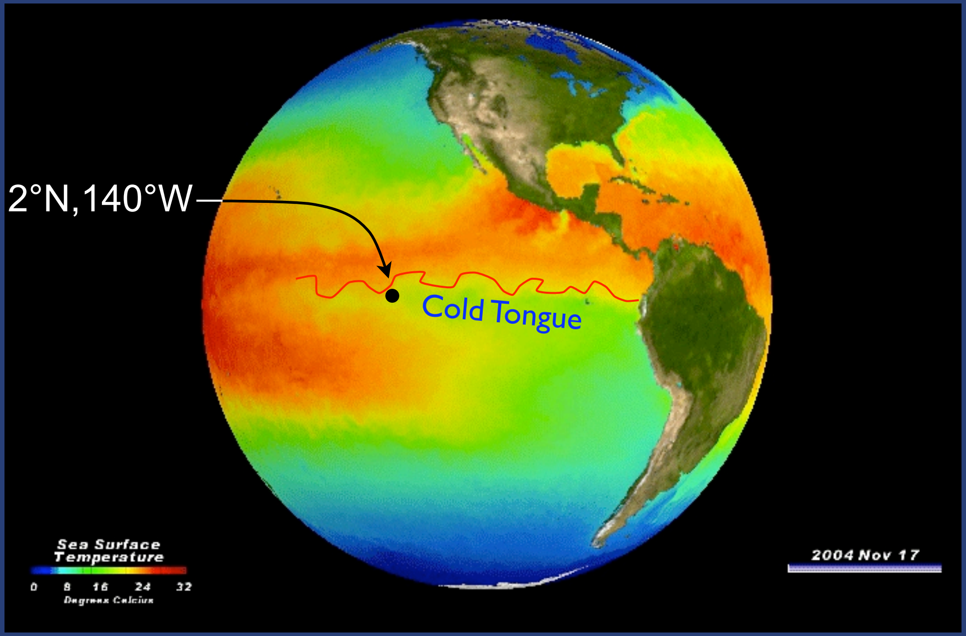
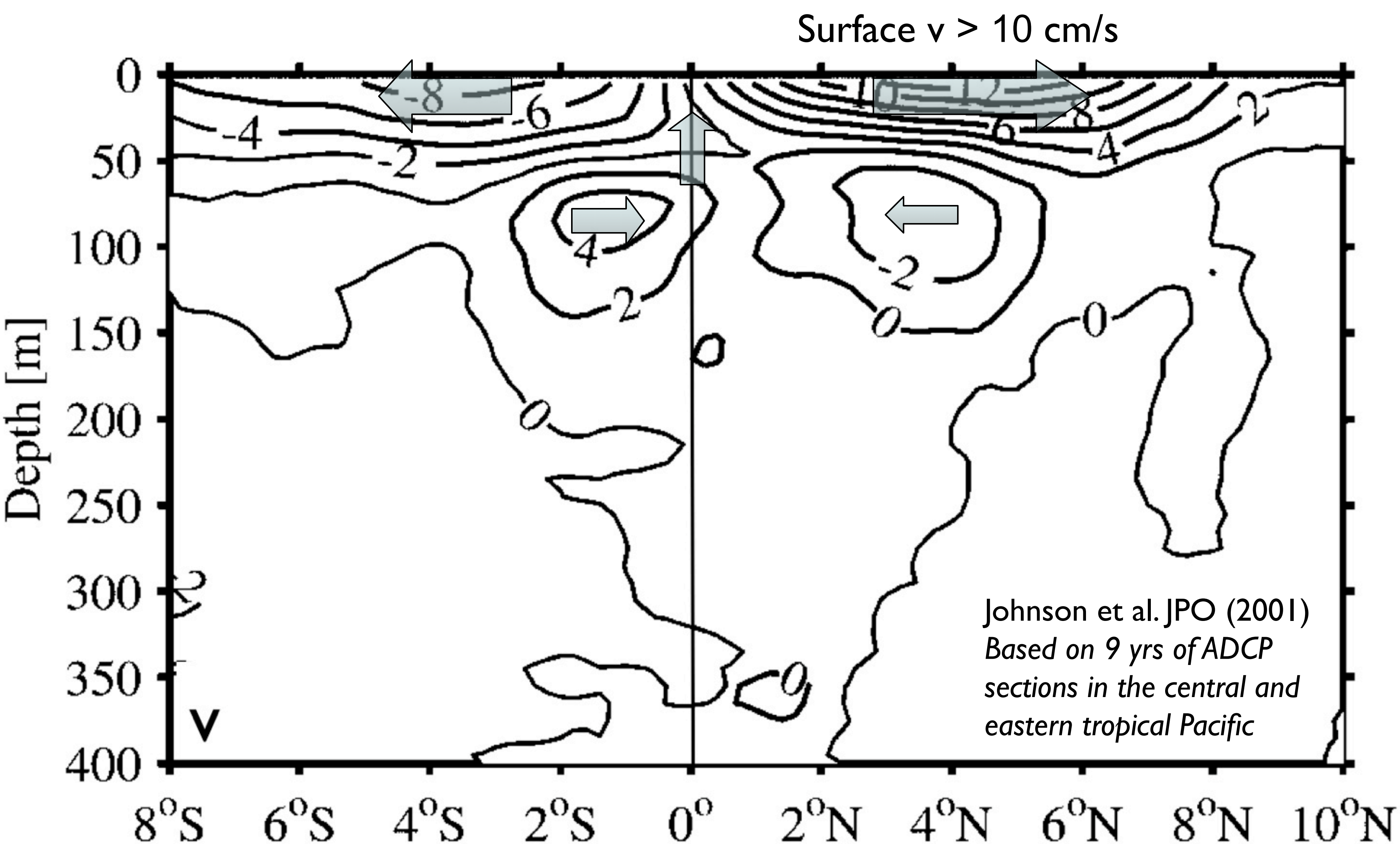
Near-surface shear flow in the tropical Pacific cold tongue front



Meghan F. Cronin
William S. Kessler
NOAA Pacific Marine Environmental Laboratory

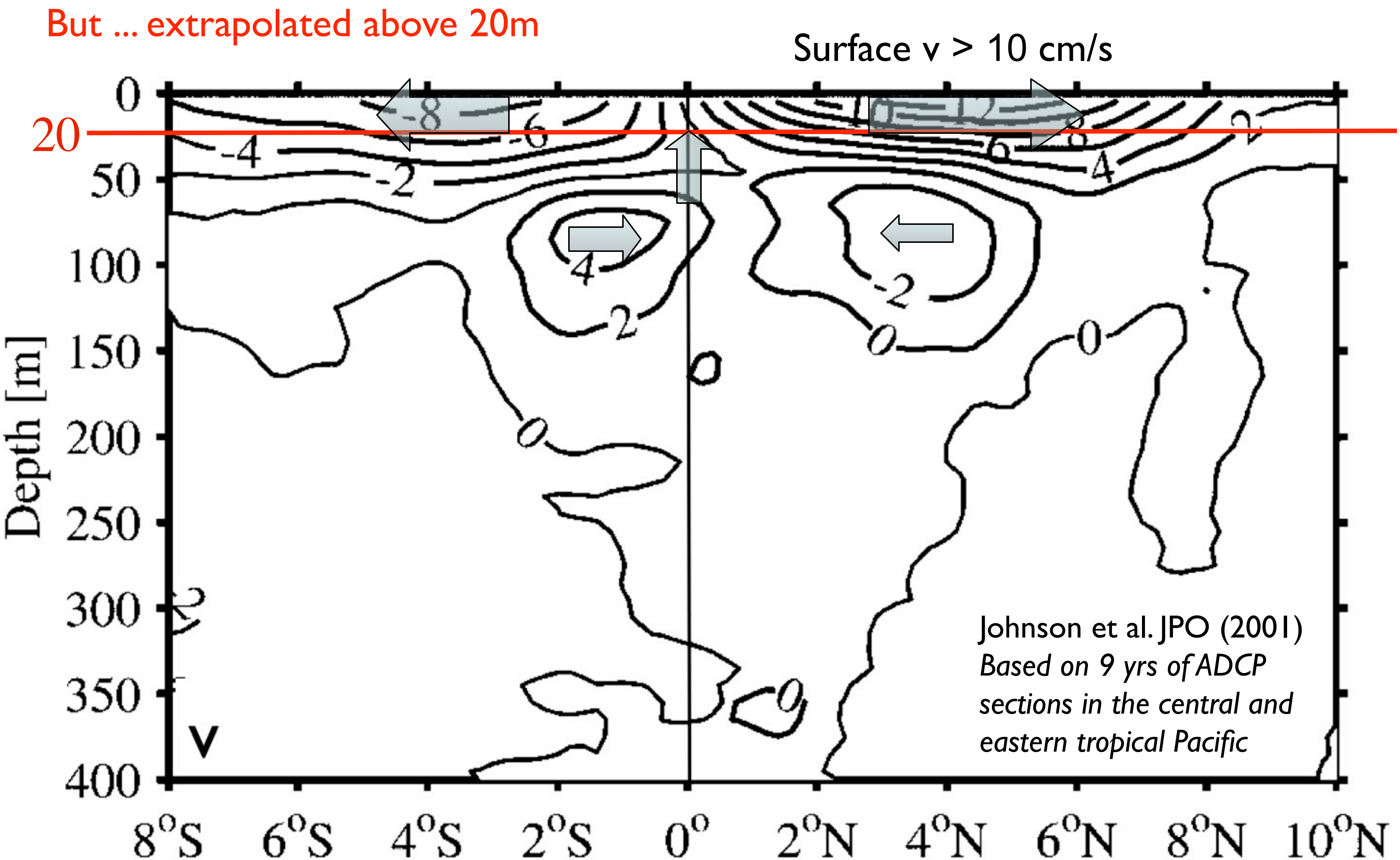
Cronin, M. F. and W. S. Kessler, 2009: Near-surface shear flow in the tropical Pacific cold tongue front. *J. Phys. Oceanogr.*, 39, 1200-1215.

The "classical" picture of equatorial circulation

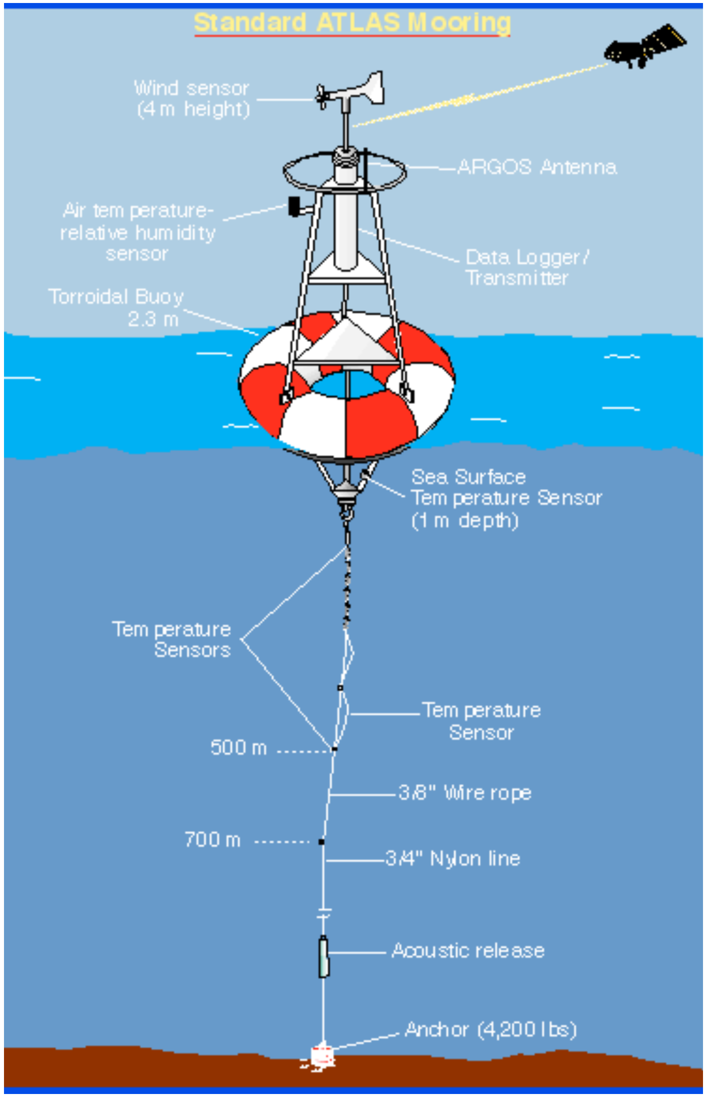


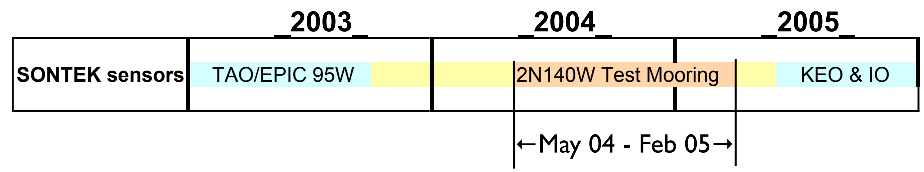
Johnson et al. JPO (2001)
Based on 9 yrs of ADCP sections in the central and eastern tropical Pacific

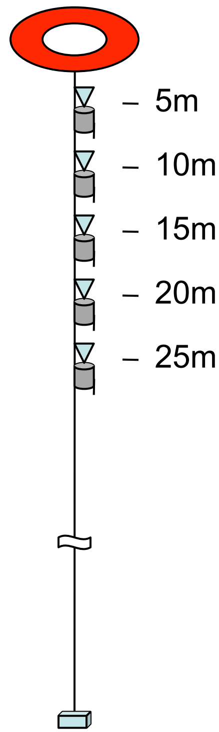
Neither shipboard, nor moored upward-looking ADCPs measure currents above 20m. *Is there shear above 25m?*



9 months of near-surface velocity and temperature: 2°N,140°W

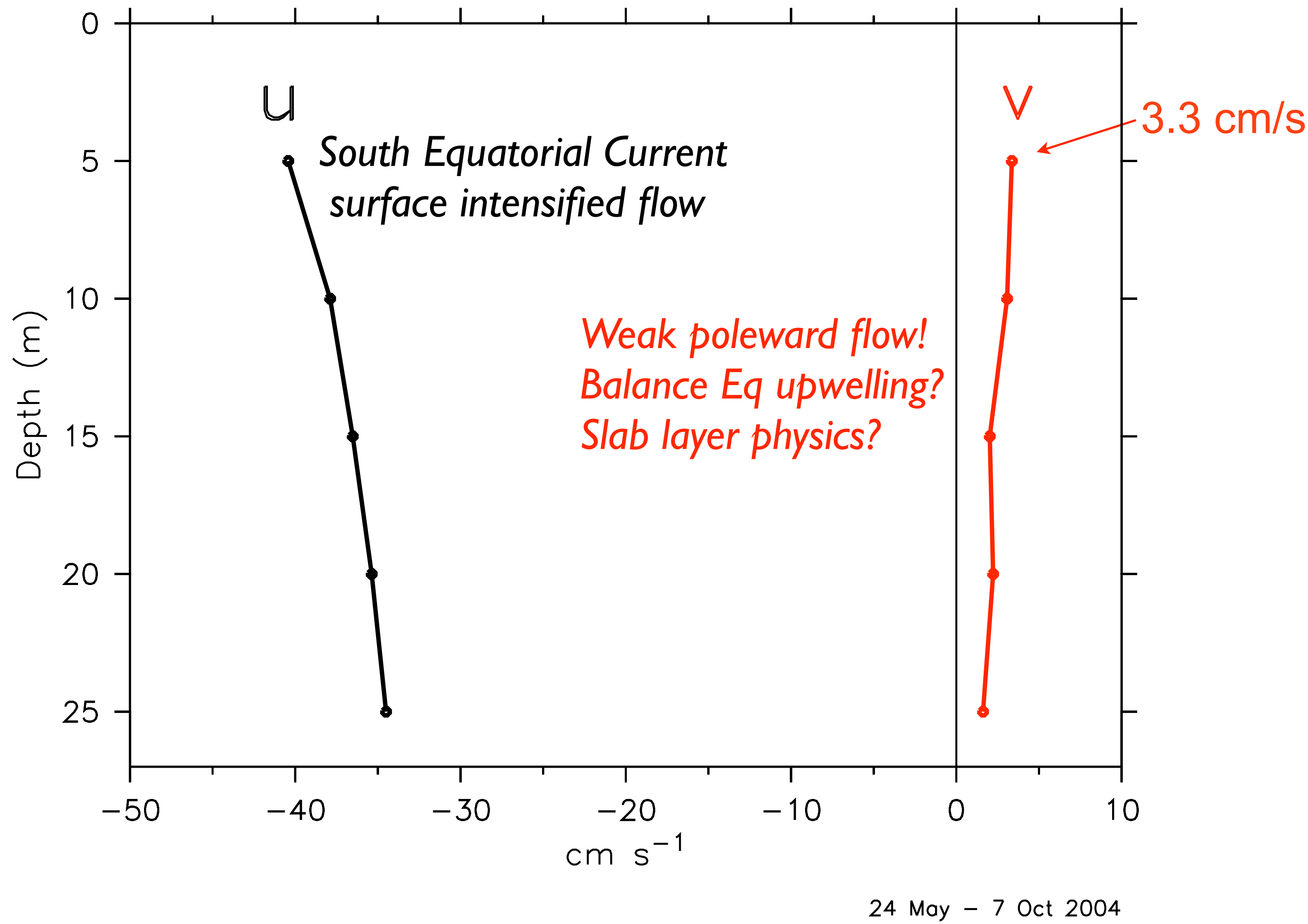




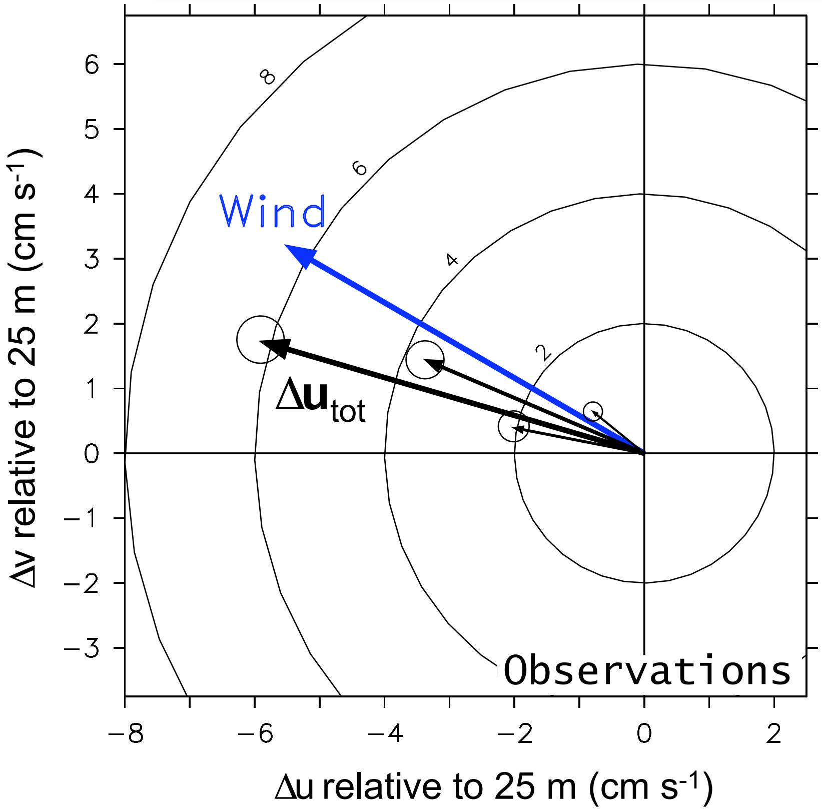


- *5 Sonteks (acoustic Doppler current meters) were placed on a test mooring near the 2°N,140°W TAO mooring.*
- *Each Sontek had a thermistor.*
- *Sampling: T and Met: 10 minutes, currents: 20 minutes.*

Mean near-surface currents at 2°N, 140°W



Observations: wind, currents



Mean for 24-May-2004 to 7-Oct-2004

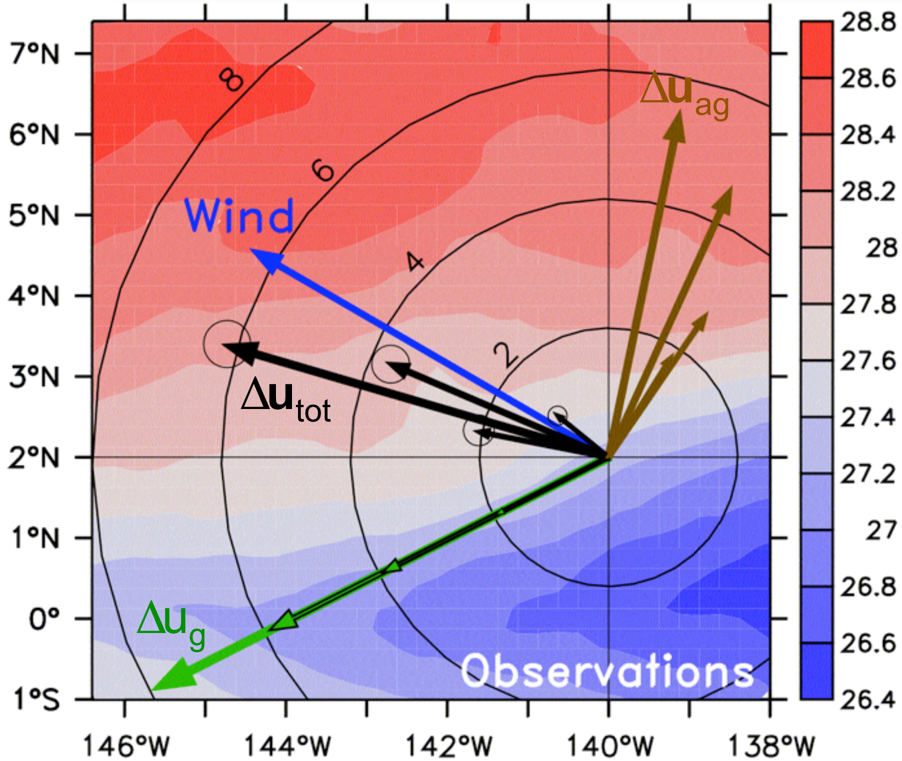
- Vectors (circular scales) are:

  Blue: wind stress (m s^{-1})

  Black: current ***shears***:

  $$\Delta \mathbf{u} = \mathbf{u} - \mathbf{u}(25\mathrm{m}) \quad (\mathrm{cm/s})$$

Observations: SST, wind, currents



Mean for 24-May-2004 to 7-Oct-2004

- Color shading is a map of SST (Mean front near mooring site)

Add current ***shears***:

u−u(25m) (cm/s)

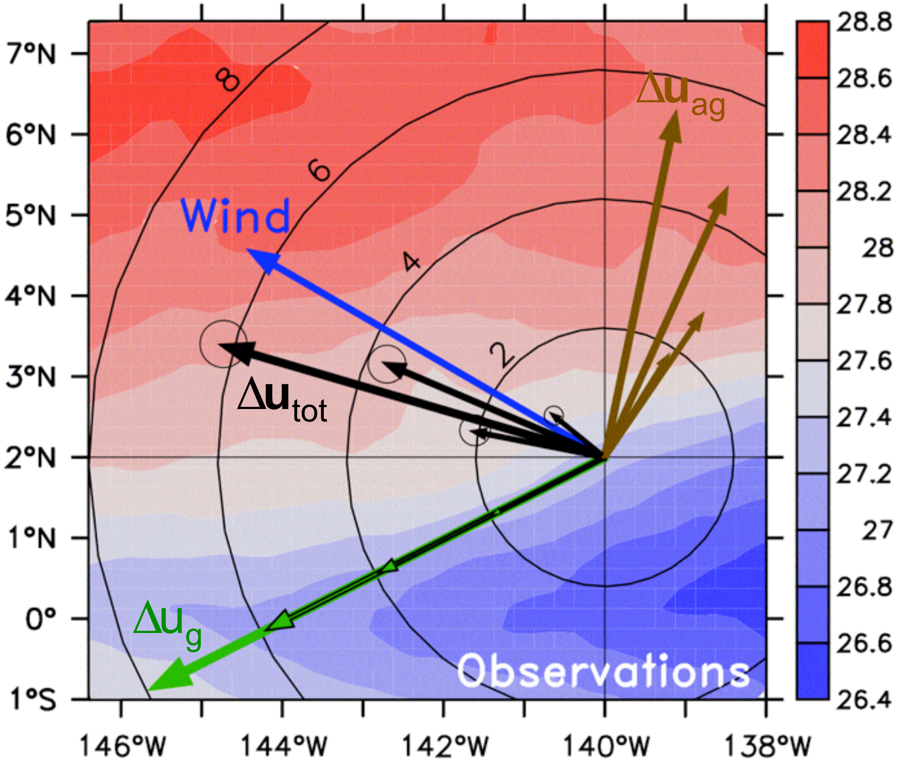
Black: Total Δu

Green: Geostrophic

Brown: Ageostrophic = $\Delta u - \Delta u_g$

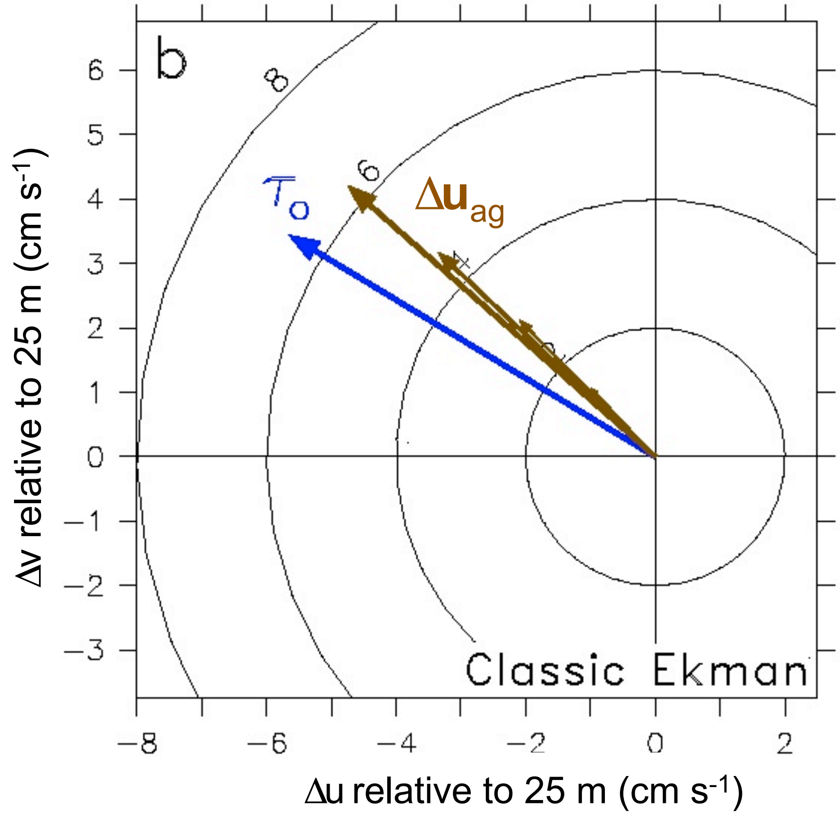
Observed ageostrophic currents relative to 25 m have Ekman-like spiral 70° to right of wind. But…

Observations



Mean for 24-May-2004 to 7-Oct-2004

Classic Ekman model



(Ekman depth D_{ek}= 80 m,
with $\nu \sim 1.6\times10^{-4}$ from observed shear)

Classic Ekman spiral has Δu_{ag} shear aligned slightly to the right of the wind stress.

“**Classic Ekman Model**” (Ekman 1905):

Assume steady, linear flow; uniform density and viscosity; driven by surface wind stress, no drag at $z = -H \sim -\infty$. Solve for $\boldsymbol{u_a}(z)$

Equation of motion:

$$ifu = -\frac{1}{\rho}\nabla P + \nu\frac{\partial^2 u}{\partial z^2}$$

Complex notation:
$\boldsymbol{u} = u + iv$
$\nabla = \partial/\partial x + i\partial/\partial y$

Boundary conditions:

$$at\ \ z = 0: \quad \frac{\partial u}{\partial z} = \frac{\tau_0}{\rho\nu}$$

$$at\ \ z = -H: \quad u = 0$$

(Uniform density: No geostrophic shear)

$$note: \quad u = u_g + u_a,\ where: \quad u_g = \frac{i}{\rho f}\nabla P \quad and \quad \frac{\partial u_g}{\partial z} = \frac{ig\alpha}{\rho f}\nabla T = 0$$

"**Classic Ekman Model**" (Ekman 1905):

Assume steady, linear flow; uniform density and viscosity; driven by surface wind stress, no drag at $z = -H \sim -\infty$. Solve for $\boldsymbol{u_a}(z)$

The familiar Classic Ekman equations:

Equation of motion:

$$ifu = -\frac{1}{\rho}\nabla P + \nu\frac{\partial^2 u}{\partial z^2} \quad \rightarrow \quad \boxed{ifu_a = \nu\frac{\partial^2 u_a}{\partial z^2}}$$

Boundary conditions:

$$at\ z = 0: \quad \frac{\partial u}{\partial z} = \frac{\tau_0}{\rho\nu} \quad \rightarrow \quad at\ z = 0: \quad \frac{\partial u_a}{\partial z} = \frac{\tau_0}{\rho\nu}$$

$$at\ z = -H: \quad u = 0 \quad \rightarrow \quad at\ z = -H: \quad u_a = 0$$

$$note: \quad u = u_g + u_a, \ where: \quad u_g = \frac{i}{\rho f}\nabla P \ \ and \ \ \frac{\partial u_g}{\partial z} = \frac{ig\alpha}{\rho f}\nabla T = 0$$

Classic Ekman Model (Ekman 1905)

Assumes steady, linear flow; with uniform density and viscosity; subject to wind stress at surface, no drag at $z = -H \sim -\infty$.

$$D_{ek} = \sqrt{\frac{2\nu}{f}}$$

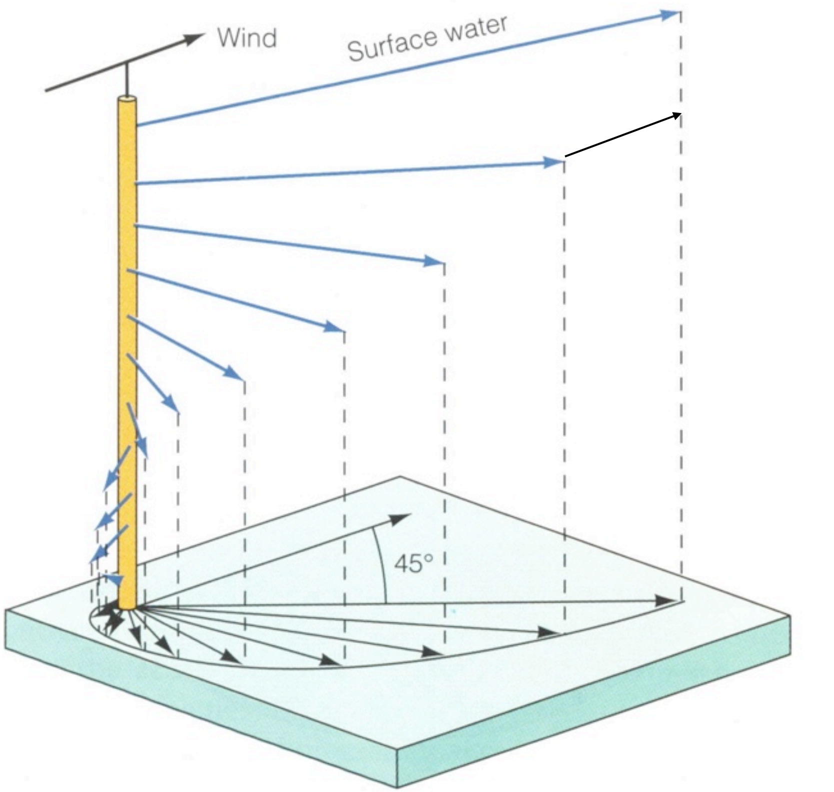
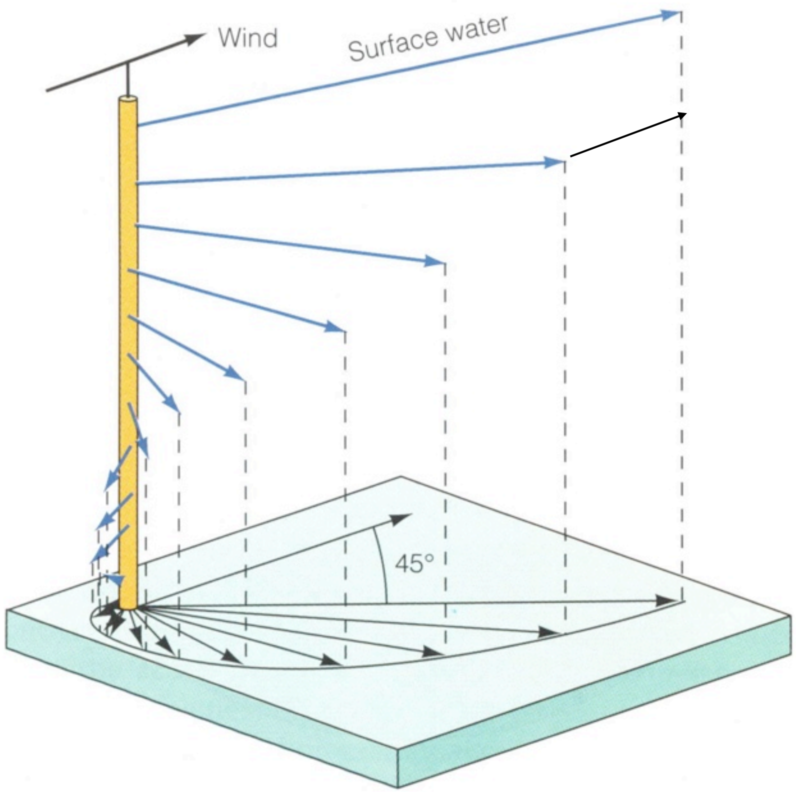
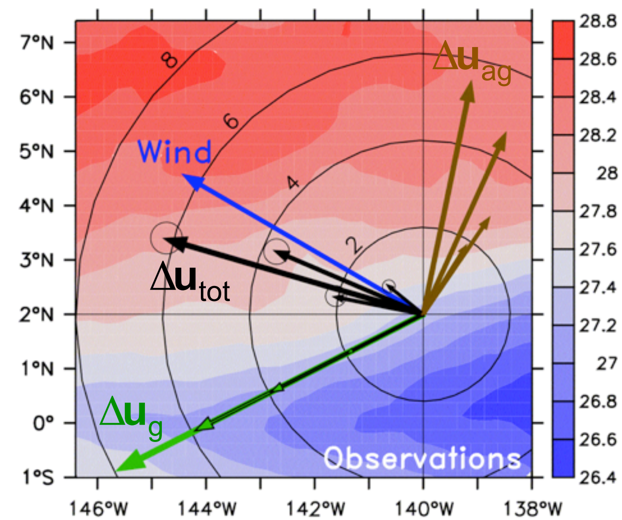


Figure from google image
http://www.eeb.ucla.edu/test/faculty/nezlin/PhysicalOceanography.htm

Classic Ekman Model (Ekman 1905)

Assumes steady, linear flow; with uniform density and viscosity; subject to wind stress at surface, no drag at $z = -H \sim -\infty$.





Geostrophic "thermal wind" shear is larger than the observed shear.

Figure from google image
http://www.eeb.ucla.edu/test/faculty/nezlin/PhysicalOceanography.htm

“Frontal Ekman Model”: Assume steady, linear flow; with uniform viscosity; driven by wind stress at surface *and geostrophic flow at $z = -H$; in a front that is uniform with depth.*

Equation of motion:

$$ifu = -\frac{1}{\rho}\nabla P + \nu\frac{\partial^2 u}{\partial z^2}$$

Boundary conditions:

$$at\ \ z = 0: \quad \frac{\partial u}{\partial z} = \frac{\tau_0}{\rho\nu}$$

$$\boxed{at\ \ z = -H: \quad u = u_g}$$

$$u = u_g + u_a, \quad where: \quad u_g = \frac{i}{\rho f}\nabla P \quad and \quad \boxed{\frac{\partial u_g}{\partial z} = \frac{ig\alpha}{\rho f}\nabla T \equiv vertically\ uniform}$$

"Frontal Ekman Model": Assume steady, linear flow; with uniform viscosity; driven by wind stress at surface *and geostrophic flow at $z = -H$; in a front that is uniform with depth.*

Equation of motion:

$$ifu = -\frac{1}{\rho}\nabla P + \nu \frac{\partial^2 u}{\partial z^2} \quad \rightarrow \quad \boxed{ifu_a = \nu \frac{\partial^2 u_a}{\partial z^2}}$$

Boundary conditions:

$\boldsymbol{u} = \boldsymbol{u}_g + \boldsymbol{u}_a$

$$at\ \ z = 0: \quad \frac{\partial \boldsymbol{u}}{\partial z} = \frac{\tau_0}{\rho\nu} \quad \rightarrow \quad at\ \ z = 0: \quad \rho\nu\frac{\partial u_a}{\partial z} = \tau_0 - \rho\nu\frac{\partial u_g}{\partial z}$$

$$at\ \ z = -H: \quad u = u_g \quad \rightarrow \quad at\ \ z = -H: \quad u_a = 0$$

"Frontal Ekman Model": Assume steady, linear flow; with uniform viscosity; driven by wind stress at surface *and geostrophic flow at $z=-H$; in a front that is uniform with depth.*

Equation of motion:

$$ifu = -\frac{1}{\rho}\nabla P + \nu\frac{\partial^2 u}{\partial z^2} \quad \rightarrow \quad \boxed{ifu_a = \nu\frac{\partial^2 u_a}{\partial z^2}}$$

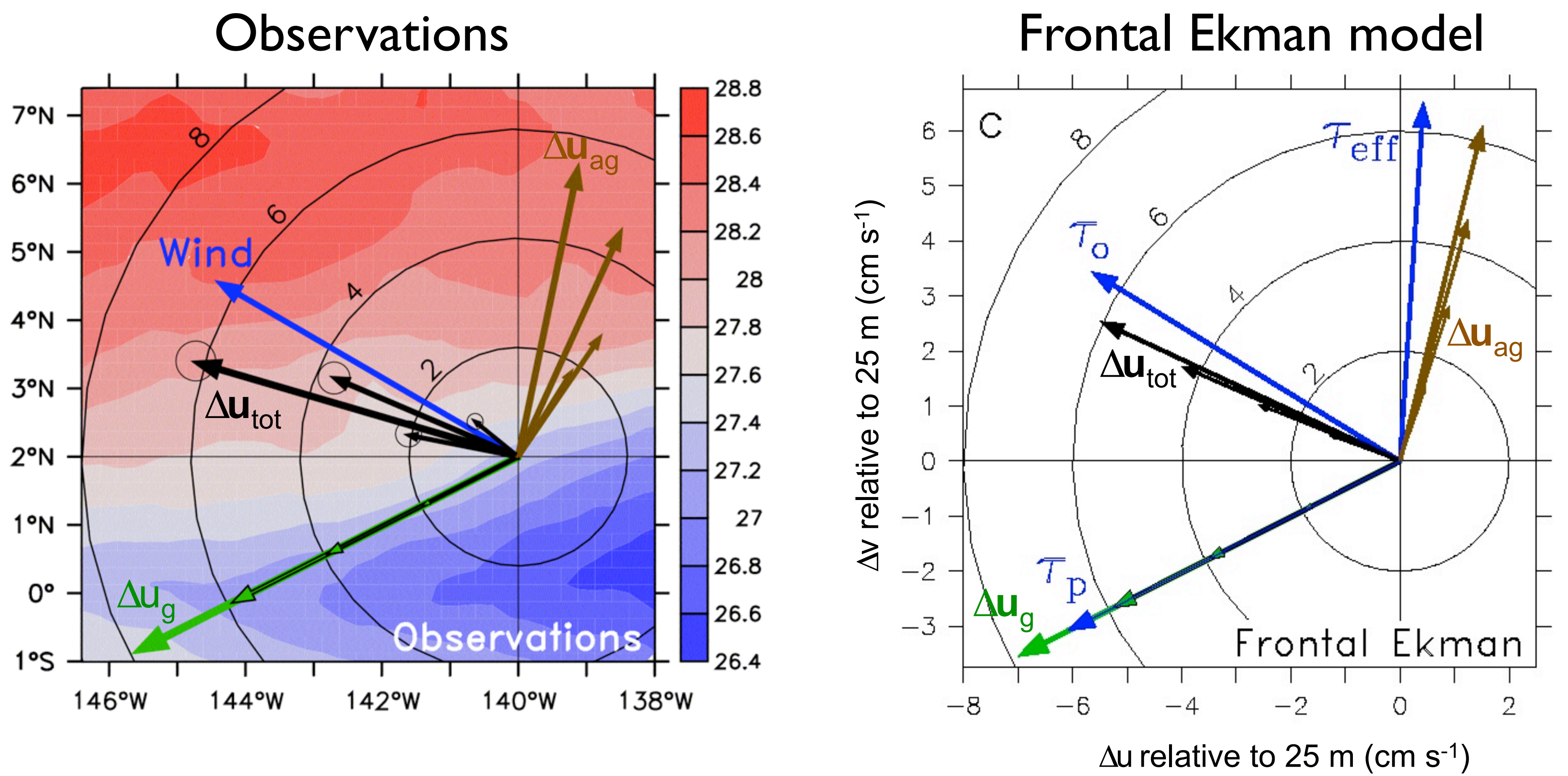
Boundary conditions:

$\boldsymbol{u} = \boldsymbol{u}_g + \boldsymbol{u}_a$

$$at\ z=0: \quad \frac{\partial \boldsymbol{u}}{\partial z} = \frac{\tau_0}{\rho\nu} \quad \rightarrow \quad at\ z=0: \quad \overbrace{\rho\nu\frac{\partial u_a}{\partial z} = \tau_0 - \underbrace{\rho\nu\frac{\partial u_g}{\partial z}}_{\tau_p}}^{\tau_{eff}}$$

$$at\ z=-H: \quad u = u_g \quad \rightarrow \quad at\ z=-H: \quad u_a = 0$$

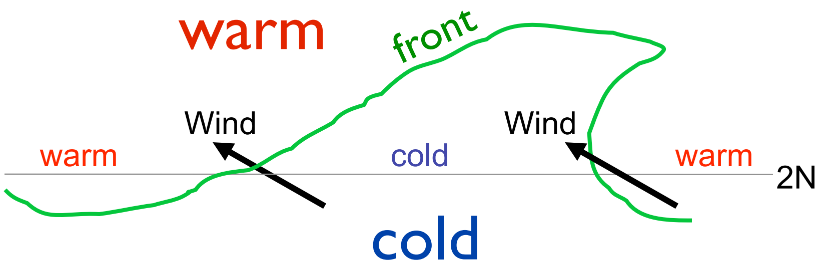
We can picture this as TWO Ekman Spirals: the usual one forced by the wind stress, and a second one required by the geostrophic shear at $z=0$. The spiral "forced" by τ_p turns towards the cold side of the front.



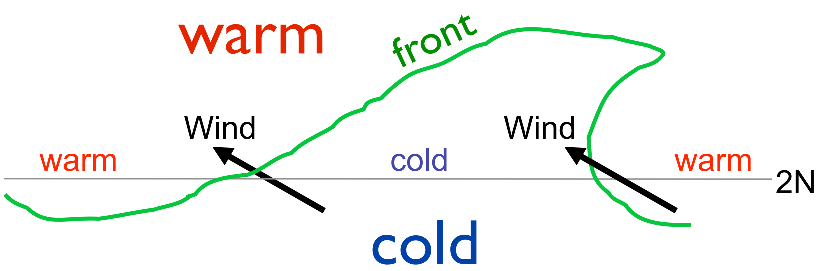
The ageostrophic Ekman response depends upon wind stress…
AND strength / orientation of the front relative to the wind: $\tau_{eff} = \tau_0 - \tau_p$

The Ekman response is reduced when winds blow along a front.

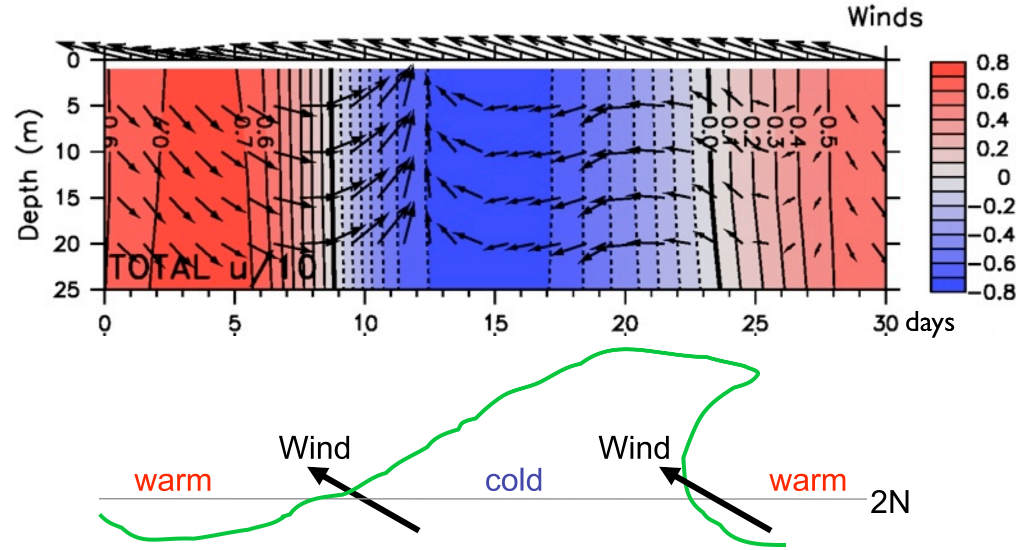
Tropical Instability Waves show how orientation of front relative to the wind affects ageostrophic shear



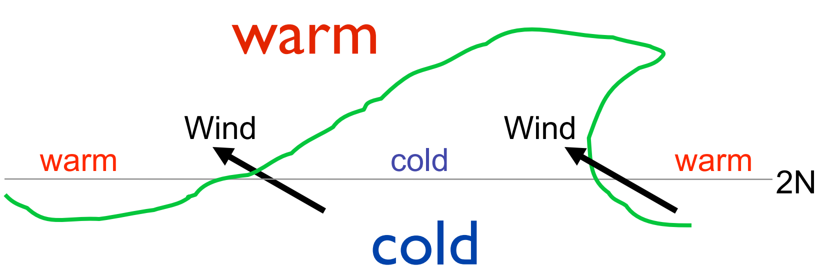
Tropical Instability Waves show how orientation of front relative to the wind affects ageostrophic shear



Composite TIW temperature and velocity



Tropical Instability Waves show how orientation of front relative to the wind affects ageostrophic shear

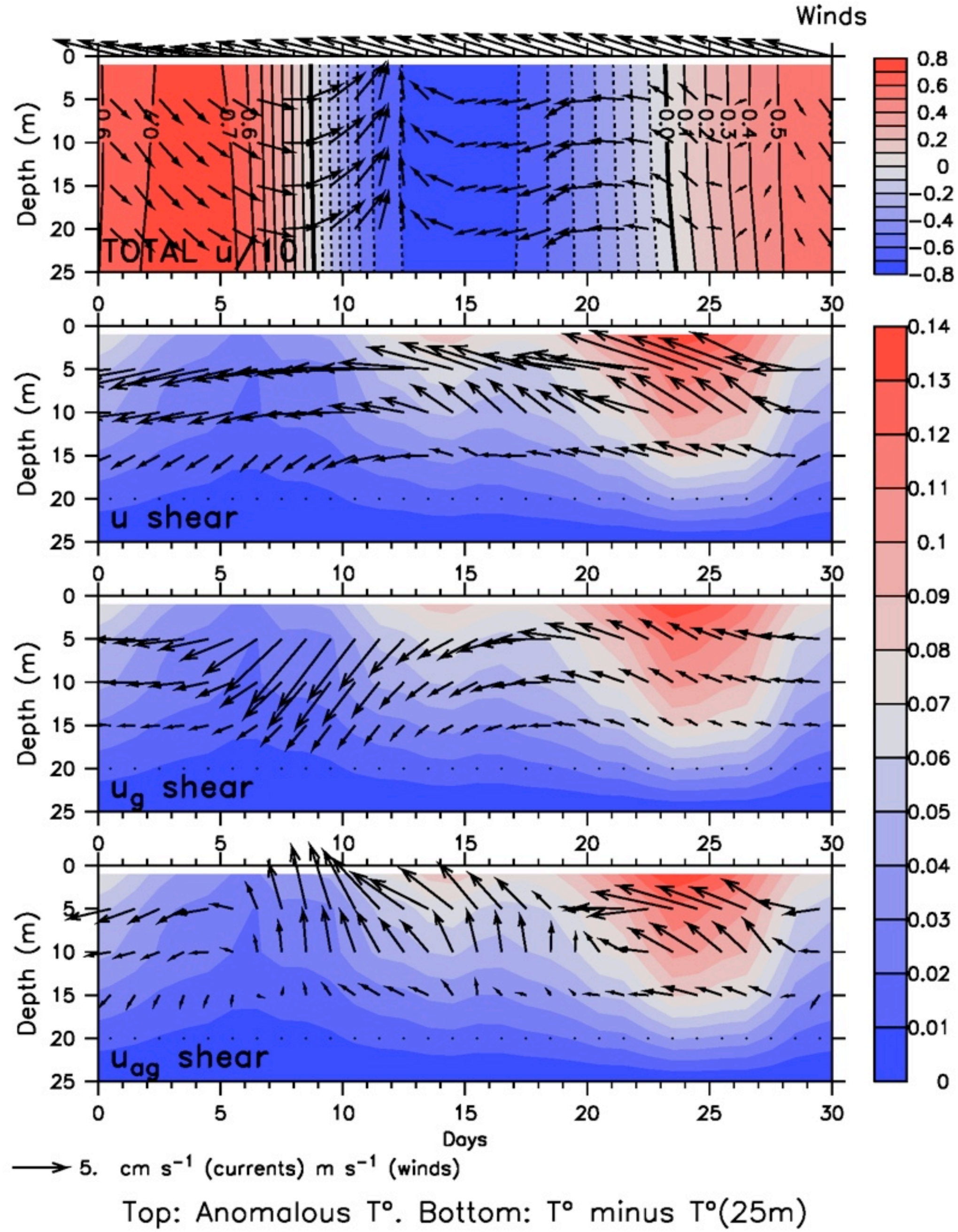


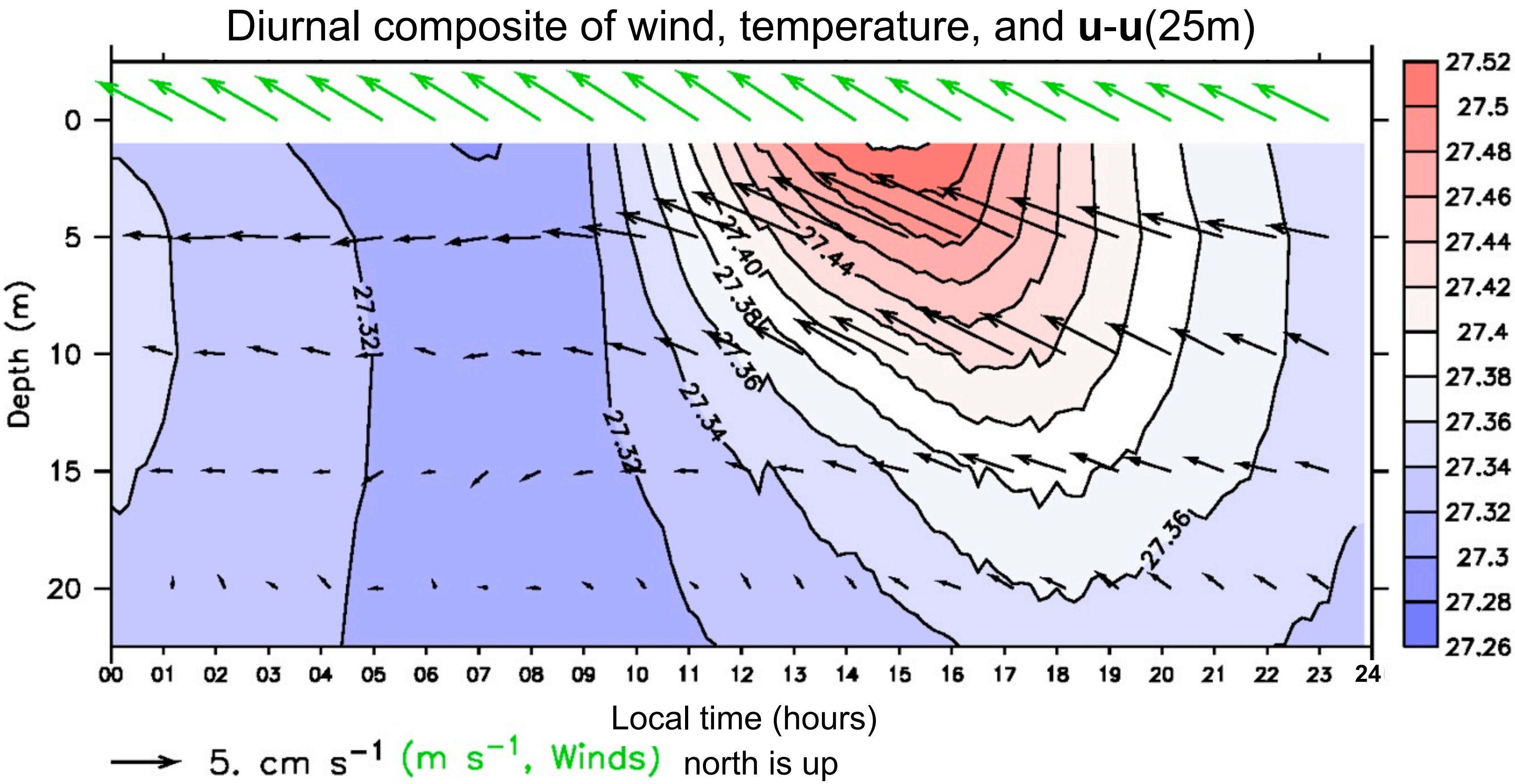
Strongest ageostrophic shear when wind blows across front

Weaker ageostrophic shear when winds are aligned with front.

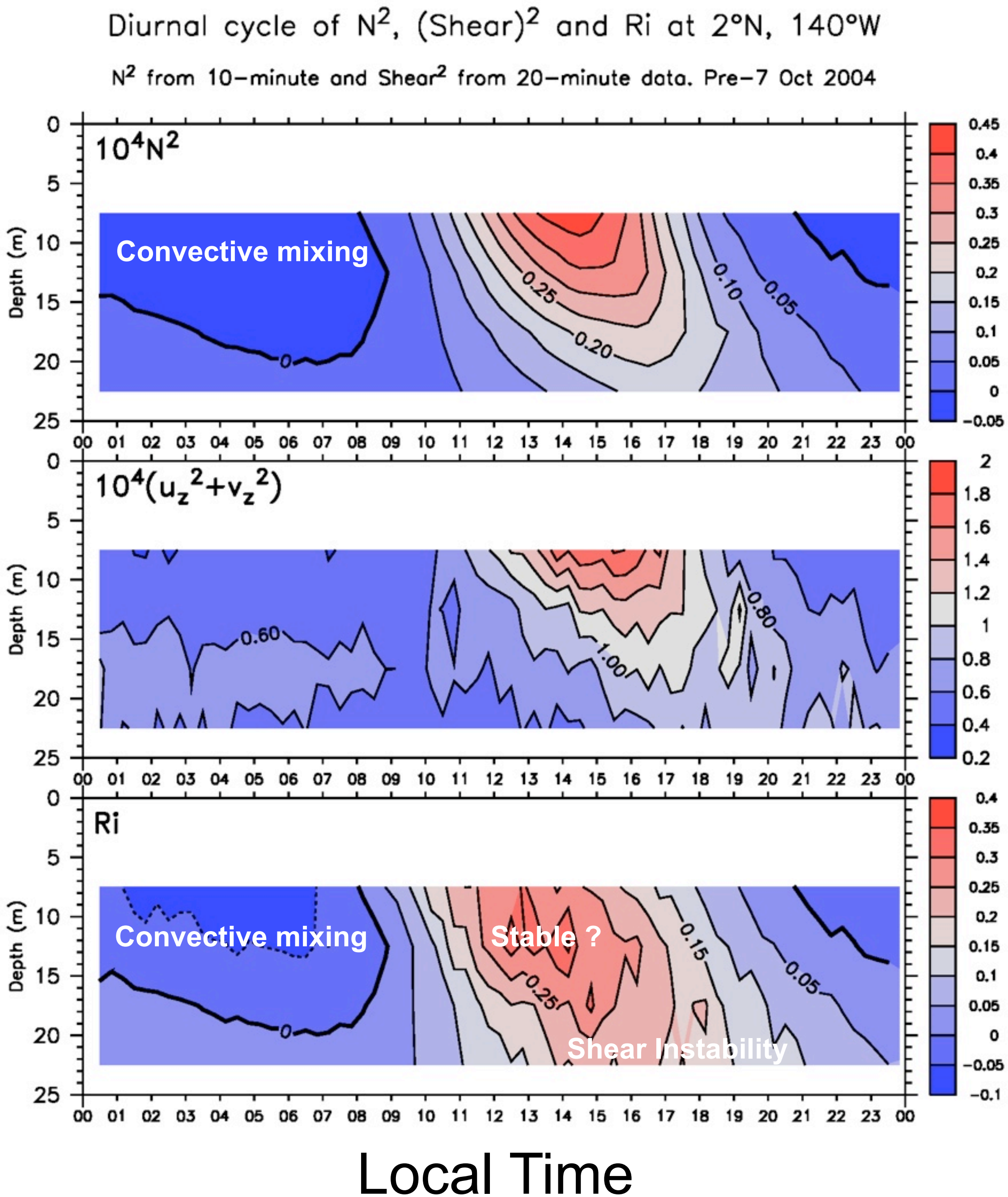
Total shear stronger when stratification is stronger.

Composite TIW temperature and velocity





- *At 1600 local, currents at 5 m are 12 cm/s stronger than at 25m and are oriented in direction of wind. Nighttime shear is weak.*
- *Even weak daytime restratification can cause diurnal jet.*



- Convective mixing down to 10-20 m for 2100-0800 local.
- Ri>0.25 (stable?) near 10m for 1200-1500 local.
- Ri<0.25 (shear instability?) propagates downward?

Viscosity is likely to be larger in the upper 25 m than below due to both nighttime convective mixing and shear instability mixing due to the diurnal jet.

Summary (thus far)

- Wind stress balances the TOTAL surface shear (combined geostrophic and ageostrophic shears).
- The effect of fronts on Ekman spiral is most pronounced at low latitudes.
- Shear is very sensitive to both the horizontal and vertical temperature distribution.

  Very weak daytime stratification (<0.2°C/25m) resulted in a diurnal jet shear of 12 cm/s / 20m at 4 PM local!

What happens at the Equator *(where f = 0)* ?

What happens if viscosity ν is not constant?

$$\tau(z) = \rho\nu(z)\frac{\partial u}{\partial z}$$

Goal: Develop a generalized Ekman model that is valid on the equator and can have a non-uniform viscosity.

We are not the first to try this ...

"Stommel (1960) Model": Assume steady, linear flow; uniform density and viscosity; driven by surface wind stress, and no stress (no shear) at $z = -H$. No flow through eastern or western boundaries. Find $P(x,y), \boldsymbol{u}(y,z)$.

Equation of motion:

$$ifu = -\frac{1}{\rho}\nabla P + \nu \frac{\partial^2 u}{\partial z^2}$$

Boundary conditions:

$$at \;\; z = 0: \quad \frac{\partial u}{\partial z} = \frac{\tau_0}{\rho \nu}$$

$$at \;\; z = -H: \quad \frac{\partial u}{\partial z} = 0$$

Stommel found a solution for total u with constant viscosity and $du_g/dz = 0$.
(Thus all shear within the layer is due to friction = stress divergence)

"Modified Stommel Model" used for OSCAR (Bonjean and Lagerloef 2002): Assume steady, linear flow; uniform viscosity in a front that is uniform with depth; driven by surface wind stress, and no shear at $z = -H$. Find $d\mathbf{u}/dz$.

Equation of motion:

$$if \frac{\partial u}{\partial z} = -\nabla b + \nu \frac{\partial^2}{\partial z^2}\left(\frac{\partial u}{\partial z}\right)$$

Boundary conditions:

$$at \;\; z = 0: \quad \frac{\partial u}{\partial z} = \frac{\tau_0}{\rho \nu}$$

$$at \;\; z = -H: \quad \frac{\partial u}{\partial z} = 0 \quad \rightarrow \quad \frac{\partial u_a}{\partial z} = -\frac{\partial u_g}{\partial z}$$

B&L '02 found an analytic solution for the vertical shear with constant viscosity.

"Modified Stommel Model" used for OSCAR (Bonjean and Lagerloef 2002): Assume steady, linear flow; uniform viscosity in a front that is uniform with depth; driven by surface wind stress, and no shear at $z = -H$. Find $d\mathbf{u}/dz$.

Equation of motion:

$$if \frac{\partial u}{\partial z} = -\nabla b + \nu \frac{\partial^2}{\partial z^2}\left(\frac{\partial u}{\partial z}\right)$$

Boundary conditions:

$$at\ \ z = 0: \quad \frac{\partial u}{\partial z} = \frac{\tau_0}{\rho \nu}$$

Realistic?

$$at\ \ z = -H: \quad \frac{\partial u}{\partial z} = 0 \quad \rightarrow \quad \frac{\partial u_a}{\partial z} = -\frac{\partial u_g}{\partial z}$$

Perhaps a more realistic way to make $\tau_{-H} = 0$ is to have viscosity = 0 at the bottom of the viscous layer, rather than insisting that shear = 0 there.

"**Generalized Ekman Model**": Assume steady, linear flow; with prescribed viscosity that decays to 0 at depth $z = -H$; subject to prescribed buoyancy gradient and wind stress.
Find $\tau(z)$: $\tau(z) = \rho\nu\, d\mathbf{u}/dz$.

Equation of motion:
(vertical shear equation)

$$if\boldsymbol{\tau} = -\rho\nu\nabla b + \nu\frac{\partial^2\boldsymbol{\tau}}{\partial z^2}$$

Boundary conditions:

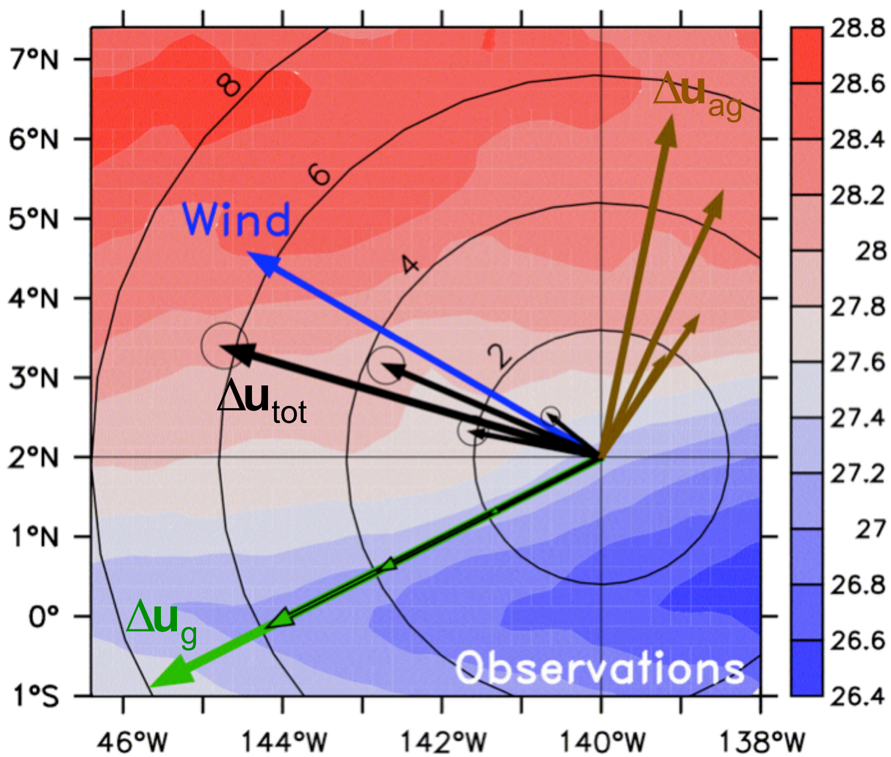
$$at\ \ z = 0: \qquad \boldsymbol{\tau} = \boldsymbol{\tau}_0$$

$$at\ \ z = -H: \qquad \boldsymbol{\tau} = 0 \quad (where\ \ \nu = 0)$$

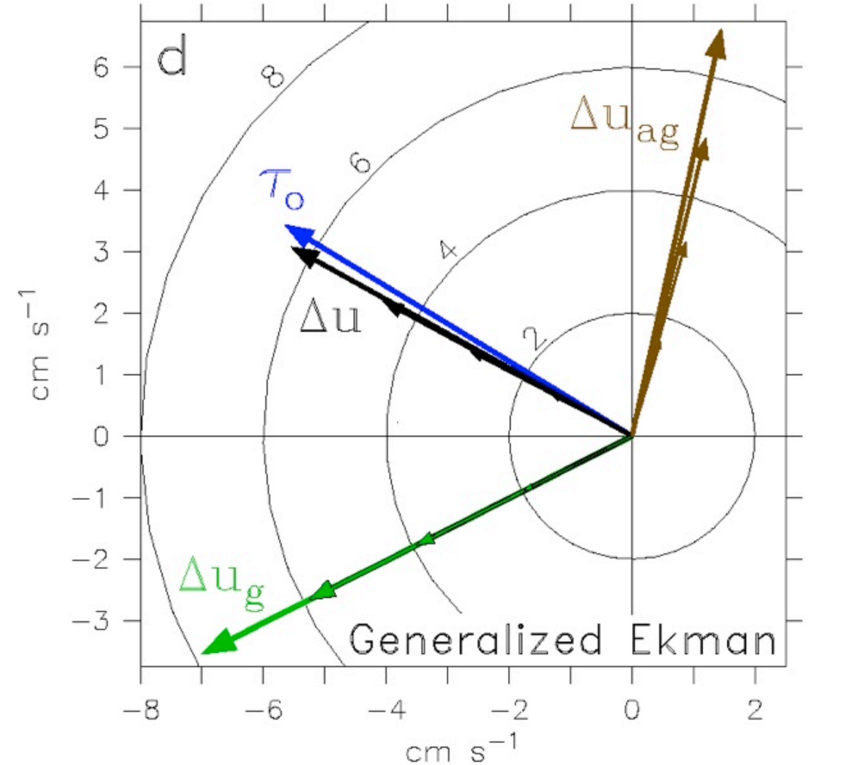
Find solution numerically.

As with the Classic Ekman model, ∇b and ν are prescribed. However, the generalized Ekman model is valid on the equator and in a frontal region, while the Classic Ekman model is not.

The Generalized Ekman model reproduces major features of the near-surface shear at 2ºN



$$\nu = \frac{1}{50}\left(e^{z/125} - e^{-250/125}\right) \text{ [from obs]}$$



Because our deepest measurement was at 25 m, we do not resolve the lower portion of the Ekman spiral where the frontal and generalized Ekman models are expected to differ.

Summary

- Wind stress balances the TOTAL surface shear (combined geostrophic and ageostrophic shears).
- The effect of fronts on Ekman spiral is most pronounced at low latitudes.
- Shear is very sensitive to both the horizontal and vertical temperature distribution.

  Very weak daytime stratification (<0.2°C/25m) resulted in a diurnal jet shear of 12 cm/s / 20m at 4 PM local!
- A generalized Ekman model is valid on the equator and in frontal regions. It requires viscosity to be zero at the bottom of the viscous layer.

Consequences

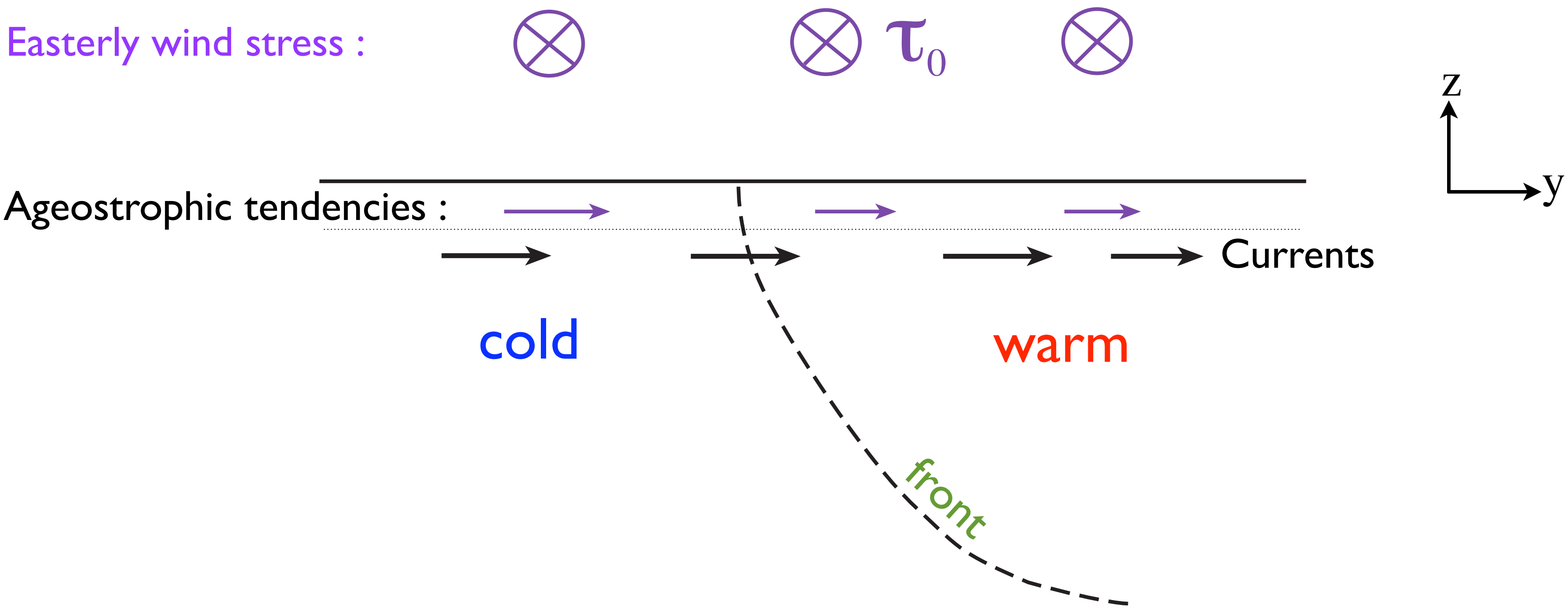
- In a frontal region, the near-surface Ekman flow is not necessarily to the right of the wind stress.
- Boundary conditions!
- If the bottom BC is realistic (i.e. ν or $u_z=0$), the integrated ageostrophic transport is as expected:

$$\text{Ekman Transport in Frontal Region} = \boldsymbol{M}_{ek} = -\frac{i\tau_0}{\rho_0 f} + \cancelto{0}{\frac{i\nu}{f}\frac{\partial \boldsymbol{u}_g}{\partial z}\Big|_{z=-H}}$$

But our "Frontal Ekman" model fails this test!

→ Implies the transmission of stress to a deeper layer.

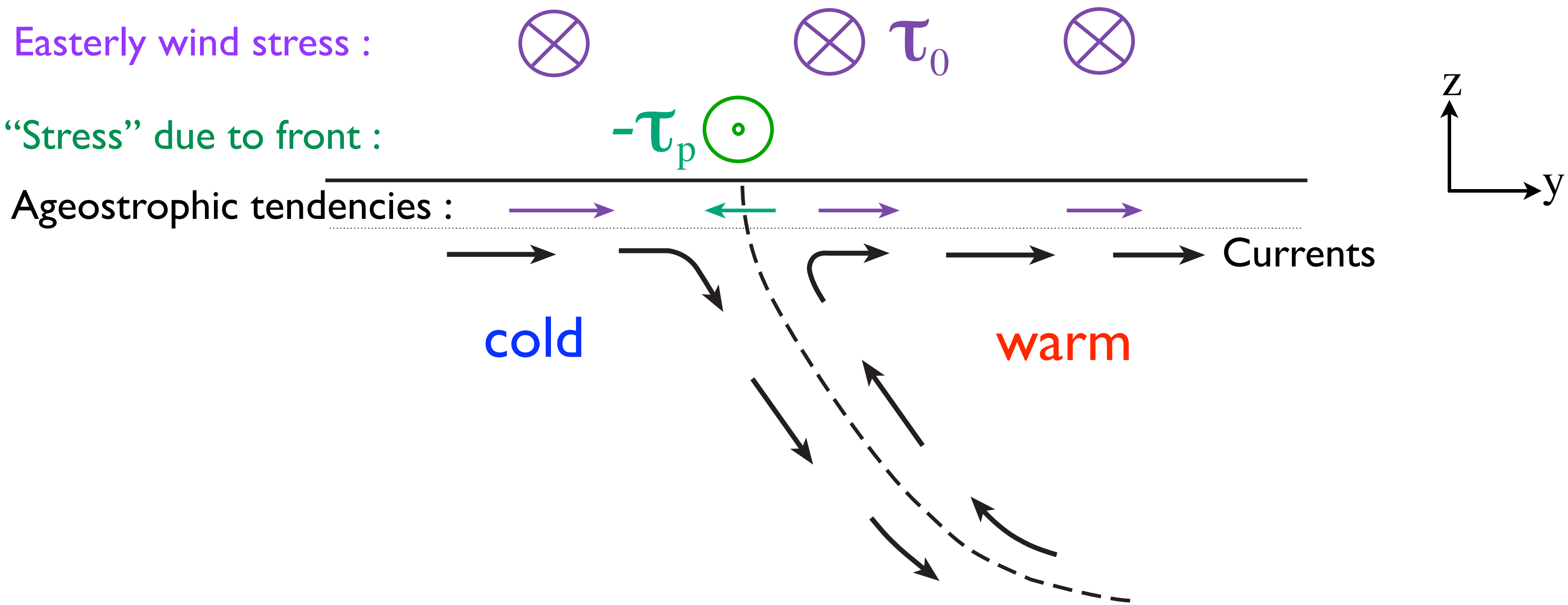
How can equatorial Ekman divergence coexist with a sharp front?



The usual picture has the diverging Ekman flow *across* the front.

But there is no mechanism to warm the water so quickly ....

How can equatorial Ekman divergence coexist with a sharp front?



At the center of the front, geostrophic shear and its Ekman transport are largest, towards the cold side (green arrow).

→ Convergence on the cold side, and divergence on the warm side. The cold diverging water slides *under* the front!

Extra

slides

follow ...