3 Exploring heat and moisture transport with atmospheric climatological data

Having explored some of the fundamental dynamics of Hadley (low rotation) and middle latitude (high rotation) regimes in the laboratory, we now use climatological observations to study them in Earth's atmosphere.
To display the data use the Climatologies/Atmosphere tab in the EsGlobe interface.

- Begin by plotting the zonally-averaged temperature (T), meridional wind (v), vertical velocity (ω) and zonal wind (u) for the month of January and July — i.e. the average of all the January's between 1982 and 2010 and likewise for July. Note that in pressure coordinates the vertical velocity is $\omega = Dp/Dt$ and so if air is moving upwards it's $\omega < 0$!)

Making use of these plots we will first discuss the tropical Hadley circula-tion. Next time we will go on to discuss the circulation in the extra-tropics.

3.1 The Hadley Circulation

3.1.1 Temperature

As discussed in our Project 3 introduction, the net radiative budget of the Earth-atmosphere system, averaged over the year, shows a net surplus of incoming radiation over the tropics and a net deficit at high latitudes. As a result of this the tropospheric temperature is higher in the tropics than at the poles, as should be evident from your Tplot.

- If the Earth were not rotating, the circulation driven by this temperature difference would be straightforward, with warm air rising in low latitudes and cold air sinking at high latitudes — see Fig.8. Can you see evidence of this in your climatological plots?

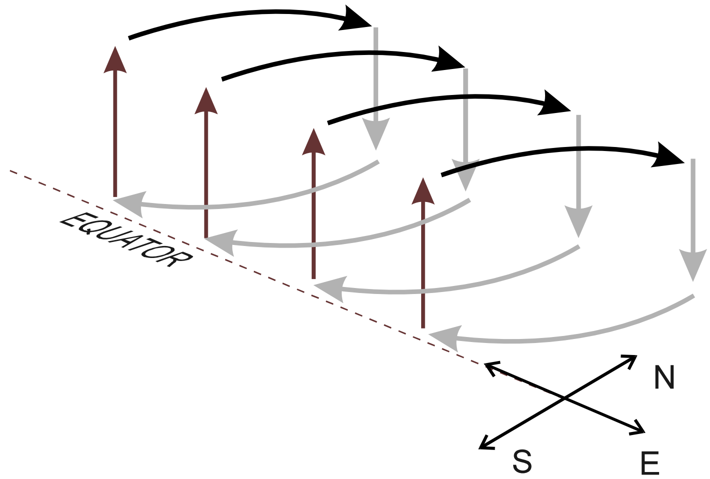

Figure 8: Schematic of the Hadley circulation (showing only the N Hem part of the circulation; there is a mirror image circulation south of the equator).

3.1.2 Meridional circulation

- From the meridional circulation (v, ω) can you observe a meridional circulation in the tropics? Note that the sinking motion is located in the subtropics rather than at high latitudes.

The circulation does not extend all the way to the pole because of angular momentum constraints due to Earth's rotation. Near the equator, however, where the Coriolis effect is small, the equatorial atmosphere acts as if the Earth is rotating slowly. How far the circulation extends polewards depends (according to theory) on many factors. From the observations we can see that it extends to roughly 30°N.

3.1.3 Zonal wind

- Inspect your zonal-average zonal wind plot and see if it is consistent with the following discussion.

Consider the upper branch of the Hadley circulation (see Fig.8). As air moves away from the equator, the Coriolis effect becomes increasingly large and, in the northern hemisphere, 'turns' the wind to the right, resulting in a westerly component to the flow. In the subtropics, air sinks (producing the desert zone and the "trade inversion" — see discussion in Chapter 8 of Marshall and Plumb) — and returns to the equator at low levels. At these low levels, the Coriolis acceleration, again turning the flow to the right in the northern hemisphere, induces easterly winds. The so-called "trade winds" and thus southeasterly in the northern hemisphere (and northeasterly in the southern hemisphere). These surface winds are not nearly as strong as in the upper troposphere, because they are damped by friction of the near-surface flow. In fact, there must also be low level westerlies somewhere because in equilibrium the net frictional drag (strictly, torque) on the entire atmosphere must be zero — if it were not the total angular momentum of the atmosphere would not be in balance. It turns out that the surface winds are westerly at the poleward edge of the circulation cell, and easterly near the equator. The state of affairs is shown in Fig.9, which is a repeat of Fig.4 in our introduction.

Schematic of the Hadley Circulation.

- Compare the schematic of the Hadley Circulation — Fig.9 — to the zonally averaged u,v and ω from your climatology and label your plots as in the schematic.

- Note that Fig.9 shows symmetry about the equator (annual-mean conditions). In reality, there are strong asymmetries that shift seasonally. Compare your January and July mean fields.

3.2 Meridional energy transport in the Hadley Circulation

Here we set out how one goes about computing the meridional energy flux achieved by the tropical Hadley Circulation from analyzed fields of atmospheric observations. Our approach directly parallels that used to estimate radial energy flux in our laboratory experiment.

We estimate the meridional heat flux by the mean meridional circulation

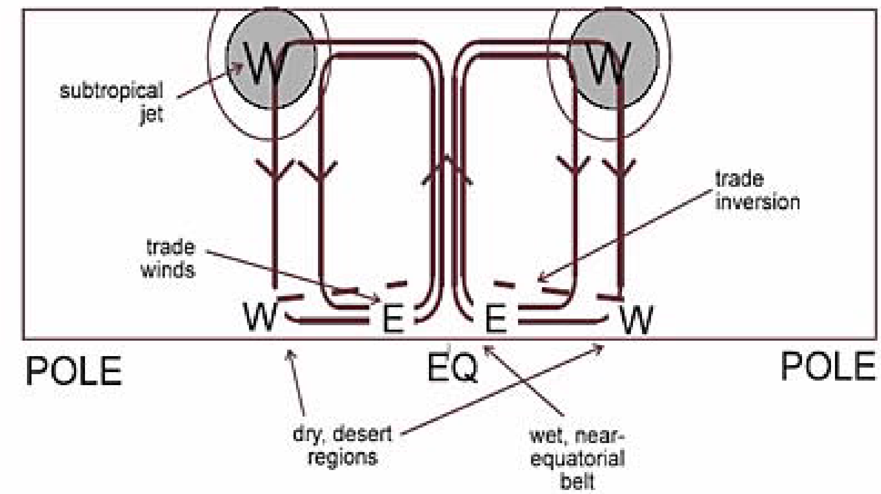

Figure 9: A schematic of the Hadley Circulation and its associated zonal flow and surface circulation. Westerlies are marked W; easterlies E.

in the atmosphere thus[1]:

$$\mathcal{H} = \rho c_p \int_0^\infty \oint \overline{v}\overline{\vartheta} dx dz = \frac{c_p}{g} \int_0^{p_s} \oint \overline{v}\overline{\vartheta} dx dp \tag{7}$$

$$= \frac{c_p}{g} \times 2\pi a \cos\varphi \int_0^{p_s} \left[\overline{v}\overline{\vartheta}\right] dp \tag{8}$$

where $\overline{v}(x,\varphi,p)$, $\overline{\vartheta}(x,\varphi,p)$ are the time-mean meridional velocity as a function of longitude, latitude and pressure, z increases upwards, x points eastward along a latitude circle, a is the radius of the Earth and φ is the latitude (in radians). In Eq.(7) we have used the hydrostatic balance to convert the vertical integral over height to one over pressure, $\rho dz = -dp/g$, and flipped the range of the vertical integral in pressure to $0 \longrightarrow p_s$ where 0 represents zero pressure at the top of the atmosphere and p_s is the surface pressure

[1] Note that in Eq.(7) we are using the 'potential temperature', ϑ, rather than ordinary temperature, T, to take in to account that air, unlike water, is a compressible fluid. See notes on 'potential temperature'.

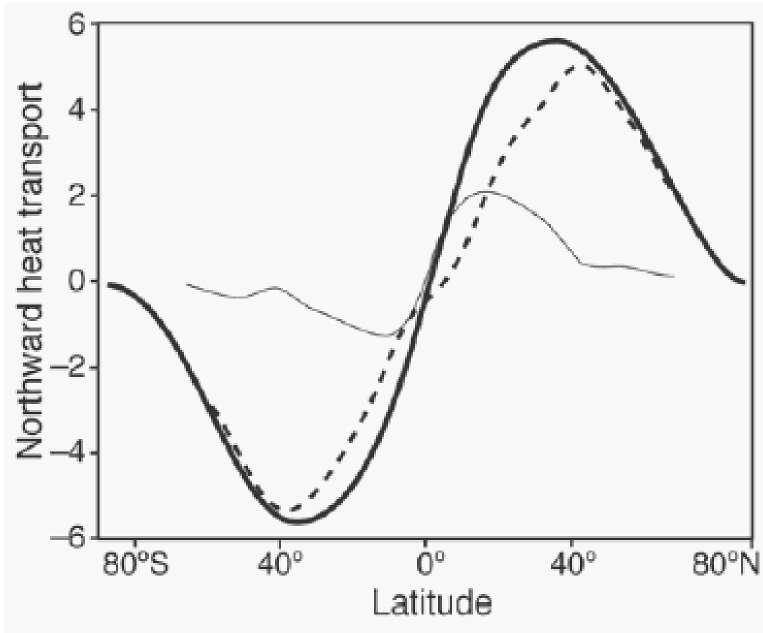

Figure 10: The ocean (thin) and atmosphere (dotted) contributions to the total northward heat flux based on NCEP Reanalysis – due to Trenberth and Caron (2001). The units of heat transport are in $PW(10^{15}W)$.

($p_s = 1000mb = 10^5\,\mathrm{Pa}$). In Eq.(8) the geometrical factor is the distance along a latitude circle at latitude φ and [ ] indicates a zonal average.

Eq.(7) is in an exact analogue of the expression used in the Hadley Circulation laboratory experiment in Section 2.3.

3.2.1 Two-layer model

Here we calculate the heatflux $\mathcal{H}$ achieved by the Hadley Circulation making use of Eq.(8) but approximating it with a two layer model, which closely parallels that used in the Hadley Circulation laboratory experiment.

On inspecting the zonally averaged meridional velocity $[v]$ and $[\theta]$ plots for January — produced using Esglobe in the troposphere up to 100mb — see Fig.11, we see that $[\theta]$ exhibits a strong north-south temperature gradients in the extra-tropics but little horizontal variation in the tropics. This is a consequence of efficient lateral mixing of temperature by the Hadley

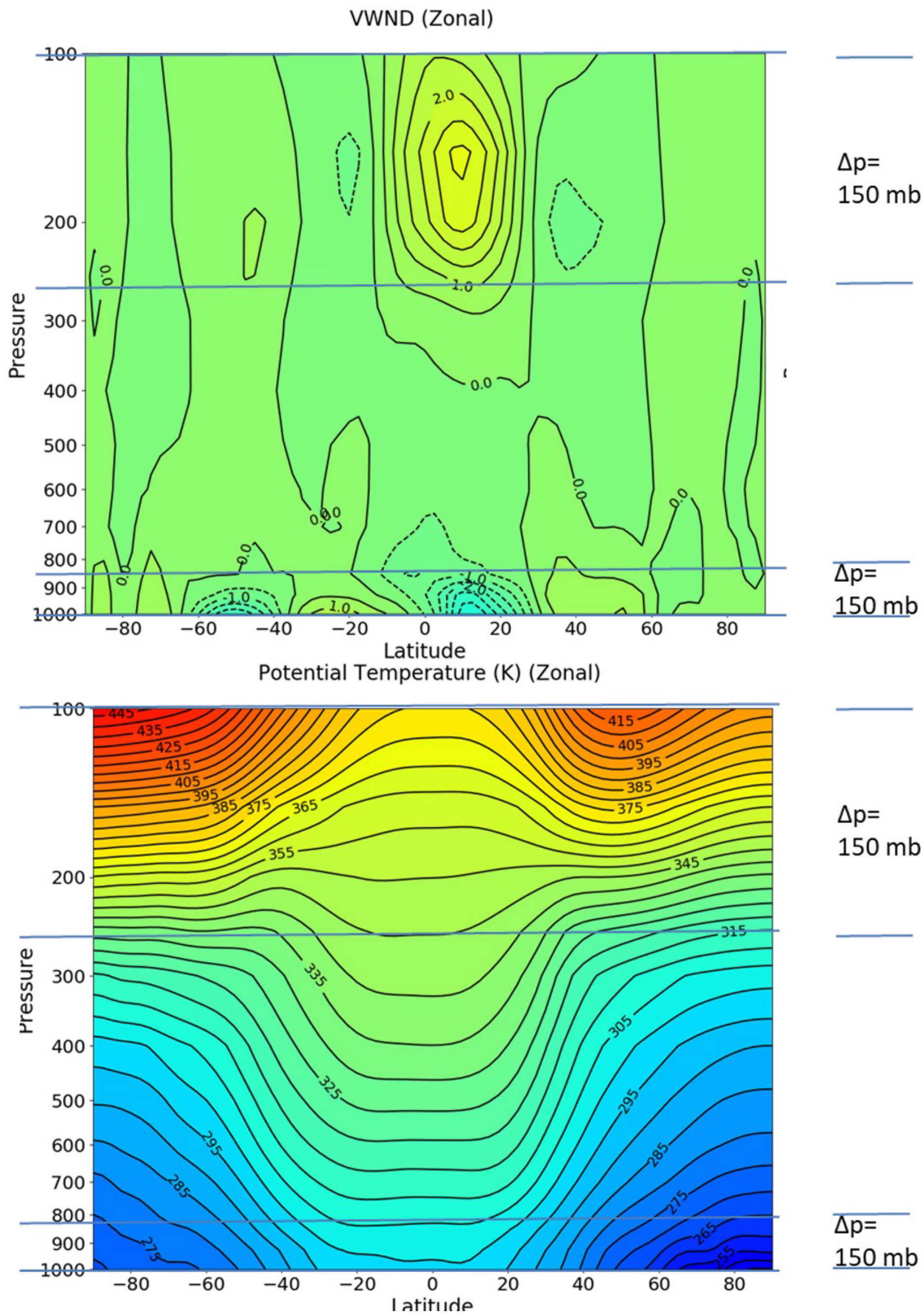

Figure 11: Zonal mean section of January-mean meridional wind $\overline{v}$ (top, in m/s) and potential temperature $\overline{\theta}$ (bottom, in K).

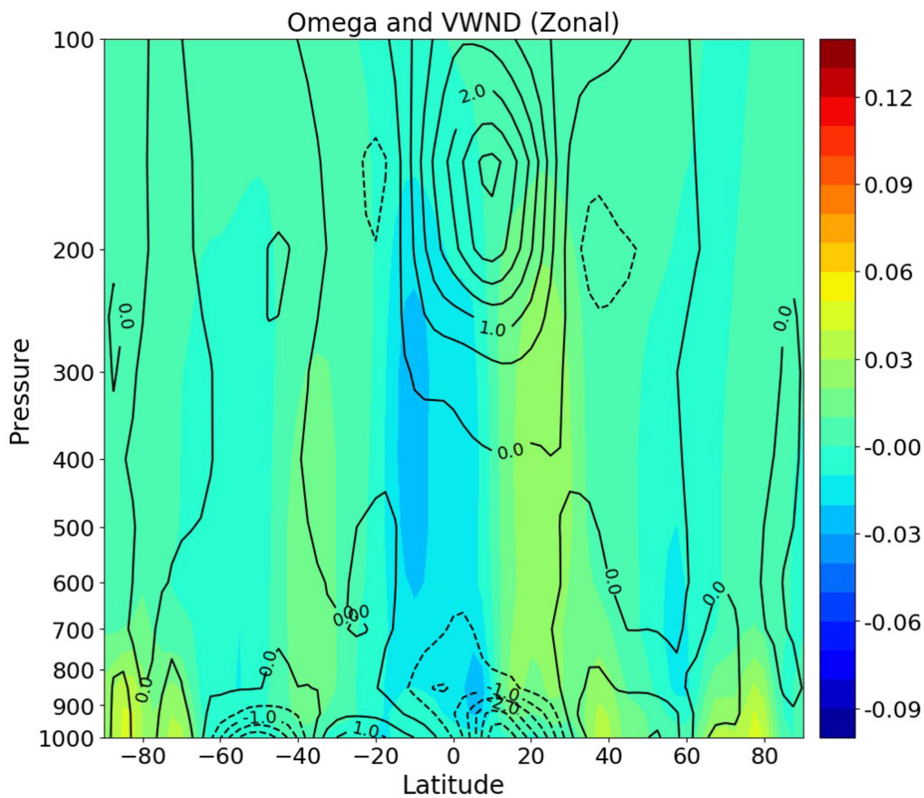

Figure 12: Zonal mean section of January mean meridional wind, $\overline{v}$ (contoured) and vertical velocity, $\overline{\omega}$ (colored). Note that $\omega < 0$ implies upward motion.

circulation. Note also that the meridional velocity v is rather small nearly everywhere except in the tropics: it is directed strongly northward in the upper troposphere centered at 150 mb and strongly southward close to the surface. In the middle regions of the troposphere $[\overline{v}]$ is very small. Fig.12 superimposes $\overline{\omega}$ and $\overline{v}$ enabling us to visualize the Hadley overturning circulation: we see one giant cell with rising air south of the equator and sinking air north of the equator.

It seems reasonable, therefore, to identify an upper and a lower layer, marked by the horizontal blue lines in Fig.11, within which the largest $[\overline{v}]$ occurs and outside of which we assume $[\overline{v}] = 0$. We can thus approximate the vertical integral in Eq.7 by:

$$\int_0^{p_s} [\overline{v}][\overline{\vartheta}]dp = ([\overline{v}]_t[\overline{\vartheta}]_t + [\overline{v}]_b[\overline{\vartheta}]_b)\Delta p = [\overline{v}]_t([\overline{\vartheta}]_t - [\overline{\vartheta}]_b)\Delta p, \tag{9}$$

where we have reasonably assumed (by inspection of Fig.11) that $[\overline{v}]_t = -[\overline{v}]_b$ and set the $\Delta p's$ of the two layers to the same value, chosen to be 150mb.

- Using Fig.11 estimate the typical $[\overline{v}]_t$, $[\overline{v}]_b$, $[\overline{\vartheta}]_t$, $[\overline{\vartheta}]_b$, Δp etc, to estimate the meridional energy flux. You will need to know that: c_p for air is $1005\,\mathrm{kg}^{-1}\,\mathrm{K}^{-1}$, the radius of the Earth is $6.3 \times 10^6\,\mathrm{m}$ etc.... Express your answer in PW$=10^{15}$ W.

You should note that your estimate is for the month of January when we expect large meridional transport from the warm (southern) hemisphere to the cold (northern) hemisphere.

- Compare and contrast your estimate with the requirement (from the Earth radiation budget) that in the annual mean the net (atmosphere and ocean) heat transport is about $2PW$ in tropical region where $1PW = 10^{15}W$ — see Fig.10. Note that Fig.10 is a plot of the total poleward energy transport including both sensible and latent heat components.

Our calculation thus far only involves the sensible heat transfer and does not include the meridional latent heat flux, which we now consider.

3.3 Role of moisture in the lateral energy transport

To compute the total energy flux we must add to Eq.(7) the latent heat flux contribution:

$$\begin{aligned}\mathcal{L} &= \frac{L}{g} \times 2\pi a \cos\varphi \int_0^{p_s} \overline{vq}\,dp \qquad (10) \\ &\simeq \frac{L}{g} \times 2\pi a \cos\varphi [\overline{v}]_b [\overline{q}]_b \Delta p.\end{aligned}$$

where we have simplified the vertical integral using our two layer model. Here $L = 2.25 \times 10^6\,\mathrm{J\,kg^{-1}}$ is the latent heat of fusion and q is the specific humidity in $\mathrm{g\,kg^{-1}}$.

From Fig.13 we see that the specific humidity, q, decreases rapidly with height, and so the equatorward flowing air in the lower troposphere carries more moisture than the poleward moving branch aloft: since $[\overline{v}]_b < 0$, $\mathcal{L}$ is directed southward, offsetting the northward sensible heat flux.

- Insert typical numbers to Eq.(10) to obtain and estimate of the latent heat flux contribution to the total energy flux. What do you notice? What is the total heat flux, sensible plus latent? How does it compare with Fig.10?

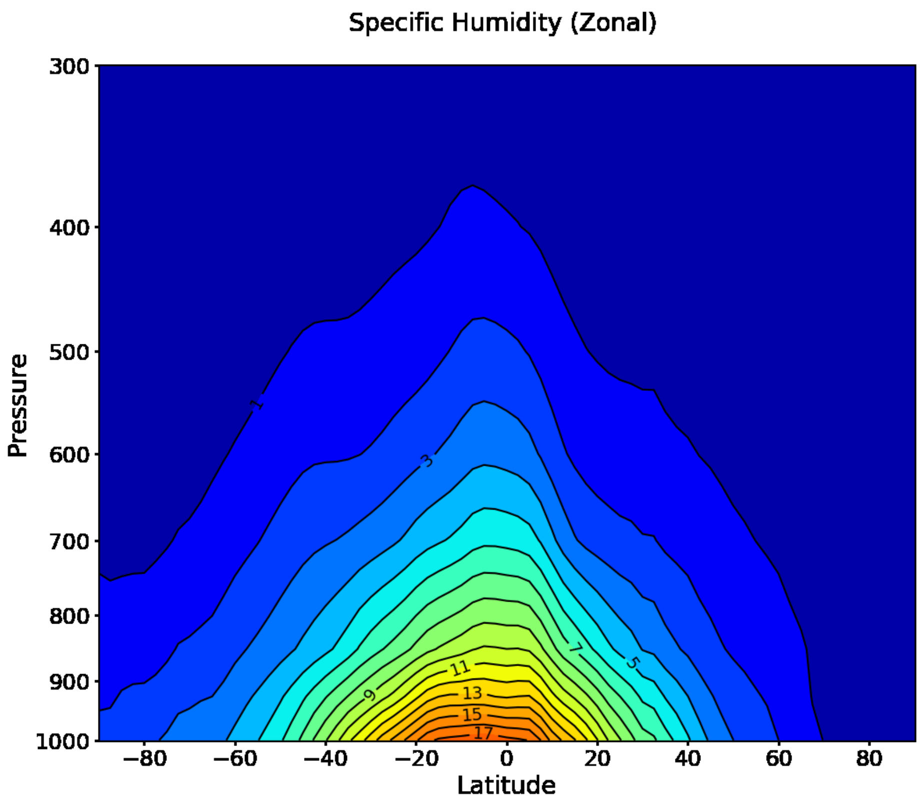

Figure 13: January mean specfic humidity $\overline{q}$ in $\mathrm{g\,kg^{-1}}$. Note that q is only available up to 300 mb, It is so small in the upper troposhere that it cannot be measured.