Baroclinic Vs Barotropic eddy saturation

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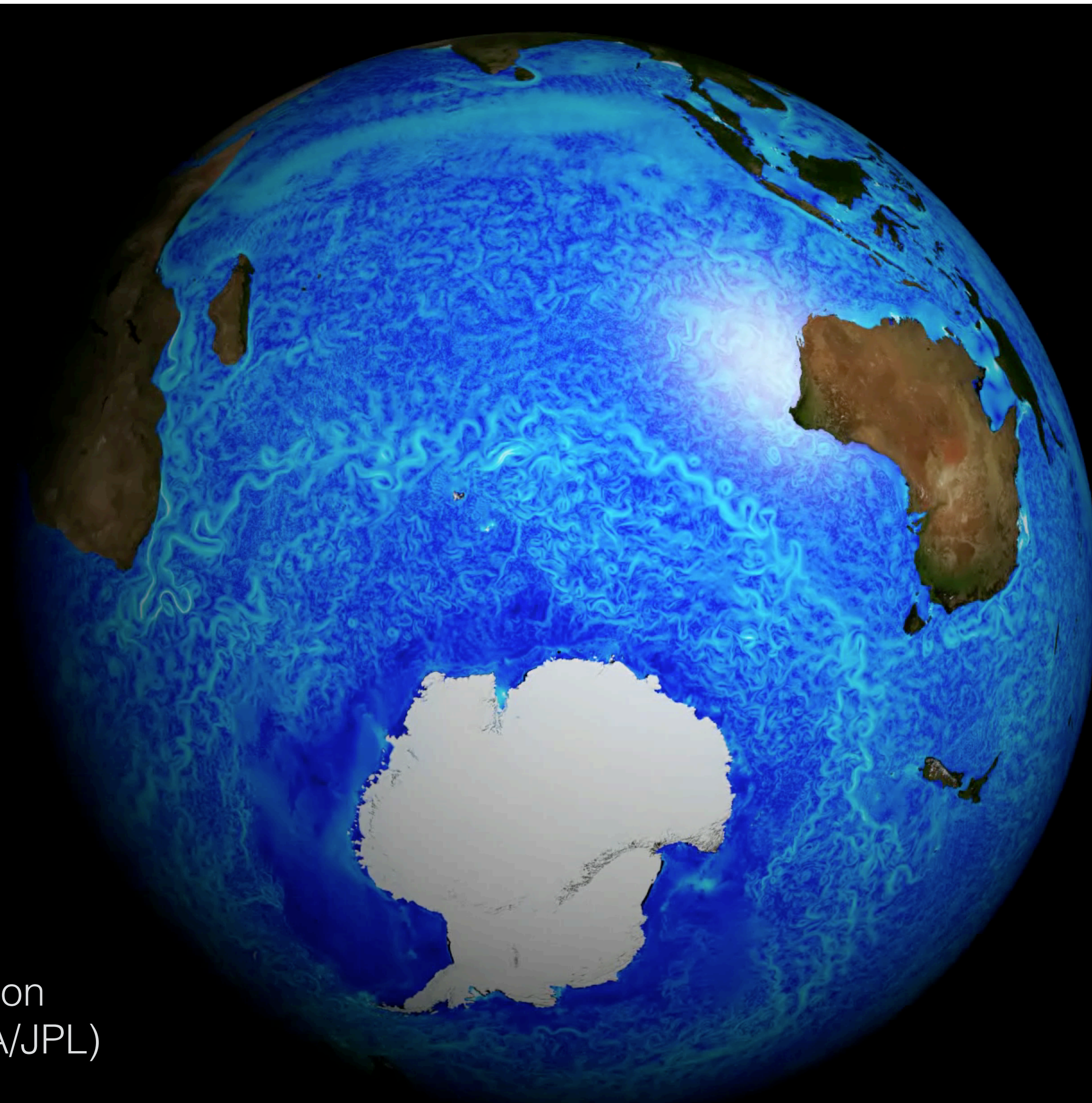
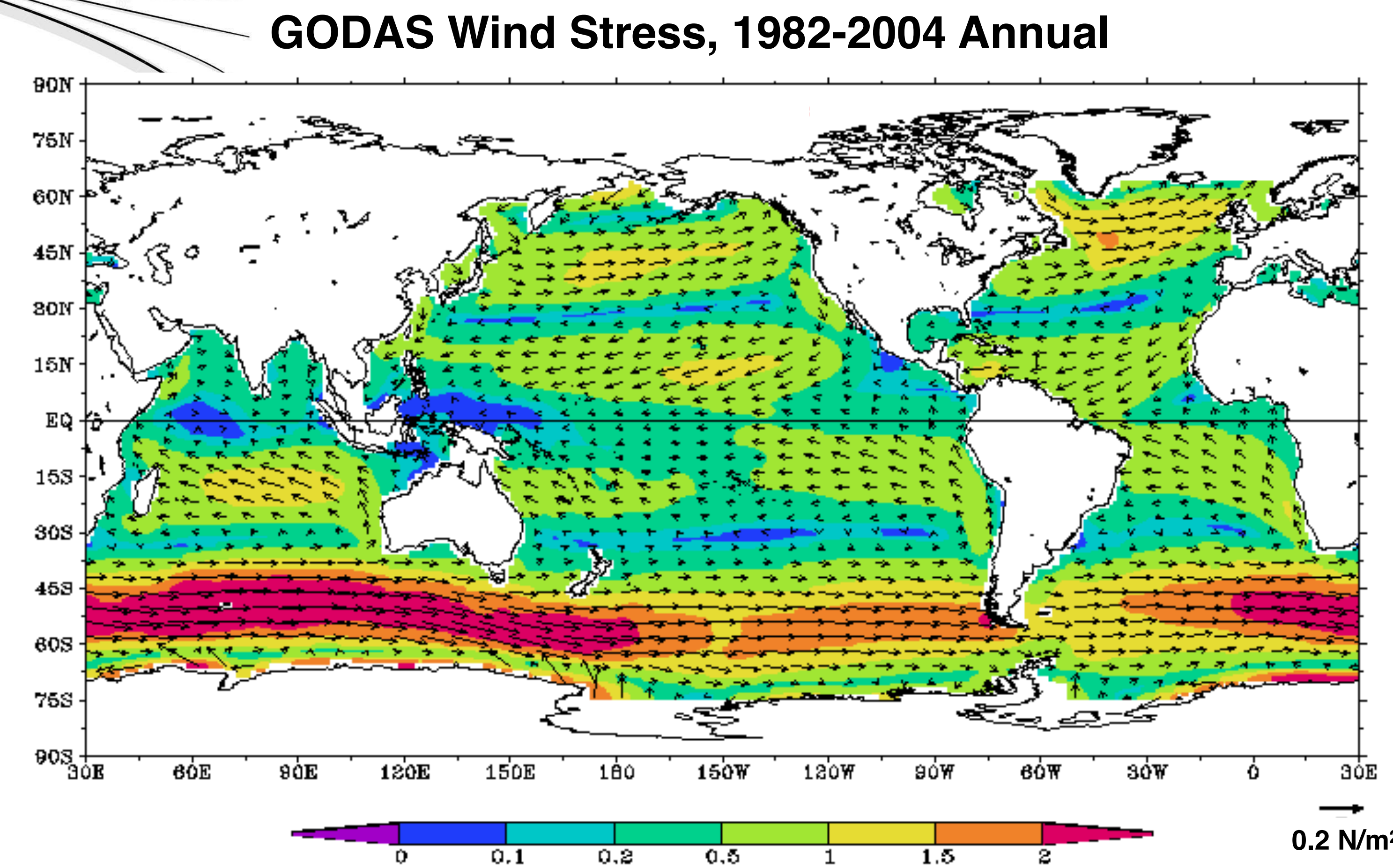


the Antarctic Circumpolar Current

LLC4320 sea surface speed animation
by C. Henze and D. Menemenlis (NASA/JPL)
1/48th degree, 90 vertical levels
MITgcm spun up from ECCO v4 state estimate

what drives the Antarctic Circumpolar Current?

Climate Prediction Center



strong westerly winds blow over the Southern Ocean
transferring momentum through wind stress at the surface

how is this momentum balanced? bottom drag?

Note on the Dynamics of the Antarctic Circumpolar Current

W.H. Munk
(101 bday on Oct 19th, 2018)

By W. H. MUNK and E. PALMÉN

1951

Abstract

Unlike all other major ocean currents, the Antarctic Circumpolar Current probably does not have sufficient frictional stress applied at its lateral boundaries to balance the wind stress. The balancing stress is probably applied at the bottom, largely where the major submarine ridges lie in the path of the current. The meridional circulation provides a mechanism for extending the current to a large enough depth to make this possible.

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start with the zonal angular momentum equation

$f(y)$ is the Coriolis parameter
$f = 2\Omega \sin\vartheta$

$$(\partial_t + u\partial_x + v\partial_y + w\partial_z)\underbrace{\left(u - \int^y f(y')\,\mathrm{d}y'\right)}_{\stackrel{\mathrm{def}}{=}a} + p_x = \tau_z$$

angular momentum

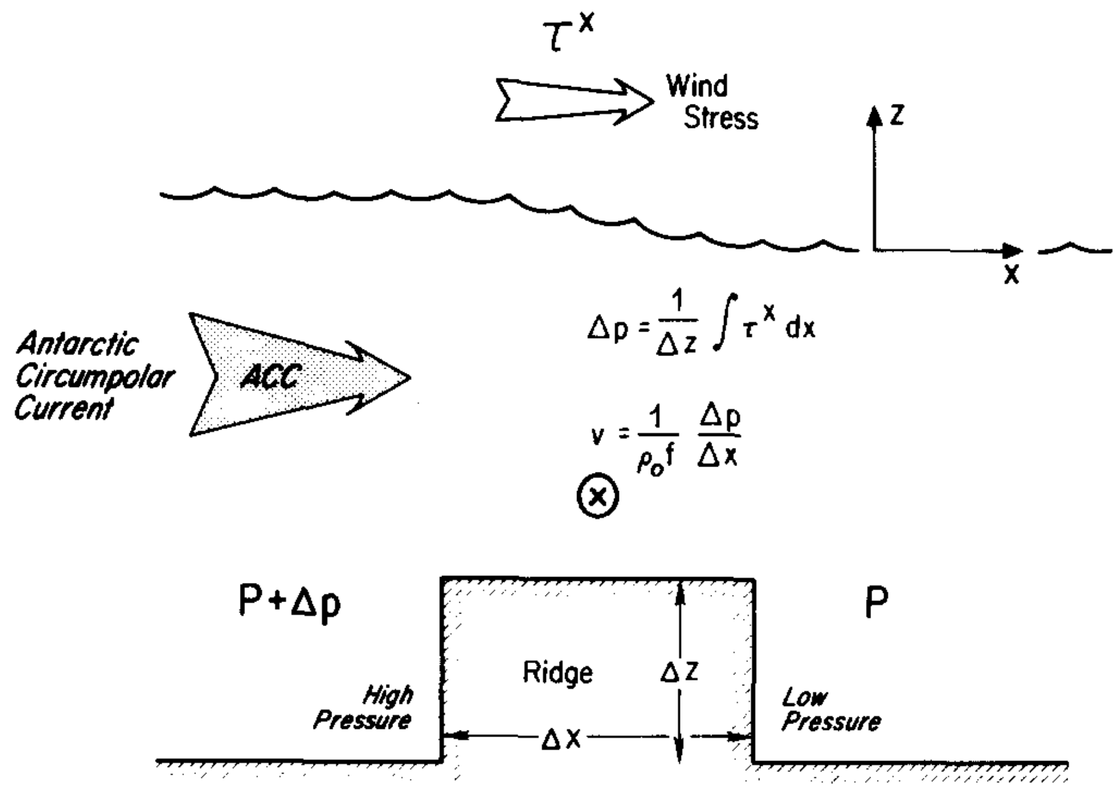
vertically integrate, top $z=0$ to bottom $z=-h(x,y)$

$$\partial_t \int_{-h}^{0} a\,\mathrm{d}z + \partial_x\left[\int_{-h}^{0} ua + p\,\mathrm{d}z\right] + \partial_y \int_{-h}^{0} va\,\mathrm{d}z =$$
$$= \underbrace{\tau(0)}_{\text{wind stress}} - \underbrace{\tau(-h)}_{\text{bottom drag}} + \underbrace{h_x p(-h)}_{\text{form stress}}$$

we've used integration by parts:

$$\int_{-h}^{0} p_x\,\mathrm{d}z = \partial_x \int_{-h}^{0} p\,\mathrm{d}z - h_x p(-h)$$

topographic form stress



Johnson & Bryden 1989

Schematic presentation of bottom form drag or mountain drag. Wind stress imparted eastward momentum in the water column is removed by the pressure difference across the ridge.

$$\partial_t \int_{-h}^{0} a \, \mathrm{d}z + \partial_x \left[\int_{-h}^{0} ua + p \, \mathrm{d}z \right] + \partial_y \int_{-h}^{0} va \, \mathrm{d}z =$$

$$= \underbrace{\tau(0)}_{\text{wind stress}} - \underbrace{\tau(-h)}_{\text{bottom drag}} + \underbrace{h_x p(-h)}_{\text{form stress}}$$

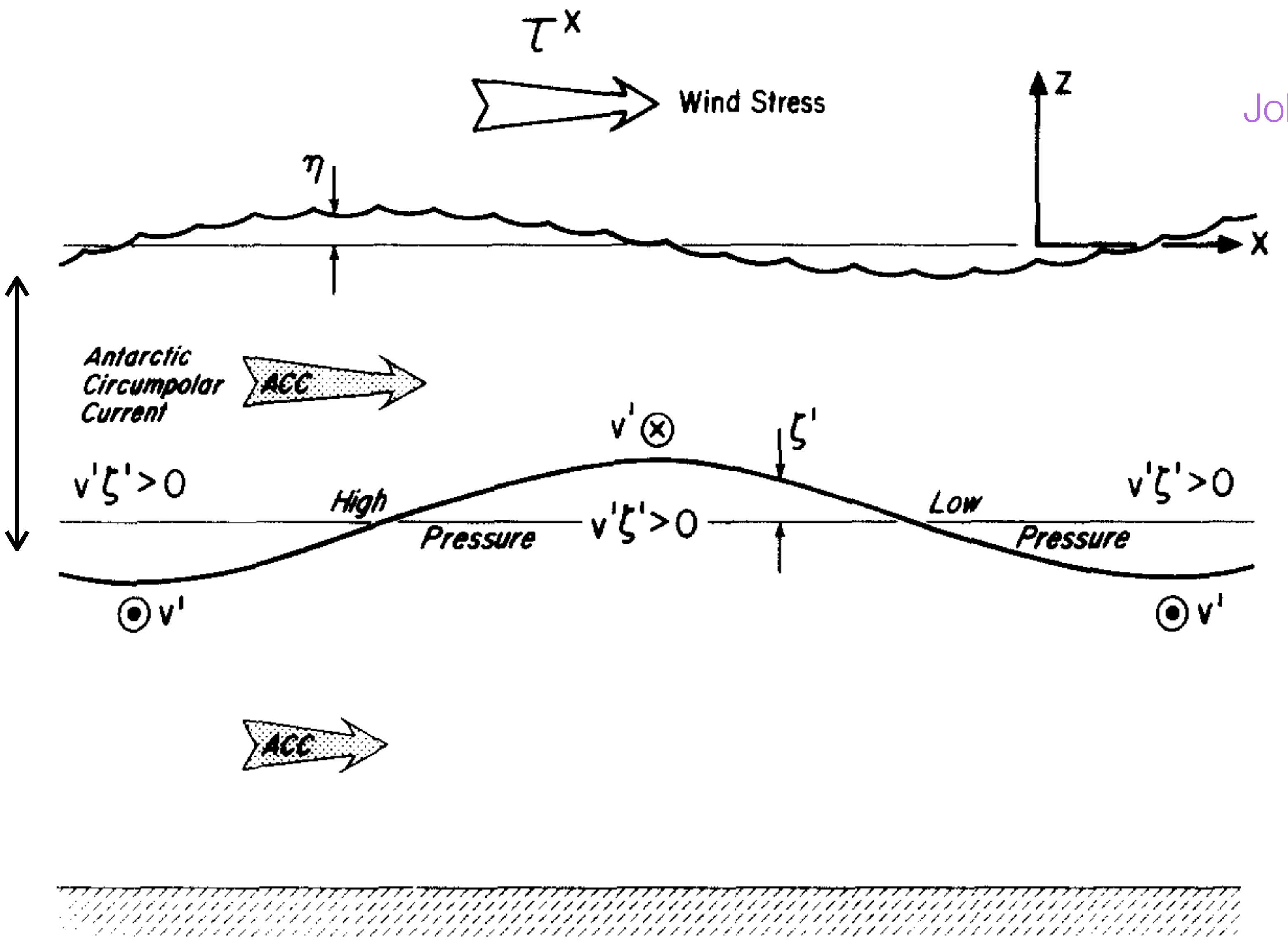
Topographic form stress is a purely **barotropic** process.

interfacial form stress

Johnson & Bryden 1989

vertically integrate from the sea-surface down to a moving buoyancy surface

(i.e., integrate within a layer of constant density)



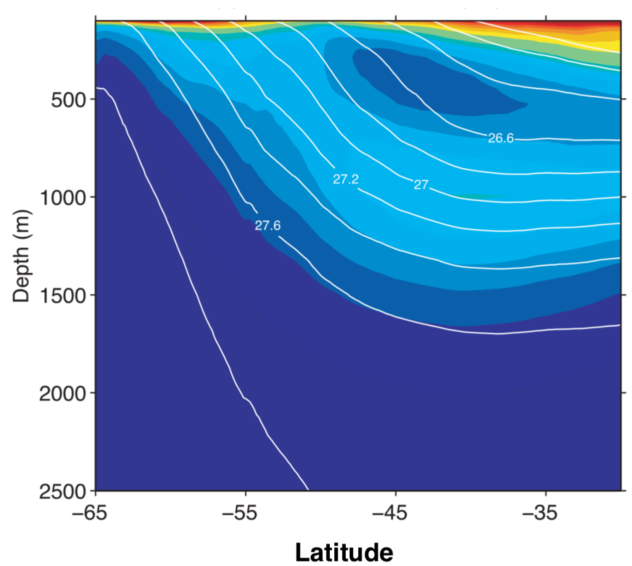
Schematic presentation of interfacial form drag. Correlations of perturbations in the interface height, ζ', and the meridional velocity, V' ($\odot$ indicating poleward flow and $\otimes$ indicating equatorward flow), which are related to pressure perturbations by geostrophy, allow the upper layer to exert an eastward force on the lower layer and the lower layer to exert a westward force on the upper layer; thus effecting a downward flux of zonal momentum.

Interfacial form stress requires **baroclinicity**.

the most popular scenario for the momentum balance

- momentum in imparted at the surface by wind,
- isopycnals slope $\longrightarrow$ baroclinic instability,
- momentum is transferred downwards by interfacial eddy form stress
- momentum reaches the bottom where is transferred to the solid Earth by topographic form stress.

Johnson & Bryden 1989



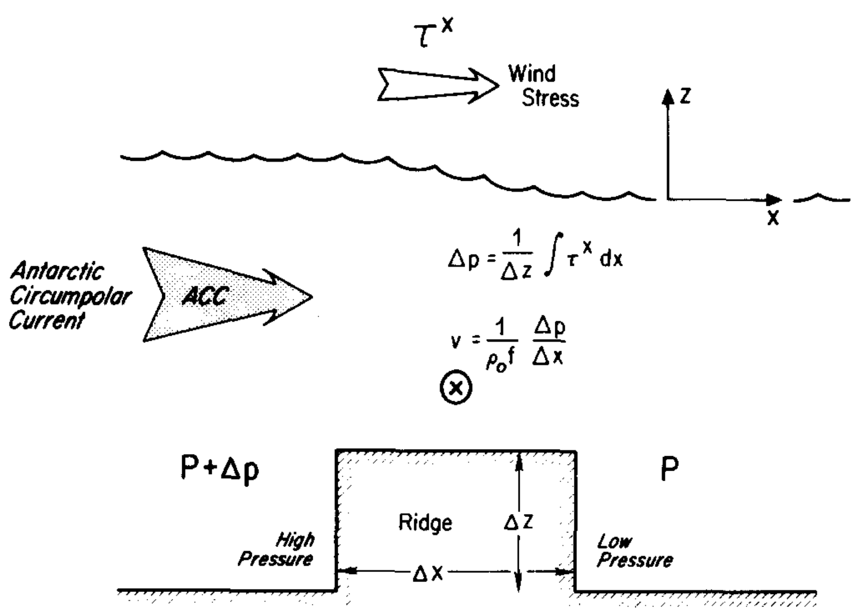
Meredith et al. 2012

$$\begin{matrix}\text{isopycnal}\\ \text{slope}\end{matrix} = \left[-\frac{\tau_s}{f\,\kappa}\right]^{1/2}$$

Marshall & Radko 2003

This baroclinic scenario sets up the ACC transport (e.g. the transport through Drake Passage).

but what about barotropic dynamics?



The sea surface pressure gradient can be *directly* communicated to the bottom.

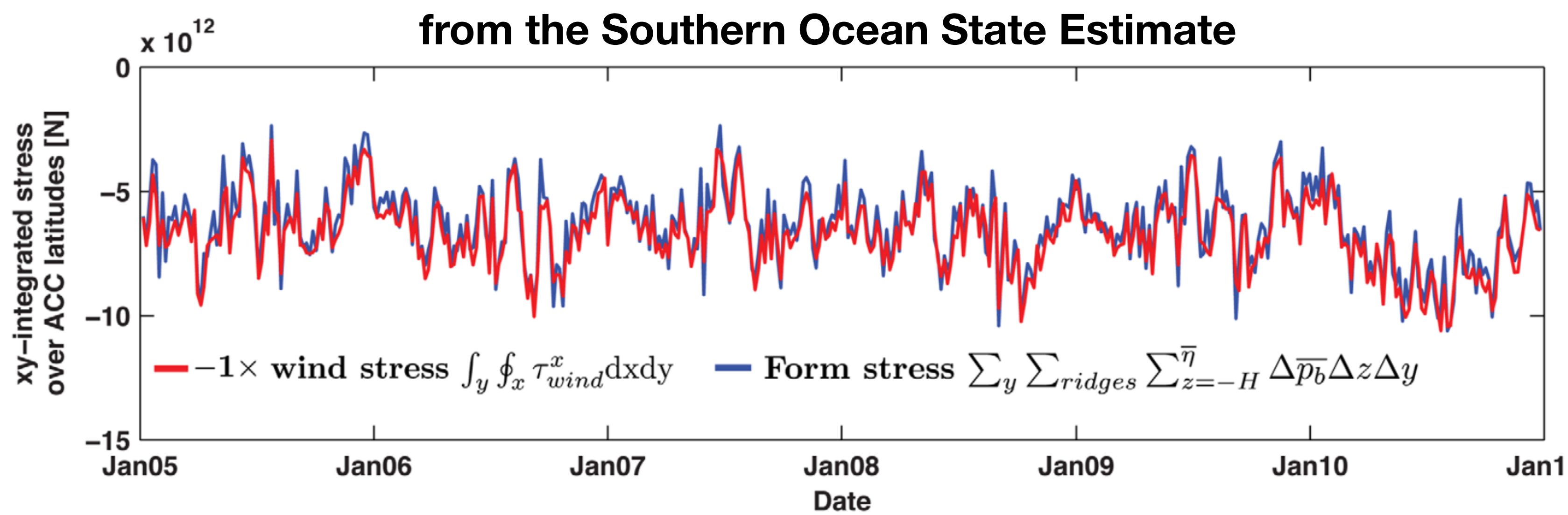
And it will be, unless compensated by internal isopycnal gradients.

Isn't barotropic "communication" much "easier"?

wind stress is *rapidly* communicated to the bottom through barotropic processes

Barotropic processes are fast (~days).

Baroclinic processes are much slower (~years).



Masich, Chereskin, and Mazloff 2015

~90% of variance in the topographic form stress signal is explained by the **0-day** time lag.

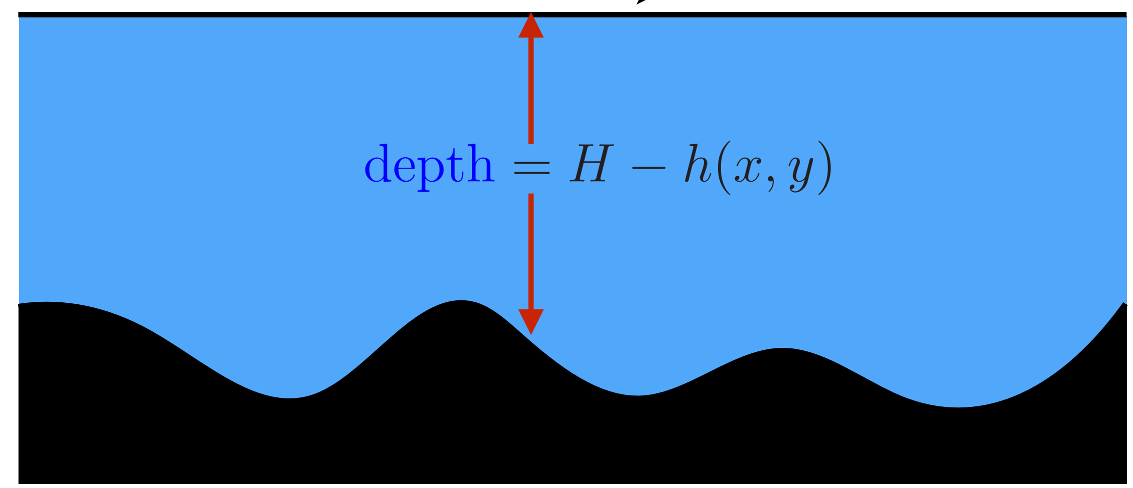
Similar statements also made by:
Straub 1993, Ward & Hogg 2011, Rintoul et al. 2014, Peña Molino et al. 2014, Donohue et al. 2016.

the plan

Revisit an old barotropic quasigeostrophic (QG) model on a beta-plane.

(Hart 1979, Davey 1980, Bretherton & Haidvogel 1976, Holloway 1987, Carnevale & Fredericksen 1987)

steady mean zonal wind stress
F



A distinctive feature of this model is a "large-scale barotropic flow" *U*(*t*).

this is the ACC

topographic potential vorticity (PV) $\eta = \dfrac{f_0 h}{H}$

QGPV $\nabla^2\psi + \eta + \beta y$

total streamfunction $-U(t)y + \psi(x, y, t)$

Study how momentum is balanced by topographic form stress and investigate the requirements for eddy saturation.

a barotropic QG model for a mid-ocean region

total streamfunction $\quad -U(t)y + \psi(x,y,t)$

QGPV $\quad \nabla^2\psi + \eta + \beta y$

Material conservation of QGPV

$$\nabla^2\psi_t + U(\nabla^2\psi + \eta)_x + \mathsf{J}(\psi, \nabla^2\psi + \eta) + \beta\psi_x = -\mu\,\nabla^2\psi + \text{hyper visc.}$$

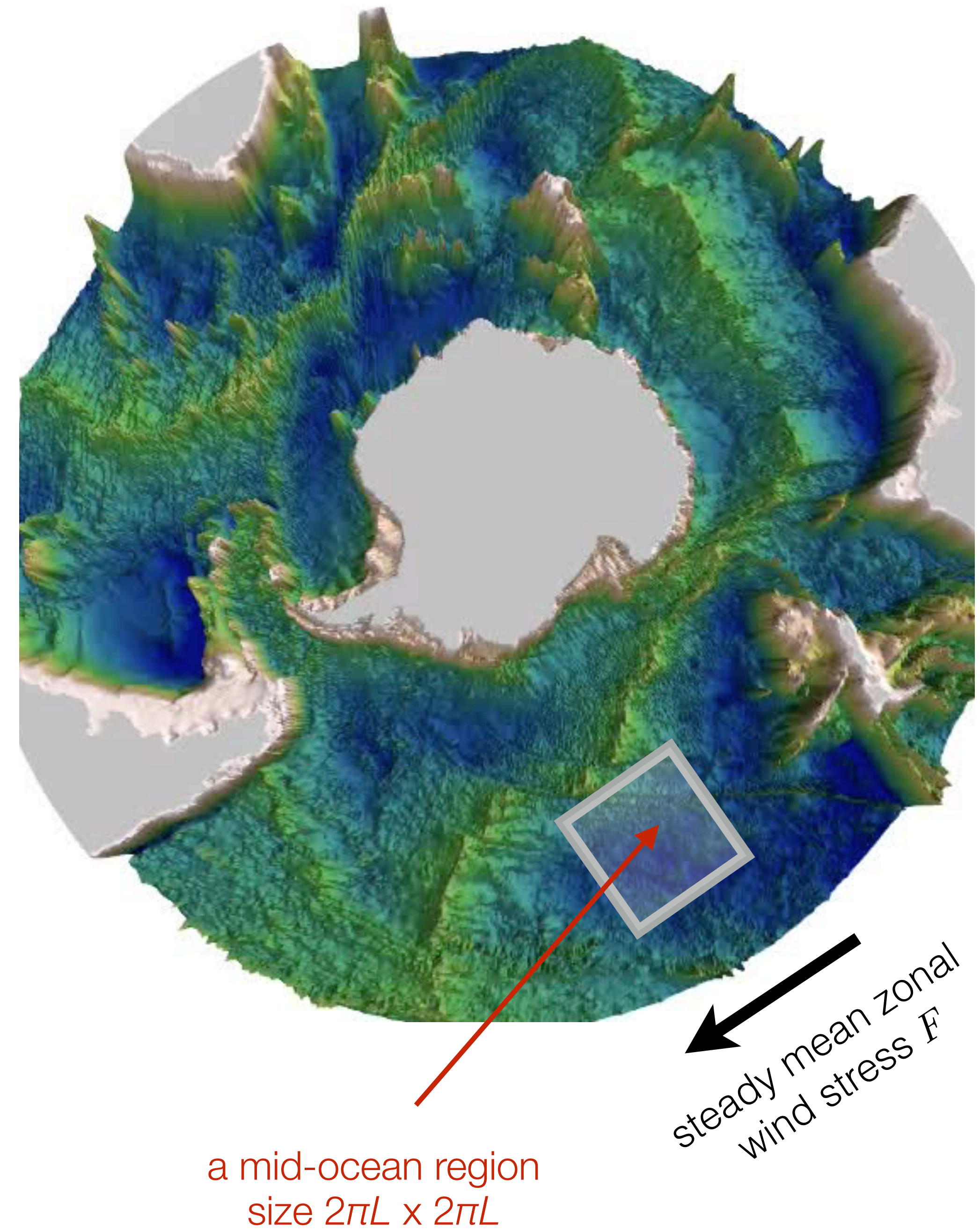
Large-scale zonal momentum

$$U_t = F - \mu U - \langle \psi \eta_x \rangle$$

topographic form stress

$\langle\ \rangle$ is domain average ; $\quad F = \dfrac{\tau_s}{\rho_0 H}$ wind stress forcing

periodic boundary conditions



(Hart 1979, Davey 1980, Bretherton & Haidvogel 1976, Holloway 1987, Carnevale & Fredericksen 1987)

the large-scale flow equation: $U_t = F - \mu U - \langle \psi \eta_x \rangle$

zonal angular momentum density: $a(x,y,z,t) = u(x,y,z,t) - \int^{y} f(y')\,\mathrm{d}y'$

vertically integrated zonal angular momentum equation

$$\partial_t \int_{-h}^{0} a\,\mathrm{d}z + \partial_x \left[\int_{-h}^{0} ua + p\,\mathrm{d}z \right] + \partial_y \int_{-h}^{0} va\,\mathrm{d}z =$$
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vertically integrated zonal angular momentum equation

$$\partial_t \int_{-h}^{0} a\,\mathrm{d}z \cancel{+ \partial_x \left[\int_{-h}^{0} ua + p\,\mathrm{d}z\right] + \partial_y \int_{-h}^{0} va\,\mathrm{d}z} =$$

$$= \underbrace{\tau(0)}_{\text{wind stress}} - \underbrace{\tau(-h)}_{\text{bottom drag}} + \underbrace{h_x p(-h)}_{\text{form stress}}$$

horizontally integrate, drop the boundary fluxes, and divide by the volume

$$U_t = F - \mu U - \langle \psi \eta_x \rangle$$

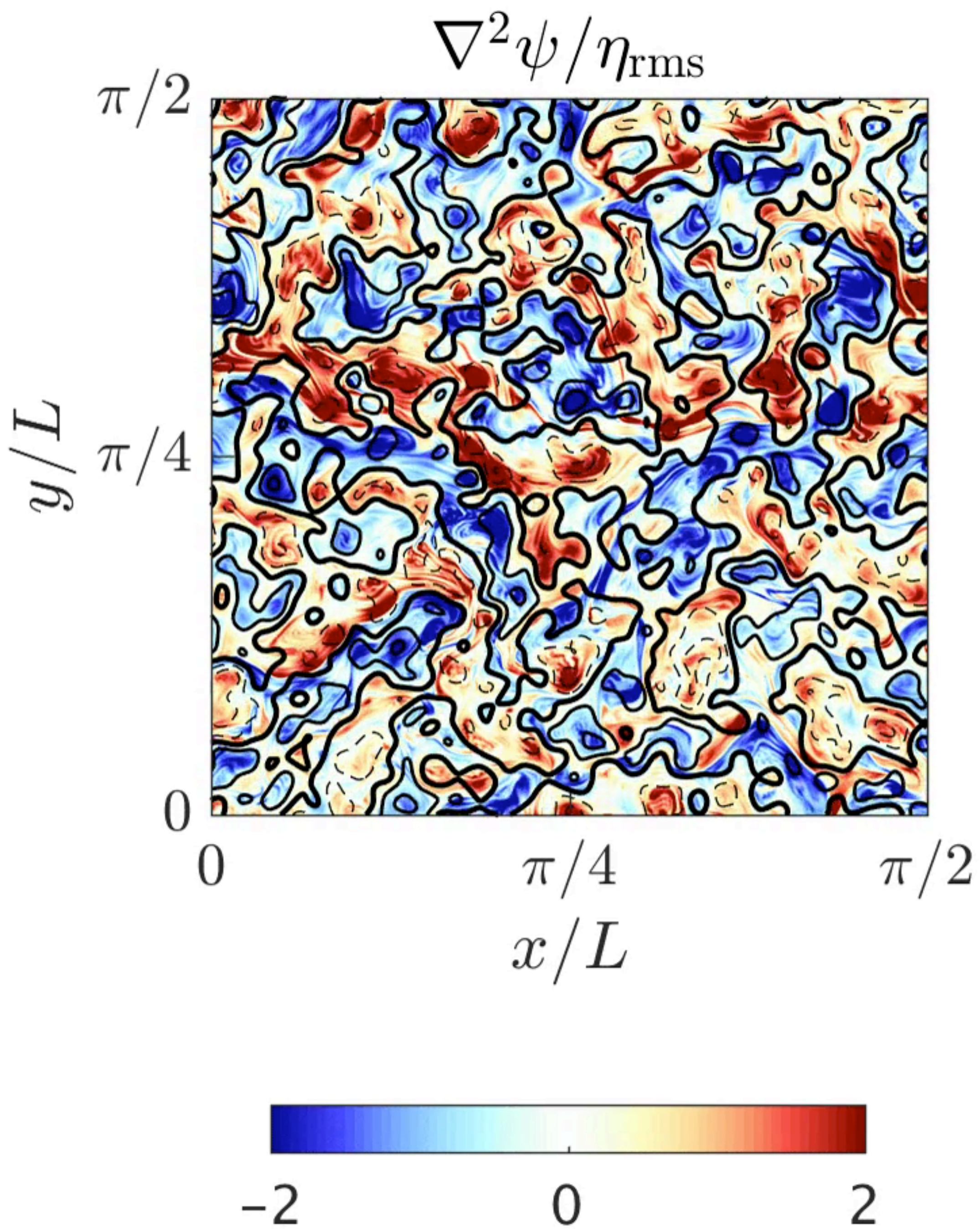
$$U(t) \stackrel{\text{def}}{=} V^{-1} \iiint u(x,y,z,t)\,\mathrm{d}V$$

vertical & horizontal integral over a mid-ocean region (**not** a zonal average)

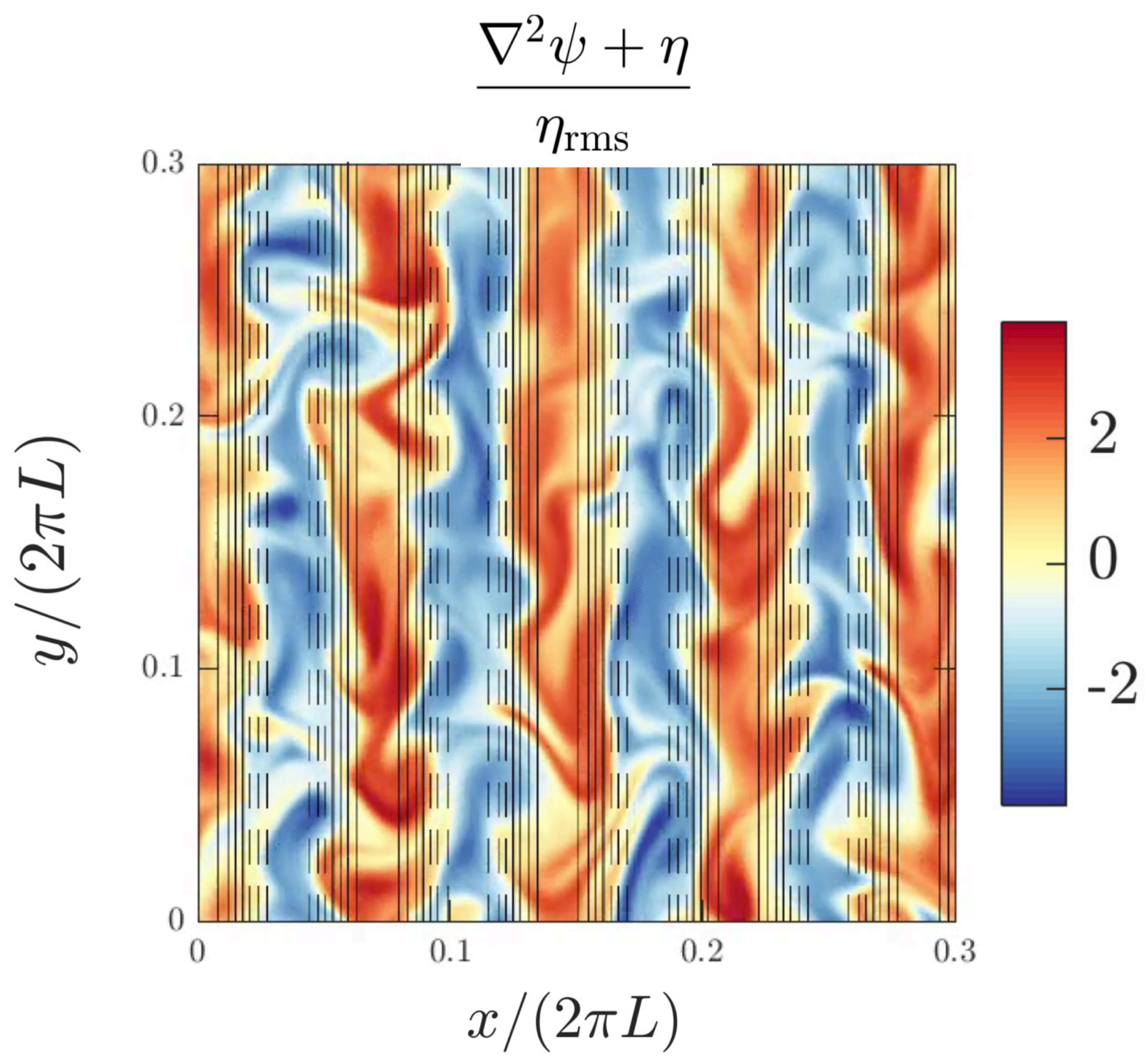
this barotropic QG model exhibits turbulence and eddies

random topography
with k^{-2} spectrum

$\mu t = 3.15$



$h \propto \cos(mx)$



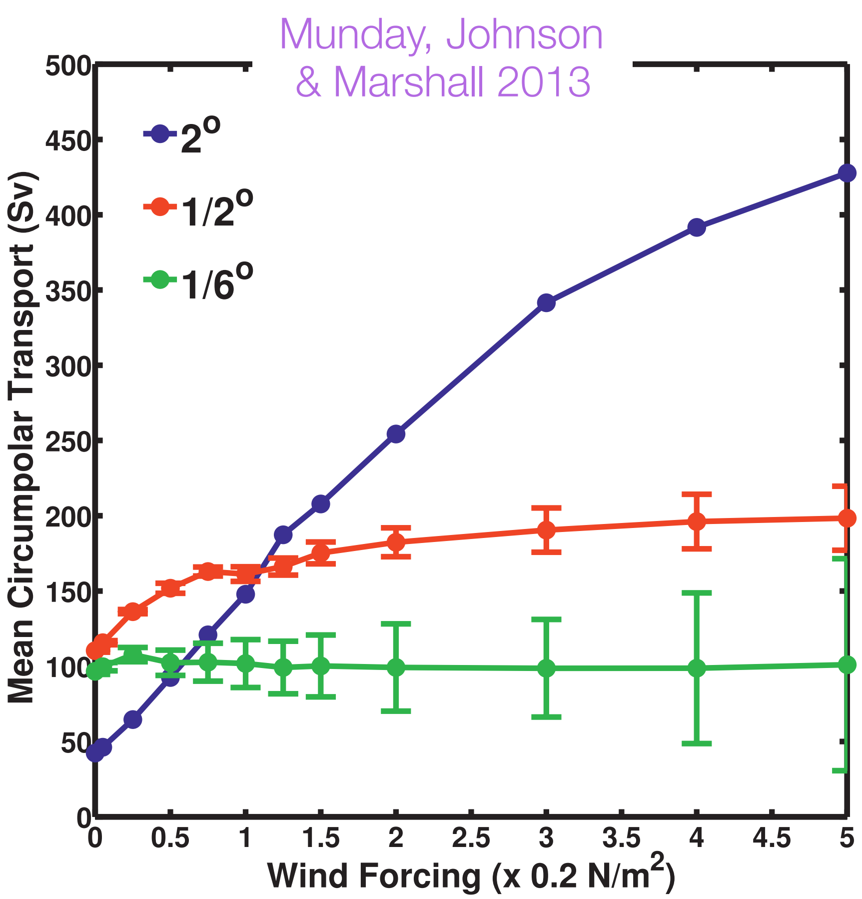
Question:

Does this barotropic QG model
show eddy saturation?

Do we need baroclinicity?
Do we even need channel walls?

but first, what is “eddy saturation”?

The *insensitivity* of the total ACC volume transport
to wind stress increase.



Eddy saturation is seen in eddy-resolving ocean models.

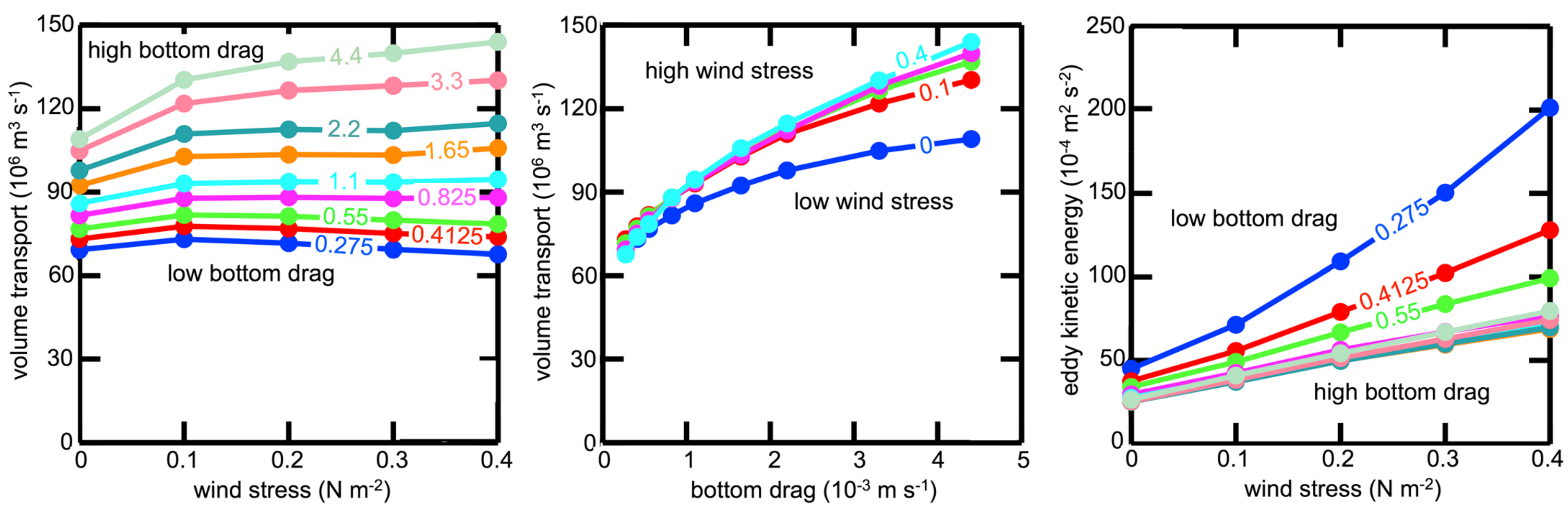
Higher resolution ⟶ eddy saturation “occurs”

Eddy saturation was theoretically predicted by Straub (1993)
but with an *entirely* baroclinic argument.
(based on vertical momentum transfer interfacial eddy form stress)

[There are many other examples: Hallberg & Gnanadesikan 2001, Tansley & Marshall 2001, Hallberg & Gnanadesikan 2006, Hogg et al. 2008, Nadeau & Straub 2009, Farneti et al. 2010, Nadeau & Straub 2012, Meredith et al. 2012, Morisson & Hogg 2013, Abernathey & Cessi 2014, Farneti et al. 2015, Nadeau & Ferrari 2015, Marshall et al. 2016.]

yet more eddy saturation

Marshall, Ambaum, Maddison, Munday & Novak 2016



increasing the bottom drag increases the transport

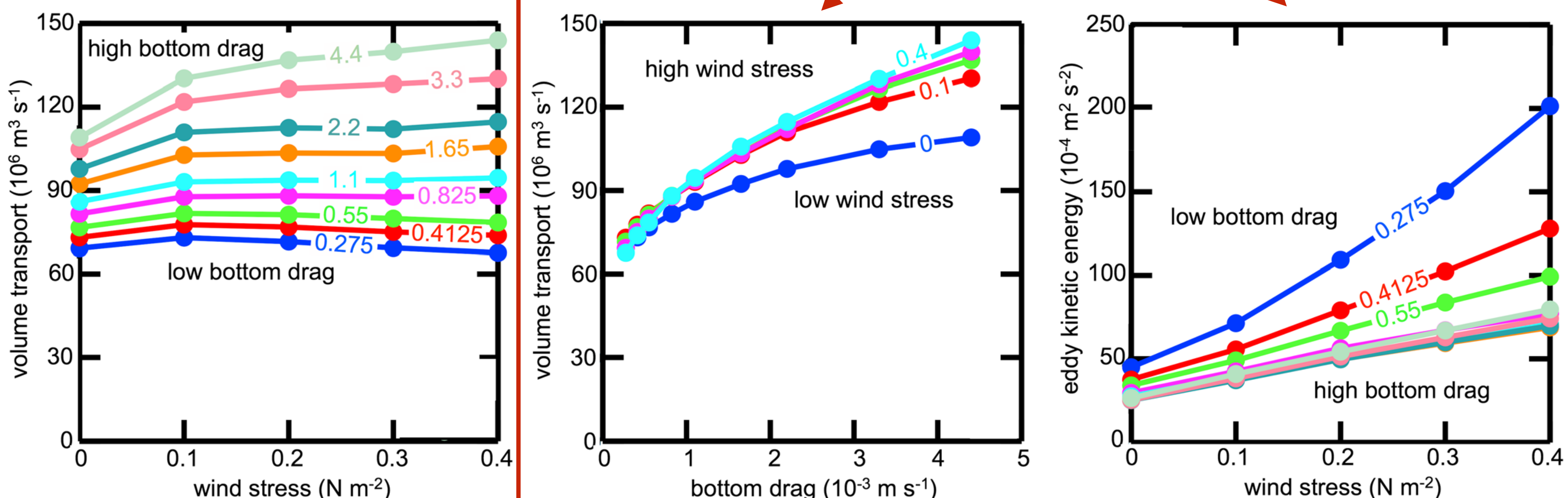
EKE varies linearly with wind stress

[See also: Hogg & Blundell 2006, Nadeau & Straub 2012, and Nadeau & Ferrari 2015.]

yet more eddy saturation

Marshall, Ambaum, Maddison, Munday & Novak 2016

further "symptoms" of eddy saturation



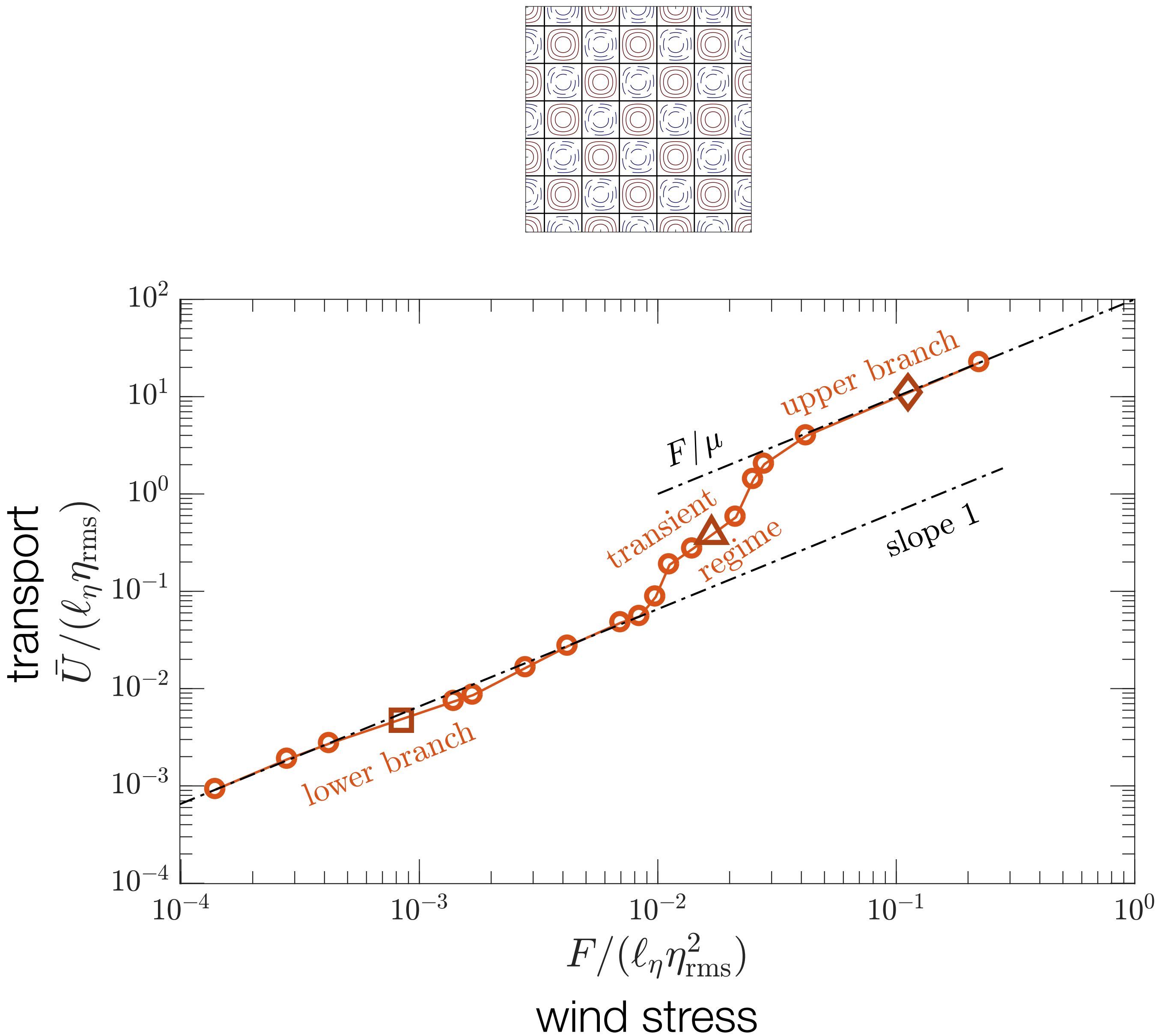
increasing the bottom drag increases the transport

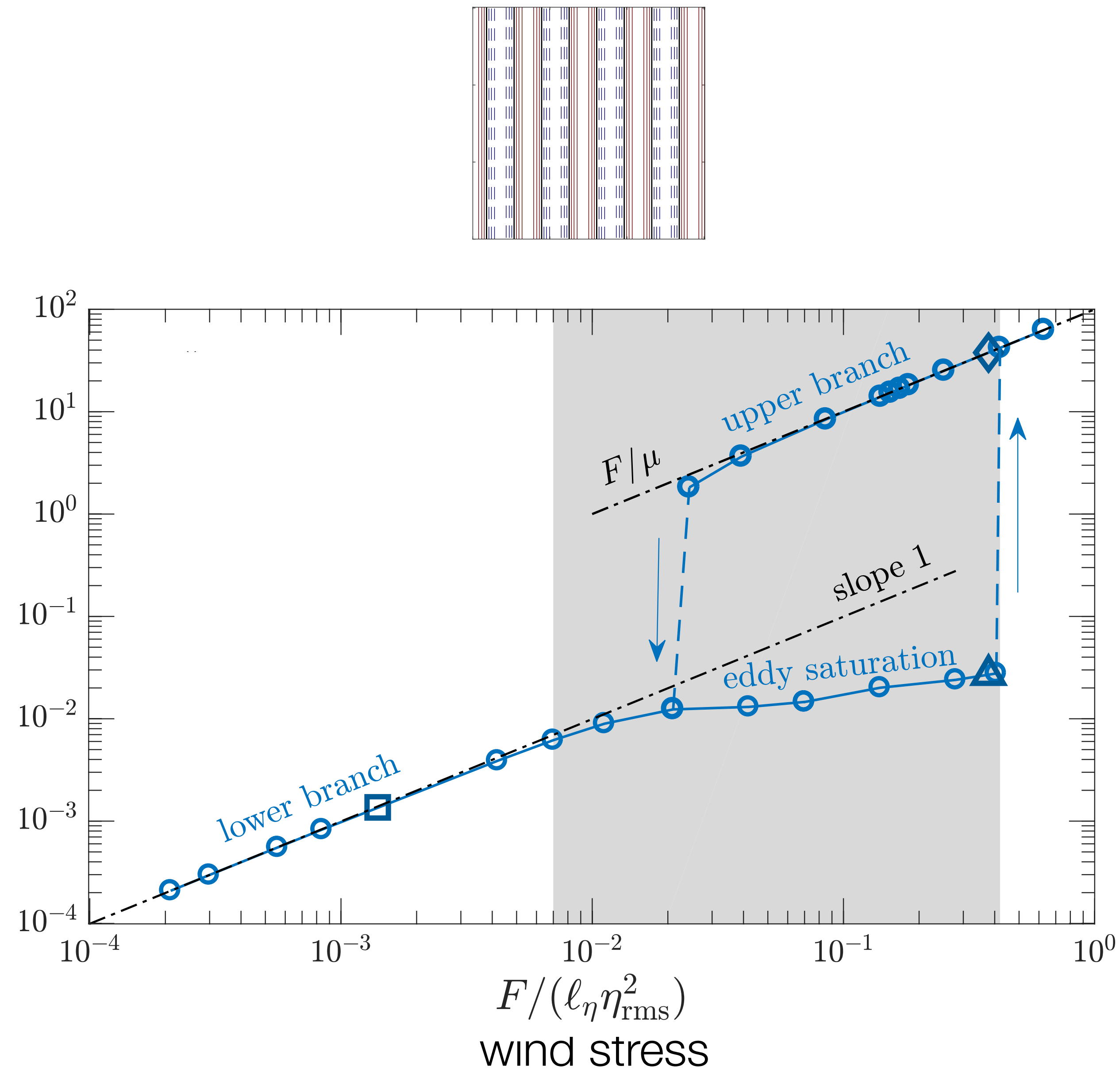
EKE varies linearly with wind stress

[See also: Hogg & Blundell 2006, Nadeau & Straub 2012, and Nadeau & Ferrari 2015.]

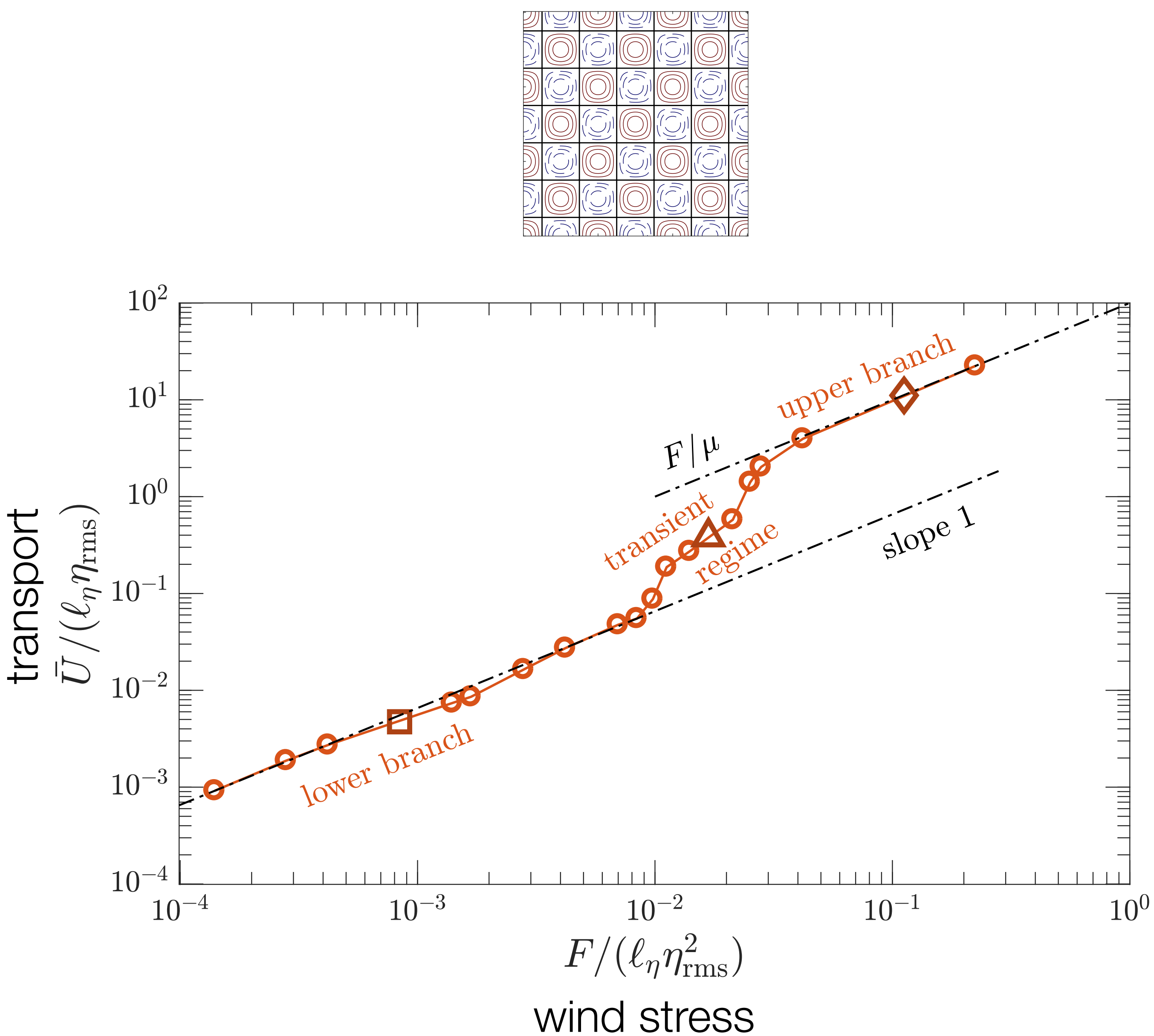
how does the transport vary with wind stress in this barotropic QG model?

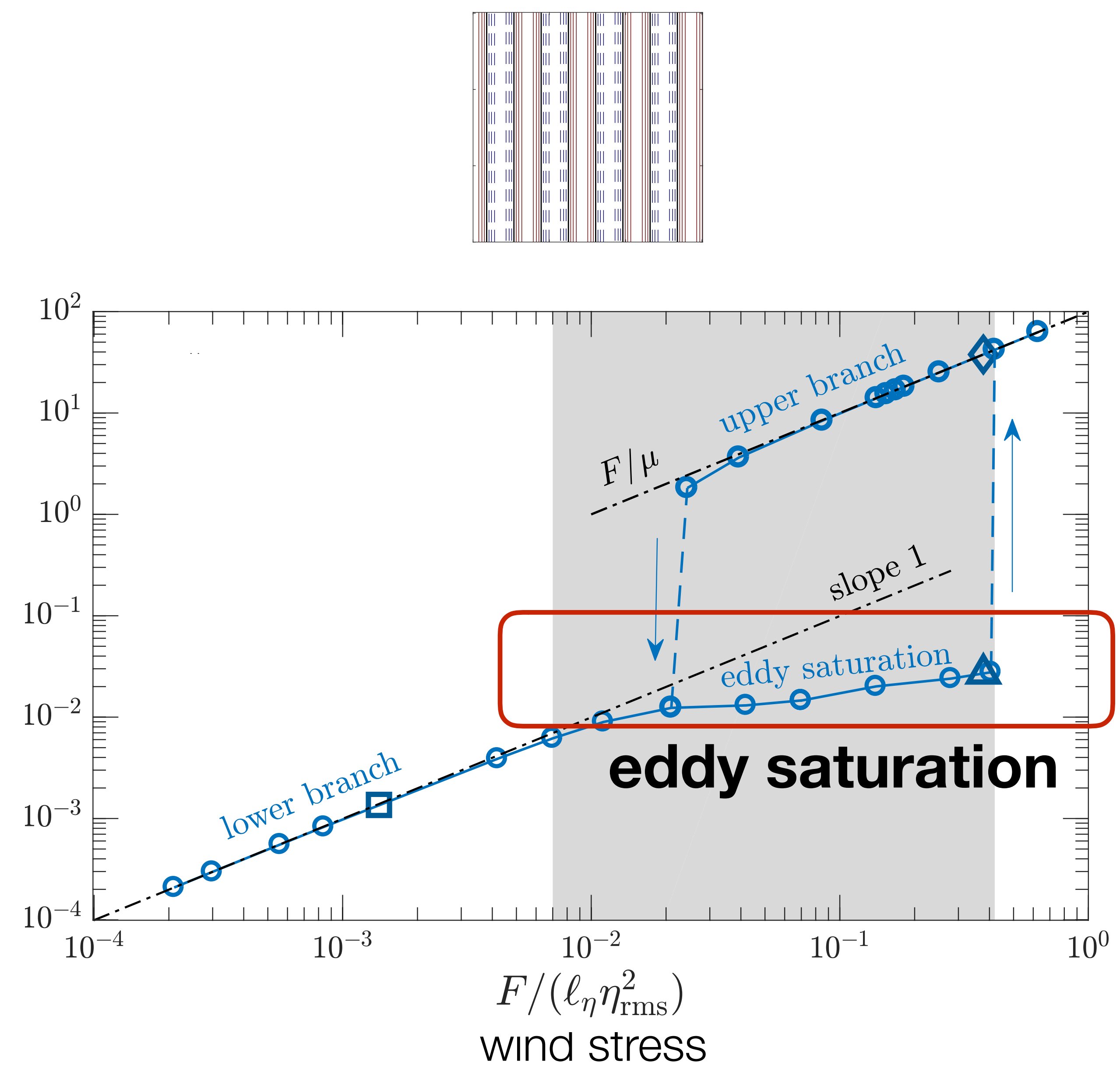
all parameters same, different topography



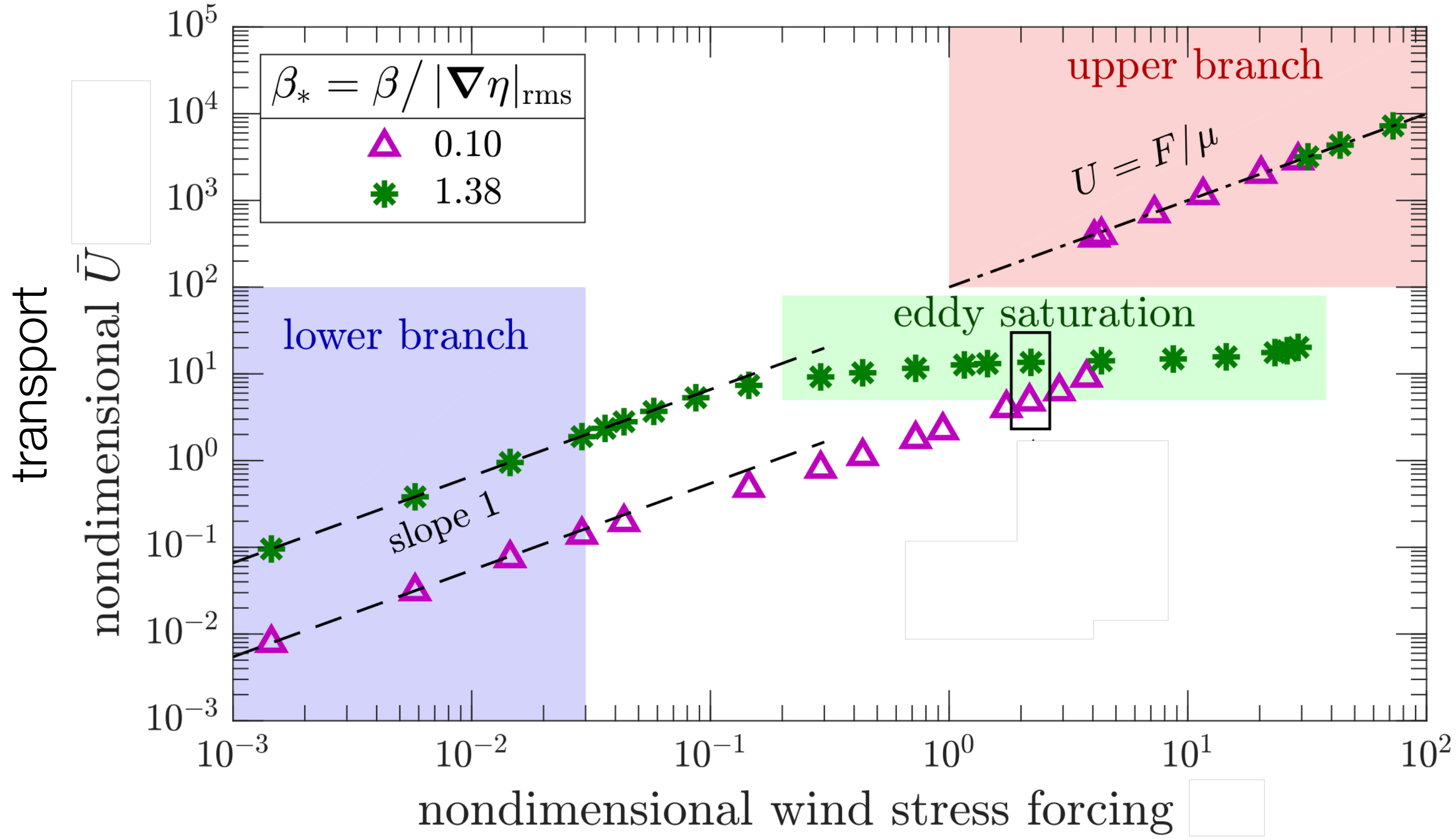


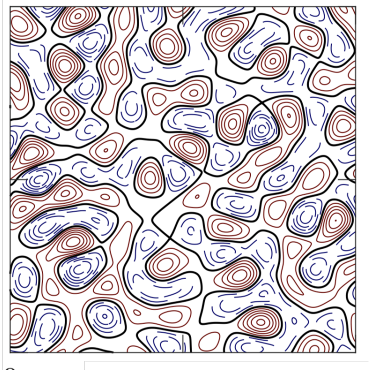
all parameters same, different topography

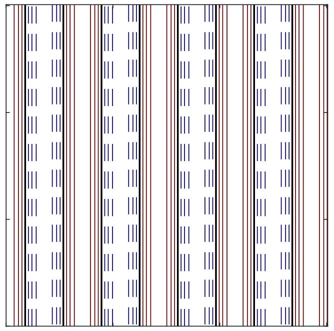




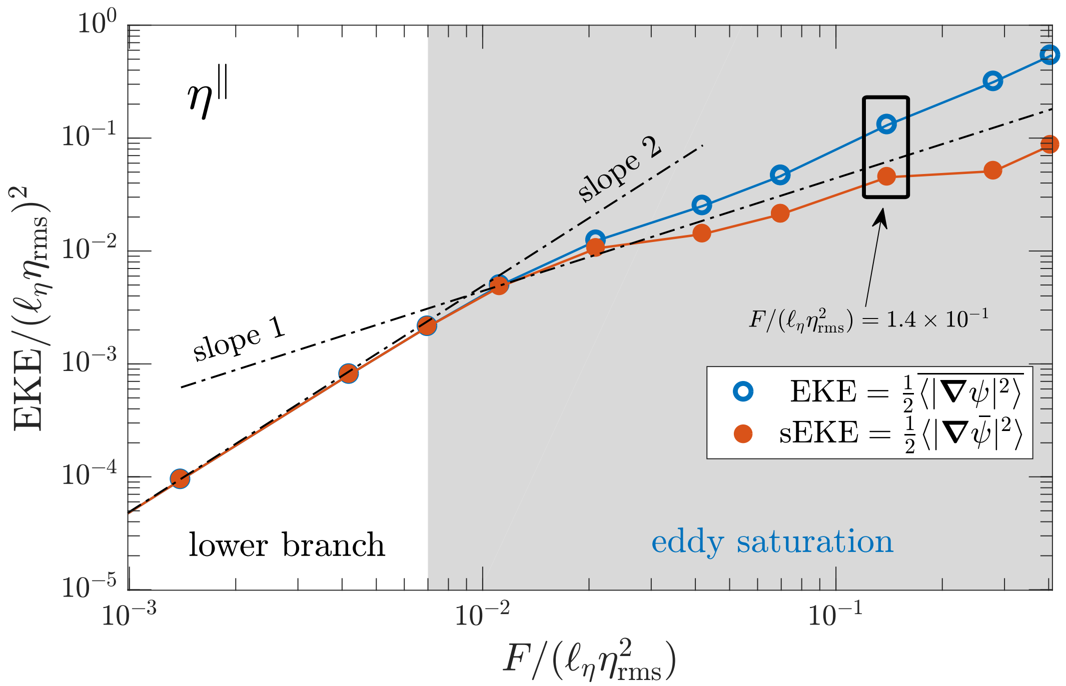
all parameters same, different value of β





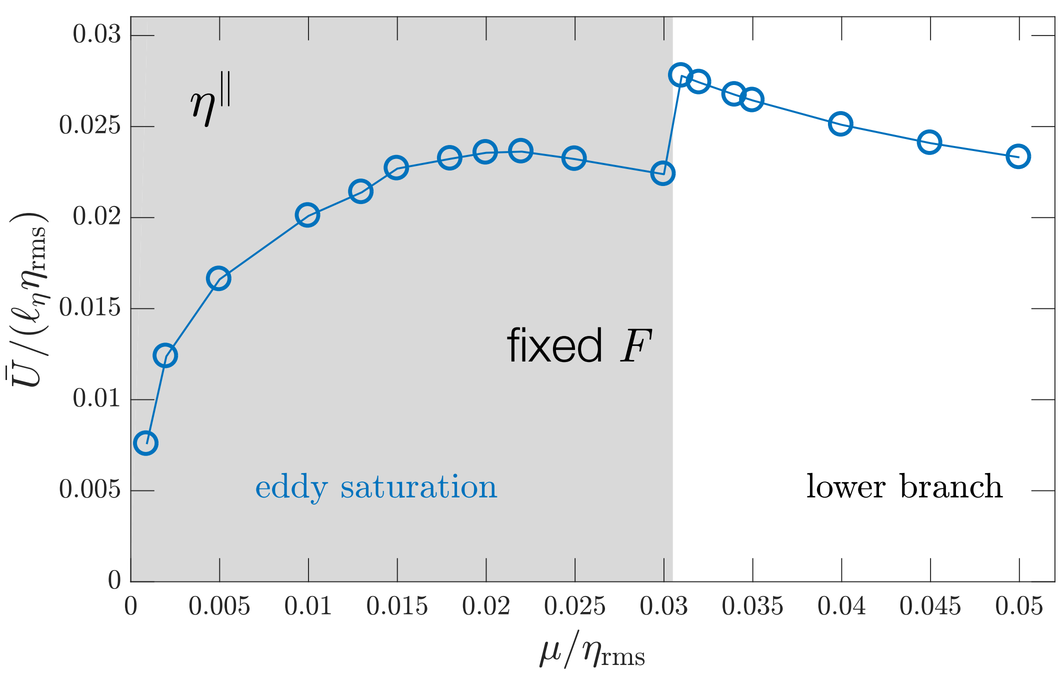


some further "symptoms" of eddy saturation



wind stress

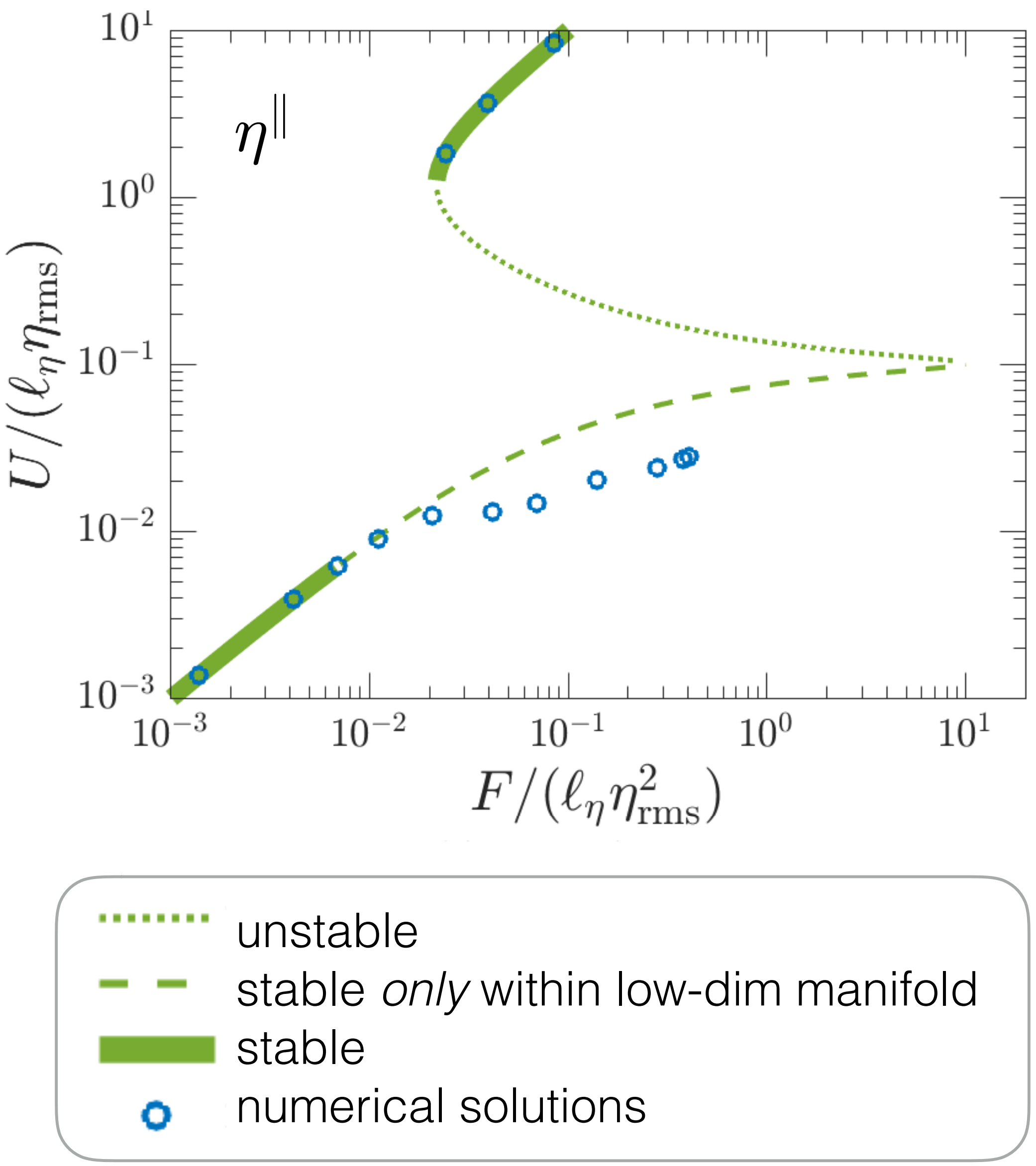
EKE grows roughly linearly with wind stress

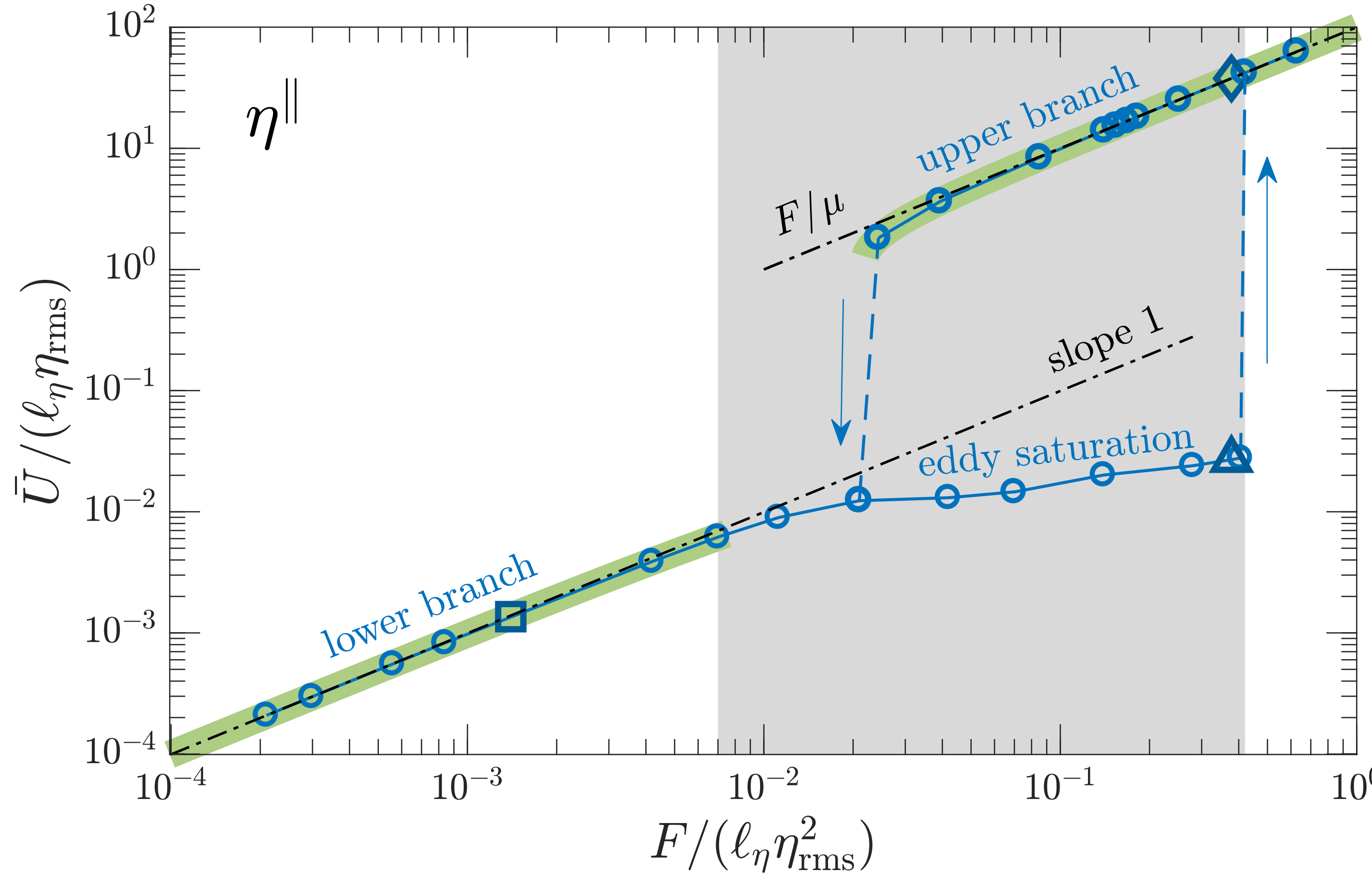


bottom drag

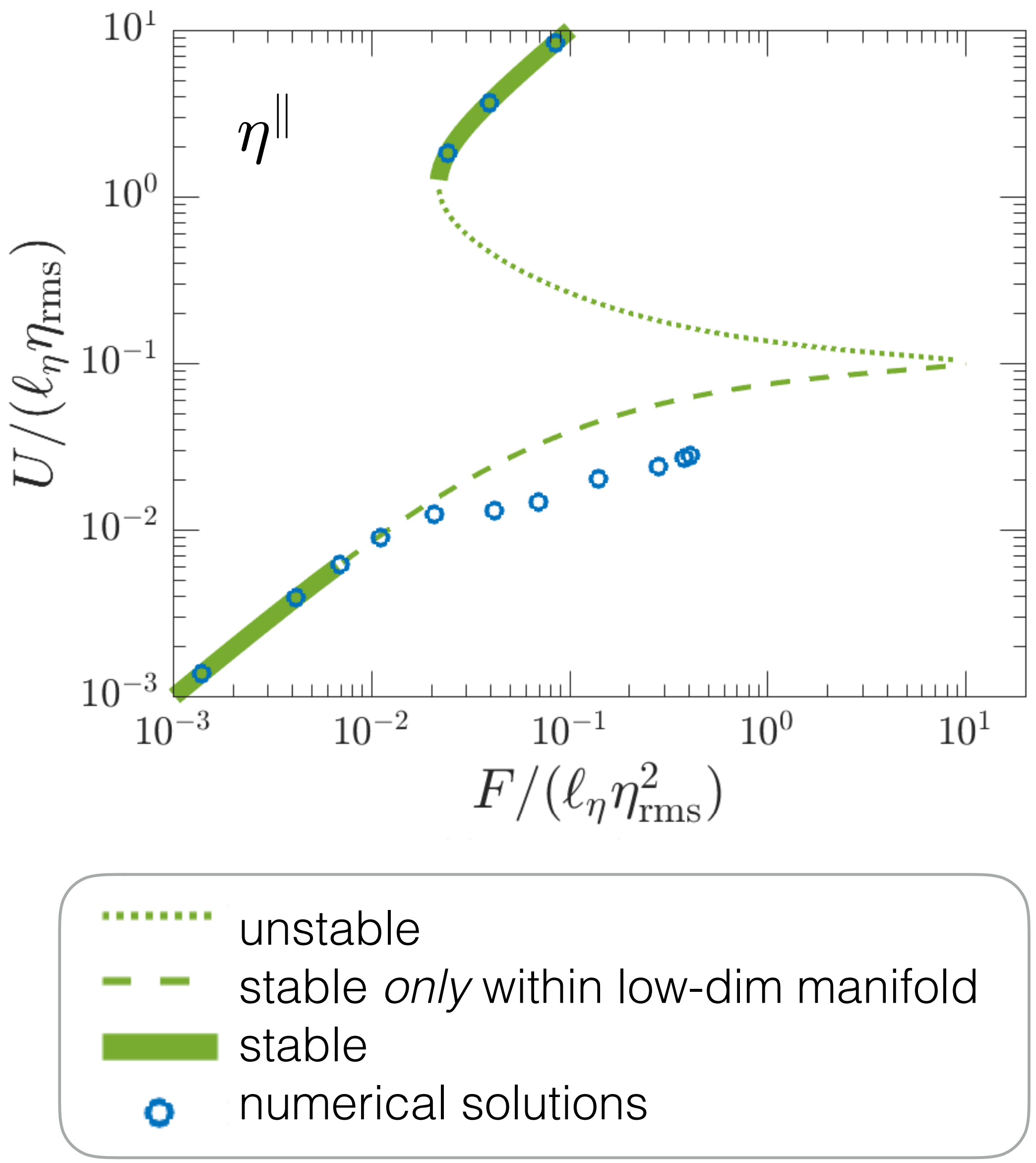
transport grows with increasing bottom drag what produces eddy saturated states in this barotropic QG model?

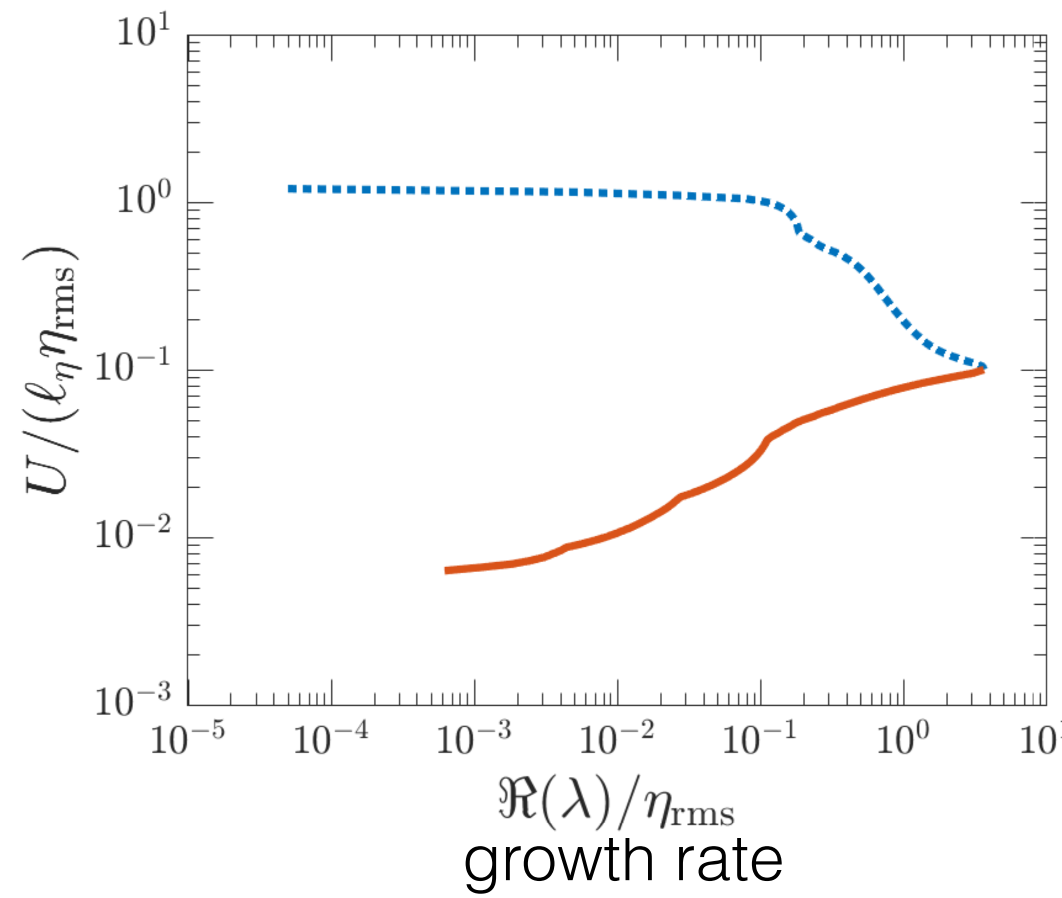
stability analysis for $\eta^{\parallel}=\eta_0 \cos(mx)$





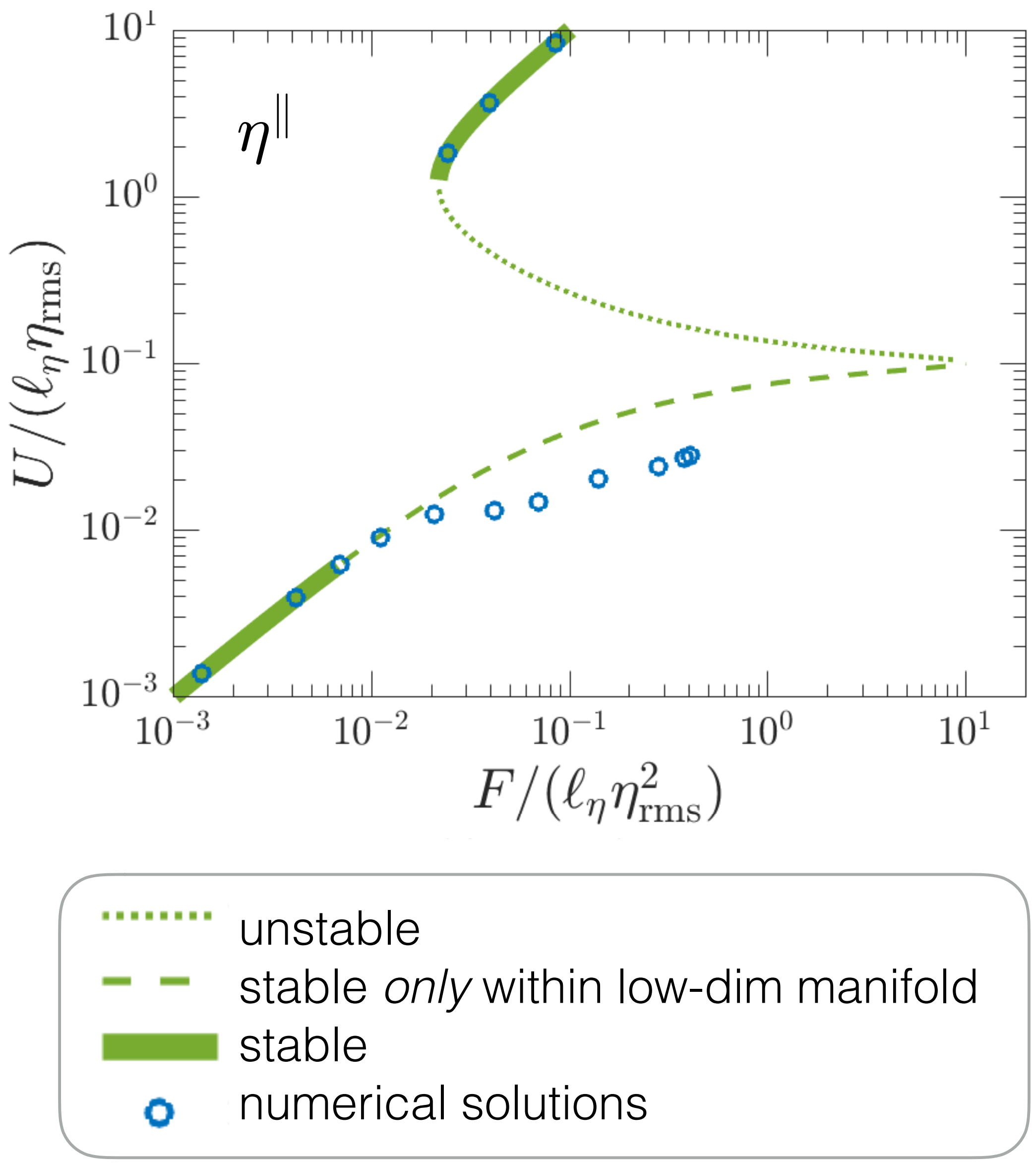
stability analysis for $\eta^{\|}=\eta_0 \cos(mx)$

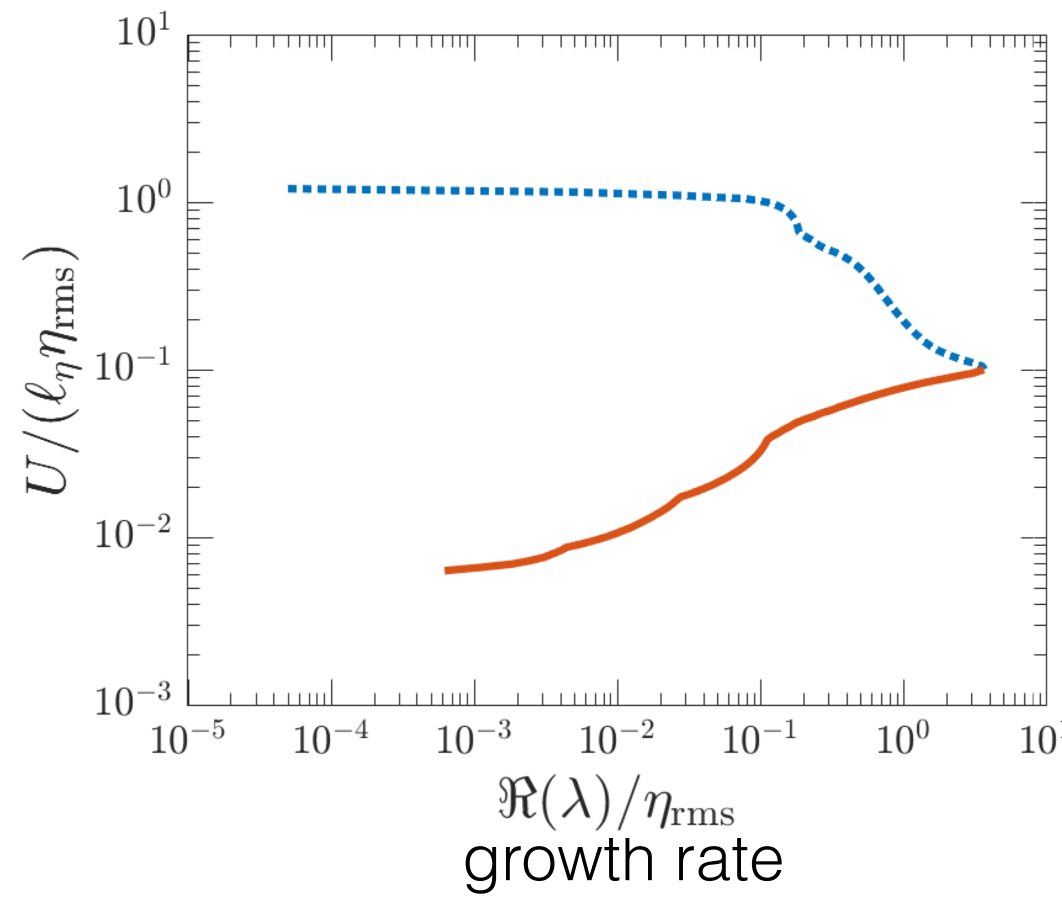




Max instability growth rate increases ~10^4 times with a 10-fold increase in U!

stability analysis for $\eta^{\|}=\eta_0 \cos(mx)$





Max instability growth rate increases ~10^4 times with a 10-fold increase in U!

Minor changes in U ⟶ large transient energy production.

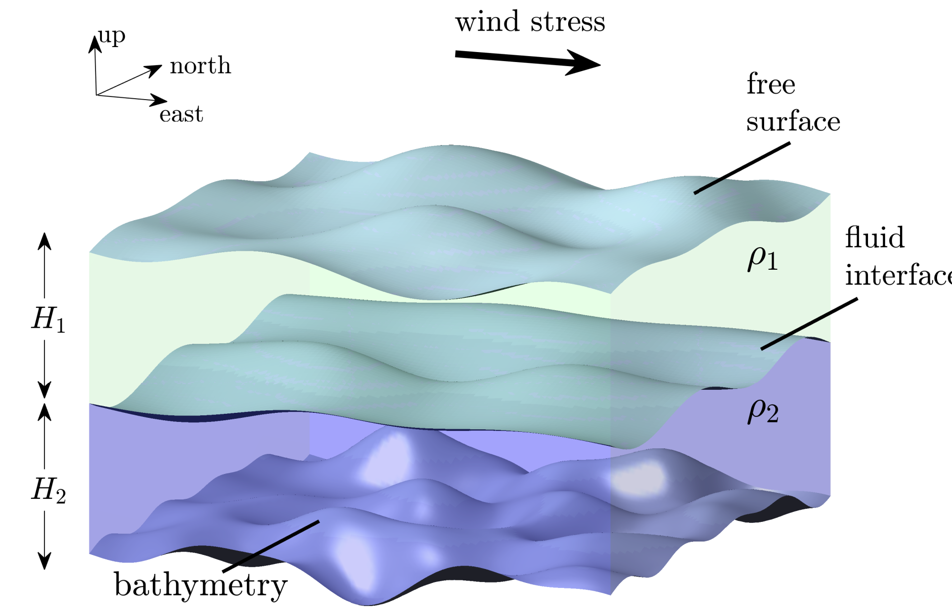
Transient eddies balance most of the momentum imparted by F ⟶ eddy saturation.

(Similarly as in the baroclinic scenario.)

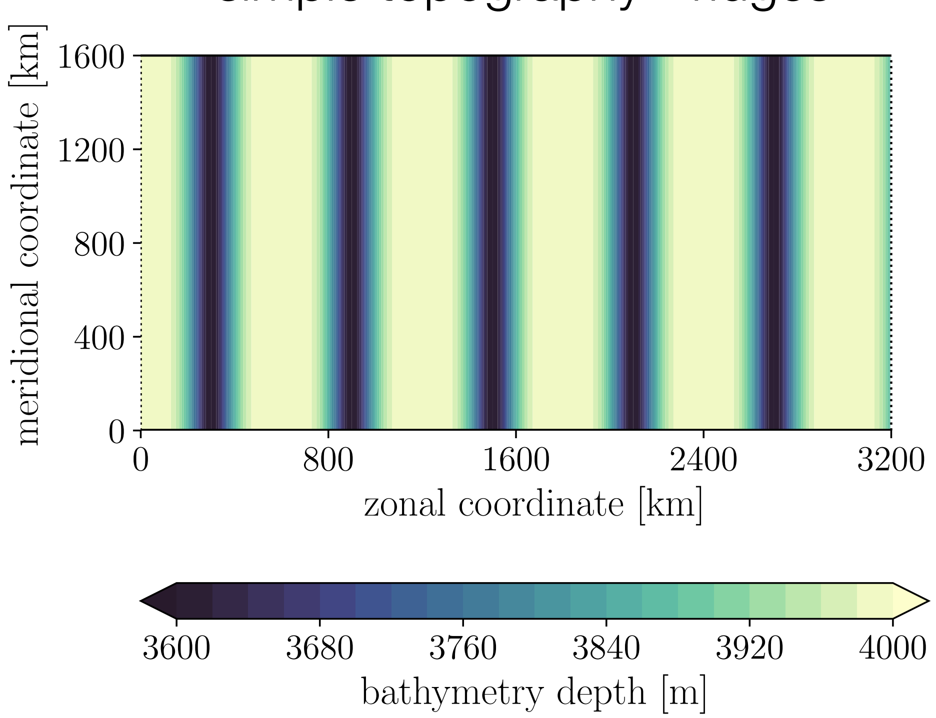
let's change page now

a setup with both BT and BC eddy saturations

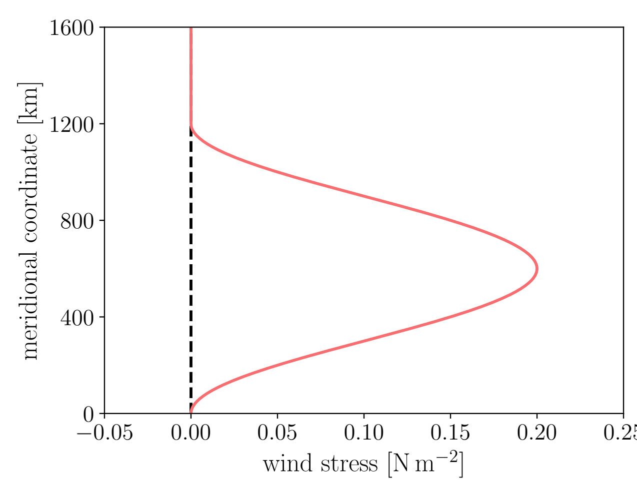
- Idealized re-entrant channel with "bumpy" bottom
- $L_x = 3200$ km, $L_y = 1600$ km, and $H = 4$ km
- beta-plane with Southern Ocean parameters
- Modest stratification (few fluid layers of constant ϱ)
- 1st Rossby radius of deformation: 15.7 km (for >1 layers)
- Modular Ocean Model v6 (MOM6) in isopycnal mode



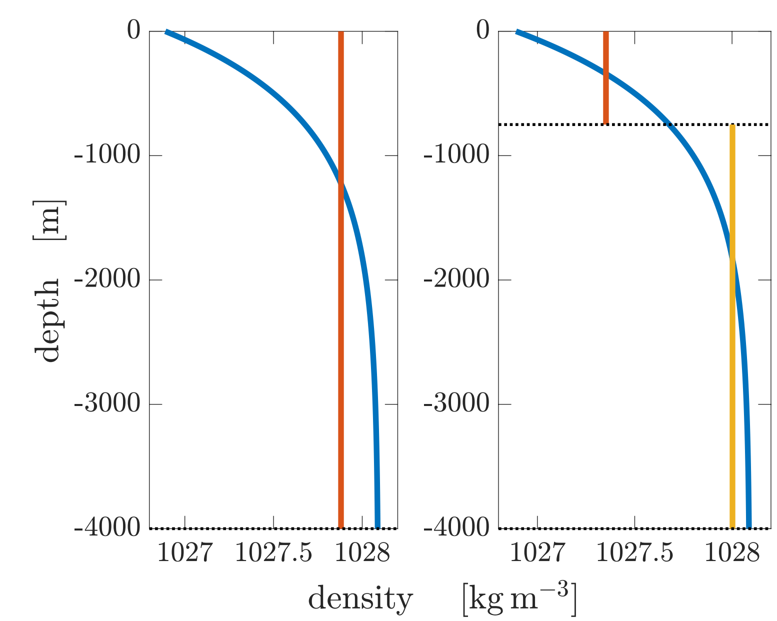
simple topography - ridges



wind stress

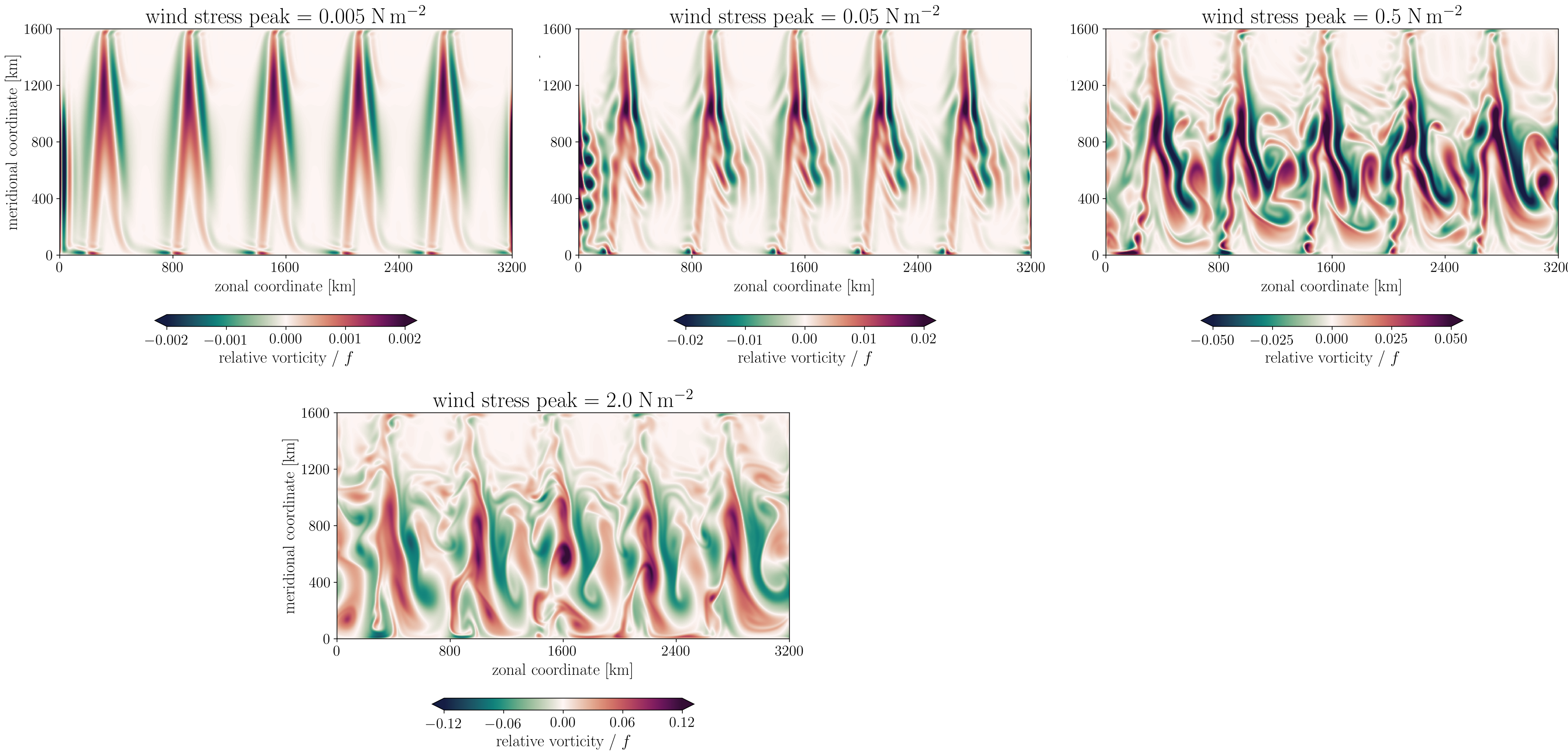


stratification



1-layer 2-layer

flow structure for 1-layer configuration



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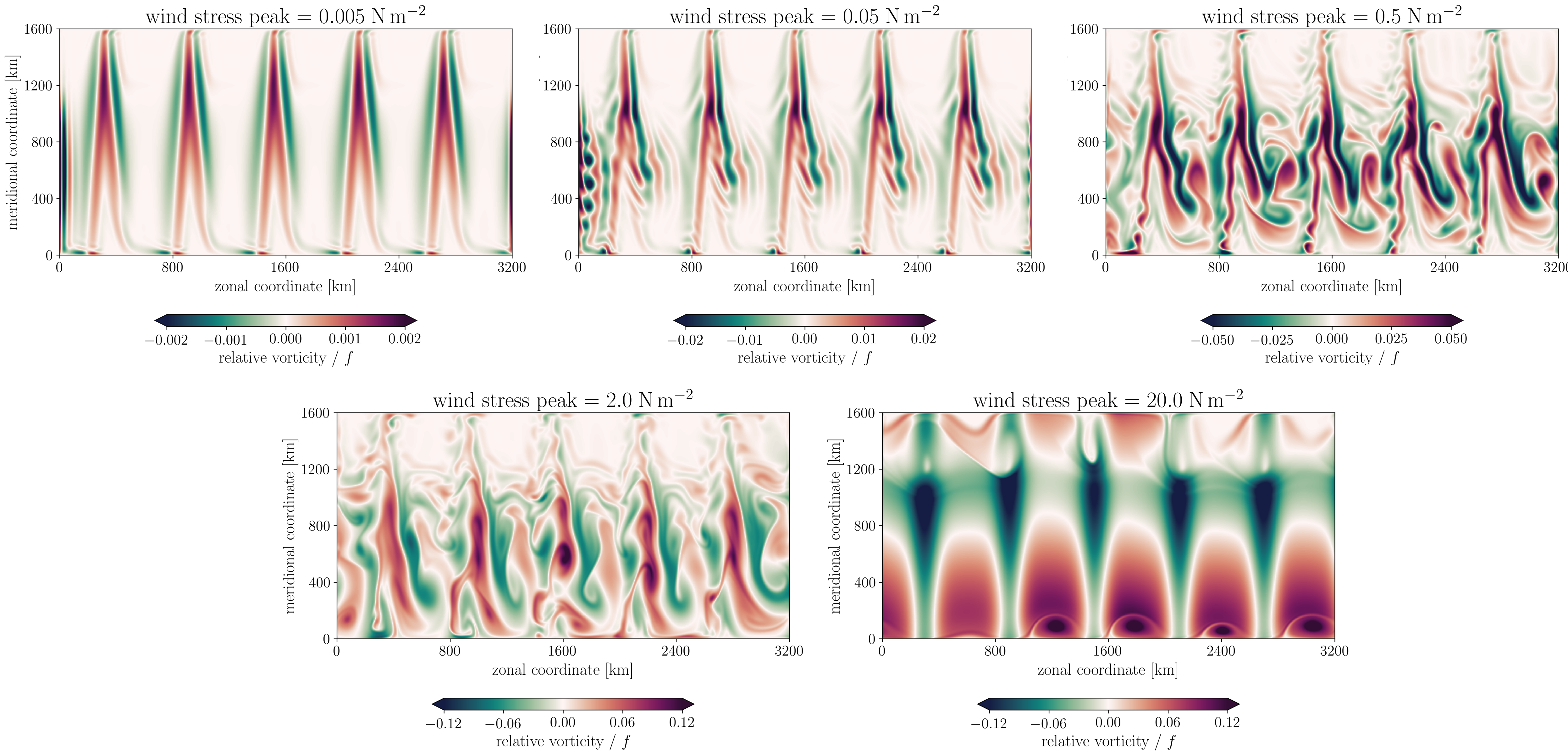
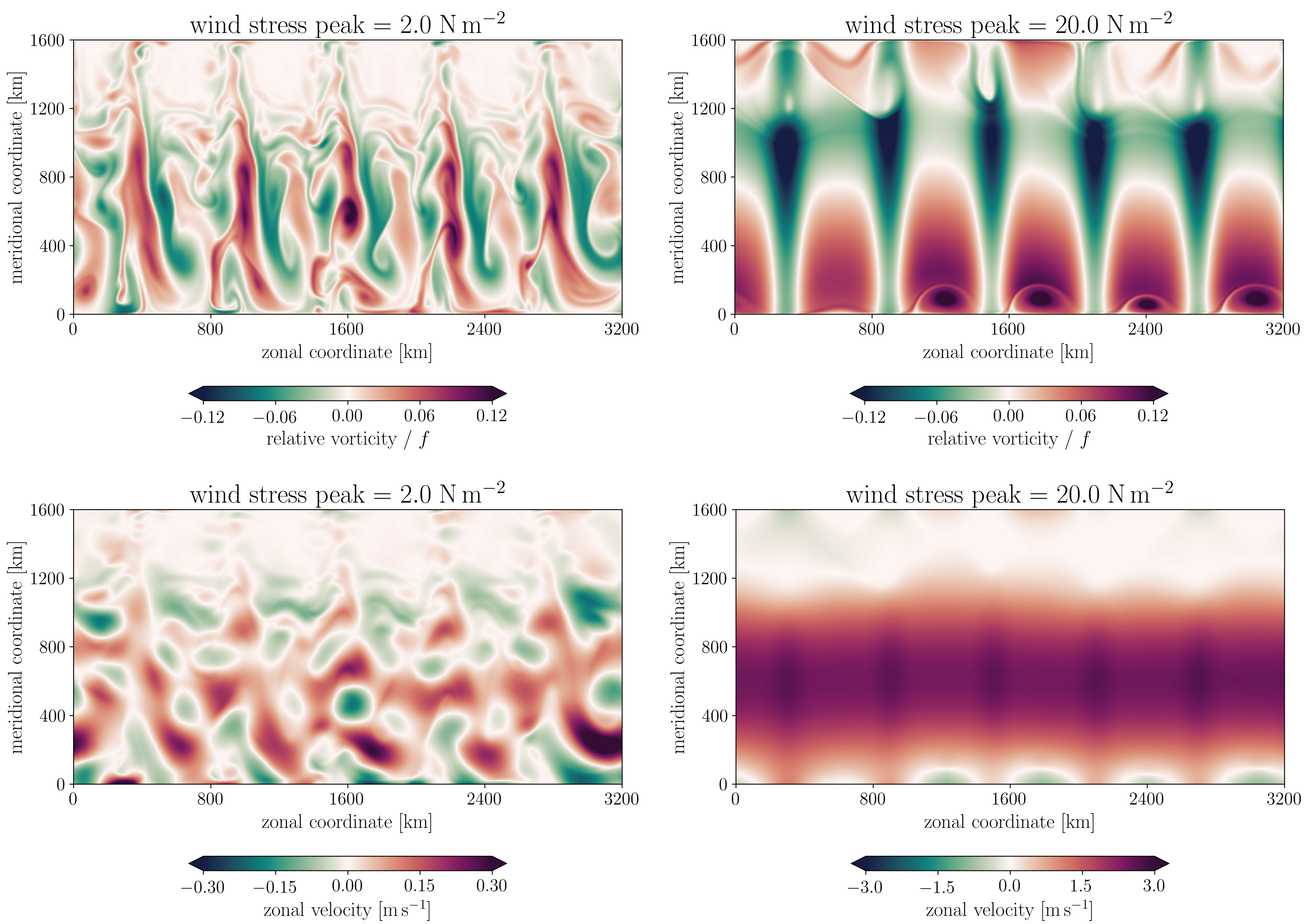


flow structure for 1-layer configuration

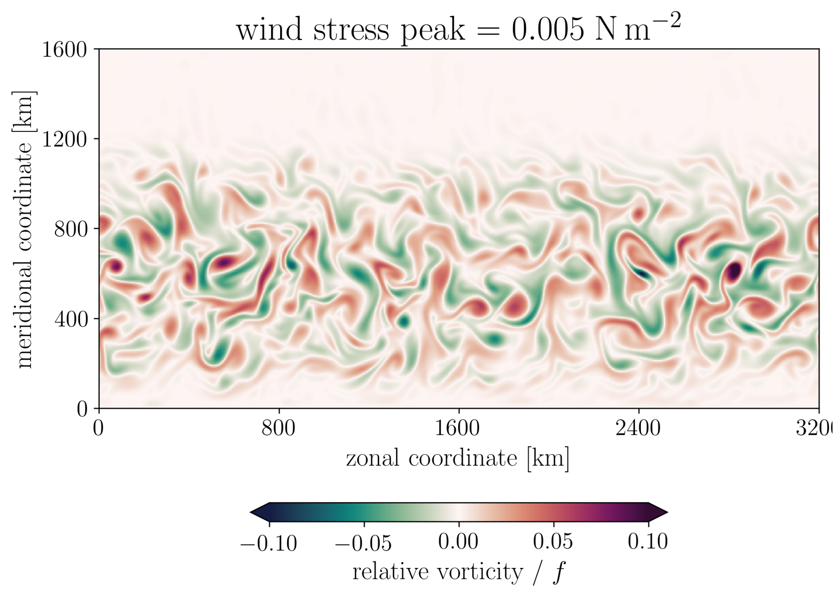
upper branch
(flow barely sees the topography)

relative vorticity

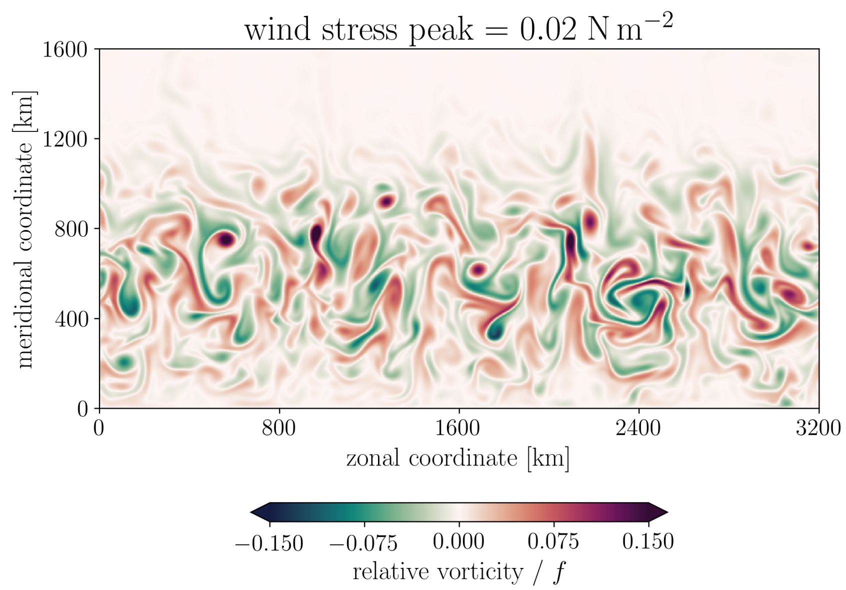
zonal flow



flow structure for 2-layer configuration



baroclinic eddies
(~200yr spinup)



baroclinic eddies
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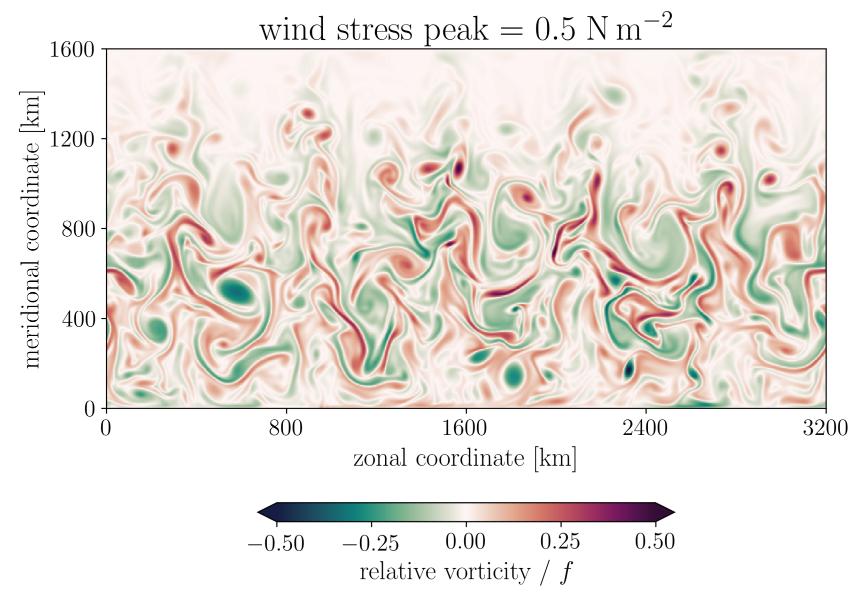
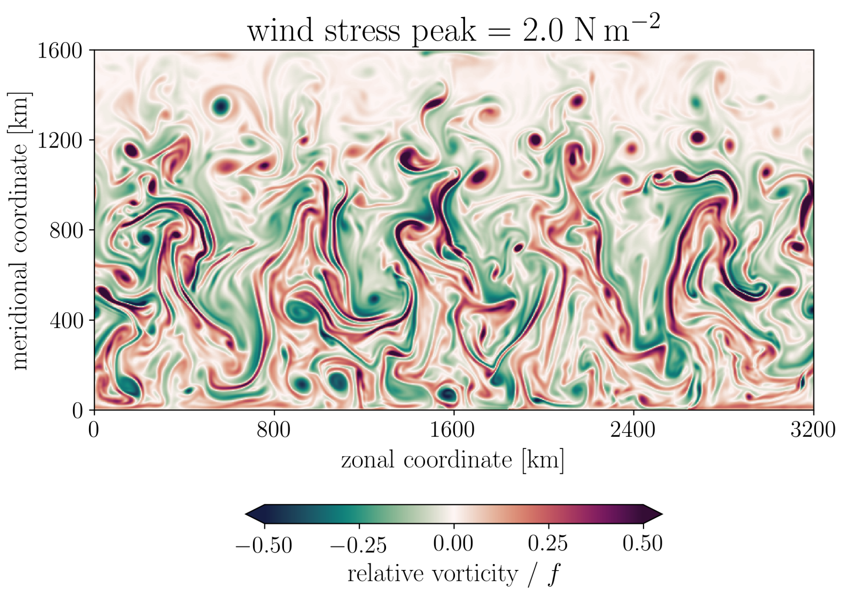
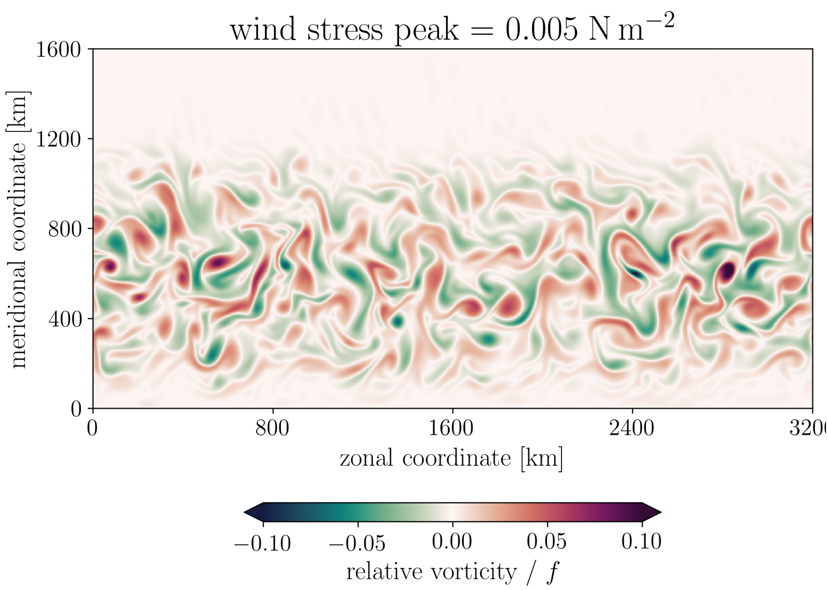


flow starts "seeing"
the topography



flow structure for 2-layer configuration



baroclinic eddies
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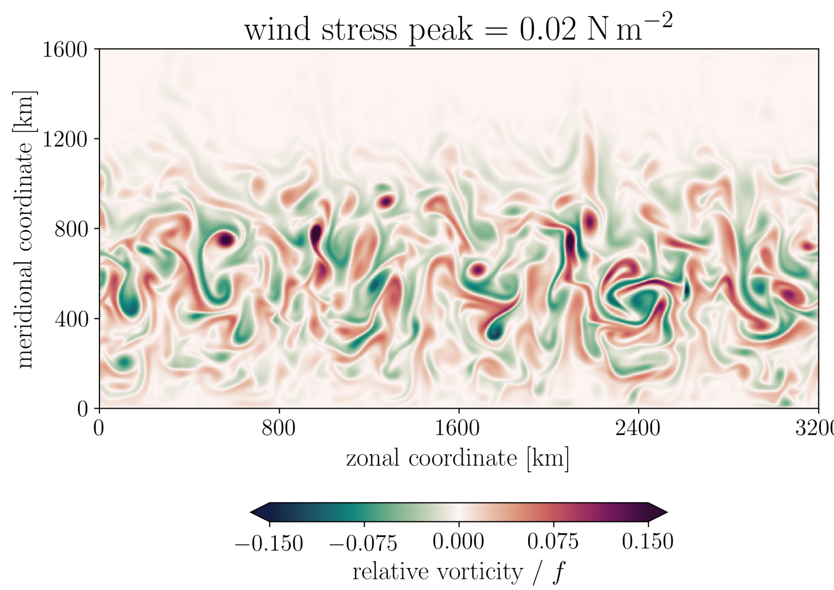
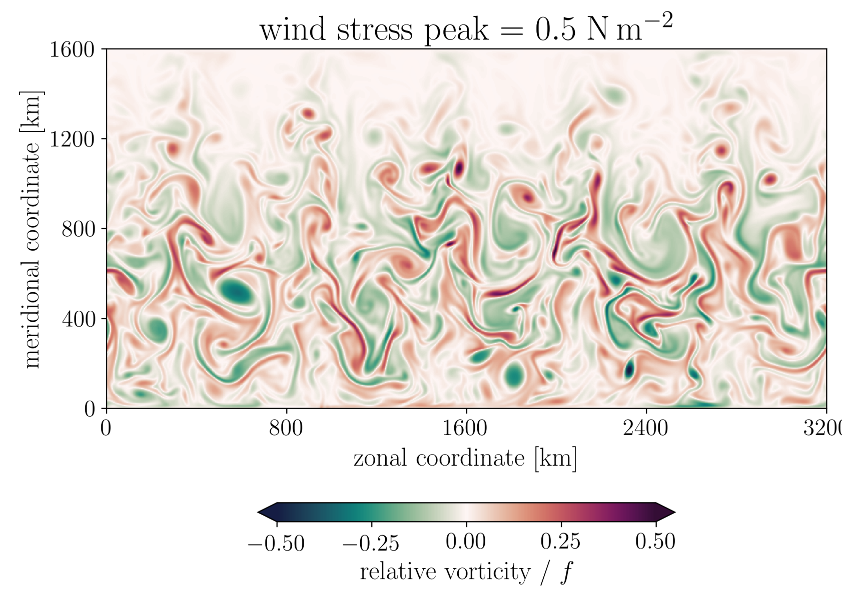
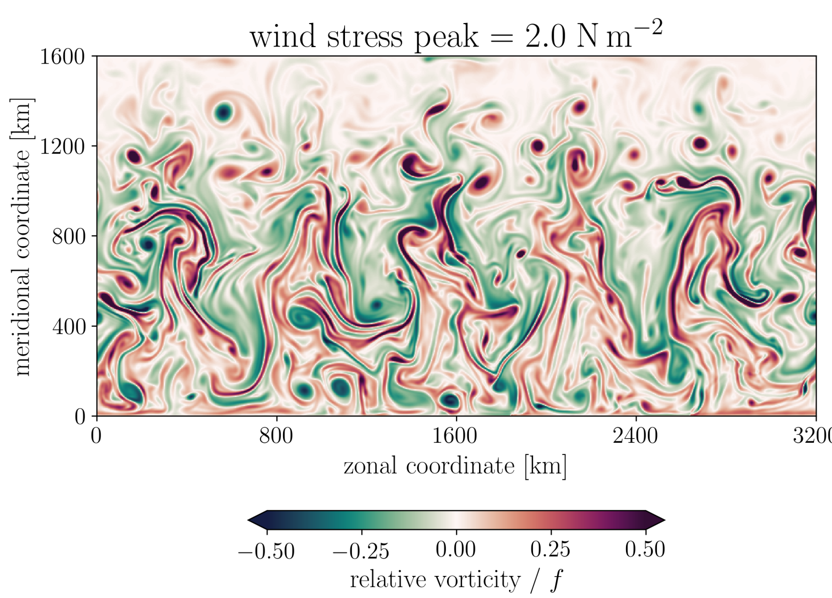


baroclinic eddies
(~50yr spinup)



flow starts "seeing"
the topography

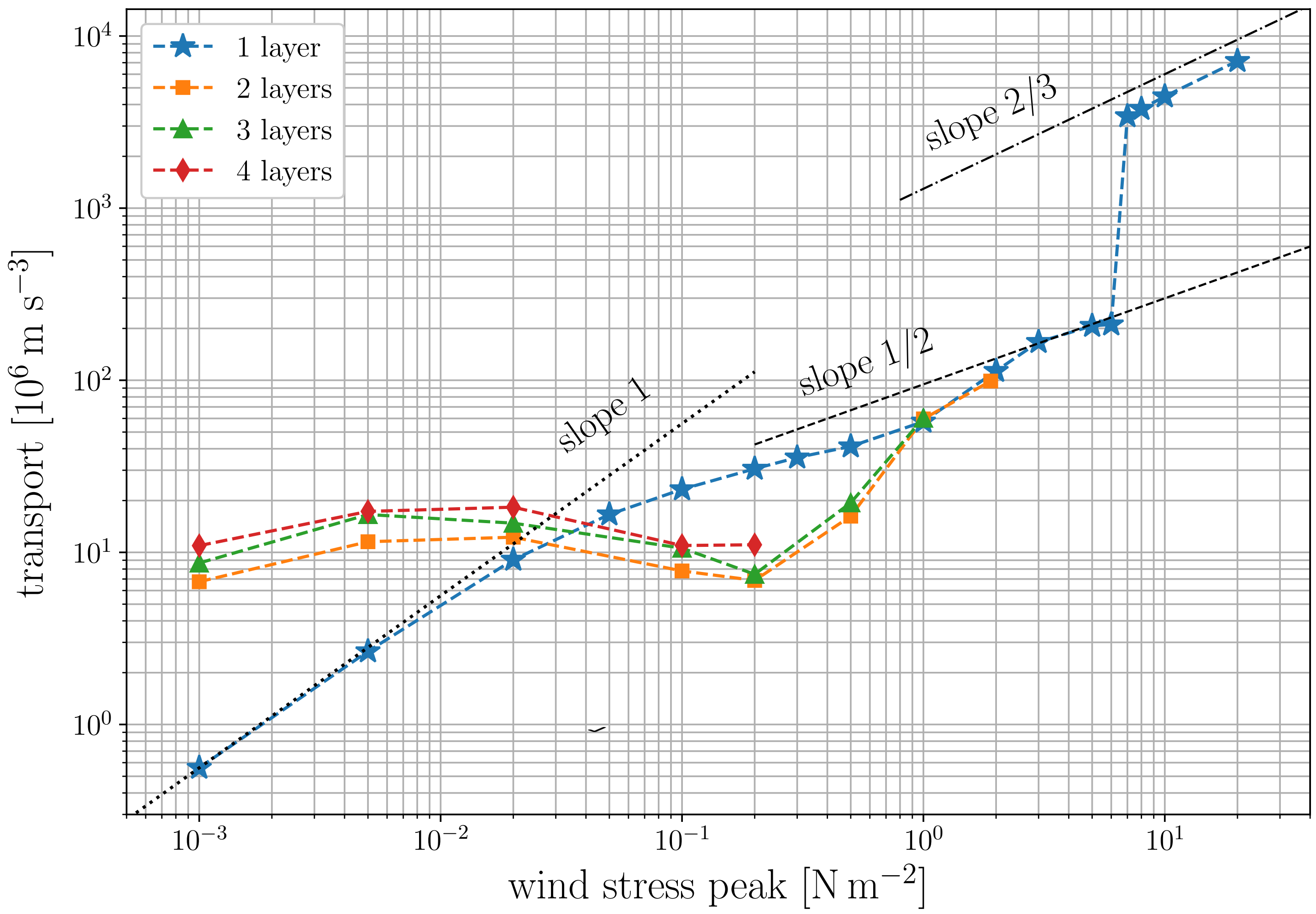


For higher wind stress values
code blows up
(blow-up related with outcropping)

Help! Anybody knows MOM6 here? :)

how does the transport vary with wind stress in this primitive-equations model?

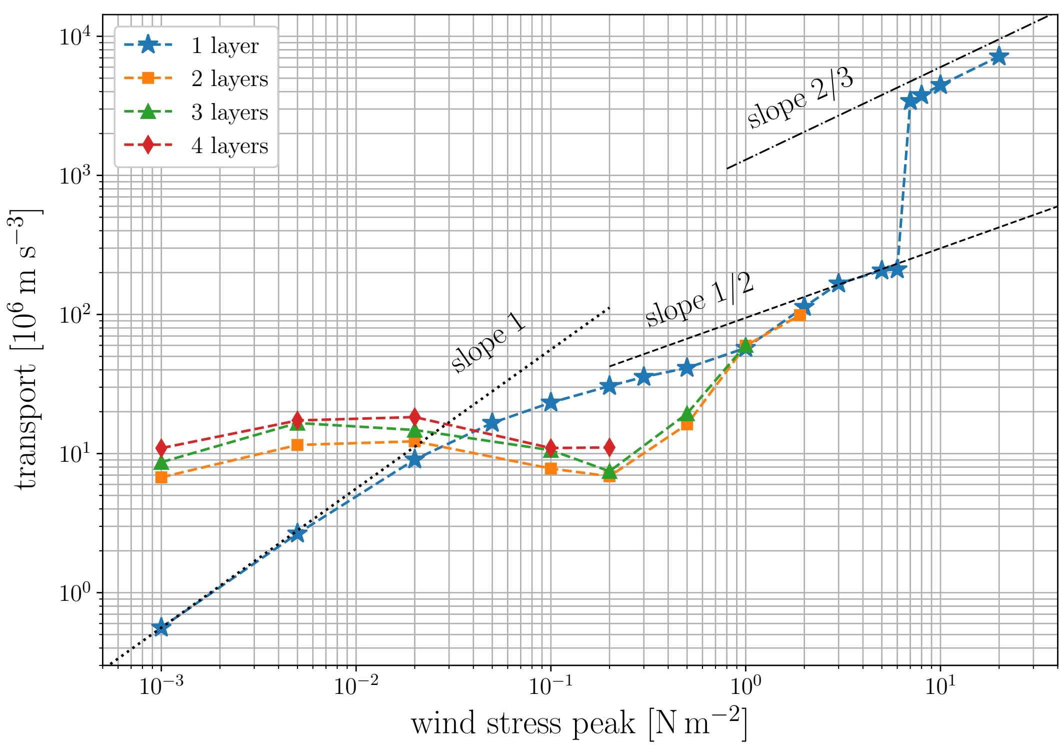
transport Vs wind stress

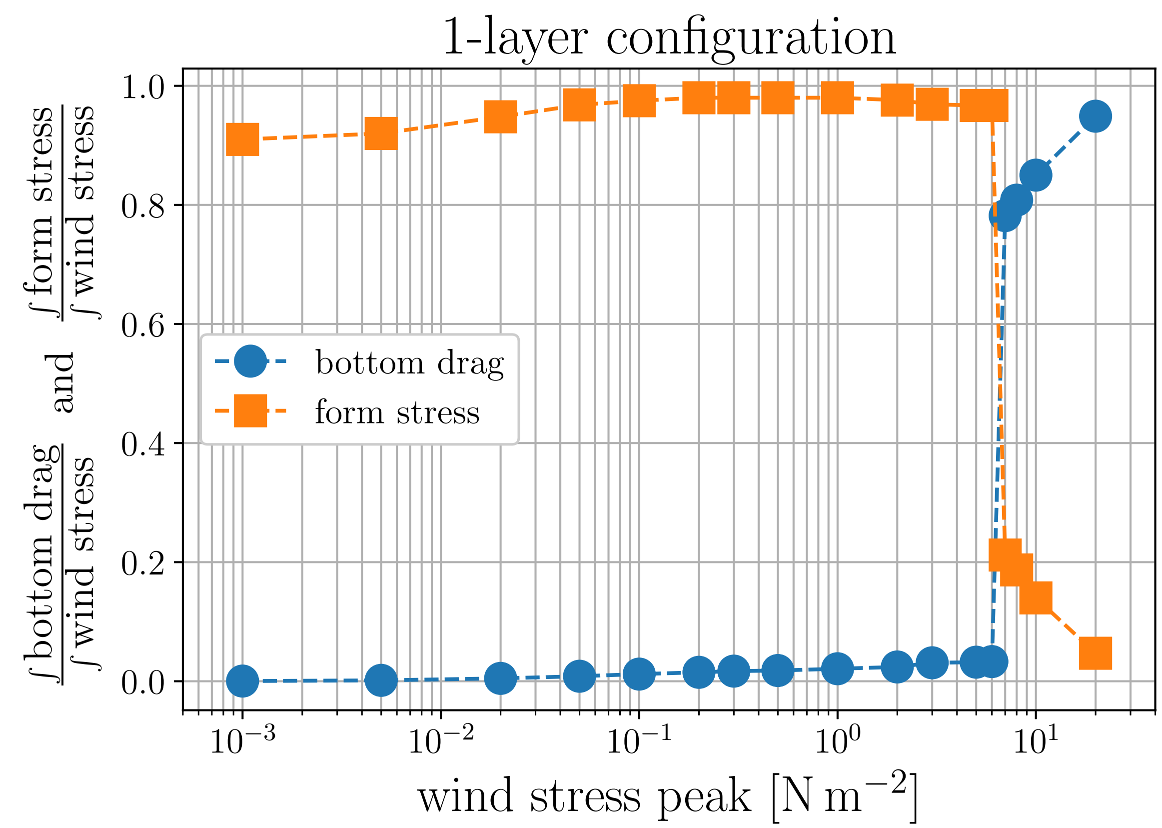


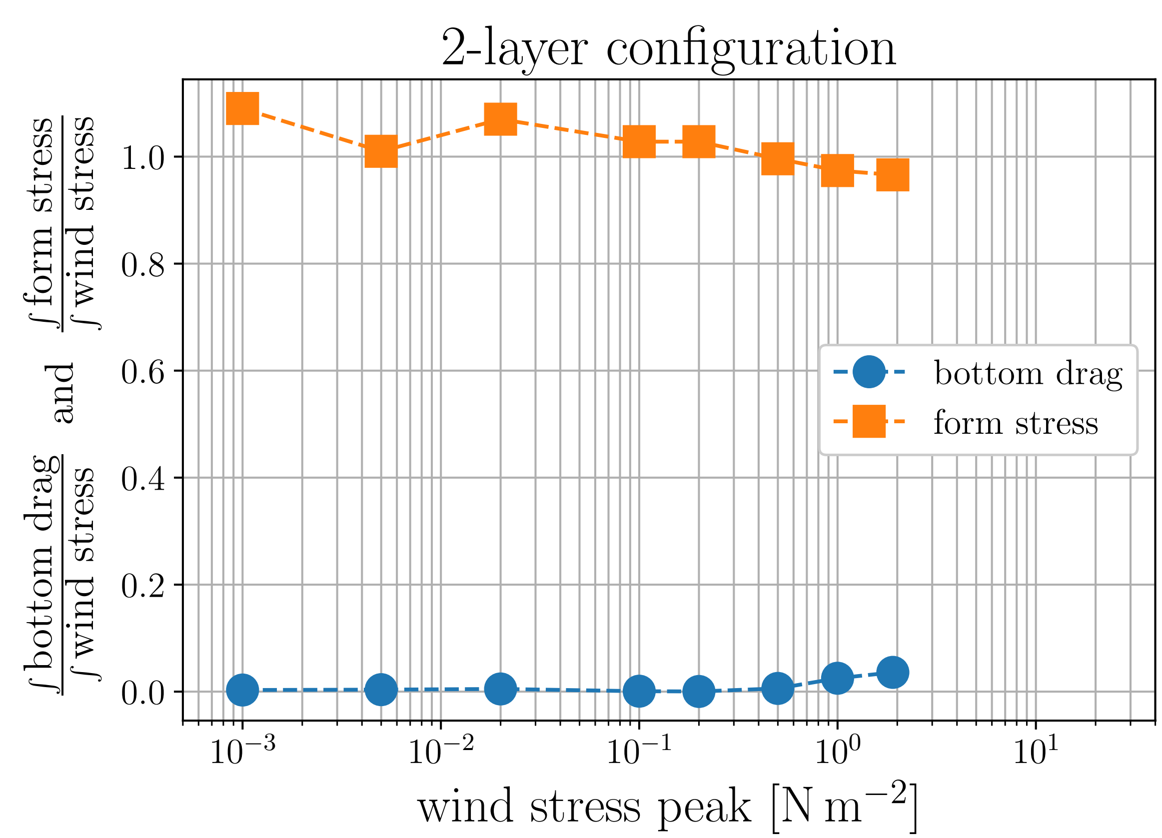
Baroclinic cases show strong eddy saturation.

The single-layer case **also** shows insensitivity to wind stress
(transport grows only about 10-fold over 100-fold wind stress increase)

how is the momentum balanced?







conclusions

This barotropic QG model shows eddy saturation.
This is surprising! All previous arguments were based on baroclinicity.

The barotropic—topographic instability is able to produce transient eddies in this model in a similar manner as baroclinic instability.

Barotropic eddy saturation "survives" in a primitive-equations multilayer channel model.

Is there a similar flow-transition bifurcation in baroclinic dynamics as in barotropic dynamics?
(can you help me with MOM6 blowups?)

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Is there a similar flow-transition bifurcation in baroclinic dynamics as in barotropic dynamics?
(can you help me with MOM6 blowups?)

This changes the way we view eddy saturation and highlights the role of topographically-induced eddies.

thank you

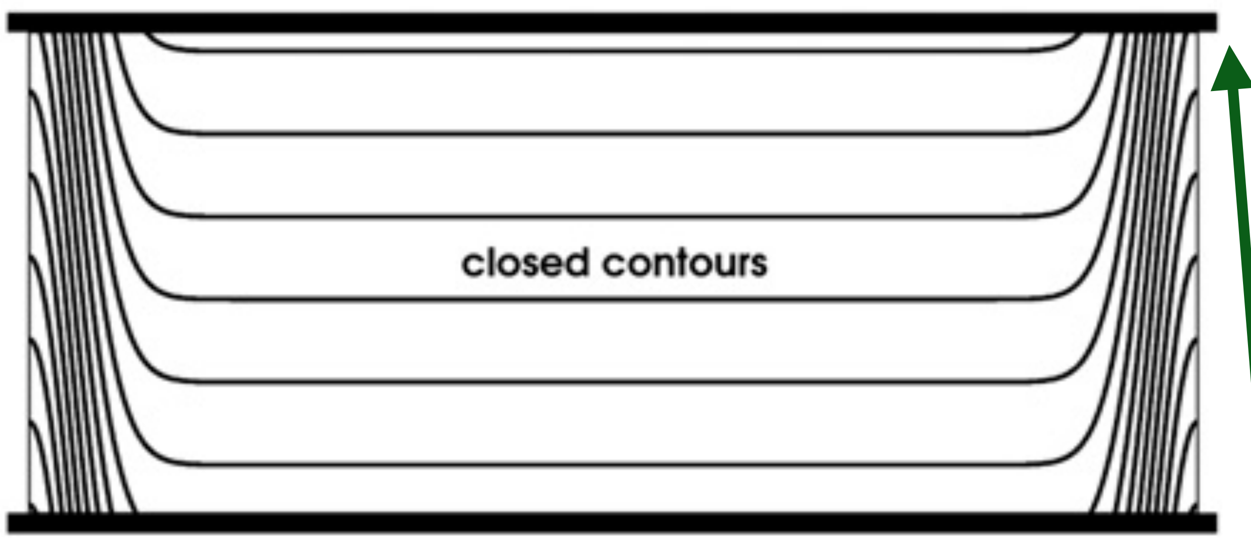
Constantinou and Young (2017). Beta-plane turbulence above monoscale topography. *J. Fluid Mech.,* **827**, 415-447.
Constantinou (2018). A barotropic model of eddy saturation. *J. Phys. Oceanogr.* **48 (2)**, 397-411.

extra slides

characterizing geostrophic contours

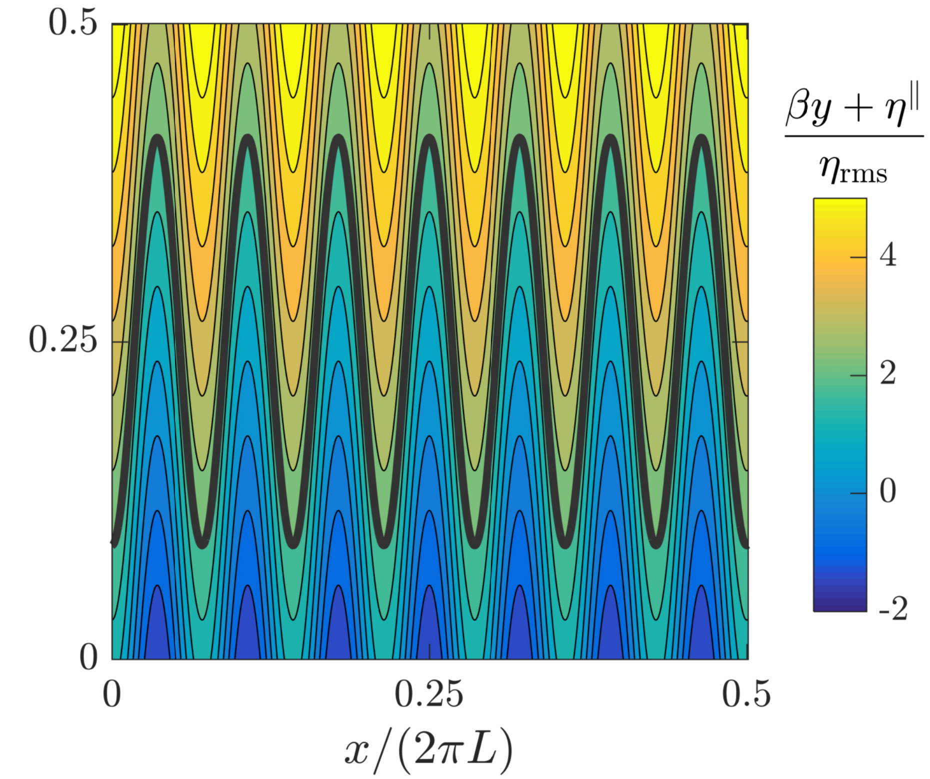
$$\beta y + \eta(x, y)$$

closed/blocked geostrophic contours



Nadeau & Ferrari 2015

open geostrophic contours



Constantinou 2017
Constantinou & Young 2017

without channel walls
both geostrophic contours look alike

decomposing the ACC transport

the time-mean zonal flow:

$$\bar{u}(x,y,z) = \underbrace{\bar{u}(x,y,z) - \bar{u}_{\mathrm{bot}}(x,y)}_{\stackrel{\mathrm{def}}{=}\bar{u}_{\mathrm{tw}}(x,y,z)} + \bar{u}_{\mathrm{bot}}(x,y)$$

"thermal wind" flow

bottom flow

$$\partial_z \bar{u} = -\partial_y \bar{b}$$

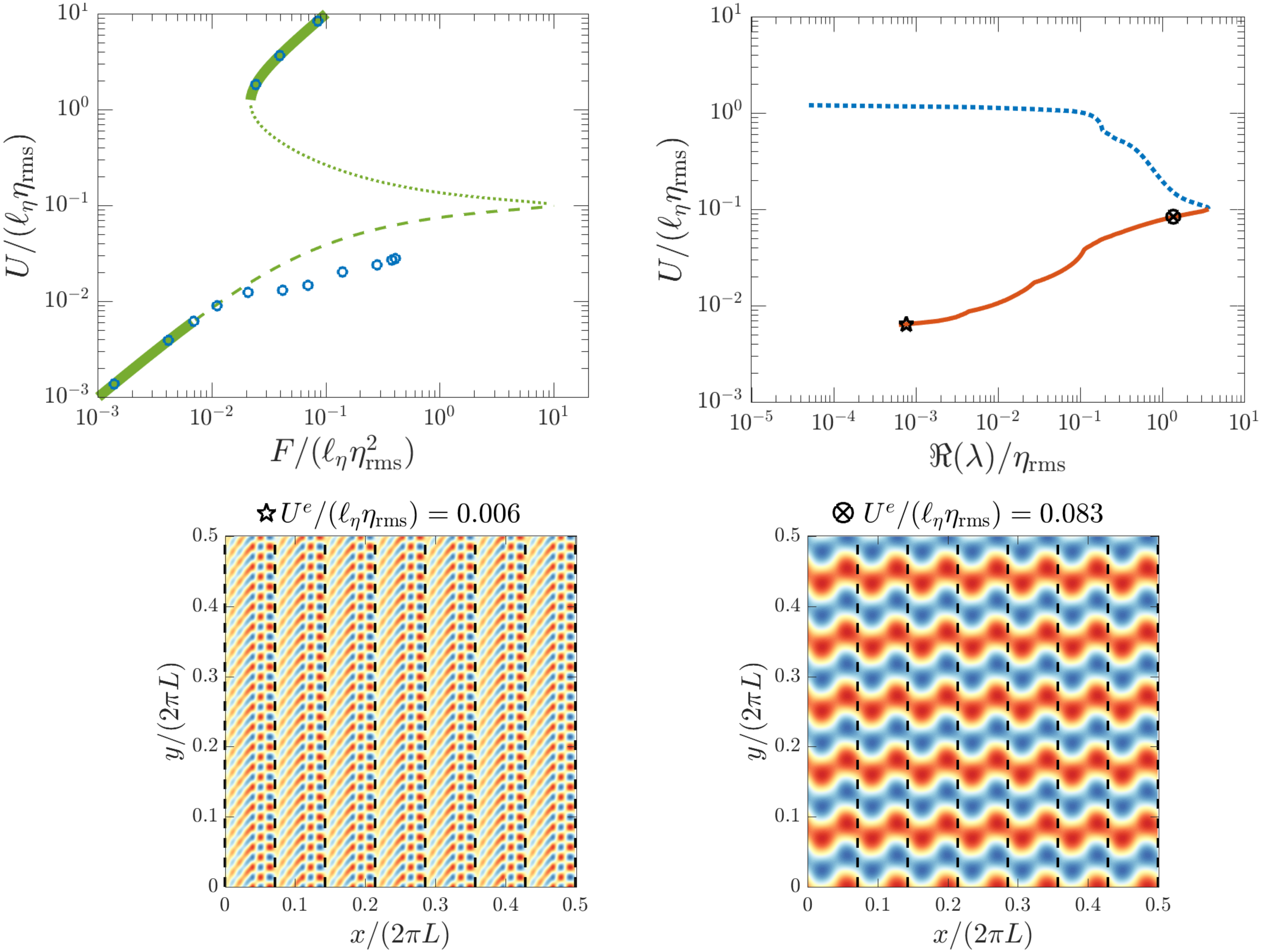
$$\underbrace{\int_{-H}^{0} \mathrm{d}z \int \mathrm{d}y \int \frac{\mathrm{d}x}{L_x}\,\bar{u}}_{\stackrel{\mathrm{def}}{=}T_{\mathrm{ACC}}} = \underbrace{\int \mathrm{d}y \int \frac{\mathrm{d}x}{L_x}\,\bar{u}_{\mathrm{bot}}}_{\stackrel{\mathrm{def}}{=}T_{\mathrm{bot}}} + \underbrace{\int_{-H}^{0} \mathrm{d}z \int \mathrm{d}y \int \frac{\mathrm{d}x}{L_x}\,\bar{u}_{\mathrm{tw}}}_{\stackrel{\mathrm{def}}{=}T_{\mathrm{tw}}}$$

total transport

bottom

"thermal wind"

not included in the barotropic QG model



$U/(\ell_\eta \eta_{\rm rms})$
$F/(\ell_\eta \eta_{\rm rms}^2)$
$\Re(\lambda)/\eta_{\rm rms}$
☆ $U^e/(\ell_\eta \eta_{\rm rms}) = 0.006$
⊗ $U^e/(\ell_\eta \eta_{\rm rms}) = 0.083$
$y/(2\pi L)$
$x/(2\pi L)$