OCN/ATM/ESS 587

Ocean circulation, dynamics and thermodynamics….

Equation of state for seawater

General *T*/*S* properties of the upper ocean

Heat balance of the upper ocean

Upper ocean circulation

Deep circulation

$\rho = \rho\,(S, T, p, \ldots\ldots)$ **[Determined from laboratory experiments]**

[*S* = salinity; *T* = temperature; *p* = pressure]

Seawater is a complex chemical system that contains traces of nearly all naturally occurring elements. The dissolved part of seawater is about 78% NaCl by mass.

The chemical composition of seawater….

Ion	‰ by weight		
chloride, Cl^-	18.980	negative ions (anions) total	= 21.861‰
sulphate, SO_4^{2-}	2.649		
bicarbonate, HCO_3^-	0.140		
bromide, Br^-	0.065		
borate, $H_2BO_3^-$	0.026		
fluoride, F^-	0.001		
sodium, Na^+	10.556	positive ions (cations) total	= 12.621‰
magnesium, Mg^{2+}	1.272		
calcium, Ca^{2+}	0.400		
potassium, K^+	0.380		
strontium, Sr^{2+}	0.013		
		overall total salinity	= 34.482‰

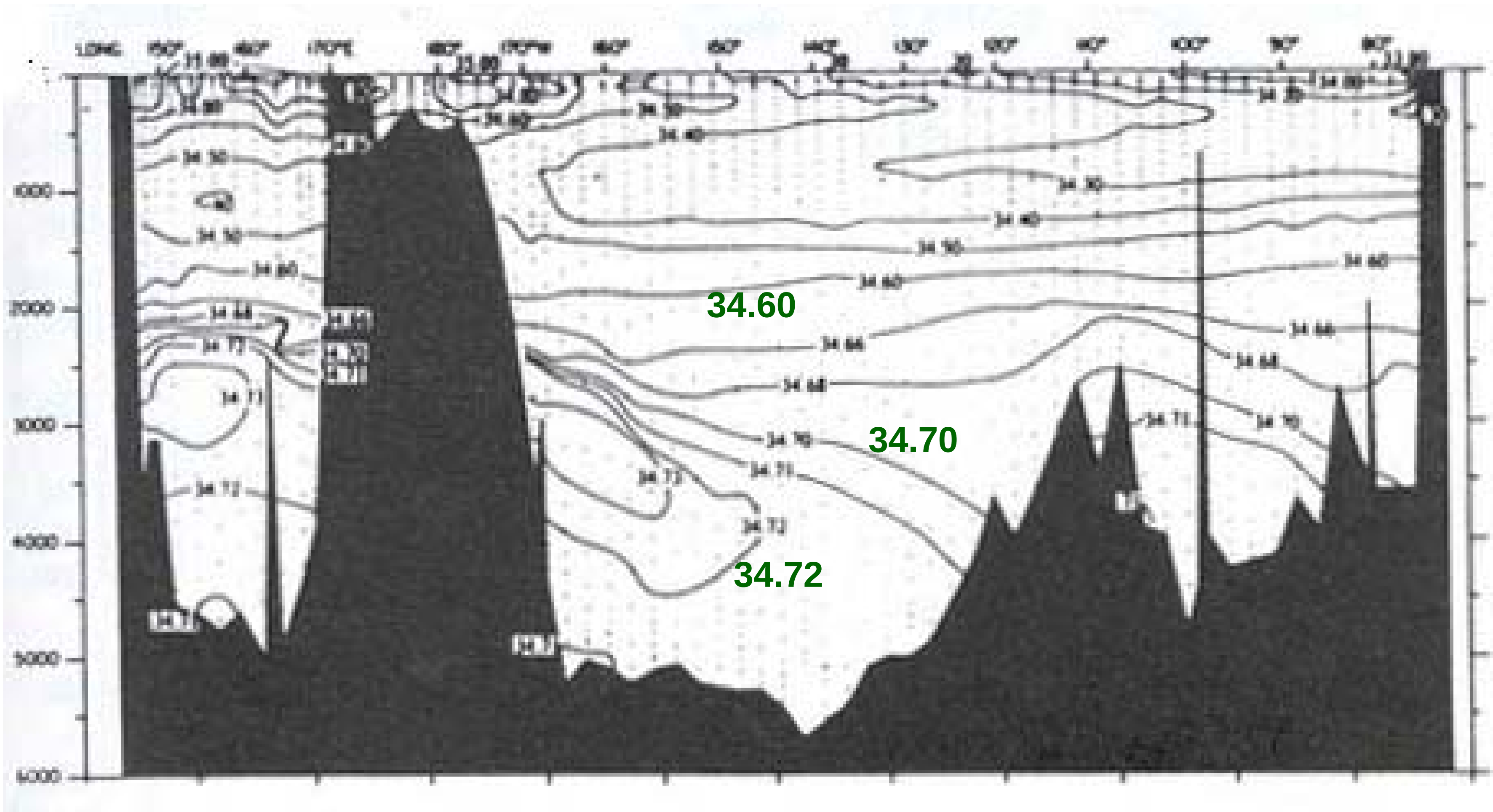
Amazingly, the proportions of the major constituents of seawater are nearly constant *everywhere*.

This constancy of proportions means that to a good approximation, only one component of seawater needs to be measured, and all of the other can be inferred from it. (what are the limitations?)

Instead of characterizing a seawater sample by knowing all of its chemical components, we can characterize the sample by a single parameter. We shall call this parameter the *salinity*.

Salinity = Mass of salt/(Mass of salt + Mass of water)

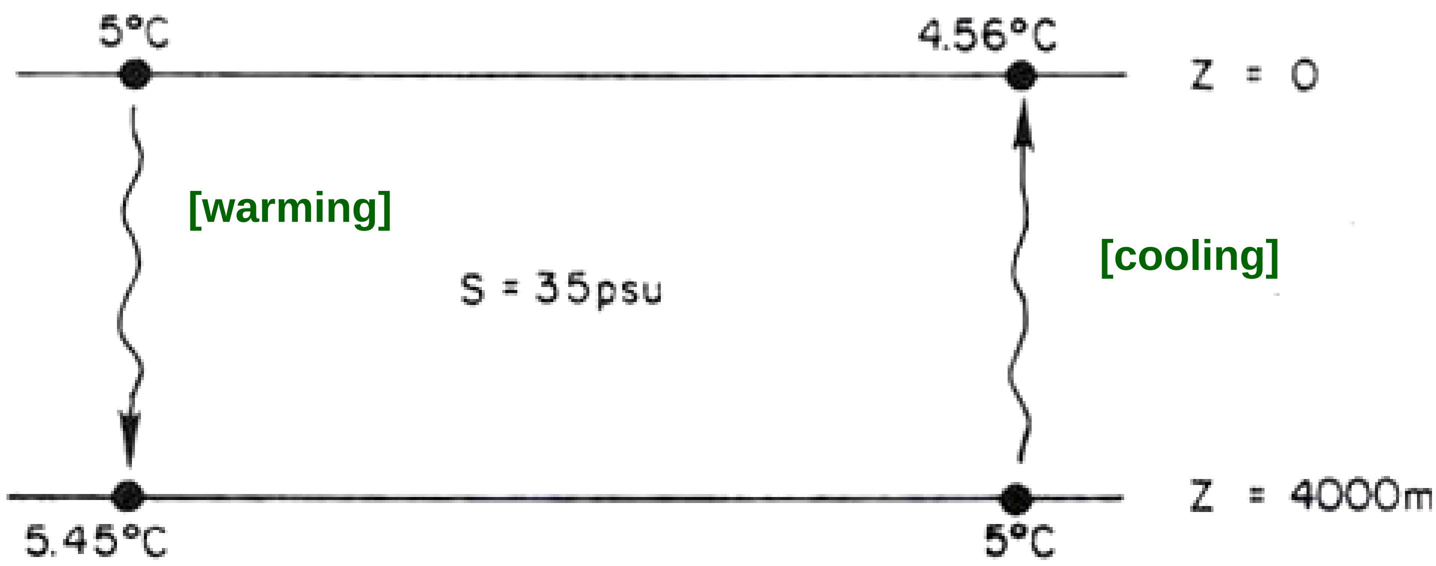
Typical ocean salinity ~ 0.035 = 35 PSU



How well do we need to measure *S*? The section of *S* along 43° S in the S. Pacific shows that there is considerable detail at signal levels ≤ 0.01 PSU.

Temperature (continued)….

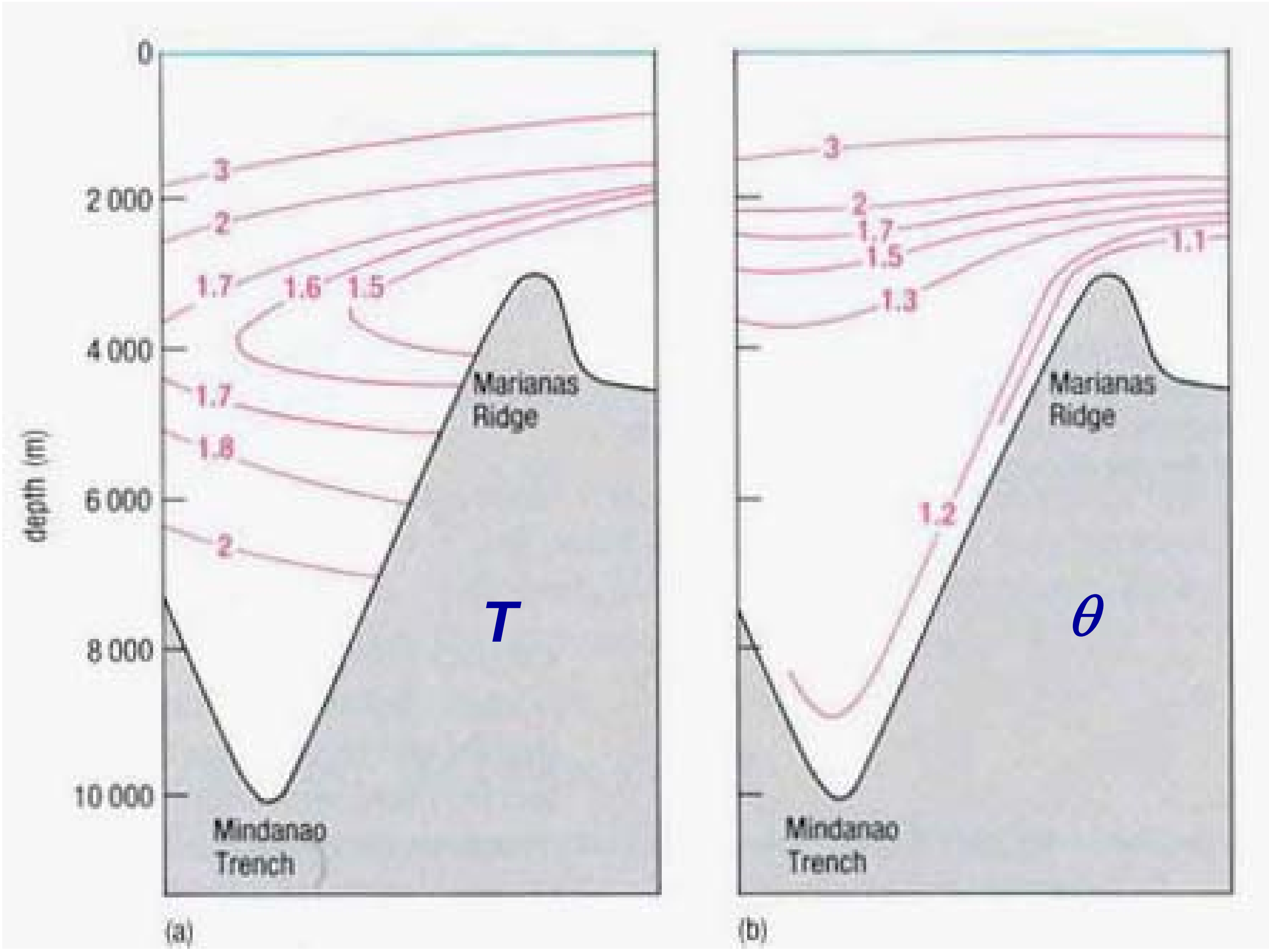
Seawater is slightly compressible. What are the implications of this?



Adiabatic **changes in temperature: changes in temperature when there are no changes in heat.**

$dQ = dq - dW$ (1st Law of thermodynamics)

Temperature (continued)….



In situ **temperature *T* increases as depth inceases.**

Potential temperature θ does not increase with depth.

Density….

$\rho = \rho\,(S, T, p)$ **is the symbolic equation of state.**

Over the water column, ρ varies by a few per cent.

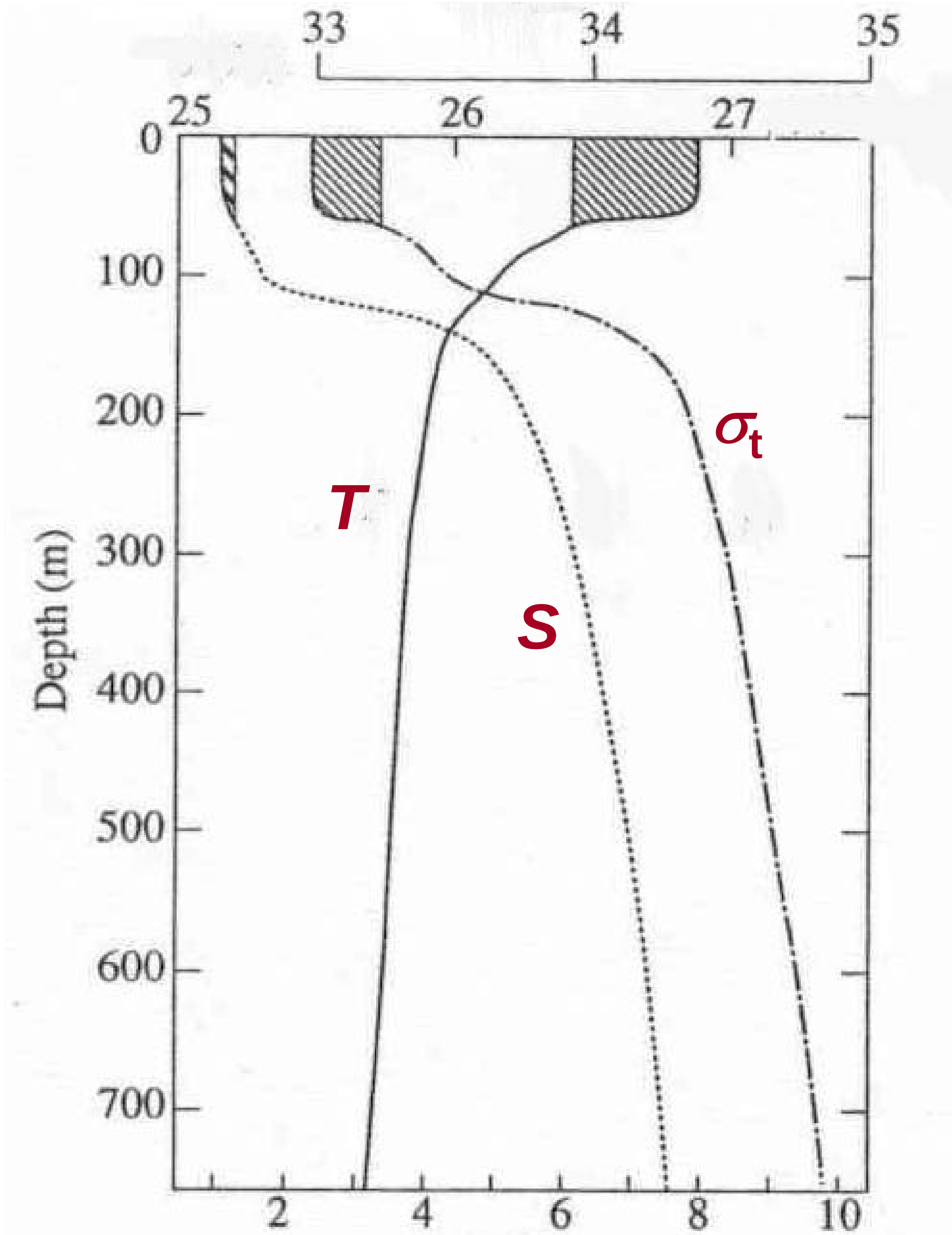
$\rho_{surf} \sim 1.021\ \mathrm{g/cm^3}$; $\rho_{bot} \sim 1.071\ \mathrm{g/cm^3}$ **[typical values]**

In the upper ocean, $\rho \sim \rho\,(S, T)$.
In the deep sea, $\rho \sim \rho\,(p)$.

Define the parameter σ as a more useful representation of density:

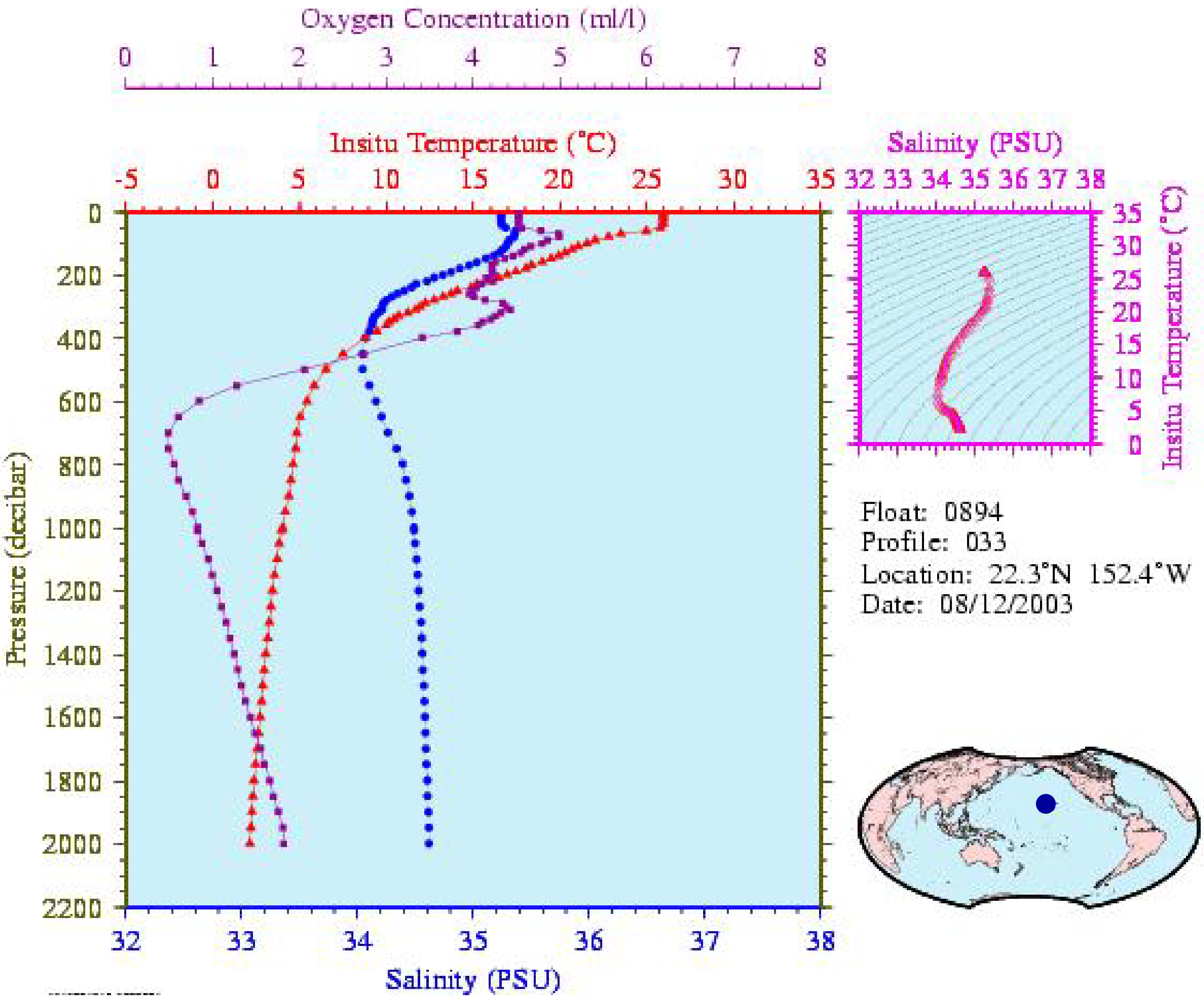
$\sigma = (\rho - 1) \times 1000$ **(cgs)** ; $\sigma = \rho - 1000$ **(mks)**

Typical profiles in the ocean….



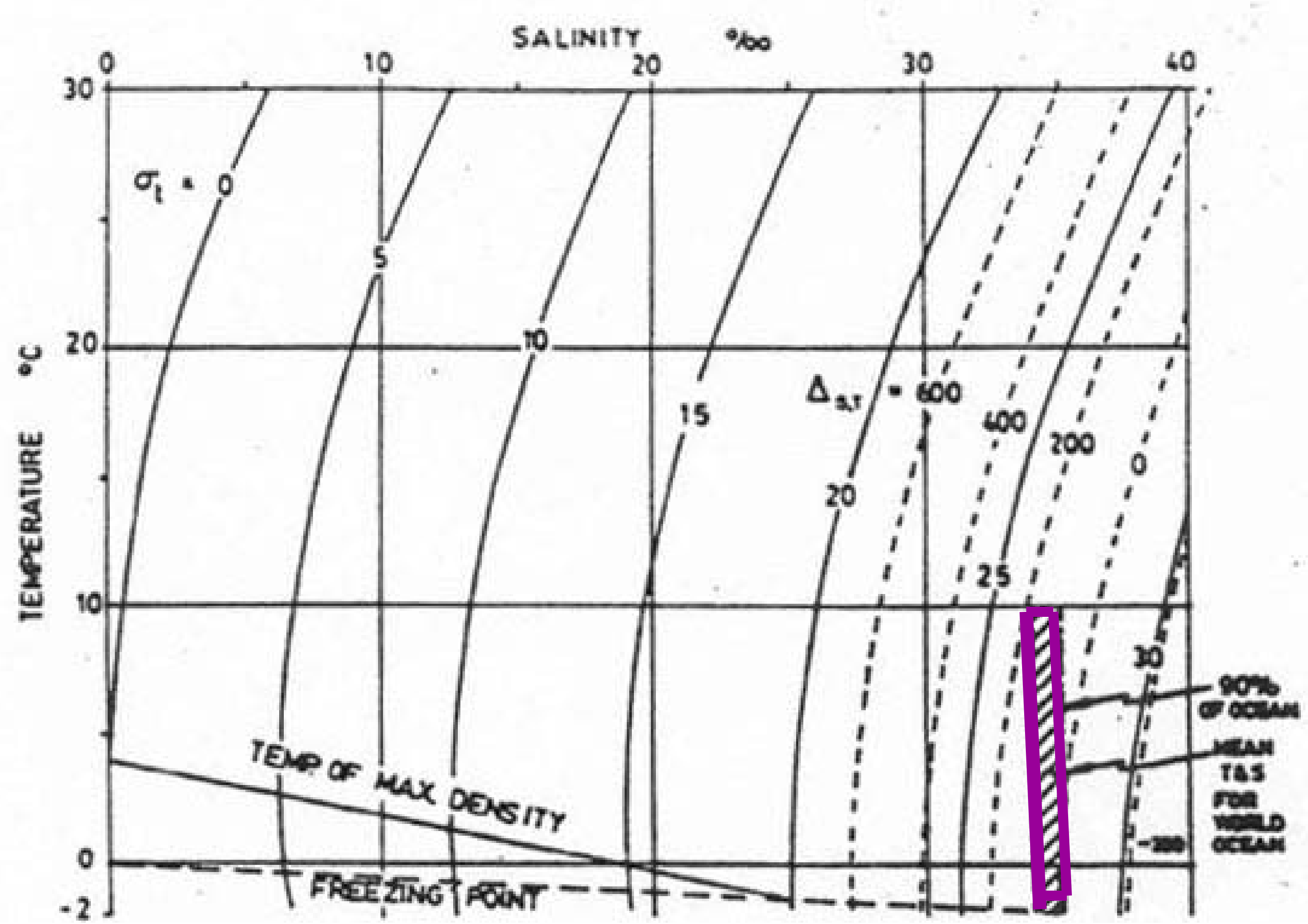
[from Ocean Station P, 50°N, 145°W]

Typical profiles (continued)….

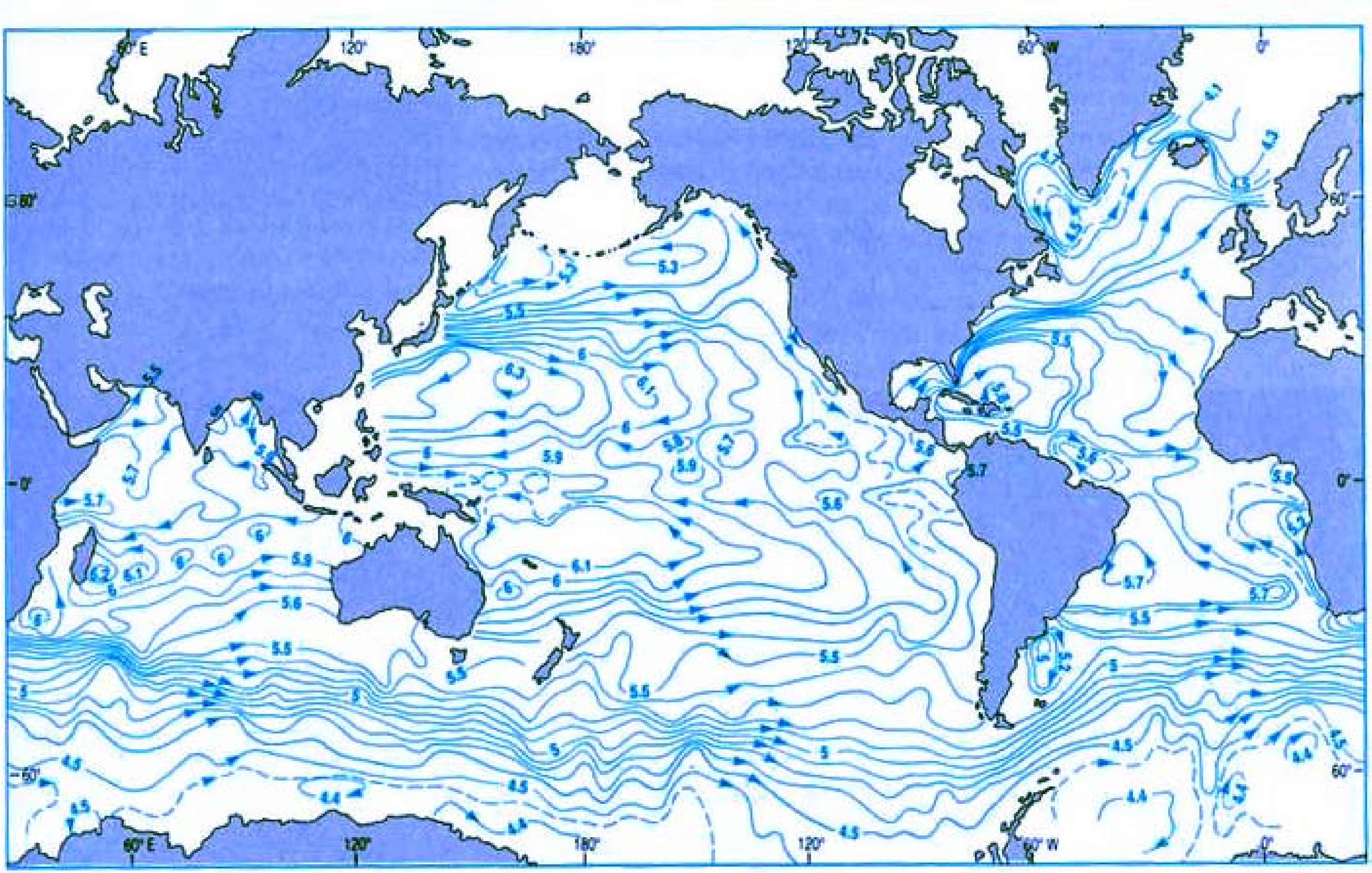


Equation of state….

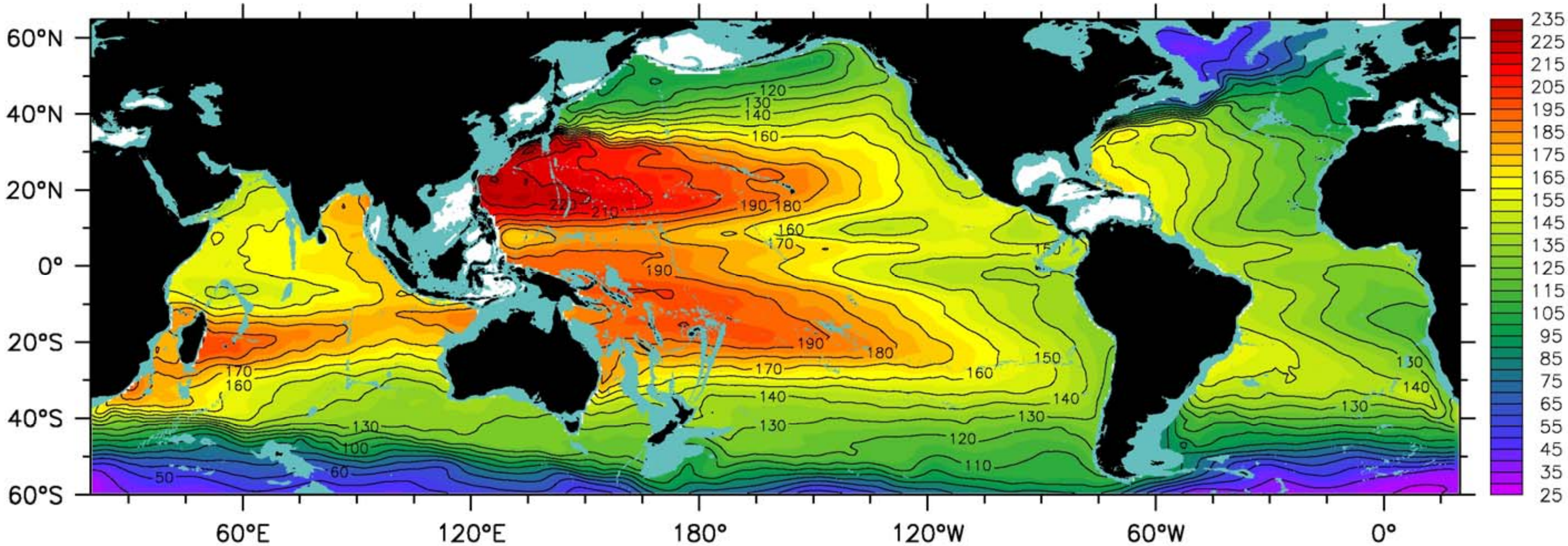
The equation of state for seawater is nonlinear. This has major implications to ocean mixing.



Ocean circulation: surface currents….

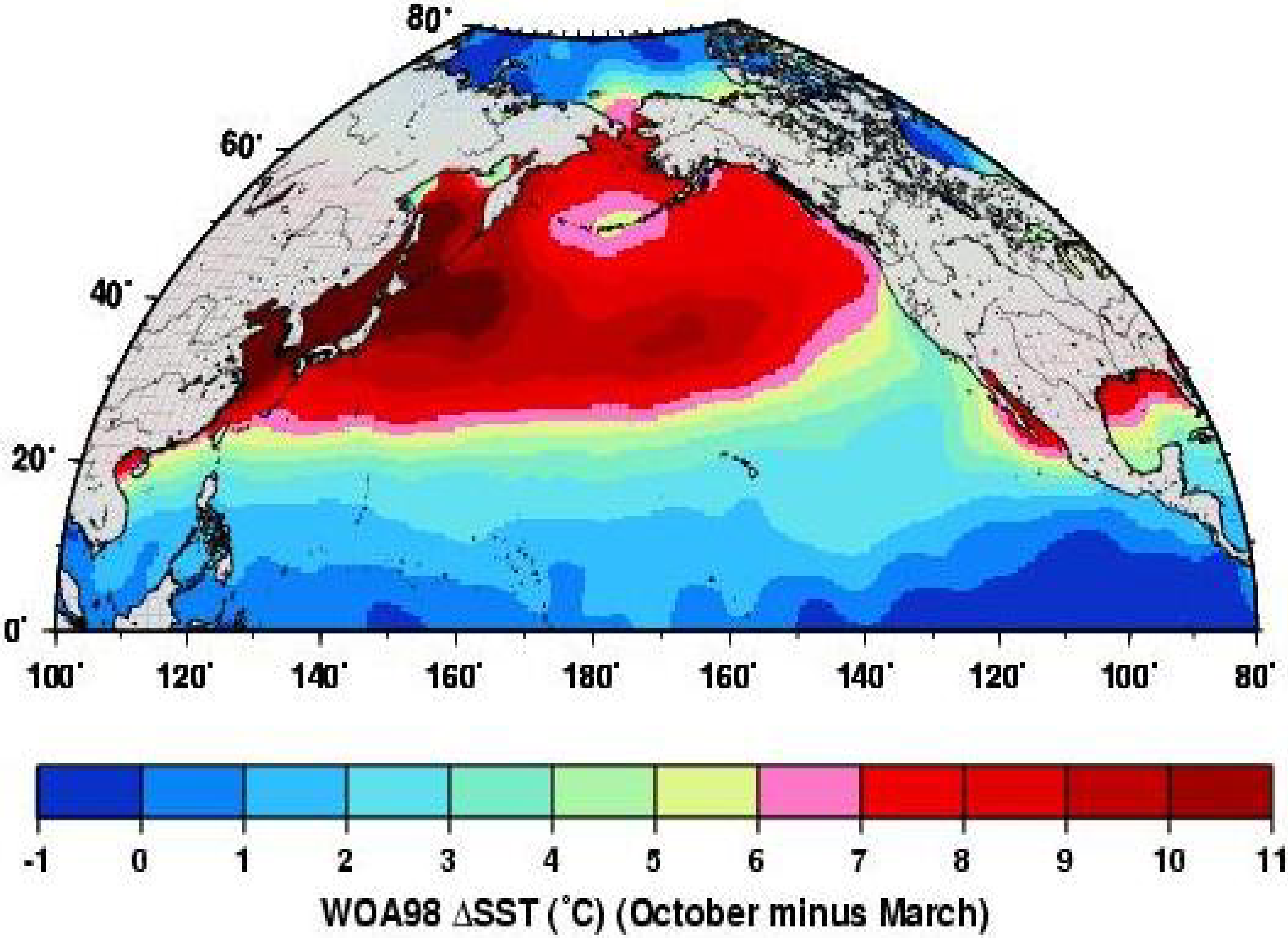


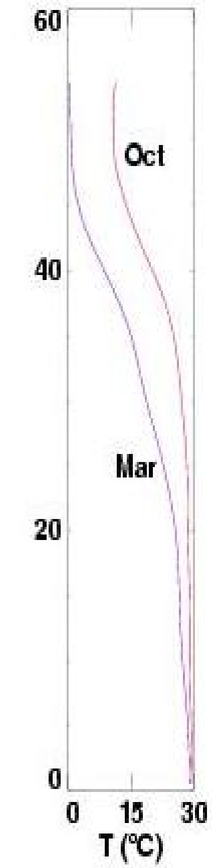
[0-500 dbar dynamic ht; maximum range ~ 2 m] **[notice E/W asymmetry]**



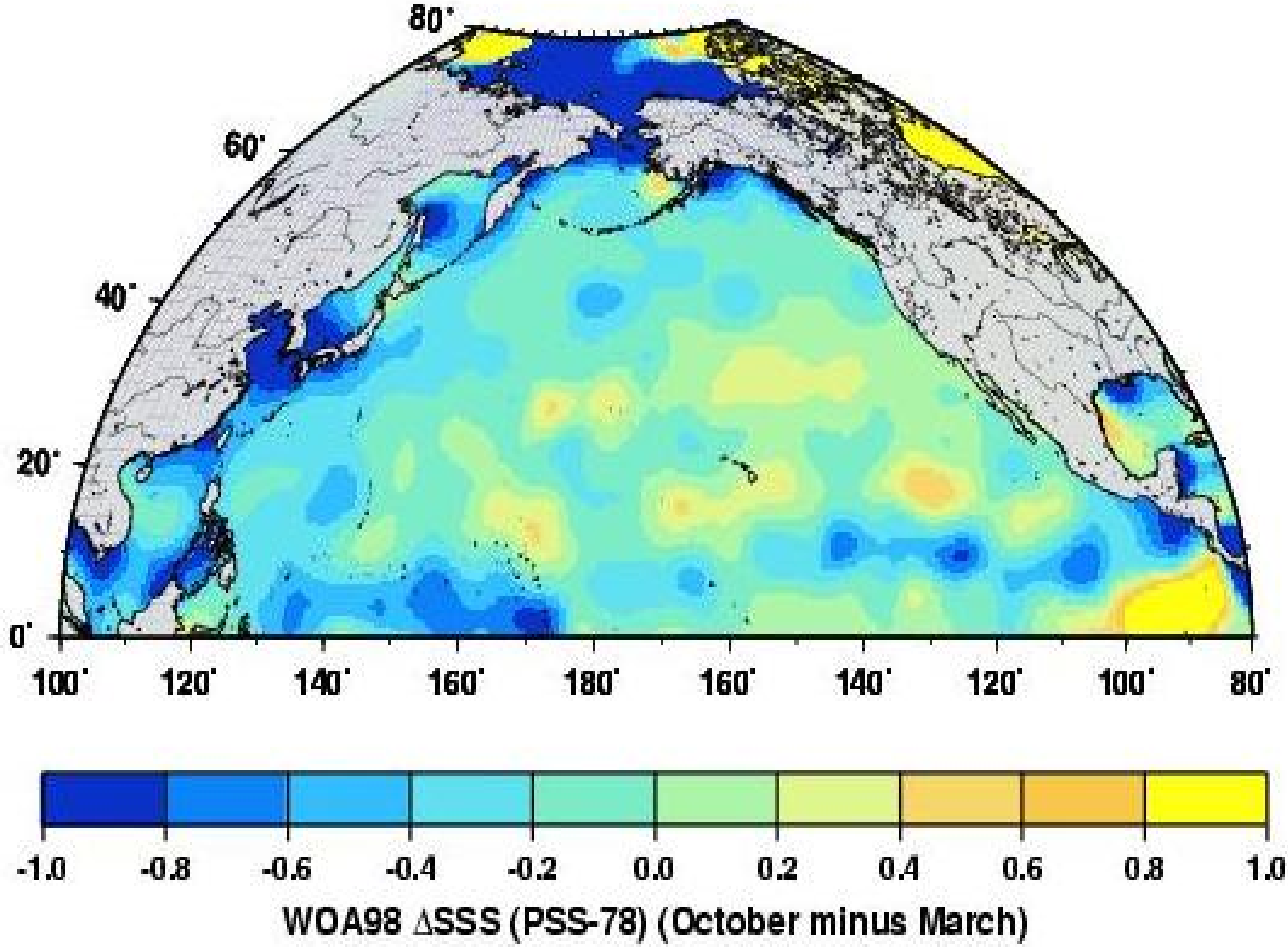
The pressure field of the upper ocean relative to a depth of 1000 dbar (~1000 m), from Argo data

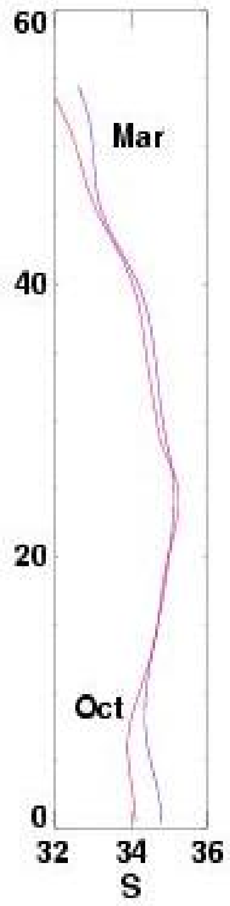
Seasonal variation in SST





Seasonal variation in SSS….





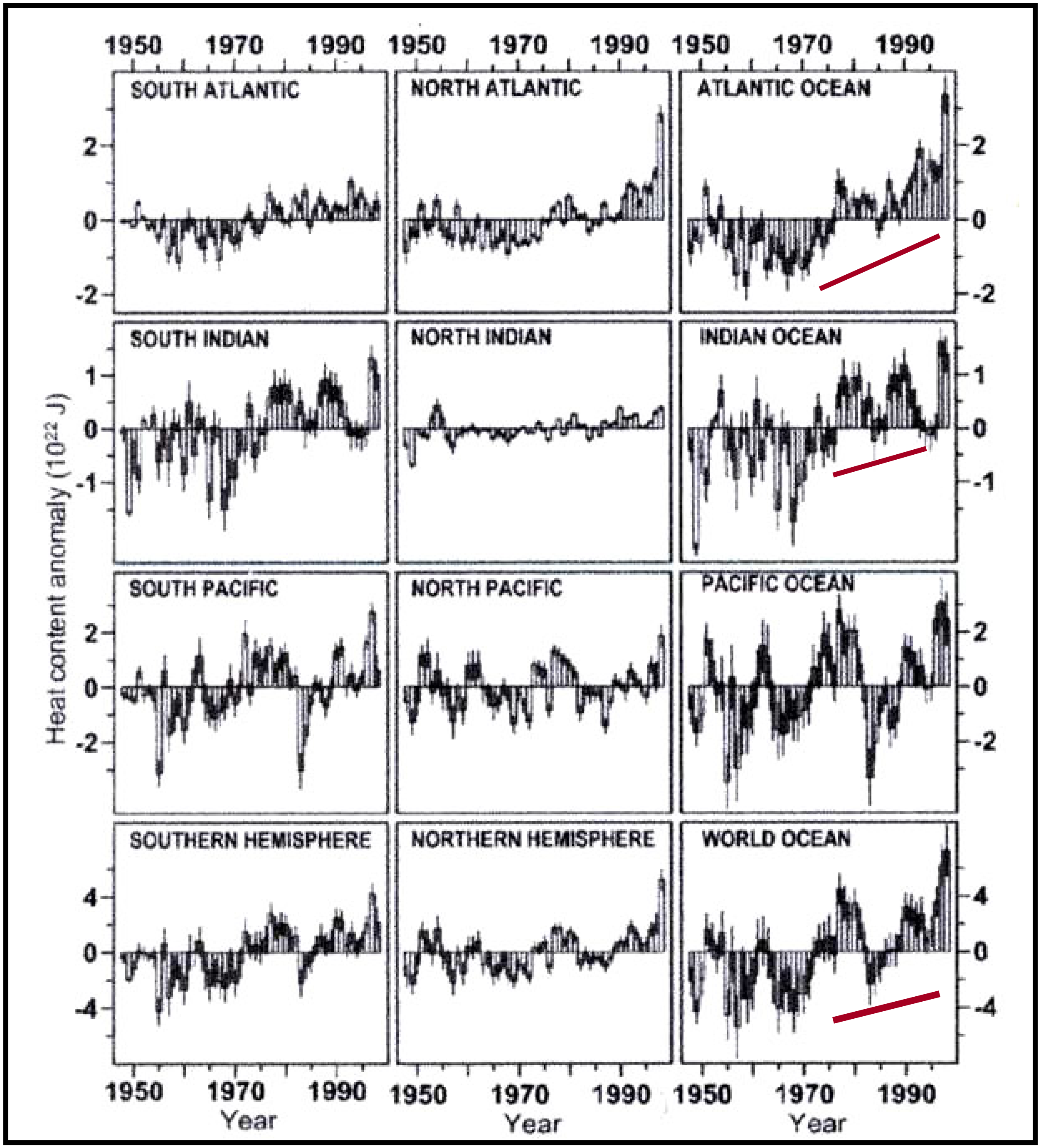
The Heat Balance of the Earth (and the ocean's role)….

Solar heating is ultimately the source of all energy at the Earth's surface, except for tides and geothermal heating.

The atmosphere and ocean exchange heat and energy in a complicated set of feedbacks.

Recall the 1st law of thermodynamics….

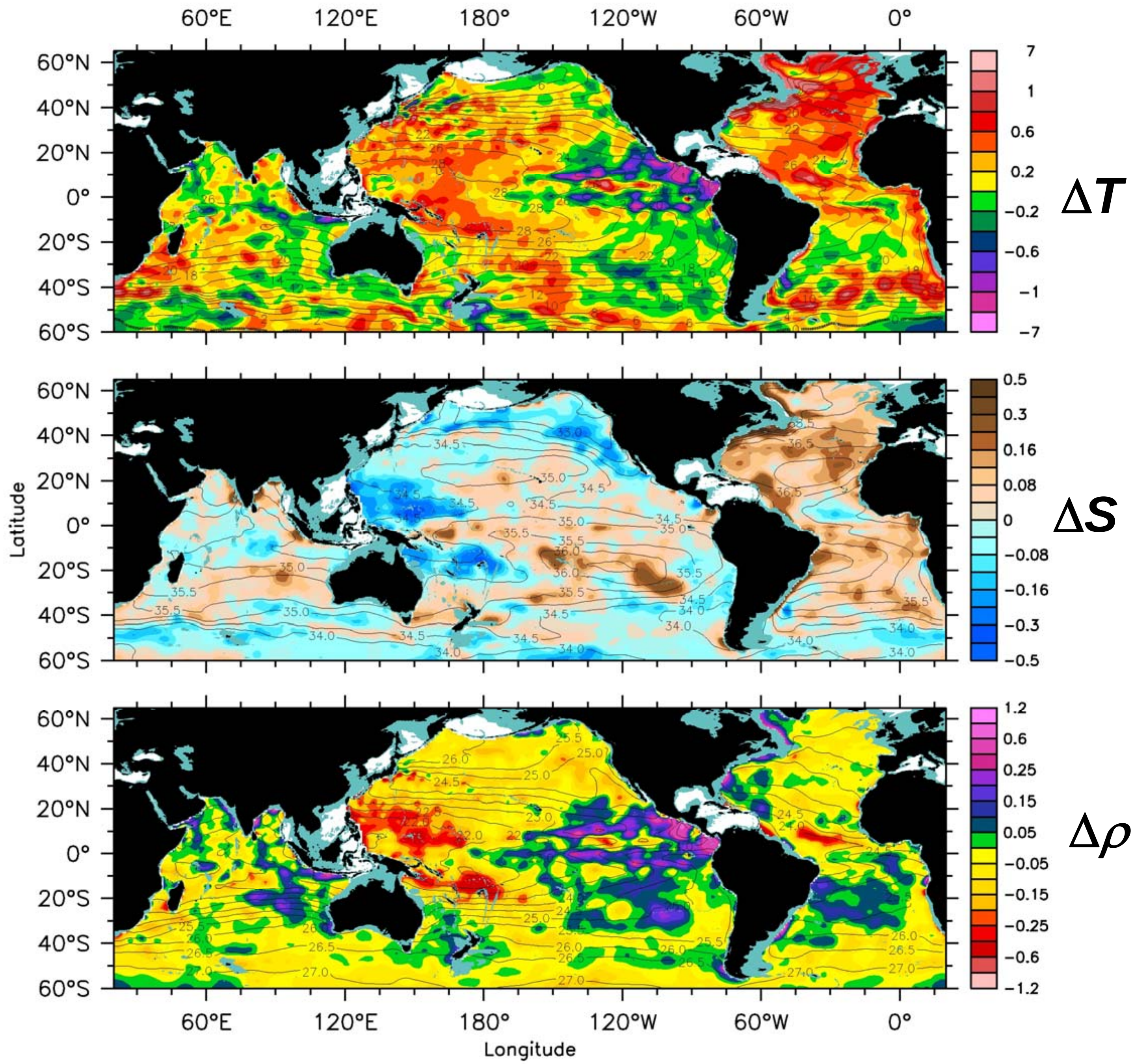
Heat is conserved.



The heat content in the upper 300 m of the ocean has varied over the past 50 years, generally increasing…where has this heat come from?

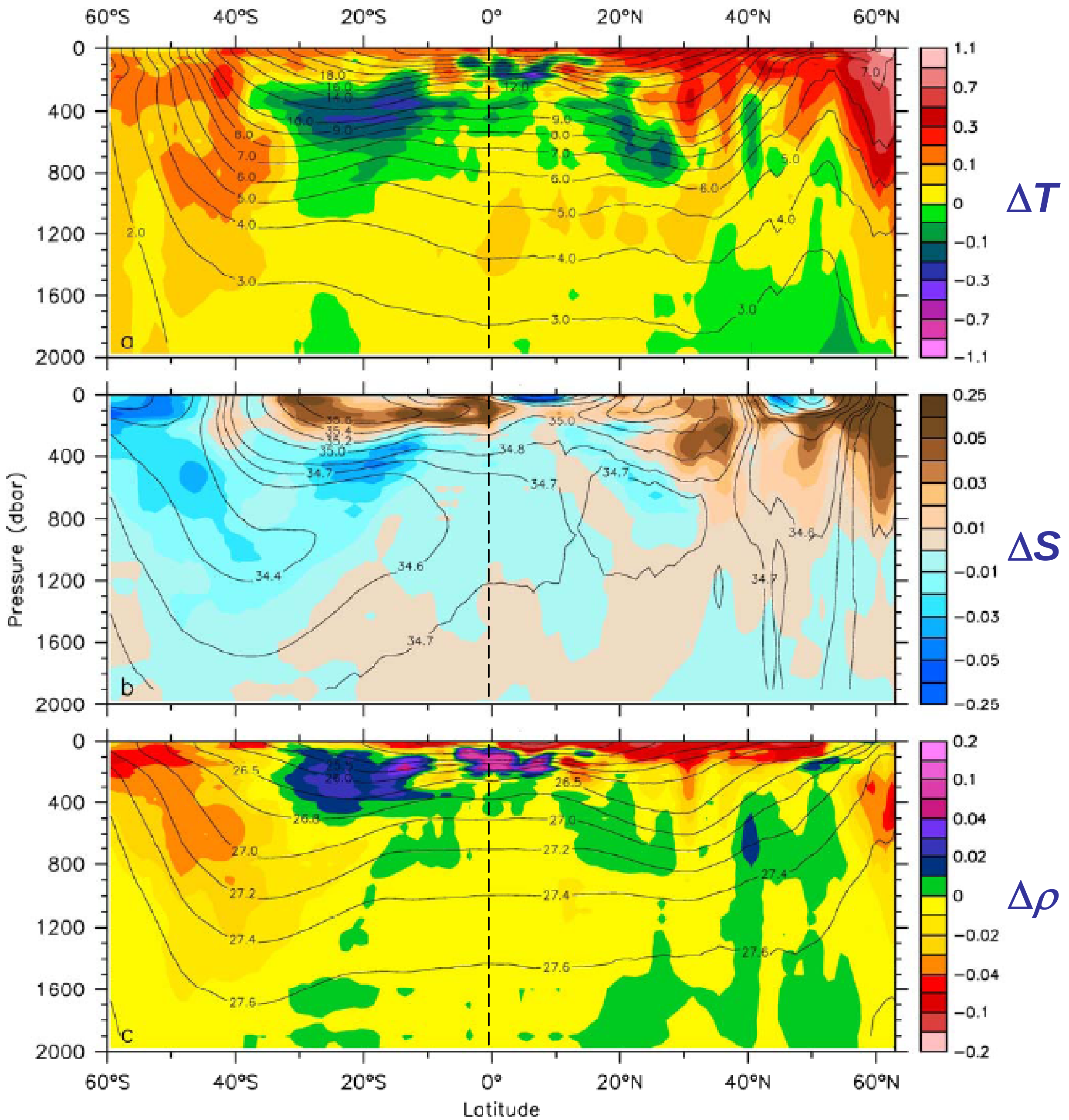
[heating rate ~ 0.6 °C/century]

[from Levitus]

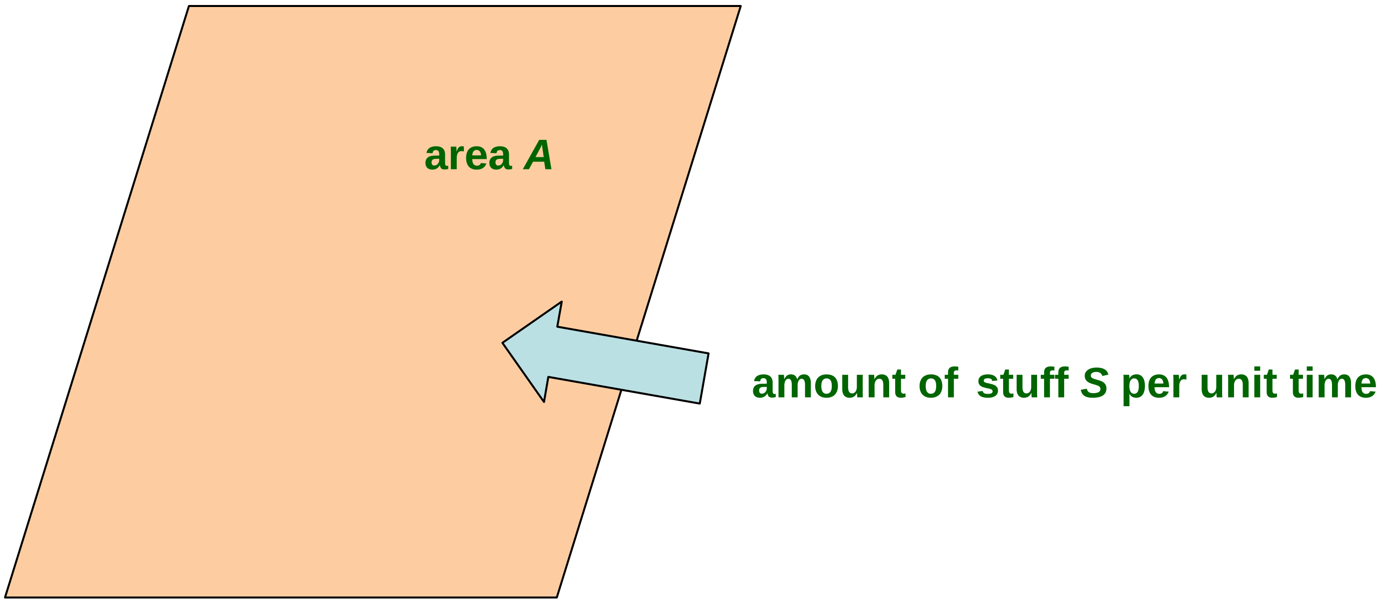


Upper ocean changes in the past 25 years

Average changes in temperature, salinity, and density in the world ocean as a function of latitude since the middle of the last century, determined using 2004-2008 Argo float observations

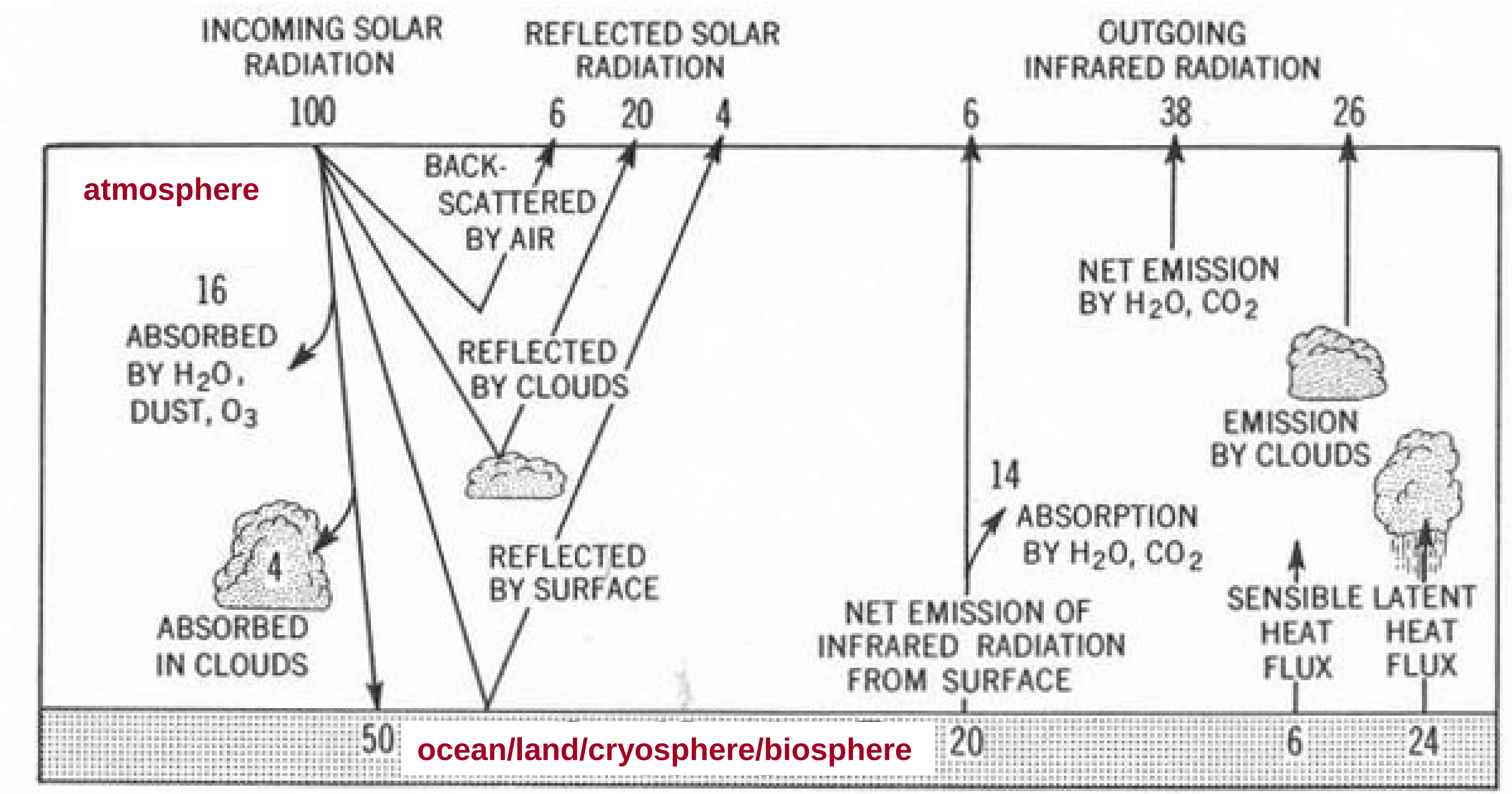


The concept of a flux….



The flux of *S* is defined as the amount of *S* crossing the area *A* per unit time; a flux is just (amount)/(area $\times$ time).

The global heat balance (schematic)….



The heat balance of the ocean….

$$Q_T = Q_S + Q_B + Q_H + Q_E + Q_V + Q_G$$

where

Q_T = net heat input to the Earth (~ 0 watts/m^2)

Q_S = direct solar input (~ +150 watts/m^2)

Q_B = black body radiation (~ –50 watts/m^2)

Q_H = sensible heat loss (~ –10 watts/m^2)

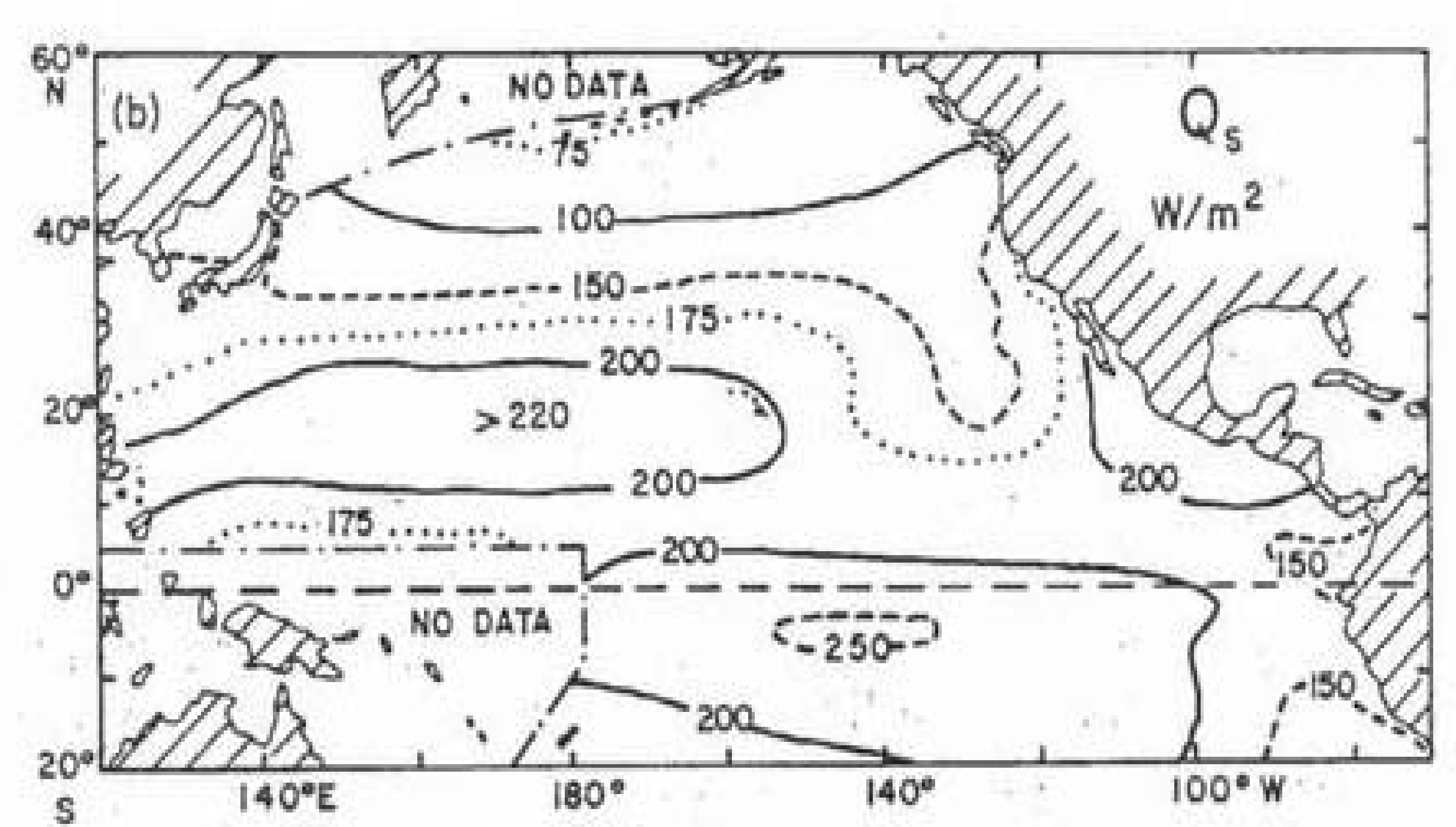
Q_E = evaporative heat loss (~ –90 watts/m^2)

Q_V = advective heat transport (0 watts/m^2)

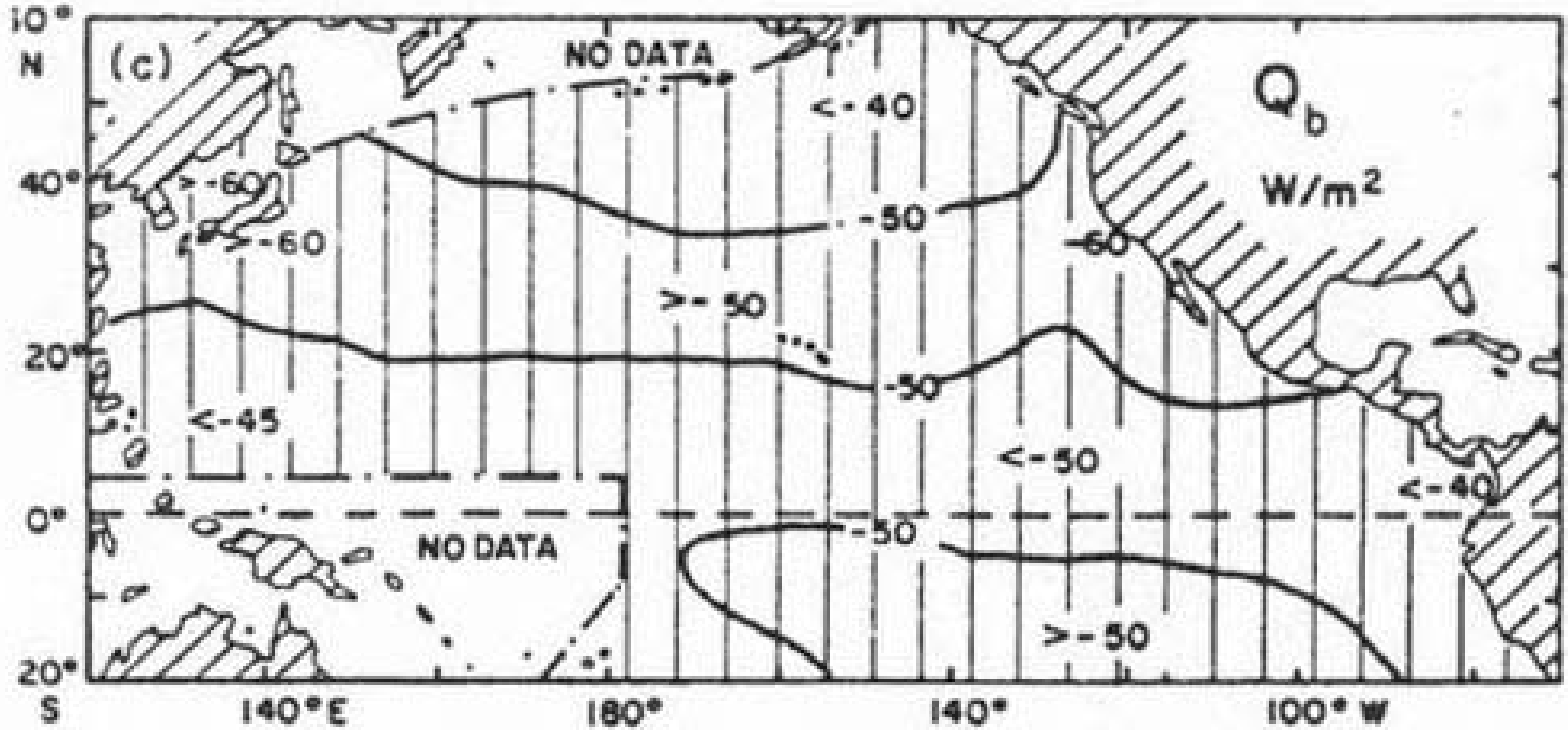
Q_G = geothermal heating (~ +0.01 watts/m^2)

[note: there are errors and uncertainties associated with each of these estimates]

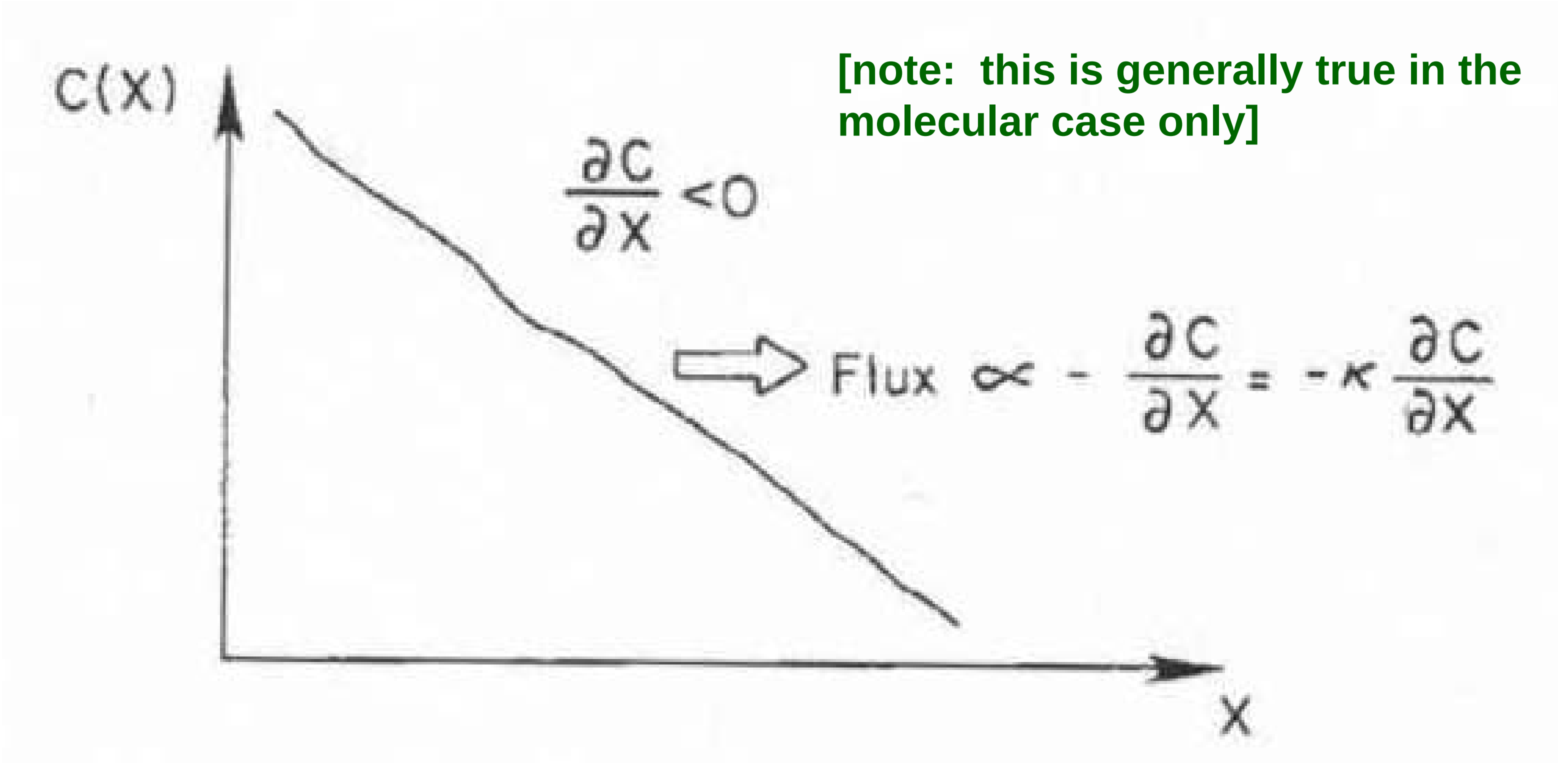
Q_S for the Pacific….



Q_B for the Pacific….



The concept of flux-gradient diffusion….



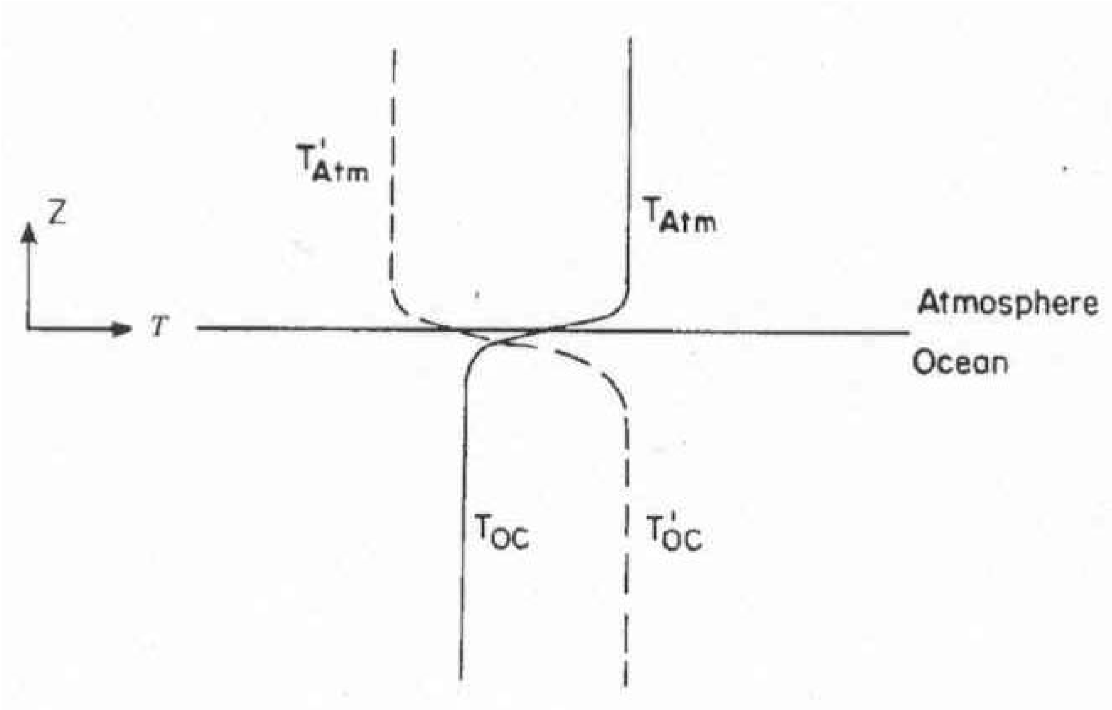
[note: this is generally true in the molecular case only]

The flux is proportional to the gradient of the concentration, with the constant of proportionality being the *diffusivity* κ.

Q_H , conductive (or sensible) heat flux….

Since the ocean is generally at a different temperature than the atmosphere, there will be a conduction of heat between them (recall the two blocks).

Fourier's Law of heat conduction:



$\boldsymbol{F \propto -\nabla T}$ **or**

$$\boldsymbol{F = -K\,\nabla T}$$

[K = diffusivity]

So, $Q_H = c\,\partial T/\partial z$,

$c = \rho C_p K$

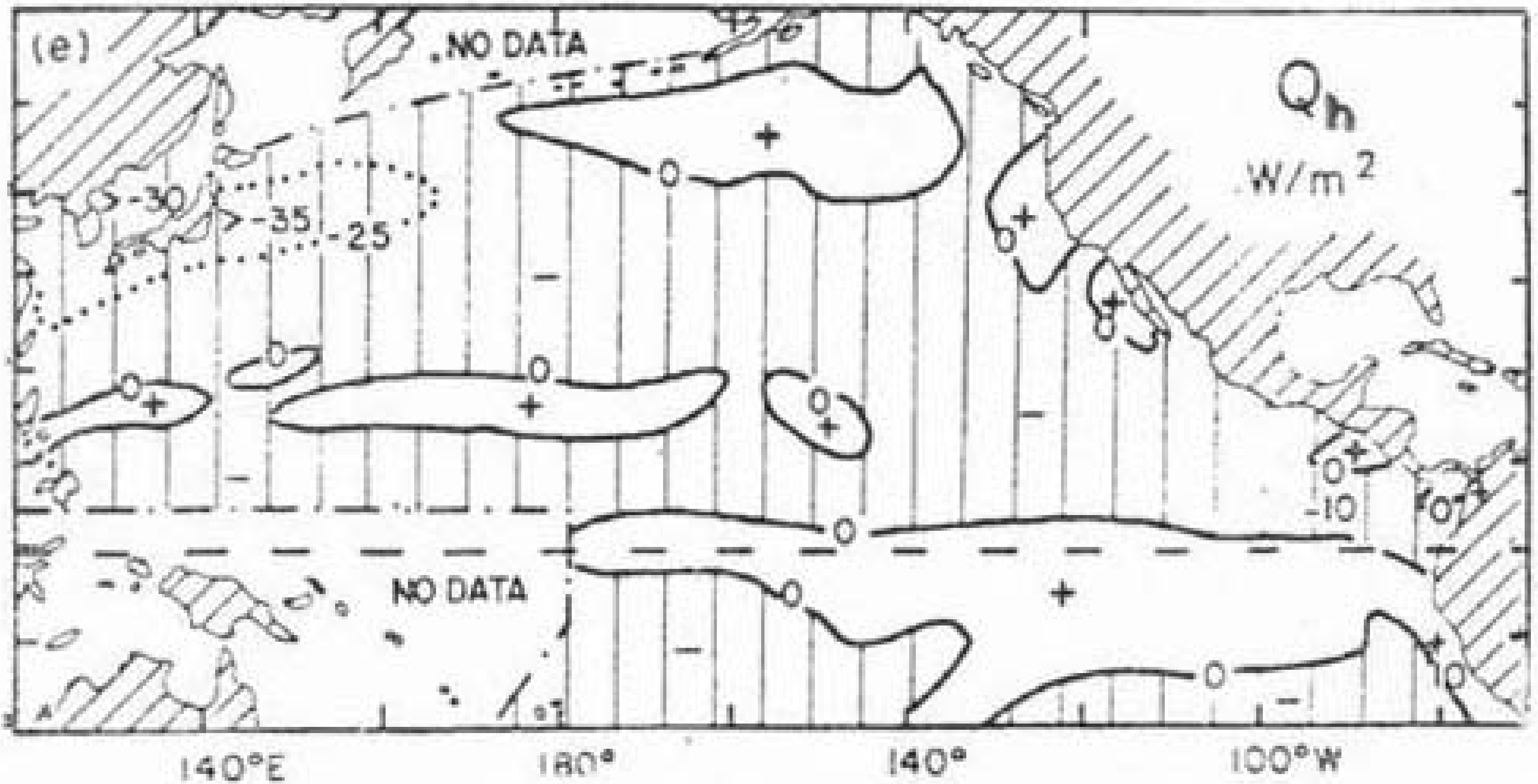
[this is an empirical result]

Q_H, continued….

Note 1: $K = K$ (T, wind, sea state, humidity…..)
$\sim 10 - 10^3$ cm^2/sec at the sea surface

Note 2: Strictly, the concept of a diffusivity *K* only applies to laminar flows. Where turbulence occurs (everywhere !), this formulation serves *as a parameterization only* ("eddy diffusivity").

Q_H for the Pacific….



[note the east-west asymmetry]

Evaporative heat flux, Q_E….

The latent heat of evaporation L for fresh water is

L = 2494 joules/kg @ 2 °C

This much energy must be supplied in order for the state transition from liquid to vapor to occur. In the oceanic case, the energy will be extracted from surface seawater in the form of heat, thus having a net cooling effect.

This evaporated seawater will appear as water vapor in the atmosphere, and the latent heat L will be released into the atmosphere upon condensation of the water vapor.

Q_E, continued….

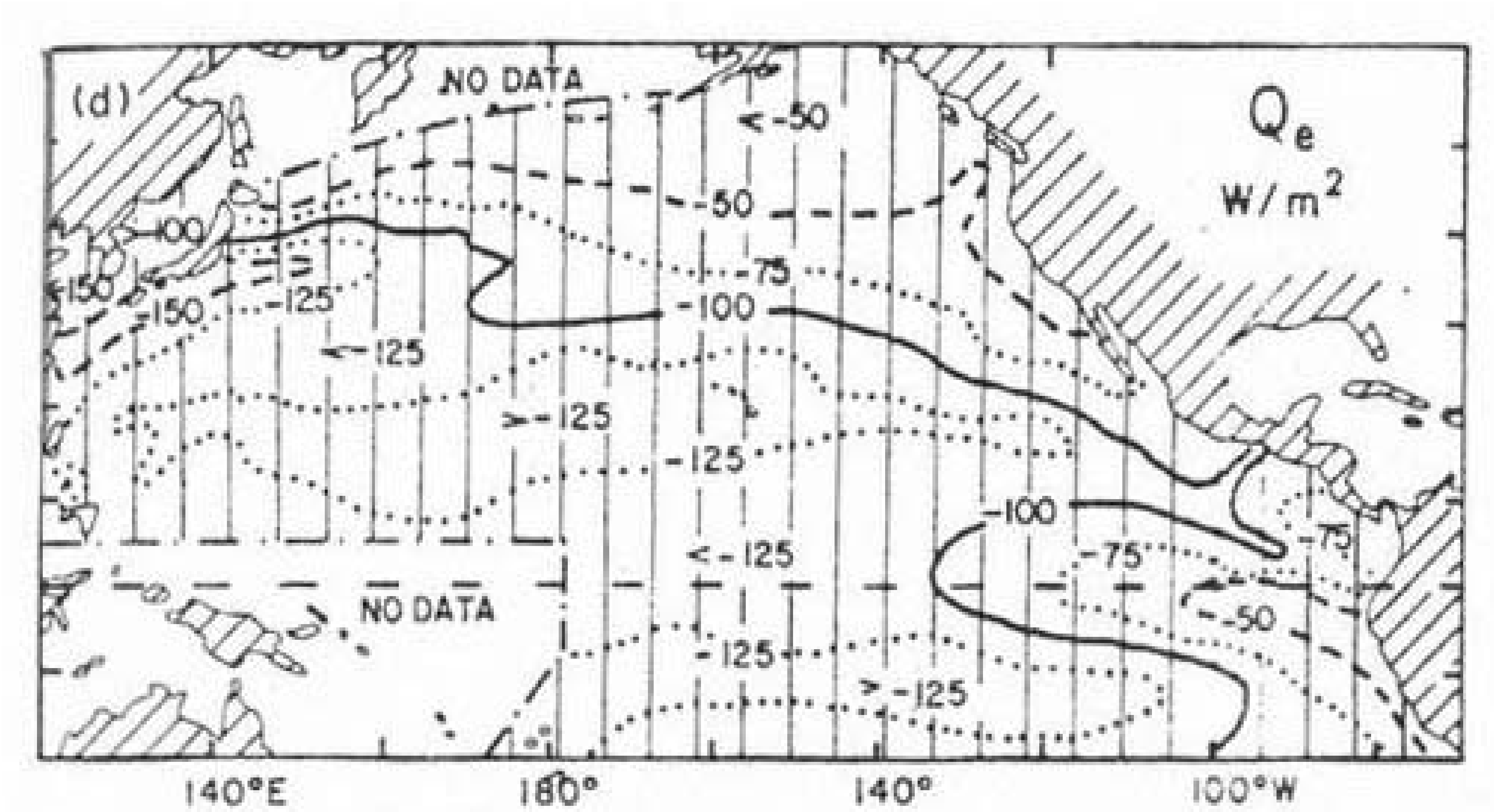
Note: Q_E is generally the largest term in the heat balance equation, after direct solar input Q_S.

How can Q_E be estimated? In general the evaporation rate and associated heat flux are a complicated function of a number of variables and can only be estimated empirically,

$$Q_E = Q_E(T_{oc}, T_{atm}, S, \text{wind, sea state, humidity,….})$$

Many estimates of the evaporative heat flux exist, using a variety of parameterizations. Estimated errors are not small, in general.

Q_E for the Pacific….



[note the east-west asymmetry]

The heat balance….

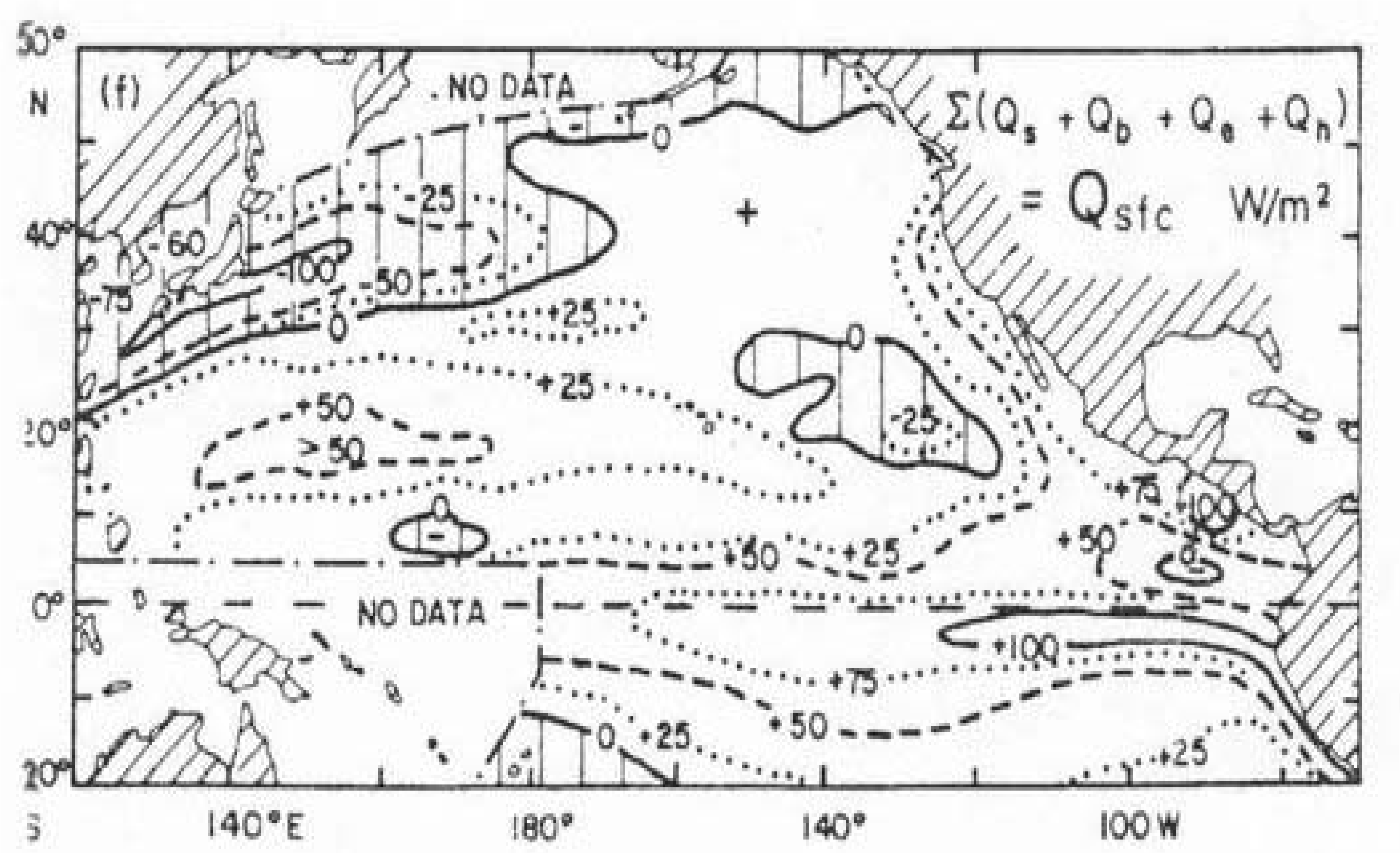
$$Q_T = Q_S + Q_B + Q_H + Q_E + Q_V + Q_G = 0$$

$$= Q_{surf} + Q_V + Q_G \approx Q_{surf} + Q_V = 0$$

where $Q_{surf} = Q_S + Q_B + Q_H + Q_E$

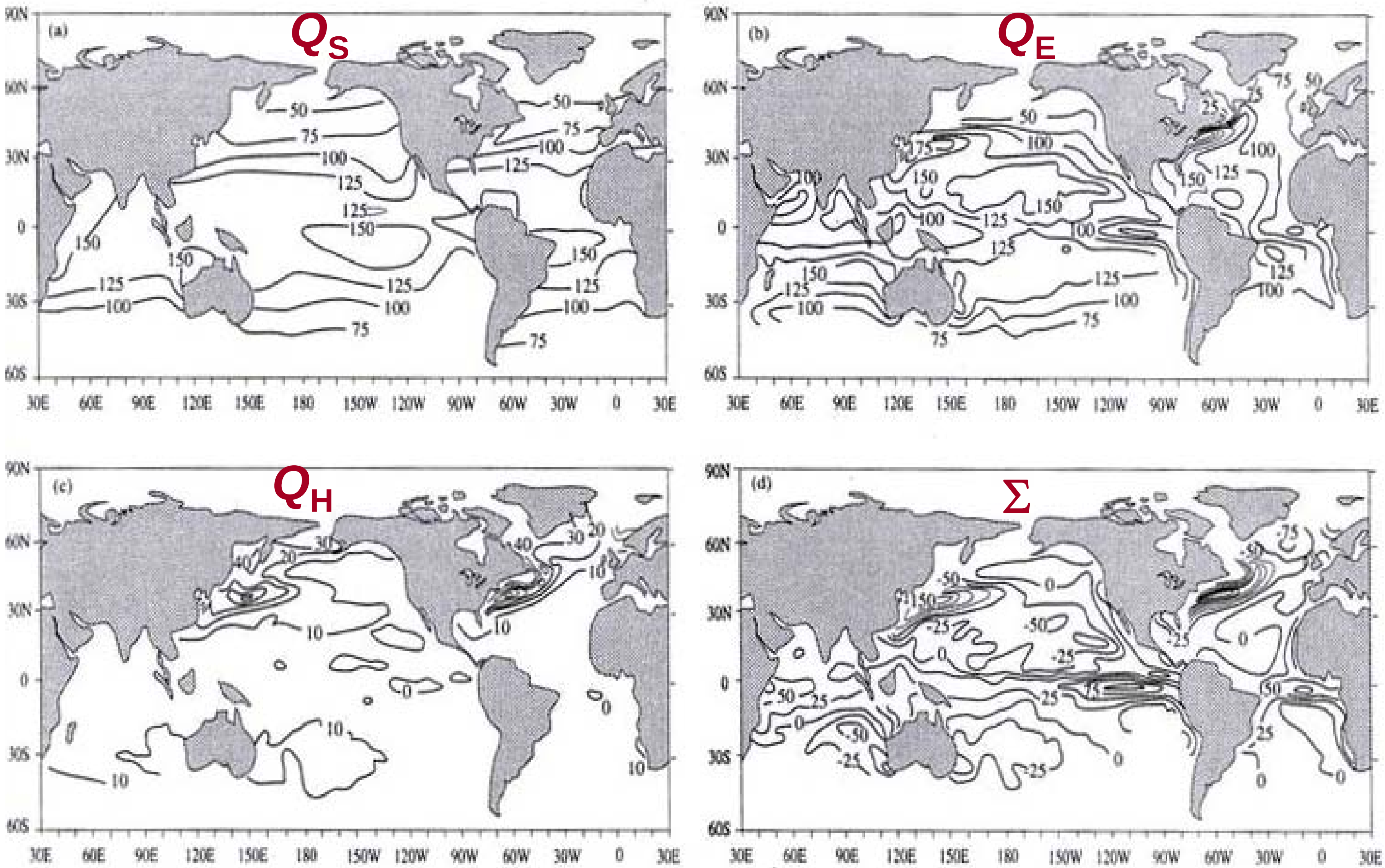
What is the total surface heat flux from the ocean, Q_{surf}?

Q_{surf} for the Pacific….



[note east-west asymmetry]

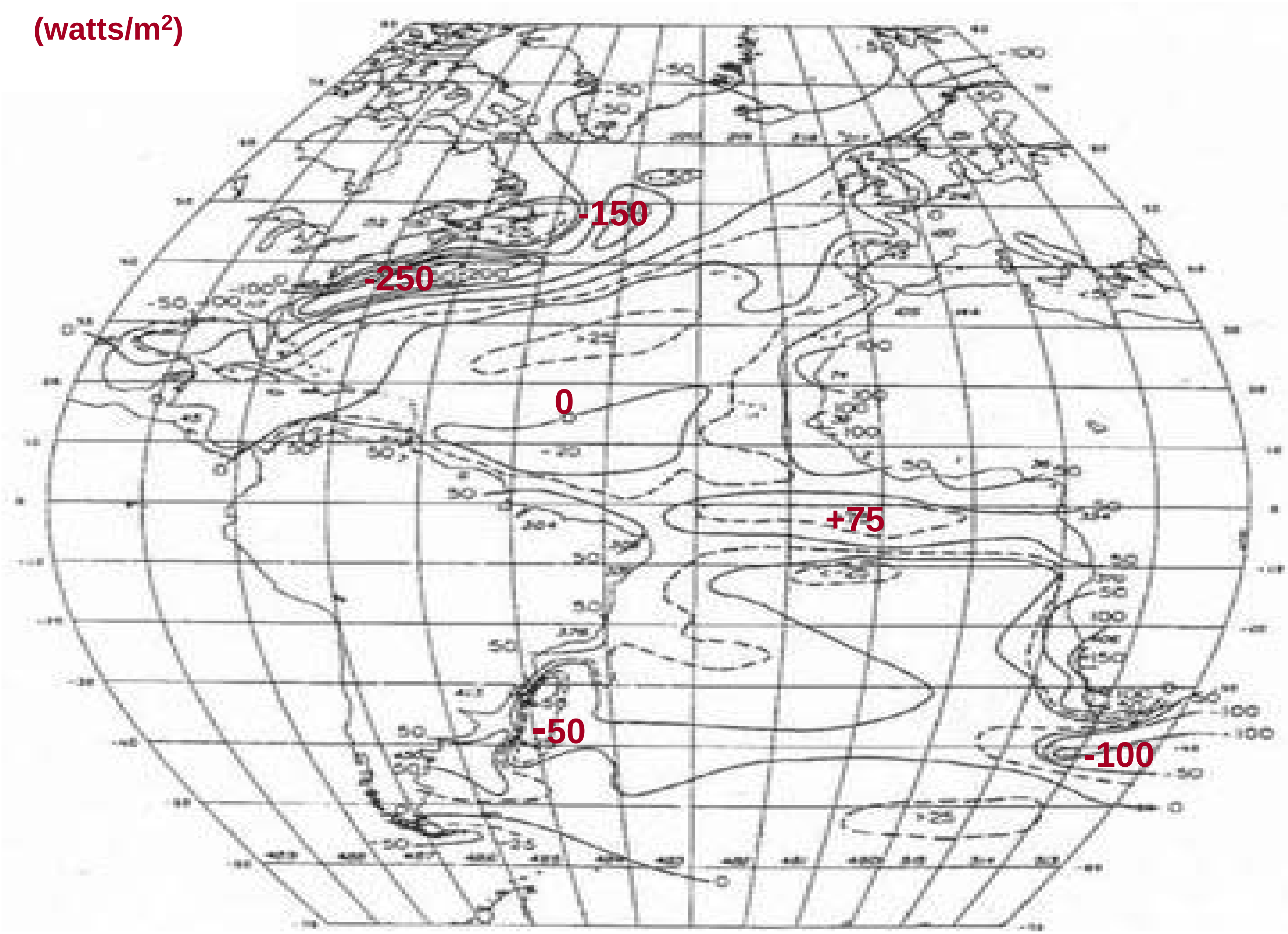
Global heating at the sea surface….



Q_{surf} for the Atlantic….

[note east-west asymmetry]

(watts/m^2)



Q_V, advective heat flux….

$$Q_{surf} + Q_V = 0$$

[Globally, $Q_V = 0$, but this might not be true locally.]

If there is a gain or loss of heat from the sea surface, then the ocean must export or import heat in order to keep the system in thermodynamical equilibrium.

Result: Ocean Circulation.

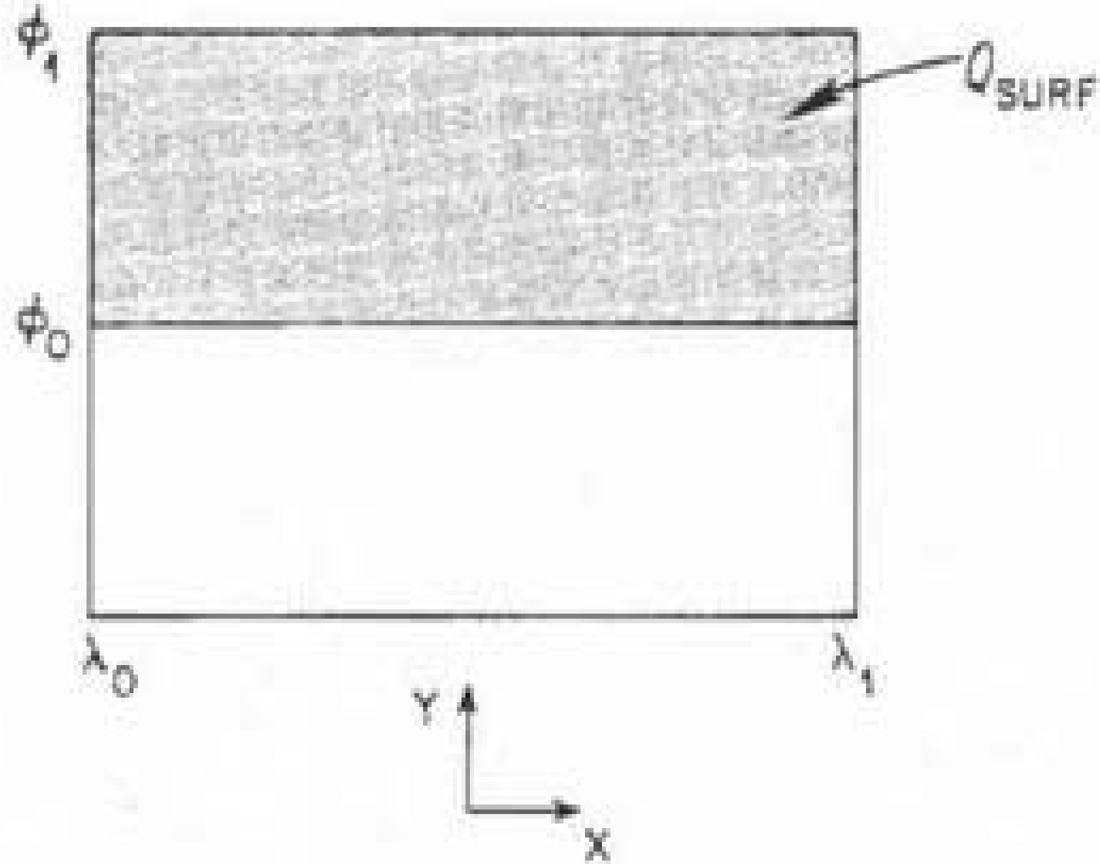
Thus,

$$\iint_{\text{ocean}} Q_{surf}\, dA = -\iint_{\text{ocean}} Q_v\, dA$$

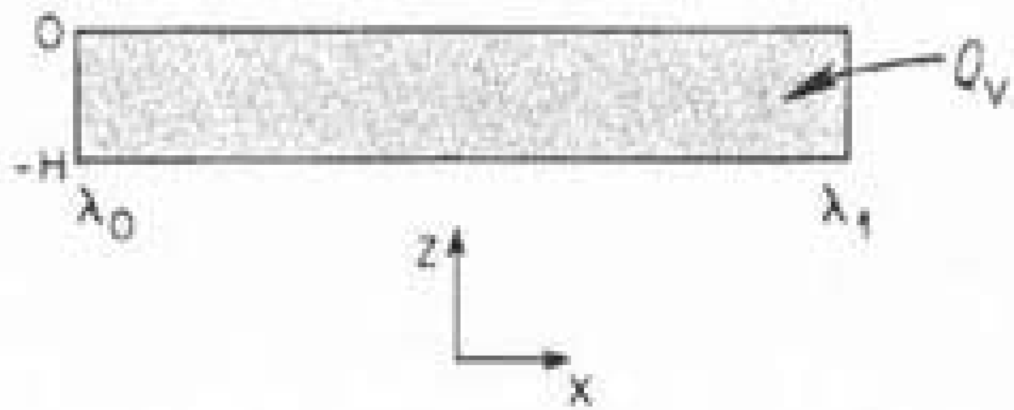
[The integration proceeds over the surface area of a sector of ocean, including the surface, sidewalls, and the bottom.]

Q_V, continued….

an ocean losing/gaining heat at the sea surface



an ocean losing/gaining heat via advection through the sides



Q_v, continued….

$$\iint_{\text{ocean}} Q_{surf}\, dA = -\iint_{\text{ocean}} Q_v\, dA$$

(1) **(2)**

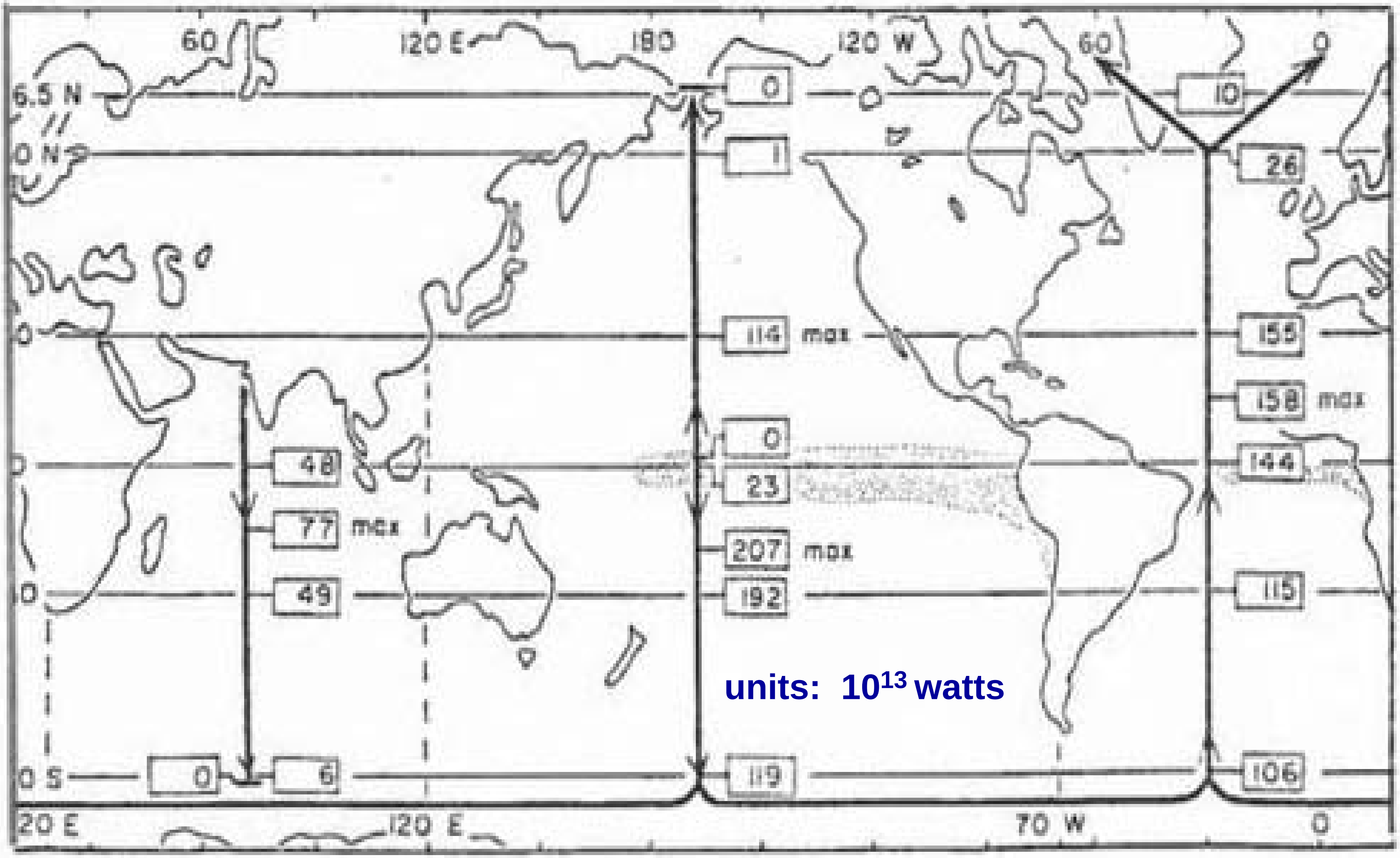
$+ \rightarrow$ **flux in,**
$- \rightarrow$ **flux out**

If (1) > 0 (ocean gains heat from the atmosphere), then (2) < 0 (ocean circulation must transport heat away)

If (1) < 0 (ocean loses heat to the atmosphere), then (2) > 0 (ocean circulation must transport heat in)

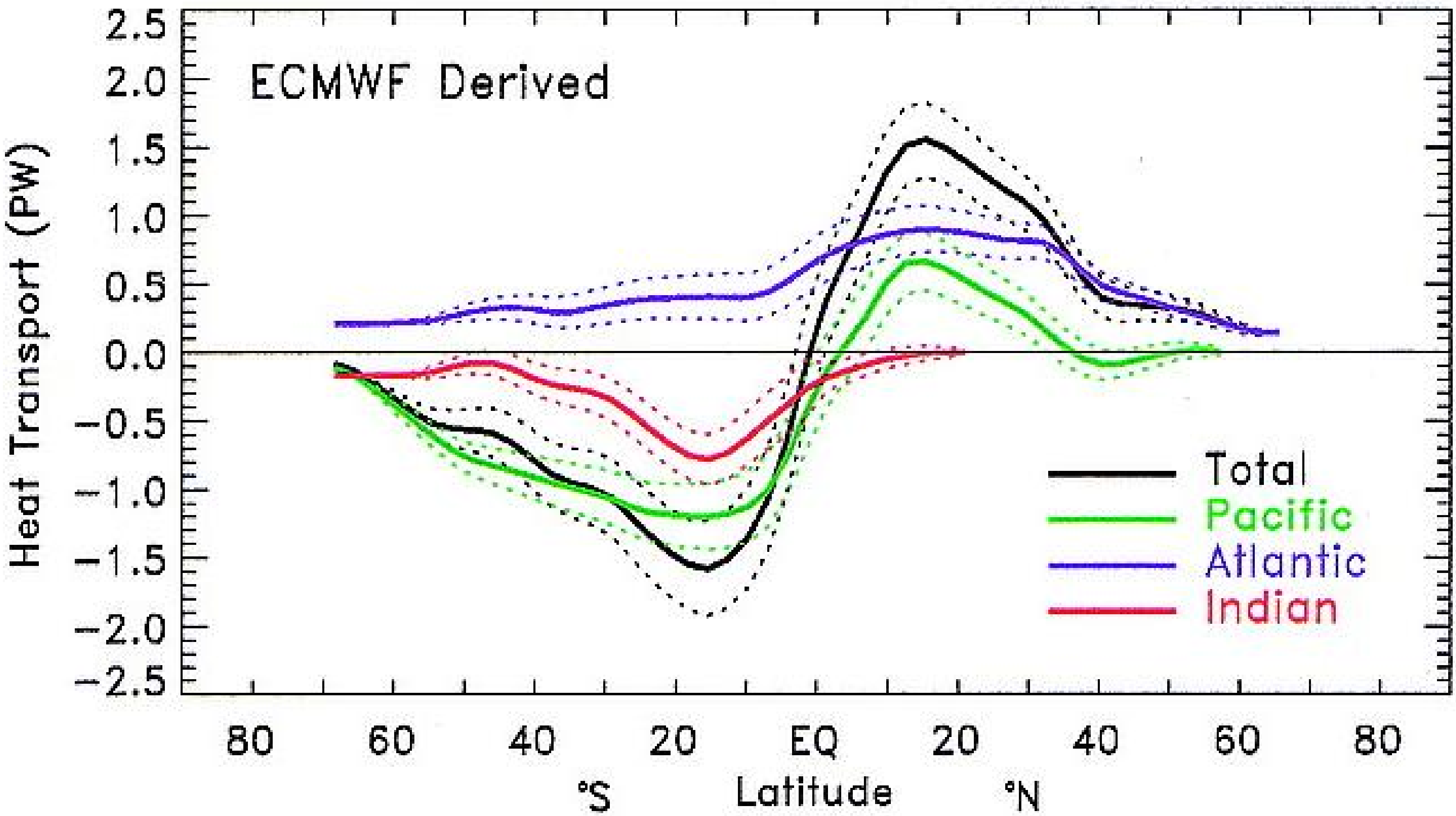
Globally, the surface heat budget for the ocean has been estimated from bulk formulae, marine observations, satellites, etc. If we have an estimate of Q_{surf}, we should be able to estimate Q_v.

Q_V continued….



Global ocean heat transport, basin integrated

Q_V, averaged along latitude by ocean….



[note: Atlantic is different]

Q_V (continued)….

Parameterization of Q_v :

Let *V* equal a velocity normal to some surface, and let *T* be the temperature of the water flowing with *V*. The heat transport associated with *V* and *T* is then

$$Q_V = \rho CVT$$

where *C* is the heat capacity and ρ the density.

Check the units:

$$Q_v = \left[\frac{M}{L^3}\right]\left[\frac{\text{heat}}{M\ {}^\circ C}\right]\left[\frac{L}{T}\right]{}^\circ C = \left[\frac{\text{heat}}{L^2 T}\right]$$

[heat/area/time, a flux by definition]

Q_V, continued….

$$Q_V = \rho C V T$$

[note: T is measured relative to some reference temperature]

Note the meaning of this parameterization:

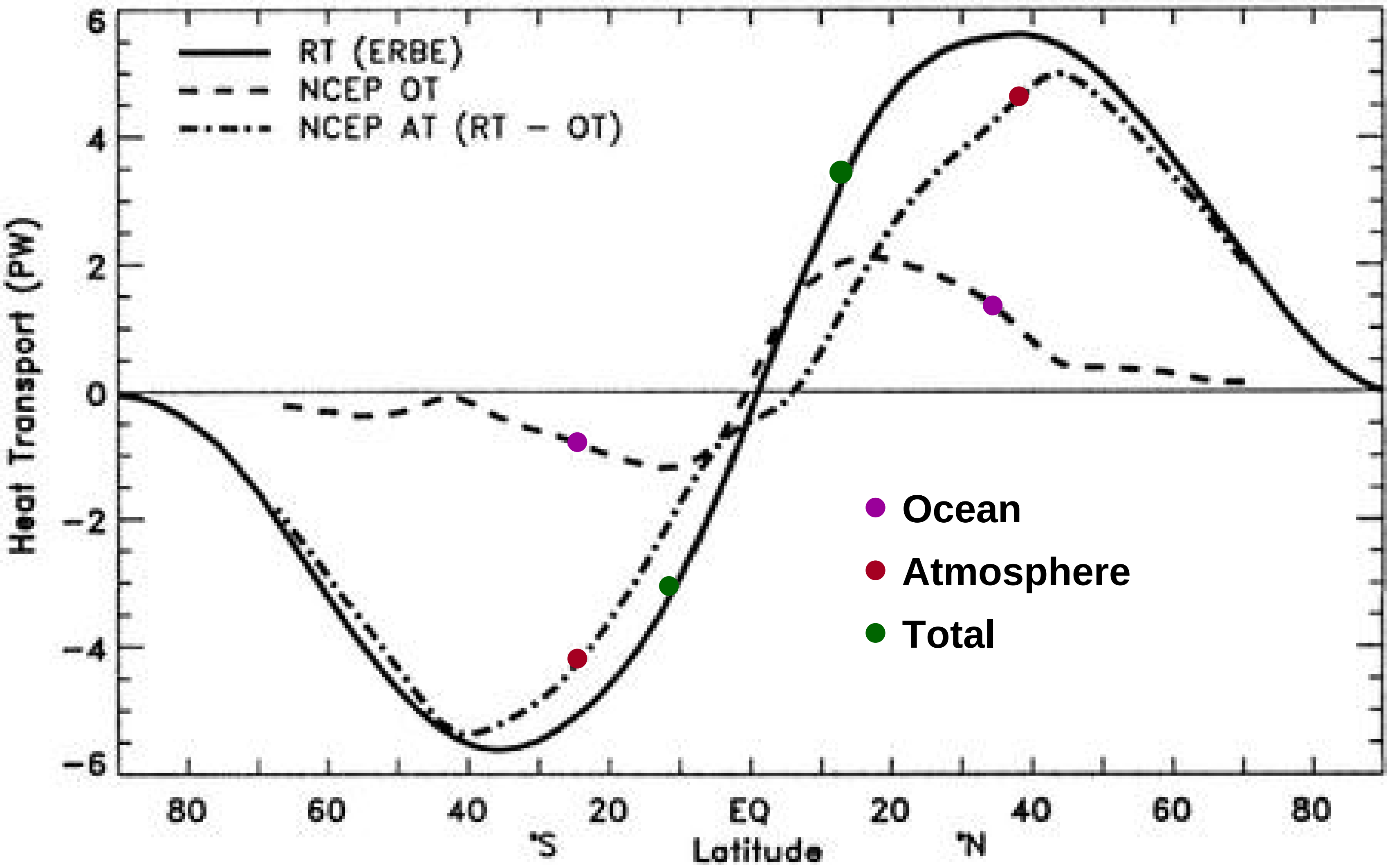
Increasing either the temperature, or the flow, will increase the heat transport.

Also note that

warm water flowing north = cold water flowing south
($T>0$, $V>0$) ($T<0$, $V<0$)

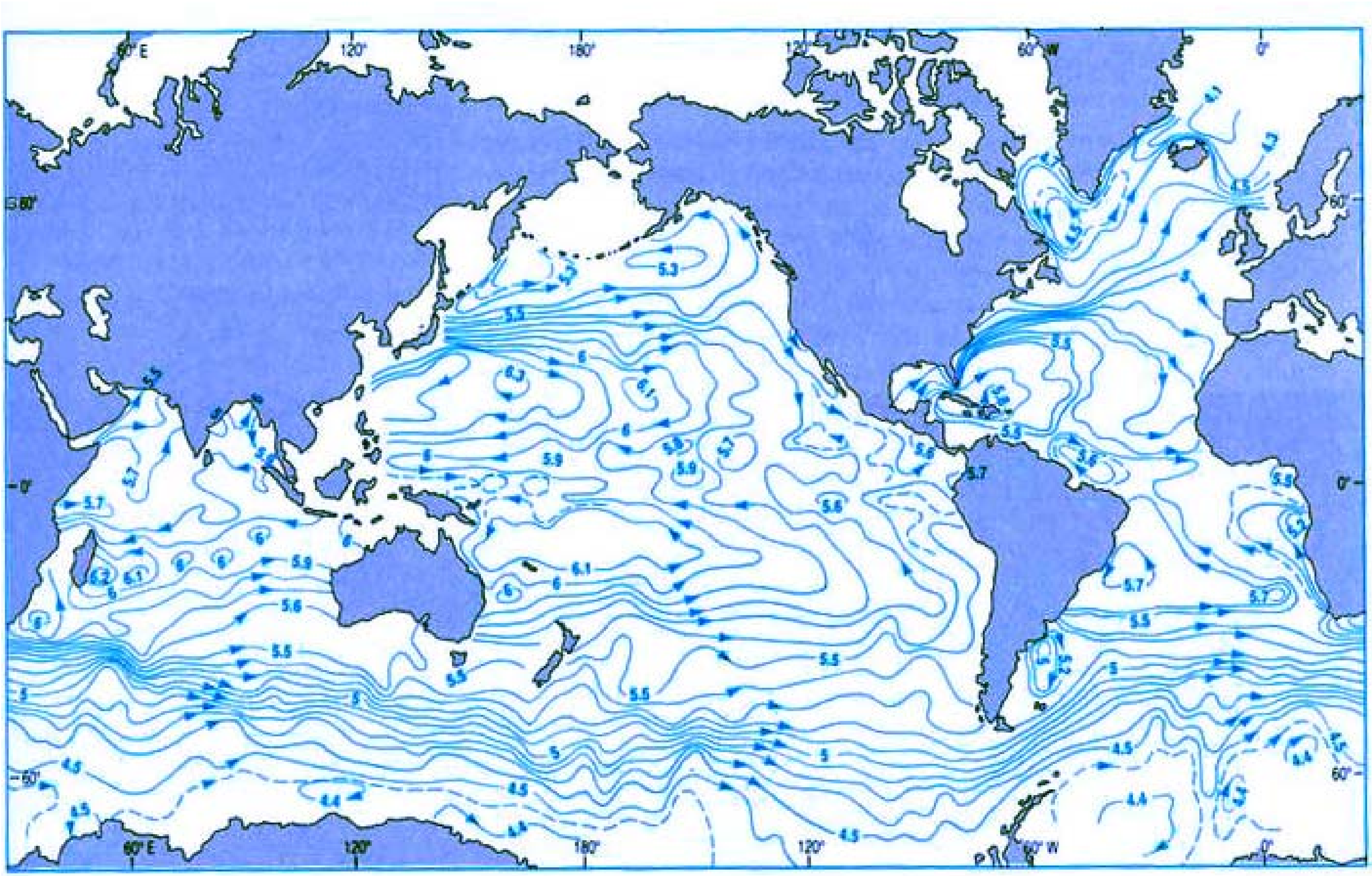
[consider the Atlantic]

Q_V, continued….

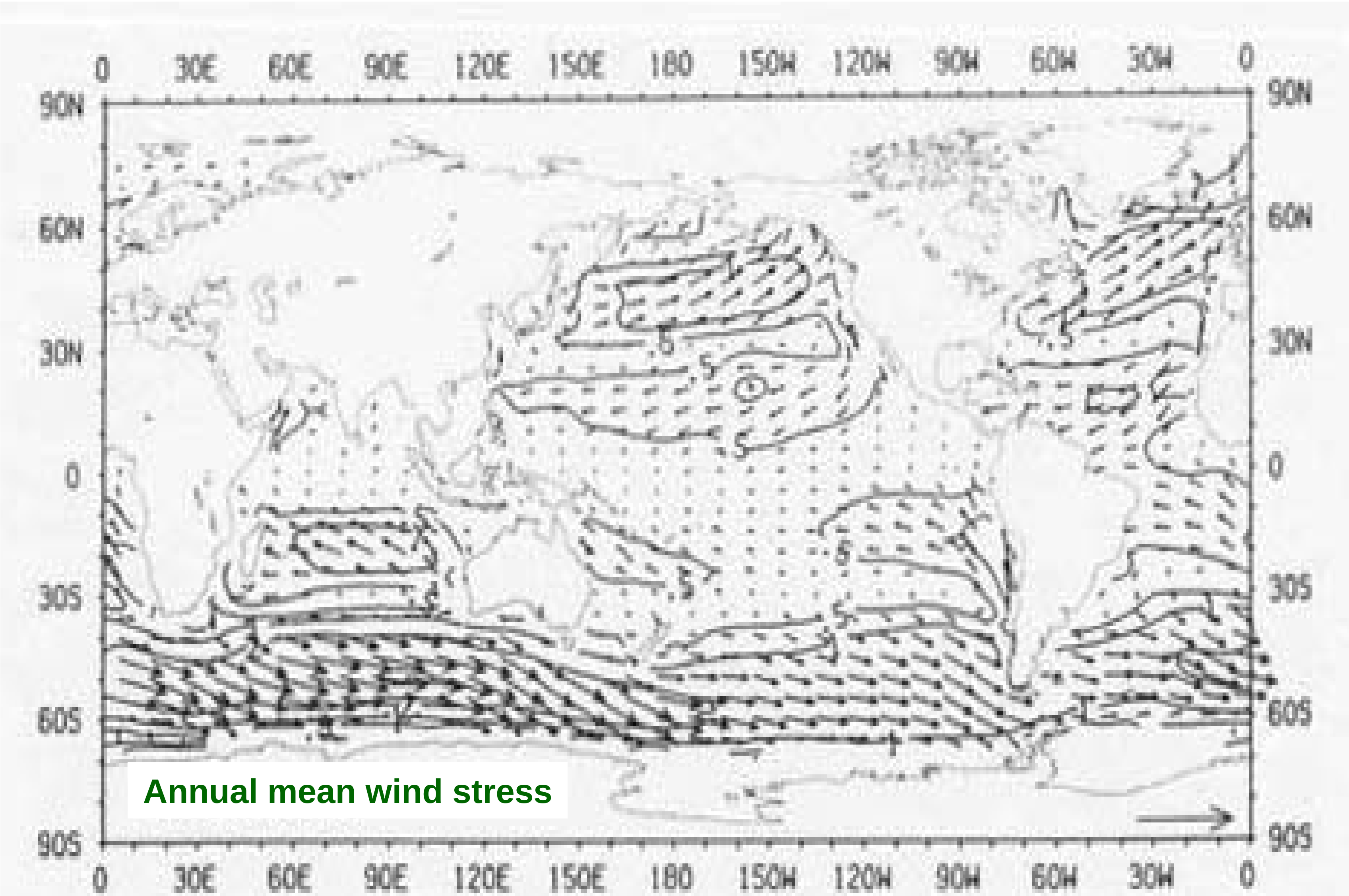


Northward heat transport by the ocean and atmosphere

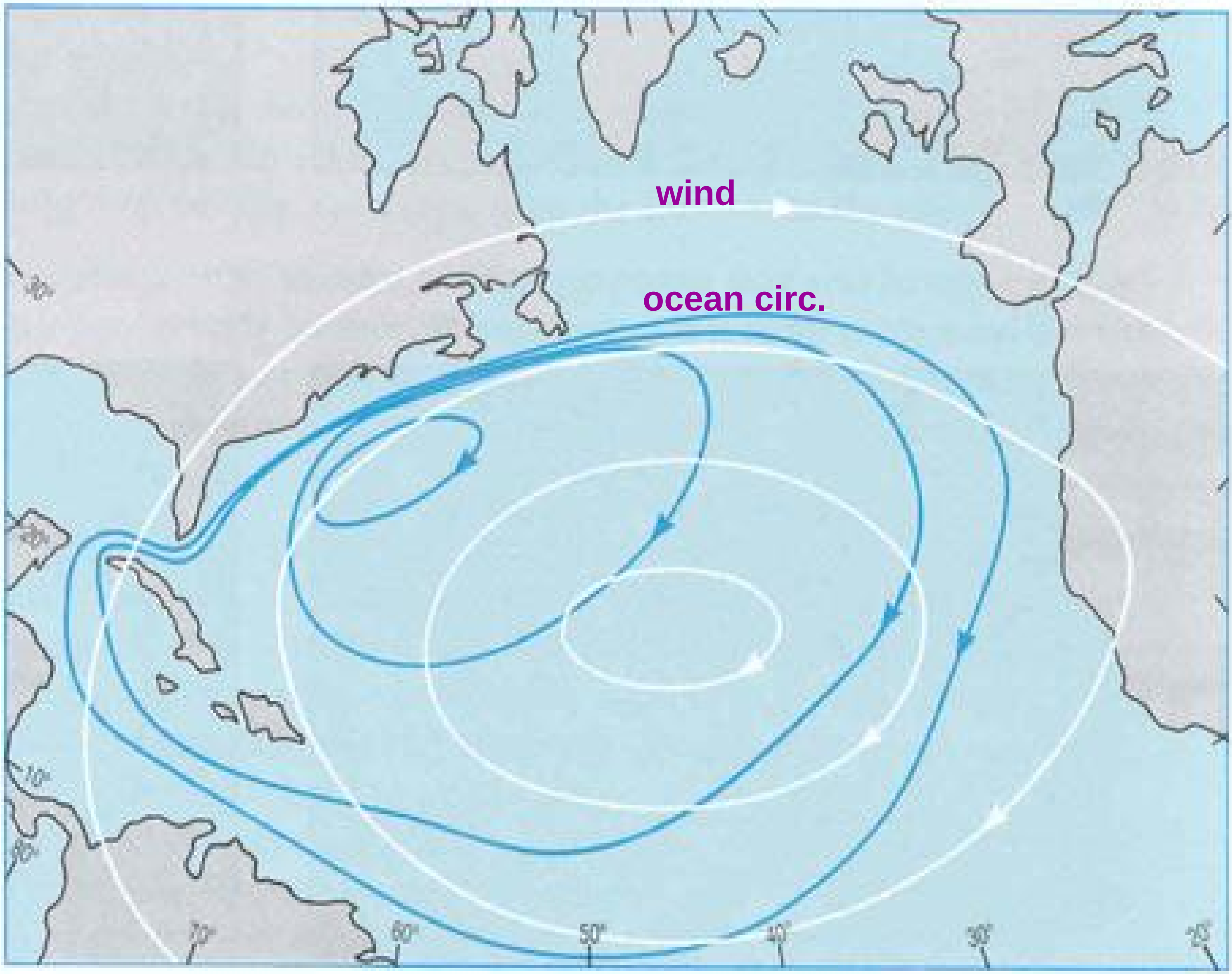
Ocean circulation: surface currents….



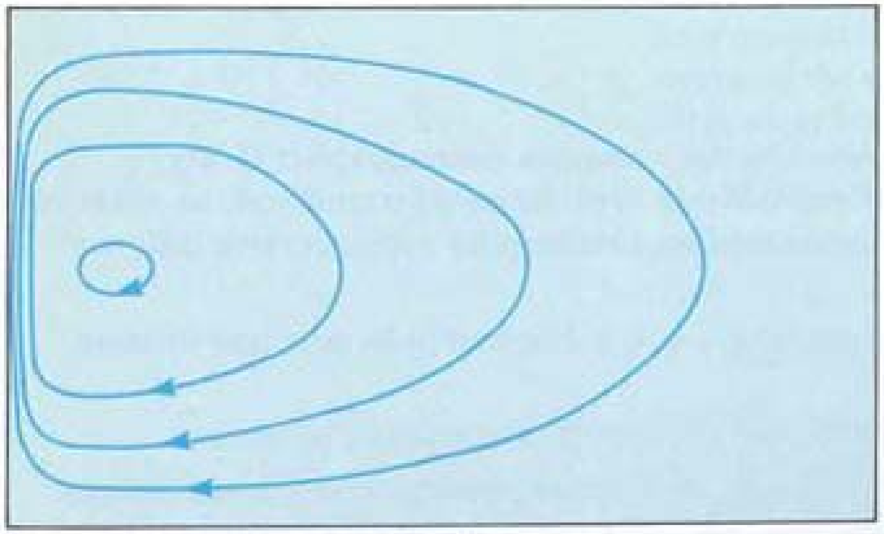
[0-500 dbar dynamic ht; maximum range ~ 2 m] **[notice E/W asymmetry]**



The upper ocean circulation is largely wind-driven.



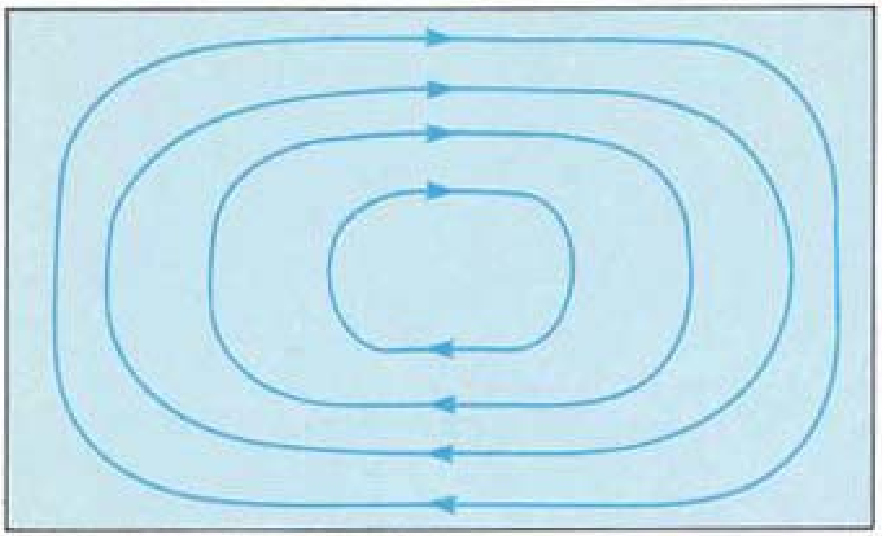
Winds: symmetric; Ocean circulation: asymmetric….why?



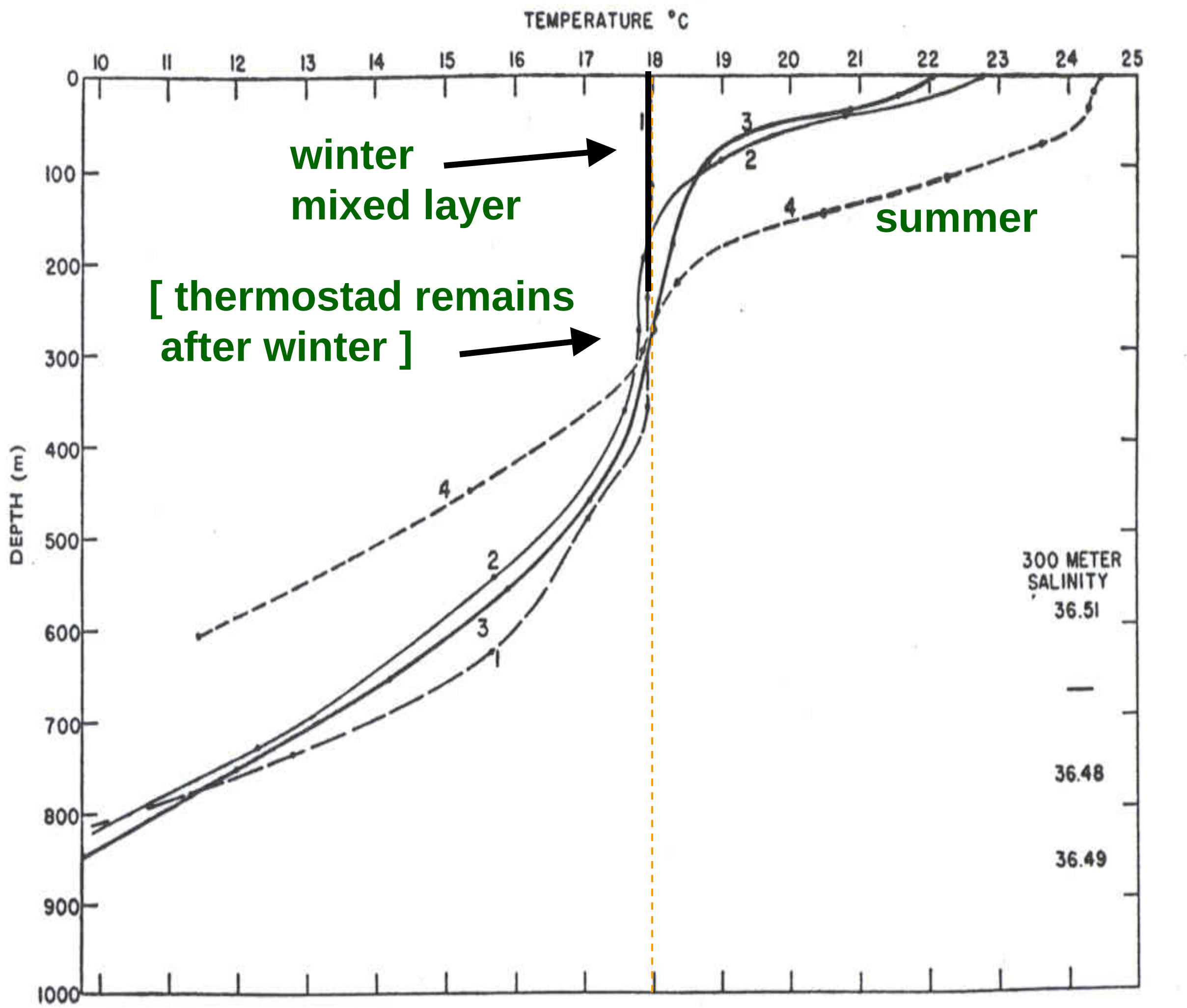
Large-scale asymmetric gyre….
possible

[See Stommel (1948) for the details]

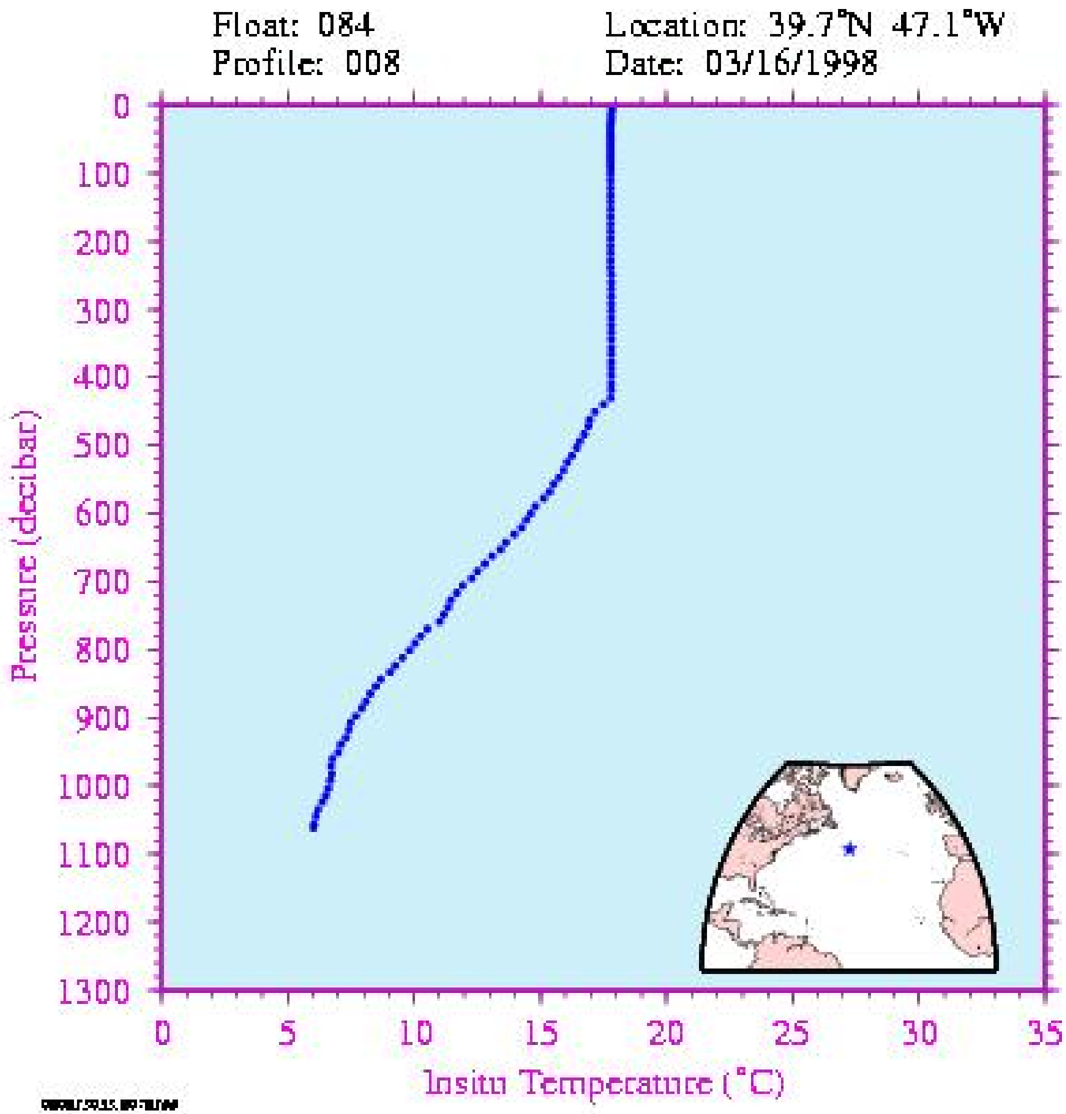
Large-scale symmetric gyre….
not possible



The surface mixed layer in the ocean….



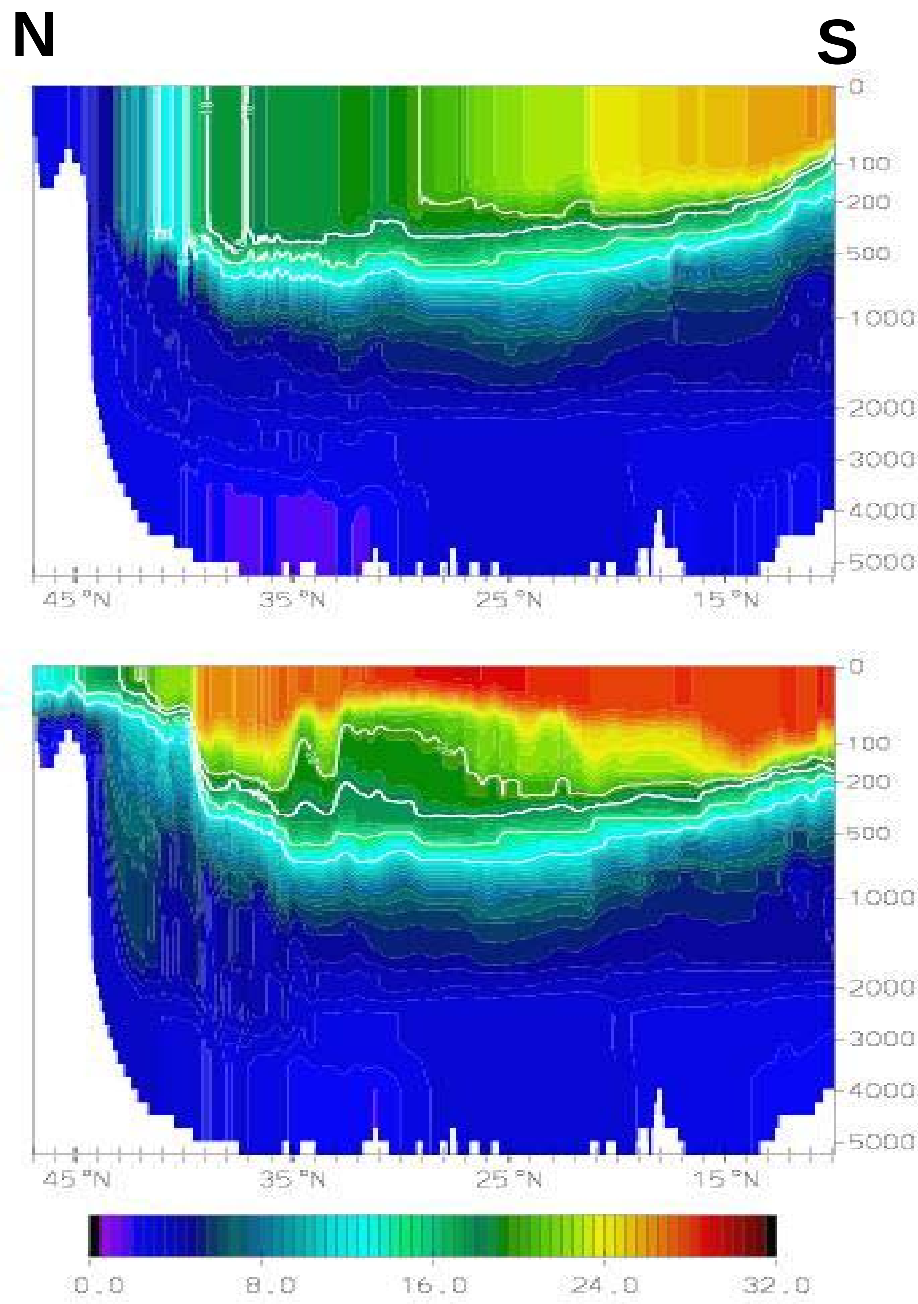
[western N. Atlantic]



Typical winter mixed layer in the western N. Atlantic, 3/98

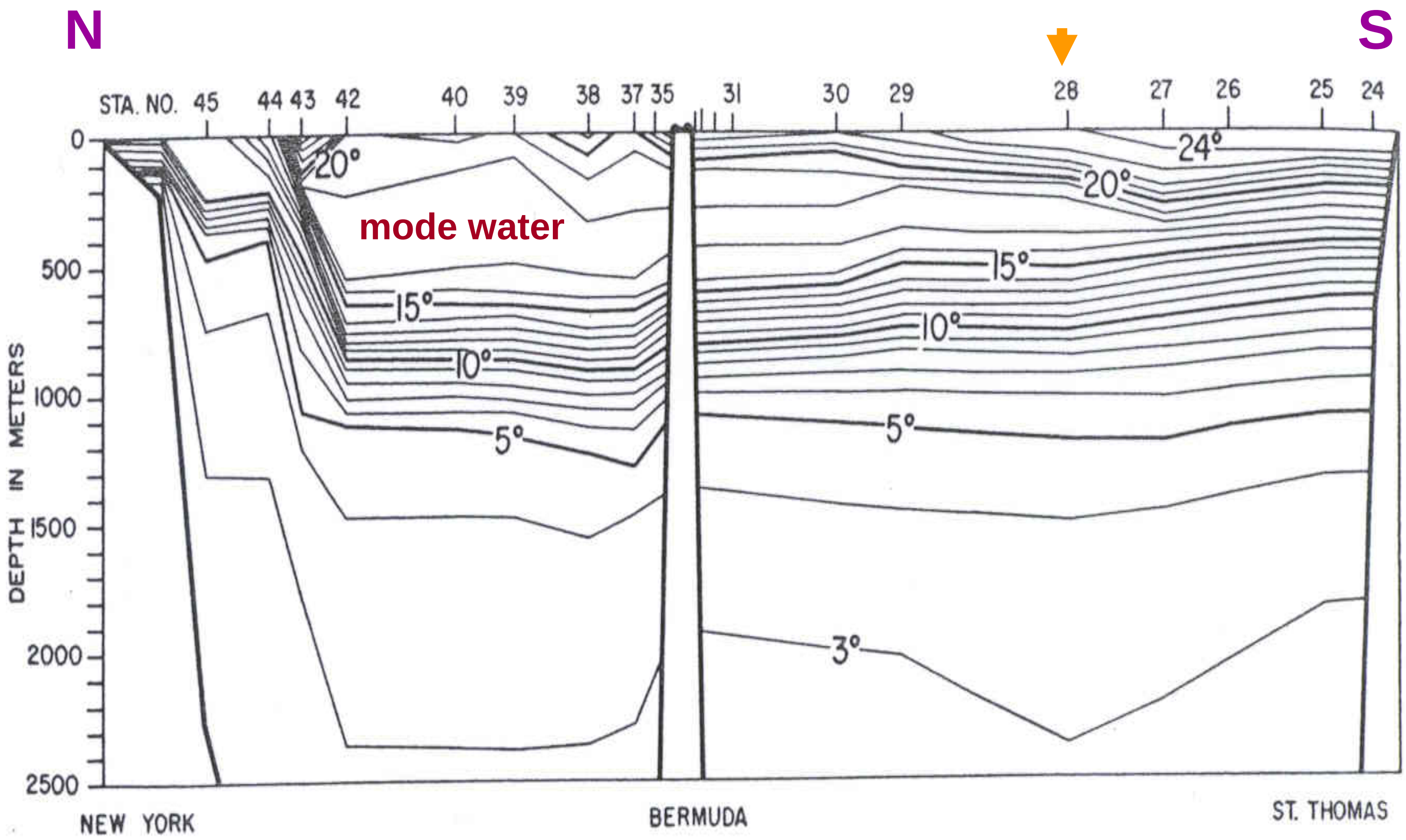
http://flux.ocean.washington.edu

Model results, western N. Atlantic mode water

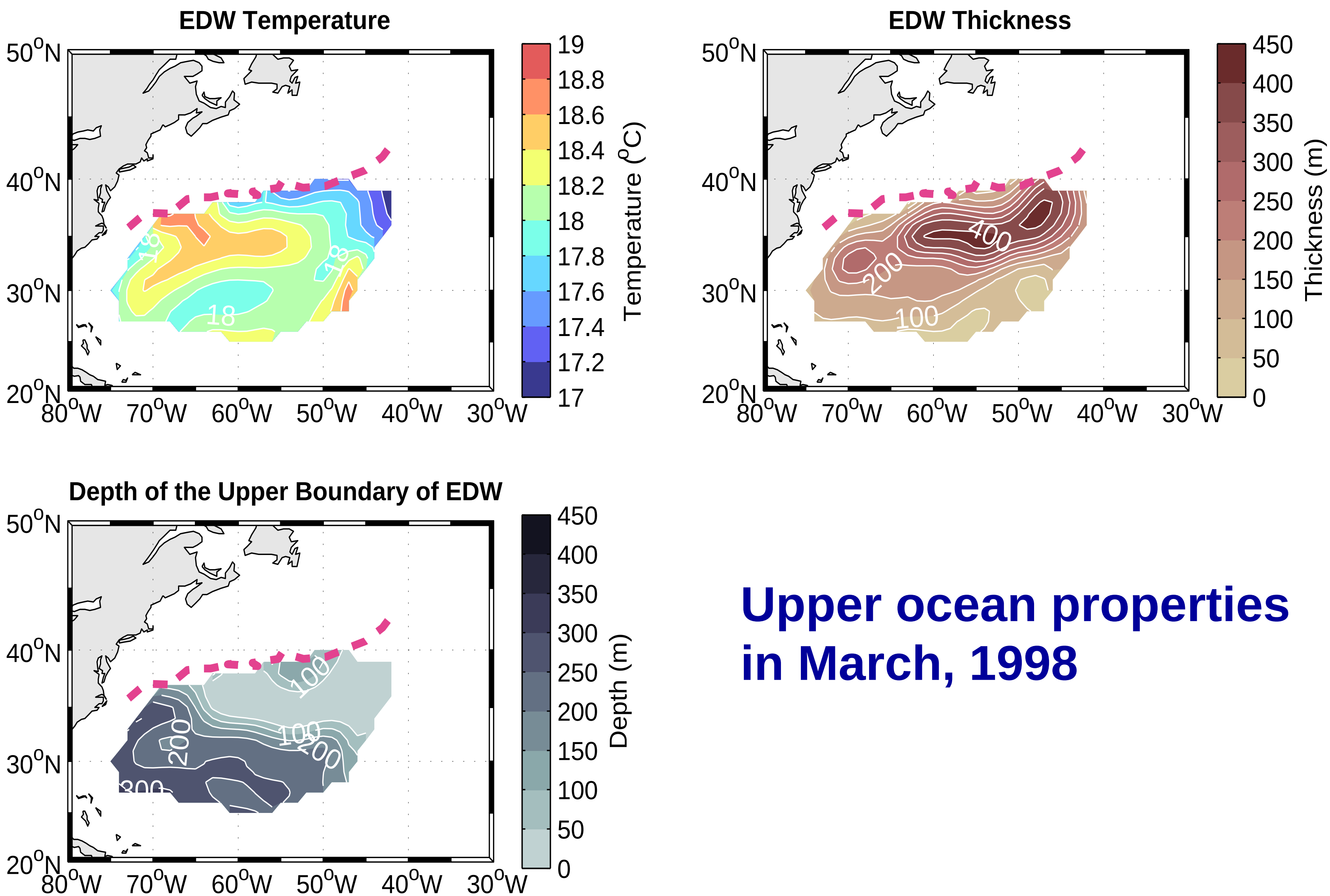


winter

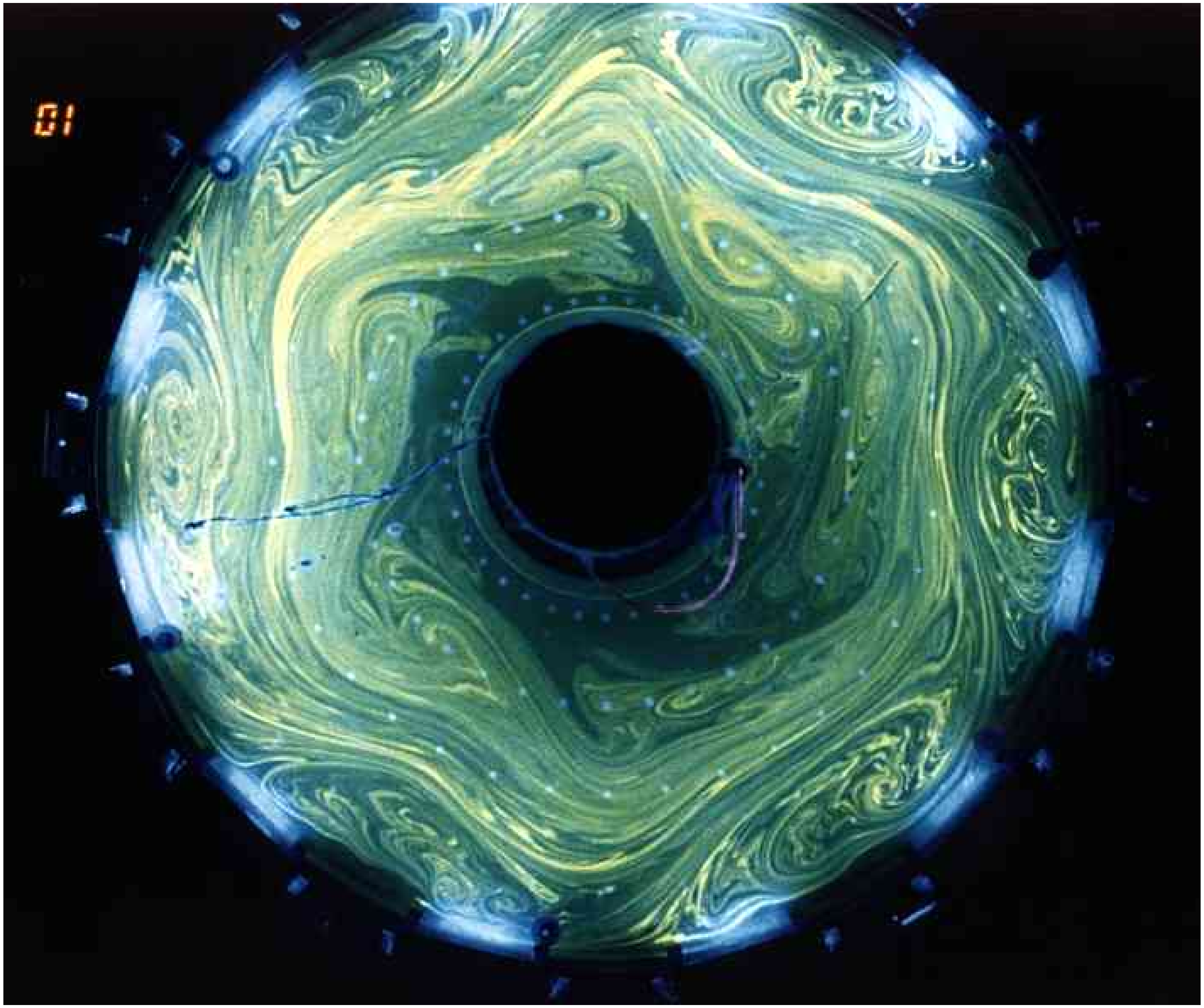
summer



Mode water, resulting from winter mixed layer, exists over much of the western N. Atlantic



Upper ocean properties in March, 1998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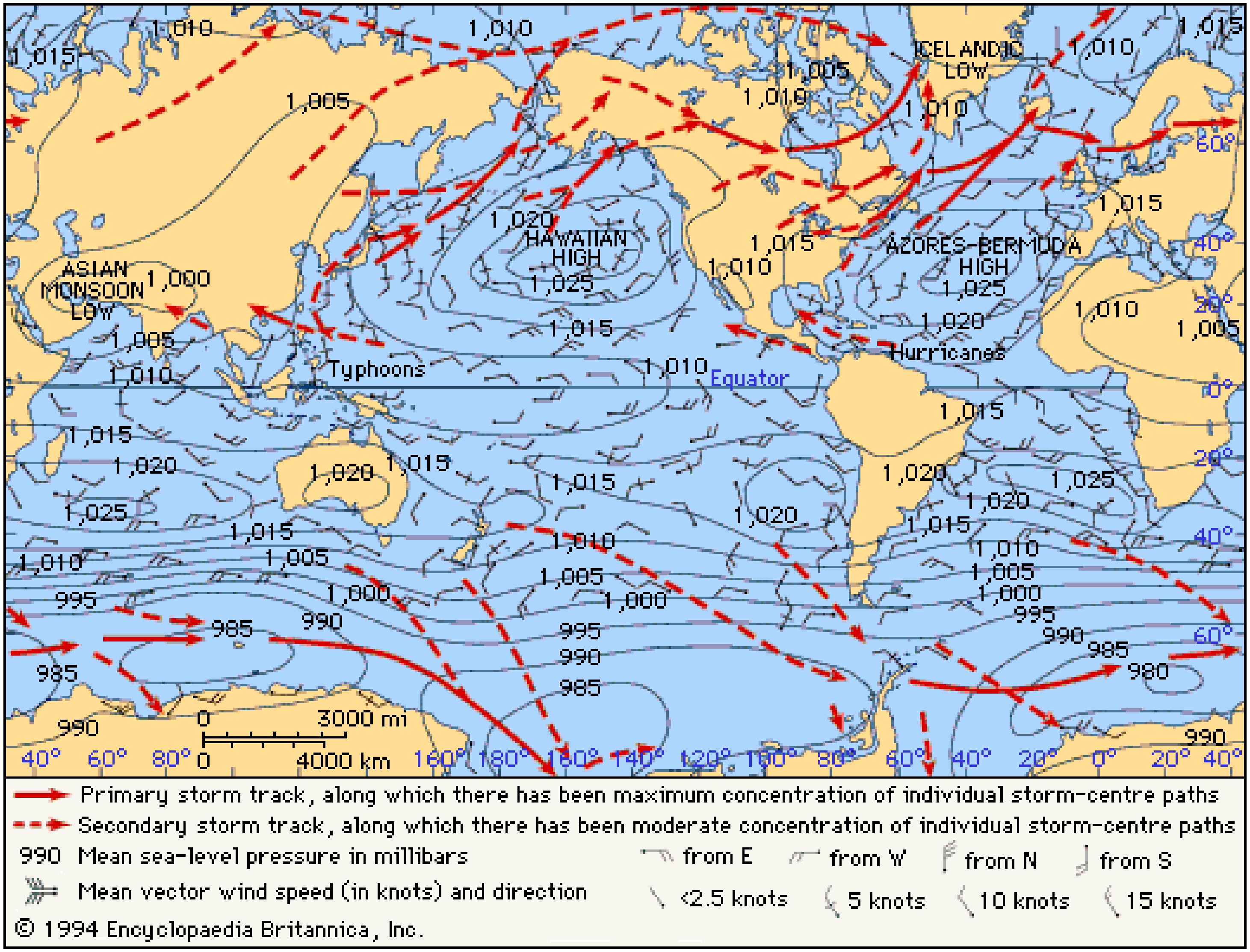
ICELANDIC LOW
HAWAIIAN HIGH
AZORES-BERMUDA HIGH
ASIAN MONSOON LOW
Typhoons
Hurricanes
Equator
1,010
1,015
1,005
1,020
1,025
1,000
995
990
985
980
60°
40°
20°
0°
0
3000 mi
4000 km
40° 60° 80° 160° 180° 160° 140° 120° 100° 80° 60° 40° 20° 0° 20° 40°
Primary storm track, along which there has been maximum concentration of individual storm-centre paths
Secondary storm track, along which there has been moderate concentration of individual storm-centre paths
990 Mean sea-level pressure in millibars
Mean vector wind speed (in knots) and direction
from E
from W
from N
from S
<2.5 knots
5 knots
10 knots
15 knots
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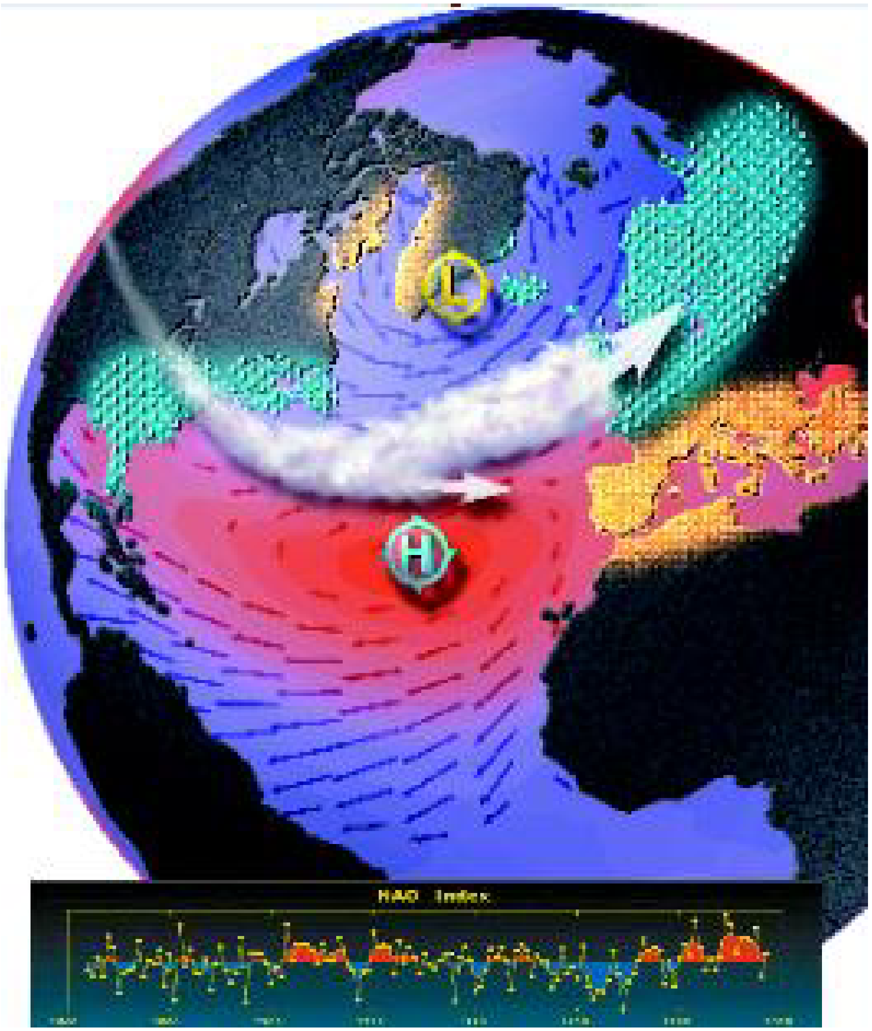
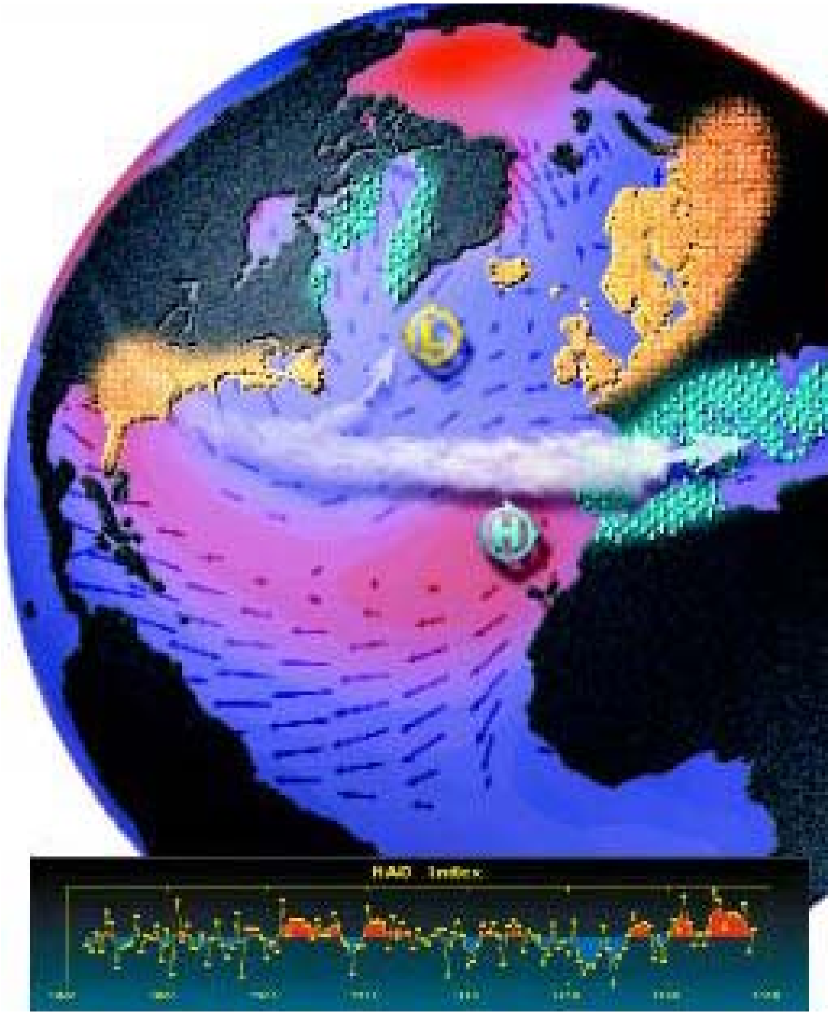
The North Atlantic Oscillation (NAO)….

⇒ Low NAO ~ Colder SST in the subtropics

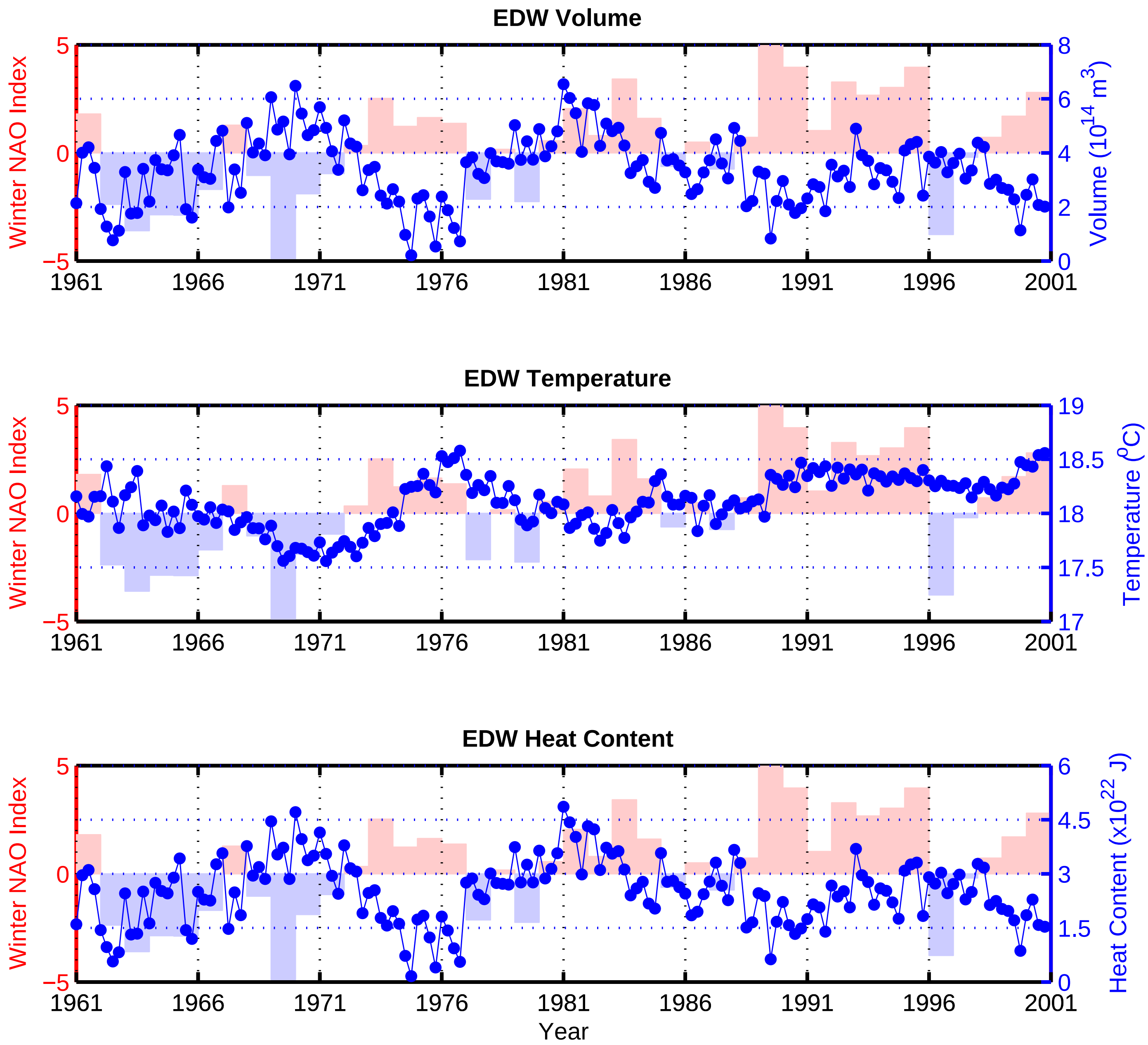
High NAO



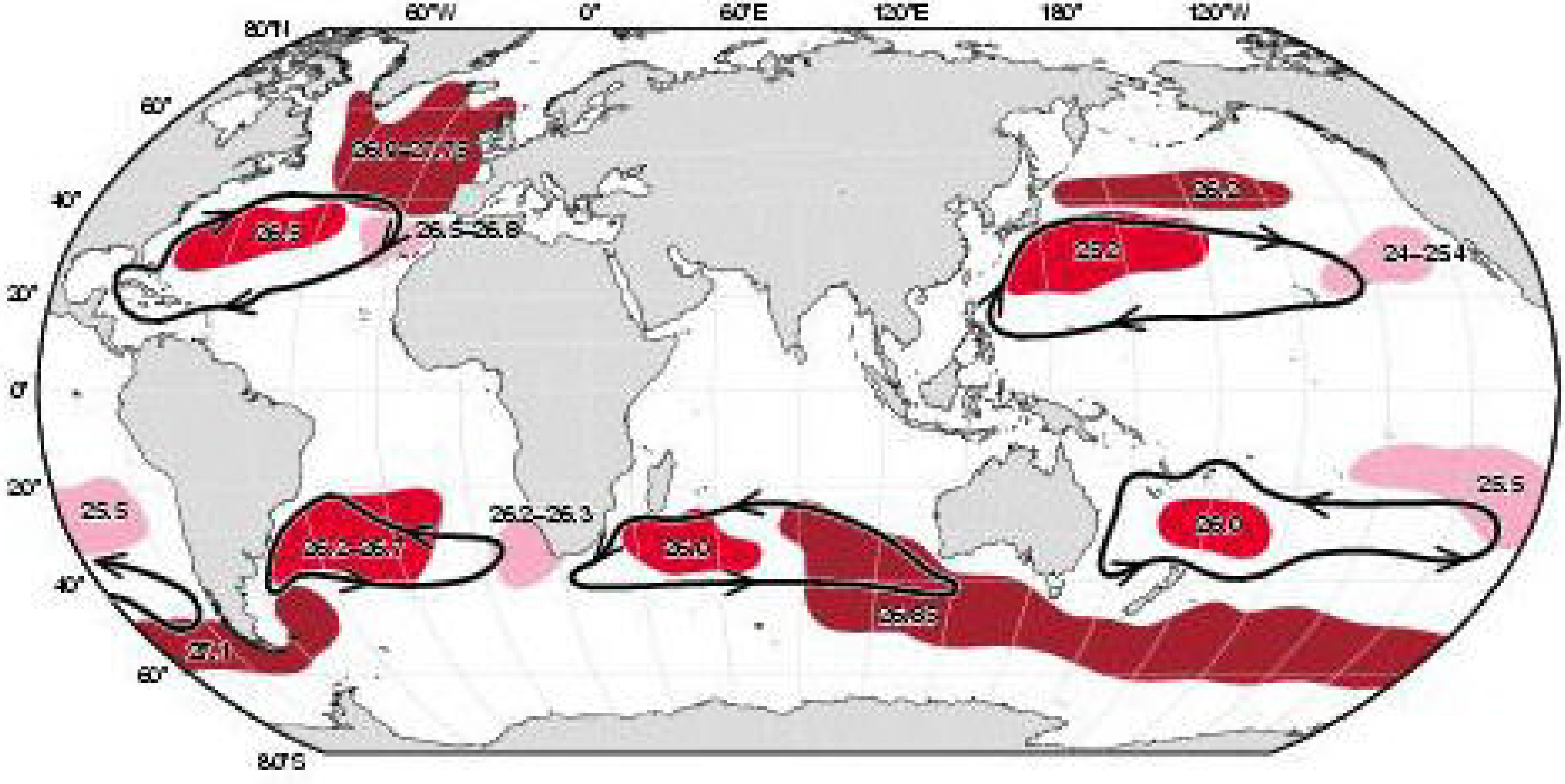
Low NAO



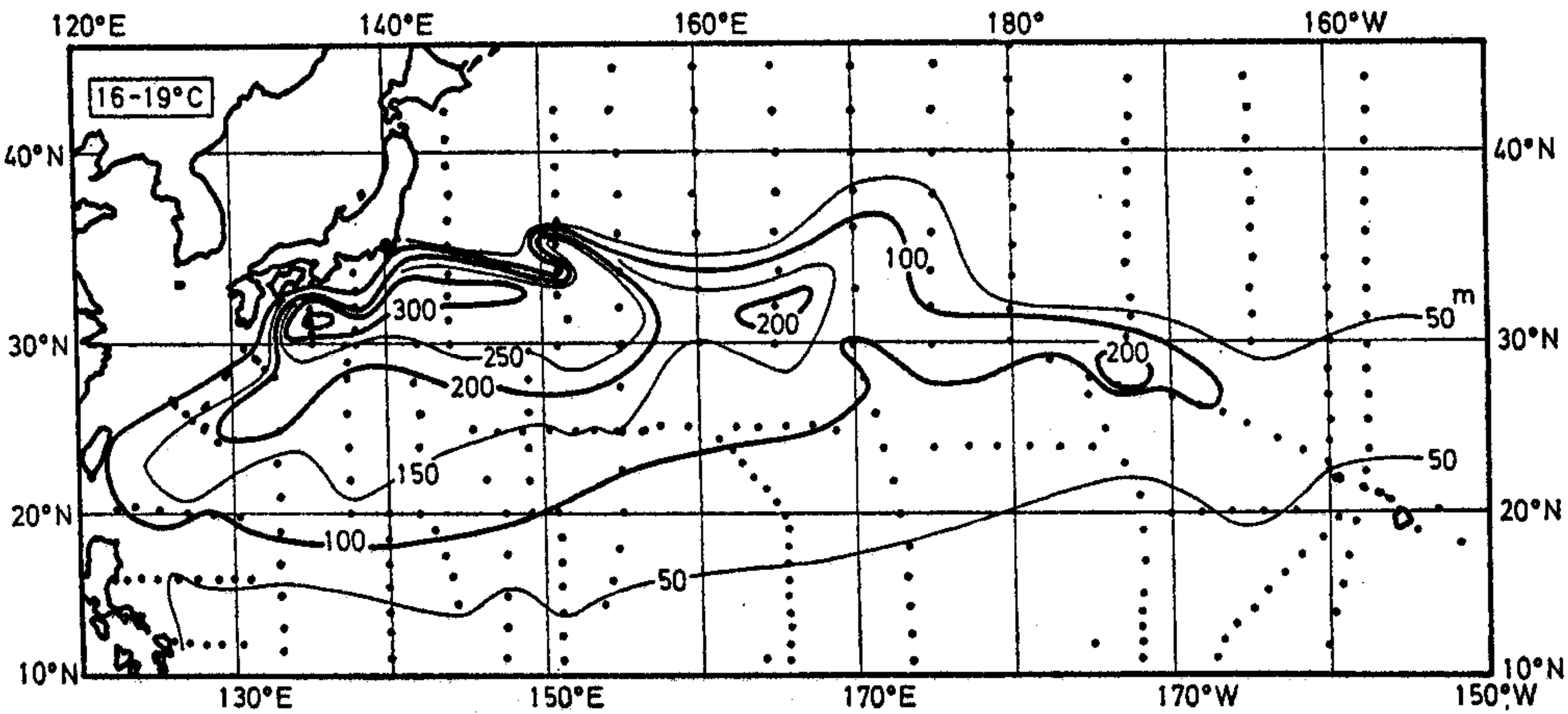
Long-term variability of western N. Atlantic mode water….



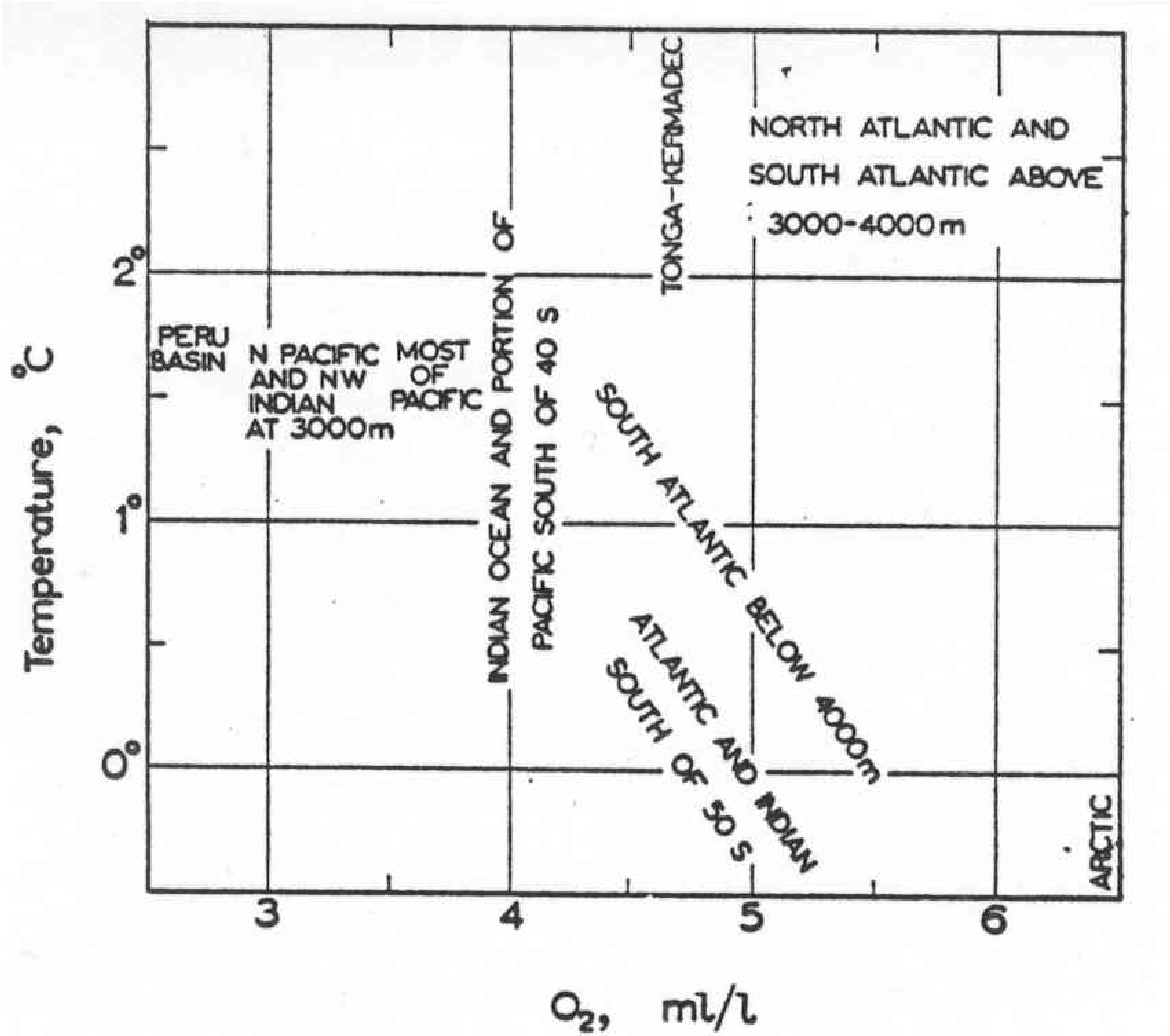
Mode waters in the world ocean….



Winter mixed layer depth in the N. Pacific….

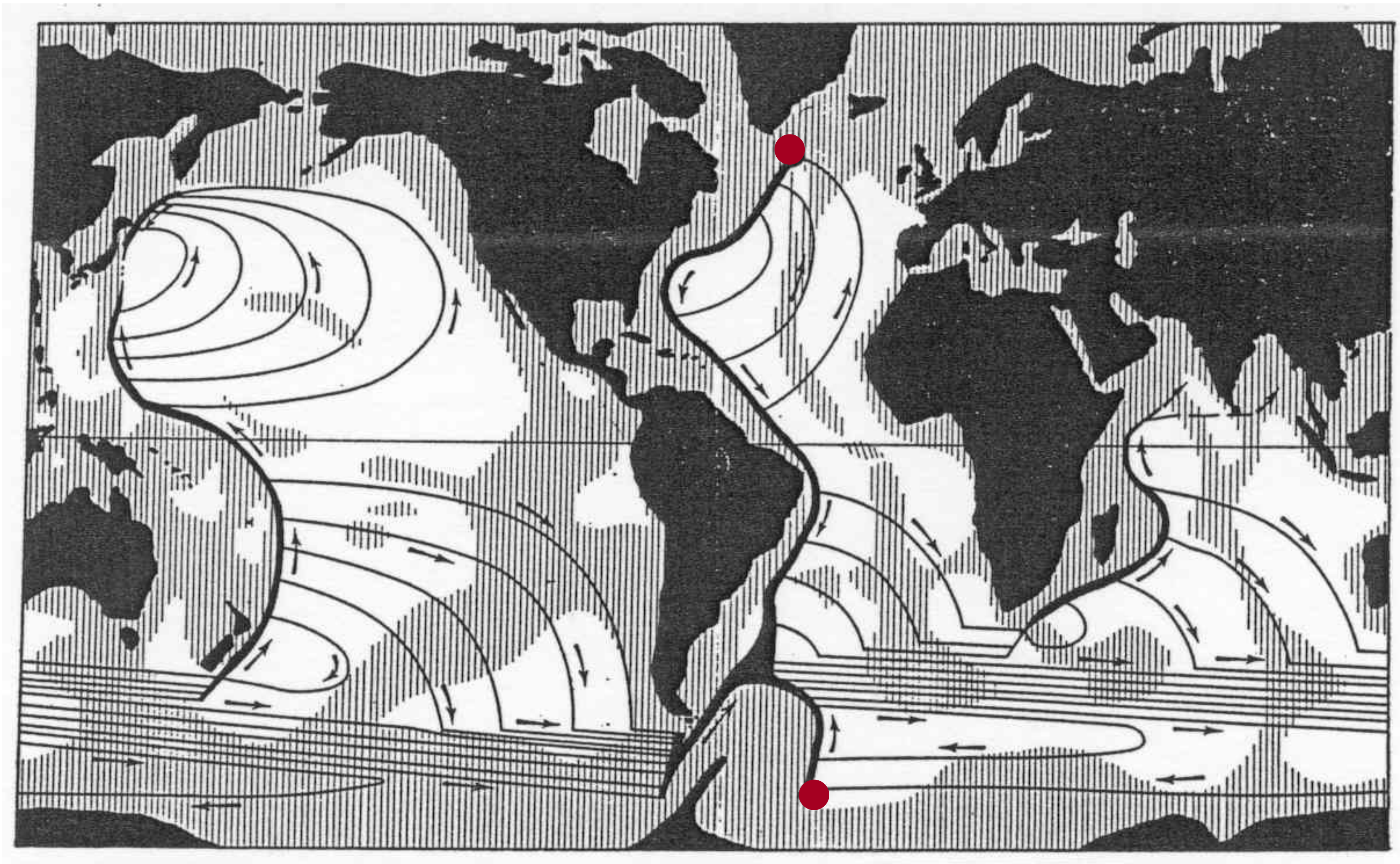


Deep circulation….

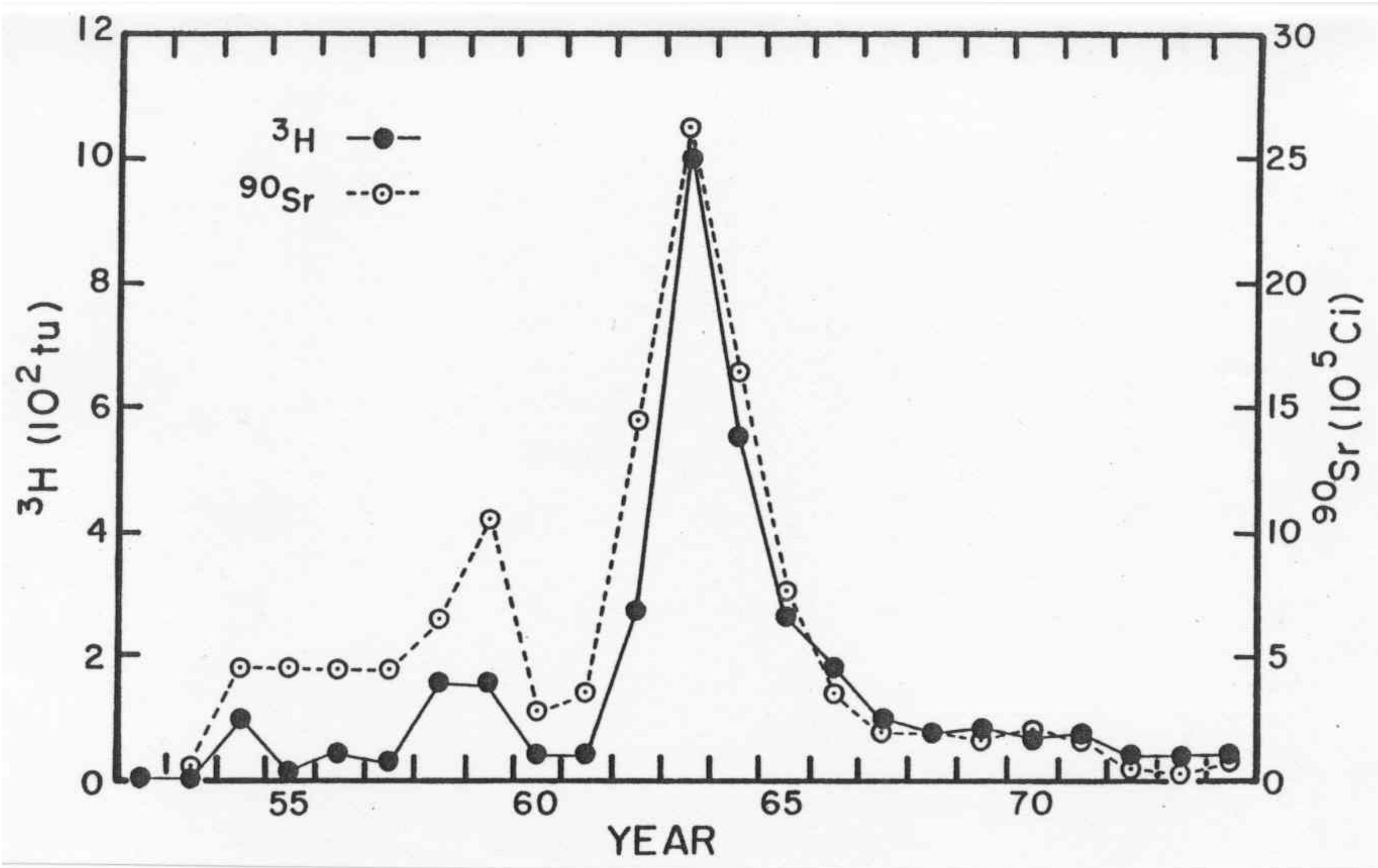


Stommel's *T*-O_2 diagram, suggesting locations of deep water sources

(1958)

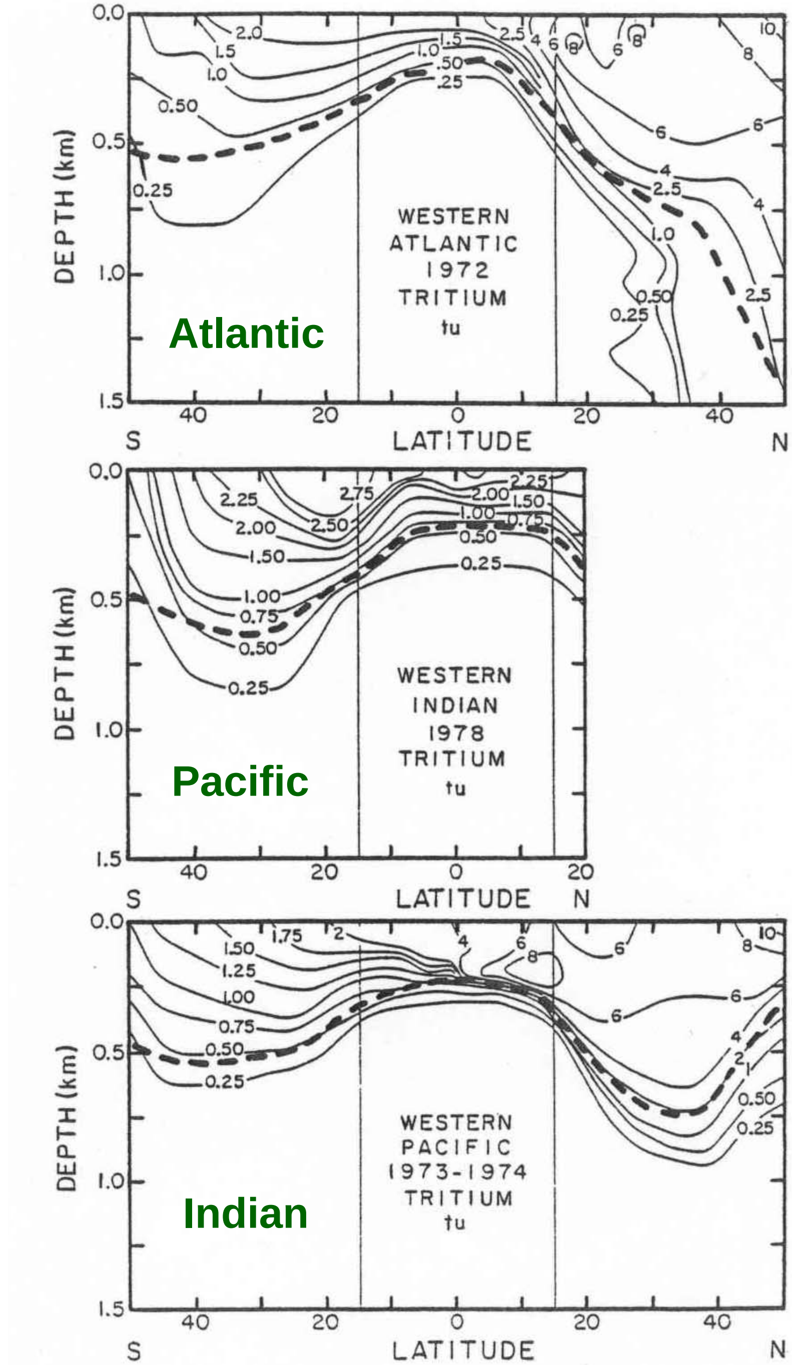


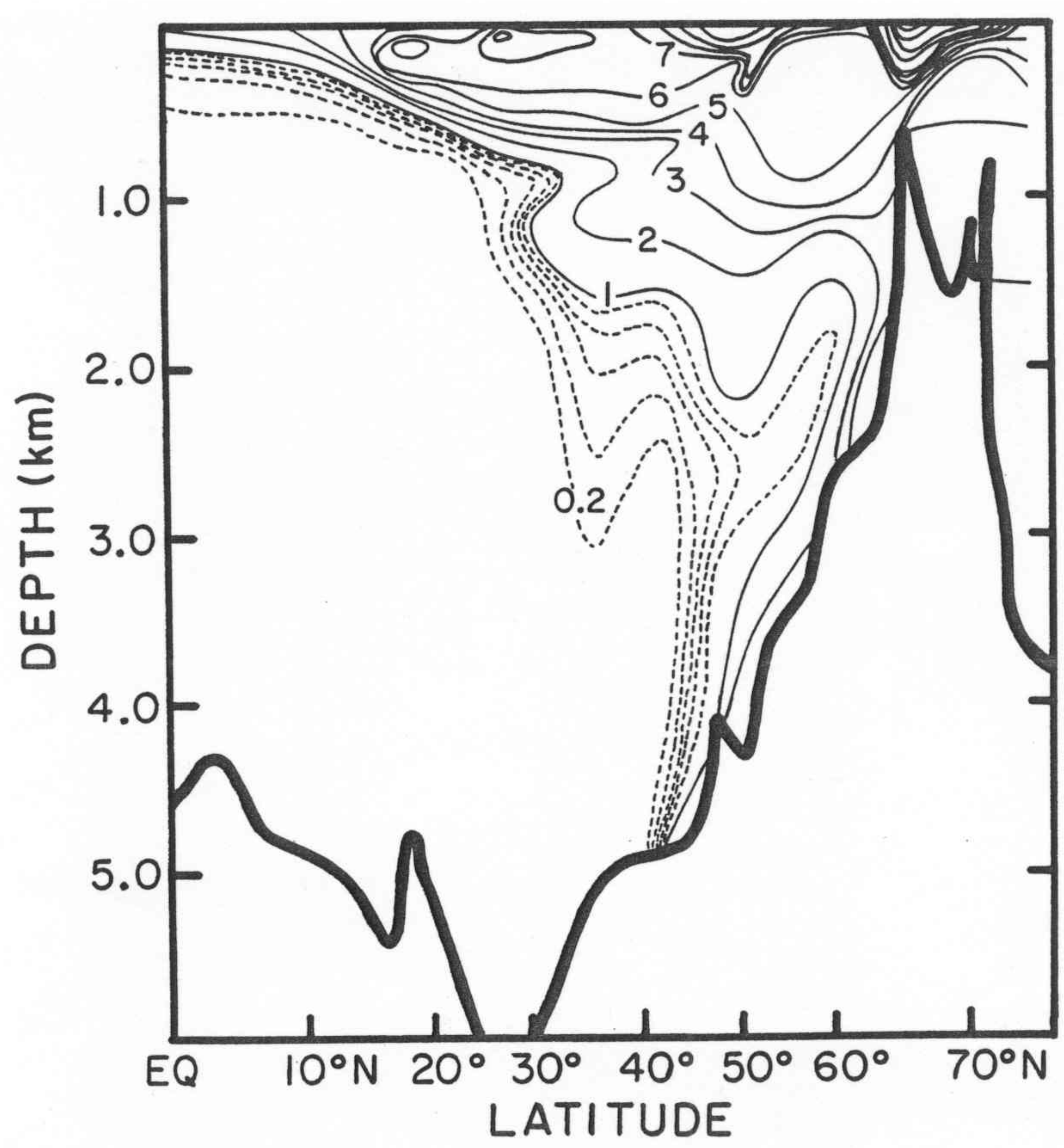
● **Stommel's deep water sources**



Source function for tritium (HTO) input to the ocean (tritium half-life ≈ 12.5 years)

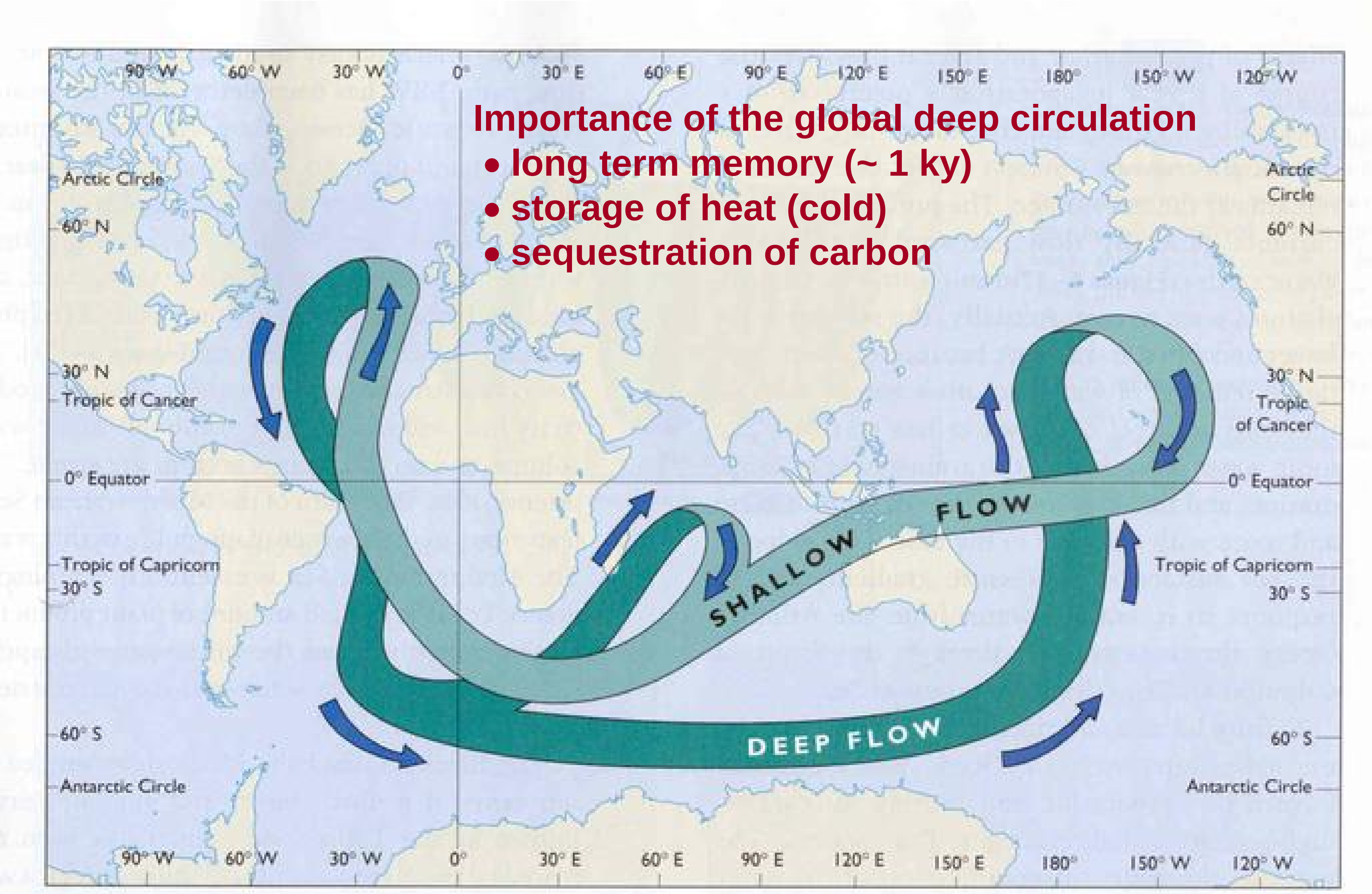
Tritium in the thermocline of the world ocean, 1970s





Tritium in the western N. Atlantic, 1972

Deep convection and the *conveyor belt*….



[apparent time scale: ~1000 years, from ^{14}C]