POLITECNICO DI MILANO

Cambiamenti Climatici e Adattamenti negli Ecosistemi e nelle Società

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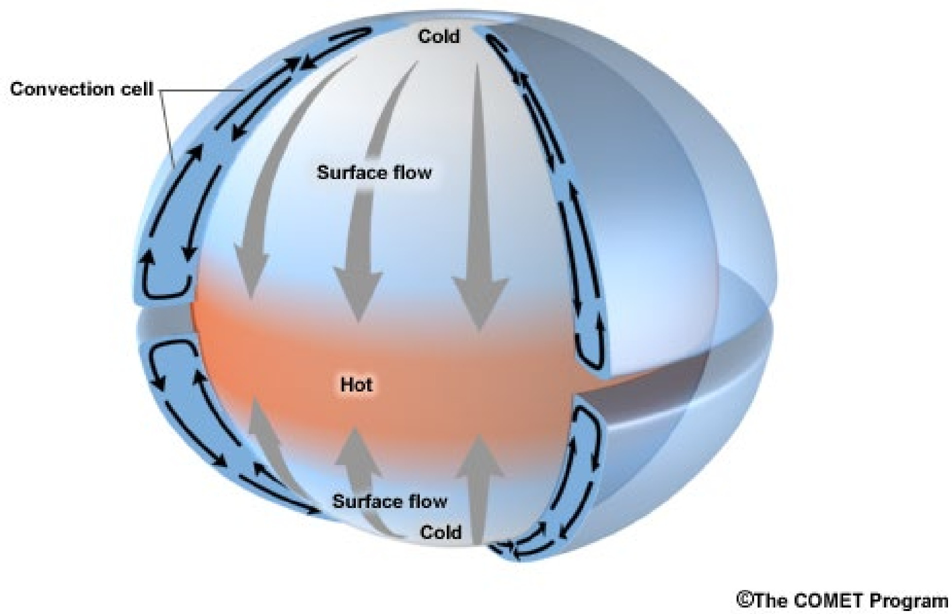
Università degli Studi di Padova

Università degli Studi di Padova

Climate Justice
Jean Monnet
Centre of Excellence

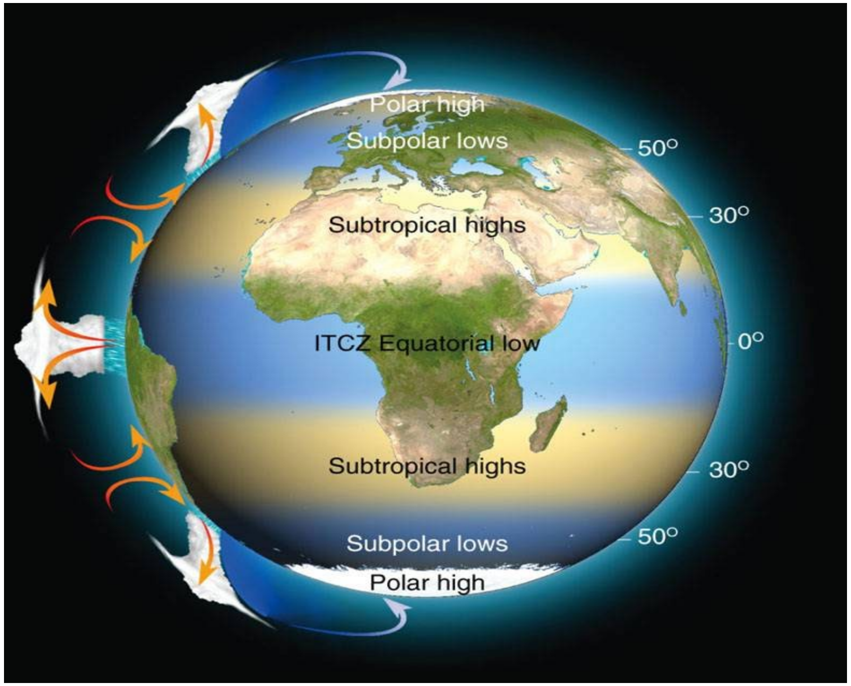
General circulation of the atmosphere

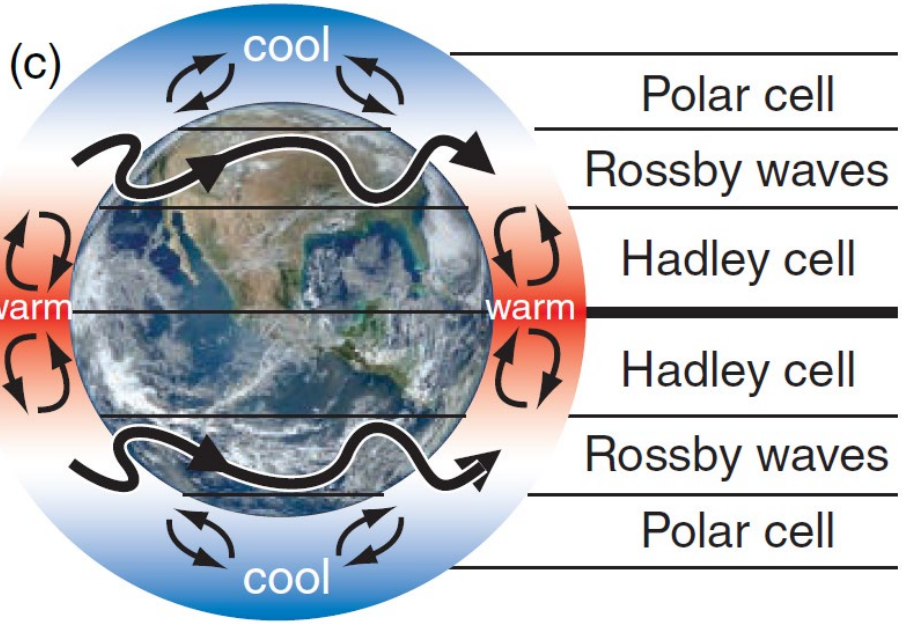
A spatial imbalance between radiative inputs and outputs exists for the earth-ocean-atmosphere system. The earth loses energy at all latitudes due to outgoing infrared (IR) radiation. Near the tropics, more solar radiation enters than IR leaves, hence, there is a net input of radiative energy. Near Earth's poles, incoming solar radiation is too weak to totally offset the IR cooling, allowing a net loss of energy.
The result is **differential heating**, creating warm equatorial air and cold polar air. This imbalance drives the global-scale **general circulation** of winds.



General circulation of the atmosphere

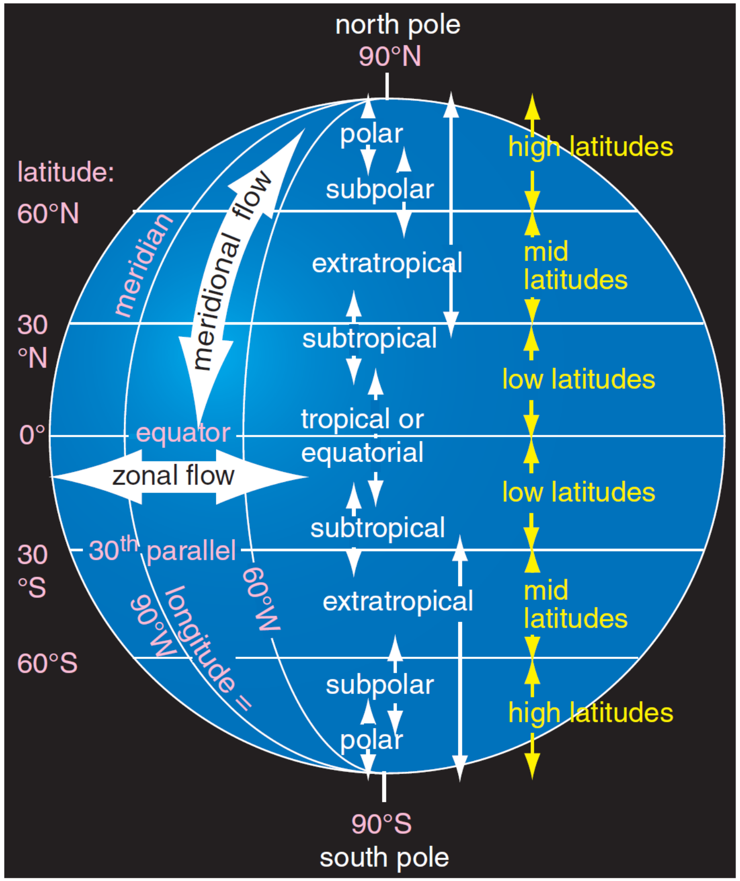
Because buoyancy causes warmer air to rise and colder air to sink, you might guess that equator to pole overturning would exist. Instead, the real general circulation has three bands of circulations in the Northern Hemisphere.





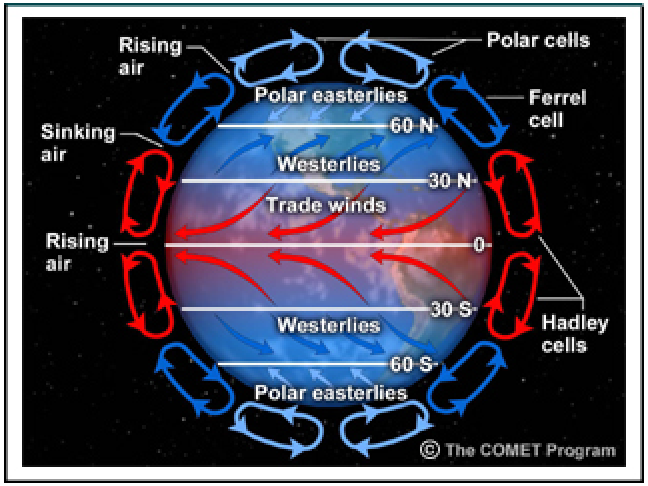
General circulation of the atmosphere

Lines of constant latitude are called **parallels**, and winds parallel to the parallels are identified as **zonal flows** (Figure). Lines of constant longitude are called **meridians**, and winds parallel to the meridians are known as **meridional flows**. Between latitudes of 30° and 60° are the **midlatitudes**. **High latitudes** are 60° to 90°, and **low latitudes** are 0° to 30°. Each 1° of latitude = 111 km. **Tropics**, **subtropics**, **subpolar**, and **polar** regions are as shown in Figure. Regions not in the tropics are called **extratropical**; namely, poleward of about 30°N and about 30°S. For example, **tropical cyclones** such as **hurricanes** are in the tropics. Low-pressure centers (**lows**, as indicated by on weather maps) outside of the tropics are called **extratropical cyclones**.



General circulation of the atmosphere

At low latitudes are broad bands of persistent easterly winds ($U \approx -7$ m s^{-1}) called **trade winds**, named because the **easterlies** allowed sailing ships to conduct transoceanic trade in the old days.
These trade winds also blow toward the equator from both hemispheres, and the equatorial belt of convergence is called the **intertropical convergence zone** (**ITCZ**).

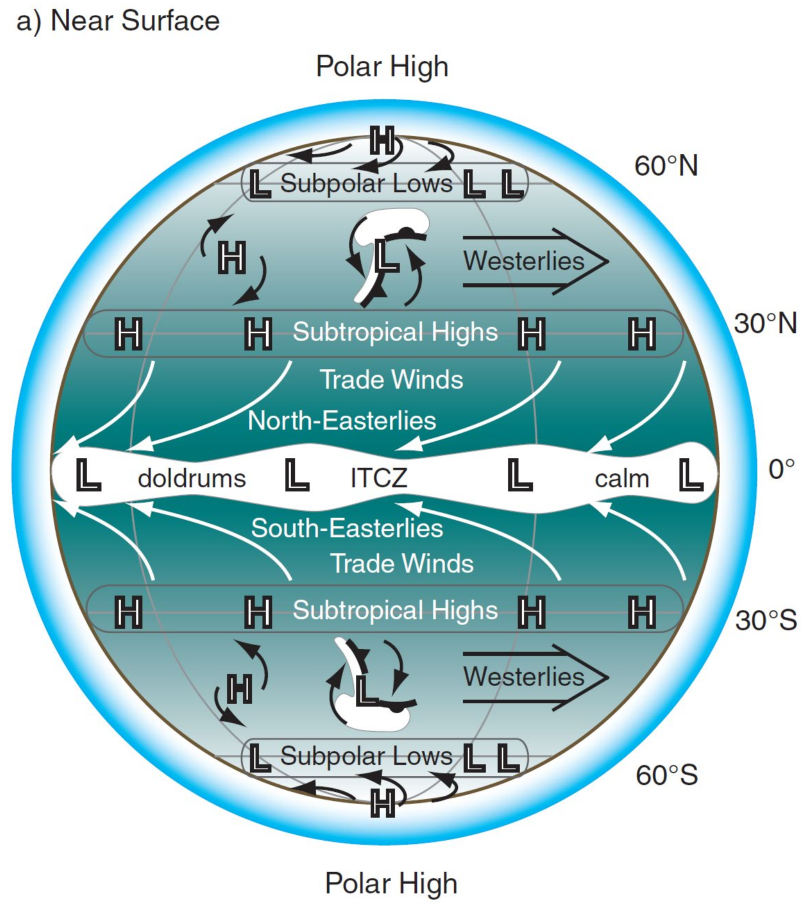


On average, the air at the ITCZ is hot and humid, with low pressure, strong upward air motion, heavy convective (thunderstorm) precipitation, and light to calm winds except in thunderstorms.
This **equatorial trough** (low-pressure belt) was called the **doldrums** by sailors whose sailing ships were becalmed there for many days.

General circulation of the atmosphere

At 30° latitude are belts of high surface pressure called **subtropical highs.** In mid-latitudes are transient centers of low pressure (**mid-latitude cyclones, L**) and high pressure (**anticyclones, H**).

Winds around lows **converge** (come together) and circulate **cyclonically** — counterclockwise in the N. Hemisphere, and clockwise in the S. Hemisphere. Winds around highs **diverge** (spread out) and rotate **anticyclonically** — clockwise in the N. Hemisphere, and counterclockwise in the S. Hemisphere. The cyclones are regions of bad weather (clouds, rain, high humidity, strong winds) and fronts. The anticyclones are regions of good weather (clear skies or fair-weather clouds, no precipitation, dry air, and light winds).

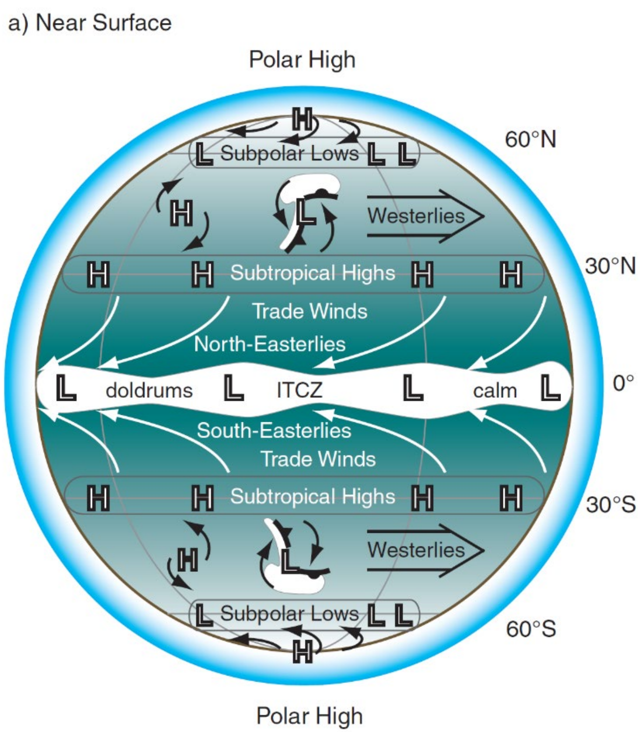


General circulation of the atmosphere

The high- and low-pressure centers move on average from west to east, driven by large-scale winds from the west. Although these **westerlies** dominate the general circulation at mid-latitudes.

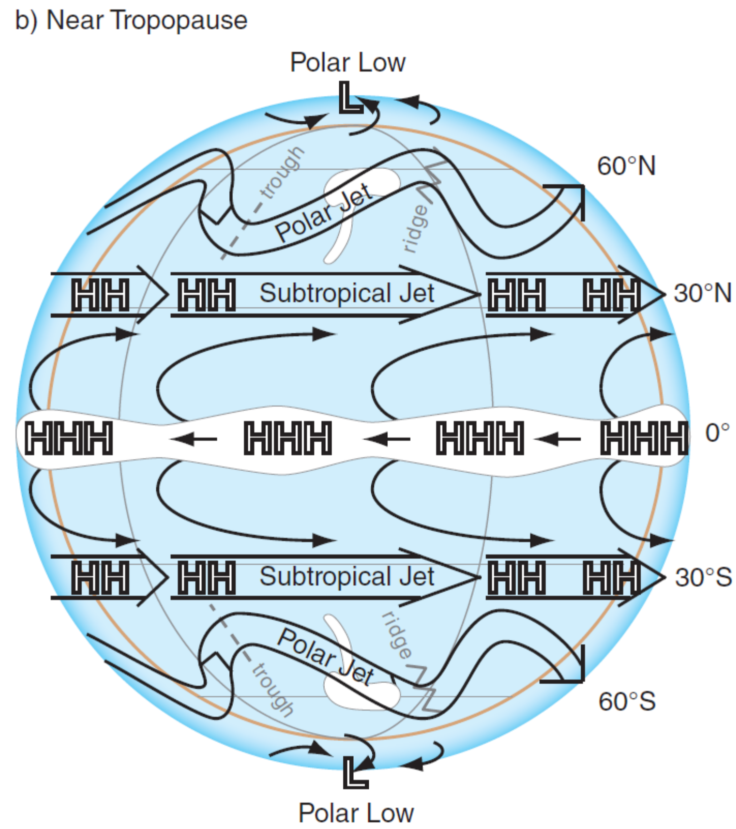
Near 60° latitude are belts of low surface pressure called **subpolar lows.**

Near each pole is a climatological region of high pressure called a **polar high.**



General circulation of the atmosphere

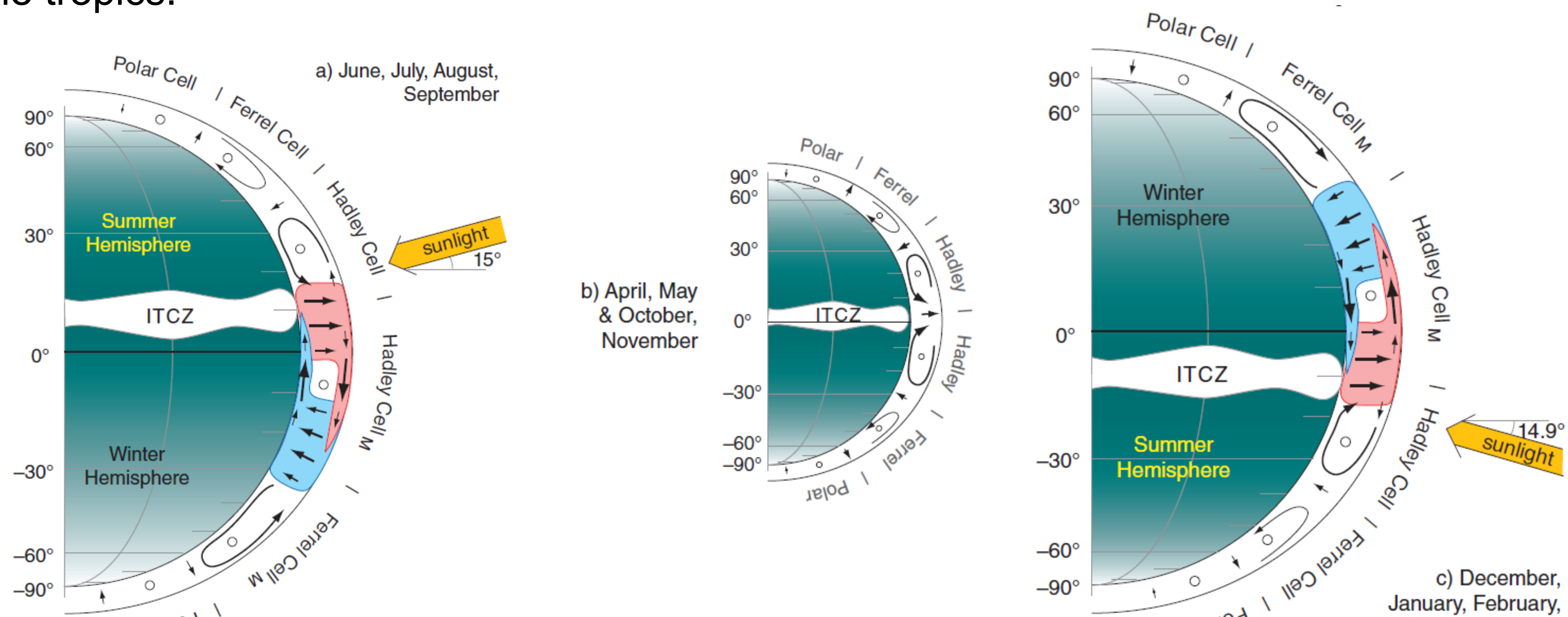
In the tropics is a belt of very strong equatorial high pressure along the tops of the ITCZ thunderstorms.
In mid-latitudes at the tropopause is another belt of strong westerly winds called the **polar jet**. The centerline of the polar jet meanders north and south, resulting in a wave-like shape called a **Rossby wave** (or **planetary wave**).
Near 30° latitude in each hemisphere is a persistent belt of strong westerly winds at the tropopause called the **subtropical jet**. This jet meanders north and south a bit. Pressure here is very high, but not as high as over the equator.



The equatorward portions of the wave are known as low-pressure **troughs**, and poleward portions are known as high-pressure **ridges**.
Near 60° at the tropopause is a belt of low to medium pressure. At each pole is a low-pressure center near the tropopause, with winds at high latitudes generally blowing from the west causing a cyclonic circulation around the **polar low.**

Vertical circulations of warm rising air in the tropics and descending air in the subtropics are called **Hadley cells** or **Hadley circulations.**

At the bottom of the Hadley cell are the trade winds. At the top, near the tropopause, are divergent winds. The updraft portion of the Hadley circulation often contains thunderstorms and heavy precipitation at the ITCZ. This vigorous convection in the troposphere causes a high tropopause (15 – 18 km altitude) and a belt of heavy rain in the tropics.

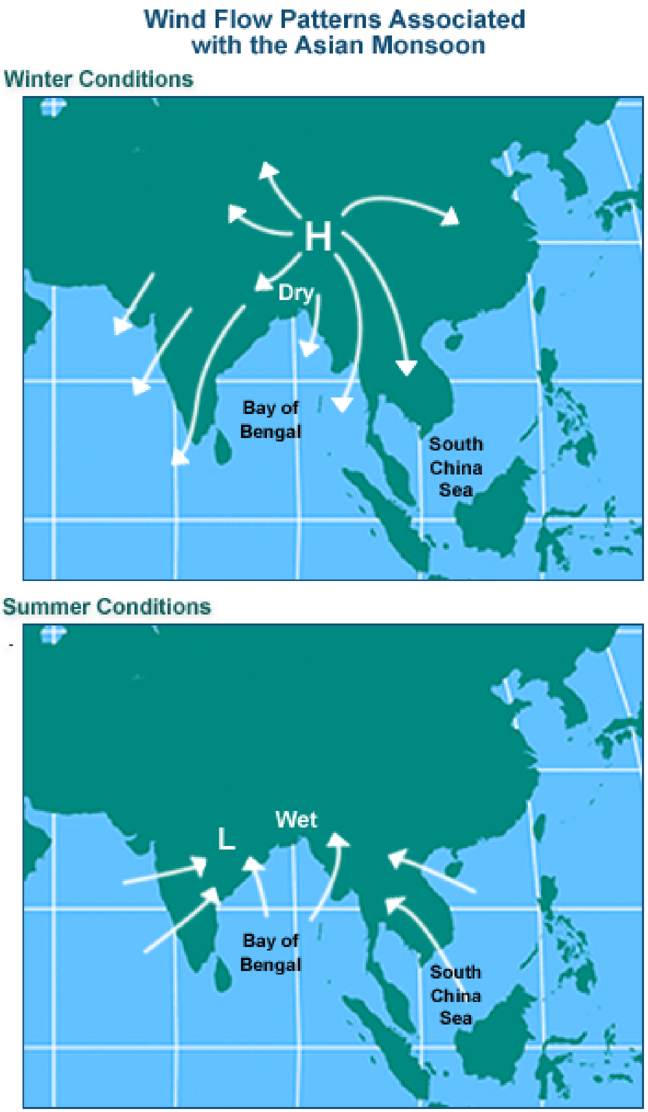


https://www.youtube.com/watch?v=PDEcAxfSYaI

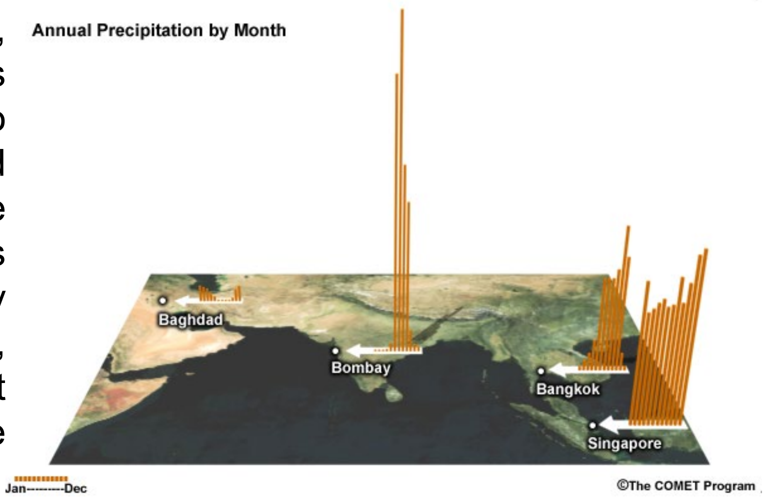
Synoptic-scale climate drivers: Monsoons

• A monsoon circulation is a seasonal change in wind direction caused by a change in the dominant atmospheric pressure pattern. The largest monsoon system is in Southeast Asia, which is dominated by synoptic-scale uplift and upslope winds in summer and downslope winds in winter.

• Monsoonal tropical cities experience copious amounts of summer rain, followed by dry winters, and parched conditions before the monsoon season hits again. The rhythms of the monsoon regime dictate the areas' weather and therefore their climate.



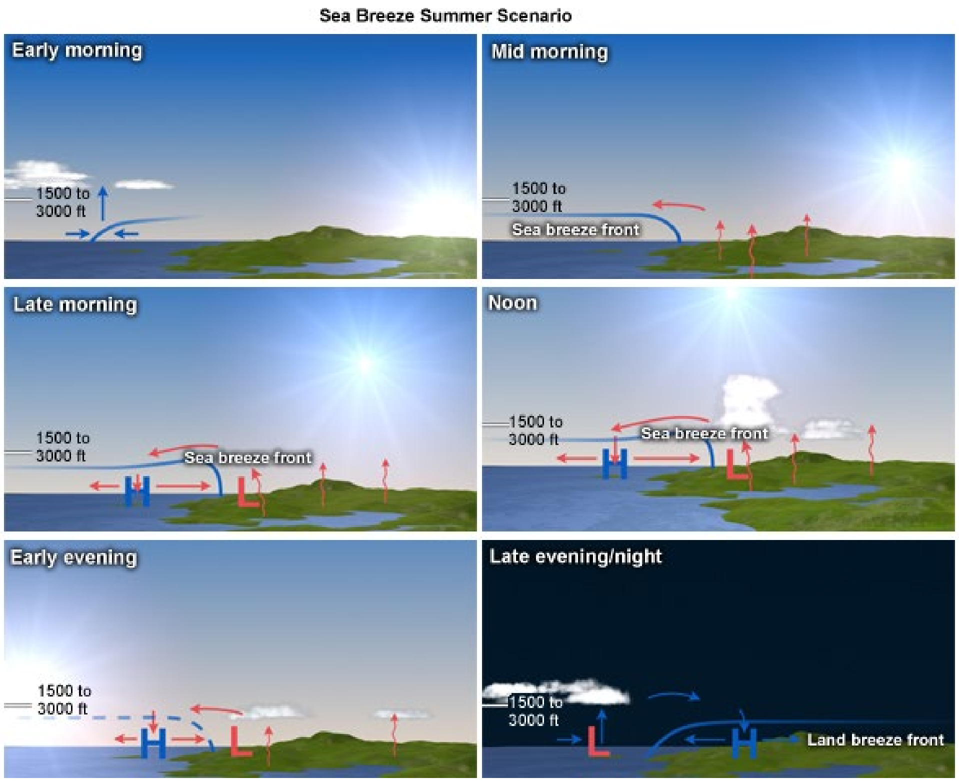
As the graphic shows, Bombay, India experiences monsoonal flow in contrast to Baghdad, Bangkok, and Singapore, which do not. The summer monsoon rains occur in Bombay primarily from June to September, while dry conditions persist throughout the rest of the year.



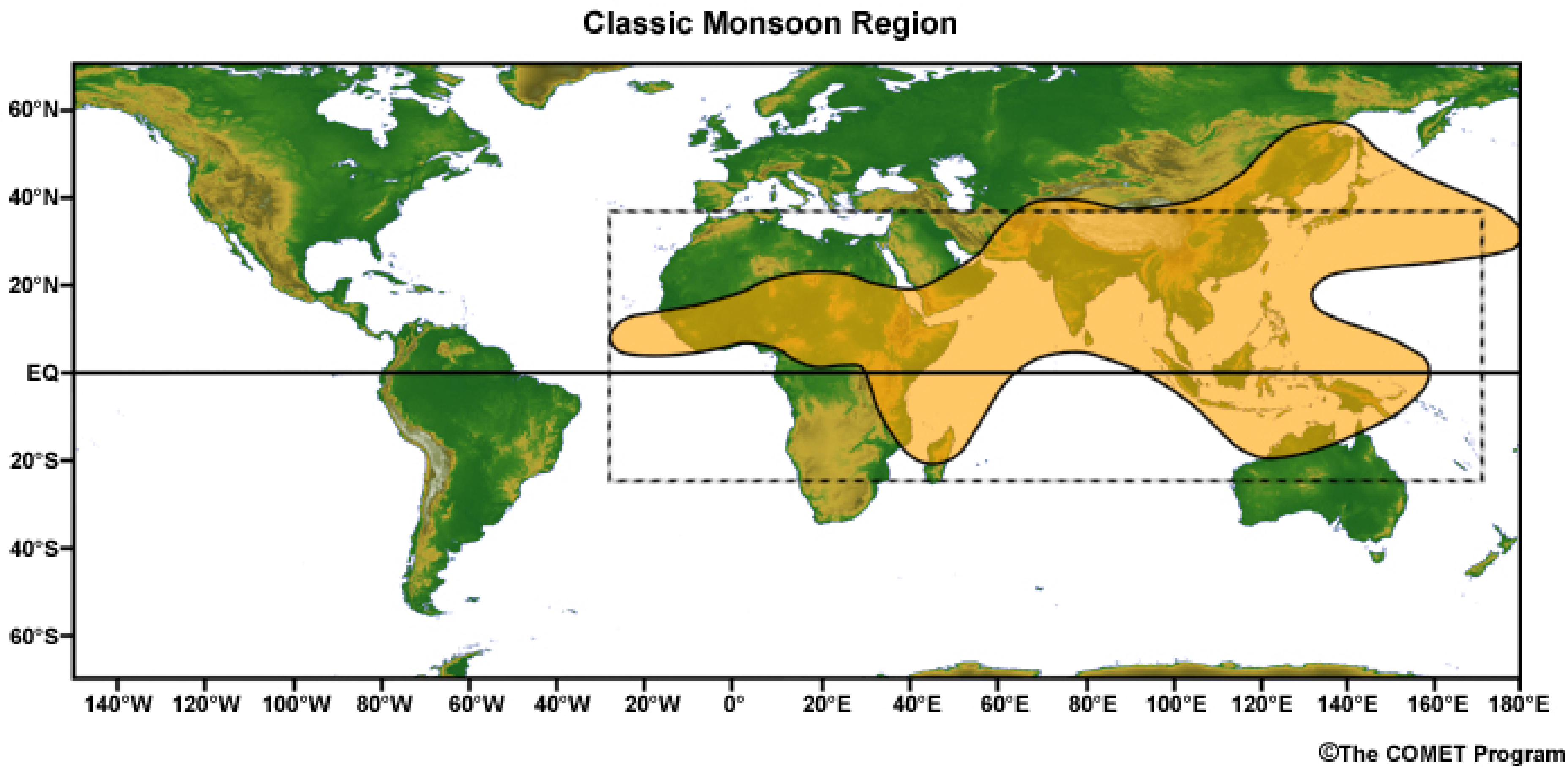
Sea/land breezes

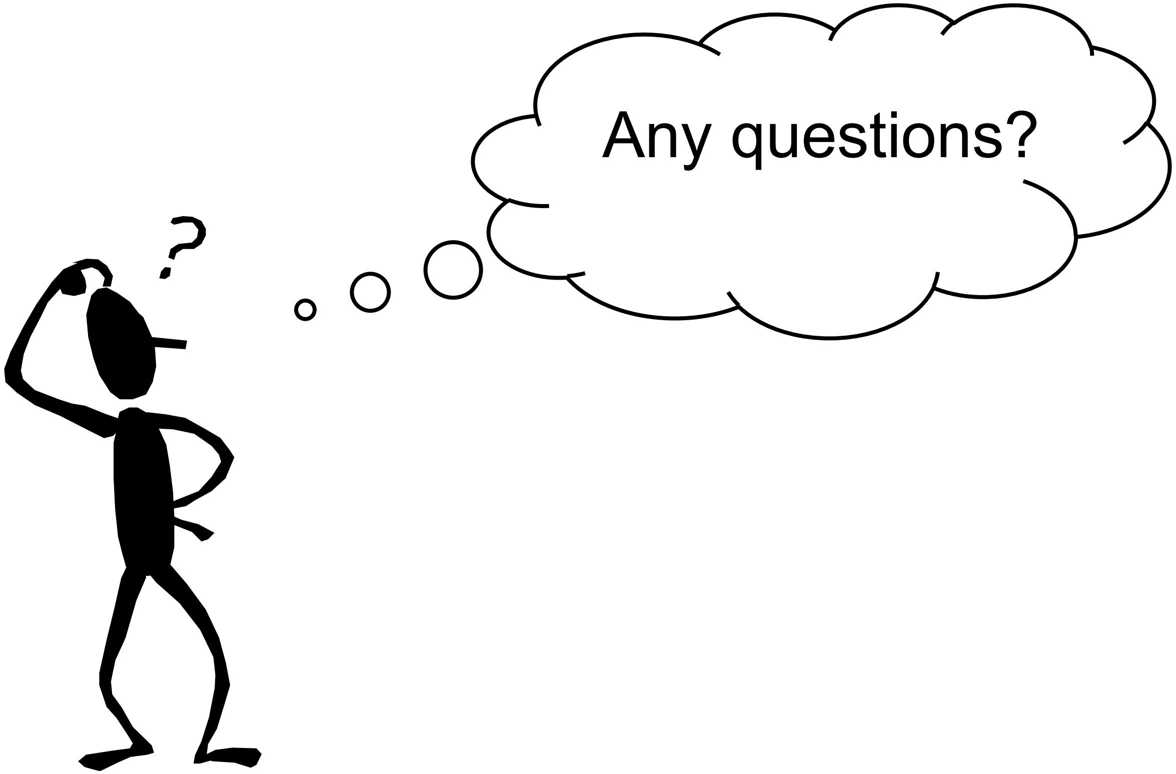
• Areas near coasts and lakes often experience sea or lake breezes as a result of heating differences between water and land surfaces. Since land heats more quickly than water during the day, the air over land warms and therefore rises, causing low-pressure, convergence, and onshore flow. The resulting circulation has return flow aloft over the water with subsiding air just offshore, which causes high pressure and divergence at the surface.

• The rising air over land helps to promote the generation of afternoon clouds and thunderstorms, while the sinking air just off the coast tends to dampen their development. Sea breeze circulations typically weaken and then reverse themselves at night when the land cools faster than the water.



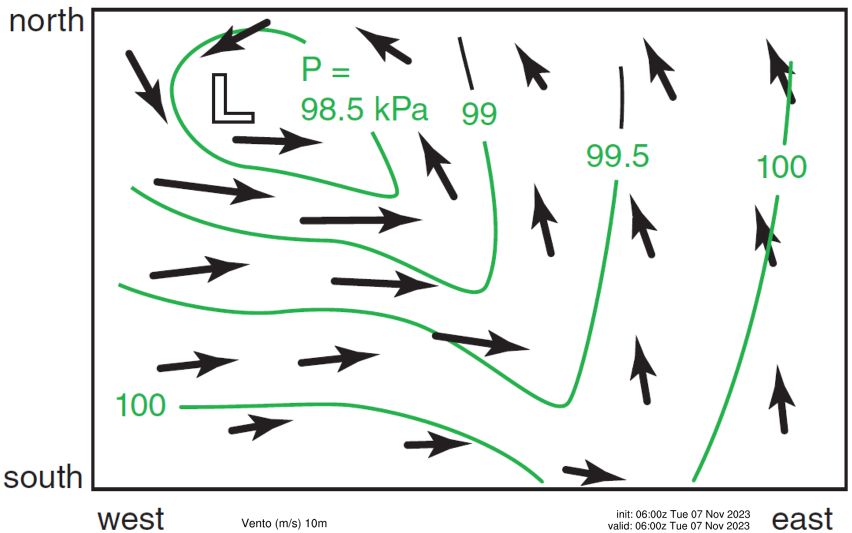
Synoptic-scale climate drivers: Monsoons

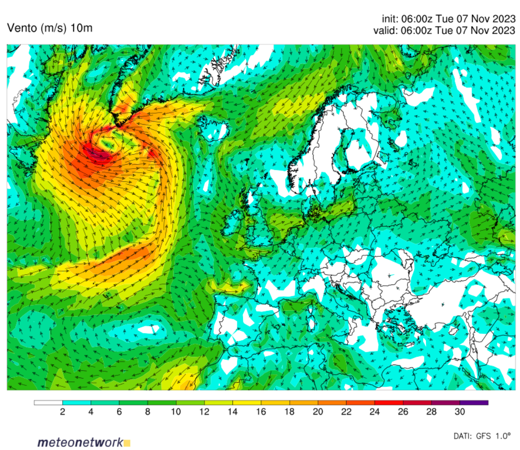


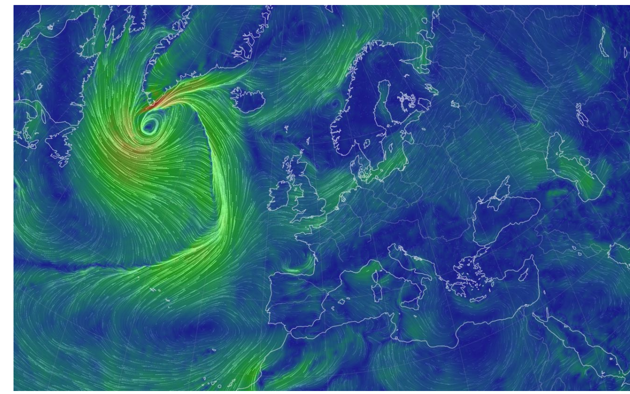

Any questions?

Atmospheric forces and wind

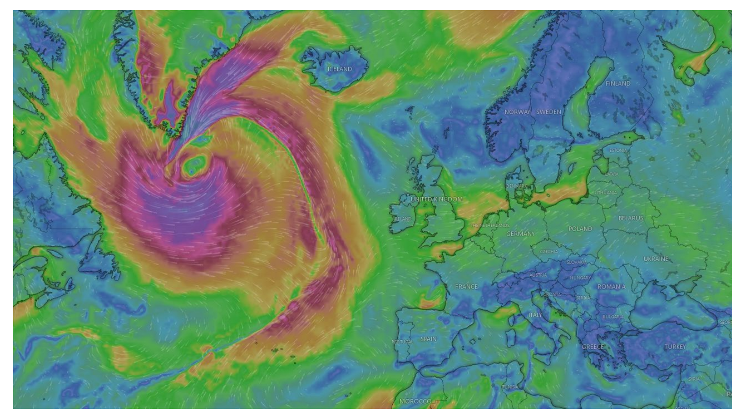
Winds are driven by forces acting on air. But these forces can be altered by heat and moisture carried by the air, resulting in a complex interplay we call weather. Newton's laws of motion describe how forces cause winds — a topic called **dynamics**.





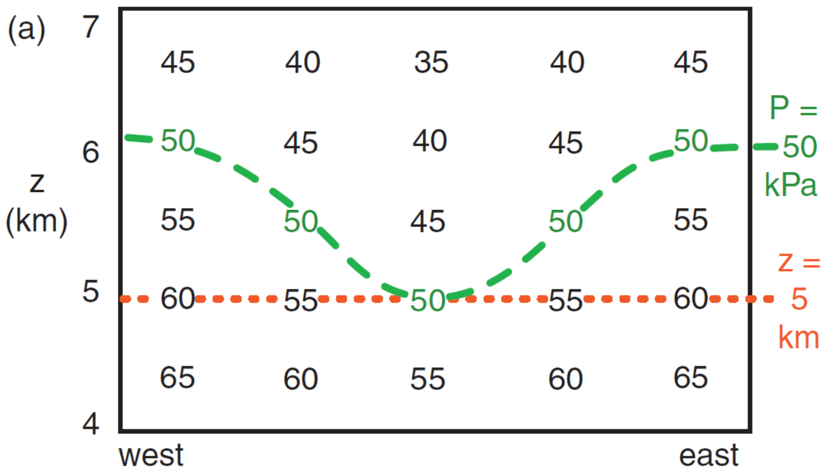


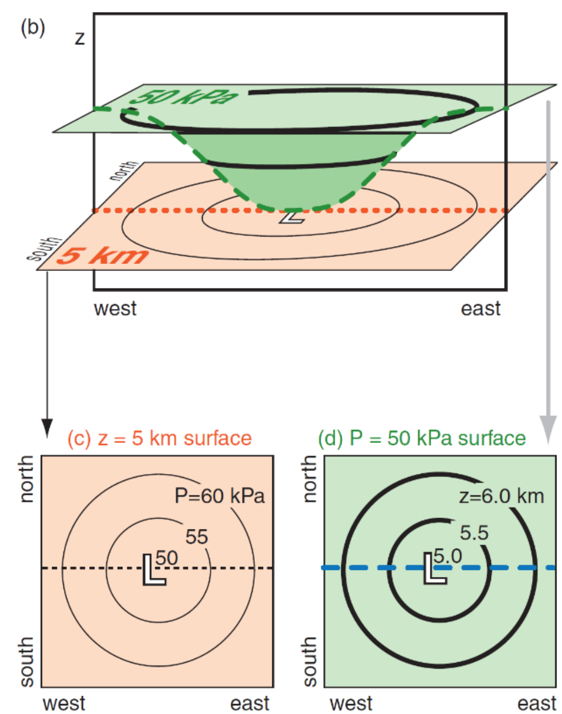
https://earth.nullschool.net/



https://www.windy.com

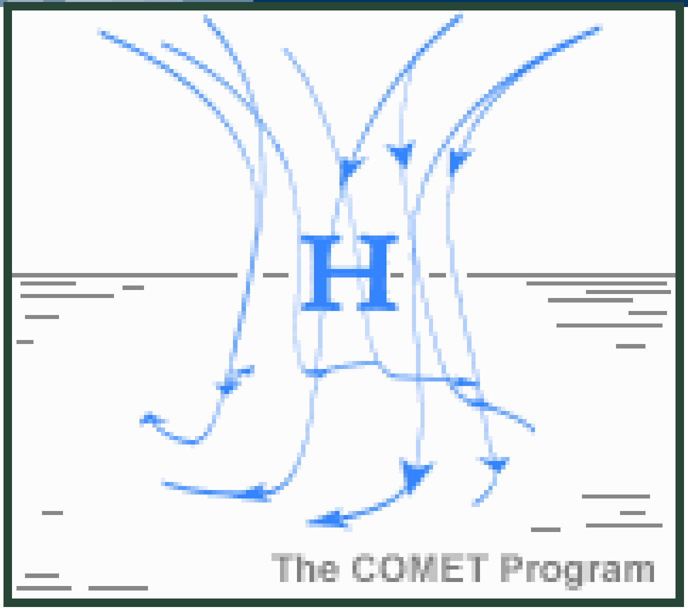
Pressure-gradient force is the most important force because it is the only one that can drive winds in the horizontal. Other horizontal forces can alter an existing wind, but cannot create a wind from calm air. We can create weather maps showing values of the pressures measured at different horizontal locations all at the same altitude, such as at mean sea-level (MSL). Such a map is called a **constant height map**. However, one of the peculiarities of meteorology is that we can also create maps on other surfaces, such as on a surface connecting points of equal pressure. This is called an **isobaric map**.

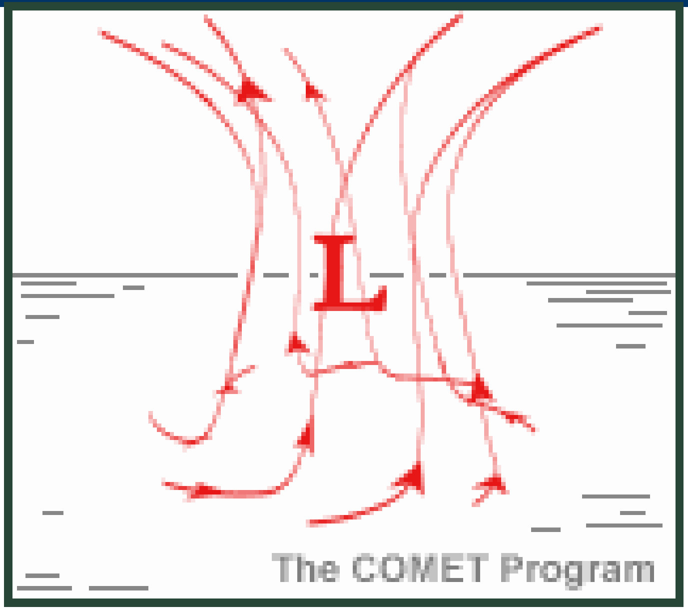


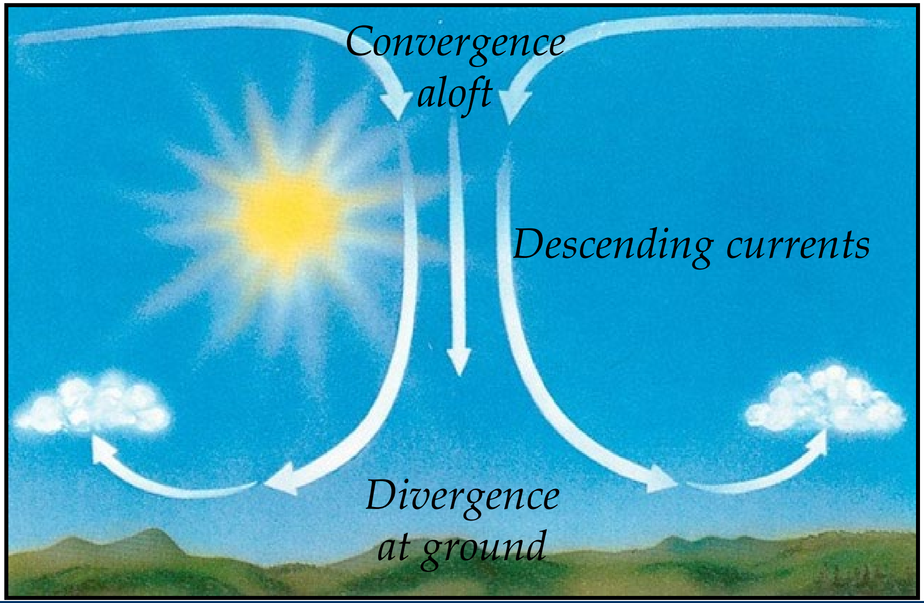


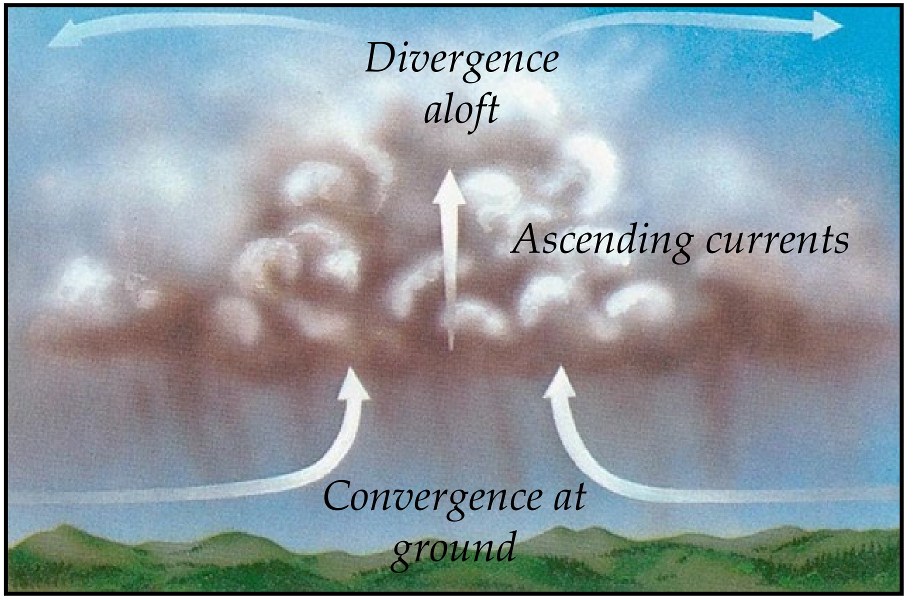
Sketch of the similarity of (c) pressures drawn on a constant height surface to (d) heights drawn on a constant pressure surface. At any one height such as z = 5 km (shown by the thin dotted line in the vertical cross section of Fig. a), pressures at one location on the map might differ from pressure at other locations.

Pressure and wind





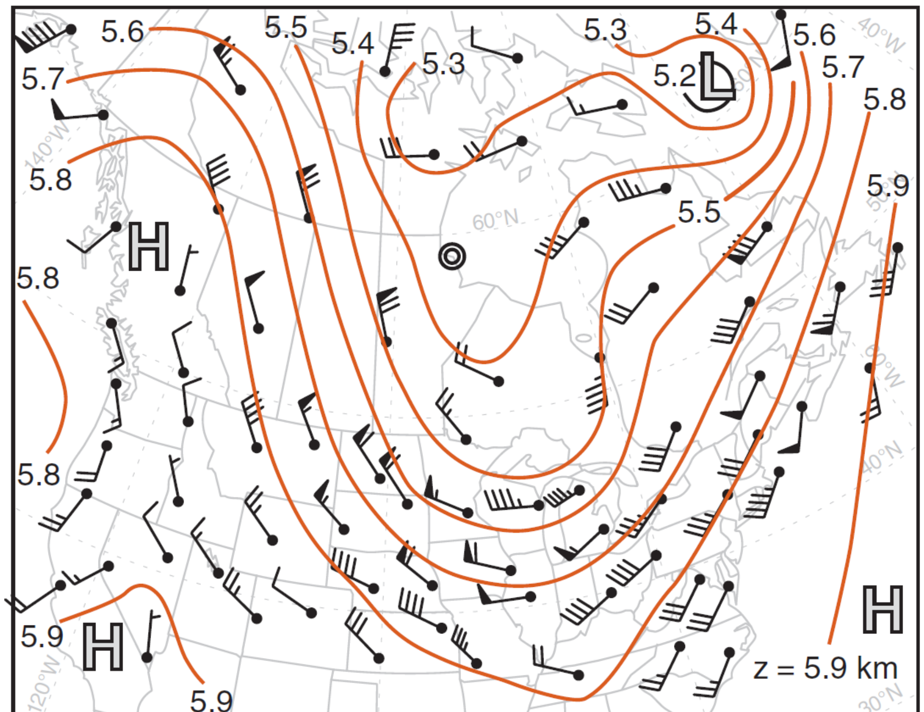




Pressure and wind

Regions on a constant-height map that have tight **packing** (close spacing) of isobars correspond to regions on isobaric maps that have tight packing of height contours, both of which are regions of strong pressure gradients that can drive strong winds.

Symbols on weather maps are like musical notes in a score — they are a shorthand notation that concisely expresses information.



50 kPa Heights (km) & Winds (knots) 12 UTC, 20 Sep 2012

Table 10-1. Interpretation of wind barbs.

Symbol	Wind Speed	Description
◎	calm	two concentric circles
	1 - 2 speed units	shaft with no barbs
	5 speed units	a half barb (half line)
	10 speed units	each full barb (full line)
	50 speed units	each pennant (triangle)

• The total speed is the sum of all barbs and pennants. For example, indicates a wind from the west at speed 75 units. Arrow tip is at the observation location.
• CAUTION: Different organizations use different speed units, such as knots, m s^{-1}, miles h^{-1}, km h^{-1}, etc. Look for a legend to explain the units. When in doubt, assume knots — the WMO standard. For unit conversion, a good approximation is 1 m s^{-1} ≈ 2 knots.

Horizontal forces

Average horizontal winds are often 100 times stronger than vertical winds, except in thunderstorms and near mountains.

Five forces contribute to net horizontal accelerations that control horizontal winds: **pressure-gradient force** (PG), **advection** (AD), **centrifugal force** (CN), **Coriolis force** (CF), and **turbulent drag** (TD):

$$\frac{F_{x\,net}}{m} = \frac{F_{x\,AD}}{m} + \frac{F_{x\,PG}}{m} + \frac{F_{x\,CN}}{m} + \frac{F_{x\,CF}}{m} + \frac{F_{x\,TD}}{m}$$

$$\frac{F_{y\,net}}{m} = \frac{F_{y\,AD}}{m} + \frac{F_{y\,PG}}{m} + \frac{F_{y\,CN}}{m} + \frac{F_{y\,CF}}{m} + \frac{F_{y\,TD}}{m}$$

While Newton's 2nd Law defines the fundamental dynamics (for a Lagrangian framework where the coordinate system follows the moving object), we cannot use it very easily because it requires a coordinate system that moves with the air. Instead, we want to apply it to a fixed location (i.e., an **Eulerian** framework), such as over your house.

$$\vec{F} = m \cdot \vec{a}$$

$$\vec{a} = \frac{\Delta \vec{V}}{\Delta t}$$

$$\vec{F} = m \cdot \frac{\Delta \vec{V}}{\Delta t}$$

$$\vec{F} = \frac{\Delta (m \cdot \vec{V})}{\Delta t}$$

$$\frac{\Delta \vec{V}}{\Delta t} = \frac{\vec{F}_{net}}{m}$$

$$\frac{\Delta V}{\Delta t} = \frac{F_{y\,net}}{m}$$

Advection

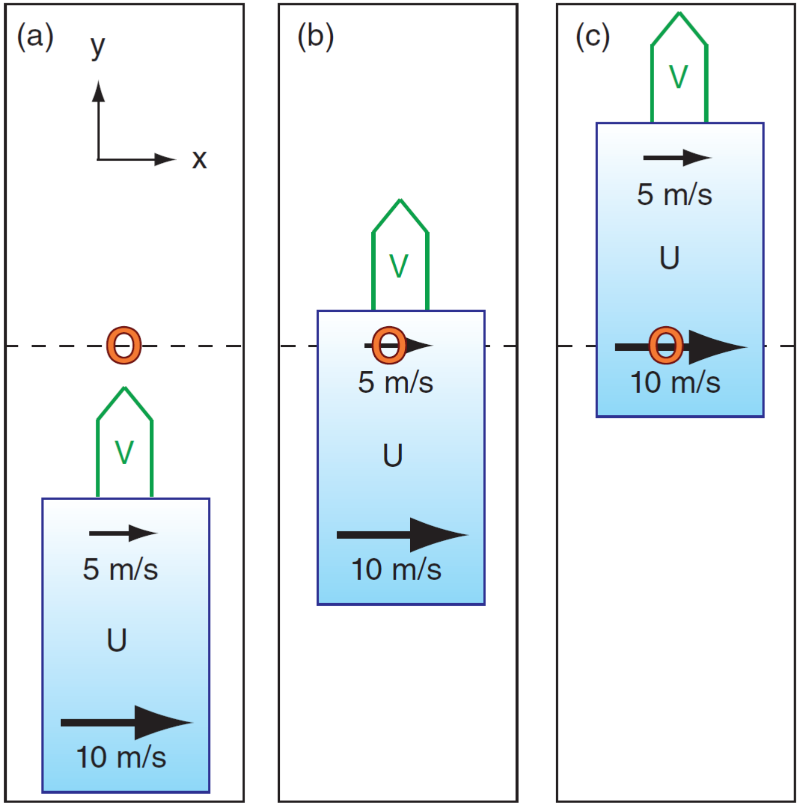
Advection is not a true force. Yet, it can cause a change of wind speed at a fixed location in Eulerian coordinates, so we will treat it like a force here. The wind moving past a point can carry **specific momentum** (i.e., momentum per unit mass). Recall that momentum is defined as mass times velocity, hence specific momentum equals the velocity (i.e., the wind) by definition. Thus, the wind can move (**advect**) different winds to your fixed location.

For advection, the horizontal force components are:

$$\frac{F_{x\,AD}}{m} = -U \cdot \frac{\Delta U}{\Delta x} - V \cdot \frac{\Delta U}{\Delta y} - W \cdot \frac{\Delta U}{\Delta z}$$

$$\frac{F_{y\,AD}}{m} = -U \cdot \frac{\Delta V}{\Delta x} - V \cdot \frac{\Delta V}{\Delta y} - W \cdot \frac{\Delta V}{\Delta z}$$

Recall that a **gradient** is defined as change across a distance, such as $\Delta V / \Delta y$. With no gradient, the wind cannot cause accelerations.



Centrifugal force

Inertia makes an air parcel try to move in a straight line. To get its path to turn requires a force in a different direction. This force, which pulls toward the inside of the turn, is called **centripetal** force. Centripetal force is the result of a net imbalance of (i.e., the nonzero vector sum of) other forces.

For mathematical convenience, we can define an apparent force, called **centrifugal** force, that is opposite to centripetal force. Namely, it points outward from the center of rotation. Centrifugal-force components are:

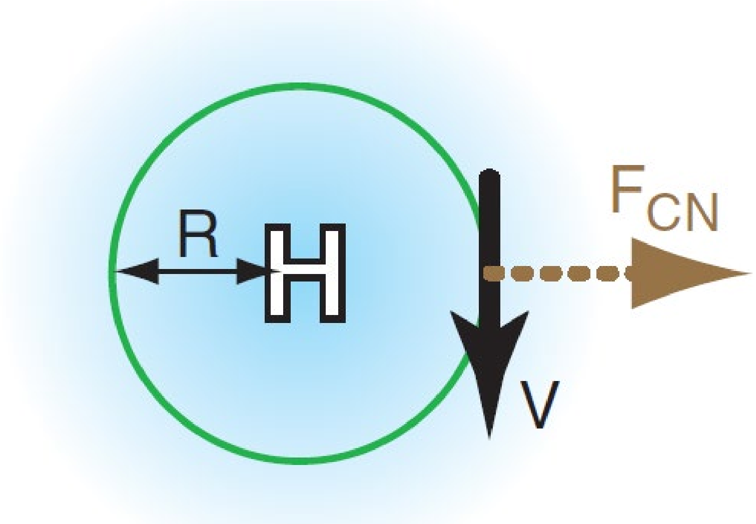
$$\frac{F_{x\,CN}}{m} = +s \cdot \frac{V \cdot M}{R}$$

$$\frac{F_{y\,CN}}{m} = -s \cdot \frac{U \cdot M}{R}$$

where $M = (U_2 + V_2)^{1/2}$ is wind speed (always positive), R is radius of curvature, and s is a sign factor from Table 10-2 as determined by the hemisphere (North or South) and synoptic pressure center (Low or High).

Table 10-2. To apply centrifugal force to separate Cartesian coordinates, a (+/–) sign factor s is required.

Hemisphere	For winds encircling a Low Pressure Center	For winds encircling a High Pressure Center
Southern	–1	+1
Northern	+1	–1



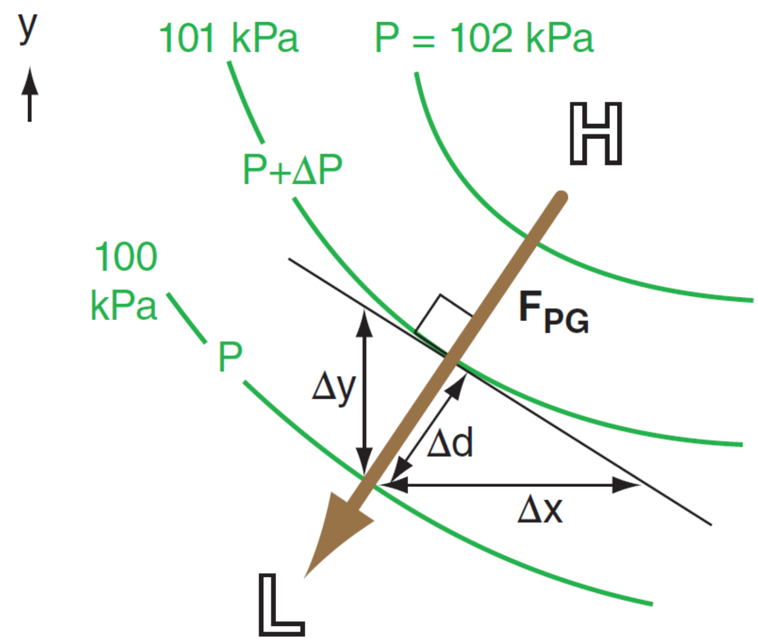
Pressure gradient force

In regions where the pressure changes with distance (i.e., a **pressure gradient**), there is a force from high to low pressure. On weather maps, this force is at right angles to the height contours or isobars, directly from high heights or high pressures to low. Greater gradients (shown by a tighter packing of isobars; i.e., smaller spacing Δ*d* between isobars on weather maps) cause greater pressure-gradient force (Figure). Pressure-gradient force is independent of wind speed, and thus can act on winds of any speed (including calm) and direction.

$$\frac{F_{x\,PG}}{m} = -\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta x}$$

$$\frac{F_{y\,PG}}{m} = -\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta y}$$

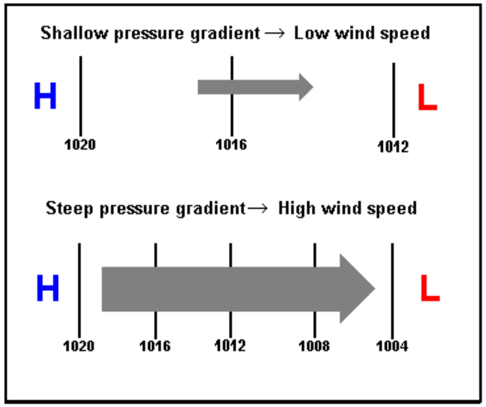
where ΔP is the pressure change across a distance of either Δx or Δy, and ρ is the density of air

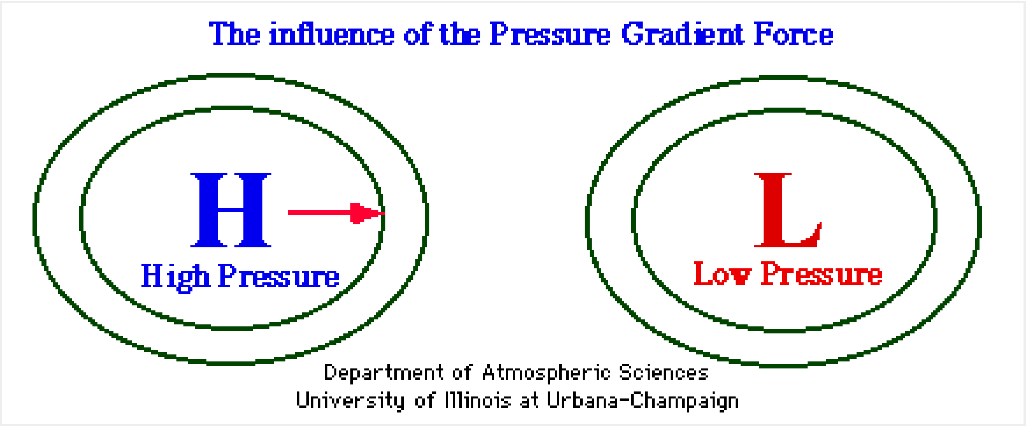


The dark arrow shows the direction of pressure-gradient force FPG from high (H) to low (L) pressure. This force is perpendicular to the isobars (solid curved green lines).

Pressure gradient force

Pressure Gradient Force directed from high to low pressure The change in pressure measured across a given distance is called a "pressure gradient".





The pressure gradient results in a net force that is directed from high to low pressure and this force is called the "pressure gradient force".
The pressure gradient force is responsible for triggering the initial movement of air.

Coriolis force

If viewed from space, earth is like a carousel!
Northern Hemisphere rotates counterclockwise
Southern Hemisphere rotates clockwise

Coriolis force

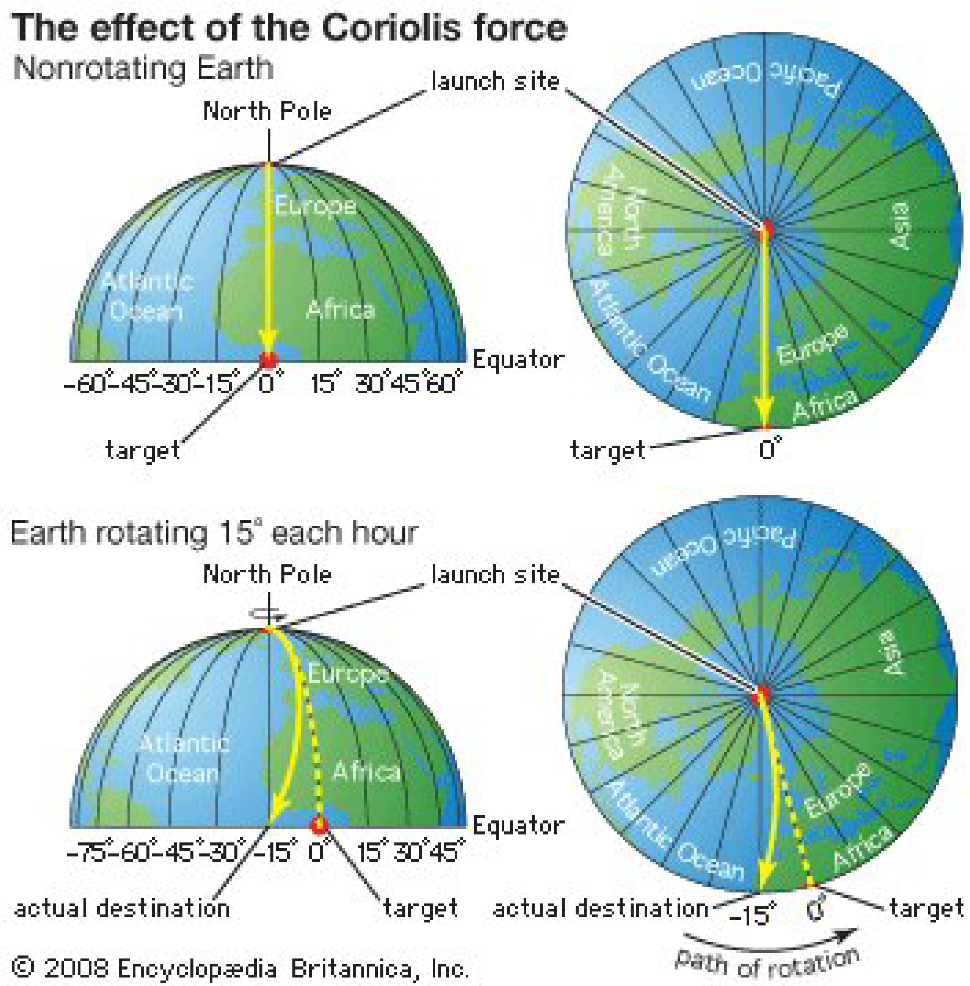
Coriolis Force an artifact of the earth's rotation. Once air has been set in motion by the pressure gradient force, it undergoes an apparent deflection from its path, as seen by an observer on the earth. This apparent deflection is called the "Coriolis force" and is a result of the earth's rotation.

As air moves from high to low pressure in the northern hemisphere, it is deflected to the right by the Coriolis force. In the southern hemisphere, air moving from high to low pressure is deflected to the left by the Coriolis force.

The amount of deflection the air makes is directly related to both the speed at which the air is moving and its latitude. Therefore, slowly blowing winds will be deflected only a small amount, while stronger winds will be deflected more. Likewise, winds blowing closer to the poles will be deflected more than winds at the same speed closer to the equator. The Coriolis force is zero right at the equator.

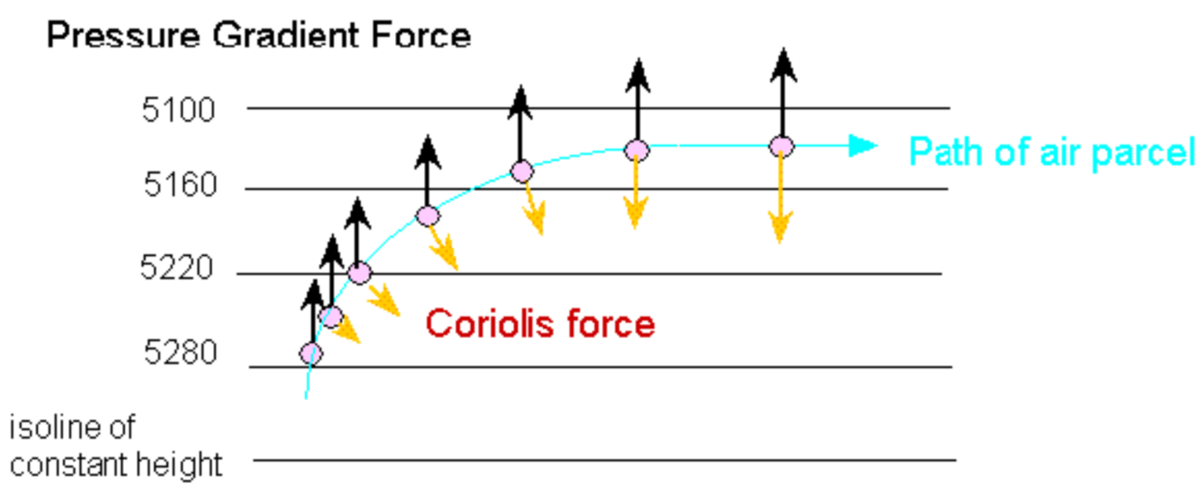
Coriolis force

https://www.metoffice.gov.uk/weather/learn-about/weather/how-weather-works/coriolis-effect



Coriolis force

Initially, since the parcel is at rest, the Coriolis force is zero. The pressure gradient results in a force that accelerates the parcel of air towards lower pressures. Initially the parcel moves in the direction of the pressure gradient force perpendicular to the isoheights. Once the parcel begins to move, the Coriolis force acts to deflect the parcel's path to the right, assuming we are in the Northern Hemisphere. As the wind speed increases because of the pressure gradient force, the Coriolis force strengthens and moves the parcel causing the parcel to travel in a curved path. Eventually, the Coriolis force and the pressure gradient force are equal and acting in opposite directions. Although the net force acting on the parcel of air is zero, the parcel continues to move because of Newton's First Law of Motion. The parcel is in motion and tends to remain in motion.



An object such as an air parcel that moves relative to the Earth experiences a **compound centrifugal force** based on the combined tangential velocities of the Earth's surface and the object. When combined with the non-vertical component of gravity, the result is called Coriolis force. This force points 90° to the right of the wind direction in the Northern Hemisphere, and 90° to the left in the S. Hemisphere.

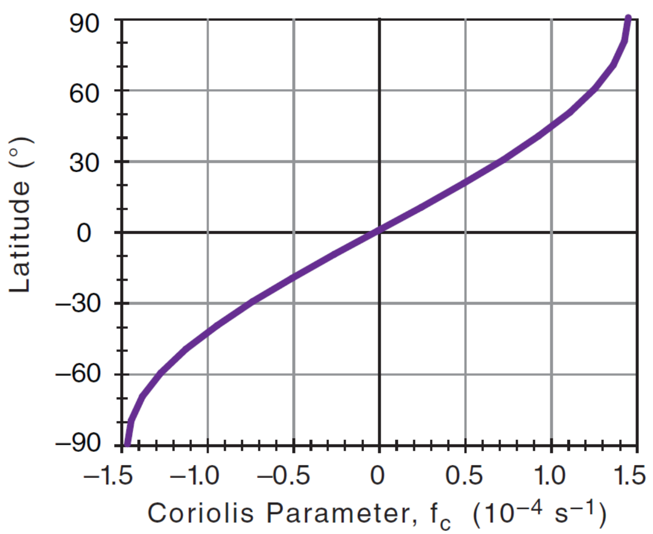
Coriolis parameter

$$f_c = 2 \cdot \Omega \cdot \sin(\phi)$$

$$\Omega = 2 \cdot \pi \,/\, P_{sidereal}$$

Psidereal = 23.9344696 h = sidereal day = period for one revolution of the Earth about its axis, relative to fixed stars

$$= 0.729\,211\,6 \times 10^{-4} \text{ radians s}^{-1}$$



where ϕ is latitude, and $2 \cdot \Omega = 1.458423 \times 10^{-4}\ s^{-1}$

Thus, the Coriolis parameter depends only on latitude.

The Coriolis force in the Northern Hemisphere is:

$$\frac{F_{x\,CF}}{m} = f_c \cdot V$$

$$\frac{F_{y\,CF}}{m} = -f_c \cdot U$$

Coriolis force

- Coriolis force causes the wind to deflect to the right of its intent path in the Northern Hemisphere and to the left in the Southern Hemisphere.

- The magnitude of Coriolis force depends on (1) the rotation of the Earth, (2) the speed of the moving object, and (3) its latitudinal location.

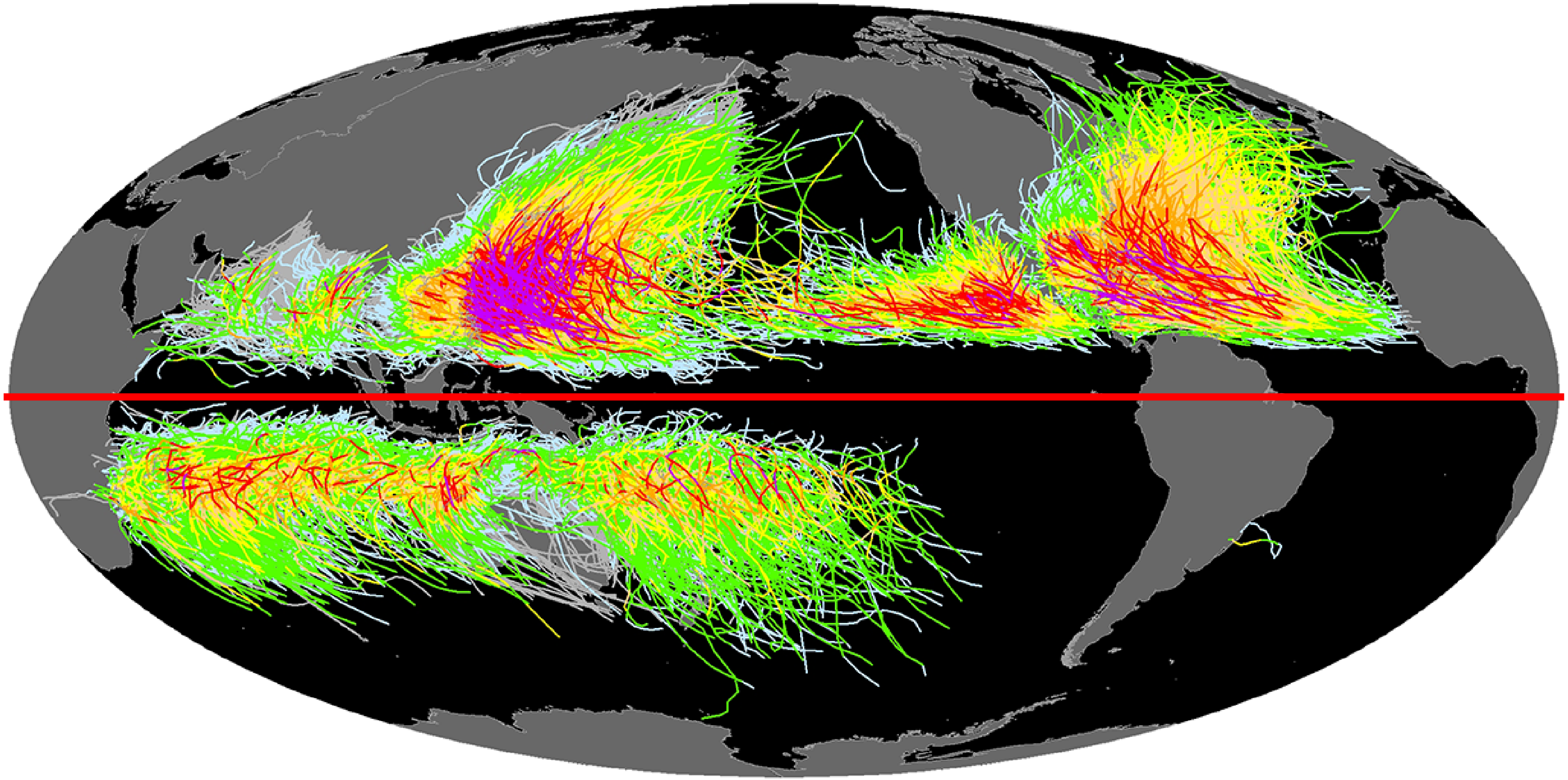
- The stronger the speed (such as wind speed), the stronger the Coriolis force.

- The higher the latitude, the stronger the Coriolis force.

- The Corioils force is zero at the equator.

- Coriolis force is one major factor that determine weather pattern.

Tracks and intensity of all tropical storms



The International Best Track Archive for Climate Stewardship (IBTrACS) stores global tropical cyclone information.

Saffir-Simpson Hurricane Wind Scale

Intensity Missing
Tropical Depression
Tropical Storm
Category 1
Category 2
Category 3
Category 4
Category 5

Surface elements such as pebbles, blades of grass, crops, trees, and buildings partially block the wind, and disturb the air that flows around them. The combined effect of these elements over an area of ground is to cause resistance to air flow, thereby slowing the wind. This resistance is called **drag.**

The net result is a drag force that is normally only felt by air in the ABL. For ABL depth z_i the drag is:

$$\frac{F_{x\,TD}}{m} = -w_T \cdot \frac{U}{z_i}$$

$$\frac{F_{y\,TD}}{m} = -w_T \cdot \frac{V}{z_i}$$

where w_T is called a turbulent **transport velocity**.

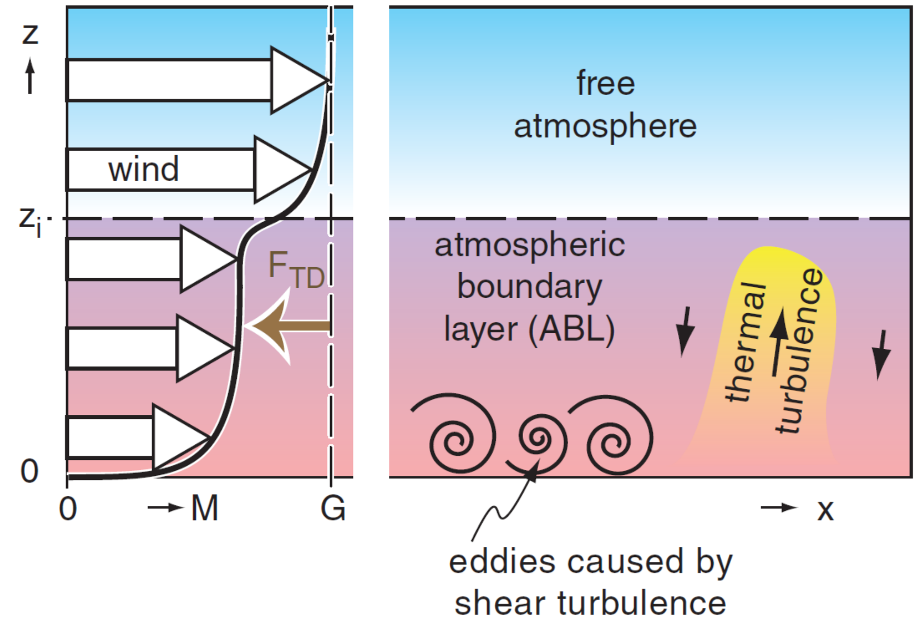
For statically **unstable** ABLs $w_T = b_D \cdot w_B$

where dimensionless factor b_D = $1.83\text{x}10^{-3}$. The **buoyancy velocity scale**, w_B

$$w_B = \left[\frac{|g| \cdot z_i}{T_{v\,ML}} \cdot (\theta_{v\,sfc} - \theta_{v\,ML}) \right]^{1/2}$$

For statically **neutral** conditions $w_T = C_D \cdot M$

where the **drag coefficient** C_D



Wind speed M (curved black line with white highlights) is slower than geostrophic G (vertical dashed line) because of turbulent drag force FTD in the atmospheric boundary layer

Summary of forces

Table 10-3. Summary of forces.

Item	Name of Force	Direction	Magnitude (N kg^{-1})	Horiz. (H) or Vert. (V)	Remarks ("item" is in column 1; H & V in col. 5)
1	**gravity**	down	$\left\|\frac{F_G}{m}\right\| = \|g\| = 9.8\ \text{m·s}^{-2}$	V	**hydrostatic equilibrium** when items 1 & 2V balance
2	**pressure gradient**	from high to low pressure	$\left\|\frac{F_{PG}}{m}\right\| = \left\|g \cdot \frac{\Delta z}{\Delta d}\right\|$	V & H	the only force that can drive horizontal winds
3	**Coriolis** (compound)	90° to right (left) of wind in Northern (Southern) Hemisphere	$\left\|\frac{F_{CF}}{m}\right\| = 2\cdot\Omega\cdot\|\sin(\phi)\cdot M\|$	H*	**geostrophic wind** when 2H and 3 balance (explained later in horiz. wind section)
4	**turbulent drag**	opposite to wind	$\left\|\frac{F_{TD}}{m}\right\| = w_T \cdot \frac{M}{z_i}$	H*	**atm. boundary-layer wind** when 2H, 3 and 4 balance (explained in horiz. wind section)
5	**centrifugal** (apparent)	away from center of curvature	$\left\|\frac{F_{CN}}{m}\right\| = \frac{M^2}{R}$	H*	centripetal = opposite of centrifugal. **Gradient wind** when 2H, 3 and 5 balance
6	**advection** (apparent)	(any)	$\left\|\frac{F_{AD}}{m}\right\| = \left\|-M\cdot\frac{\Delta U}{\Delta d} - \dots\right\|$	V & H	neither creates nor destroys momentum; just moves it

**Horizontal is the direction we will focus on. However, Coriolis force has a small vertical component for zonal winds. Turbulent drag can exist in the vertical for rising or sinking air, but has completely different form than the boundary-layer drag given above. Centrifugal force can exist in the vertical for vortices with horizontal axes. Note: units N kg^{-1} = m·s^{-2}.*

Combining the forces into Newton's Second Law of Motion gives simplified equations of horizontal motion.

$$\frac{\Delta U}{\Delta t} = -U\frac{\Delta U}{\Delta x} - V\frac{\Delta U}{\Delta y} - W\frac{\Delta U}{\Delta z} - \frac{1}{\rho}\cdot\frac{\Delta P}{\Delta x} + f_c\cdot V - w_T\cdot\frac{U}{z_i}$$

$$\underbrace{\frac{\Delta V}{\Delta t}}_{\textit{tendency}} = \underbrace{-U\frac{\Delta V}{\Delta x} - V\frac{\Delta V}{\Delta y} - W\frac{\Delta V}{\Delta z}}_{\textit{advection}} \underbrace{- \frac{1}{\rho}\cdot\frac{\Delta P}{\Delta y}}_{\textit{pressure gradient}} \underbrace{- f_c\cdot U}_{\textit{Coriolis}} \underbrace{- w_T\cdot\frac{V}{z_i}}_{\textit{turbulent drag}}$$

These are the forecast equations for wind. For special conditions where steady winds around a circle are anticipated, centrifugal force can be included.

Horizontal winds

Caution: Steady state means no further change to the non-zero winds. Do not assume the winds are zero. Under certain idealized conditions, some of the forces in the equations of motion are small enough to be neglected. For these situations, theoretical steady-state winds can be found based on only the remaining larger-magnitude forces. These theoretical winds are given special names, as listed in Table 10-4.

Table 10-4. Names of idealized steady-state horizontal winds, and the forces that govern them.

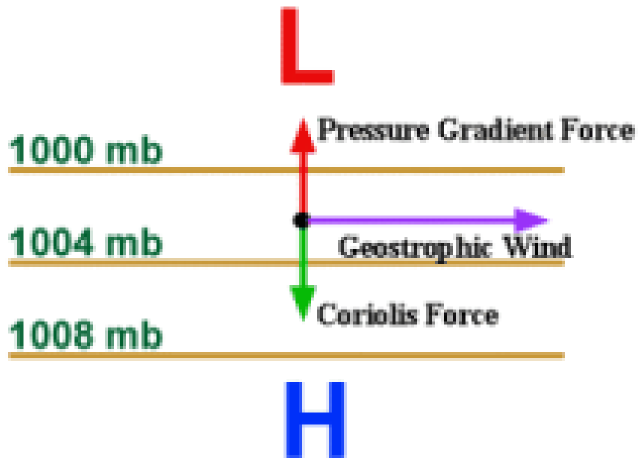
$$0 = -U\frac{\Delta U}{\Delta x} - \frac{1}{\rho}\cdot\frac{\Delta P}{\Delta x} + f_c\cdot V - w_T\cdot\frac{U}{z_i} + s\frac{V\cdot M}{R}$$

Forces:	*pressure gradient*	*Coriolis*	*turbulent drag*	*centrifugal*
Wind Name				
Geostrophic	•	•		
Gradient	•	•		•
Atm.Bound. Layer	•	•	•	
ABL Gradient	•	•	•	•
Cyclostrophic	•			•
Inertial		•		•
Antitriptic	•		•	

Geostrophic winds

An air parcel initially at rest will move from high pressure to low pressure because of the pressure gradient force (PGF). However, as that air parcel begins to move, it is deflected by the Coriolis force to the right in the northern hemisphere (to the left on the southern hemisphere). As the wind gains speed, the deflection increases until the Coriolis force equals the pressure gradient force. At this point, the wind will be blowing parallel to the isobars. When this happens, the wind is referred to as **geostrophic**.

Winds in nature are rarely exactly geostrophic, but to a good approximation, the winds in the upper troposphere can be close. This is because winds are only considered truly geostrophic when the isobars are straight and there are no other forces acting on it and these conditions just aren't found too often in nature.



Geostrophic winds

For special conditions where the only forces are Coriolis and pressure-gradient , the resulting steady-state wind is called the geostrophic wind, with components (*U*g , *V*g).

$$0 = -\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta x} + f_c \cdot V$$

$$0 = -\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta y} - f_c \cdot U$$

Define $U \equiv U_g$ and $V \equiv V_g$ in the equations above, and then solve for these wind components

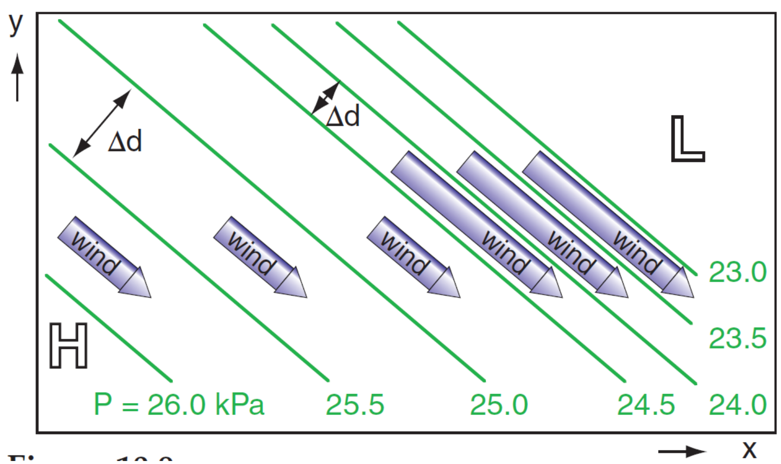
$$U_g = -\frac{1}{\rho \cdot f_c}\cdot\frac{\Delta P}{\Delta y}$$

$$V_g = +\frac{1}{\rho \cdot f_c}\cdot\frac{\Delta P}{\Delta x}$$

where fc = (1.4584×10^{-4} s^{-1}) · sin(*latitude*) is the Coriolis parameter, ρ is air density, and $\Delta P/\Delta x$ and $\Delta P/\Delta y$ are the horizontal pressure gradients

Real winds are nearly geostrophic at locations where isobars or height contours are relatively straight, for altitudes above the atmospheric boundary layer. Geostrophic winds are fast where isobars are packed closer together. The geostrophic wind direction is parallel to the height contours or isobars.

In the N. (S.) hemisphere the wind direction is such that low pressure is to the wind's left (right).

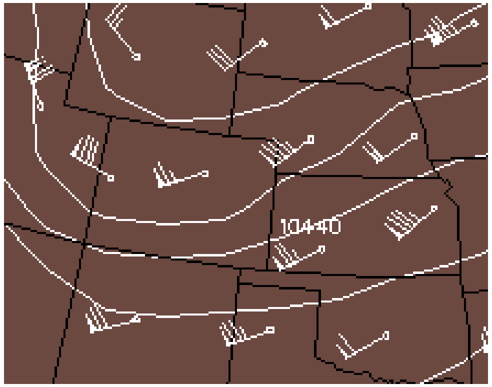


Gradient winds

If there is no turbulent drag, then winds tend to blow parallel to isobar lines or height-contour lines even if those lines are curved. However, if the lines curve around a low-pressure center (in either hemisphere), then the wind speeds are **subgeostrophic** (i.e., slower than the theoretical geostrophic wind speed). For lines curving around high-pressure centers, wind speeds are **supergeostrophic** (faster than theoretical geostrophic winds). These theoretical winds following curved isobars or height contours are known as **gradient winds.**

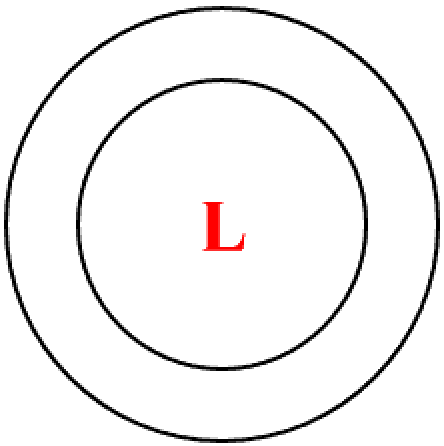
Gradient winds differ from geostrophic winds because Coriolis force F_{CF} and pressure-gradient force F_{PG} do not balance, resulting in a non-zero net force F_{net}. This net force is called centripetal force, and is what causes the wind to continually change direction as it goes around a circle.

Geostrophic winds exist in locations where there are no frictional forces and the isobars are straight. However, such locations are quite rare. Isobars are almost always curved and are very rarely evenly spaced. This changes the geostrophic winds so that they are no longer geostrophic but are instead in **gradient wind balance**. They still blow parallel to the isobars, but are no longer balanced by only the pressure gradient and Coriolis forces, and do not have the same velocity as geostrophic winds.



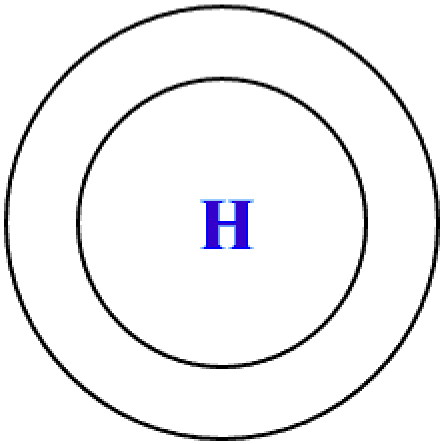
Gradient winds: subgeostrophic

In the diagram below at point A, the parcel of air will move straight north. The pressur gradient and Coriolis force are present, but when the isobars are curved, there is a third force - the centrifugal force. This apparent force, pushes objects away from the center of a circle. The centrifugal force alters the original two-force balance and creates the non-geostrophic gradient wind.
In this case, the centrifugal force acts in the same direction as the Coriolis force. As the parcel moves north, it moves slightly away from the center - decreases the centrifugal force. The pressure gradient force becomes slightly more dominant and the parcel moves back to the original radius. This allows the gradient wind to blow parallel to the isobars.
Since the pressure gradient force doesn't change, and all the forces must balance, the Coriolis force becomes weaker. This in turn decreases the overall wind speed. This is where the gradient wind differs from the geostrophic winds. **In this case of a low pressure system or trough, the gradient wind blows parallel to the isobars at a less than geostrophic (subgeostrophic) speed**.



Gradient winds: supergeostrophic

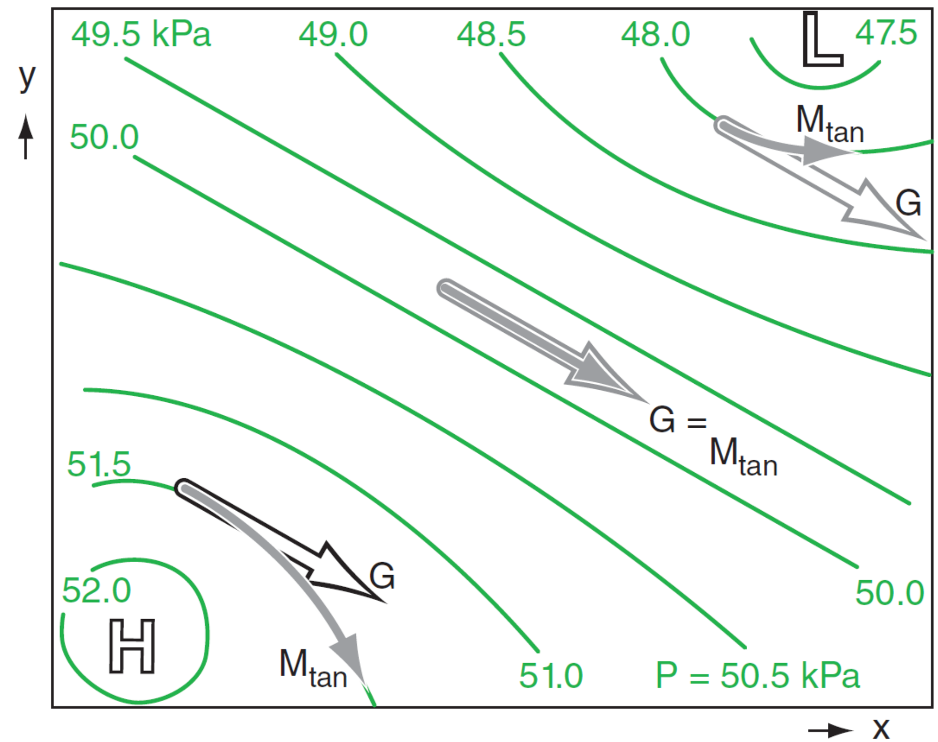
This also applies to high-pressure systems as well. In this case, again starting from point A, the geostrophic wind will blow straight south. This time the centrifugal force is pushing in the same direction as the pressure gradient force, and when it gets slightly further away from the center, the centrifugal force again reduces, but this time that makes the Coriolis Force more dominant and the air parcel will move back to its original radius - again with the end result being wind blowing parallel to the isobars.
Since the pressure gradient force still doesn't change, the Coriolis force must again adjust to balance the forces. However now it becomes stronger, which in turn increases the overall wind speed. **This means that in a high pressure system or ridge, the gradient wind blows parallel to the isobars faster than geostrophic (supergeostrophic) speed.**



By describing this change in direction as causing an apparent force (centrifugal), we can find the equations that define a steady-state gradient wind:

$$0 = -\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta x} + f_c\cdot V + s\cdot\frac{V\cdot M}{R}$$

$$0 = \underbrace{-\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta y}}_{\textit{pressure gradient}} \underbrace{- f_c\cdot U}_{\textit{Coriolis}} \underbrace{- s\cdot\frac{U\cdot M}{R}}_{\textit{centrifugal}}$$



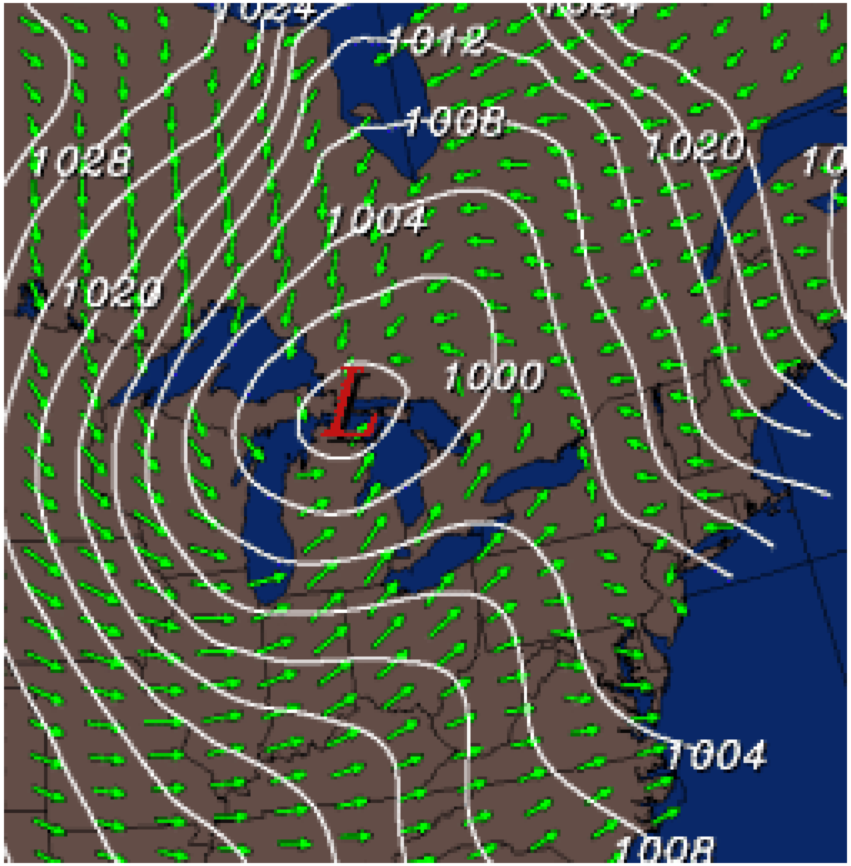
Effect of friction

Winds near the surface, winds affected by friction.

Geostrophic wind blows parallel to the isobars because the Coriolis force and pressure gradient force are in balance. However it should be realized that the actual wind is not always geostrophic - especially near the surface.

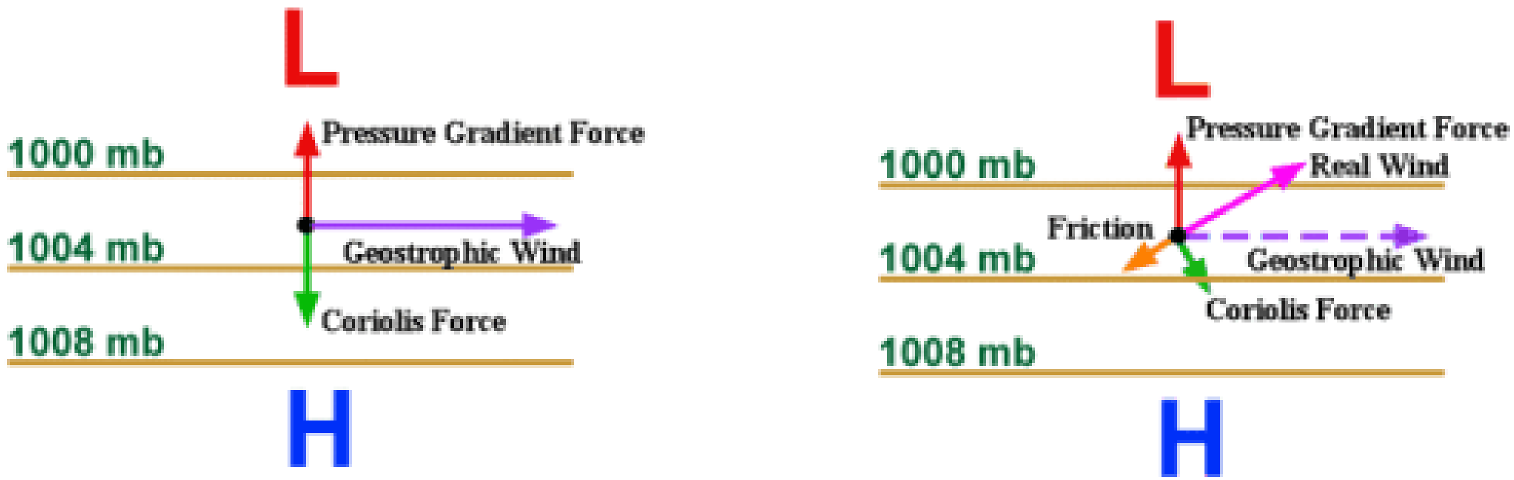
The surface of the Earth exerts a frictional drag on the air blowing just above it. This friction can act to change the wind's direction and slow it down - keeping it from blowing as fast as the wind aloft.

Actually, the difference in terrain conditions directly affects how much friction is exerted. For example, a calm ocean surface is pretty smooth, so the wind blowing over it does not move up, down, and around any features. By contrast, hills and forests force the wind to slow down and/or change direction much more.



Atmospheric-Boundary-Layer winds: effect of friction

As we move higher, surface features affect the wind less until the wind is indeed geostrophic. This level is considered the top of the boundary (or friction) layer.



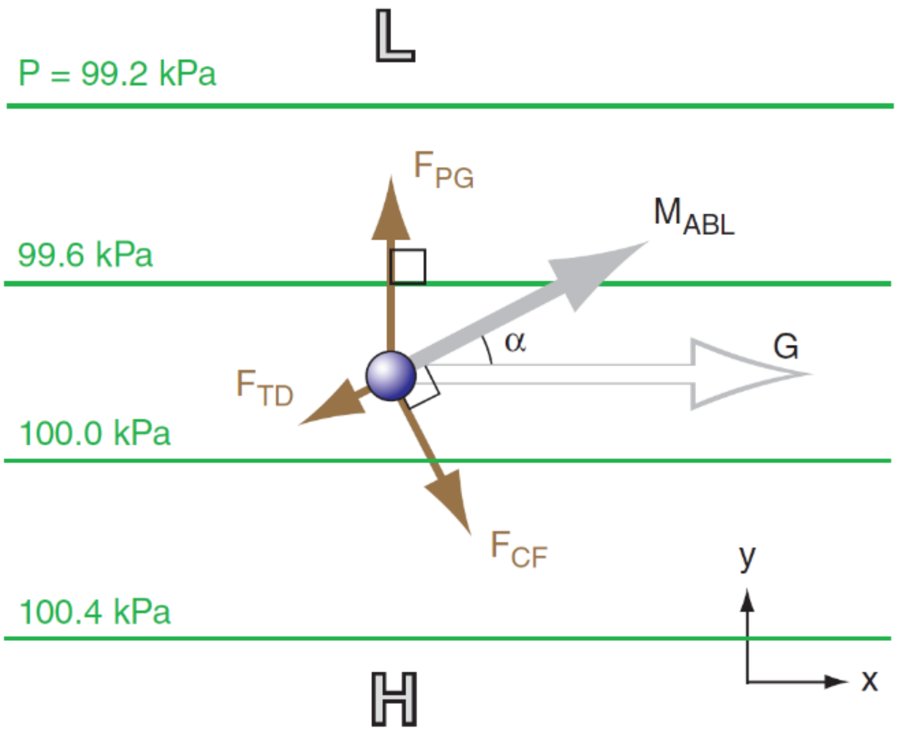
In the friction layer, the turbulent friction that the Earth exerts on the air slows the wind down. This slowing causes the wind to be not geostrophic. As we look at the diagram above, this slowing down reduces the Coriolis force, and the pressure gradient force becomes more dominant. As a result, the total wind deflects slightly towards lower pressure. The amount of deflection the surface wind has with respect to the geostrophic wind above depends on the roughness of the terrain. Meteorologists call the difference between the total and geostrophic winds **ageostrophic winds**.

Atmospheric-Boundary-Layer winds : effect of friction

If you add turbulent drag to winds that would have been geostrophic, the result is a **subgeostrophic** (slower-than-geostrophic) wind that crosses the isobars at angle (α) (Figure). This condition is found in the atmospheric boundary layer (ABL) where the isobars are straight. The force balance at steady state is:

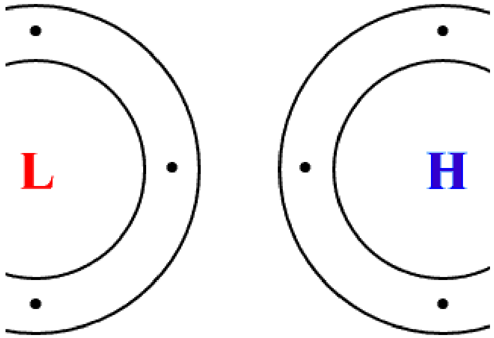
$$0 = -\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta x} + f_c \cdot V - w_T \cdot \frac{U}{z_i}$$

$$0 = -\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta y} - f_c \cdot U - w_T \cdot \frac{V}{z_i}$$



Atmospheric-Boundary-Layer Gradient winds: effect of friction

If we look at low and high-pressure systems, we can see this mechanism at work. Here in this exmple below, the winds would, without friction effects, be moving counter-clockwise around the center of the low in the northern hemisphere. However, when the surface friction is accounted for, the wind slows down, and therefore the Coriolis force weakens and the pressure gradient force becomes dominant, resulting in the spiraling of air into the center of a low pressure system and away from the center of the high pressure system. This causes convergence in the center of the low pressure system at the surface. It is this surface convergence which leads to rising air which can create clouds and even cause rain and storms to form.



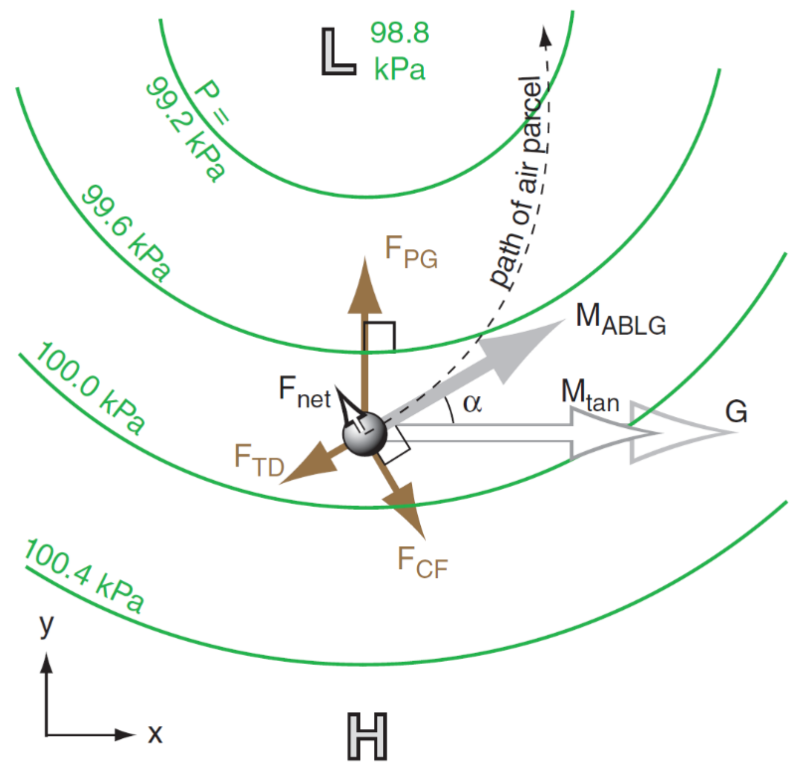
At the same time, wind flows around a northern hemisphere high-pressure system in a clockwise manner, but when frictional effects are introduced the wind again slows down, and the Coriolis force reduces and the pressure gradient force becomes dominant. In this case, though, the pressure gradient is outward from the center of the high, so the result is that surface wind spirals away from the center. This causes divergence in the center of the high pressure system at the surface. This surface divergence causes sinking motion which supresses cloud development and gives us clear skies.

Atmospheric-Boundary-Layer Gradient winds: effect of friction

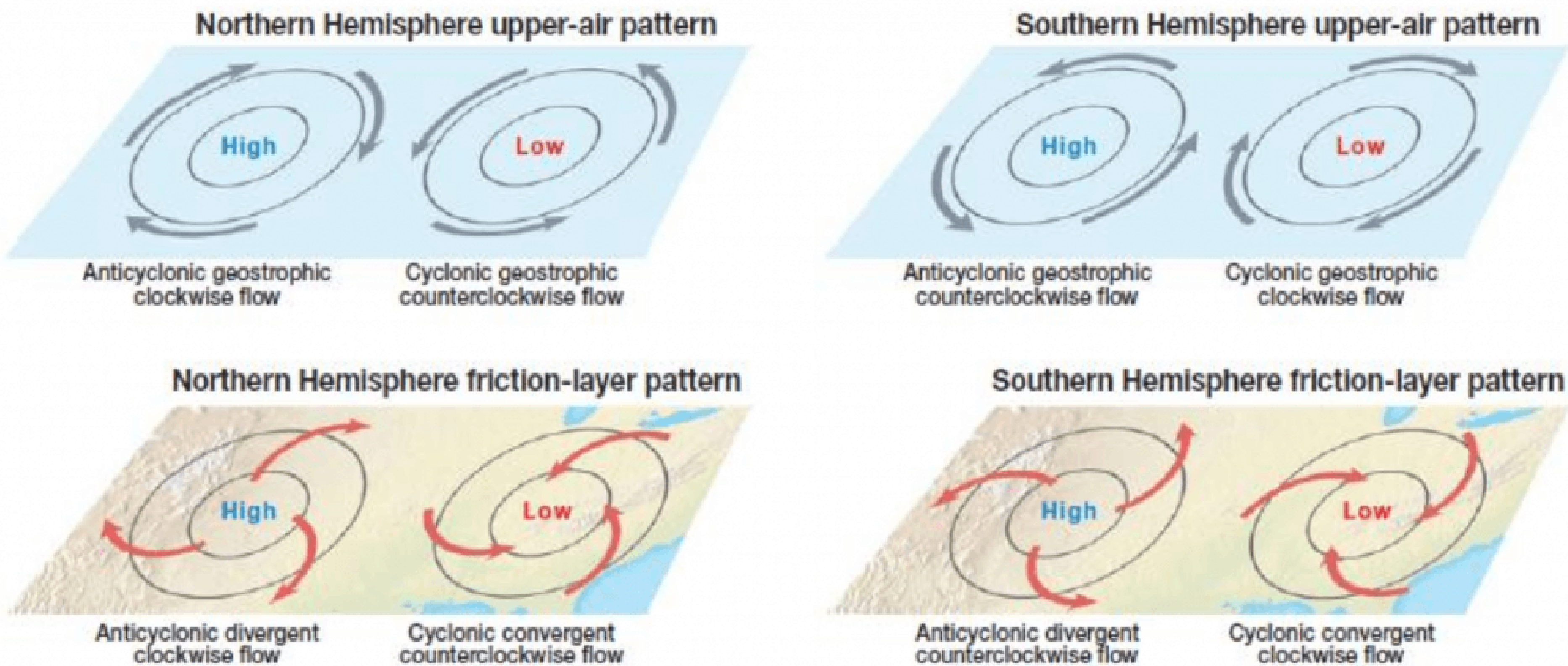
For curved isobars in the atmospheric boundary layer (ABL), there is an imbalance of the following forces: Coriolis, pressure-gradient, and drag. This imbalance is a centripetal force that makes ABL air spiral outward from highs and inward toward lows

$$0 = -\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta x} + f_c \cdot V - w_T \cdot \frac{U}{z_i} + s \cdot \frac{V \cdot M}{R}$$

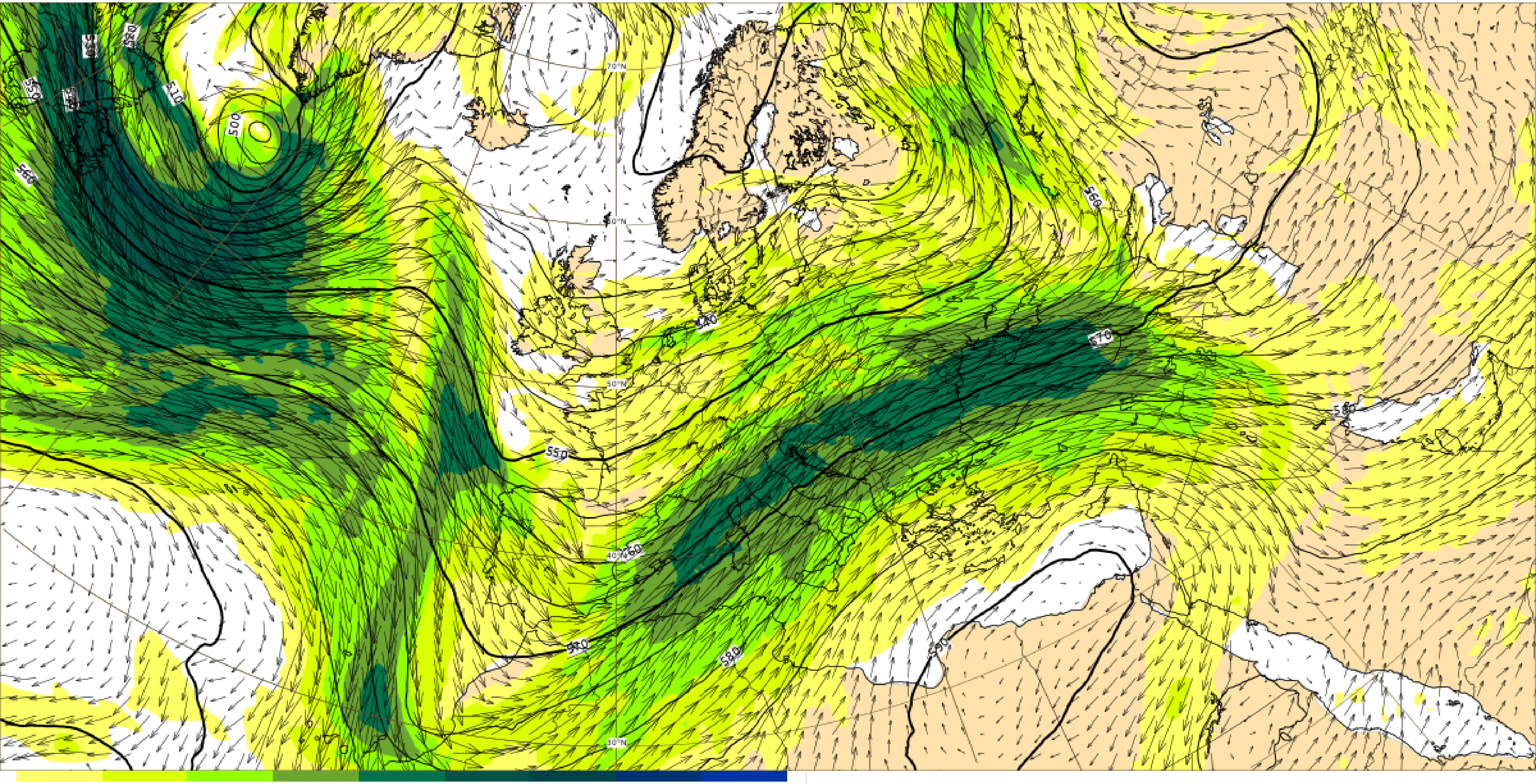
$$0 = \underbrace{-\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta y}}_{\text{pressure gradient}} \underbrace{- f_c \cdot U}_{\text{Coriolis}} \underbrace{- w_T \cdot \frac{V}{z_i}}_{\text{turbulent drag}} \underbrace{- s \cdot \frac{U \cdot M}{R}}_{\text{centrifugal}}$$



Effect of friction



Today's weather situation



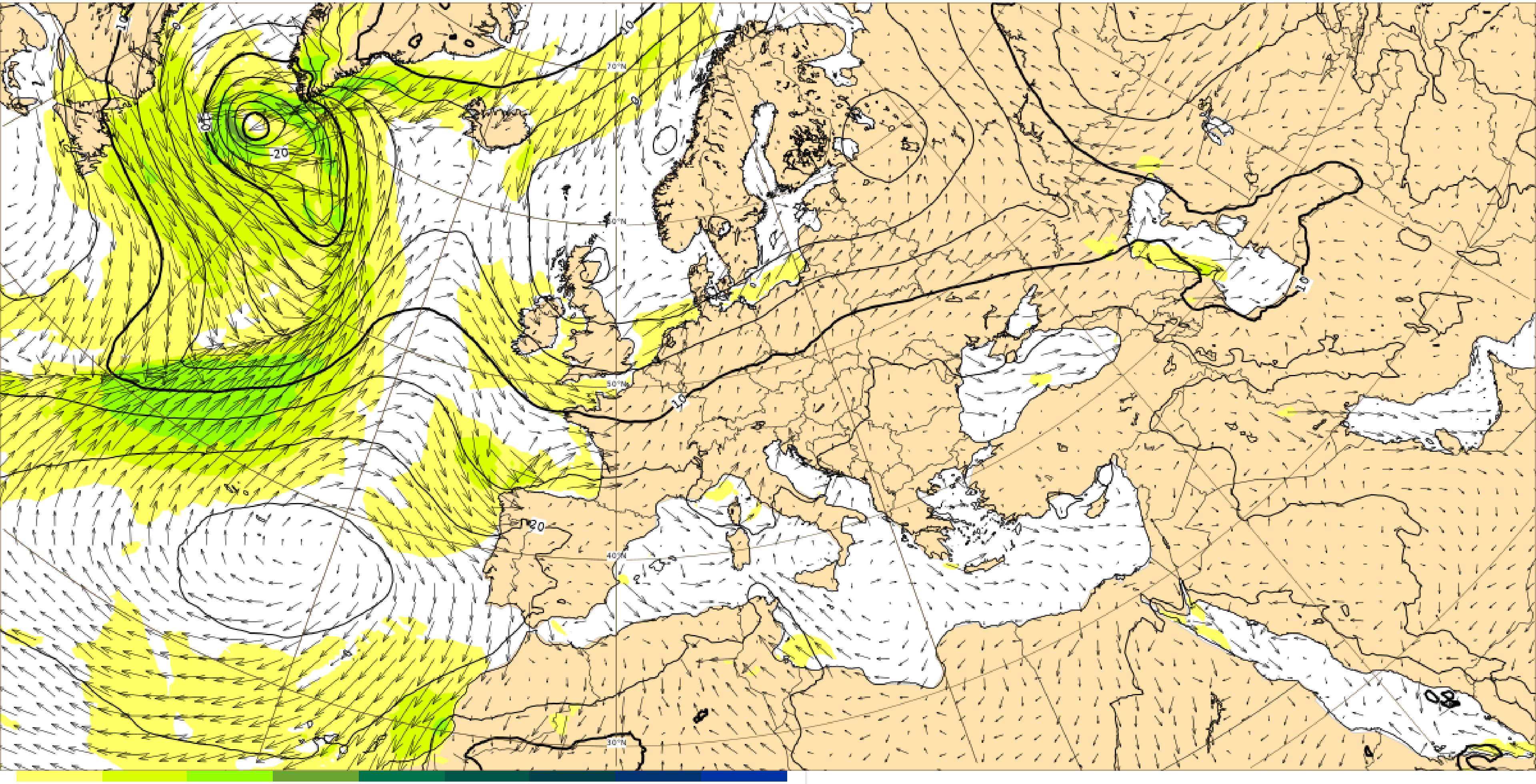
Geopotential at 500 hPa (dam)

Wind at 500 hPa ()

Black wind arrows

500 hPa

Today's weather situation



Geopotential at 500 hPa (dam)

Wind at 500 hPa ()

Black wind arrows

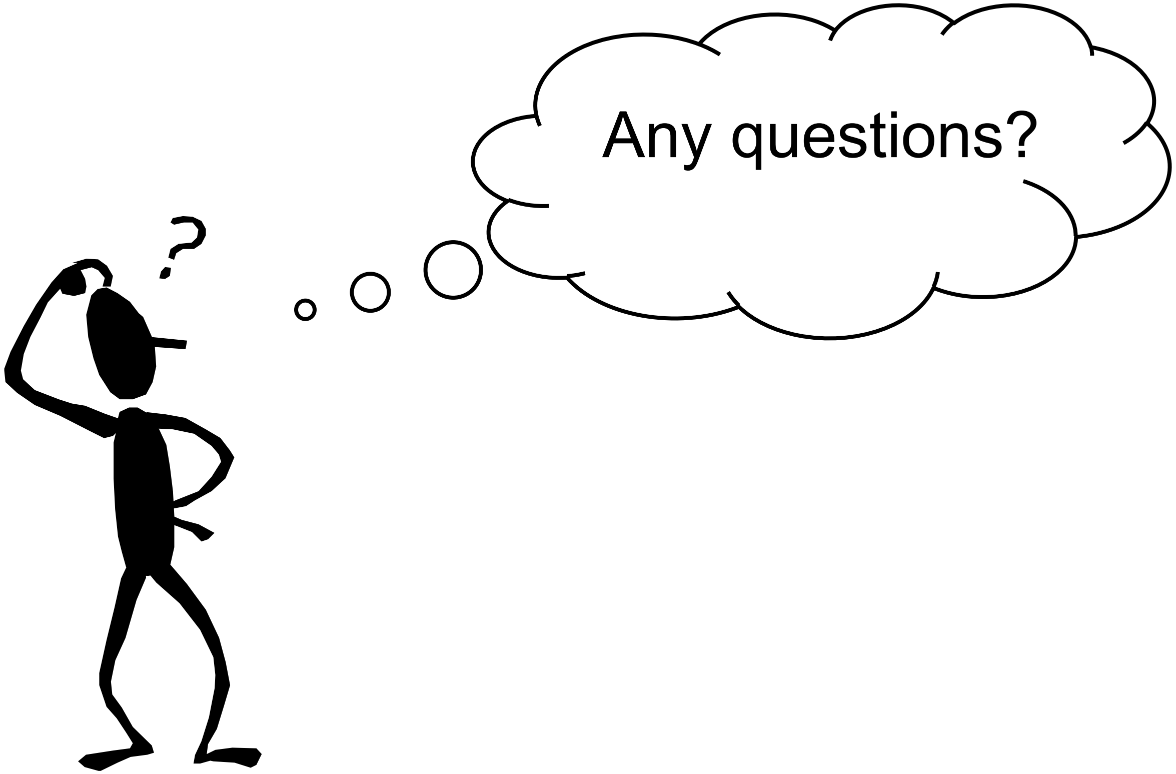
1000 hPa

Summary of horizontal winds

Table 10-5. Summary of horizontal winds**.

Item	Name of Wind	Forces	Direction	Magnitude	Where Observed
1	**geostrophic**	pressure-gradient, Coriolis	parallel to straight isobars with Low pressure to the wind's left*	faster where isobars are closer together. $G = \left\| \frac{g}{f_c} \cdot \frac{\Delta z}{\Delta d} \right\|$	aloft in regions where isobars are nearly straight
2	**gradient**	pressure-gradient, Coriolis, centrifugal	similar to geostrophic wind, but following curved isobars. Clockwise* around Highs, counterclockwise* around Lows.	slower than geostrophic around Lows, faster than geostrophic around Highs	aloft in regions where isobars are curved
3	**atmospheric boundary layer**	pressure-gradient, Coriolis, drag	similar to geostrophic wind, but crosses isobars at small angle toward Low pressure	slower than geostrophic (i.e., **subgeostrophic**)	near the ground in regions where isobars are nearly straight
4	**atmospheric boundary-layer gradient**	pressure-gradient, Coriolis, drag, centrifugal	similar to gradient wind, but crosses isobars at small angle toward Low pressure	slower than gradient wind speed	near the ground in regions where isobars are curved
5	**cyclostrophic**	pressure-gradient, centrifugal	either clockwise or counterclockwise around strong vortices of small diameter	stronger for lower pressure in the vortex center	tornadoes, waterspouts (& sometimes in the eye-wall of hurricanes)
6	**inertial**	Coriolis, centrifugal	anticyclonic circular rotation	coasts at constant speed equal to its initial speed	ocean-surface currents

** For Northern Hemisphere. Direction is opposite in Southern Hemisphere. ** Antitriptic winds are unphysical; not listed here.*


Any questions?