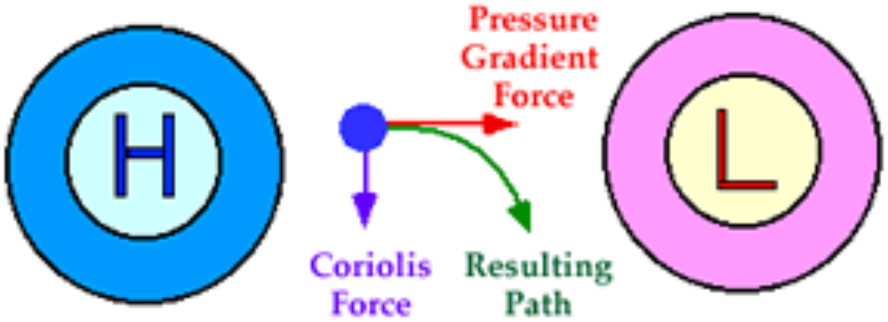
Surface winds

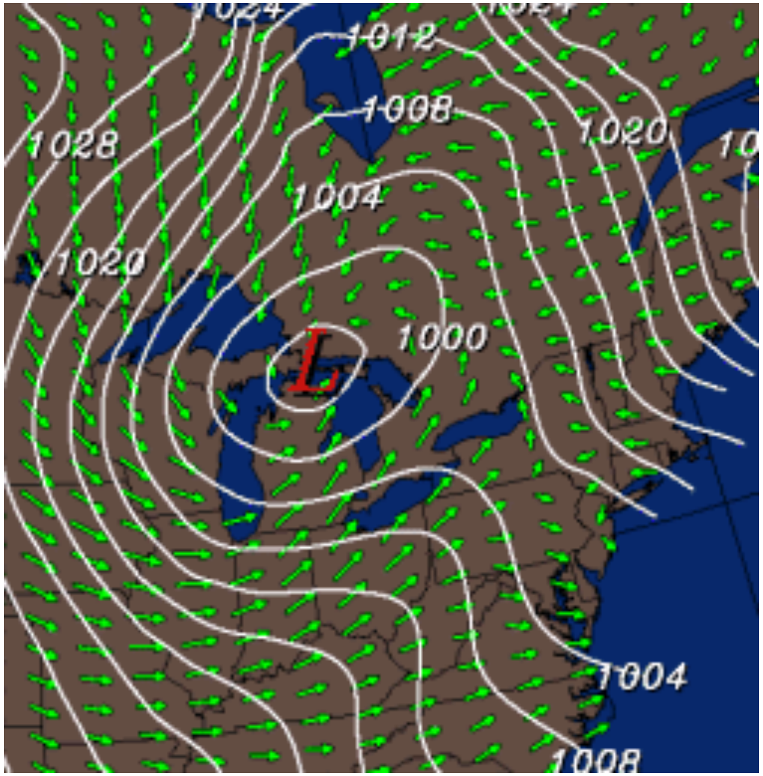
An air parcel initially at rest will move from high pressure to low pressure (**pressure gradient force**)



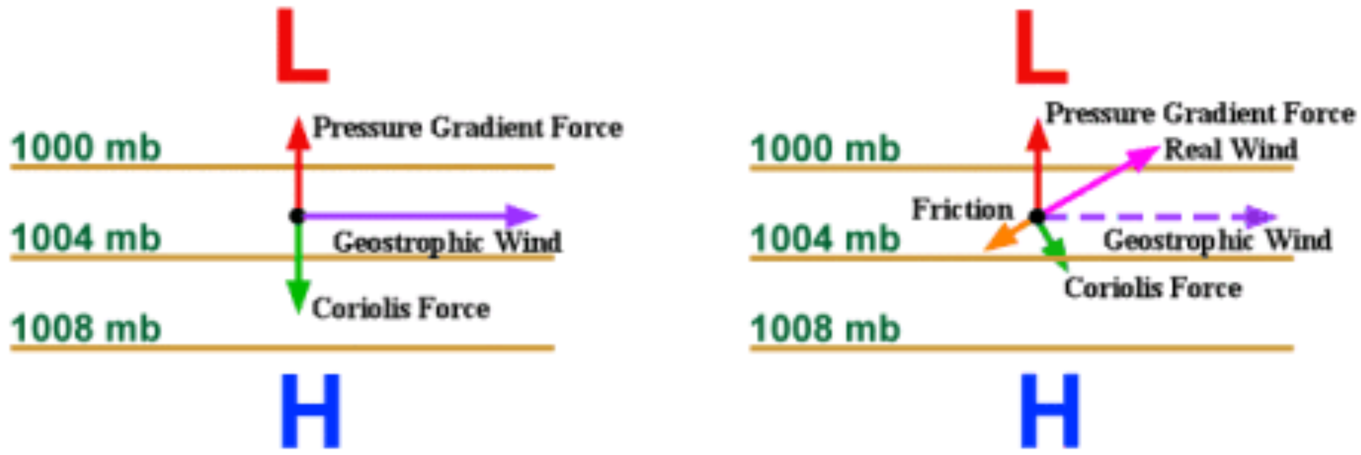
Geostrophic wind blows parallel to the isobars because the **Coriolis force** and **pressure gradient force** are in balance.

Close to geostrophy, but

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.



Effect of friction



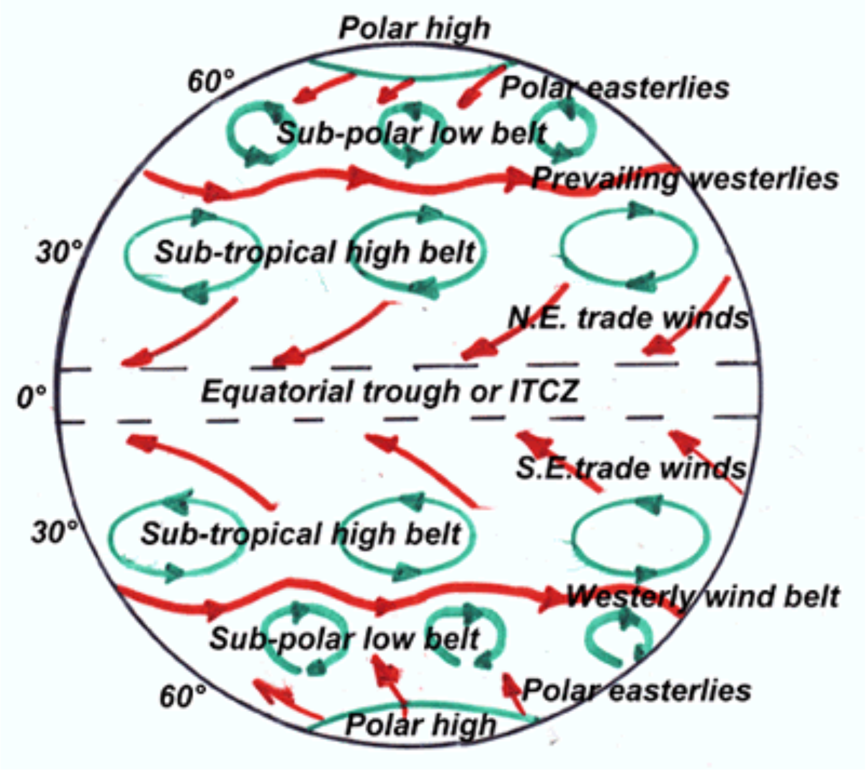
In the friction layer, the **turbulent friction slows the wind down**

This slowing causes the **wind to be not geostrophic**

This **reduces the Coriolis force**

pressure gradient force becomes more dominant

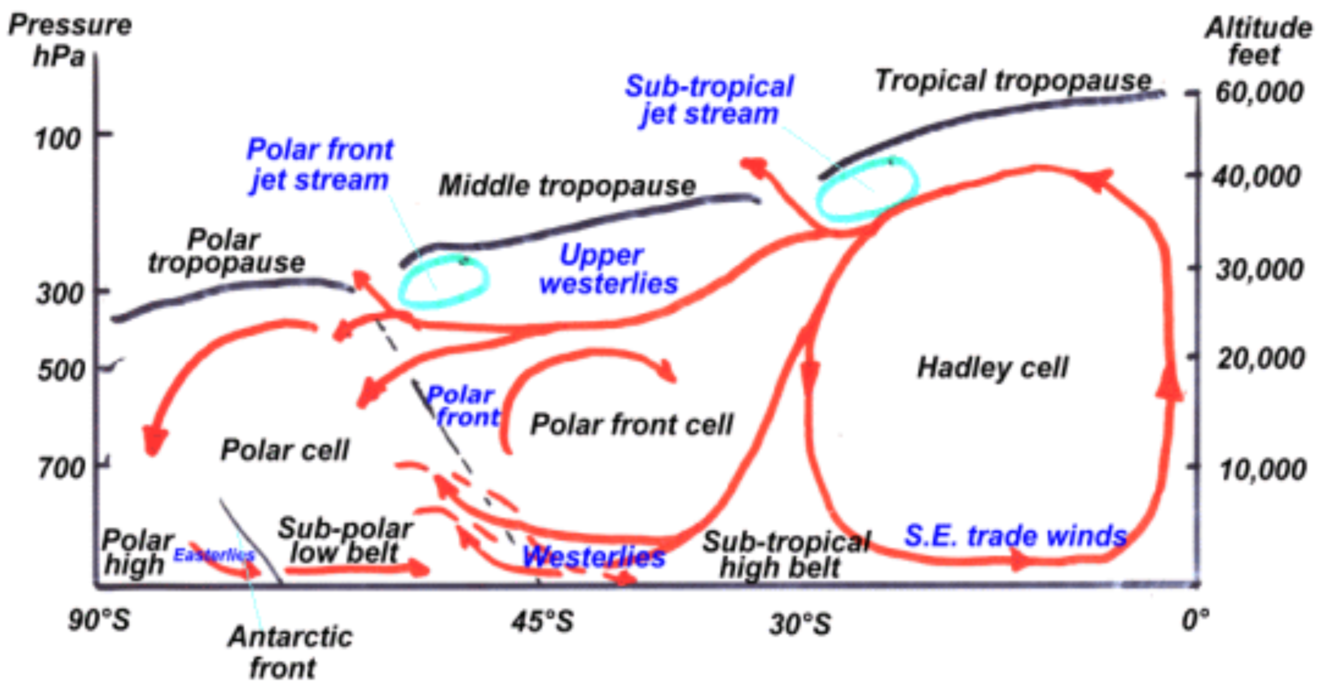
General global circulation



The atmosphere is rotating in the same direction as the Earth:

westerly winds move faster and easterly winds move slower than the Earth's surface.

Tropospheric Circulation

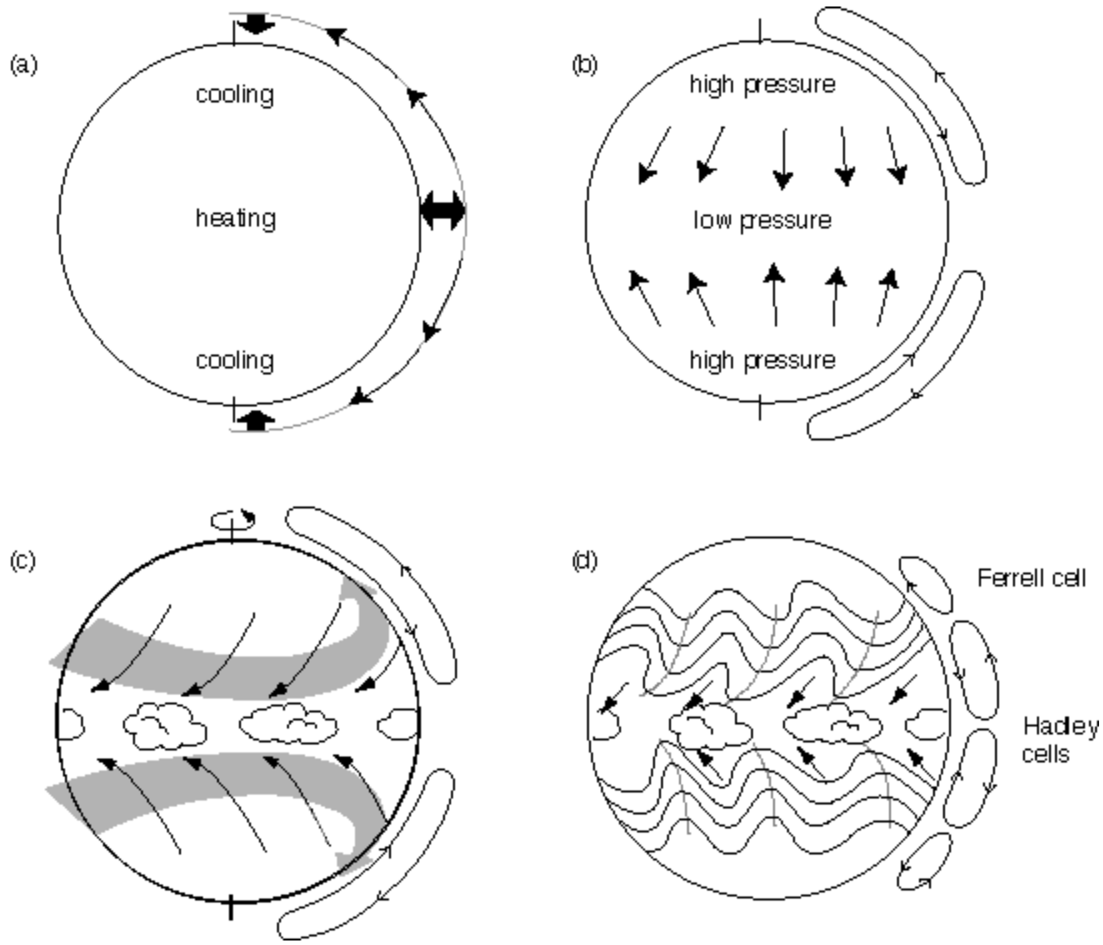


Intertropical Convergence Zone (ITCZ) and the Hadley cell

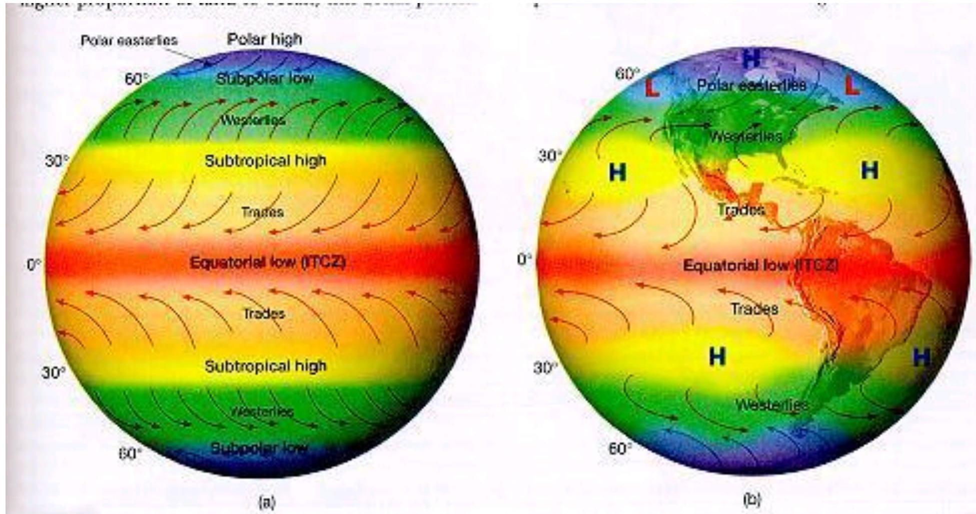
George Hadley [1685-1768], a British meteorologist, fromulated trade wind theory

50% of the Earth's surface 30° N - 30° S: Hadley cells directly affect half the globe

Why doesn't the Earth's circulation have a single cell?



circulation



(a) An imaginary uniform Earth with idealized zonal (continuous) pressure belts. (b) The real Earth with disruptions of the zonal pattern caused by large landmasses. These disruptions break up pressure zones into semipermanent high- and low-pressure cells.

H and L

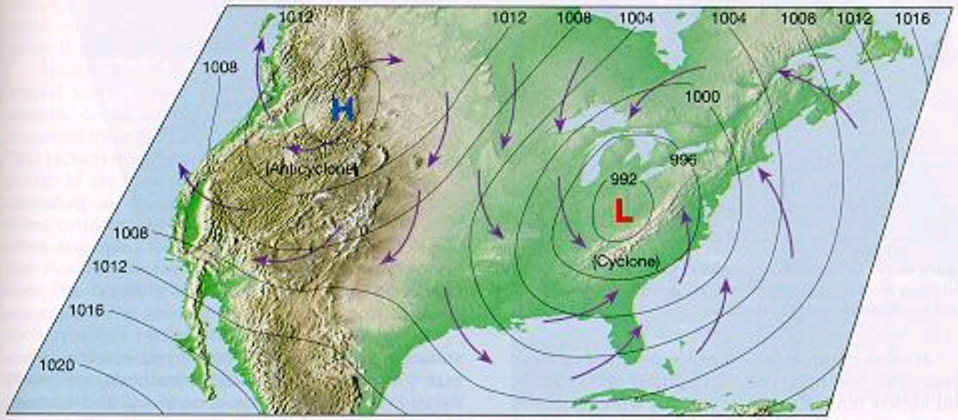


Figure 6–17 Cyclonic and anticyclonic winds in the Northern Hemisphere. Arrows show the winds blowing inward and counterclockwise around a low, and outward and clockwise around a high.

H and L

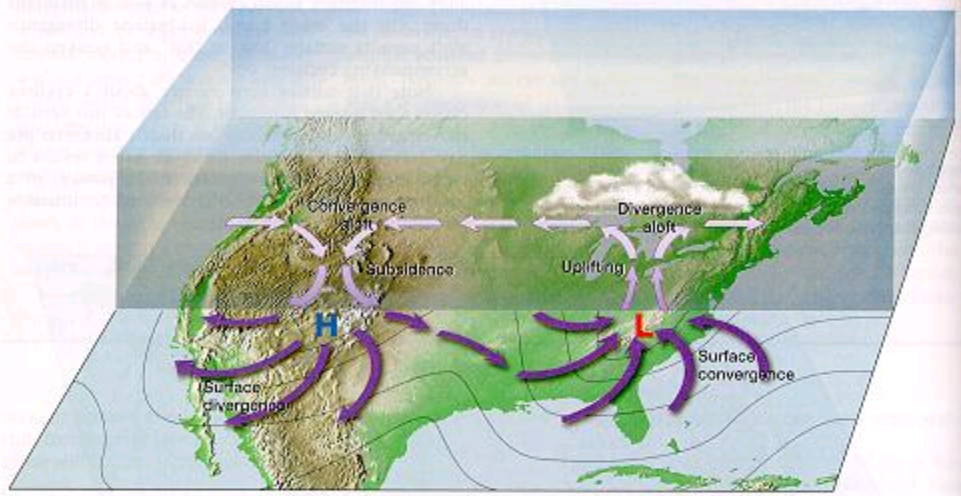
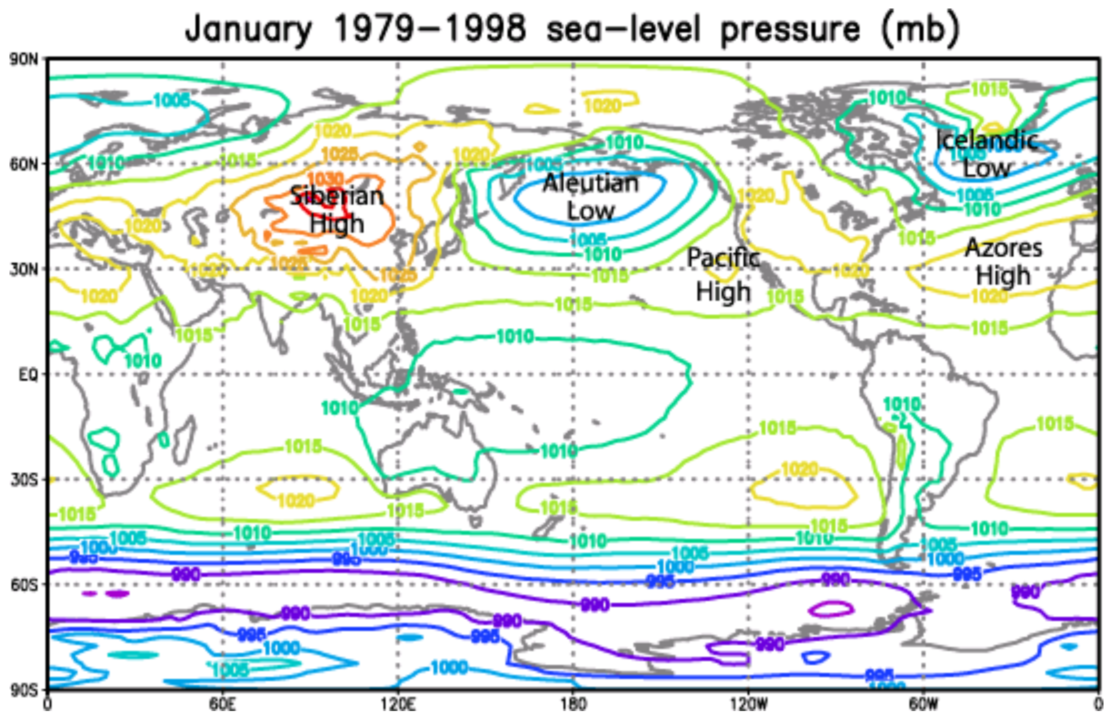


Figure 6–19 Airflow associated with surface cyclones and anticyclones. A low, or cyclone, has converging surface winds and rising air causing cloudy conditions. A high, or anticyclone, has diverging surface winds and descending air, which leads to clear skies and fair weather.

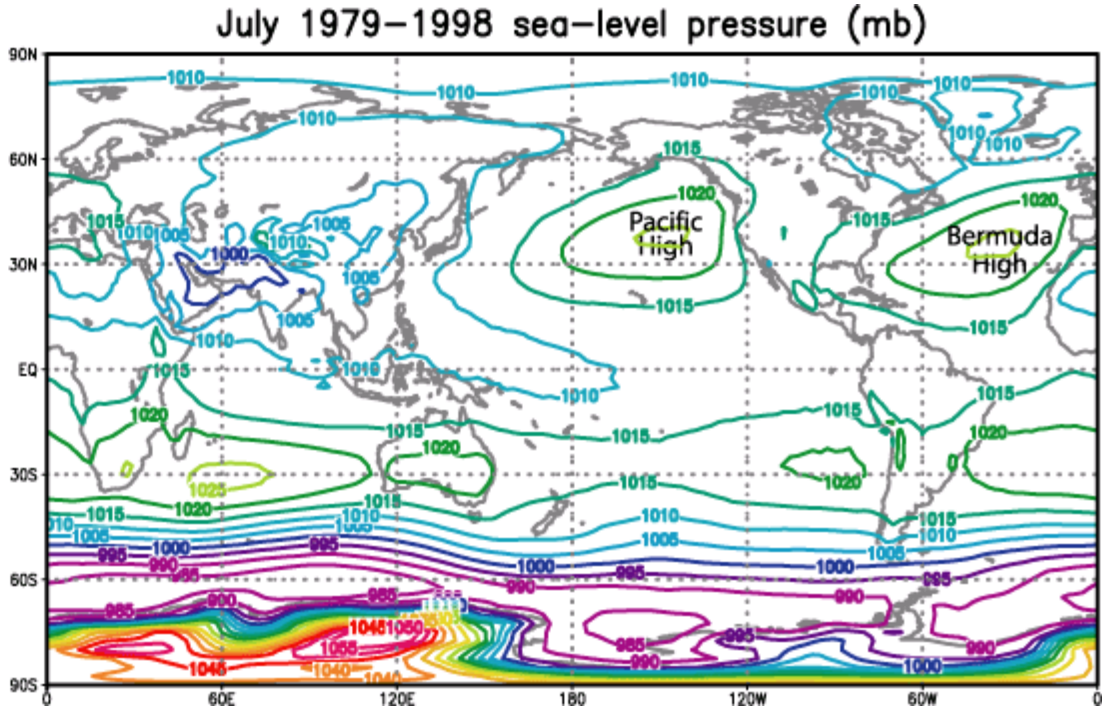
The pressure patterns are not stationary.

They change daily and seasonally.

Climatology

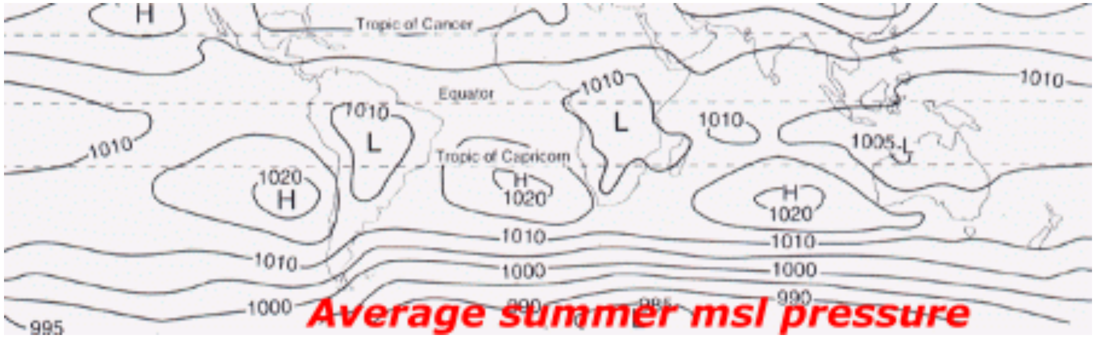


Climatology



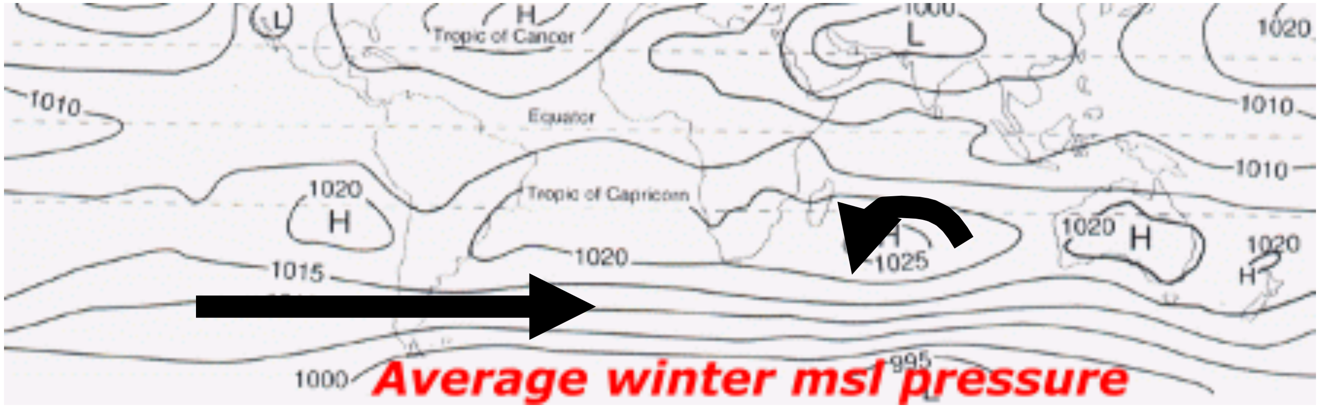
The sub-tropical anticyclones

The subsiding high level air of the Hadley cell: persistent sub-tropical high pressure belt, or ridge, encircling the globe between 30° S - 50° S. Within the belt there are three semi-permanent year-round high pressure centres in the Indian, Pacific and South Atlantic Oceans.

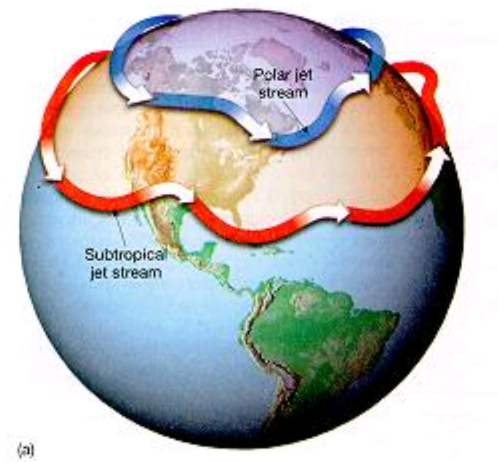


The sub-tropical anticyclones

In winter the high pressure belt moves northward.



Rossby waves and the westerly wind belt



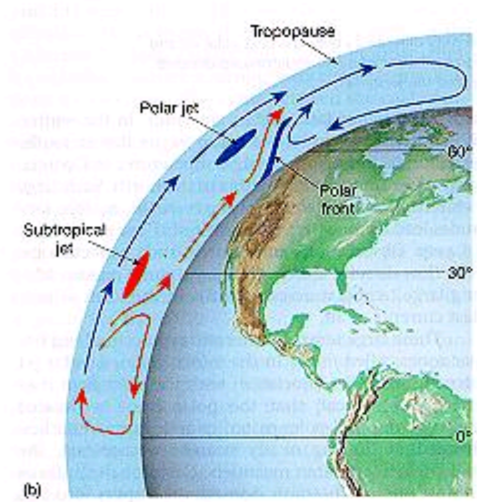
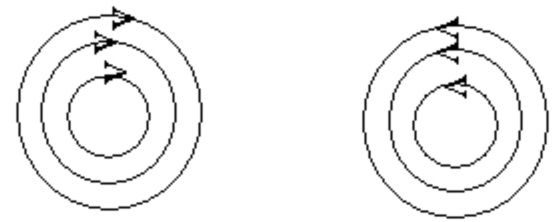


Figure 7–10 Jet streams. (a) Approximate positions of the polar and subtropical jet streams. Note that these fast-moving currents are generally not continuous around the entire globe. (b) A cross-sectional view of the polar and subtropical jets.

The jet stream is closely linked to the position of **Rossby waves**.

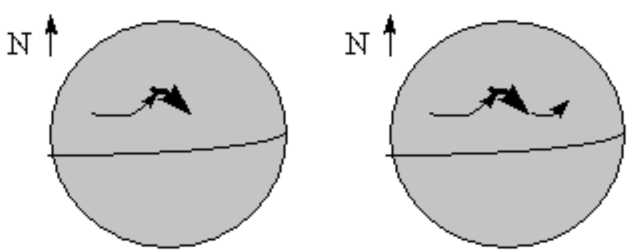
Rossby waves

Vorticity - the tendency to spin about an axis



negative vorticity (anticyclones in NH)   positive vorticity (cyclones in NH)

On the spinnng Earth there is vorticity from the Earth's spin (**planetary vorticity**) and local vorticity due to cyclonic/anticyclonic behaviour (**relative vorticity**)
The absolute vorticity is conserved: *zeta + f = constant*   *(f = 2 \omega sin \phi)*



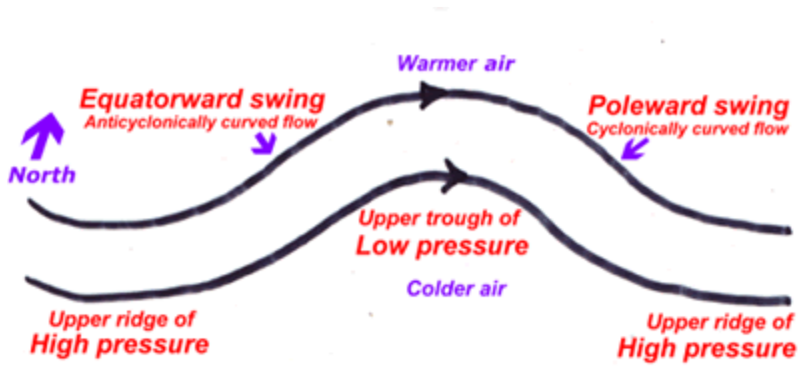
f increases => ζ decreases   f decreases => ζ increases

Oscillations: Rossby waves

Topographic Rossby waves: standing wave fixed to a permanent forcing location

Rossby waves and the westerly wind belt

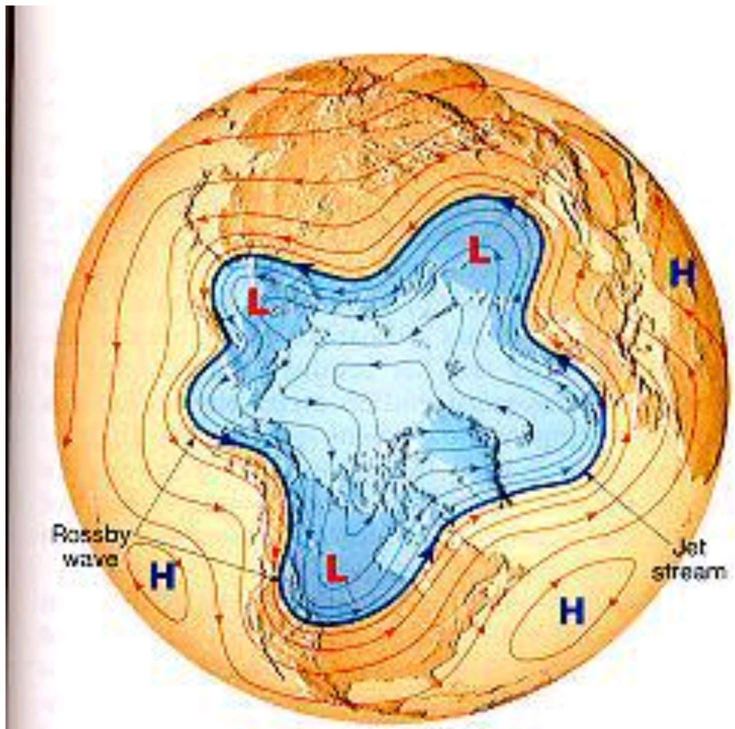
large-scale meanders of the mid-latitude jet stream.
Here Southern Hemisphere



Extend: 30° of latitude

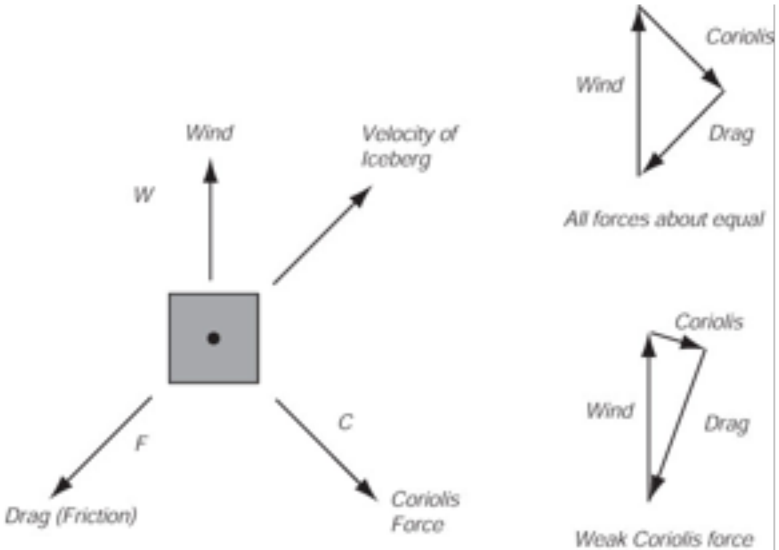
Zeta + f = const.

Atmosphere: Large-scale meanders



Idealized air flow of the westerlies at the 500-millibar level. The five long-wavelength undulations, called *Rossby waves*, compose this flow. The jet stream is the fast core of this wavy flow.

Nansen's Qualitative Arguments



Fridtjof Nansen noticed that wind tended to blow icebergs 20° – 40° to the right of the wind in the Arctic.

Nansen argued that three forces must be important:

1) Wind Stress W

2) Friction F (otherwise the iceberg would move as fast as the wind)
Drag must be opposite the direction of the ice's velocity

3) Coriolis Force C.
Coriolis force must be perpendicular to the velocity

The forces must balance for steady flow: **W + F + C =** 0

Fridtjof Wedel-Jarlsberg Nansen (10 October 1861 – 13 May 1930) was a Norwegian explorer, scientist, diplomat, humanitarian and Nobel laureate.

In his youth a champion skier and ice skater, he led the first crossing of the Greenland interior in 1888, and won international fame after reaching a record northern latitude of 86° 13' during his North Pole expedition of 1893–96.

In the final decade of his life Nansen devoted himself primarily to the League of Nations, following his appointment in 1921 as the League's High Commissioner for Refugees. In 1922 he was awarded the **Nobel Peace Prize** for his work on behalf of the displaced victims of the First World War and related conflicts.

Fridtjof Nansen's 1888 route across Greenland

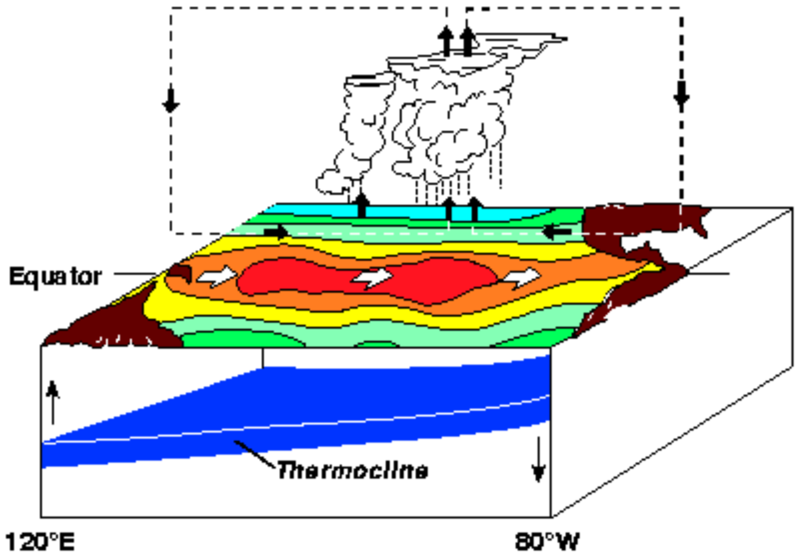
Blue: Dotted line is the ship *Jason*'s journey from Iceland to near Sermilik fjord continuous blue: two small boats trying to reach the coast.

Red: Planned journey from Sermilik northwest to Christianhaab (today known as Qasigiannguit).

Green: Nansen's actual journey across Greenland from Umivik fjord to Gothaab (Nuuk).

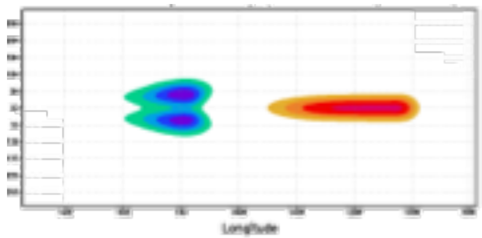
Film2

El Niño Conditions

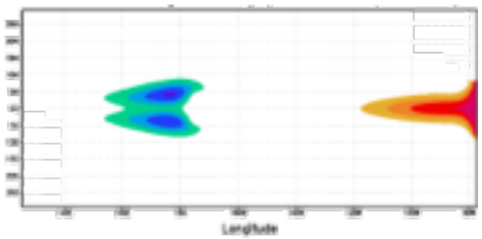


Kelvin wave (3 m/s, 70 days) Rossby wave (1m/s, 210 days)

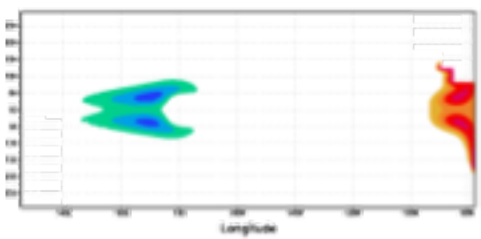
25 days



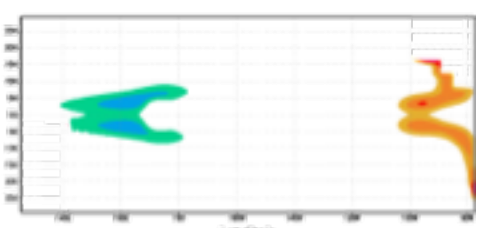
50 days



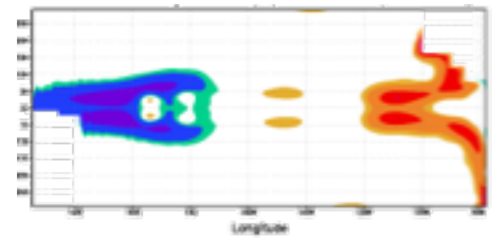
75 days



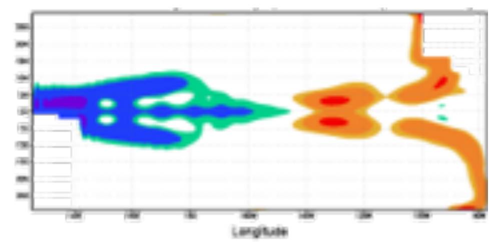
100 days



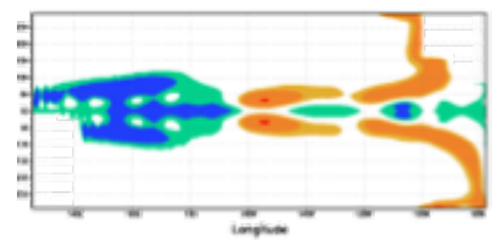
125 days



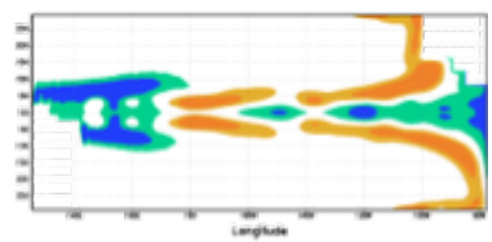
175 days



225 days



275 days



sea surface height anomalies

Ekman Mass Transport

- Integral of the Ekman Velocities down to a depth d:

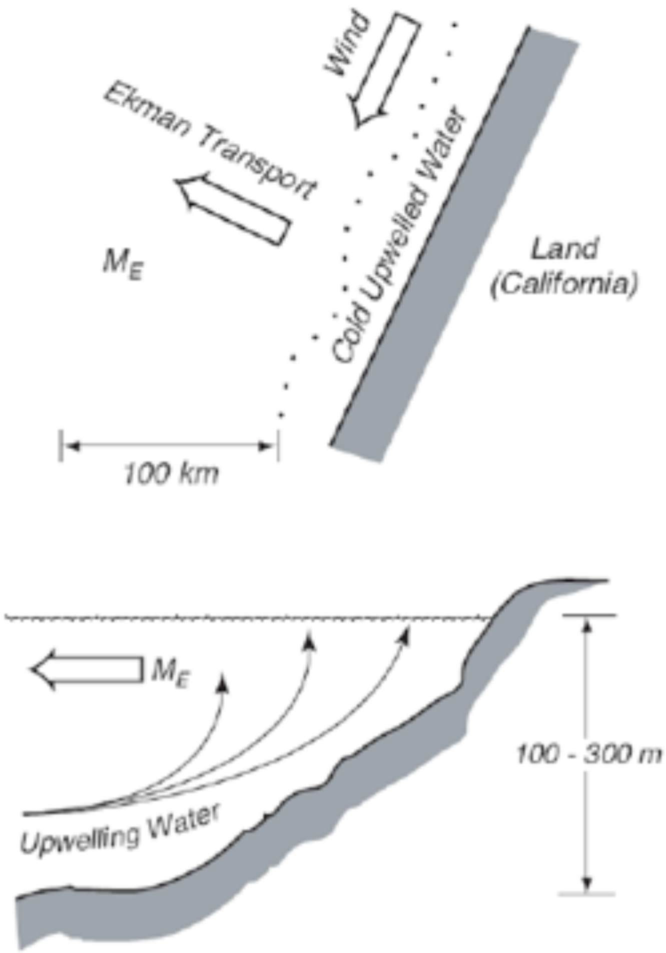
$$M_{Ex} = \int_{-d}^{0} \rho U_E \, dz, \qquad M_{Ey} = \int_{-d}^{0} \rho V_E \, dz$$

- Ekman transport relates the surface wind stress:

$$f\, M_{Ey} = -T_{xz}(0)$$
$$f\, M_{Ex} = \quad T_{yz}(0)$$

- Mass transport is perpendicular to wind stress
- In the northern hemisphere, *f* is positive, and the mass transport is in the *x* direction, to the east.

Coastal Upwelling



Courtesy of Prof. Robert Stewart. Used with permission.

Source: Introduction to Physical Oceanography, http://oceanworld.tamu.edu/home/course_book.htm

- Upwelling enhances biological productivity, which feeds fisheries.
- Cold upwelled water alters local weather. Weather onshore of regions of upwelling tend to have fog, low stratus clouds, a stable stratified atmosphere, little convection, and little rain.
- Spatial variability of transports in the open ocean leads to upwelling and downwelling, which leads to redistribution of mass in the ocean, which leads to wind-driven geostrophic currents via *Ekman pumping*.

Ekman Pumping

- The horizontal variability of the wind blowing on the sea surface leads to horizontal variability of the Ekman transports.
- Because mass must be conserved, the spatial variability of the transports must lead to vertical velocities at the top of the Ekman layer.
- To calculate this velocity, we first integrate the continuity equation in the vertical direction:

$$\rho \int_{-d}^{0} \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) dz = 0$$

$$\frac{\partial}{\partial x} \int_{-d}^{0} \rho\, u\, dz + \frac{\partial}{\partial y} \int_{-d}^{0} \rho\, v\, dz = -\rho \int_{-d}^{0} \frac{\partial w}{\partial z}\, dz$$

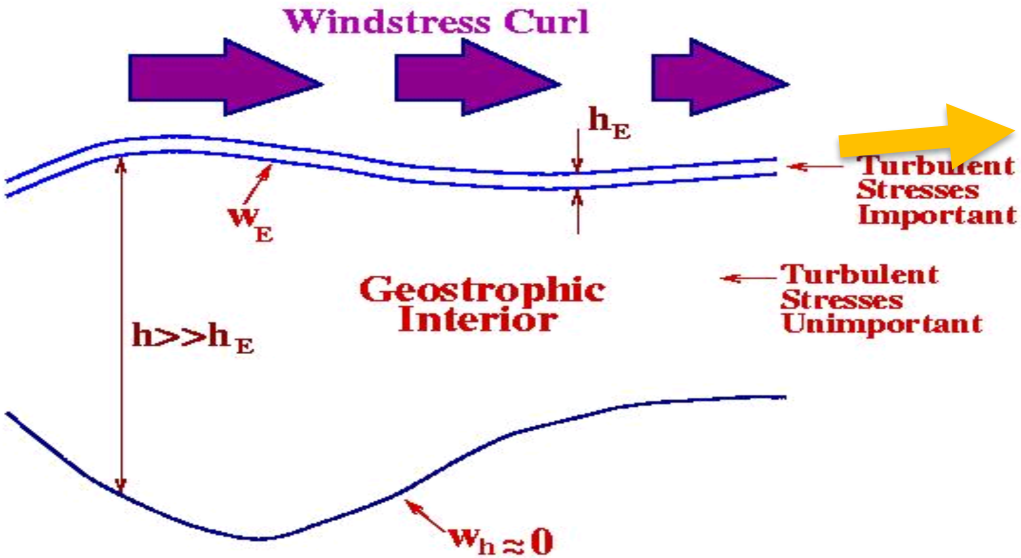
$$\frac{\partial M_{Ex}}{\partial x} + \frac{\partial M_{Ey}}{\partial y} = -\rho \left[w(0) - w(-d) \right]$$

Ekman v. Reality

- Inertial currents dominate
- Flow is nearly independent of depth within the mixed layer on time periods on the order of the inertial period (i.e. the mixed layer moves like a slab)
- Current shear is strongest at the top of the thermocline
- Flow averaged over many inertial periods is almost exactly that calculated by Ekman
- Ekman depth is typically on target with experiments, but velocities are often as much as half the calculated value
- Angle between wind and flow at surface depends on latitude and is near 45 degrees at mid-latitudes

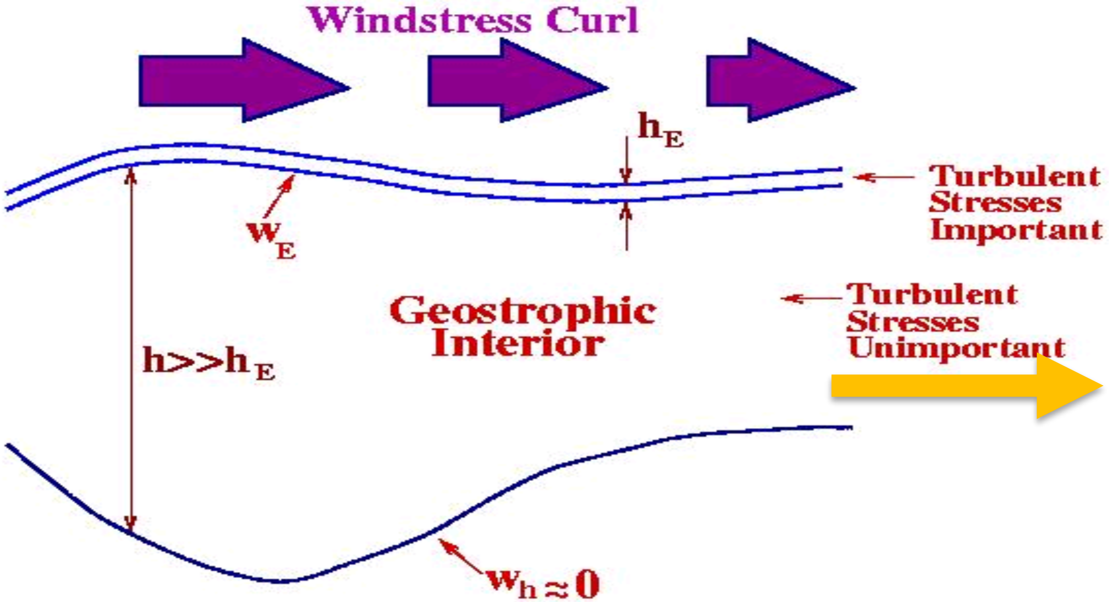
Ekman

- the vertical flow from the surface Ekman layer into the geostrophic interior is



$$w_E = \frac{1}{\rho_0 f}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right)$$

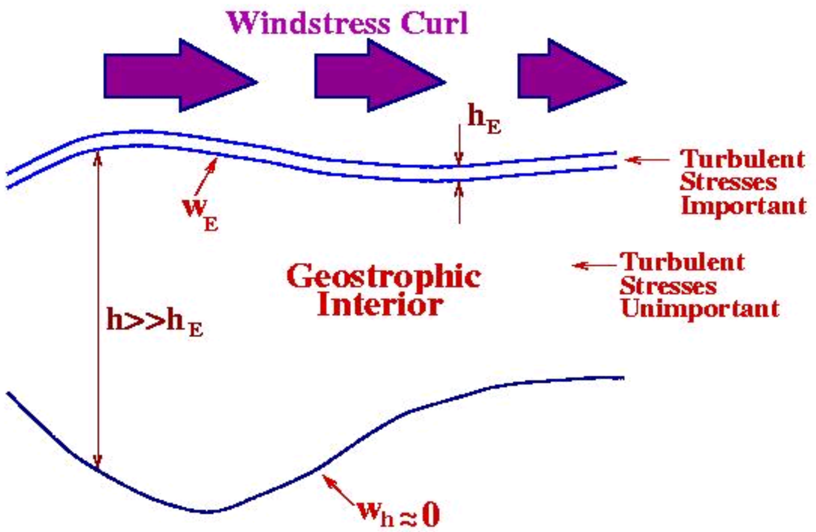
Sverdrup Balance



$$\beta \int_{\text{bot}}^{\text{top}} v \, dz = f \, w\Big|_{\text{bot}}^{\text{top}}$$

Geostrophic balance

Sverdrup Balance



$$V \equiv \int_{\text{bot}}^{\text{top}} v \, dz$$

$$w_{\text{top}} = w_E = \frac{1}{\rho_0 f}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right)$$

$$\beta V = \frac{1}{\rho_0}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right)$$

… relates the integral meridional flow *throughout the vertical extent* of the treated layer to the local windstress curl.

Sverdrup Balance

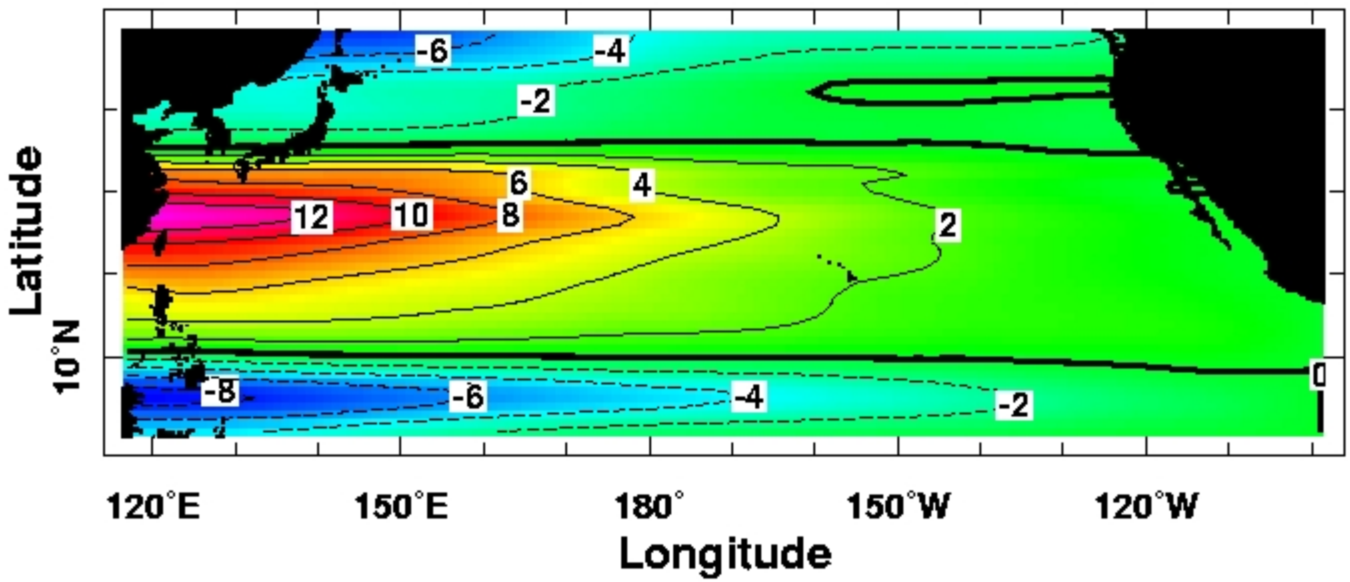
$$\beta V = \frac{1}{\rho_0}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right)$$

Sverdrup Stream function

$$\frac{\partial \psi}{\partial x} = \frac{1}{\beta \rho_0}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right)$$

$$\psi(x\ ) = \frac{1}{\beta \rho_0}\int_{x_E}^{x}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right)\, dx$$

$$\psi(x,y) = \frac{1}{\beta\rho_0}\int_{x_E}^{x}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right)\, dx$$



Being that the curl is negative throughout the subtropics, it follows that the meridional flux must be everywhere equatorward. But such a situation, if sustained, will progressively empty the midlatitude oceans, while piling-up more and more water along the Equator; a clear physical impossibility!

There must be somewhere a return poleward flow that `drains' the Equatorial region while replenishing the midlatitude missing volume.

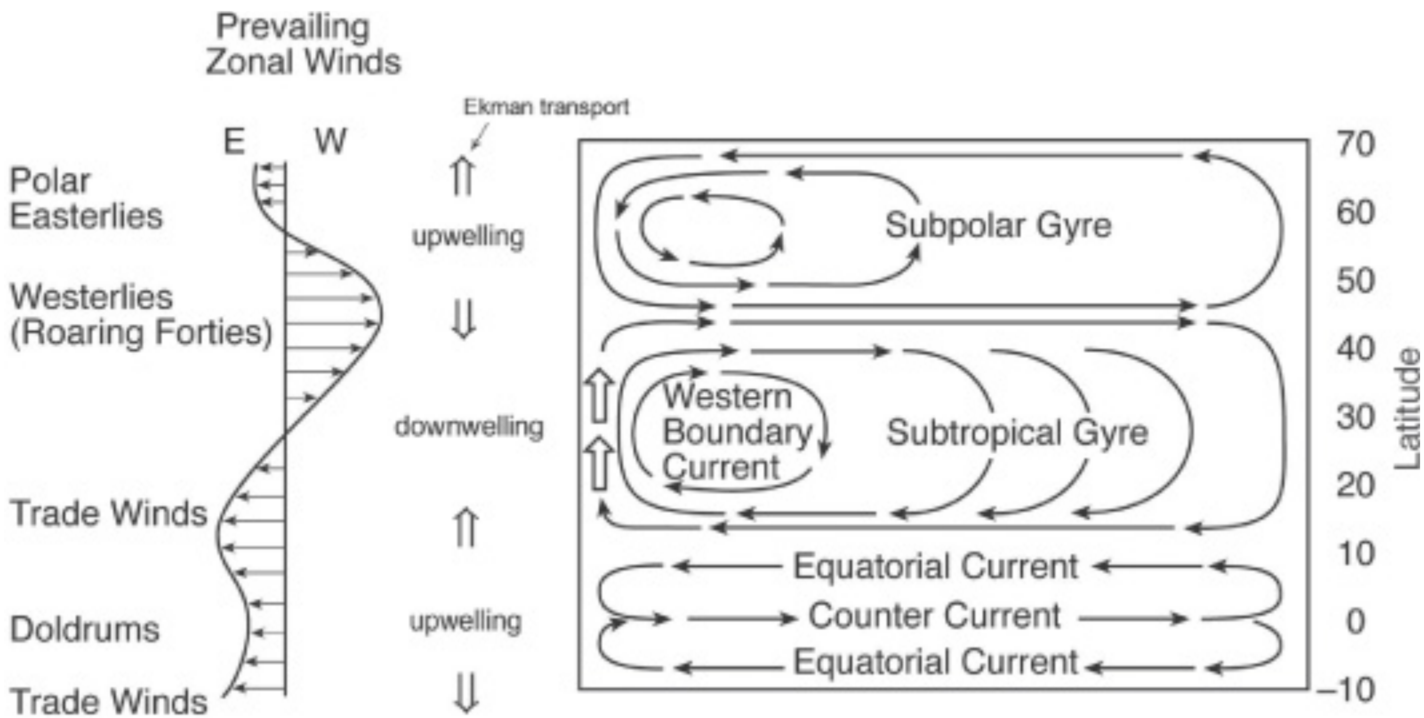


Figure 10.20: Schematic diagram showing the classification of ocean gyres and major ocean current systems and their relation to the prevailing zonal winds. The pattern of Ekman transport and regions of upwelling and downwelling are also marked.

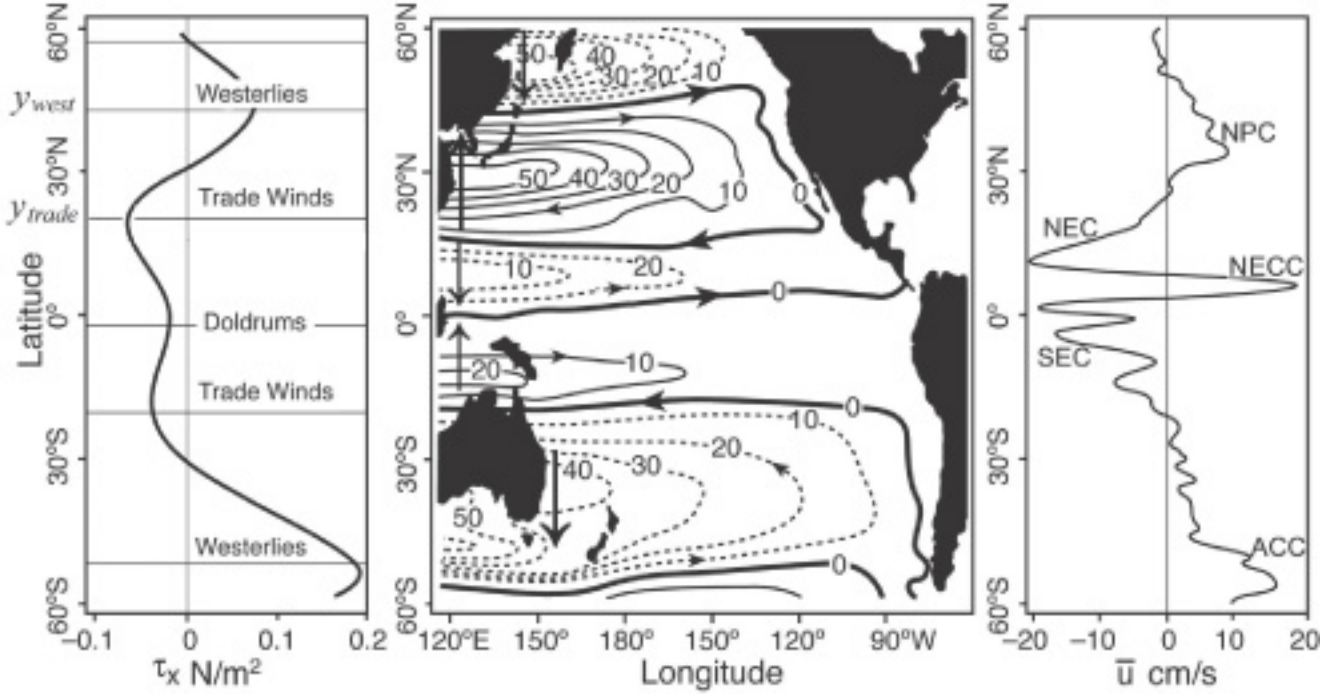
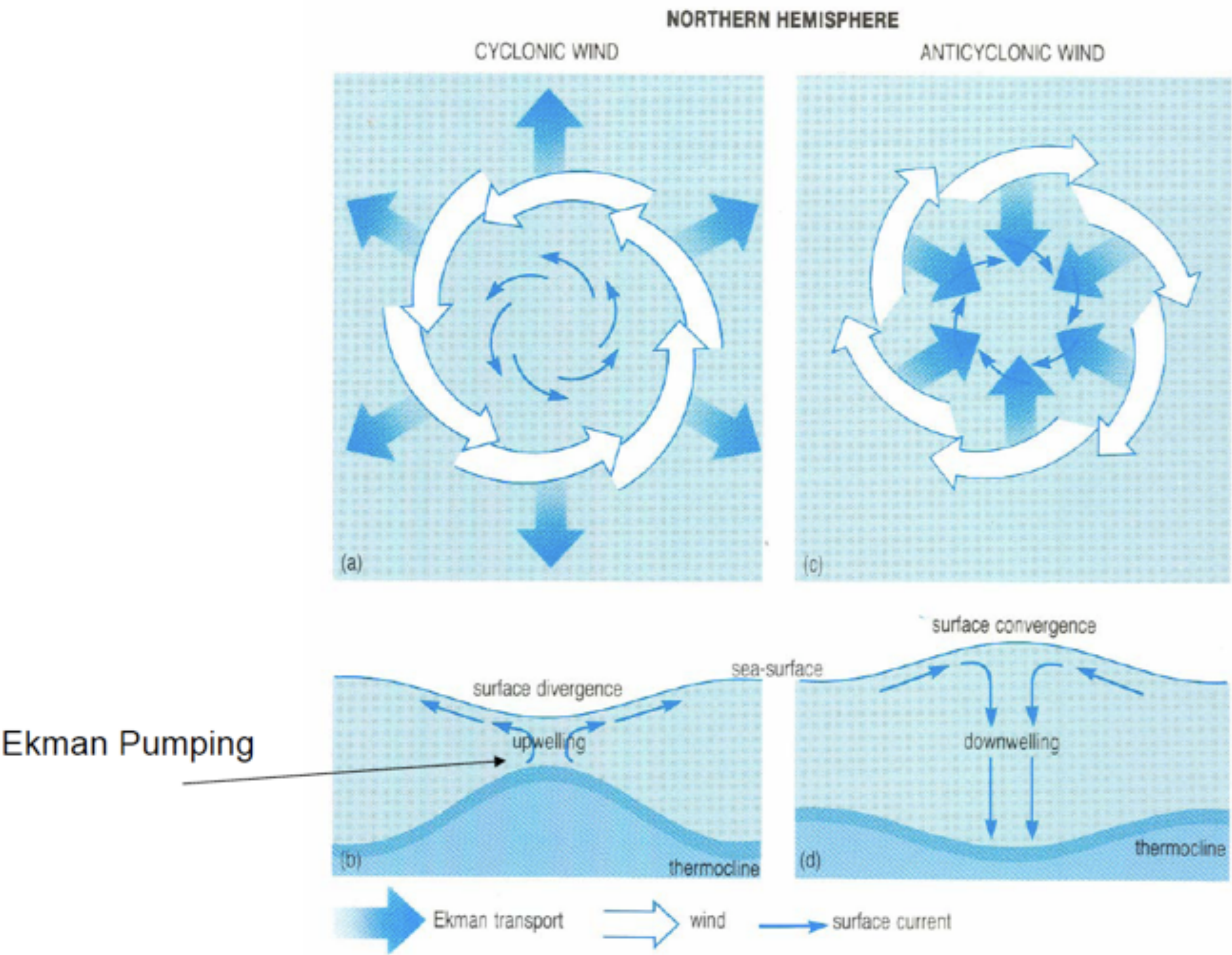
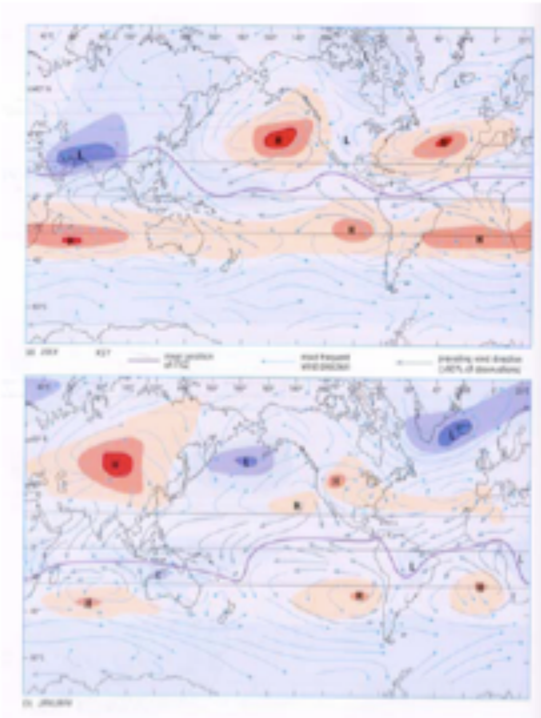
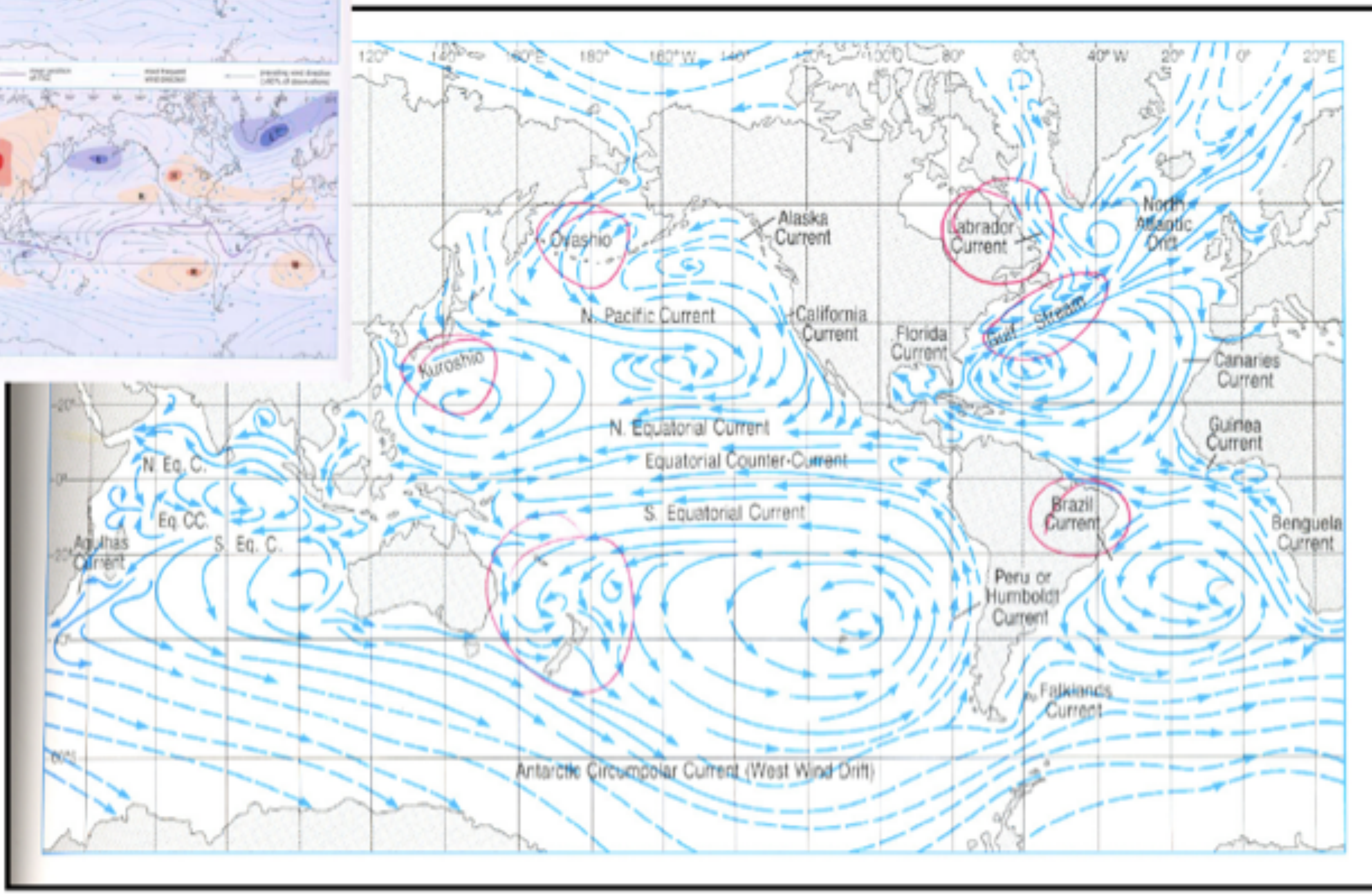


Figure 10.21: (left) The zonal-average of the zonal wind stress over the Pacific Ocean. (middle) The Sverdrup transport stream function (in $Sv = 10^6\ m^3\ s^{-1}$) obtained by evaluation of Eq. 10-20 using climatological wind stresses, Fig. 10.2. Note that no account has been made of islands; we have just integrated right through them. The transport of the western boundary currents (marked by the $N \leftrightarrow S$ arrows) can be read off from Ψ_{west_bdy}. (right) The zonal-average zonal current over the Pacific obtained from surface drifter data shown in Fig. 9.14. Key features corresponding to Fig. 9.13 are indicated.

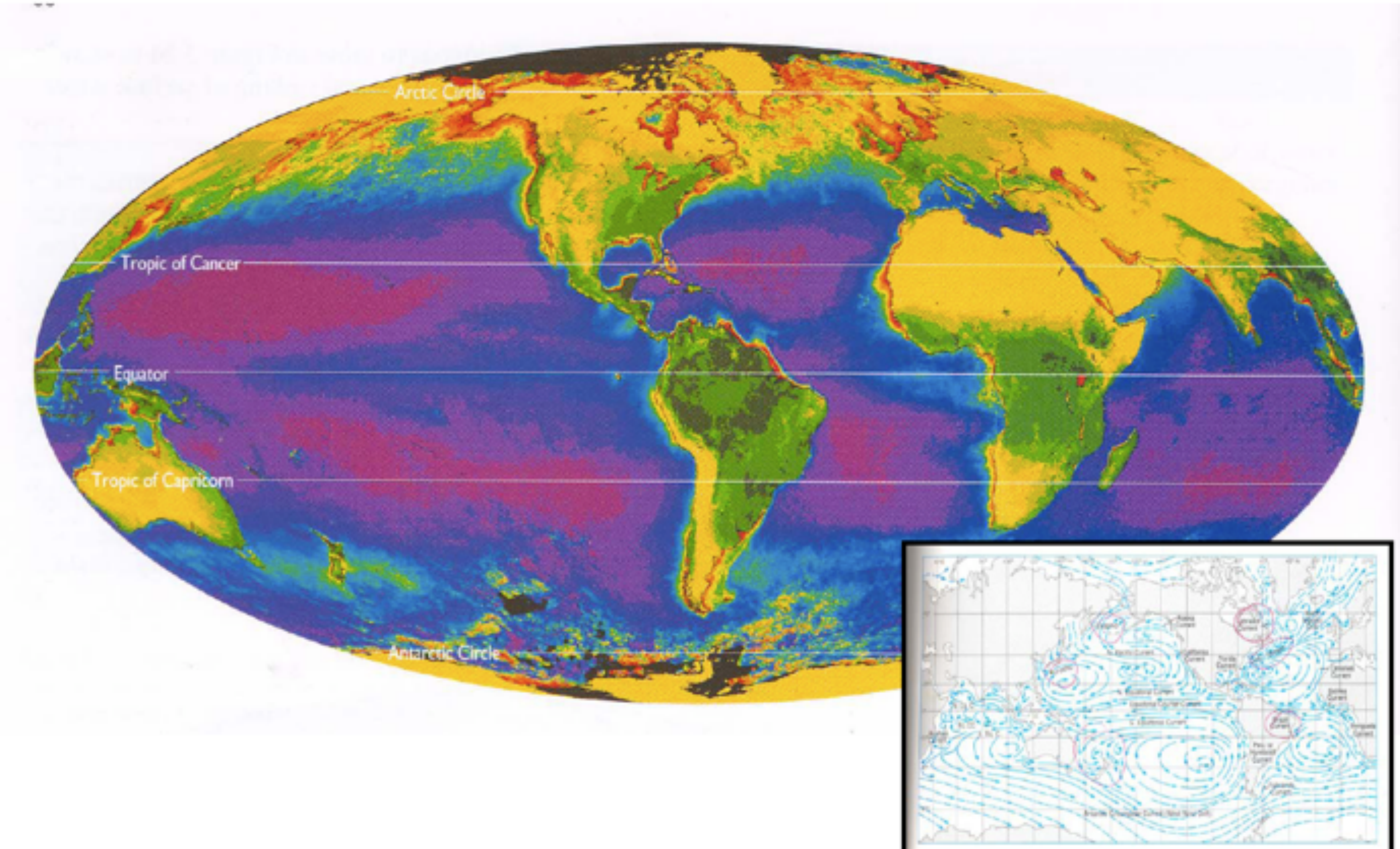

NORTHERN HEMISPHERE
CYCLONIC WIND
ANTICYCLONIC WIND
(a)
(c)
surface convergence
sea-surface
surface divergence
upwelling
downwelling
Ekman Pumping
(b)
thermocline
(d)
thermocline
Ekman transport
wind
surface current

Subtropical/Subpolar gyres

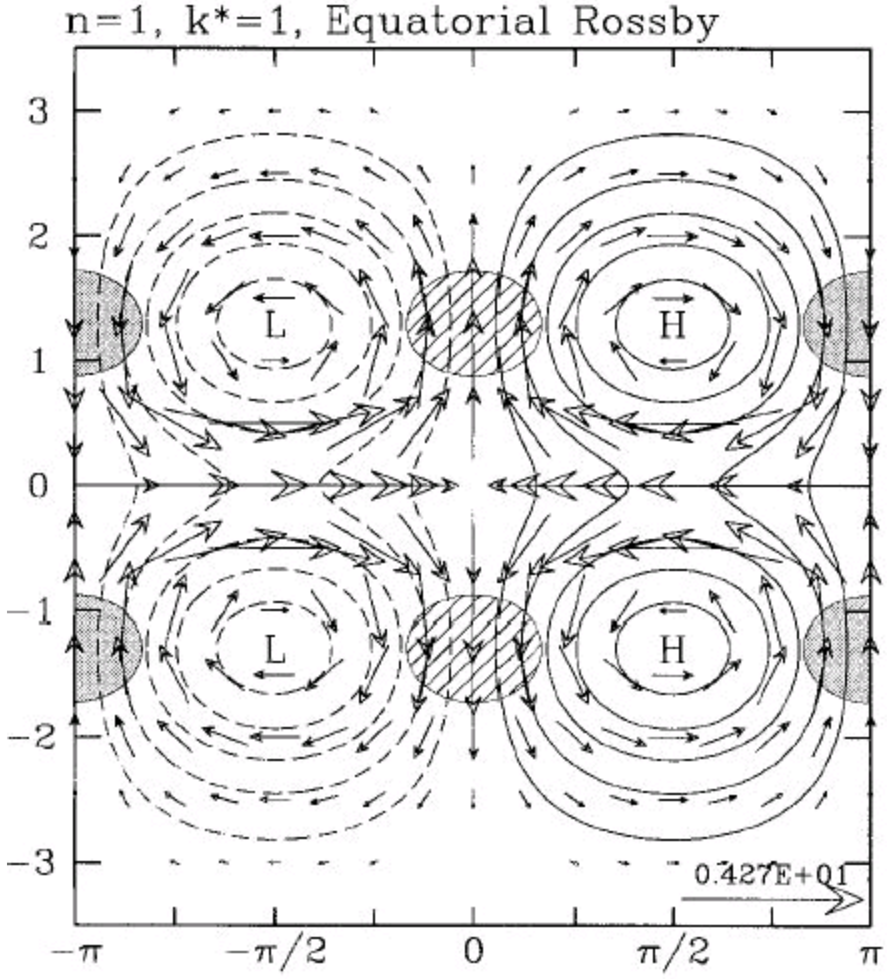




Distribution of Primary Production: red represents regions of high productivity and purple indicating areas of low productivity

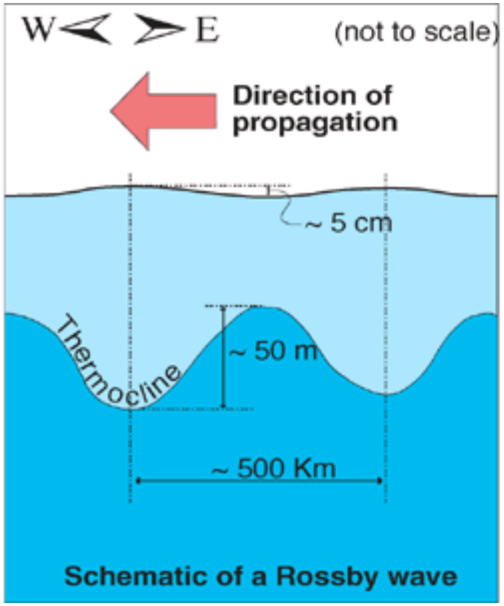


N=1, k=1 Rossby wave



Rossby waves: Ocean

Schematic view "first-mode baroclinic" Rossby wave



Existence in the oceans (Carl-Gustav Rossby, 1930s) has been only indirectly confirmed before the advent of satellite oceanography.

Why is it so difficult to observe them?

It is the big difference in the horizontal and vertical scale of these waves which makes them so difficult to observe.

speed varies with latitude and increases equatorward, order of just a few cm/s

Shallow Water Model

Case b=0

$$
\begin{aligned}
\frac{\partial u}{\partial t} + u\,\frac{\partial u}{\partial x} + v\,\frac{\partial u}{\partial y} - f v &= -g\,\frac{\partial h}{\partial x}\\
\frac{\partial v}{\partial t} + u\,\frac{\partial v}{\partial x} + v\,\frac{\partial v}{\partial y} + f u &= -g\,\frac{\partial h}{\partial y}\\
\frac{\partial h}{\partial t} + \frac{\partial}{\partial x}(hu) + \frac{\partial}{\partial y}(hv) &= 0.
\end{aligned}
$$

This is a formulation that we will encounter in layered models !

Shallow Water Model

Case b=0

$$\frac{\partial u}{\partial t} \qquad - f v = - g \frac{\partial h}{\partial x}$$

$$\frac{\partial v}{\partial t} \qquad + f u = - g \frac{\partial h}{\partial y}$$

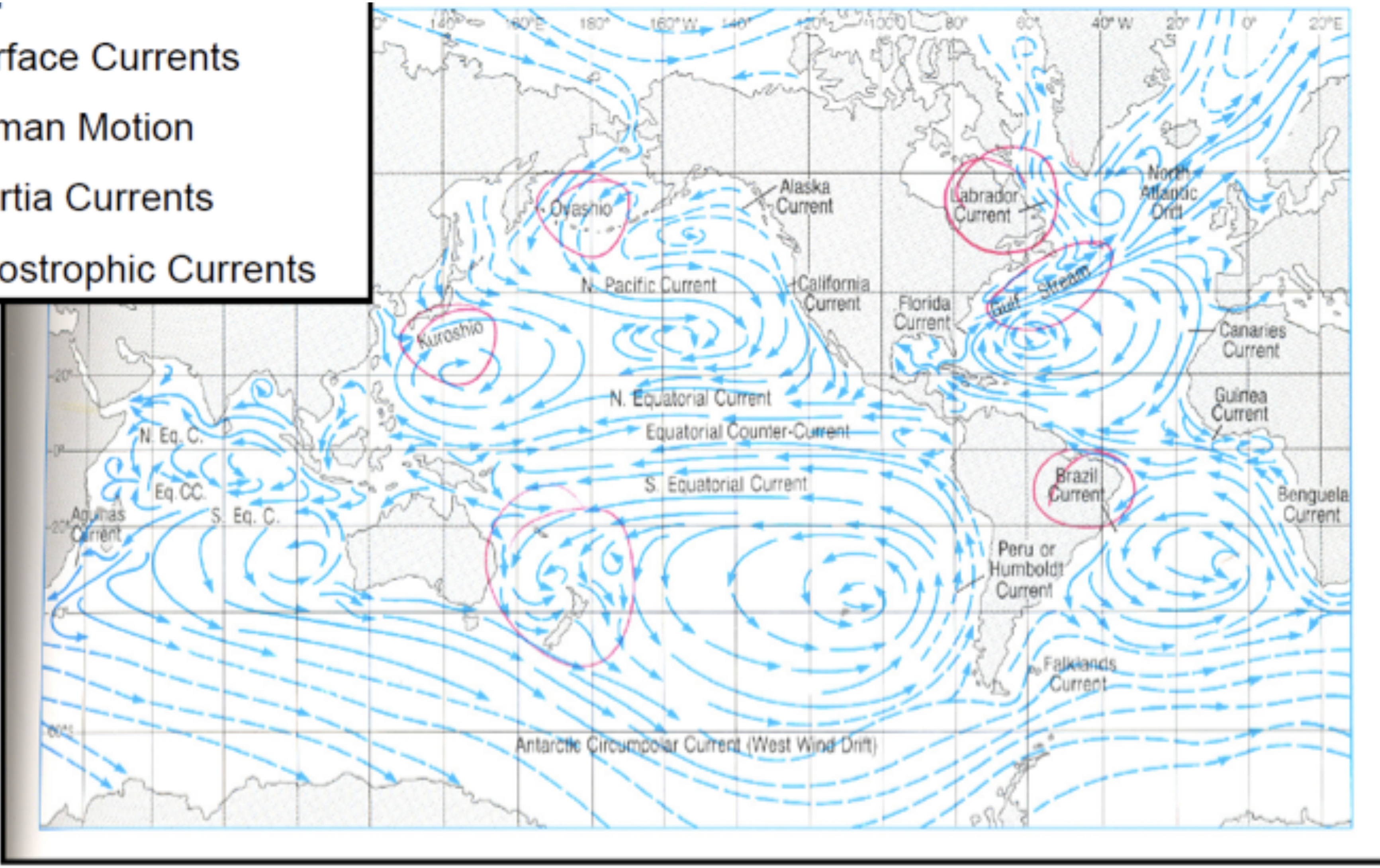
$$\frac{\partial h}{\partial t} + \frac{\partial}{\partial x}(hu) + \frac{\partial}{\partial y}(hv) = 0.$$

Surface Currents

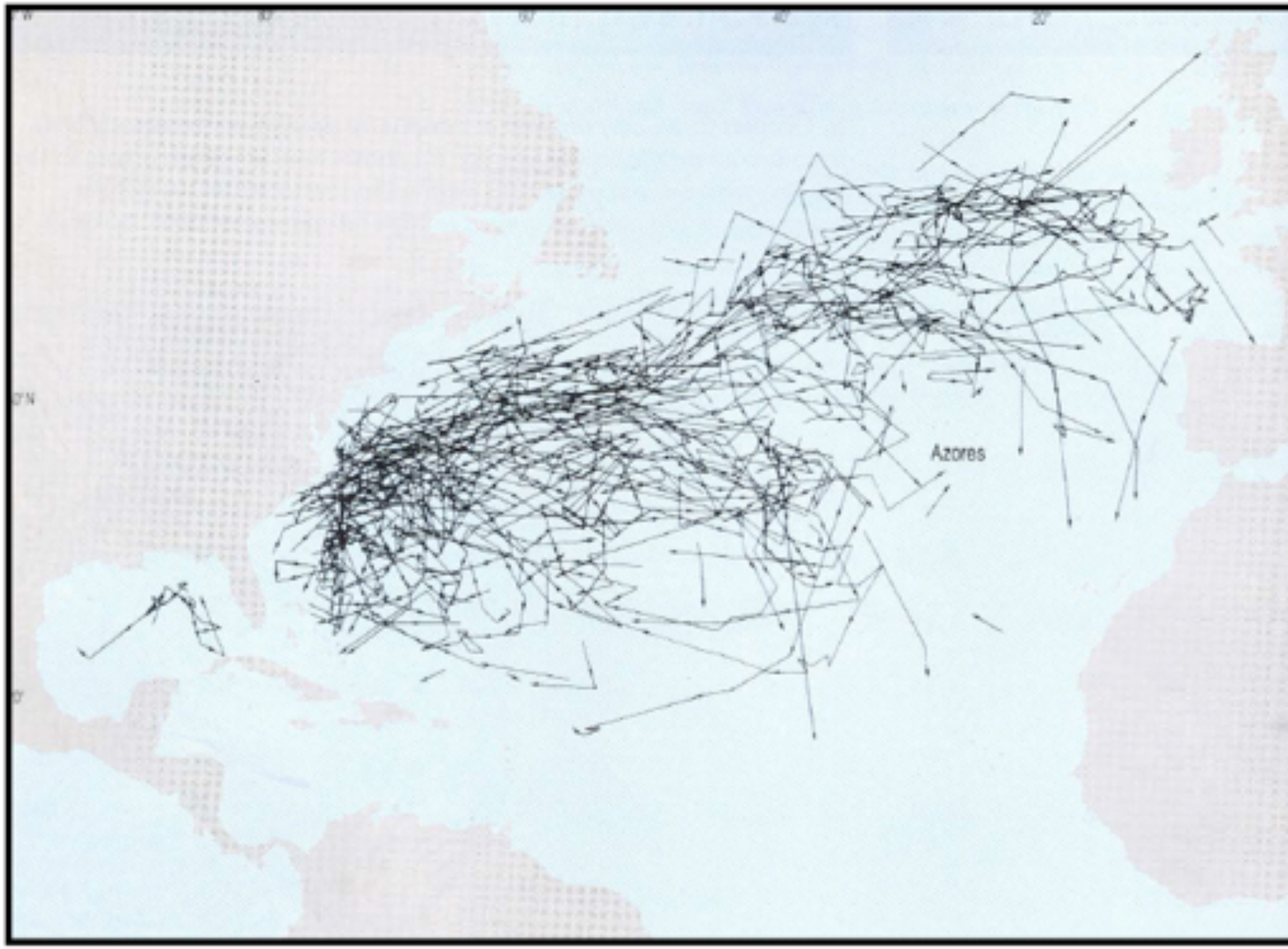
Ekman Motion

Inertia Currents

Geostrophic Currents



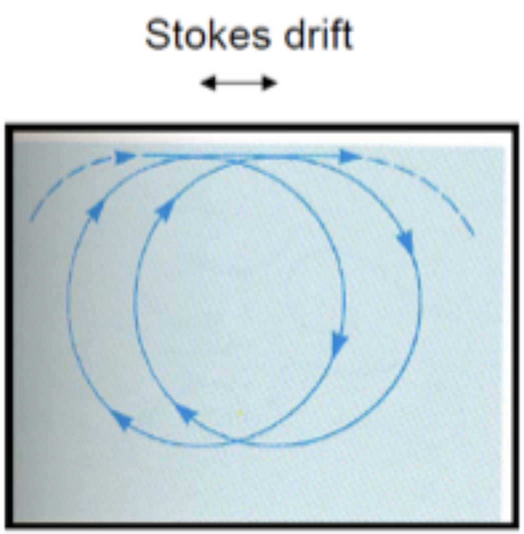
The global surface current system to some extent reflects the surface wind field, but ocean currents are constrained by continental boundaries and current systems are often characterized by gyral circulaitons



Maps of wind and current flows represent average conditions only; at any one time the actual flow at a given point might be markedly different from that shown

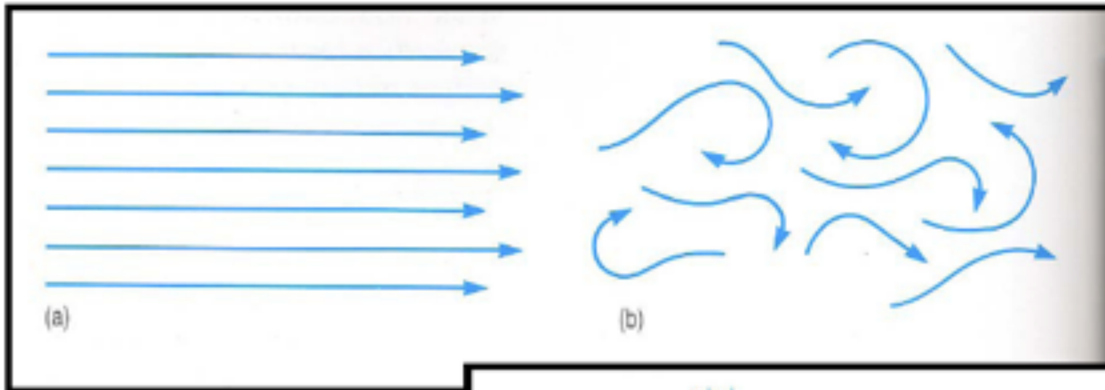
The frictional force caused by the action of wind on the sea-surface is known as the wind stress. Its magnitude is proportional to the square of the wind speed; it is also dependent on the roughness of the sea-surface and conditions in the overlying atmosphere.

$$\tau = c^*W^2$$



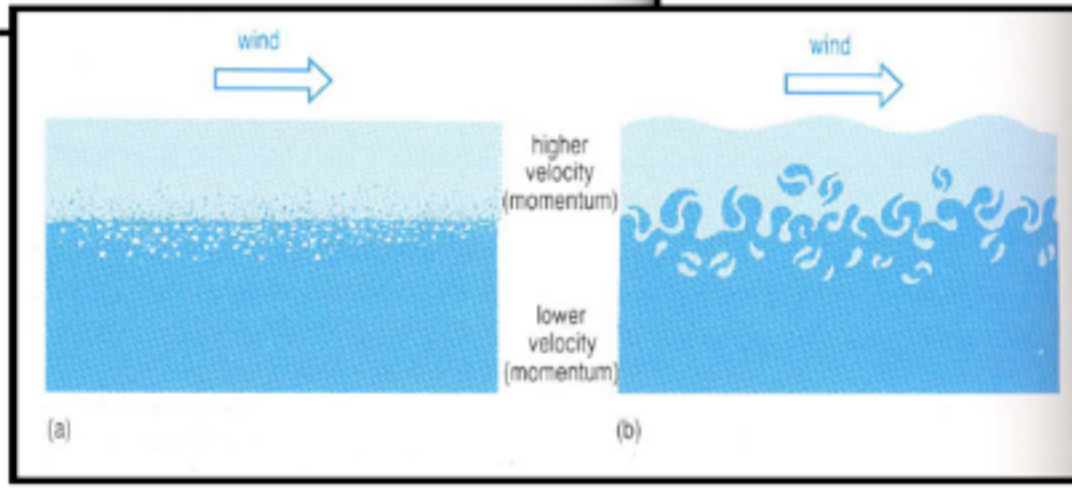
Energy from the atmosphere can go toward building waves and/or generating currents

Surface currents ~= 3% of the wind speed



turbulent flow -->
transfer by "**eddy viscosity**" - a
property of the flow

laminar flow -->
transfer by **molecular viscosity** - a property
of the fluid

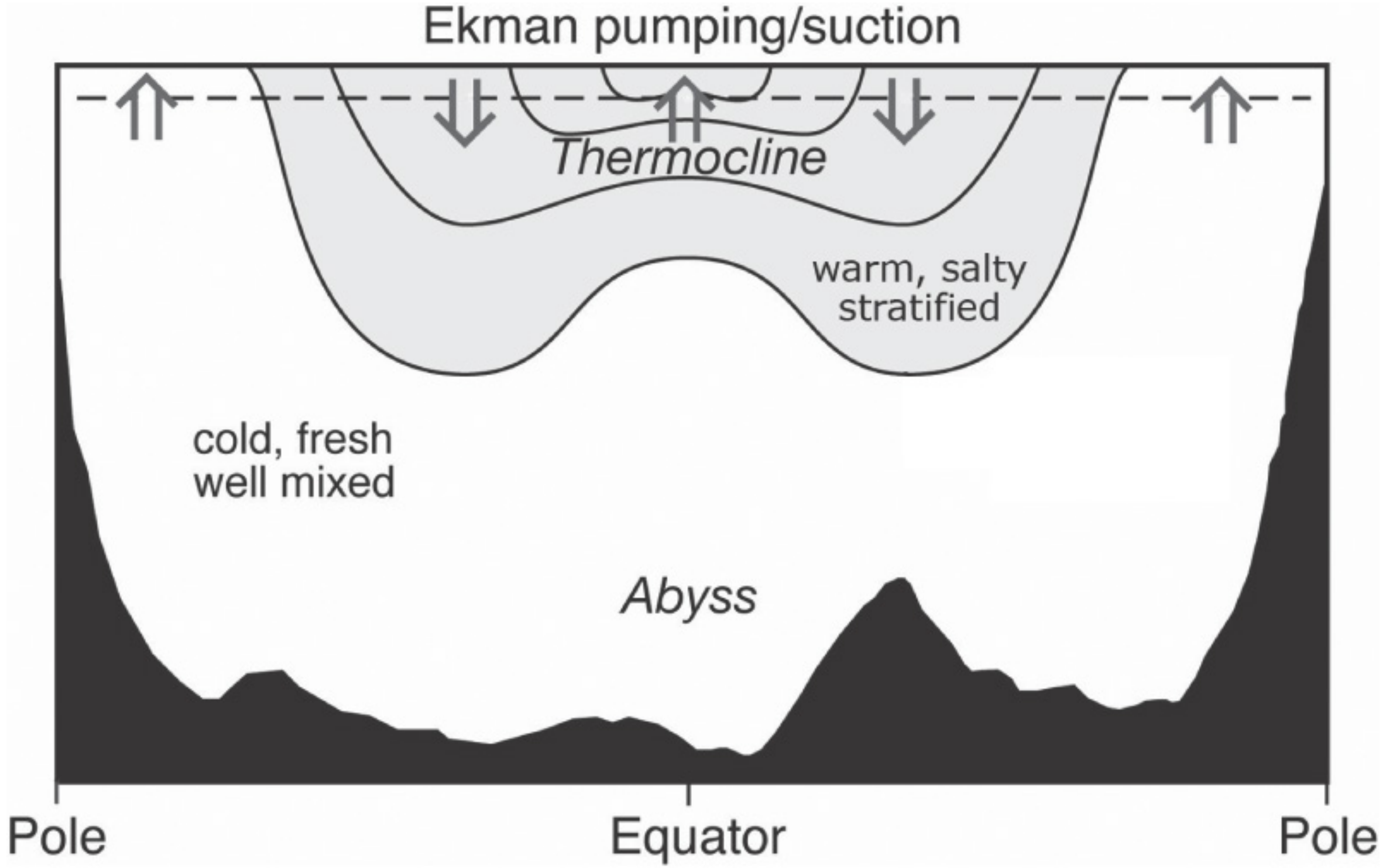


Motion is transmitted downward through frictional coupling caused by turbulence.
Coefficient of friction important for ocean studies = coefficient of **eddy viscosity**

Az = $10^{-2} – 10^{2}$ kg/(ms) Ah = $10^{4} – 10^{8}$ kg/(ms)

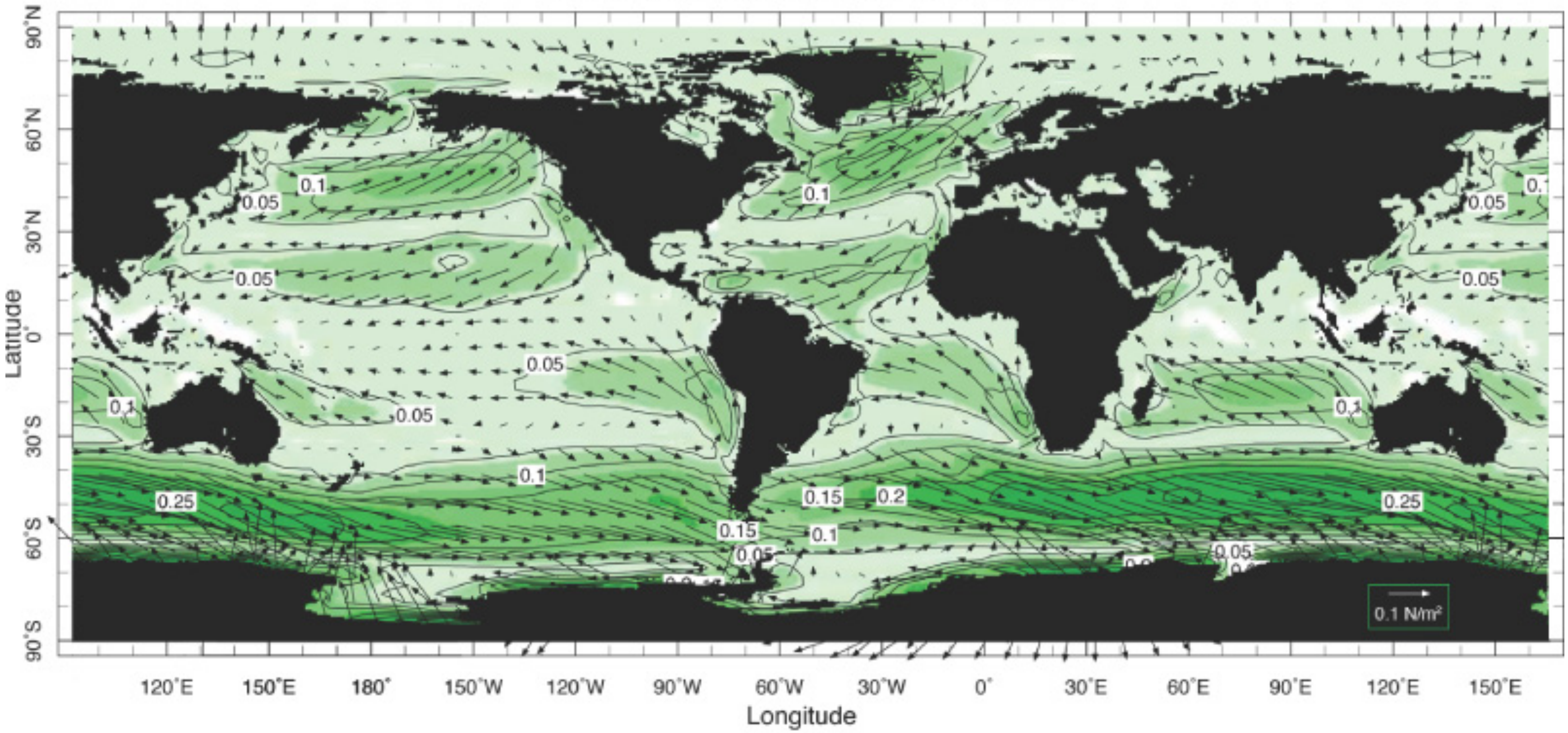
Magnitude of eddy viscosity depends on stratification of the water column - the stronger the (stable) stratification the more turbulent mixing gets suppressed. Conversely, in a homogeneous fluid turbulence mixes very efficiently.

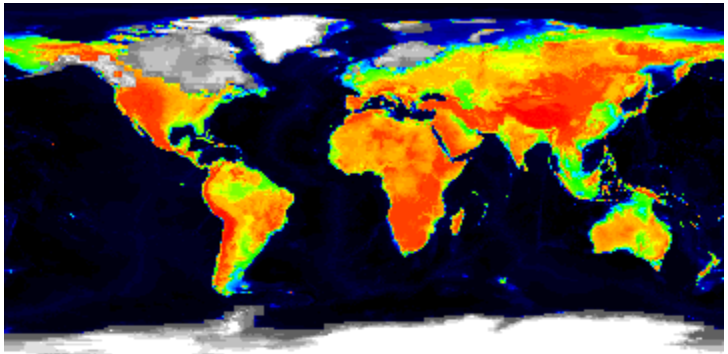
Ocean layers

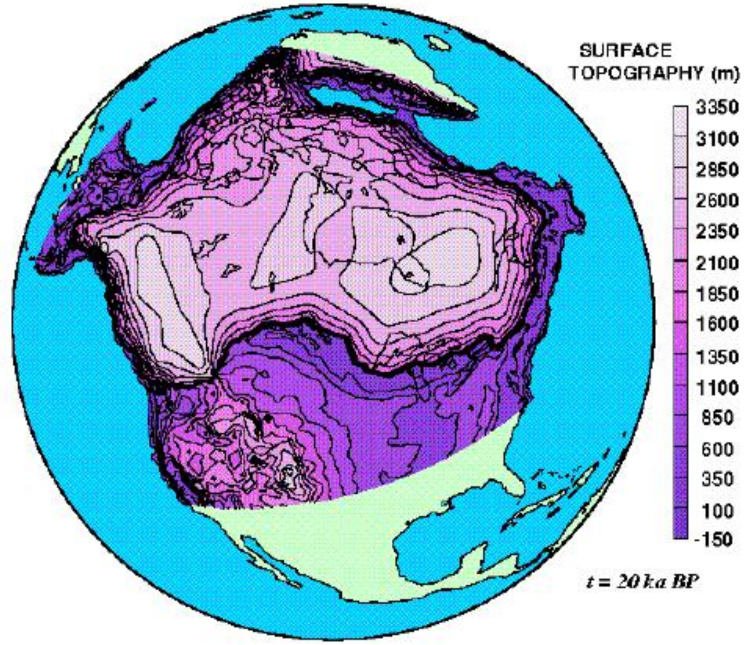


- Steady winds blowing on the sea surface produce a thin, horizontal boundary layer, the *Ekman layer.*
- *thin: a few-hundred meters thick, compared with the depth of the water in the deep ocean.*
- *A similar boundary layer exists at the bottom of the ocean and at the bottom of the atmosphere just above the sea surface, the planetary boundary layer or frictional layer .*

Surface Wind Stress (N/m²)

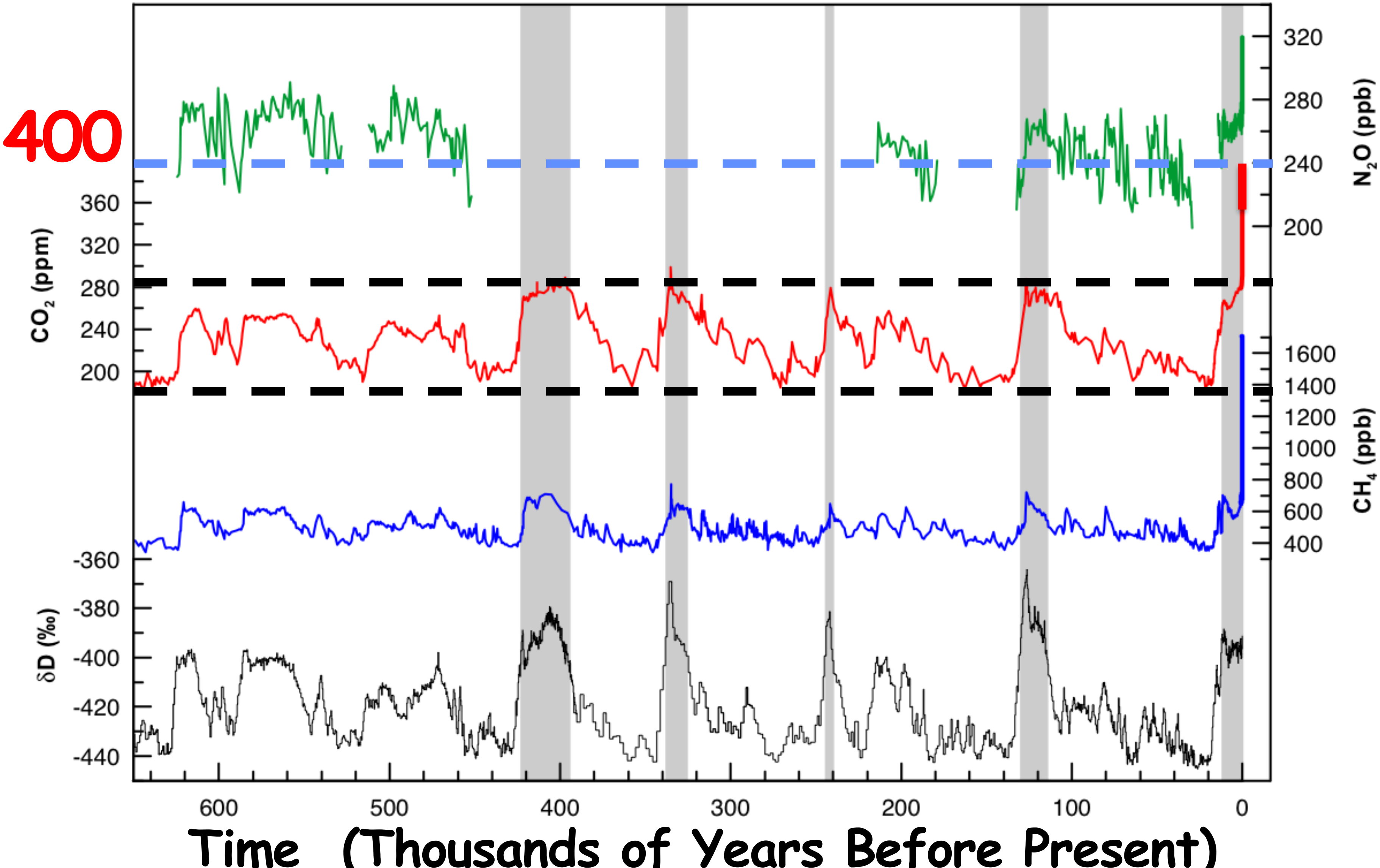




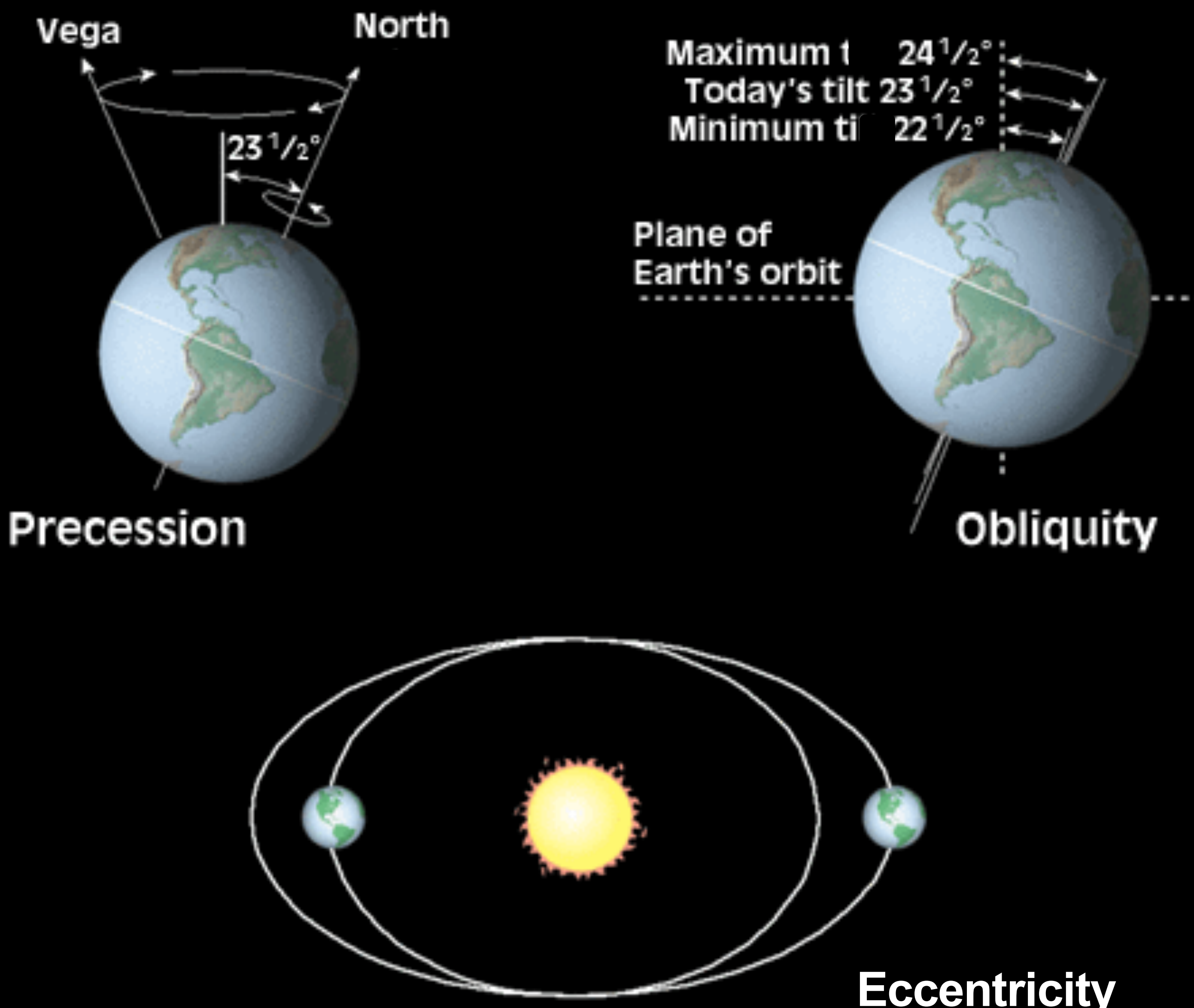


Deglaciation

Atmospheric Gas Concentrations from Ice Cores



EPICA 2008

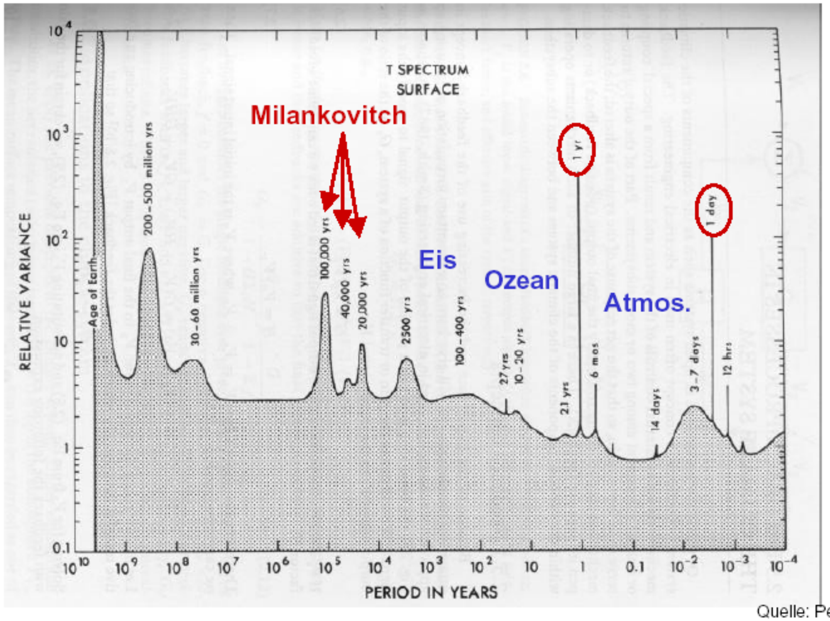


Vega
North
23 1/2°
Precession
Maximum t 24 1/2°
Today's tilt 23 1/2°
Minimum ti 22 1/2°
Plane of
Earth's orbit
Obliquity
Eccentricity

Orbital focing

- ~20.000, ~40.000, ~100.000 years
- 0.5, 1 year

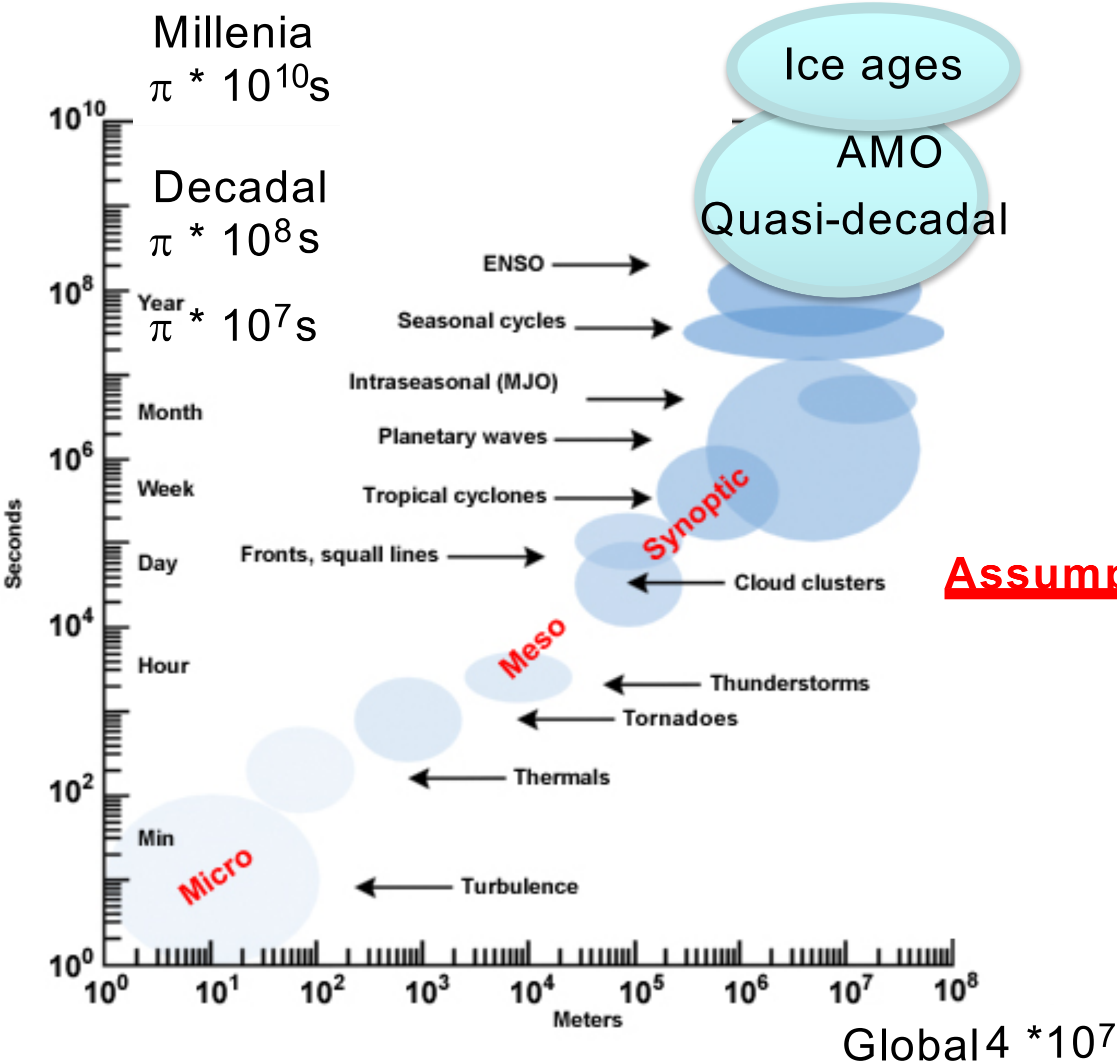
- Geometry of the Sun-Earth configuration



Quelle: Peixoto & Oort

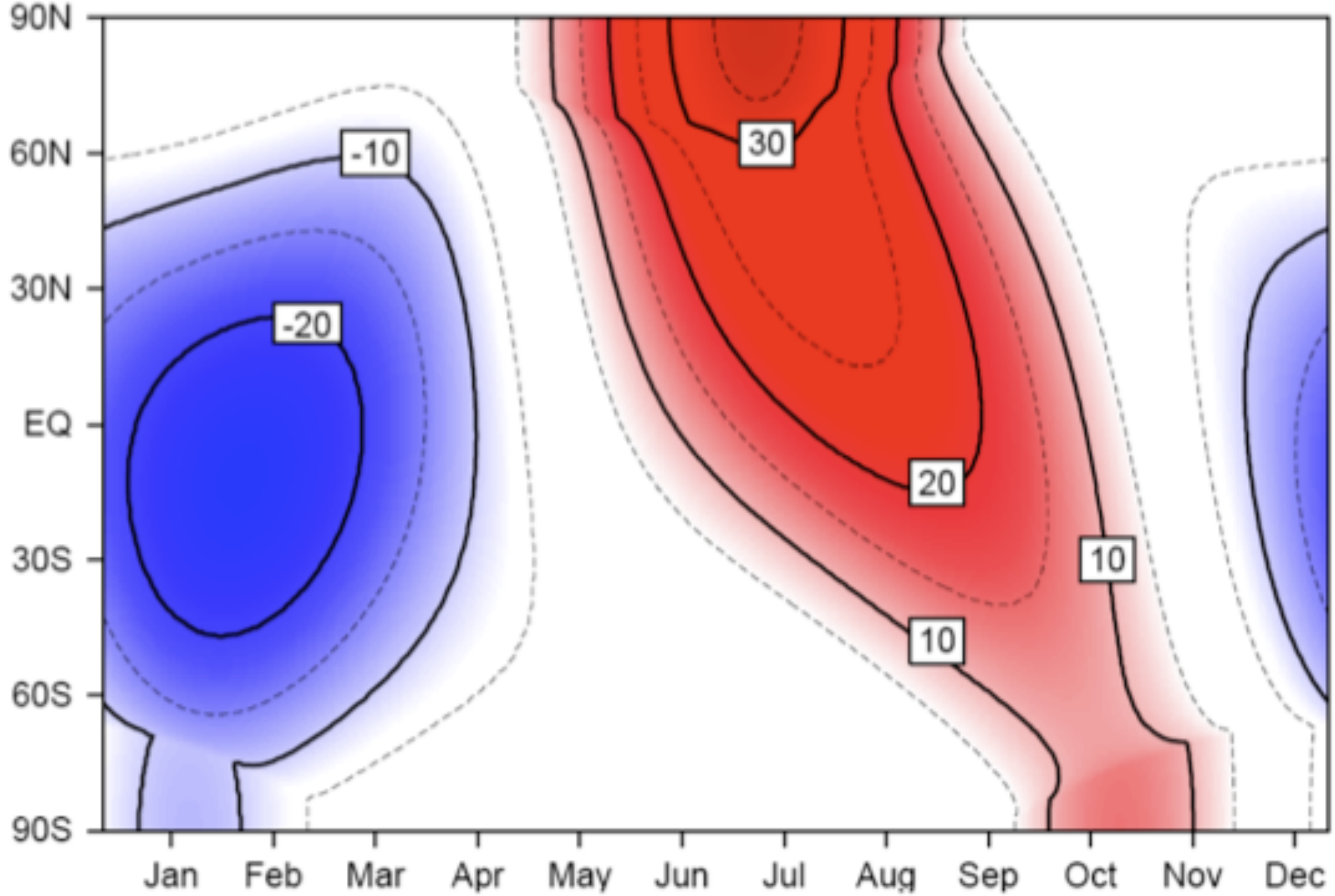
Spatio-Temporal Scales

Dissipative Systems (as atmosphere & ocean) cannot maintain large gradients on long time scales



Assumption: Spatial || temporal Scales

Insolation (6k minus present)



Statistics

covariance is a measure of how much two random variables change together

Covariance (cross, auto)

$$\gamma(\Delta) = E\Big((x(t) - \bar{x})\, (y(t+\Delta) - \bar{y}) \Big)$$

e.g. coral   e.g. meteorol. data

$$\operatorname{cov}(X,Y) = \frac{1}{n}\sum_{i=1}^{n}(x_i - E(X))(y_i - E(Y)).$$

Correlation (cross, auto)

$$\rho_{xy} = \frac{\gamma(\Delta)}{\text{normalized}}$$

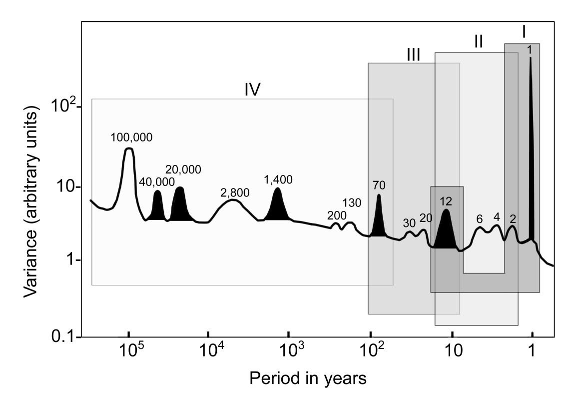
measures the tendency of x (t) and y (t) to covary, between -1 and 1

Spectrum (cross, auto)

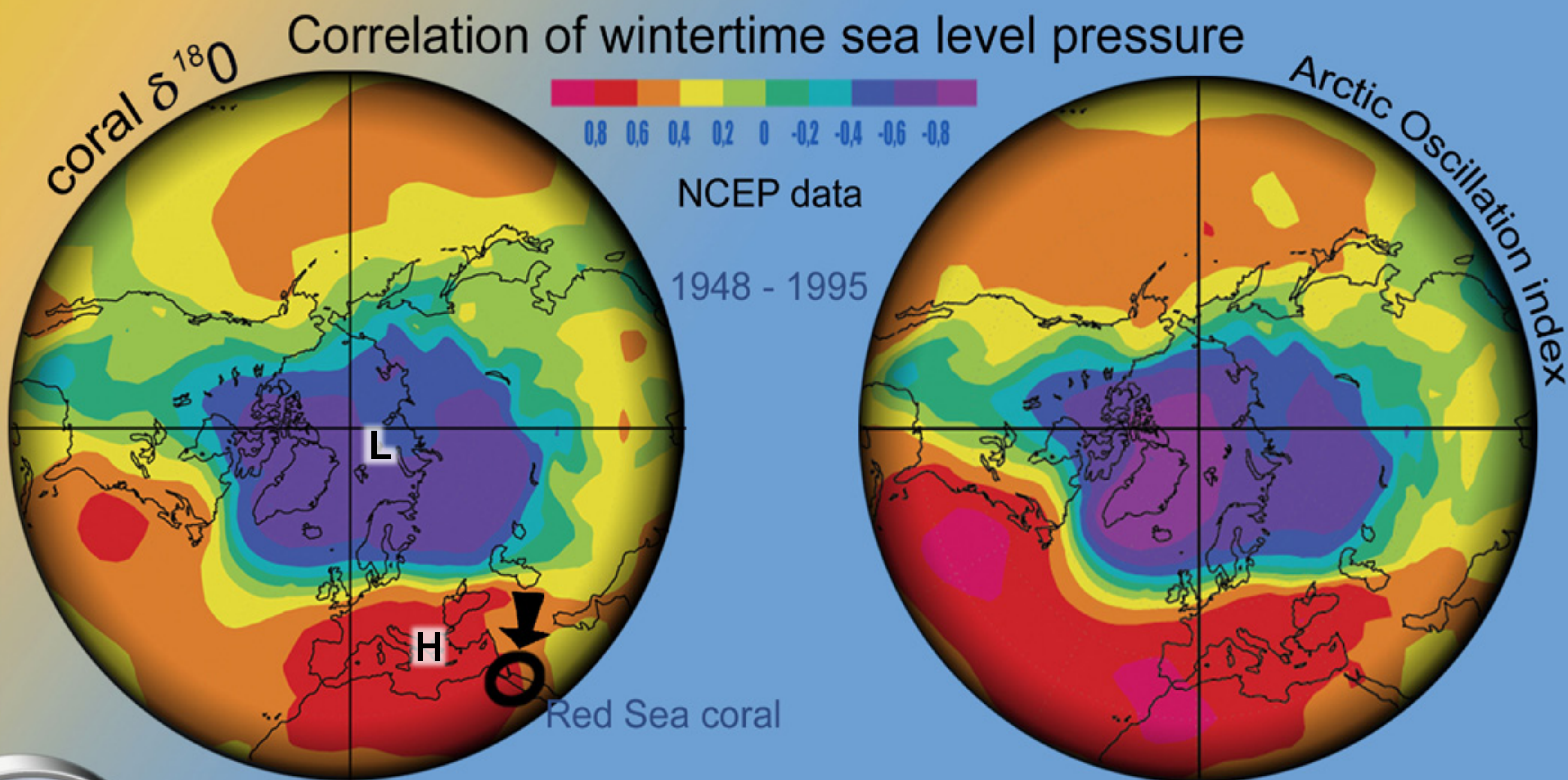
(spectral density)

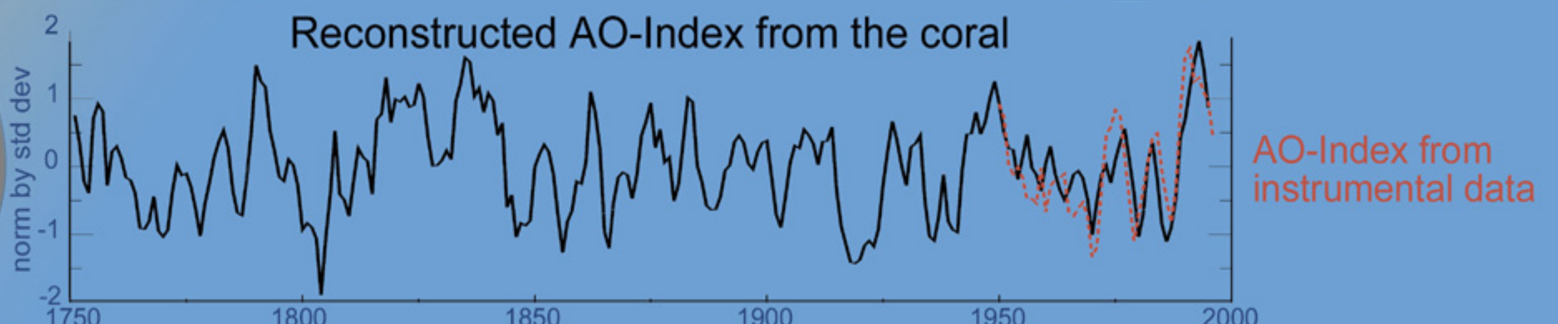
$$\Gamma(\omega) = \sum_{\Delta=-\infty}^{\infty} \gamma(\Delta)\, e^{-2\pi i\Delta}$$

measures variance



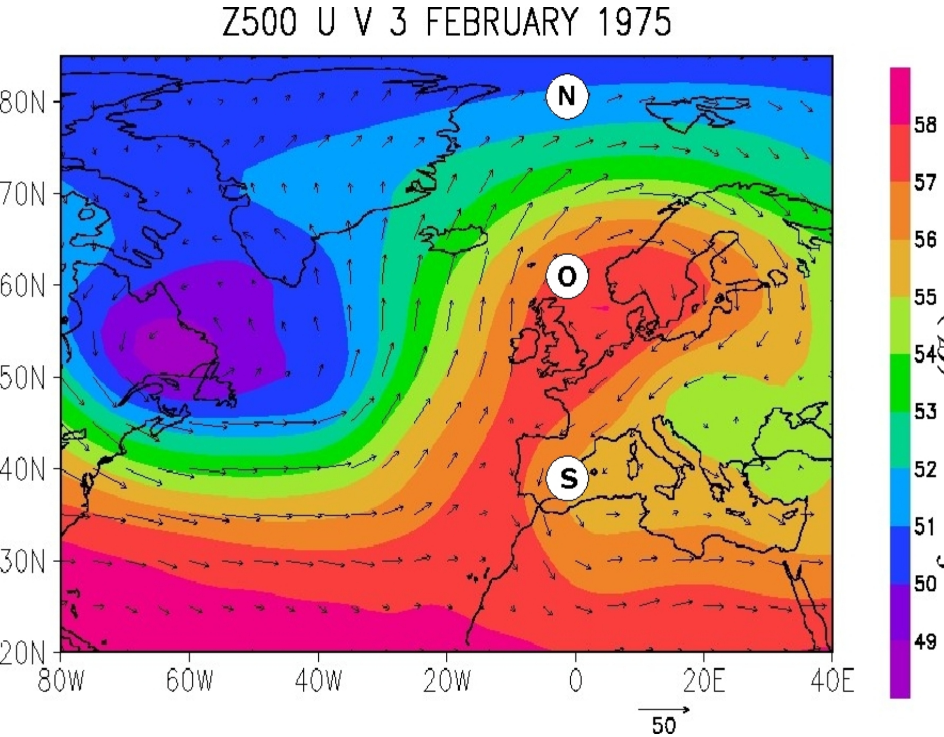
ARCTIC OSCILLATION SIGNATURE IN A RED SEA CORAL

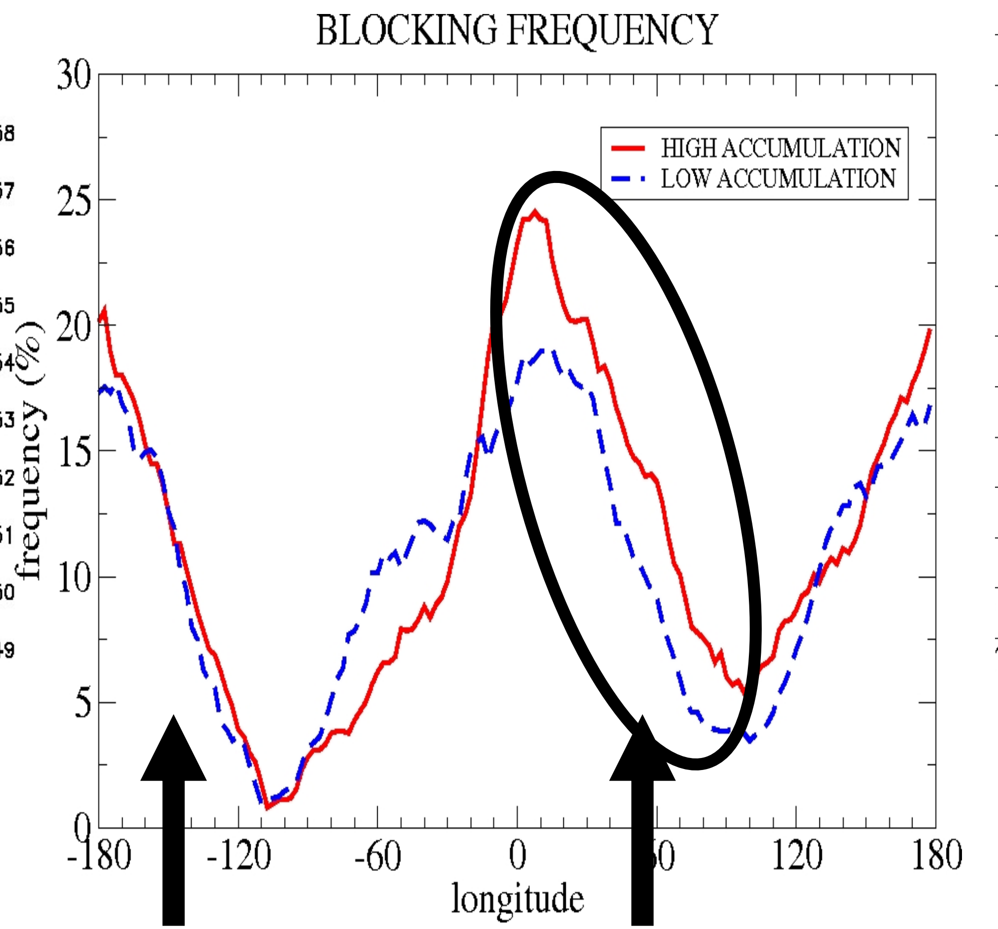




Rimbu, Lohmann, Felis, Pätzold; GRL 2001

Atmospheric Blocking





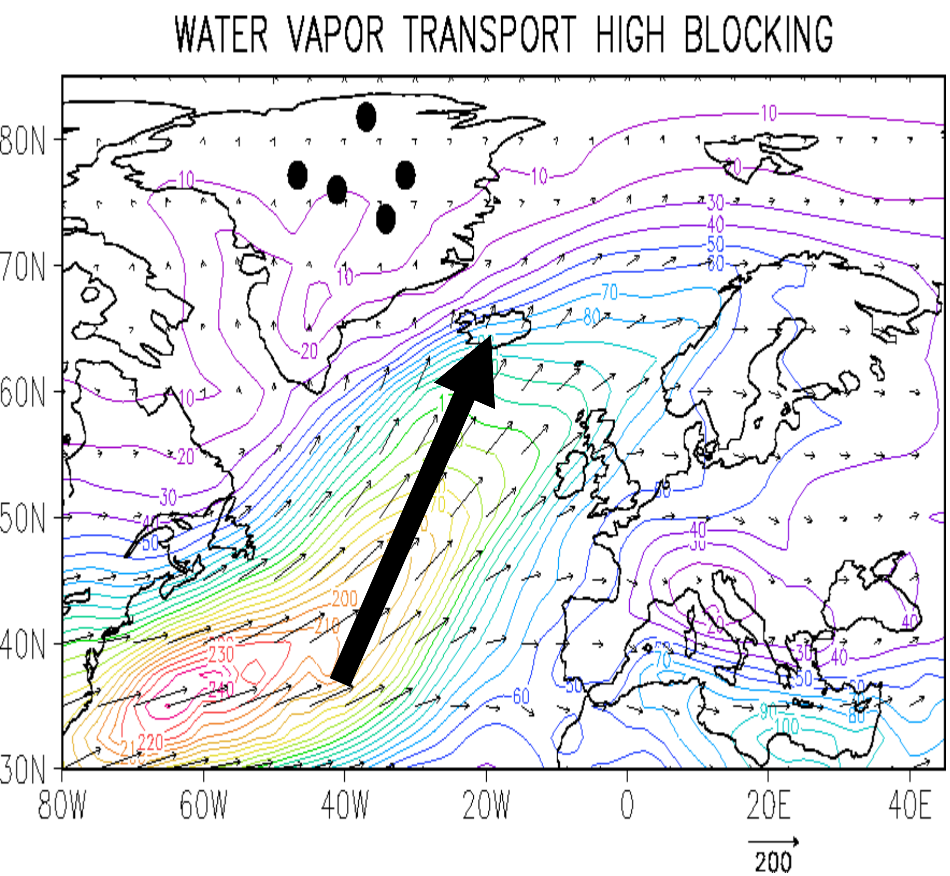
Z500 (NCEP): 1948-2005 (57 winters)
TM index (Tibaldi and Molteni,1990)

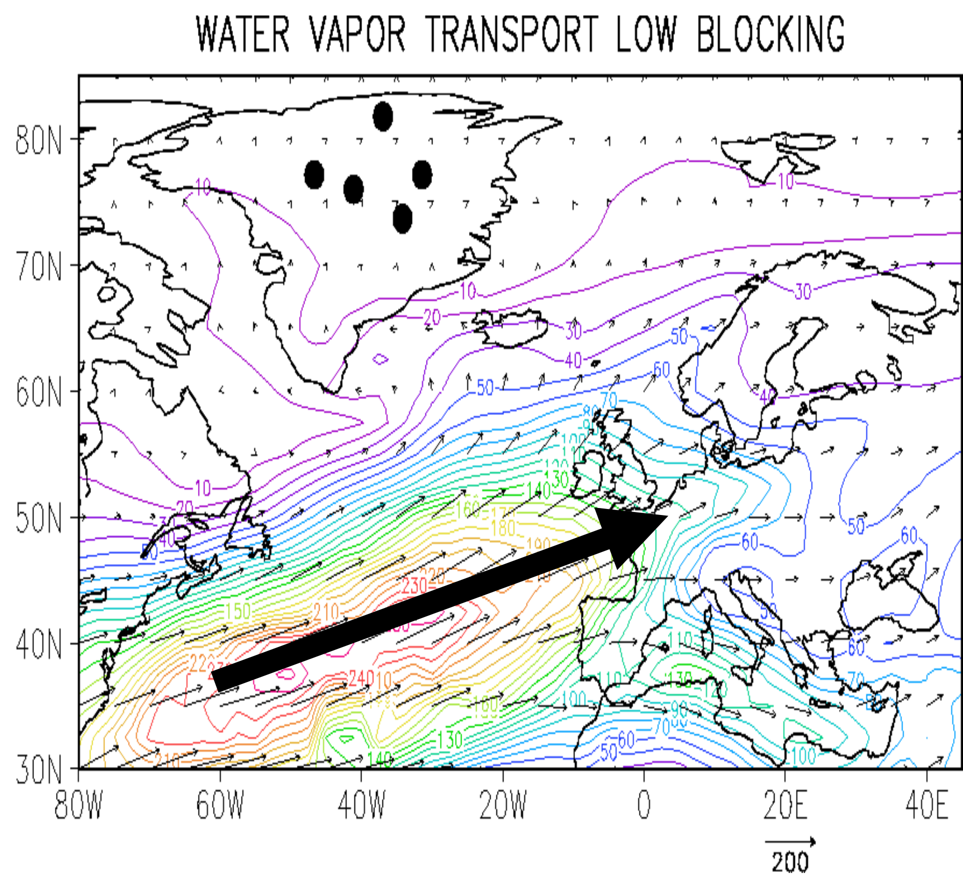
Pacific

Euro-Atlantic

sectors

WATER VAPOR TRANSPORT





ENHANCED MOISTURE TRANSPORT TOWARD GREENLAND DURING HIGH BLOCKING ACTIVITY IN 20ºW - 20ºE SECTOR

http://climexp.knmi.nl

1) Monthly climate indices (temp, precip, ...)

Calculate different regions on the world
(home town, Bremerhaven 53° N, 8° E)

2) Correlation with temperature, precipiation, SLP

3) Explain the teleconnections for different seasons

---

4) Modes of climate variability (global temperature)

Select a monthly time series

Climate indices

Select a time series by clicking on the name		
ENSO	absolute NINO12, NINO3, NINO3.4, NINO4, relative NINO12, NINO3, NINO3.4, NINO4 (1880-now, ERSST v4, relative is relative to 20S-20N, i.e., without global warming, recommended)	i
	NINO12, NINO3, NINO3.4, NINO4 (1870-now, HadISST1)	i
	NINO12, NINO3, NINO3.4, NINO4 (1856-1981 Kaplan, 1982-now NCEP OISSTv2)	i
	SOI (1866-now, Jones)	i
	SOI (1882-now, NCEP)	i
	Precipitation Niño indices: GPCC land , CMORPH satellite	i
	MEI (1950-now, NOAA/ESRL/PSD)	i
	Warm Water Volume (5°S-5°N, 120°E-80°W, 1980-now, PMEL/TAO)	i
	WWV (5°S-5°N, 120°E-80°W, 1960-now, POAMA/PEODAS)	i
	temperature averaged to 300m (130°E-80°W, 1979-now, GODAS)	i
NAO	NAO Gibraltar-Stykkisholmur (1821-now, Jones)	i
	NAO Azores-Stykkisholmur (1865-2002, data from Jones)	i
	NAO (pattern-based, 1950-now, CPC)	i
	NAO reconstruction (1658-2001, Luterbacher)	i
SNAO	Summer NAO from NCEP/NCAR (1948-now), UCAR (1899-now), 20C (1871-2008) SLP	i
AO	Arctic Oscillation derived from SLP (1899-2002) and derived from SAT (1851-1997, Thompson, Colorado State)	i
	Arctic Oscillation (1950-now, NCEP/CPC)	i
AMO	Atlantic Multidecadal Oscillation derived from HadSST (1850-now) and derived from ERSST (1880-now) SST 25°-60°N, 7°-75°W minus regression on Tglobal	i
	Atlantic Multidecadal Oscillation derived from HadSST (1850-now) and derived from ERSST (1880-now) SST EQ-60°N, 0°-80°W minus SST 60°S-60°N	i
AMOC	Atlantic Meridional Overturning Circulation: ECMWF S3 (1961-2005)	
Teleconnection patterns	East Atlantic, East Atlantic/Western Russia, Scandinavia and Polar/Eurasia patterns (1950-now, CPC)	i

Select a time series

- Daily station data
- Daily climate indices
- Monthly station data
- Monthly climate indices
- Annual climate indices
- View, upload your time series

Select a field

- Daily fields
- Monthly observations
- Monthly reanalysis fields
- Monthly and seasonal historical reconstructions
- Monthly seasonal hindcasts
- Monthly decadal hindcasts
- Monthly CMIP3+ scenario runs
- Monthly CMIP5 scenario runs
- Annual CMIP5 extremes
- Monthly CORDEX scenario runs
- Attribution runs
- External data (ensembles, ncep, enact, soda, ecmwf, ...)
- View, upload your field