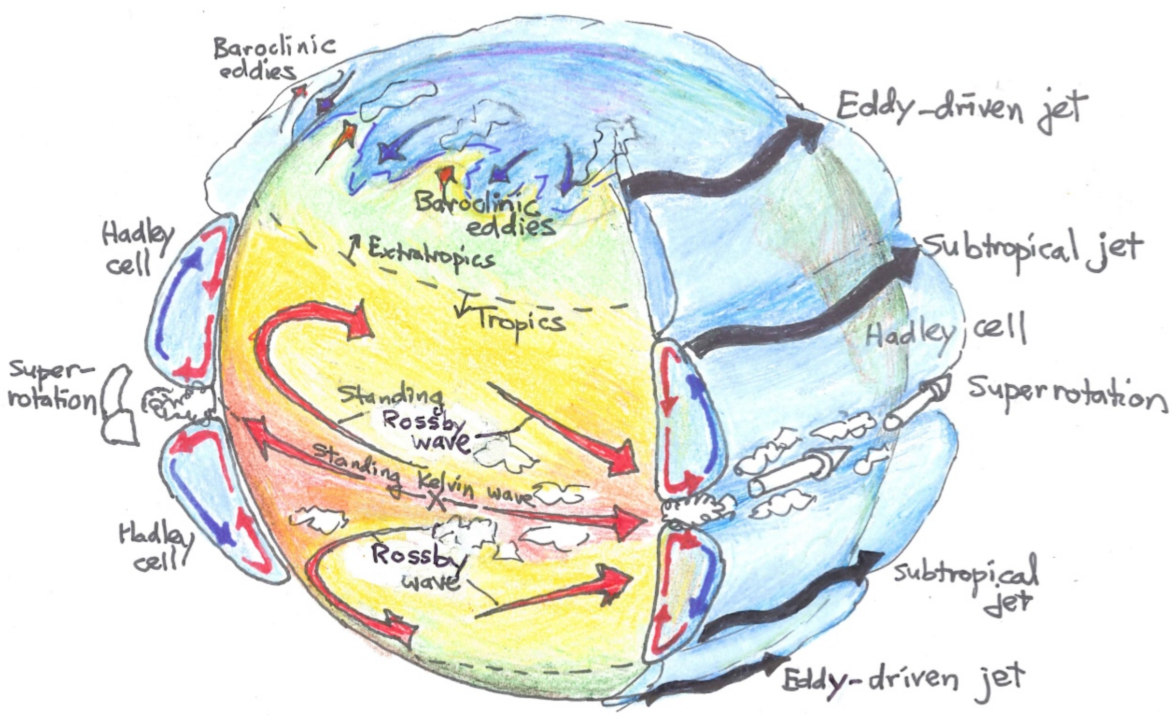
Contents

Chapter 1

Dynamical basics, part 1

Here we survey the basic dynamical regimes and mechanisms that occur in planetary atmospheres. Before presenting any theory, let's describe qualitatively the kind of regimes that exist. This shows the atmospheric circulation of a generic terrestrial exoplanet:



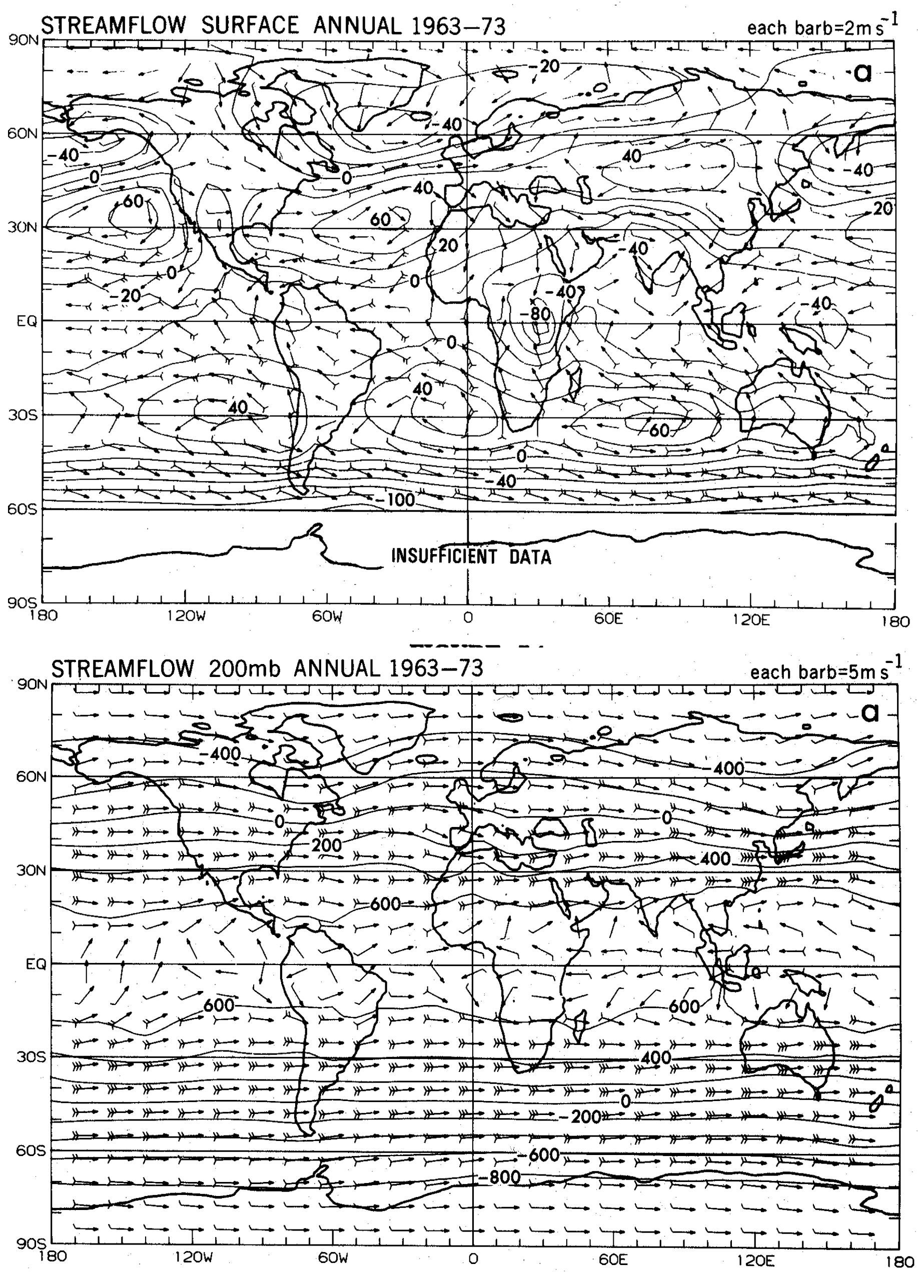
Showman et al. (2013, "Atmospheric circulation of terrestrial exoplanets," in *Comparative Climatology of Terrestrial Planets*, Univ. Arizona Press)

There are two basic regimes:

- **Tropics:** In the tropics is the Hadley circulation, whereby air rises near the equator, moves poleward in the upper troposphere, sinks at ∼20–30° latitude, and returns to the equator near the surface. This is an efficient means of transporting heat and contributes to the relative constant temperatures with latitude in the tropics. However, Earth's rotation prohibits the Hadley cell from extending to the poles. The poleward movement of air in the upper branch of the Hadley cell causes an eastward deflection due to the Coriolis force, which leads to the so-called *subtropical jets.*
- **Extratropics:** Poleward of ∼30°, dynamical instabilities lead to warm

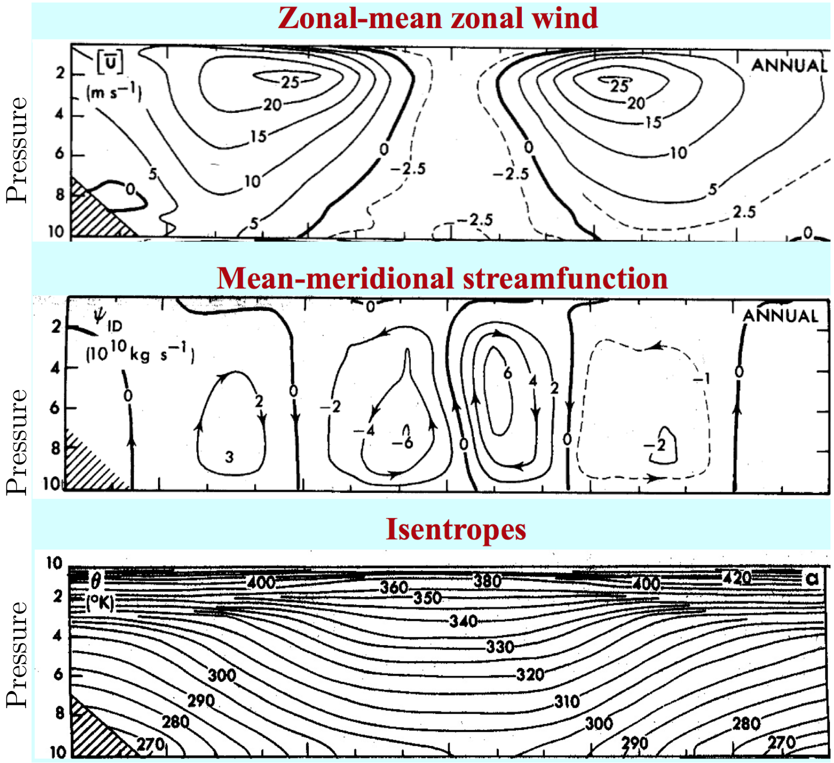
and cold eddies (∼1000-3000 km across in the case of Earth) that transport heat poleward. This is called the *baroclinic zone* and the instability responsible for producing these eddies is called *baroclinic instability.* This is not such an efficient heat-transport mechanism and so the latitudinal temperature gradient is larger in midlatitudes. The eddies converge momentum into their latitude region, generating the so-called *eddy-driven jet.*

Let's look at a few observations for Earth. This shows the annual-mean surface and 200 mbar (∼tropopause) winds. Note weak equatorial winds, moving slightly westward at the surface. Midlatitude winds are eastward and get very strong in the upper troposphere. This is the signature of **jet streams**.



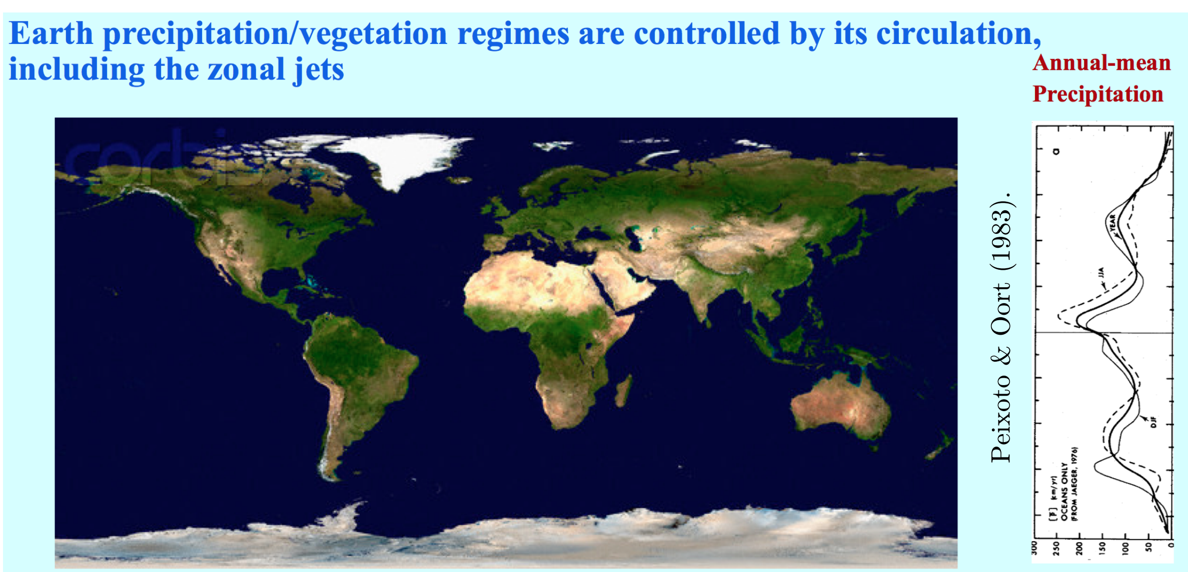
Peixoto & Oort (1992)

This shows vertical structure of east-west ("zonal") wind, the streamfunction of the so-called meridional circulation (i.e., the circulation in the latitude-height plane), and the structure of the longitudinal-mean isentropes:

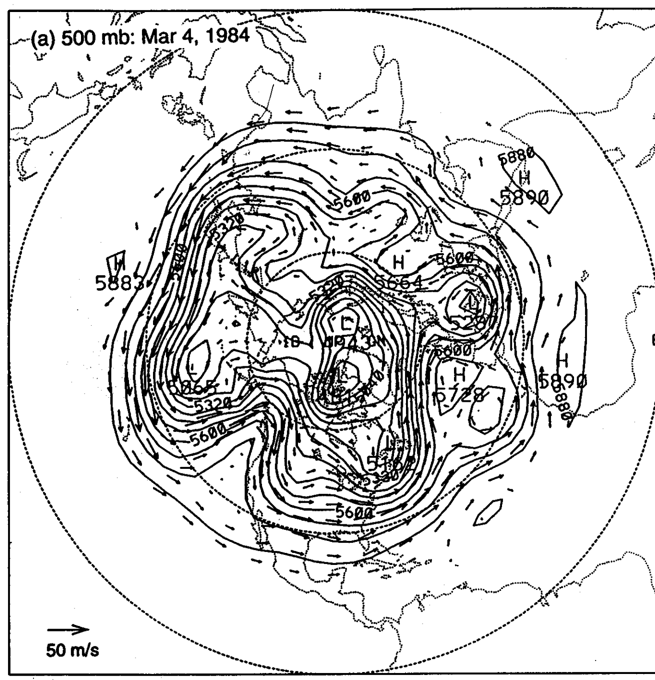


From Peixoto & Oort (1992).

This circulation pattern controls the geographic distribution of vegetation. The abundant rainfall associated with the rising branch of the Hadley cell leads to the tropical rainforests (Brazil, central Africa, southeast Asia). The dearth of rainfall associated with the descending branch of the Hadley cell leads to the abundant deserts at ∼20–30° latitude (the Sahara, central Asia, the American Southwest, Australia, and South Africa). The midlatitude peak in cloudiness and precipitation associated with the baroclinic zone leads to abundant temperate forests in midlatitudes.

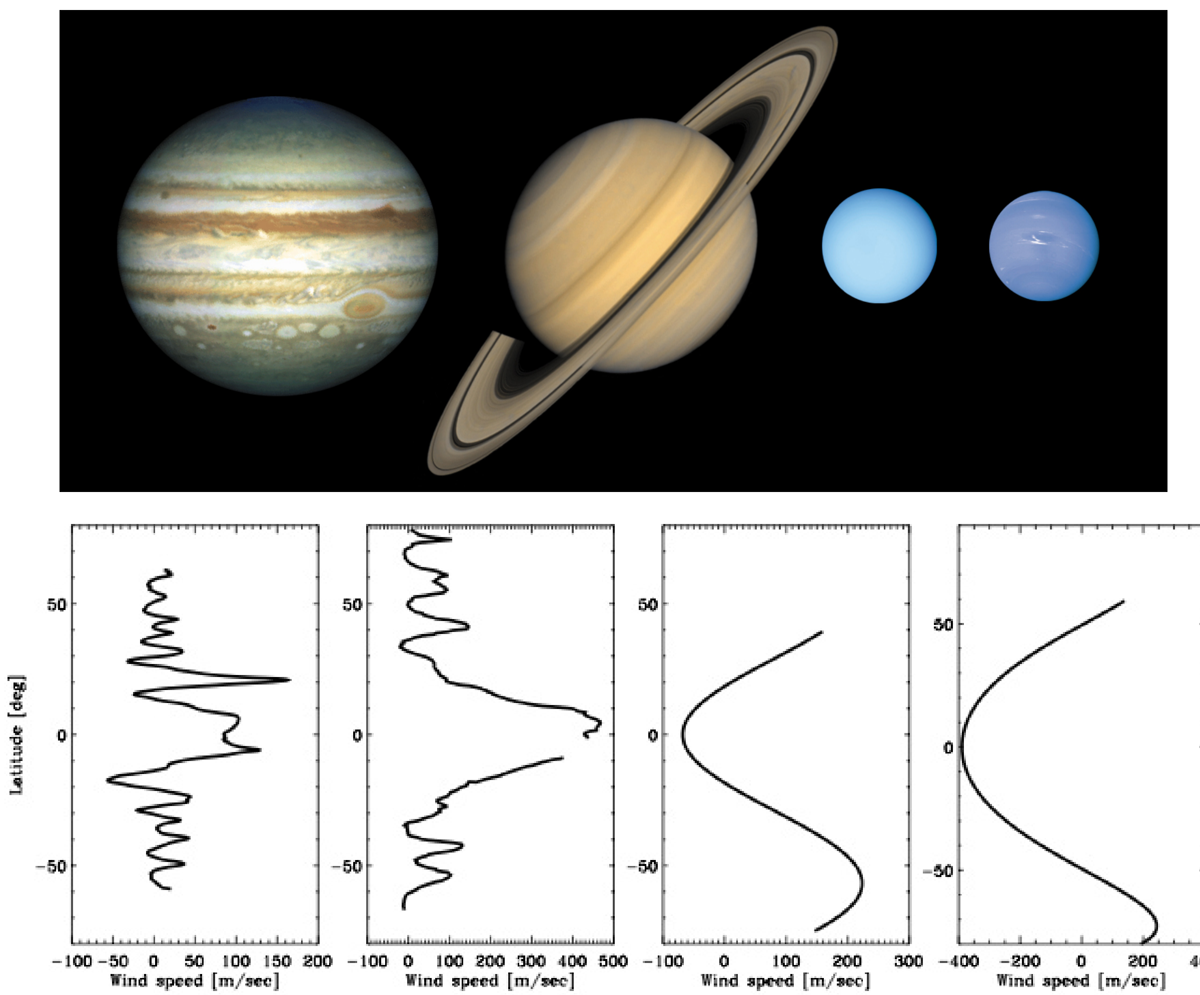


Poleward heat transport in the baroclinic zone involves continual growth and decay of eddies. These eddies cause the fronts and multi-thousand-km-long arc-shaped clouds seen in weather maps and satellite images at high latitudes and cause most winter weather in the U.S., Europe, and other midlatitude locations. They also cause the meanders on the jet stream seen in the geopotential heights of the 500-mbar surface:



Salby (1996)

Now let's look at the circulation of the giant planets.



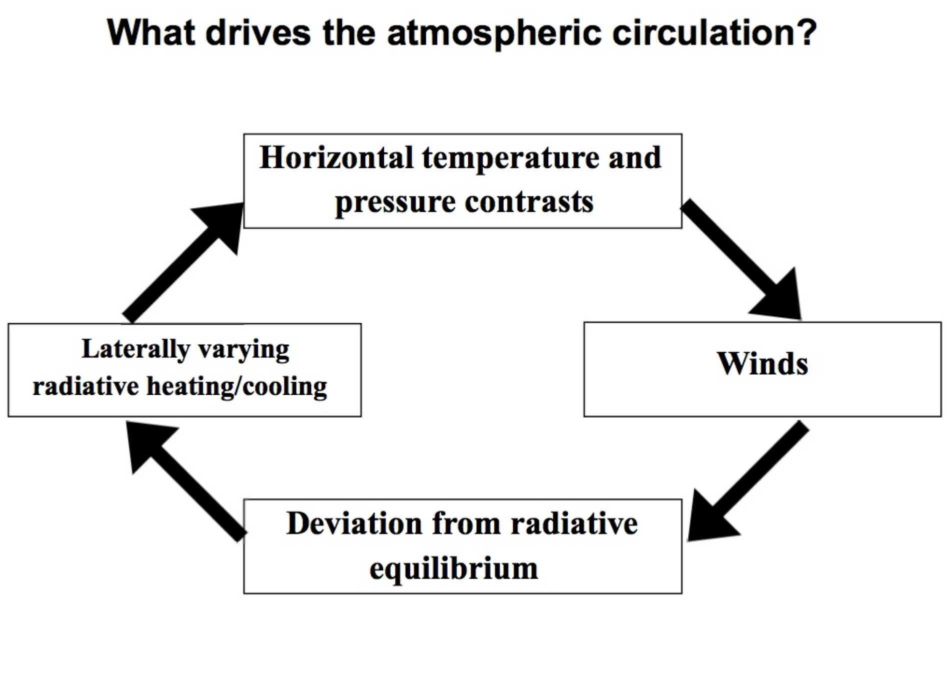
Showman, Menou & Cho (2010).

Jupiter, Saturn, Uranus, and Neptune have radii of 11, 9, 4, and 4 times that of Earth. They are completely fluid planets with no solid surfaces and thus challenge our understanding of how atmospheric circulation should behave in the absence of a surface. They are also rapid rotators, with rotation periods ranging from 10 hours on Jupiter to 17 hours for Uranus. Cloud tracking shows that the circulation is dominated by zonal jets with speeds peaking at $\sim 200\,\mathrm{m\,sec^{-1}}$ on Jupiter and $\sim 400\,\mathrm{m\,sec^{-1}}$ or more on Saturn and Neptune. These clouds are at pressures of ~ 1 bar within a factor of a few. How winds can be so fast despite the weak solar forcing is not well understood.

1.1 What drives the atmospheric circulation?

Atmospheric circulation is fundamentally a coupled radiation-dynamics problem. In the absence of circulation, the vertical temperature profiles would relax into a local radiative equilibrium, which would be very hot on the dayside and cold at the poles/nightside. The radiative heating rate would everywhere be zero. When a circulation operates, however, horizontal temperature contrasts imply horizontal pressure contrasts, which drive winds. The winds in turn drive the atmosphere away from radiative equilibrium by transporting heat from hot regions to cold regions (e.g., from the equator to the poles on Earth). This deviation from radiative equilibrium allows net radiative heating and cooling to occur, thus maintaining the horizontal temperature and pressure contrasts that drive the winds. *Spatial contrasts in thermodynamic heating/cooling therefore drive the circulation, yet it is the existence of the circulation that gives rise to these heating/cooling patterns in the first place.*

The coupled nature of the problem can be illustrated with this flow chart, which represents an infinite loop of cause-and-effect:[1]



[1]See Showman, Menou & Cho (2010), "Atmospheric circulation of exoplanets," in *Exoplanets* [S. Seager, Ed.], Univ. Arizona Press; online at arXiv 0911.3170.

On the Earth, for example, the equator and poles are not in radiative equilibrium. The equator is subject to net heating, the poles to net cooling, and it is the mean latitudinal heat transport that is both responsible for and driven by these net imbalances.

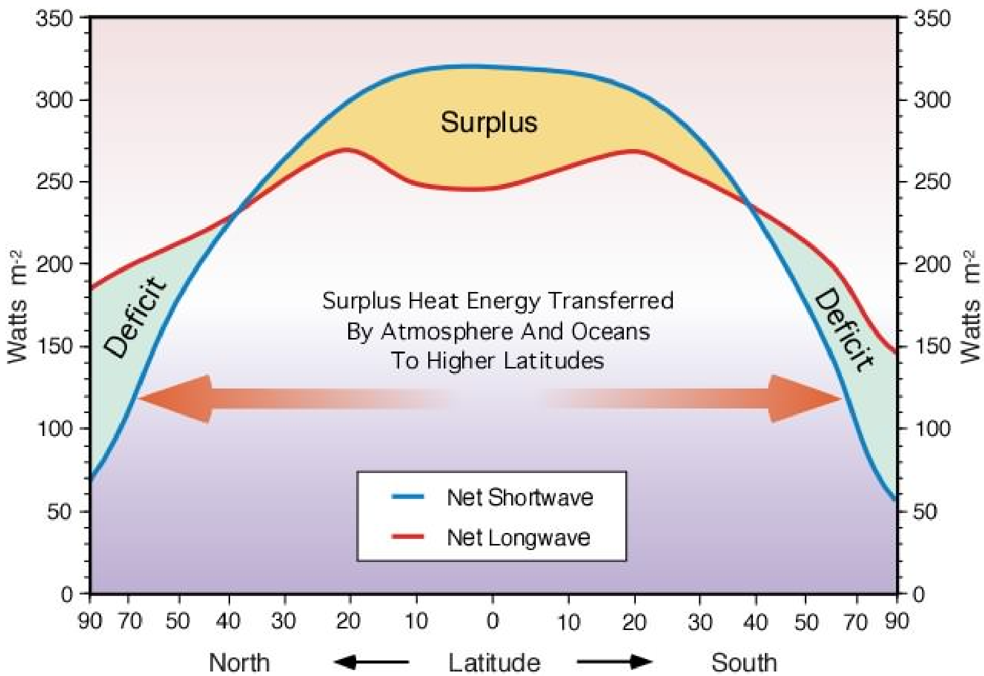


Figure 1.1: Earth's energy balance vs latitude, averaged over the year. Blue is absorbed sunlight and red is OLR. Because the circulation mutes the latitudinal temperature contrasts, the OLR exhibits weaker latitude variation than absorbed sunlight, and this is what allows the net heating at the equator and net cooling at the poles—which in turn drives the circulation. Data are an annual-average for 1987 obtained from the NASA Earth Radiation Budget Experiment (ERBE) project. (Source: M. Pidwirny, www.physicalgeography.net.)

1.2 Conservation laws

Atmospheric circulation is fundamentally controlled by the laws governing the conservation of momentum, mass, and energy.

Conservation of momentum: This is essentially $F = ma$ applied to a fluid. The acceleration of a fluid parcel, which is just the rate of change of the parcel's velocity with time, $\frac{d\mathbf{v}}{dt}$, is the sum of the forces per mass that act on the parcel:

$$\boxed{\frac{d\mathbf{v}}{dt} = -\frac{1}{\rho}\nabla p + \mathbf{g} - 2\Omega \times \mathbf{v} + \mathcal{F}}$$

where p is pressure, ρ is density, $\mathbf{g}$ is gravity, Ω is rotation rate, $\mathbf{v}$ is velocity, and $\mathcal{F}$ is friction. On the right-side, the forces that act to accelerate the parcel are the pressure-gradient force, gravity, Coriolis force, and friction. The Coriolis force acts perpendicularly to both the planetary rotation vector, Ω, and to the velocity, $\mathbf{v}$. An important rule to remember is that, in the horizontal force balance, the Coriolis force acts to the *right* of the fluid motion in the northern hemisphere and to the *left* of the fluid motion in the southern hemisphere.[2] [3] Friction is generally small except within a few cm of a planetary surface. However, small-scale turbulent mixing can cause accelerations of the large-scale flow that in some cases exhibit qualitative similarities to friction. In models that do not explicitly resolve such small-scale turbulence, this is often represented with a frictional term.

In a Cartesian frame, where x is eastward, y northward, and z upward, the momentum equations are

$$\boxed{\begin{aligned}\frac{du}{dt} &= -\frac{1}{\rho}\frac{\partial p}{\partial x} + 2\Omega v \sin\phi - 2\Omega w \cos\phi + \mathcal{F}_x \\ \frac{dv}{dt} &= -\frac{1}{\rho}\frac{\partial p}{\partial y} - 2\Omega u \sin\phi + \mathcal{F}_y \\ \frac{dw}{dt} &= -\frac{1}{\rho}\frac{\partial p}{\partial z} - g + 2\Omega u \cos\phi + \mathcal{F}_z\end{aligned}}$$

where u, v, and w are the eastward, northward, and upward velocity components, ϕ is latitude, and $\mathcal{F}_x$, $\mathcal{F}_y$, and $\mathcal{F}_z$ are the eastward, northward, and vertical frictional forces.

Conservation of mass: Also called the *continuity equation,* this is

$$\boxed{\frac{1}{\rho}\frac{d\rho}{dt} + \nabla\cdot\mathbf{v} = 0}$$

which simply states that fractional changes in the volume of a fluid parcel (i.e. nonzero divergence) must be accompanied by changes in the density. For order of magnitude considerations, it often suffices to treat the fluid as incompressible, leading to

$$\boxed{\nabla\cdot\mathbf{v} = 0}$$

Conservation of thermal energy: This is just the first law of thermodynamics, which is

$$\boxed{c_p\frac{dT}{dt} = Q + \frac{1}{\rho}\frac{dp}{dt}}$$

[2]Northern and southern hemispheres are here defined as those where $\Omega\cdot\mathbf{k} > 0$ and < 0, respectively, where $\mathbf{k}$ is the local vertical (upward) unit vector.

[3]On a spherical planet, the Coriolis force can intuitively be understood as representing two physical effects. When a fluid parcel moves poleward (equatorward), conservation of angular momentum about the planetary rotation axis tends to deflect the parcel eastward (westward). When a fluid parcel moves eastward (westward) relative to the planetary surface, there is an acceleration away from (toward) the rotation axis. It can be seen that, in both cases, the force is to the right (left) of the motion in the northern (southern) hemisphere.

or alternately

$$\boxed{c_p \frac{d\theta}{dt} = \frac{\theta}{T} Q}$$

where θ is potential temperature, $\theta = T(p_0/p)^{R/c_p}$, where p_0 is a reference pressure. Q is thermodynamic heating due to radiation, conduction, latent heating, etc.

Together, this set constitutes five equations (the east-west, north-south, and vertical momentum equations, continuity equation, and thermodynamic energy equation) in six unknowns: the three velocity components (u, v, w), ρ, p, and T (or alternately θ). The system is closed by specifying an equation of state, typically ideal gas.

In all of the above equations, the time derivatives on the left-hand sides give the time rate of change following an individual air parcel as it moves around. This is called the **material** or **total derivative** and, for any scalar quantity ψ, is given by

$$\boxed{\frac{d\psi}{dt} = \frac{\partial \psi}{\partial t} + \mathbf{v} \cdot \nabla \psi}$$

The term $\mathbf{v} \cdot \nabla \psi$ is called the *advection term*, because it represents the advection (transport) of ψ by the flow.

1.3 Large-scale circulation: Basic force balances

We now consider some basic aspects of the large-scale circulation. In this section we consider the dominant force balances, in the vertical and horizontal momentum equations, for typical large-scale circulations. We do this by estimating the magnitudes of the terms in the equations, allowing small terms to be thrown out and exposing the dominant balances.

1.3.1 Vertical momentum balance: hydostaticity

Atmospheres are shallow, with horizontal scales greatly exceeding vertical ones. Consider the magnitudes of terms in the vertical momentum equation for a typical terrestrial planet. Let U, W, L, H, τ, and P be the characteristic magnitudes of horizontal and vertical velocity, horizontal and vertical length scale, evolution time scale for the circulation, and pressure change across vertical scale H. Terms in the vertical momentum equations have magnitudes

$$
\begin{array}{ccccc}
\dfrac{\partial w}{\partial t} + \mathbf{v}\cdot\nabla w = & -\dfrac{1}{\rho}\dfrac{\partial p}{\partial z} & - g & + 2\Omega u\cos\phi \\
\dfrac{W}{\tau} \quad \dfrac{UW}{L} & \dfrac{P}{\rho H} & g & \Omega U \\
10^{-7} \quad 10^{-7} & 10 & 10 & 10^{-3}
\end{array}
$$

where the numerical estimates are in $\mathrm{m\,sec^{-2}}$ and use $H \sim 10\,\mathrm{km}$, $L \sim 10^3\,\mathrm{km}$, $\tau \sim 10^5\,\mathrm{sec}$, $U \sim 10\,\mathrm{m\,sec^{-1}}$, $W \sim 10^{-2}\,\mathrm{m\,sec^{-1}}$, and $P \sim 1\,\mathrm{bar}$, relevant to Earth.

Clearly, hydrostatic balance dominates. But much of this is a static balance that has nothing to do with meteorological motions. What we want to do is subtract off the static, background state and look at the balances associated solely with meteorological motions. Define $p = p_0(z) + p'$ and $\rho = \rho_0(z) + \rho'$, where $p_0(z)$ and $\rho_0(z)$ are the time-independent basic-state pressure and density and, by construction, $\frac{\partial p_0}{\partial z} \equiv -\rho_0 g$. Primed quantities are the deviations from this basic state caused by dynamics. Substituting these expressions into the momentum equation, we obtain

$$
\begin{array}{cccc}
\dfrac{\partial w}{\partial t} + \mathbf{v}\cdot\nabla w = & -\dfrac{1}{\rho}\left(\dfrac{\partial p'}{\partial z} + \rho' g\right) & & + 2\Omega u\cos\phi \\
\dfrac{W}{\tau} \quad \dfrac{UW}{L} & \dfrac{\Delta P}{\rho H} & \dfrac{\Delta\rho}{\rho} g & \Omega U \\
10^{-7} \quad 10^{-7} & 0.1 & 0.1 & 10^{-3}
\end{array}
$$

where the basic-state hydrostatic balance has been subtracted off. The terms in parentheses give only the flow-induced contributions to the vertical pressure gradient and weight (they go to zero in a static atmosphere). In the order-of-magnitude estimate, $\Delta p \sim 1000\,\mathrm{Pa}$ and $\frac{\Delta\rho}{\rho} \sim 10^{-2}$ are the characteristic pressure and fractional density perturbations associated with midlatitude large-scale meteorology on Earth, and the numerical values are again in $\mathrm{m\,sec^{-2}}$. The terms in the parentheses on the right side still dominate by several orders of magnitude. Thus, even after we cancel out the mean (hydrostatic) pressure gradient and gravity, we find that the meteorologically induced *pressure perturbations* are in hydrostatic balance with the *density perturbations*:

$$\frac{\partial p'}{\partial z} + \rho' g \approx 0$$

Therefore, vertical accelerations are negligible in the force balance and, for the large-scale circulation, we can replace the vertical momentum equation with "local" hydrostatic balance:

$$\boxed{\frac{\partial p}{\partial z} = -\rho g}$$

This analysis can be formalized, and doing so shows that local hydrostatic balance is a good assumption provided that (i) $L^2/H^2 \gg 1$, and (ii) the fluid is stably stratified (see Vallis 2006, *Atmospheric and oceanic fluid dynamics,* pp. 80-84). Note that this does *NOT* mean vertical motions do not occur: they still occur, but the accelerations are simply so small that we do not need to keep track of them in the vertical momentum equation.

1.3.2 Horizontal momentum balance on rapidly rotating planets: geostrophy and the importance of rotation

A similar scale analysis of the horizontal momentum equation can be performed as follows:

$$\begin{array}{cccc} \dfrac{\partial v}{\partial t} + & \mathbf{v}\cdot\nabla v = & -\dfrac{1}{\rho}\dfrac{\partial p}{\partial y} & - 2\Omega u \sin\phi \\ \dfrac{U}{\tau} & \dfrac{U^2}{L} & \dfrac{\Delta P}{\rho L} & \Omega U \\ 10^{-4} & 10^{-4} & 10^{-3} & 10^{-3} \end{array}$$

where again the numbers are $\mathrm{m\,sec}^{-2}$. The scaling analysis indicates that the time-derivative and advection terms are significantly smaller than the pressure gradient and Coriolis terms, which approximately balance. This can be formalized by defining the **Rossby number**, which is just the ratio of horizontal advection to Coriolis accelerations:

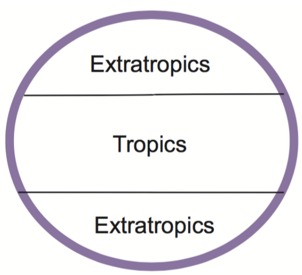
$$\boxed{Ro = \frac{U}{fL}}$$

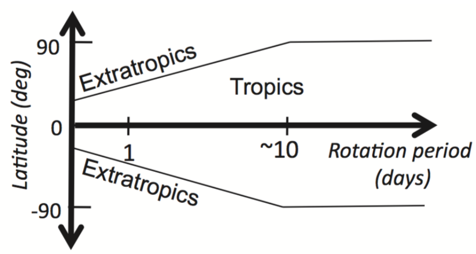
where $f \equiv 2\Omega \sin\phi$ is called the "Coriolis parameter." Thus, whenver $Ro \ll 1$, the dominant horizontal momentum balance is between pressure-gradient and Coriolis accelerations. This is called **geostrophic balance.**

The dynamics at $Ro << 1$ are significantly different from those at $Ro \gtrsim 1$. So we can use this to define regimes:

- **Extratropics:** Regime of $Ro \ll 1$. Rotationally dominated (geostrophically balanced to leading order). In the presence of latitudinally varying stellar forcing, can develop significant temperature differences and baroclinic instabilities.

- **Tropics:** Regime of $Ro \gtrsim 1$. Dynamics are ageostrophic, though rotation still usually plays a role. Hadley circulation and wave adjustment mechanisms tend to maintain small temperature differences.





This gives estimates for the midlatitudes:

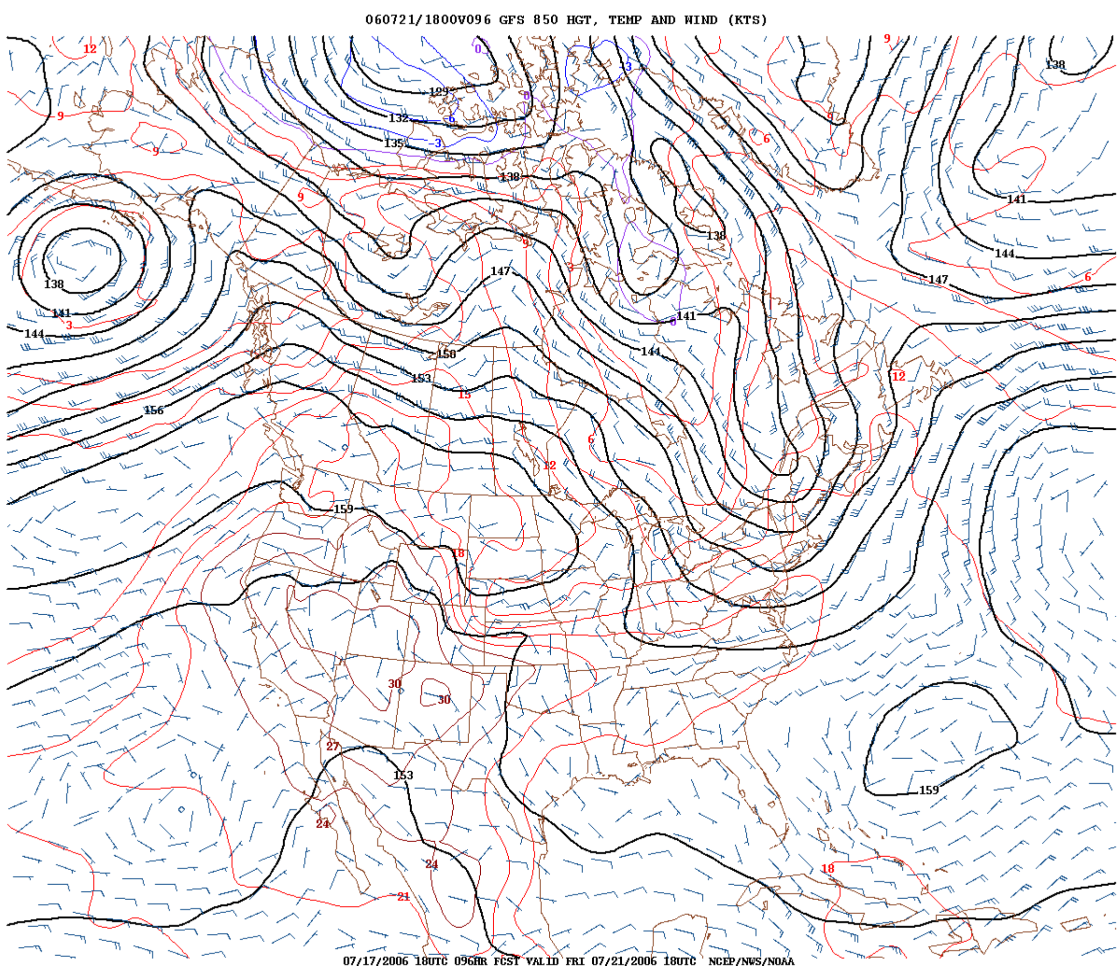
	U (m/sec)	Rotation period	Ω (sec^{-1})	L (km)	Ro
Venus	10	243 days	3×10^{-7}	3000	10
Earth	10	24 hours	7.3×10^{-5}	1000	0.1
Mars	10	24.6 hours	7.1×10^{-5}	1000	0.1
Titan	10	16 days	4.5×10^{-6}	1000	2
Jupiter	30	10 hours	1.7×10^{-4}	5000	0.02
Saturn	150	11 hours	1.65×10^{-4}	10^4	0.05
Uranus	100	17 hours	9.7×10^{-5}	10^4	0.1
Neptune	200	16 hours	1.1×10^{-4}	10^4	0.1
HD 189733b	1000?	2.2 days	3×10^{-5}	50,000?	0.5
Brown dwarf	30-300?	~3-5 hours	4×10^{-4}	10^4?	0.01–0.1
Earth-sized planet in habitable zone of M star	10?	30 days	2×10^{-6}	3000?	~2?

We can see that Earth, Mars, Jupiter, Saturn, Uranus, Neptune, and brown dwarfs exhibit geostrophic balance at mid-to-high latitudes, but of course $f \to 0$ at the equator so geostrophy breaks down there. Slowly rotating planets like Titan and Venus have $Ro \gtrsim 1$ across the whole planet, at least in the upper troposphere and above where winds are fast. Tidally locked hot Jupiters, due to faster wind speeds and modestly slower rotation than Earth, may have $Ro \sim 1$, while tidally locked terrestrial planets in the habitable zones of M dwarfs also have $Ro \gtrsim 1$ due to their slow rotation periods.

When geostrophy holds, we can write the approximate horizontal momentum balance as

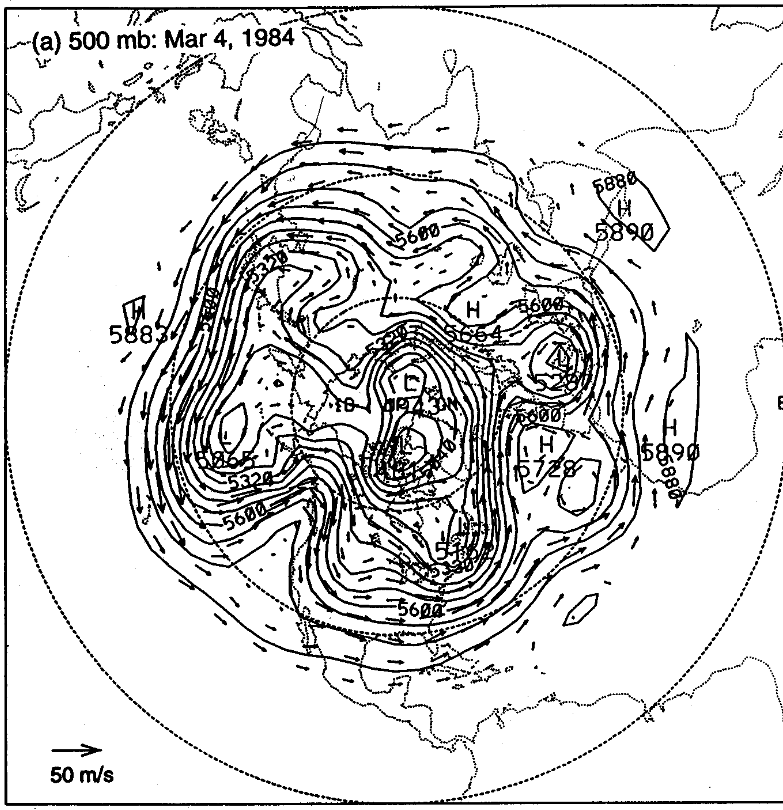
$$\boxed{\begin{aligned} fu &\approx -\frac{1}{\rho}\frac{\partial p}{\partial y} \\ fv &\approx \frac{1}{\rho}\frac{\partial p}{\partial x} \end{aligned}}$$

What does this mean? Geostrophy implies that the horizontal winds tend to flow parallel to the intersection of isobars with a horizontal plane. This is why the pressure field is so useful on a midlatitude weather map: the winds simply flow along isobars, so one knows the approximate wind field even if only the isobars, but not winds, are plotted. Importantly, this phenomenon differs drastically from common intuition, in which air flows from high to low pressure. In the midlatitudes of rapidly rotating planets, nature does not abhor a vacuum but rather organizes a cyclone around it.

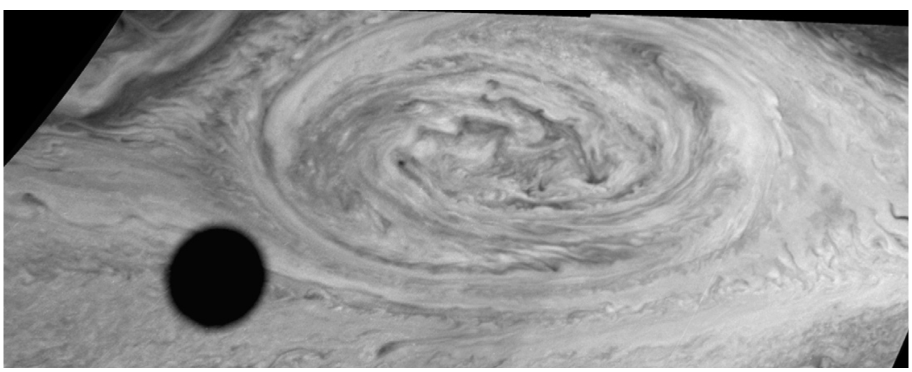


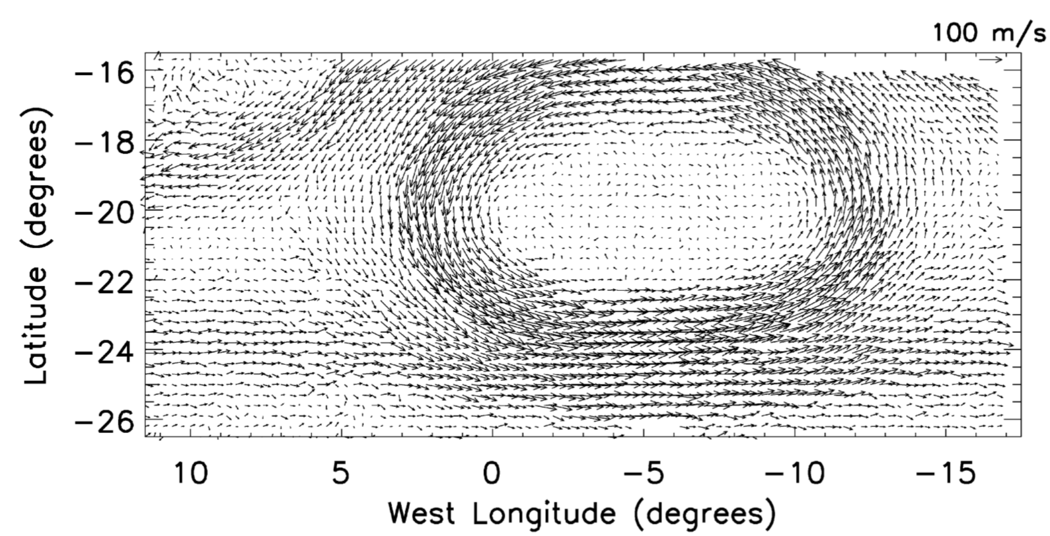
Isobars (black), isotherms (red) and winds (arrows) near 850 mbar level, in the bottom part of the troposphere above the boundary layer.

Another example, this one from the 500-mbar level. This shows the upper tropospheric jet stream and the baroclinic eddies (associated with baroclinic instabilities) as seen from over the north pole. For March 4, 1984 (Salby 1996).

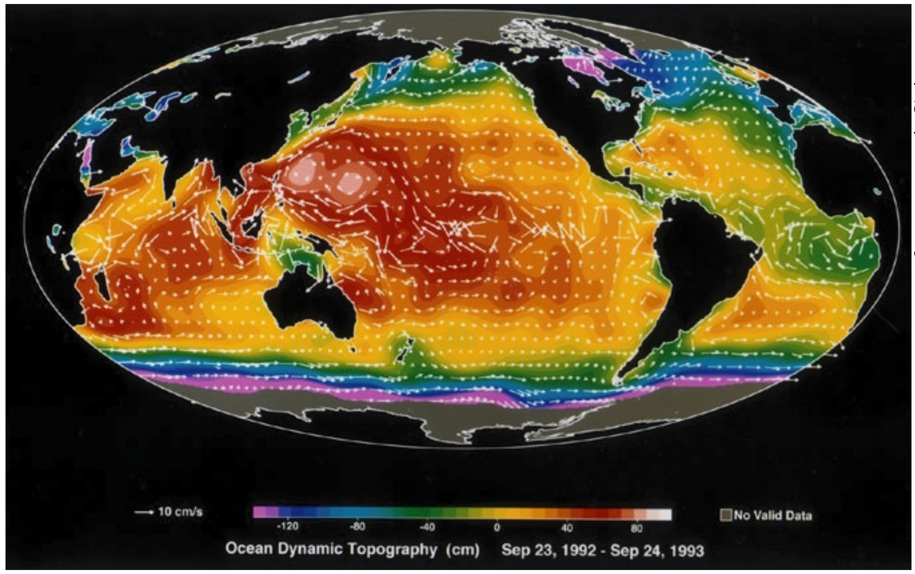


Jupiter's Great Red Spot, from Choi et al. (2007, *Icarus* **188,** 35-46).





Geostrophy also explains why currents generally flow parallel to the contours of time-mean ocean surface topography (when waves and tides are subtracted off): the horizontal pressure gradients just below the ocean surface are proportional to the topographic gradients. Satellite measurements of global ocean topography therefore determine the surface currents.

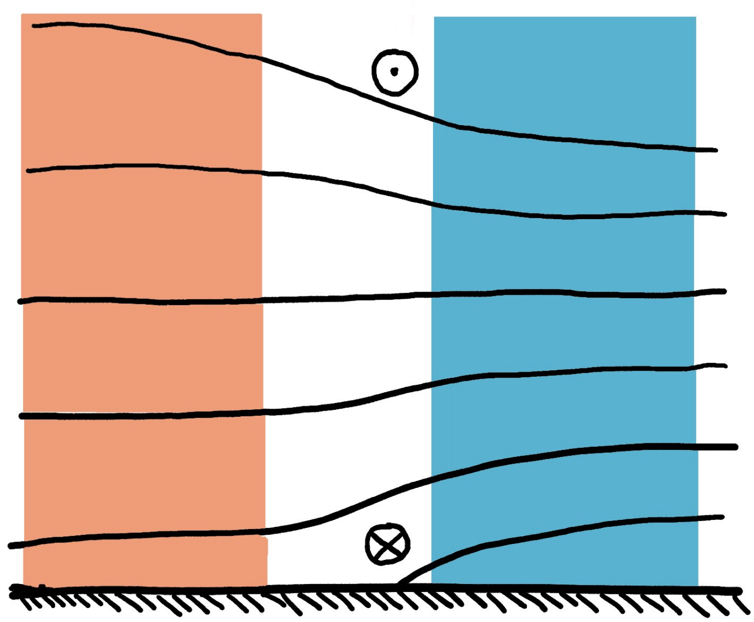


Measurements from NASA Topex-Poseidon spacecraft

Vertical Structure of geostrophy: The thermal wind

When $Ro \ll 1$, geostrophy generally holds at all heights except near the surface, where friction becomes strong. This leads to a tight coupling between the winds and temperatures, or more precisely between horizontal temperature gradients and vertical gradients of the horizontal wind.

To illustrate, imagine a geostrophically balanced atmosphere in the presence of a horizontal temperature gradient. In the warm air column, the air is less dense than in the cold air column. Because the mass per area contained between isobars p_1 and p_2 is $\frac{p_1 - p_2}{g}$ in hydrostatic balance, *the isobars are more widely spaced in the warm air column than the cold air column.* As the diagram makes evident, this changes the magnitude, and potentially even the sign, of the horizontal pressure-gradient force as you consider different altitudes. When these pressure gradients are geostrophically balanced, this leads to strong variations in the horizontal wind with altitude.



Thus, in a geostrophic atmosphere, *horizontal temperature gradients imply vertical shear of the horizontal wind.* This is called the **thermal-wind relationship**. As shown in the diagram, this allows a weather system to be a cyclone at the surface and an anticyclone at the tropopause (or vice versa).

Doing this for an atmosphere is complicated by the fact that density is not constant, but the situation becomes simple if we convert from height to pressure as a vertical coordinate. To do so, express a pressure increment

$$dp = \left(\frac{\partial p}{\partial x}\right)_z dx + \left(\frac{\partial p}{\partial y}\right)_z dy + \left(\frac{\partial p}{\partial z}\right)_x dz$$

which implies that

$$\begin{aligned}\left(\frac{\partial p}{\partial x}\right)_z &= -\left(\frac{\partial p}{\partial z}\right)_{x,y}\left(\frac{\partial z}{\partial x}\right)_p = \rho g\left(\frac{\partial z}{\partial x}\right)_p = \rho\left(\frac{\partial \Phi}{\partial x}\right)_p \\ \left(\frac{\partial p}{\partial y}\right)_z &= -\left(\frac{\partial p}{\partial z}\right)_{x,y}\left(\frac{\partial z}{\partial y}\right)_p = \rho g\left(\frac{\partial z}{\partial y}\right)_p = \rho\left(\frac{\partial \Phi}{\partial y}\right)_p\end{aligned}$$

where hydrostatic balance has been used between the second and third steps and Φ is the gravitational potential (often called "geopotential" in the dynamics literature). This implies that, in the horizontal momentum equations, the pressure gradient force can simply be represented as gradients of geopotential on isobars rather than gradients of pressure on constant-height surfaces. Geostrophy is then just $fu = -\left(\frac{\partial \Phi}{\partial y}\right)_p$ and $fv = \left(\frac{\partial \Phi}{\partial x}\right)_p$. Differentiating this with pressure,

$$\begin{aligned}f\frac{\partial u}{\partial p} &= -\frac{\partial}{\partial y}\left(\frac{\partial \Phi}{\partial p}\right) = \frac{\partial}{\partial y}\left(\frac{RT}{p}\right)_p \\ f\frac{\partial v}{\partial p} &= \frac{\partial}{\partial x}\left(\frac{\partial \Phi}{\partial p}\right) = -\frac{\partial}{\partial x}\left(\frac{RT}{p}\right)_p\end{aligned} \qquad (1.1)$$

where we have used the fact that, in pressure coordinates, hydrostatic balance is $\frac{\partial \Phi}{\partial p} = -\frac{1}{\rho}$, and we have expressed $\rho = \frac{p}{RT}$ using ideal gas. Absorbing the pressure on the righthand side into the derivative, and assuming R is constant, yields the thermal-wind equations for an ideal-gas atmosphere:

$$\boxed{\begin{aligned}f\frac{\partial u}{\partial \ln p} &= R\left(\frac{\partial T}{\partial y}\right)_p \\ f\frac{\partial v}{\partial \ln p} &= -R\left(\frac{\partial T}{\partial x}\right)_p\end{aligned}}$$

Here are some examples. Deep ocean currents are generally weak, and the fast surface currents exist in a thermal-wind balance with horizontal temperature gradients. This shows the oceanic Gulf Stream and rings (~100-km sized vortices that spin off the Gulf Stream):

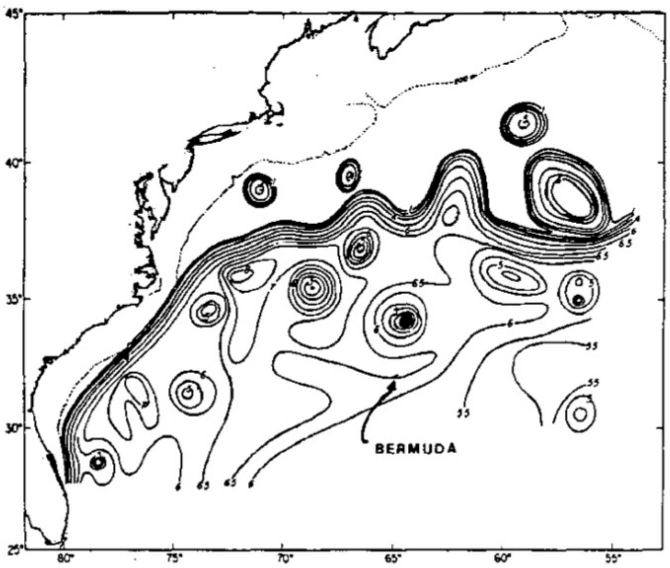


Figure 1 The topography of the 15°C isotherm (depth in 100's of meters) in the northwestern Atlantic for the period March 16–July 9, 1975 (from Richardson et al 1978). The Gulf Stream is indicated by the abrupt northward shoaling of the isotherm, extending from the southwest up the US coast and then off to the east. Four anticyclonic, warm-core rings are found north of the Gulf Stream, while nine cyclonic, cold-core rings are observed to the south.

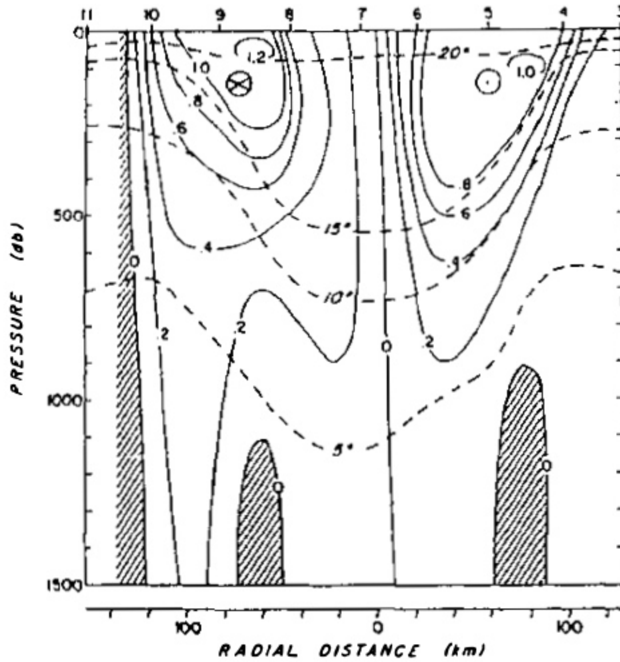
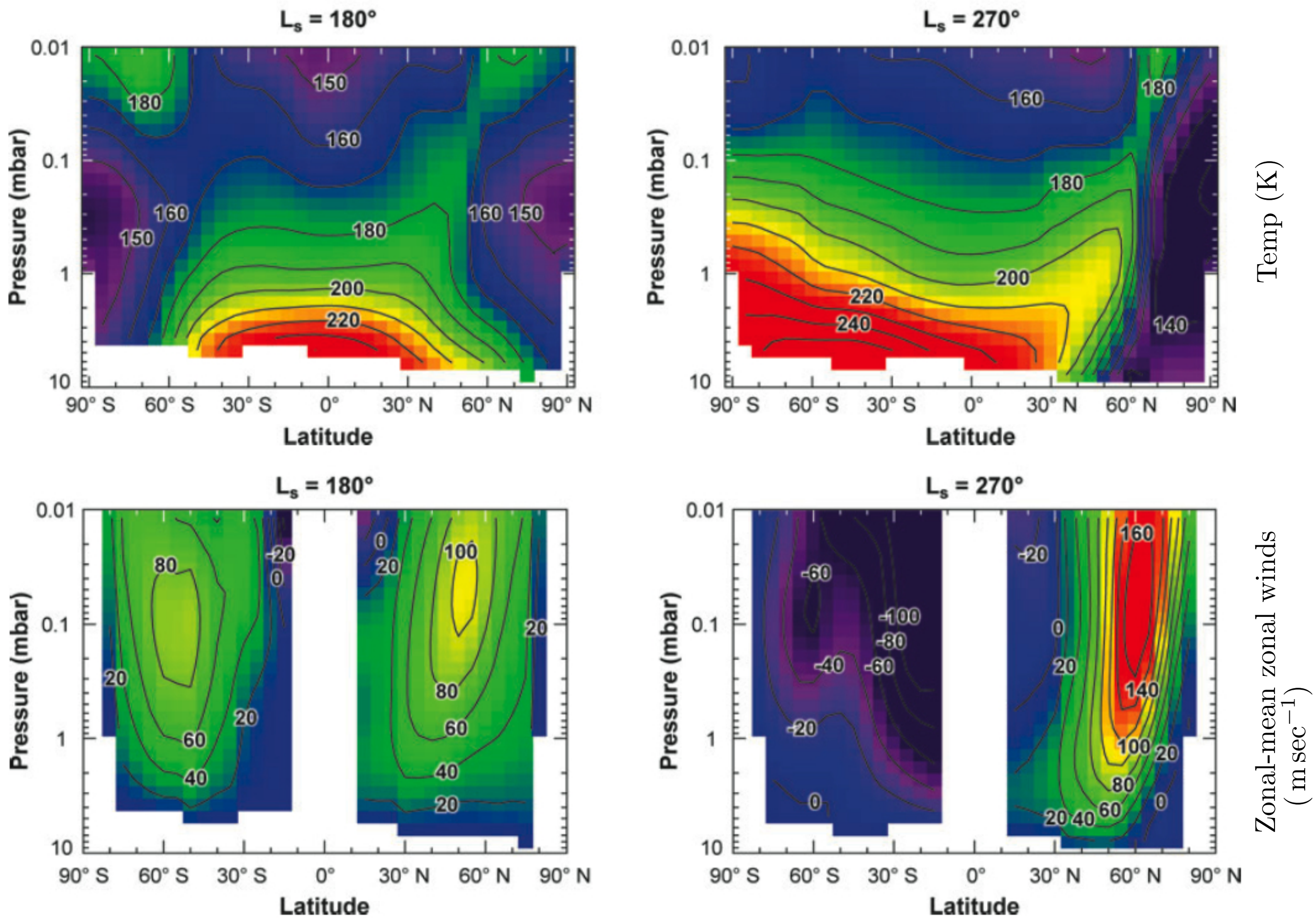


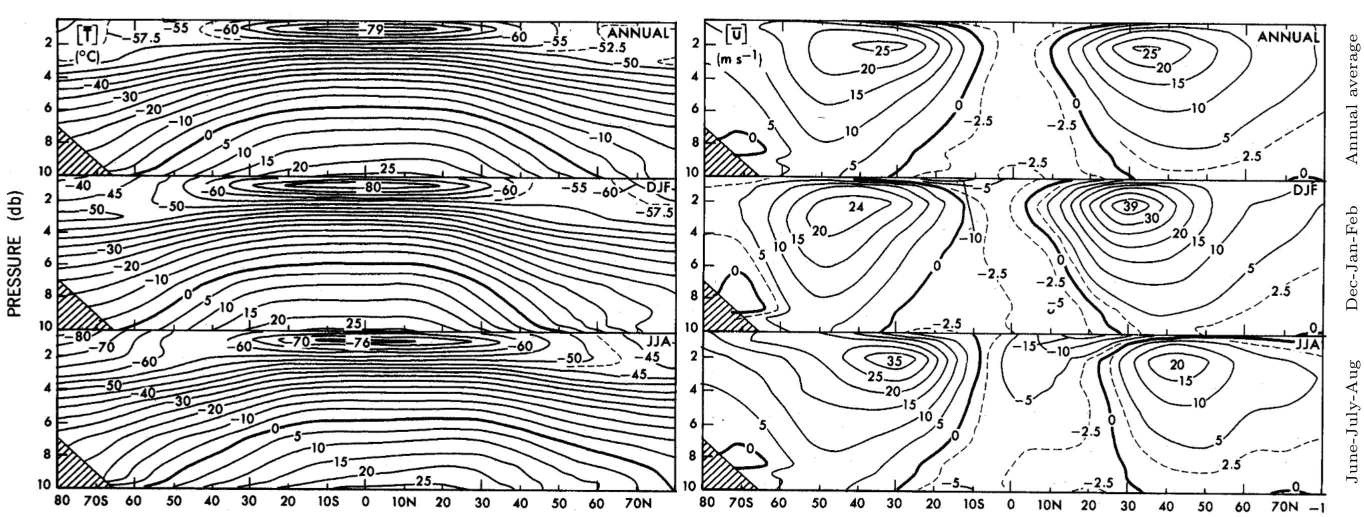
Figure 1.2: A warm-core ring showing isotherms (dashed) and contours of azimuthal current speed (solid), with direction indicated.

Olson (1991), *Ann. Rev. Earth Planet. Sci.* **19**, 283-311

Mars: Top panels show zonal- and seasonal-mean temperature (K), measured by the Mars Global Surveyor spacecraft (Smith 2008). Bottom panels show zonal- and seasonal-mean zonal winds (m/sec) derived from the temperatures assuming thermal-wind balance and zero wind at the surface. Left column is equinox and right column is northern winter solstice. Notice (i) that greatest vertical wind shear occurs at latitudes where temperature gradients are strongest, and (ii) that predominant wind speed in midlatitudes is eastward.

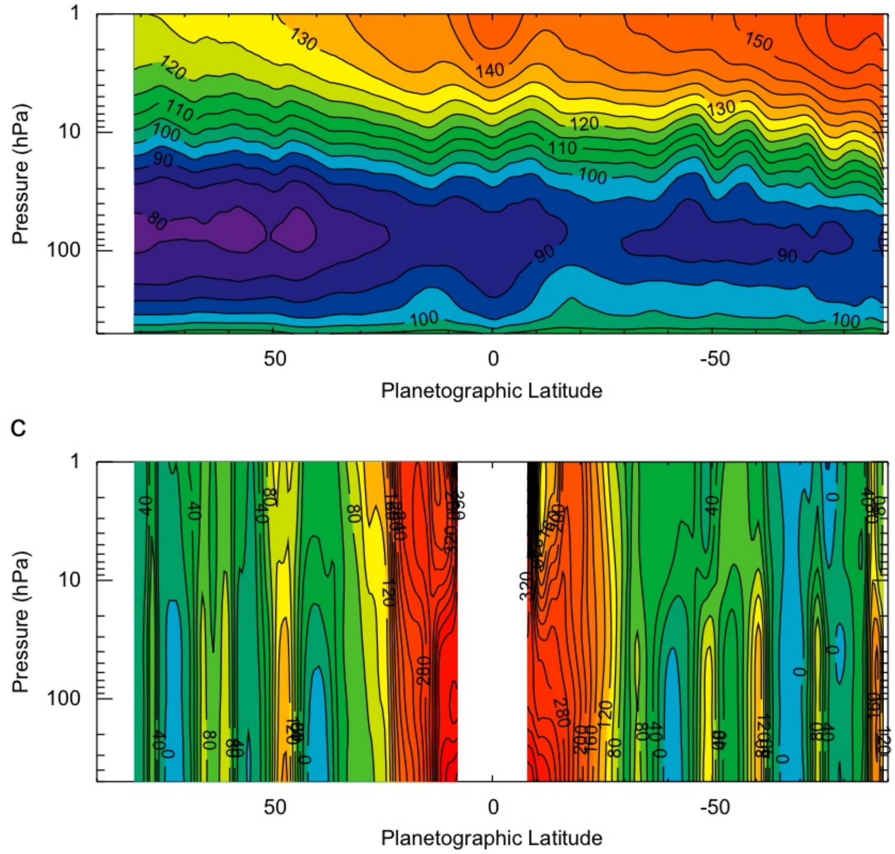


Earth: Left panels show zonal- and time-mean temperatures (°C) and right panels show zonal- and time-mean zonal winds (m/sec) from observations (Peixoto & Oort 1992). Again notice that (i) peak wind shears occur at latitudes with peak temperature gradients, and (ii) predominant wind direction is eastward.



On the giant planets, temperatures above the clouds generally exhibit a correlation with jet shear (cold in anticyclonic regions and warm in cyclonic regions), implying through thermal wind that, in the upper troposphere and lower stratosphere, the jets tend to decay with altitude above the clouds. Some exceptions exist. This shows Saturn, but a similar picture holds for the other three giant planets. The details of why this occurs are not well understood.

Read et al. (2009), *Planet. Space Sci.* **57,** 1682-1698



It is important to emphasize that thermal-wind balance is not a *mechanism* to explain the generation and maintenance of particular wind and temperature patterns; rather, it is simply a statement of force balance. All we really know from thermal-wind balance is that, wherever there are large horizontal temperature gradients, there will be large vertical wind shears. There exist an infinite number of possible configurations for the wind and temperatures in such a balance, so one needs additional physics to determine what controls the particular temperature/wind pattern on any given planet. Still, insofar as the time-mean horizontal temperature gradients are predominantly north-south, with the equator being hotter than the pole and the greatest temperature gradients in midlatitudes, thermal-wind balance makes several useful statements:

- It shows why the time-mean zonal (east-west) winds are stronger than meridional (north-south) winds. In other words, given the temperature distribution, thermal-wind balance provides a partial explanation for the zonally "banded" appearance of the winds on planets.
- Given that temperature gradients peak in midlatitudes, thermal-wind helps explain why the zonal wind shear—hence the upper-tropospheric zonal winds themselves—are greater at midlatitudes than in the tropics or at the poles on Earth and Mars. This is the latitude of the jet streams, so thermal-wind balance would appear to play a role in showing how the winds increase with height in the jet streams.

- Since the equator is generally hotter than the poles on terrestrial planets, thermal-wind balance implies that the zonal wind *increases* (i.e., becomes more eastward) with altitude. Assuming the surface winds are weak, this explains the predominantly eastward nature of the tropospheric winds, especially in midlatitudes.

It is also worth emphasizing that, since thermal-wind balance only constrains the vertical wind *shear*, and not the wind itself, it says nothing about what controls the wind at the surface.

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