaporation and therefore less heat entering the ocean across the surface. Similarly, cooler water will have less evaporation and therefore more heat entering the ocean. Clearly, therefore, the heat entering the surface of the ocean tends to *oppose* the temperature changes.

If we consider some water near the ocean surface, the temperature can change if it is heated by heat entering the ocean through the surface, if it mixes with warm or cold water entering from the sides, or if it is cooled by water entering from below (Figure 1.8).

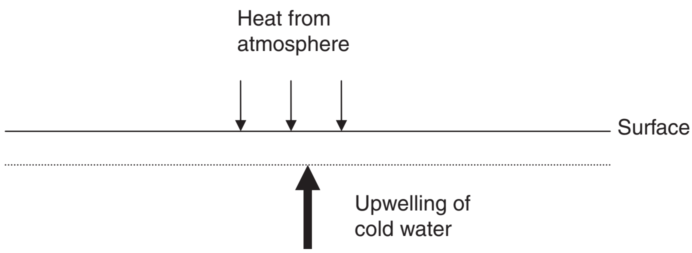

Figure 1.8. Schematic of vertical heat inputs into the tropical ocean mixed layer.

In the eastern Pacific, water at the surface is constantly cooled by water upwelled from below the thermocline and this cooling is opposed by heat entering through the surface. In the western Pacific, the temperature of the water is determined by the interactions with the atmosphere. It is approximately in equilibrium with the atmosphere, and is neither cooled from below nor heated from the atmosphere above. If the upwelling in the east were to decrease to a new, but smaller, steady value, not as much cold water would be brought up from below and the heat entering from the surface would warm the water until it reached a new not as cool temperature – this would be a warm SST anomaly. The evaporation would increase, the heat entering though the surface would decrease, and the water near the surface would reach a new warmer equilibrium; cooled not as much from below and heated not as much from above. The water could also warm if it mixed with warmer water from the west or perhaps from the north. In either case, warm SST anomalies would be associated with more evaporation and therefore less heating of the ocean surface from above.

1.5.2 The process by which warm SST anchors regions of persistent precipitation

Warm regions tend to have lighter air above these regions, as the air is warmed by the surface. Warm air is light and since surface pressure is the total weight of the air above it, the surface pressure tends to be lower above warm tropical regions. Air from surrounding higher pressure regions rushes in and is warmed, moistened and raised. Rising air condenses and the heat of condensation raises the air further. If the underlying SST is warm enough, about 28 °C or 29 °C, the clouds can reach to the top of the atmosphere (the tropopause) and regions of deep cumulonimbus convection result. The average amount of rain that falls is equal to the local evaporation plus the amount of moisture that converges into the region. Moisture is confined mostly to the lowest 1 or 2 kilometers of the atmosphere, so it is the low-level moist air that converges into the region that provides the additional moisture for the rainfall. The overall picture may be sketched as in Figure 1.9.