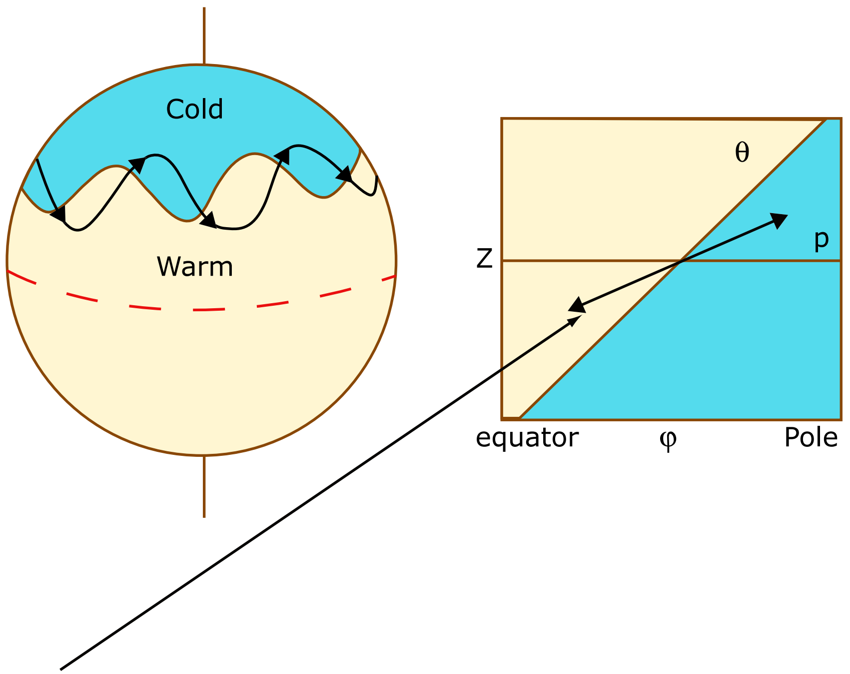
Circulation of the Atmosphere

Atmospheric Heat Transport



Warm air upward and poleward
Cold air downward and equatorward

Efficient poleward heat transport

Image by MIT OpenCourseWare.

Steady Flow:

$$\nabla \bullet \left[F_{rad} \hat{k} + F_{conv} \hat{k} + \rho \mathbf{V} E \right] = 0,$$

$$where$$

$$E \equiv c_p T + gz + L_v q + \frac{1}{2} | \mathbf{V} |^2$$

Integrate from surface to top of atmosphere:

$$\nabla_2 \bullet \overline{\rho \mathbf{V} E} + F_{rad_{TOA}} - \left(F_{rad} + F_{conv} \right)_{surface} = 0$$

What causes lateral enthalpy transport by atmosphere?

1: Large-scale, quasi-steady overturning motion in the Tropics,

2: Eddies with horizontal dimensions of ~ 3000 km in middle and high latitudes

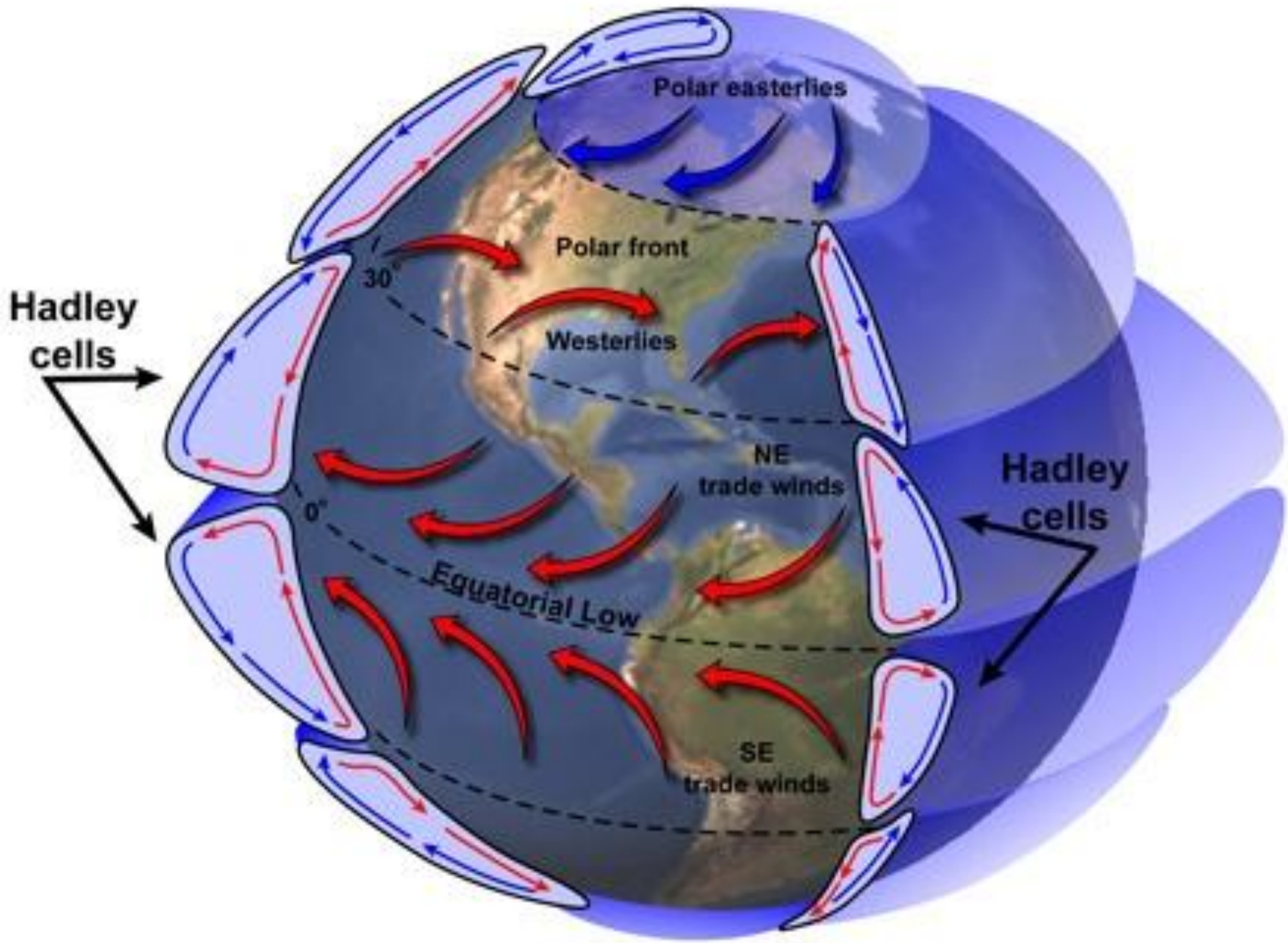


Image courtesy of NASA.

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Please see the second image on page
http://www.saeonmarine.co.za/SADCOFunStuff/OceanCurrents.htm

First consider a hypothetical planet like Earth, but with no continents and no seasons and for which the only friction acting on the atmosphere is at the surface.

This planet has an exact nonlinear equilibrium solution for the flow of the atmosphere, characterized by:

1. Every column is in radiative-convective equilibrium,

2. Wind vanishes at planet's surface

3. Horizontal pressure gradients balanced by *Coriolis accelerations*

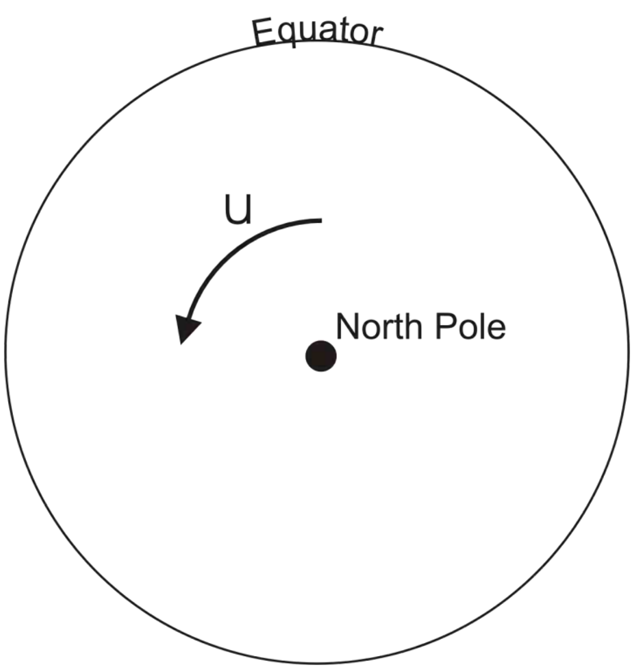
Hydrostatic balance:

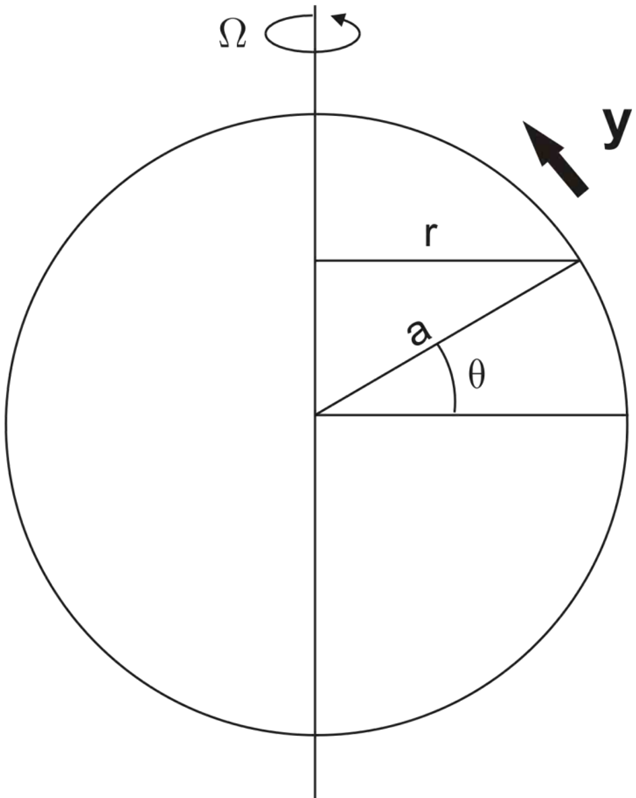
$$\frac{\partial p}{\partial z} = -\rho g$$

Horizontal force balance in *inertial* reference frame:

$$\frac{du}{dt} = -\alpha \frac{\partial p}{\partial x}, \qquad \frac{dv}{dt} = -\alpha \frac{\partial p}{\partial y}$$

Rotating reference frame of Earth:





$$\frac{dv}{dt} = -\alpha \frac{\partial p}{\partial y} - \frac{u^2}{r}\sin\theta$$

$$u = \Omega a \cos\theta + u_{rel}, \qquad r = a\cos\theta$$

$$\rightarrow \frac{dv}{dt} = -\alpha \frac{\partial p}{\partial y} - \Omega^2 a \cos\theta \sin\theta - 2\Omega \sin\theta u_{rel} - \frac{u_{rel}^2}{a}\tan\theta$$

Bracketed term absorbed into definition of gravity:

$$\frac{dv}{dt} = -\alpha \frac{\partial p}{\partial y} - 2\Omega \sin\theta u_{rel} - \frac{u_{rel}^2}{a}\tan\theta$$

$$\cong -\alpha \frac{\partial p}{\partial y} - 2\Omega \sin\theta u_{rel}$$

$$\equiv -\alpha \frac{\partial p}{\partial y} - f u_{rel}, \quad where\ f \equiv 2\Omega \sin\theta$$

Geostrophic Balance

$$\alpha \frac{\partial p}{\partial y} = -f u_{rel}, \quad where\ f \equiv 2\Omega \sin\theta$$

Similarly,

$$\alpha \frac{\partial p}{\partial x} = f v_{rel}$$

$$\alpha \frac{\partial p}{\partial y} = -fu_{rel} \qquad \textit{geostrophic}$$

$$\alpha \frac{\partial p}{\partial z} = -g \qquad \textit{hydrostatic}$$

$Eliminate\ p:$

$$f\frac{\partial u}{\partial z} = -g\left(\frac{\partial \ln(\alpha)}{\partial y}\right) = -g\left(\frac{\partial \ln(T)}{\partial y}\right) \quad \textit{Thermal wind}$$

Zonal wind increases with altitude if temperature decreases toward pole

Exact Solution

- Every individual column of the atmosphere and the surface beneath it are in radiative-convective equilibrium
- Surface pressure is constant
- Pressure above the surface decreases poleward
- Pressure gradient in geostrophic balance with a west wind

Two potential problems with this solution:

1. Not enough angular momentum available for required west-east wind,

2. Equilibrium solution may be unstable

Angular momentum per unit mass:

$$M = a\cos\theta\left(\Omega a\cos\theta + u\right)$$

$$a = radius\ of\ earth$$

$$\theta = latitude$$

$$\Omega = angular\ velocity\ of\ earth$$

$$u = west - east\ wind\ speed$$

Violation results in large-scale overturning circulation, known as the Hadley Circulation, that transports heat poleward and drives surface entropy gradient back toward its critical value

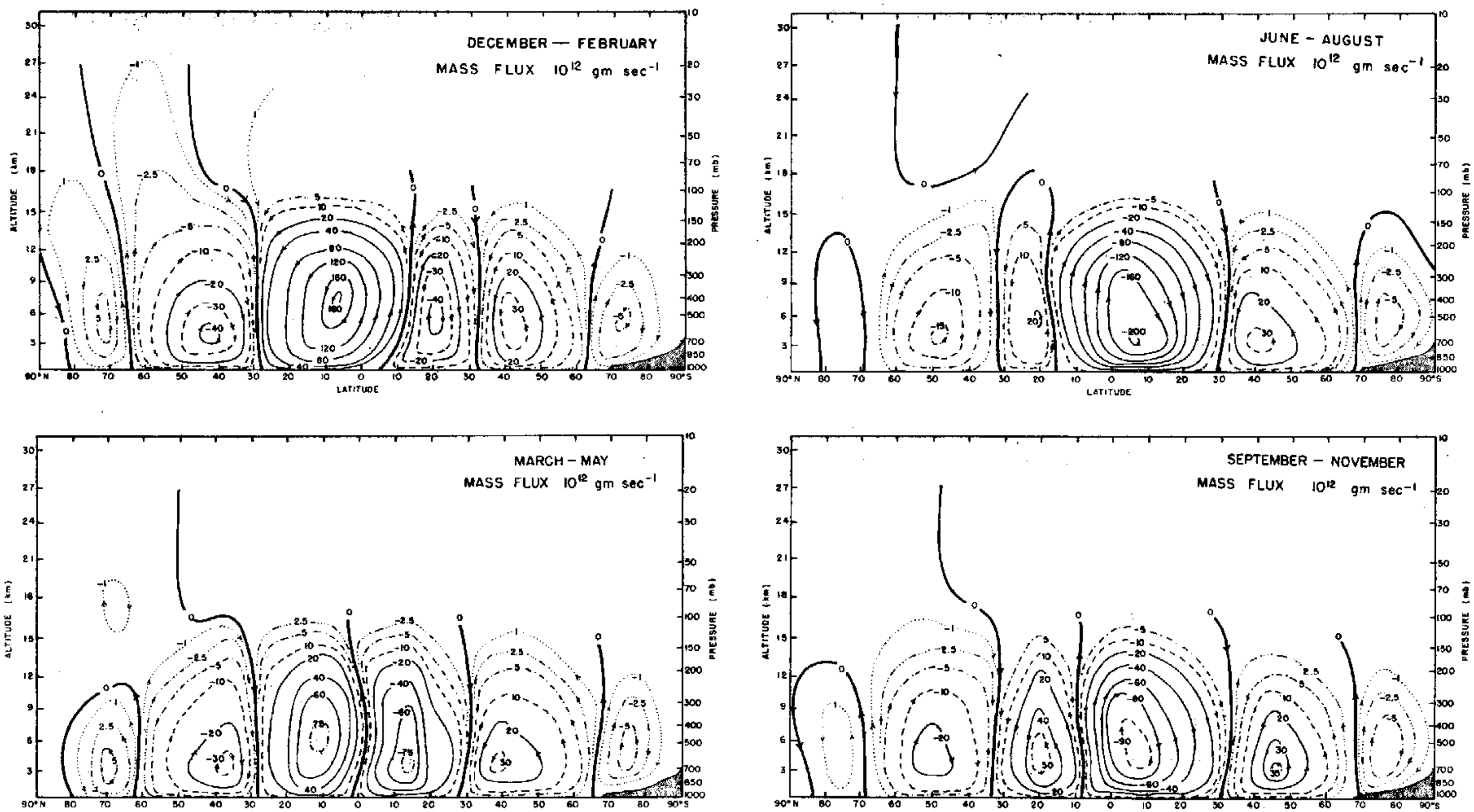


Figure 3.19

Concept of eddy fluxes:

$$\nabla \bullet \overline{\rho \mathbf{V} E} + F_{rad_{TOA}} - \left(F_{rad} + F_{conv} \right)_{surface} = 0$$

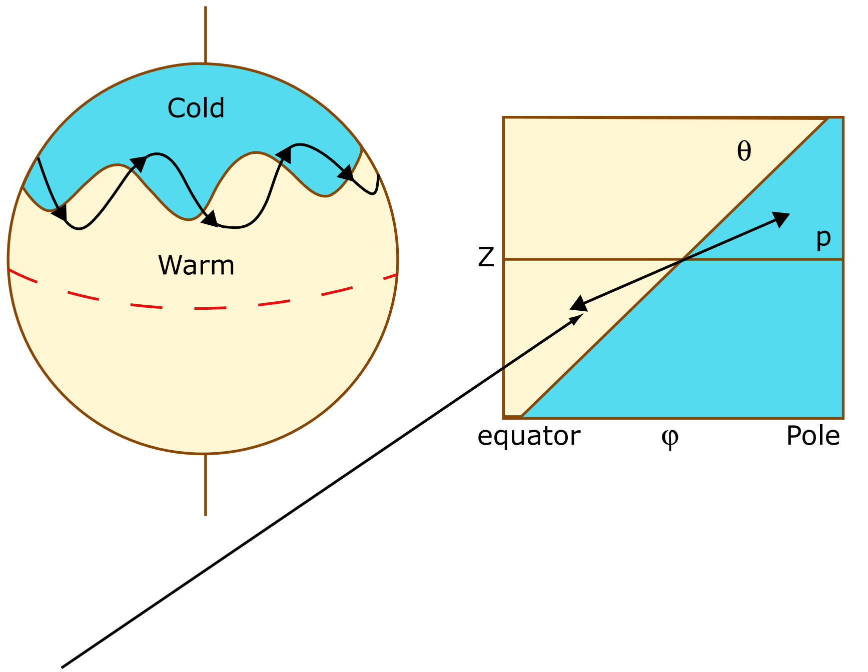
$$\rho \mathbf{V} = \{\rho \mathbf{V}\} + \rho \mathbf{V}',$$

$$E = \{E\} + E',$$

$$\text{where } \{X\} \equiv \frac{1}{2\pi} \int_0^{2\pi} X \, d\lambda$$

$$\rightarrow \nabla \bullet \left[\overline{\{\rho \mathbf{V}' E'\}} + \overline{\{\rho \mathbf{V}\}\{E\}} \right] + F_{rad_{TOA}} - \left(F_{rad} + F_{conv} \right)_{surface} = 0$$

Atmospheric Heat Transport



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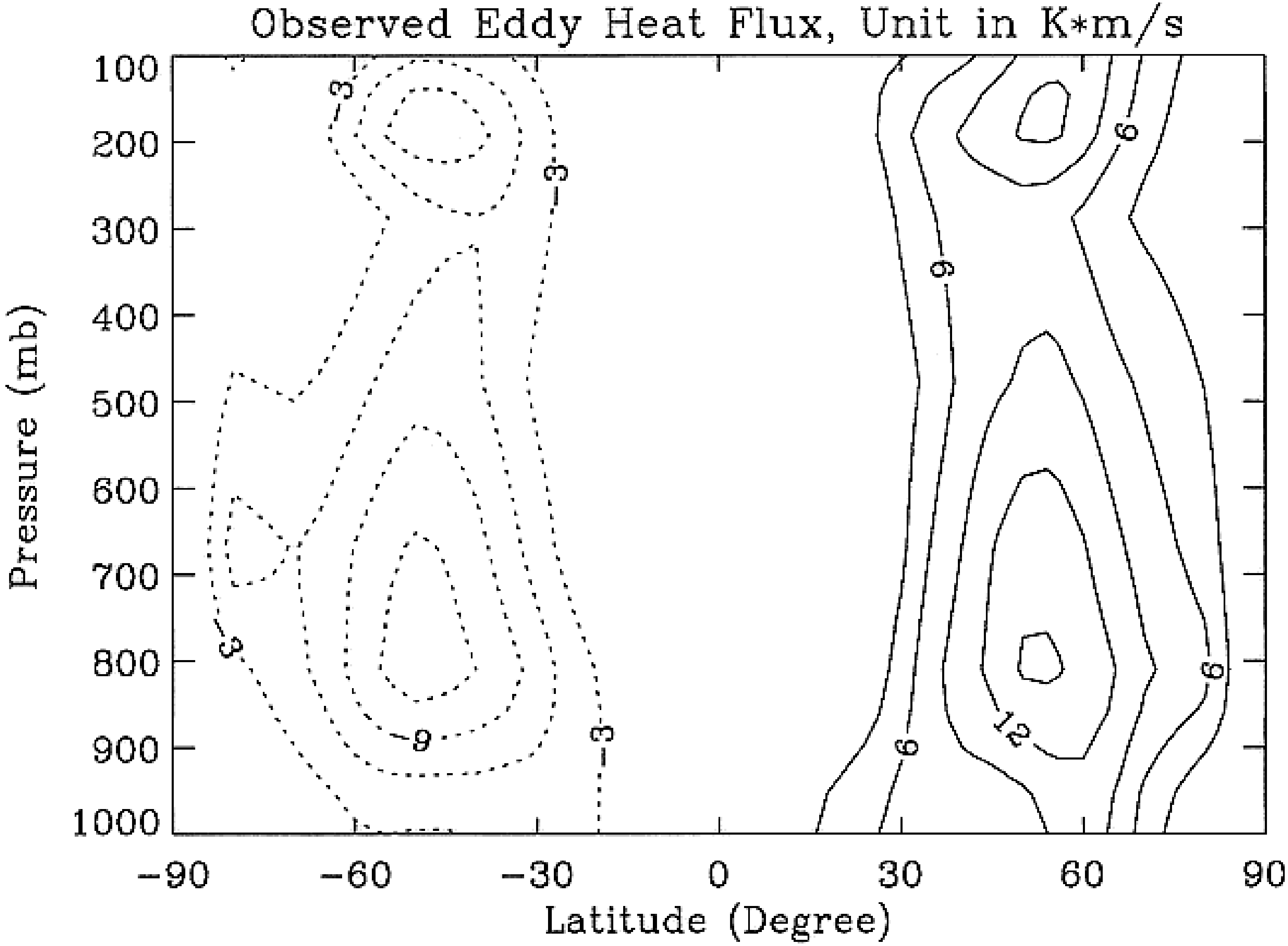
Efficient poleward heat transport

Image by MIT OpenCourseWare.



Observed annual mean eddy heat flux, after Oort and Peixoto, 1983

Eddy heat fluxes not efficient enough to prevent temperature gradients from developing

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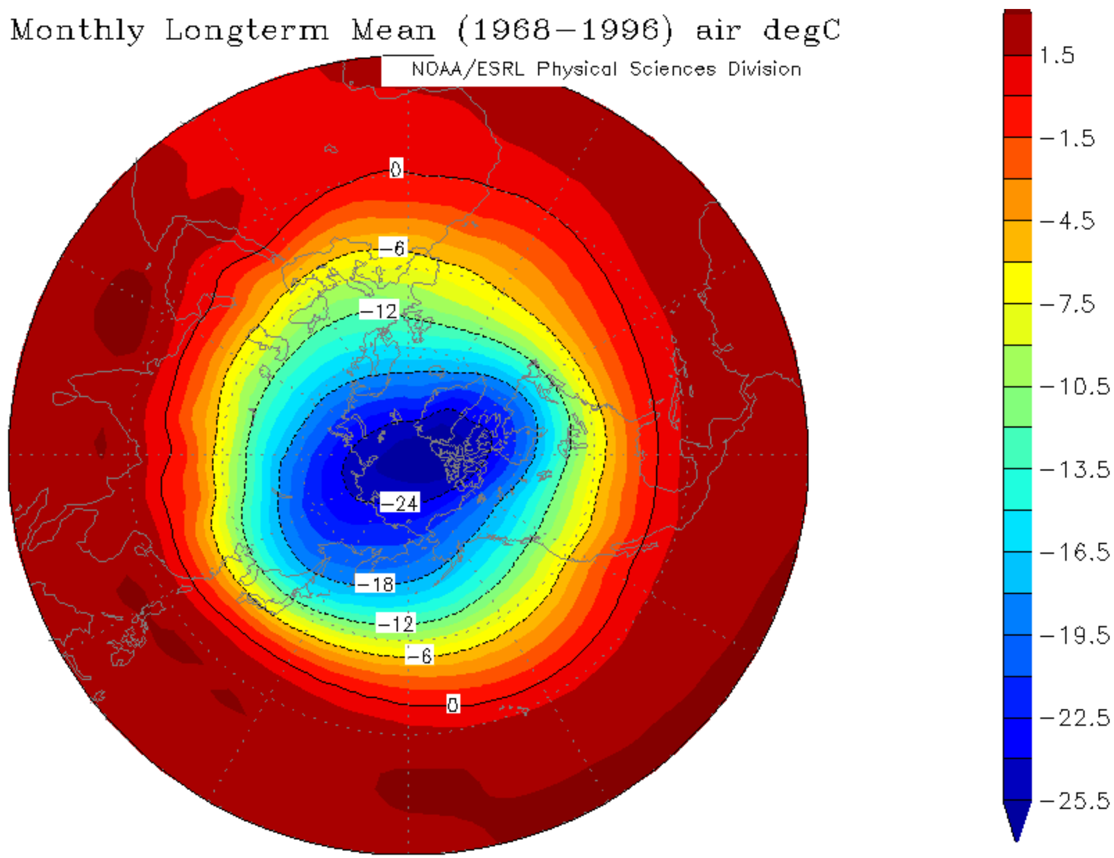
Please see Figure 1.6 in the book
Hartmann, Dennis L. 1994. *Global physical climatology*.
San Diego: Academic Press. ISBN: 9780123285300.

What controls the temperature gradient in middle and high latitudes?



Annual mean temperature, northern hemisphere, about 4 km altitude

Image courtesy of NOAA.

Issues

- Temperature gradient controlled by eddies of horizontal dimensions ~ 3000 km
- Familiar highs and lows on weather maps
- Eddy physics not simple
- Concept of criticality does not apply...critical T gradient = 0 (not observed)
- While eddies try to wipe out T gradient, they do not succeed

Example of surface pressure distribution

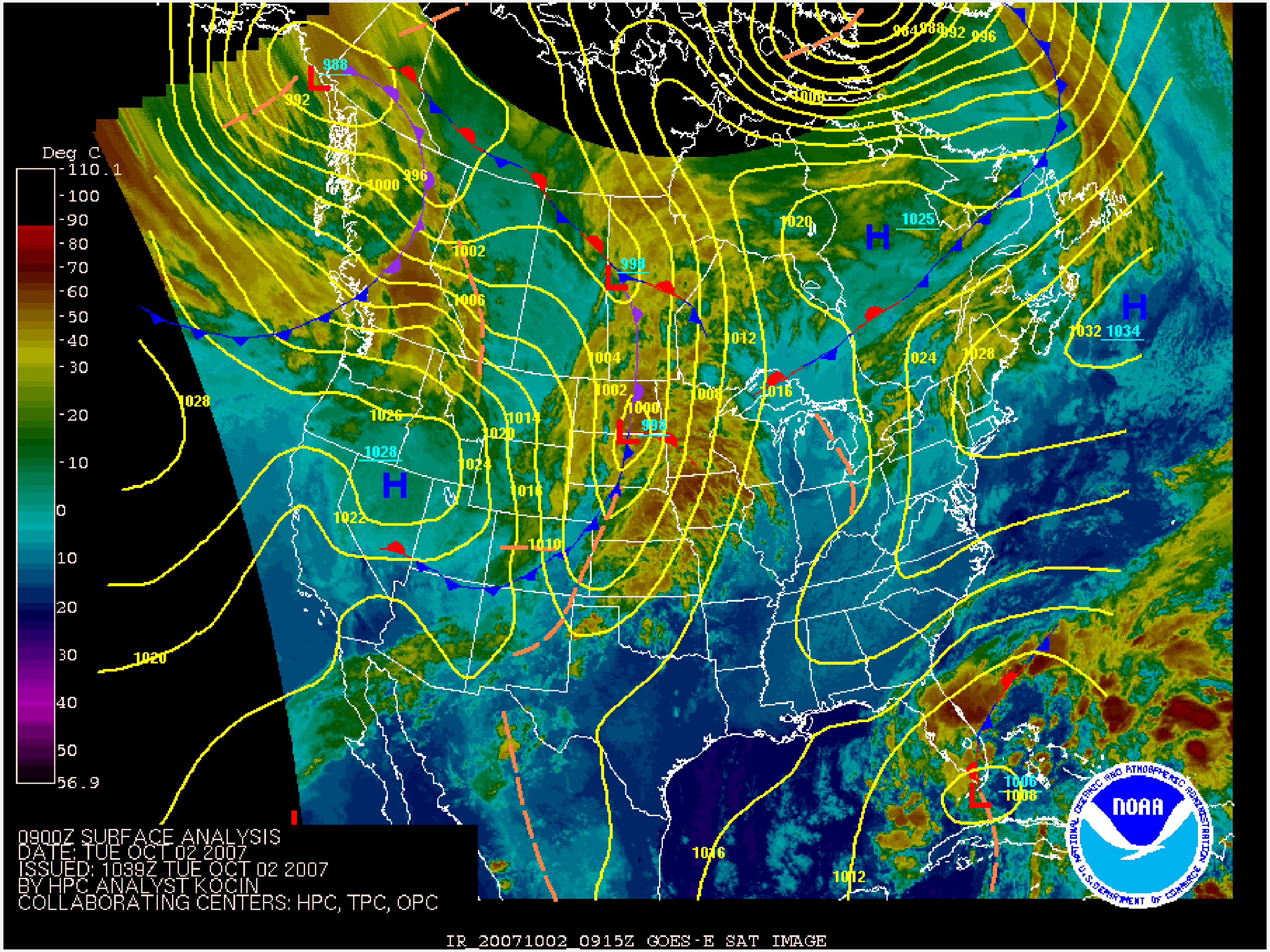


Image courtesy of NOAA.

The Oceans and Climate

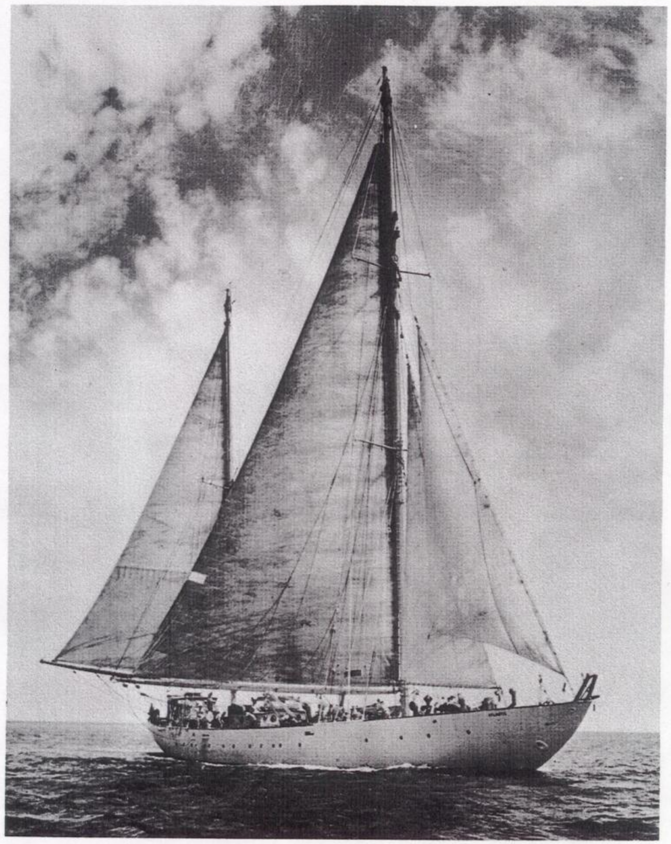
Major elements of the connection of the oceans with the climate system:

(1) The ocean has almost all of the water on the planet
(2) It has an immense heat capacity compared to the atmosphere
(3) It retains a memory of past disturbances that can extend to thousands of years

Consequently and in addition:
(3) It exchanges energy with the atmosphere (heat, moisture) and *transports it* in very large amounts
(4) It absorbs, stores, and ejects carbon dioxide in very large amounts
(5) It is the site of a large fraction of the biological activity on Earth.
(6) It is a major component of the biogeochemical cycles (nitrogen, phosphorous, etc.) on Earth
(7) When frozen, it can undergo a very large albedo change

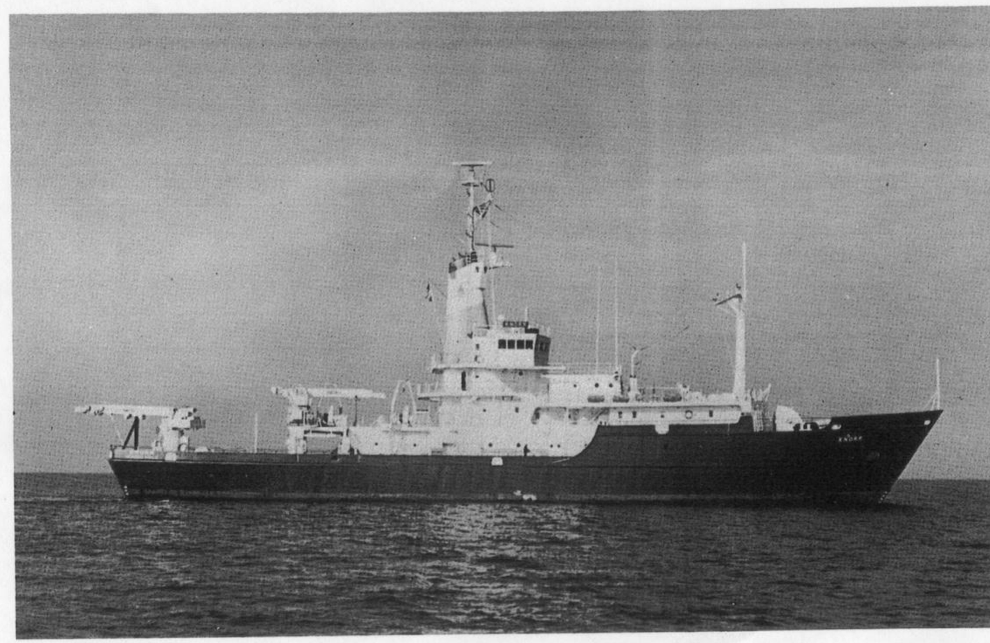
- Similarities to atmosphere:
  - Global-scale fluid on a rotating earth
  - Governing equations nearly identical
- Major differences:
  - Continental barriers to east-west motions
  - Ocean virtually opaque to radiation
  - The ocean is heated (and cooled) at its upper surface
  - No equivalent in the ocean of moist convection (although salinity introduces some analogous issues)
  - Much more difficult to observe the ocean

How does one determine what the ocean does? Dynamically, the ocean is much like the atmosphere, but the problem of observing it is radically different, and the difference in observational technologies has strongly influenced inferences about the ocean circulation. It is essential to understand what is really known, and what is surmised from limited data.



R/V *Atlantis I*

A 142-foot steel ketch built in Copenhagen in 1931 for Woods Hole Oceanographic Institution. In 1964 *Atlantis* was sold to the Consejo Nacional de Investigaciones Cientificas y Technicas, Buenos Aires, Repulica de Argentina, where she was renamed *El Austral.*



R/V *Knorr* (AGOR 15)

Built in 1970 by Defoe Shipbuilding Company of Bay City, Michigan, the *Knorr,* a sister ship of the R/V *Melville* (AGOR 14), features the use of cycloidal propellers for propulsion, position keeping, and maneuverability. The Navy has assigned her to Woods Hole Oceanographic Institution.

Length: 245' Displacement: 2,075 tons

Image courtesy of Woods Hole Oceanographic Institution. Used with permission.

Hanging a Nansen bottle on a wire before lowering it into the sea.

Effect of Long Memory

Suppose that the ocean is just a non-moving slab that stores heat coming in from the atmosphere and gives it back when required. Such a system might be modeled by a simple equation like

$$\kappa \frac{\partial T}{\partial t} = q$$

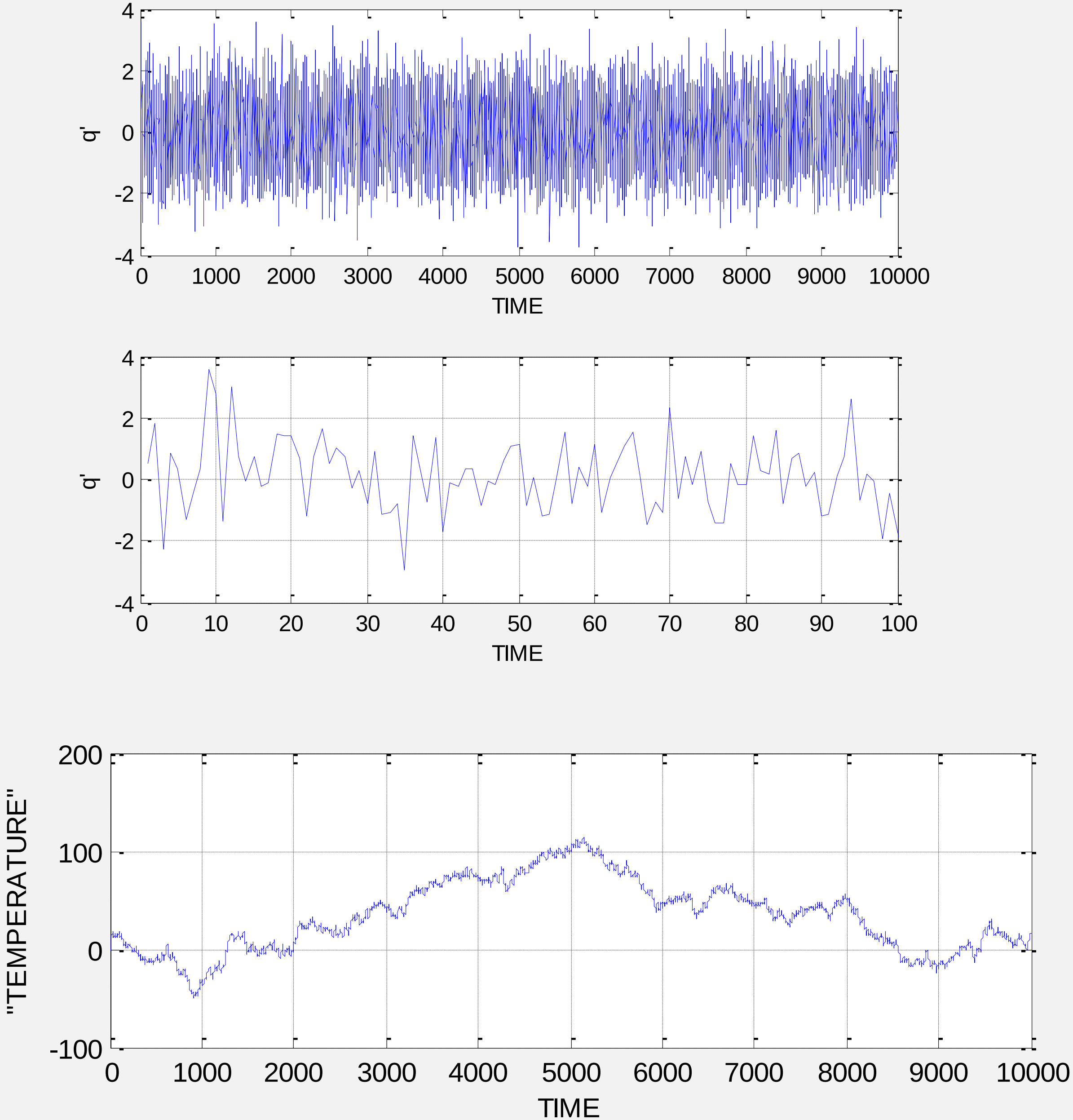
Re-write it as a simple finite difference,

$$T((n+1)\Delta t) - T(n\Delta t) = \frac{\Delta t}{\kappa} q(n\Delta t)$$

or,

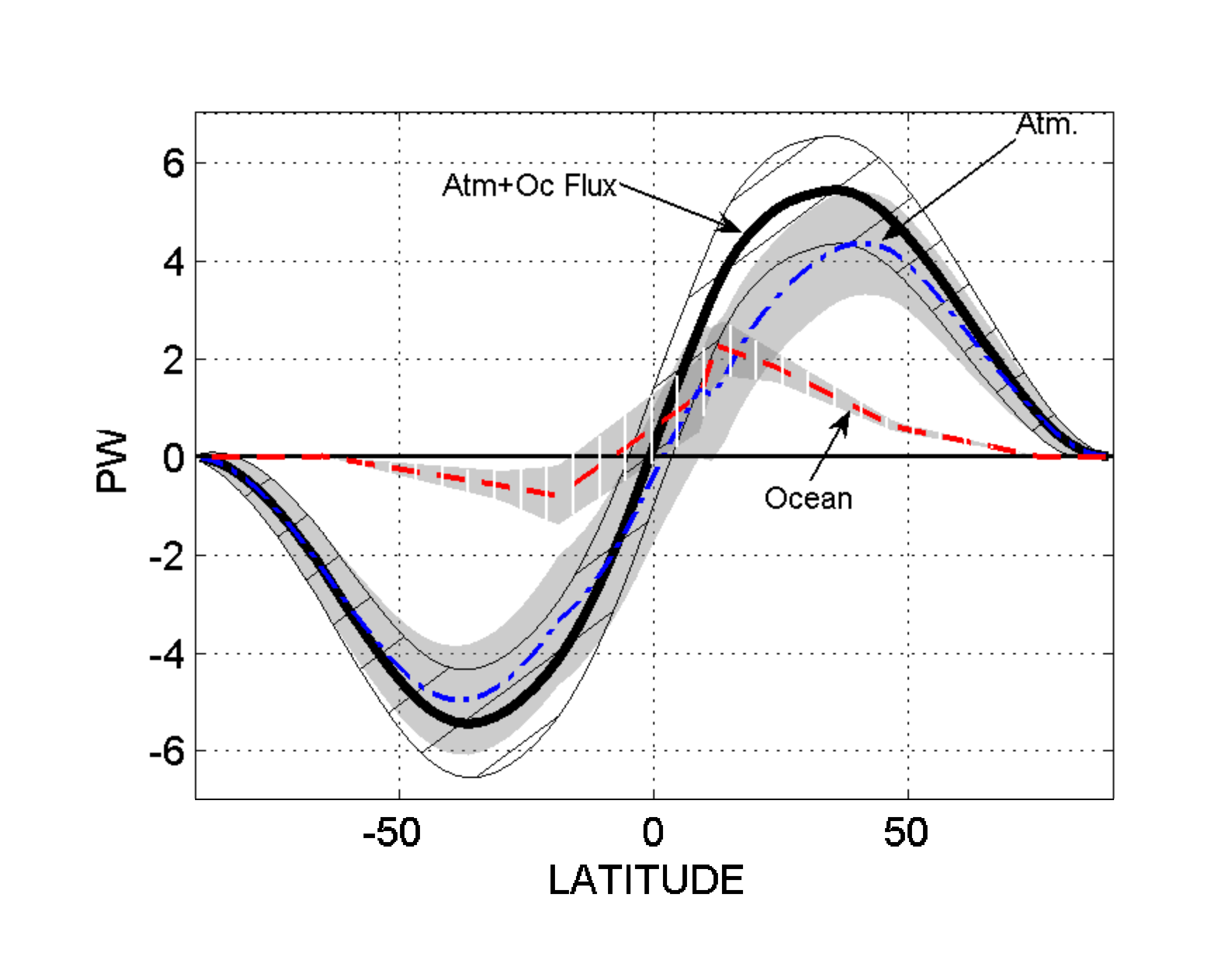
$$T((n+1)\Delta t) = T(n\Delta t) + q'(n\Delta t)$$

where we think of q' as being “weather noise” and treat it as completely random. Easy then to compute $T((n+1)\Delta t)$ from $T(n\Delta t)$.



q'
4
2
0
-2
-4
0
1000
2000
3000
4000
5000
6000
7000
8000
9000
10000
TIME
q'
4
2
0
-2
-4
0
10
20
30
40
50
60
70
80
90
100
TIME
"TEMPERATURE"
200
100
0
-100
0
1000
2000
3000
4000
5000
6000
7000
8000
9000
10000
TIME

1 PW= 1 PetaWatt=10^{15} W



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Example of what the ocean does---meridional transport of heat away from the equator. Does this change? Can it change?

Wind-Driven Circulation

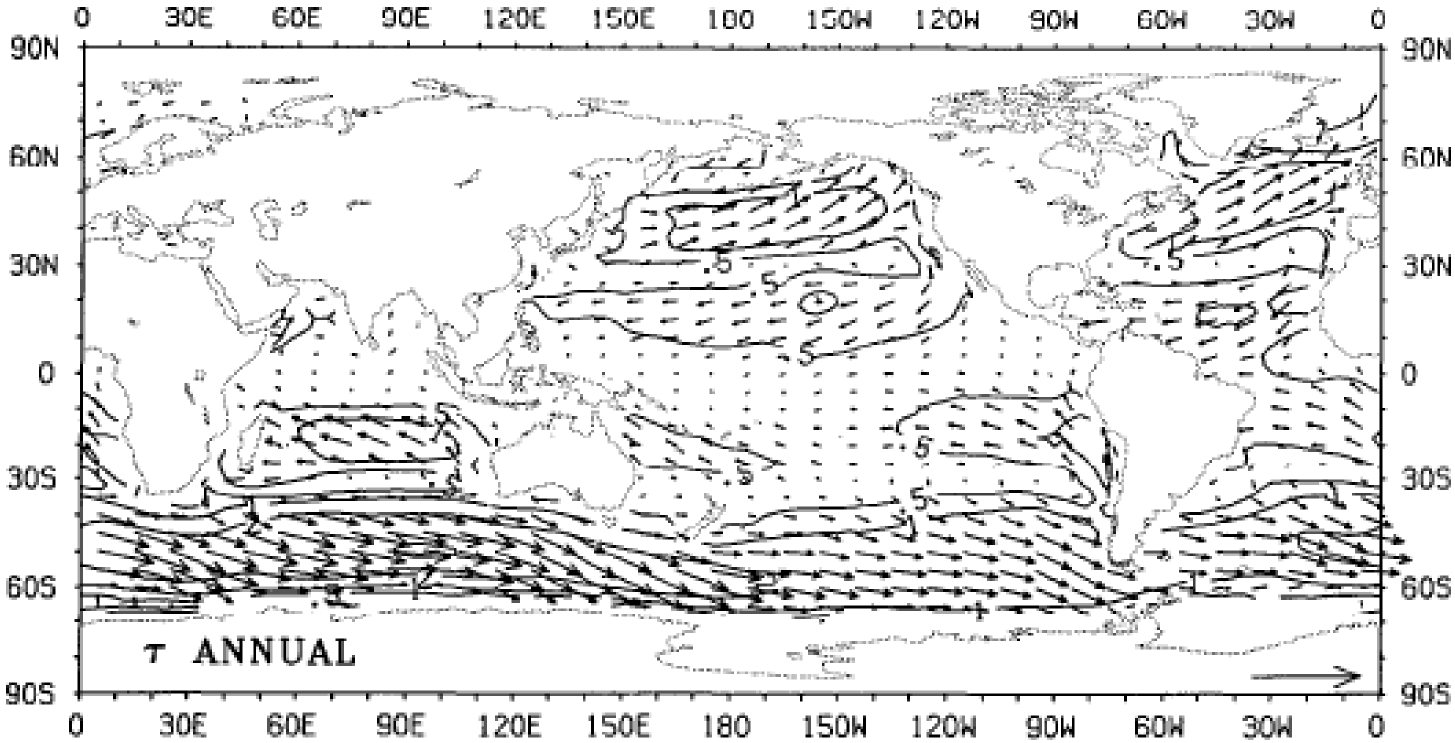

FIG. 1. Annual mean wind stress over the global oceans depicted as vectors. The arrow at bottom right corresponds to 5 dyn cm^{-2} and contours of magnitude of 0.5, 1, 2 and 3 dyn cm^{-2} are plotted.

Climatological wind field over the ocean.
Trenberth et al. (1990, JPO) from ECMWF. Patterns of dominant westerlies and easterlies. Low-latitude easterlies are called “trade-winds”.

Deep Overturning

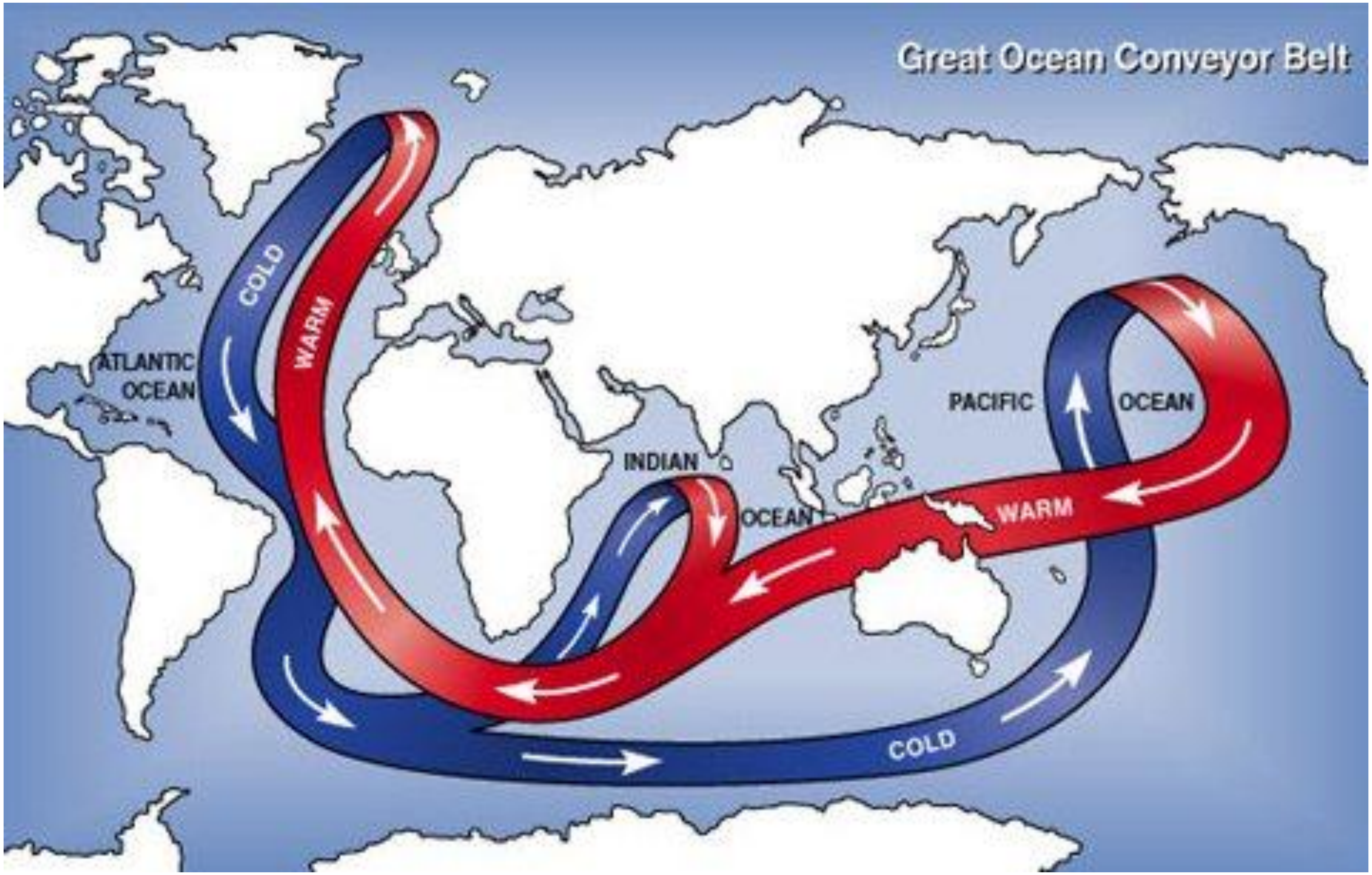

Image courtesy of Argonne National Laboratory.

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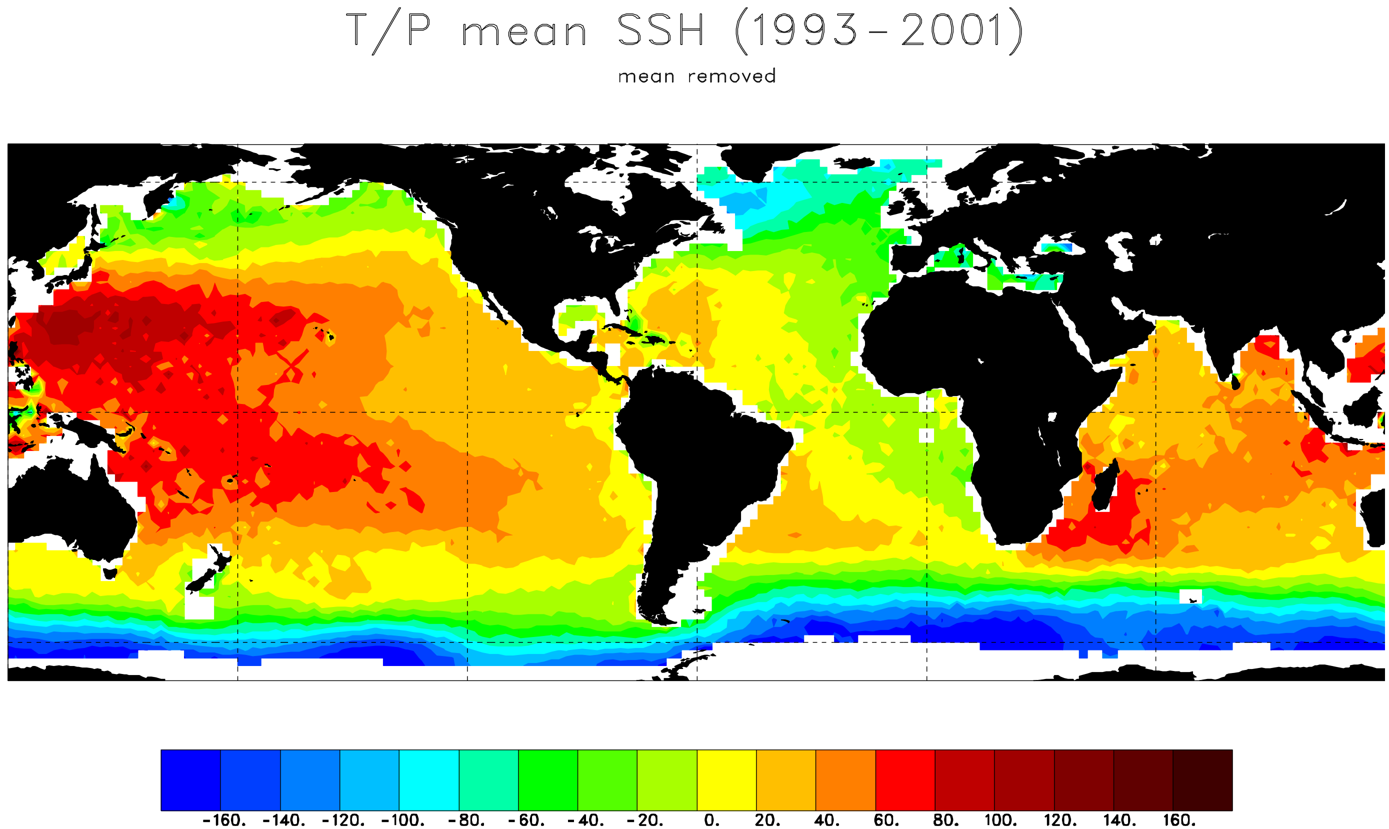
Please see Figure 1-2-6 on page http://www.tpub.com/weather3/1-24.htm.

By 1992 it had become technically possible to measure the absolute topography of the sea surface from space using orbiting altimeters. There are still major problems, many having to do with the inability to determine the shape of the Earth accurately enough, but the system is adequate to show the major inferred oceanic features:



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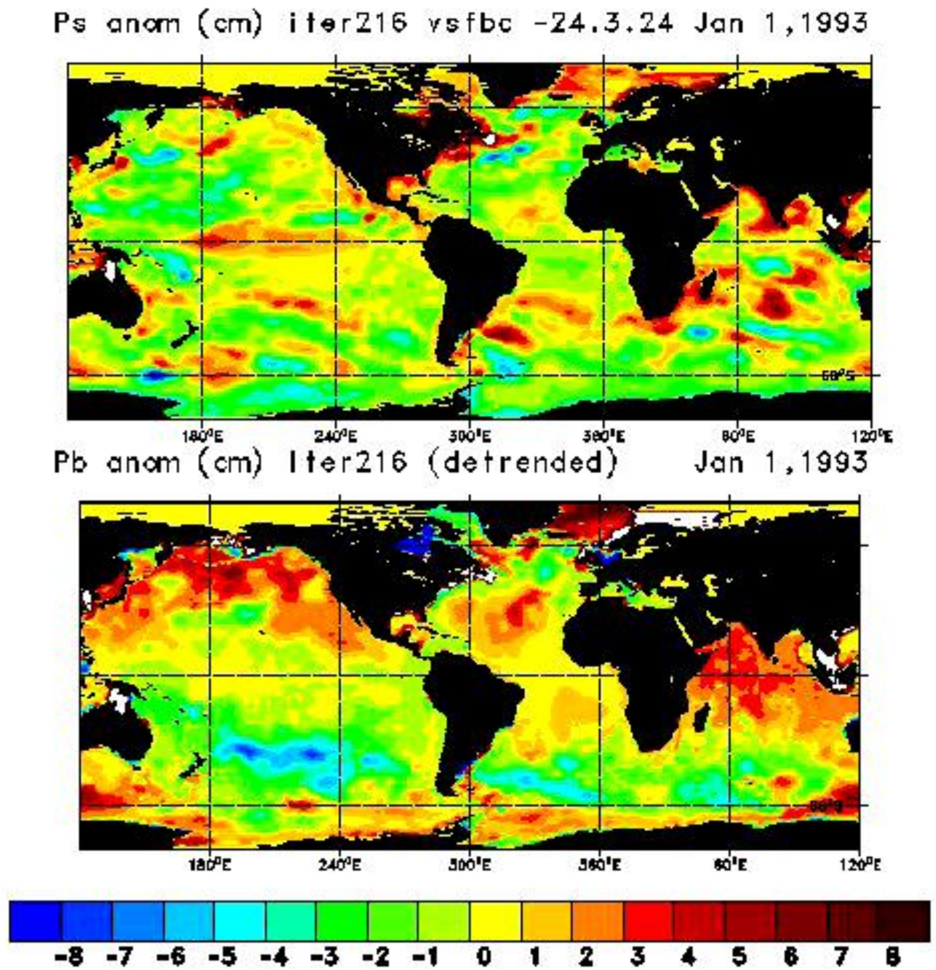
Heights are in centimeters relative to a surface defined to be a gravitational equipotential.

An animation. See http://ecco-group.org



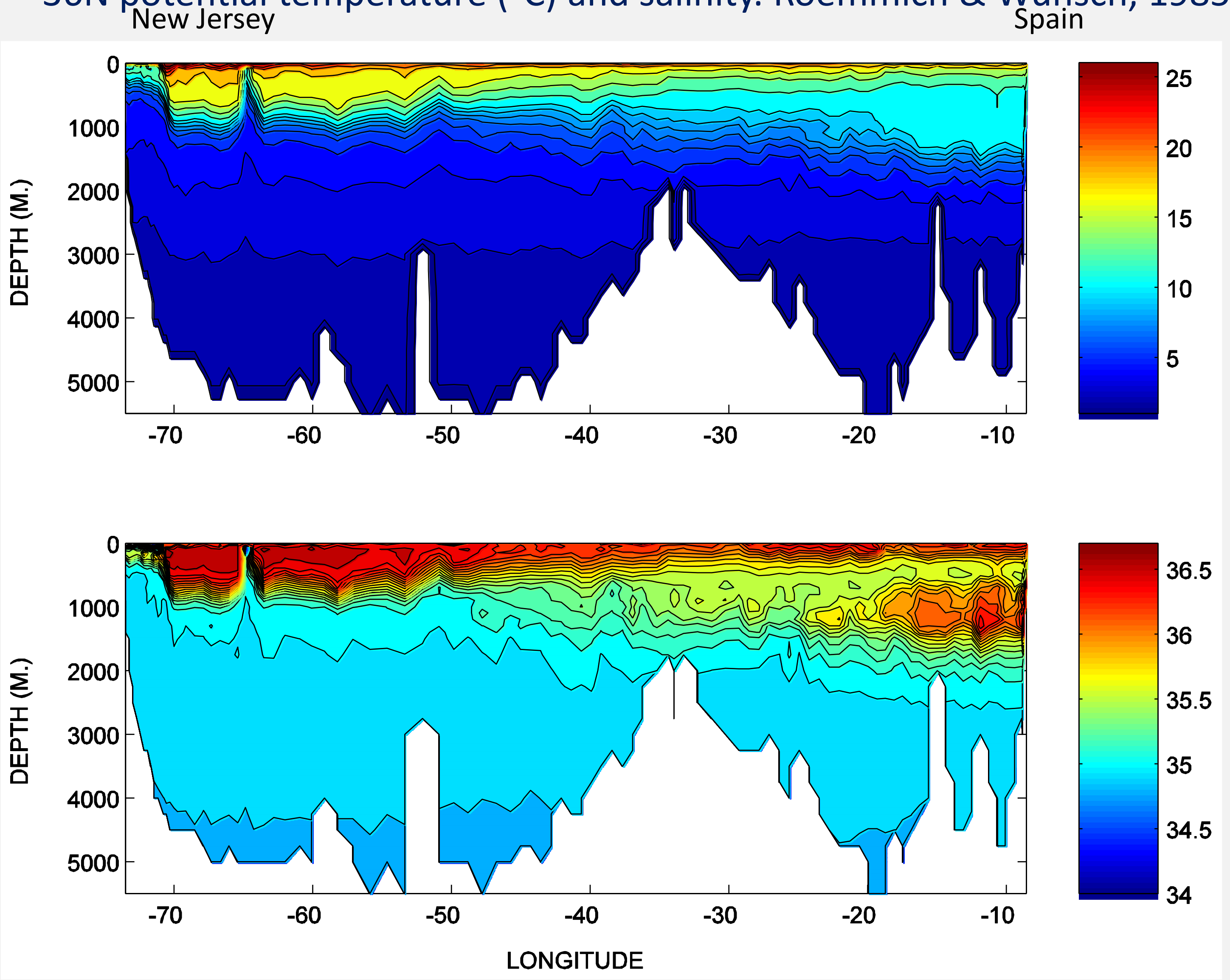
surface elevation, cms

bottom pressure, cms

Image courtesy of Carl Wunsch. Used with permission.

From Estimating the Circulation and Climate of the Ocean (ECCO) consortium Results.

36N potential temperature (°C) and salinity. Roemmich & Wunsch, 1985



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Salinity is formally dimensionless, but is very close to being mass in parts/thousand.

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The image is from the World Ocean Circulation Experiment (WOCE). Global coverage from about 1992-1997.

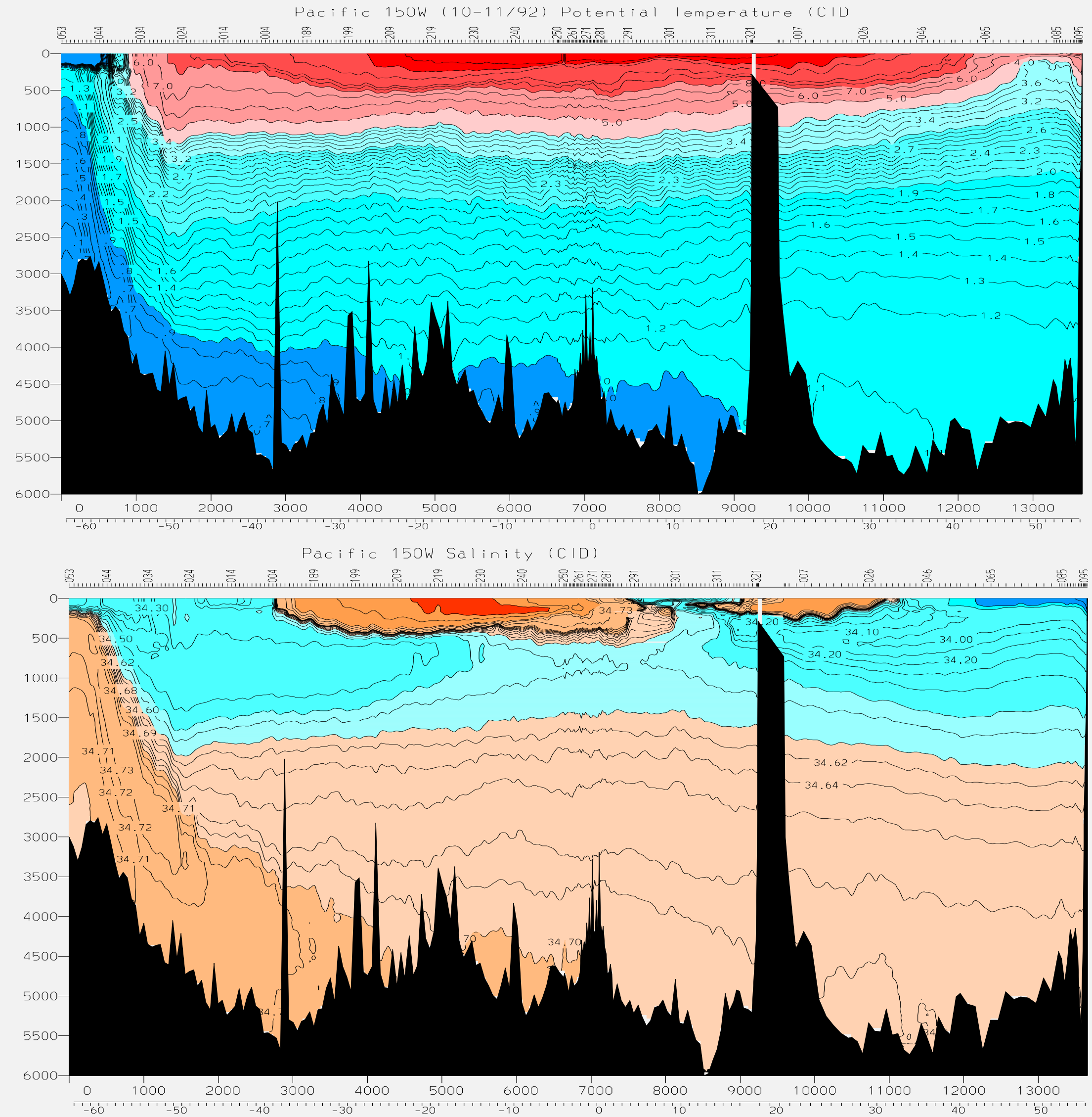
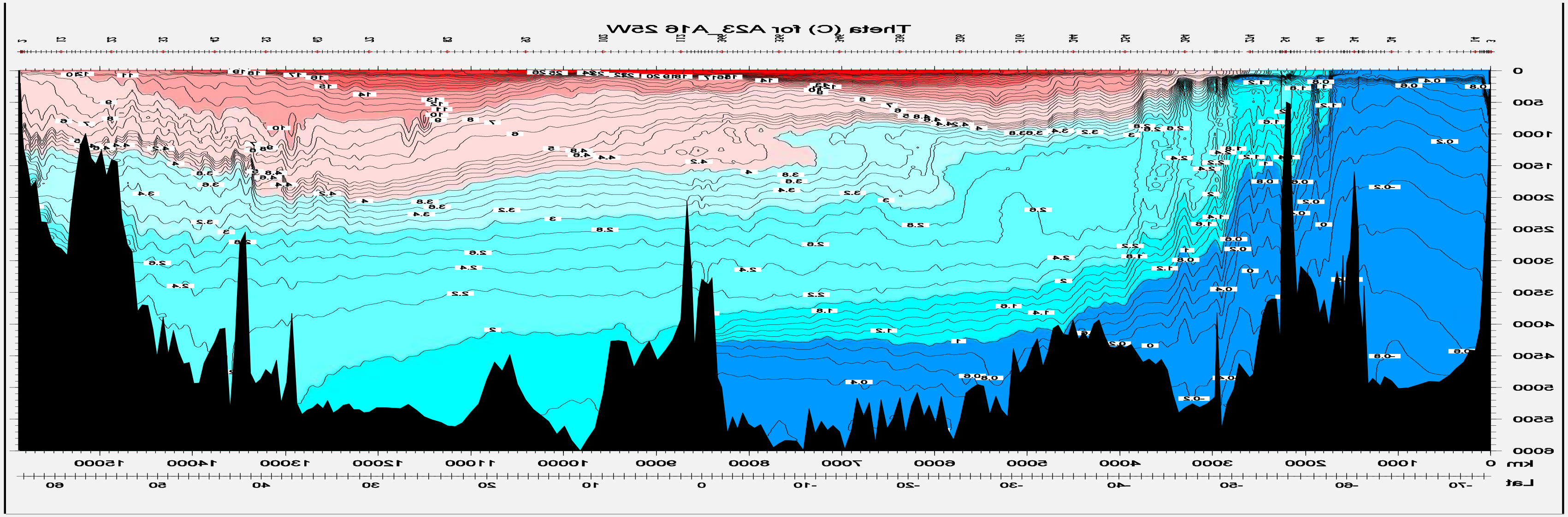
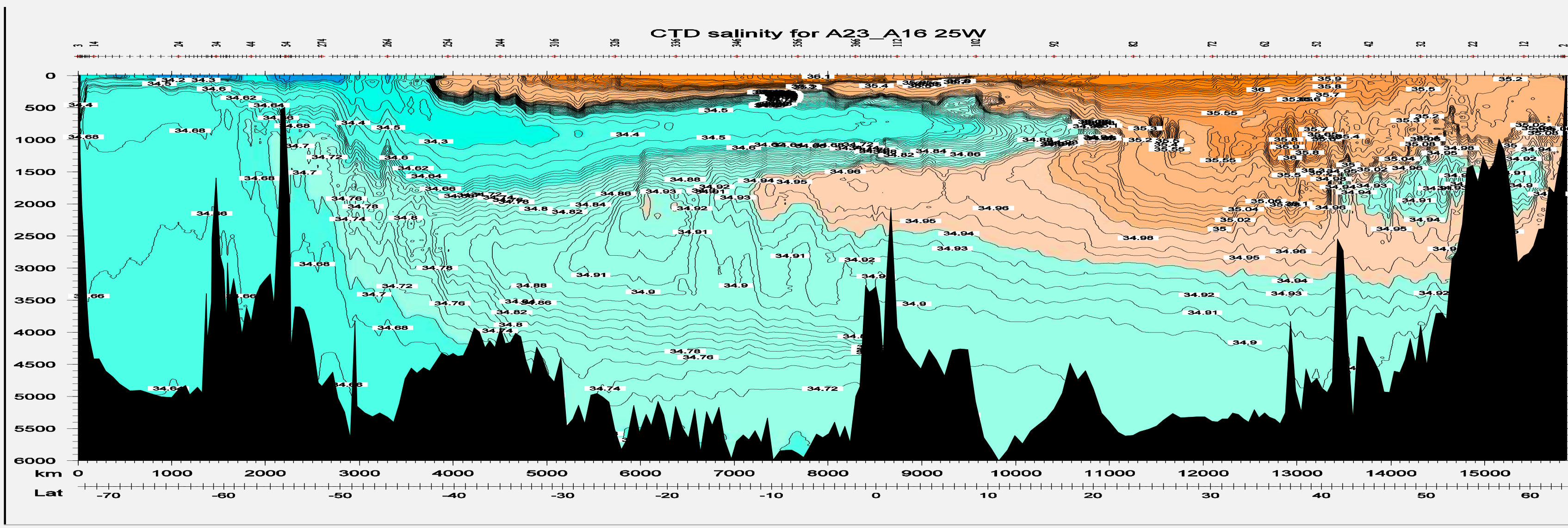


Image courtesy of NOAA.

A6, A23 30W in N. Atlantic Pot. Temp and Salinity





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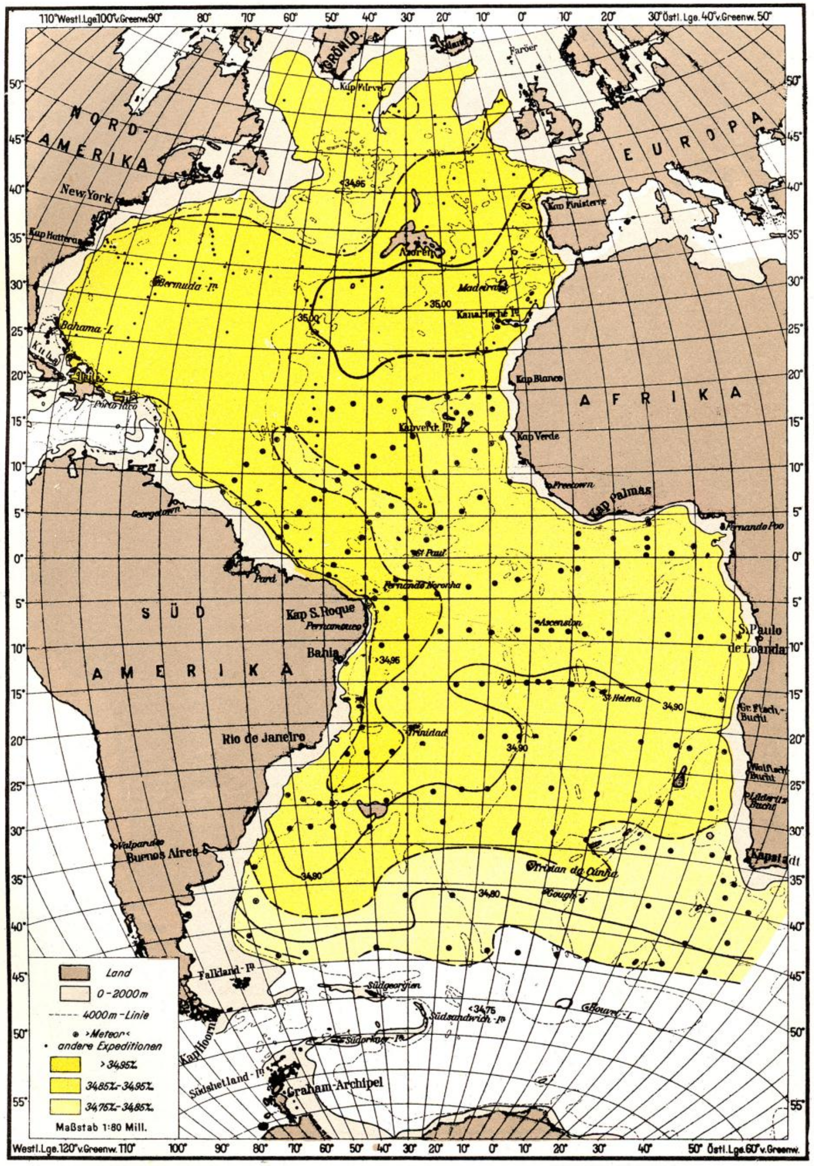
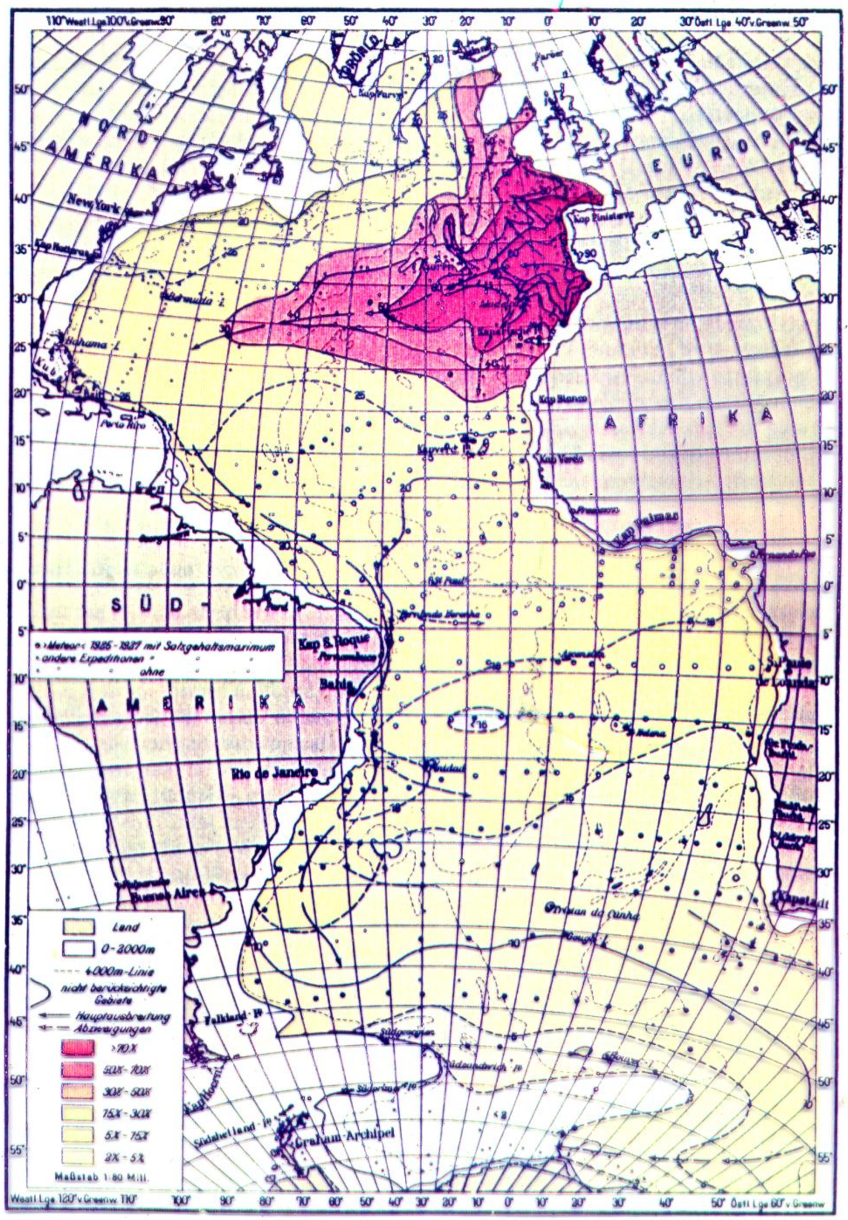


Salinity (‰) of the core layer (intermediate oxygen maximum) of the Middle North Atlantic Deep Water.



Spreading of the Upper North Atlantic Deep Water in the core layer (intermediate salinity maximum), represented by the percentage of the Mediterranean component M.

Image courtesy of US government.

Two views of the meridional overturning circulation, showing vertical cross-sections from pole to pole. In each panel, the dashed lines denotes the base of the mixed layer and the thin sold lines are isopycnal surfaces. In the first view (top panel), Ekman pumping forces fluid down in one hemisphere and up in the other, and flow below the mixed layer is adiabatic. Since all density sources and sinks are in the mixed layer an thus approximately at the same pressure, buoyancy does no net work on the circulation. In the second view (bottom panel), cold water sinks in either or both hemispheres at high latitudes and returns by flowing across isopycnal surfaces on the warm side of the system. Downward turbulent mixing of warm water must balance upward advection by the mean flow in the return branch of the circulation. Net, column-integrated heating must occur in the Tropics in this case, while most of the cooling is applied in the mixed layer. The induced poleward heat flux cannot extend poleward of the isopycnal surface (heavy curve) corresponding to the greatest depth to which mixing extends in the Tropics . These two views are not mutually exclusive.