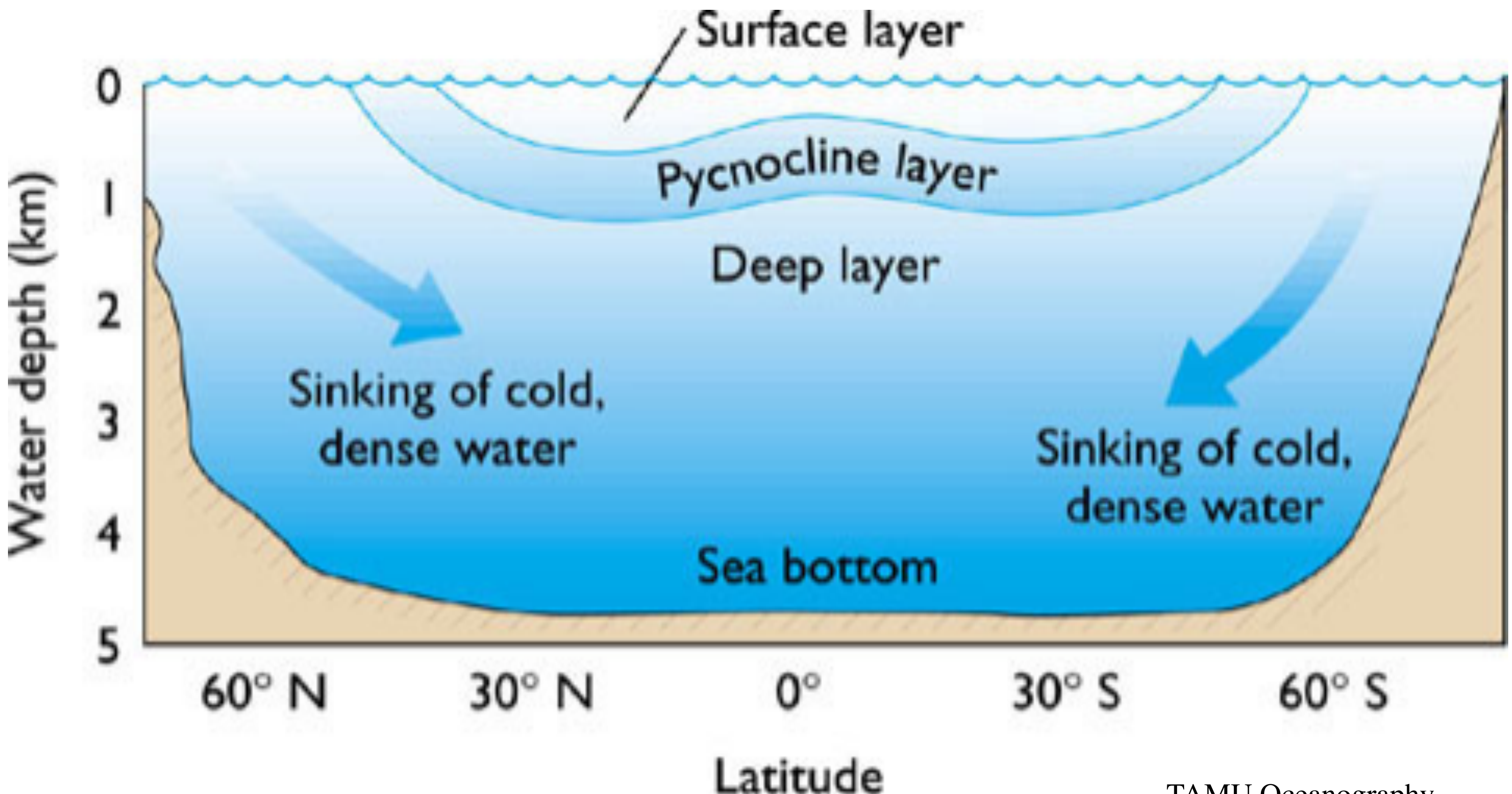
Abyssal Circulation

AOS660, Prof. McKinley, Fall 2013

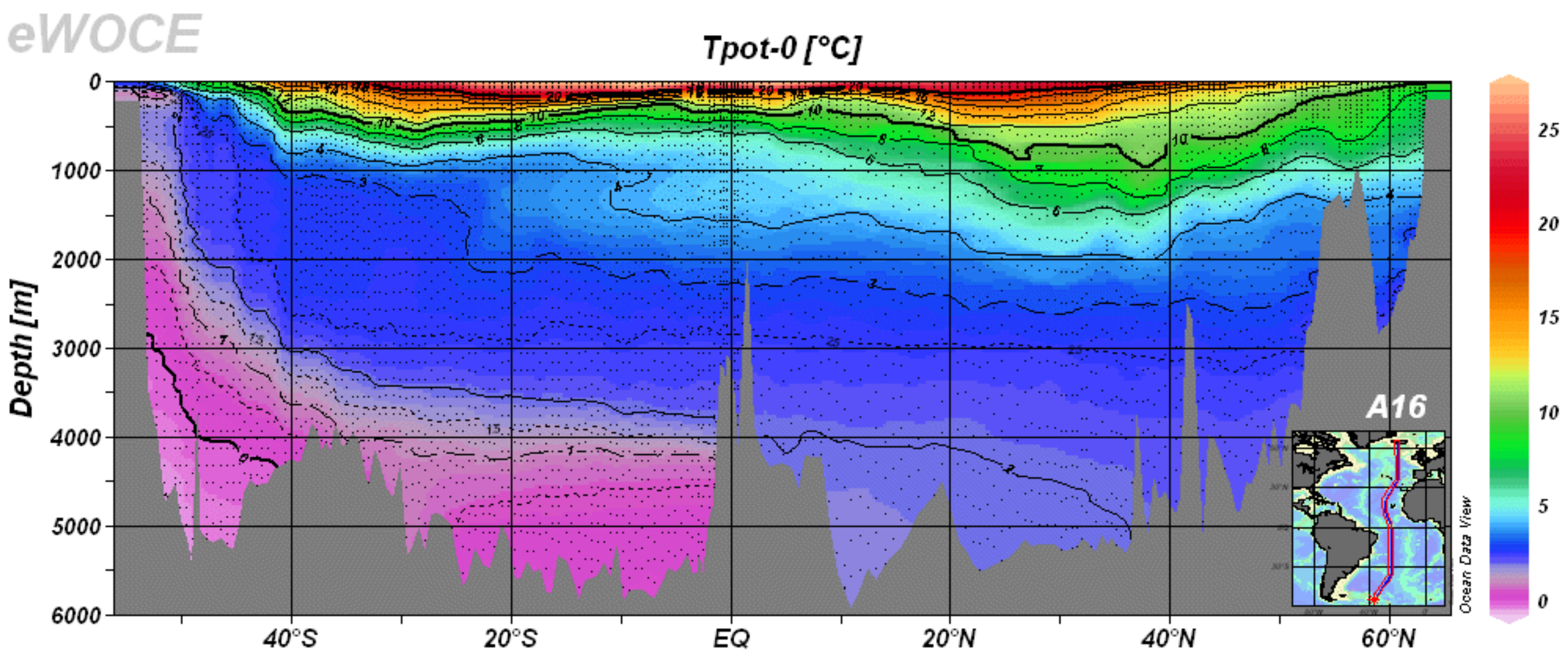
- Abyssal Circulation = Thermohaline Circulation
- Θ-S diagrams
- Sverdrup flow
  - Deep Western Boundary Currents (DWBC)
- Improvements to our scientific understanding
  - How fast is the Vertical Mixing?
    - Munk's classical calculation
    - Inhomogeneity of mixing
  - How coherent is the DWBC?
- Heat and Freshwater Transport

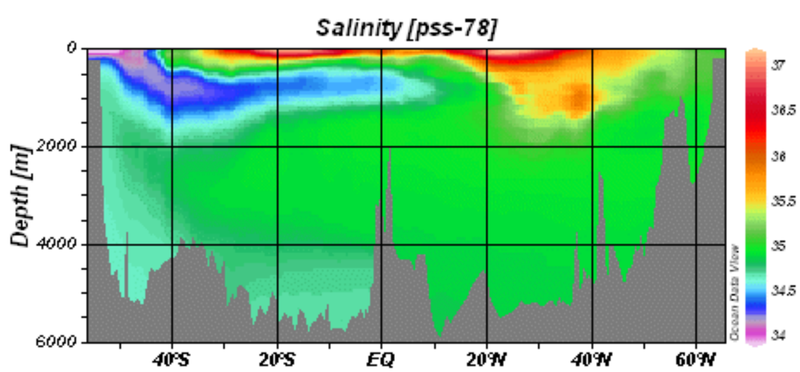
Density Cross-Section

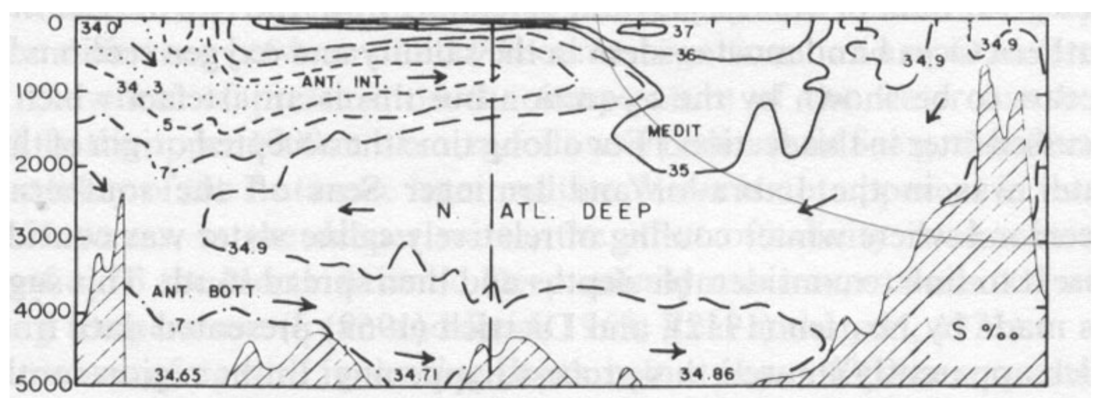


TAMU Oceanography

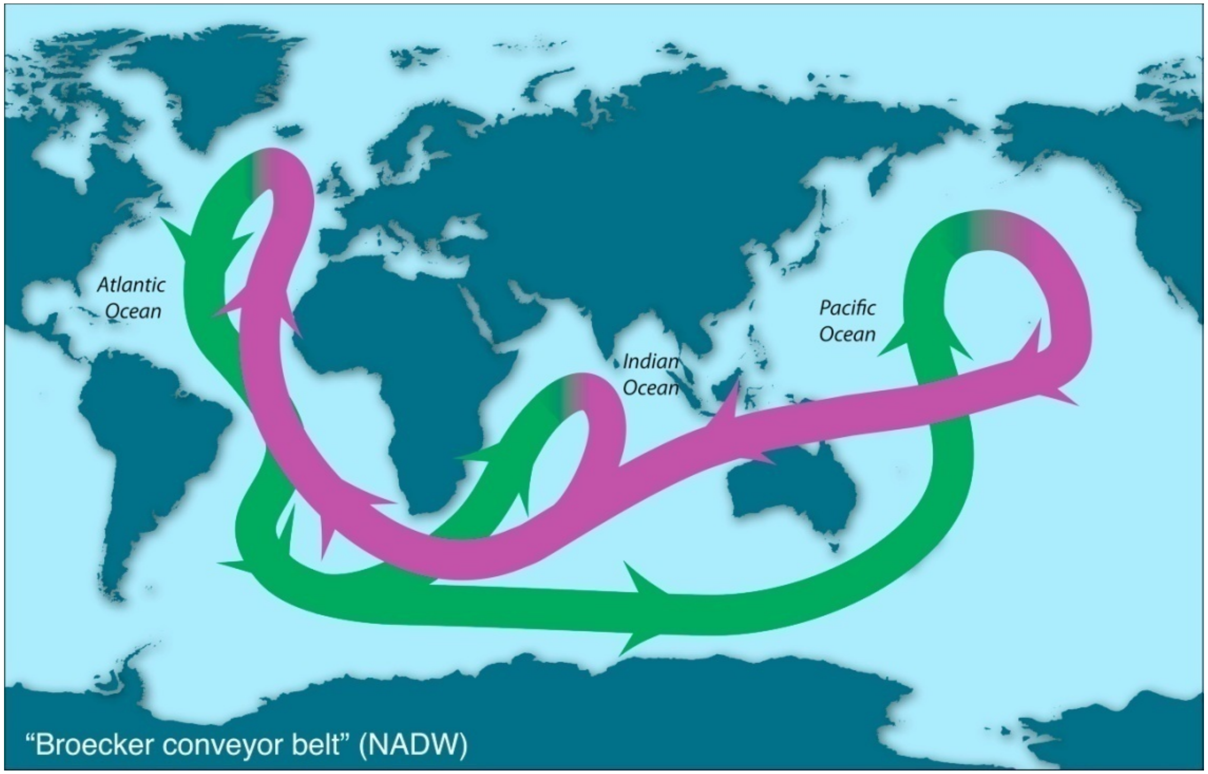
An oceanographic section







Thermohaline, simplified, 1980's



Simplified global NADW cell, which retains sinking only somewhere adjacent to the northern North Atlantic and upwelling only in the Indian and Pacific Oceans. See text for usefulness of, and also issues with, this popularization of the global circulation, which does not include any Southern Ocean processes. *Source: After Broecker (1987).*

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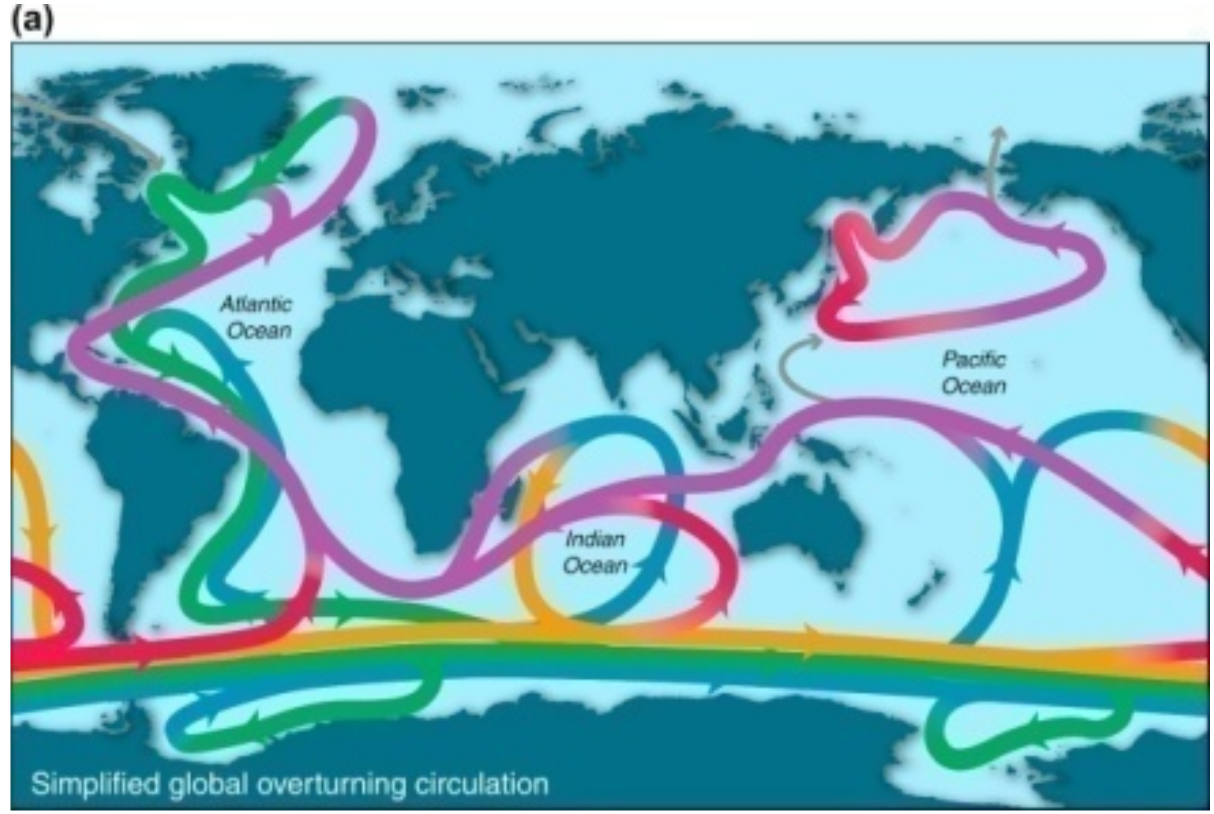
FIGURE 14.10

Thermohaline, still simplified, 2000's



Global overturning circulation schematics. (a) The NADW and AABW global cells and the NPIW cell. (b) Overturn from a Southern Ocean perspective. *Source: After Gordon (1991), Schmitz (1996b), and Lumpkin and Speer (2007).* (c) Two-dimensional schematic of the interconnected NADW, ID
W, PDW, and AABW cells. The schematics do not accurately depict locations of sinking or the broad geographic scale of upwelling. Colors: surface water (purple), intermediate and Southern Ocean mode water (red), PDW/IDW/UCDW (orange), NADW (green), AABW (blue). See Figure S14.1 on the textbook Web site for a complete set of diagrams. This figure can also be found in the color insert. *Source: From Talley (2011).*

FIGURE 14.11a

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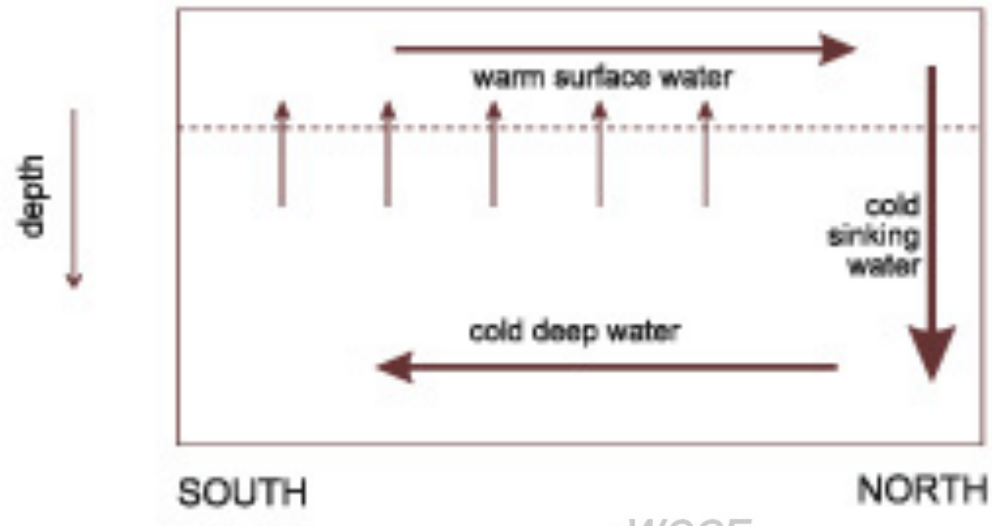
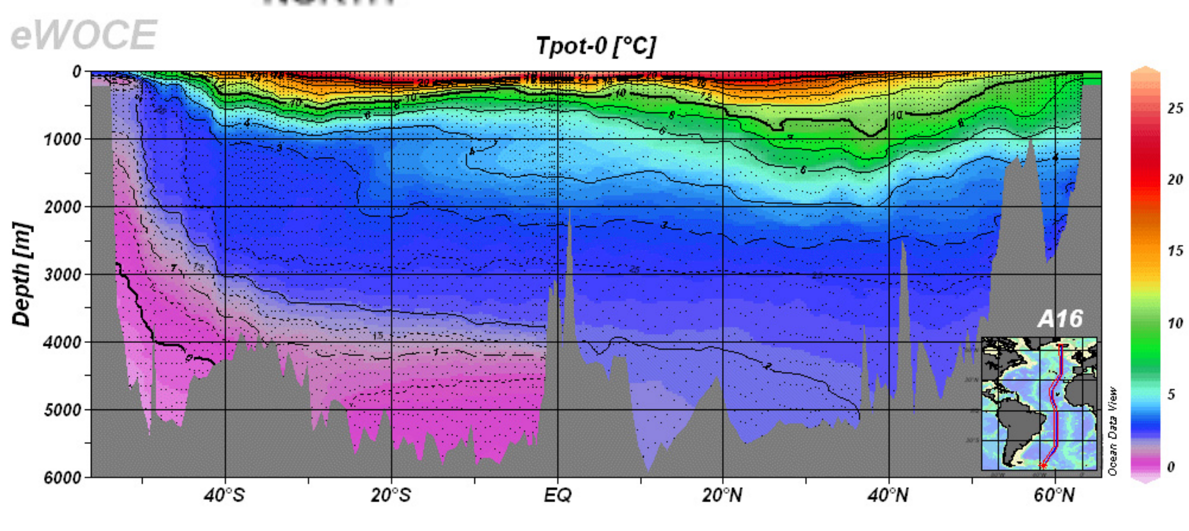
Global zonal average of THC



Figure



Thermohaline Circulation

- Buoyancy forcing in key high latitude locations
  - Heat loss driving buoyancy loss dominates (cooling)
  - Freshwater input causes buoyancy gain
  - In some locations, brine rejection causes important buoyancy loss

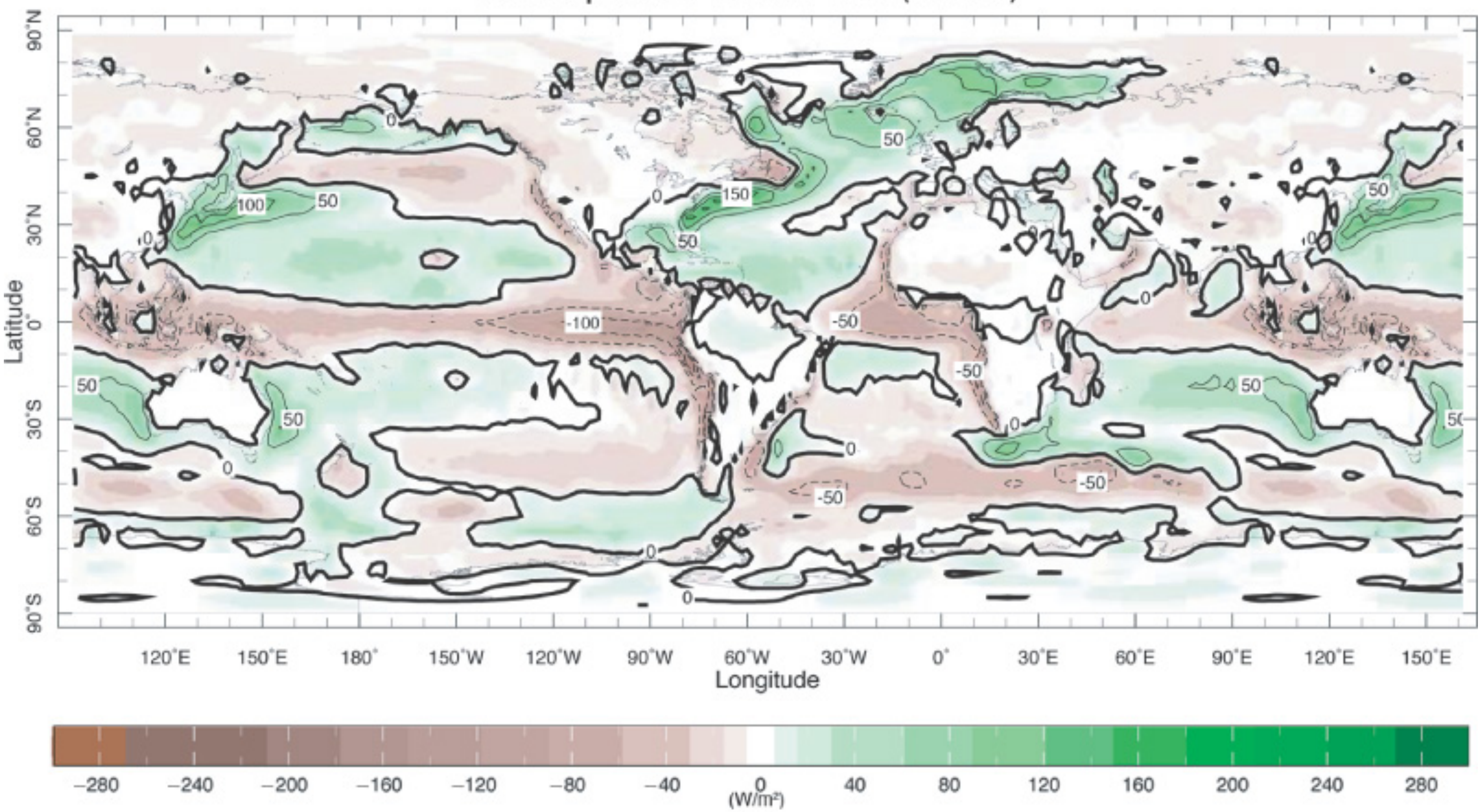
- Why doesn't it sink in Pacific?
  - Strong stratification promoted by large freshwater inputs

Buoyancy

$$b = -g\left(\frac{\rho - \rho_e}{\rho}\right)$$

$$\rho = \rho(T, S)$$

Net Upward Heat Flux (W/m^2)

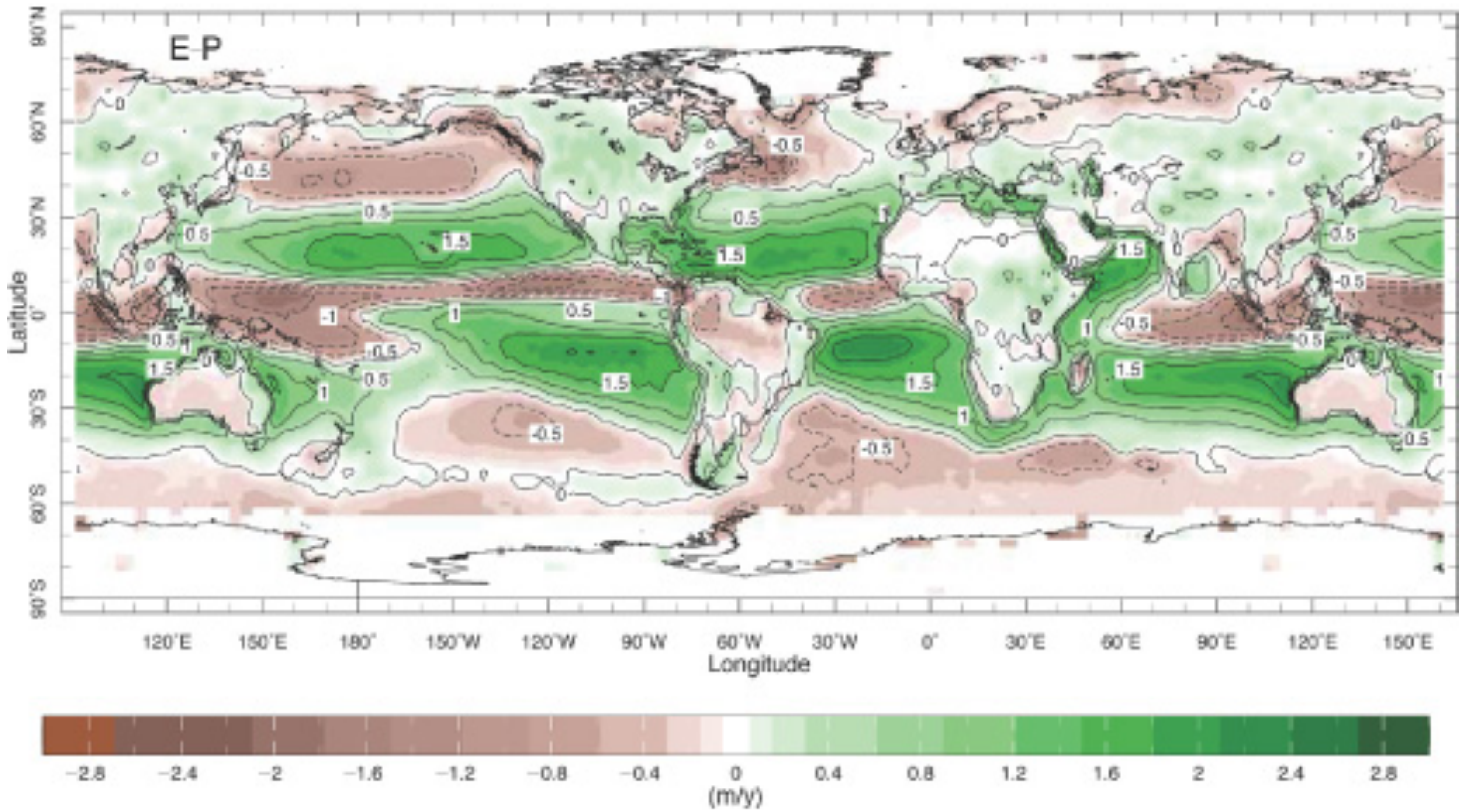


(From Kalnay et al. (1996).)

Net Salt Flux

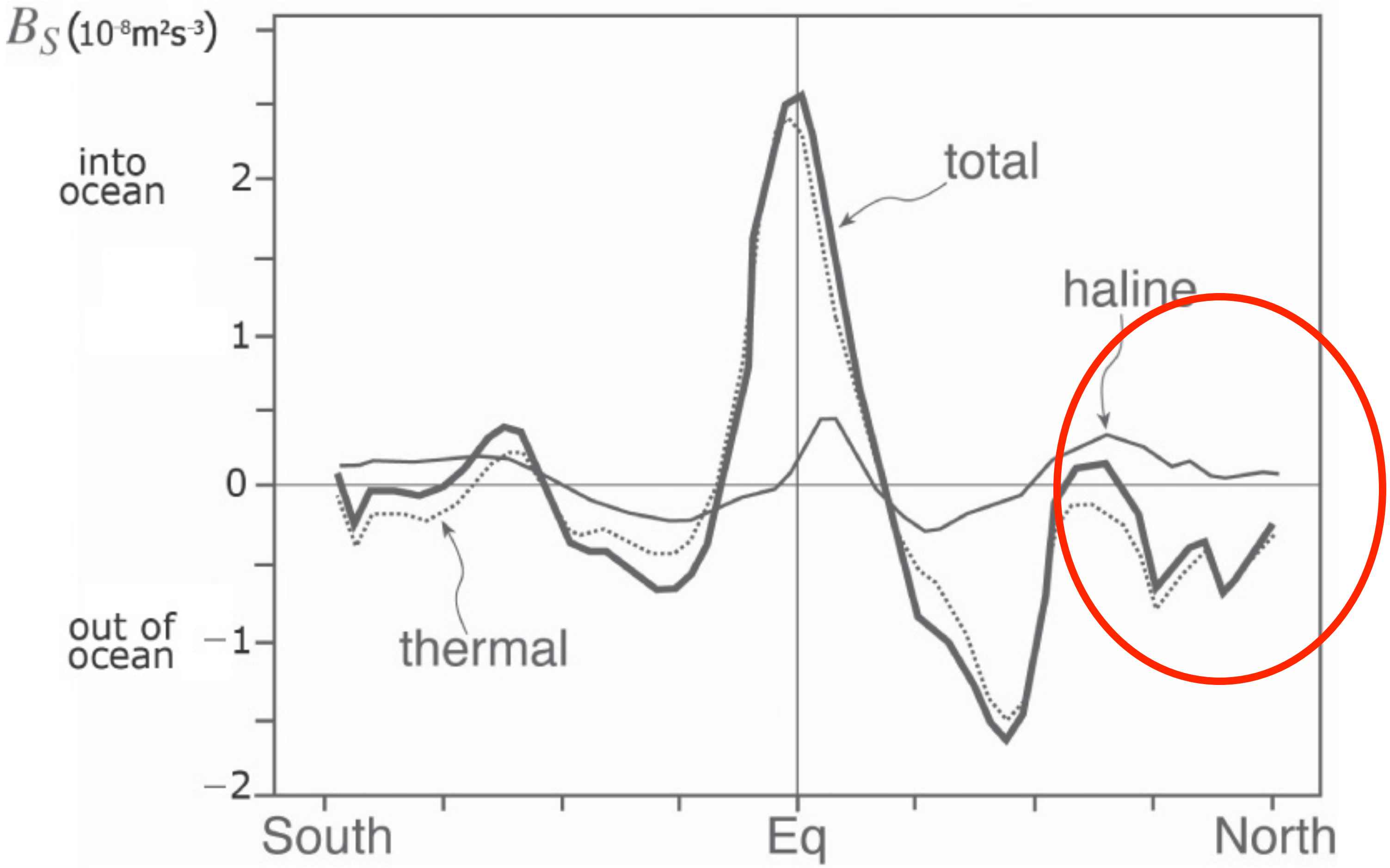
Evaporation - Precipitation

E-P



(From Kalnay et al. (1996).)

Net Buoyancy LOSS at high latitudes, GAIN at low latitudes

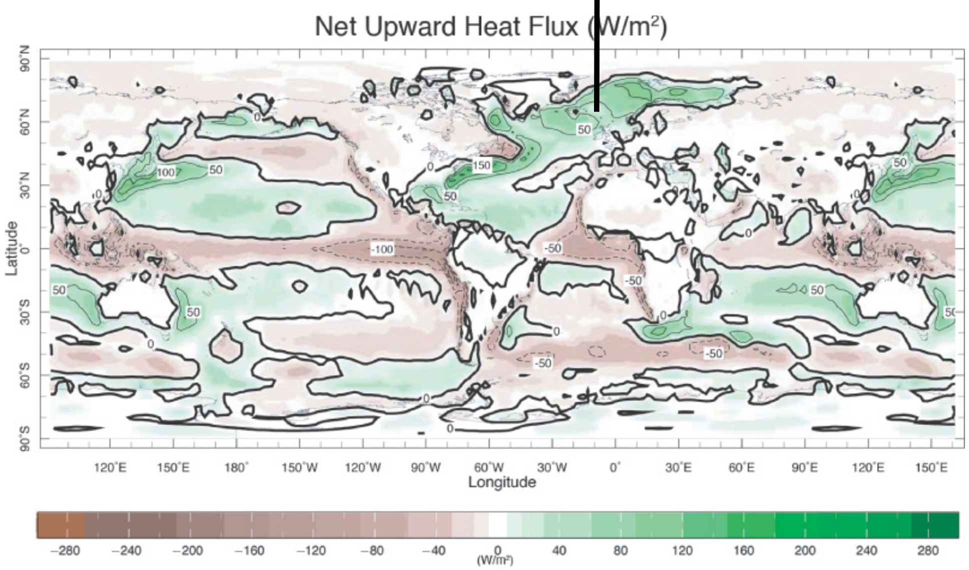


(Data from Kalnay et al. (1996).)

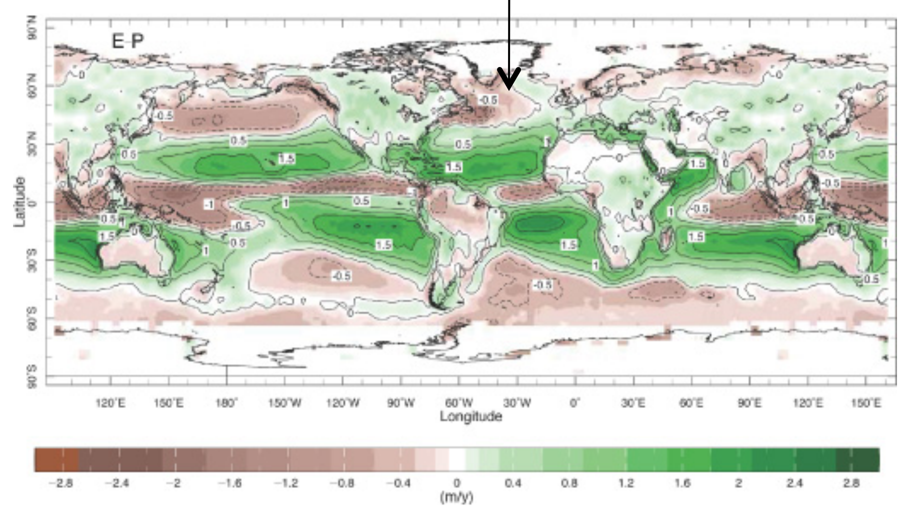
N. Atlantic: Competing Effects

Buoyancy Loss

Buoyancy Gain
(net Precip)

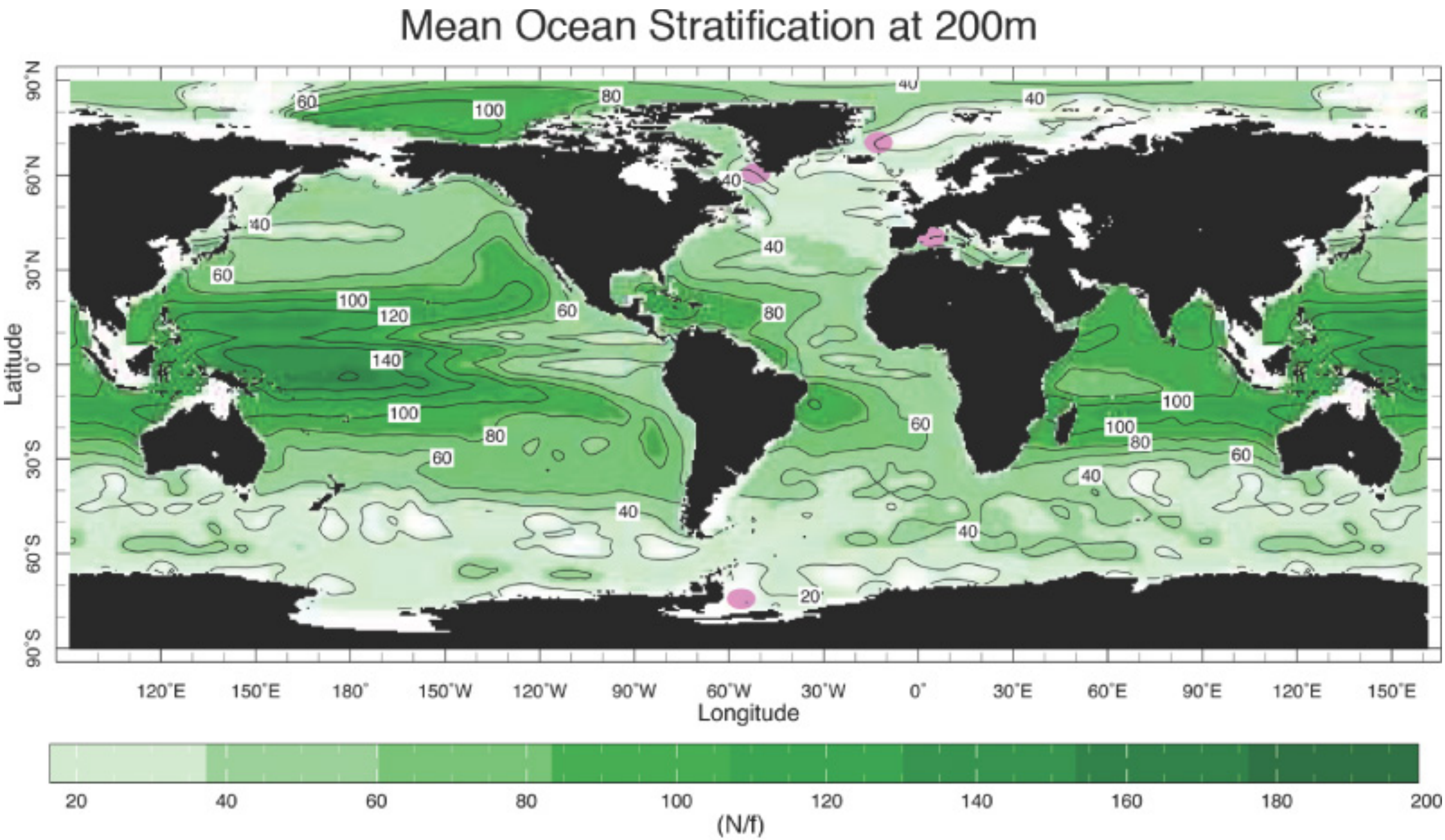


(From Kalnay et al. (1996).)



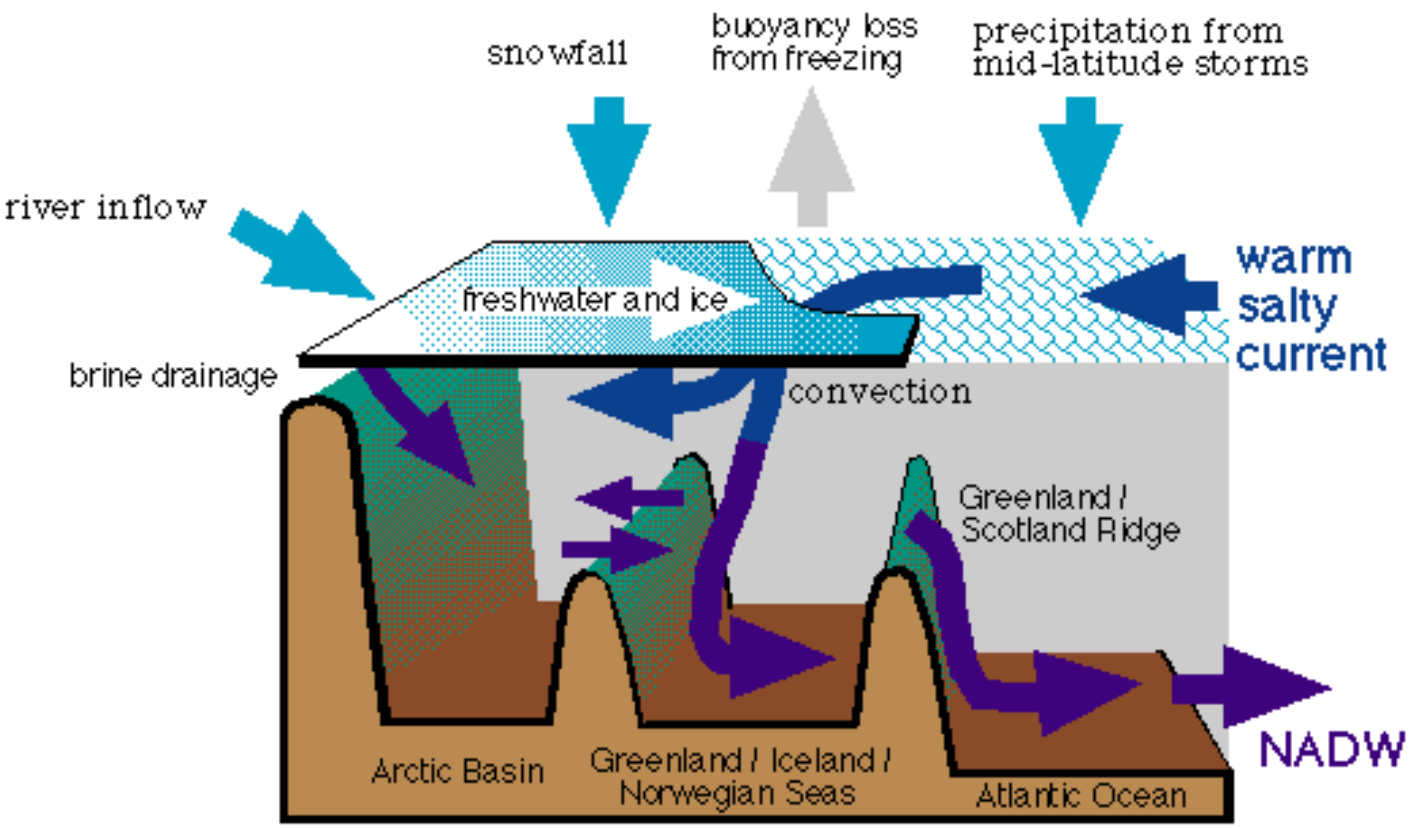
(From Kalnay et al. (1996).)

N/f


Mean Ocean Stratification at 200m
Latitude
90°N
60°N
30°N
0°
30°S
60°S
90°S
120°E
150°E
180°
150°W
120°W
90°W
60°W
30°W
0°
30°E
60°E
90°E
120°E
150°E
Longitude
20
40
60
80
100
120
140
160
180
200
(N/f)

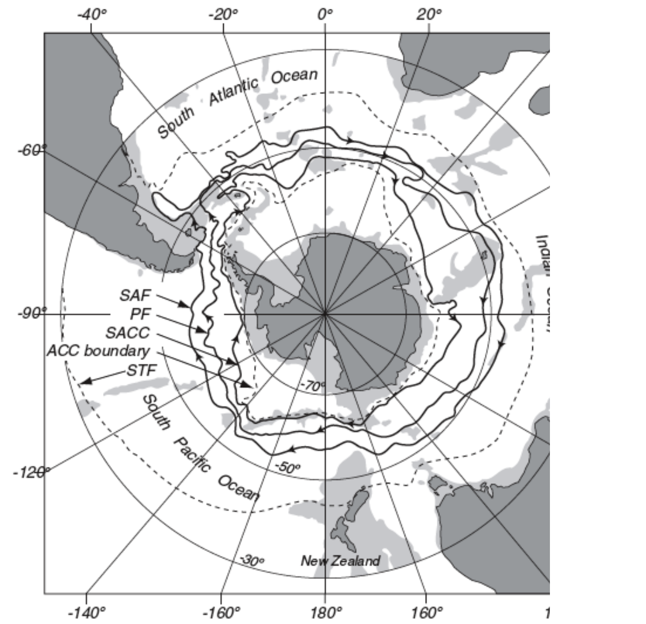
Deep water formation is, in fact, quite localized and deep water properties get set by these local conditions

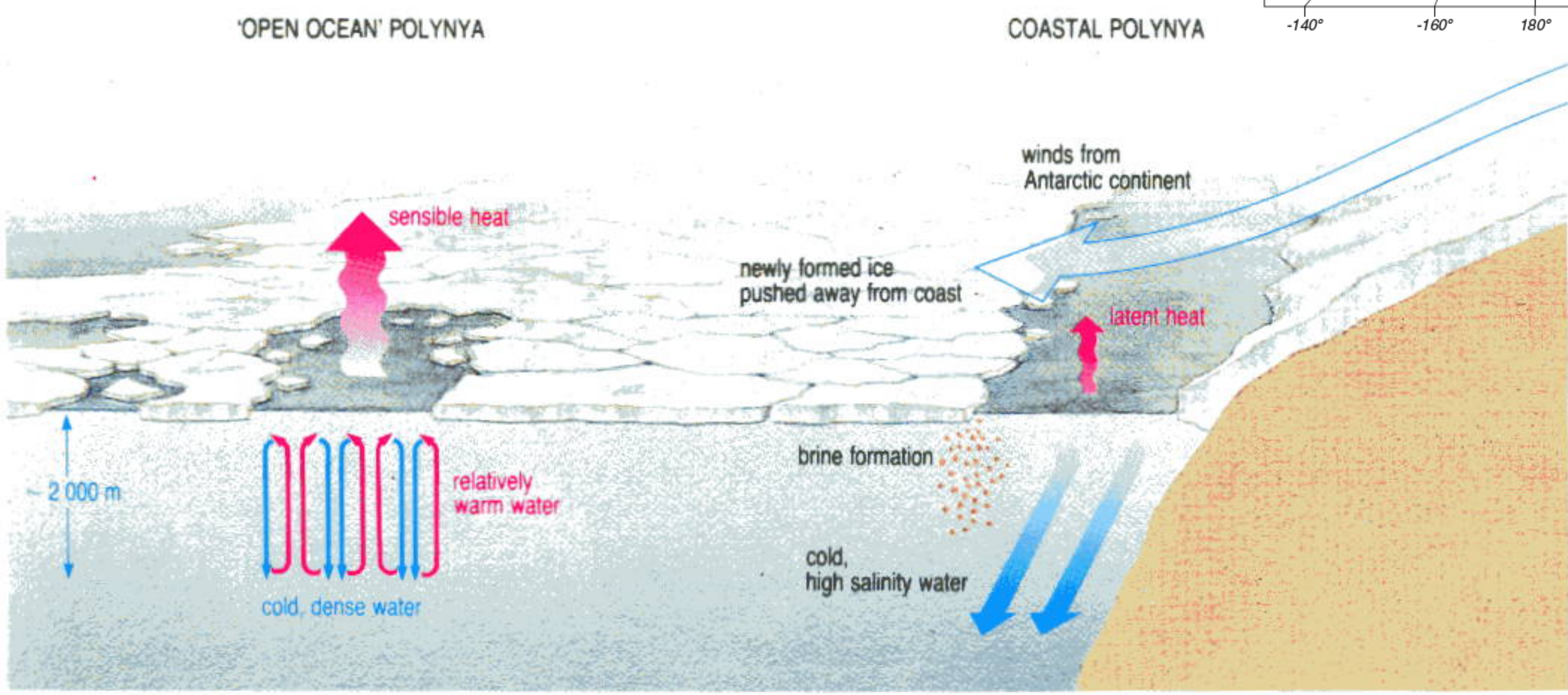
NADW formation



M. Tomczak and S. J. Godfrey

Polynyna – Sea Ice "factory"



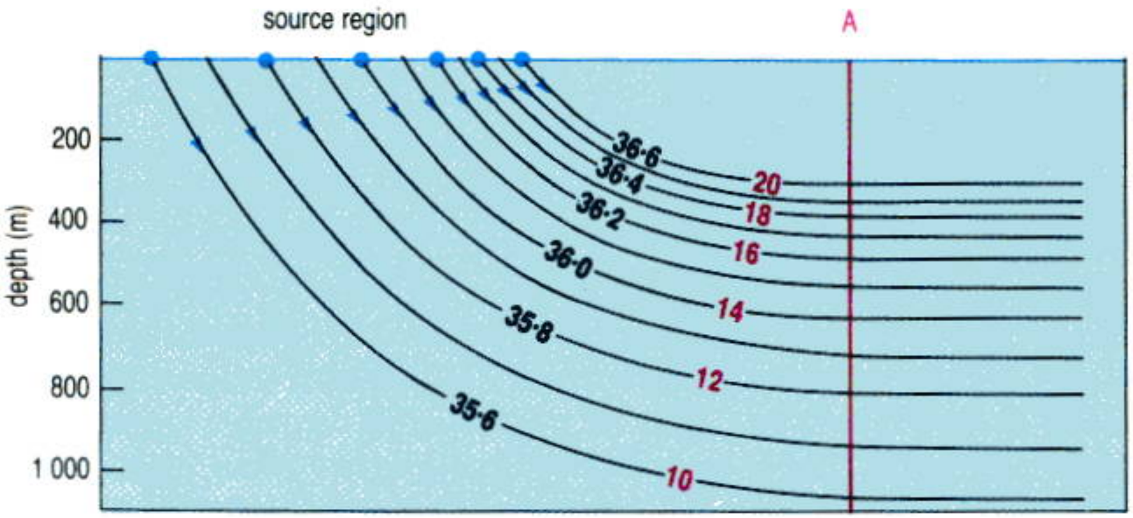


Watermass Analysis

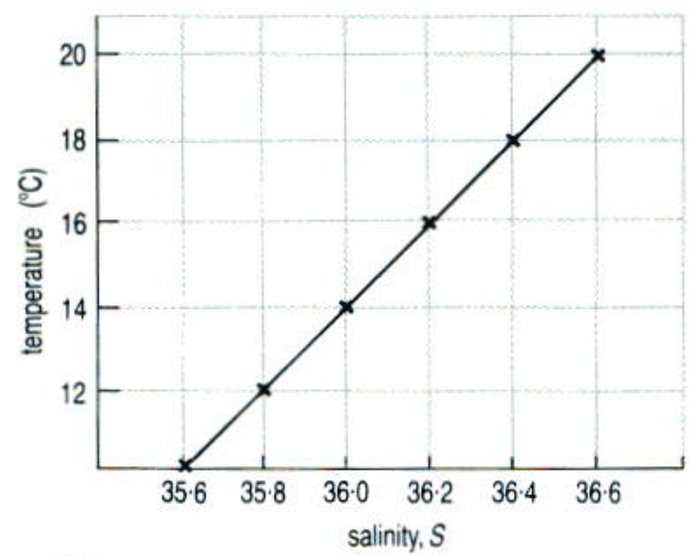
In traditional oceanography, Θ-S profiles were observed in the deep ocean. From this, one would like to:

- Understand where deep water originated
- Understand something about the flow to the point of observation
- And thus be able to trace out the ocean structure and flow from sparse data

If formation only (no transformation / mixing) – a vertical cast shows stratification

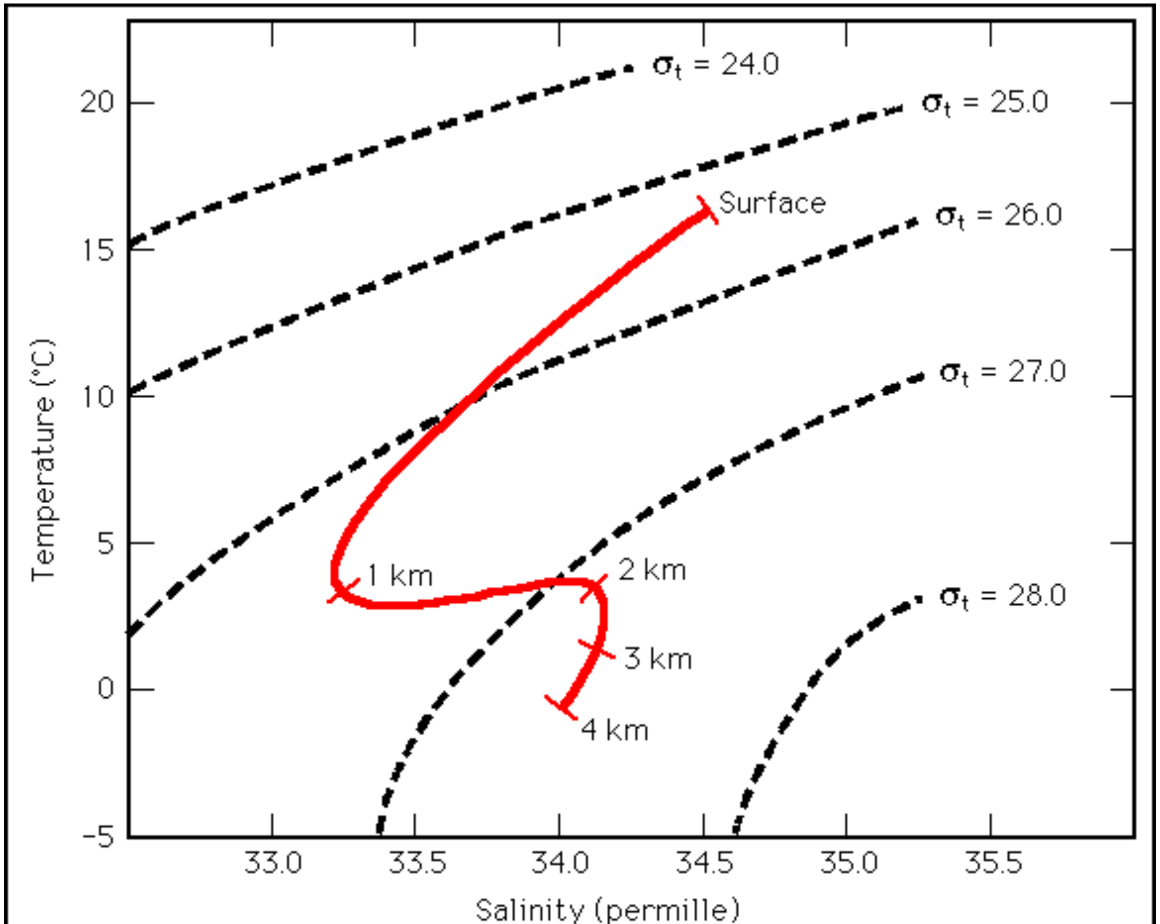


(a)



(b)

In reality, looks more like this:



T-S diagram (After Stowe, 1996)

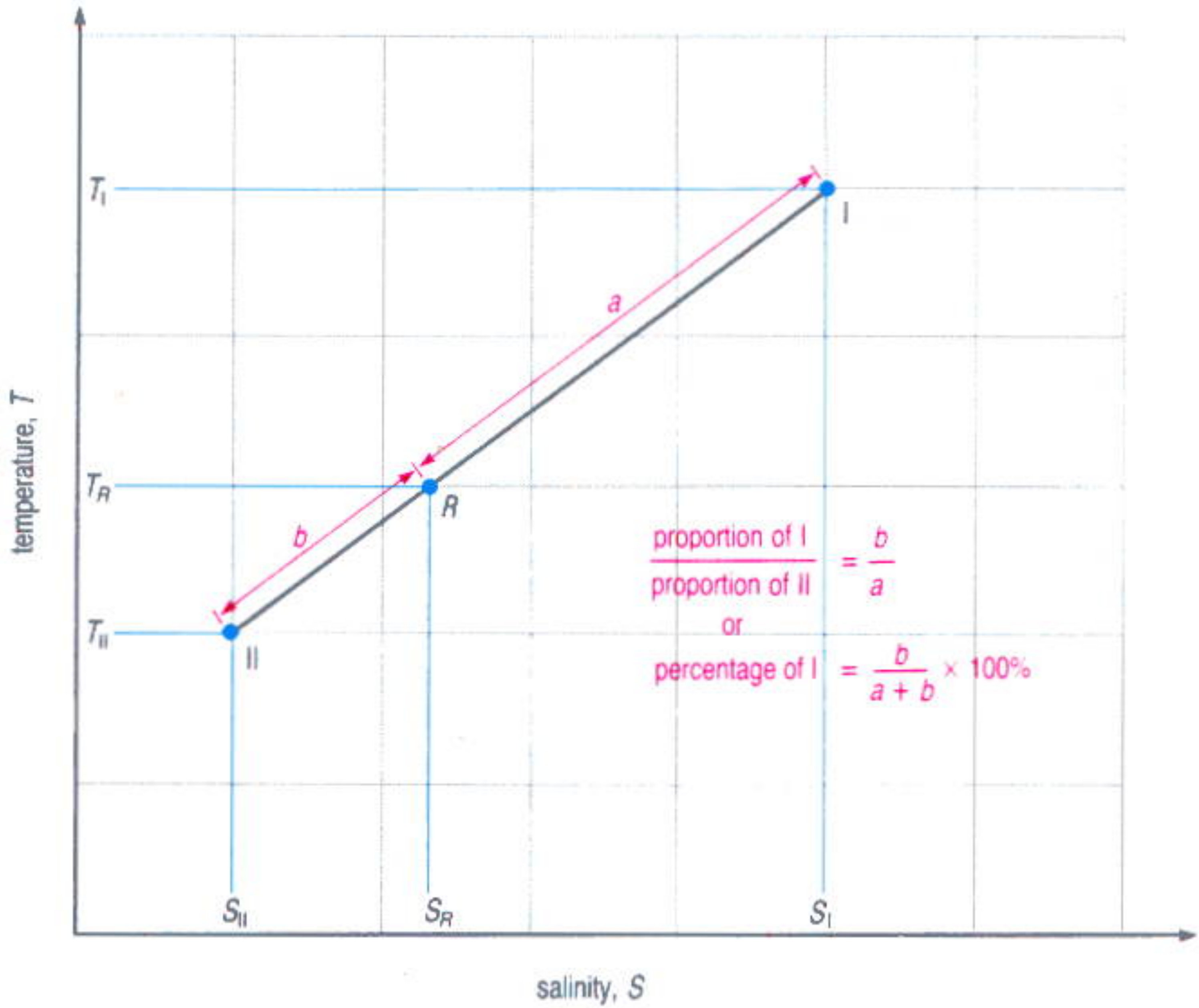
Θ-S diagrams basics

- Since Θ-S are characteristic of the water masses where they are formed, these properties can be used to determine where observed masses originated

- AND Mixing of water types results in predictable changes in density
  - Densification because density contours are non-linear

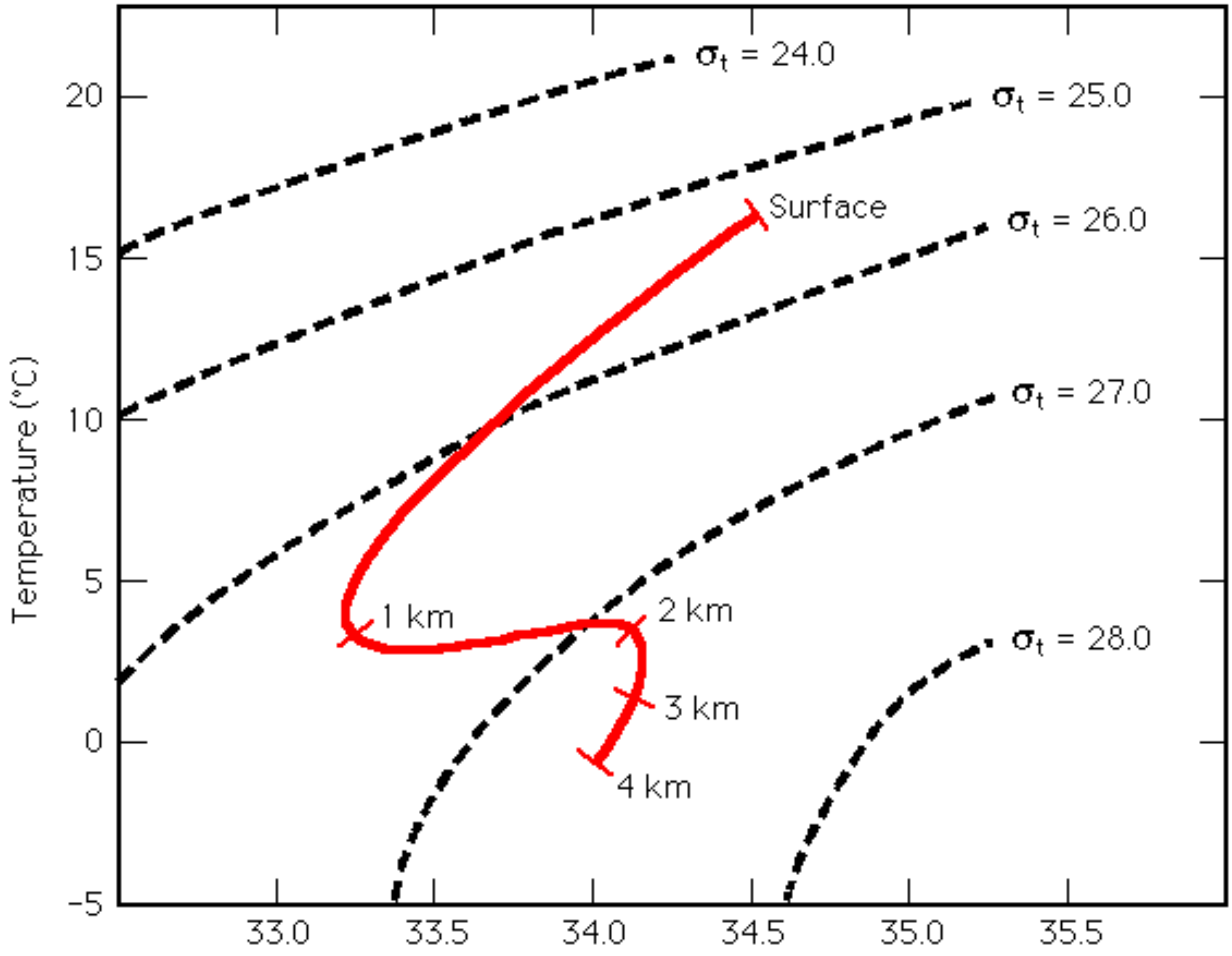
Mixing of 2 watermasses

A straight line on Θ-S diagram can be drawn…

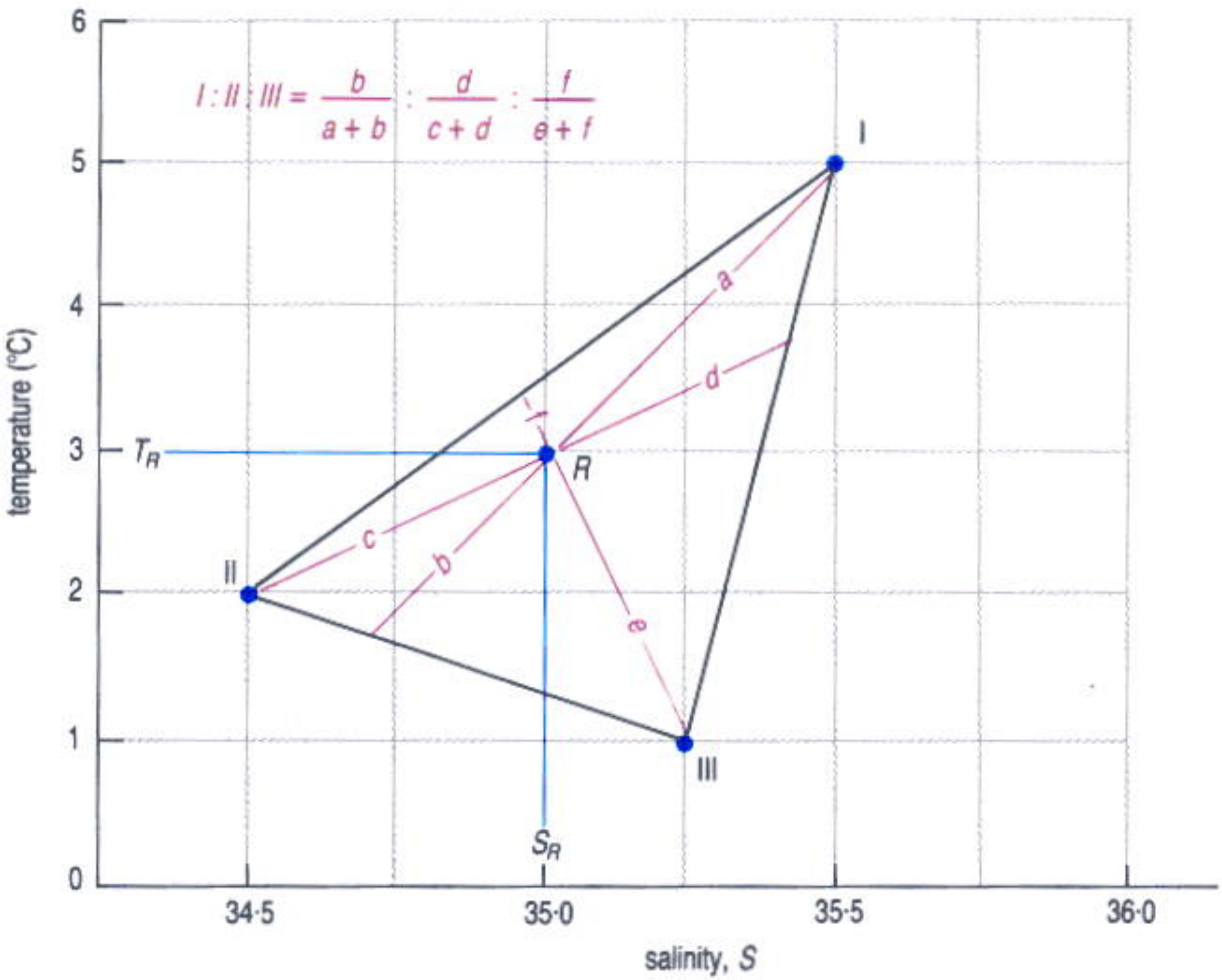


temperature, T
T_I
T_R
T_{II}
I
R
II
a
b
$\frac{\text{proportion of I}}{\text{proportion of II}} = \frac{b}{a}$
or
$\text{percentage of I} = \frac{b}{a+b} \times 100\%$
S_{II}
S_R
S_I
salinity, S

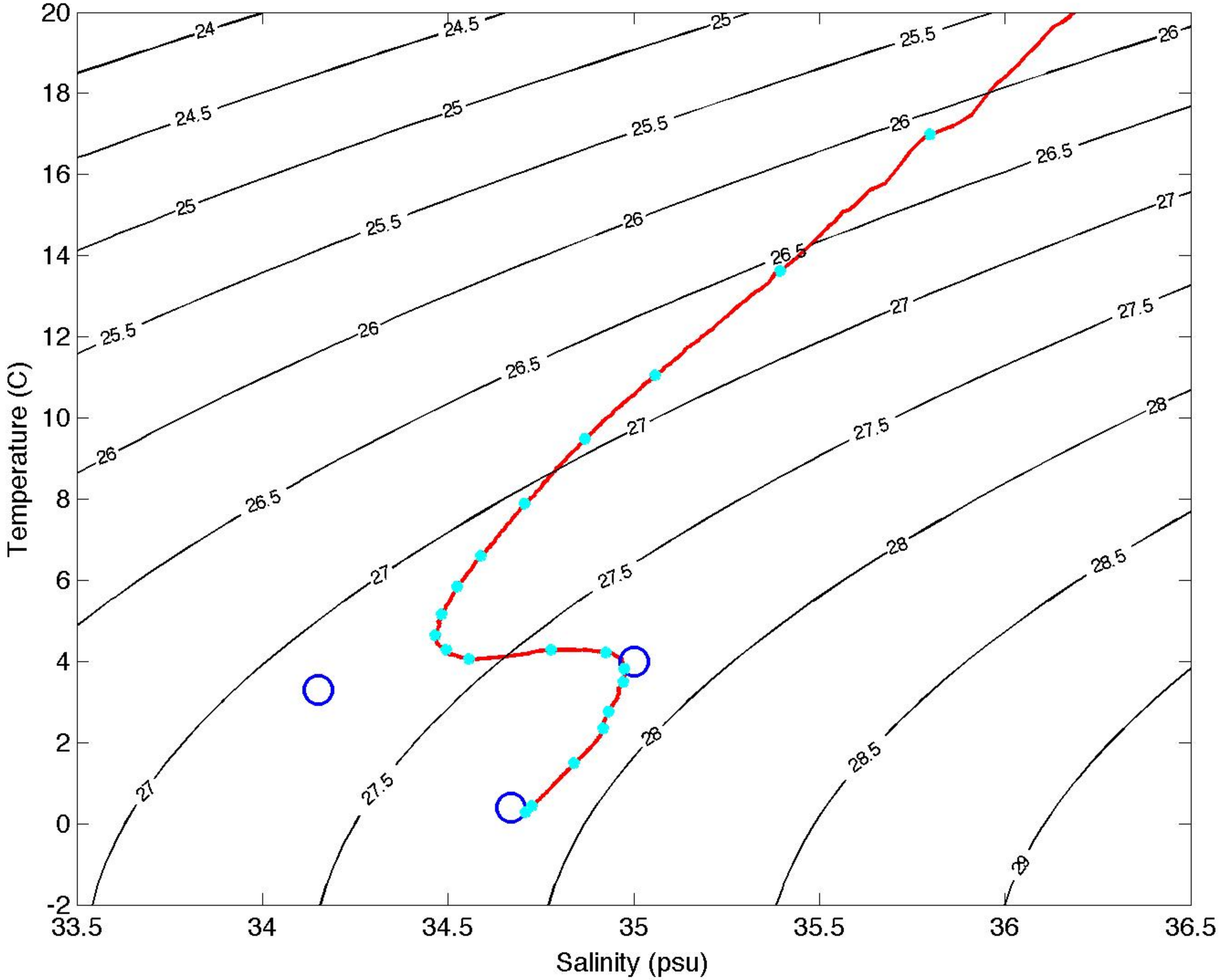
How about 3 end members?



T-S diagram (After Stowe, 1996)

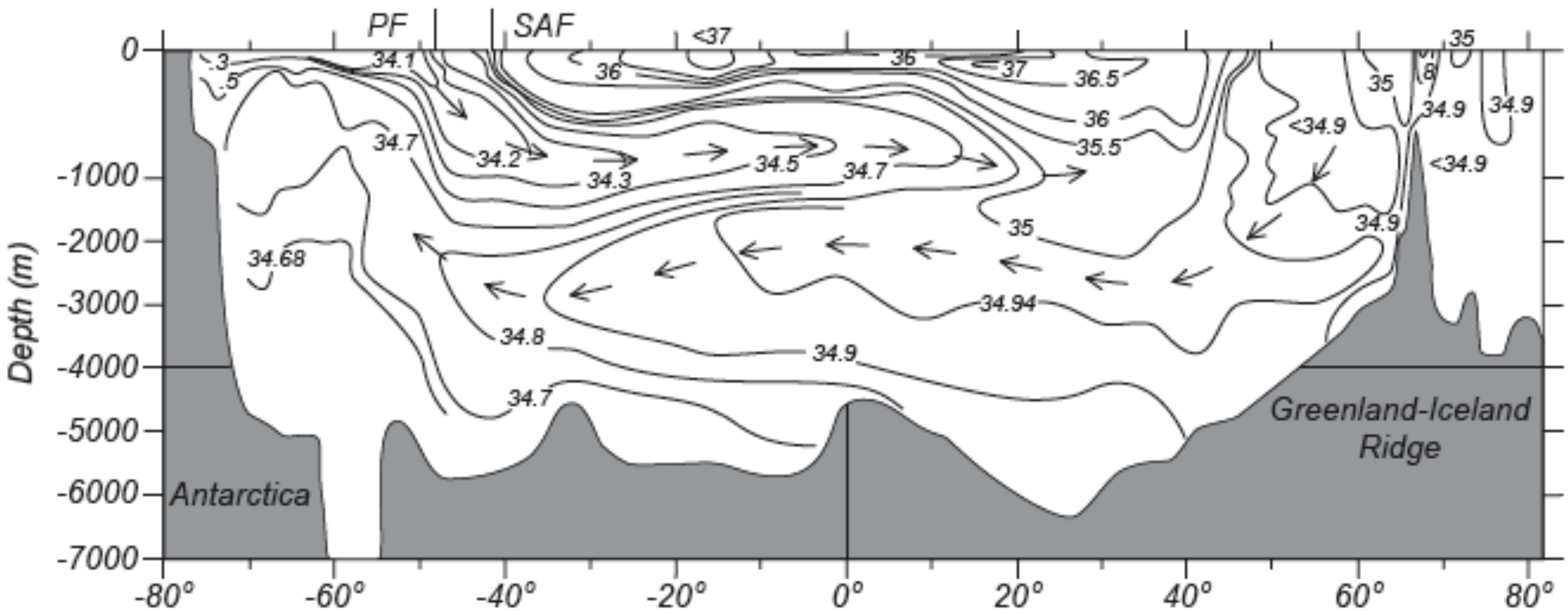


$I : II : III = \frac{b}{a+b} : \frac{d}{c+d} : \frac{f}{e+f}$
I
II
III
R
T_R
S_R
a
b
c
d
e
f
temperature (°C)
salinity, S
0
1
2
3
4
5
6
34·5
35·0
35·5
36·0



Temperature (C)
Salinity (psu)
20
18
16
14
12
10
8
6
4
2
0
-2
33.5
34
34.5
35
35.5
36
36.5

Watermass Analysis - useful to idenitfy waters of common origin. But not dynamically complete



- “implied” flow along tongue of core
- In fact, motion is along a Deep Western Boundary Current and then spreads into interior

How does interior flow move?

On long timescales, there is a steady source from high latitudes, and no concentrated loss

Thus, the deep ocean level must rise / upwell slowly everywhere

Away from surface and boundaries, ocean is very geostrophic

$$fv = \frac{1}{\rho}\frac{\partial p}{\partial x}$$

$$fu = -\frac{1}{\rho}\frac{\partial p}{\partial y}$$

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

To derive Sverdrup flow, (1) relax the horizontally non-divergent requirement of geostrophy and (2) $df/dy \neq 0$

To derive Sverdrup flow, (1) relax the horizontally non-divergent requirement of geostrophy and (2) df/dy $\neq 0$

$$\beta v_g = f \frac{\partial w_a}{\partial z}$$

Now integrate vertically

$$\beta V = f w_a$$

Rotating Tank Experiment: Thermohaline Circulation

http://www-paoc.mit.edu/labweb/lab14/gfdxiv.htm

Schematic of Abyssal Circulation

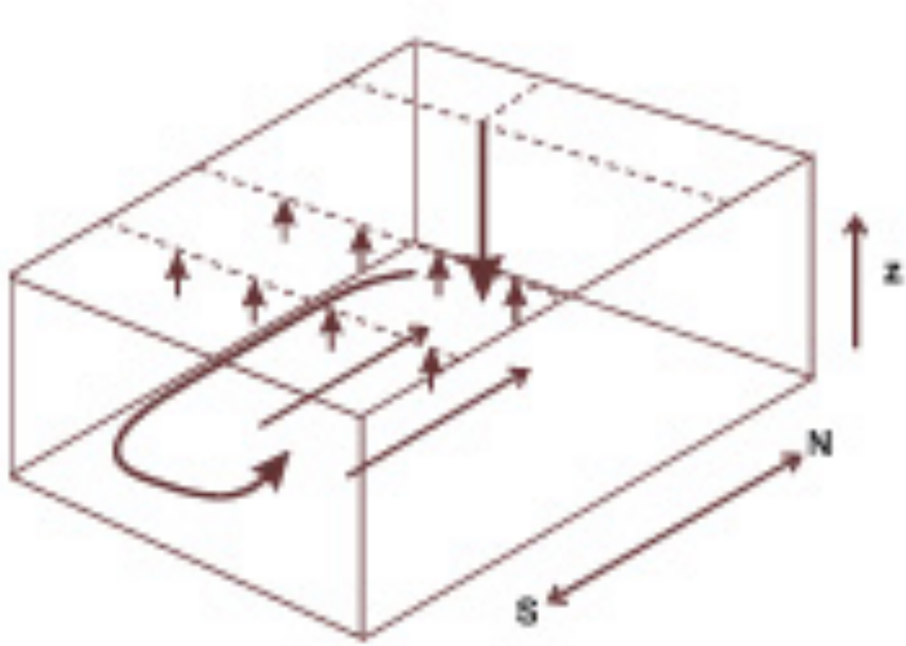
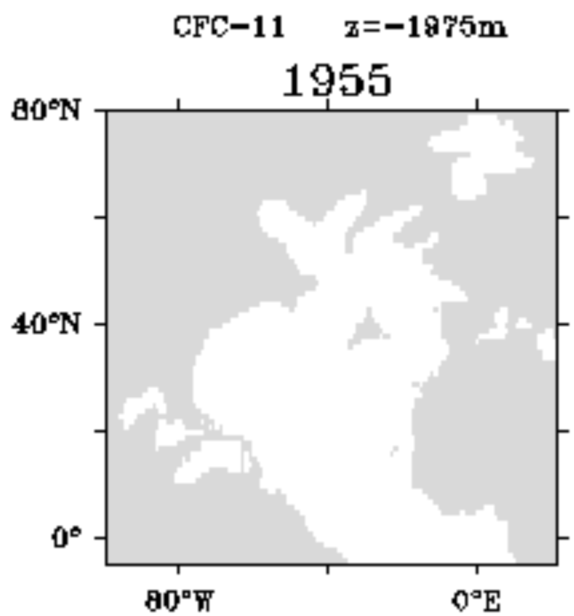


Figure 11.9, Marshall and Plumb, 2003

CFC-11 Invading the North Atlantic Ocean at 2000m in an ocean model



http://puddle.mit.edu/~mick/cfcall.html

Smethie et al. 2000 - CFCs

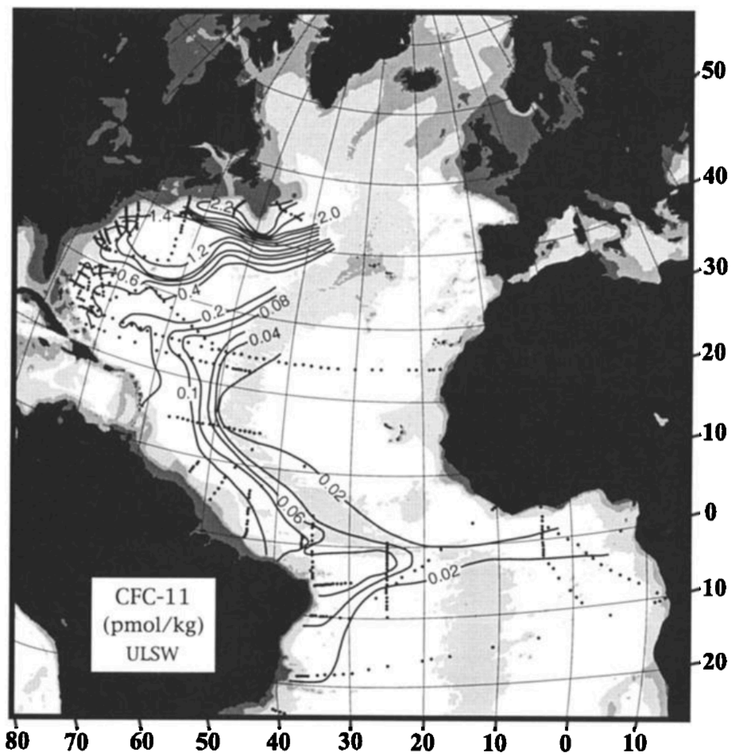
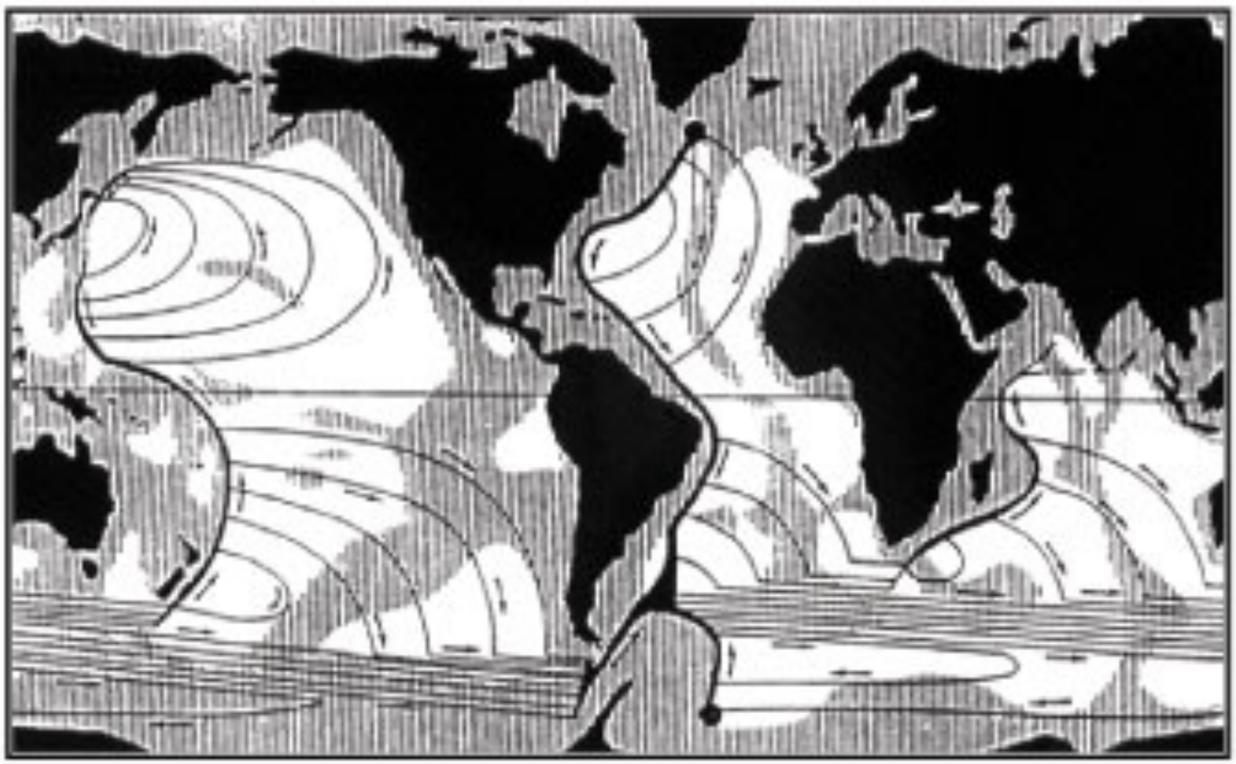
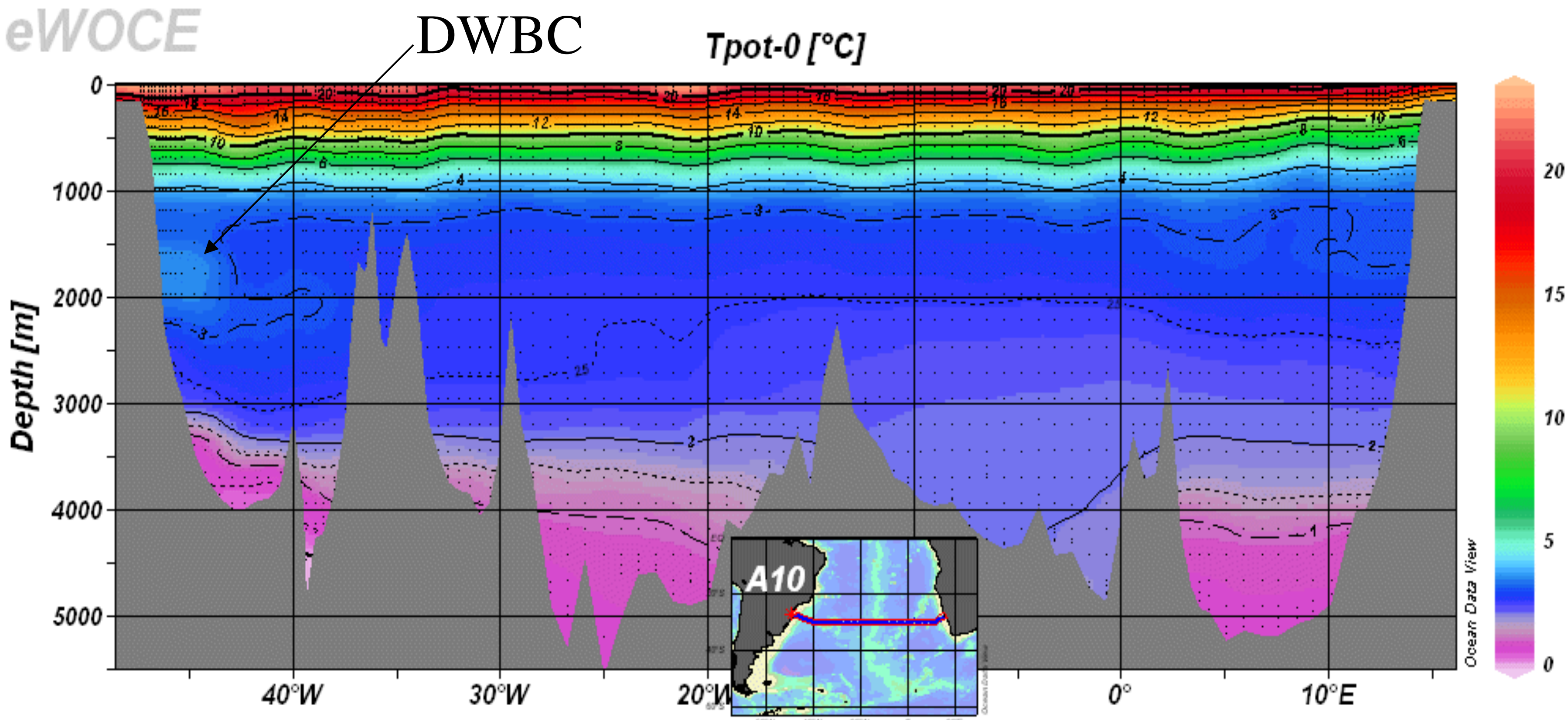


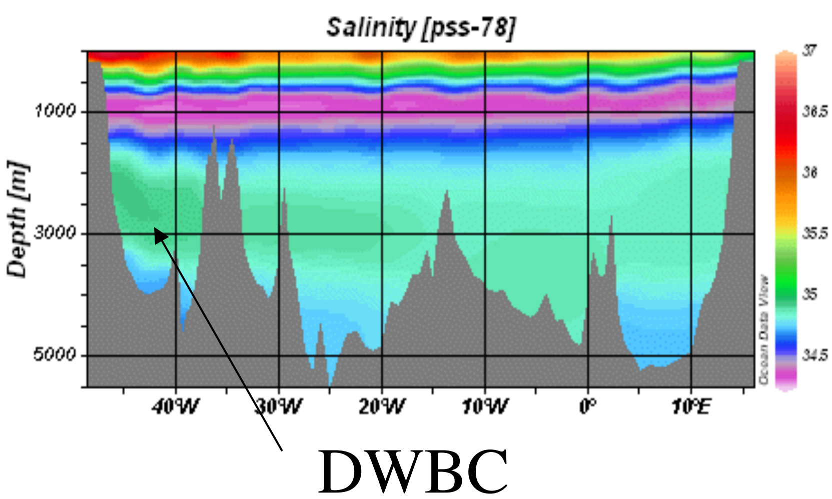
Figure 5a. Lateral map of the maximum CFC-11 concentration in Upper Labrador Sea Water (ULSW). See Table 1 for data sources. These data have been adjusted to a common date of 1990 as described in the text.

Deep Western Boundary Currents



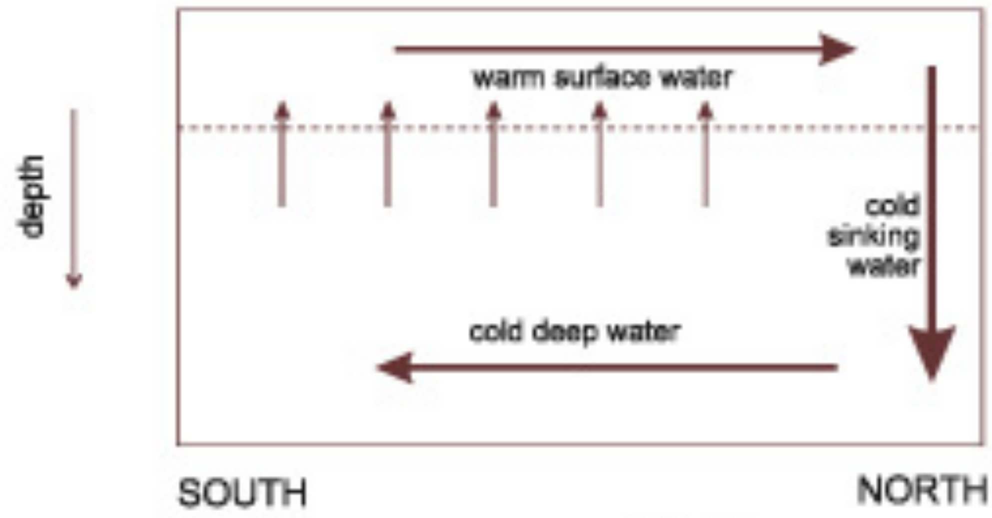
Stommel, 1958



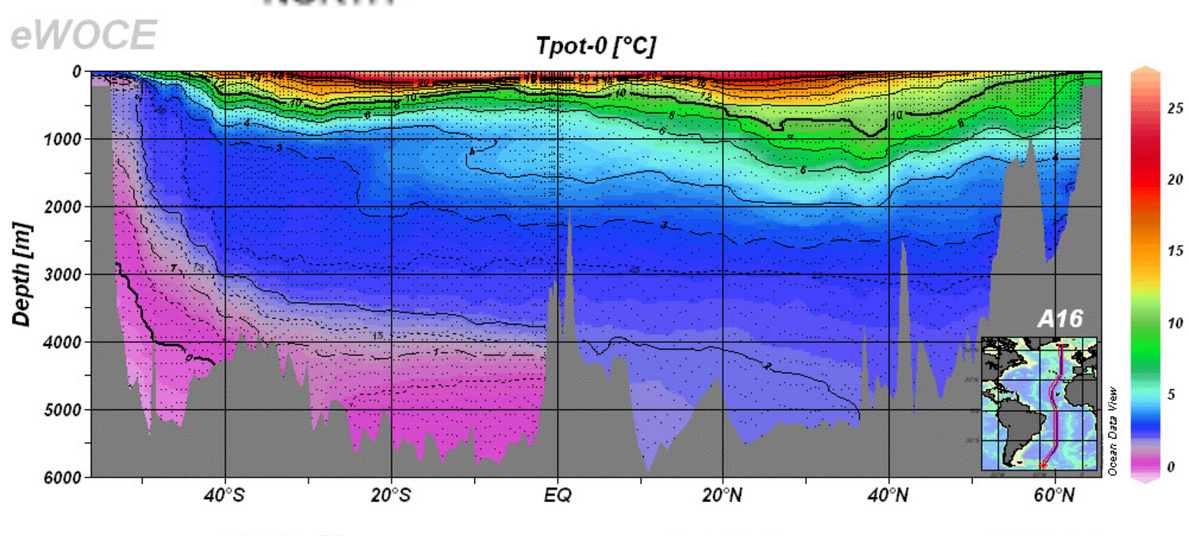


Observations of the Deep western boundary current in the Atlantic

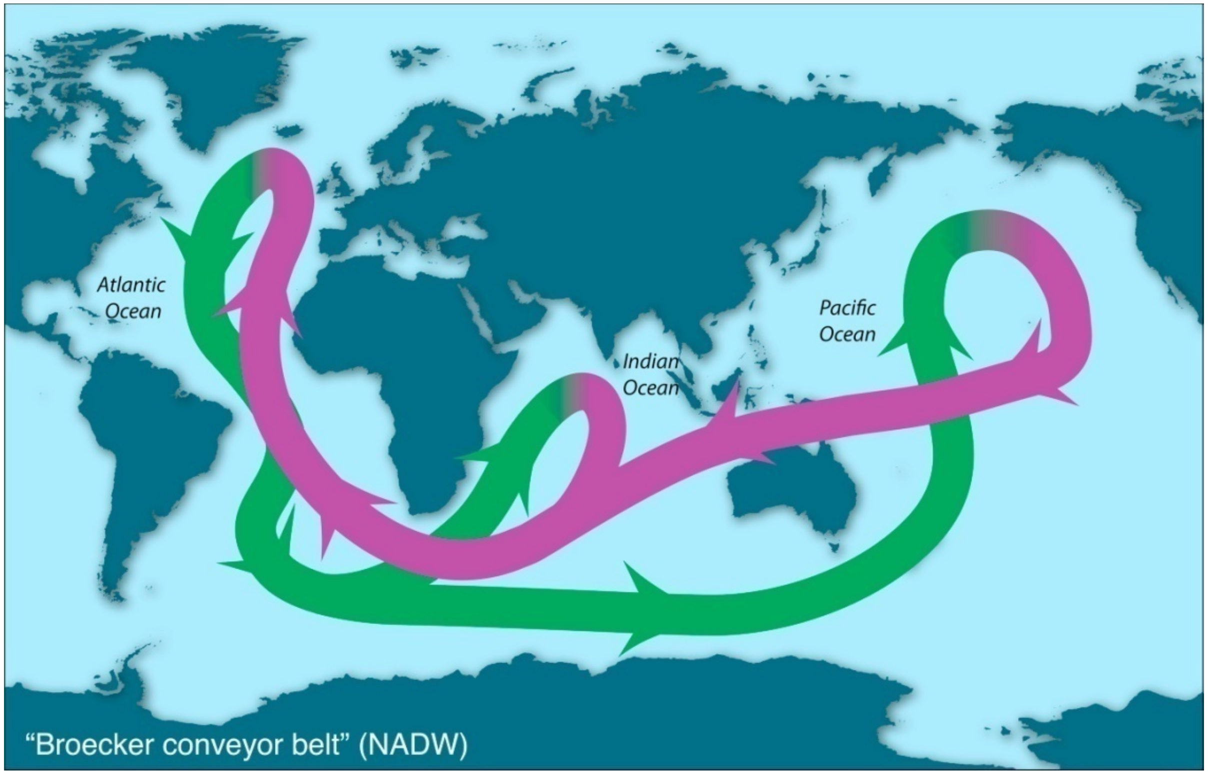
Global zonal average of THC



Figure



Thermohaline, simplified, 1980's



Simplified global NADW cell, which retains sinking only somewhere adjacent to the northern North Atlantic and upwelling only in the Indian and Pacific Oceans. See text for usefulness of, and also issues with, this popularization of the global circulation, which does not include any Southern Ocean processes. *Source: After Broecker (1987).*

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FIGURE 14.10

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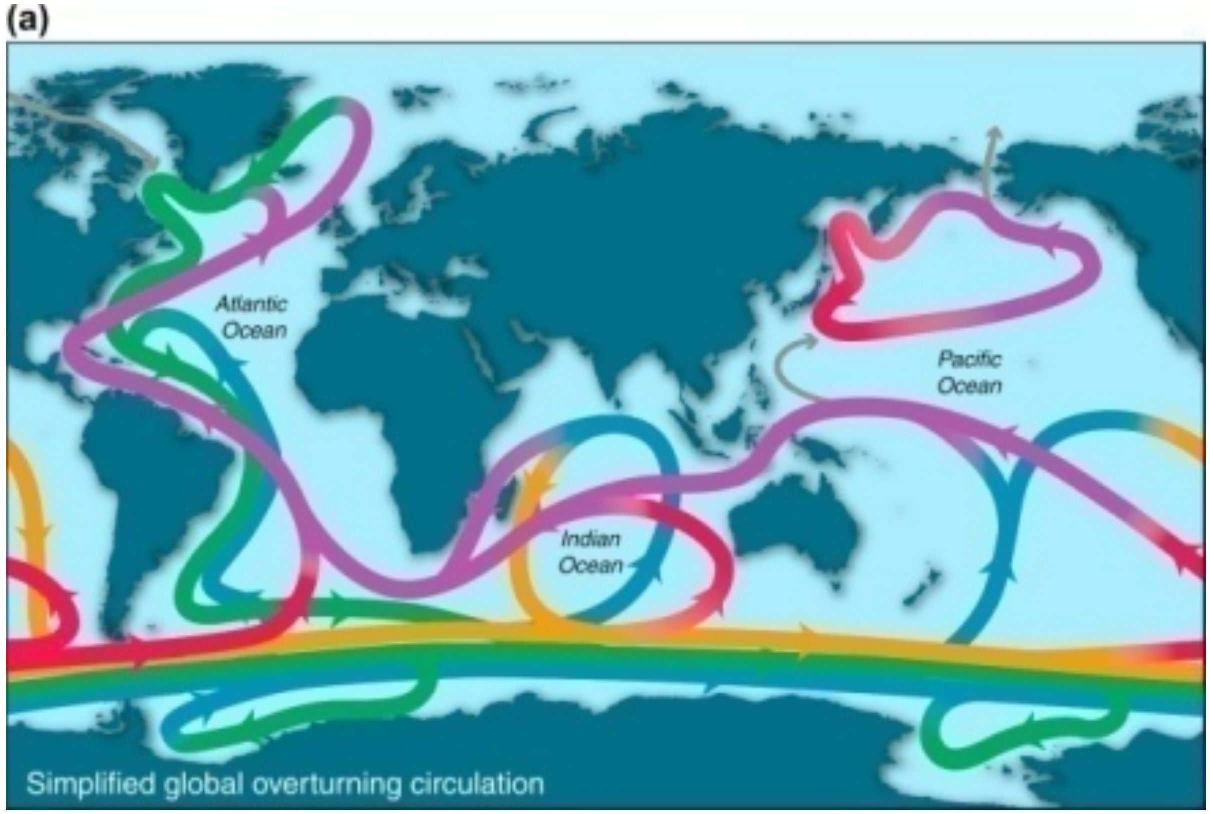

Global overturning circulation schematics. (a) The NADW and AABW global cells and the NPIW cell. (b) Overturn from a Southern Ocean perspective. *Source: After Gordon (1991), Schmitz (1996b), and Lumpkin and Speer (2007).* (c) Two-dimensional schematic of the interconnected NADW, ID
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FIGURE 14.11a

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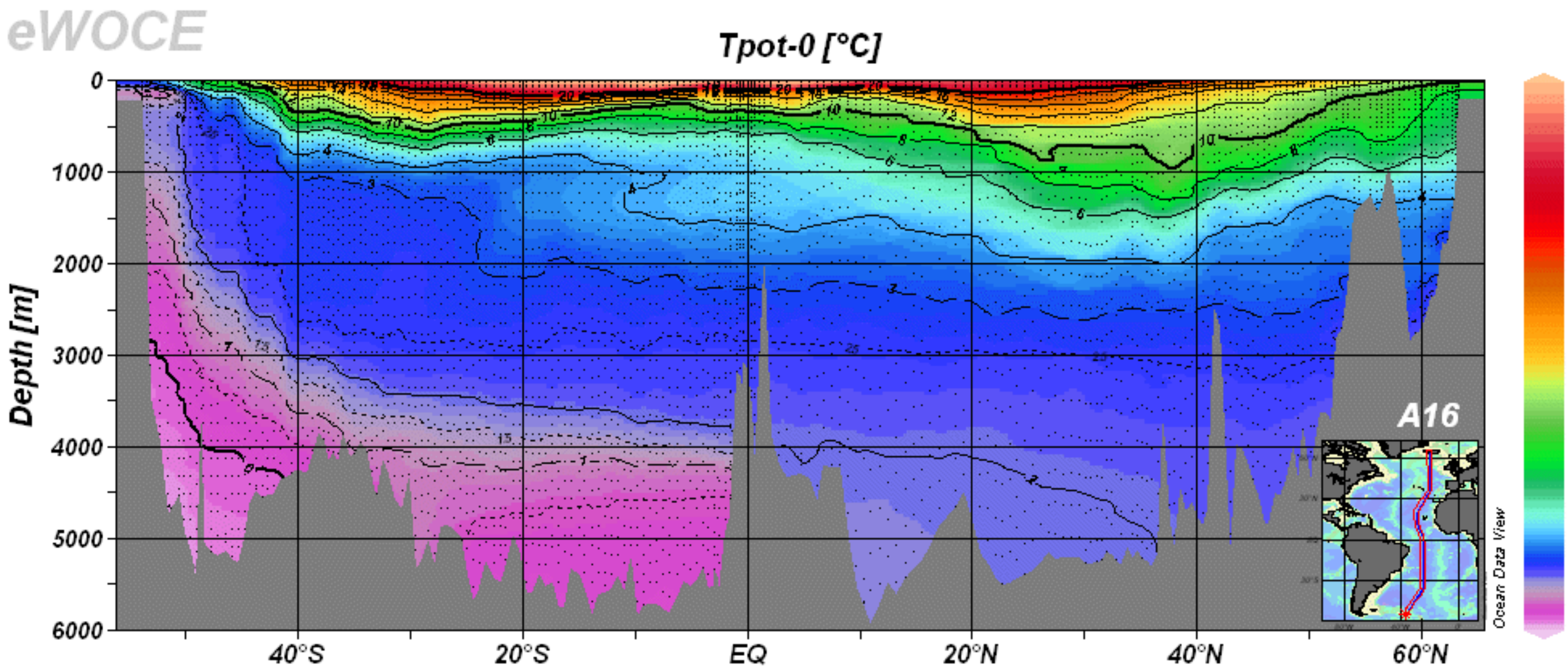
Mixing?

- Mixing between these water masses must occur
- Also, mixing needed to return water supplied “down” during deep convection
  – Turbulence
    - But little energy to drive at depth
    - Evidence that much larger near topography, where tidal energy dissipated and internal waves break (Ledwell et al. – Paper 5)
  – Salt fingering
    - Because molecular diffusion of heat > salt
- Slow, but critical to the large scale overturning

How large is the vertical mixing?

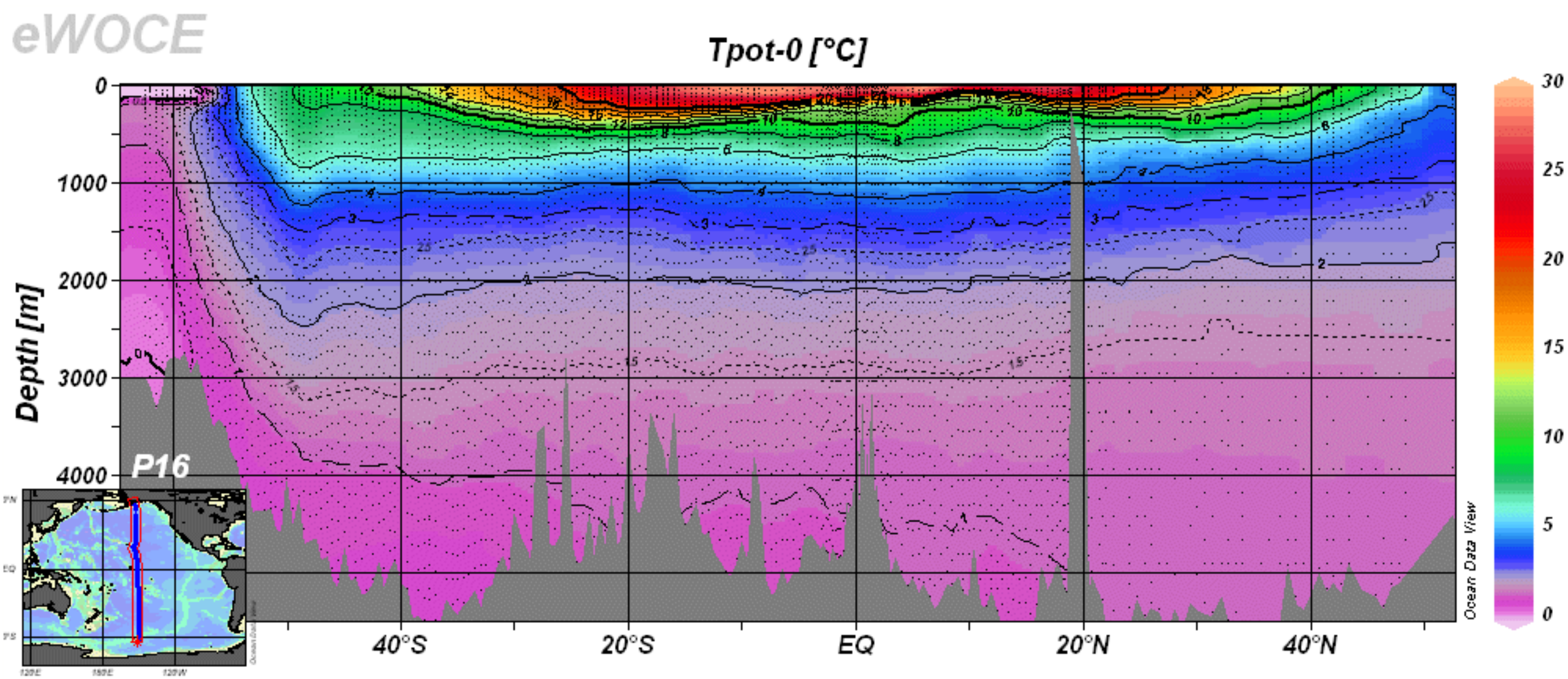
Munk's first calculation

Atlantic



eWOCE
Tpot-0 [°C]
Depth [m]
0
1000
2000
3000
4000
5000
6000
40°S
20°S
EQ
20°N
40°N
60°N
A16
Ocean Data View
25
20
15
10
5
0

Pacific


eWOCE
Tpot-0 [°C]
Depth [m]
0
1000
2000
3000
4000
40°S
20°S
EQ
20°N
40°N
P16
30
25
20
15
10
5
0
Ocean Data View

Balance equation

$$\frac{\partial T}{\partial t} + w\frac{\partial T}{\partial z} = \frac{\partial}{\partial z}(A_z\frac{\partial T}{\partial z}) + S$$

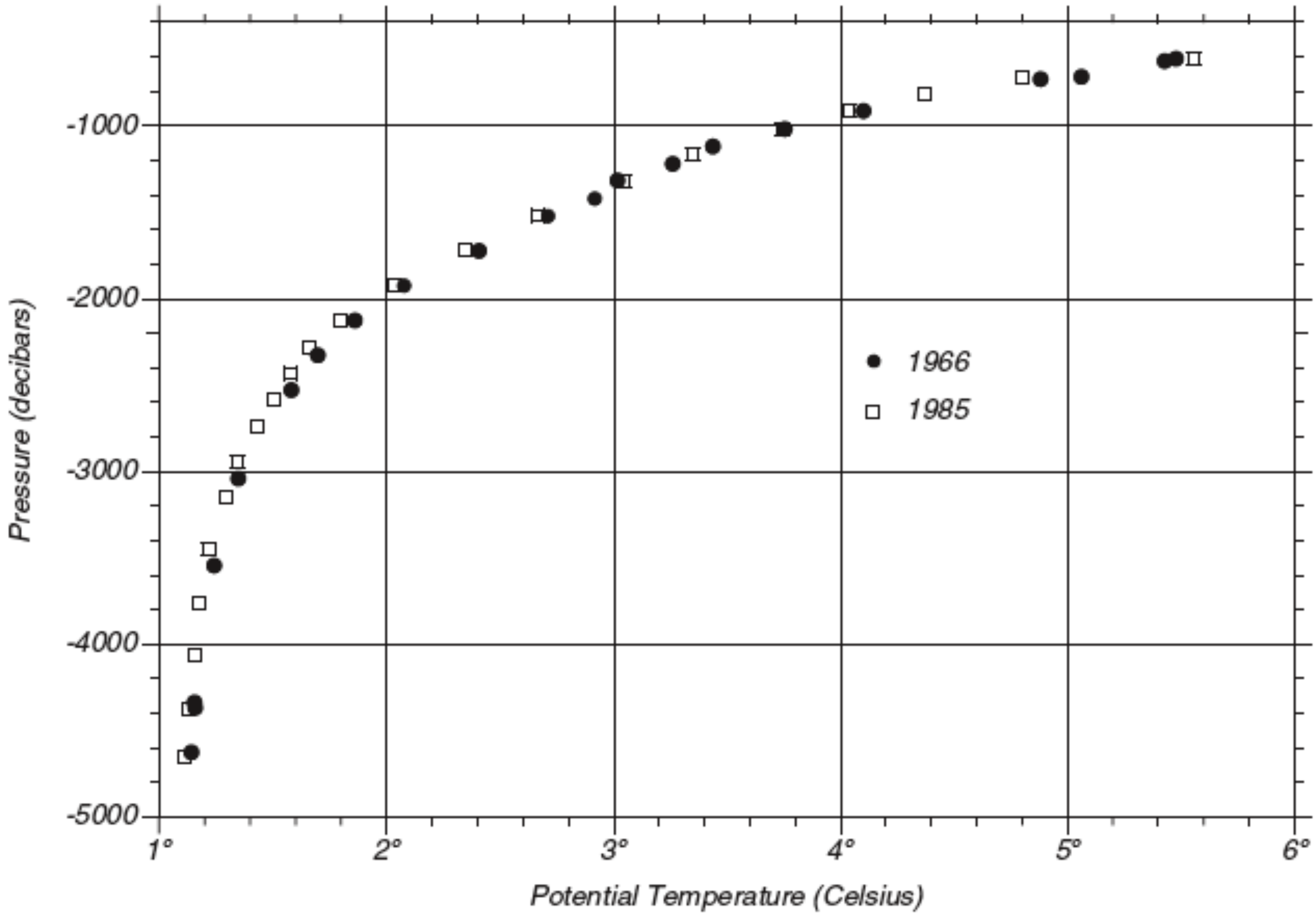


Figure 8.4 Potential temperature measured as a function of depth (pressure) near 24.7°N, 161.4°W in the central North Pacific by the *Yaquina* in 1966 (•), and by the *Thompson* in 1985 (□). Data from *Atlas of Ocean Sections* produced by Swift, Rhines, and Schlitzer.

In steady state, with no source (i.e. away from surface)

$$w\frac{\partial T}{\partial z} = \frac{\partial}{\partial z}(A_z\frac{\partial T}{\partial z})$$

$$T = T_o e^{-z/H}$$

$$H = \frac{A_z}{w}$$

Munk (1966)

- H from data (930m, observed thermocline depth)
- Using ^{14}C, w = 1.2 cm/d = 4.4 m/yr
- RESULT -> $A_z = 1.3 \times 10^{-4}$ m^2/s

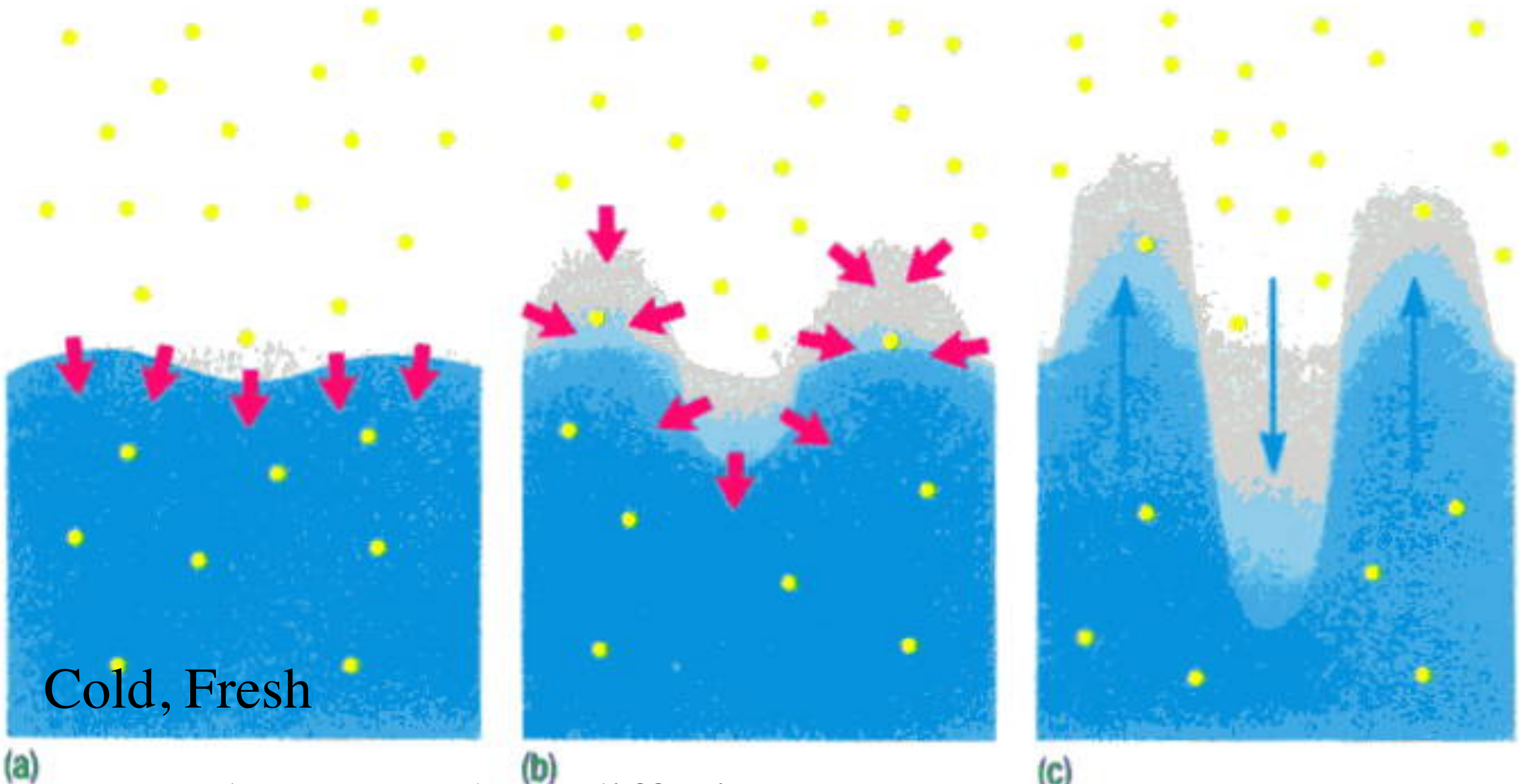
- If consider area of whole ocean, suggests 25-30 Sv (m^3/s) of upwelling, not dissimilar to deep water formation rates
- Standard model for 30 years -- this is a globally constant value

Salt Fingering

Warm, salty

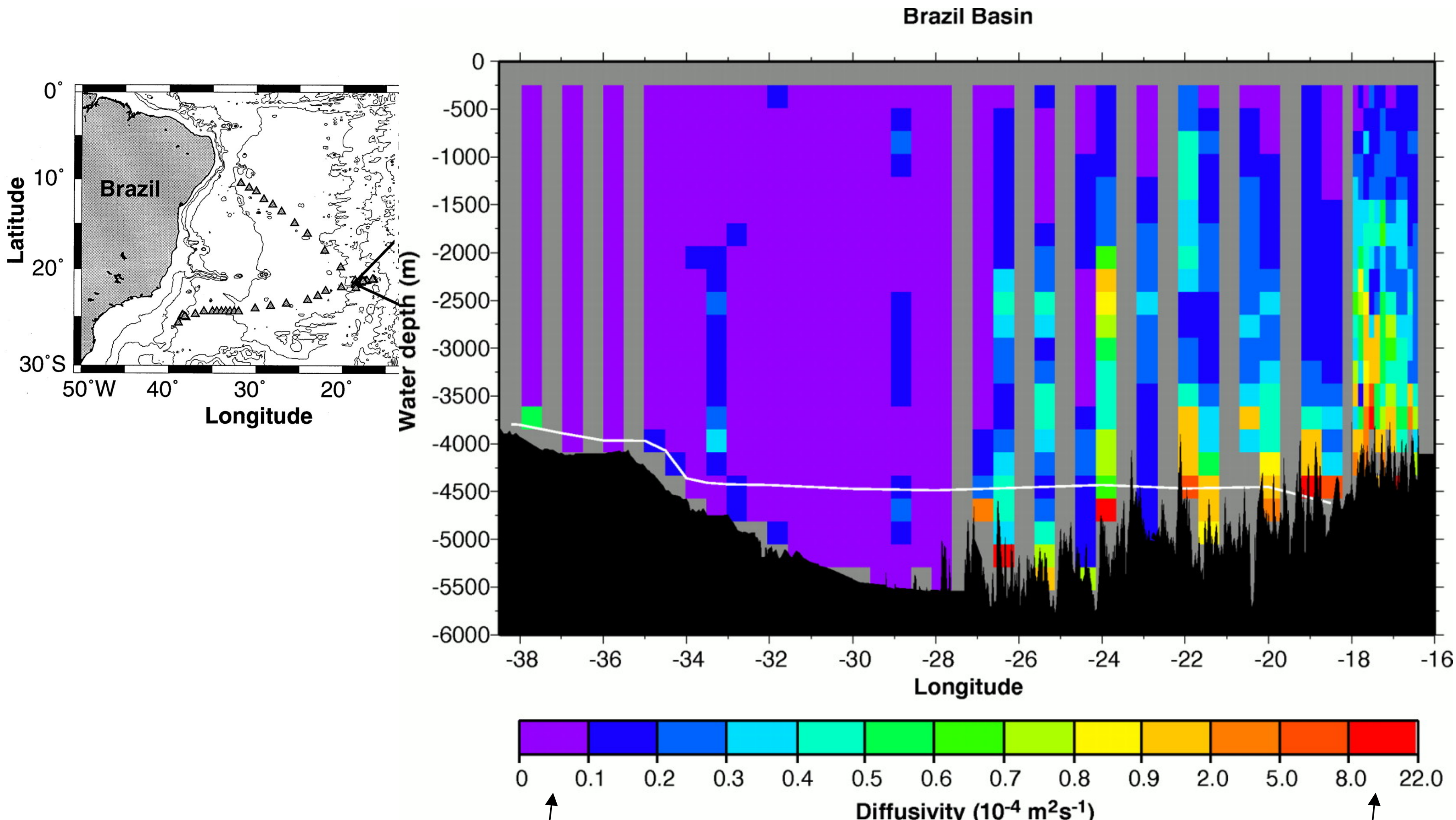
Diffusion heat > diffusion salt

“Fingers” cool, salty; warmer, fresh -> mix



Red arrows = heat diffusion

But we’ve learned in recent years that vertical mixing is quite inhomogeneous



Interior: $A_z \sim 10^{-5}$ m^2/s

Near bathymetry: $A_z \sim 10^{-3}$ m^2/s

Polzin et al. 1997

Parameterizing Mixing as an Eddy Viscosity

- Viscosity is the tendency of a fluid to resist shear

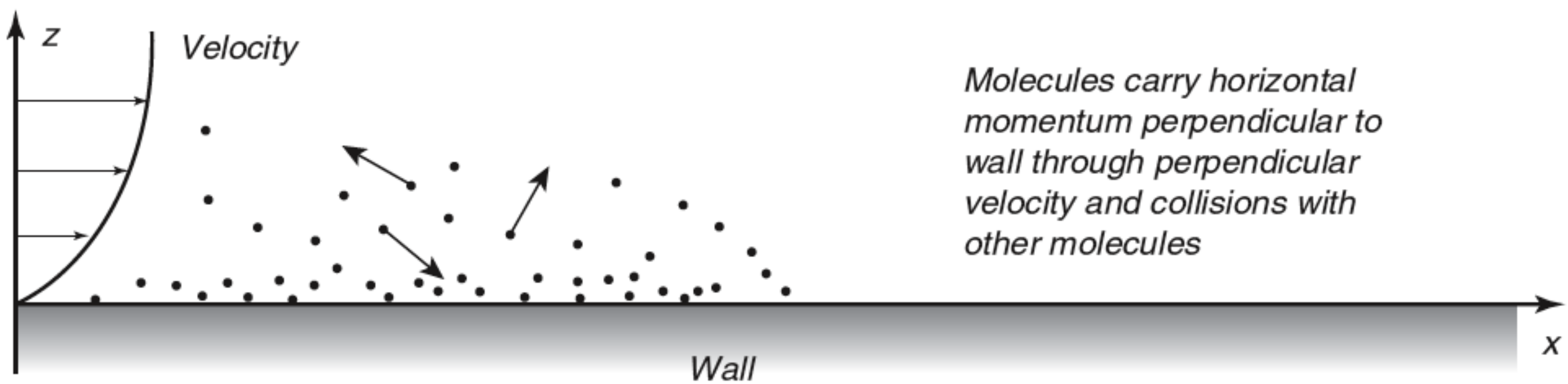


Figure 8.1 Molecules colliding with the wall and with each other transfer momentum from the fluid to the wall, slowing the fluid velocity.

Parameterizing Mixing as an Eddy Viscosity

- Viscosity is the tendency of a fluid to resist shear
- <u>Molecular viscosity</u> is only important within a few mm of boundary because molecules travel only a few μm and thus can only transfer momentum that far

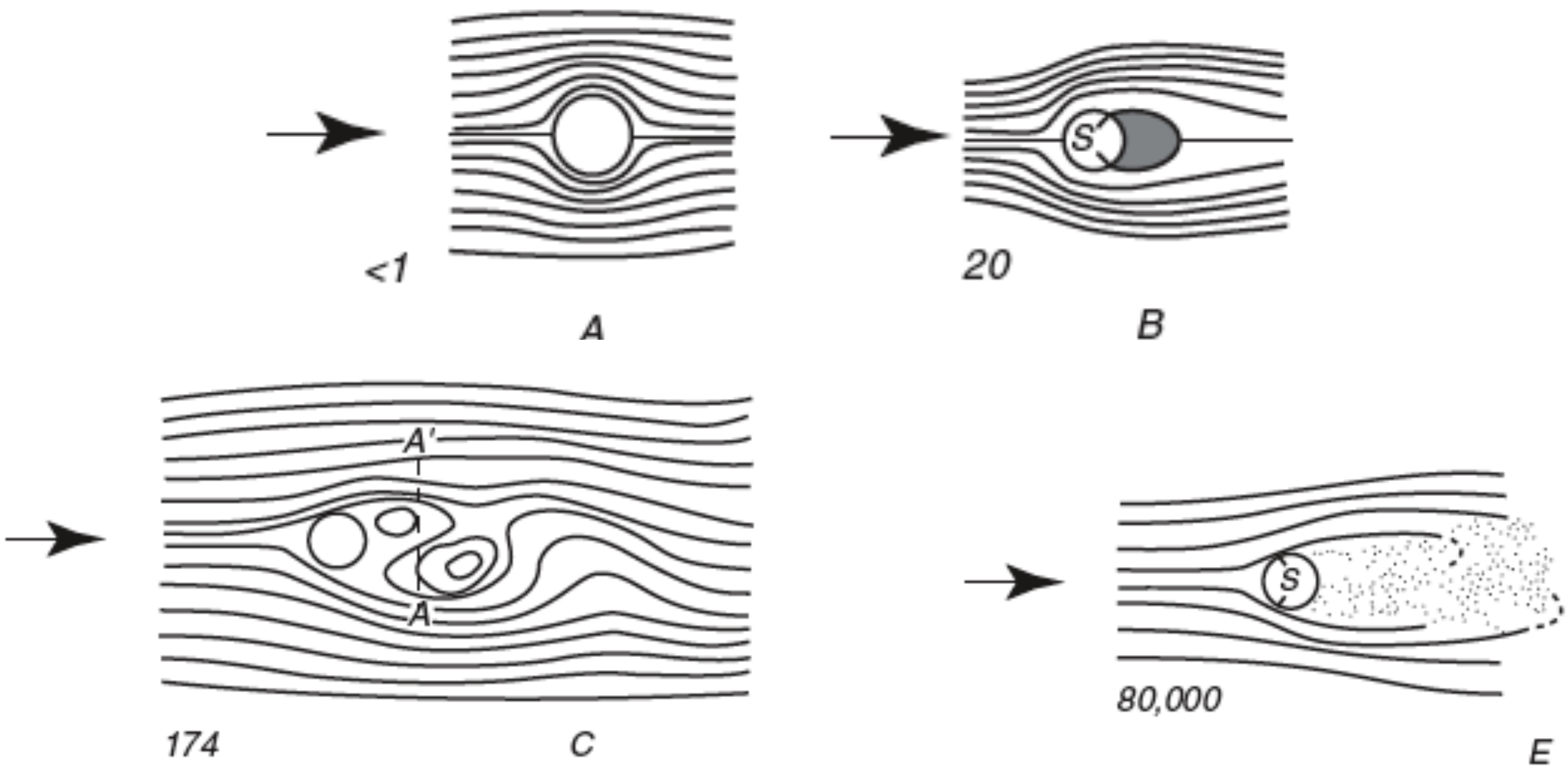
Navier Stokes

$$\frac{D\vec{u}}{Dt} = -\frac{1}{\rho}\nabla p - 2\vec{\Omega} \times \vec{u} + \vec{g} + \vec{F}$$

Reynolds number

$$R_e = \frac{UL}{\nu}$$

Flow at different Reynolds



Open ocean, large scale scaling, what is Re?

Do nonlinear or viscous terms dominate?

- $U \sim 10^{-1}$ m/s
- $L \sim 10^{6}$ m
- $\nu = 10^{-6}$ m^2/s

Parameterizing Mixing as an Eddy Viscosity

- Viscosity is the tendency of a fluid to resist shear
- <u>Molecular viscosity</u> is only important within a few mm of boundary because molecules travel only a few um and thus can only transfer momentum that far
- <u>Eddy viscosity</u> (= the tendency of turbulence to transfer momentum) can be derived by analogy to molecular viscosity
  - This is much more important to geophysical flows

Eddy mixing in the ocean

How big are A_x, A_y and A_z?

Horizontal mixing

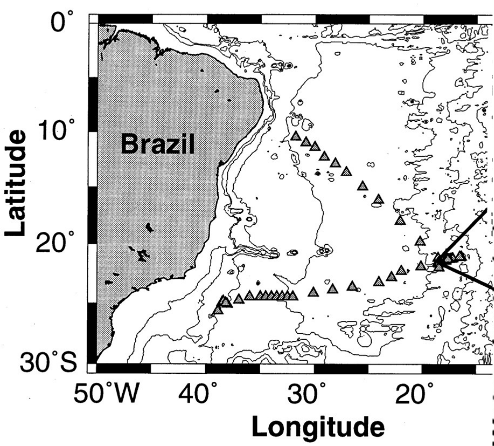
- Along isopycnals, mixing is efficient
  - No need to work against buoyancy force!
- $UL \sim A_x$
  - ι.ε. we can estimate the mixing by eddies from data. This done with observations of tracers (for L) and current (for U)
- For geostrophic eddy ($L \sim 100$km, $U \sim 10^{-2}$ m/s), $A_x \sim 10^3$ m^2/s

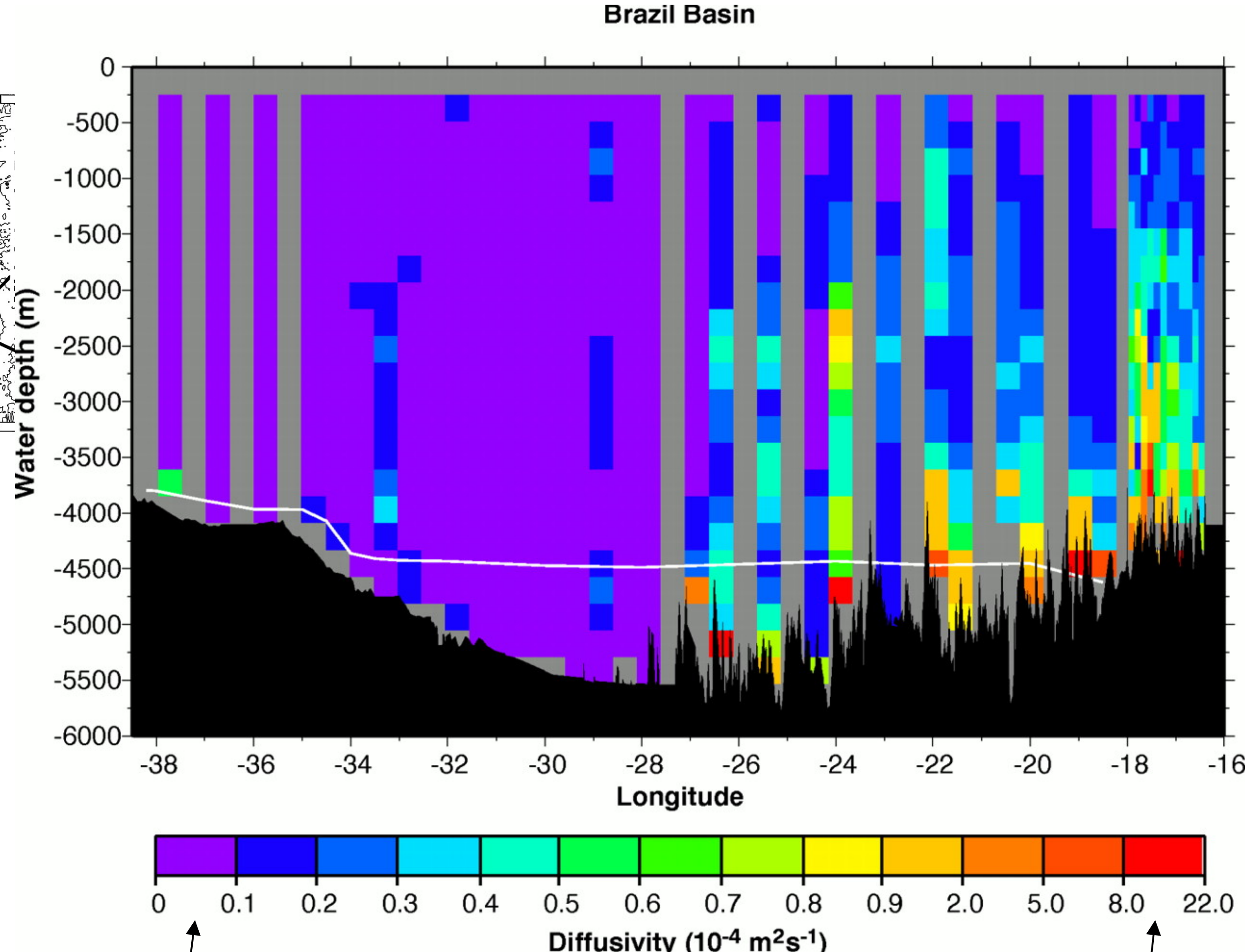
Comparison of eddy mixing

large scale flows

- Horizontal mixing: $A_x \sim 10^3$ m^2/s
  - Along isopycnals

- Vertical mixing: $A_z \sim 10^{-5}$ -10^{-3} m^2/s
  - Weak in interior
  - Stronger with waves and tides over bathymetry





Interior: $A_z \sim 10^{-5}$ m^2/s

Near bathymetry: $A_z \sim 10^{-3}$ m^2/s

Polzin et al. 1997

New paradigm for ocean vertical mixing

- Weak interior mixing
- Strong mixing over bathymetry, driven by deep tides
- Propagation of mixing energy via internal waves