Atmospheric General Circulation

Weston Anderson

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1 Introduction

In this section we're going to describe some large-scale features of the climate. We begin with a description of the energy balance of the earth as motivation for then describing how this energy is transported around the earth. We'll provide a brief introduction to major zonal and meridional features of climate, and describe what makes a climate perpetually wet or dry.

2 Energy balance of the atmosphere

As we described in our notes on radiation, the net energy balance of the earth is simply that incoming (shortwave) solar radiation at the top of the atmosphere (R_{toa}) must be balanced by outgoing (longwave) terrestrial radiation (R_{out}). However, although the earth as a whole is in radiative balance, there are a number of ways that energy moves from one part of the system to another. Here we'll briefly describe the atmosphere's role in transporting energy around the earth system. Unlike the ocean, the heat capacity of the atmosphere is negligible over an annual average. So in the following discussion we'll neglect any storage of heat in the atmosphere.

We can first approximate the net radiation (R_a) added to an atmospheric column as the difference between the incoming solar radiation at the top of the atmosphere and the outgoing longwave radiation emitted to space. Energy may also be added to the atmosphere in the form of of sensible heat (SH) or latent heat (LP; usually latent heat of vaporization released by precipitation). Finally, any residual energy in the atmospheric column must be transported by large scale circulations, which we will call an atmospheric energy flux divergence (ΔF_a).

$$R_a + LP + SH = \Delta F_a$$

A useful way to conceptualize this energy balance on a global scale is to plot the zonally averaged quantity of each of term, as is done in Figure 1. This will tell us something about how energy is transported from the equator to the poles, which is necessary because incoming solar radiation is greater at the equator as compared to the poles (see notes on radiation and on the Hadley cell for a more detailed description of this).

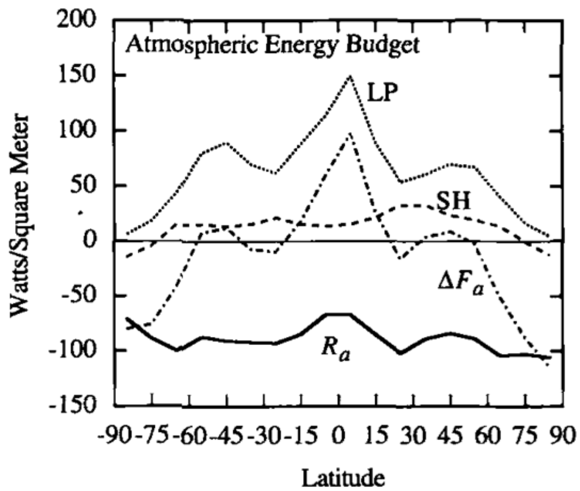


Figure 1: Credit: Figure 6.1 of (Hartmann, 2015). Distribution of zonally averaged terms in the atmospheric energy balance

We can tell from the above figure that sensible heating plays a relatively minor role in the atmospheric energy balance. We can also see that the net radiation is relatively constant over the entire earth. This means that the dominant balance of energy in the atmosphere is between the latent heat of vaporization and divergence of energy associated with large scale atmospheric motions. Let's consider the meridional structure of each of these terms.

Latent heat of vaporization associated with precipitation is everywhere positive, meaning that latent heat adds a tremendous amount of heat to the atmosphere. This term is greatest in the zonal bands that are wettest, such as the tropics (where we have constant precipitation) followed by the midlattitudes. Latent heating is at a minimum in the subtropics, which contain many of the world's deserts, and at the poles, which are so cold that the atmosphere holds little water vapor.

The divergence of energy in the atmosphere looks similar to the latent heat of vaporization. The two most prominent features of the energy flux divergence is the export of energy from the tropics and the import of energy to the poles. This is an important result of the general circulation of the atmosphere, and one that we will explore in more depth below.

In the following sections we will break the general circulation into three terms: one corresponding to the zonal mean circulation (in space and time), one corresponding to deviations from that mean in time (transient eddies) and one corresponding to longitudinal deviations from that zonal mean in space (stationary eddies).

3 Mean circulations

3.1 Zonally averaged circulations

We'll begin with a description of the zonally averaged circulation. The meridional component of this circulation refers to the Hadley and Ferrel cells, which act to transport water and energy from the equator to the poles. For a more detailed description of the Hadley Cell and its structure, see the 'Overturning Circulations' notes. The zonal mean wind tells us about the location and strength of the jet streams in the atmosphere. It's important to note here that in general the zonal winds on earth (the jet streams) are much stronger than the velocities of meridional winds, and hundreds of times stronger than the vertical velocities in the atmosphere. Never-the-less, the mean meridional circulations play an important role in the climate system.

3.1.1 Hadley and Ferrel cells

To get a picture of the mean meridional circulation (MMC), we can define the northward mass stream function.

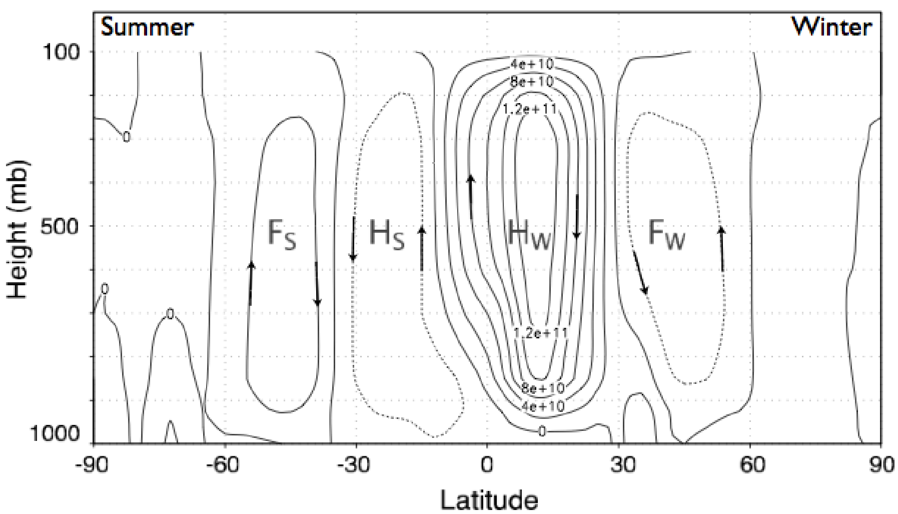


Fig. 11.3 The observed, zonally averaged, meridional overturning circulation of the atmosphere, in units of kg s^{-1}, averaged over December–January–February (DJF). In each hemisphere note the presence of a direct *Hadley Cell* (H_W and H_S in winter and summer) with rising motion near the equator, descending motion in the subtropics, and an indirect *Ferrel Cell* (F_W and F_S) at mid-latitudes. There are also hints of a weak direct cell at high latitudes. The winter Hadley Cell is far stronger than the summer one.

Figure 2: Credit: Vallis (2006)

In Figure 2, we have plotted that stream function for northern hemisphere winter. We have four main cells, two Hadley cells centered near the equator and two Ferrel cells centered in the midlatitudes. To zeroth order, we can explain this circulation in terms of incoming solar radiation, and equator-to-pole temperature gradients. Insolation at the equator drives rising motions. Here, for southern hemisphere summer, that rising motion is slightly south of the equator following the insolation maximum. The equator-to-pole temperature gradient then drives the strength of the cell, meaning that the winter hemisphere cells are stronger than the summer hemisphere cells (because the winter hemisphere temperature gradient is greater). Below we can look at the Hadley cell in northern hemisphere winter, summer and the annual average. Note the rising motions migrating from just south of the equator in NH winter to just north of the equator in NH summer. The annually averaged streamfunctions, although often the picture of the Hadley cell that we are given, is somewhat deceiving: it depicts two cells of equal strength, when in reality the Hadley cell is dominated by a single cross-equatorial cell flowing from summer hemisphere to winter hemisphere.

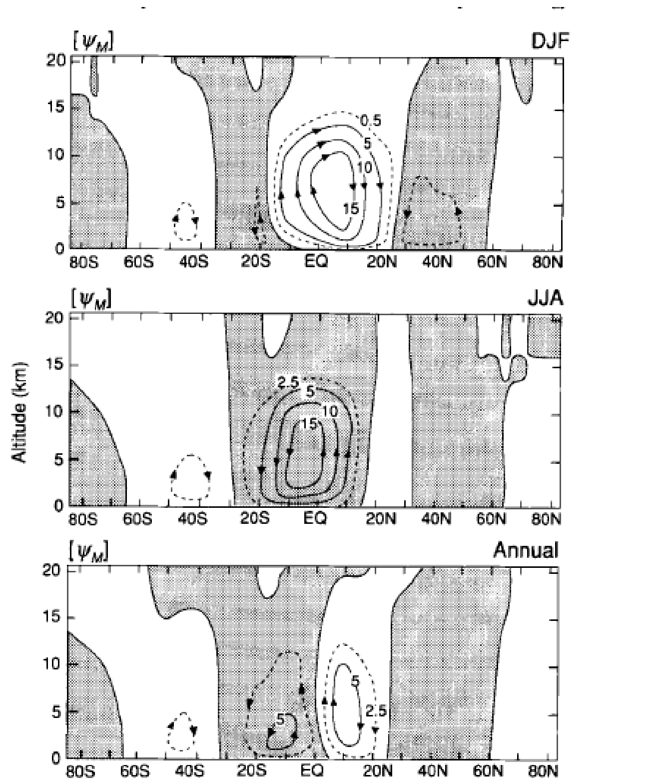


Fig. 6.5 Latitude–height cross section of the mean meridional mass circulation. Shaded values are negative; units are 10^{10} kg s^{-1}. [Data from Oort (1983).]

Figure 3: Credit: (Hartmann, 2015)

In the Hadley cell warm air rises on the equator and cool air sinks in the subtropics, we call this a *thermally direct* circulation. In the Ferrel cell relatively warmer air is sinking while further poleward colder air is rising. This is a thermally indirect circulation and is the byproduct of strong eddy motions in these latitudes. Let's now take a minute to understand why the Hadley cell is important, even if the velocities involved are many times smaller than the zonal velocities associated with midlatitude jet streams

The flux of mass and therefore moisture associated with the mean meridional circulation is largest near the equator, where warm moist air converges, rises, and results in heavy precipitation. Following convection, drier air travels poleward aloft, sinks in the subtropics and flows back to the equator at the surface, drawing moisture from the surface and into the atmosphere along the way.

This circulation is transporting both moisture and energy. In general, we divide energy fluxes into three main categories: 1. the energy associated with the temperature and heat capacity of air (sensible), 2. the gravitational potential energy associated with the vertical height of the air mass (potential) and 3. the energy associated with moisture in air and the latent heat of vaporization (latent). Sensible heat can be thought of as dry heat, present even when there is no moisture in the air. Latent heat is the heat of vaporization associated with precipitation. Water requires energy to transition from the liquid to gas phase transition (think boiling water), and therefore releases energy in the gas to liquid phase transition (precipitation). Taken as a whole, there three types of energy constitute *Moist Static Energy*, which is a conserved variable, so long as the composition of the air parcel (including moisture) remains the same.

Moist static energy$= c_pT + gz + Lq =$sensible + potential + latent

Above c_p is the specific heat of air, T is air temperature, g is the gravitational constant, z is height, L is the latent heat of vaporization of water, and q is the specific humidity of air (how much moisture is in the air). Moist static energy increases in the vertical, so the poleward motion of air aloft in the Hadley cell and equatorward motion at the surface results in a net transport of energy towards the pole. Mean meridional cells transport energy polewards, but not particularly effectively outside of the tropics as we can see in Figure 4.

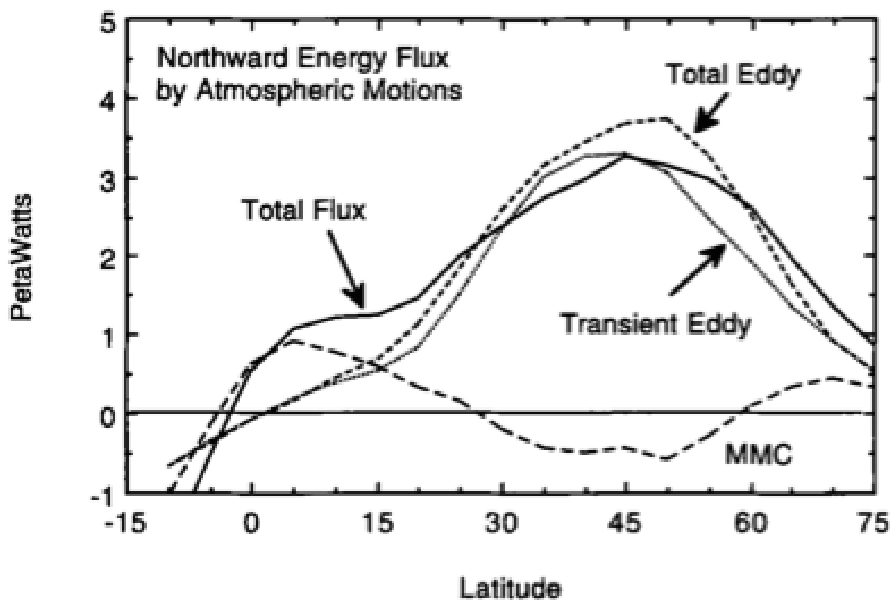


Fig. 6.11 Annual average northward energy flux plotted versus latitude in the Northern Hemisphere. Units are 10^{15} W. Mean meridional circulation (MMC). [Data from Oort (1971). Reprinted with permission from the American Meteorological Society.]

Figure 4: Credit: (Hartmann, 2015)

So what transports moisture and energy outside of the tropics? Well, as you may have inferred from our discussion of the Ferrel cells and of the above graph, eddy motions become important in the middle to upper latitudes. We'll discuss these eddies in detail in a moment, but first let's talk about how the zonal winds vary with latitude.

3.1.2 Midlatitude jet streams

The equator-to-pole temperature gradient, along with the Thermal Wind relation, can tell us a great deal about the structure of zonal winds in the atmosphere (for the math, see notes on Geostrophy and Thermal Wind). Without going into the math, however, we can say that the Thermal Wind equation relates the vertical wind shear to meridional gradients in temperature. In the troposphere, temperature falls monotonically with latitude. In the troposphere, meridional temperature gradients are largest at the edge of the subtropics, which leads to zonal jets with maximums in these latitudes (see Figure 6 below). The temperature gradient (and therefore zonal jet) is stronger in the winter hemisphere than the summer because peak insolation is shifted towards the summer hemisphere so the winter pole receives virtually no direct sunlight. These jet streams peak in strength near the poleward termination of the Hadley cell. If Figure 6 seems a little abstract, Figure 5 may be a more tractable way of connecting the location of the jet streams to the Hadley cells. Note, however that in this figure we are looking again at the annually and zonally averaged circulation.

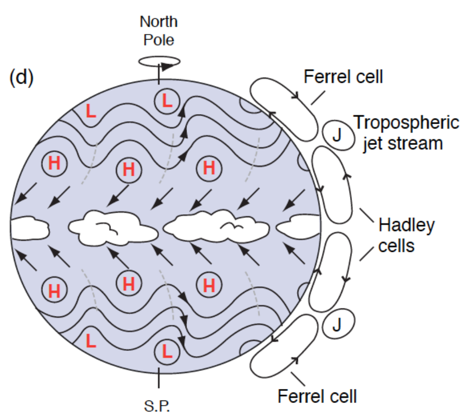


Figure 5: Schematic of the annual mean zonally symmetric general circulation. H represents high pressure, L represents low pressure systems.
Credit: **?**

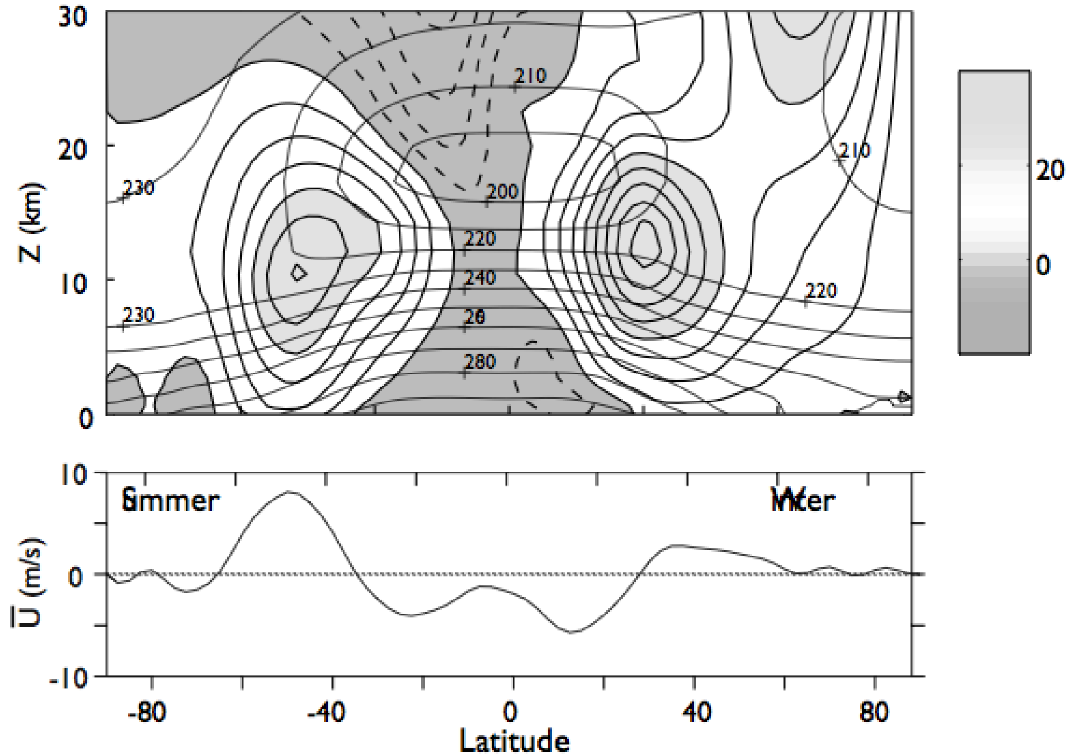


Figure 6: Top panel: filled contours are the zonally averaged zonal wind, outlines overlayed on these contours are contours of temperature. Bottom panel: zonally averaged zonal winds at the surface
Credit: Vallis (2006)

Above the troposphere (starting somewhere between 8 and 16 km above the sea level) is the stratosphere, where temperature increases with height. In the stratosphere the meridional temperature gradient is reversed, meaning the winds that strengthened from the surface will from this point weaken with height.

At the surface, winds alternate from easterly near the equator to westerly in the midlatitudes and finally easterly again at high latitudes. Surface winds are stronger in the southern hemisphere due to a lack of drag from continental land masses. And while only one season is shown here, surface winds within a given hemisphere are stronger in the winter as compared to the summer season (corresponding to the stronger jet aloft).

3.2 Meridionally averaged circulations

3.2.1 The Walker cell

In the tropics the mean circulation in the east-west direction (zonal) is referred to as the Walker Circulation (see Figure 7 below).

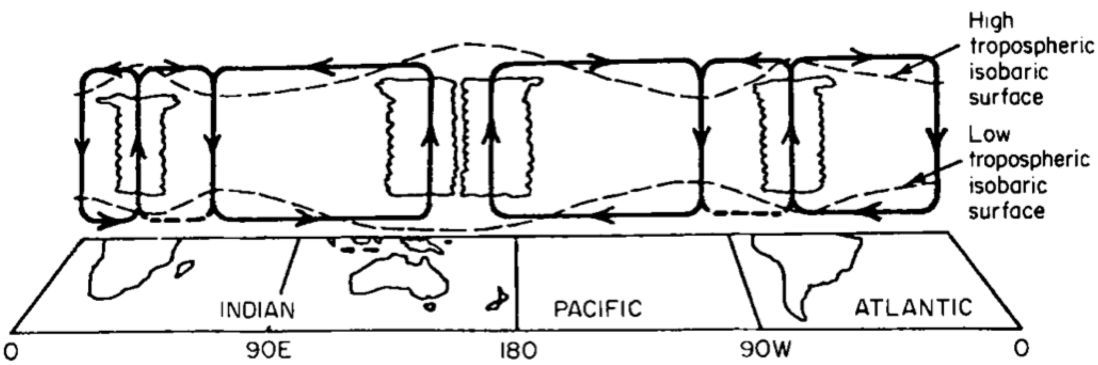


Fig. 6.22 Schematic view of the east-west Walker circulation along the equator indicating low-level convergence in regions of convection where mean upward motion occurs. [From Webster (1983).]

Figure 7: Credit: (Hartmann, 2015)

This circulation is driven by rising motions over Africa, South America and the Maritime Continent (comprised of Indonesia, Malaysia and New Guinea). This circulation isn't driven by gradients of insolation – as the meridional circulations were by the equator-to-pole temperature gradient – but it does relate to contrasts between wet, cool regions and relatively warm areas. We'll discuss many of these regions in more depth when we discuss monsoonal climates. But for now, let's return to meridional transport of energy and water to discuss the role of eddy motions in the atmosphere.

4 Eddy transport of heat and moisture

We can conceptualize eddy motion as being atmospheric motions that deviate from the zonal and time mean meridional circulation. These eddies can therefore be anomalies in time (we call these transient eddies because the develop and decay) or they can be anomalies in space (we call these stationary eddies because they are largely fixed in space).

4.1 Transient eddies and storm tracks

Transient eddies are familiar to most of us as synoptic scale storms (1000s of kms) that pass over the continental US in about a week. This can also be referred to as baroclinic instability (see notes on equations of motion for more on this). The general structure of these eddies is depicted below in Figure 8. In terms of transport of energy and moisture (as well as momentum, which we will come back to later), it is important to note that the streamlines of the eddy are out of phase with the thermal anomalies. This means that although net transport across the dashed line is zero, northward transport tends to be associated with warm surface anomalies while southward transport is associated with cold surface anomalies, which leads to a poleward transport of heat and energy by these eddies.

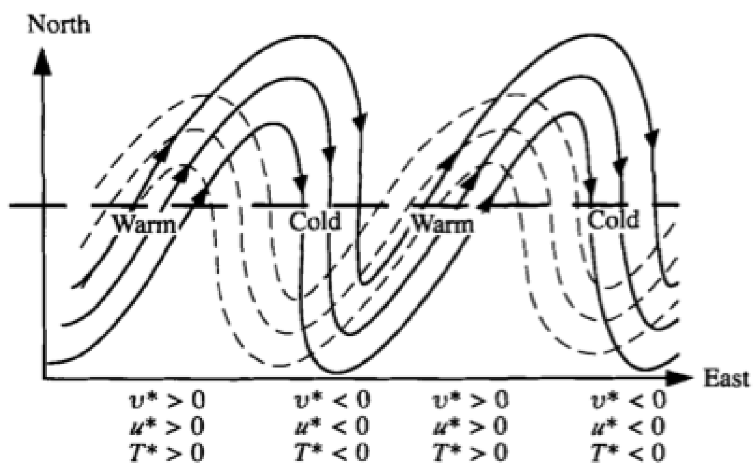


Fig. 6.6 Schematic of the streamlines (solid) and isotherms (dashed) associated with a large-scale atmospheric disturbance in midlatitudes of the Northern Hemisphere. Arrows along the streamline contour indicate the direction of wind velocity. The streamlines correspond approximately to lines of constant pressure, since the winds are nearly geostrophic. The signs of the deviations of the wind components from their zonal-average values are shown to illustrate that the NE–SW tilt of the streamlines indicates a northward zonal momentum transport, and the westward phase shift of the temperature wave relative to the pressure wave gives a northward heat transport.

Figure 8: Credit: (Hartmann, 2015)

Transient eddies don't occur in equal amounts everywhere. They relate to atmospheric instabilities, and so become important in the midlatitudes. We can see in Figure 9 that polewards of 30 N/S transient eddies become the primary driver of energy transport. Equatorward of this line the mean meridional circulation dominates energy transfer. The latitudes of maximum transport of energy by transient eddies is referred to as the 'storm tracks'.

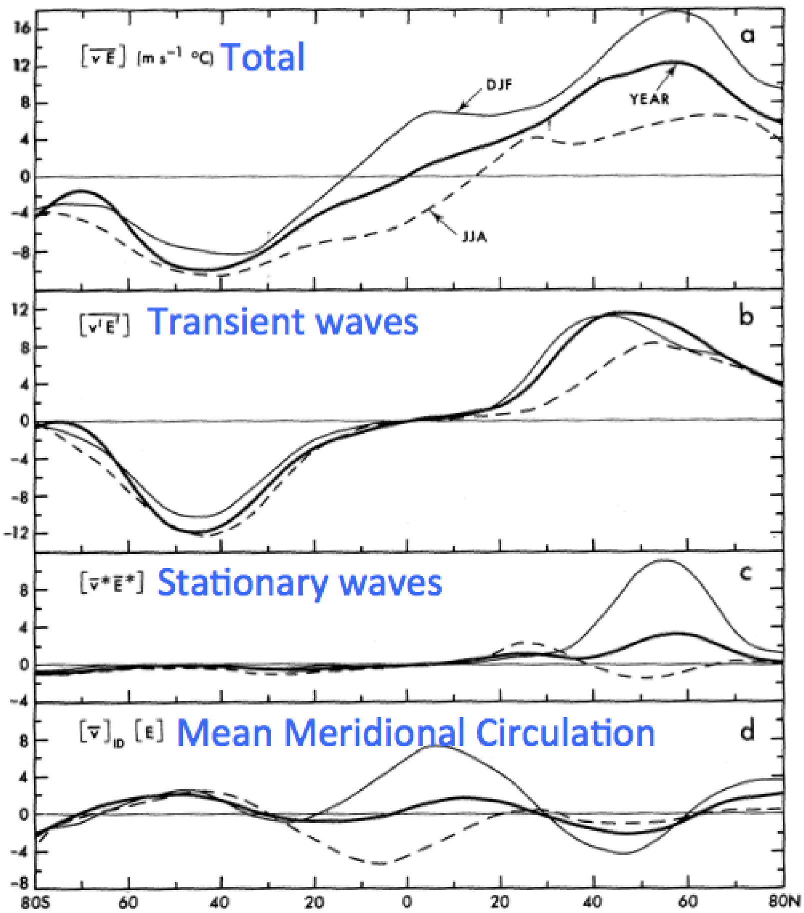


FIGURE 13.11. Meridional profiles of the zonal- and vertical-mean northward transport of energy by all motions (a), transient eddies (b), stationary eddies (c), and mean meridional circulations (d) in °C m s^{-1} [to convert to units of 10^{15} W multiply values by $2\pi R \cos\phi\ c_p(p_0/g) \approx 0.4 \cos\phi$; from Oort and Peixoto, 1983].

Peixoto and Oort *Physics of Climate* 1992

Figure 9: Credit: Taken from slides of Ron Miller, NASA GISS, which was in turn adapted from (Peixoto and Oort, 1992)

The storm tracks are not zonally symmetric, but instead are greatest where there is a zonal wind flowing from dry, cold land to warm, moist oceans. This tends to happen as air moves eastward off the Asian continent and over the Pacific ocean, and where air moves eastward off North America and over the Atlantic ocean.

Finally, we can see from Figure 9 that we have discussed the mean meridional circulation and transient eddies but not stationary eddies, which are important in the Northern Hemisphere (particularly in winter).

4.2 Stationary waves

Stationary eddies are driven by zonal anomalies of elevation (i.e. mountains) and temperature (land-sea contrasts). Both of these forcings are most pronounced in the Northern Hemisphere, which contains the majority of the earth's land masses. This explains why stationary eddies are important in the Northern Hemisphere but less so in the Southern Hemisphere, as shown in Figure 9. But why the seasonal dependence? Well, orographic forcing is one means of producing stationary eddies (and indeed this does depend on the seasonally varying background flow) but stationary eddies are also produced by land-sea temperature contrasts. And because oceans have a much larger heat capacity than does the land, the land experiences much larger seasonal swings in temperature, which means that continents tend to be warmer than the oceans in the summer and cooler than the oceans in the winter. Cold flow off of the continents in the wintertime move over the warm Gulf Stream to induce large baroclinicity. For more on stationary waves see the notes on atmospheric Rossby waves.

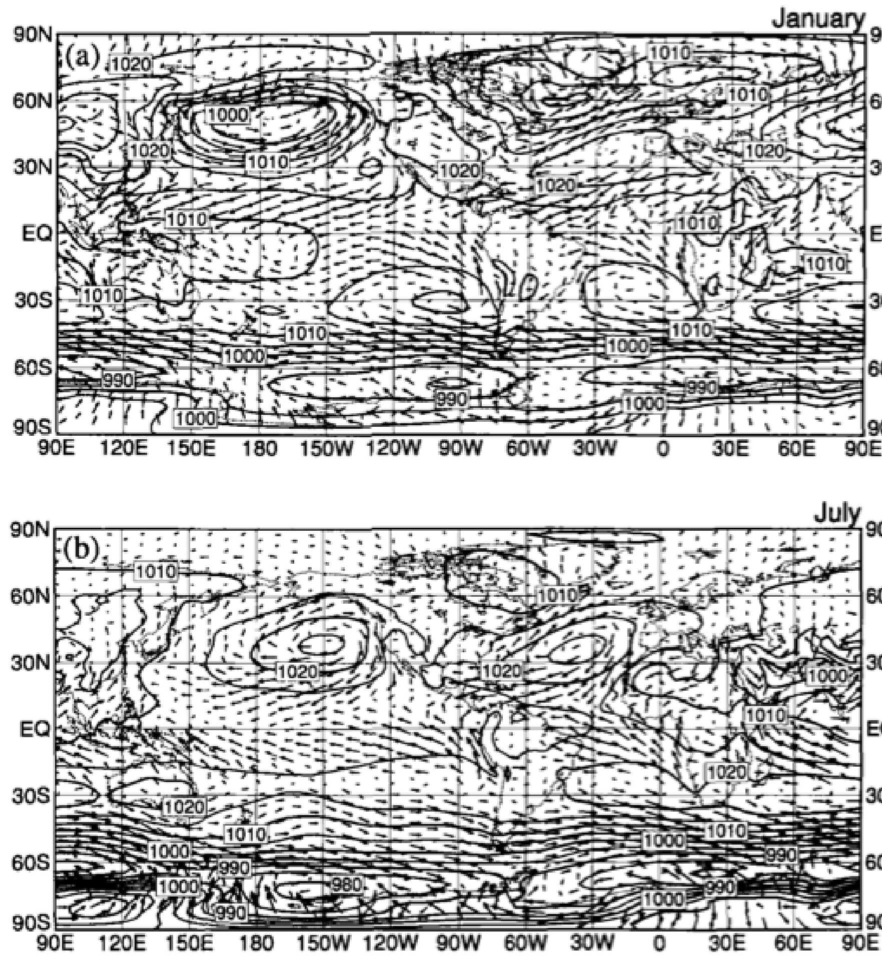
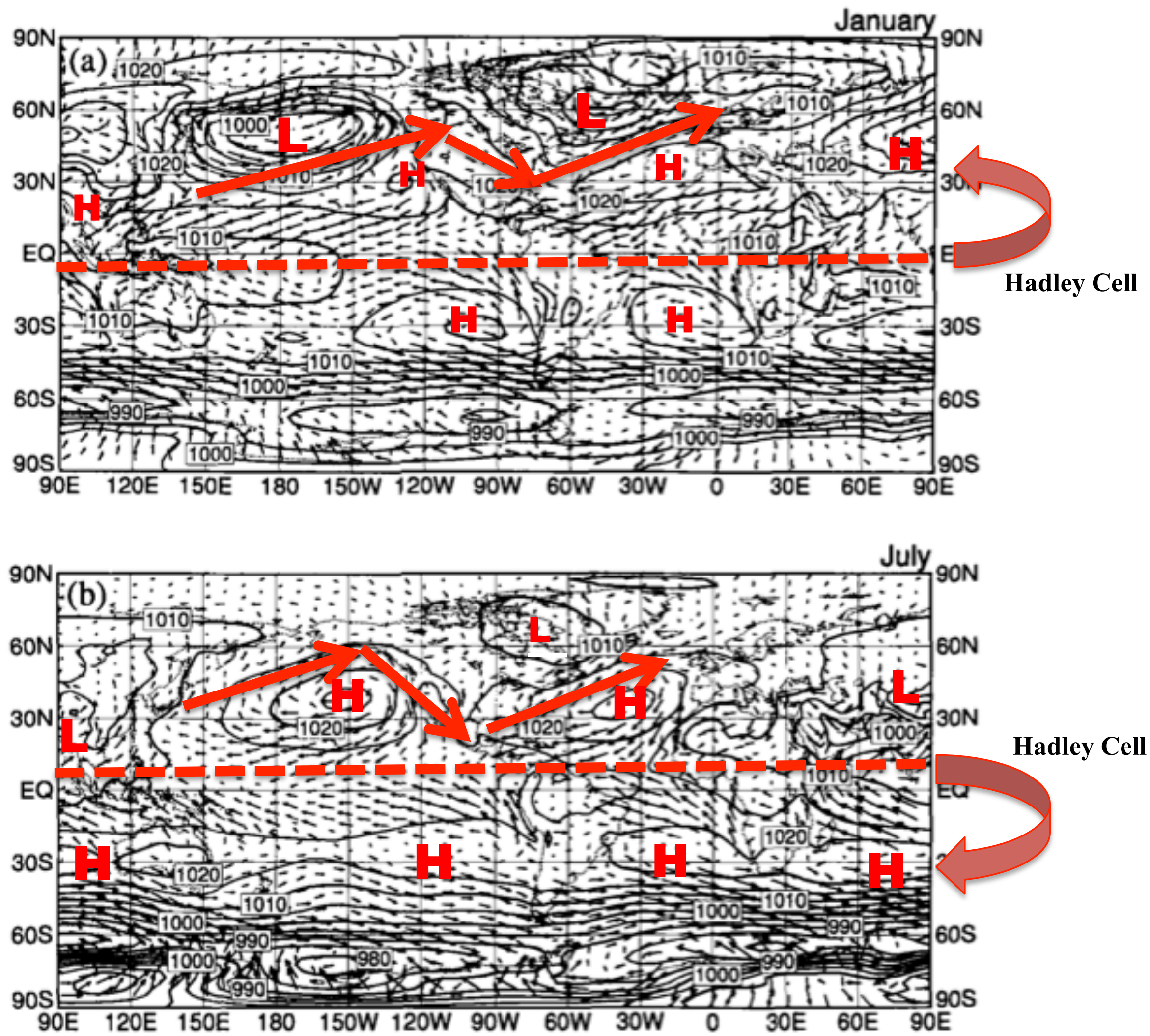


Fig. 6.18 Maps of mean sea-level pressure for (a) January and (b) July. Wind vectors for the 1000-mb level are superimposed. Data are 1980–87 analyses from a forecast model. Contour interval is 5 mb and largest vector represents a wind speed of 12 m s^{-1}.

Figure 10: Credit: (Hartmann, 2015)



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4.3 Origin of stationary waves (and associated storm tracks)

The stationary waves may therefore be a combination of any number of factors, including

- **Orography** excites downstream troughs. Additionally, air that deflects around orography creates a pool of cold air to the northeast of the mountain (advection of cold air that was deflected to the north returning southward) and warm, precipitating air on the leeward side of the mountain (as winds that were deflected to the south return northwards and rise along sloping isentropes to induce precipitation). So the local effects of mountains reinforce the SW-NE tilt of precipitation / storm tracks.

- **Land-ocean contrasts** create differences in drag and moisture availability as well as temperature contrasts.

- **SST fronts** such as the Gulf Stream and North Atlantic Drift create closely spaced and diffuse isentropes, respectively.

We can now revisit Figure 10, which provides an observational estimate of the storm tracks in DJF. We can see that they have a decidedly SW-NE tilt, and that the baroclinicity is greatest over the western boundary currents of the oceans, downstream of the Rockies and Himalayas.

5 What makes a climate wet or dry?

Wet climates typically coincide with frequent precipitation, which is to say where humid air is uplifted. This is common in the tropics, where an abundance of insolation and a convergence of moisture leads to convection. On a seasonal scale, we can also follow the Intertropical Convergence Zone (ITCZ), which is a band of precipitation that circles the earth. The ITCZ is not stationary, but rather migrates north and south roughly with the solar zenith.

5.1 Monsoons

Monssonal climates are climates that have a pronounced wet and dry season. The word ‘monsoon’ derives from an Arabic and/or Hindi word for ‘season’. Wet climates are often a result of the migrating ITCZ, but are also a product of local orography trapping moisture over a continent.

The Himalayas, for example. play a crucial role in the Asian Monsoon. Although it used to be thought that the Tibetan plateau acted as a ‘heat pump’, creating warm air aloft that led to a low pressure system that brought in moist air during the monsoon, the idea has since been discredited. Current thinking is that when warm moist air flows north from the Indian Ocean it is shielded from the relatively cold dry air coming off the Asian continent to the North. The moist air eventually precipitates and releases latent heat that warms the air, which rises to create a low pressure system that is fed by moist inflow (and again shielded from dry inflow by the Himalayas), completing the positive feedback.

As with the Himalayas, orography plays a crucial role in the wet climate of the Amazon. Humid easterly winds off the Atlantic travel over the Amazon and are trapped from continuing to the Pacific without precipitating. The latent heat released as a result of precipitation warms the air and causes it to rise, leading to a low pressure system that pulls in more moist air from the Atlantic. This positive feedback pumps in an incredible amount of moisture to the Amazon basin.

5.2 Deserts

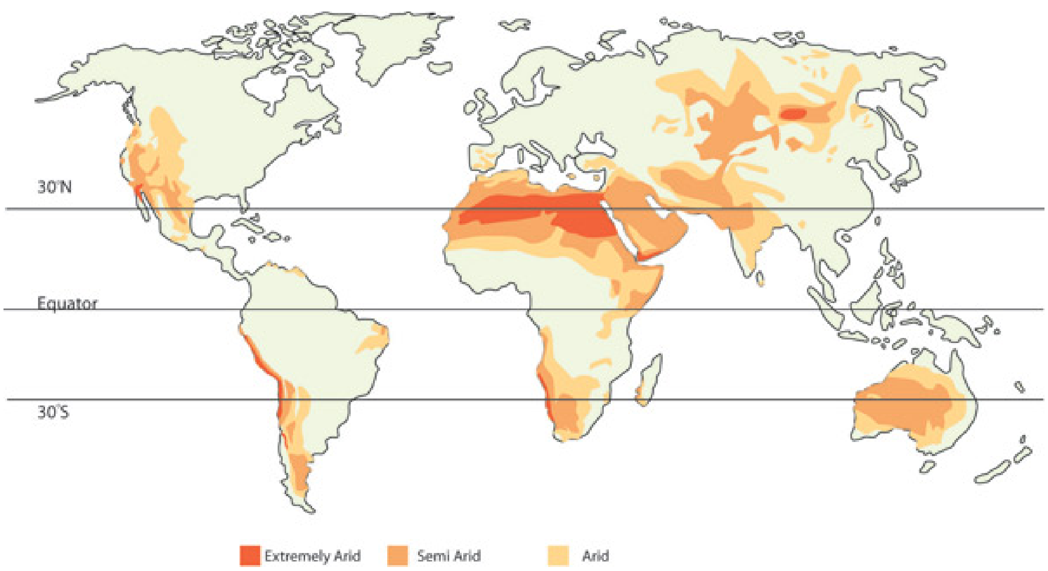

Figure 12:

A climate may be dry due to a few different mechanisms, all of which relate to large-scale atmospheric dynamics. Figure 12 demonstrates where we find deserts in the world (usually on the western side of continents in the subtropics). Rainfall can be suppressed by large scale atmospheric descent, such as in the descending branch of the Hadley cell. This happens in the subtropics (20-40 latitude in both hemispheres), where dry air aloft at the poleward end of the Hadley Cell sinks to the surface and flows equatorward, exporting moisture with it. Subsidence may also be associated with the rising or descending branch of stationary eddies (poleward flow implies rising and precipitation; equatorward flow implies descent and drying). Persistent subsidence can also be caused by orography. Climates downwind of mountain ranges that run perpendicular to the prevailing wind tend to experience persistent large-scale descent, and therefore be exceedingly dry. The North American desert is an example of this. Finally, climates near cold sea surface temperatures may also lack frequent precipitation. The relatively warm air cools as a result of sensible heat exchange with the ocean, which makes the air at the surface cooler and denser than the air above it, forcing the atmosphere into a particularly stable state that inhibits convection. Desert climates experience large diurnal fluctuations in temperature as a result of the radiative heating and cooling of the land surface with little moisture in the air or clouds aloft to emit longwave back to the surface.

Putting these factors together then, we would expect deserts to be most prevalent where we have large-scale descent associated with the Hadley cell (in the subtropics, see Fig. 2) and where we have equatorward summertime flow associated with stationary eddies (see Fig. 10) and where we lack a monsoon circulation.

6 What will happen with Climate Change?

The earth's mean climate is expected to change in a number of ways as a result of increasing greenhouse gasses. Here we step through a few of the expected changes.

6.1 Radiation budget

The most immediate change will be to the radiative balance of the atmosphere and the land surface. Here we'll distinguish the transient responses from the equilibrium responses of the climate system in a 'two-layer' conceptual model that consists of a surface layer (the ocean) and an atmospheric layer. If we first consider the ocean we need to keep two factors in mind: (1) the ocean mixed layer has a relatively high heat capacity, meaning it can store a fair amount of energy but will also take time to heat up, and (2) the ocean mixed layer will (slowly) exchange heat with the deep ocean via diffusion. As we consider the surface response, we also need to keep in mind how the surface fluxes (latent and sensible heat) will respond to forcing at each timescale. We now turn to Figure 13 to illustrate the immediate, intermediate, and final response of the radiative budget.

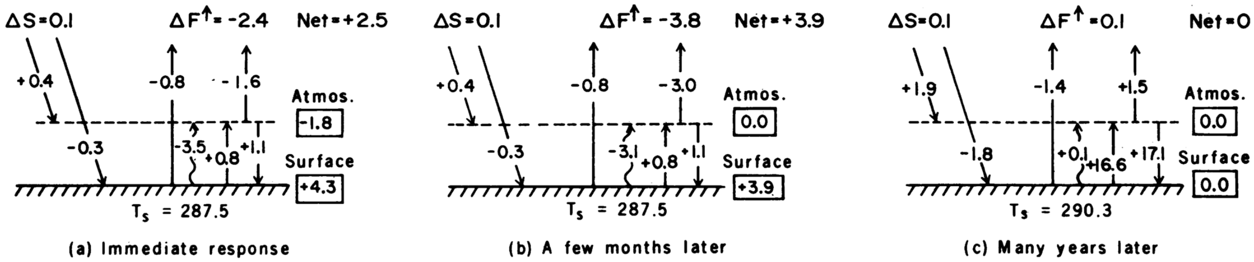

Fig. 4. Change of fluxes (watts per square meter) in the 1-D RC model when atmospheric CO_2 is doubled (from 300 to 600 ppm). Symbols: ΔS, change in solar radiation absorbed by the atmosphere and surface; $\Delta F \uparrow$, change in outward thermal radiation at top of the atmosphere. The wavy line represents convective flux; other fluxes are radiative.

Figure 13: Figure credit: Hansen et al. (1981)

With these factors in mind we can consider how this simple model will respond to a doubling of CO_2. The immediate response will be (i) the added CO_2 raises the atmospheric emission point and therefore reduces emission to space (higher, colder, less OLR). (ii) the troposphere warms (increase in CO_2) and the stratosphere cools (b/c dominant balance is O_3 absorption of SW [no change] and LW emission by CO_2 [increases]). This warming of the troposphere increases the flux from the atmosphere to the ground. (iii) the warming troposphere warms the surface if we assume the two are coupled by a linear lapse rate. But then by energy conservation, the surface latent heat flux must decrease.

The intermediate response (a few months), the atmosphere has come into equilibrium but the earth surface has not. This means that the stratosphere is in equilibrium but the tropospheric temperature will continue to change because it is convectively coupled to the land surface. During this stage energy is gained by the surface and used to heat the ocean. The final response (years later) will be to increase the surface temperature by $\sim 2.7°C$, about half of which is due directly to the CO_2 greenhouse effect and half of which is the water-vapor feedback.

References

James Hansen, Donald Johnson, Andrew Lacis, Sergej Lebedeff, Pius Lee, David Rind, and Gary Russell. Climate impact of increasing atmospheric carbon dioxide. *Science*, 213(4511):957–966, 1981.

Dennis L Hartmann. *Global physical climatology*, volume 103. Newnes, 2015.

José Pinto Peixoto and Abraham H Oort. Physics of climate. 1992.

Geoffrey K Vallis. *Atmospheric and oceanic fluid dynamics: fundamentals and large-scale circulation.* Cambridge University Press, 2006.