Plumas beta lineales forzadas por el viento: aplicaciones a la Contracorriente Hawaiana

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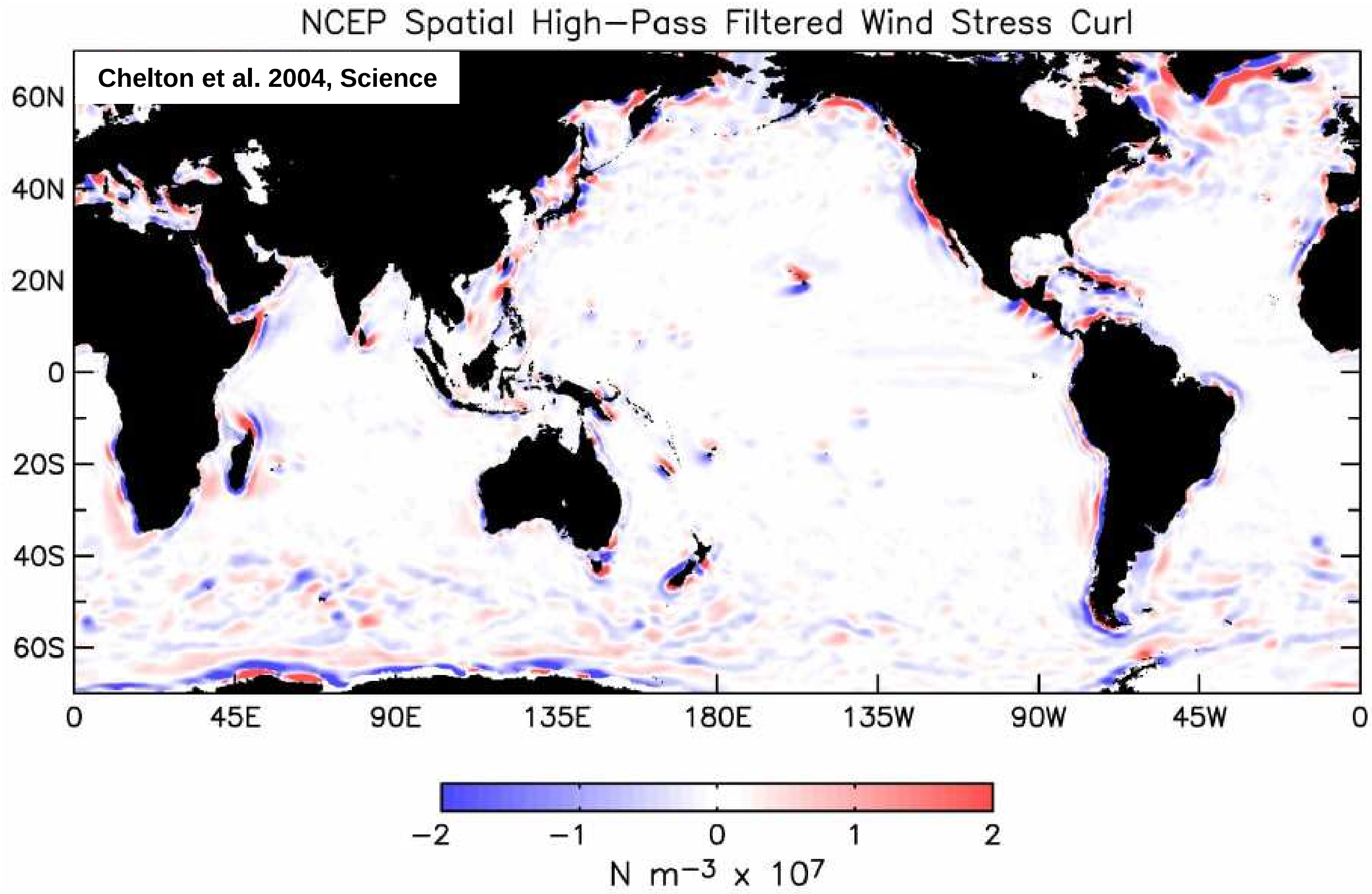
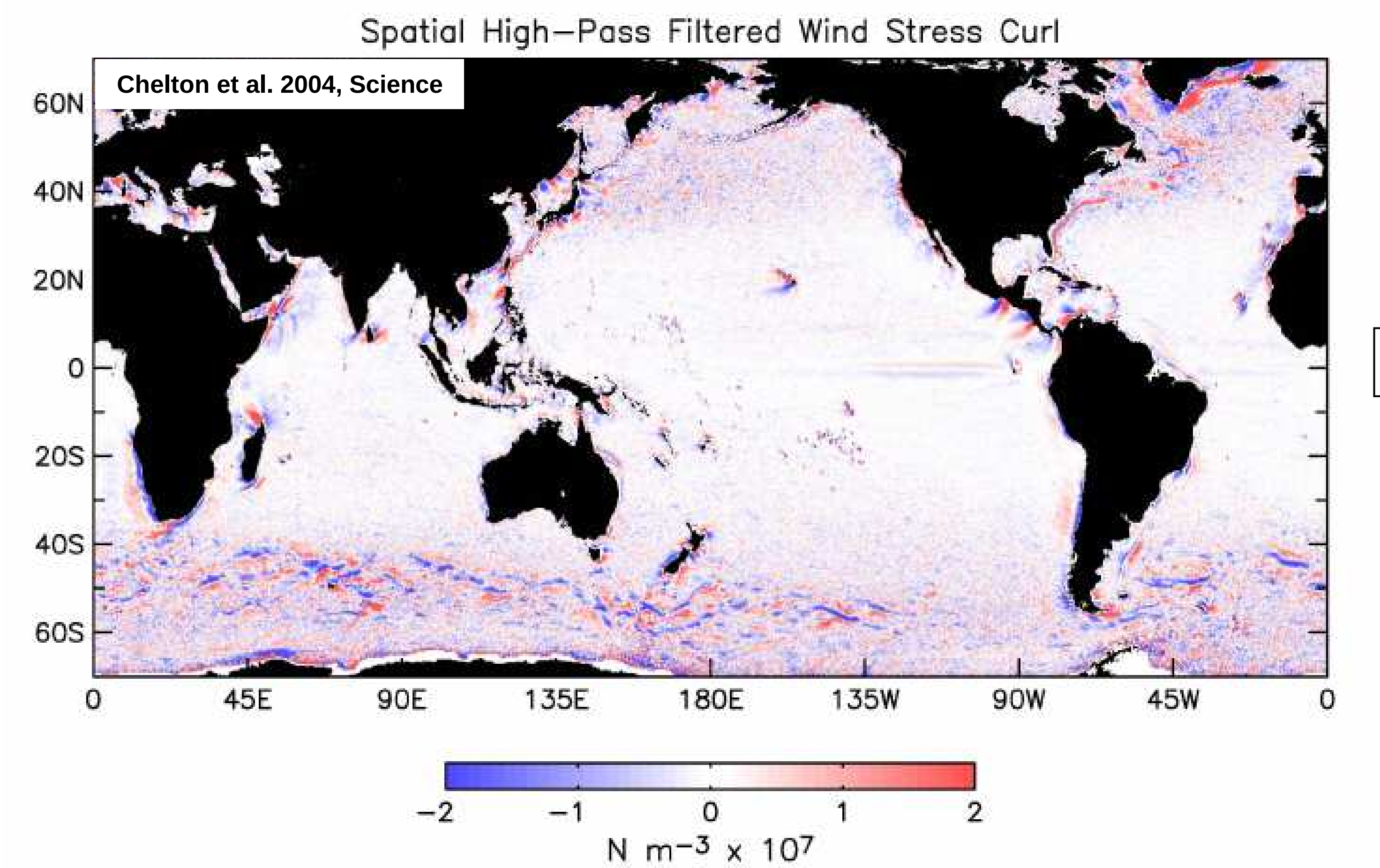
Small-scale wind stress curl (WSC)



4-year average

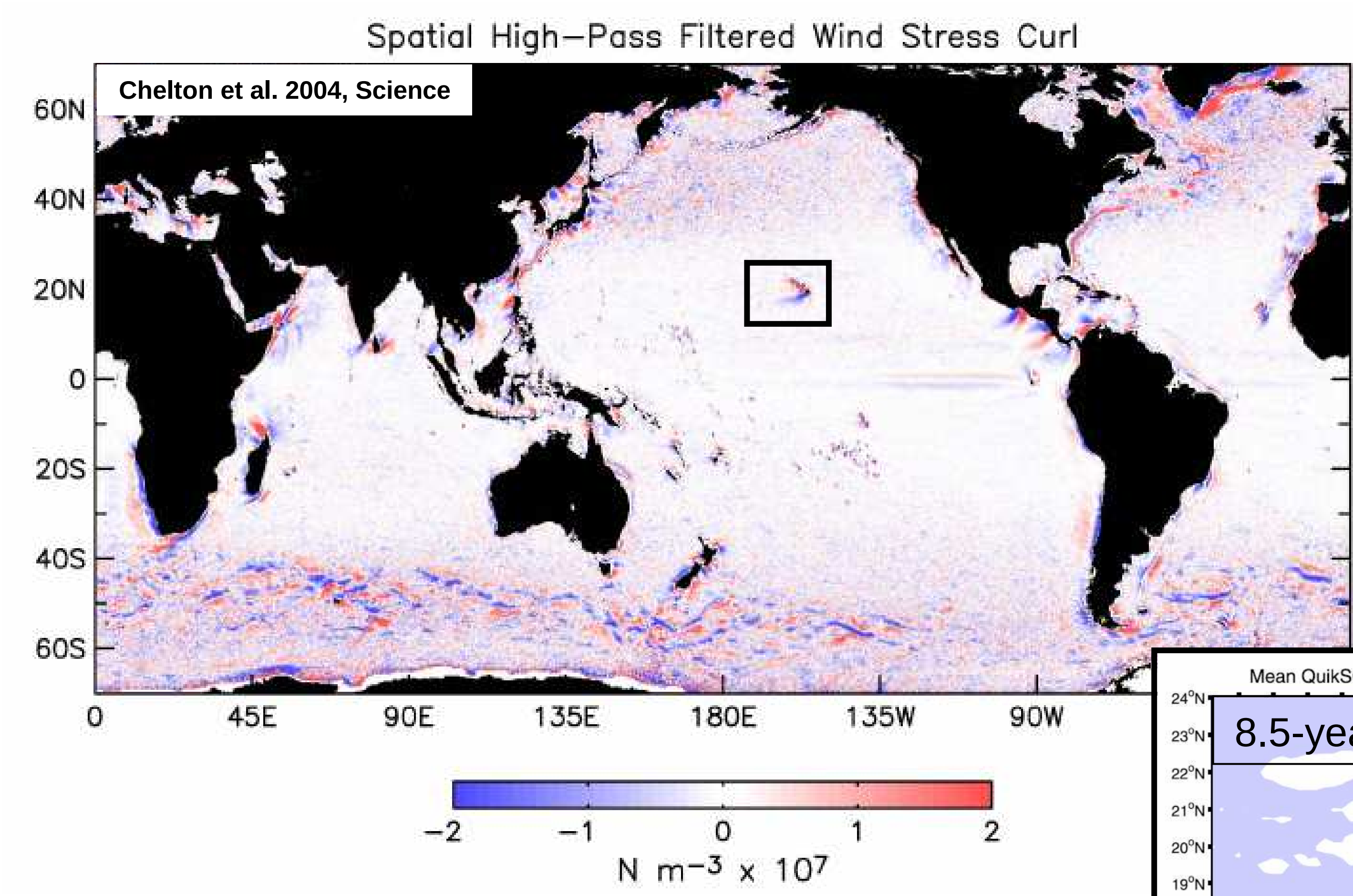
Small-scale wind stress curl (WSC)



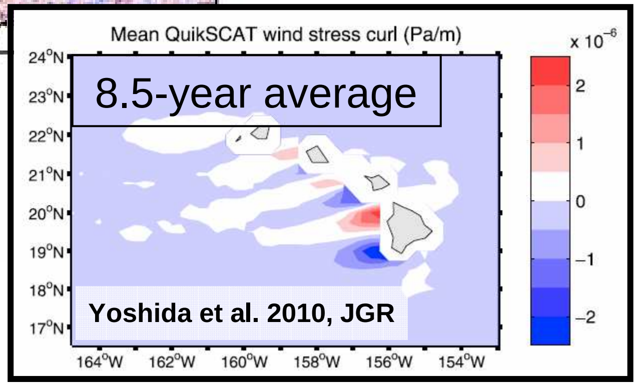
4-year average

- **Persistent small-scale WSC**: fronts, currents, orography (e.g., tall islands).
- **Patterns**: dipoles, monopoles, bands, etc.
- **Ocean dynamical response?**

Ocean response: example of the Hawaiian Lee Countercurrent (HLCC)



4-year average

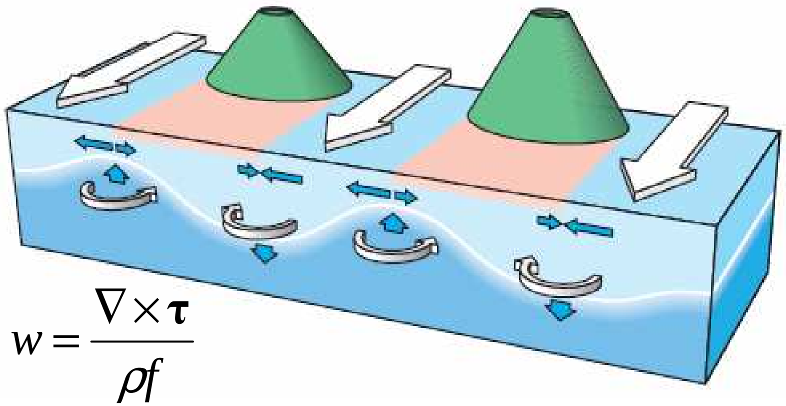


- Trade-wind blocking by Hawaii mountains → WSC **dipoles**.
- Dipole in Big Island lee drives **HLCC** via **Sverdrup** dynamics (e.g., Xie et al. 2001, Science).

HLCC: curl-driven zonal jet

Ekman flow: w

Chavanne et al. 2002, CJRS



$$w = \frac{\nabla \times \boldsymbol{\tau}}{\rho f}$$

Ekman pumping

- **WSC** drives **Ekman pumping** / suction.
- **Thermocline** is lifted / depressed.
- Cyclonic / anticyclonic **eddies** (e.g., Calil et al. 2008, DSR).
- **Rossby waves** propagate anomalies (e.g., Sasaki et al. 2010, OD).

Sverdrup flow: V

$$\frac{\partial \xi}{\partial t} + \beta v = f\,\frac{\partial w}{\partial z} + \frac{1}{\rho}\frac{\partial \nabla \times \boldsymbol{\tau}}{\partial z}$$

Baroclinic Vorticity Equation

- Vorticity produced by **vortex stretching** + **planetary vorticity** + **turbulent stress**.
- **Barotropic vorticity equation** (flat bottom): vortex stretching vanishes.

$$V = \frac{\nabla \times \boldsymbol{\tau}}{\rho \beta}$$

Sverdrup Balance

- **Meridional transport** driven by **WSC** (at steady state).

Volume conservation: U

$$\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} = 0$$

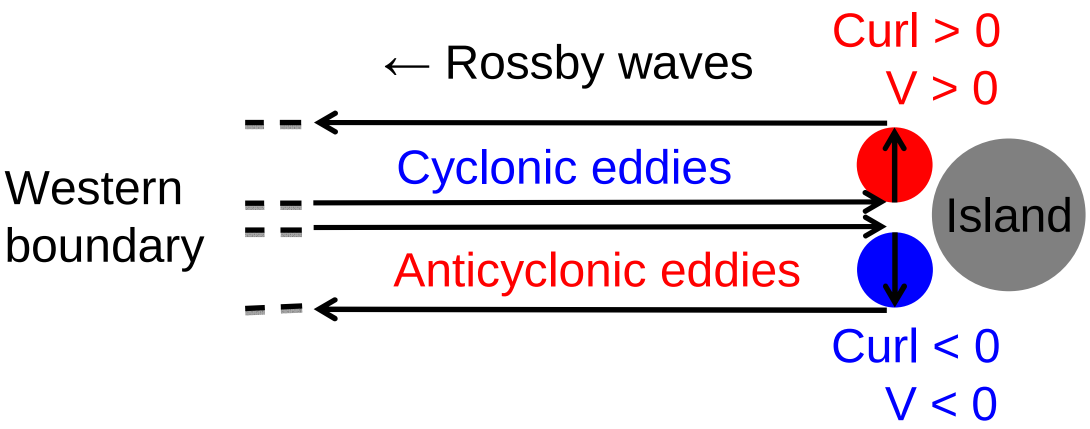
Barotropic Continuity Equation

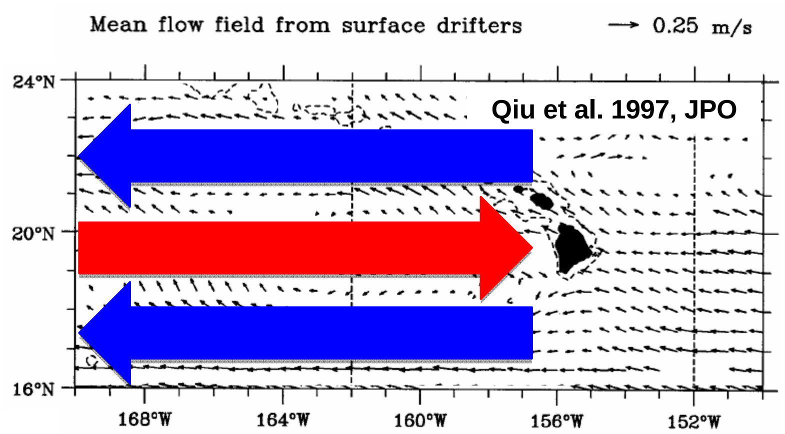
- **Nondivergent barotropic flow**.
- **Zonal transport** to the **west**.

$$U = -\int_{x_e}^{x} \frac{\partial V}{\partial y}\,dx$$

Integration from eastern boundary

HLCC: wind-forced β-plume



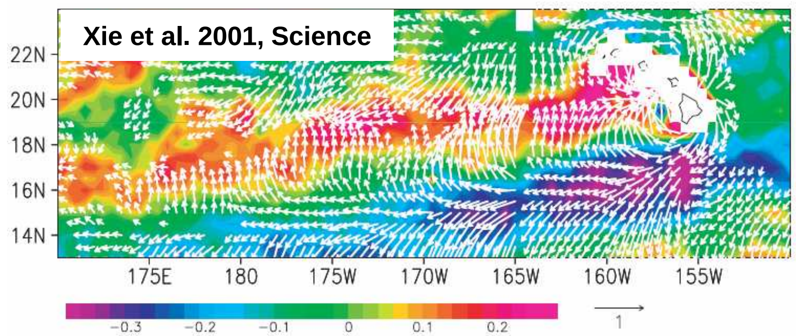


- **Elongated double-gyre** west of Hawaii.
- **HLCC**: **narrow eastward jet** embedded in broad North Equatorial Current (NEC).

- ***β*-plume** (Rhines 1994, Chaos) = **Sverdrup gyre** driven by **compact vorticity source** (momentum, heat, mass).

- **HLCC = wind-forced *β*-plume** (Jia et al. 2011, JGR).

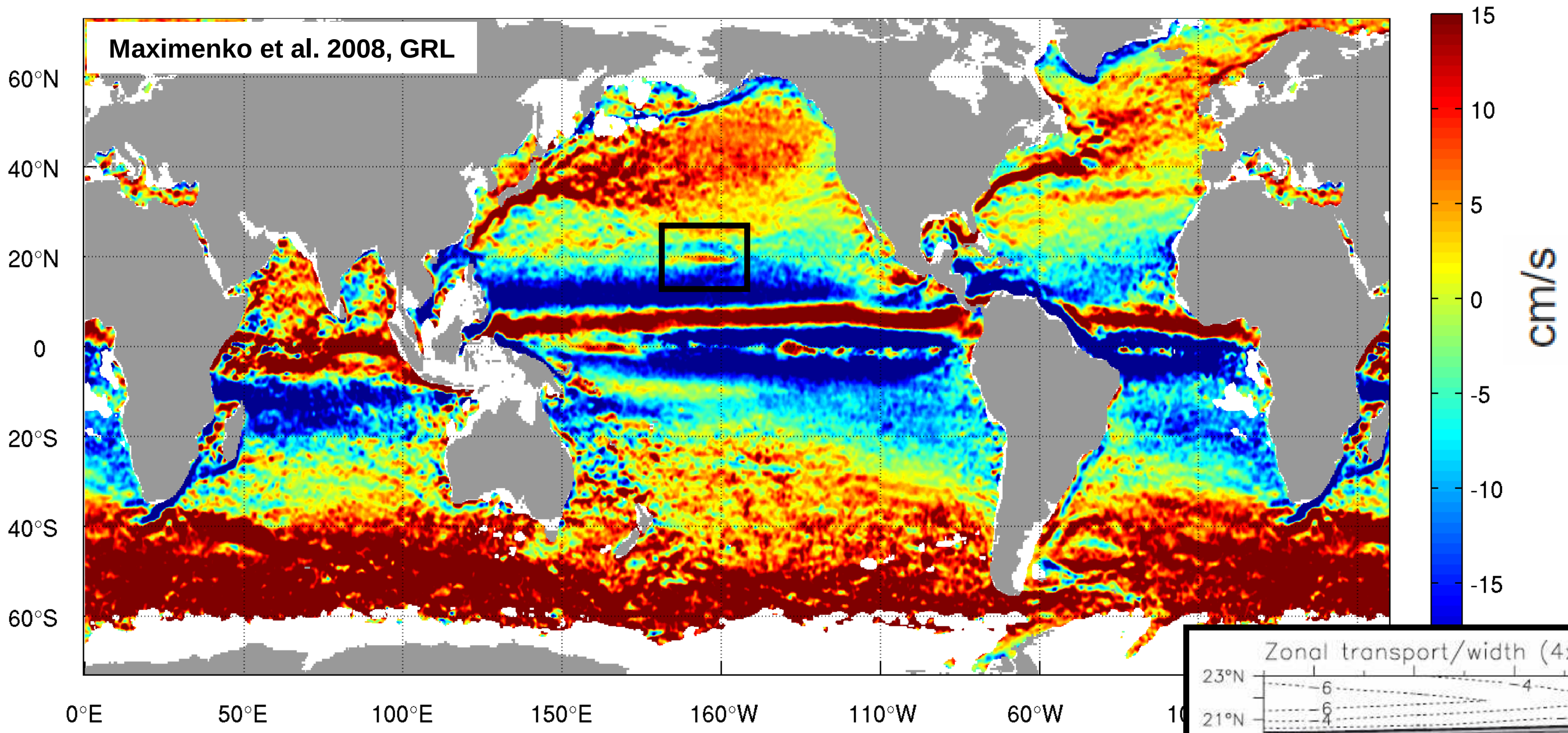
- **Other mechanisms**: **air-sea coupling**, island-induced modified large-scale flow (Qiu Durland 2002, JPO), mode water intrusions (Sasaki et al. 2012, JO), etc.



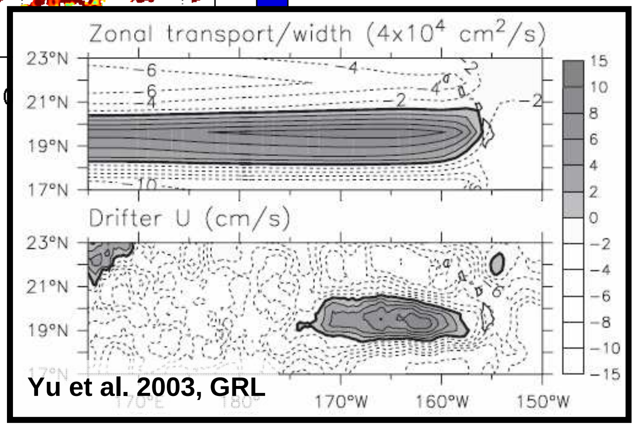
Spatial-filtered SST & wind

- HLCC advects warm SST → **far-field WSC dipole** (Xie et al. 2001, Science; Hafner Xie 2003, JAS; Sakamoto et al. 2004, GRL; Sasaki Nonaka 2006, GRL, etc.).

HLCC early termination



Mean surface zonal currents



- Zonal transport should reach western boundary, but **surface jet does not extend beyond 180ºE** (Chavanne et al. 2002, CJRS; Yu et al. 2003, GRL).
- **Horizontal dissipation by mesoscale eddies** → HLCC early termination (Yu et al. 2003).
- Underlying assumption: **equivalent-barotropic** vertical structure.

$$u(z) = \frac{U}{d_{th}} e^{z/d_{th}}$$

Chavanne et al. 2002, CJRS

Baroclinic β-plumes and the HLCC

- What is the **baroclinic structure** of *β*-plumes induced by localized WSC? HLCC?

- What is the **sensitivity** of the ocean circulation to the **scale of the WSC forcing**? HLCC?

- **Are eddies necessary** for the early termination of *β*-plumes? HLCC?

Outline

1. **Linear baroclinic *β*-plumes**: idealized model and theory

2. **HLCC vertical structure**: model results and observations

Conclusion

Idealized wind-forced β-plume: model set-up

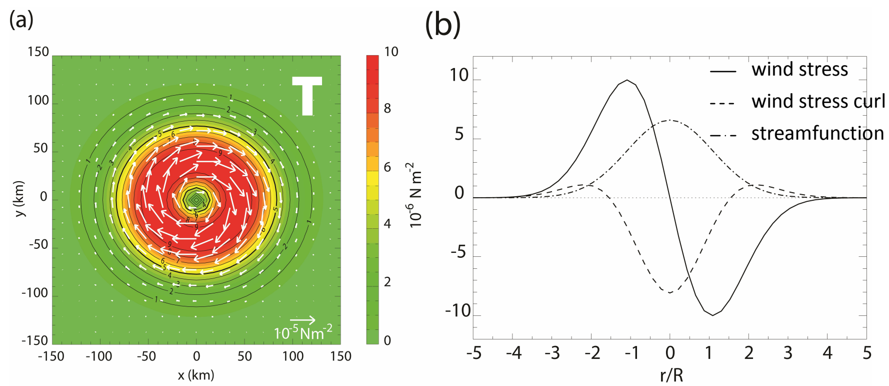
- **ROMS** (Shchepetkin McWilliams 2005, OM): free surface, hydrostatic model, resolves the primitive equations using stretched σ-coordinates.
- **Idealized subtropical ocean** (20ºN - 40ºN, 60º zonal) with **flat bottom** (H=4000m).
- **Eddy-resolving** (1/12º) with **32 vertical levels**.
- **Uniform initial stratification** $N^2(z)$ from World Ocean Atlas 2009 (typical of N Pacific gyre).
- **Constant vertical diffusion + viscosity**: $\kappa=10^{-5}m^2s^{-1}$, $\nu=10^{-4}m^2s^{-1}$.
- Simulation started from rest and run for **30 years** with 20 min time step (20 s for the barotropic mode). **Steady state after 20 years**. Outputs are saved every 5 days.
- Surface forcing: **localized steady wind vortex** in the center of the domain

  → **WSC quasi-monopole** (basic physics).
- **Linear regime**: **weak wind** $\tau_{max}=10^{-5}Nm^{-2}$.

Linear β-plume: steady-state barotropic solution

Wind: steady anticyclonic vortex   (*R* = 40 km)

$$\psi_a = R\tau_{\max}\sqrt{e}\,\exp\left(-\frac{x^2+y^2}{2R^2}\right), \quad \boldsymbol{\tau} = \mathbf{k}\times\nabla\psi_a$$



Meridional transport: Sverdrup balance

$$\overline{V} = \frac{\nabla\times\boldsymbol{\tau}}{\rho\beta}$$

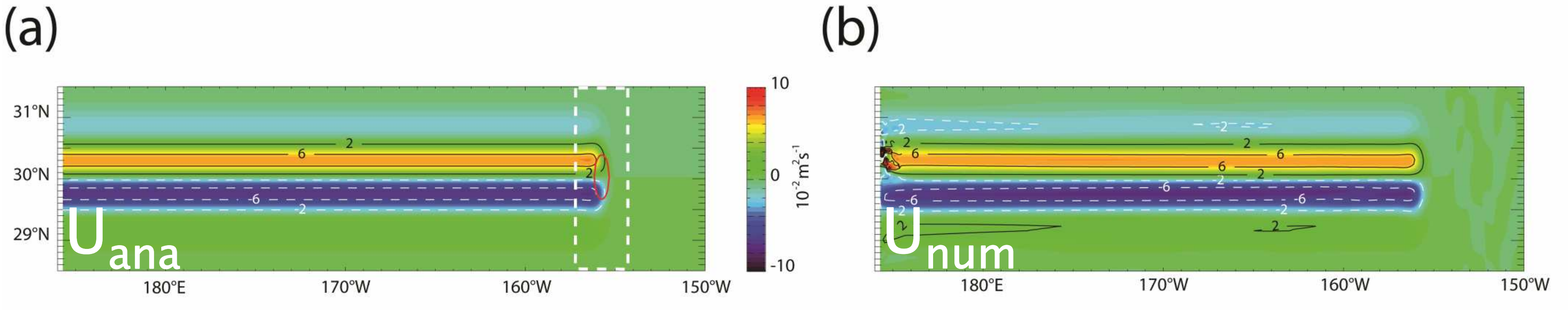
, ρ = 1025 kg.L^{-1}

$\beta \approx 1.98\ 10^{-11}\ s^{-1}m^{-1}$

Zonal transport: continuity equation

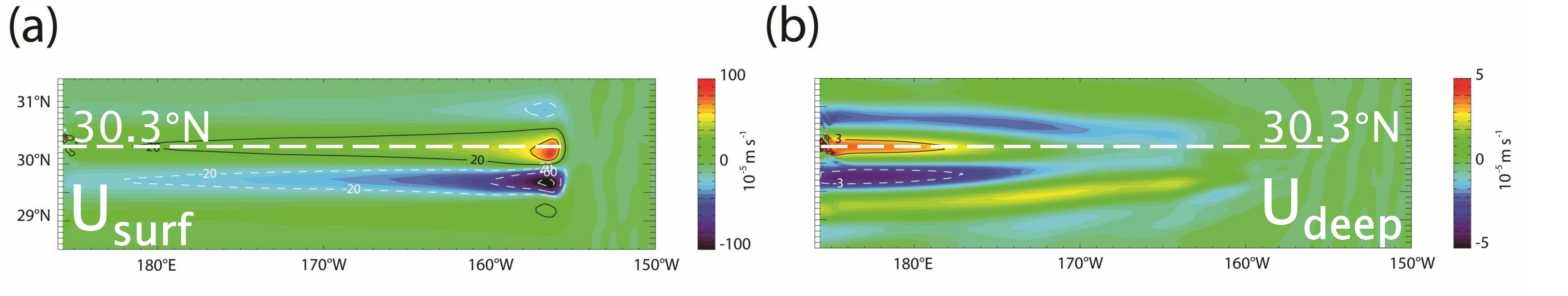
$$\overline{U} = -\int_{x_e}^{x}\frac{\partial\overline{V}}{\partial y}dx = \frac{\tau_{\max}\sqrt{e}}{\rho R^4\beta}\,y.e^{-\frac{y^2}{2R^2}}\left\{\sqrt{\frac{\pi}{2}}\left(3R^2-y^2\right).\left[erf\left(\frac{x_e}{\sqrt{2}R}\right)-erf\left(\frac{x}{\sqrt{2}R}\right)\right]-R\left[x.e^{-\frac{x^2}{2R^2}}-x_e.e^{-\frac{x_e^2}{2R^2}}\right]\right\}$$

, $x_e \approx 2890$ km



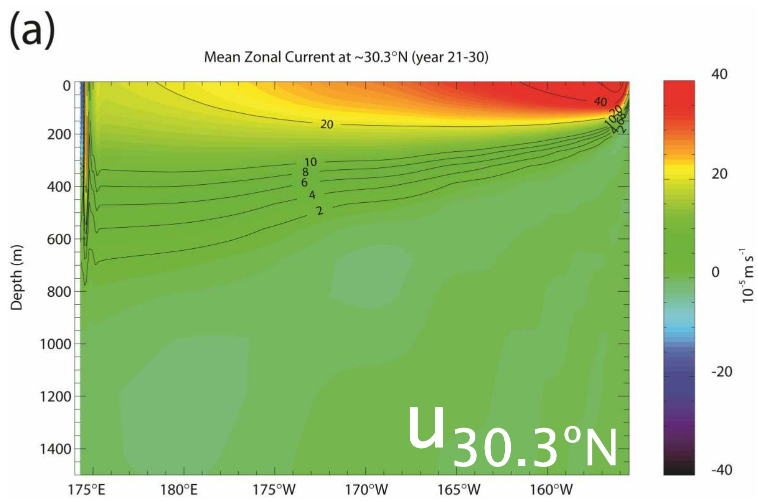
- **Good agreement** between analytical and numerical solutions.
- **1 anticyclonic cell** (2 jets) + **2 weak cyclonic cells** = 2+2 **x-independent zonal jets**.

Linear β-plume: Vertical structure



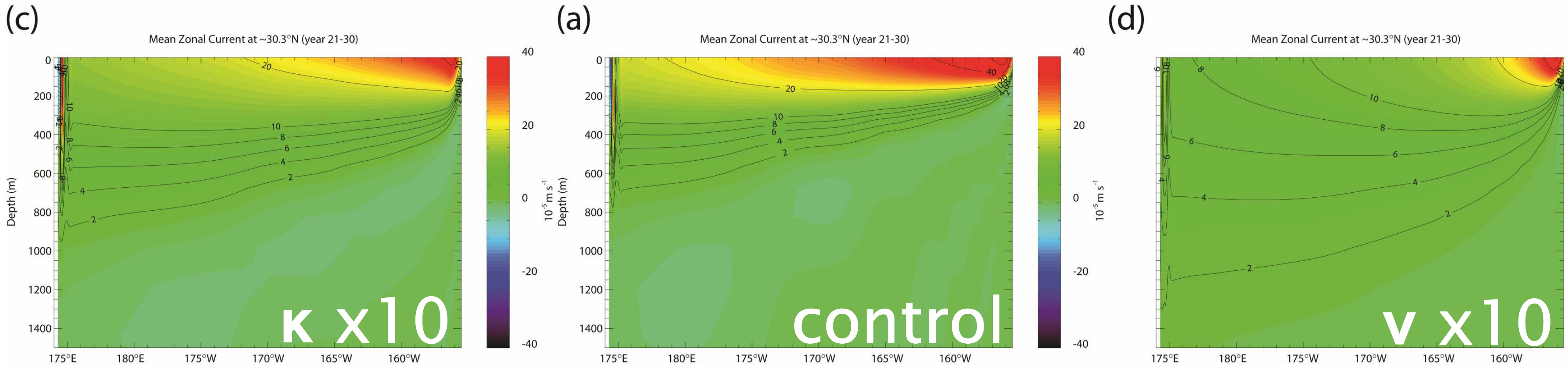
- In contrast with barotropic flow, **surface jets decay westward**.

- **Deepening of *β*-plume** lower boundary + emergence of **deep flow far away from forcing**.



- Zonal change in baroclinic structure: **damping of baroclinic Rossby waves?** (especially higher-order modes)

Linear β-plume: Vertical mixing effects



- ➢ **Zonal scales are smaller when vertical mixing is increased.**
- ➢ **Stronger sensitivity to viscosity compared to diffusion.**
- ➢ Usually vertical viscosity effects are very weak. Why are they dominant here?

Linear continuously-stratified (LCS) model

Linearized primitive equations (McCreary 1981, PTRA)

$$\frac{\partial u}{\partial t} - f\upsilon + \frac{1}{\rho_0}\frac{\partial p}{\partial x} = \frac{\partial}{\partial z}\left(\nu \frac{\partial u}{\partial z}\right) \qquad \frac{\partial u}{\partial x} + \frac{\partial \upsilon}{\partial y} + \frac{\partial w}{\partial z} = 0$$

$$\frac{\partial \upsilon}{\partial t} + fu + \frac{1}{\rho_0}\frac{\partial p}{\partial y} = \frac{\partial}{\partial z}\left(\nu \frac{\partial \upsilon}{\partial z}\right) \qquad \frac{\partial p}{\partial z} = -\rho g$$

$$\frac{\partial \rho}{\partial t} - \frac{\rho_0 N^2}{g} w = \frac{\partial^2}{\partial z^2}(\kappa\rho)$$

Vertical mode decomposition

$$u(x,y,z,t) = \sum_{n=1}^{+\infty} u_n(x,y,t)\psi_n(z) \qquad w(x,y,z,t) = \sum_{n=1}^{+\infty} w_n(x,y,t)\int_{-H}^{z}\psi_n(z)dz$$

$$\upsilon(x,y,z,t) = \sum_{n=1}^{+\infty} \upsilon_n(x,y,t)\psi_n(z) \qquad \rho(x,y,z,t) = \sum_{n=1}^{+\infty} \rho_n(x,y,t)\frac{\partial \psi_n(z)}{\partial z}$$

$$p(x,y,z,t) = \sum_{n=1}^{+\infty} p_n(x,y,t)\psi_n(z)$$

$$\frac{\partial}{\partial z}\left(\frac{1}{N^2(z)}\frac{\partial \psi_n(z)}{\partial z}\right) = -\frac{1}{c_n^2}\psi_n(z)$$

Linear continuously-stratified (LCS) model (2)

Steady-state **baroclinic mode primitive equations** (McCreary 1981, PTRA)

$$\frac{\nu N^2}{c_n^2} u_n - f\upsilon_n + \frac{1}{\rho_0}\frac{\partial p_n}{\partial x} = F_n \quad , \; F_n = \tau_x \Big/ \rho_0 \int_{-H}^{0} \psi_n^2 dz$$

$$\frac{\nu N^2}{c_n^2} \upsilon_n + f u_n + \frac{1}{\rho_0}\frac{\partial p_n}{\partial y} = G_n \quad , \; G_n = \tau_y \Big/ \rho_0 \int_{-H}^{0} \psi_n^2 dz$$

$$\frac{\kappa N^2}{\rho_0 c_n^4} p_n + \frac{\partial u_n}{\partial x} + \frac{\partial \upsilon_n}{\partial y} = 0$$

$$w_n = -\frac{\kappa N^2}{\rho_0 c_n^4} p_n$$

$$p_n = -\rho_n g$$

Steady-state baroclinic mode **quasi-geostrophic potential vorticity equation**

$$\underbrace{\beta \frac{\partial p_n}{\partial x}}_{\beta\text{ effect}} = \underbrace{\frac{\kappa N^2 f^2}{c_n^4} p_n}_{\text{Vortex stretching}} \underbrace{- \frac{\nu N^2}{c_n^2}\left(\frac{\partial^2 p_n}{\partial x^2} + \frac{\partial^2 p_n}{\partial y^2}\right)}_{\text{Curl (viscosity)}} + \underbrace{\rho_b f \left(\frac{\partial G_n}{\partial x} - \frac{\partial F_n}{\partial y}\right)}_{\text{WSC}}$$

Linear continuously-stratified (LCS) model (3)

- Ratio of *κ*-term [*Vortex stretching*] over *ν*-term [*Curl(viscosity)*]:

$$M_n = \frac{\kappa f^2}{c_n^2} p_n \Big/ \nu\left(\frac{\partial^2 p_n}{\partial x^2} + \frac{\partial^2 p_n}{\partial y^2}\right) \sim \frac{\kappa f^2}{c_n^2} \Big/ \nu\left(\frac{1}{L_x^2} + \frac{(2\pi)^2}{L_y^2}\right) \approx \frac{\kappa f^2 L_y^2}{4\pi^2 \nu c_n^2} = \frac{4R^2}{\pi^2 \sigma R_n^2}$$

with $\sigma = \frac{\nu}{\kappa}$ (Prandtl number), $R_n = \frac{c_n}{f} = \frac{c_1}{nf}$ (Rossby radius), and $L_y = 4R$ (y-wavelength)

- For **large-scale flow**, $R >> R_n \pi\sqrt{\sigma}/2$ and $M_n >> 1$:

$$\Rightarrow \quad \beta \frac{\partial p_n}{\partial x} = \underline{\frac{\kappa N^2 f^2}{c_n^4} p_n} + \rho_b f\left(\frac{\partial G_n}{\partial x} - \frac{\partial F_n}{\partial y}\right)$$

Vortex stretching

- For **small-scale WSC**, $R << R_n \pi\sqrt{\sigma}/2$ and $M_n << 1$:

$$\Rightarrow \quad \beta \frac{\partial p_n}{\partial x} = \underline{-\frac{\nu N^2}{c_n^2}\left(\frac{\partial^2 p_n}{\partial x^2} + \frac{\partial^2 p_n}{\partial y^2}\right)} + \rho_b f\left(\frac{\partial G_n}{\partial x} - \frac{\partial F_n}{\partial y}\right)$$

Curl (viscosity)

Baroclinic mode damping by viscosity/diffusion

Viscosity vs. Diffusion: dependence on mode number

$$\boxed{M_n = \frac{4R^2}{\pi^2 \sigma R_1^2} n^2 = M_1 n^2}$$

$\Longrightarrow$ Even when M_1<<1, for $n \geq n_0$ high enough M_n≥1

- For **small-scale WSC**, viscosity (diffusion) damps the lower- (higher-) order modes

  ↳ In ROMS, $M_1 \approx 0.04$ and $n_0 \approx 6$ **explain the stronger sensitivity to viscosity**

- For **large-scale flow**, diffusion damps all the baroclinic modes

Viscosity vs. Diffusion: decay length scales

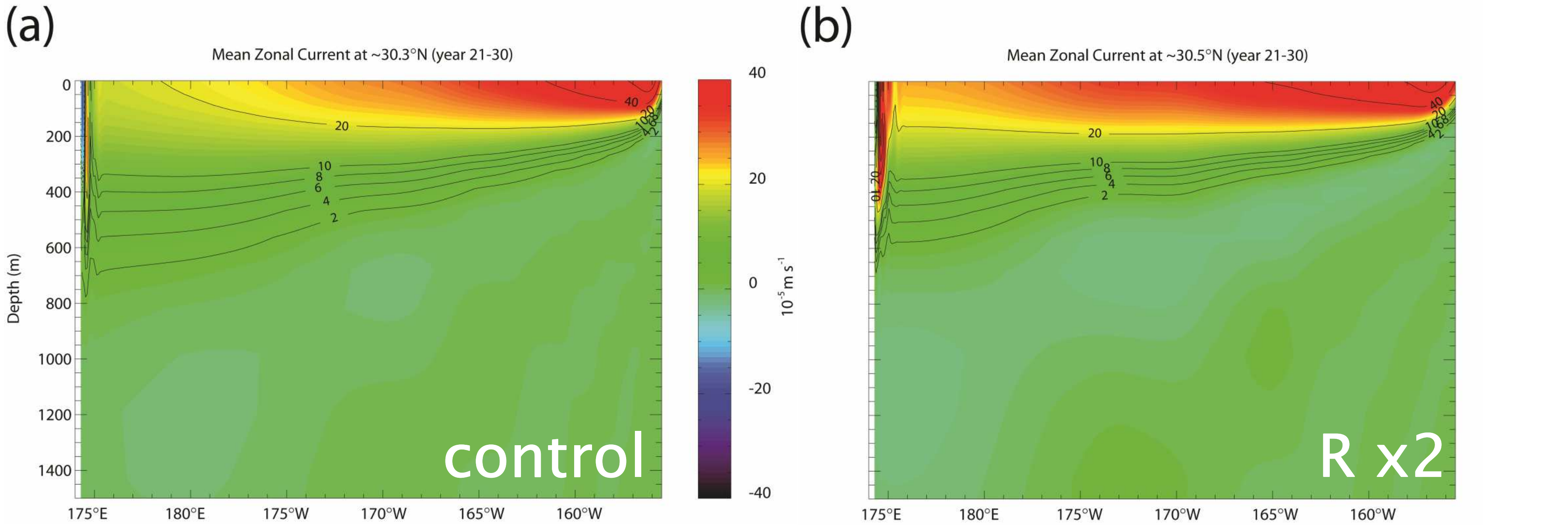
- Viscosity: $$\boxed{L_\nu = \frac{4\beta R^2 c_1^2}{\pi^2 \nu n^2 N^2}}$$

- Diffusion: $$\boxed{L_\kappa = \frac{\beta c_1^4}{\kappa f^2 n^4 N^2}}$$

- Smaller scales with enhanced mixing.
- Smaller scales for higher-order modes.
- **Only L_ν varies with R** because *Curl(viscosity)* acts on vorticity perturbation while *Vortex stretching* acts on pressure perturbation.

Sensitivity to forcing scale

Wind: R x 2 , $\mathbf{T}_{max}$ x 2



$$L_\nu = \frac{4\beta R^2 c_1^2}{\pi^2 \nu n^2 N^2}$$

- **Baroclinic x-scale increases with WSC y-scale** in agreement with theory.

Outline

1. **Linear baroclinic *β*-plumes**: idealized model and theory

2. **HLCC vertical structure**: model results and observations

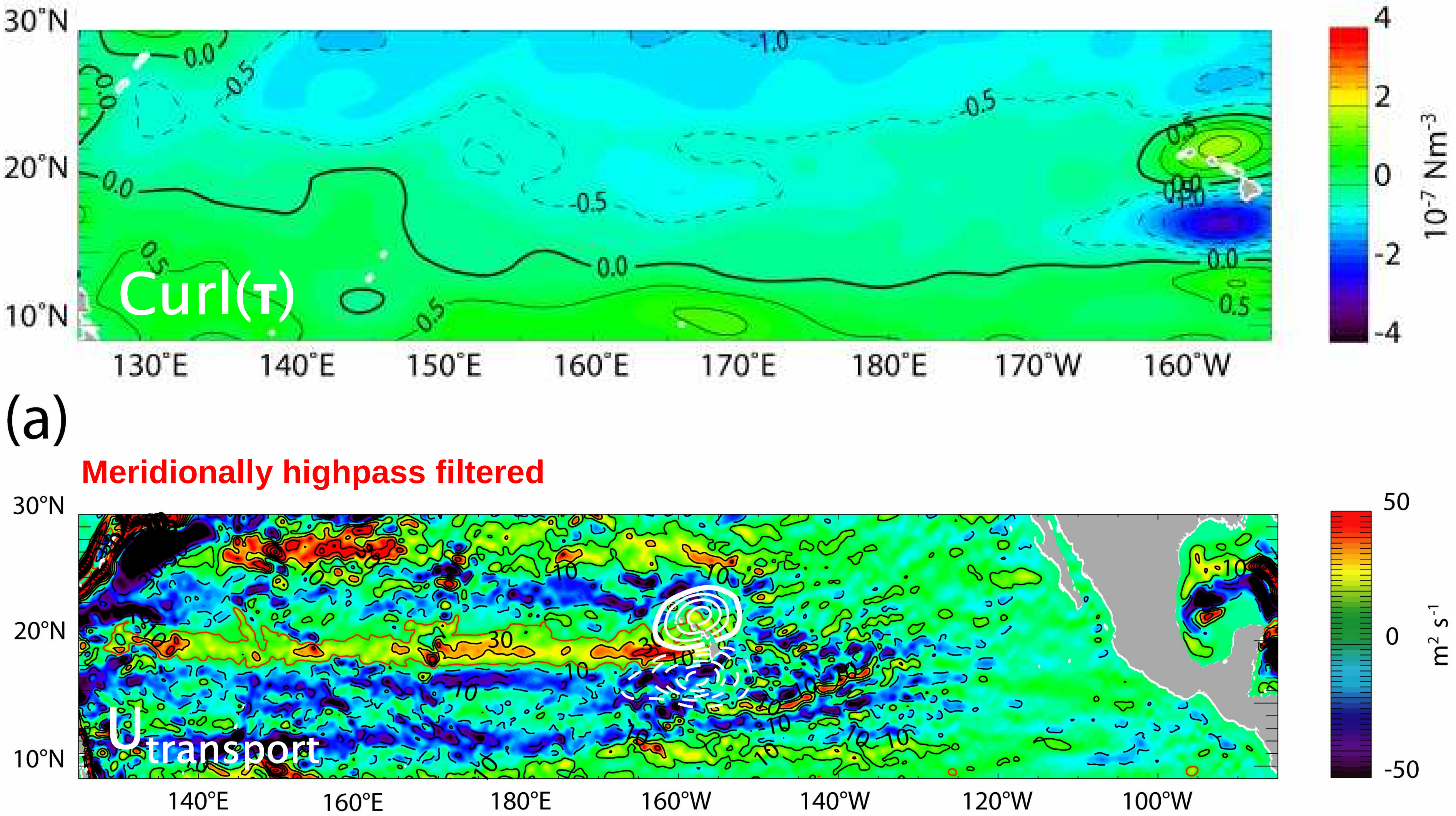
Conclusion

Hawaiian Lee Countercurrent: model and data

- **OFES** (Masumoto et al. 2004, JES): **global ocean model** based on GFDL MOM3.
- **Eddy-resolving** (1/10º) with **54 vertical levels**. **Bottom topography** is from OCCAM 1/30º.
- **KPP vertical mixing** scheme (Large et al. 1994, RG).
- **Study region = (sub)tropical North Pacific Ocean** (9ºN - 30ºN, 125ºE - 85ºW).
- **2 simulations** analyzed over 1999-2008 (**10 years**): **OFES-N** (NCEP forcing) and **OFES-Q** (QuikSCAT forcing), similarly to Sasaki Nonaka 2006, GRL. Monthly means are used here.

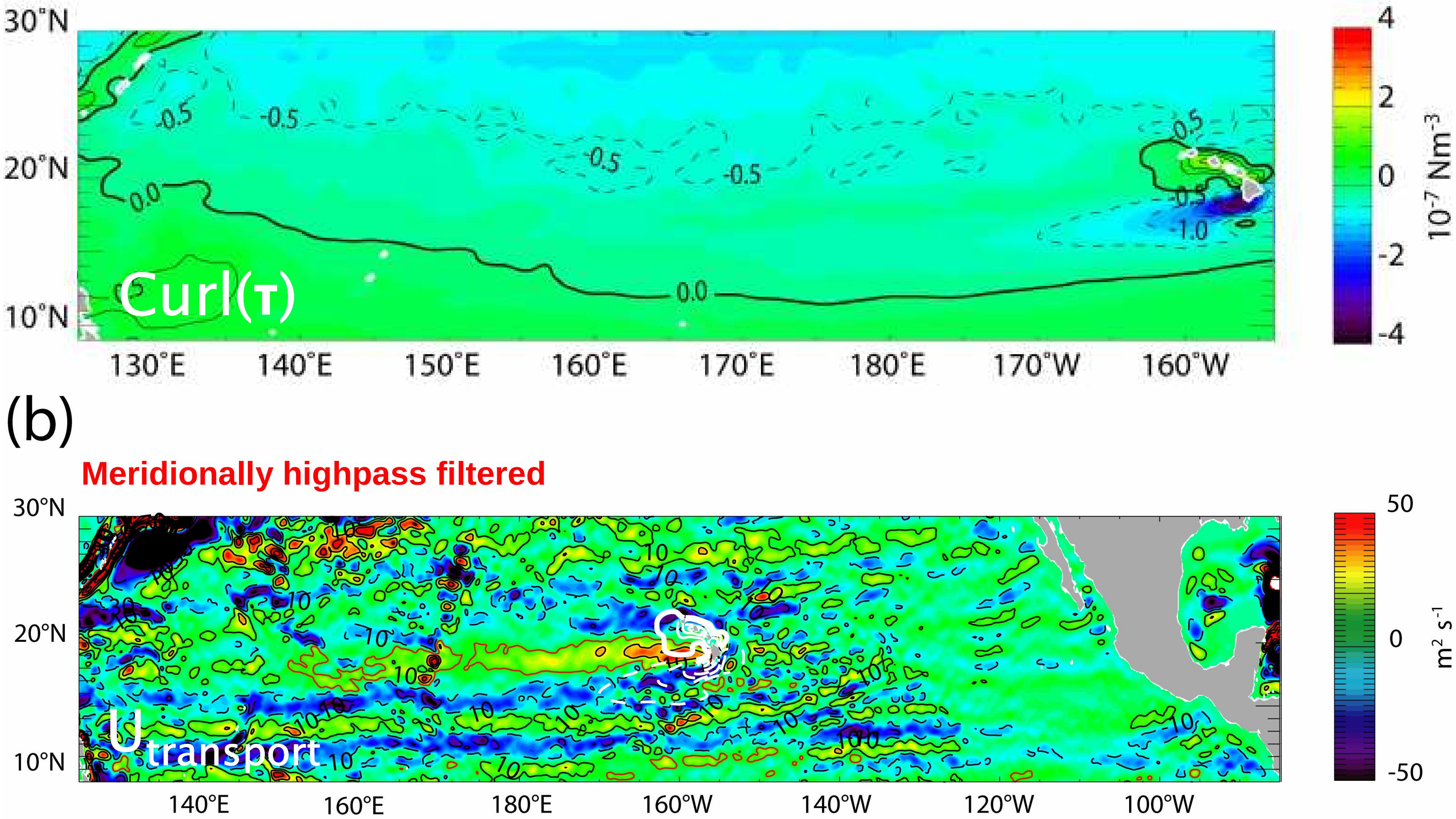
- Lebedev et al. 2007, IPRC = **surface** and **deep currents** estimated from trajectories of 4284 **ARGO** floats over 1997-2007 (**11 years**).
- **Surface velocities** = linearly regressed from float coordinates fixed by satellite.
- **Deep velocities** estimated from float displacements during submerged phase of the cycle.

HLCC in OFES: NCEP forcing



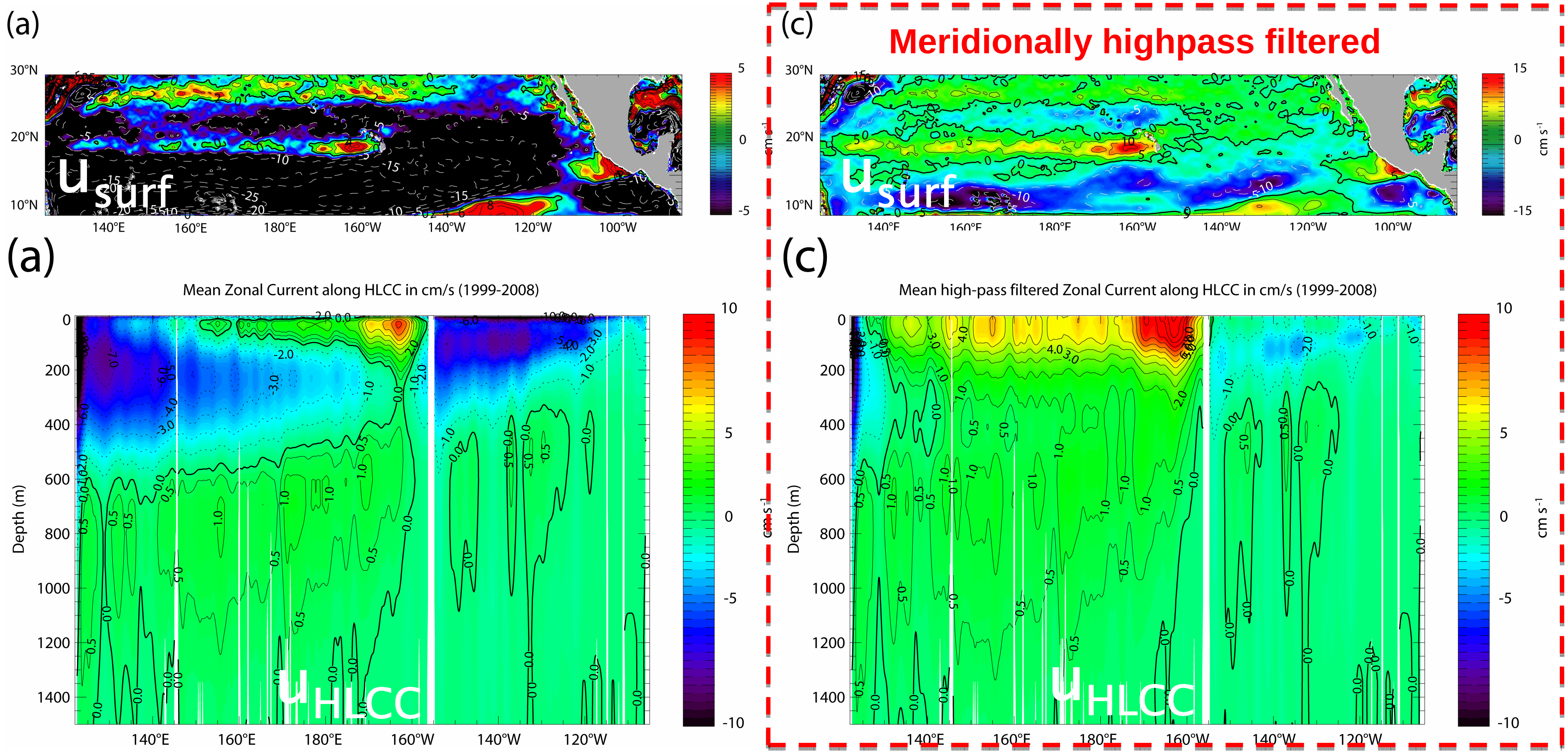
- **HLCC = *β*-plume** forced by WSC **dipole** around Hawaii.
- **Transport is ~ x-independent** (at least east of 170ºE).

HLCC in OFES: QuikSCAT forcing



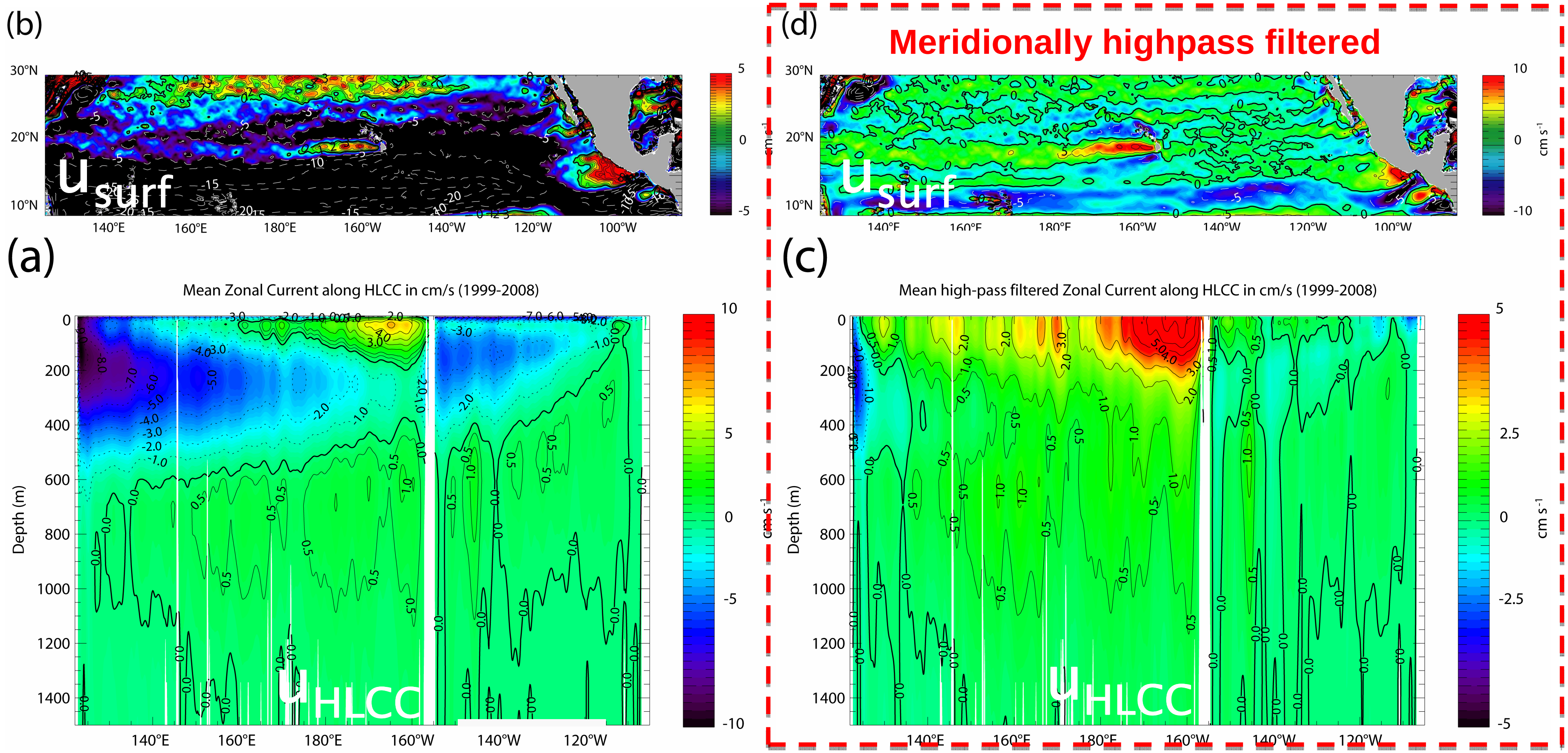
- **Forcing scale = smaller** => **narrower HLCC** meridional scale.
- **Transport also decays:** due to far-field wind or eddy dissipation?

HLCC in OFES: NCEP forcing



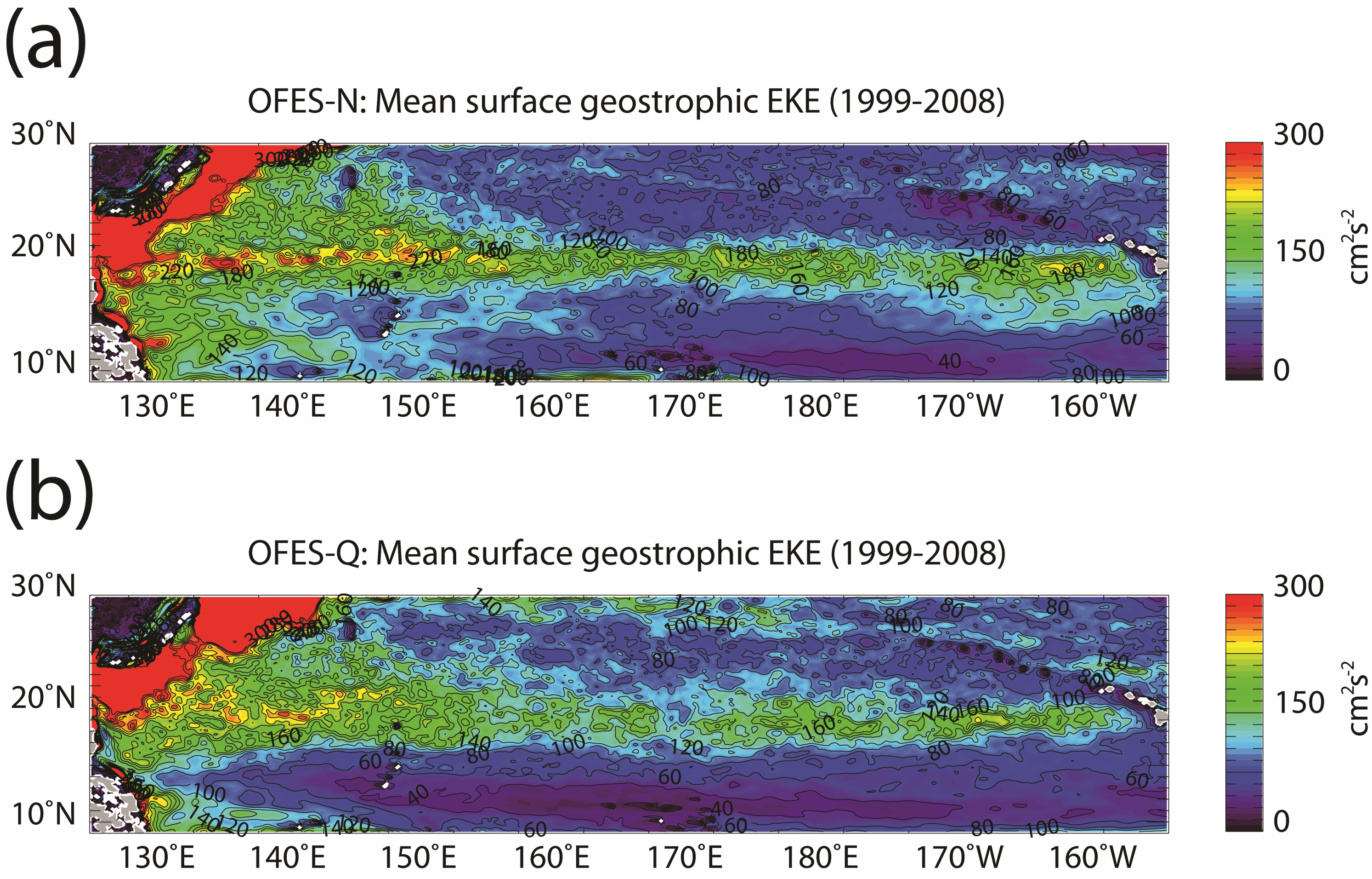
- **Surface current** extends to **~155ºE**, but **weaker west of ~170ºW**.
- **Baroclinic** flow = NEC + ***β*-plume**: surface decay, westward deepening, emergence of deep flow.

HLCC in OFES: QuikSCAT forcing



- **Surface current decay** scale = **shorter**: consistent with idealized model.
- **Baroclinic flow**: **less consistent** with idealized β-plume. Due to eddies and/or far-field wind?

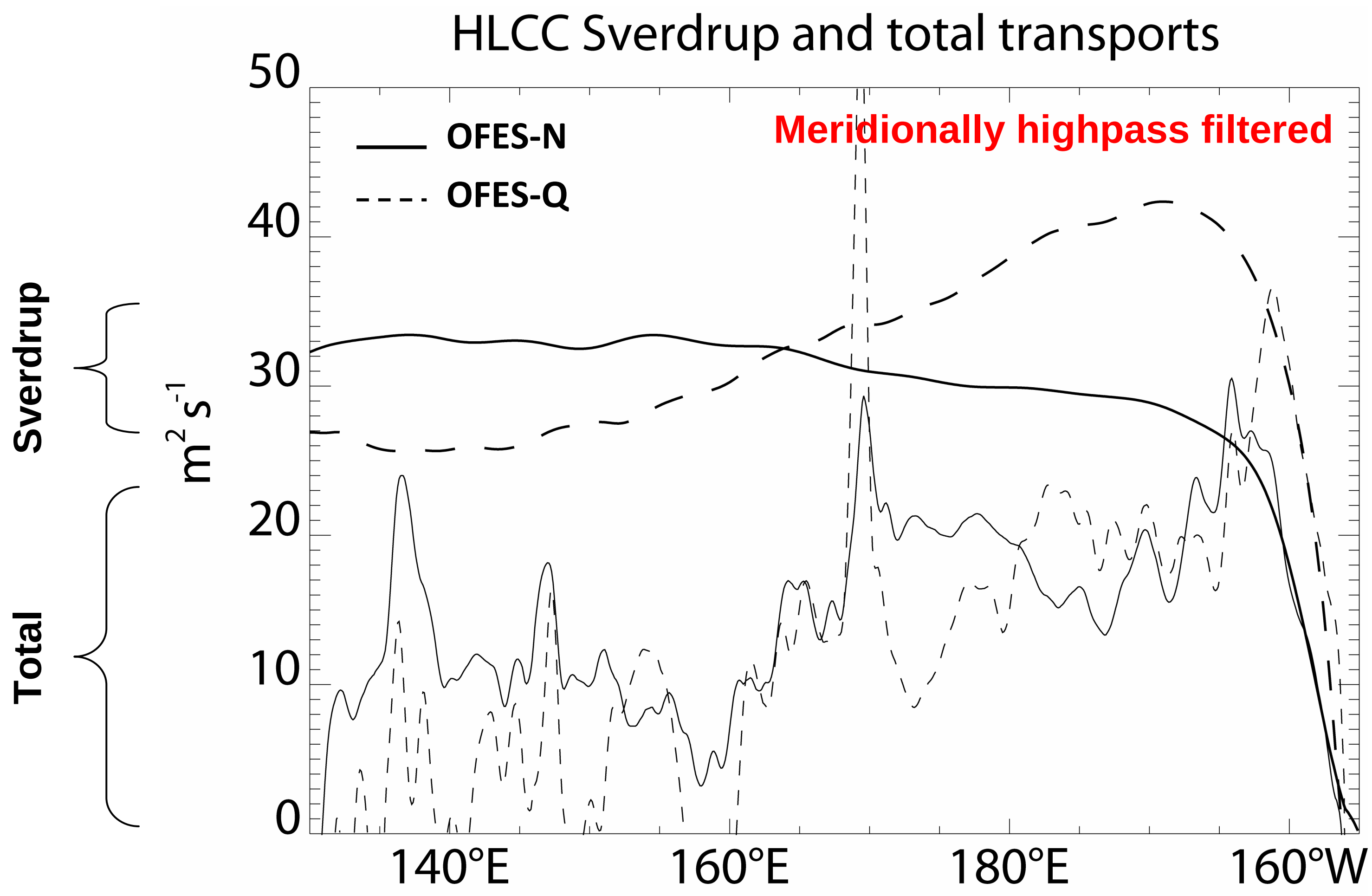
Transport decay: eddies?



- **High EKE along the HLCC axis** due to eddies shed in the island lee and generated by baroclinic/barotropic instabilities in the far field (Calil et al. 2008, Yoshida et al. 2011, etc.).

- **Similar EKE levels in OFES-N/Q** (slightly higher in OFES-N) suggest **eddies may not be responsible** for differences in barotropic transport.

Transport decay: Sverdrup flow?



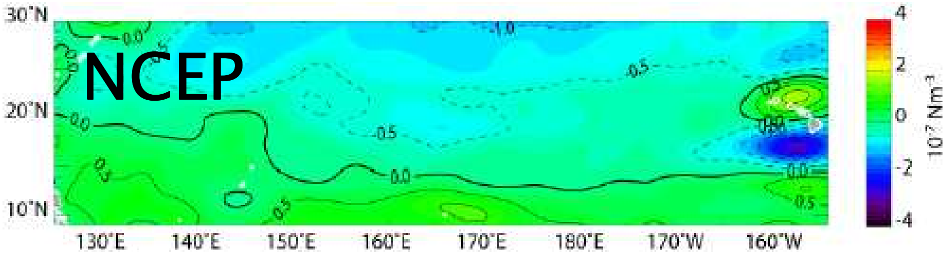
- NCEP WSC favors an ~x-independent transport west of Hawaii, whereas **QuikSCAT WSC likely contributes to the faster-decaying transport**.
- This is likely due to the effect of **air-sea interaction** in the far field with **tilted WSC dipole and HLCC**. Possibly also contributes to **discrepancy with idealized vertical structure**.

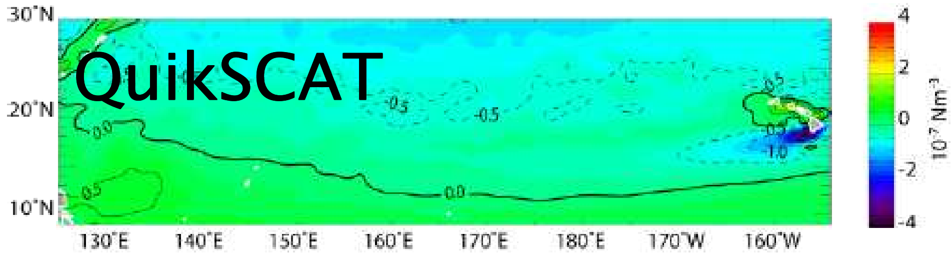
HLCC in OFES / observations: viscosity or diffusion?

$$M_n = \frac{4R^2}{\pi^2 \sigma R_n^2}$$

σ ~ 10

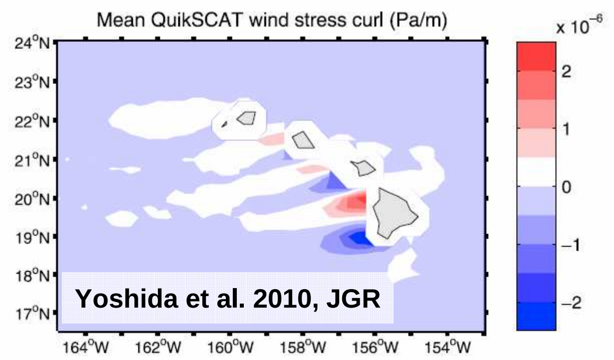
R_1 ~ 60 km





R ~ 200 km $\Rightarrow$ M_1 ~ 0.45, n_0 ~ 2

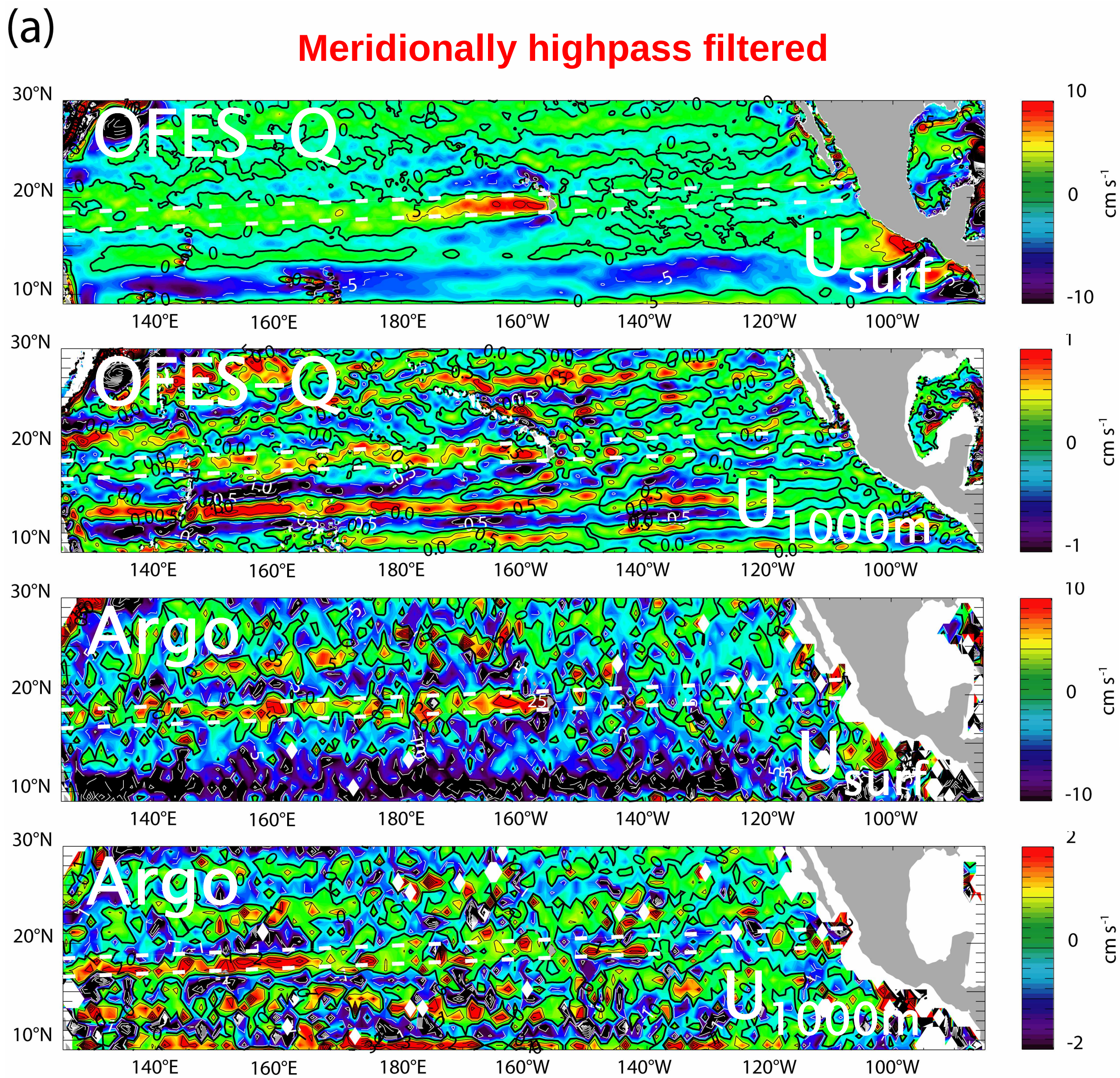
R ~ 100 km $\Rightarrow$ M_1 ~ 0.11, n_0 ~ 4



R ~ 40 km $\Rightarrow$ M_1 ~ 0.02, n_0 ~ 8 (ROMS) M_1 ~ 0.04, n_0 ~ 6

- Estimates derived from observed winds suggest **viscosity plays a significant role**.
- Wind and mixing in idealized runs are reasonable (perhaps even more than those in OFES).

HLCC in the real ocean?



Argo consistent w/ deepening:

- **max** near **Hawaii @ 0m**
- near **145-165ºE @ 1000m**

Caution: scarce / noisy data... but is the only available data! (perhaps)

OFES-Q ≠deep flow increase but air-sea coupling may be poorly represented

(Chelton Xie 2010)

Conclusion

- An idealized linear primitive-equation model and an analytical linear continuously-stratified model show that for **small-scale wind forcing**, **vertical viscosity** (and diffusion) **damps baroclinic Rossby waves**, resulting in a ***β*-plume westward thickening** with **decay of surface zonal jets** and **emergence of deep flow**. Barotropic flow is in Sverdrup balance and zonal transport is x-independent.

- **Zonal change** in baroclinic flow occurs over **shorter** (longer) **distances** for **smaller** (larger) **meridional forcing scales**, if forcing smaller than 1st Rossby radius.

- **Eddies** are not necessary for the early termination of *β*-plumes.

Conclusion (2)

- High-resolution OGCM simulations suggest that the **Hawaiian Lee Countercurrent may have baroclinic and barotropic structures consistent with idealized linear *β*-plumes**. Model surface HLCC also has **similar sensitivity to scale of WSC** in lee of Hawaii.

- Specific roles of far-field **air-sea interaction** and **eddy fluxes** need to be addressed.

- **A possible deep extension of HLCC is found** for the first time in **ARGO float trajectory data** and in OGCM simulations.

Gracias!

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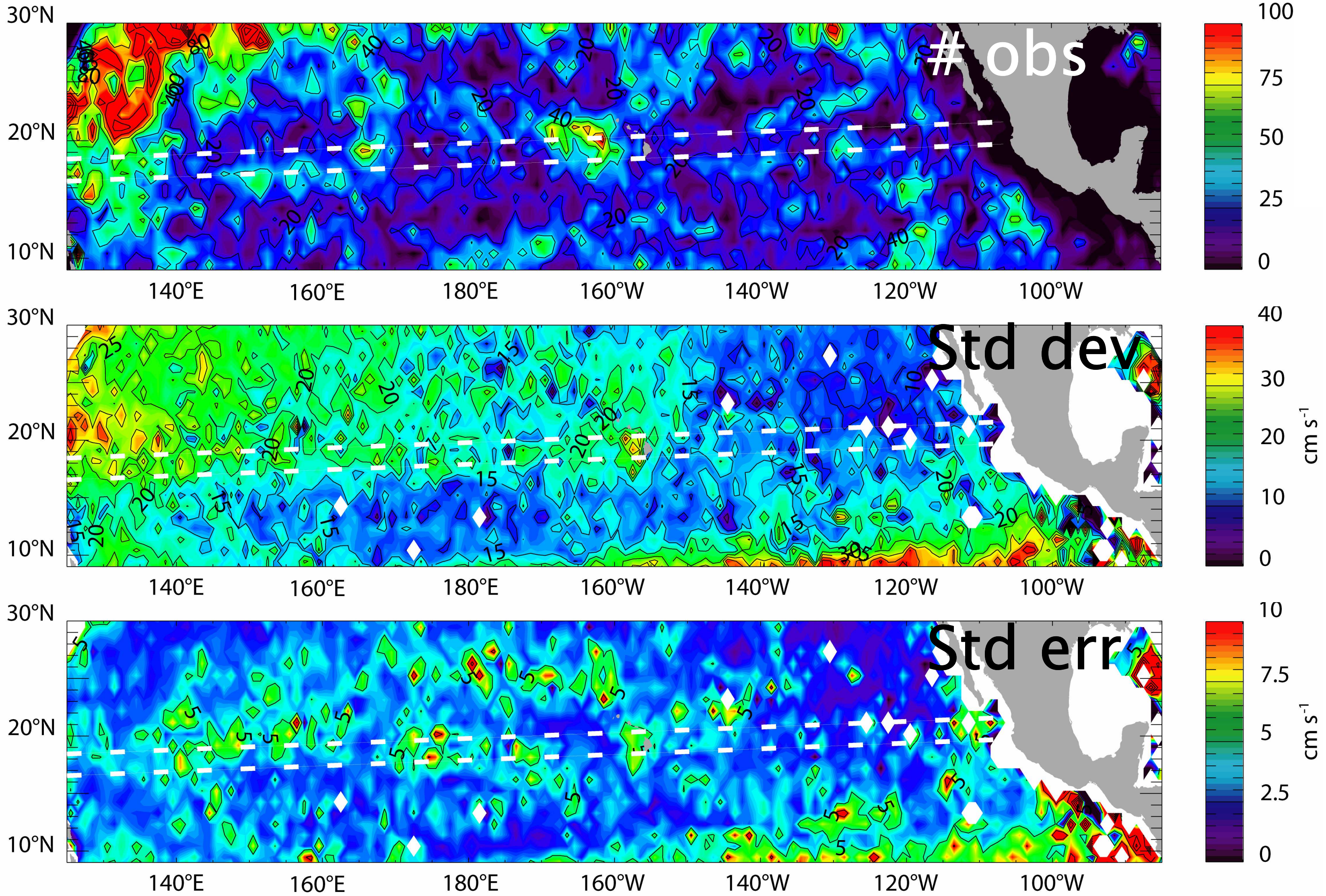
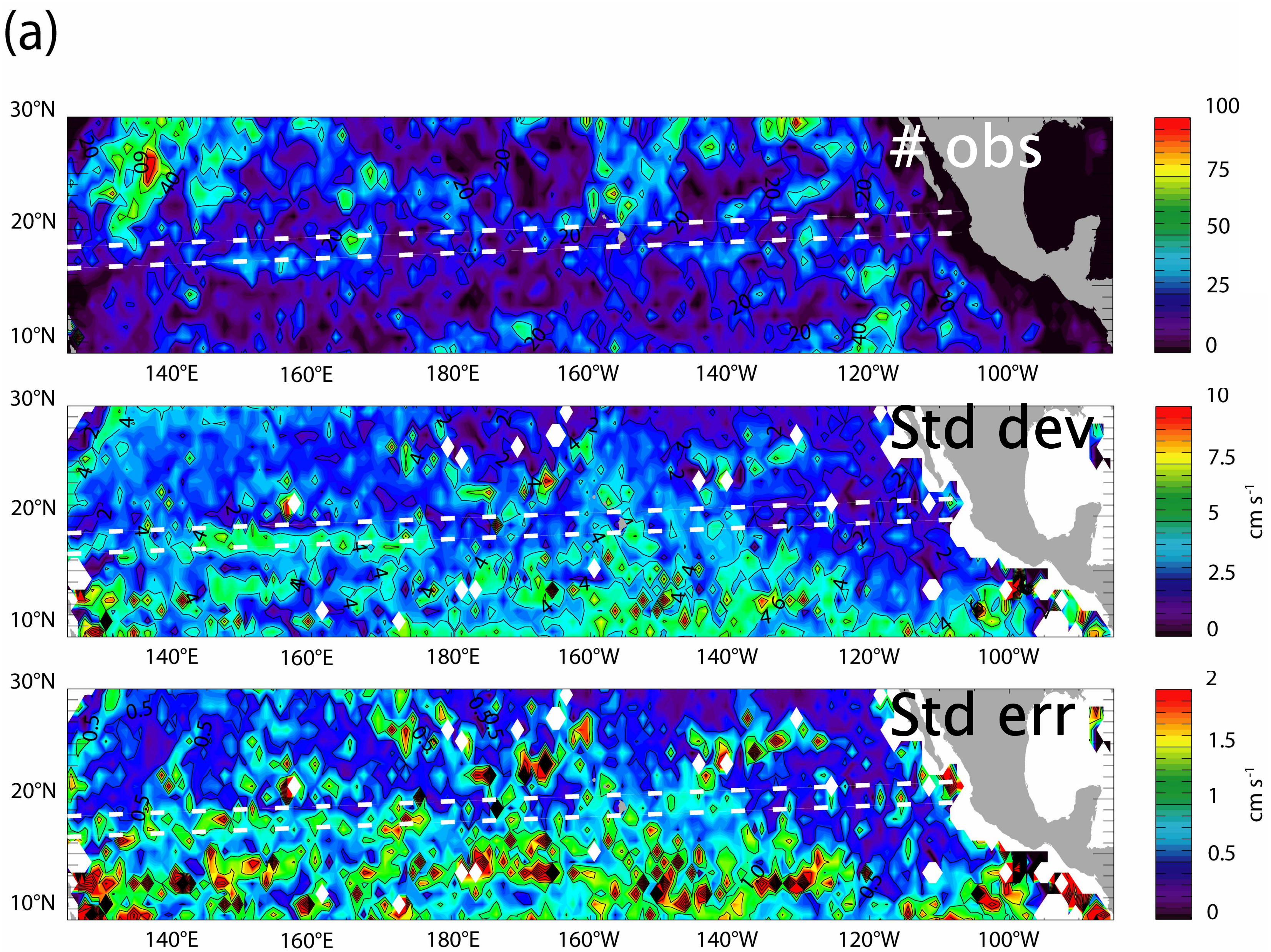
Standard error in Argo data @ 0 m

(a)

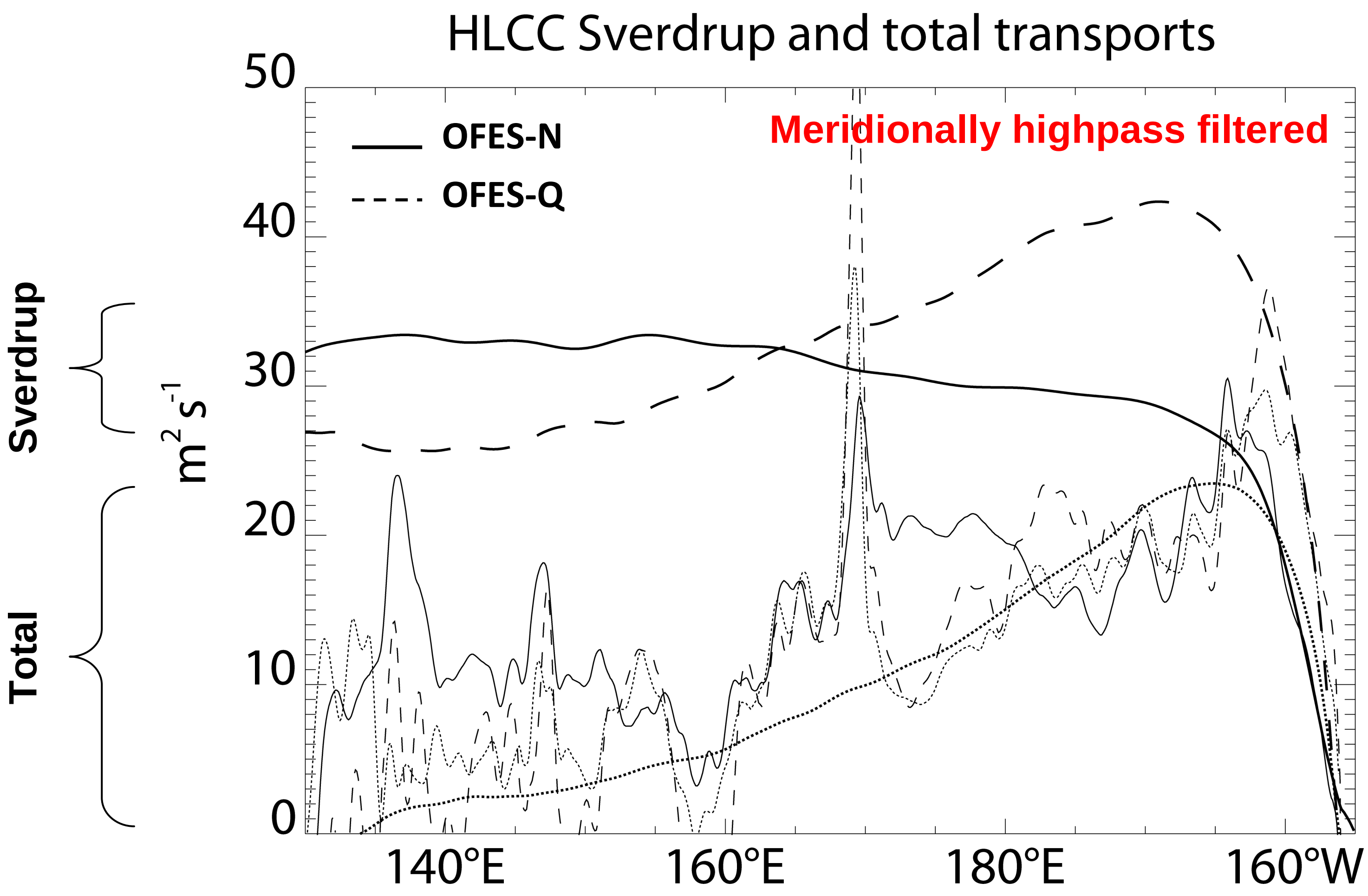


Standard error in Argo data @ 1000 m

(a)



Transport decay: Sverdrup flow?



- NCEP WSC favors an ~x-independent transport west of Hawaii, whereas **QuikSCAT WSC likely contributes to the faster-decaying transport**.
- This is likely due to the effect of **air-sea interaction** in the far field with **tilted WSC dipole and HLCC**. Possibly also contributes to **discrepancy with idealized vertical structure**.