Chapter 8

The equatorial atmosphere and ocean

8.1 Tropical meteorological maps

In the extratropics, we have got used to summarizing synoptic meteorological situations through the use of maps of pressure (at fixed height) or of height (at fixed pressure). Such maps are useful because the geostrophic wind relationship allows us to determine wind speed and direction from this information alone. However, this kind of map is much less useful in the tropics, for two reasons:

1. Pressure (or height) variations become weak in the tropics. From the geostrophic relationship

$$\mathbf{u} = \frac{1}{f\rho}\mathbf{k} \times \nabla p ,$$

   the typical magnitude of wind speed is $U \sim \delta p / (f\rho L)$, where δp is the magnitude of pressure variations over distance L. Near the equator, $f = 2\Omega \sin\varphi \to 0$; since (as a matter of observation, as well as physical common sense) U does not become infinite, we must have $\delta p \to 0$. So *pressure variations are weak in the tropics*: so there is little to plot on a pressure map.

2. The validity of the geostrophic relationship requires that the Rossby number $R = U/fL$ be small. As $f \to 0$ in the tropics, this becomes

less valid. So, even if we did plot maps of the (weak) tropical pressure structure, we could not use these maps to deduce winds as we do in middle latitudes.

3. A corollary of the previous point is that tropical winds are not necessarily even approximately nondivergent.

The upshot of all this is that pressure (or height maps) are not as much use in the tropics. To see winds, it is more revealing to plot the winds directly, either (nowadays) as vector (arrow) wind plots, or as maps of:

1. **Streamlines**—lines that follow the direction of the flow. Note that these are not *contours*, of streamfunction or of anything else. Their spacing is of no significance and, unlike contour plots, these lines can converge into, or diverge out of, a point where the flow is divergent (or convergent).

2. **Isotachs**—contours of wind speed.

Such maps take some getting used to; we shall see some examples in what follows.

8.2 The Trade Wind circulation

The climatological mean upper and lower tropospheric winds in Jan and July are shown in Figs 8.1 and 8.2. Note the presence, across the Pacific and Atlantic oceans just north of the equator, and across the Indian ocean south of the equator in January, of lines of *low-level convergence* in the flow. These regions are known collectively as the **intertropical convergence zone (ITCZ)**, which is associated with frequent and extensive deep convection. A schematic is shown in Fig. 8.3 of the low level flow over the Pacific east of the date line. (The Atlantic region is similar, also the Indian Ocean region in northern winter, except that the latitude is reversed). A typical latitude-height cross-section is shown in Fig. 8.4.

The zone of deep cumulonimbus (Cb) and rainfall is located near the maximum sea surface temperature (SST) but is significantly narrower (we'll see the SST distribution in Fig. 8.8, below). The subsidence near the subtropical highs produces a low level inversion—the **trade inversion**, capping

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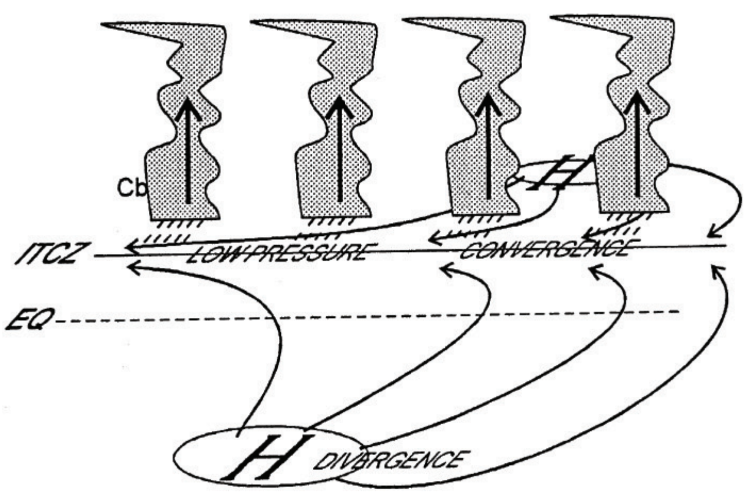
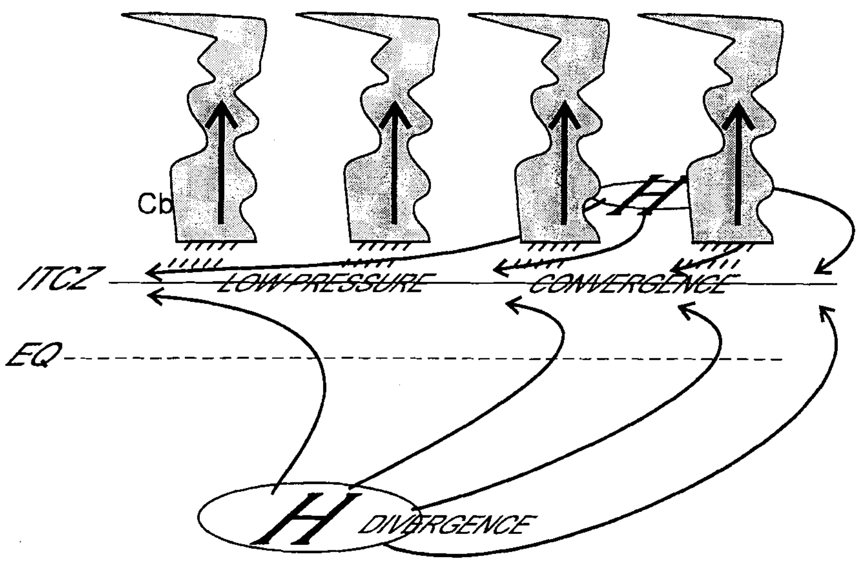


Figure 8.3: Schematic of the trade wind/ITCZ circulation



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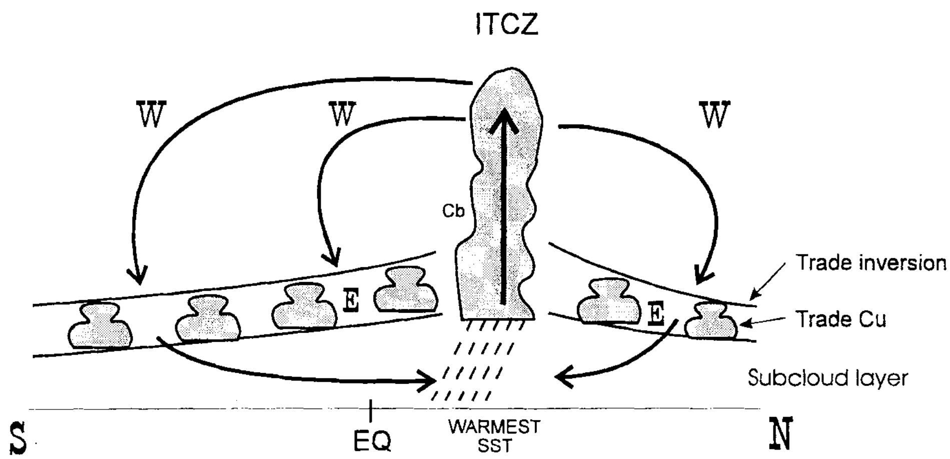

Figure 8.4: Schematic N-S cross-section of the ITCZ.

a layer of shallow "**trade cumulus**", over the ocean; **deserts** over land. Because these structures are long in the E-W direction, it seems sensible to try (at first) to understand the flow in terms of an axisymmetric (no E-W variation) theory (such as that for the tropical Hadley circulation, of which these structures form a part). However, the presence of upper-level westerlies over the equatorial regions in N winter and immediately south of the equator in N summer is inconsistent with this, as we will now see.

For inviscid flow under zonal symmetry, angular momentum, whose density is

$$m = \Omega a^2 \cos^2 \varphi + ua\cos\varphi \ ,$$

is conserved. Now, suppose all the ascending air within the ITCZ leaves the boundary layer at altitude φ_0, and that, because of boundary layer friction, $u \approx 0$ within the boundary layer. Then the angular momentum density of the air ascending within the ITCZ is

$$m_0 = \Omega a^2 \cos^2 \varphi_0 \ .$$

Once the air has left the boundary layer, friction becomes negligible and angular momentum is conserved, so that $m = m_0$ all along the streamlines of the flow until it subsides and enters the boudary layer again. In the upper troposphere, therefore, $m = m_0$, whence

$$\Omega a^2 \cos^2 \varphi + ua\cos\varphi = \Omega a^2 \cos^2 \varphi_0 \ ,$$

i.e.,

$$u = \Omega a \frac{[\cos^2 \varphi_0 - \cos^2 \varphi]}{\cos\varphi} \ .$$

Specifically, at the equator $\varphi = 0$,

$$u_{eq} = -\Omega a \sin^2 \varphi_0 \ .$$

Thus, if $\varphi_0 = 0$, $u_{eq} = 0$; otherwise, $u_{eq} < 0$, *i.e.*, easterly.

Thus, while an axisymmetric model does represent many features of the observed tropical circulation, it is clearly incomplete.

8.3 The "Walker Circulation" of the equatorial Pacific atmosphere

8.3.1 Observations; the atmosphere

To understand the failure of axisymmetric models, we need to consider the zonally asymmetric nature of the tropical circulation and the processes driving it. The observed situation is summarized in the following maps of rainfall (Fig. 8.5), and OLR [1] (Fig. 8.6). Note the following:

1. Latent heat release over the Pacific (illustrated in rainfall and/or OLR maps) is concentrated in the ITCZ, in the South Pacific convergence (SPC), over continental tropical America and, especially, in the far western equatorial Pacific around Indonesia and Melanesia.

2. There are upper tropospheric westerlies / lower tropospheric easterlies over the equatorial Pacific and the opposite pattern (in N winter) over Indian ocean.

3. The circulation in the equatorial (x-z) plane (Fig. 8.7) shows upwelling over the west Pacific and S America but downwelling over the east Pacific (and east Atlantic)—a suggestion of an overturning circulation in the longitude-height plane, driven by localized thermal driving over Indonesia? (*cf.*, circulation in the latitude-height plane—the Hadley circulation—driven by localized heating in the ITCZ.)

8.3.2 Observations; the ocean

Why is the west Pacific wet, the east Pacific dry? The distribution and seasonal variation of sea surface temperature (SST) is shown in Fig. 8.8). Note how the rainfall / OLR pattern mirrors (to a reasonable approximation) the location of the warm water. Note also the narrow strip of cold water in

[1] Outgoing Longwave Radiation (OLR) is measured from satellites; it is the total IR emission from the Earth. Assuming balckbody radiation, the OLR is a measure of the temperature of the emitting layer. For radiation from the surface or from low cloud, the emitting layer is warm, and OLR high. For radiation from high clouds—and only the tops of deep convective clouds are opaque enough—the emitting layer is cold. Thus, low OLR corresponds to regions of deep convection and thus of heavy rainfall. In fact, there is a good quantitative correspondence between rainfall and negative OLR anomalies.

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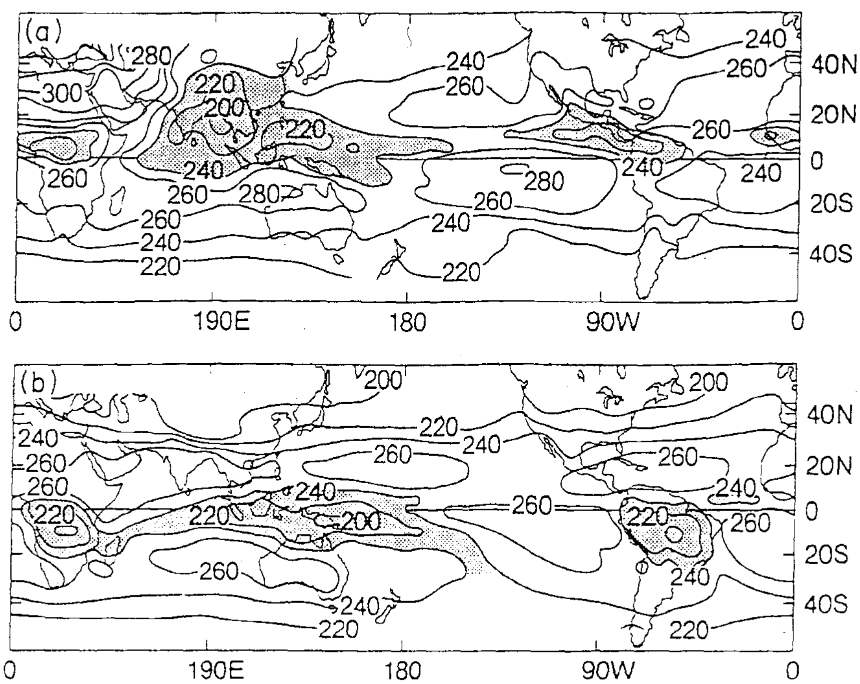

Figure 8.6: Outgoing longwave radiation (OLR) in January and July. (Contour interval: $50 \mathrm{Wm}^{-2}$.)

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the equatorial east Pacific, separating bands of warmer water to the north and south. Note the similar pattern of SST in the Atlantic.

Why does the SST look like this? The equatorial ocean is subjected to a primarily easterly wind stress. In response to this wind stress, the primary balance of the mixed layer to a westward wind stress is simply geostrophic:

$$-fv = \tau/\rho H$$

where τ is the wind stress and H the depth of the mixed layer. Thus, there is (for $\tau < 0$) a northward Ekman flow north of the equator, and a southward flow to the south. This induces *upwelling* along the equator, which brings cold deep water to the surface. This is illustrated in Fig. 8.9.

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But this is not the whole story; the westward wind stress also drives a westward surface flow (because the upwelling raises the thermocline near the equator, requiring a westward flow in geostrophic balance). The surfaces waters are thus advected westward, becoming warmer because of heat input from above and because the westward flow produces a convergence in the western ocean which opposes the tendency for upwelling there. So, though the thermocline is elevated in the east and SST is cold there, it is depressed in the west where the SST is much greater. This is shown in the temperature cross-section of Fig. 8.10.

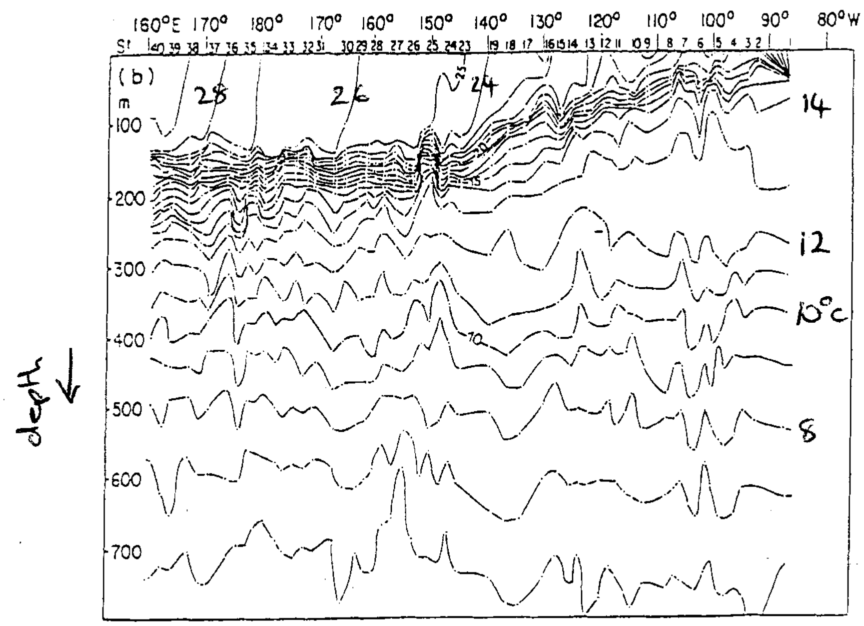

Figure 8.10: Longitude-depth cross-section of temperature across the equatorail Pacific ocean.

Note also that, because of the EW boundaries, there is a "piling up" of water in the west which produces an E-W pressure gradient; below the surface (away from the direct effects of the wind stress) this pressure gradient drives an *eastward* "jet", the **equatorial undercurrent**. This current (and other aspects of the circulation) is confined tightly to the equator—even though the wind stress distribution is broad—because of the governing equatorial dynamics.

8.3.3 Theory of the Walker circulation

In the simplest picture of the zonally symmetric Hadley circulation—such as we depicted in Fig. 8.4—we consider the circulation driven by a latitudinally localized, but zonally symmetric, heating. As we have seen, the dominant heating—latent heating, implicit in the rainfall and OLR patterns—is far from zonally symmetric, having a strong maximum in the far western equatorial Pacific. This suggests an analogous, simple model for the Walker circulation—the circulation driven by a longitudinally localized heating on the equator.

Since the prevailing winds are weak near the equator, we neglect any background flow. If we assume the problem to be linear (which will be accurate for a sufficiently small amplitude disturbance) then the problem becomes one of wave propagation (see Fig. 8.11): the only way the circulation can extend

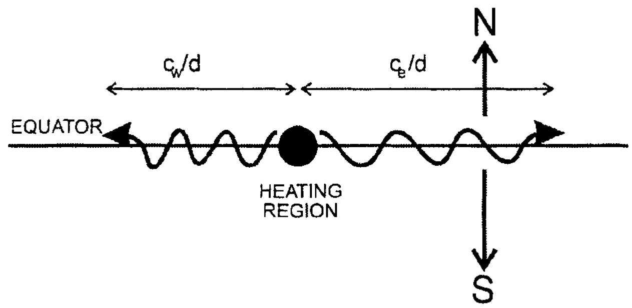

Figure 8.11:

beyond the localized heating region is through propagation of information. As the information propagates, it will be dissipated, at rate d, say. If the fastest wave propagation (group velocity) eastward is c_e, then we expect the information to propagate a distance c_e/d before being dissipated; therefore the forced circulation should extend a distance c_e/d east of the forcing. Similarly, the circulation will extend a distance c_w/d to the west, where c_w is the fastest westward wave speed. Rossby wave dynamics has taught us not to expect that $c_e = c_w$.

In fact, equatorial dynamics are rather special. Rossby waves still exits there, though they are a special type that are confined to an equatorial "wave guide"; like middle latitude Rossby waves, the long waves (which are the most important) have westward group velocity. In addition, however, there is a new (to us) and rather special class of waves that exist within this wave guide, known as **equatorial Kelvin waves.**

Consider a shallow water sysytem of mean depth D; with the addition of rotation,

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} - fv = -g\frac{\partial h}{\partial x} .$$

Note that the second term vanishes under our assumption of linearity about a motionless basic state. Now, in our midlatitude analyses we have used the "beta-plane" approximation. We can do the same here, writing (for $\varphi \ll 1$)

$$f = 2\Omega \sin \varphi \simeq 2\Omega\varphi = \beta_0 y$$

where $y = a\varphi$ and

$$\beta_0 = \frac{2\Omega}{a} = 2.29 \times 10^{-11}\text{m}^{-1}\text{s}^{-1}$$

is the equatorial beta parameter. Making this substitution, and adding the y-momentum equation and the linearized continuity equation,

$$\begin{aligned} \frac{\partial u}{\partial t} - \beta_0 yv &= -g\frac{\partial h}{\partial x} ; \\ \frac{\partial v}{\partial t} + \beta_0 yu &= -g\frac{\partial h}{\partial y} ; \\ \frac{\partial h}{\partial t} + D\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) &= 0 . \end{aligned} \tag{8.1}$$

Equations (8.1) actually describe many kinds of wave motion, including Rossby waves, but there is one special kind of solution that turns out to be very important. Let's look for solutions that have $v = 0$, i.e., flow that is exactly E-W and vertical. The second of (8.1) immediately gives

$$\beta_0 yu = -g\frac{\partial h}{\partial y} ; \tag{8.2}$$

there is geostrophic balance in the NS direction, despite the fact that we are at the equator. The first of (8.1) gives

$$\frac{\partial u}{\partial t} = -g\frac{\partial h}{\partial x} , \tag{8.3}$$

which is precisely the equation for small amplitude nonrotating motions—the Coriolis term has dropped out of the $u-$equation because $v = 0$. Now, for solutions of the form

$$\begin{pmatrix} u(x,y,t) \\ h(x,y,t) \end{pmatrix} = Re \begin{pmatrix} U(y) \\ H(y) \end{pmatrix} e^{ik(x-ct)} ,$$

(8.3) gives

$$cU = gH ,$$

whence (8.2) gives

$$\frac{dH}{dy} = -\frac{\beta_0}{c} y H .$$

This has solutions

$$H(y) \sim \exp\left(-\frac{\beta_0}{2c} y^2\right)$$

which is a physically reasonable solution that satisfies boundedness as $y \to \pm\infty$ only if $c > 0$. So these motions are trapped near the equator, with a characteristic length scale $L = \sqrt{c/\beta_0}$, and propagate *eastward.* These are equatorial Kelvin waves. For the kinds of baroclinic motions we are interested in here[2], $c \sim 20\text{ms}^{-1}$ and $L \sim 1000$km (about 10° of latitude).

So, we have Rossby waves to transmit information westward and Kelvin waves to transmit it eastward. For a dissipation rate of $d = 1/(5\text{days})$, the

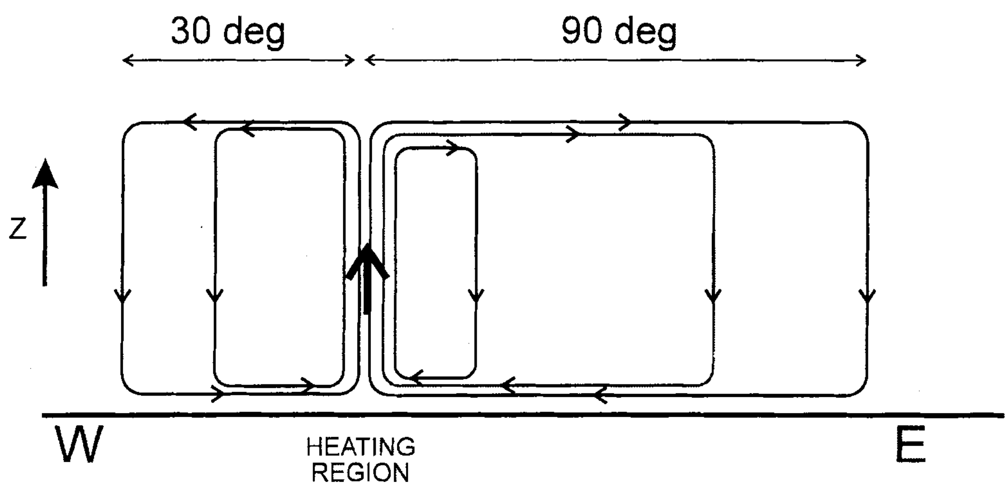

Figure 8.12:

Kelvin waves will reach a distance of $c/d \simeq 10000$km, about 90° of longitude. It turns out that the fastest Rossby wave group velocity is only one-third that of the Kelvin wave so, in the context of Fig. 8.11, the circulation extends only

[2]Kelvin waves are nondispersive, so this value is the same for phase and group velocities.

one-third as far to the west as it does to the east. So we expect a circulation something like that shown in Fig. 8.12.

The horizontal structure, from a complete linear solution, is shown in the Fig. 8.13 (u, v and p shown are for the lower level reponse; upper levels are

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opposite in sign). Note:

- the different structures and different zonal length scales (factor of 3) of the Kelvin and Rossby components to east and west.

- The twin low level cyclones / upper level anticyclones straddling the equator just west of forcing.