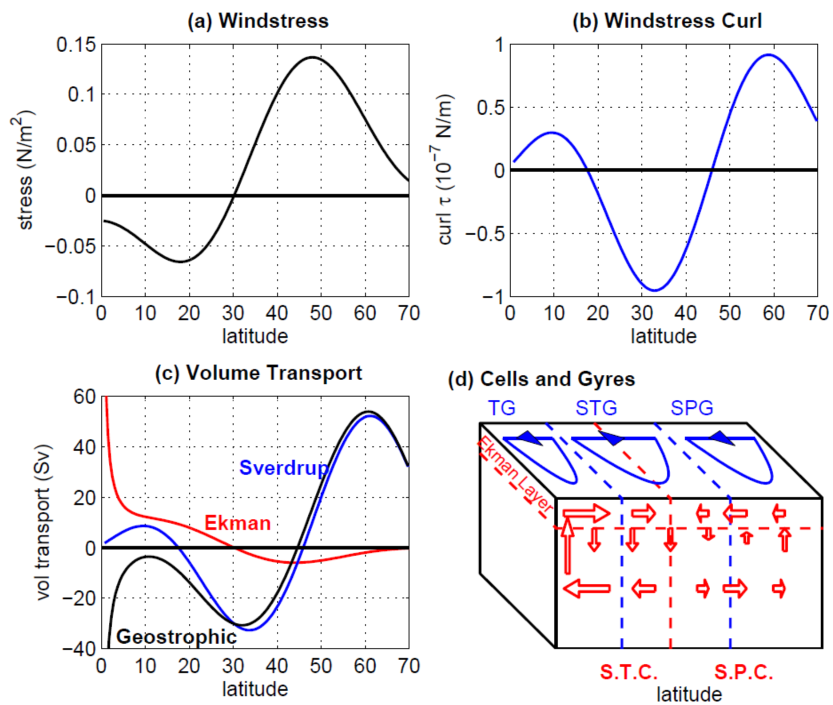
George Mason University
CLIM 752 Ocean General Circulation

Introduction



Reading: *Ocean Circulation in Three Dimensions,* Chapters I and II

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CLIM 752 Ocean General Circulation: Goals

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You have already seen:
- Description of major ocean gyres
- Linear theory of wind-driven barotropic circulation
  (Sverdrup, Stommel, Munk)
- Description of deep meridional overturning and water masses
- Stommel-Arons theory of horizontal flow patterns in deep ocean

Some questions this class will address:
- How does the ocean circulation affect climate?
- What does the 3D wind-driven circulation look like?
- How do we deduce this 3D flow from first principles?
- How are subduction regions and equatorial upwelling linked?
- What determines the strength of the deep meridional overturning?
- Can the meridional overturning undergo catastrophic change?

CLIM 752 Outline of Topics

1. Intro: Dynamics, Kinematics, Heat Transport
2. Review of wind-driven gyres: observation and theory
3. Three-dimensional barotropic flow
4. Surface Properties and Mixed Layer
5. Subduction and Luyten-Pedlosky-Stommel theory
6. The Equatorial Undercurrent
7. Shallow Overturning: Tropical and Subtropical Cells
8. Eddies and Turbulence
9. Deep Meridional Overturning Observations
10. Deep Meridional Overturning Models and Theory
11. Southern Ocean and Antarctic Circumpolar Current
12. Multiple states and time variability of the deep merid overturning

Grades

30% of grade: problem set every week or two
40% of grade: final exam
30% of grade: paper on topic relevant to course (1000-1500 words)

Reference Material

Required Reading

Klinger and Haine: *Ocean Circulation in Three Dimensions*, Cambridge University Press, Draft at http://mason.gmu.edu/~bklinger/bookhome.html

Recommended Reading

Pedlosky, 1996: *Ocean Circulation Theory*, Springer-Verlag

Schmittner, Chiang, and Hemming, eds., 2007: *Ocean Circulation, Mechanisms and Impacts,* American Geophysical Union

Siedler, Church, and Gould, eds., 2001: *Ocean Circulation and Climate*, Academic Press

Tomczak and Godfrey, 1994: *Regional Oceanography*, Pergamon Press

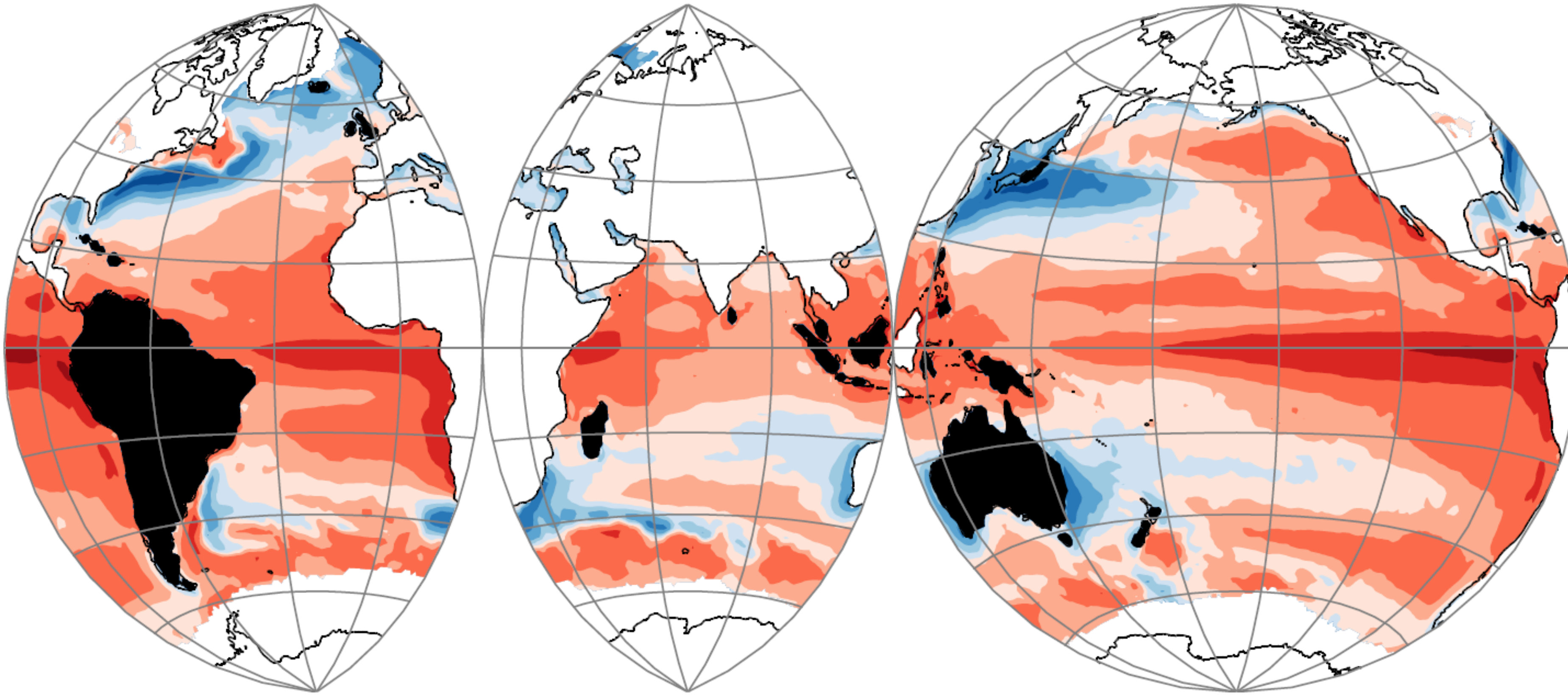
Van Aken, 2007: *The Oceanic Thermohaline Circulation, An Introduction*, Springer

Some ways ocean circulation affects climate:

- **biological productivity → gas exchange**
- **direct transport of gases (e.g., the carbon cycle)**
- **ice → albedo**
- **heat exchange with atmosphere**

Annual Average Net Heat Flux Into Ocean

contours at 0, ±20, 40, 80, 160, 320 W/m^2

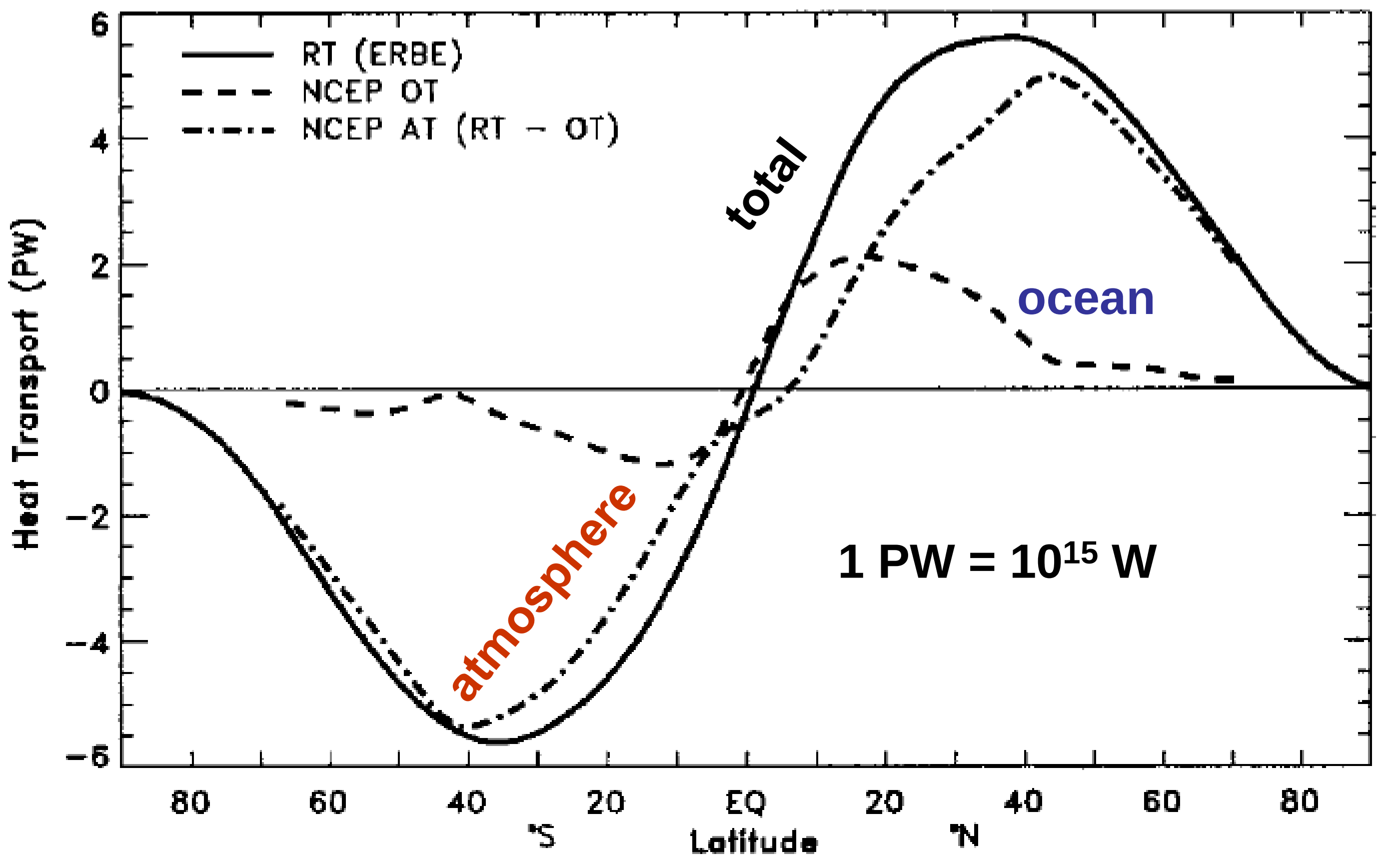


Heat exchange with atmosphere indicates ocean influence on climate

- Heat flux magnitudes comparable to solar irradiance.
- Ocean absorbs heat at equator, releases heat at mid-high latitudes
- Also zonal structure: heat absorption in east, emission in west
- Differences between oceans and between N and S hemispheres

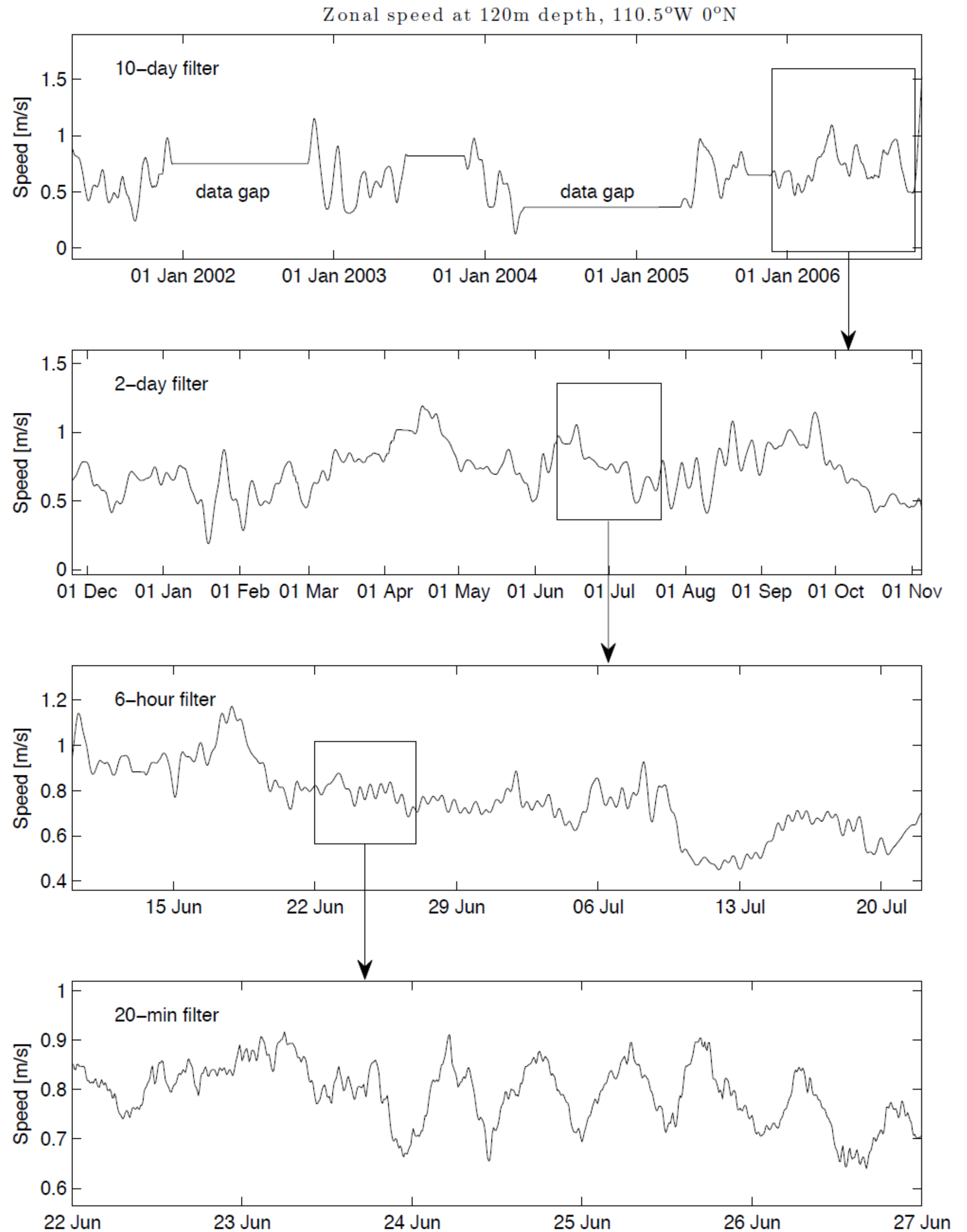
Ocean and Atmospheric Meridional Heat Transport

Ocean clearly important in tropics
Latest observations ➔ smaller ocean role at high latitudes
(still significant uncertainties)

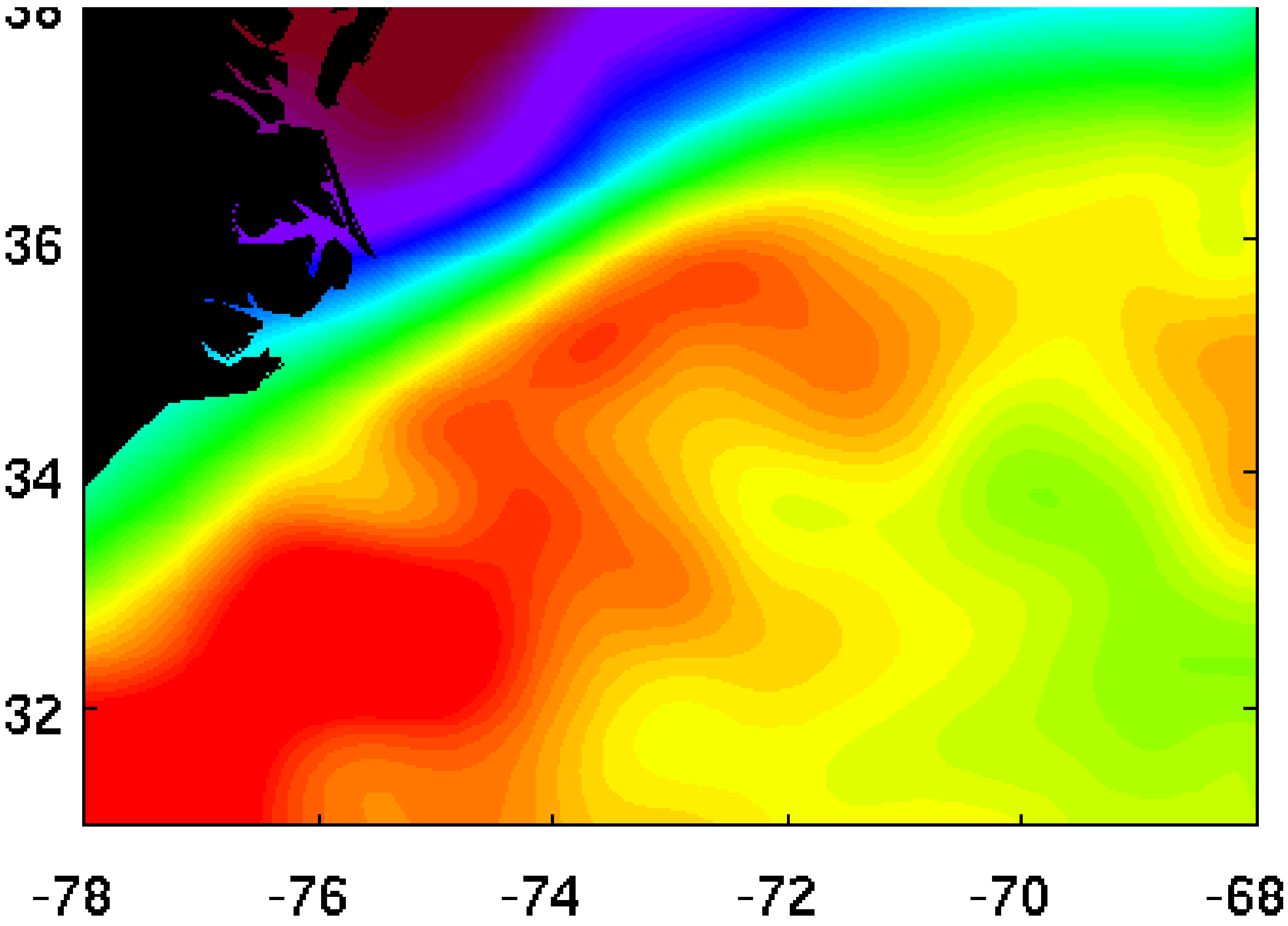


Trenbirth and Caron, 2001: J Clim, 3433-3443

Ocean Features Occur on Wide Space and Time Scales

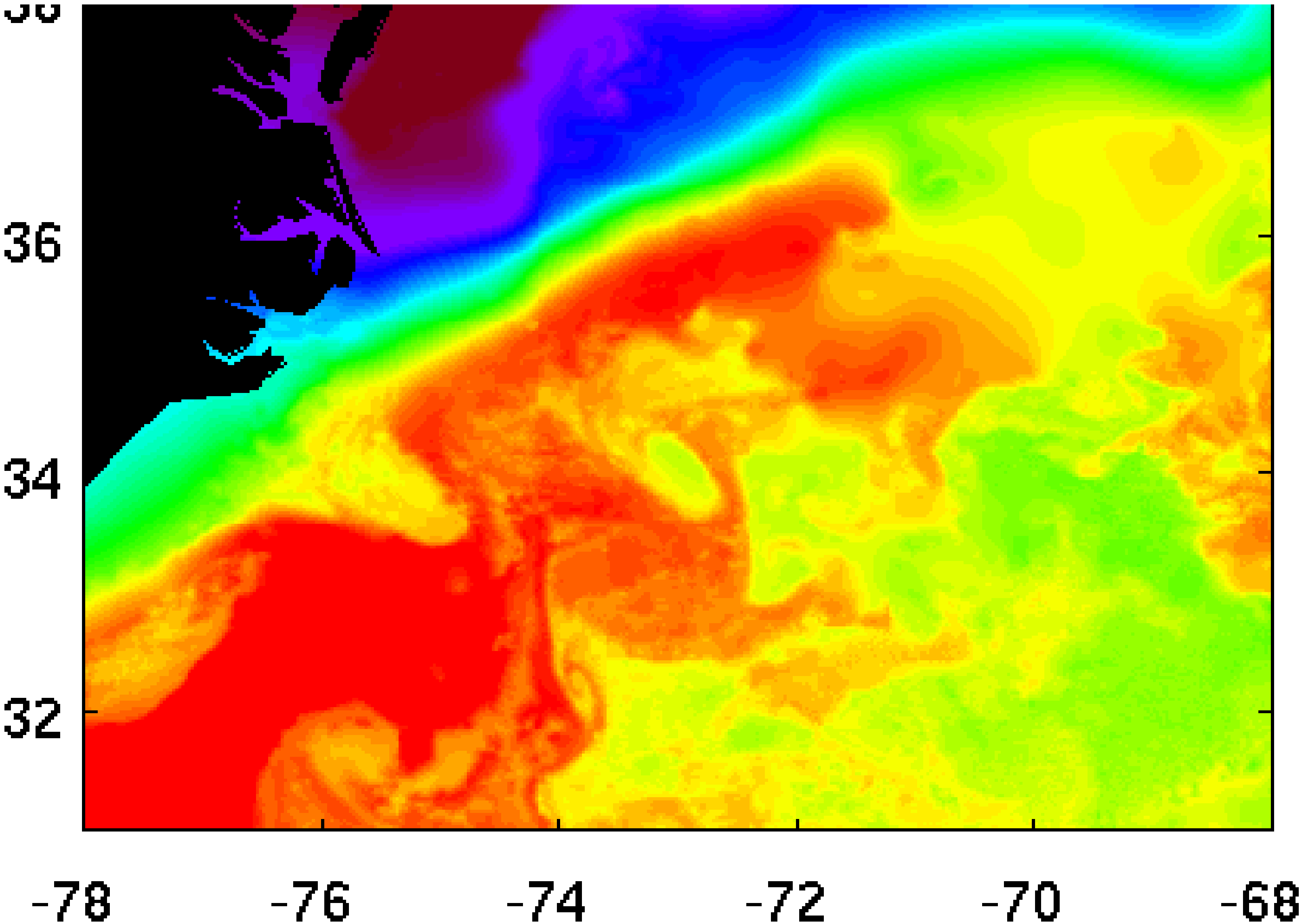


Different spatial scales also reveal different structures



http://podaac.jpl.nasa.gov/Multi-scale_Ultra-high_Resolution_MUR-SST

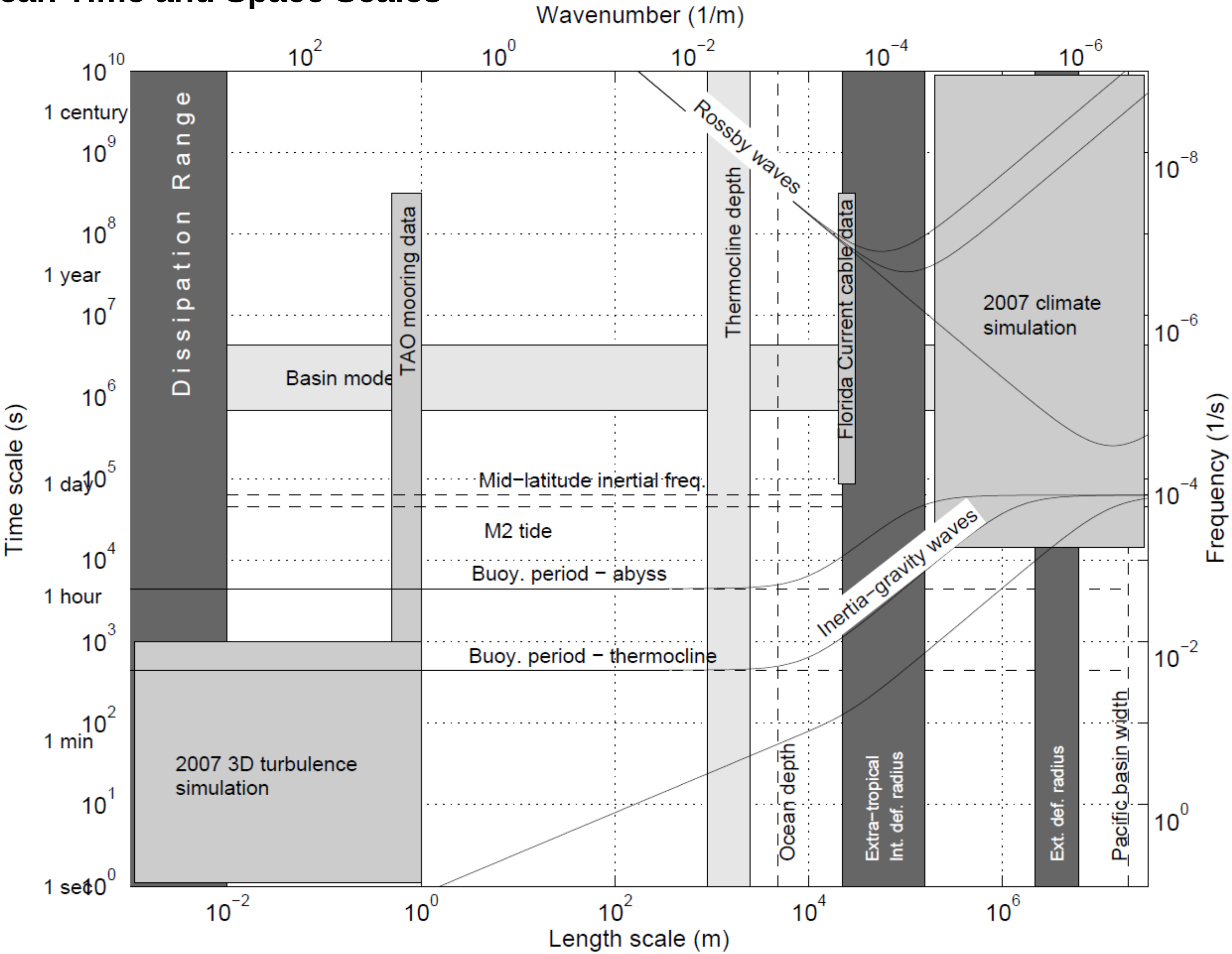
MUR L6 – 64 km resolution

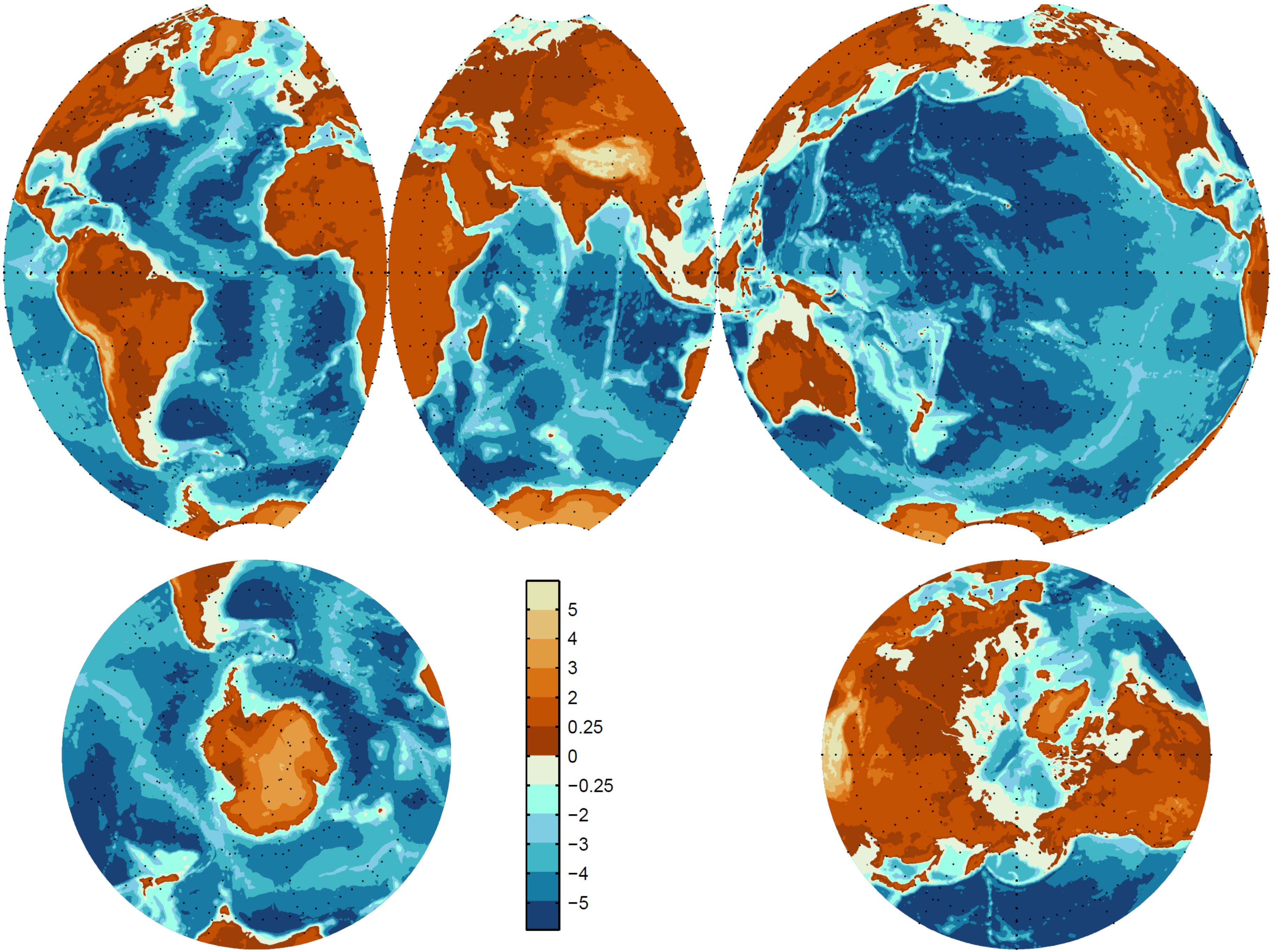


http://podaac.jpl.nasa.gov/Multi-scale_Ultra-high_Resolution_MUR-SST

MUR L12 – 1 km resolution

Ocean Time and Space Scales




5
4
3
2
0.25
0
−0.25
−2
−3
−4
−5

Difficulties in Measuring Ocean Flow

Almost no remote sensing below surface
➔Point measurements (thermal wind, floats, current meters)
➔ aliasing from interannual, seasonal, and mesoscale
(exception: acoustic tomography
theoretically promising, not used so much so far)

Velocity/mass-transport dilemma:
(small velocity error) X (over large depth range) = big transport error

Satellite altimeters have greatly improved in last decade
but O(few cm) errors in variability and somewhat larger in mean

WOCE (World Ocean Circulation Experiment)
greatly expanded data in last 10 yr
attempted transition from qualitative to quantitative
partially successful?

A Quick Review of Some Ocean Dynamics

0. Equations of Motion

$$\underline{\text{Momentum equation}}: \quad \frac{D\mathbf{u}}{Dt} = -\frac{1}{\rho_0}\nabla p - \frac{\rho}{\rho_0} g\hat{\mathbf{z}} - 2\boldsymbol{\Omega}\times\mathbf{u} + \nabla\cdot\mathbf{D_u}(\mathbf{u})$$

$$\underline{\text{Incompressibility}}: \quad \nabla\cdot\mathbf{u} = 0$$

$$\underline{\text{Heat conservation}}: \quad \frac{D\theta}{Dt} = \nabla\cdot\mathbf{D}_\theta(\theta) + \mathcal{F}_\theta$$

$$\underline{\text{Salt conservation}}: \quad \frac{DS}{Dt} = \nabla\cdot\mathbf{D}_S(S)$$

$$\underline{\text{Equation of state}}: \quad \rho = \rho\left(\theta, S, z\right),$$

…and Boundary Conditions

Variable	**Solid bndries** (usual)	**Surface bndry** (usual)
$\vec{u}\cdot\hat{n}$ (normal vel.)	No flow	No flow
$\vec{u}\cdot\hat{t}$ (tangent vel.)	No-slip, others	Given flux (=wind stress)
S salinity	No flux	S or flux may be given
θ temperature	No flux	θ or heat flux given

2. Decomposition of flow in rotating system

$$\frac{\partial \mathbf{u}_H}{\partial t} + \mathbf{u}_H \cdot \nabla \mathbf{u}_H = -\frac{1}{\rho_0}\underbrace{\nabla_H p - f\hat{\mathbf{z}}}_{\text{geostrophy}}\overbrace{\times\, \mathbf{u}_H + \frac{\partial}{\partial z}\left(\nu_v \frac{\partial \mathbf{u}_H}{\partial z}\right)}^{\text{Ekman}} + \nabla_H \cdot (\nu_h \nabla \mathbf{u}_H) + w\frac{\partial \mathbf{u}_H}{\partial z}.$$

Rotating sphere with period T and rotation rate $\Omega = 2\pi/T$ with θ = latitude

Coriolis parameter: $f = 2\Omega \sin(\theta)$

General circulation (speed U, horizontal length scale L)
usually concerned with small Rossby Number $\quad Ro = \frac{U}{fL} \ll 1$

A useful decomposition of ocean flow:

$$\vec{u} = \vec{u}_{\text{Geostrophic}} + \vec{u}_{\text{Ekman}} + \vec{u}_{\text{Other}}$$

Ekman component only important near surface (and sometimes near bottom), given by

$$\int \vec{u_E} dz = \frac{\vec{\tau} \times \hat{z}}{f\rho}$$

For large-scale flow, generally ignore $\vec{u}_{\text{O}}$ due to nonlinear terms, time-variations, and viscosity.

2. Hydrostatic Balance

$$\frac{\partial p}{\partial z} + \rho g = 0. \quad \Longrightarrow \quad p = -\int_z^\eta g\rho dz.$$

constant reference density ρ_0 and varying part $\rho'(\mathbf{x}, t)$,

$$\rho = \rho_0 + \rho'. \quad \Longrightarrow \quad p = p_0(z) + p'(\mathbf{x}, t)$$

$p_0(z \leq 0) = -g\rho_0 z$ is a background pressure that always increases with depth

p' is the **dynamic pressure** $\Longrightarrow$ $p'(\mathbf{x}) = \int_0^\eta g\rho_0 dz + \int_z^\eta g\rho' dz.$

"barotropic" **"baroclinic"**

Use these terms w/ caution.
Different authors use them
somewhat differently.

3. Components of Geostrophy

$$f\hat{\mathbf{z}} \times \mathbf{u}_G = -\frac{1}{\rho}\nabla_h p,$$

$$f\mathbf{u}_G = \hat{\mathbf{z}} \times \frac{1}{\rho}\nabla_h p.$$

Or in Cartesian coordinates

$$\begin{cases} -fv_G = -\dfrac{1}{\rho}\dfrac{\partial p}{\partial x} \\ fu_G = -\dfrac{1}{\rho}\dfrac{\partial p}{\partial y}, \end{cases}$$

Using previous decomposition of pressure ➔
and taking $\rho(x,y,z) \approx \rho_0$

$$p'(\mathbf{x}) = \int_0^\eta g\rho_0 dz + \int_z^\eta g\rho' dz.$$

$$\vec{u}_G = \vec{u}_S(x,y) + \vec{u}_\rho(x,y,z) = \frac{1}{f}\hat{z} \times g\left[\nabla\eta + \frac{1}{\rho_0}\nabla\left(\int_z^0 \rho' dz\right)\right]$$

Combine geostrophy w/ hydrostatic relation ($\partial p/\partial z = -\rho g$)

$$f\hat{\mathbf{z}} \times \frac{\partial \mathbf{u}}{\partial z} = \frac{g}{\rho}\nabla_p \rho',$$

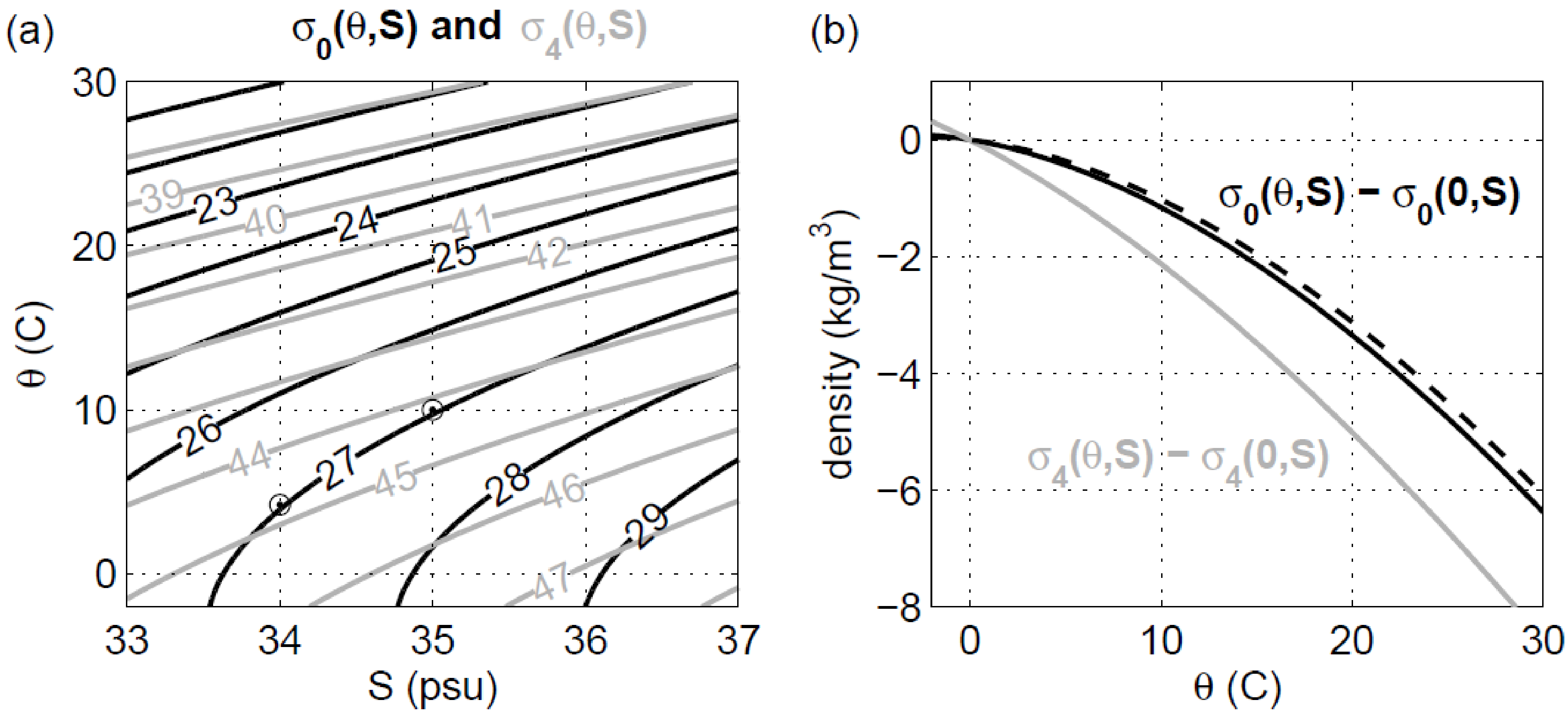
$$f\frac{\partial \mathbf{u}}{\partial z} = -\frac{g}{\rho}\hat{\mathbf{z}} \times \nabla_p \rho',$$

So if we know ρ but not η, we can still get geostrophic vertical shear

Don’t forget:

- sea surface slope does not ALWAYS balance isopycnal slope
- surface slope gives surface current, NOT depth-avg current
  ($\int\rho_x\, dz \neq 0$ for instance)

Density of Sea Water



$$\sigma_0 = [\rho(\theta, S, 0 \text{ db}) - 1000] \ \text{kgm}^{-3}.$$

$$\rho = \rho_0 + \rho_1(p) - \alpha(\theta, S, p)\theta + \beta(\theta, S, p)S$$

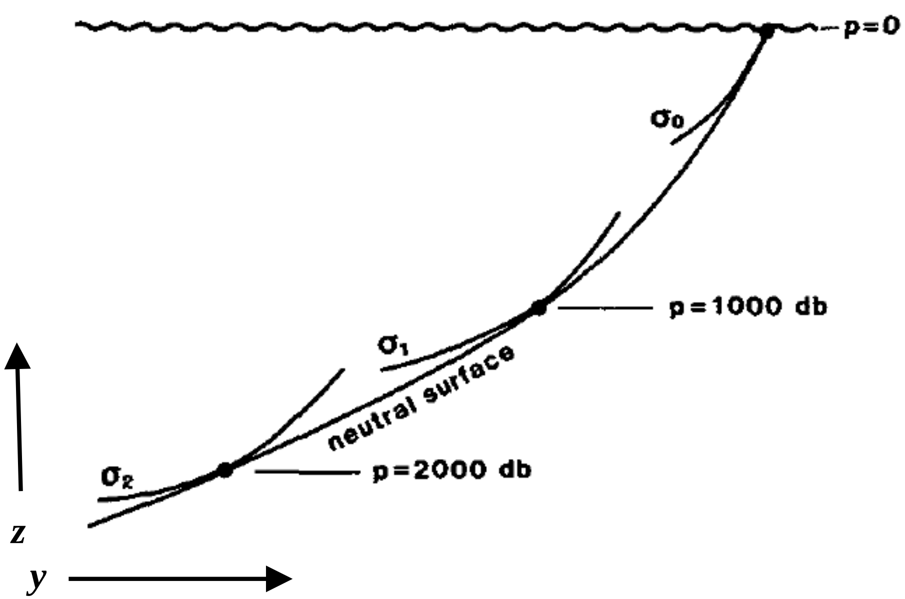
$\beta \approx .8$ kg m^{-3} psu^{-1} to an accuracy of about 10%.

McDougall, 1987: Neutral Surfaces, *JPO*, 17, 1950 – 1964.

Potential density allows us to compare ρ parcels at 2 different depths.
But which depth? Parcel 1, parcel 2, or somewhere inbetween?
If they have different θ and S,

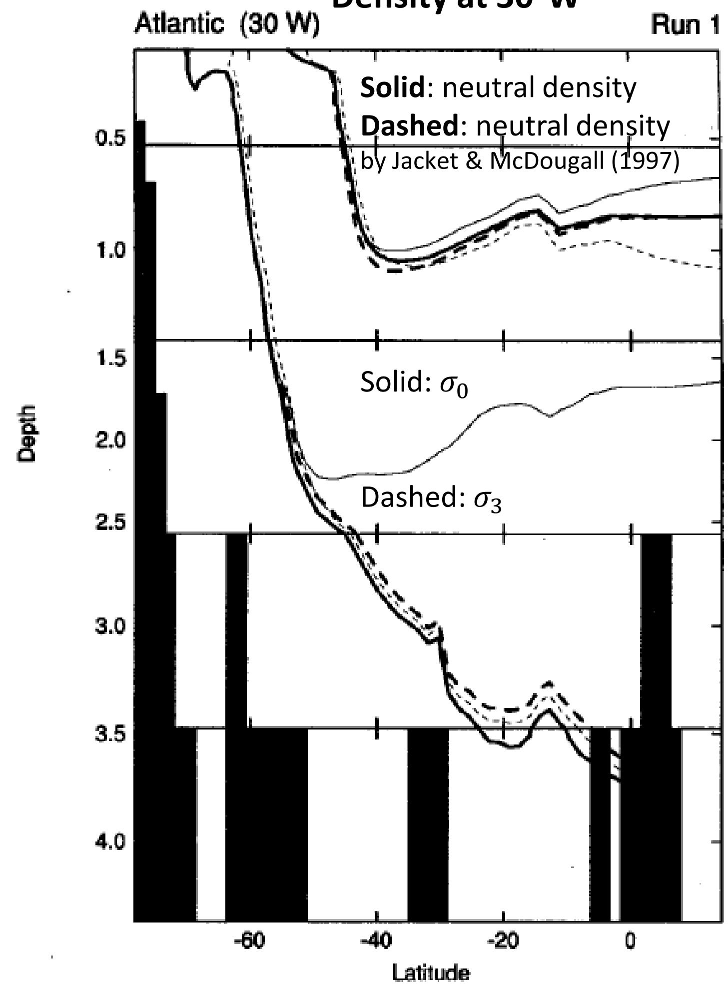
- #1 might be denser at some level
- #2 might be denser at another level

Solution: construct **neutral surface** linking two regions of ocean.
At every point, surface is tangent to potential density surface **Referenced to that depth**.
[Warning: don't confuse this with ***in situ* density**]
This is surface a parcel would slide along **if no change in** (θ, S) **.**



OK, but how much does this matter really?
Usually not too much.
Sometimes a lot.

Density at 30°W



Hirst, Jackett & McDougall, 1996: JPO