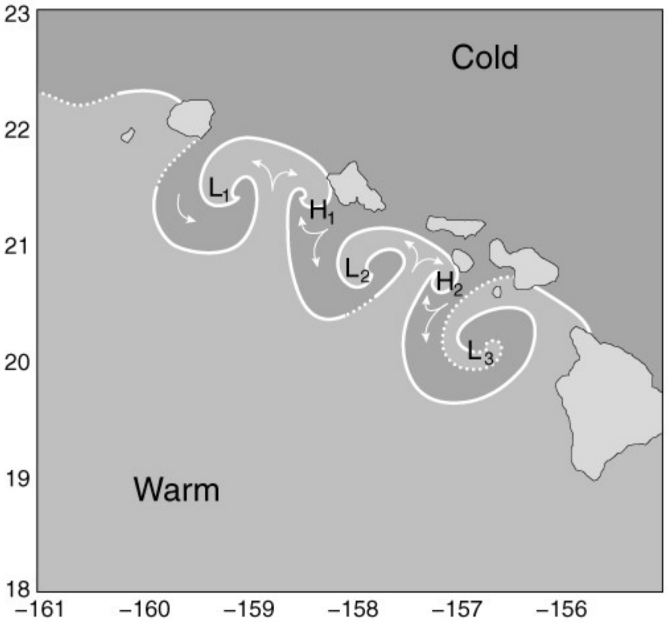
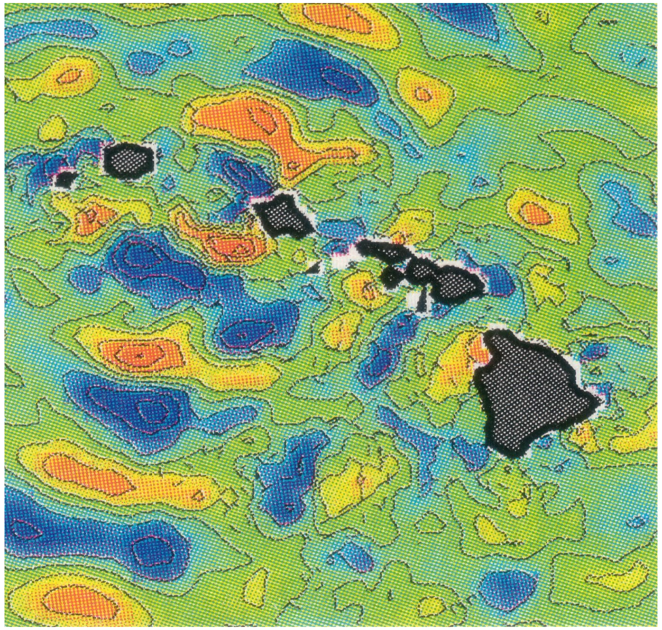
Observations of the Impact of Mesoscale Currents on Internal Tide Propagation



Cédric
Chavanne



Ph.D., University of Hawaii

Dissertation Committee:

Pierre Flament (Chairperson), Eric Firing,

John Learned, Doug Luther, Mark Merrifield

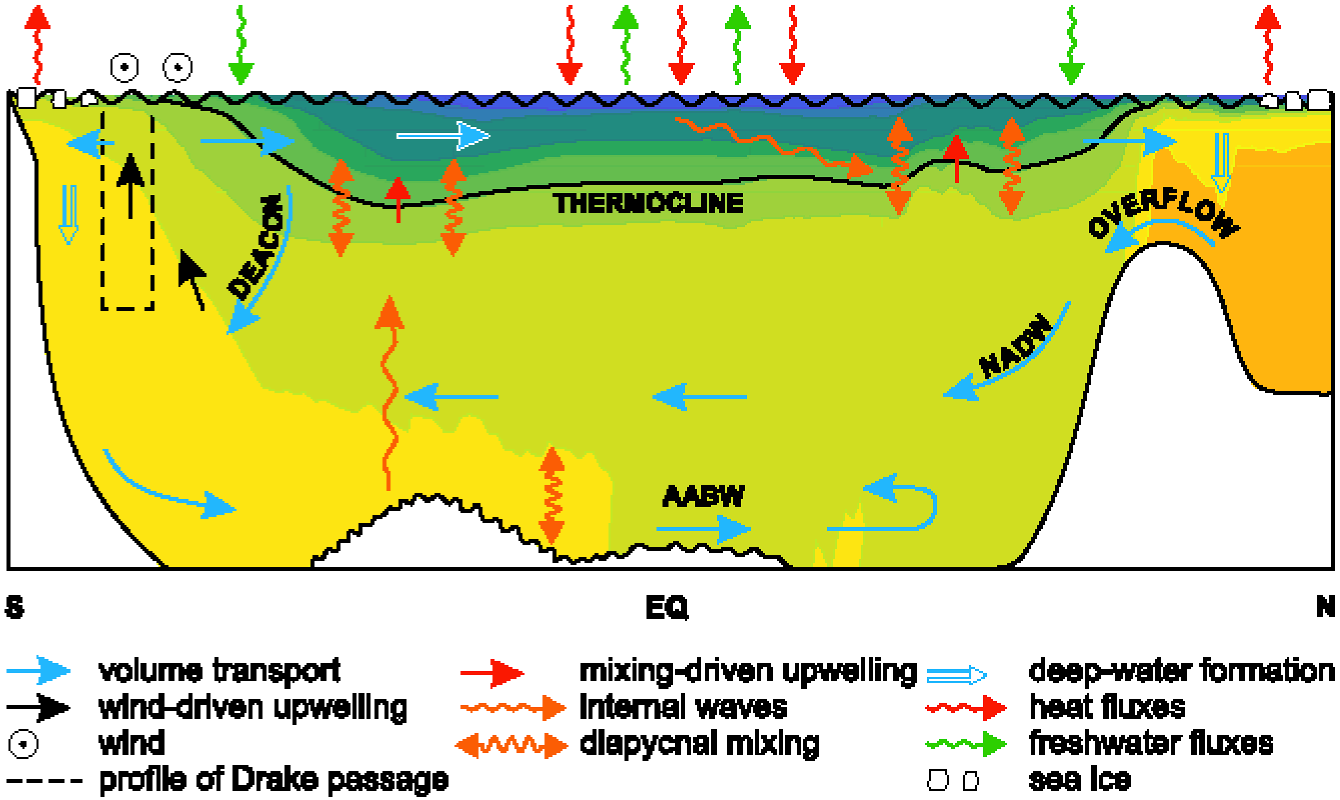
Why do internal tides matter ?

- often dominant energy signal
- ocean mixing
  - larval and pollutants dispersal
  - marine productivity
  - global climate

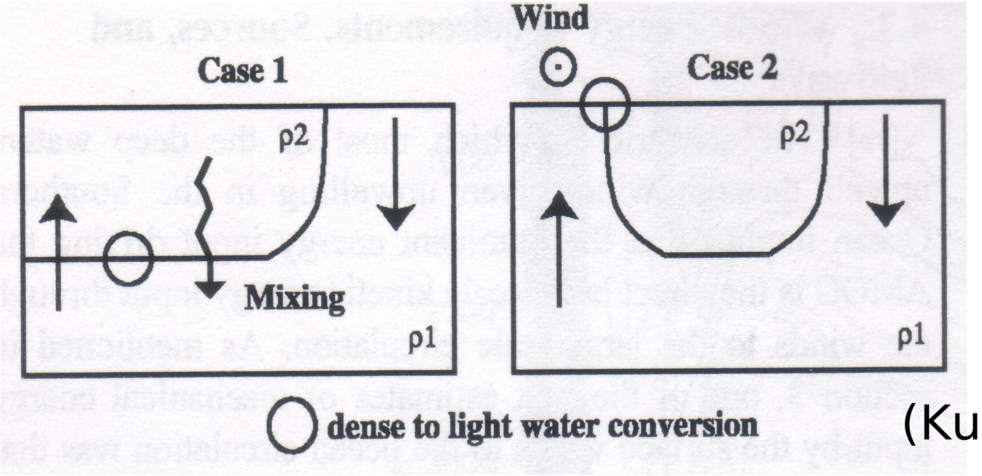
Relevance to climate predictions

- sensitivity of earth’s climate to vertical mixing in ocean
- sensitivity of ocean circulation to geography of mixing
- physically-based parameterization of diapycnal mixing in OGCMs

Meridional Overturning Circulation



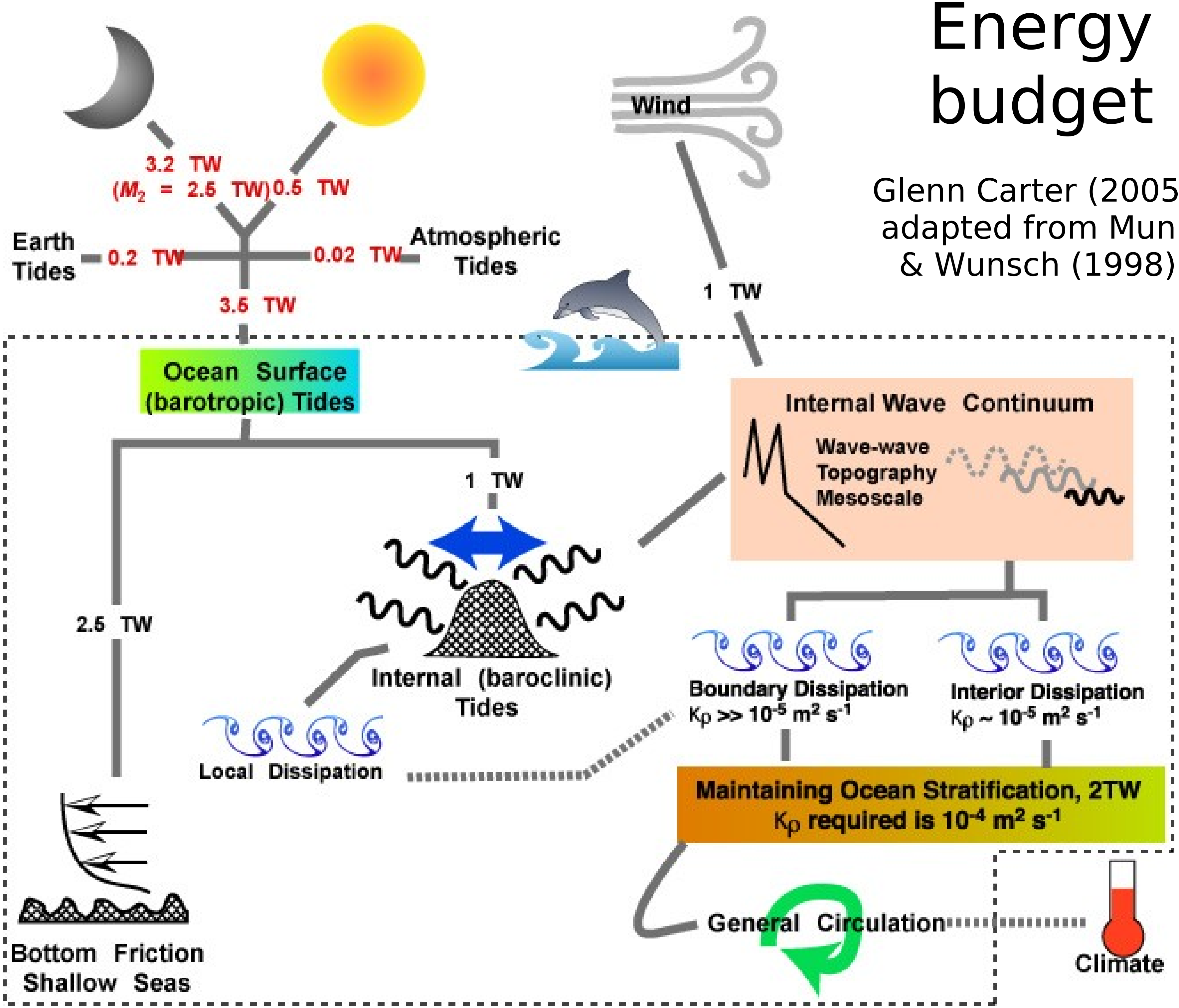
Driving Mechanisms



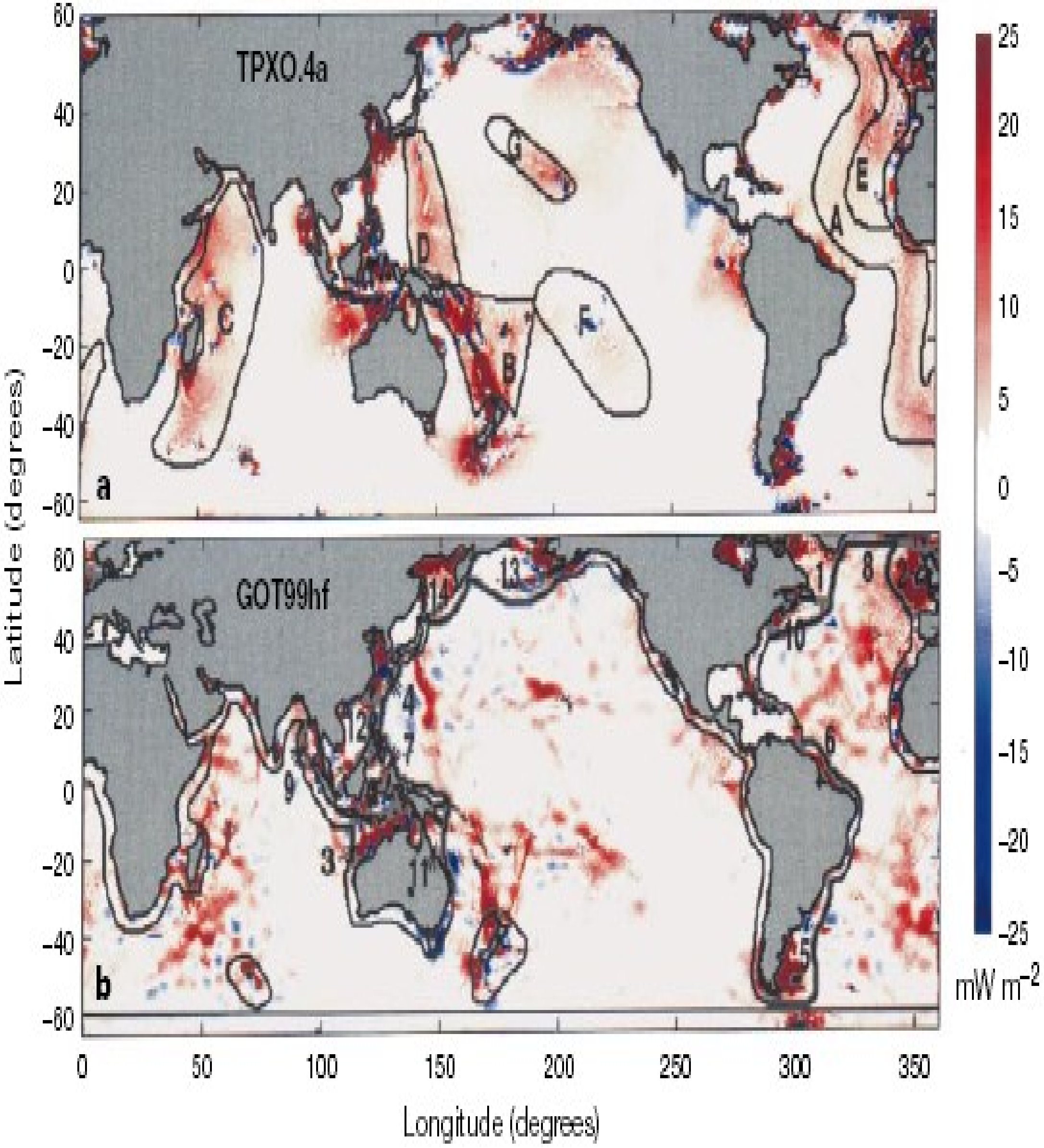
(Kuhlbrodt et al., 2007)

Energy budget

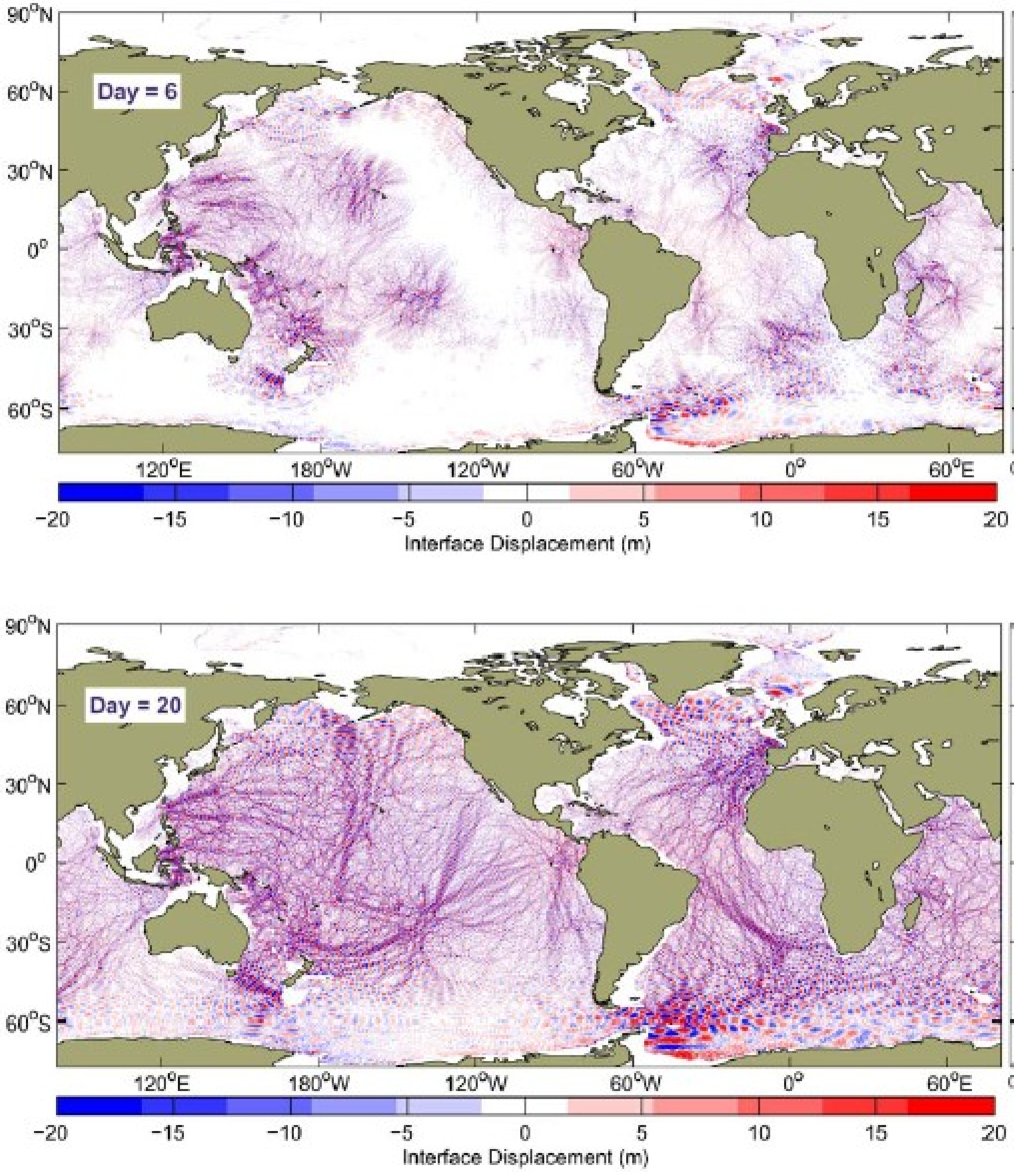
Glenn Carter (2005), adapted from Munk & Wunsch (1998)



Tidal dissipation and radiation



(Egbert & Ray, 2000)



(Simmons et al., 2004)

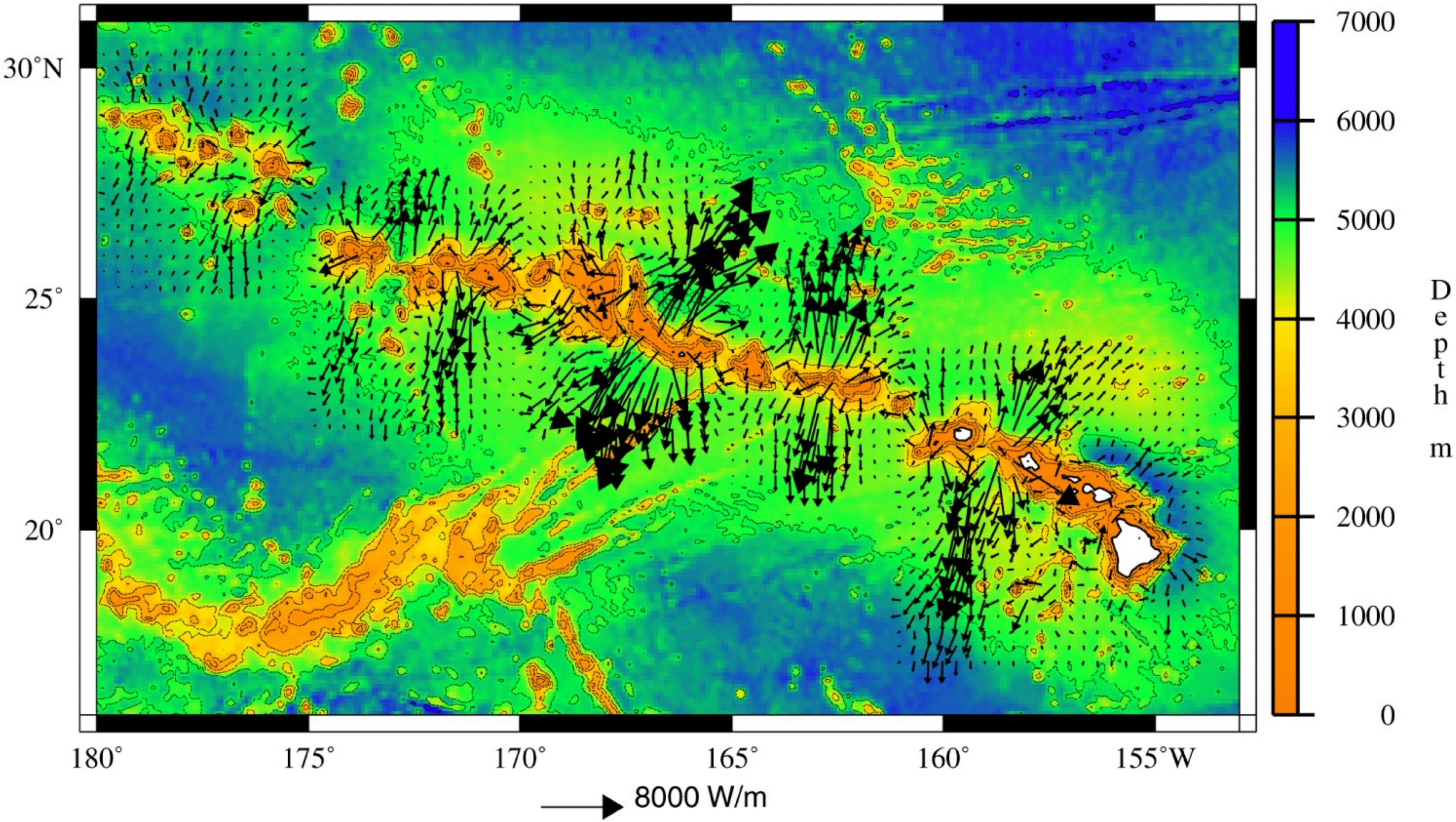
Hawaiian Ocean Mixing Experiment (HOME)

- determine if mid-ocean sites such as Hawaii are significant contributors to global mixing
- quantify tidal energy budget for an isolated abrupt topographic feature
- determine the principal mechanisms which transfer energy from large scale flows to turbulent motions

→ generalize results to global ocean

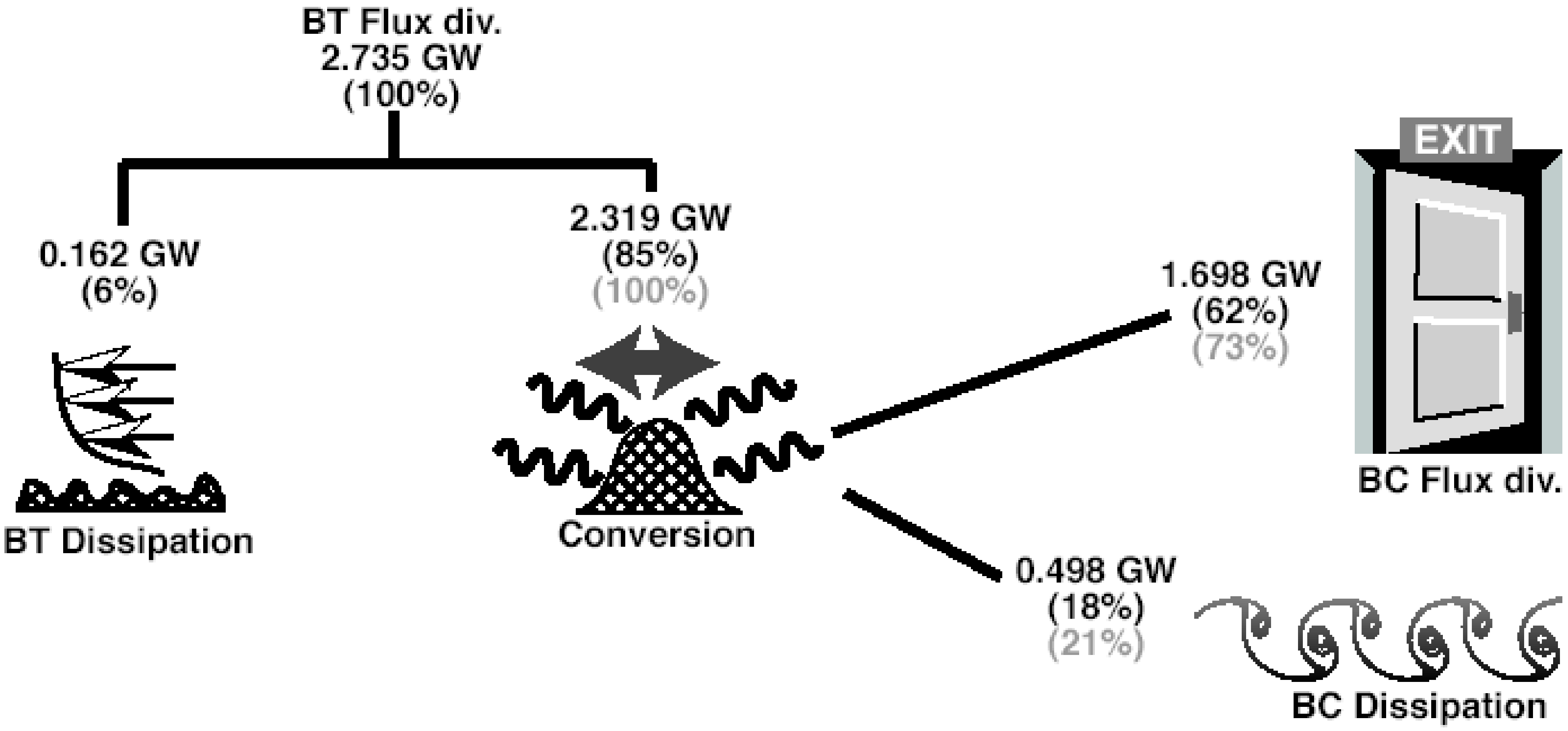
→ improve parameterisation of mixing in numerical models

Depth-integrated M_2 baroclinic energy fluxes



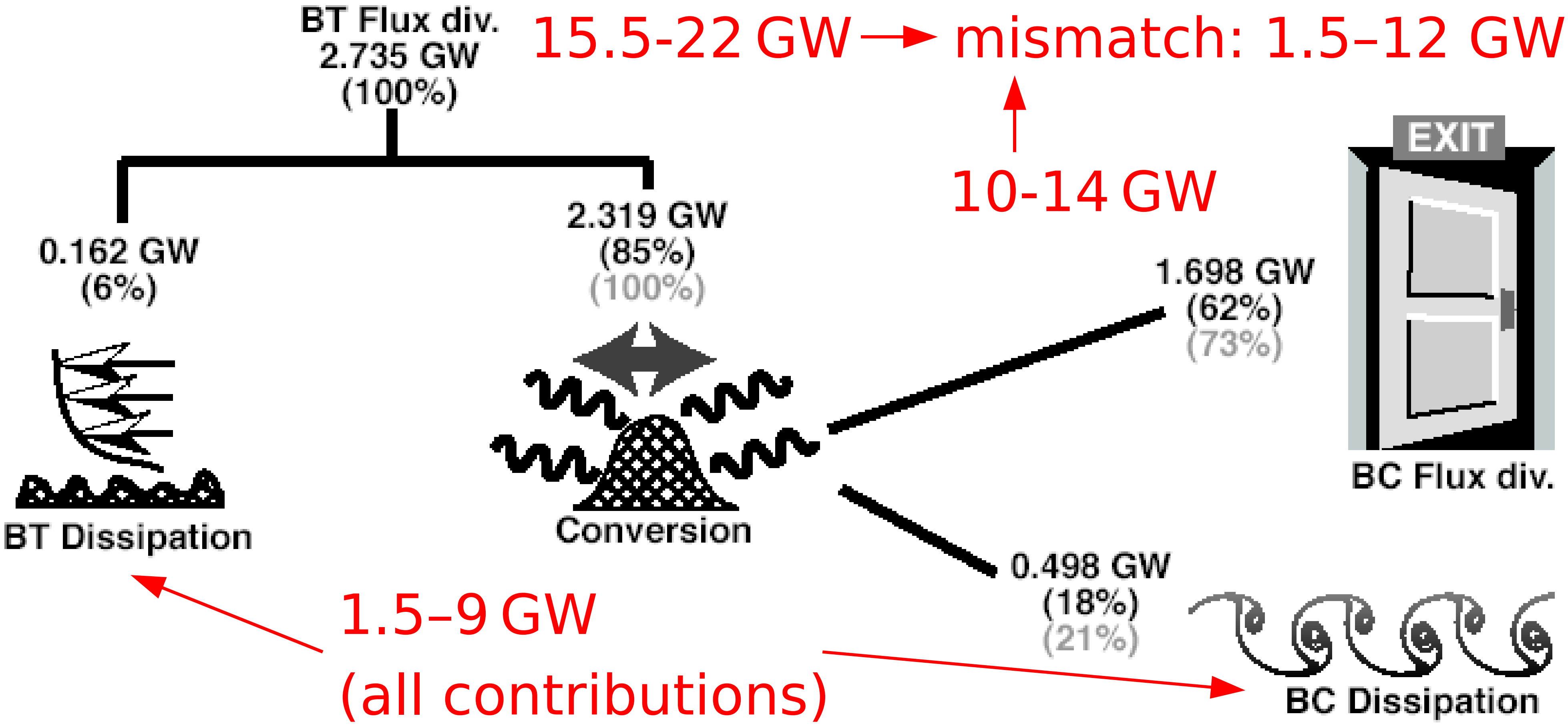
(Merrifield et al., 2001)

M_2 Energy Budget for Hawai`i



Carter et al. (2007): model budget for Kaua`i-Maui

M_2 Energy Budget for Hawai`i



15.5-22 GW → mismatch: 1.5–12 GW

10-14 GW

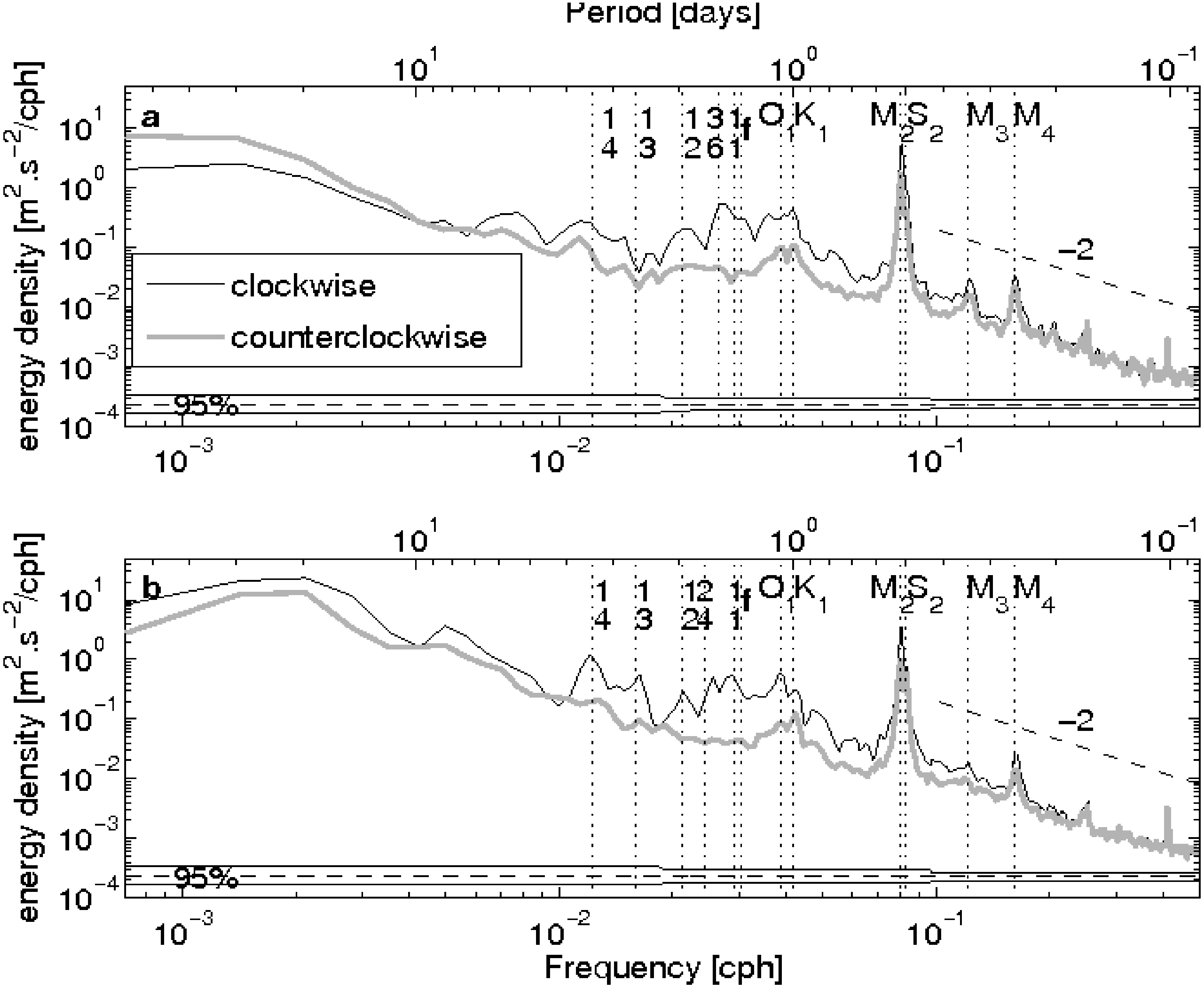
1.5–9 GW

(all contributions)

Carter et al. (2007): model budget for Kaua`i-Maui

Modeled and observed for entire Hawaiian Ridge

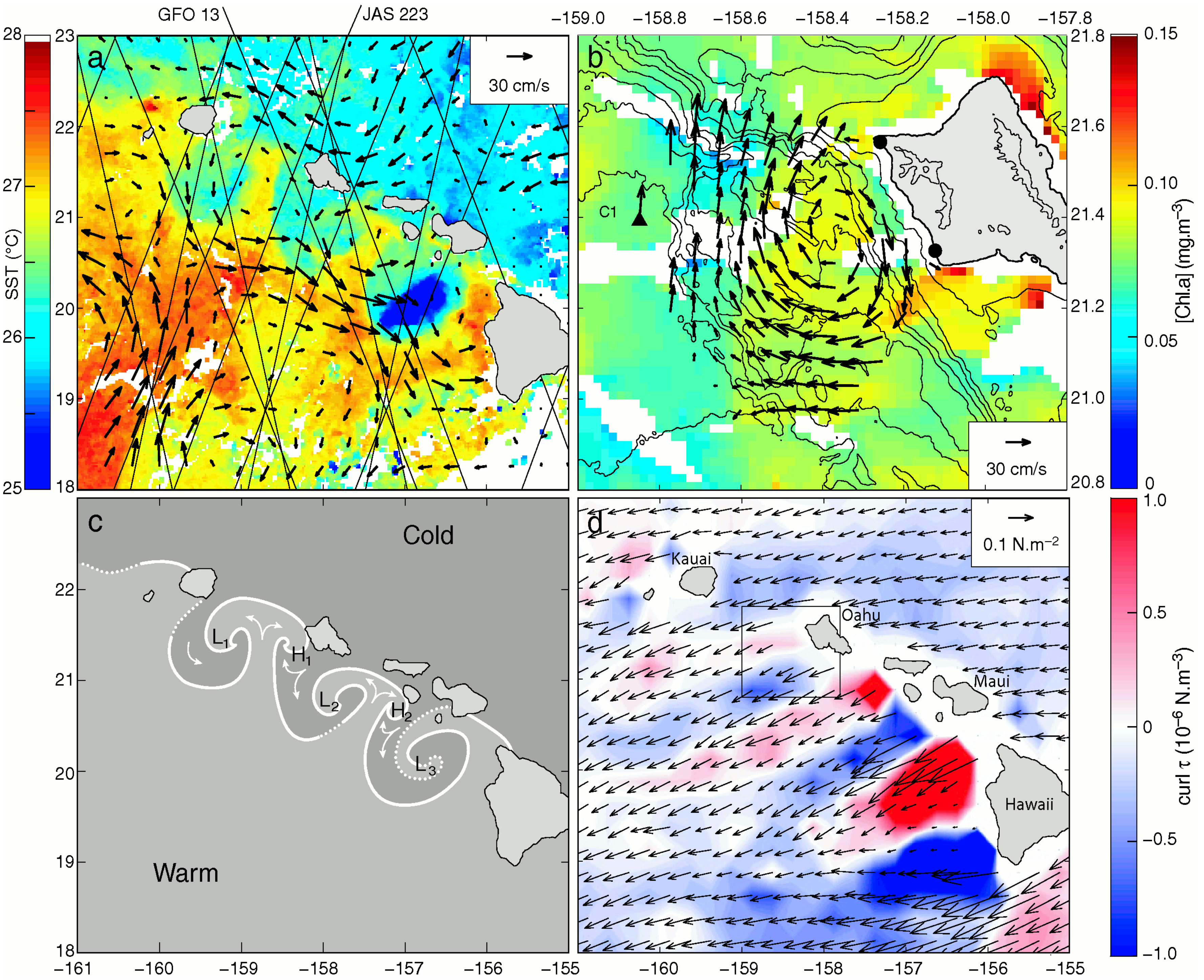
Surface currents power spectra



Fall 2002

Spring 2003

Mesoscale circulation



Questions

- What is the impact of mesoscale currents on internal tide propagation ?
- What are the implications for tidal energy budgets, and abyssal mixing ?

Methodology

- ray tracing: requires horizontal and vertical structures of mesoscale currents and stratification
- observations: HF-radios (horizontal) and moored ADCPs (vertical)
- stratification inferred from thermal wind balance
- comparisons with numerical predictions of tides in ocean at rest

Presentation Overview

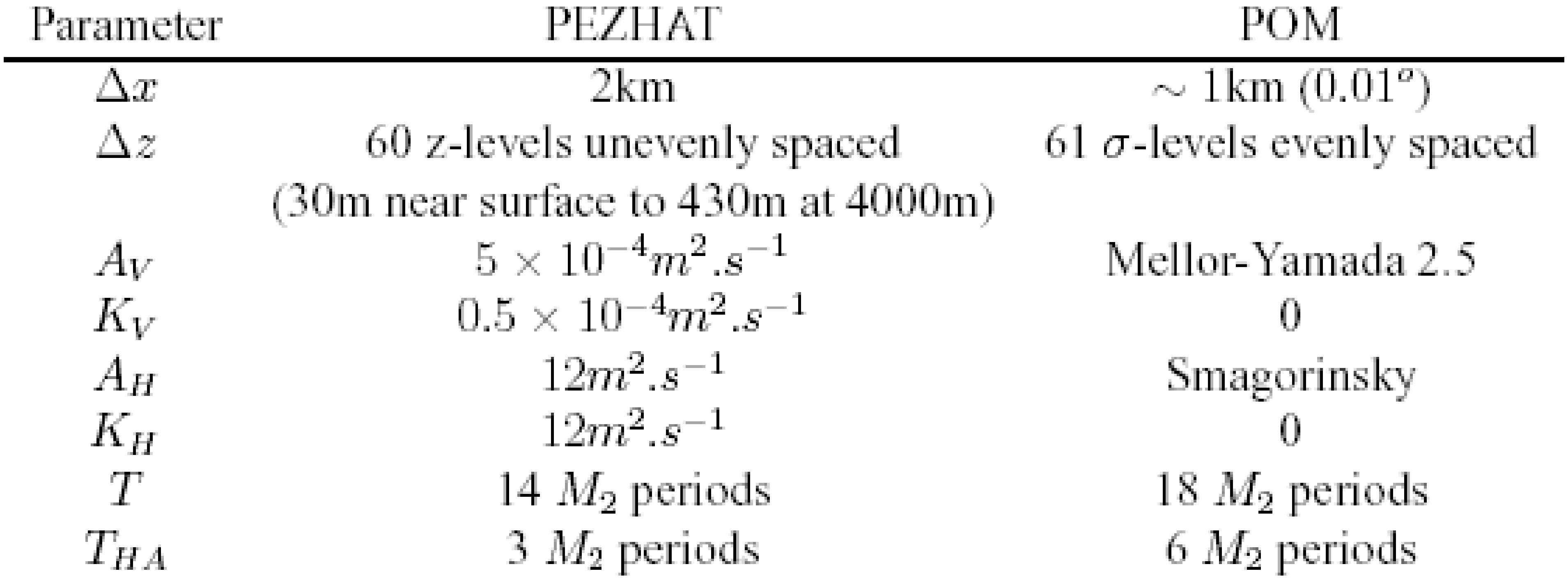
1) Tidal currents

2) Mesoscale currents

3) Propagation of internal tides through mesoscale currents

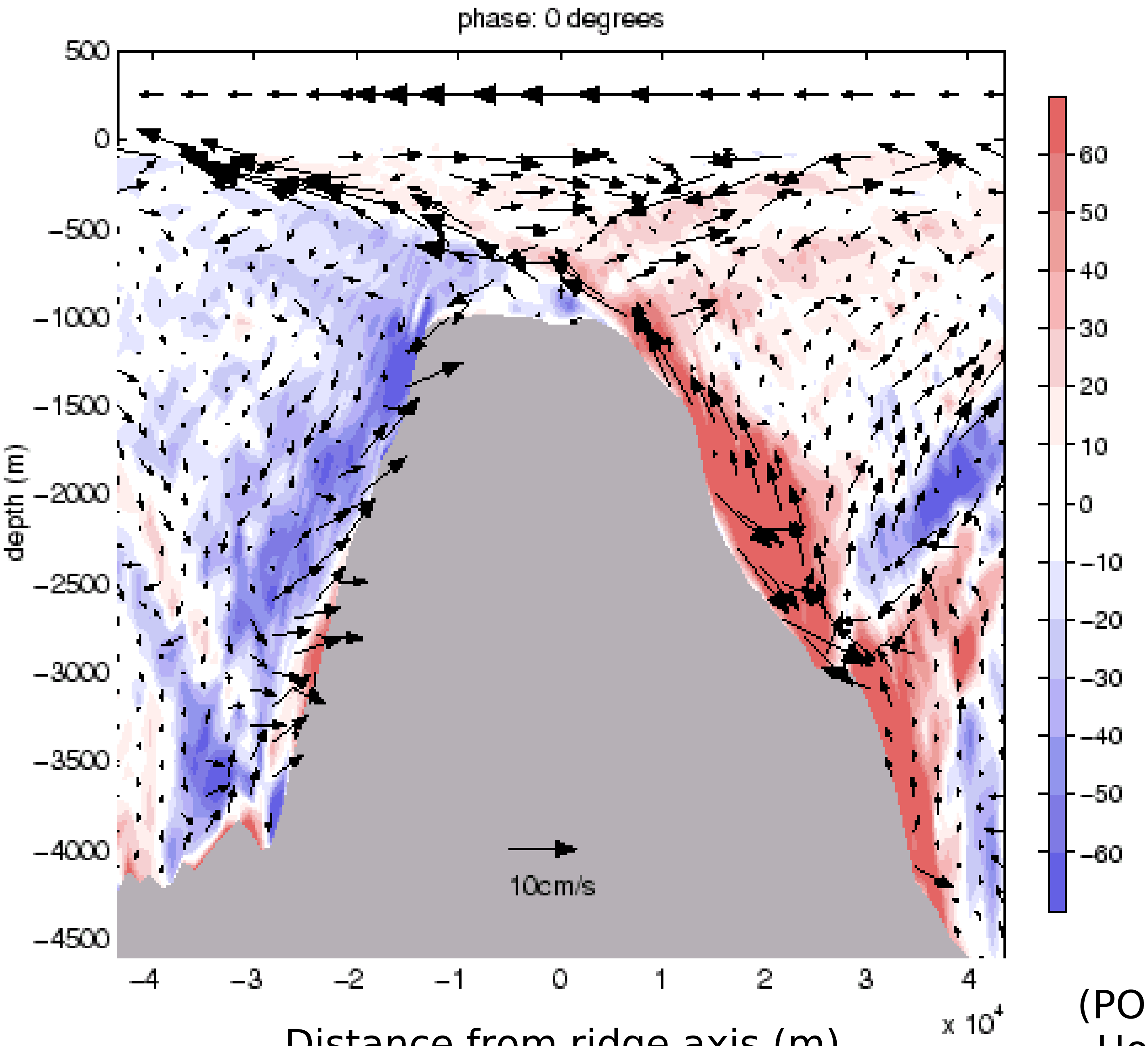
Numerical models of the tides

- **POM**: Princeton Ocean Model (Carter et al., 2007), nonlinear primitive equations
- **PEZHAT**: Primitive Equation Z-coordinate - Harmonic Analysis Tides (Zaron & Egbert, 2006), linearized



Parameter	PEZHAT	POM
Δx	2km	$\sim$ 1km (0.01^o)
Δz	60 z-levels unevenly spaced (30m near surface to 430m at 4000m)	61 σ-levels evenly spaced
A_V	$5 \times 10^{-4} m^2.s^{-1}$	Mellor-Yamada 2.5
K_V	$0.5 \times 10^{-4} m^2.s^{-1}$	0
A_H	$12 m^2.s^{-1}$	Smagorinsky
K_H	$12 m^2.s^{-1}$	0
T	14 M_2 periods	18 M_2 periods
T_{HA}	3 M_2 periods	6 M_2 periods

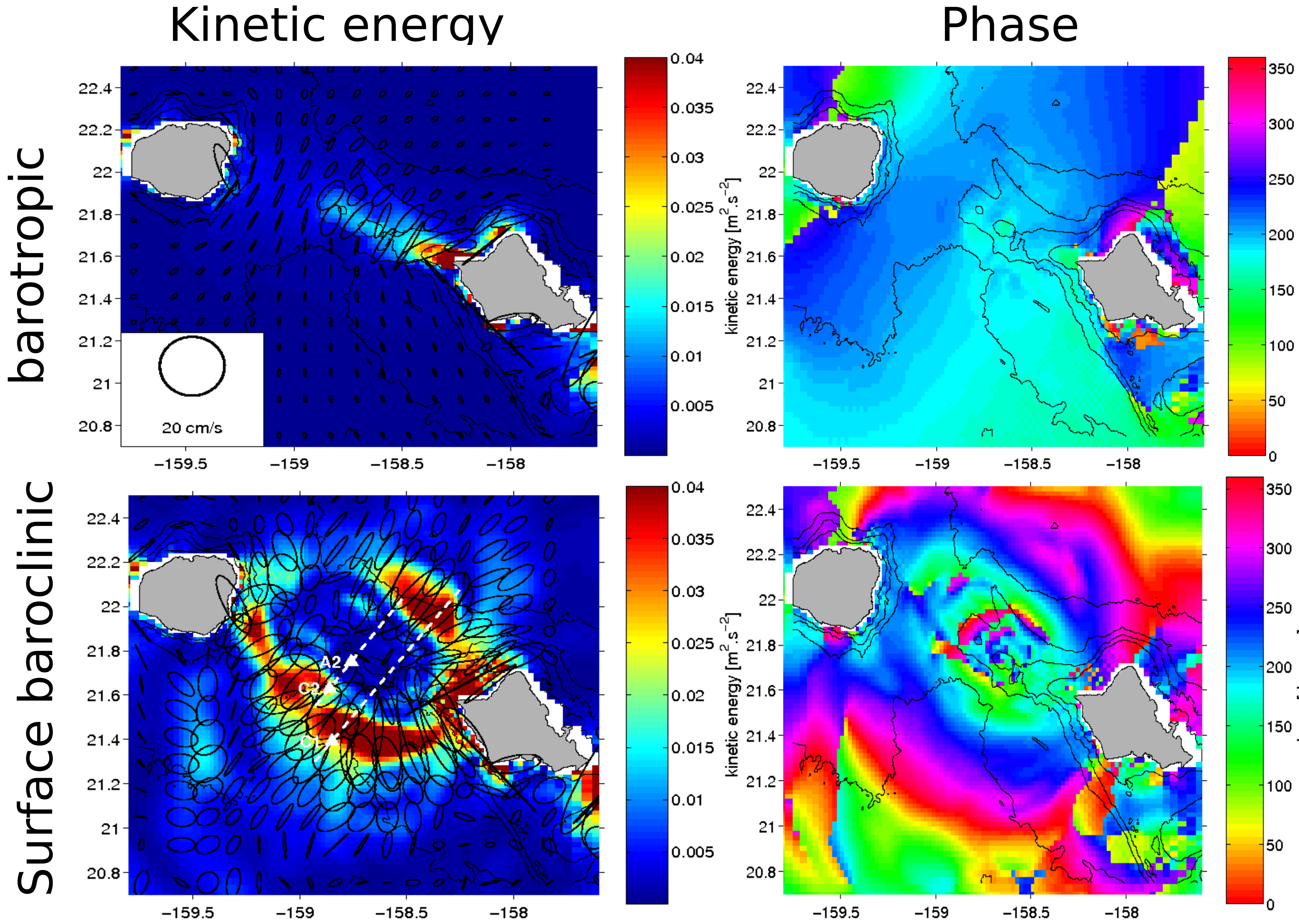
Internal tides generation



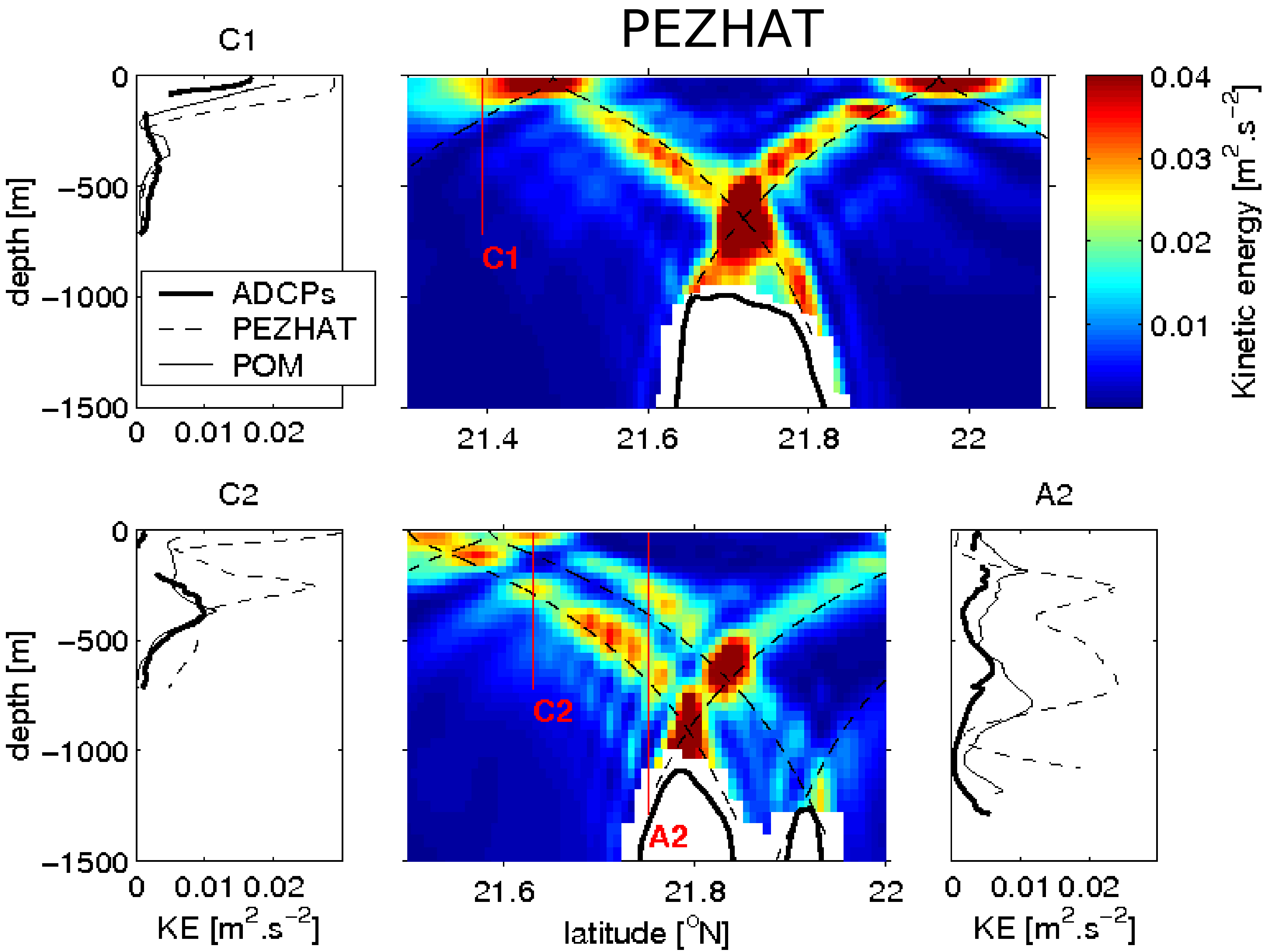
Distance from ridge axis (m)

(POM, Merrifield & Holloway, 2002)

PEZHAT currents



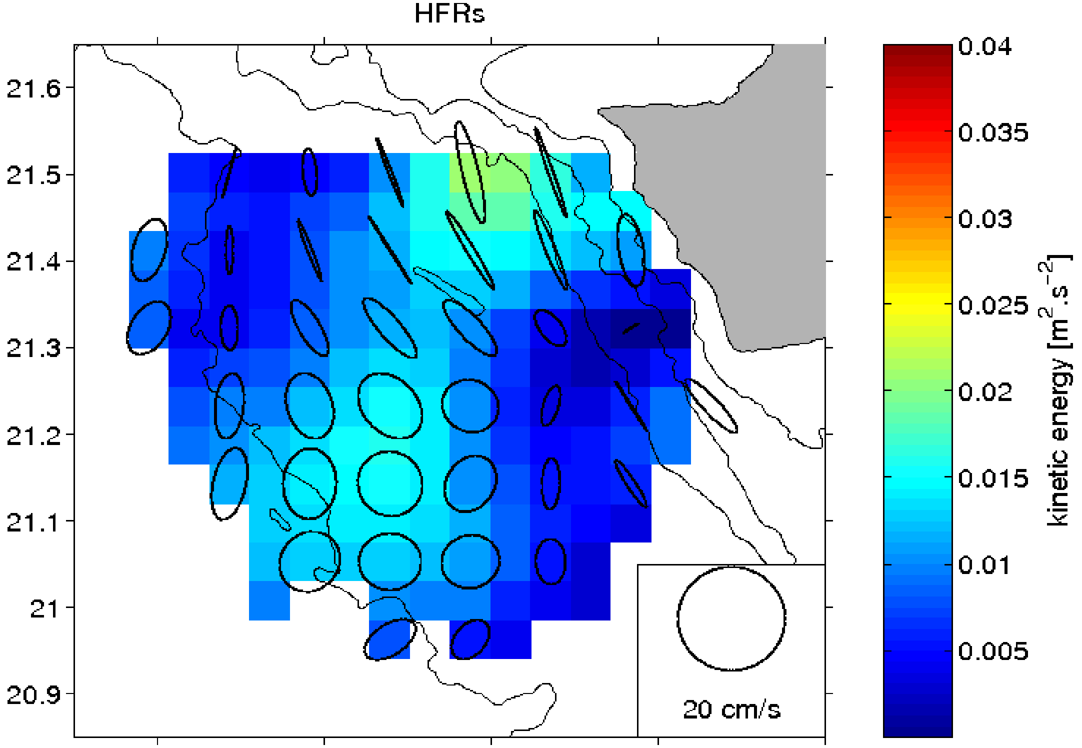
Vertical structure of KE



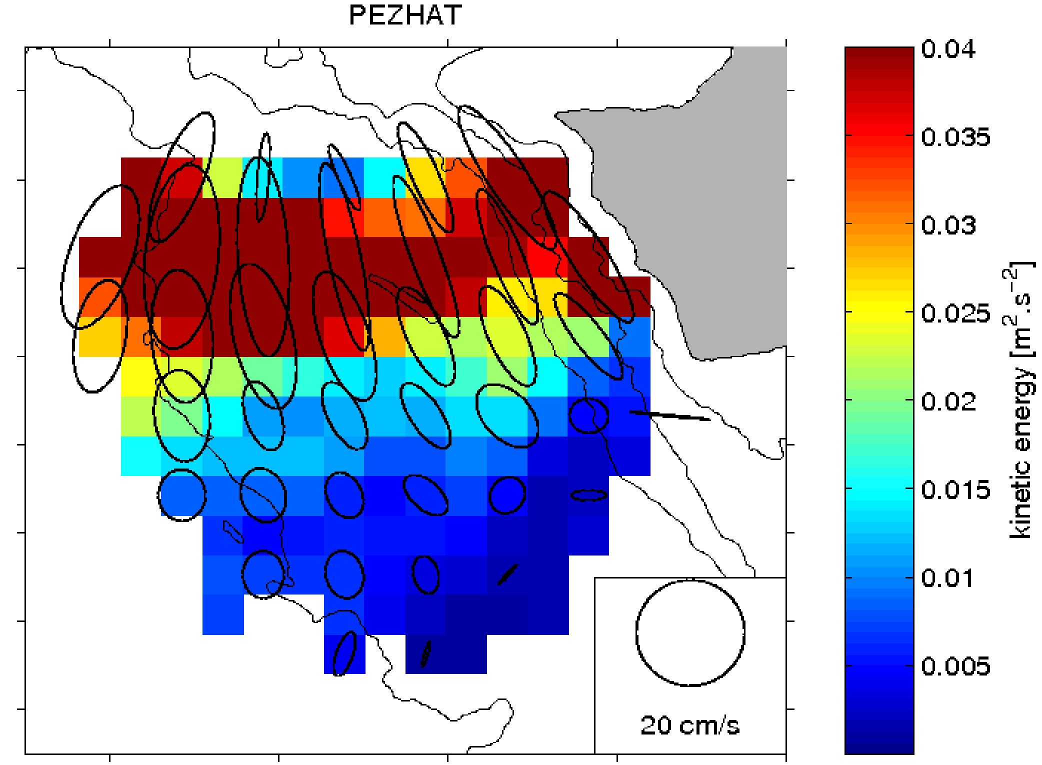
Surface kinetic energy

$$KE_{PEZ} / KE_{HFR} = 2.7$$

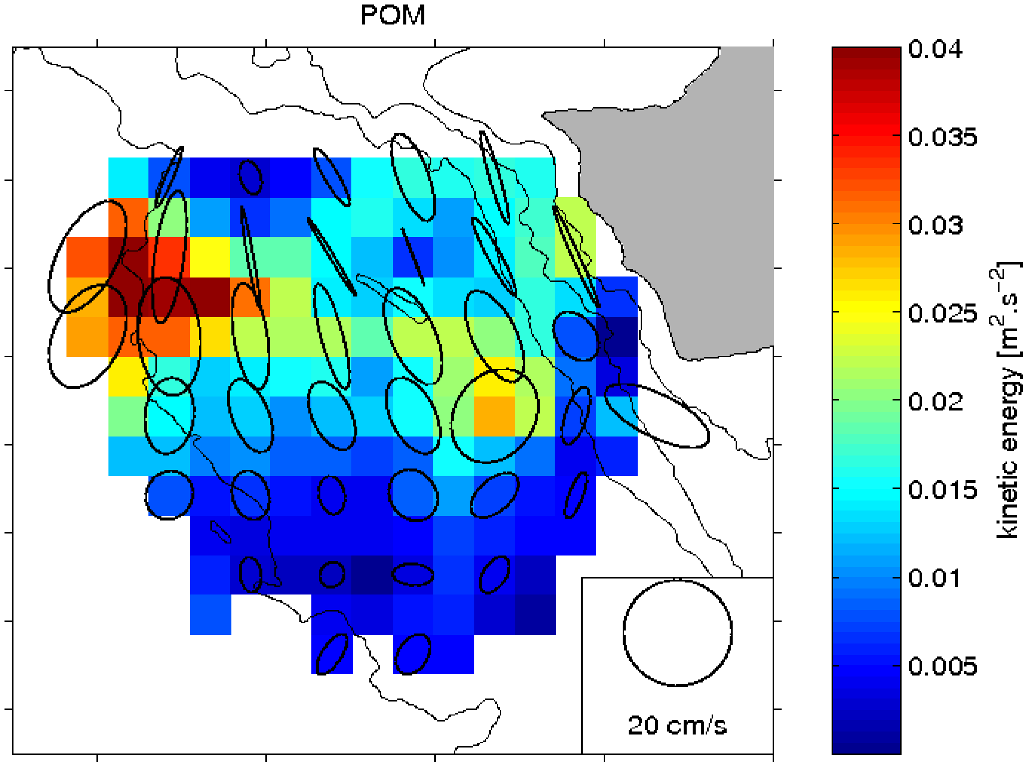
$$KE_{POM} / KE_{HFR} = 1.4$$



HFRs



PEZHAT

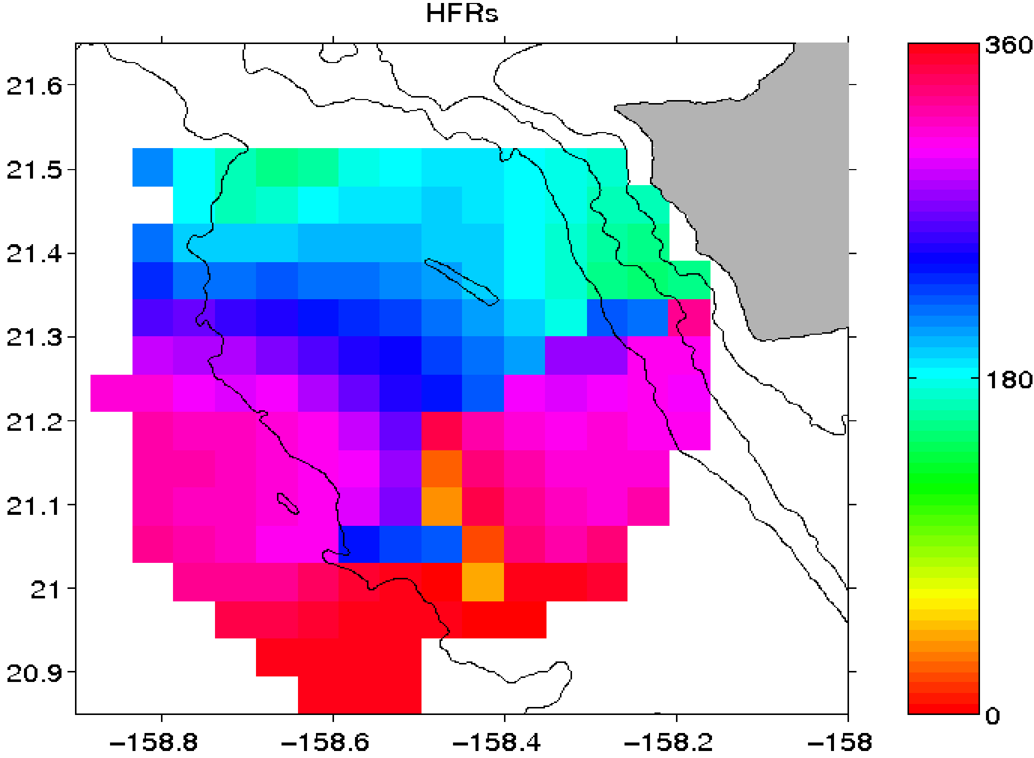


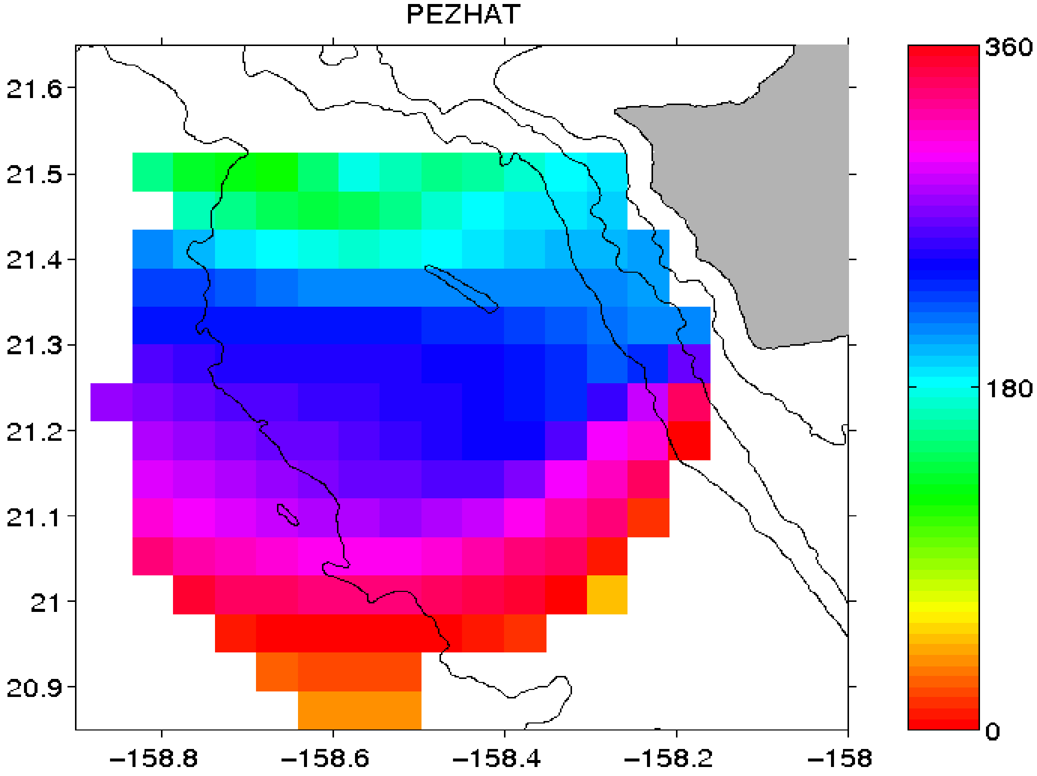
POM

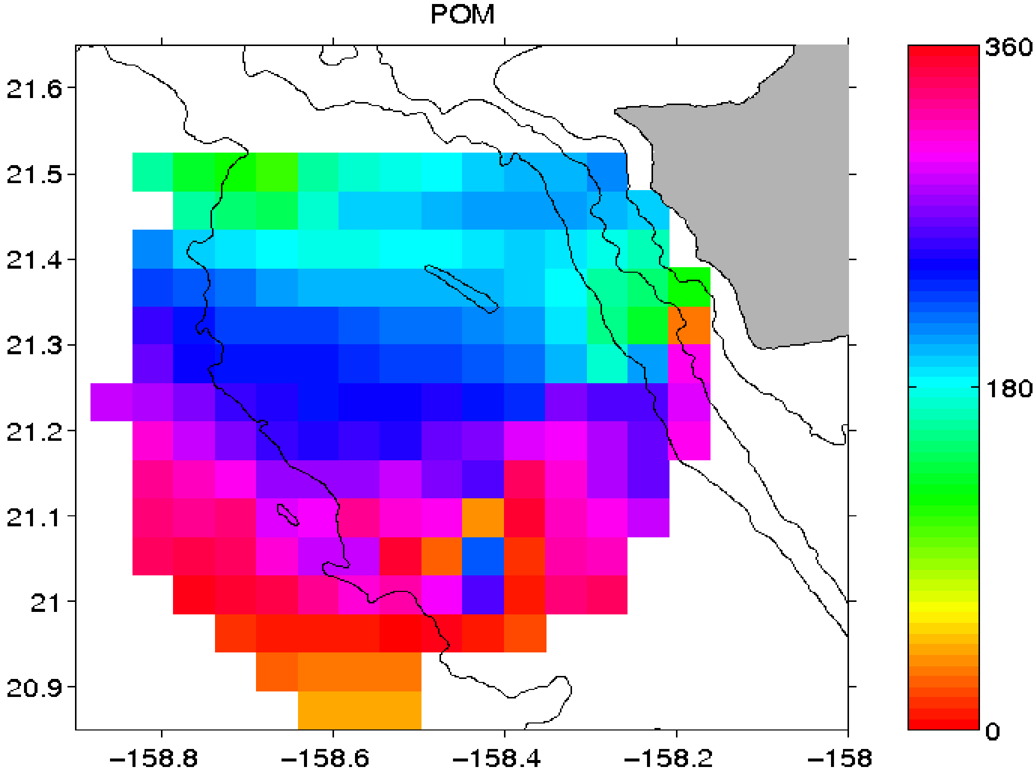
Surface phase

$$\lambda_{HFR} = 123\text{ km}$$

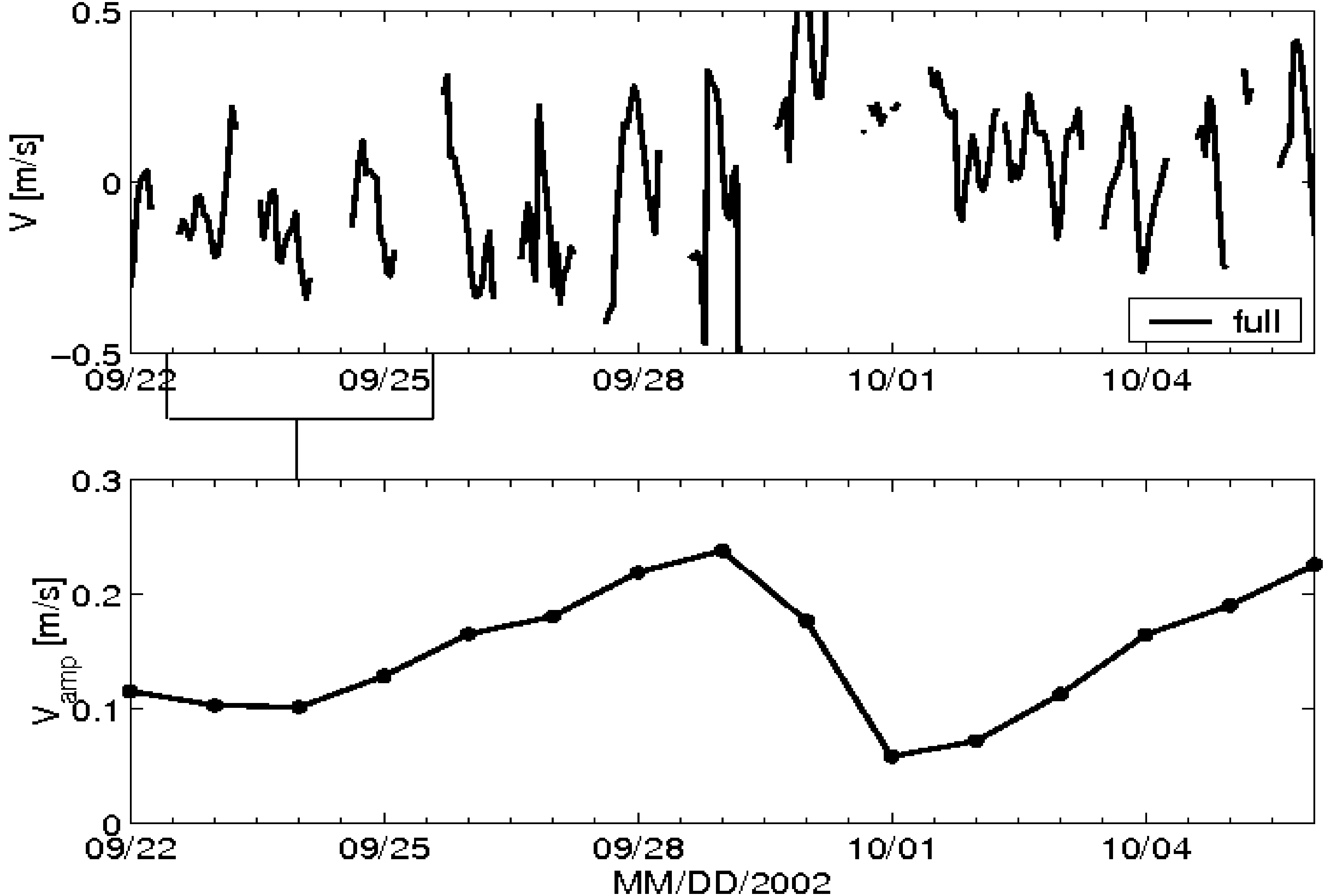
$$\lambda_{PEZ} = \lambda_{POM} = 105\text{ km}$$


HFRs

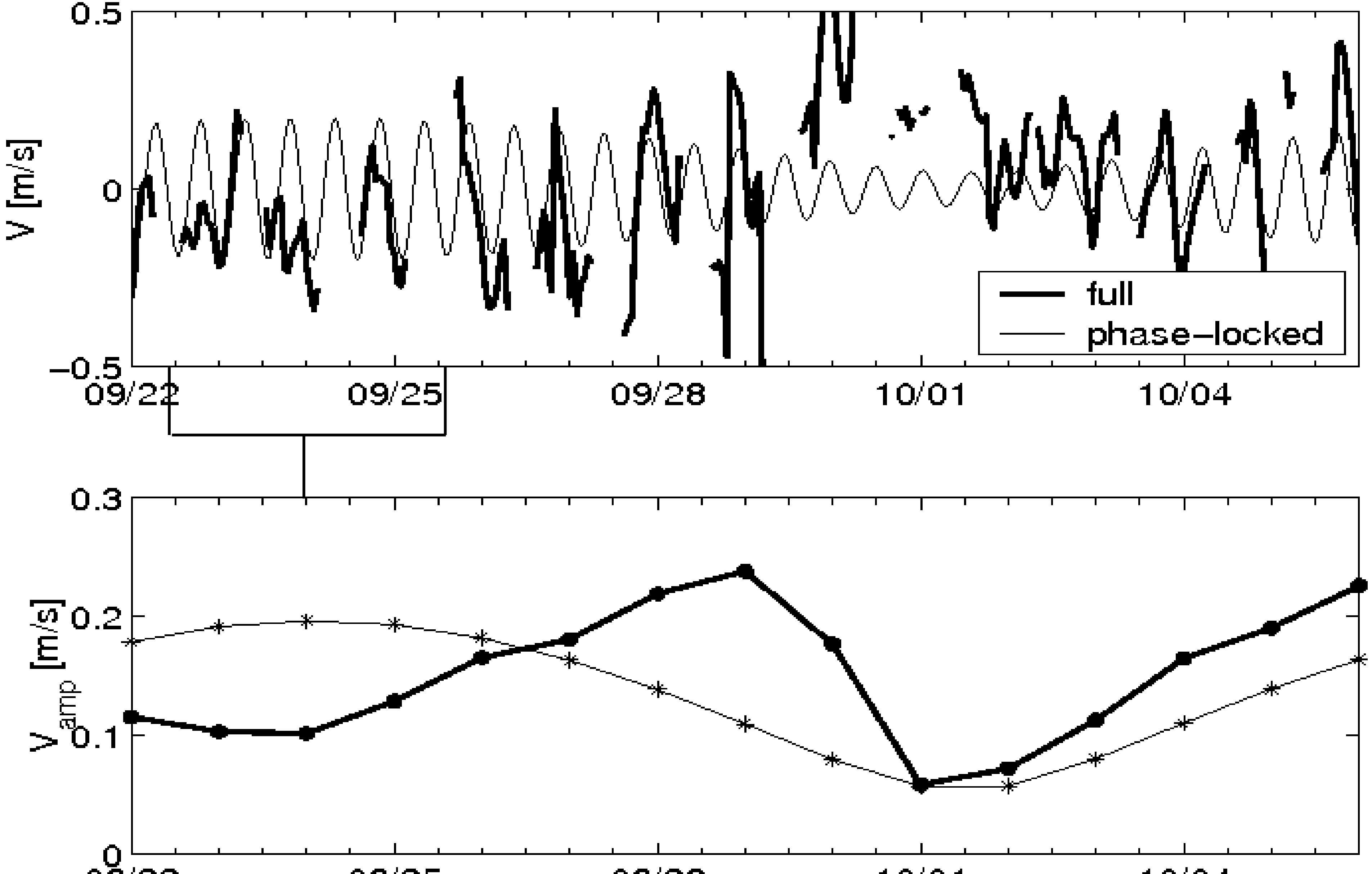

PEZHAT


POM

Complex demodulation: example

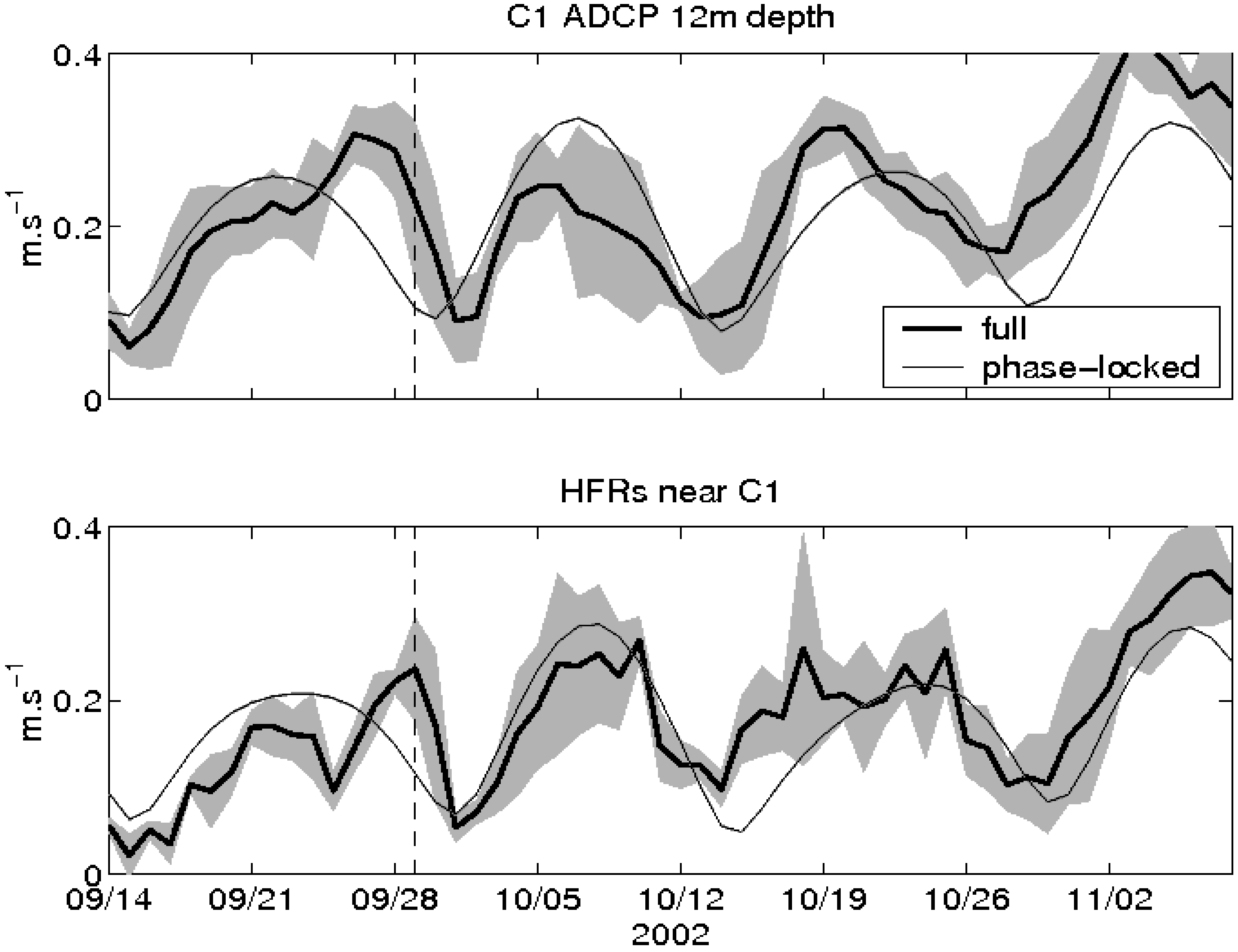


Complex demodulation: example



Temporal variations at C1

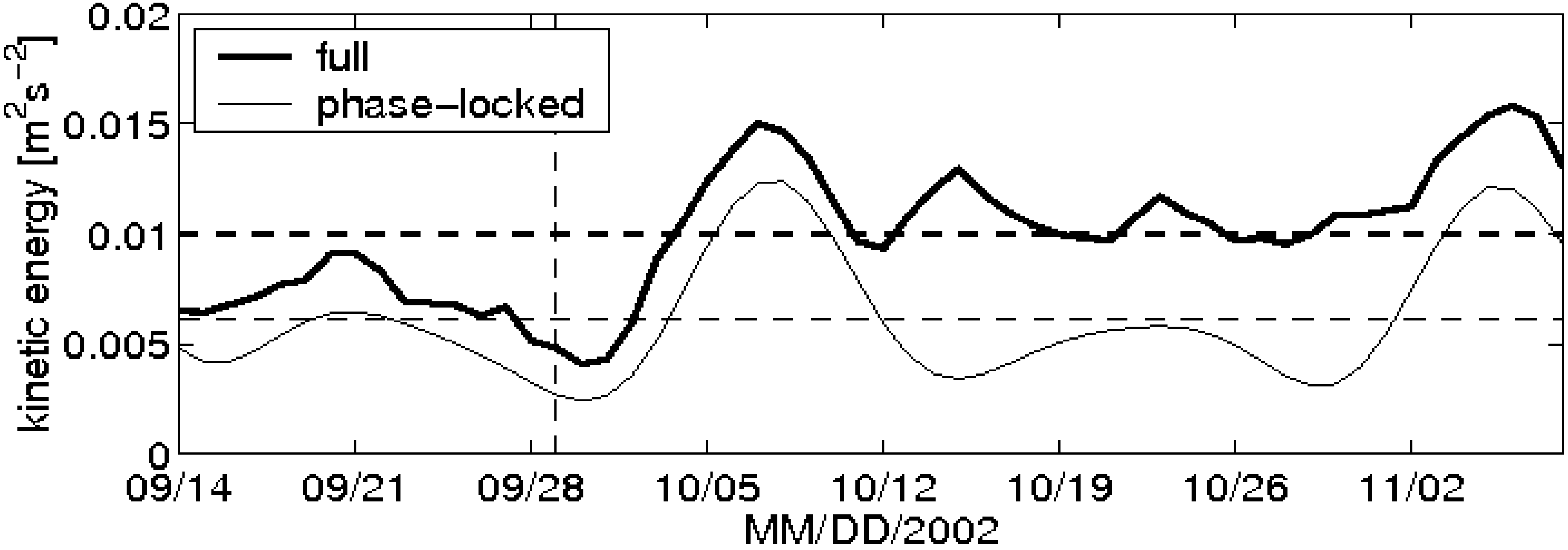
Major axis amplitude



Spatially-averaged kinetic energy evolution

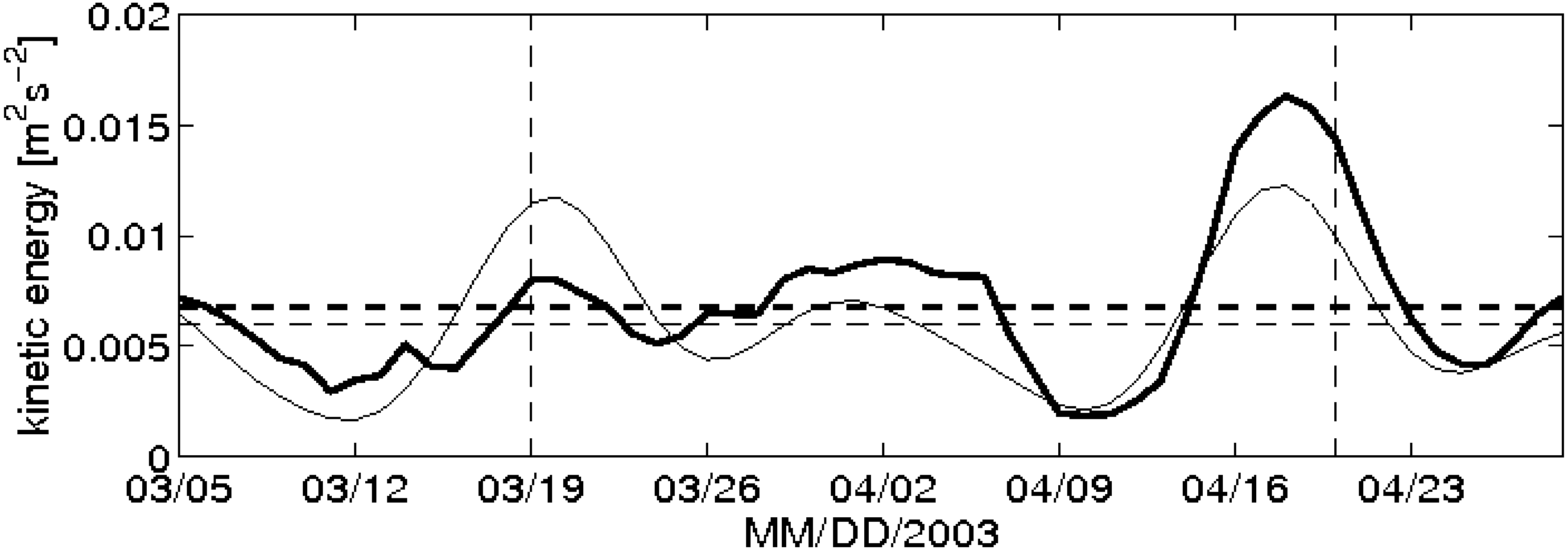
KE_{full}

KE_{lock}
=
1.6



KE_{full}

KE_{lock}
=
1.1

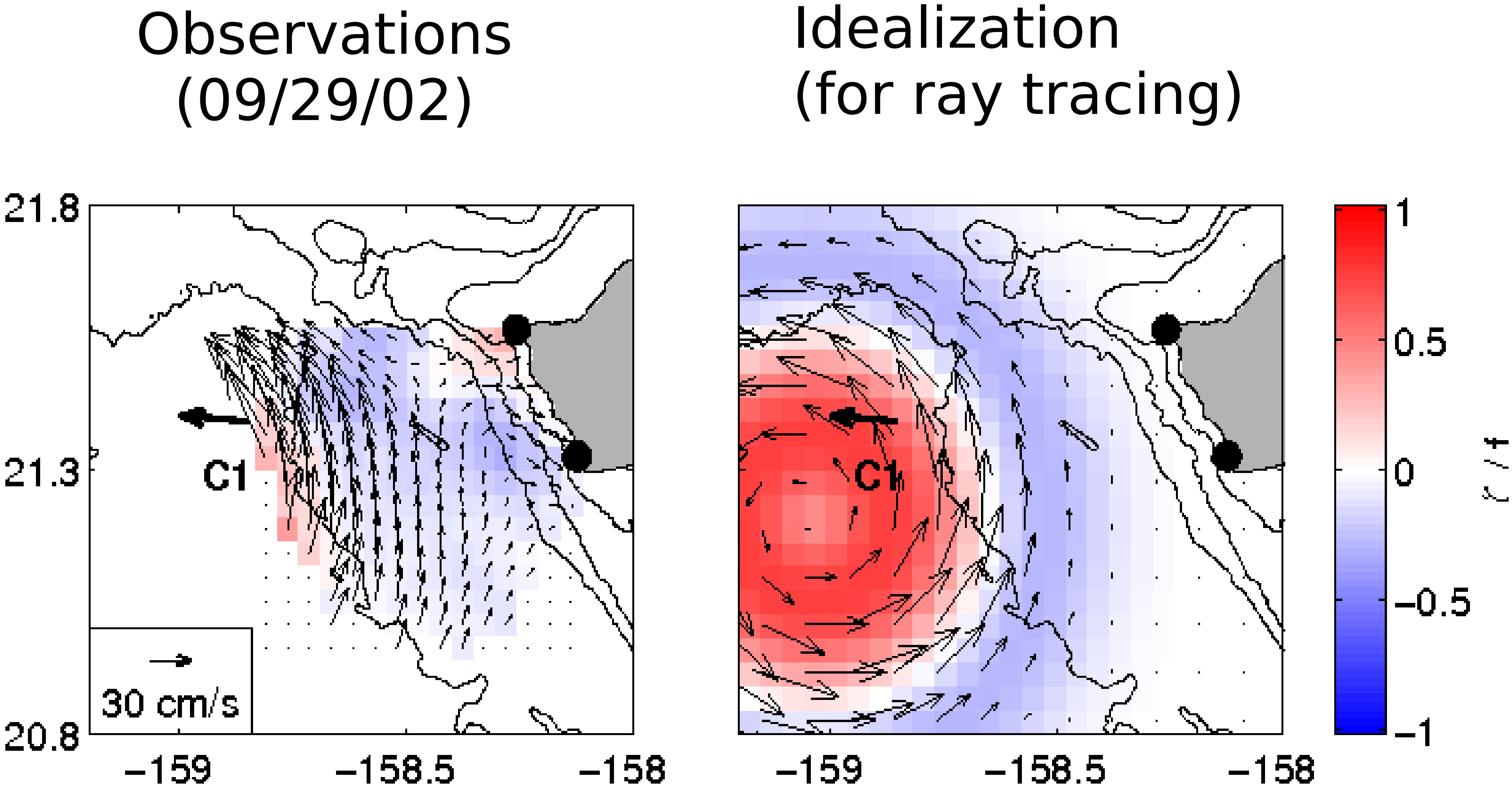


2. Mesoscale currents

a) Cyclone

b) Vorticity waves

a) Cyclone: horizontal structure

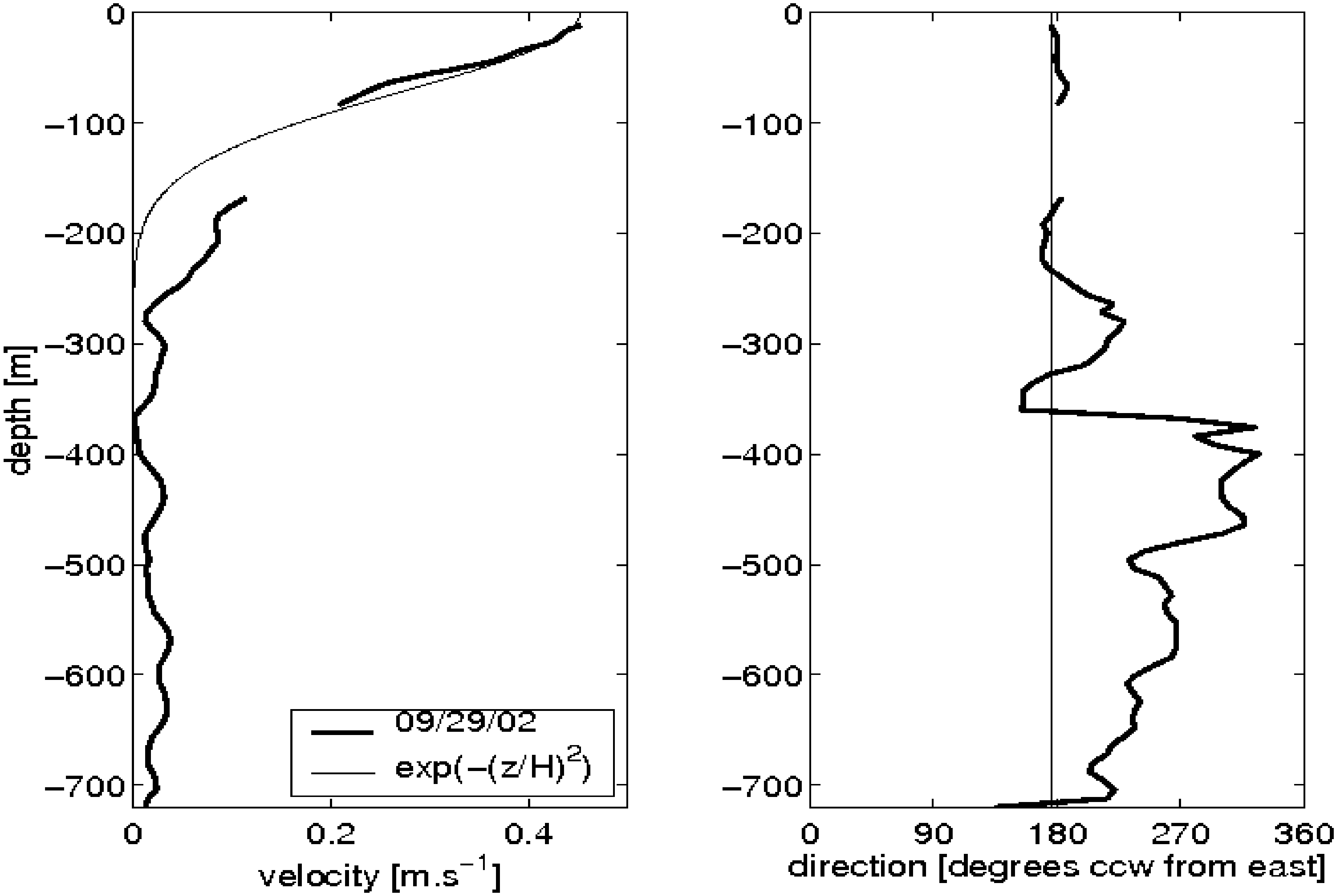


Observations
(09/29/02)

Idealization
(for ray tracing)

R ~ 30 km

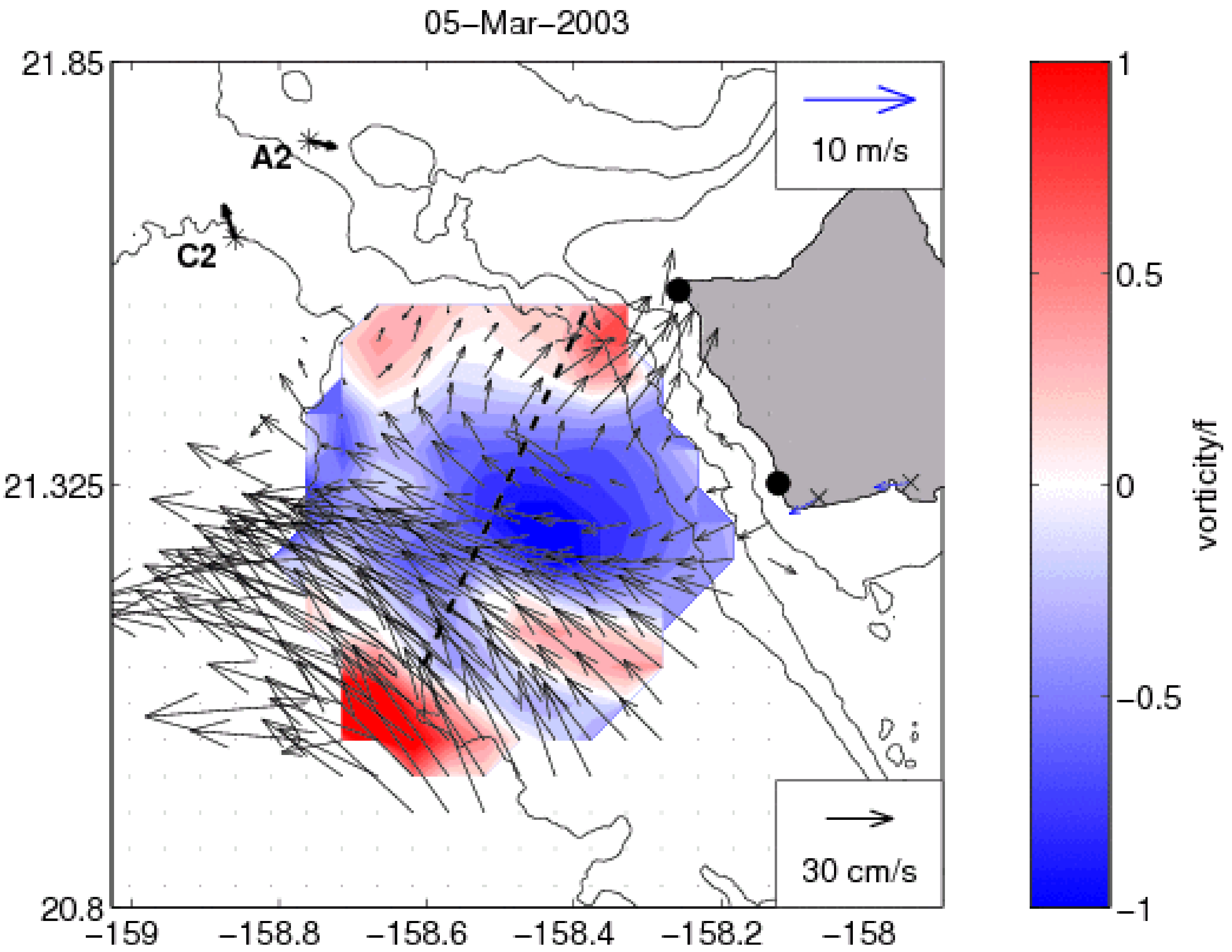
a) Cyclone: vertical structure



velocity

direction

b) Vorticity waves: propagation

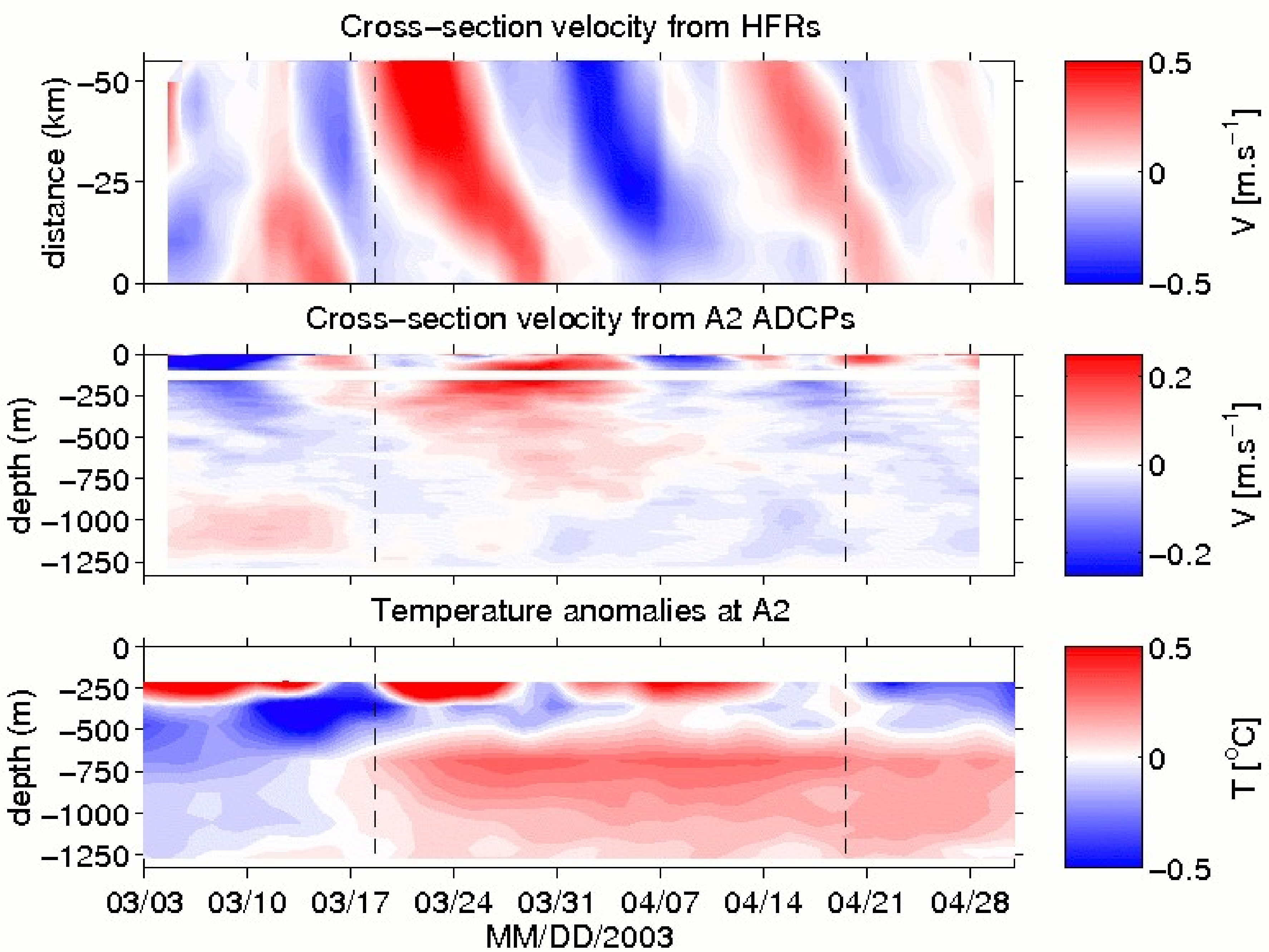


b) Vorticity waves: propagation

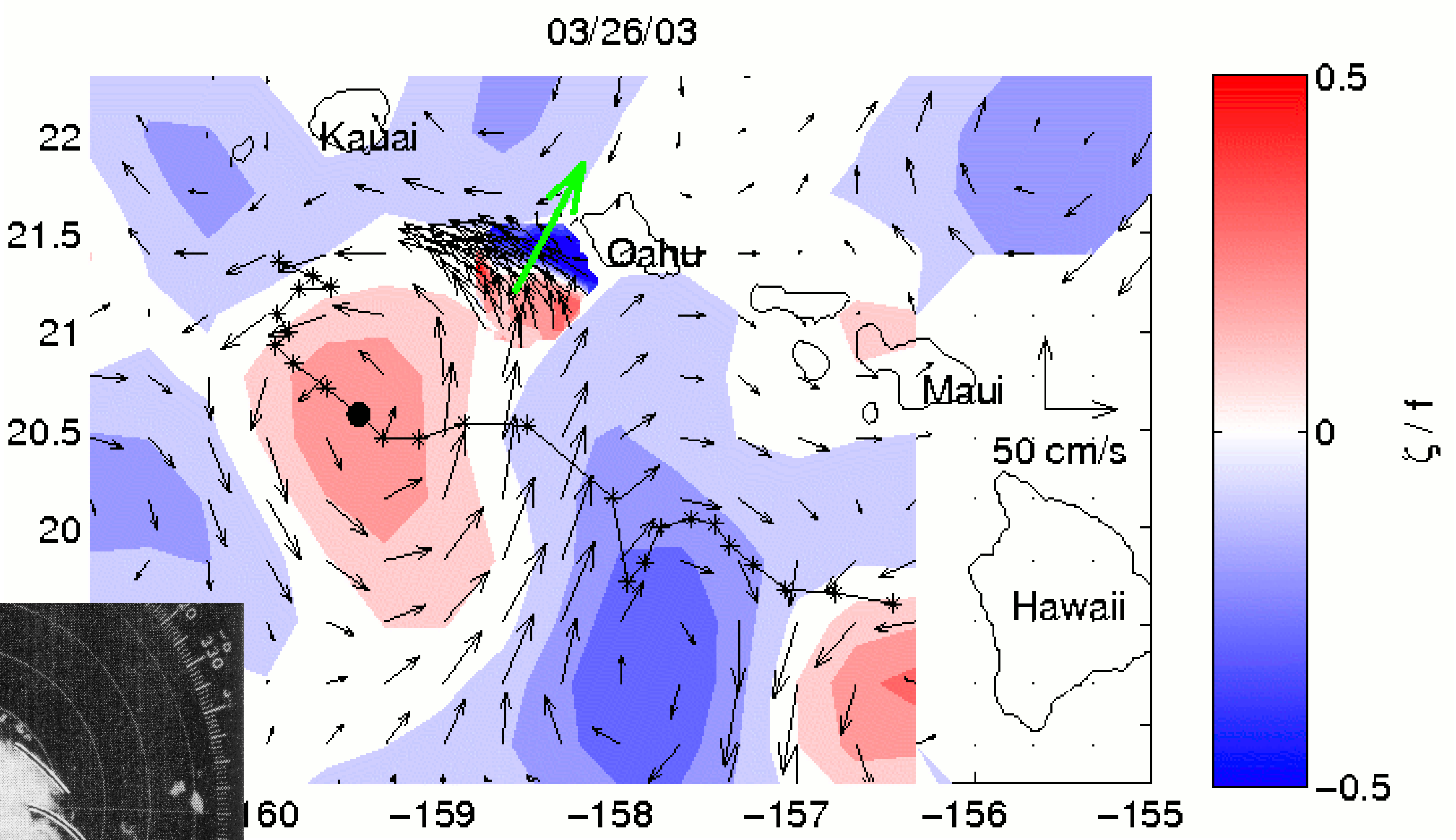
HFRs

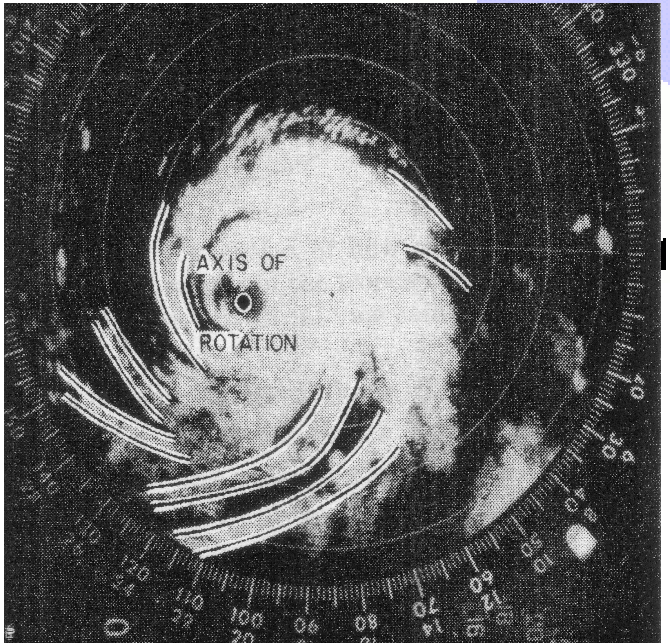
A2

A2



b) Vorticity wave = Vortex Rossby wave





Vortex Rossby waves in a hurricane
(MacDonald, 1966)

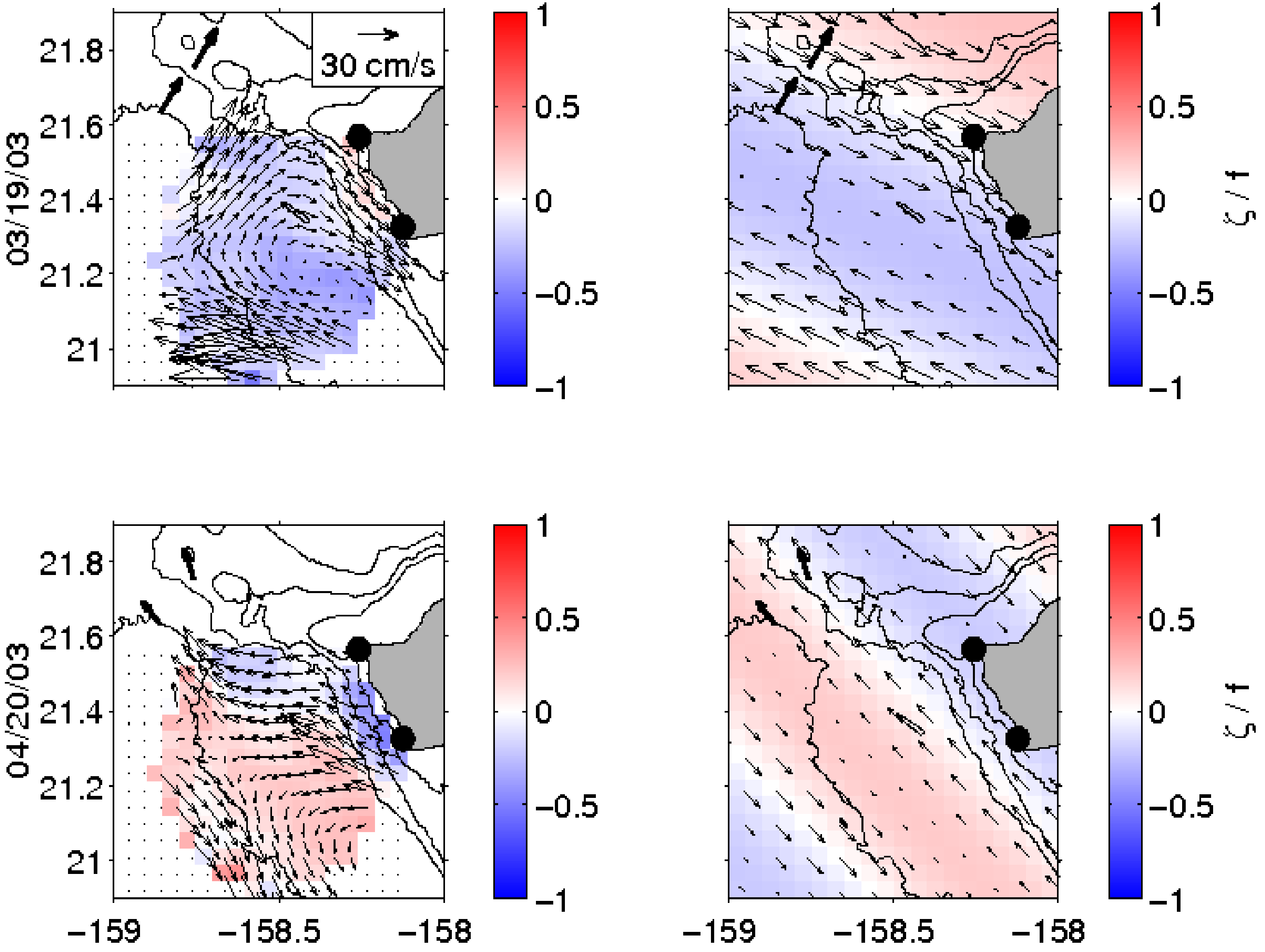
b) Vorticity waves: horizontal structure

Observations

Idealization
(for ray tracing)

03/19/03
λ ~ 150 km

04/20/03
λ ~ 100 km



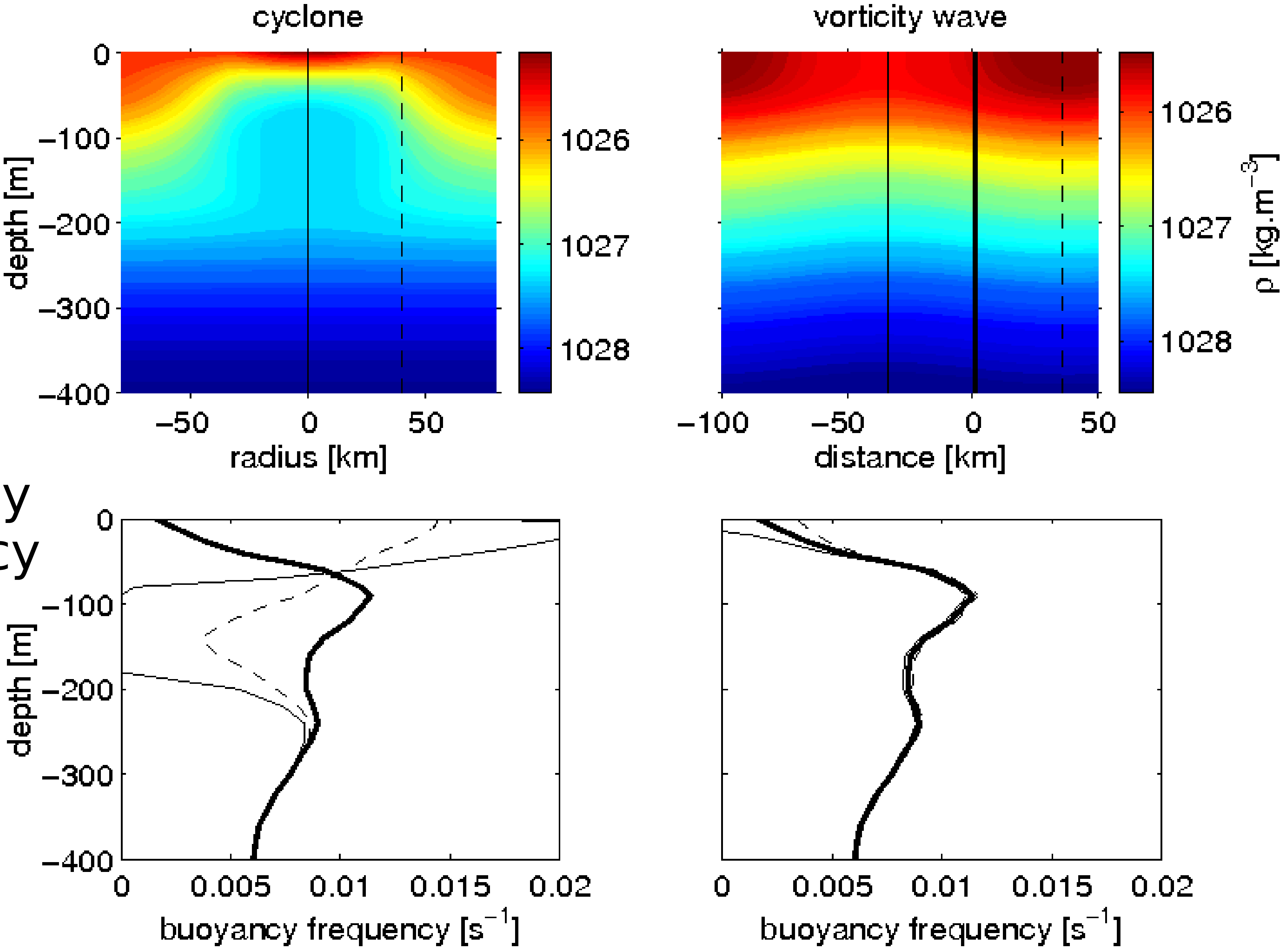
Idealized stratification

a) Cyclone

b) Vorticity wave

density

buoyancy frequency



3. Ray tracing

- wave packet: $\psi(\mathbf{x}, t) = a(\mathbf{x}, t)\ \exp^{i\theta(\mathbf{x},t)}$

- local wavenumber: $\mathbf{k}(\mathbf{x}, t) := \dfrac{\partial\theta}{\partial\mathbf{x}}$

- local frequency: $\omega(\mathbf{x}, t) := -\dfrac{\partial\theta}{\partial t}$

- local dispersion relation: $\omega = \Omega(\mathbf{k}; \mathbf{x}, t)$

- ray equations:

$$\frac{d\mathbf{x}}{dt} = \mathbf{C_g} = \frac{\partial\Omega}{\partial\mathbf{k}}$$

$$\frac{d\mathbf{k}}{dt} = \mathbf{r} = -\frac{\partial\Omega}{\partial\mathbf{x}}$$

Ray tracing

- Doppler shift:

$$\omega = \Omega(\mathbf{k}, \mathbf{x}, t) = \omega_0 + \mathbf{k}.\mathbf{U}$$

- intrinsic frequency:
  (Kunze, 1985)

$$\omega_0^2 = f_{eff}^2 + N_{eff}^2 \frac{k_h^2}{k_v^2}$$

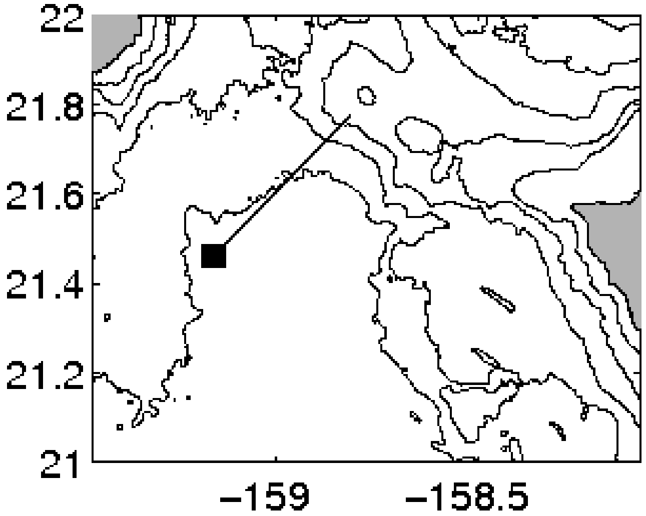
- effective Coriolis frequency*:*

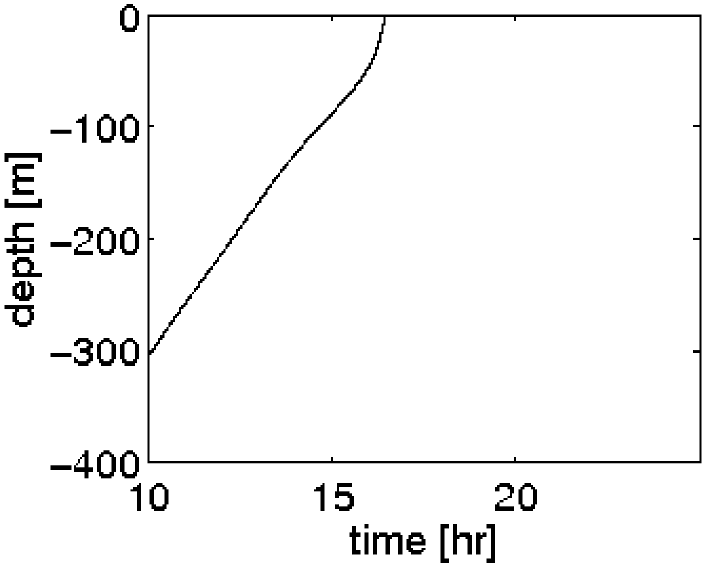
$$f_{eff} = f + \zeta/2 + \ldots$$

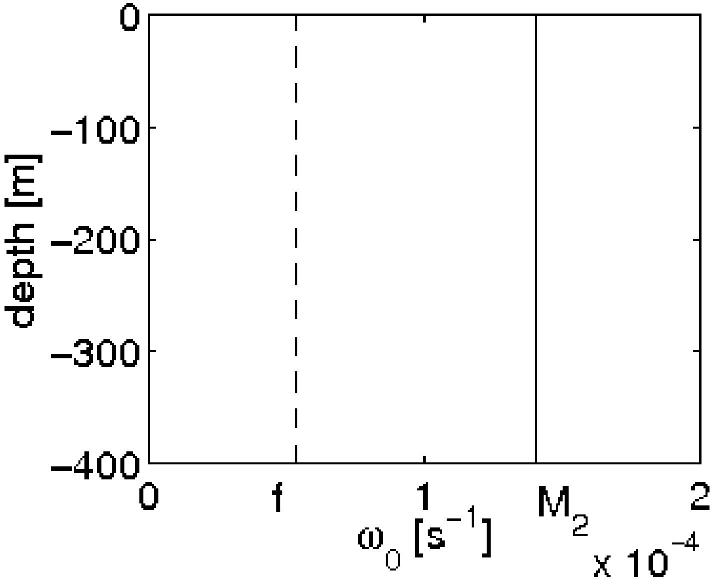
- effective buoyancy frequency:

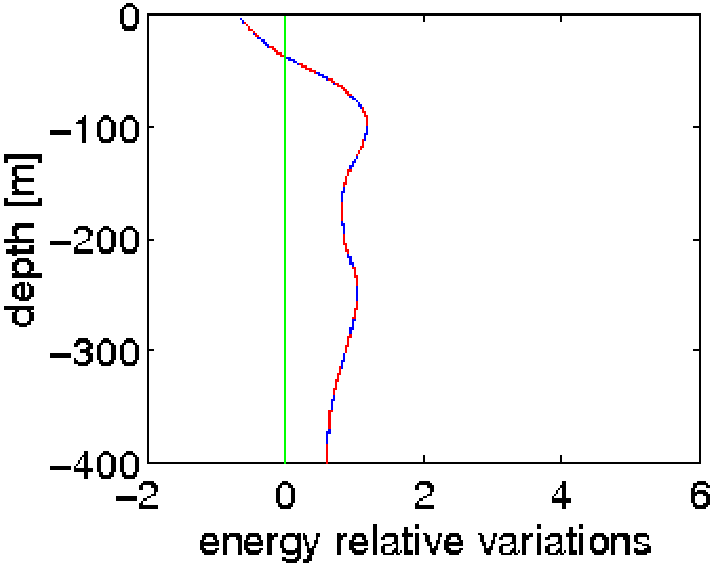
$$N^2_{\mathrm{eff}} = N^2 + \ldots$$

Example: without currents



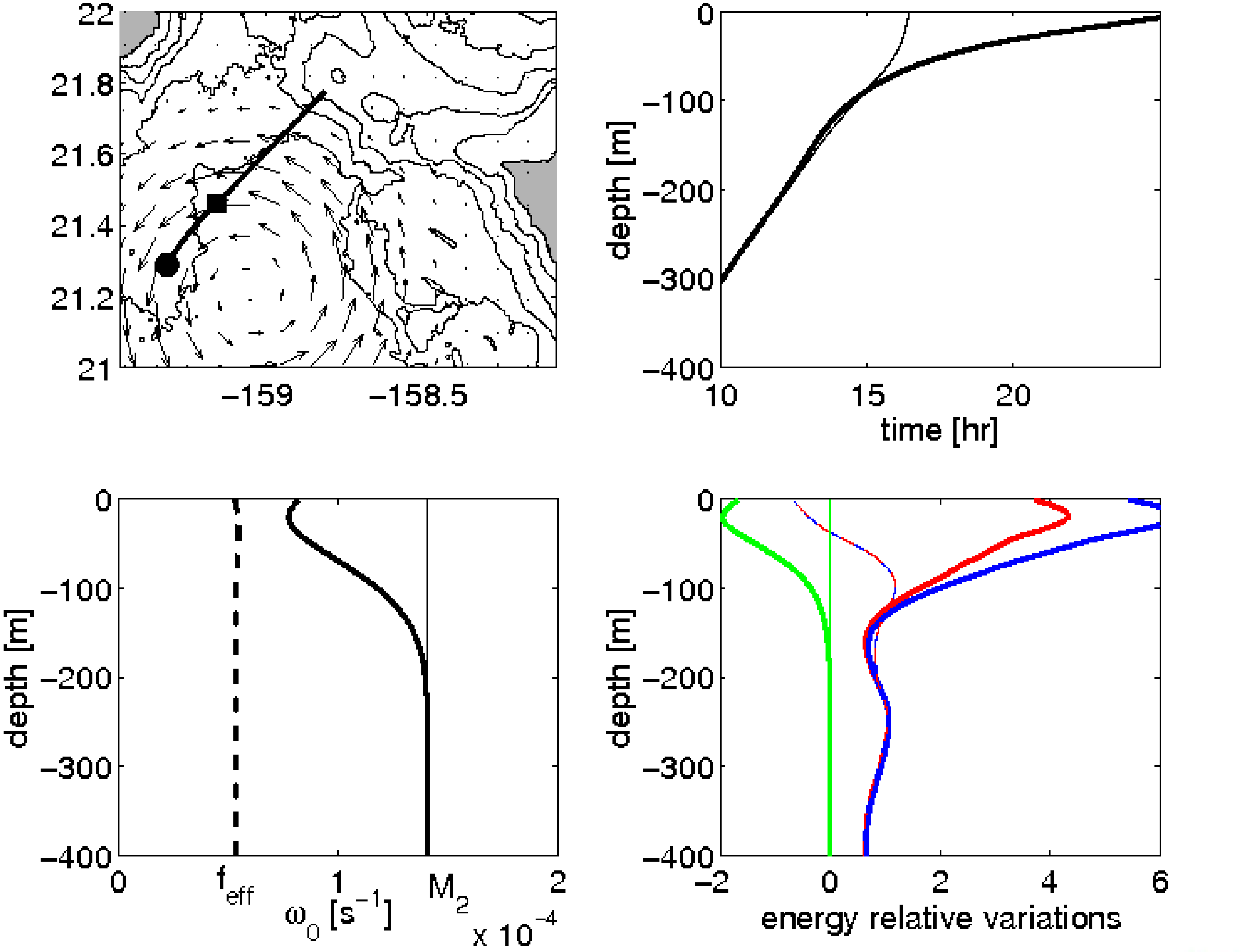






$$\frac{dE}{dt} = -E\nabla .C_g + \frac{E}{\omega_0}\frac{d\omega_0}{dt}$$

Example: with currents

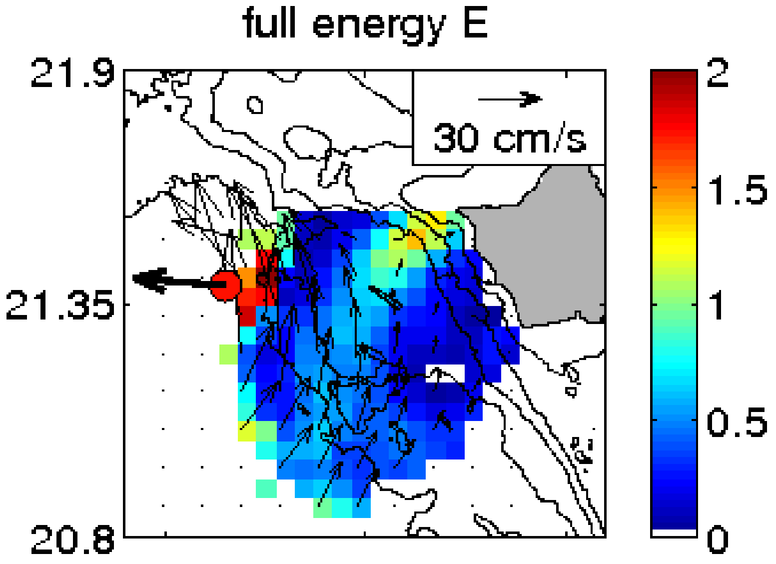


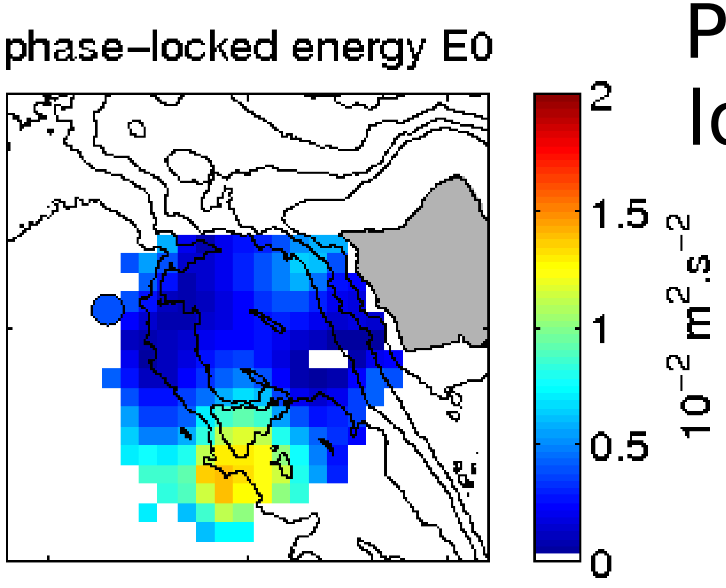
$$\frac{dE}{dt} = -E\nabla.C_g + \frac{E}{\omega_0}\frac{d\omega_0}{dt}$$

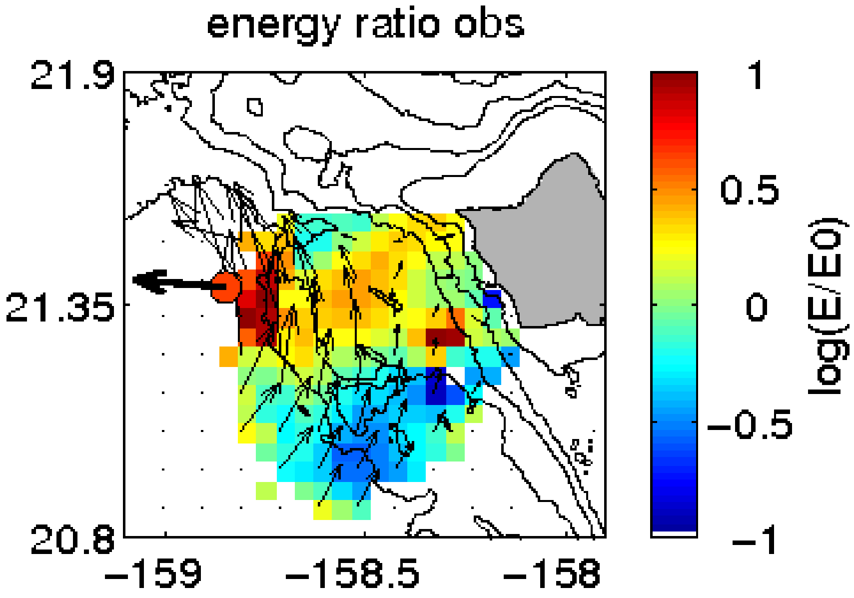
Surface kinetic energy

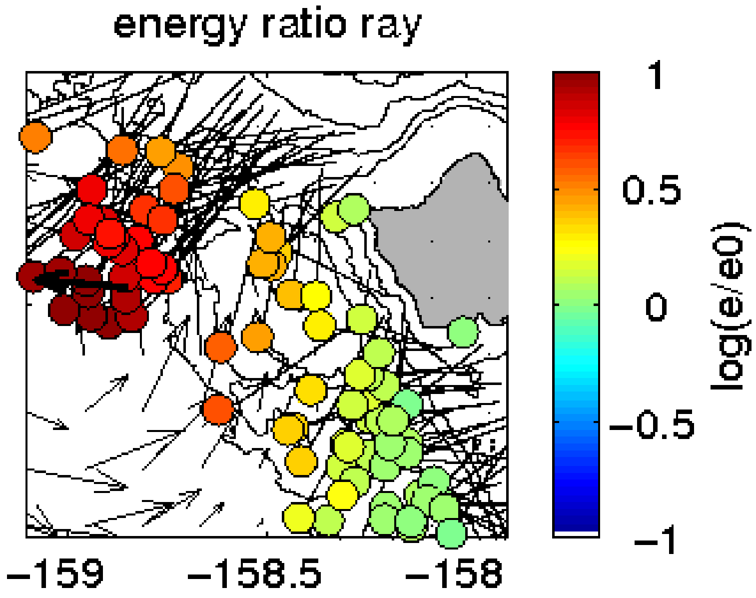
“full”

Phase-locked


full energy E


phase–locked energy E0

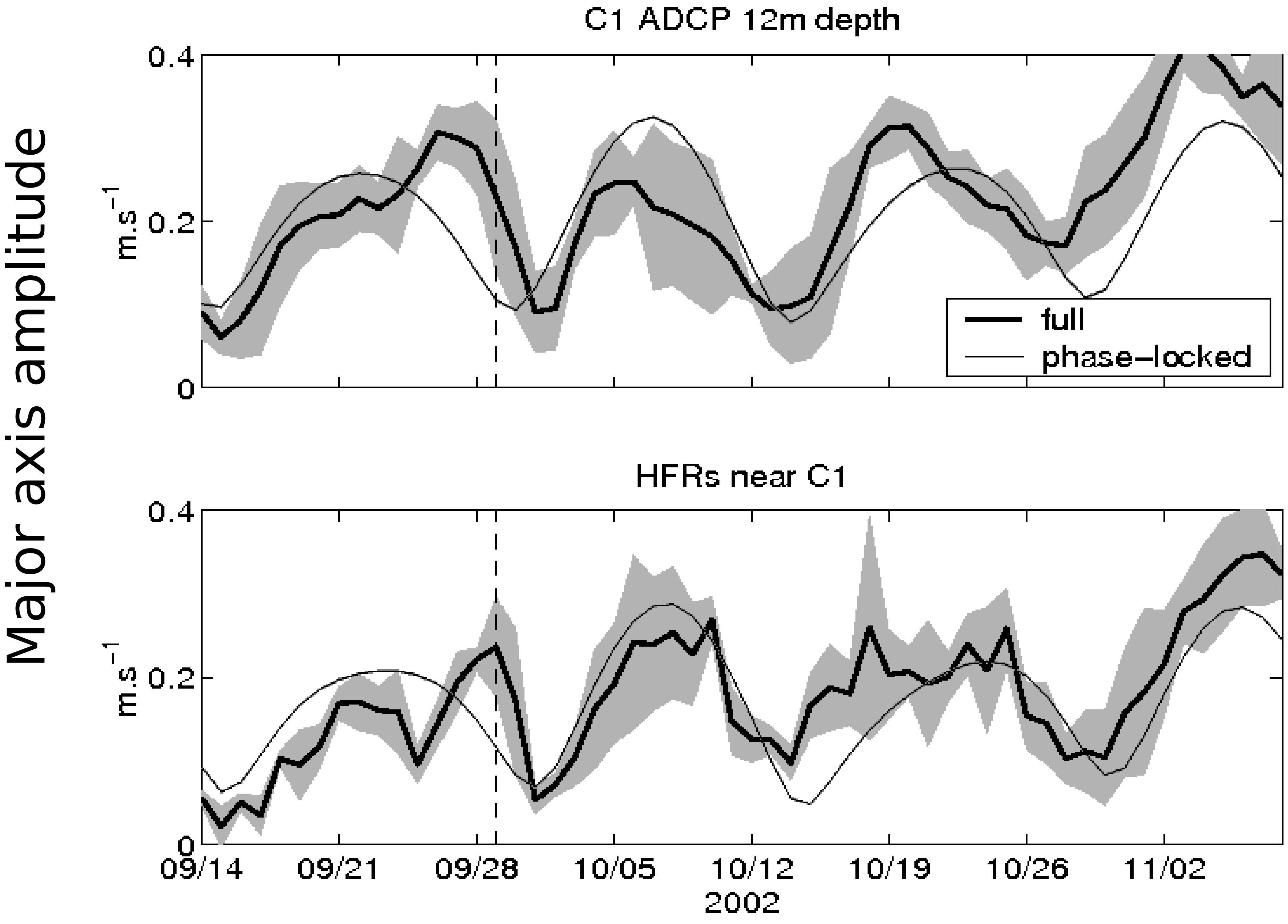

energy ratio obs


energy ratio ray

Full / phase-locked

Currents / No currents

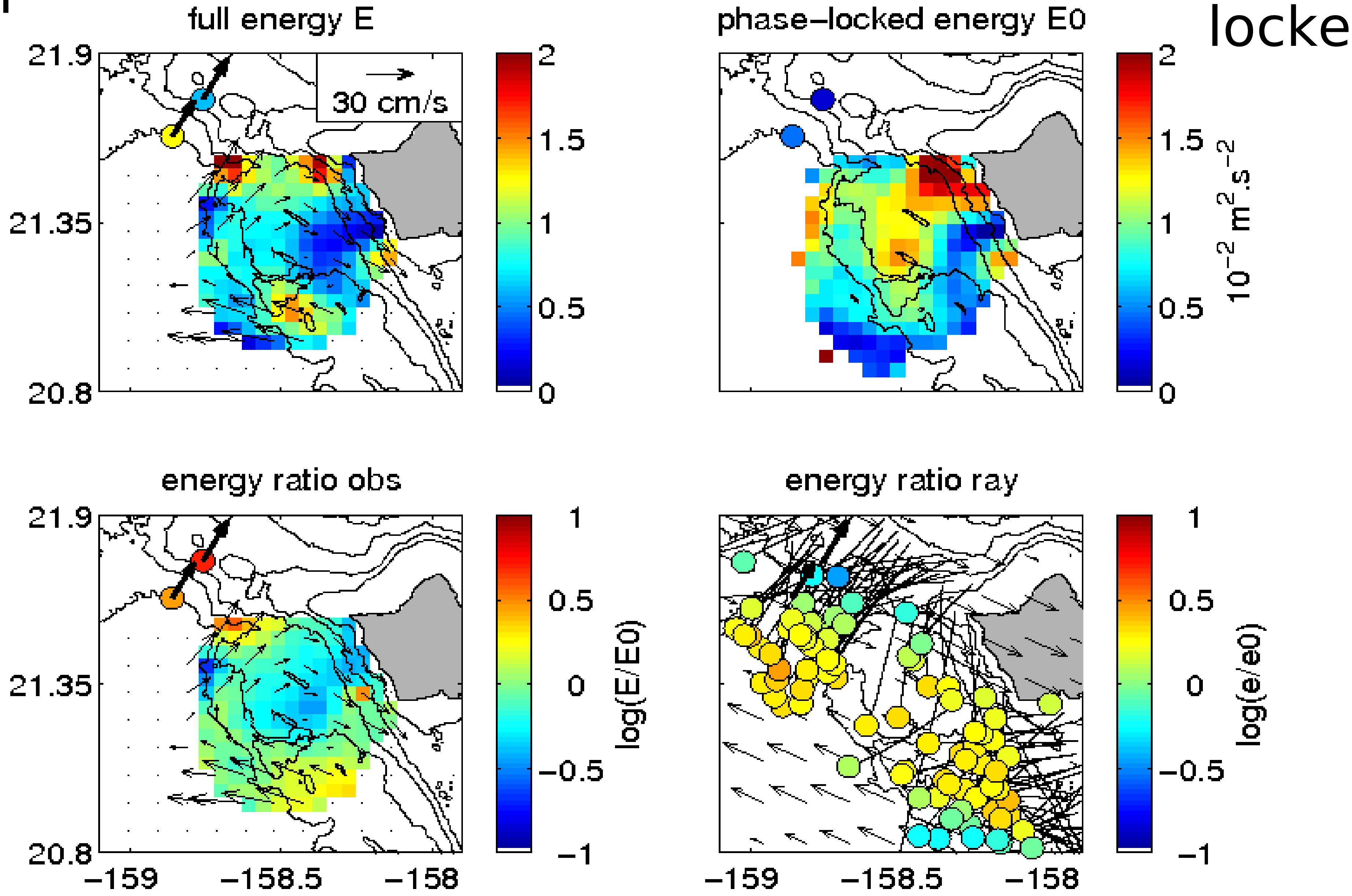
Temporal variations at C1



VRW on March 19, 2003

“full”

Phase-locked



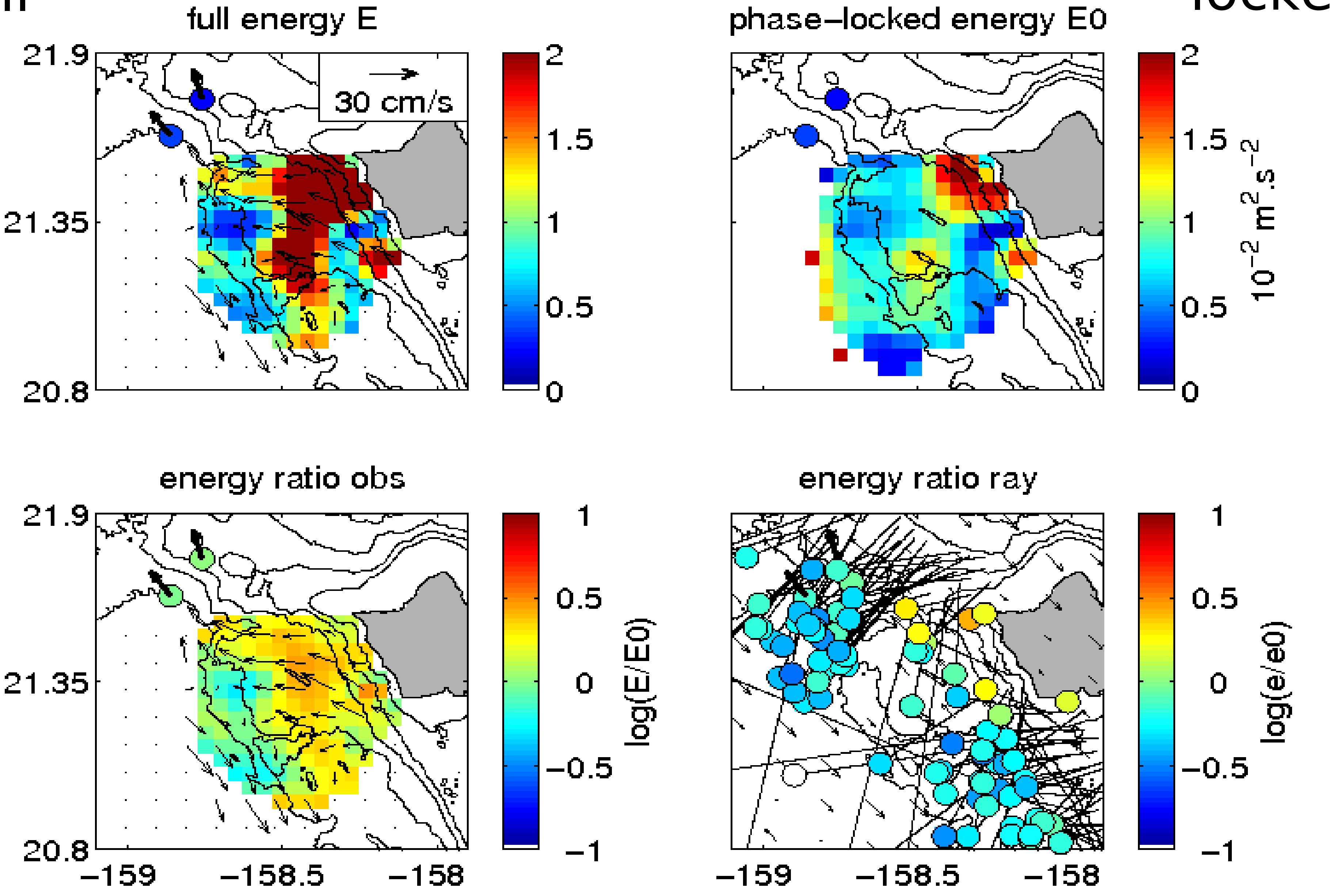
Full / phase-locked

Currents / No currents

VRW on April 20, 2003

“full”

Phase-locked



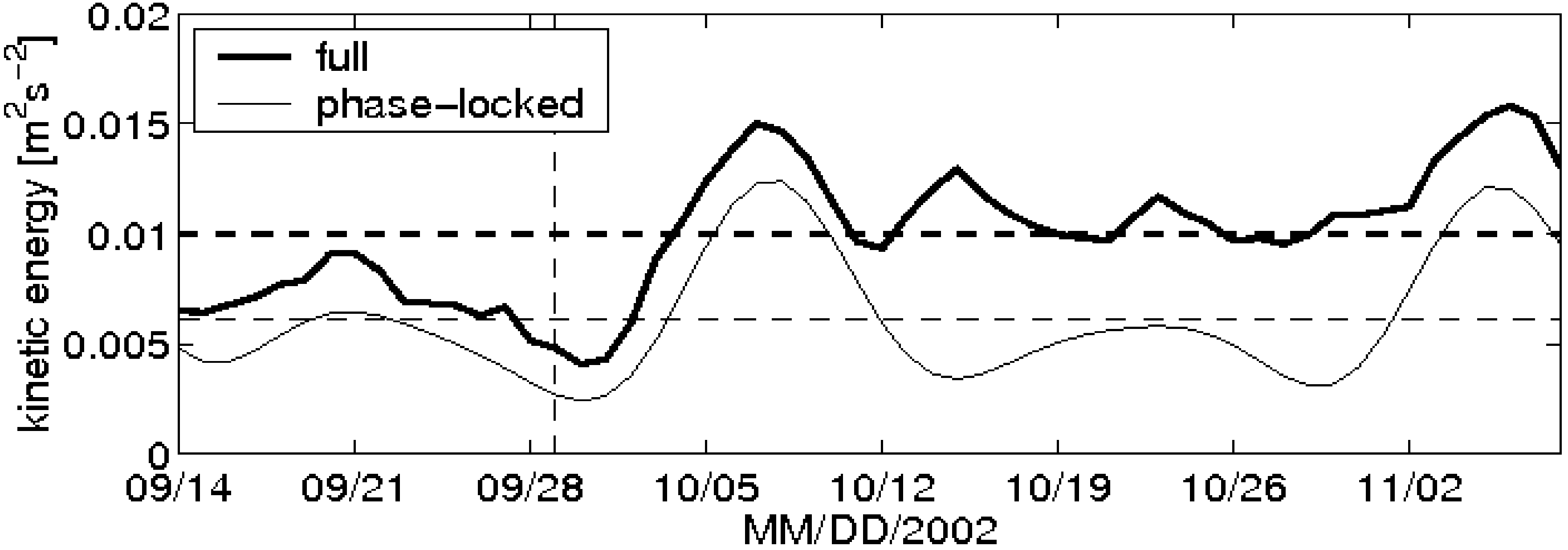
Full / phase-locked

Currents / No currents

Spatially-averaged kinetic energy evolution

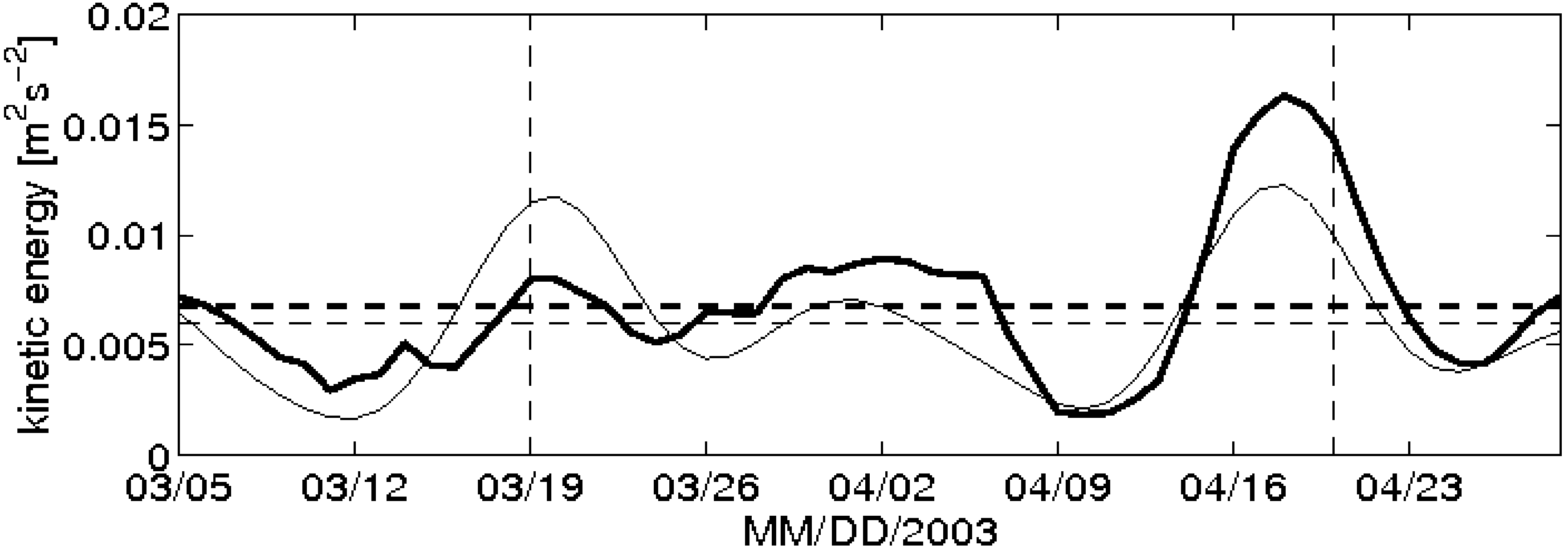
KE_{full}

KE_{lock}
=
1.6



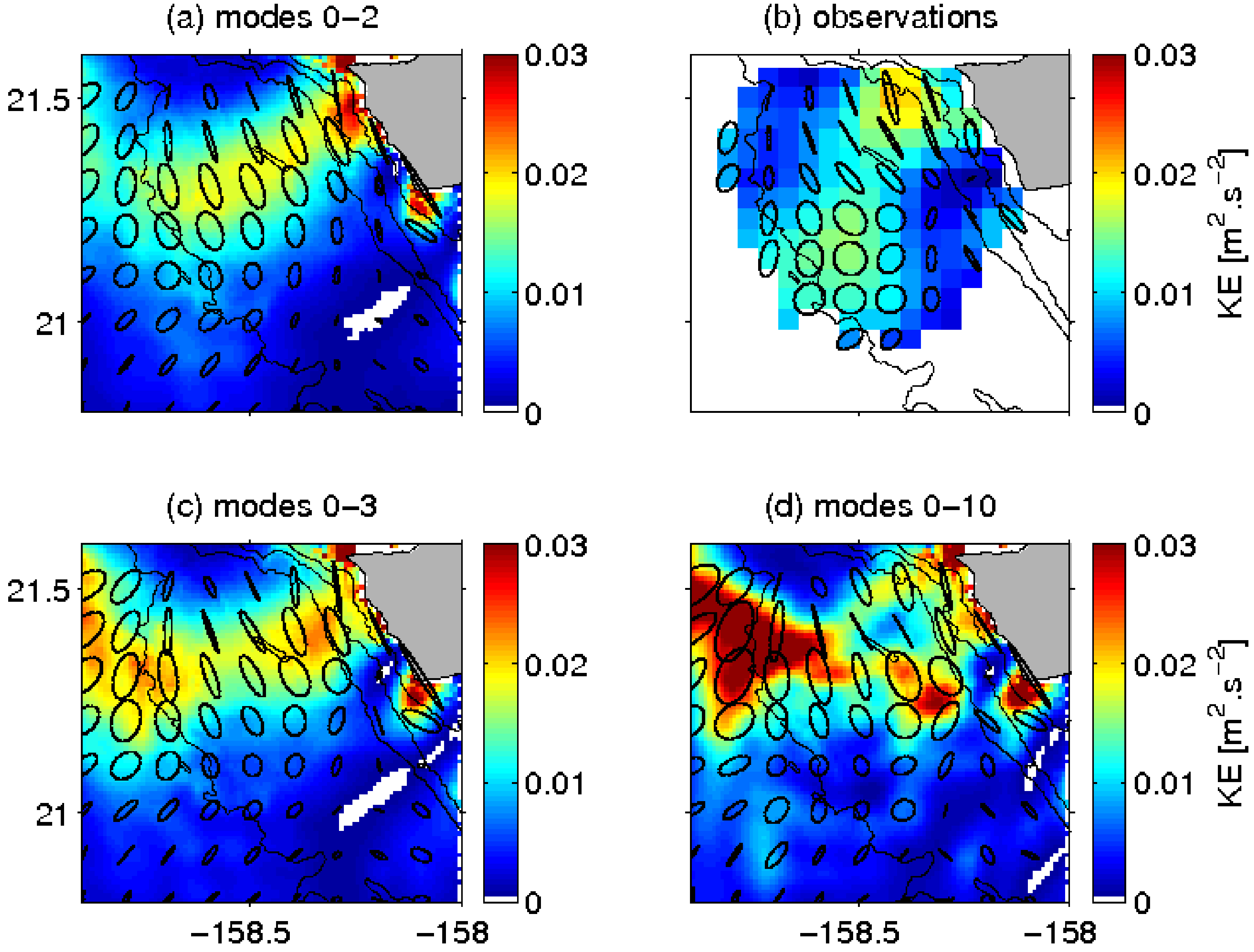
KE_{full}

KE_{lock}
=
1.1



Vertical modes filtering

Model (POM) | Observations



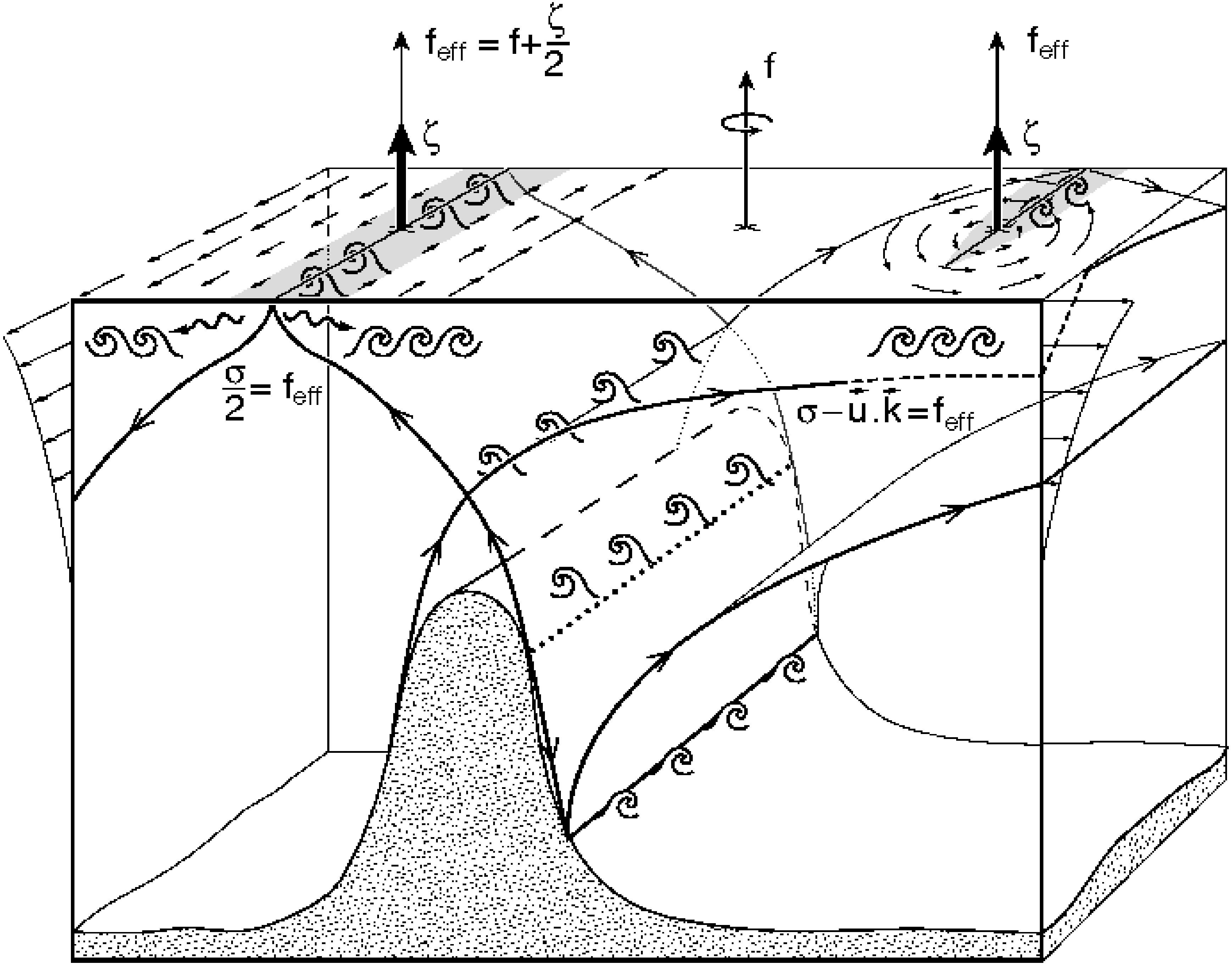
Conclusions

- Energetic mesoscale and submesoscale features (Ro~1 cyclones and anticyclones, fronts, vortex Rossby waves), not resolved by altimetry
- Mesoscale currents refract, Doppler-shift, and exchange energy with internal tidal beams at their first surface reflexion in the Kauai Channel
- Phase and amplitude modulations lead to low-pass filtering of vertical modes when harmonically analyzed over long periods of time

Implications for tidal energy budget

- barotropic energy loss:
  well constrained by assimilation of phase-locked observations (e.g. altimetry)
- baroclinic energy flux radiation:
  assimilation of phase-locked observations should be considered as lower bounds;
  assimilation of HF-radio phase-locked M2 currents into PEZHAT leads to ~10% decrease
- What about locally “dissipated” energy ?

Where should dissipation occur ?



Final remarks

- Less energy available for deep mixing ?
- Garrett and St Laurent (2002):

  “It may well be that the behavior of the surface layer is the most important oceanic component for the climate system. This behavior depends on mixing processes at the base of the surface layer. Internal waves generated by the wind, and possibly also by the tide, drive mixing at the base of the surface layer.”

Adrien
Desoria

Cédric
Chavanne

Julie
Deshayes

Joël
Benito

Good morning everyone. I would like to start by thanking all and each of you for serving on this committee.
Today, I am going to present my dissertation proposal :
“Dynamics of wave-Induced flow over very rough boundaries”

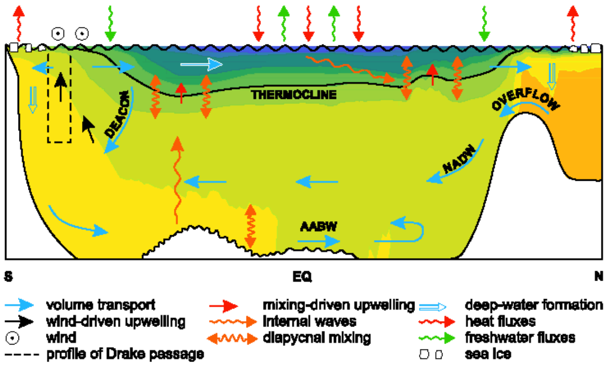
Why do internal tides matter ?

- often dominant energy signal
- ocean mixing
  - larval and pollutants dispersal
  - marine productivity
  - global climate

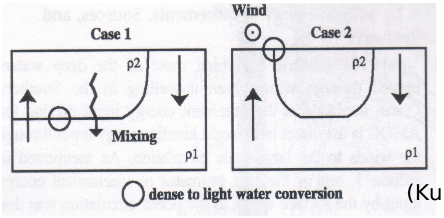
Relevance to climate predictions

- sensitivity of earth’s climate to vertical mixing in ocean
- sensitivity of ocean circulation to geography of mixing
- physically-based parameterization of diapycnal mixing in OGCMs

Meridional Overturning Circulation



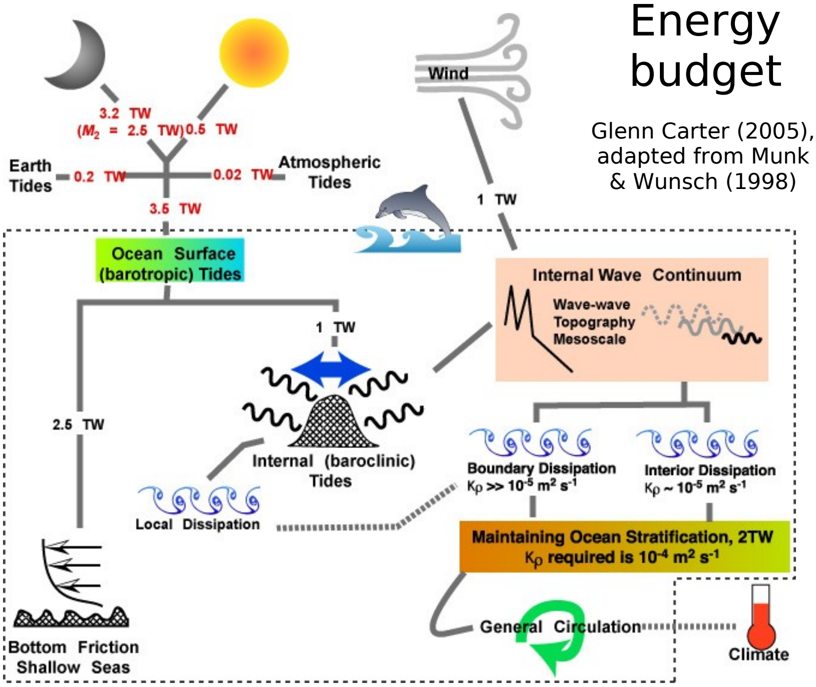
Driving Mechanisms



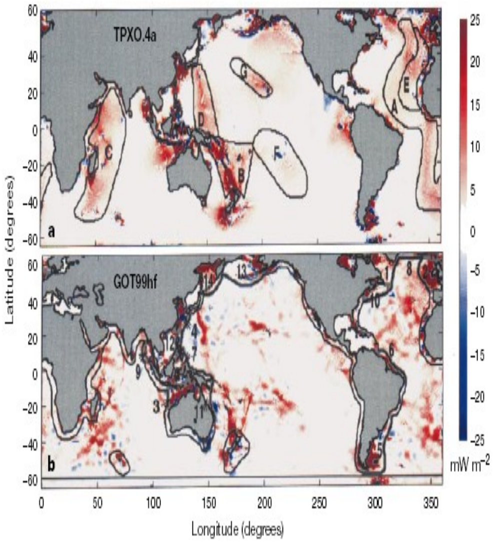
(Kuhlbrodt et al., 2007)

Energy budget

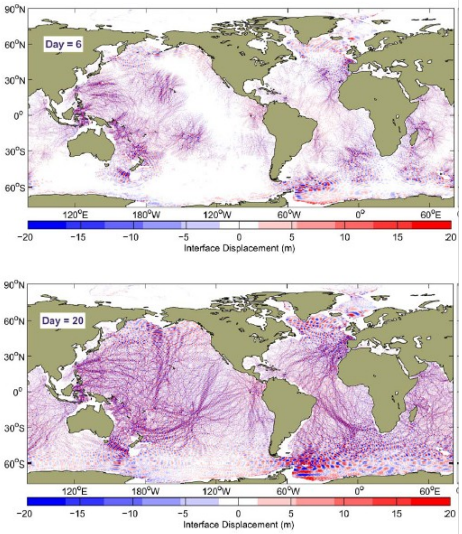
Glenn Carter (2005), adapted from Munk & Wunsch (1998)



Tidal dissipation and radiation



(Egbert & Ray, 2000)



(Simmons et al., 2004)

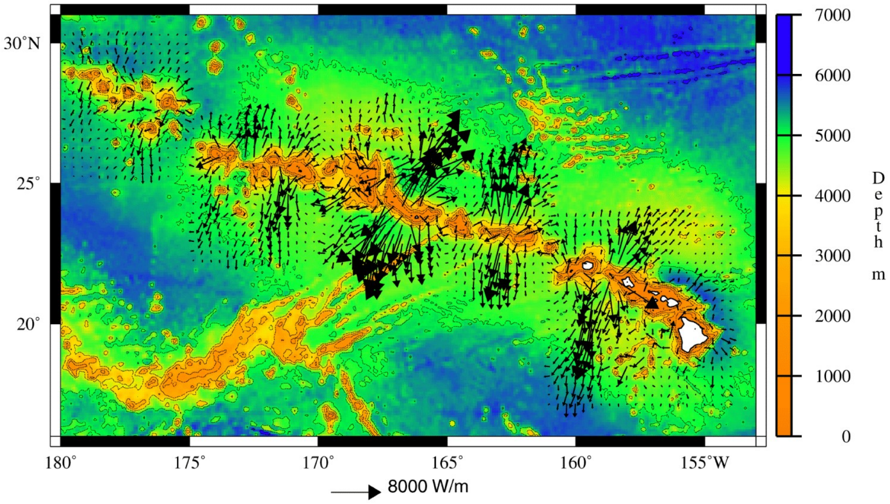
Hawaiian Ocean Mixing Experiment (HOME)

- determine if mid-ocean sites such as Hawaii are significant contributors to global mixing
- quantify tidal energy budget for an isolated abrupt topographic feature
- determine the principal mechanisms which transfer energy from large scale flows to turbulent motions

→ generalize results to global ocean

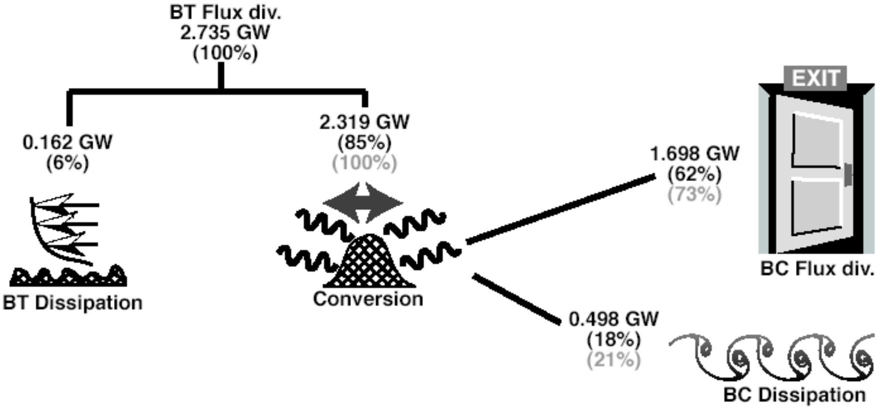
→ improve parameterisation of mixing in numerical models

Depth-integrated M_2 baroclinic energy fluxes



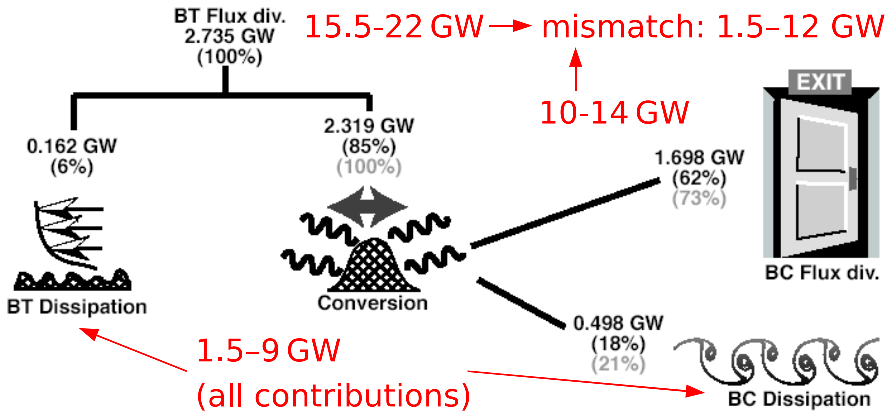
(Merrifield et al., 2001)

M_2 Energy Budget for Hawai`i



Carter et al. (2007): model budget for Kaua`i-Maui

M_2 Energy Budget for Hawai`i



15.5-22 GW → mismatch: 1.5–12 GW

10-14 GW

1.5–9 GW
(all contributions)

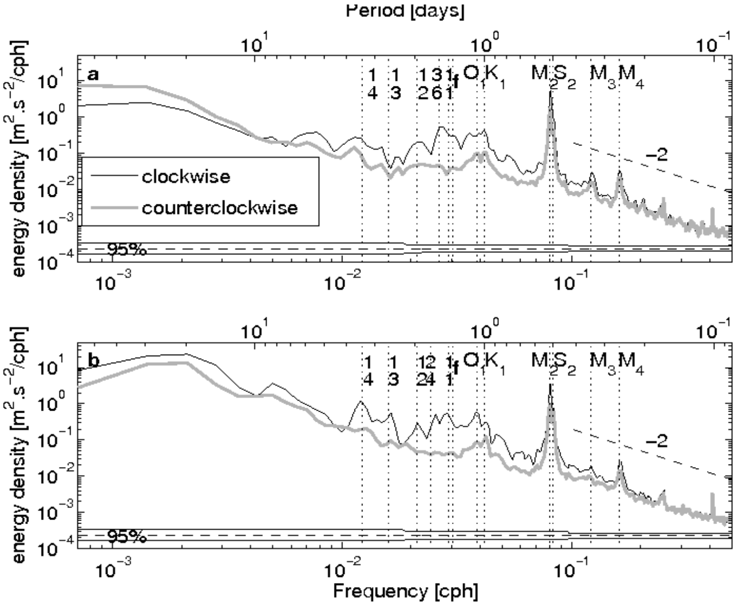
Carter et al. (2007): model budget for Kaua`i-Maui

Modeled and observed for entire Hawaiian Ridge

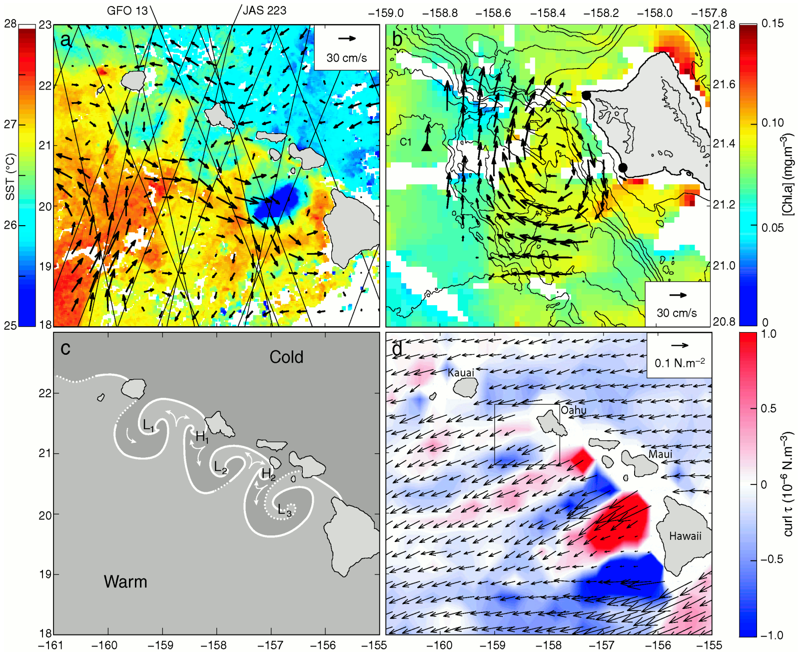
Surface currents power spectra

Fall 2002

Spring 2003



Mesoscale circulation



Questions

- What is the impact of mesoscale currents on internal tide propagation ?
- What are the implications for tidal energy budgets, and abyssal mixing ?

I will start this presentation with a little bit of background and go over the motivations of this research. I will then outline the objectives of my thesis and the specific questions I will attempt to answer.
I will present the methods we used to achieve these goals in the following section and give you a glimpse of some of my results.
Finally we will see where I am going from this…

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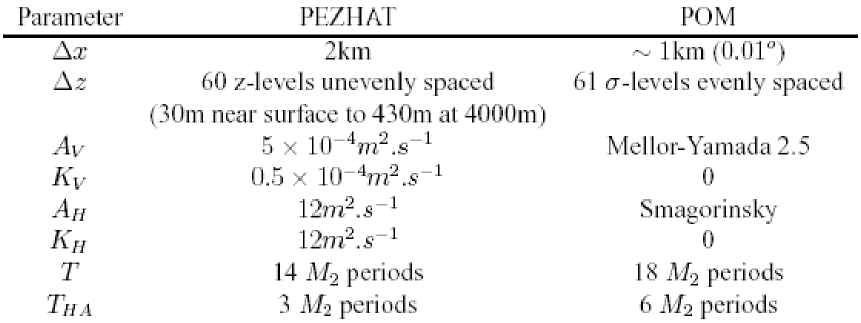
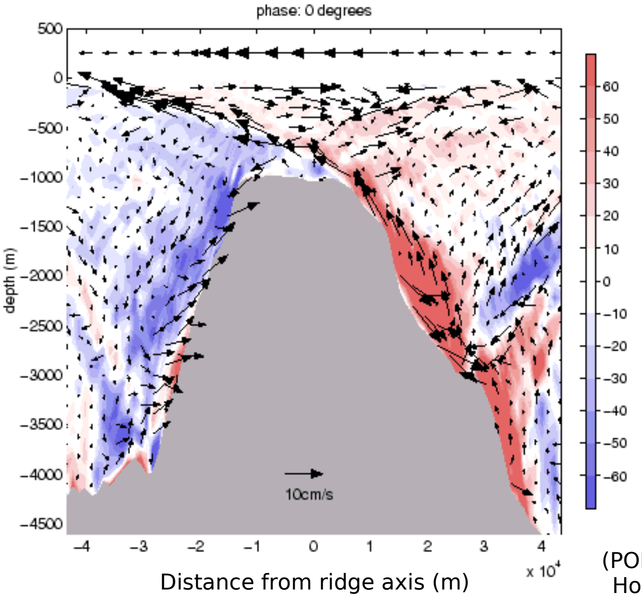
Numerical models of the tides

- **POM**: Princeton Ocean Model (Carter et al., 2007), nonlinear primitive equations
- **PEZHAT**: Primitive Equation Z-coordinate - Harmonic Analysis Tides (Zaron & Egbert, 2006), linearized

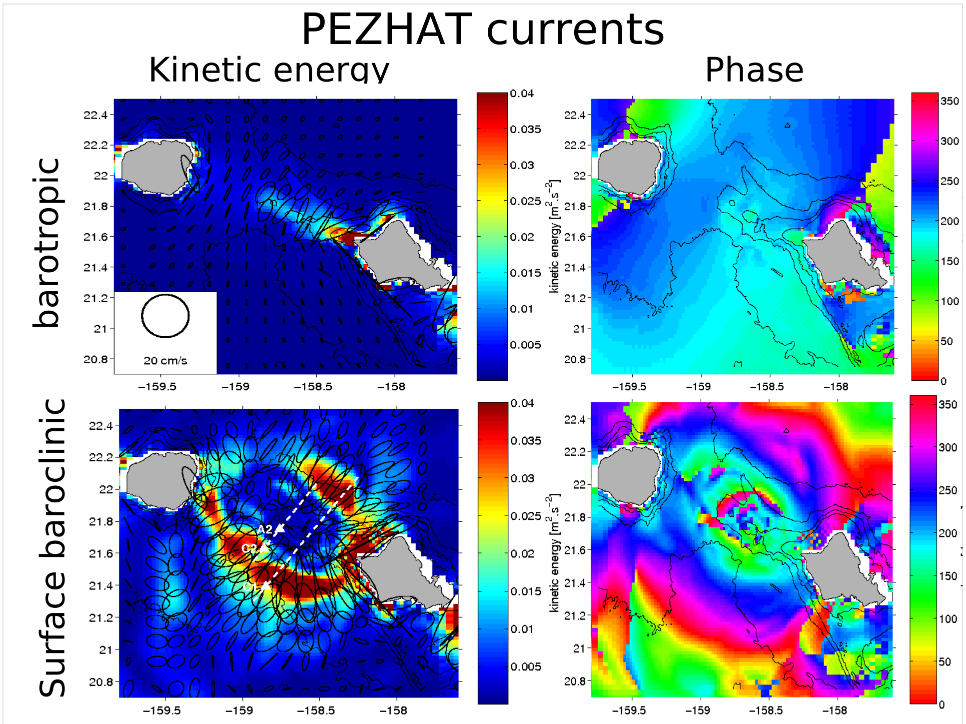


Parameter	PEZHAT	POM
Δx	2km	$\sim$ 1km (0.01^o)
Δz	60 z-levels unevenly spaced (30m near surface to 430m at 4000m)	61 σ-levels evenly spaced
A_V	$5 \times 10^{-4} m^2.s^{-1}$	Mellor-Yamada 2.5
K_V	$0.5 \times 10^{-4} m^2.s^{-1}$	0
A_H	$12 m^2.s^{-1}$	Smagorinsky
K_H	$12 m^2.s^{-1}$	0
T	14 M_2 periods	18 M_2 periods
T_{HA}	3 M_2 periods	6 M_2 periods

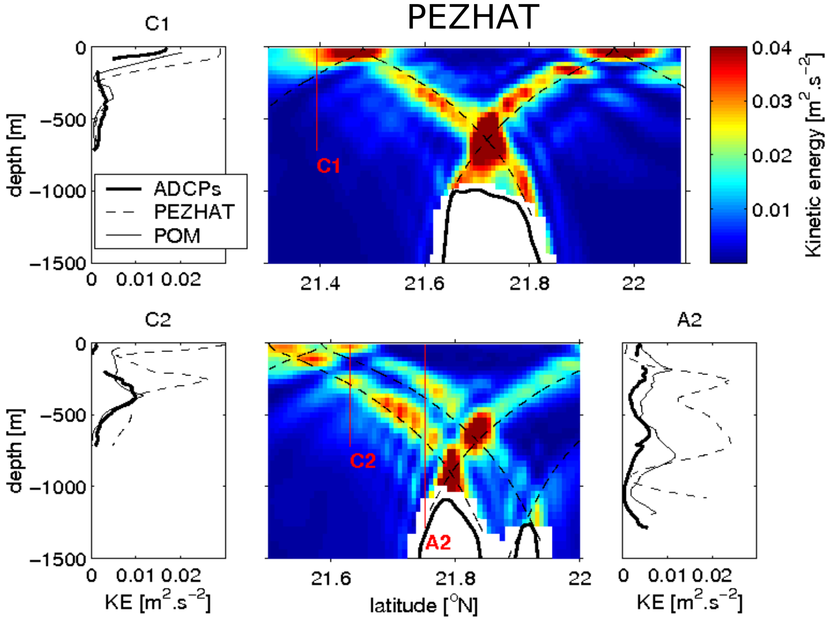
Internal tides generation



(POM, Merrifield & Holloway, 2002)


PEZHAT currents
Kinetic energy
Phase
barotropic
Surface baroclinic
20 cm/s
A2
C2
kinetic energy [m^2.s^{-2}]
22.4
22.2
22
21.8
21.6
21.4
21.2
21
20.8
−159.5
−159
−158.5
−158
0.04
0.035
0.03
0.025
0.02
0.015
0.01
0.005
350
300
250
200
150
100
50
0

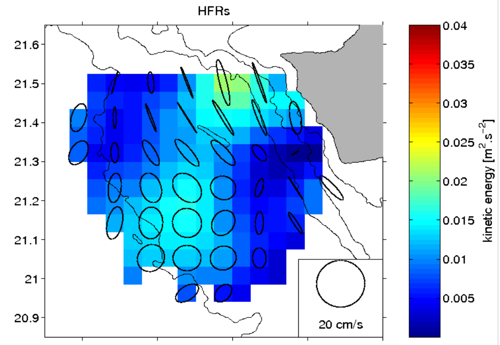
Vertical structure of KE

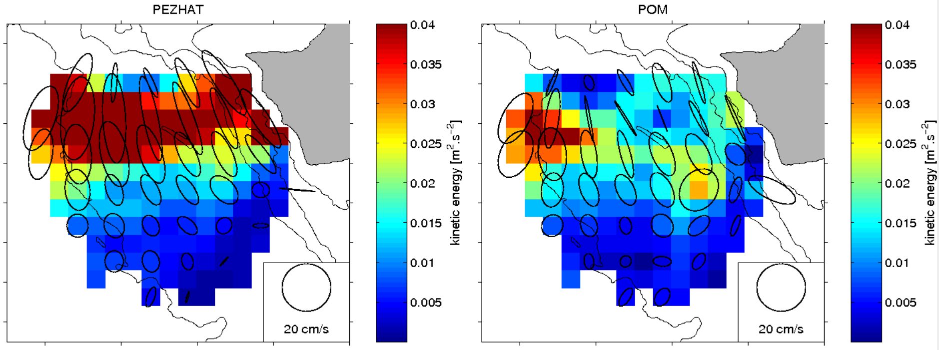


Surface kinetic energy

$KE_{PEZ} / KE_{HFR} = 2.7$

$KE_{POM} / KE_{HFR} = 1.4$

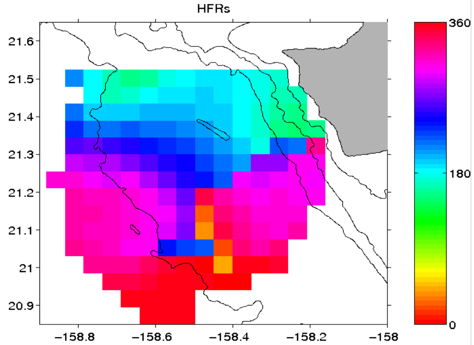


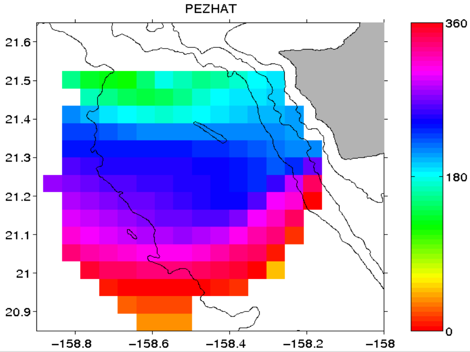


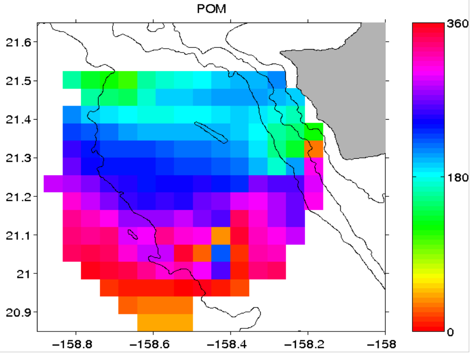
Surface phase

$$\lambda_{HFR} = 123 \text{ km}$$

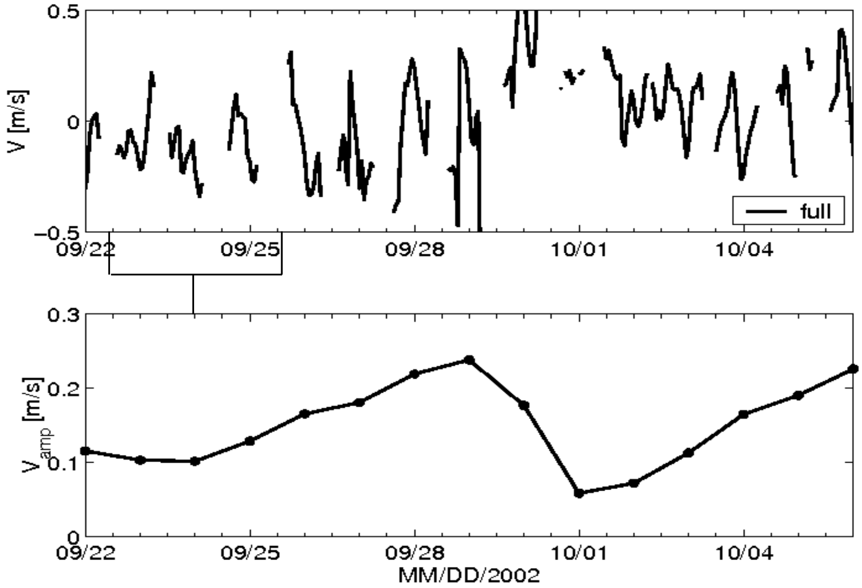
$$\lambda_{PEZ} = \lambda_{POM} = 105 \text{ km}$$



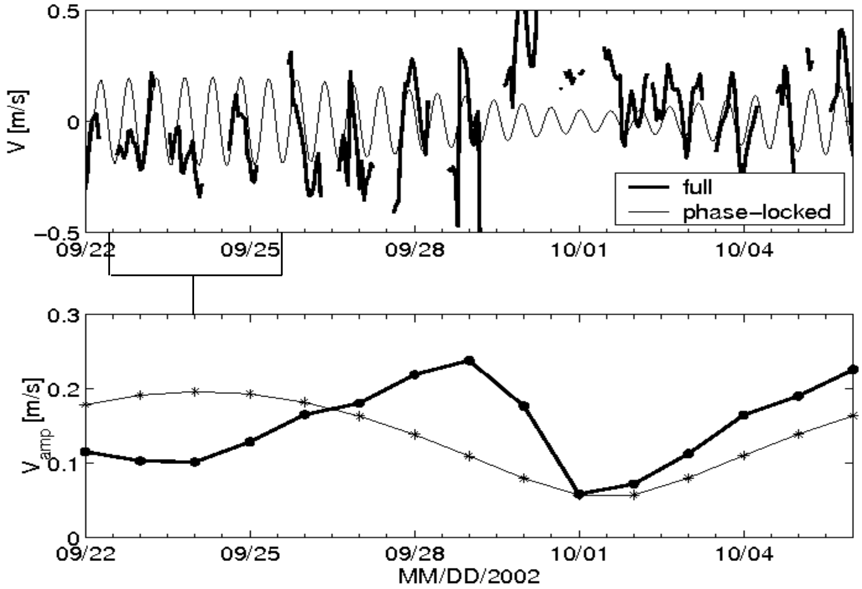




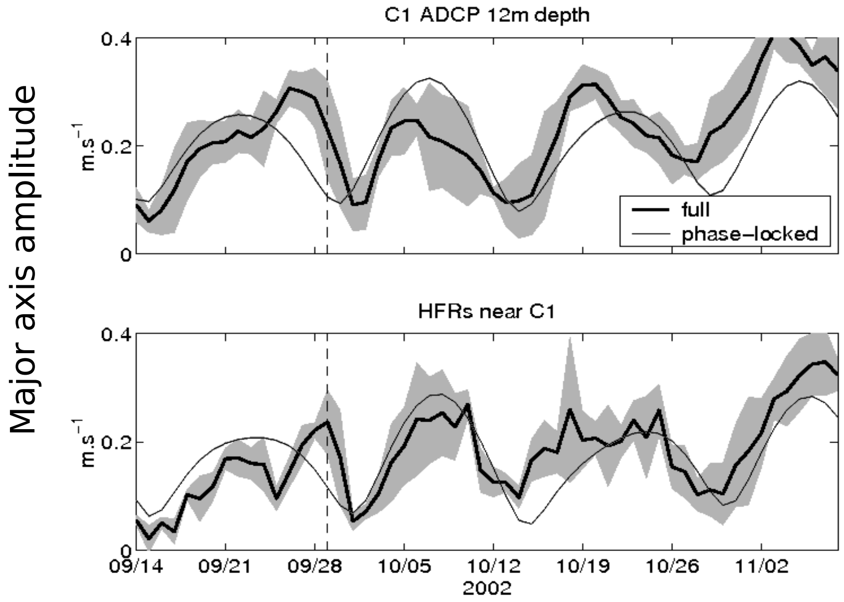
Complex demodulation: example



Complex demodulation: example



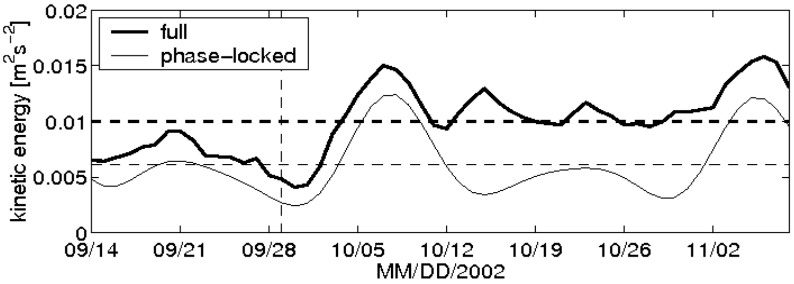
Temporal variations at C1



Spatially-averaged kinetic energy evolution

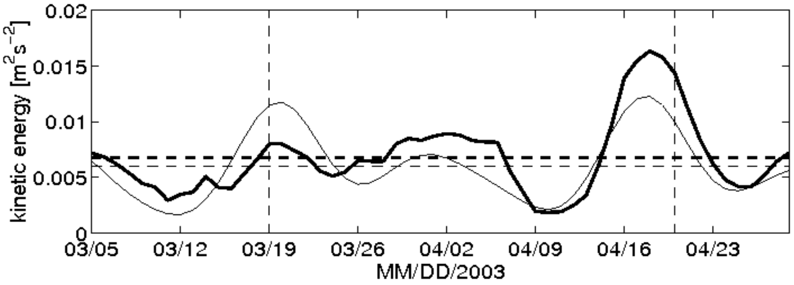
$\mathrm{KE}_{\mathrm{full}}$

$\mathrm{KE}_{\mathrm{lock}}$
=
1.6



$\mathrm{KE}_{\mathrm{full}}$

$\mathrm{KE}_{\mathrm{lock}}$
=
1.1



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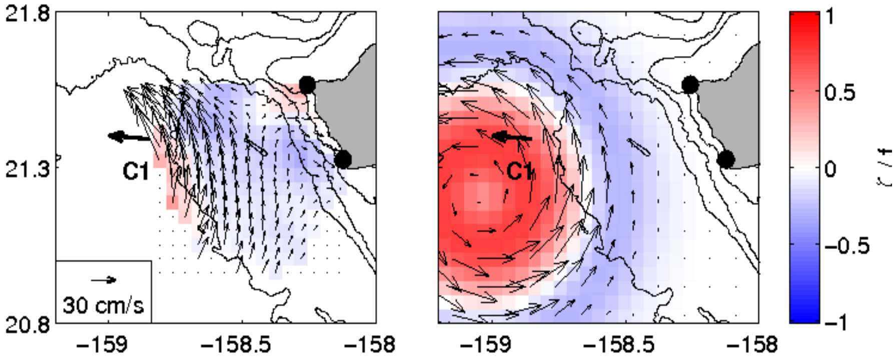
a) Cyclone: horizontal structure

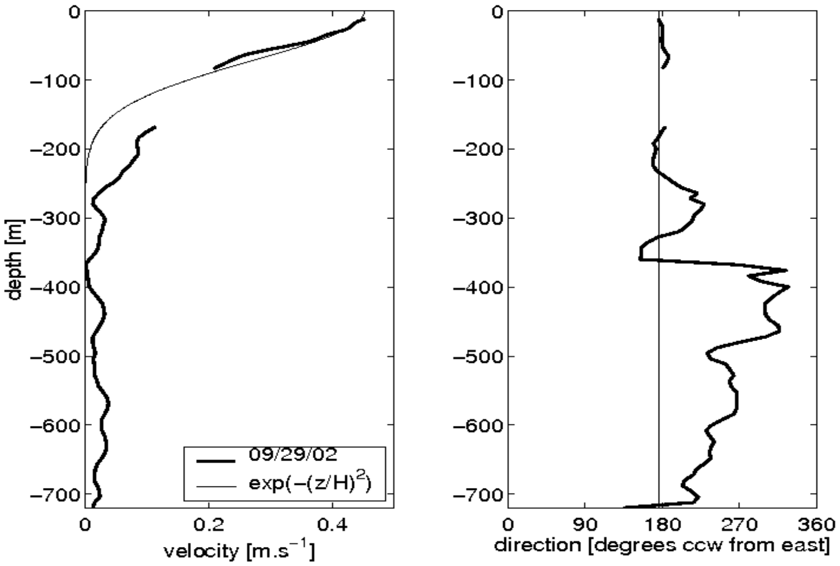
Observations
(09/29/02)

Idealization
(for ray tracing)



R ~ 30 km

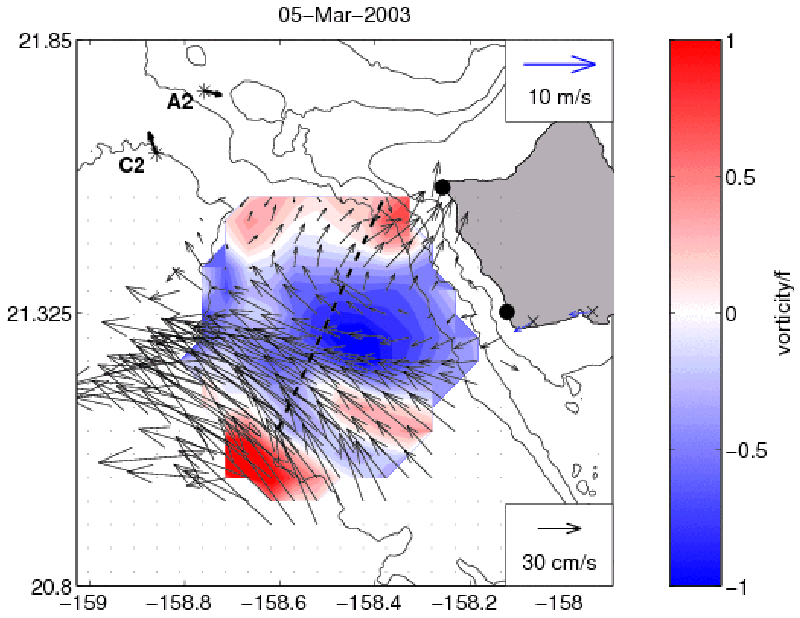
a) Cyclone: vertical structure



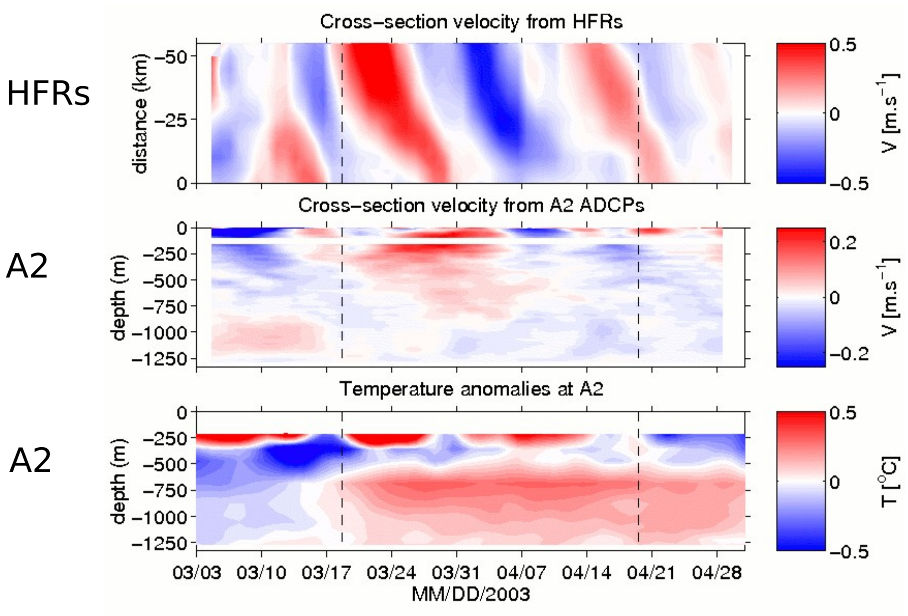
velocity

direction

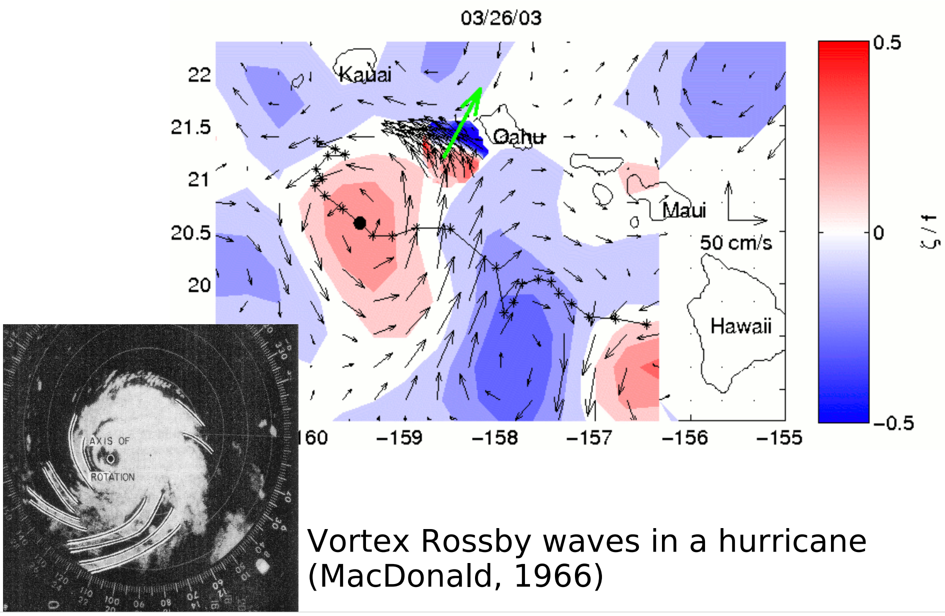
b) Vorticity waves: propagation



b) Vorticity waves: propagation



b) Vorticity wave = Vortex Rossby wave



Vortex Rossby waves in a hurricane
(MacDonald, 1966)

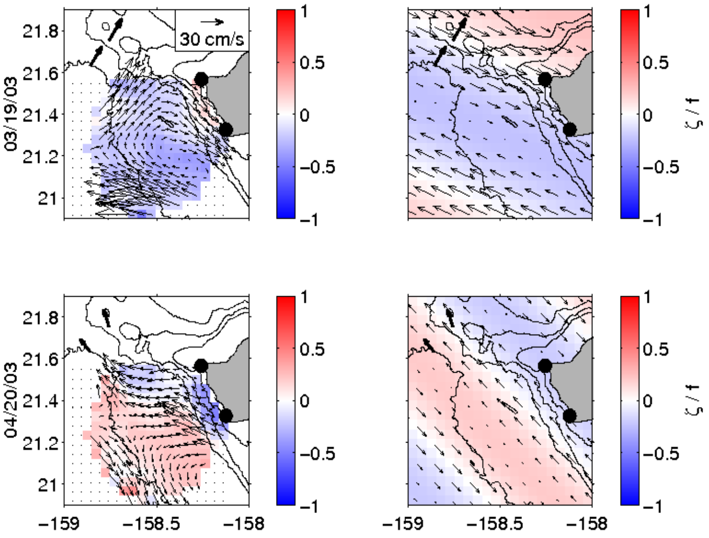
b) Vorticity waves: horizontal structure

Observations

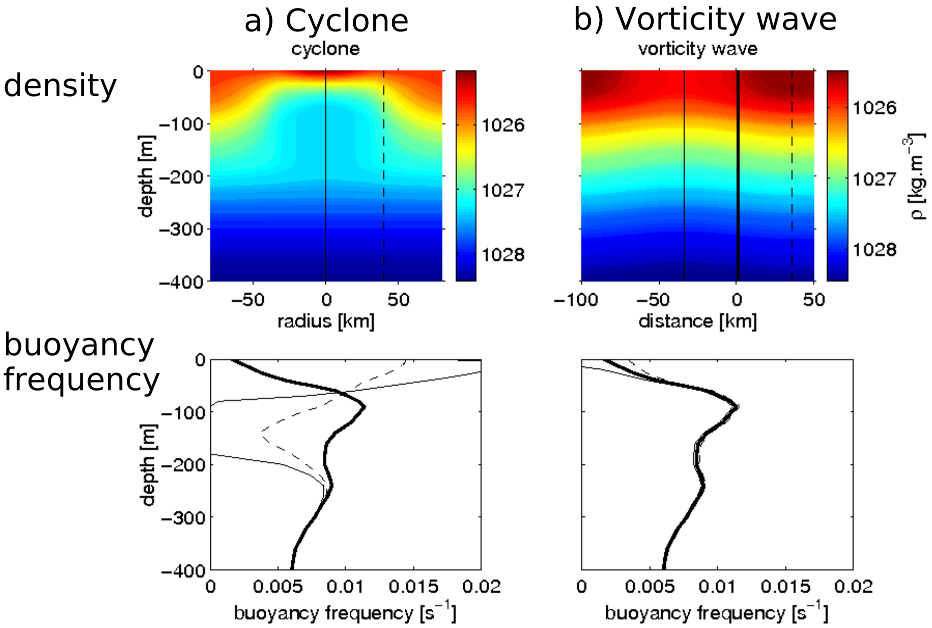
Idealization
(for ray tracing)

03/19/03
λ ~ 150 km

04/20/03
λ ~ 100 km



Idealized stratification



a) Cyclone

b) Vorticity wave

density

buoyancy frequency

3. Ray tracing

- wave packet: $\psi(\mathbf{x}, t) = a(\mathbf{x}, t)\ \exp^{i\theta(\mathbf{x},t)}$

- local wavenumber: $\mathbf{k}(\mathbf{x}, t) := \frac{\partial \theta}{\partial \mathbf{x}}$

- local frequency: $\omega(\mathbf{x}, t) := -\frac{\partial \theta}{\partial t}$

- local dispersion relation: $\omega = \Omega(\mathbf{k}; \mathbf{x}, t)$

- ray equations: $\frac{d\mathbf{x}}{dt} = \mathbf{C_g} = \frac{\partial \Omega}{\partial \mathbf{k}}$

$$\frac{d\mathbf{k}}{dt} = \mathbf{r} = -\frac{\partial \Omega}{\partial \mathbf{x}}$$

Ray tracing

- Doppler shift: $\omega = \Omega(\mathbf{k}, \mathbf{x}, t) = \omega_0 + \mathbf{k}.\mathbf{U}$

- intrinsic frequency: $\omega_0^2 = f_{eff}^2 + N_{eff}^2 \frac{k_h^2}{k_v^2}$
  (Kunze, 1985)

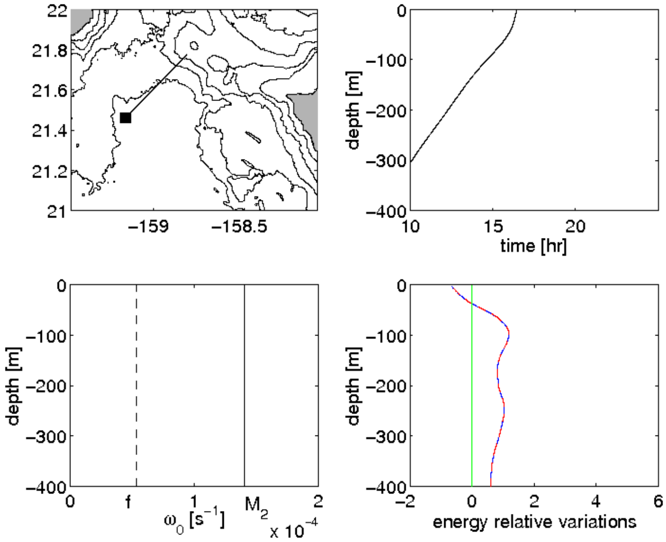
- effective Coriolis frequency:

  $f_{eff} = f + \cancel{\zeta/2 + \ldots}$

- effective buoyancy frequency:

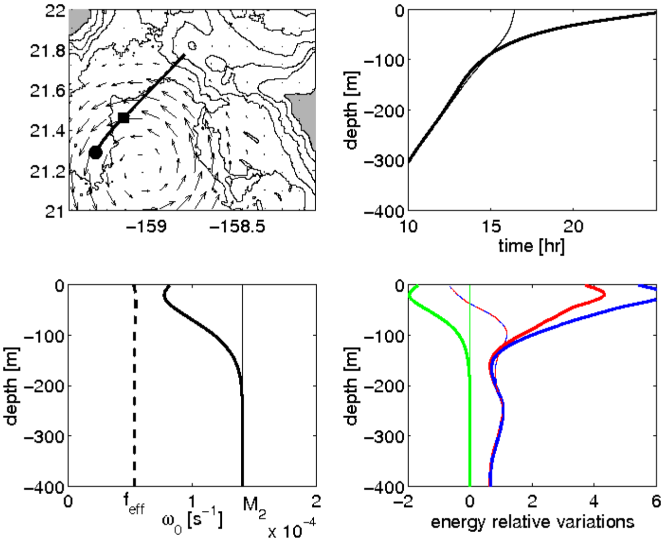
  $N^2_{eff} = N^2 + \ldots$

Example: without currents



$$\frac{dE}{dt} = -E\nabla.C_g + \frac{E}{\omega_0}\frac{d\omega_0}{dt}$$

Example: with currents

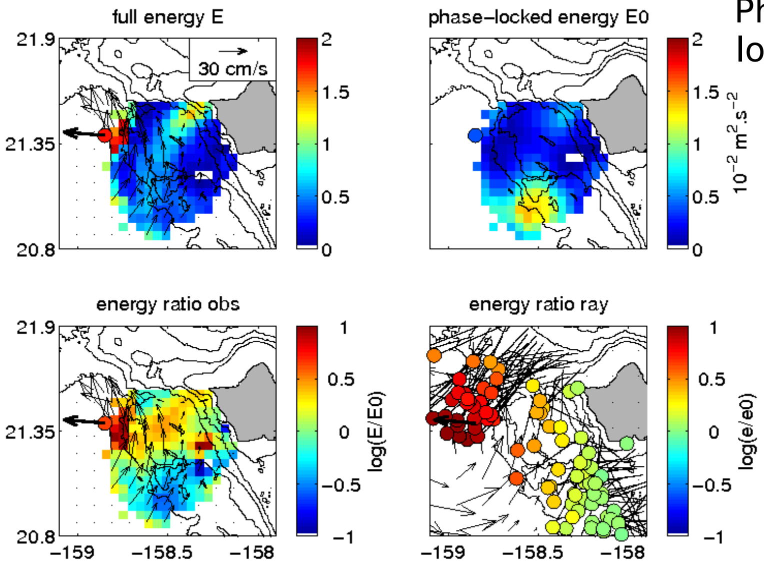


$$\frac{dE}{dt} = -E\nabla . C_g + \frac{E}{\omega_0}\frac{d\omega_0}{dt}$$

Surface kinetic energy

“full”

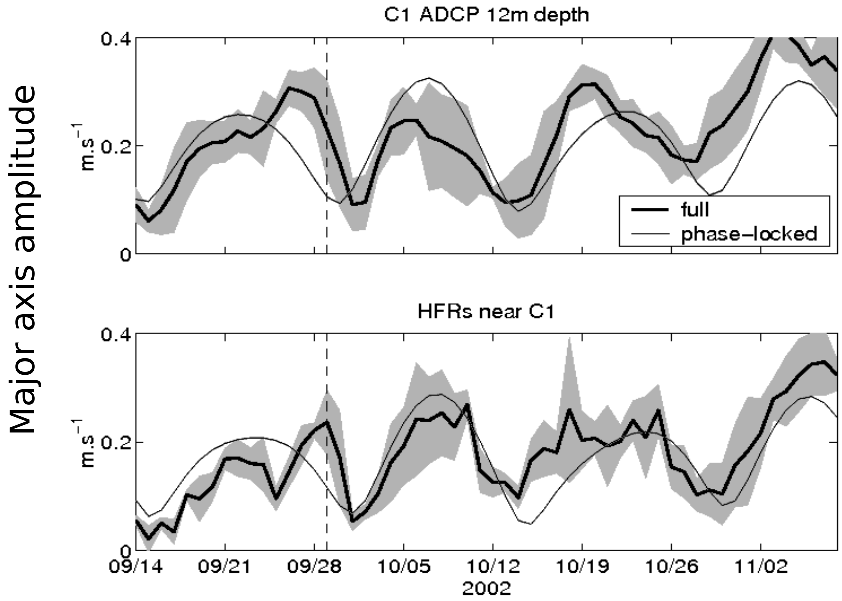
Phase-
locked



Full / phase-locked

Currents / No currents

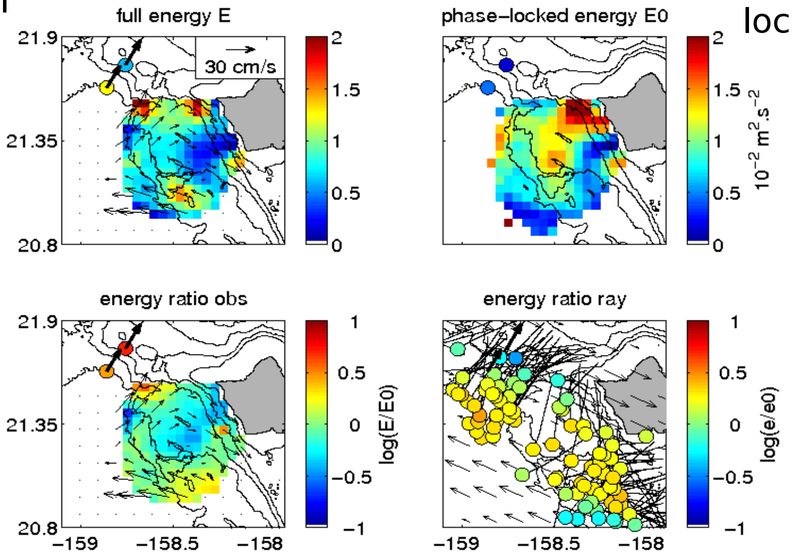
Temporal variations at C1



VRW on March 19, 2003

“full”

Phase-locked



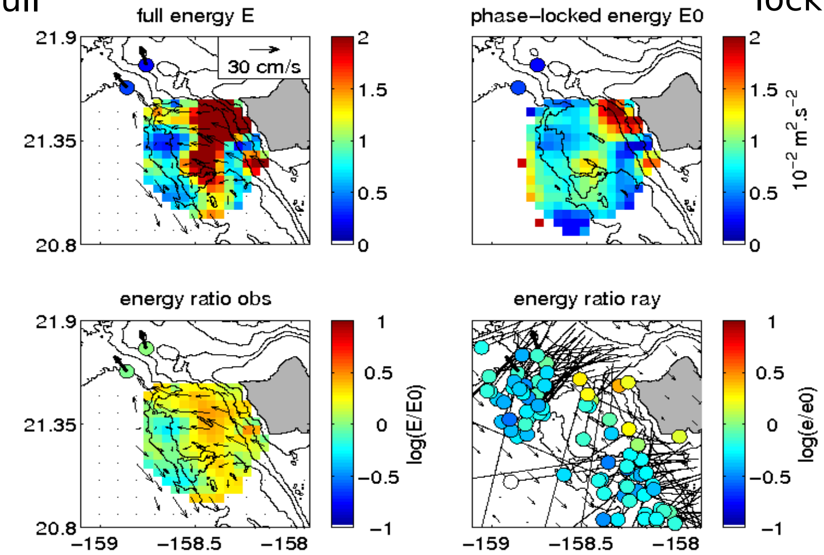
Full / phase-locked

Currents / No currents

VRW on April 20, 2003

“full”

Phase-locked



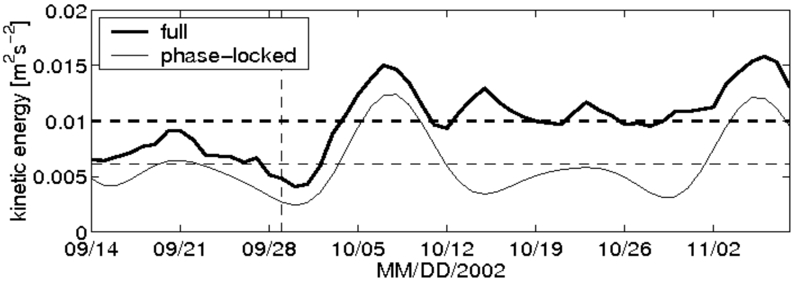
Full / phase-locked

Currents / No currents

Spatially-averaged kinetic energy evolution

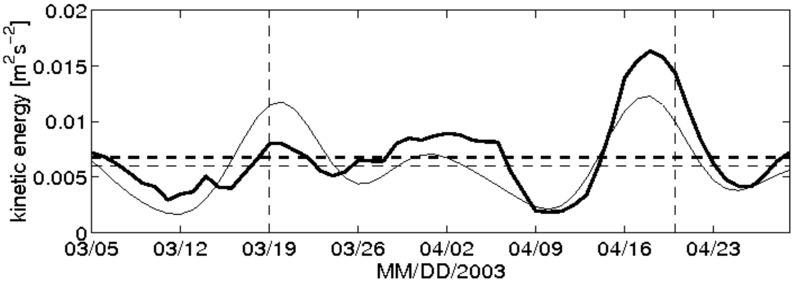
KE_{full}

KE_{lock}
=
1.6

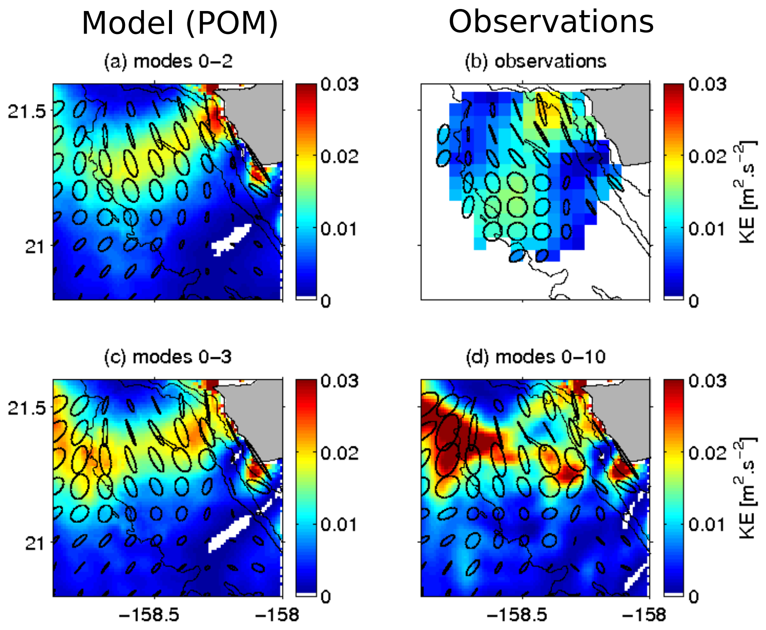


KE_{full}

KE_{lock}
=
1.1



Vertical modes filtering



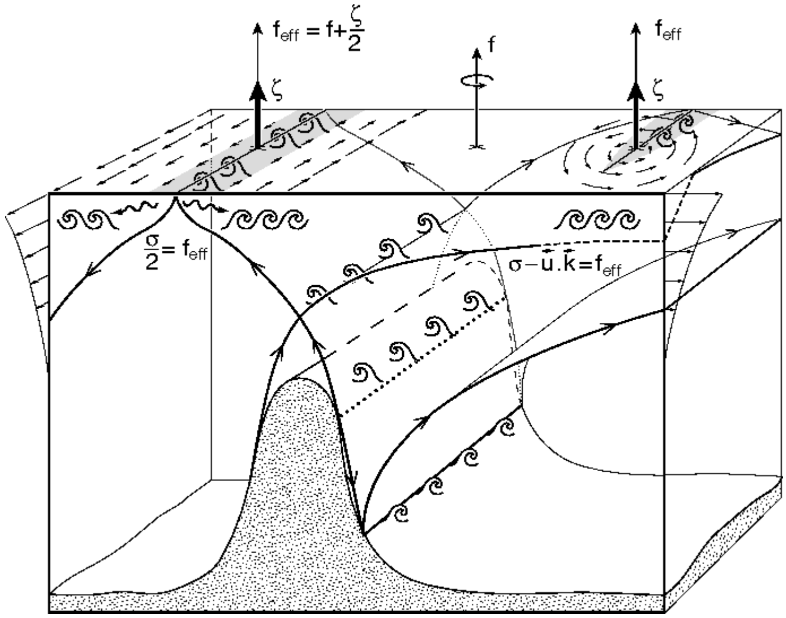
Conclusions

- Energetic mesoscale and submesoscale features (Ro~1 cyclones and anticyclones, fronts, vortex Rossby waves), not resolved by altimetry
- Mesoscale currents refract, Doppler-shift, and exchange energy with internal tidal beams at their first surface reflexion in the Kauai Channel
- Phase and amplitude modulations lead to low-pass filtering of vertical modes when harmonically analyzed over long periods of time

Implications for tidal energy budget

- barotropic energy loss:
  well constrained by assimilation of phase-locked observations (e.g. altimetry)
- baroclinic energy flux radiation:
  assimilation of phase-locked observations should be considered as lower bounds;
  assimilation of HF-radio phase-locked M2 currents into PEZHAT leads to ~10% decrease
- What about locally “dissipated” energy ?

Where should dissipation occur ?



Final remarks

- Less energy available for deep mixing ?
- Garrett and St Laurent (2002):

  “It may well be that the behavior of the surface layer is the most important oceanic component for the climate system. This behavior depends on mixing processes at the base of the surface layer. Internal waves generated by the wind, and possibly also by the tide, drive mixing at the base of the surface layer.”

Julie
Deshayes

Adrien
Desoria

Joël
Benito

Cédric
Chavanne