Aspects of wind stress / SST coupling at the oceanic mesoscale

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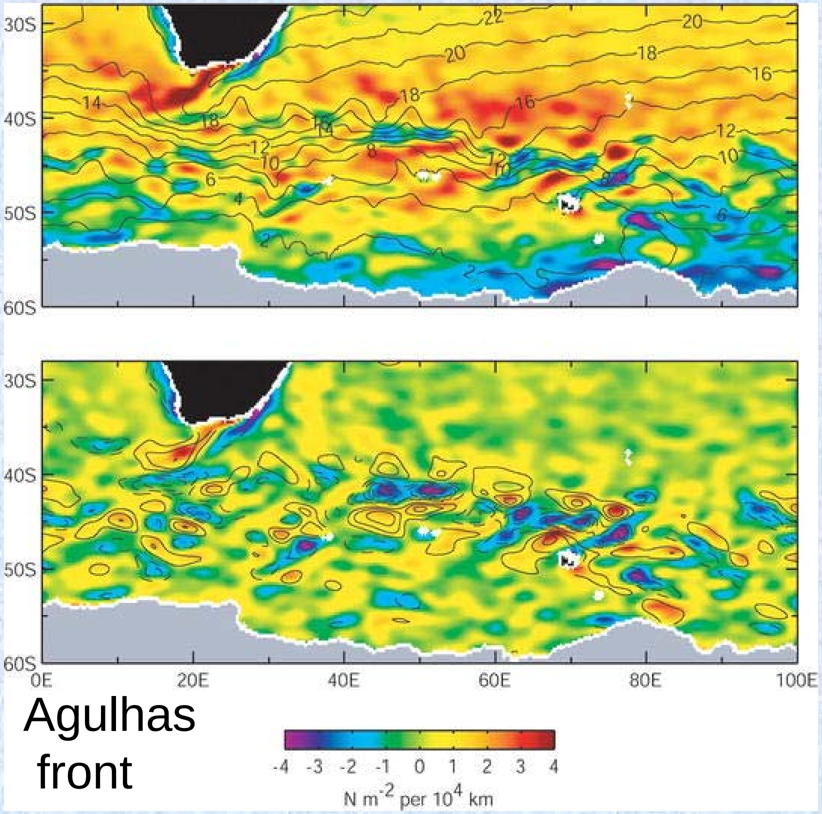
Two research areas:

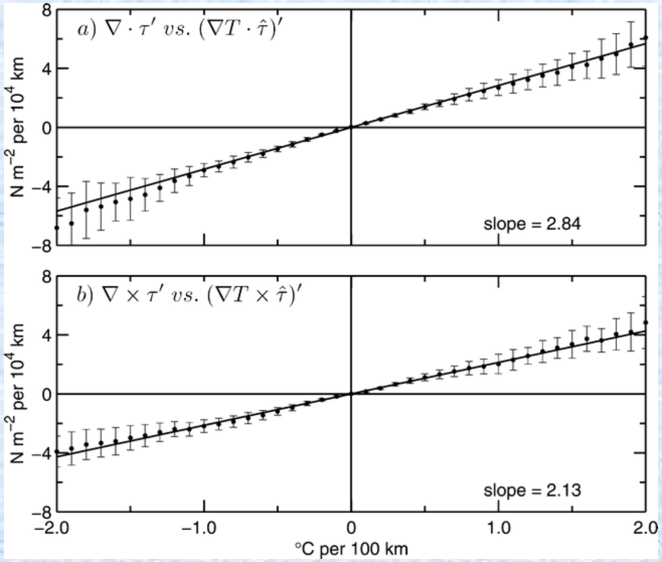
What is the mechanism for the coupling ?

What are the consequences for the ocean circulation ?

Satellite data indicates a close correlation between SST and surface wind stress

(O'Neill et al 2005)





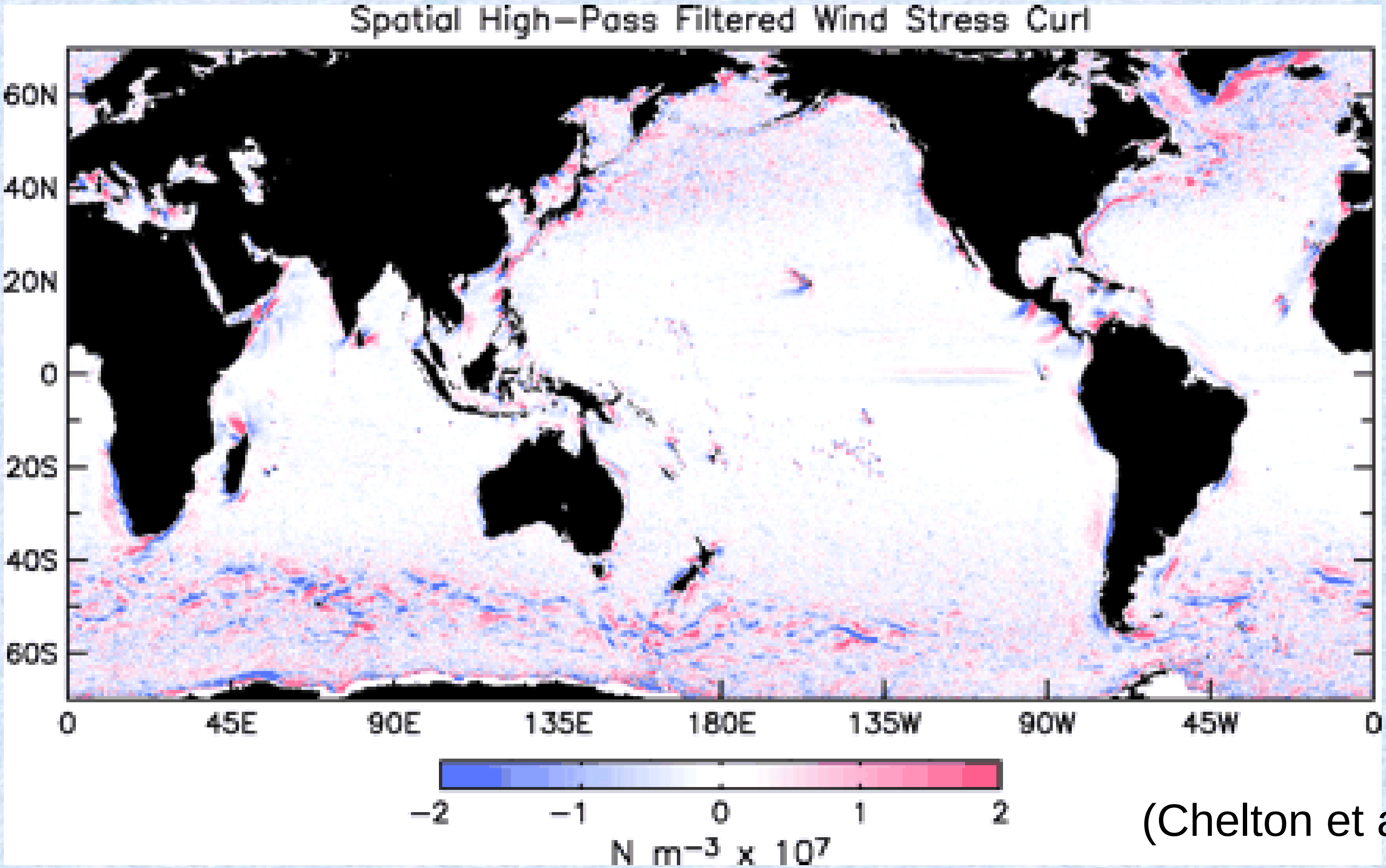
binned wind stress divergence (upper) and curl (lower) versus SST gradient (~ 0.02 N m^{-2} C^{-1})

Upper panel:
1 year average QuikSCAT wind stress curl and SST

wind stress curl is linearly proportional to the SST gradient

Lower panel:
high pass filter wind stress curl and SST

Similar correlations are found at all major ocean fronts



(Chelton et al. 2004)

Increased wind over warm water gives rise to strong wind stress curl in the vicinity of ocean fronts in both mean and synoptic data

(Strong wind curl is also found in lee of islands and mountain passes)

Several mechanisms have been proposed to explain the observed correlation between SST and surface winds

Adjustment to pressure gradients in the vicinity of the front

Lindzen and Nigam 1987, Wai and Stage 1989, Small et al. 2003
Song and Cornillon 2006

Downward mixing of momentum from aloft

Sweet et al. 1981, Wallace et al. 1989, Skyllingstad et al. 2006

Large-scale adjustment of the boundary layer

Samelson et al. 2006

The dominant mechanism seems to vary depending on application and region.

Variables include: latitude, wind strength, direction, frontal strength.

There is no general understanding of what controls the momentum budgets or strength of coupling in the vicinity of SST gradients.

Consider a very simple 2-D model configuration with a wind blowing from either the cold to warm or warm to cold side of an ocean front

The model used is the Coupled Ocean Atmosphere Mesoscale Prediction System (COAMPS), with the ROMS ocean model and coupling by Perlin et al. (2006)

Initial conditions:
specify change in SST and width of front L_f
specify adiabatic lapse rate in atmosphere and surface temperature
specify initial geostrophically balanced cross-front wind

Run the model for 48 hours, until fields are at steady state

Diagnose change in surface winds, boundary layer properties, coupling strength, momentum balances

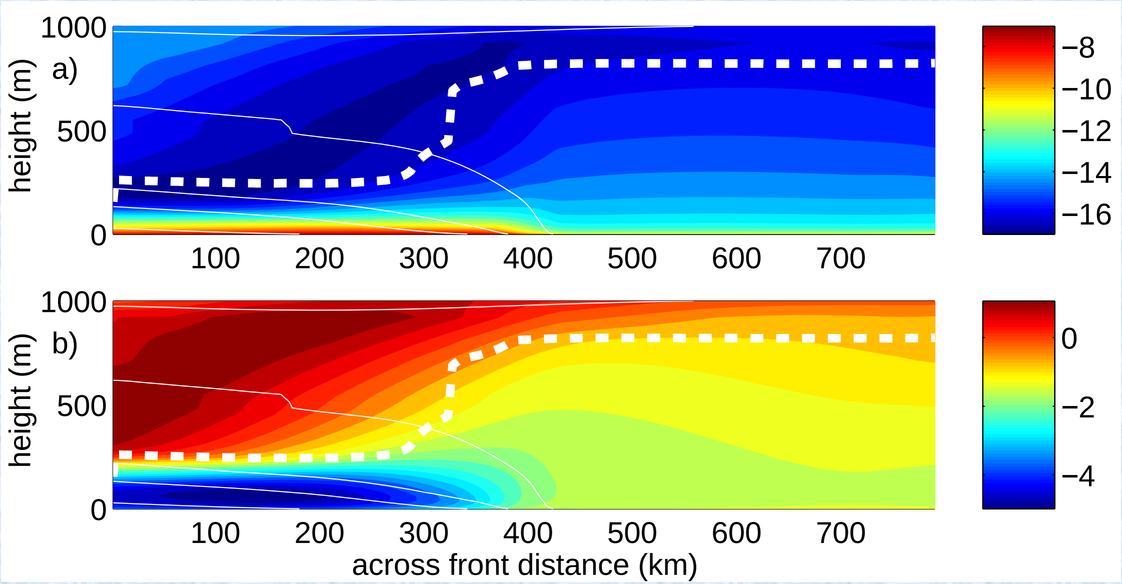
Sensitivity studies varying ocean state, atmospheric state, subgrid turbulent mixing

Example of warm to cold flow (right to left)

initial winds -15 m/s, $\Delta T=5$ front is centered at 400 km, e-folding width 30 km

cross-front wind

along-front wind



dashed line boundary layer height

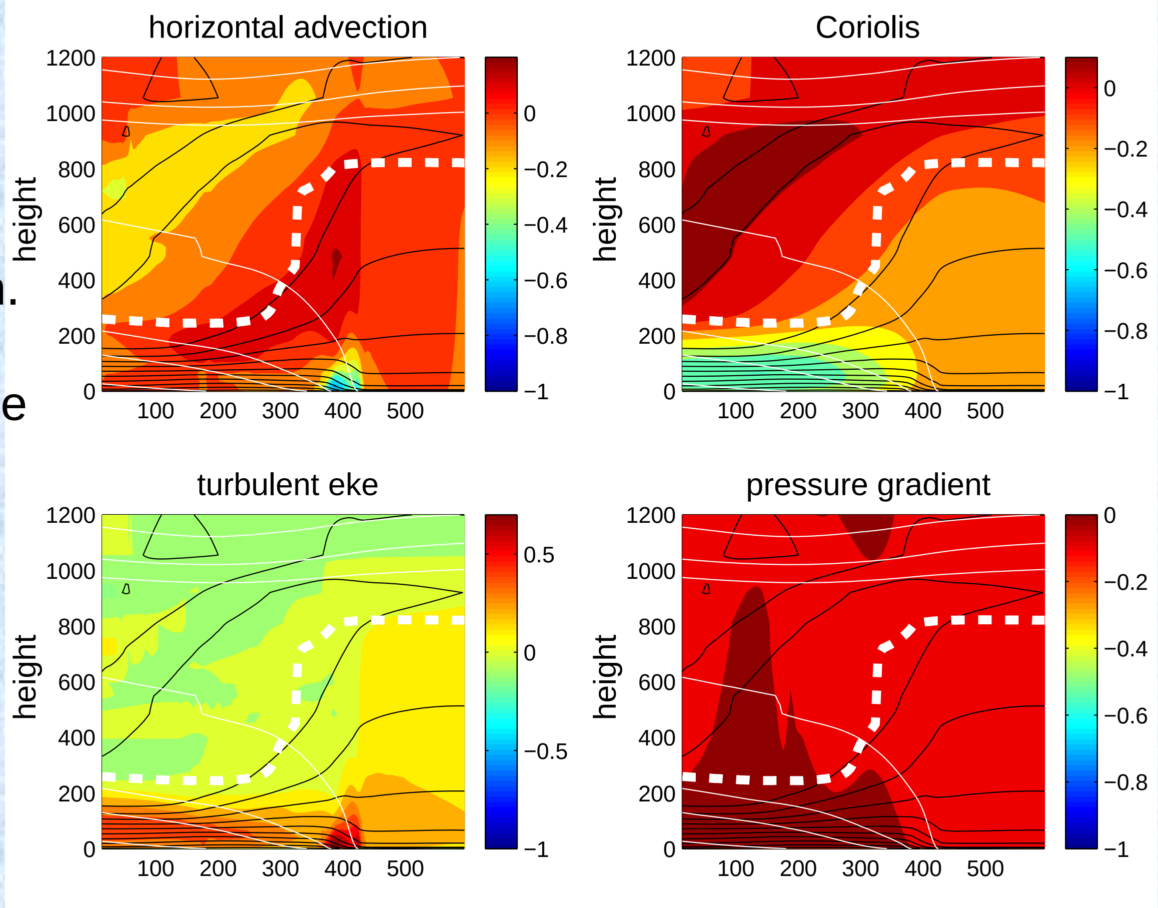
thin lines are potential temp

Essentially one-dimensional balance upwind of the ocean front

Boundary layer rapidly shallows over cold water, internal boundary layer develops (Mahrt et al. 2004, Skyllingstad et al. 2006)

Cross-front wind slows near surface, along-front wind veers strongly at the surface, begins turning aloft

Momentum budgets



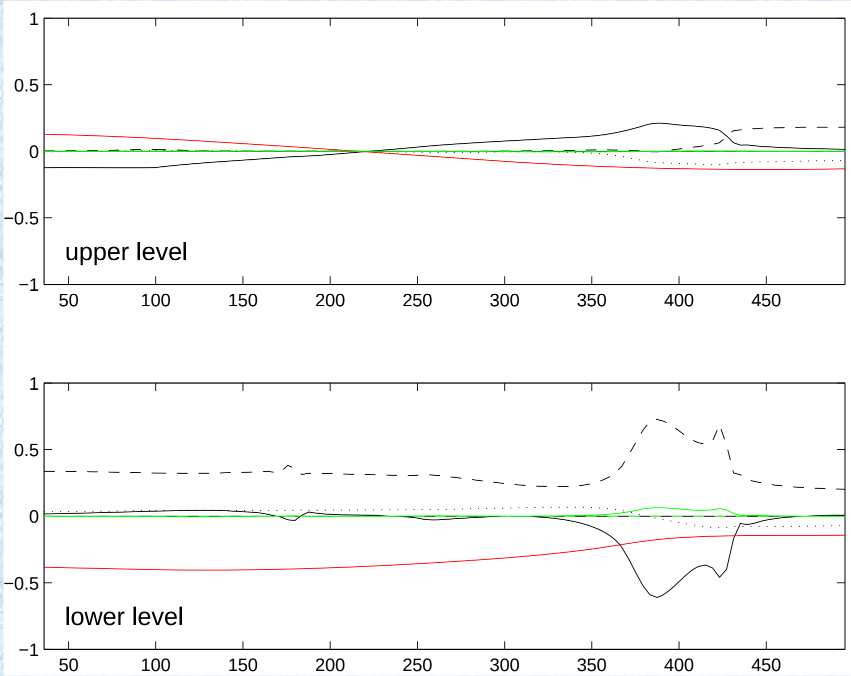
Colors : terms in zonal mom eqn.
Solid lines : zonal velocity
Thin white : potential temperature
Dashed white : boundary layer

Momentum balance away from the front is between vertical mixing and Coriolis
Although more complicated, this is an Ekman-like balance

Near the front: surface deceleration is balanced by change in TKE

Upper level: wind changes balanced by Coriolis, “inertial lee wave”

Consider the balances in upper and lower boundary layer



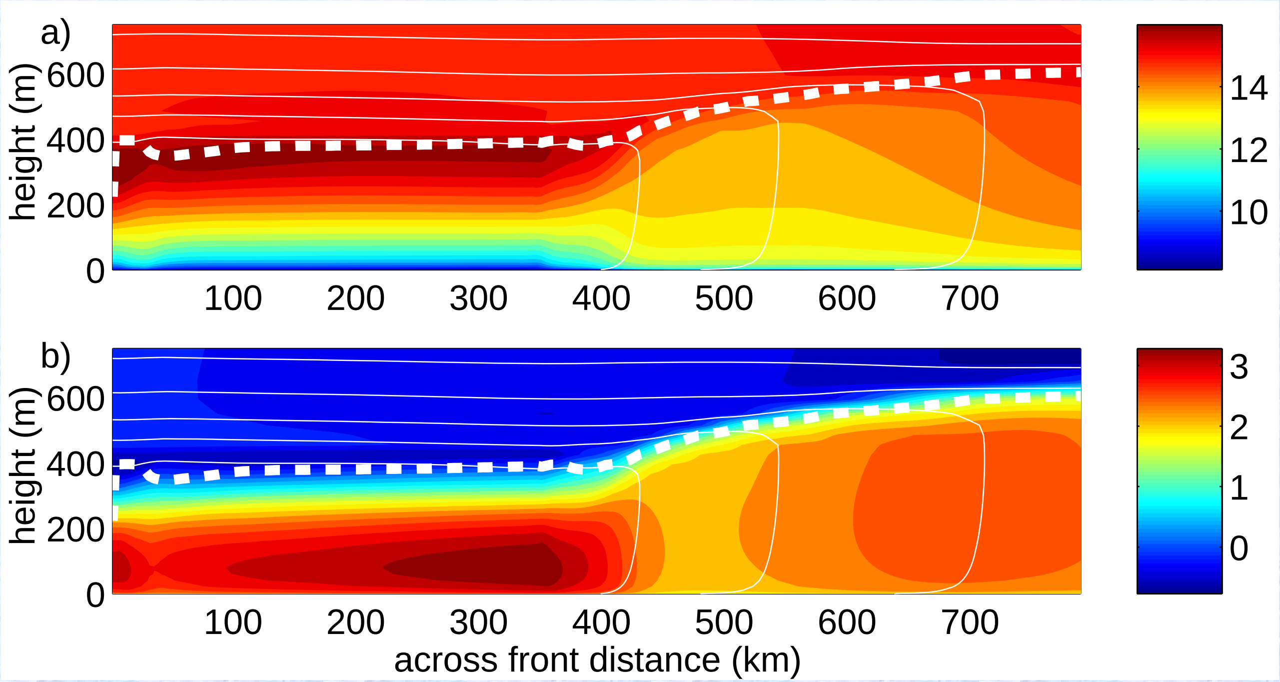
Black : zonal acceleration
Red : Coriolis
Black dashed : turbulent mixing
Black dotted : pressure gradient
Green : vertical advection

Upper level : TKE rapidly drops to zero outside boundary layer
nothing to balance this change, Coriolis accelerates zonal flow,
change in U drives meridional flow through Coriolis term in V mom eqn
results in a standing inertial lee wave downwind of the ocean front
wavelength = $2\pi U/f$ = 1000 km for 15 m/s wind at mid latitudes

Lower level : rapid increase in TKE slows surface winds
change in U eventually changes Coriolis, balances TKE downwind

Example of cold to warm flow (left to right)

initial winds 15 m/s, ΔT=5 front is centered at 400 km, e-folding width 30 km



Ekman-like balance upwind of the ocean front

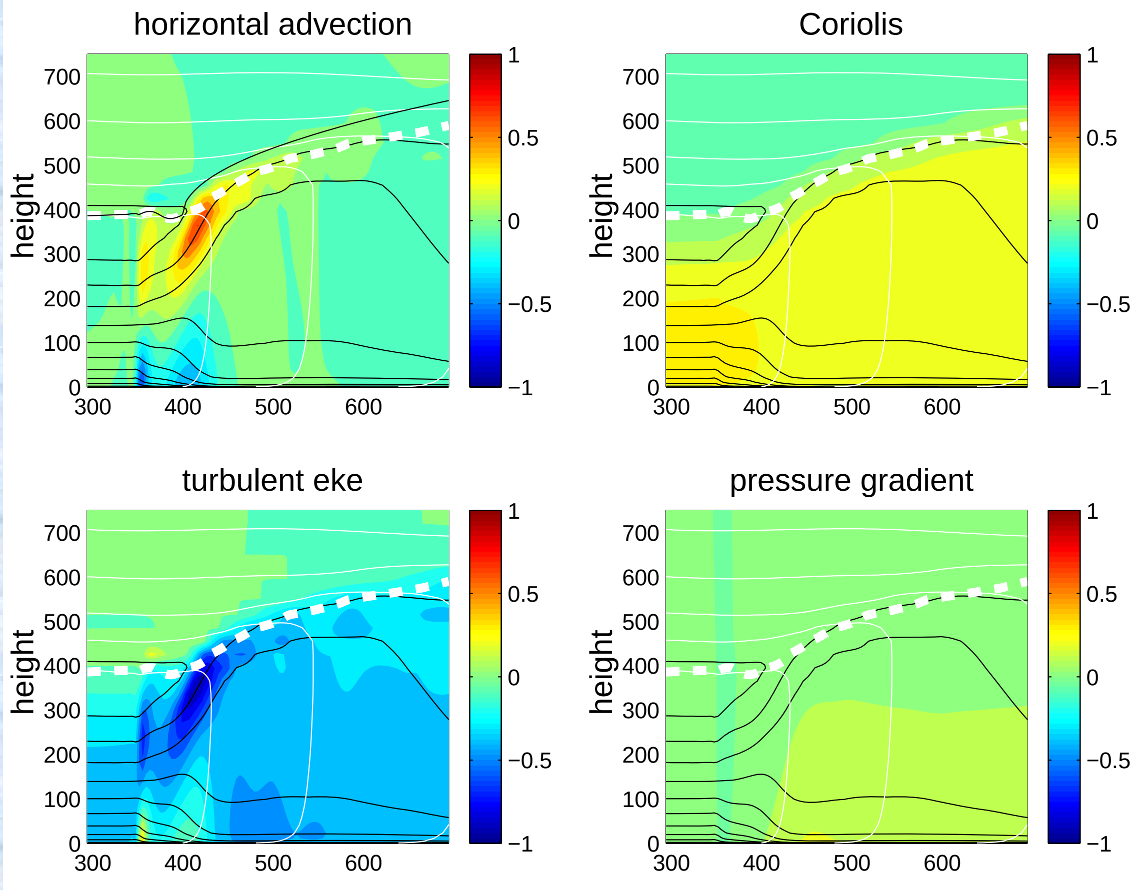
Near surface flow accelerates over the ocean front by several m/s, upper level flow decelerates, similar to observed

Boundary layer gradually grows over warm water on length scale much larger than the frontal width, very different from warm to cold flow

Advection of temperature is important, large adjustment length scale

Momentum budget

Colors : terms in zonal mom eqn.
Solid lines : zonal velocity
Thin white : potential temperature
Dashed white : boundary layer

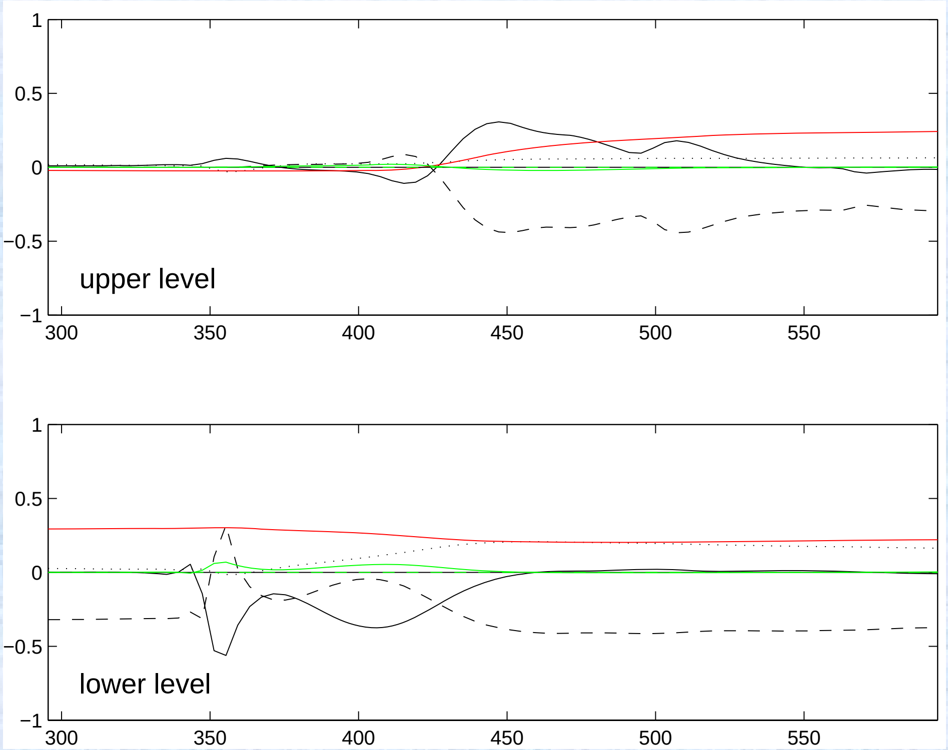


Momentum balance away from the front is between turbulent mixing and Coriolis (Ekman-like)

Change in zonal momentum again balanced by change in turbulent mixing in the vicinity of the front

Consider the balances in upper and lower boundary layer

Black : zonal acceleration
Red : Coriolis
Black dashed : turbulent mixing
Black dotted : pressure gradient
Green : vertical advection

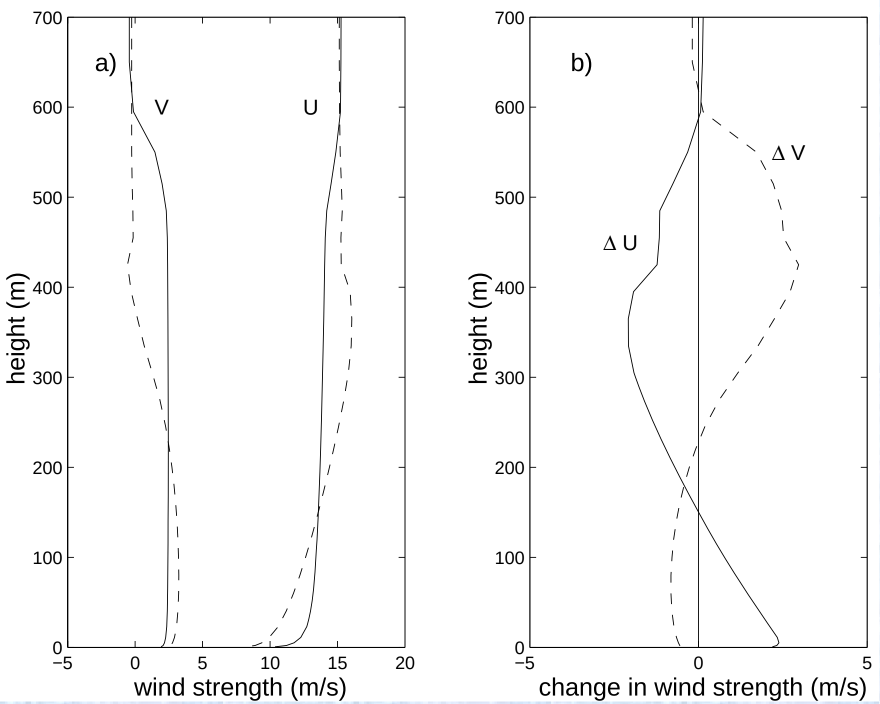


Upper level : TKE grows, decelerates the zonal flow. Coriolis increases in response over somewhat larger length scale than the front

Lower level : rapid decrease in TKE, but still acting to slow near surface winds acceleration of surface winds results from imbalanced Coriolis term not from vertical mixing of momentum or pressure gradients

Ekman-like balances are found both upwind and downwind of the front

The acceleration of low level winds looks a lot like the deceleration of upper level winds. Is this simply downward mixing of momentum from aloft?



velocity profiles
dashed : upwind
solid : downwind

change in velocities across front

gain in zonal momentum in the lower boundary layer (182 m^2/s) is similar to loss of meridional momentum in lower boundary layer (140 m^2/s)

but

is much less than the loss of zonal momentum in the upper boundary layer (513 m^2/s) or gain in meridional momentum (635 m^2/s)

Changes in momentum are achieved through horizontal exchanges via the Coriolis term, not vertical exchange

Several length (or time) scales define the flow regimes

L_f is the frontal width

$L_i = U/f$ is the scale over which the Coriolis term can change

$L_P = UH^2/K$ is the scale over which the pressure term can change

if H is Ekman layer, $L_P = L_i P_r$, where P_r is Prandtl number

For typical mid-latitude, strong wind regimes : $L_f < L_i, L_P$

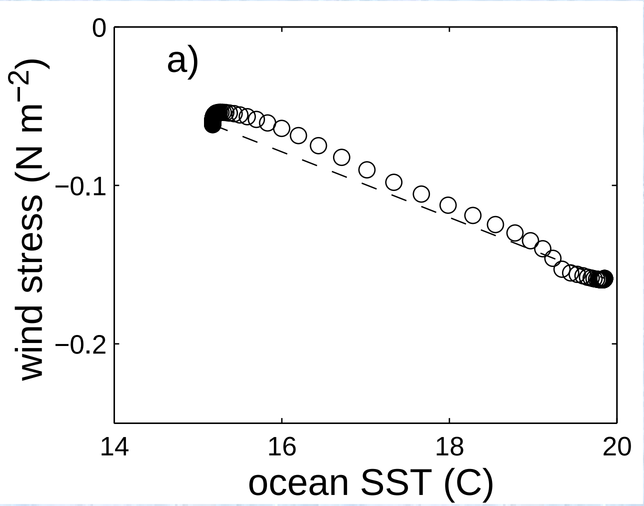
Turbulent mixing changes on length scale of L_f , essentially 1-D because horizontal advection of TKE is small

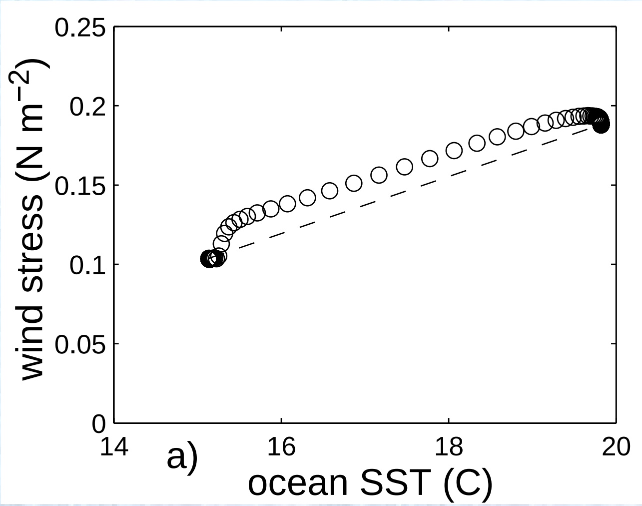
This means that the Coriolis term is unbalanced near the front.

The momentum for acceleration/deceleration of the winds in the boundary layer does not come from vertical mixing or pressure gradients, but instead from the along front flow

Calculations with L_i, $L_P = L_f$ support this conclusion

How strong is the coupling in the model ?





warm to cold

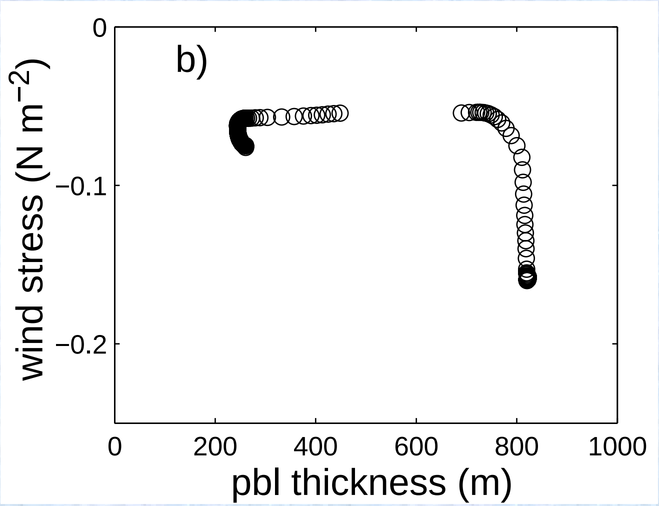
cold to warm

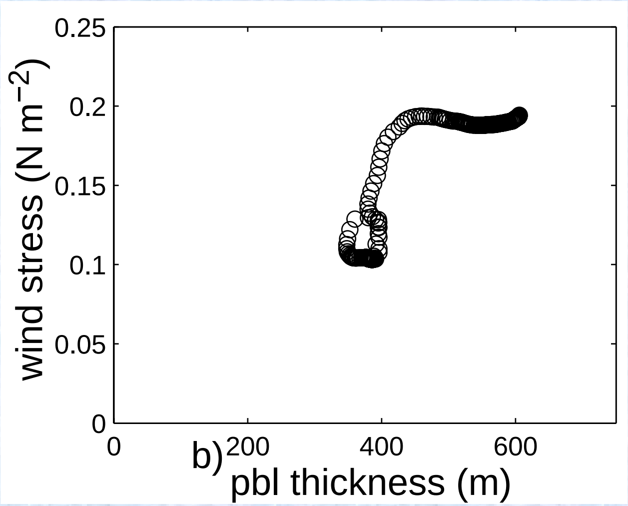
Stress is linearly related to SST

Change in stress takes place in vicinity of the front

Slope is consistent with observations, 0.024 N m^{-2} C^{-1}
and 0.02 N m^{-2} C^{-1}

Is stress linearly related to boundary layer thickness ?





warm to cold

cold to warm

Most of the change in stress occurs at constant boundary layer thickness

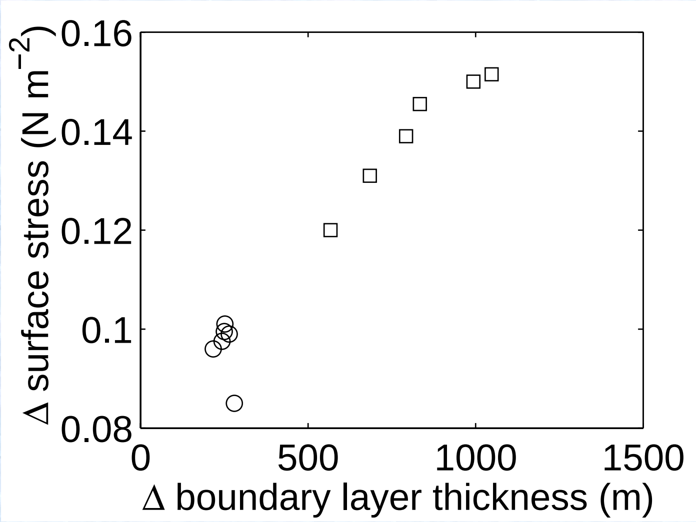
Most of the change in boundary layer thickness occurs at ~ constant stress

The linear relation of Samelson et al. (2006) does not hold in the vicinity of the front. This is not surprising since their theory assumes a steady balance.

What if we look away from the transition region ?

Consider 5 cases warm to cold and cold to warm in which SST is increased by 1^0C.

This results in deeper boundary layers on both sides of the front (200 km)



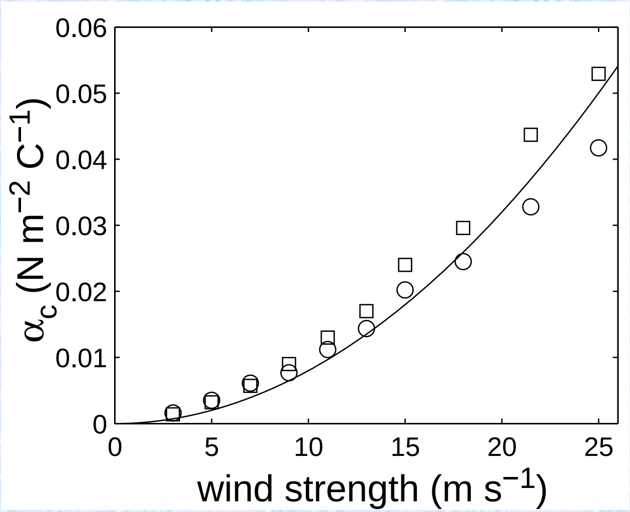
The change in stress is nearly linearly related to the change in boundary layer thickness, consistent with Samelson et al. (2006) and the basic Ekman layer balance

What controls the strength of the coupling $\alpha_c = \Delta\tau/\Delta T$?

- Series of calculations with various terms in the TKE equation turned on and off indicate that coupling is primarily due to changes in vertical mixing of momentum as a result of a decrease in the flux Richardson number

- The stability dependency of the drag coefficient is only 10% effect

-The strongest effect is found for the strength of the cross front wind

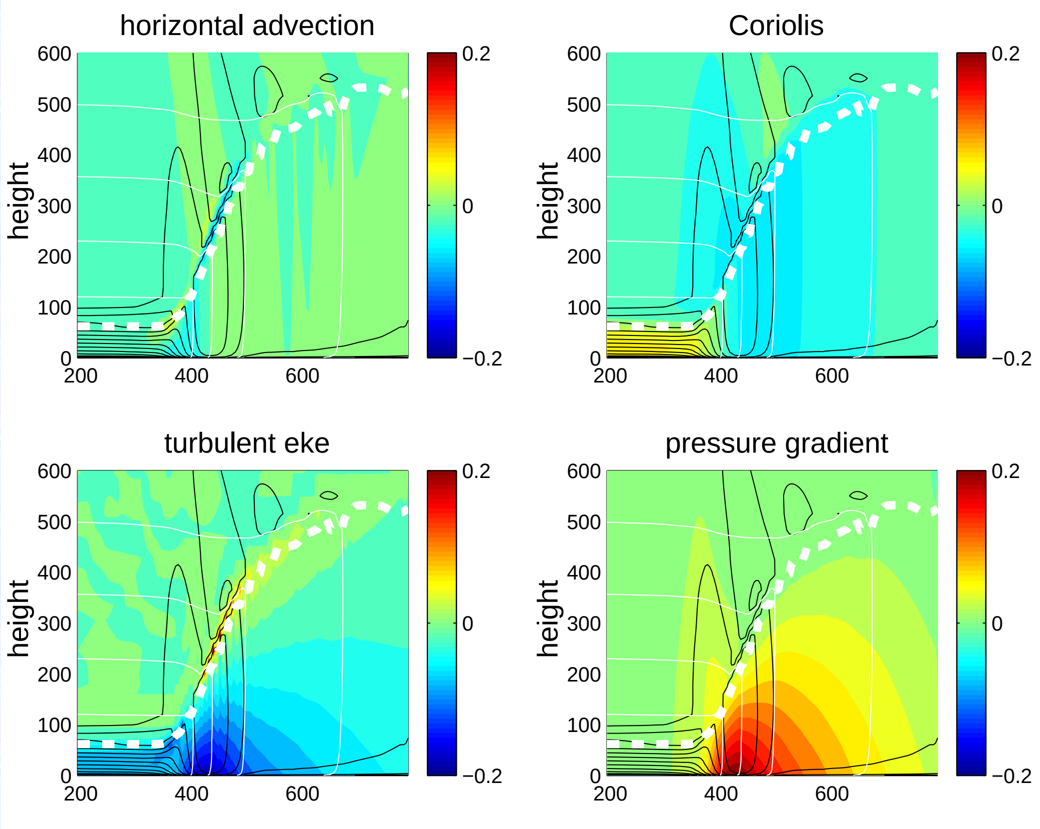


squares : warm to cold flow
circles : cold to warm flow

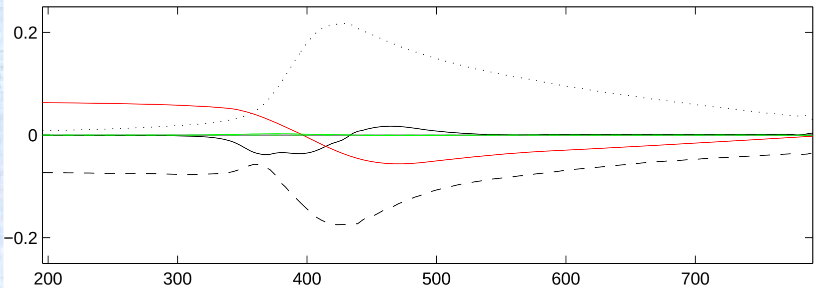
Nearly quadratic dependence for both warm to cold and cold to warm flows

Stronger coupling for warm to cold is due to the internal boundary layer over cold water (lower stress there)

What happens when $L_f = L_i$, L_P (Weak winds or wide fronts) ?



Try U = 3 m/s, L_i = 30 km

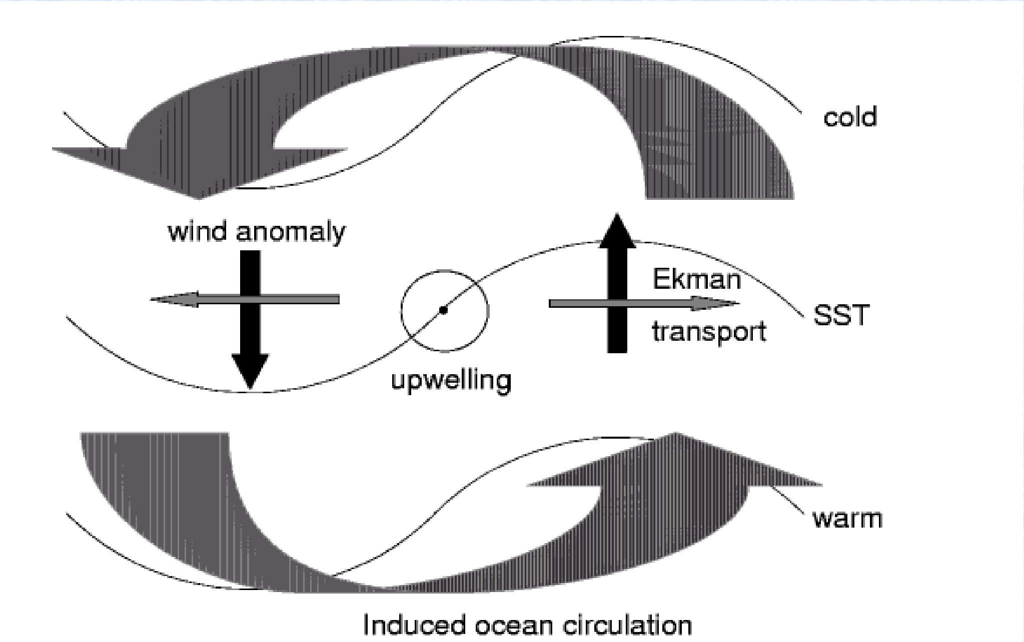


Low level momentum balances

Pressure term is now important in momentum balance

Both pressure and Coriolis change in length scale of the front and both contribute to acceleration of low level winds

What are the consequences of this coupling for the oceanic circulation ?



Consider a wind blowing across an ocean front with a wave-like perturbation

- winds over warm water will be stronger than winds over cold water
- Ekman transport over warm water is stronger than over cold water
- this gives rise to upwelling between troughs and crests and
  downwelling between crests and troughs
- the ocean will adjust to this Ekman pumping/suction by developing relative vorticity

- suggests warm to cold wind may enhance the growth rate of baroclinic waves
  (and cold to warm wind may reduce growth rate)

Formally pose the problem in the simplest context

$$\tau = \tau_0 + \alpha_c T'$$

Assume that the wind stress anomaly is proportional to SST anomaly with coupling α_c

$$\tau_x = \alpha_c T'_x$$

Wind stress curl will then be proportional to temperature gradient (for simplicity, consider meridional winds only)

$$\boxed{w_e = \frac{\alpha_c T'_x}{\rho_0 f}}$$

and the Ekman pumping rate will be proportional to the temp gradient

Consider now a baroclinic, quasigeostrophic flow on an f-plane with uniform vertical shear, no horizontal shear (Eady problem) subject to this Ekman pumping at the surface

$$\left(\frac{\partial}{\partial t}+v\bullet\nabla\right)\left(\nabla^2+\frac{1}{B}\frac{\partial^2}{\partial z^2}\right)\psi=0$$

Nondimensional PV equation for fluid interior, ψ is QG streamfunction

$$\left(\frac{\partial}{\partial t}+v\bullet\nabla\right)\psi_z+\delta\psi_{xz}=0 \qquad z=1$$

$$\left(\frac{\partial}{\partial t}+v\bullet\nabla\right)\psi_z=0 \qquad z=0$$

Surface and lower boundary conditions

Upper boundary condition δ term represents effect of wind stress coupling, which is proportional to zonal temperature gradient

Consider this term in more detail :

The upper boundary condition states that the horizonal advection of temperature balances vertical advection due to Ekman pumping.

Strength of vertical advection is indicated by nondimensional parameter δ.

$$\delta = \frac{\alpha_c N^2}{\alpha g f U}$$

α_c coupling coefficient
N^2 stratification
f Coriolis parameter
U horizontal velocity
α thermal expansion coef
g gravity

Effect is strongest for :

strong coupling coefficient α_c
large stratification N^2
low latitudes
weak background flows

To obtain an analytic solution :

Assume streamfunction consists of uniform background flow with a small perturbation $\psi = -yz + \phi$

Further assume that the perturbation is wave-like in x

$$\phi = F(z)e^{i(kx-\sigma t)}$$

Has solution :

$$\phi = Ae^{i(kx-\sigma t)}\left(\sinh(\mu z) - \eta\cosh(\mu z)\right)$$

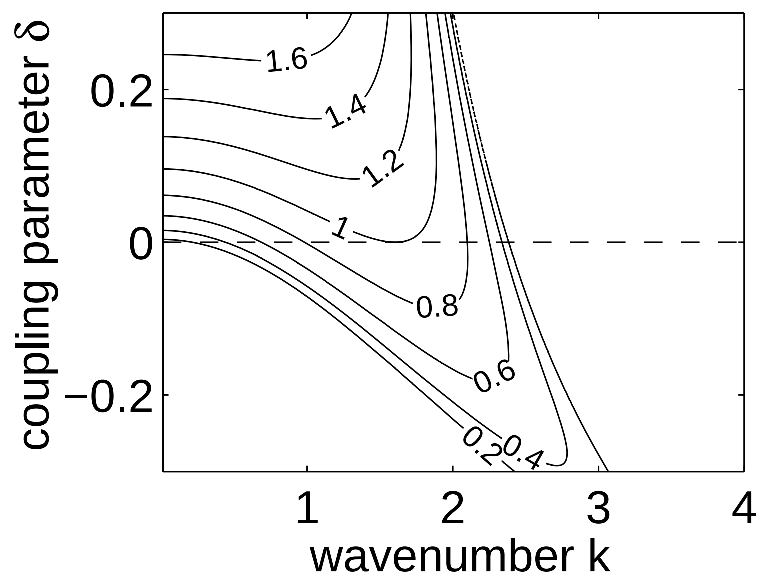
$$\eta = \frac{\mu}{2}(1+\delta) \pm \left[\frac{\mu^2}{4}\left(1+\delta^2\right) - \frac{\mu(1+\delta) - \tanh\mu}{\tanh\mu}\right]^{1/2}$$

$$\mu = kB^{1/2}$$

Affects the phase speed and growth rate of unstable waves

Eady solution is recovered for $\delta = 0$

Growth rate for various δ and wavenumber



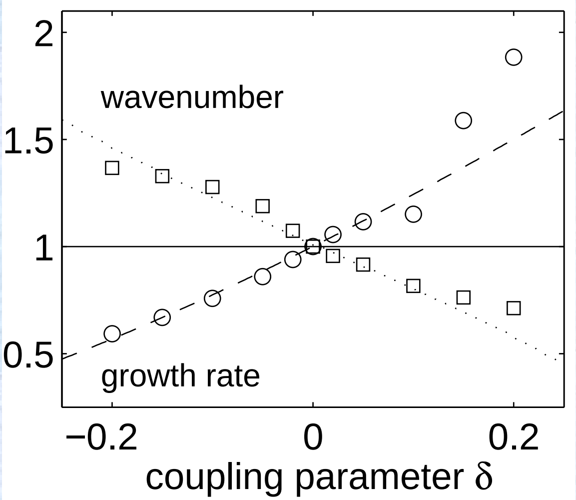
Coupling destabilizes long waves for $\delta > 0$
(warm to cold flow)
stabilizes long waves for $\delta < 0$
(cold to warm flow)

Also affects the wavelength of most unstable waves

Growth rate can change by 50% for realistic values of δ

Test theory with ocean circulation model (ROMS)

Initialize the model with uniformly sheared zonal flow in a channel
force at the surface with wind stress proportional to temperature anomaly
vary the strength of coupling and diagnose wave growth rate and wavelength



growth rate (circles) and
wavenumber (squares)
for a series of model calculations,
normalized by Eady results

Generally good agreement between model (symbols) and theory (lines)
warm to cold flow ($\delta>0$) enhances growth rate, longer waves
cold to warm flow ($\delta<0$) reduced growth rate, shorter waves

How important is this coupling for various regions ?

$$\delta = \frac{\alpha_c N^2}{\alpha g f U}$$

α_c = 0.02 Kg/m s^2 C (O'Neill et al. 2005)
α = 0.2 Kg/m^3 C (thermal expansion coef)
g = 10 m^2/s

	Gulf Stream	Shelf Break Front	Tropical Instability waves
N^2 (s^{-2})	10^{-4}	$1\text{-}4\times10^{-4}$	10^{-4}
f (s^{-1})	10^{-4}	10^{-4}	4×10^{-6}
U (m/s)	1	0.2	0.5
δ	0.01	0.07-0.27	0.5

Coupling is most effective for low latitude, strongly stratified flows

Not likely to be important for separated mid-lat boundary currents such as the Gulf Stream or Kuroshio

Summary

Explored various aspects of coupling between sea surface temperature and atmospheric winds on the scales of the oceanic internal deformation radius

On small scales ($L < U/f$), the acceleration/deceleration of the near surface winds over SST fronts results from an imbalance of the Coriolis term, the frontal region serves to transition the atmospheric boundary layer between approx 1-d Ekman layers

This coupling is driven primarily by changes in TKE as a result of changes in the flux Richardson number and increases quadratically with wind speed

Such coupling can influence baroclinic instability in the ocean, most strongly for low latitude, stratified flows