Physical oceanography
Lecture 1
Introduction to the ocean general circulation
Thermodynamics and dynamics

ICTP: The General Circulation of the Atmosphere and Oceans: A Modern Perspective

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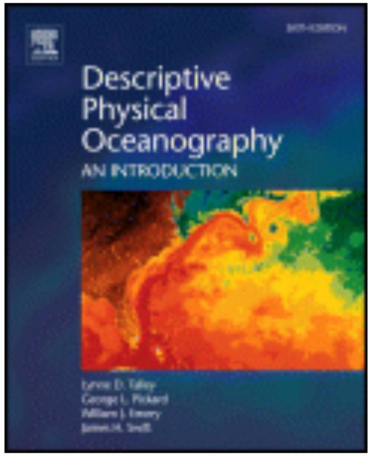
Monday July 11, 2011

Descriptive Physical Oceanography: An Introduction (6th edition)
Talley, Pickard, Emery, Swift (2011) (Elsevier)



Just published in April: hard copy and ebook, extensive web materials, including open access chapters on Dynamics, Climate and the Oceans, adjacent seas including the Mediterranean

Background, in-depth coverage of all topics can be found in DPO

http://www-pord.ucsd.edu/~ltalley/DPO/ has links to all materials

1. Thermodynamics

- Equation of state of seawater (dependence of density on pressure, temperature, salinity)
- External forcings for ocean density
- Adiabatic compression effects: potential temperature and potential density
- Typical distributions of temperature, salinity, density
  in physical space and using density as a vertical coordinate

1. Thermodynamics: Pressure

Pressure is a force per unit area

Units of pressure: dyne/cm^2 and N/m^2

1 Pascal = 1 N/m^2

1 bar = 10^6 dynes/cm^2 = 10^5 N/m^2

approximately the atmospheric pressure at sea level

1 atmosphere = 1000 millibar = 1 bar

1 dbar = 0.1 bar

For motions slower and larger than gravity waves, the ocean is in hydrostatic balance to a very good approximation

Vertical momentum equation reduces to

$$Dw/Dt \approx 0 = -(\Delta p/\Delta z) - \rho g$$

For Δz = 1 meter, density ρ ~1025 kg/m^3, and g = 9.8 m/s^2, we get

$\Delta p = \rho\, g\, \Delta z$ = (1025 kg/m^3)(9.8 m/s^2)(1 m) =

10045 kg/(m s^2) = 0.10045 bar = 1.0045 dbar

1. Thermodynamics: Temperature and heat

Temperature: standard concept

Ocean range: freezing point (about -1.8°C) to about 30°C

(Freezing point is < 0°C because of salt content)

Temperature is changed externally by heating and cooling.

Potential temperature θ: water is slightly compressible. Remove passive effects of compression on temperature.

θ is the temperature a parcel of water has if moved adiabatically (without heat exchanges or mixing) to the sea surface.

Heat: units are Joules

Heat change: units are Joules/sec = Watts

Heat content per unit volume (q in J/m^3):

$dq/dT = \rho\, c_p$ where c_p is specific heat

Typical seawater values: c_p ~3850 J/kg °C; ρ ~ 1025 kg/m^3

1. Thermodynamics: Temperature and heat

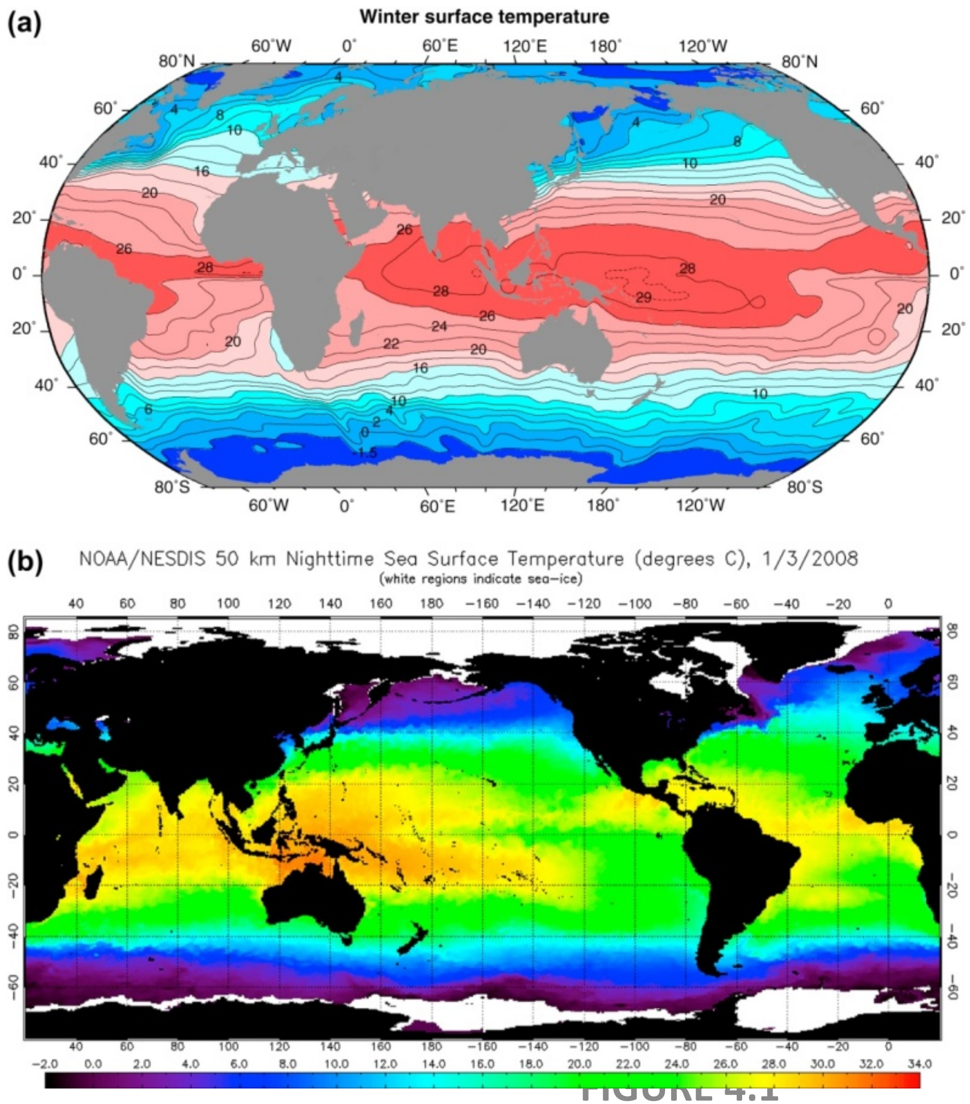
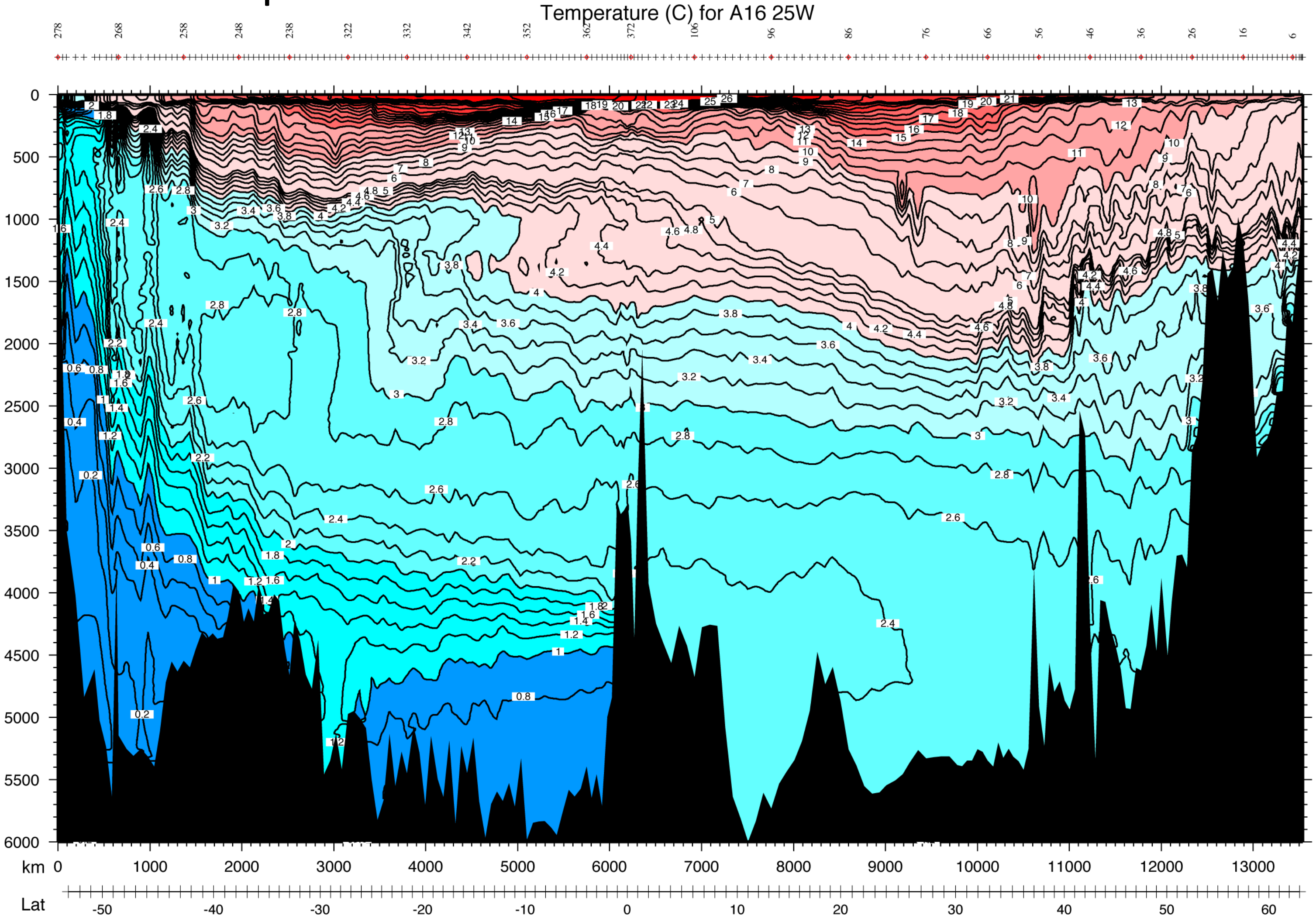


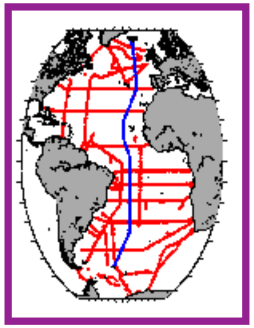
FIGURE 4.1

(a) Surface temperature (°C) of the oceans in winter (January, February, March north of the equator; July, August, September south of the equator) based on averaged (climatological) data from Levitus and Boyer (1994). (b) Satellite infrared sea surface temperature (°C; nighttime only), averaged to 50 km and 1 week, for January 3, 2008. White is sea ice. (See Figure S4.1 from the online supplementary material for this image and an image from July 3, 2008, both in color).

1. Thermodynamics: Temperature and heat

Atlantic temperature section

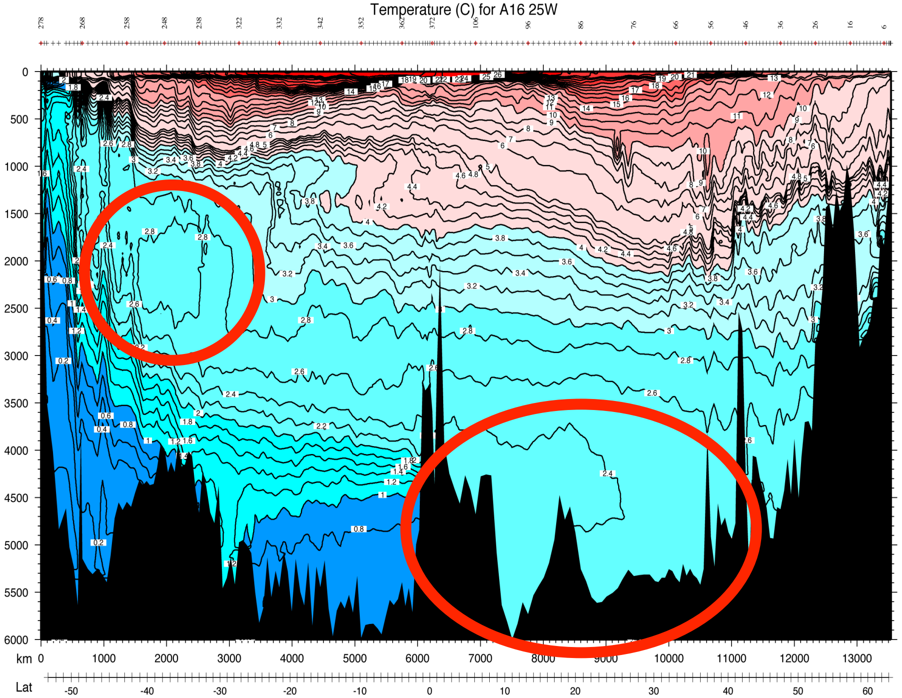




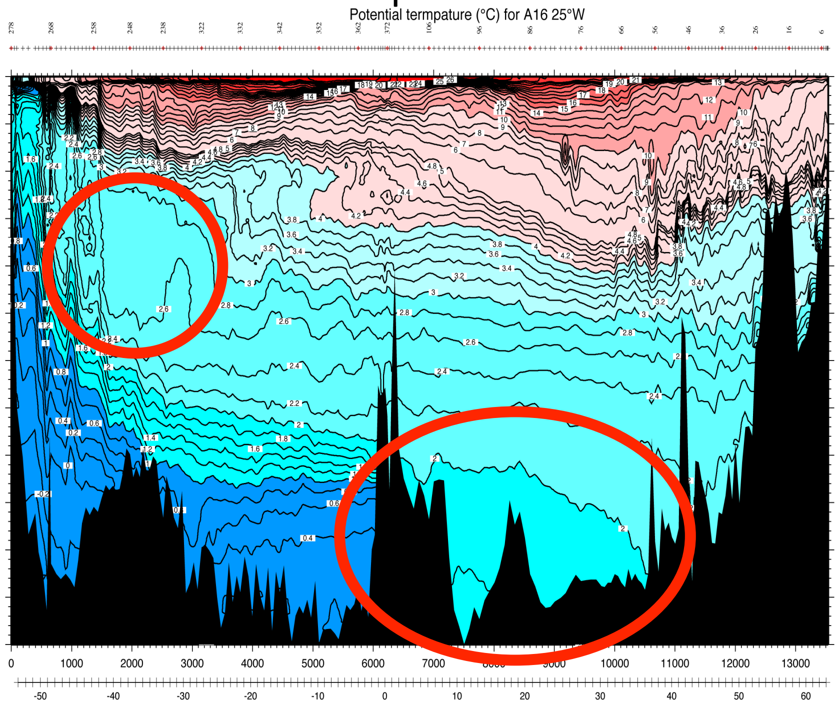
1. Thermodynamics: Temperature and heat

Atlantic temperature and potential temperature sections for contrast

Temperature

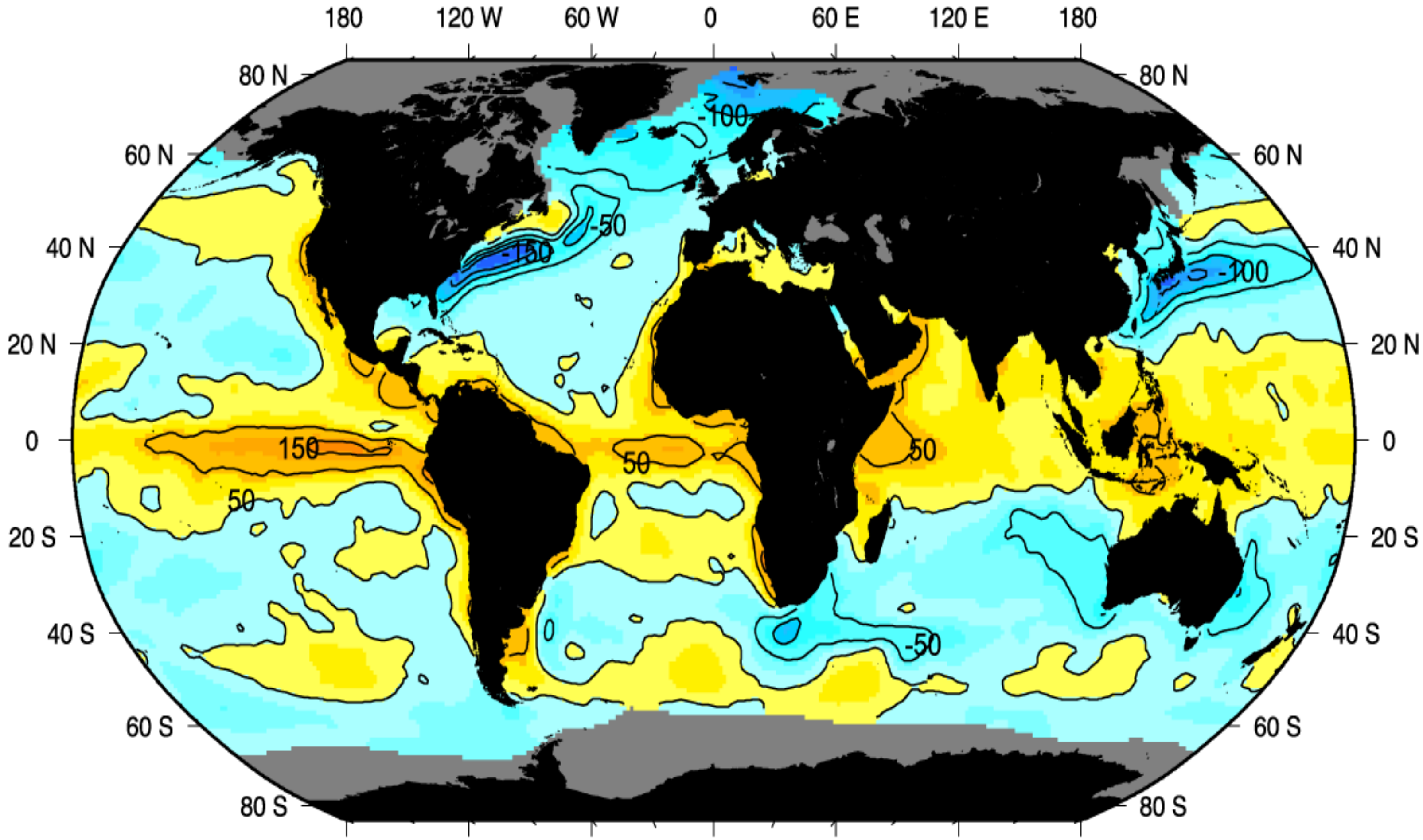


Potential temperature



1. Thermodynamics: Temperature and heat

Surface heat flux (W/m^2) into ocean (annual mean)



DPO Figure 5.12

1. Thermodynamics: Salinity

Salinity: mass of dissolved matter (expressed in grams) in a kilogram of seawater = Absolute salinity

•Units are parts per thousand (o/oo) or "psu" (practical salinity units), or unitless (mass/mass)

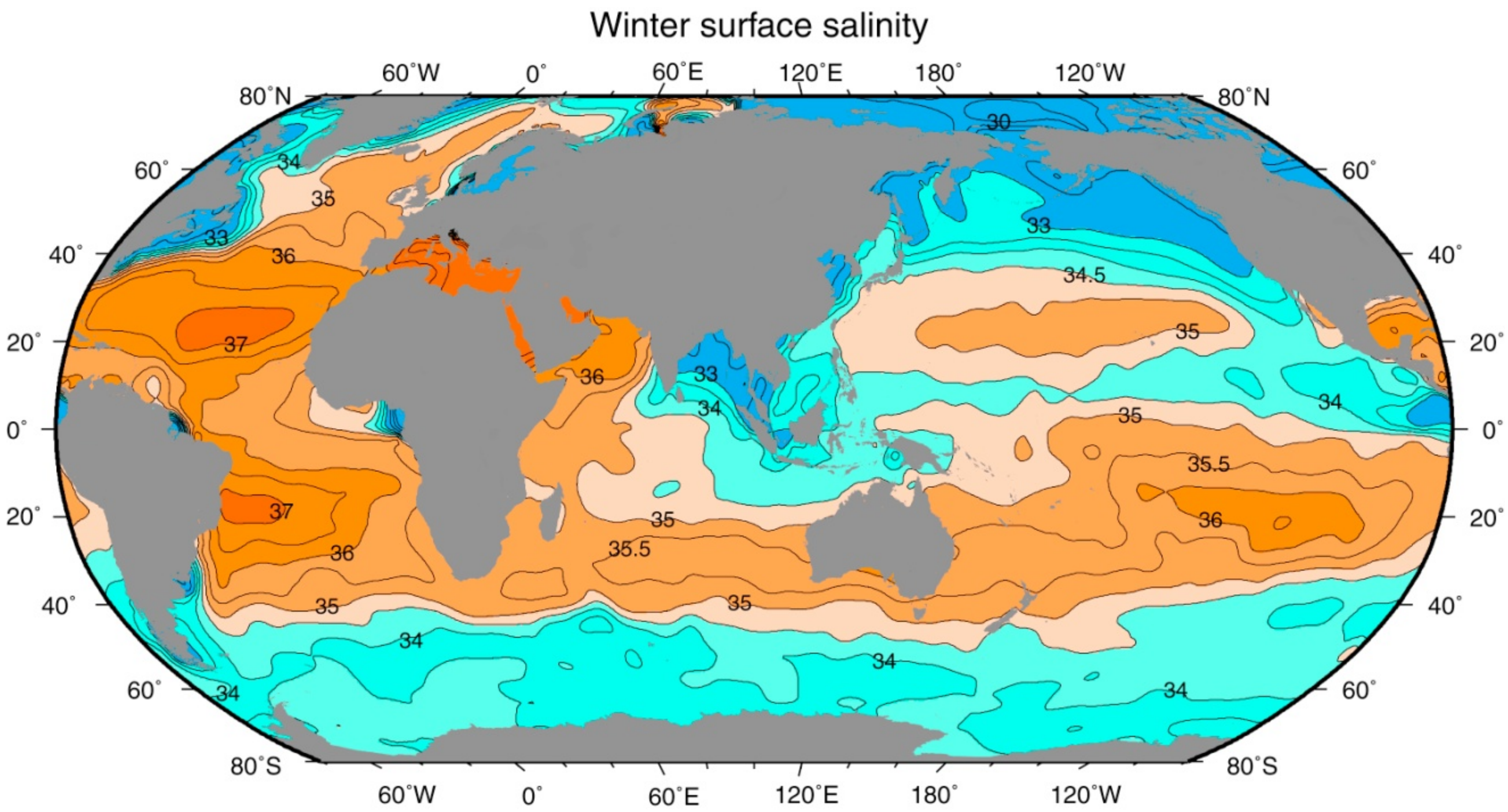
• "Law of equal proportions": the constituents of sea salt are present in almost equal proportion everywhere in the ocean because the ocean is well mixed on the time scale of weathering (source of most salts)

•Typical ocean salinity is 34 to 36 (i.e. 34 to 36 gm salt/kg seawater); total range in open ocean is about 30 (sea ice regions) to 40 (Red Sea)

•Salinity is changed by DILUTION with freshwater (from precipitation, evaporation, runoff). The total amount of salt in the ocean does not change on non-geological time scales.

•New international definition of salinity in 2010. Measurements and archived observations are still to be based on the conductivity methods (standard in 1978), but corrections to salinity are to be applied henceforth.

1. Thermodynamics: Salinity



Surface salinity (psu) in winter (January, February, and March north of the equator; July, August, and September south of the equator) based on averaged (climatological) data from Levitus et al. (1994b).

FIGURE 4.15

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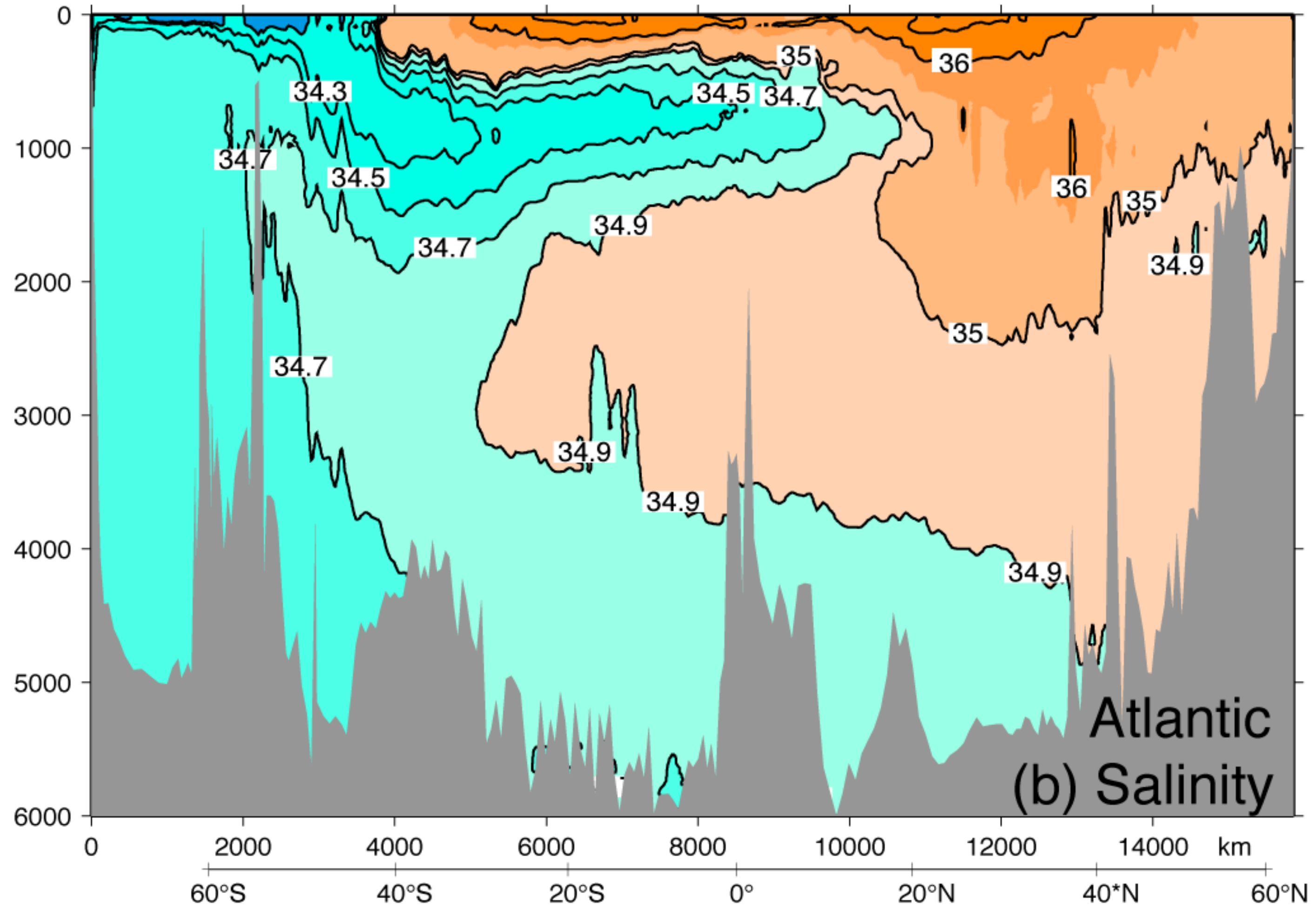
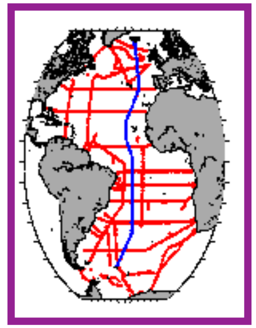
1. Thermodynamics: Salinity

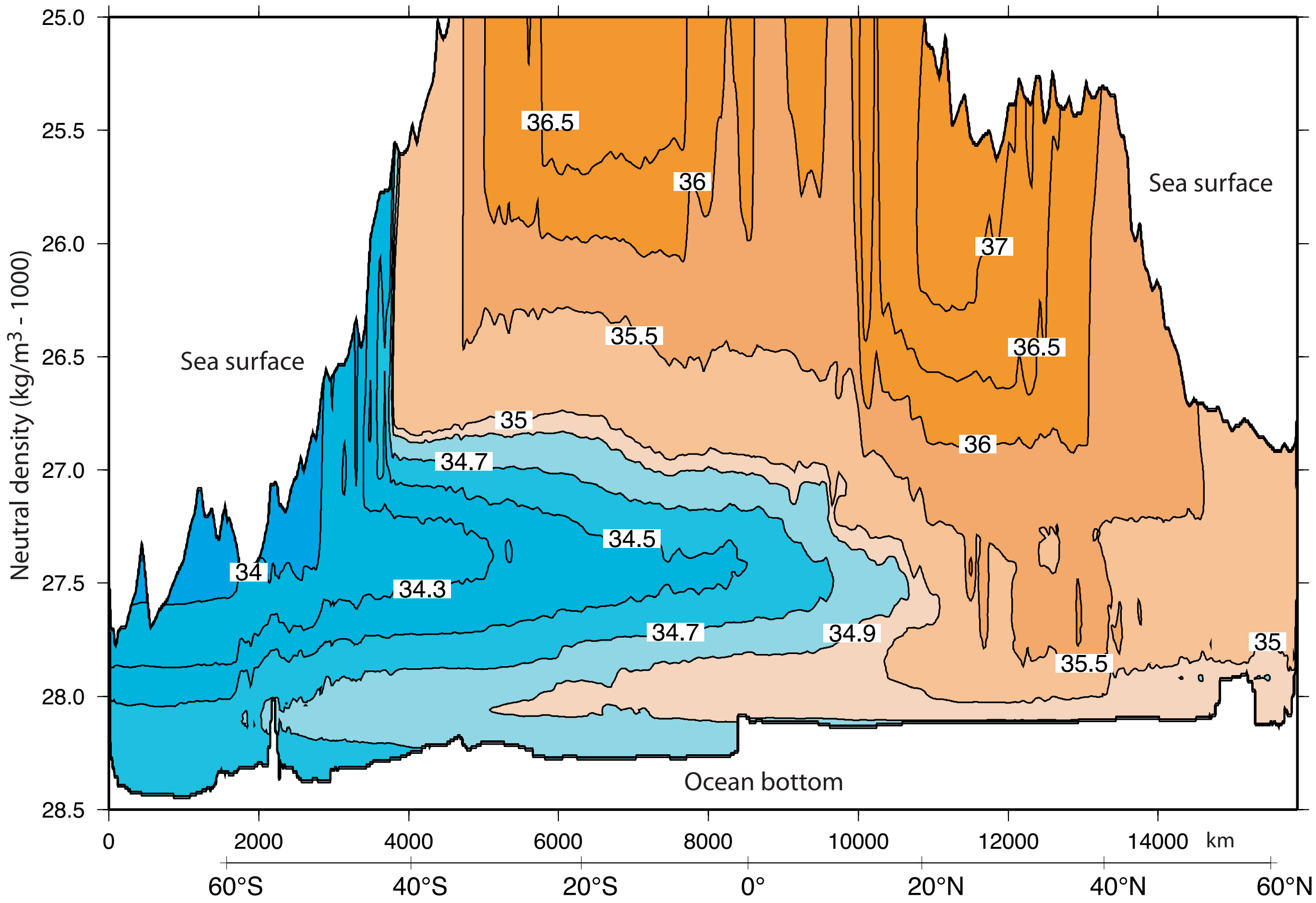
Atlantic salinity section

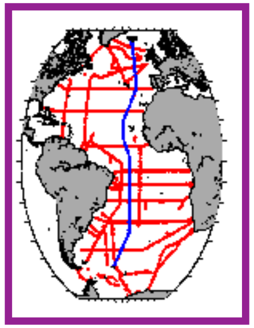




1. Thermodynamics: Salinity

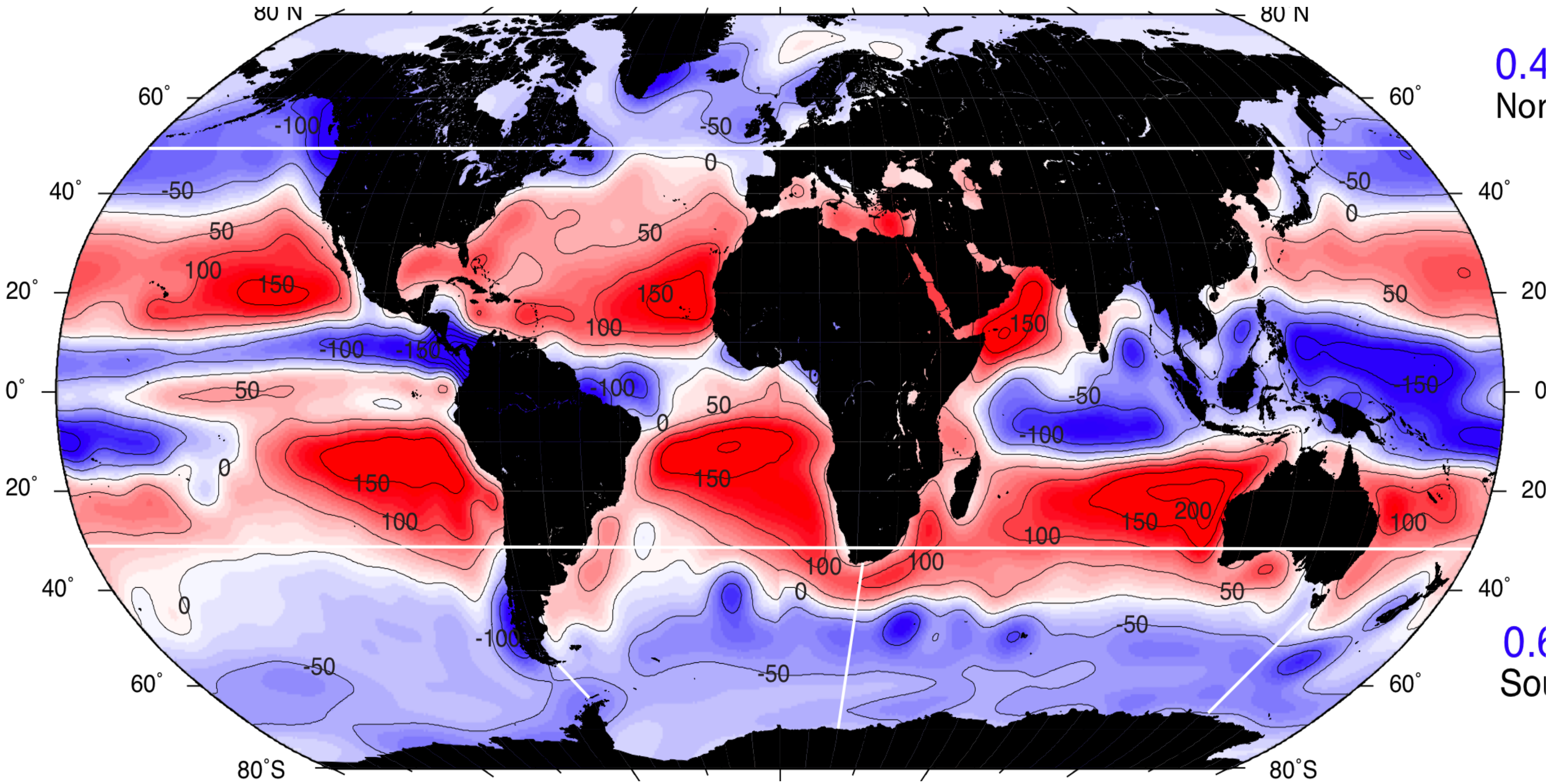
Atlantic salinity section with neutral density as the vertical coordinate (oceanography equivalent to isentropic coordinates)





1. Thermodynamics: Salinity

What sets salinity? Precipitation + runoff minus evaporation (cm/yr)



Salinity is set by freshwater inputs and exports since the total amount of salt in the ocean is constant, except on the longest geological timescales

NCEP climatology DPO 5.4

1. Thermodynamics: density

Equation of state (EOS) of seawater

Seawater density is determined empirically (in the lab)

$\rho = \rho(S, T, p)$ Units are mass/volume (kg/m^3)

The EOS is nonlinear; modelers often approximate it as linear.

Most commonly used EOS: EOS80

Newest EOS: TEOS-10 (IOC, SCOR, IAPSO, 2010)

ρ ranges from about 1020 kg/m^3 at the sea surface to 1050 kg/m^3 at bottom of ocean, mainly due to compression (pressure dependence)

We often express density as

$\sigma (S, T, p) = \rho(S, T, p) - 1000 \ kg/m^3$

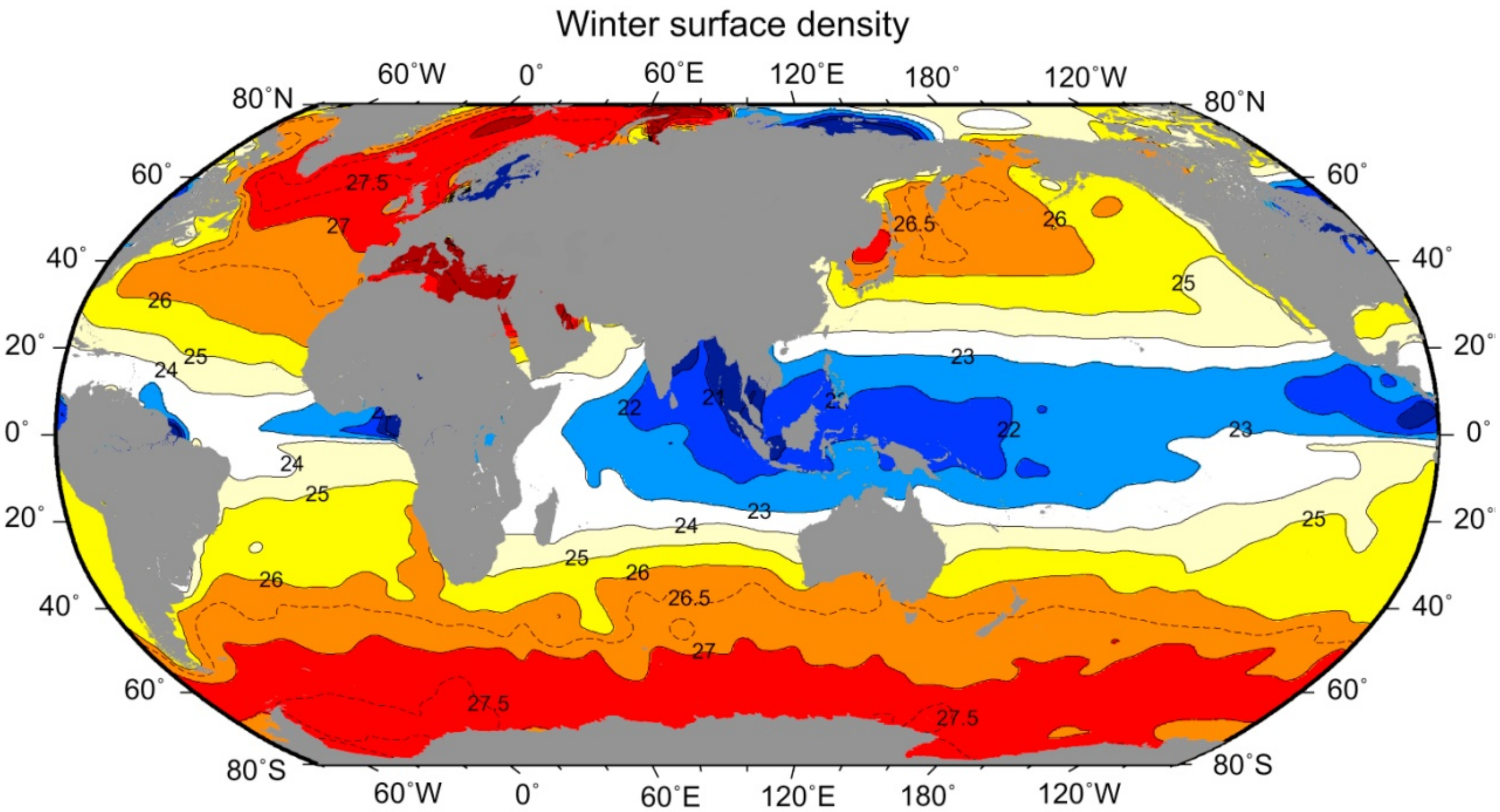
Thus σ ranges from about 20 to 50 kg/m^3

1. Thermodynamics: density

Potential density: Remove effect of mechanical compression. Create potential density profile by finding the density a water parcel has if moved adiabatically to a reference pressure.

Compressibility depends on temperature: cold water is more compressible than warm. Therefore two parcels (one cold/fresh and the other warm/salty) that might be the SAME density at their actual pressure (for example, at 4000 dbar), would not have the same density at the sea surface (0 dbar). This results in all kinds of complications; choice of reference pressure therefore depends entirely on what you want to do (your application).

1. Thermodynamics: density



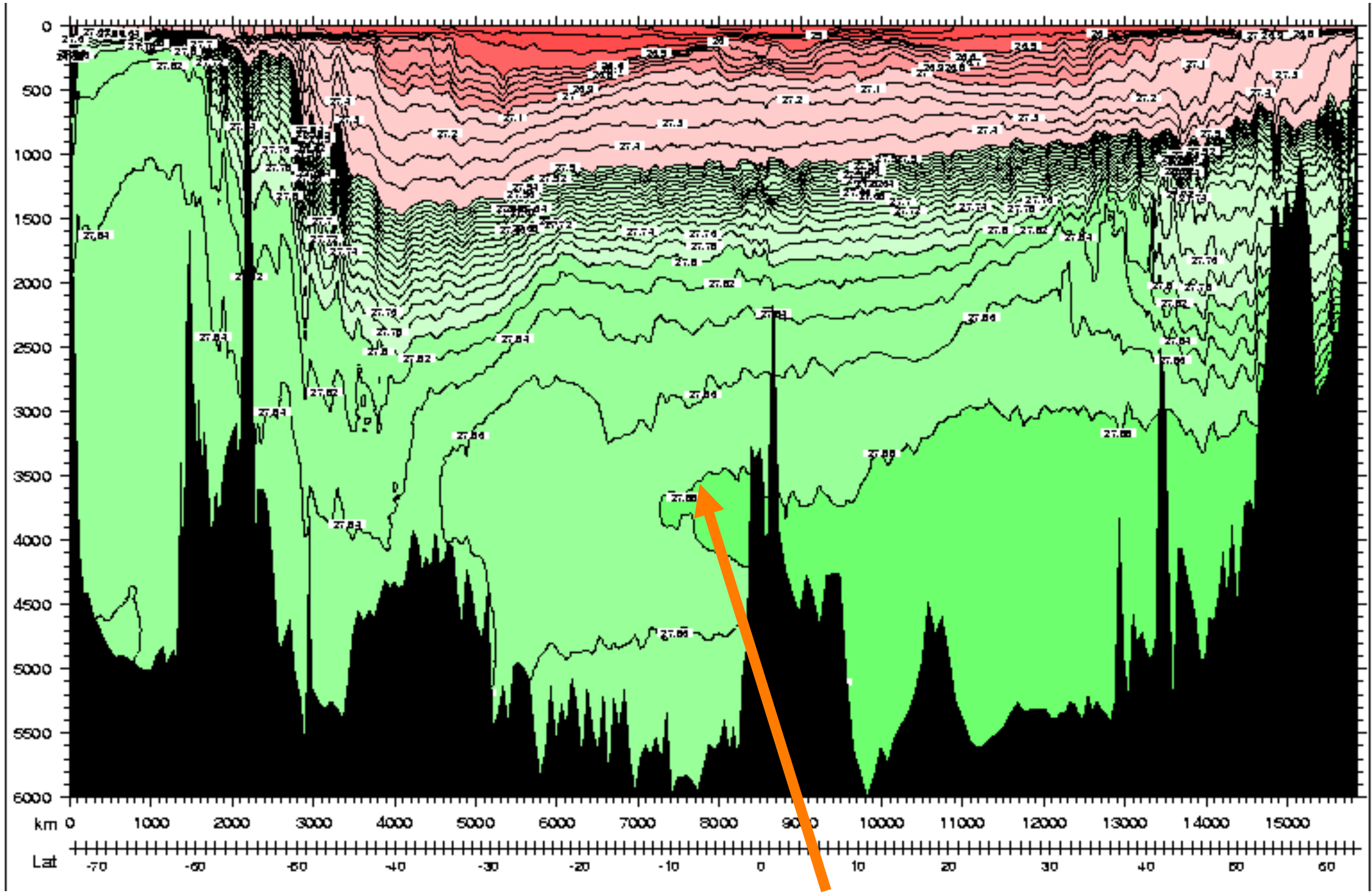
Surface density $\sigma\theta$ (kg m^{-3}) in winter (January, February, and March north of the equator; July, August, and September south of the equator) based on averaged (climatological) data from Levitus and Boyer (1994) and Levitus et al. (1994b).

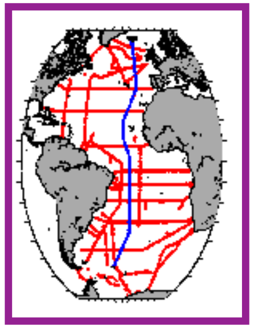
FIGURE 4.19

TALLEY

1. Thermodynamics: density

Atlantic section of potential density referenced to 0 dbar (sea surface): σ_θ

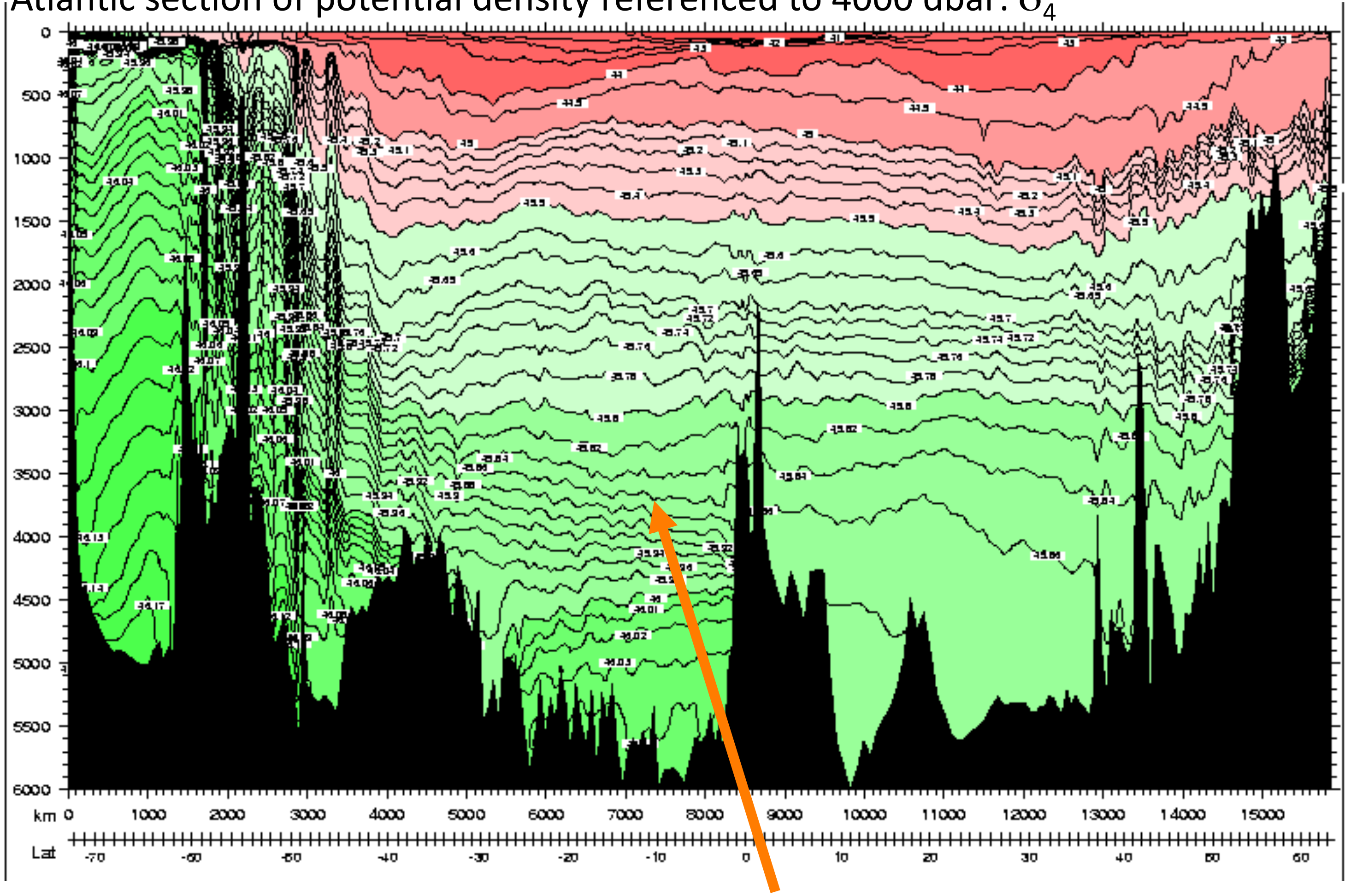


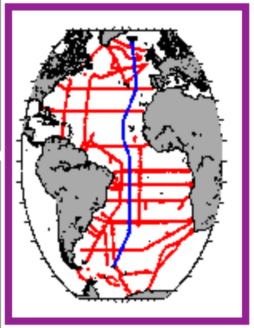


Need to use deeper reference pressures to check local vertical stability (e.g. σ_4)

1. Thermodynamics: density

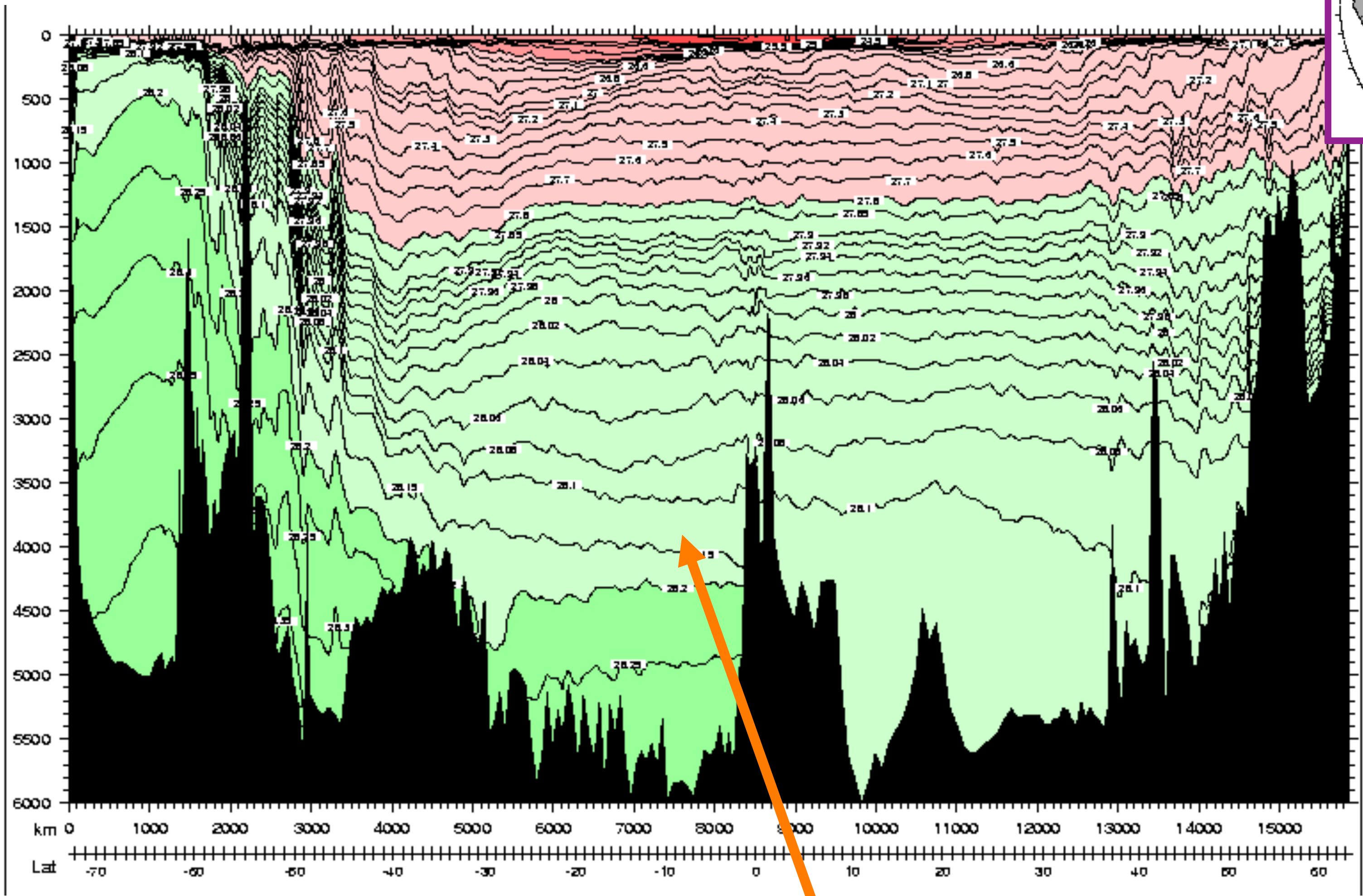
Atlantic section of potential density referenced to 4000 dbar: σ_4

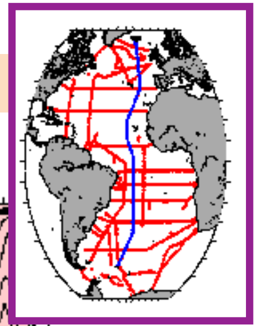




Potential density σ_θ inversion vanishes with use of deeper reference (σ_4): in fact, extremely stable!!

Atlantic section of “neutral density”: γ^n





Jackett and McDougall (1997)

1. Thermodynamics: density

- External forcings for ocean density: heating/cooling and precipitation/evaporation/runoff/brine rejection
- Buoyancy flux: buoyancy is 1/density
- When buoyancy increases, density decreases
- Heating increases buoyancy
- Rain increases buoyancy

- From these maps, you can see that the buoyancy is mostly changed by heat flux

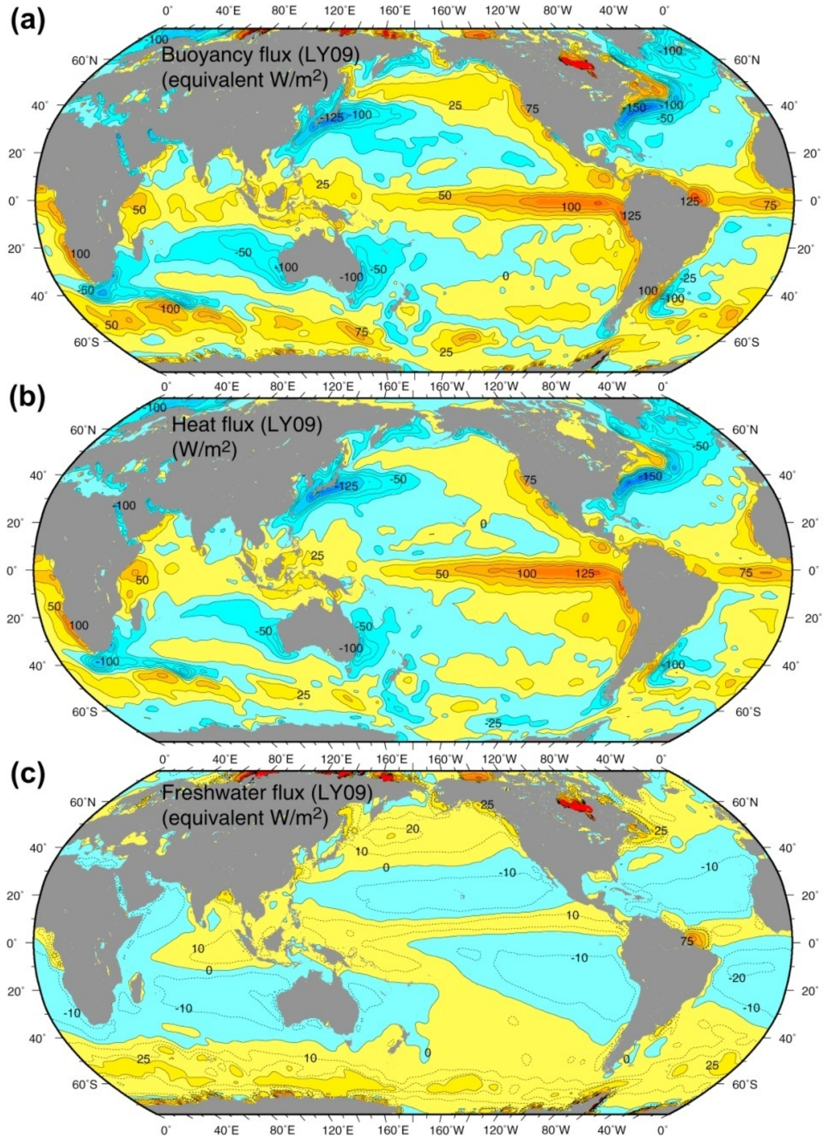


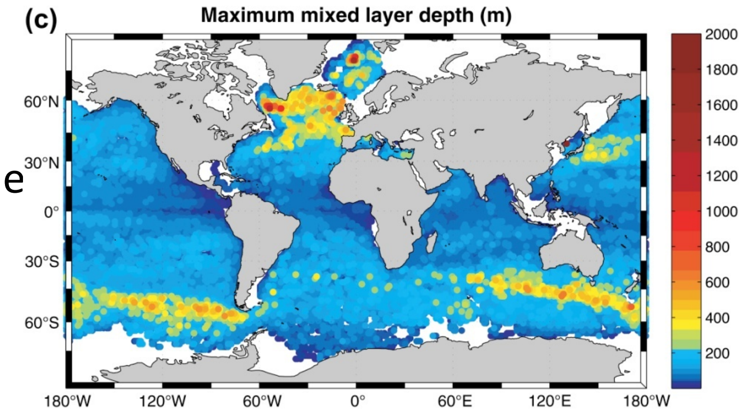
FIGURE S5.8

TALLEY

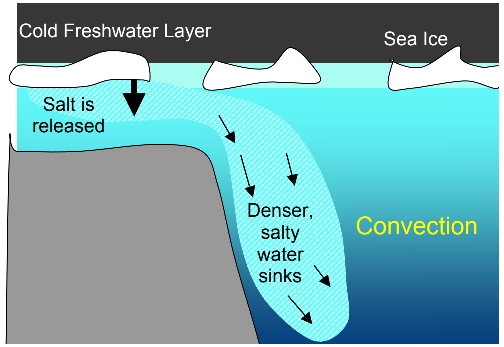
1. Thermodynamics: density

- Mechanisms to change ocean temperature, salinity, density

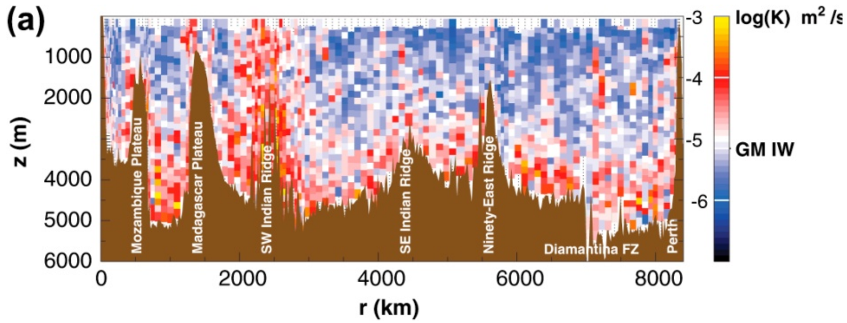
- Through the mixed layer: convection and restratification (external buoyancy flux through the sea surface)



- Brine rejection (external buoyancy flux that creates sea ice, removal of salt from ice)



- Internal mixing (diapycnal and along-isopycnal diffusion): spatially and temporally dependent diffusivities – much stronger at ocean bottom



2. Dynamics

- Governing equations (momentum, mass, S, T, EOS)
- Coriolis force
- Geostrophic balance
- Ekman balance
- Potential vorticity evolution

2. Dynamics

Governing equations

- Mass conservation (continuity) (no holes)

- Force balance: Newton's Law ($\vec{F} = m\vec{a}$) (3 equations)
- Equation of state (for oceanography, dependence of density on temperature, salinity and pressure) (1 equation)
- Equations for temperature and salinity change in terms of external forcing, or alternatively an equation for density change in terms of external forcing (2 equations)

2. Dynamics

Momentum equations with Earth's rotation

Horizontal (x) (west-east)

acceleration + advection + Coriolis = pressure gradient force + viscous term

Horizontal (y) (south-north)

acceleration + advection + Coriolis = pressure gradient force + viscous term

Vertical (z) (down-up)

acceleration +advection (+ neglected very small Coriolis) =

pressure gradient force + effective gravity (including centrifugal force) + viscous term

2. Dynamics

Momentum equations with rotation
Coordinates are (x, y, z) (east, north, up)
Velocities are (u, v, w)

x: $\partial u/\partial t + u\, \partial u/\partial x + v\, \partial u/\partial y + w\, \partial u/\partial z - fv = -(1/\rho)\partial p/\partial x$
$+ \partial/\partial x(A_H \partial u/\partial x) + \partial/\partial y(A_H \partial u/\partial y) + \partial/\partial z(A_V \partial u/\partial z)$ (7.11a)

y: $\partial v/\partial t + u\, \partial v/\partial x + v\, \partial v/\partial y + w\, \partial v/\partial z + fu = -(1/\rho)\partial p/\partial y$
$+ \partial/\partial x(A_H \partial v/\partial x) + \partial/\partial y(A_H \partial v/\partial y) + \partial/\partial z(A_V \partial v/\partial z)$ (7.11b)

z: $\partial w/\partial t + u\, \partial w/\partial x + v\, \partial w/\partial y + w\, \partial w/\partial z$ (+ neglected small Coriolis) $= -(1/\rho)\partial p/\partial z - g$
$+ \partial/\partial x(A_H \partial w/\partial x) + \partial/\partial y(A_H \partial w/\partial y) + \partial/\partial z(A_V \partial w/\partial z)$ (7.11c)

f is the Coriolis parameter $f = 2\Omega \sin(\text{latitude})$
g is gravitational acceleration
Eddy viscosity rather than molecular viscosity;
A_H is horizontal and A_V is vertical eddy viscosity

2. Dynamics

Coriolis force

Deflection relative to Earth's rotating surface, to the right in the northern hemisphere, to the left in the southern hemisphere

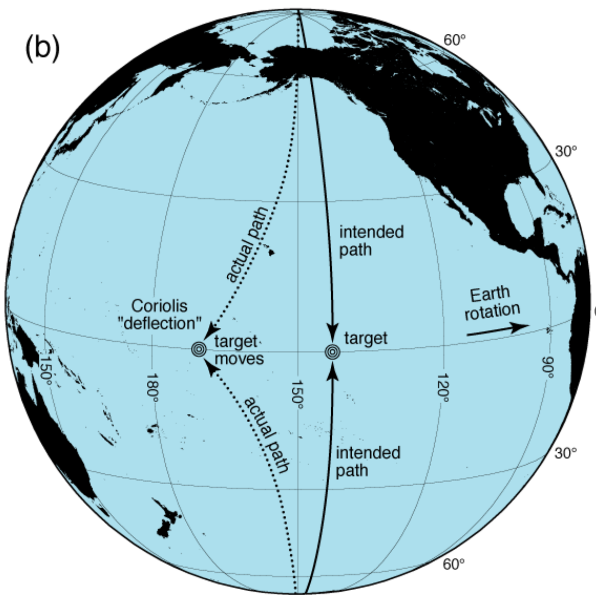


FIGURE S7.2

Inertial motion observations:

Surface drifters after passing of a storm in the NE Pacific
(d'Asaro et al., 1995)

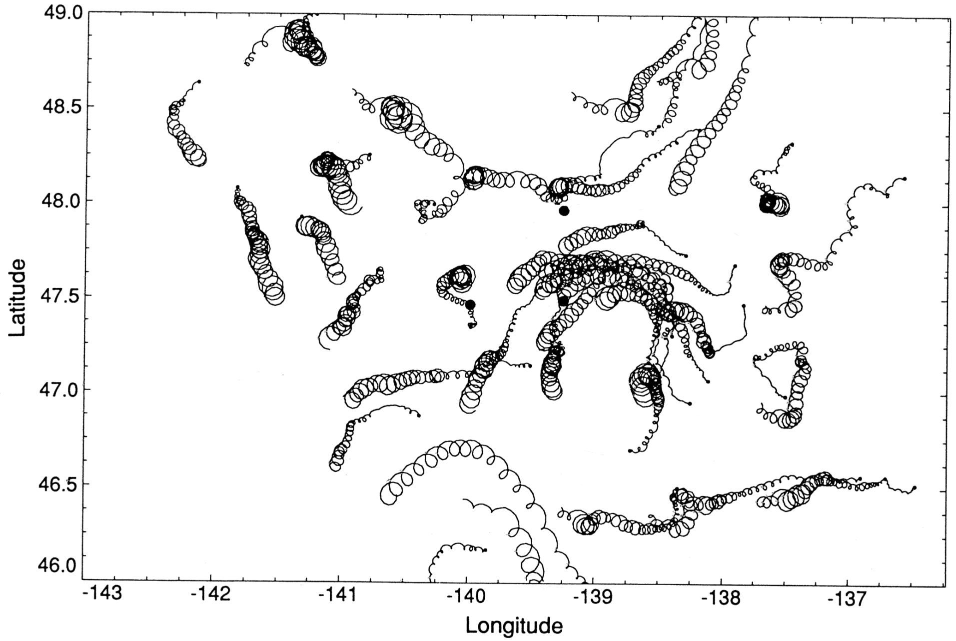
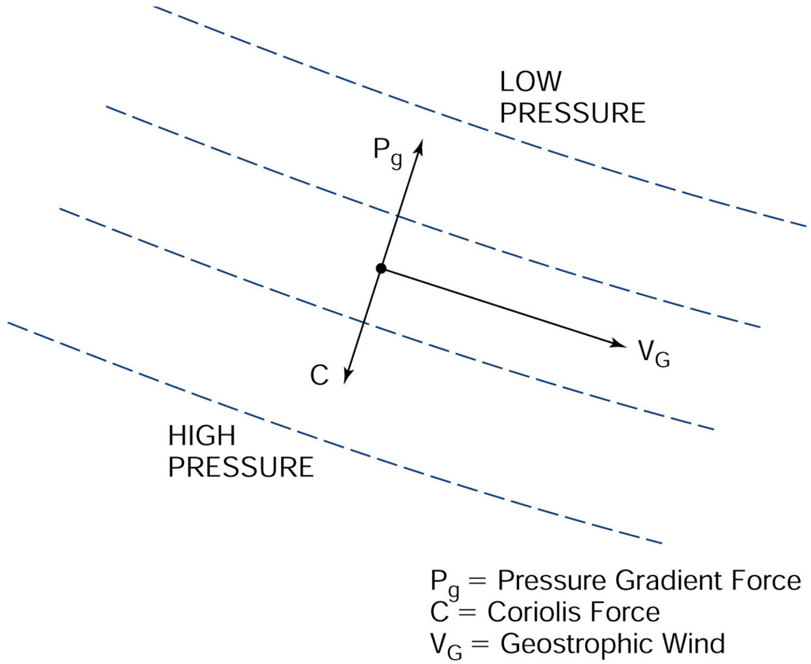


FIGURE S7.8

2. Dynamics

Geostrophic balance: pressure gradient force balanced by Coriolis force



The approximation (small Rossby number, small aspect ration, small Ekman number)

Horizontal (x): Coriolis = pressure gradient force

$$- fv = - (1/\rho)\partial p/\partial x$$

Horizontal (y): Coriolis = pressure gradient force

$$fu = - (1/\rho)\partial p/\partial y$$

(and hydrostatic balance:

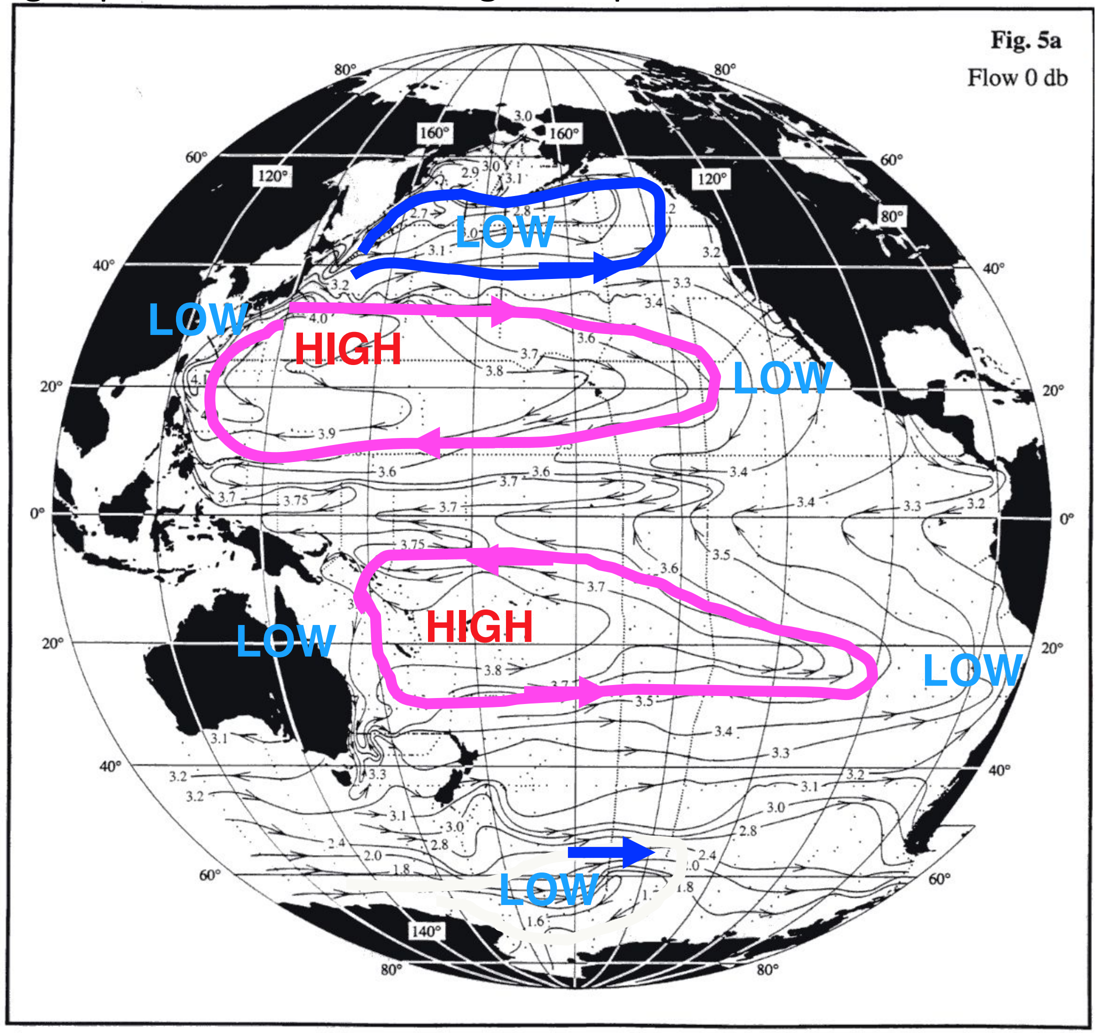
Vertical (z): 0 = pressure gradient force + effective gravity

$$0 = - (1/\rho)\partial p/\partial z - g$$

2. Dynamics

Example from ocean of geostrophic flow

Surface height/pressure and surface geostrophic circulation



Circulation is counterclockwise around the Low (cyclonic) and clockwise around the High (anticyclonic).

(northern hemisphere)

Southern Hem:

Cyclonic is clockwise, etc..

Reid, 1997

2. Dynamics

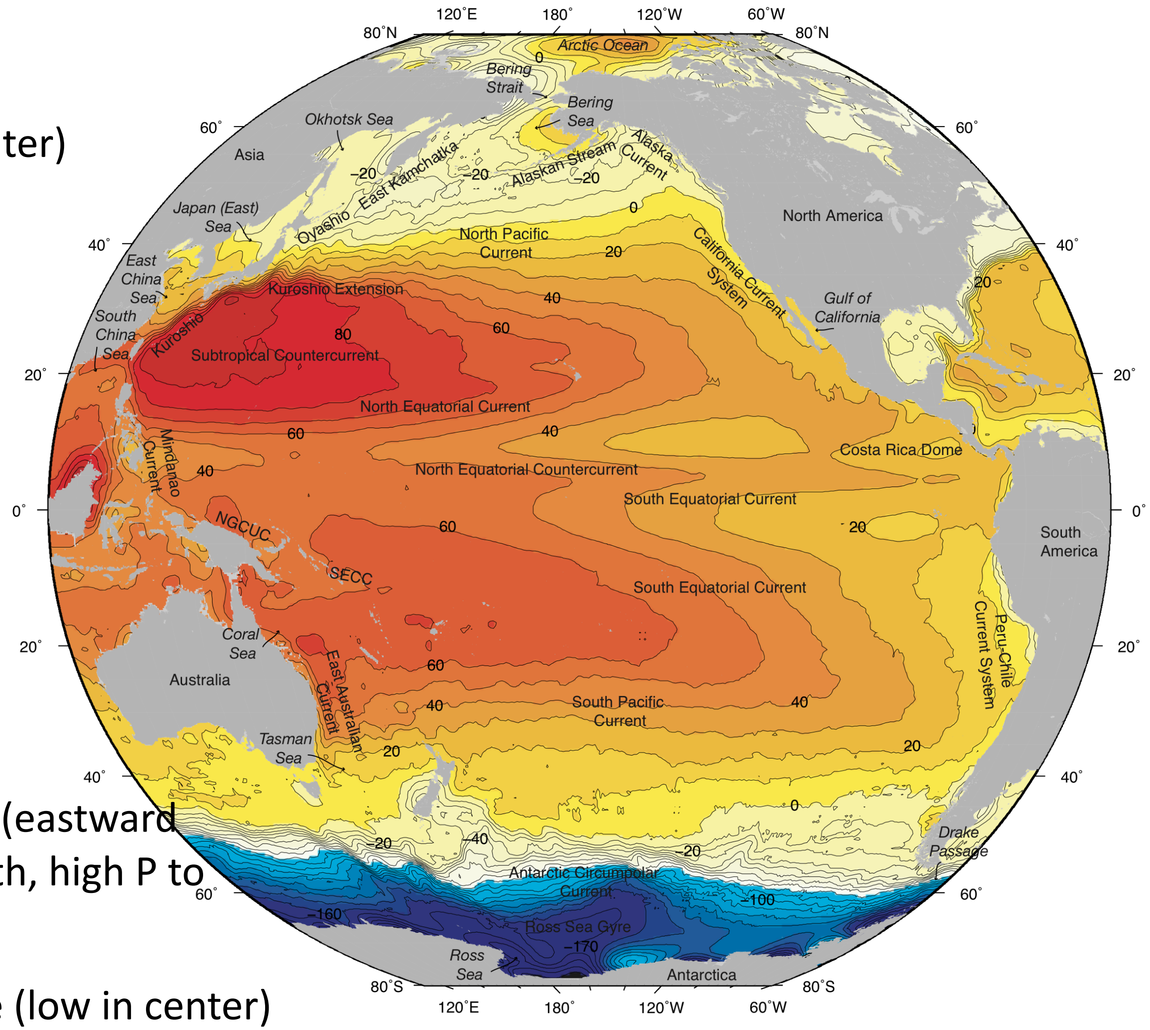
Sea surface height in the Pacific. (Using Niiler et al., 2003, surface heights based on drifters)

Cyclonic gyre (low in center)

Anticyclonic gyres (high in center)

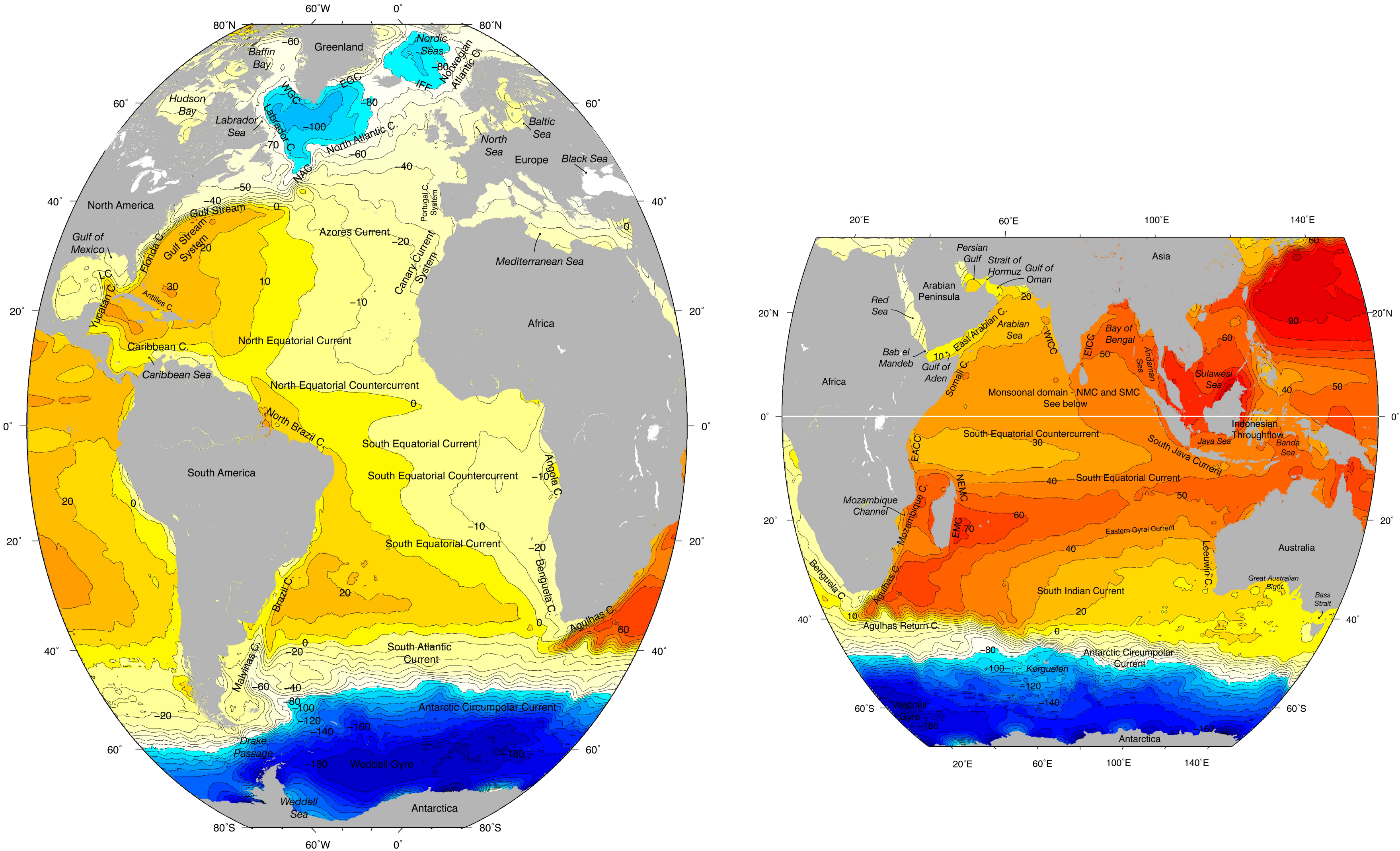
Antarctic Circumpolar Current (eastward geostrophic flow, low P to south, high P to north)

Cyclonic gyre (low in center)



2. Dynamics

Sea surface height. (Using Niiler et al., 2003, surface heights based on drifters)



2. Dynamics: Thermal wind and geostrophic method

Thermal wind relation:

$$-f\partial v/\partial z = (g/\rho_o)\partial\rho/\partial x$$
$$f\partial u/\partial z = (g/\rho_o)\partial\rho/\partial y$$

Relates vertical shear to horizontal density variations. Variants of this constitute the “geostrophic method” or “dynamical method” for estimating geostrophic velocities from density profiles.

Requires measurement or estimate of velocity at one depth. This has traditionally been the trickiest part of large-scale circulation estimation.

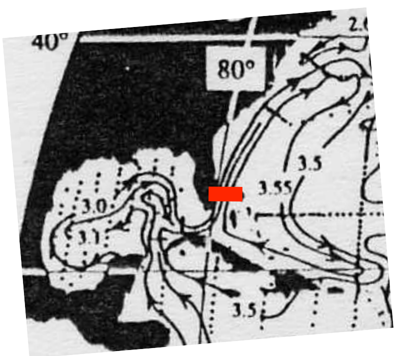
“Reference level velocity”: determined through overall mass balances, excellent observed velocities (floats or drifters) but carefully matched with synoptic observations (e.g. through data assimilation or inverse methods). This used to be called setting the “level of no motion”, but it is really finding the “level of known motion”

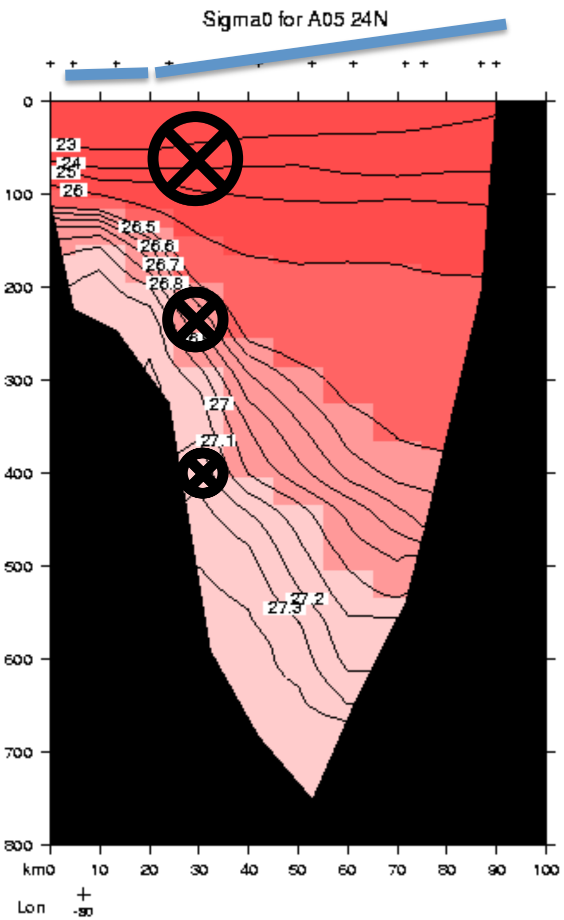
2. Dynamics: Thermal wind and geostrophic method

Gulf Stream in Florida Strait:

Potential density section and velocity section.

Inferred sea surface height is low inshore and high offshore





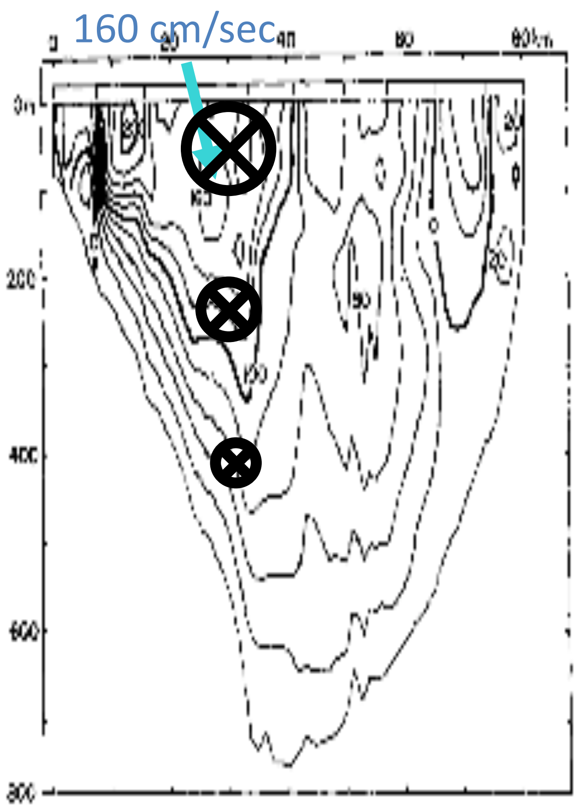
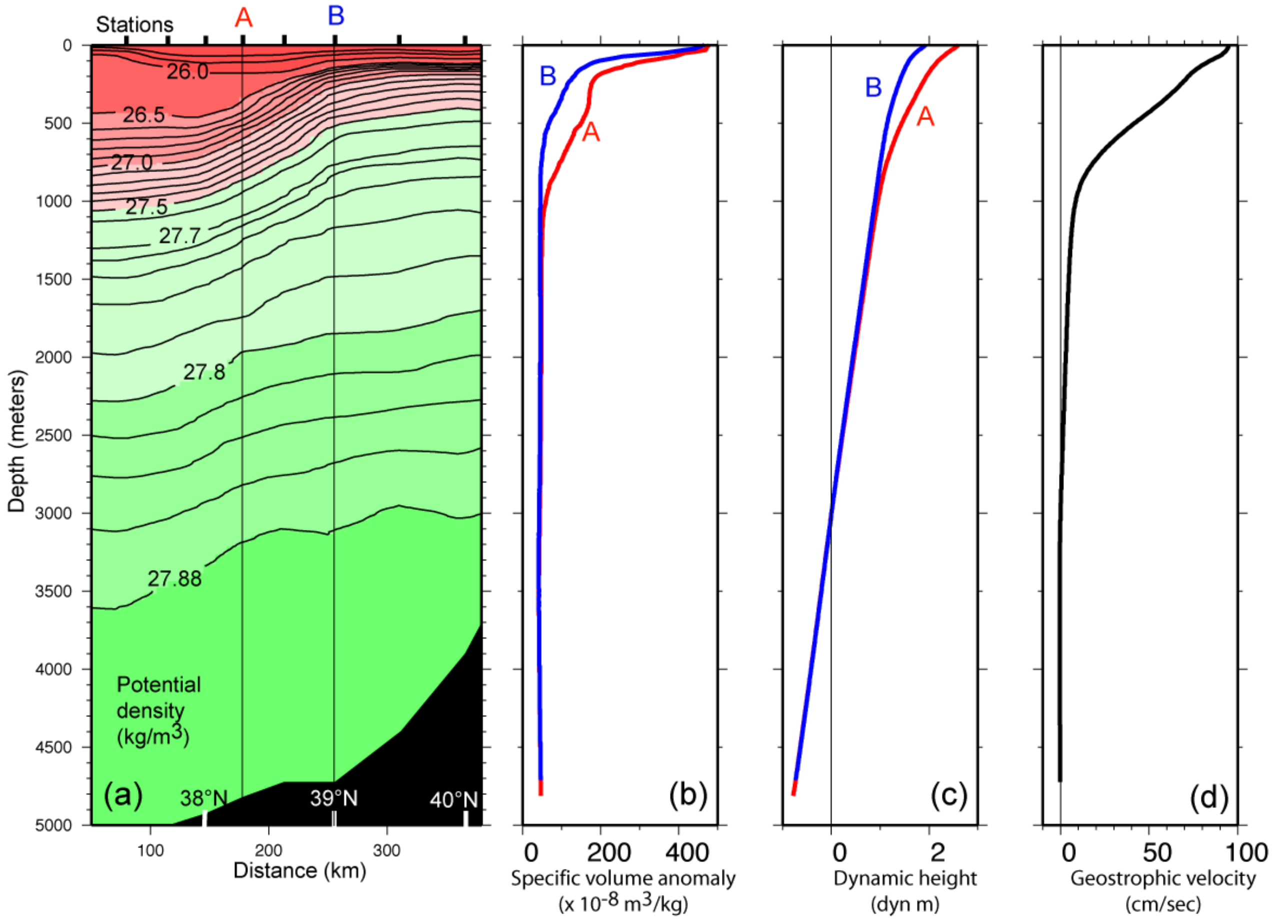


FIG. 7. Geostrophic velocity relative to the bottom. Positive is northward. The contour interval is 20 cm s^{-1}.

Roemmich, 1983

2 Dynamics: thermal wind and geostrophic method



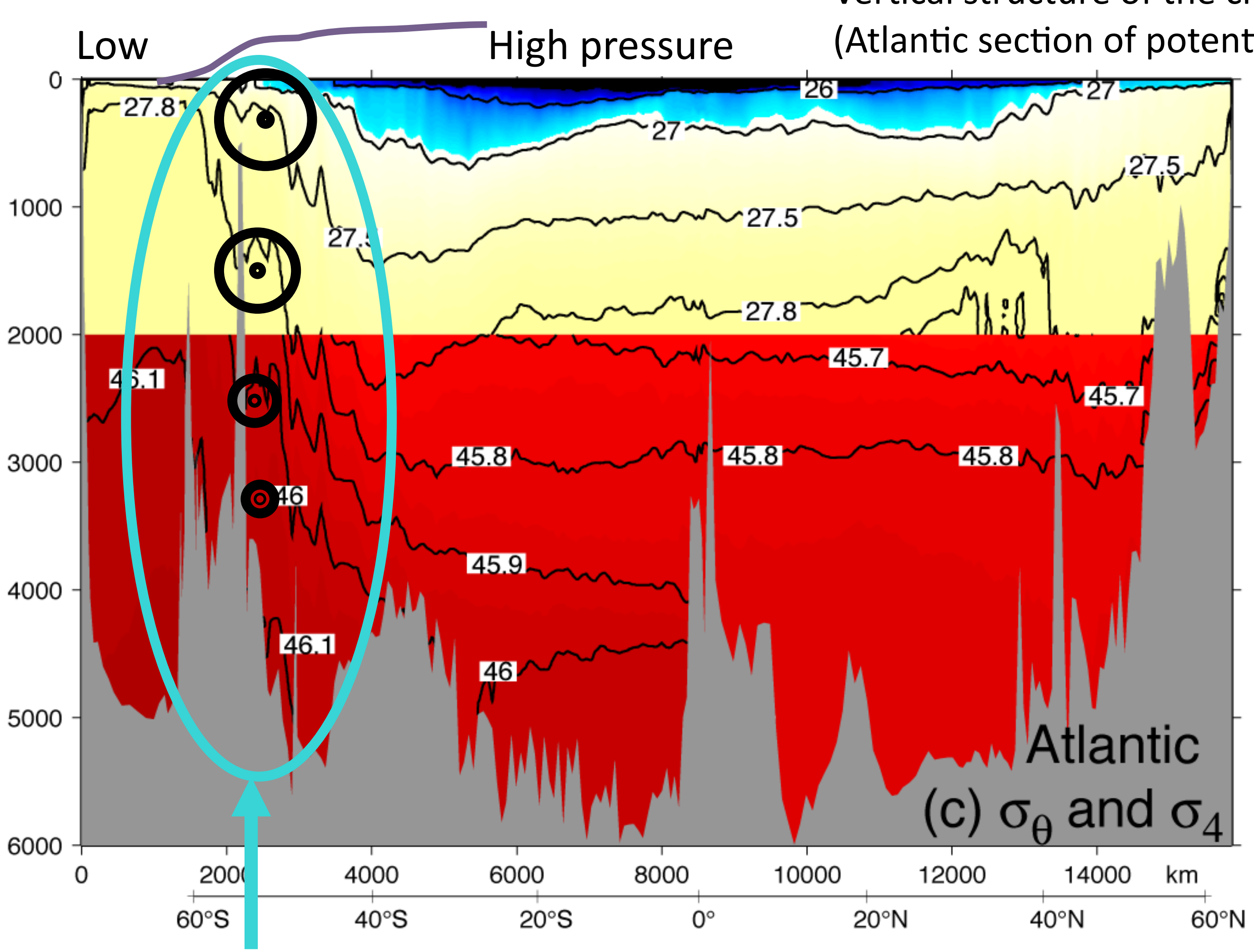
In this example, compute geostrophic velocities relative to 0 cm/sec at 3000 dbar. If we KNOW v at 3000 dbar, then just add it to the whole profile

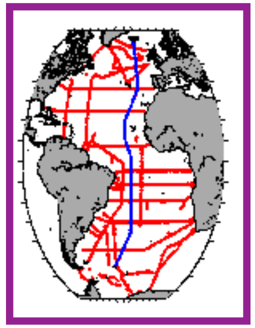
DPO Fig. 7.11

Gulf Stream density, dynamic height and geostrophic velocity

2. Dynamics: thermal wind and geostrophic method

Vertical structure of the circulation
(Atlantic section of potential density)





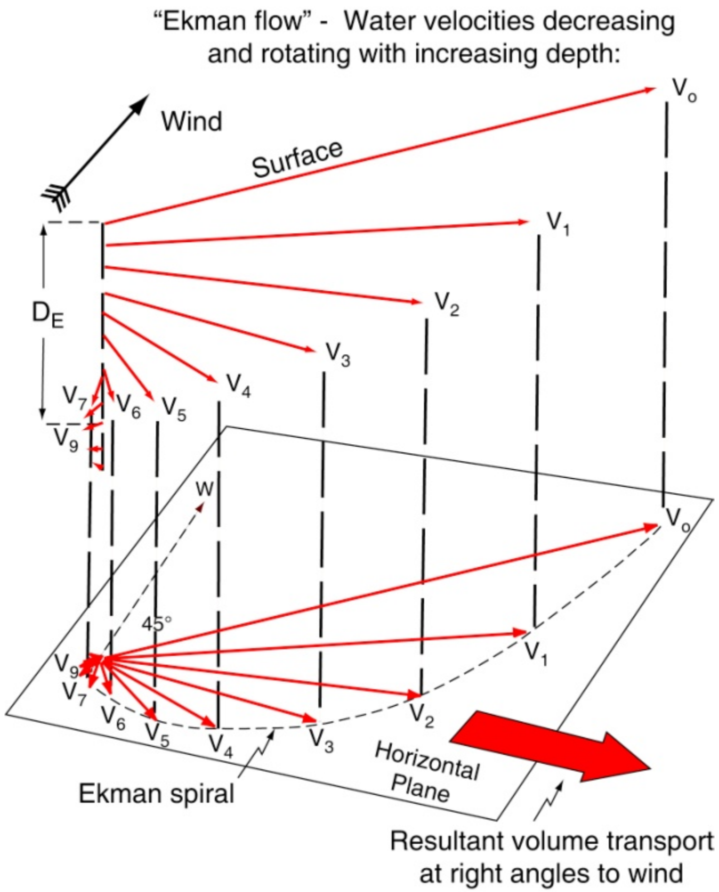
Antarctic Circumpolar Current

DPO Fig. 4.10

2. Dynamics: Ekman balance

Ekman balance

- Frictional balance in the surface layer: driven by wind stress, subject to Coriolis force



In the surface boundary layer, the approximate force balance is

x: $-fv = \partial/\partial z(A_V \partial u/\partial z)$

y: $fu = \partial/\partial z(A_V \partial v/\partial z)$

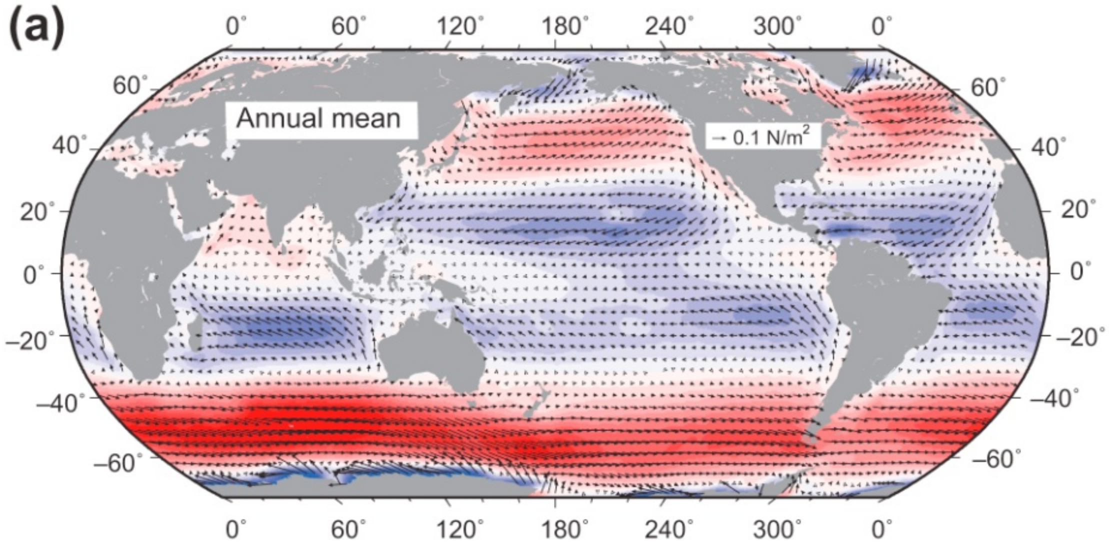
Integrate vertically to obtain the Ekman layer "transport", which depends only on the wind stress and not on the eddy viscosity magnitude or distribution

x: $-fV_{EK} = A_V \partial u/\partial z = \tau^{(x)} / \rho$

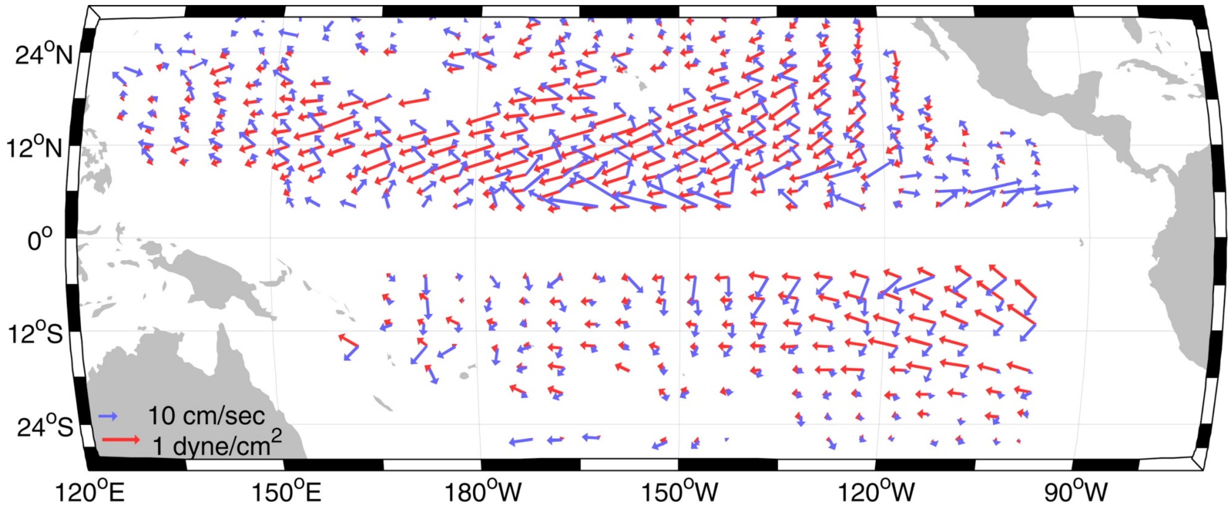
y: $fU_{EK} = A_V \partial v/\partial z = \tau^{(y)} / \rho$

2. Dynamics

Ekman balance: observed response to wind stress



Annual mean wind stress (NCEP)
Color is zonal stress



Mean currents at 15 m (blue) plotted with mean wind stress (red)
(Ralph and Niiler, 1999)