# Mixing and the Tropical Oceans

Marcus Dengler

Roadmap: Mixing and the Tropical Oceans

▸ Motivation: Stirring and mixing of a tracer

- • Eddies and stirring
- • Turbulent mixing

▸ Seasonal cycle of the tropical ocean

▸ Eddies in the equatorial oceans

▸ Turbulent mixing within the Equatorial Undercurrent

▸ Mixed layer heat budget

▸ Summary

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Stirring and mixing of a tracer
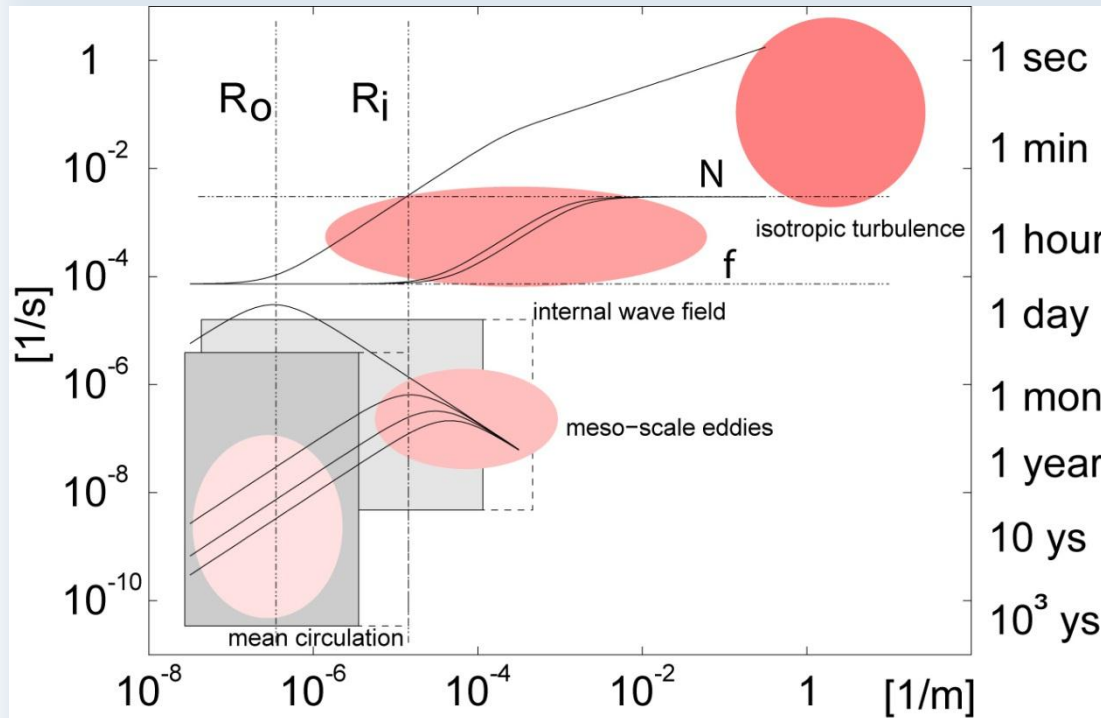
(Olbers, Willebrand and Eden, 2012)

▸ mixing in the ocean is achieved by molecular diffusion at the scales of isotropic turbulence.

▸ Smallest fluctuations:
velocity   2-5 cm
temperature   1 cm
salinity   1 mm

▸ micro-scale gradients for molecular forces to act are generated by
  • meso-scale eddies acting on a predominantly isopycnal mean gradient
  • turbulent motions acting on mean and meso-scale gradients

Steering by eddies

Snapshot of satellite sea surface height anomaly

Eddies acting on a mean isopycnal tracer gradient
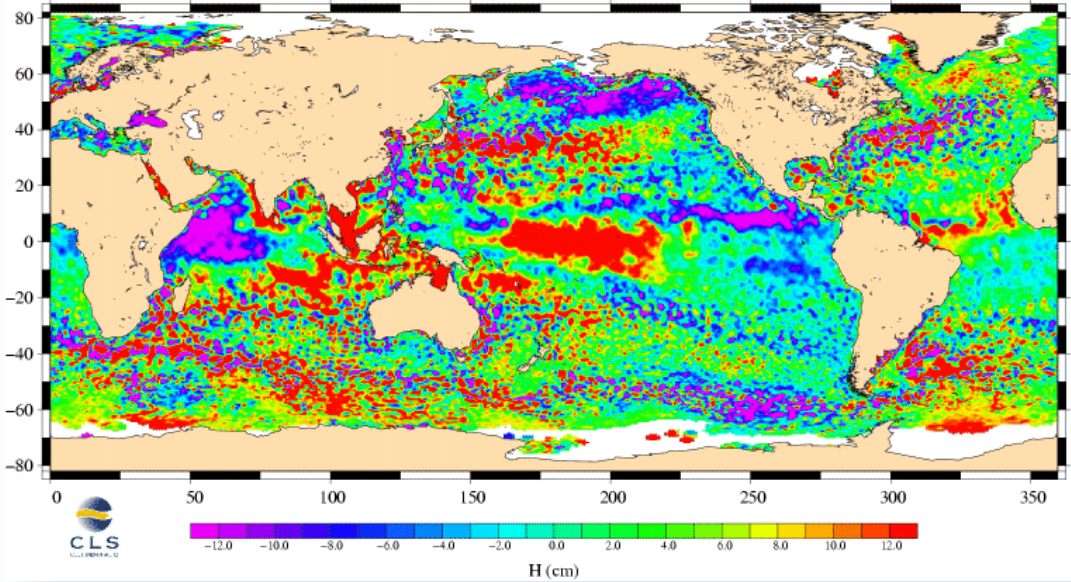
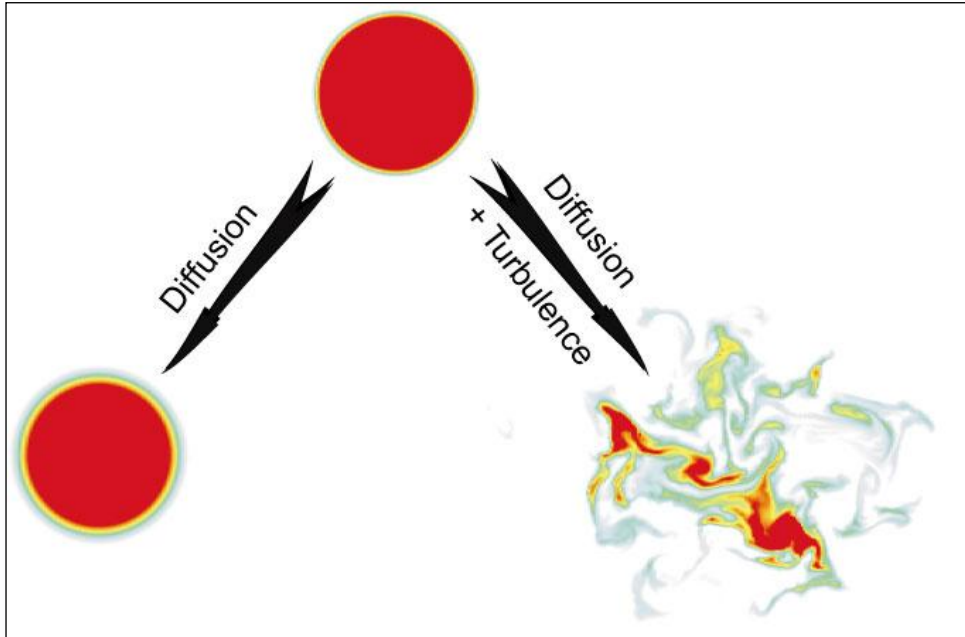
- horizontal scales of a few km to several 1000km
- mixes a tracer predominately along isopycnals (steering)
- steering creates contrast between water masses and enhances variability of the temperature - salinity (T-S) relationship
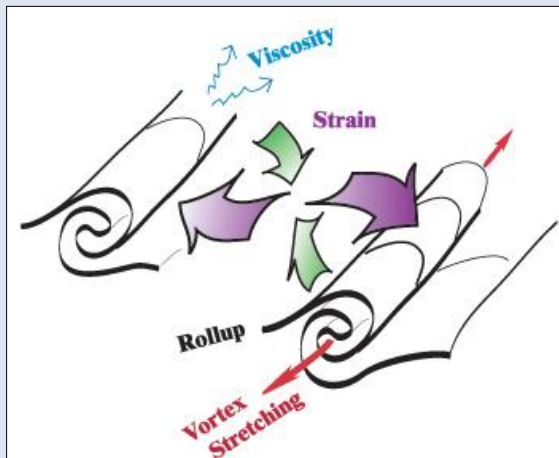
Turbulence and mixing

- ▶ largest horizontal scales of a few 100m and vertical scales of 50m, smallest scales of about 1cm (isotropic)
- ▶ turbulent motions predominately mix across isopynals
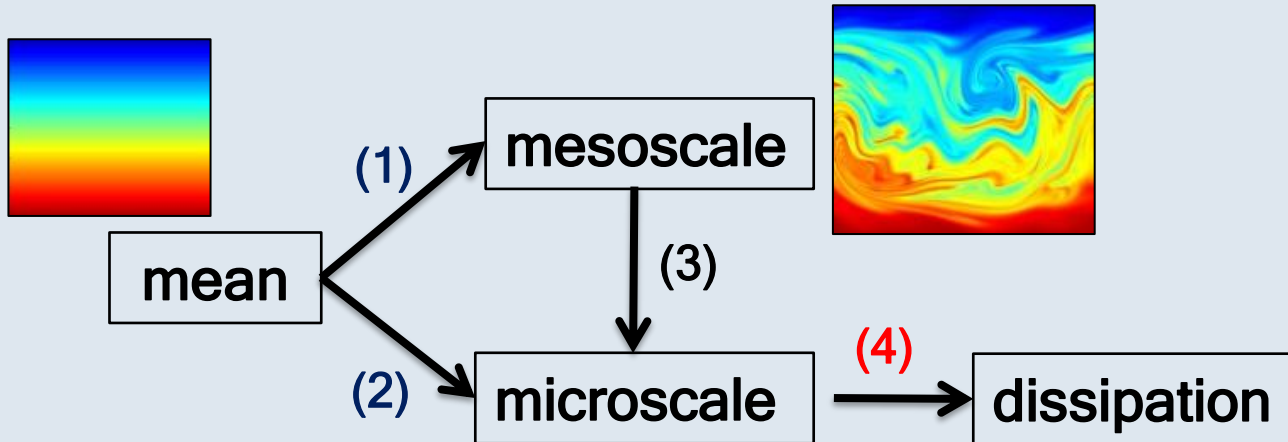- ▶ turbulence acts to homogenize different water masses and tightens the T-S relation

Concept: Triple decomposition

Transport equation of temperature variance using:

$$\theta = \theta_m + \theta_e + \theta_t$$

m - mean, e - eddy scale, t - turbulence scale

(extended Osborn-Cox model, Joyce, 1977; Davis, 1994; Garrett, 2001; Ferrari and Polzin, 2005)

(1) **variance production** by mesoscale eddies (along isopycnals)
(2) **variance production** by turbulence (across isopycnals)
(3) **transfer of variance** (mesoscale ➔ microscale) by turbulence
(4) **variance dissipation** on molecular scale

Concept requires scale separation which is not given in the ocean

Eddy vs. turbulence variance production

$$\langle u_e \theta_e \rangle \cdot \nabla_n \theta_m + \langle w_t \theta_t \rangle \cdot \nabla_n^\perp \theta_m = \kappa_C \langle |\nabla \theta_t|^2 \rangle / 2 \equiv \langle \chi \rangle / 2$$

production by eddies   production by turbulence   dissipation of variance

∇_n denotes along-neutral-surface gradients
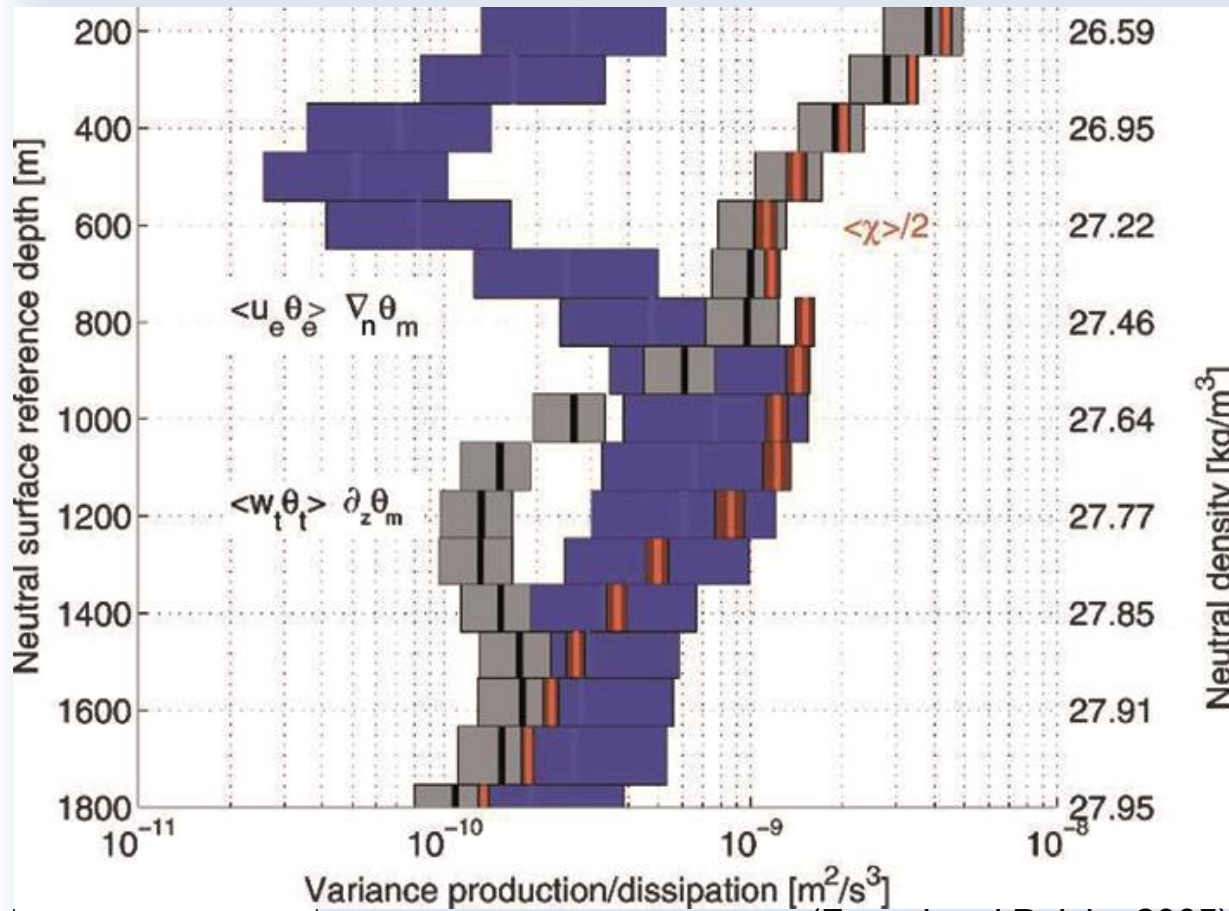

(Ferrari and Polzin, 2005)

▸ evaluated from data collected in the eastern subtropical North Atlantic

▸ turbulent temperature variance production equates dissipation above 800m and below 1400m

▸ stirring by Mediterranean Water eddies dominates production between 800 and 1400m depth

7

Turbulence and Mixing
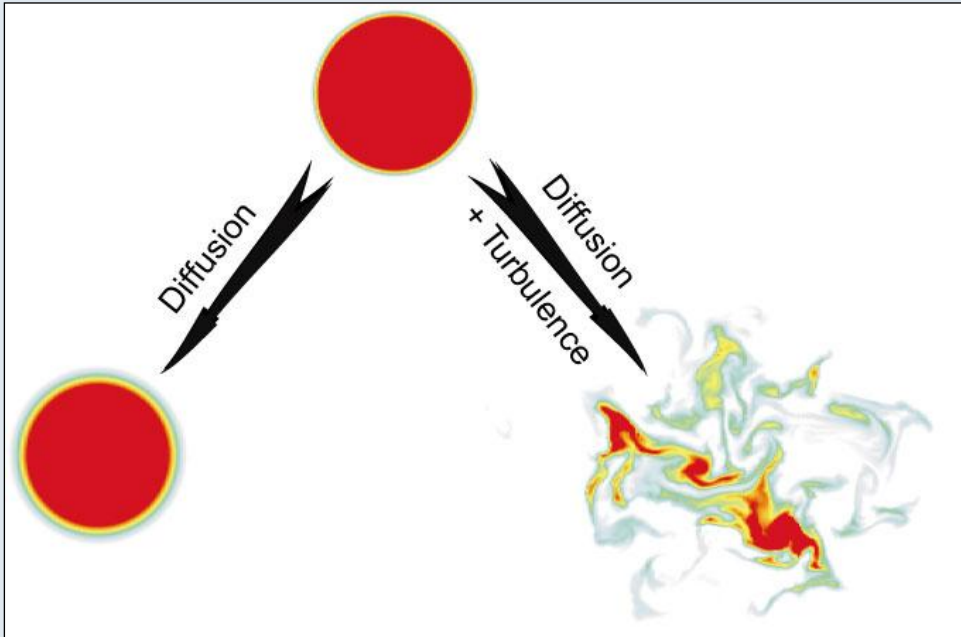
- ▸ turbulence is a fluid regime characterized by chaotic, stochastic property change
- ▸ its flow field is often rationalized as a superposition of many small eddies (billows)
- ▸ momentum/energy is supplied to turbulence from internal waves or larger scale flow predominately through Kelvin-Helmholtz Instabilities
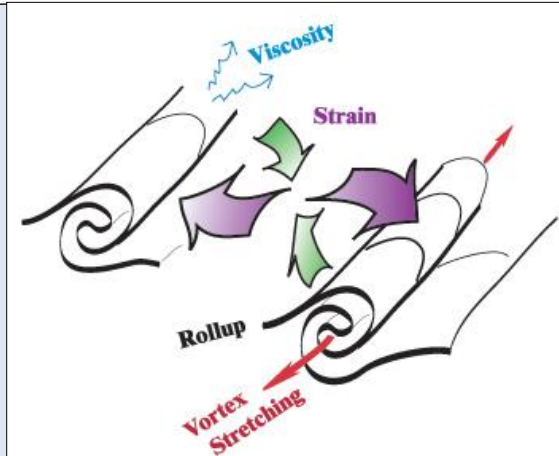
Shear Instability (Kelvin-Helmholtz Instability)
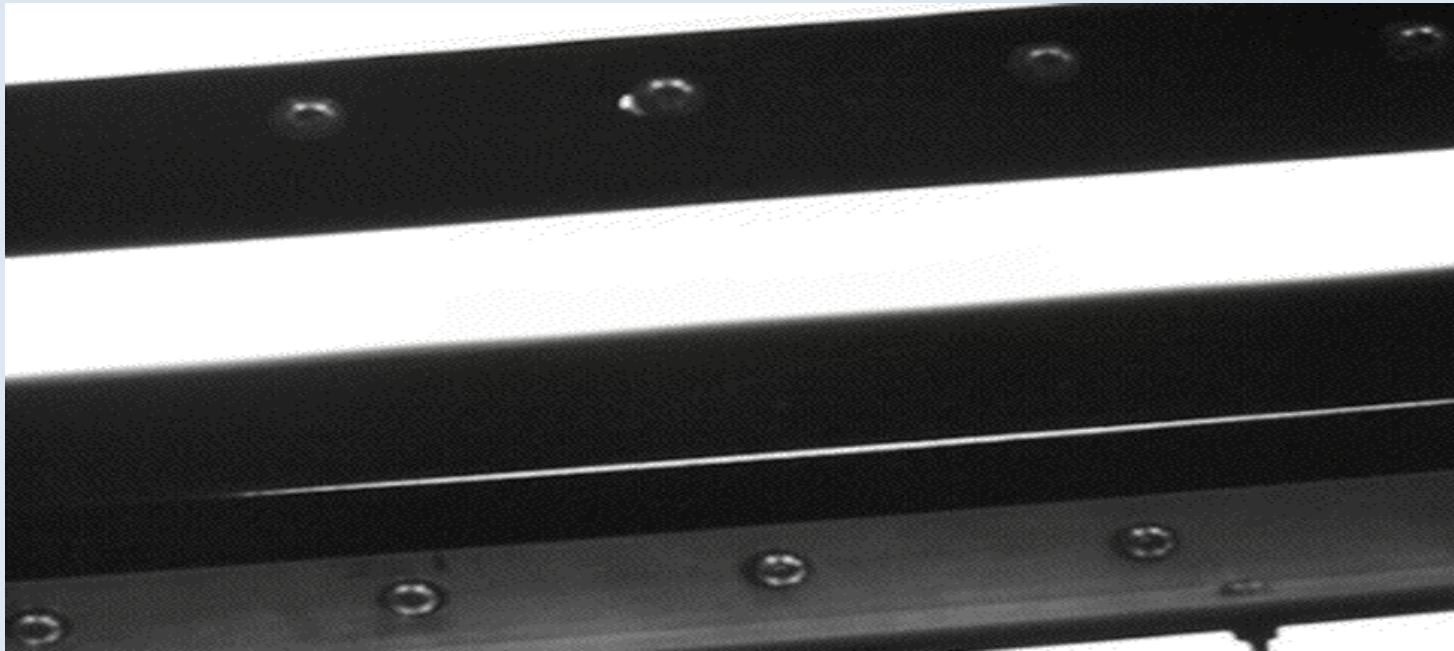
Flow in the white layer is faster than flow in the dark layer

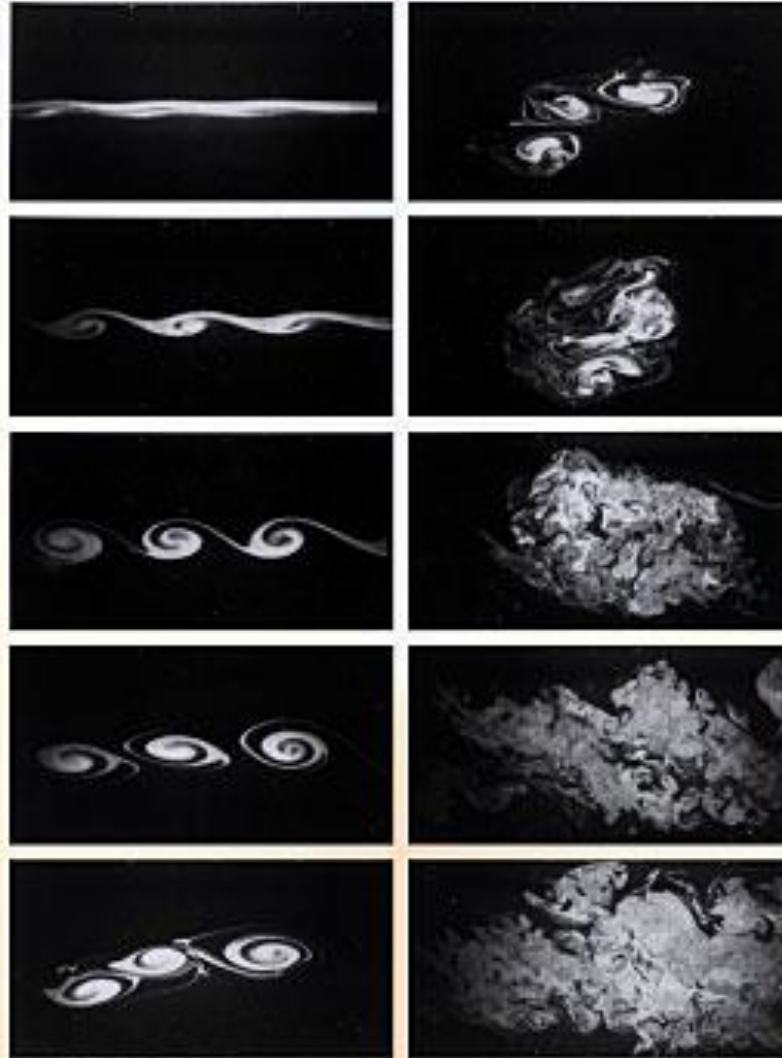Necessary condition for Kelvin-Helmholtz Instability to develop:

$$\text{Ri} = \frac{N^2}{S^2} < 0.25$$

Ri – Richardson Number

$N^2 = \frac{g}{\rho}\frac{\partial \rho}{\partial z}$ stratification, $S^2 = \left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2$ shear of background flow

Kelvin-Helmholtz Instability

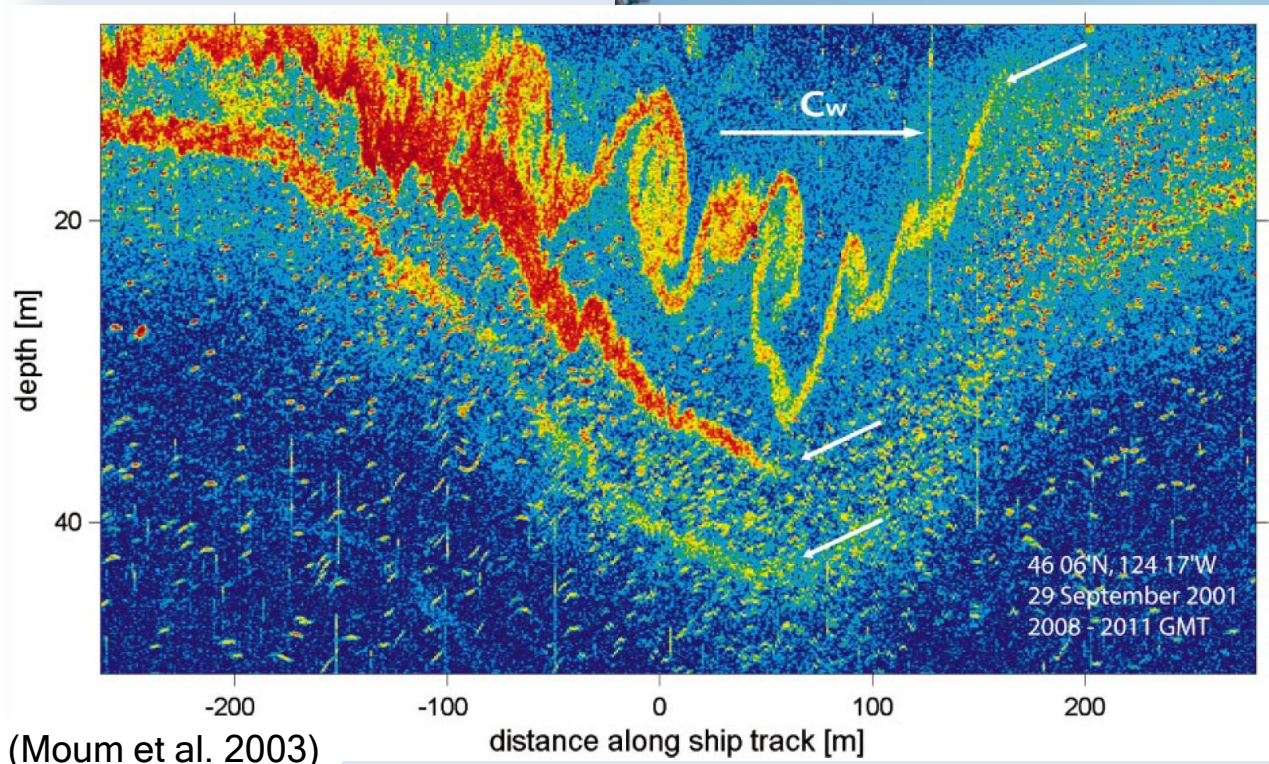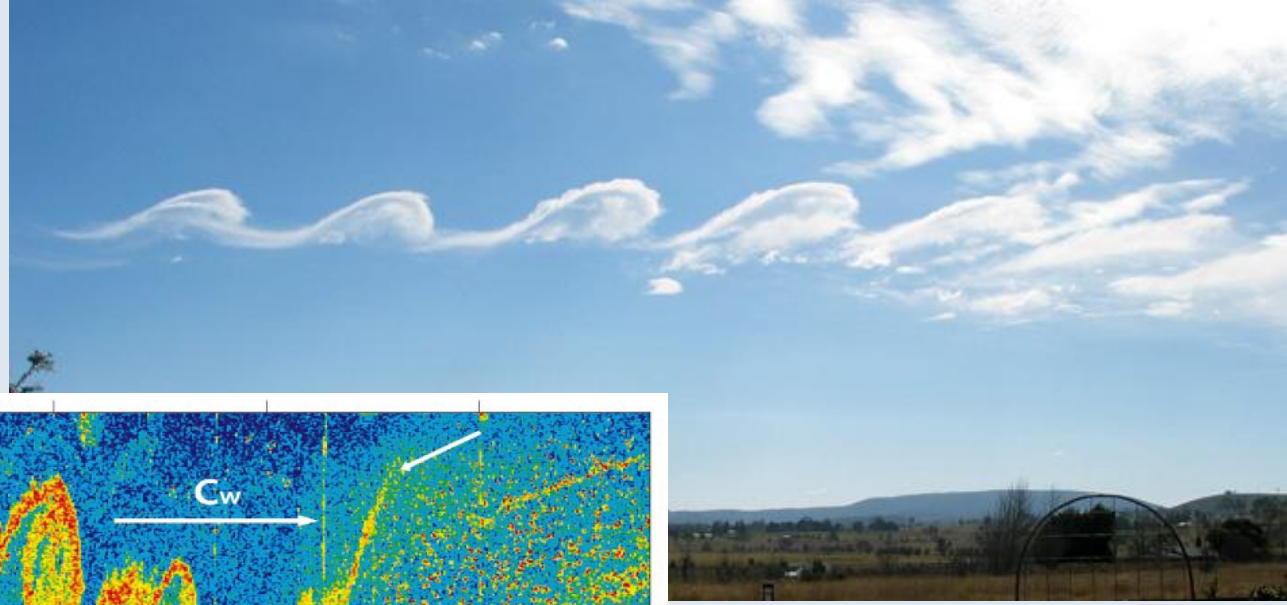Velocity shear generates Kelvin-Helmholtz vortices (billows, eddies)

Break down of the large billow structure into small turbulent billows (few m to 1 cm) and chaotic turbulent flow.

At very small scales, these billows are isotropic (<20 cm).
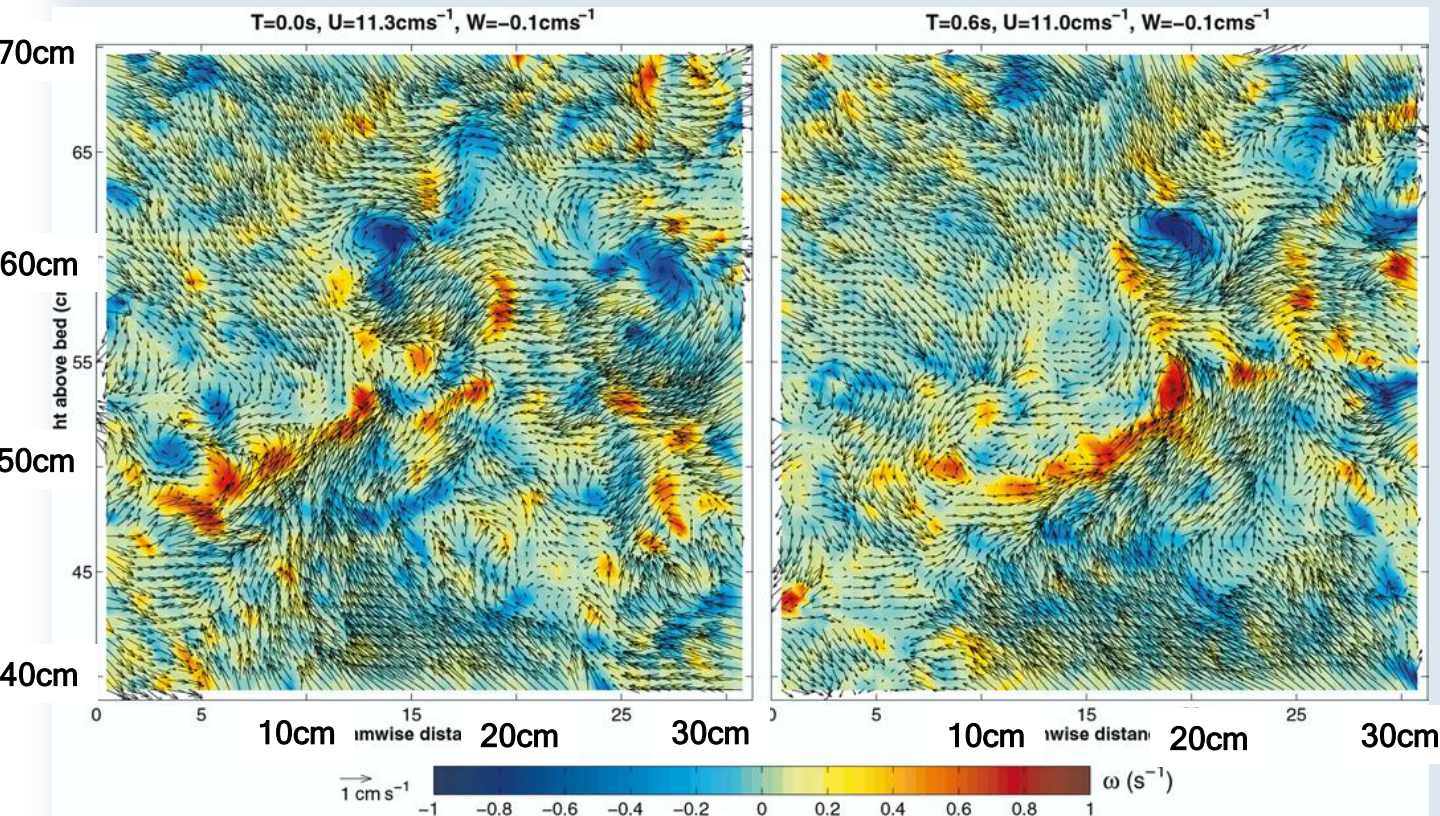
Kelvin-Helmholtz Instability

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Clouds visualizing the formation of Kelvin-Helmholtz Instability in the atmosphere

Strong shear in a long internal wave (400m horizontal scale) leads to Kelvin-Helmholtz instabilities (30m horizontal and 10m vertical scale)

depth [m]

C_w

46 06'N, 124 17'W
29 September 2001
2008 - 2011 GMT

distance along ship track [m]

(Moum et al. 2003)

The oceans' smallest movements

High-resolution instantaneous ocean velocities (30cm x 30cm)

- ▸ structure of small turbulent motions are also eddy-like

- ▸ there is a smallest eddy scale (i.e. there are no tiny eddies)

(Nimmo Smith et al., 2002)

Two sample instantaneous velocity vector maps from particle imaging velocimetry. Vectors are shown with the mean velocities removed, and are superimposed on the vorticity fields. The two maps are of the same sample area, but separated in time by 0.6 s

Turbulence Cascade

Wavenumber spectrum of turbulence
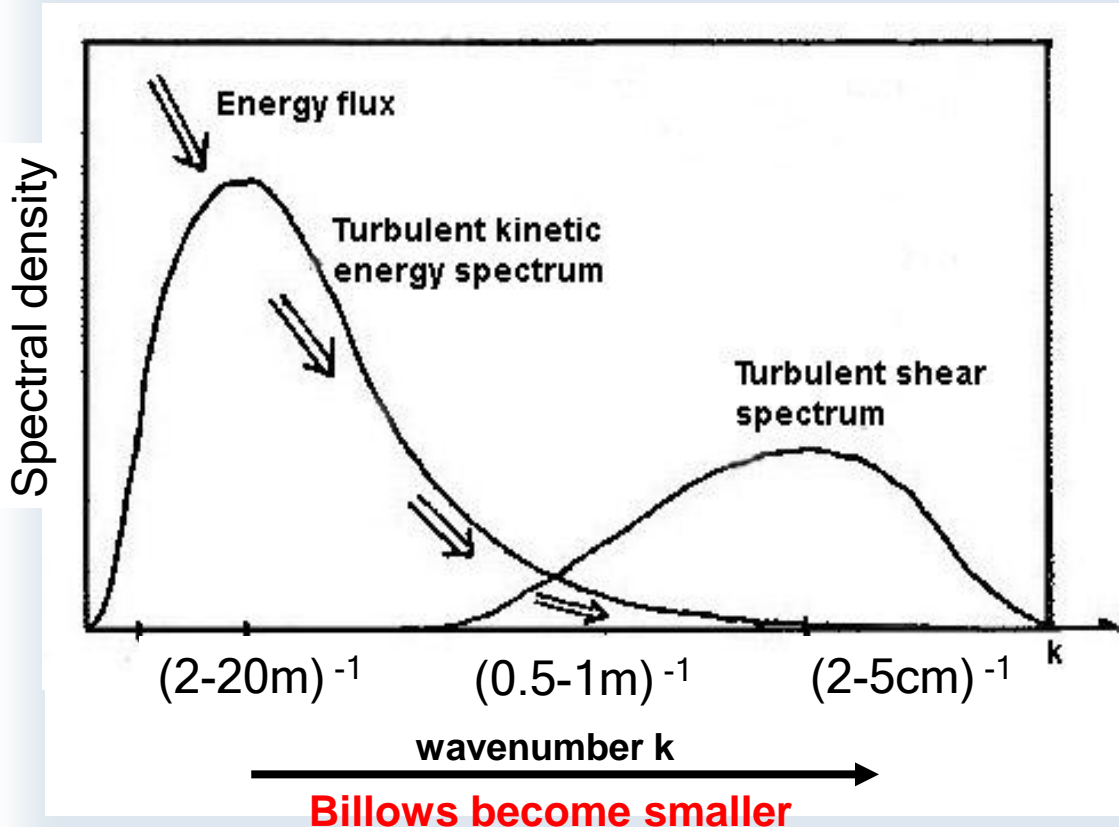
- instabilities of the background flow flux energy into largest turbulent billows (1m-40m)

- smaller turbulent billows are generated from interaction of larger billows (energy flux to smaller scales)

- viscous forces act on small billows having large shear passing their energy to molecular movements
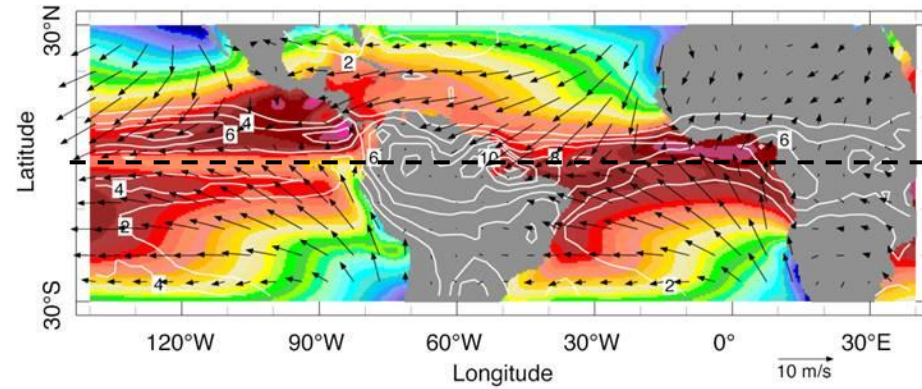
Roadmap: Mixing and the Tropical Oceans

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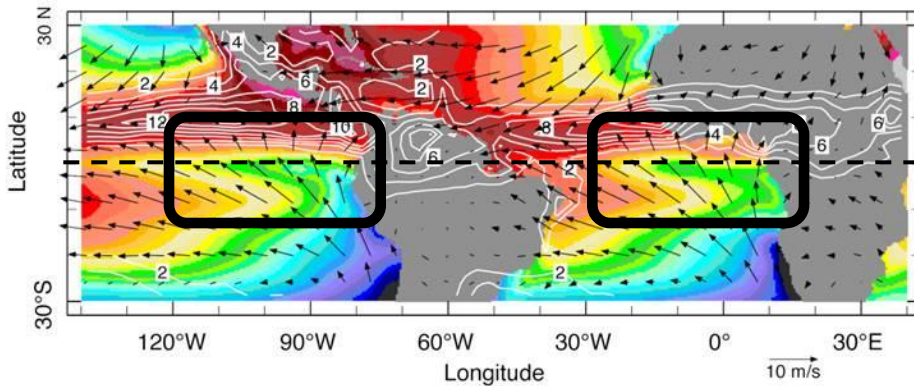
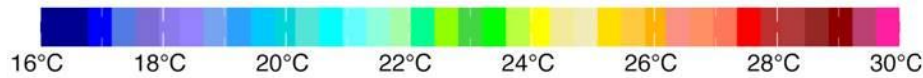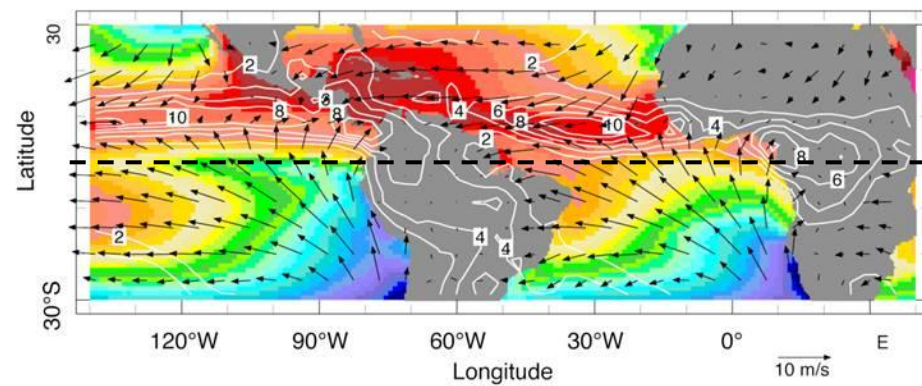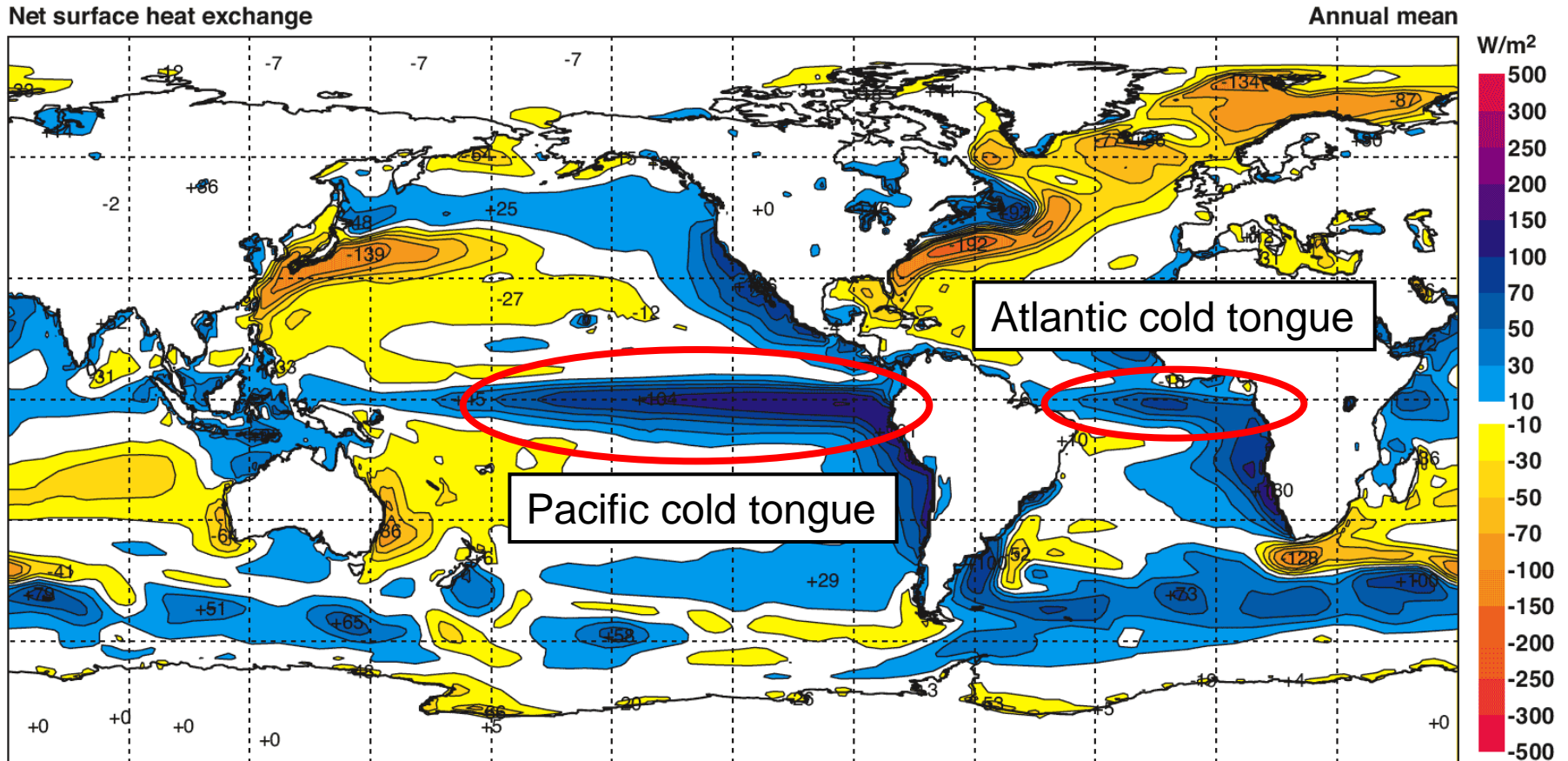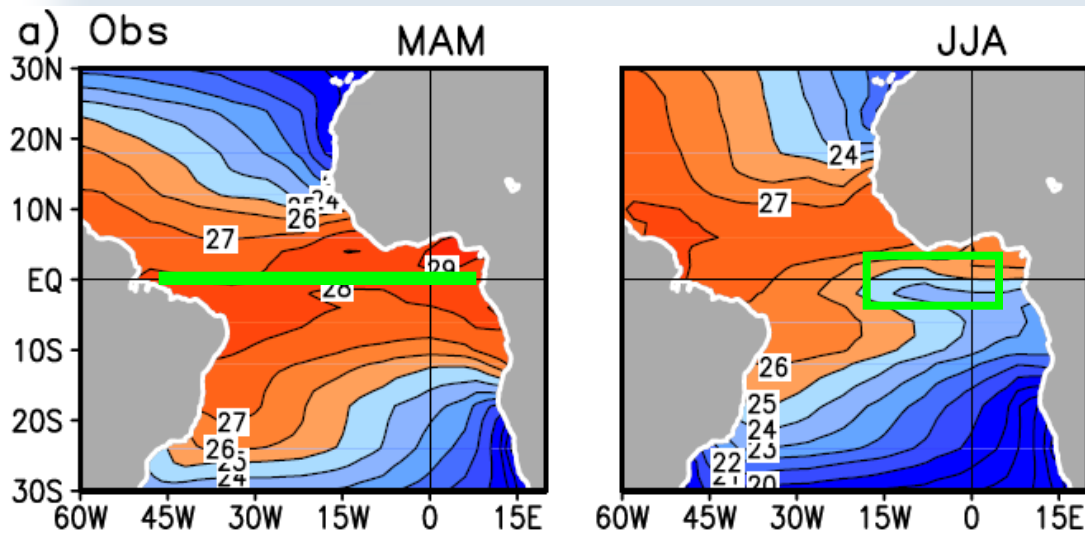# Seasonal cycle of tropical marine ITCZ, SST and wind

Net surface heat fluxes
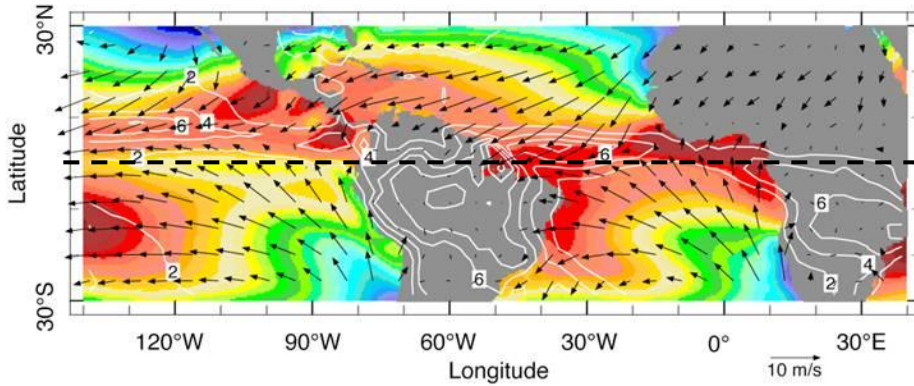
Annual-mean heat flux Q through the sea surface in Wm^{-2}
calculated from the ECMWF 40-year reanalysis (Kallberg et al., 2005)

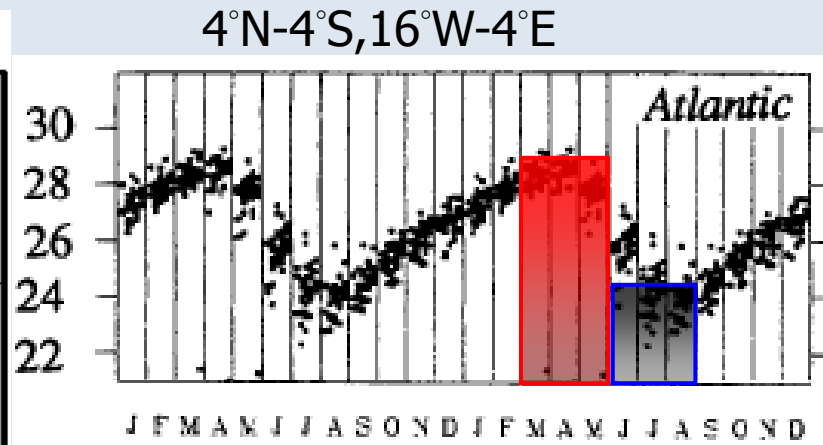# Surface winds in the Atlantic
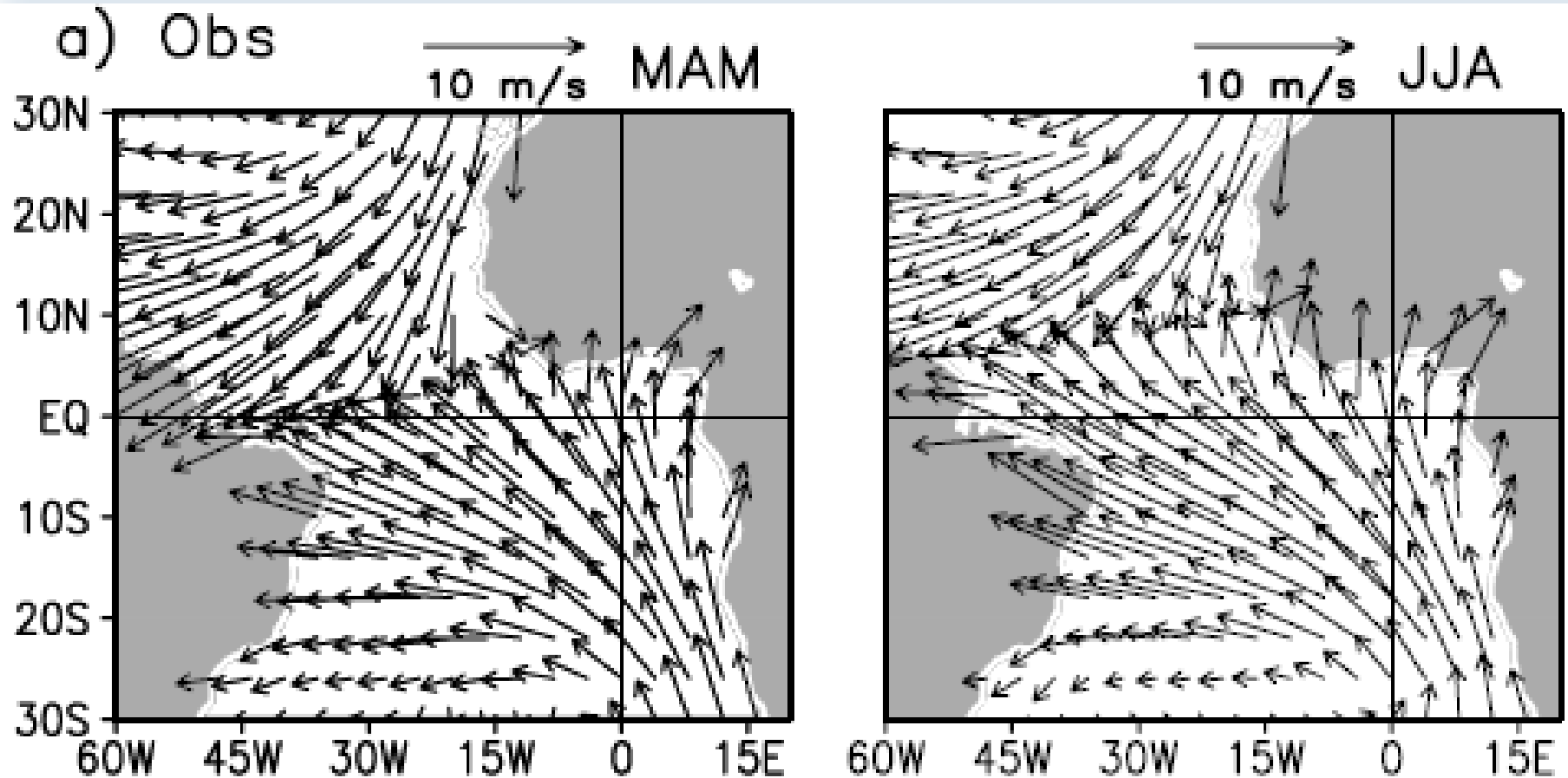
Equatorial southeastern trade winds strengthen in Jun-Jul

Equatorial SST and thermocline depth

a) Obs — MAM / JJA maps

Warmest in MAM; coldest in JJA

4°N-4°S,16°W-4°E

Scatter diagram of monthly mean SST

(Mitchell and Wallace 1992)

MAM   JJA
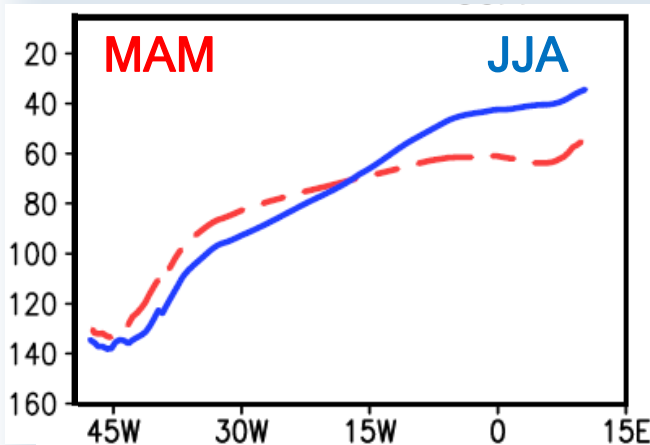
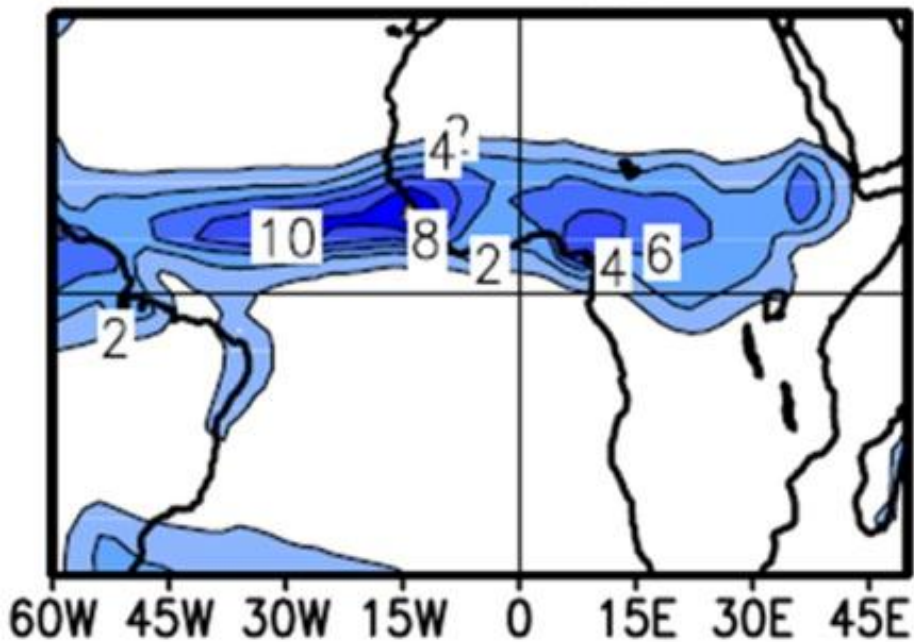Thermocline flat and "deep" in MAM and steep and "shallow" in JJA

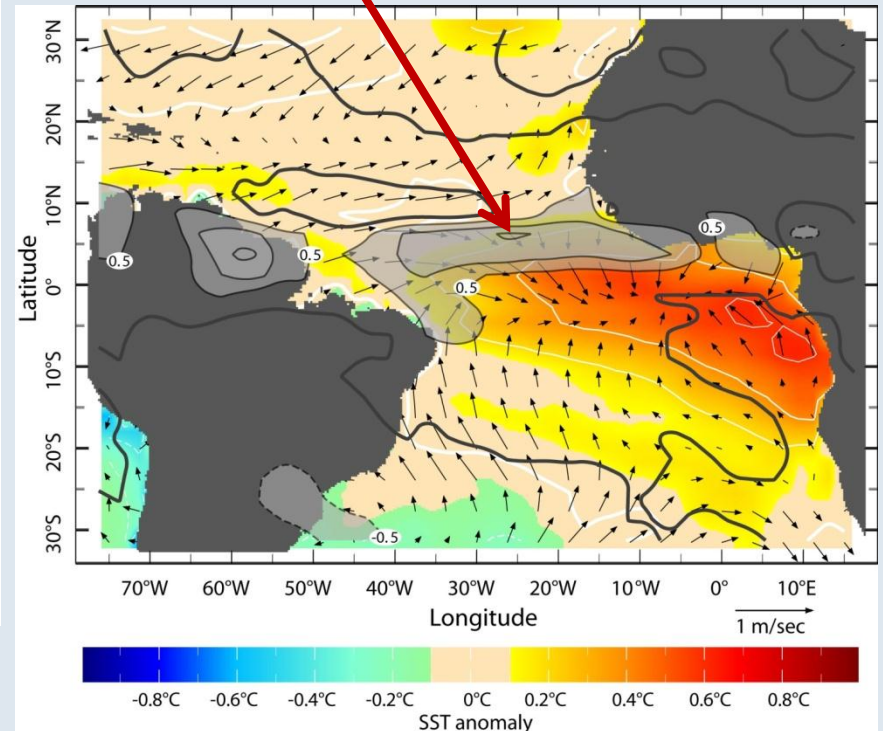Depth of thermocline (averaged between 3°S-2°N)

Interannual SST and Climate Variability

Recall Peter Brandt's lecture: Interannual variability of ACT SSTs is tied to interannual variability of rainfall over the adjacent continents

Average precipitation [mm/day] during boreal summer

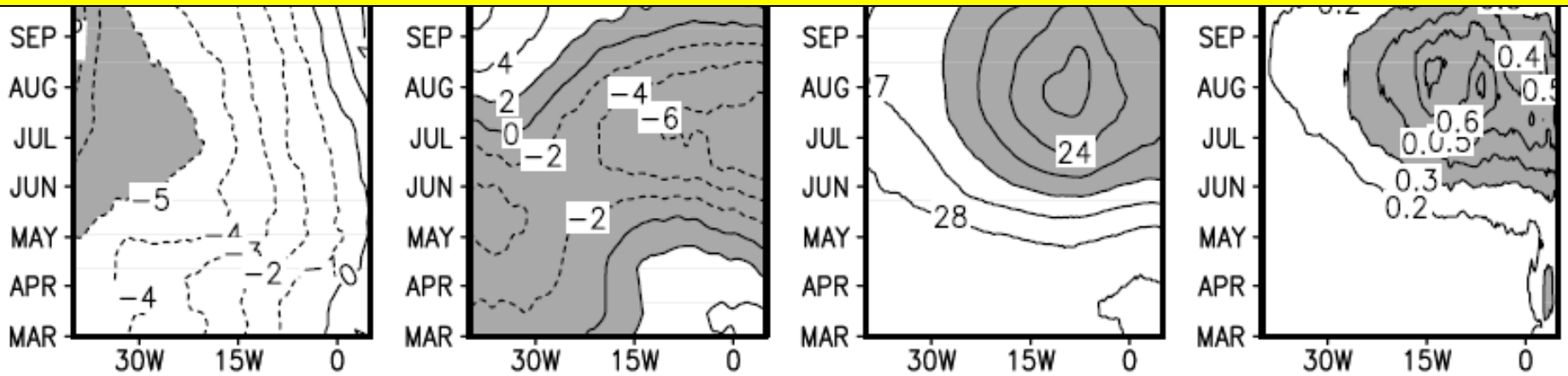First EOF of interannual variability of boreal summer precipitation [mm/day]

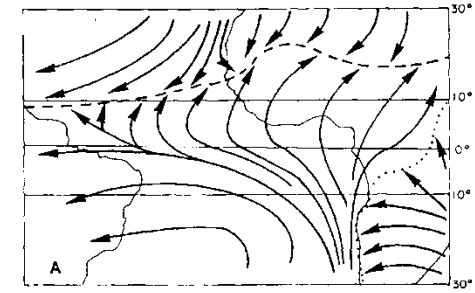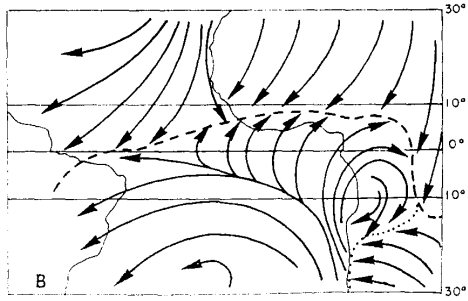# Seasonal cycle of "equatorial upwelling"

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Seasonal cycle of equatorial (a) zonal winds, (b) sea surface height, (c) sea surface temperature, and (d) chlorophyll-a
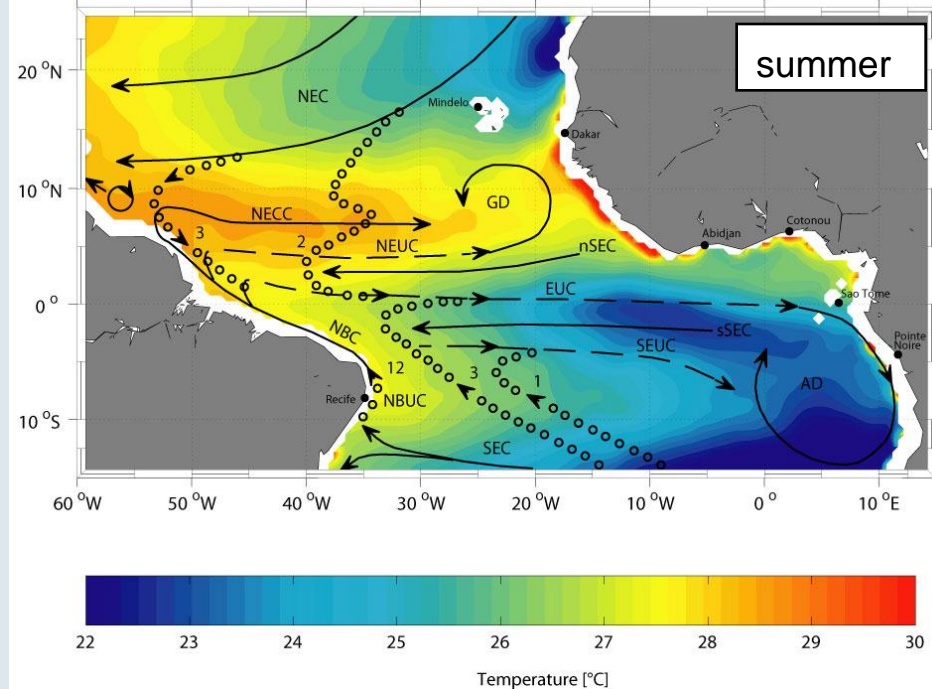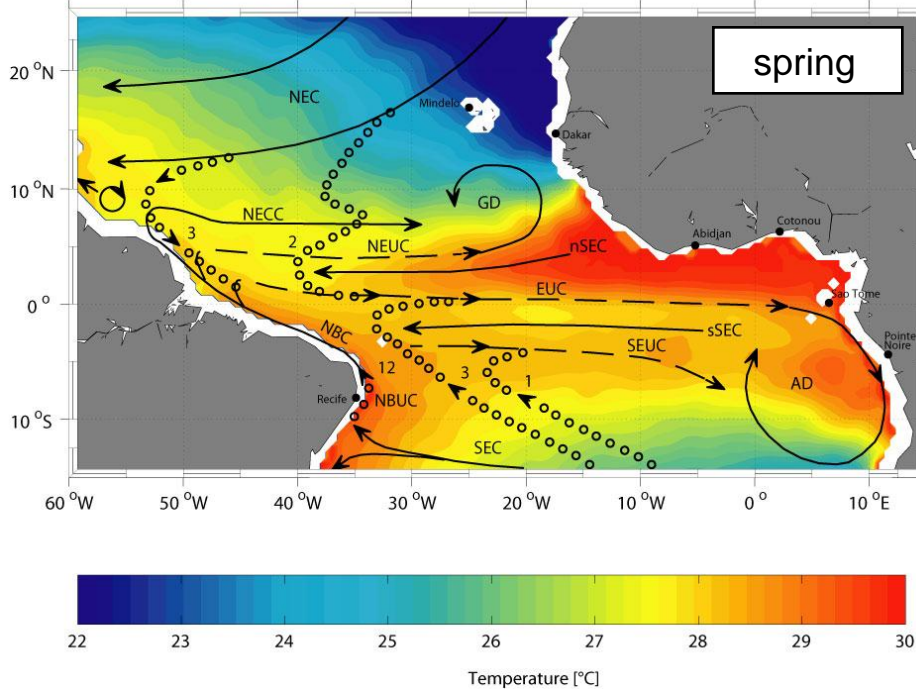

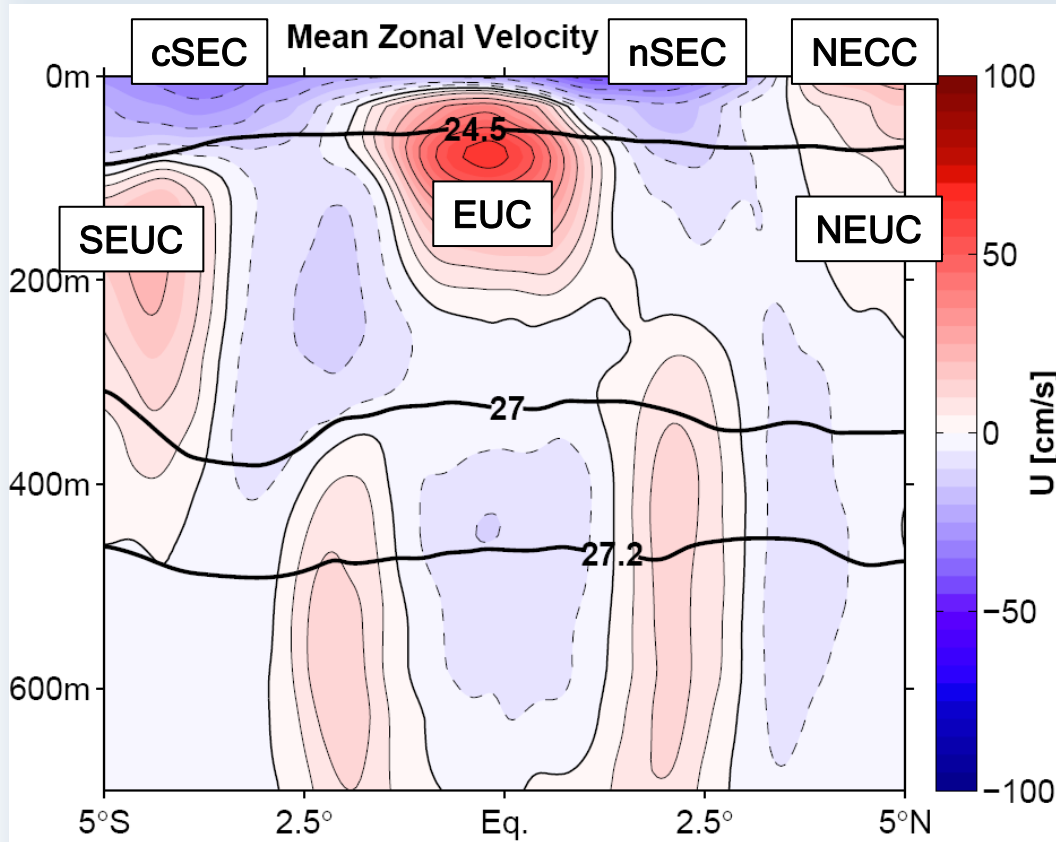## Why does the mixed layer cool during boreal summer?

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(Grodsky et al. 2008)

Circulation and SST variability

Atlantic STC and TRMM–TMI SST March – April 2005 — spring

Atlantic STC and TRMM–TMI SST June, July, August 2004 — summer

Eastward undercurrents (EUC, SEUC, NEUC) supply recently subducted waters from the western boundary to the upwelling regions

———— surface currents

– – – undercurrents

Equatorial circulation

Mean Zonal Velocity

cSEC | nSEC | NECC | EUC | SEUC | NEUC

24.5 | 27 | 27.2

5°S | 2.5° | Eq. | 2.5° | 5°N

U [cm/s]

The zonal current system at the equator cause elevated meridional and vertical velocity gradients

westward currents:
cSEC - central South Equatorial Current
nSEC - northern South Equatorial Current

eastward:
EUC -   Equatorial Undercurrent
NECC - North Equatorial Countercurrent
SEUC - South Equatorial Undercurrent
NEUC - North Equatorial Undercurrent

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Meso-scale eddy variability
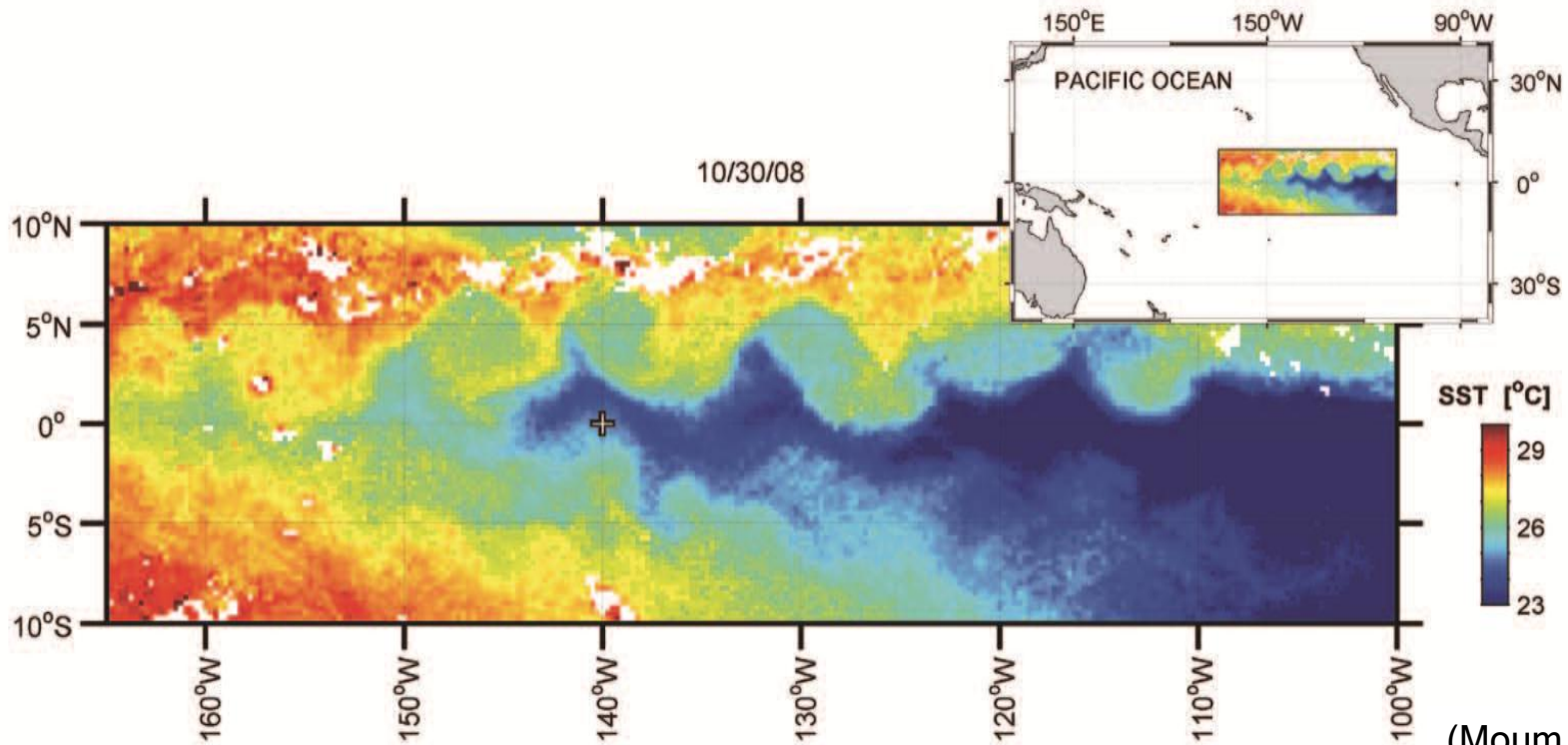
(Moum et al. 2009)

Snap shot of sea surface temperature from the equatorial Pacific (30 October 2008). Surface signature of tropical instability noted as cusps north of equator at 10 degrees zonal wavelength.
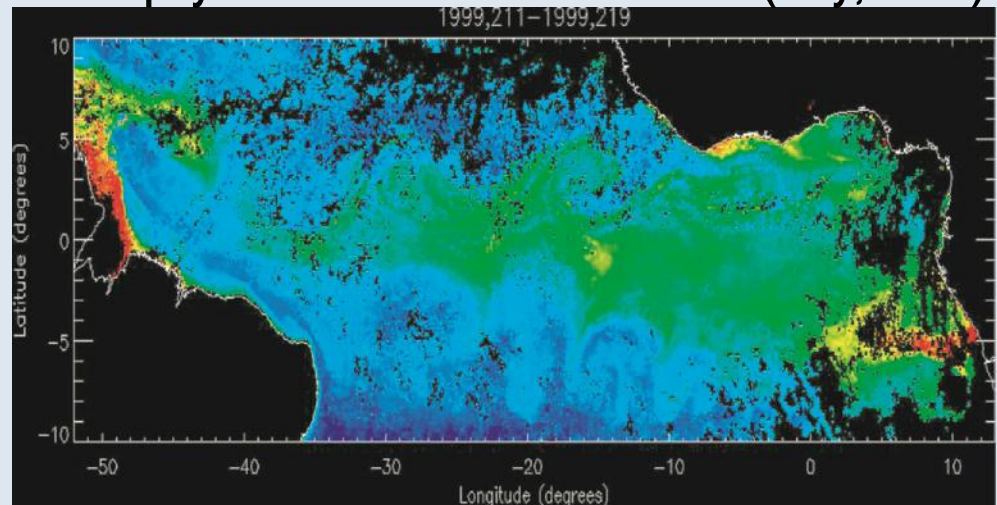
Tropical instability waves

- intraseasonal fluctuations with periods between 20-60 days and wavelength of 400-1500 km

- propagate westward with velocities of 25-50 cms^{-1}

- pronounced between May and October
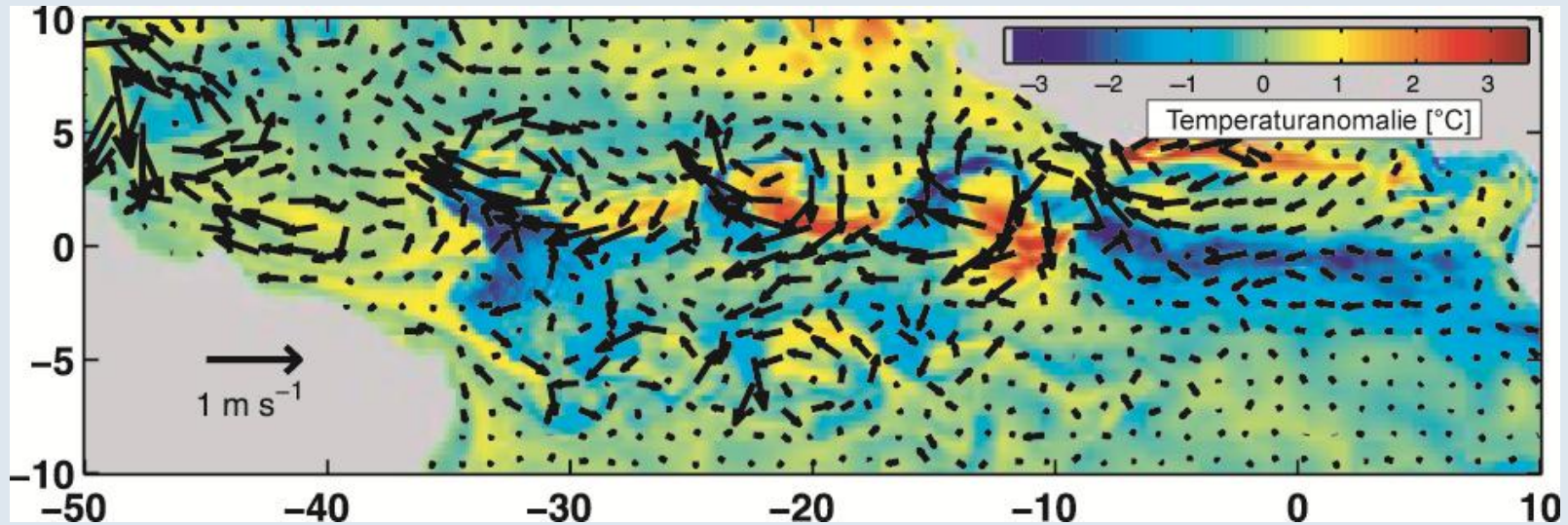
Atlantic's sea surface temperature (TRMM-SSTs)

19JUL – 27JUL2002
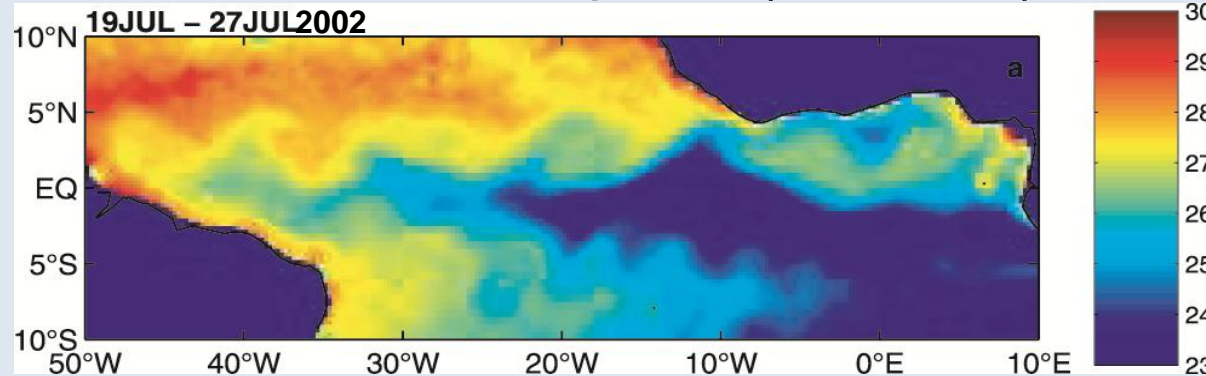
(Grodsky et al., 2005)

Chlorophyll distribution from Sea WIFS (July, 1999)

1999,211–1999,219

(Jochum et al., 2003)

Tropical instability waves in models

Snapshot of anomalous mixed layer currents and temperature
(MICOM model)

(Foltz et al., 2004)

Similar to the observations, TIWs in models are represented as
anticyclonic eddies that propagate westward.

TIW composite (Pacific)

▸ velocity measurements from ships and trajectories from floats were used to construct a synoptic data grid of a TIW at 140°W in the Pacific.

▸ the anticyclonic vortex had a diameter of 500-600km and translated westward with at 30 cms^{-1} (1/4° per day).

▸ Associated currents were in the order of 40 cms^{-1}
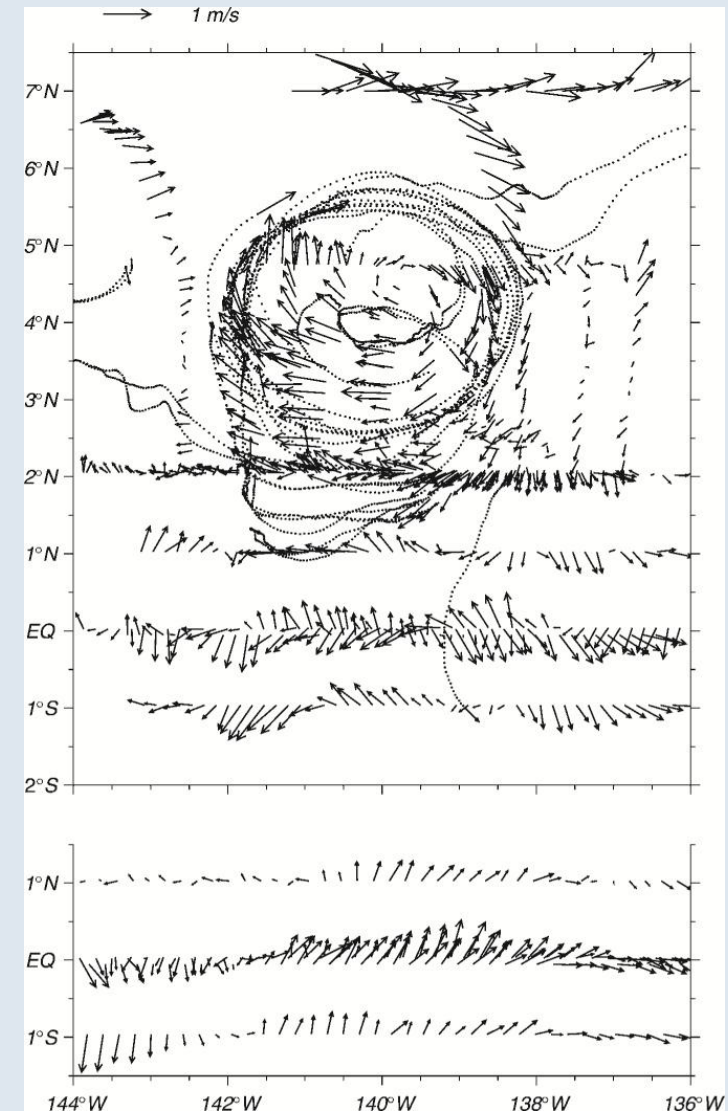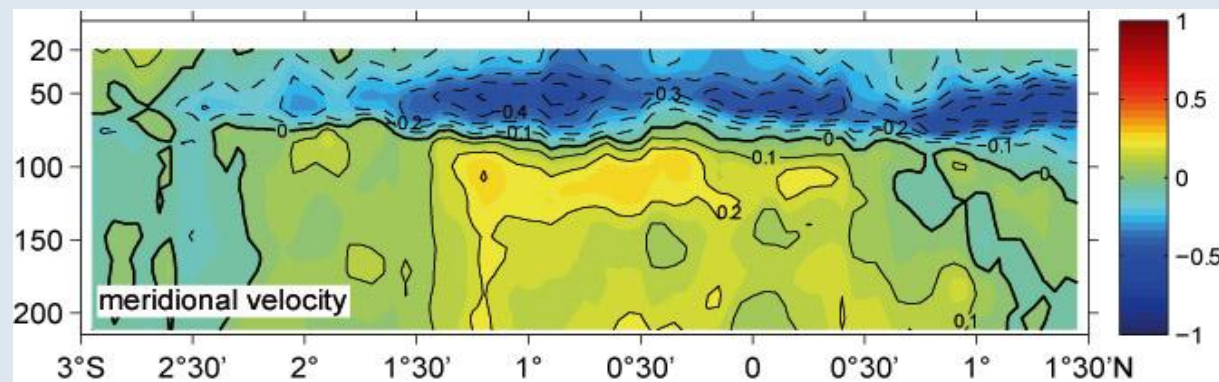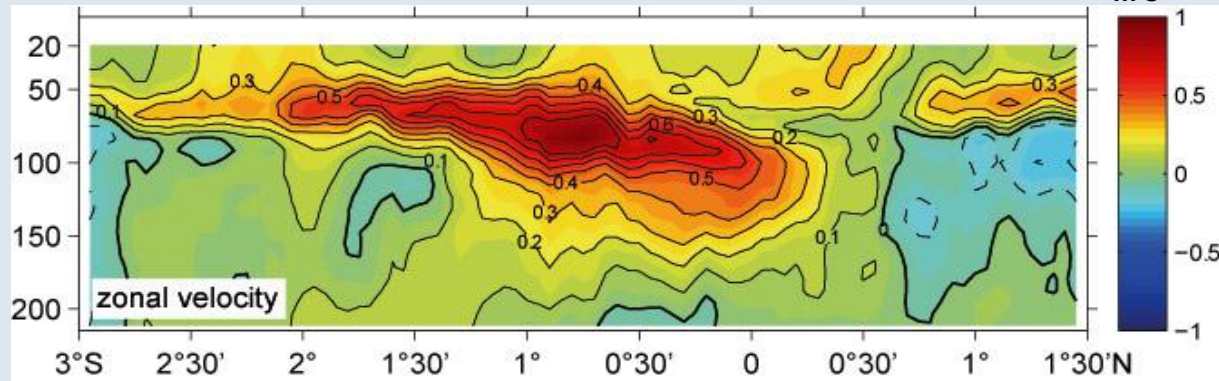

(Kennan and Flament, 2000)

Tropical Instability Waves (Atlantic)

SST and near surface currents
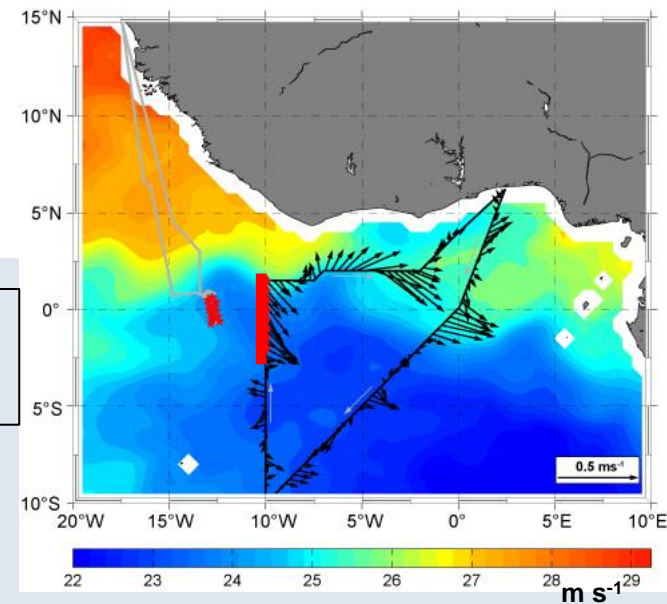
- ▸ Meridional velocity section along 10°W showing structure at the equator

- ▸ In meridional velocity flow reversal at greater depth (>70m) indicates baroclinic structure.
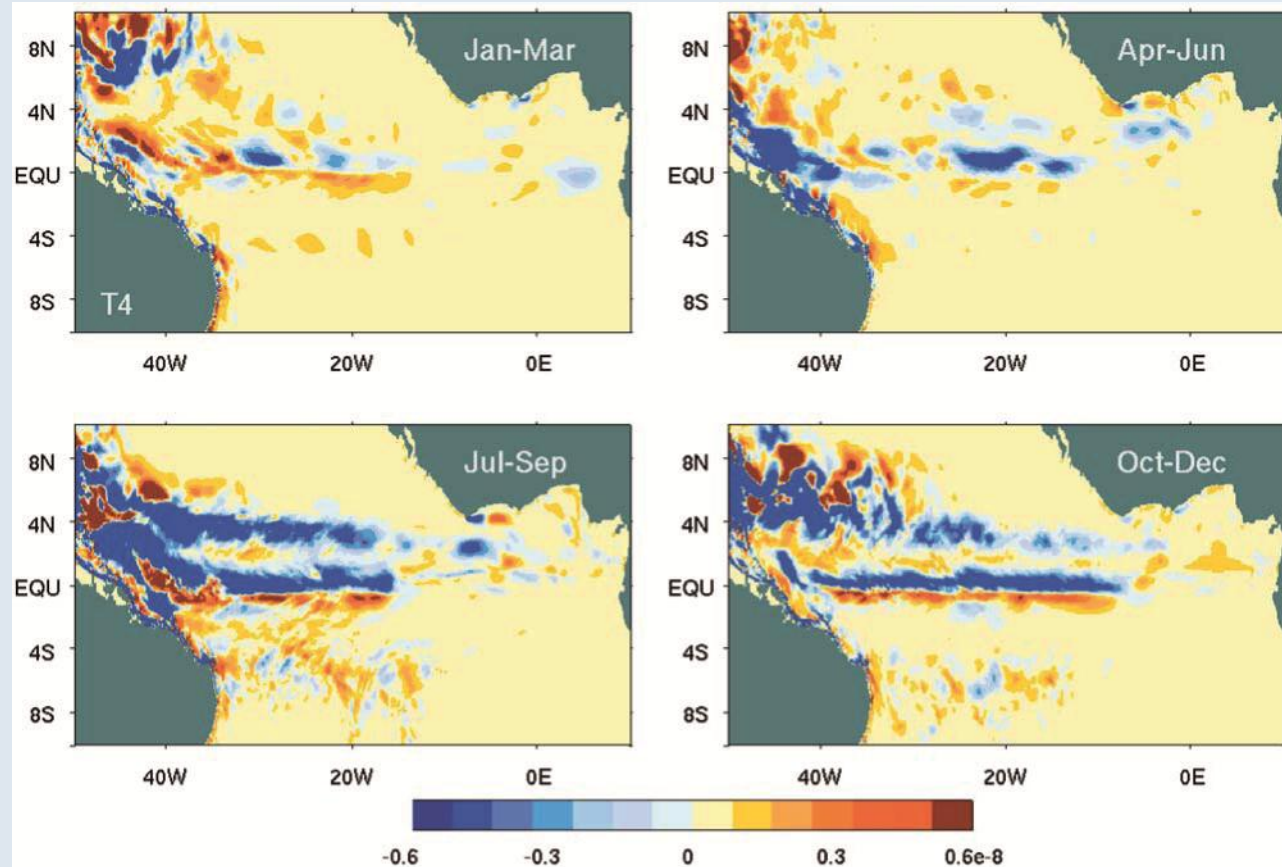

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TIW generation: Barotropic instability

▶ Strength of barotropic and baroclinic instability can evaluated from numerical models.

▶ Barotropic instability in the eddy kinetic energy budget:

$$S = -\overline{\mathbf{u}'u'} \cdot \nabla\overline{U} - \overline{\mathbf{u}'v'} \cdot \nabla\overline{V}$$

$\overline{U}, \overline{V}$ -mean velocity

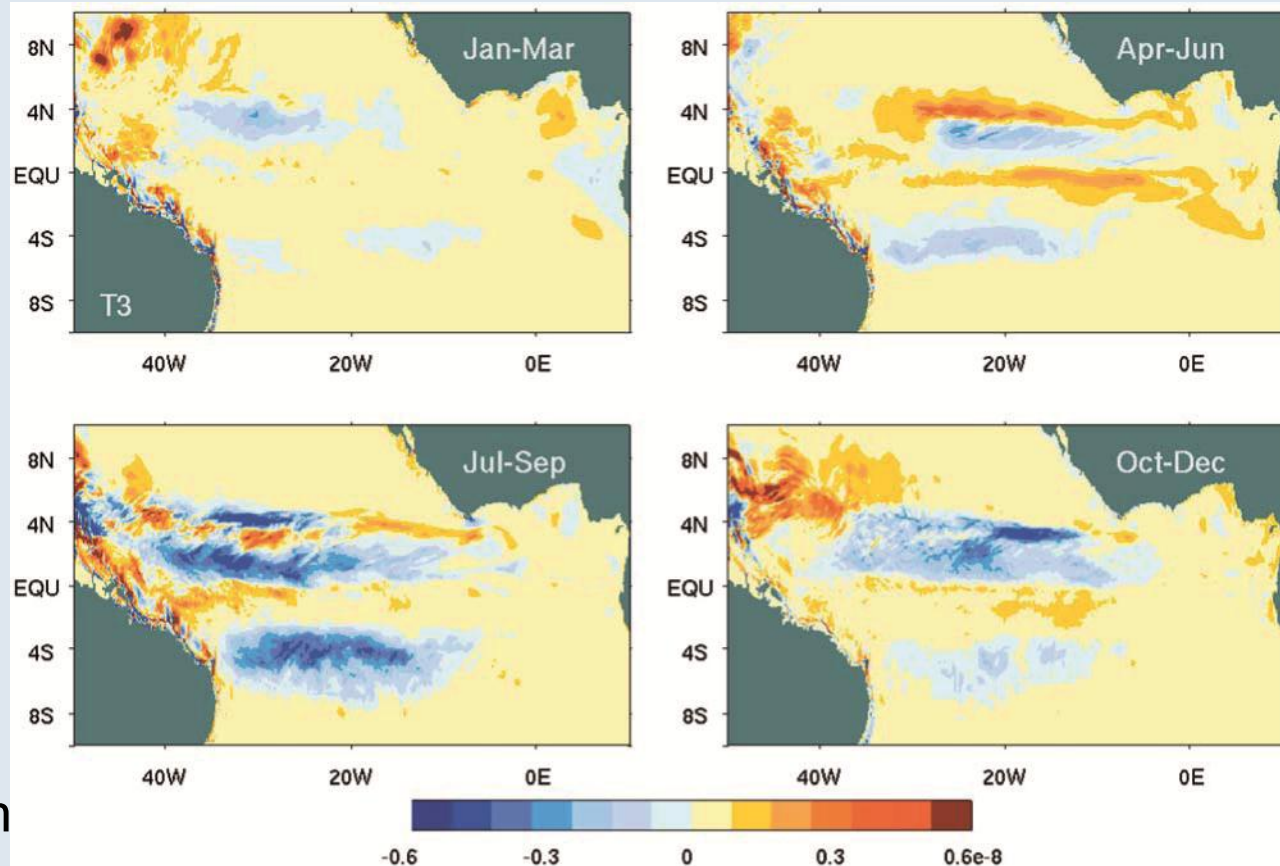u', v' -deviation from seasonal mean

Seasonal maps of the barotropic instability production rate S in [$m^2 s^{-3}$] from a model simulation and averaged over the top 50 m. Negative sign denotes kinetic energy transfer from mean to eddies.

(von Schuckmann et al., 2009)

TIW generation: Baroclinic instability

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▸ Baroclinic instability in the eddy kinetic energy budget:

$$\overline{b'w'}$$

b′ - buoyancy deviation from mean

w′- vertical velocity deviation

Seasonal maps of the baroclinic instability production rate $\overline{b'w'}$ in [m^2s^{-3}] derived from a model simulation and averaged over the top 50 m. Negative sign denotes transfer into the fluctuation.

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▸ Barotropic instability is elevated in the region of strong horizontal velocity shear between EUC and nSEC and between nSEC and NECC

▸ Baroclinic instability is elevated in the region of strong vertical shear (SEUC, upper EUC)
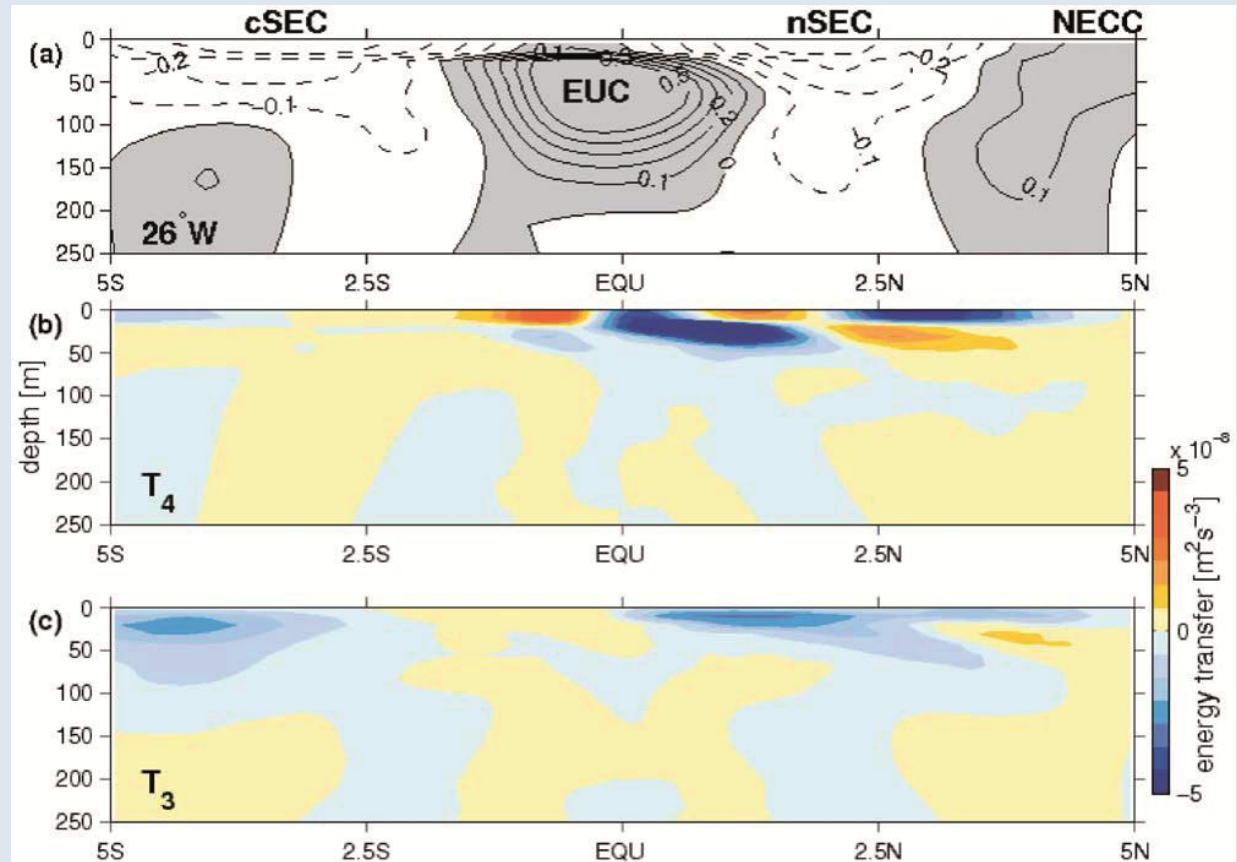
(a) Mean zonal velocity along 26W
(b) Barotropic production rate S averaged from 30°W to 10°W. Negative signs denote transfer into EKE.
(c) Corresponding average of the baroclinic production term b'w'. Negative signs denote transfer into EKE.

SSH variability and equatorial waves

- SSH variability (3°N-3°S) can be separated into an equatorial-symmetric and -antisymmetric component

- symmetric variability exhibits characteristics of an first meridional mode Rossby waves

- antisymmetric variability shows enhanced energy for baroclinic (3rd or high vertical mode) mixed Rossby-Gravity waves (Yanai waves)
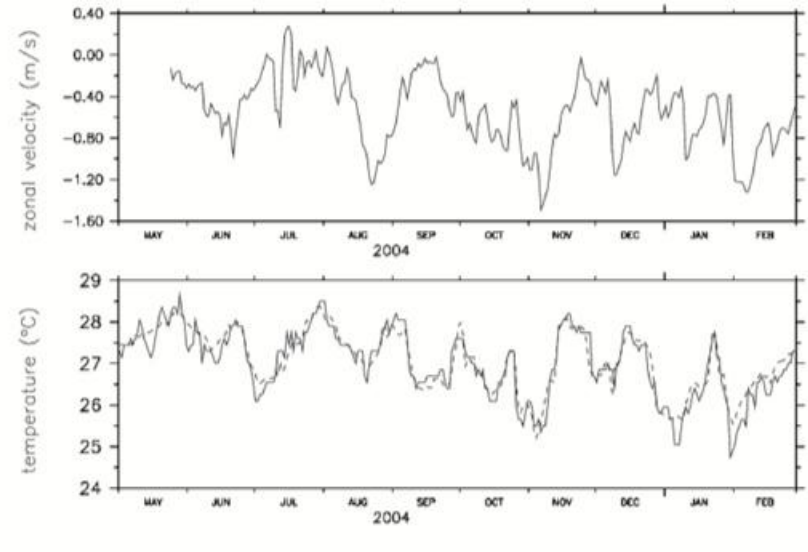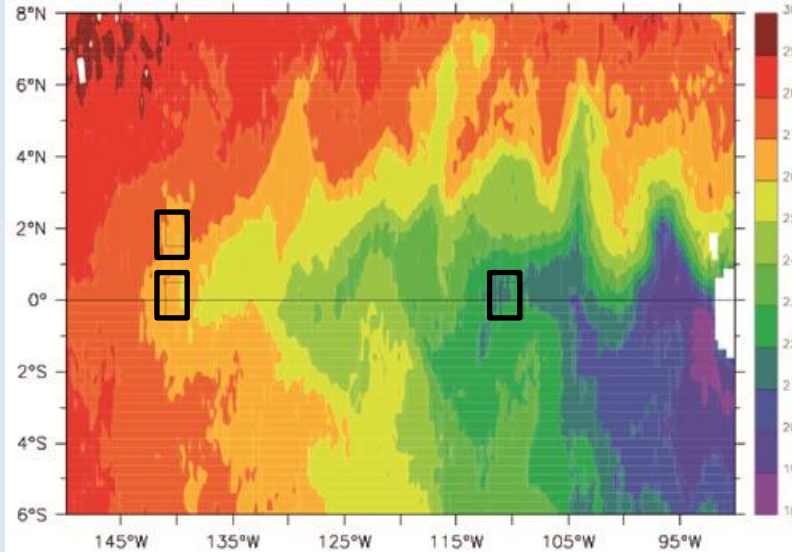
symmetric SSH variability

anti-symmetric SSH variability
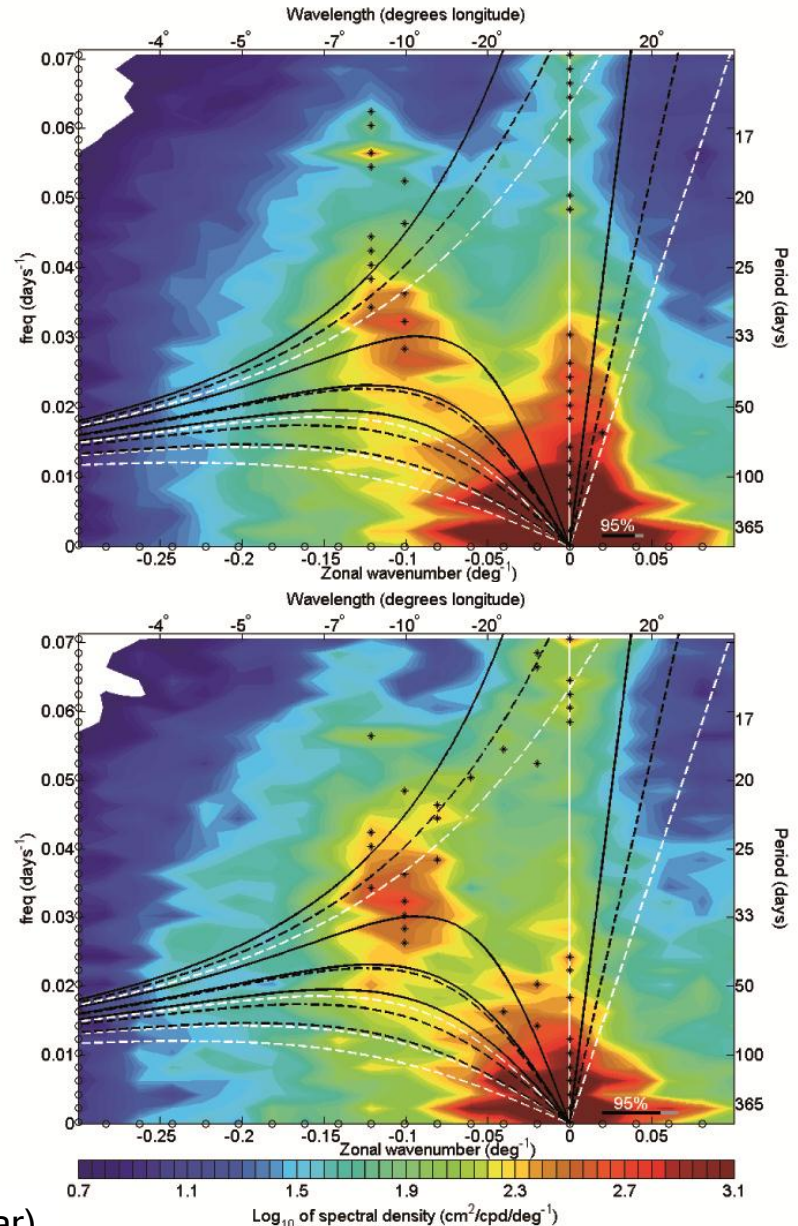
(pers. comm. T. Farrar)

TIW-steering of heat in the mixed layer

(Jochum et. al. 2007)

Zonal velocity at 10m depth from 2°N, 140°W (top)
Temperature from TMI (solid line) and at 20 m depth (bottom)

▸ In the presents of horizontal gradients of temperature (e.g. during boreal summer when the cold tongue is pronounced), TIWs temporarily displace warm water into the region of cold water and visa versa.

▸ When the warm water has cooled when it is returned to its initial position after the TIW has passed, a net heat flux into the cold water region has occurred that can be expressed as $\left\langle \overline{U' \cdot \partial_h T'} \right\rangle \neq 0$

U'-TIW velocity, T'-Temperature fluctuations

TIW-steering of heat in the mixed layer
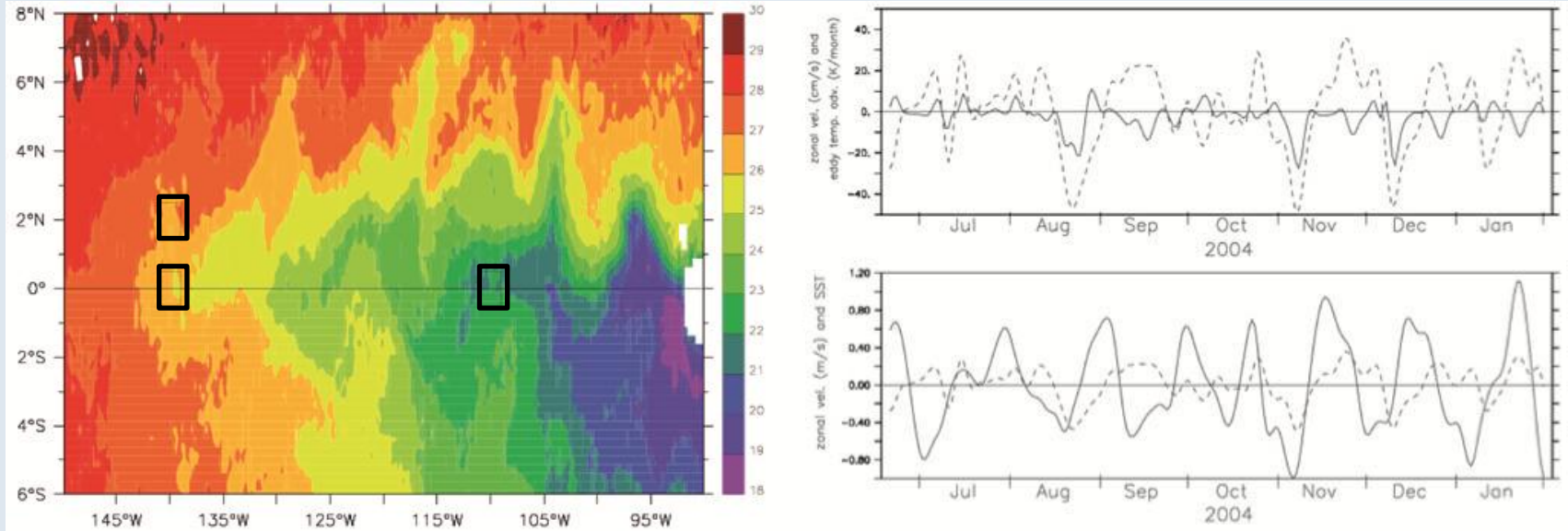
Band-pass filtered zonal velocity (broken line) at 10m depth and zonal eddy temperature advection (top)
Band-passed filtered zonal velocity (broken line) and SST (bottom)

▸ Observations suggest that steering by TIWs leads to a net meridional and to a lesser extend zonal flux of heat into the cold tongue

▸ The heat flux due to the TIW is in the order of 1°-2°/month in meridional and 0.5°-1°/month in zonal direction

(Jochum et. al. 2007)

Advective and eddy heat fluxes in the mixed layer (model)

GEOMAR
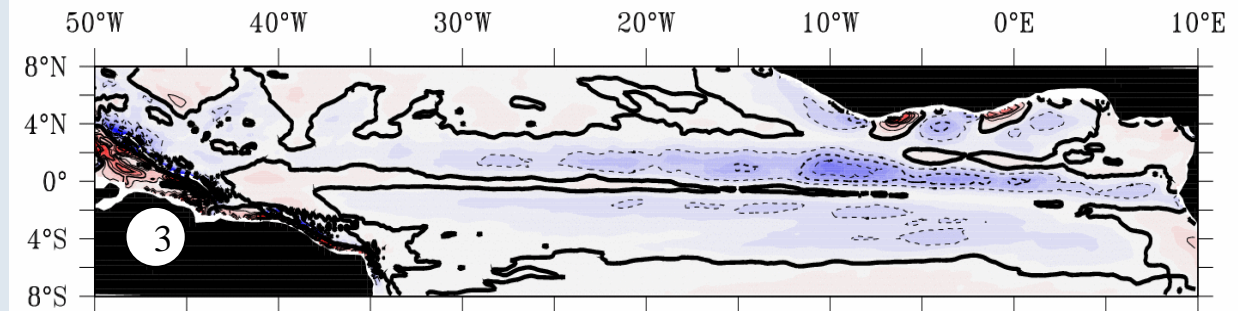
annual mean trends by different processes (°C/month)

zonal advection of heat by low frequency currents
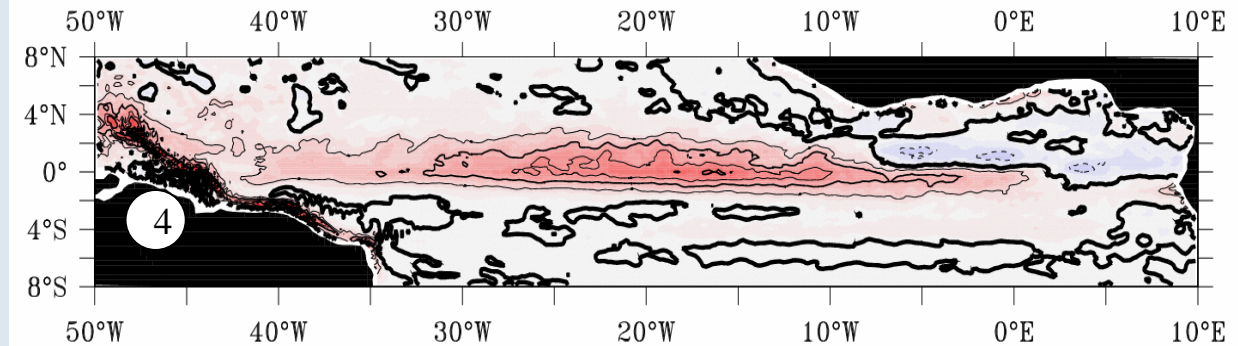
$$\langle -\bar{u}\,\partial_x\bar{T}\rangle$$

meridional advection by low frequency currents

$$\langle -\bar{v}\,\partial_y\bar{T}\rangle$$
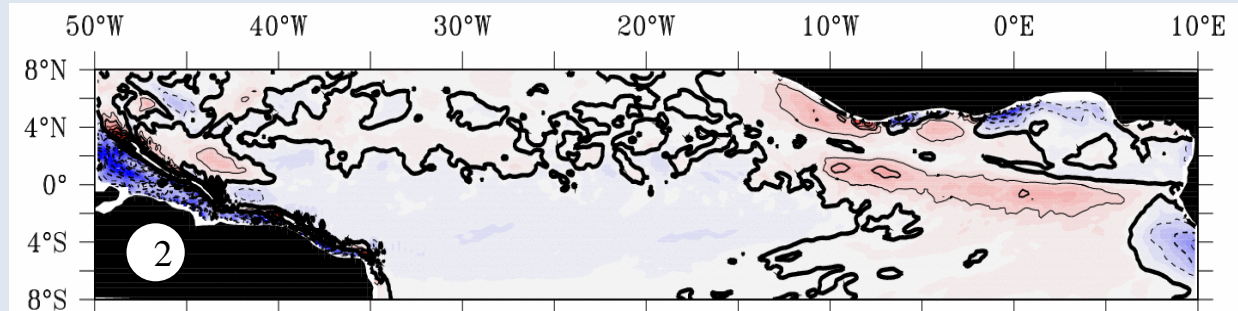
high frequencies advection (<35 days), steering of TIWs

$$-\langle\overline{u'\partial_x T'}\rangle - \langle\overline{v'\,\partial_y T'}\rangle + \langle D_1\rangle$$

D_1-lateral diffusion

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(Peter et al., 2006)

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Early turbulence observations within the equatorial belt

▶ **Equatorial Pacific**

- Microstructure measurements from the upper thermocline in the late 70's and 80's revealed deep-cycle turbulence (Gregg et al., 1985, Moum and Caldwell, 1985)

- Numerous measurement programs since the late 70's

(Peters et al, 1995)

▶ **Equatorial Atlantic**

- 8 microstructure profiles sampled in 1976 indicated enhanced mixing above the EUC core (Crawford and Osborn, 1979)

- Le Noroit cruise by WHOI in late autumn 1994 to the eastern flank of the Mid-Atlantic Ridge (Romache Fracture Zone)

Ocean-microstructure observatories
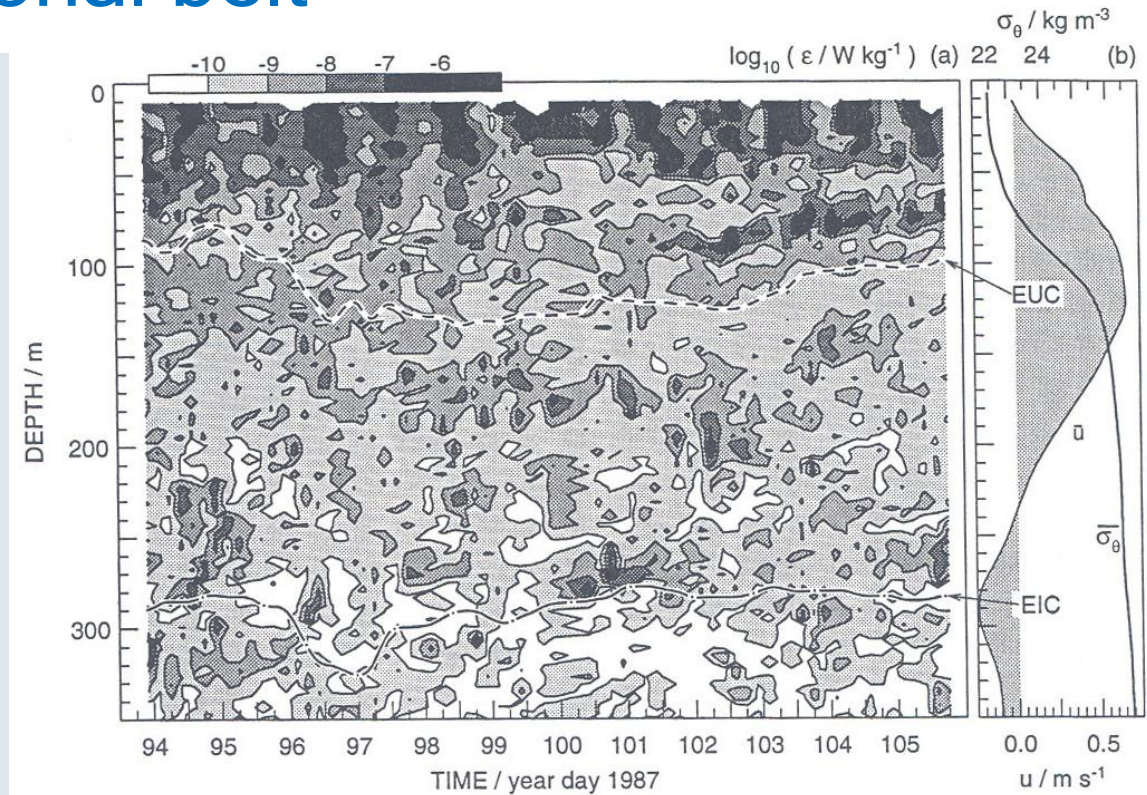
Ship-based microstructure systems
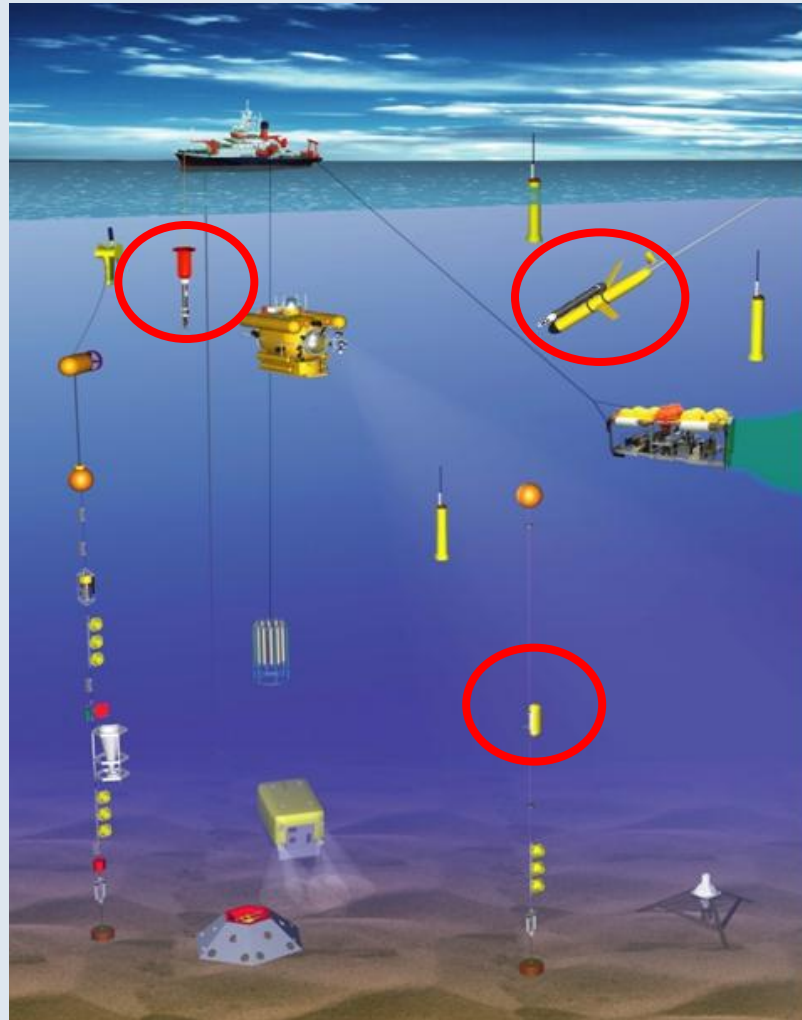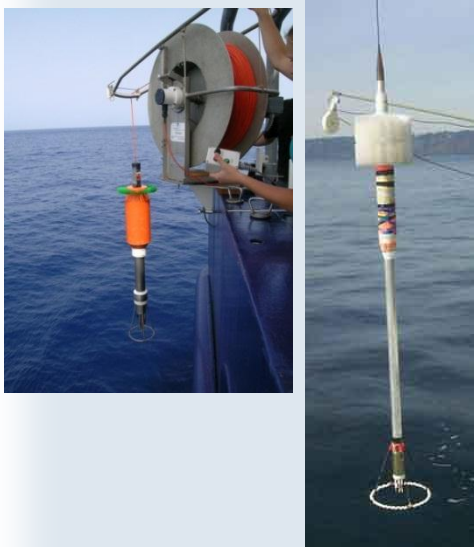
Autonomous microstructure platforms

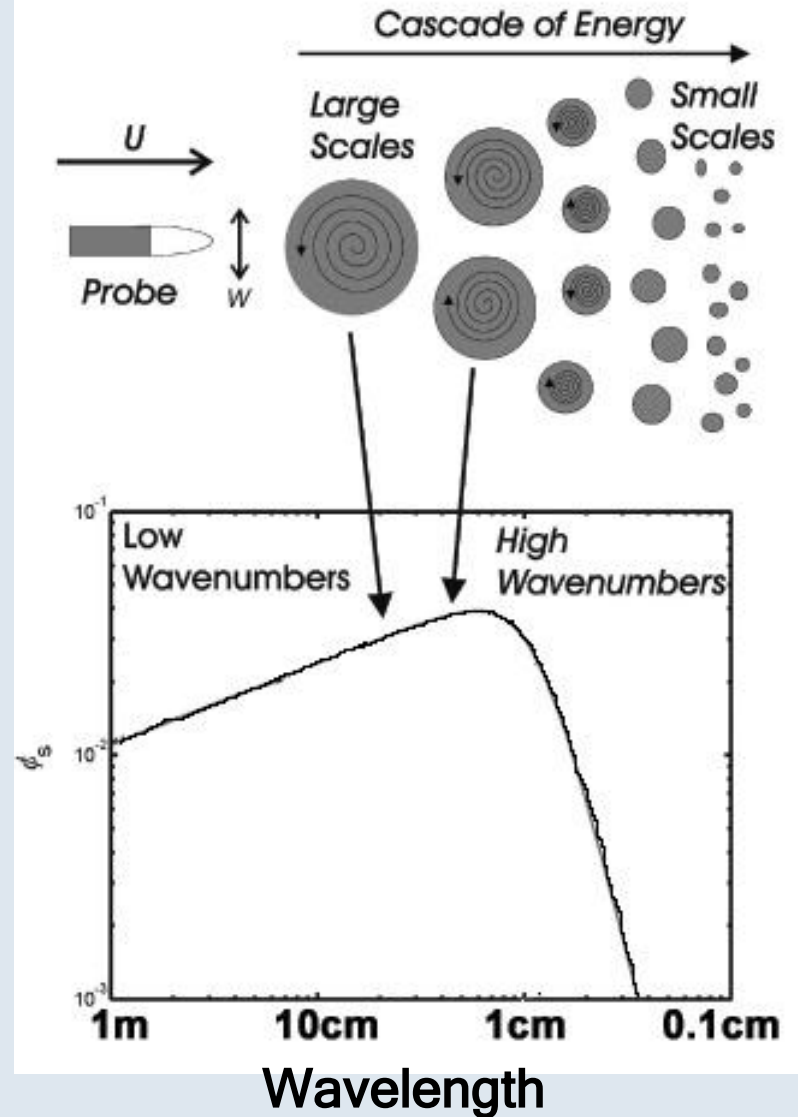**Instrumentation for moorings (χ-pods)**

Microstructure shear sensors
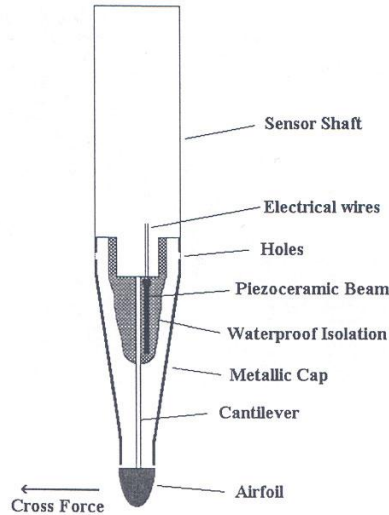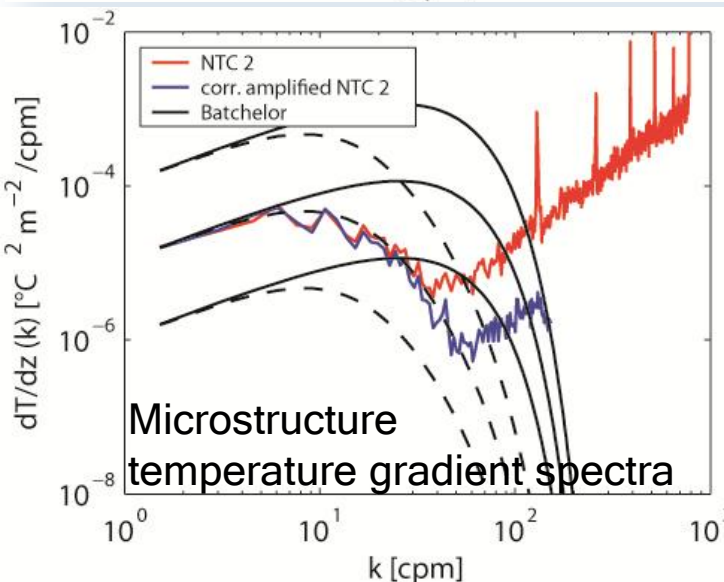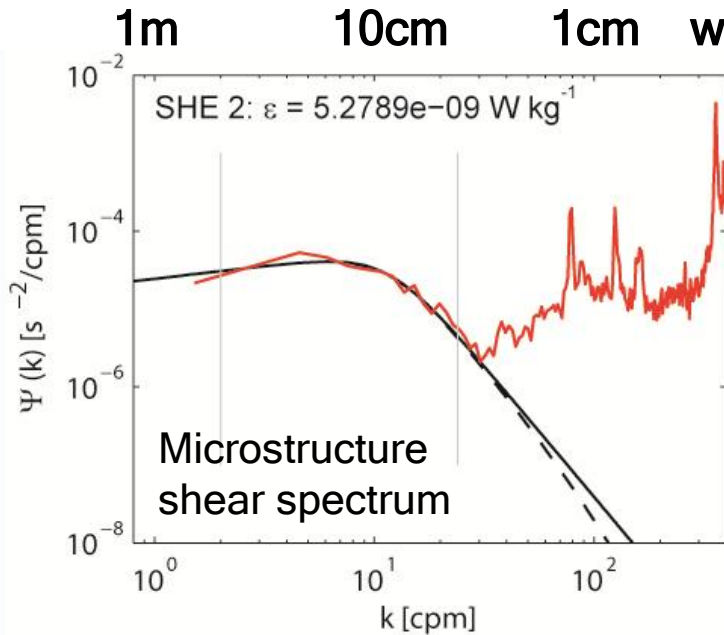

Piezoceramic beam translates tangential force into a potential difference

Sampling frequency 500-1000Hz

Wavelength
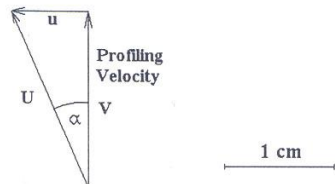
Microstructure data treatment

1m   10cm   1cm   wavelength

Microstructure shear spectrum

Microstructure temperature gradient spectra

Dissipation rate of turbulent kinetic energy is determined from microstructure shear:

$$\varepsilon = \nu \overline{\left(\frac{\partial u'_j}{\partial x_i}\right)^2} = 7.5\nu \overline{\left(\frac{\partial u}{\partial z}\right)^2}$$
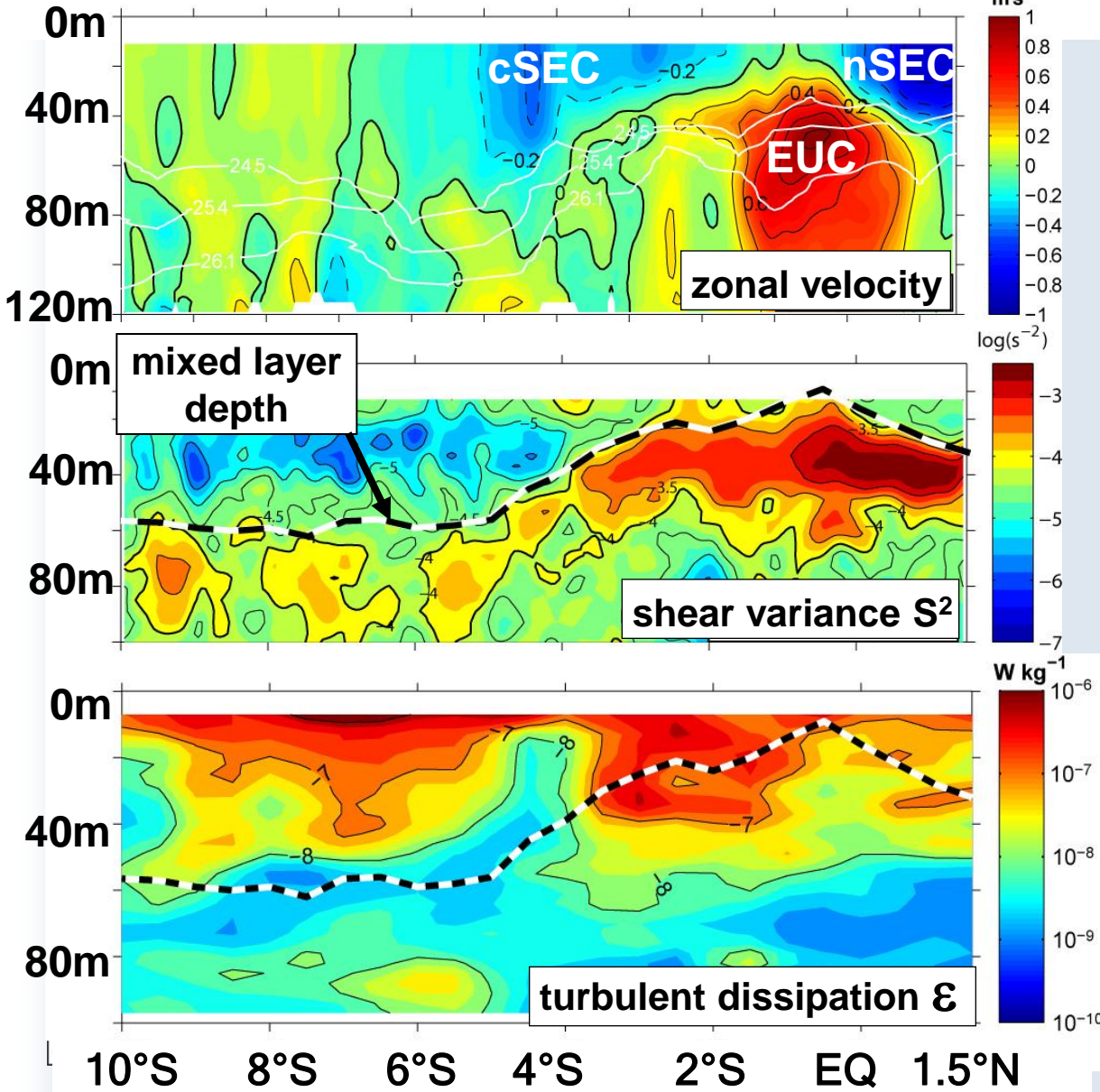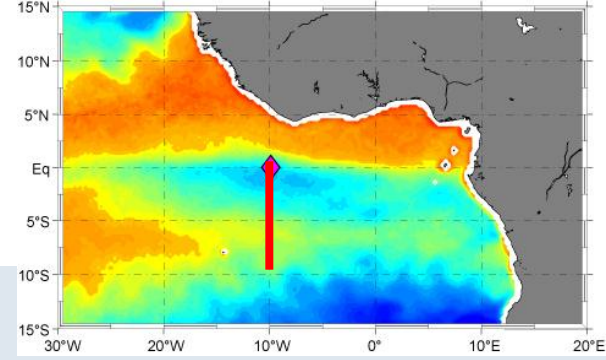
Dissipation rate of temperature variance can be fast-thermistor data:

$$\chi = 2D_T \overline{\left(\frac{\partial \theta'}{\partial x_i}\right)^2} = 6D_T \overline{\left(\frac{\partial \theta'}{\partial z}\right)^2}$$

Eddy diffusivities for buoyancy:

$$K_\rho = \Gamma \frac{\varepsilon}{N^2}, \quad \Gamma = R_f/(1-R_f) \approx 0.2$$

(Osborn, 1980)

# Latitudinal distribution of upper ocean turbulence along 10°W

- ▸ elevated vertical shear of horizontal velocity at the base of the mixed layer extends from 3°S to 1.5°N

- ▸ elevated turbulence levels below mixed layer are found between 3°S and 1°N

- ▸ little mixing in stratified layer below MLD south of 4°N

(Hummels et al., 2013)

# Ship-board microstructure measurements (2005-2011)

SST [°C]
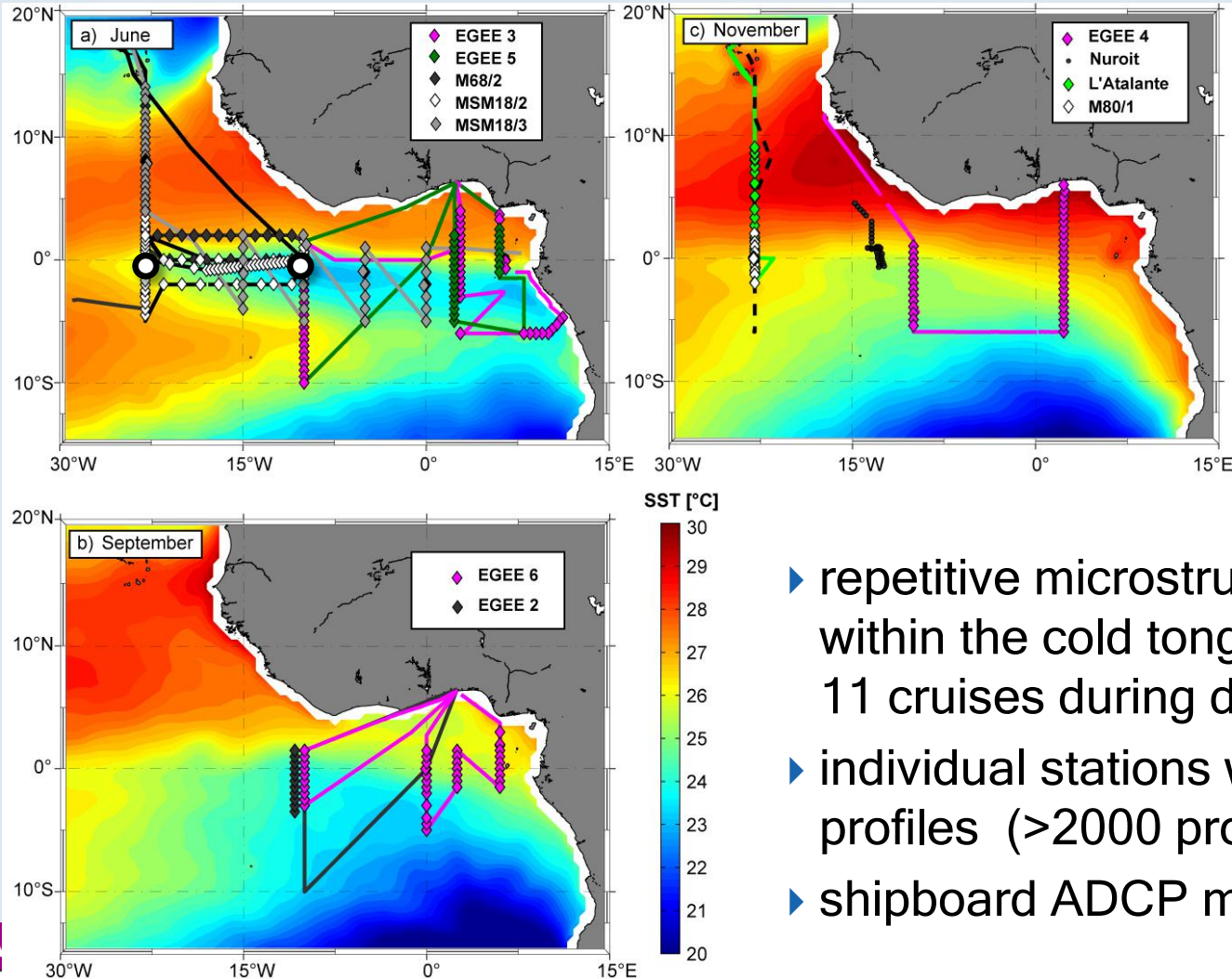
- ▸ repetitive microstructure sections within the cold tongue region from 11 cruises during different seasons
- ▸ individual stations with at least 3 profiles (>2000 profiles)
- ▸ shipboard ADCP measurements

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Christian-Albrechts-Universität zu Kiel
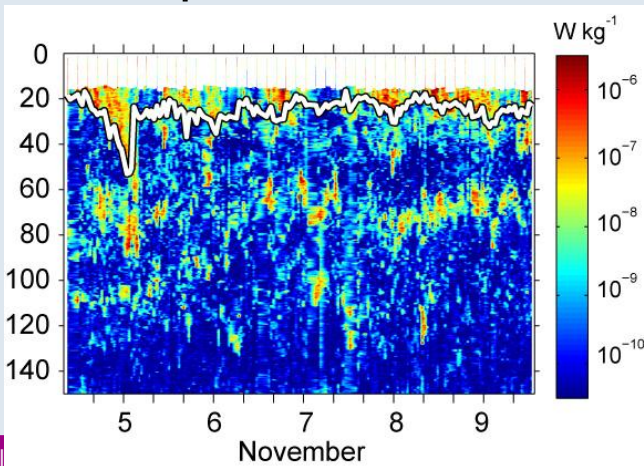
Time series of ε from a MicroRider/glider package

10°W

23°W

W kg^{-1}
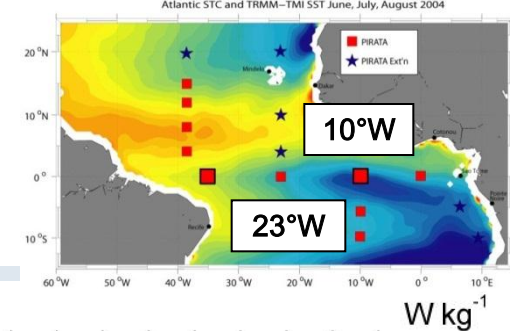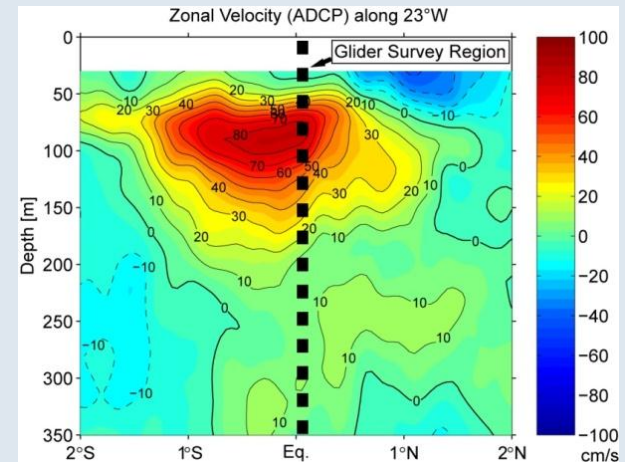
10°W, Equ., May-Jul. 2011
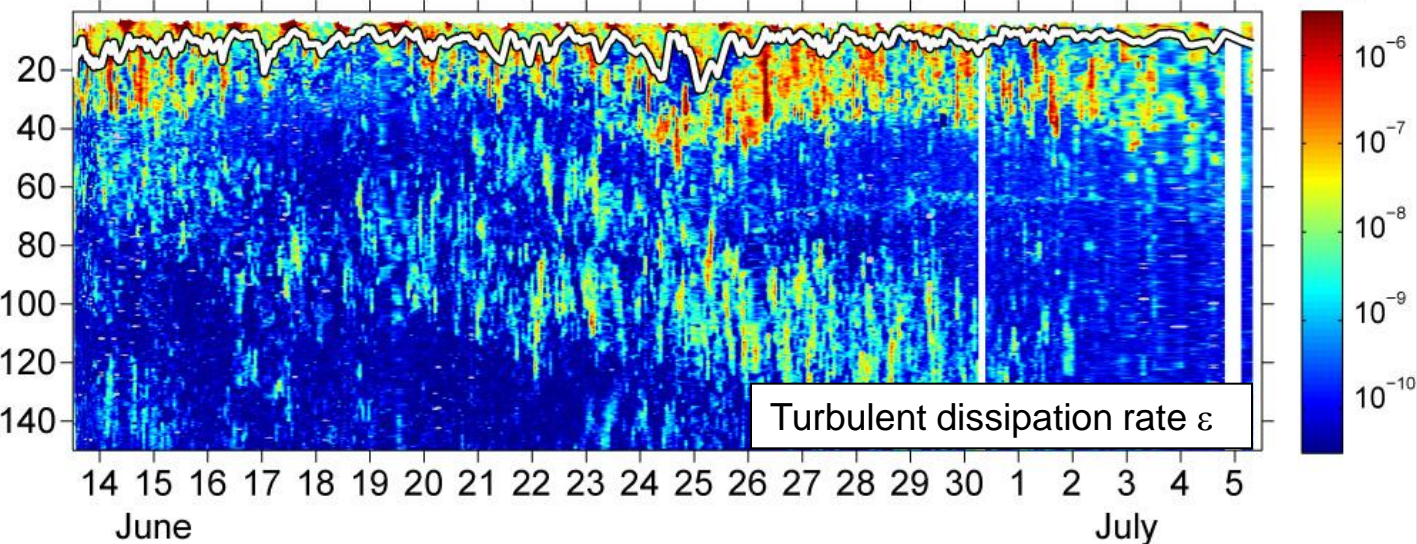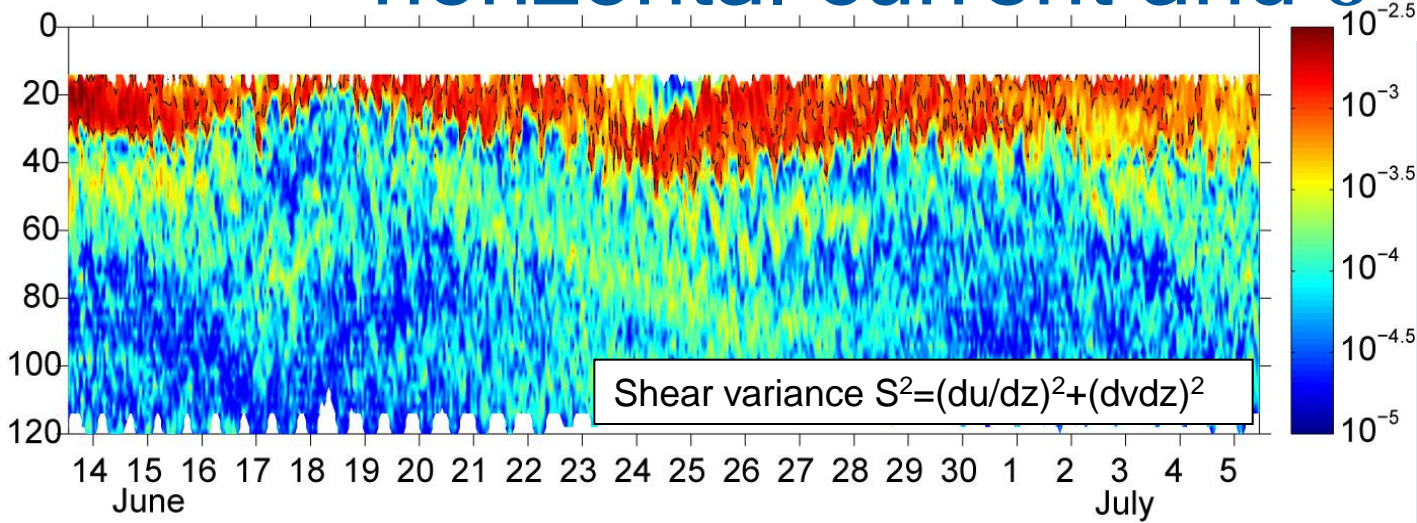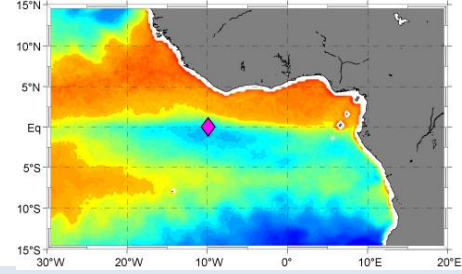
Mixed layer depth
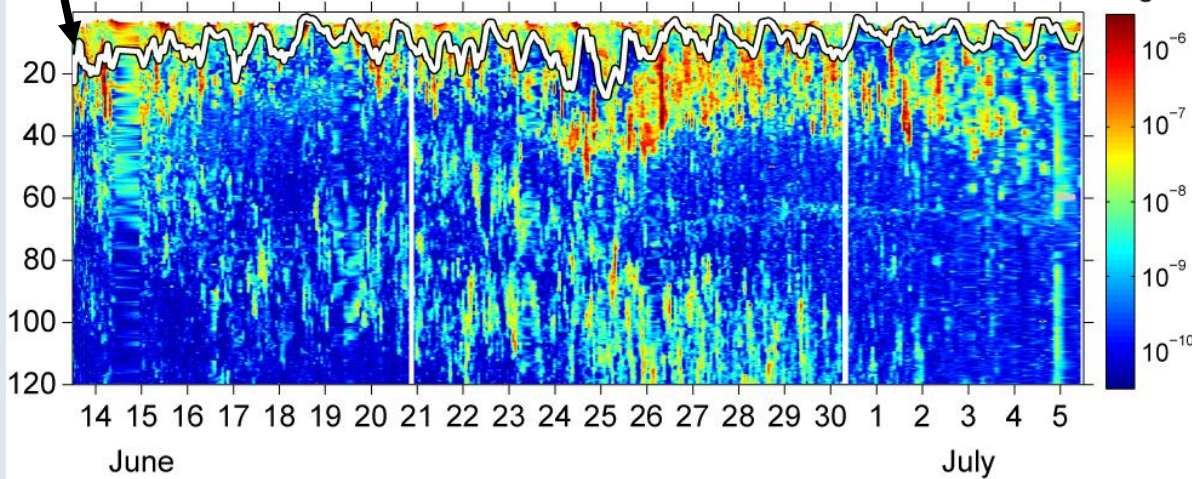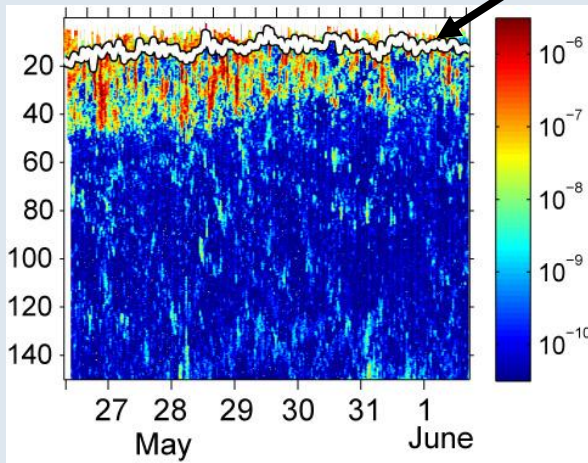
23°W, Equ. Nov. 2009

bursts of elevated turbulence in the upper thermocline occur sporadically and last up to a few hours

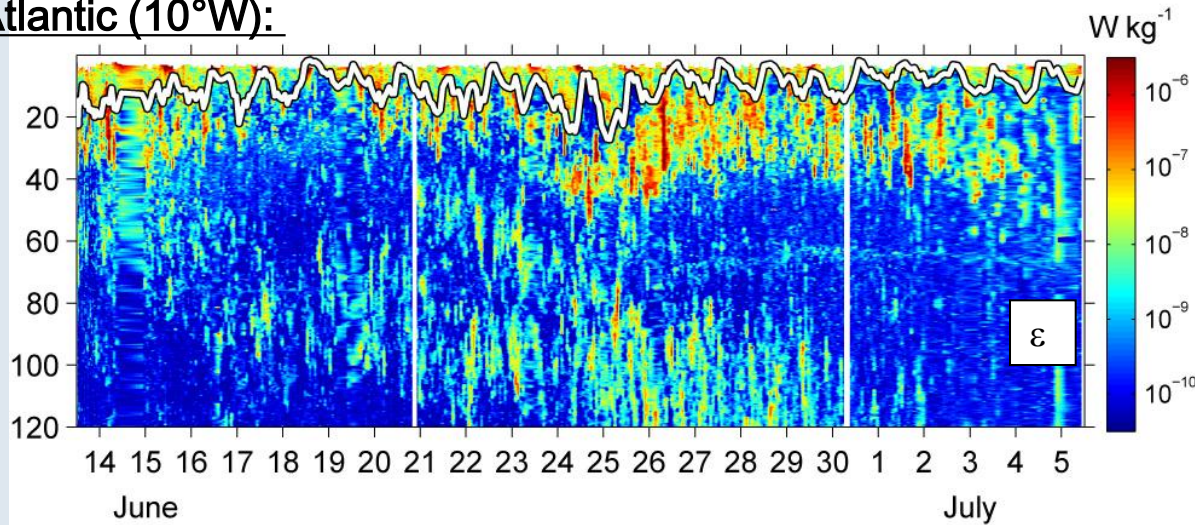# Vertical shear of horizontal current and ε

Shear variance $S^2=(du/dz)^2+(dvdz)^2$

Turbulent dissipation rate ε

▸ elevated dissipation rates coincide with elevated shear variance

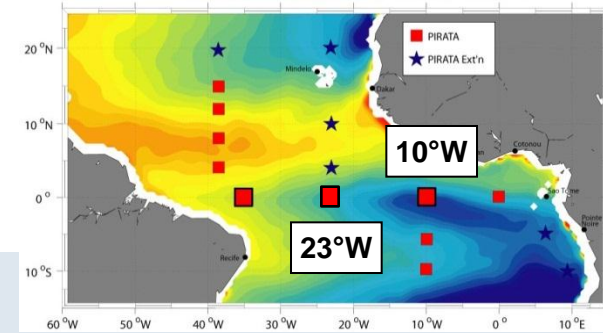# Equatorial turbulent dissipation rate time series

Atlantic (10°W):

Pacific (140°W):

Mixed layer depth

EUC core

(Moum et al., 2009)

▸ elevated mixing region extends deeper in the Pacific

▸ diurnal variability dominant in the upper thermocline
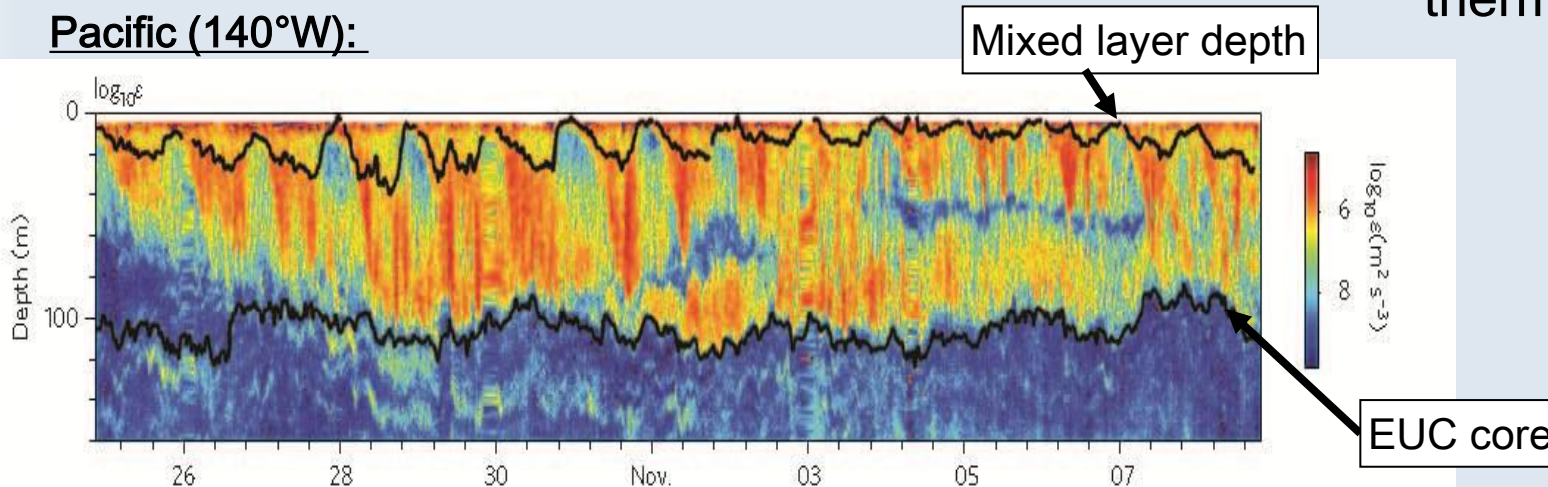
Diurnal cycle of temperature and ε
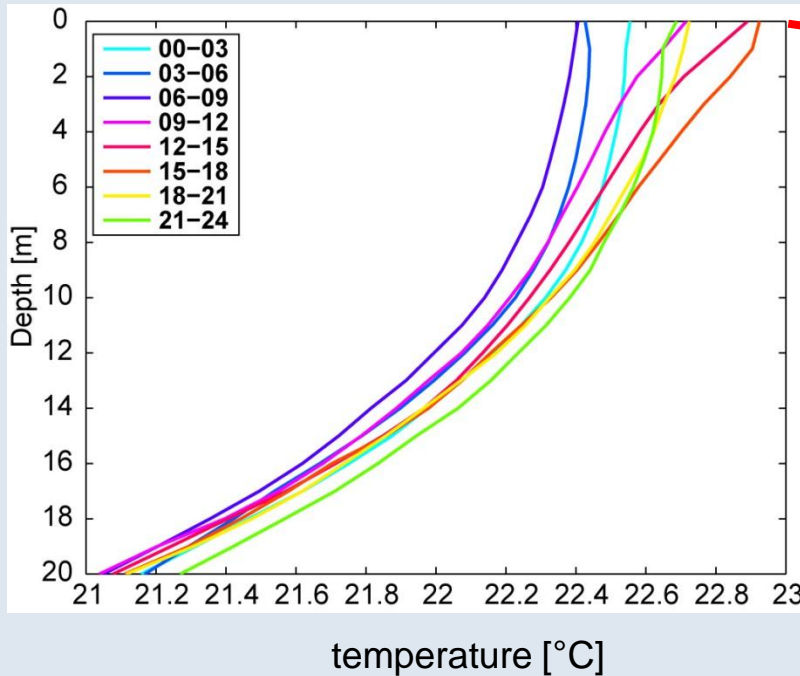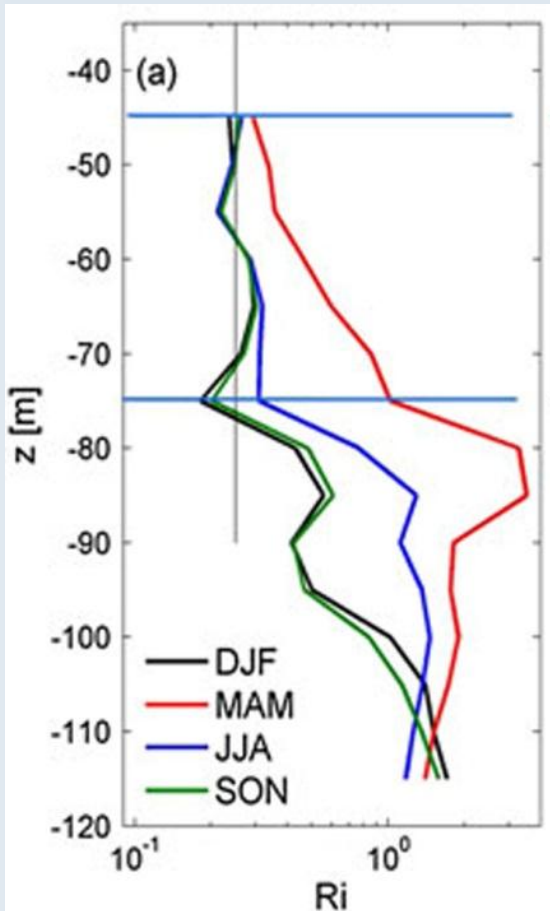

June, 10°W:

temperature [°C]
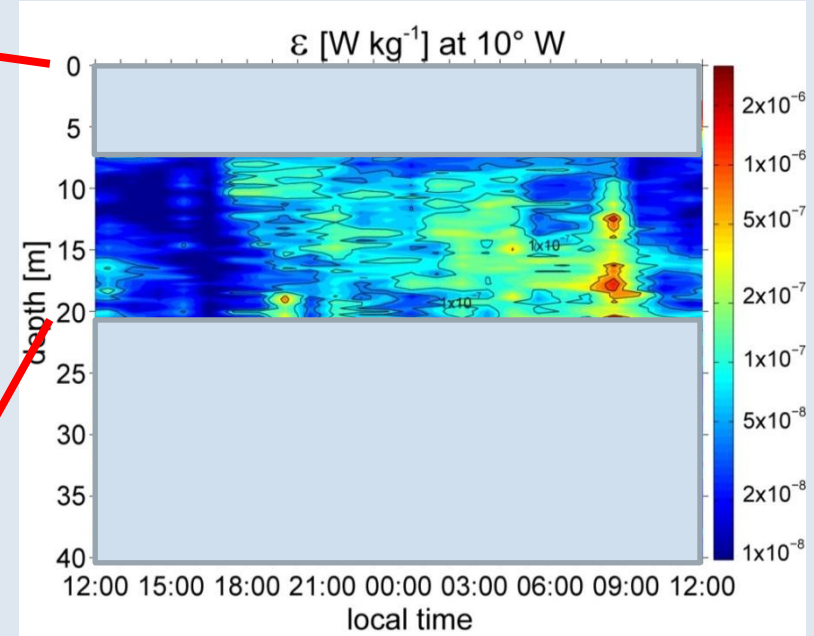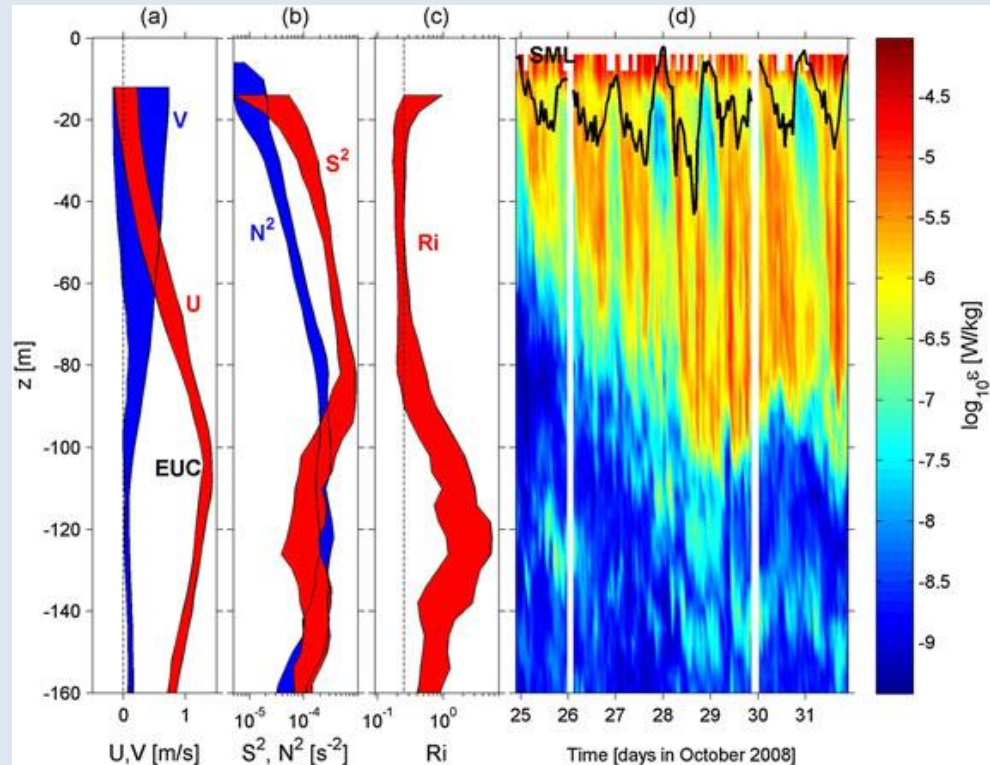

▸ turbulence in the stratified upper ocean shows a 100-fold increase during night time (21:00-9:00) in the equatorial Atlantic

Marginal Ri-instability of the EUC

(Smyth et al., 2013)

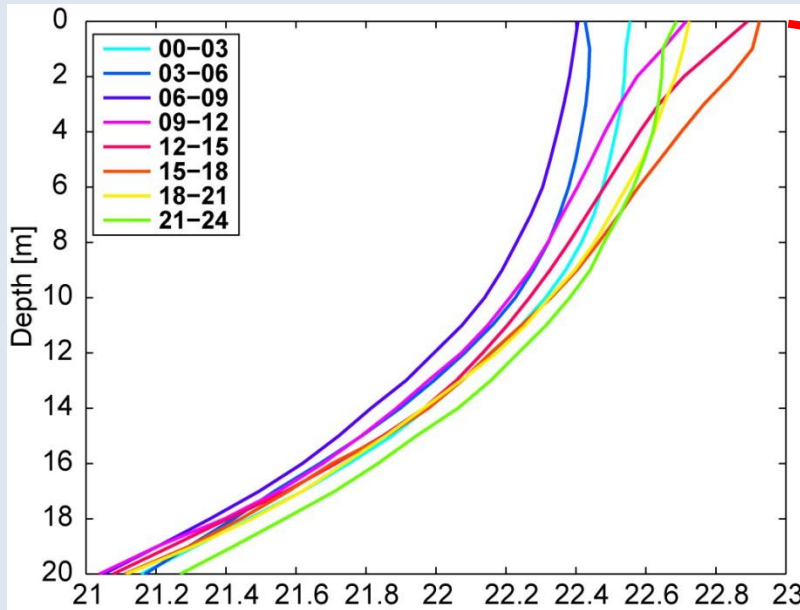<u>Recall:</u> Necessary condition for Kelvin-Helmholtz Instability to develop:

$$Ri = \frac{N^2}{S^2} < 0.25$$

▸ During most of the season (except in spring) the Richardson Number (Ri) above the core of the EUC and below the mixed layer is 0.25, i.e. marginally unstable!
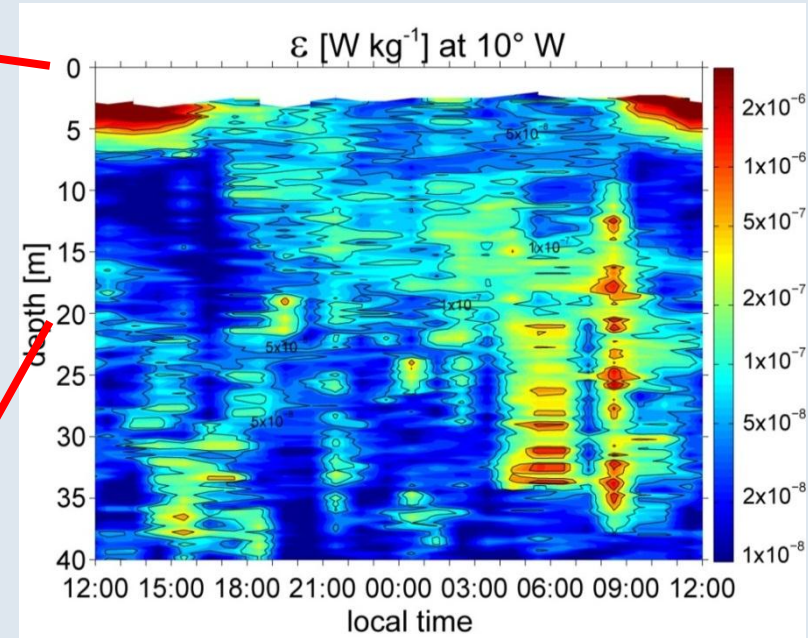
Christian-Albrechts-Universität zu Kiel
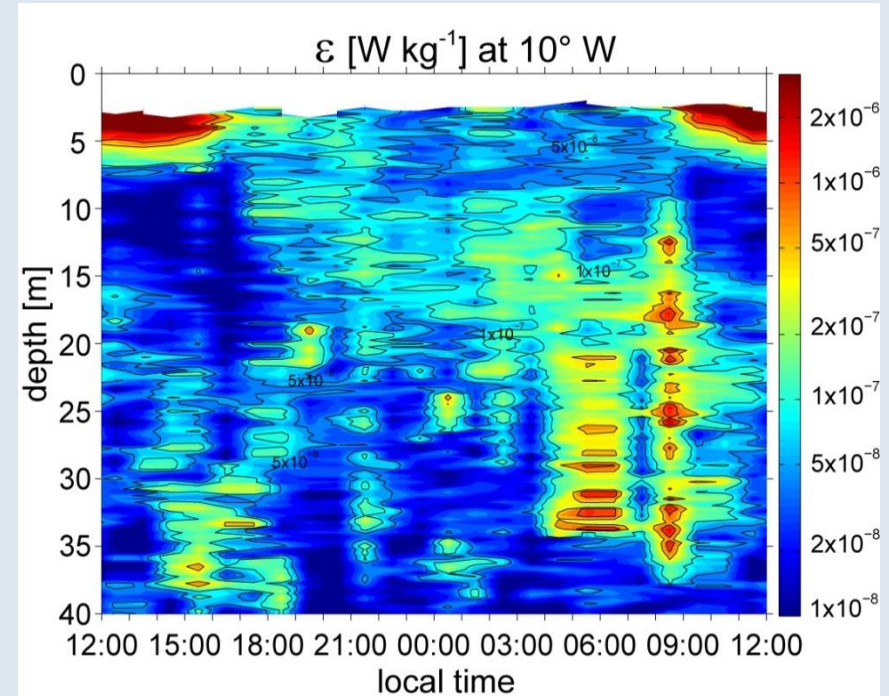
Diurnal cycle of temperature and ε

June, 10°W:

temperature [°C]

▸ Solar heating causes daily cycle of temperature in the upper few meters (2-6m) of the ocean exceeding 0.5°C.

▸ Intense mixing occurring in the mixed layer during local daytime (10:00-16:00)

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Christian-Albrechts-Universität zu Kiel

Diurnal cycle of stratification and ε
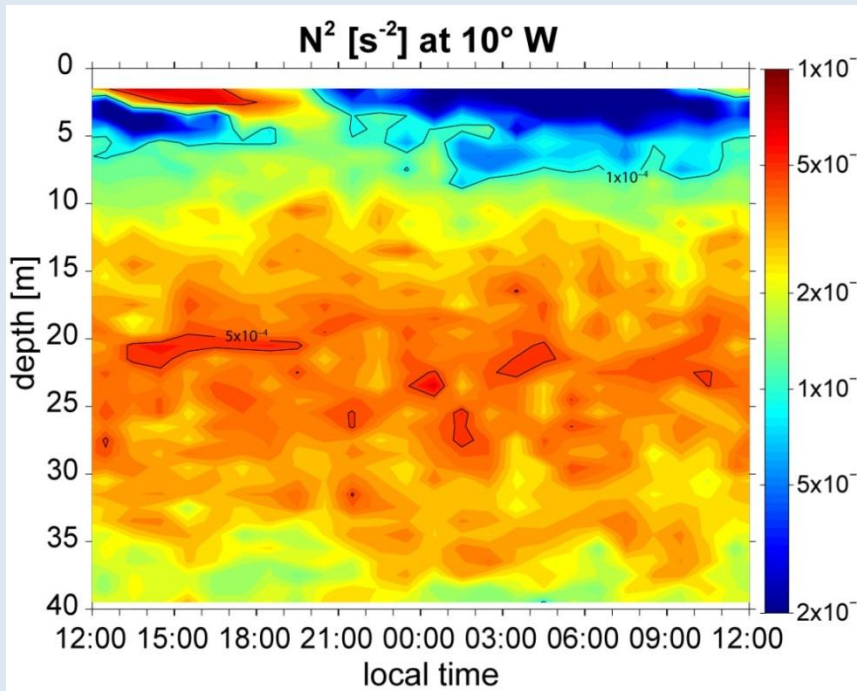
- ▸ Intense daytime mixing occurs during period of strong stratification with the upper mixed layer

- ▸ Near-surface stratification inhibits vertical turbulent momentum flux and thus traps momentum input by winds to the top few meters generating strong shear below.

Deep cycle turbulence in the Pacific (140°W)
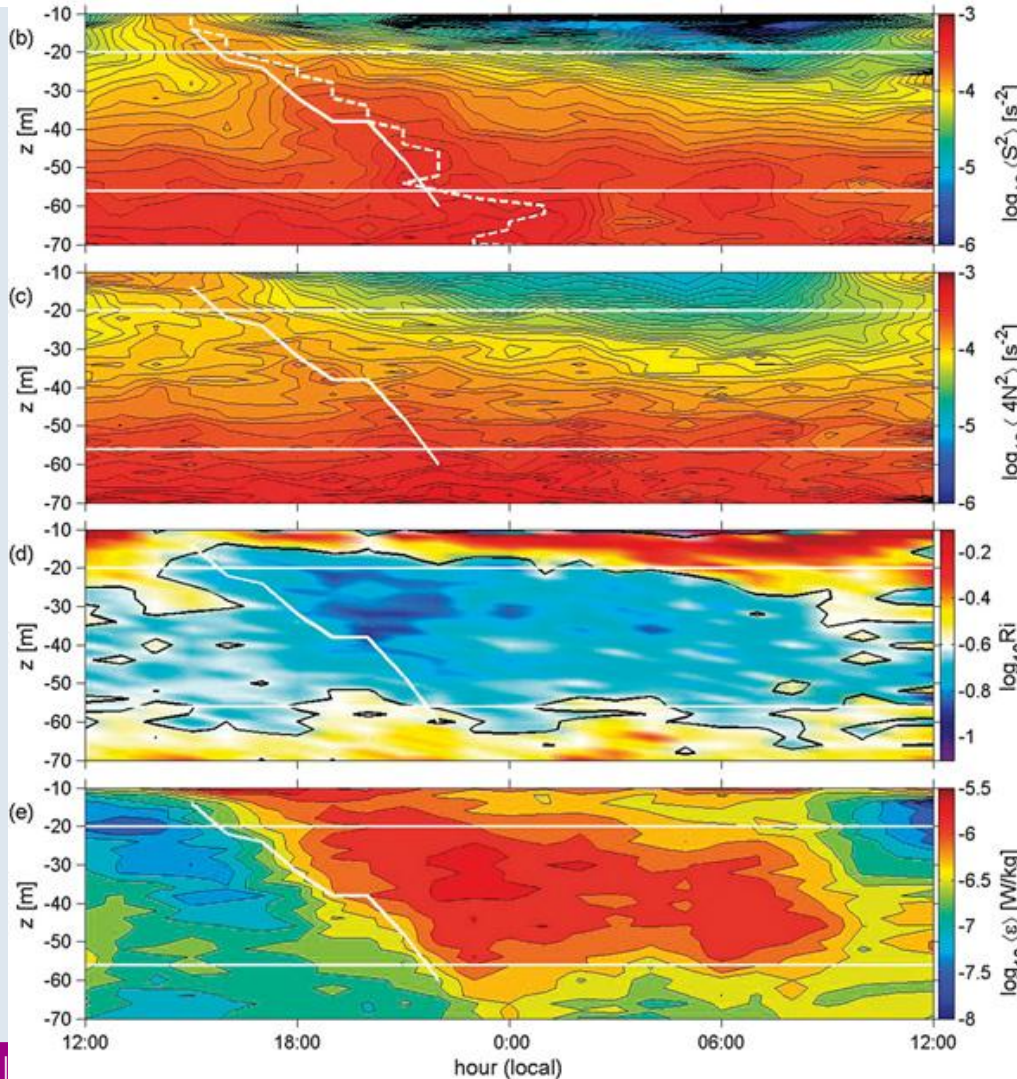
Shear variance
$S^2=(du/dz)^2+(dvdz)^2$

Stratification $4N^2$

Richardson Number ($\log_{10}$)

Turbulent dissipation rate

▸ Descending stratification and shear in the afternoon observed in the Pacific.

▸ Elevated mixing in the upper thermocline occurs after the shear layer passes a particular depth level

(Smyth et al., 2013b)

High resolution model simulation

▸ Dissipation rate of turbulent kinetic energy from a very-high resolution model simulations (LES) ($\Delta x, y$ = 1.25m and Δz = 0.25m)

(Pham et al., 2013)

High resolution model simulation
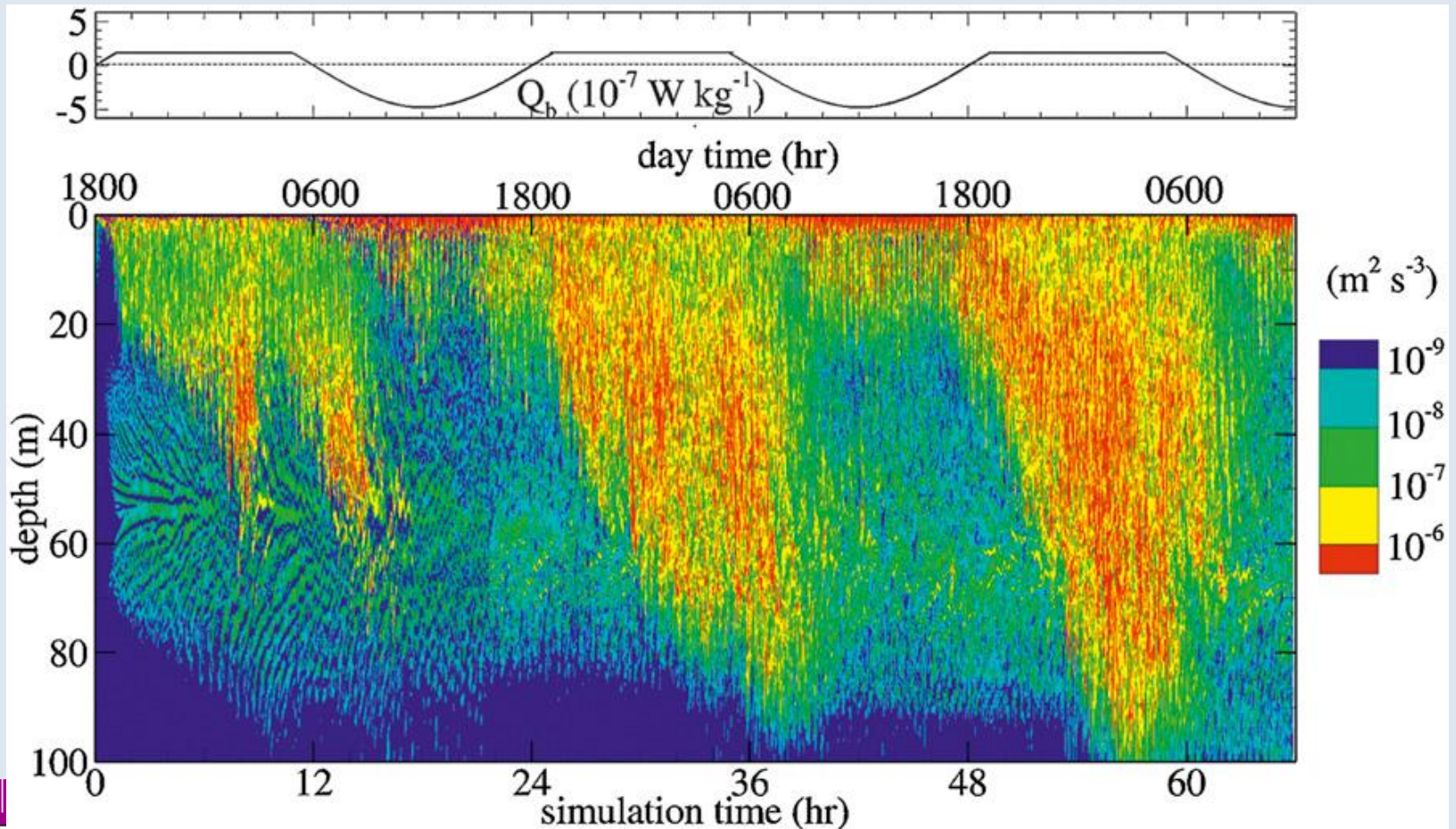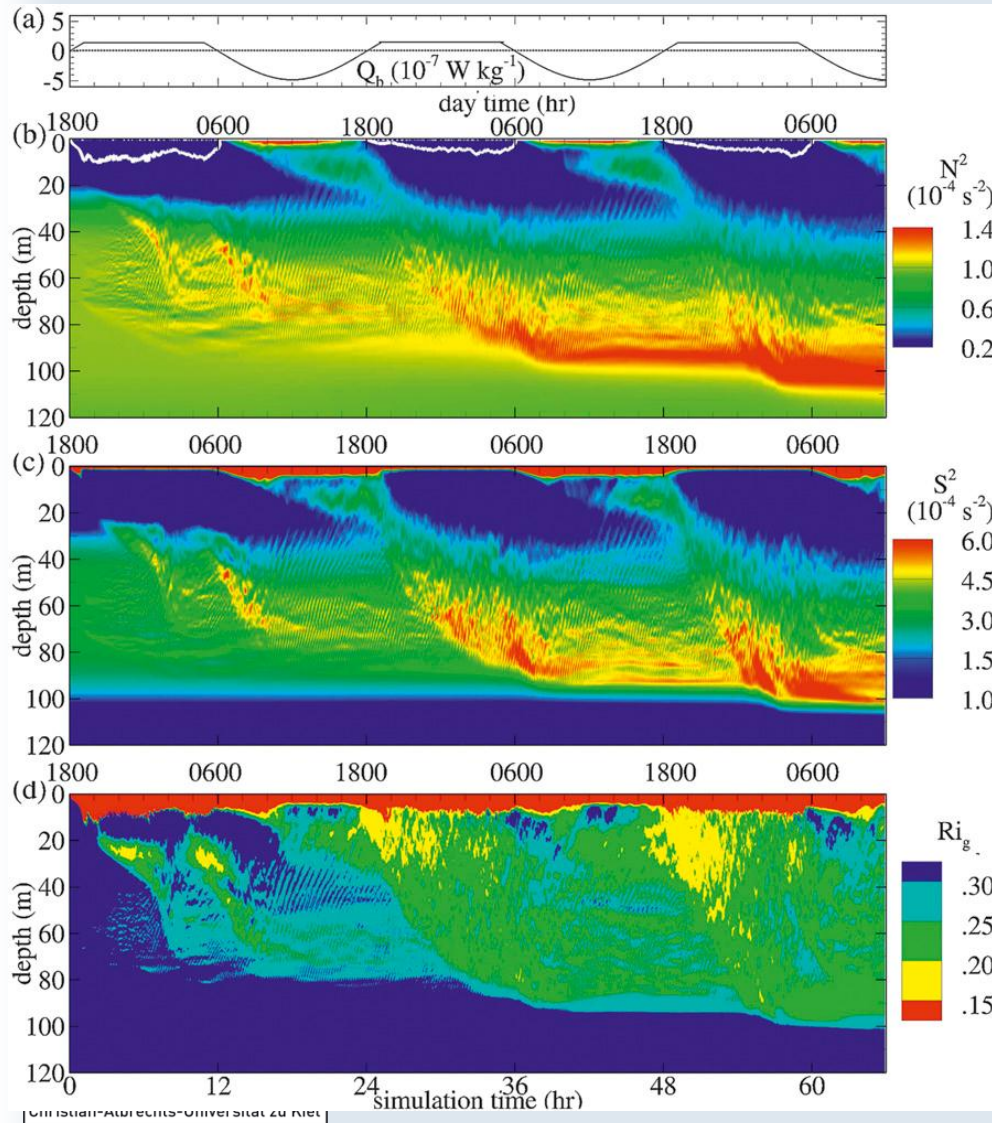
- ▸ Very-high resolution model simulations (LES) also show downward propagation of shear layer
- ▸ Daytime trapped surface current between 0-4m depth has velocities of 1.5 m/s (Pham, pers. comm)

(Pham et al., 2013)

Turbulent heat flux into the deeper ocean at the equator

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Atlantic

▸ Average heat loss of the mixed layer due to turbulence:

May (6 days) ˜ 80 W/m²

June (21 day) ˜ 60 W/m²

Nov.(6 days) ˜ 15 W/m²

▸ Mixed layer heat is redistributed to the upper 40m of the thermocline

$$J_h = K\rho \frac{\partial T}{\partial z} = \frac{\Gamma \varepsilon}{N^2}\frac{\partial T}{\partial z}$$
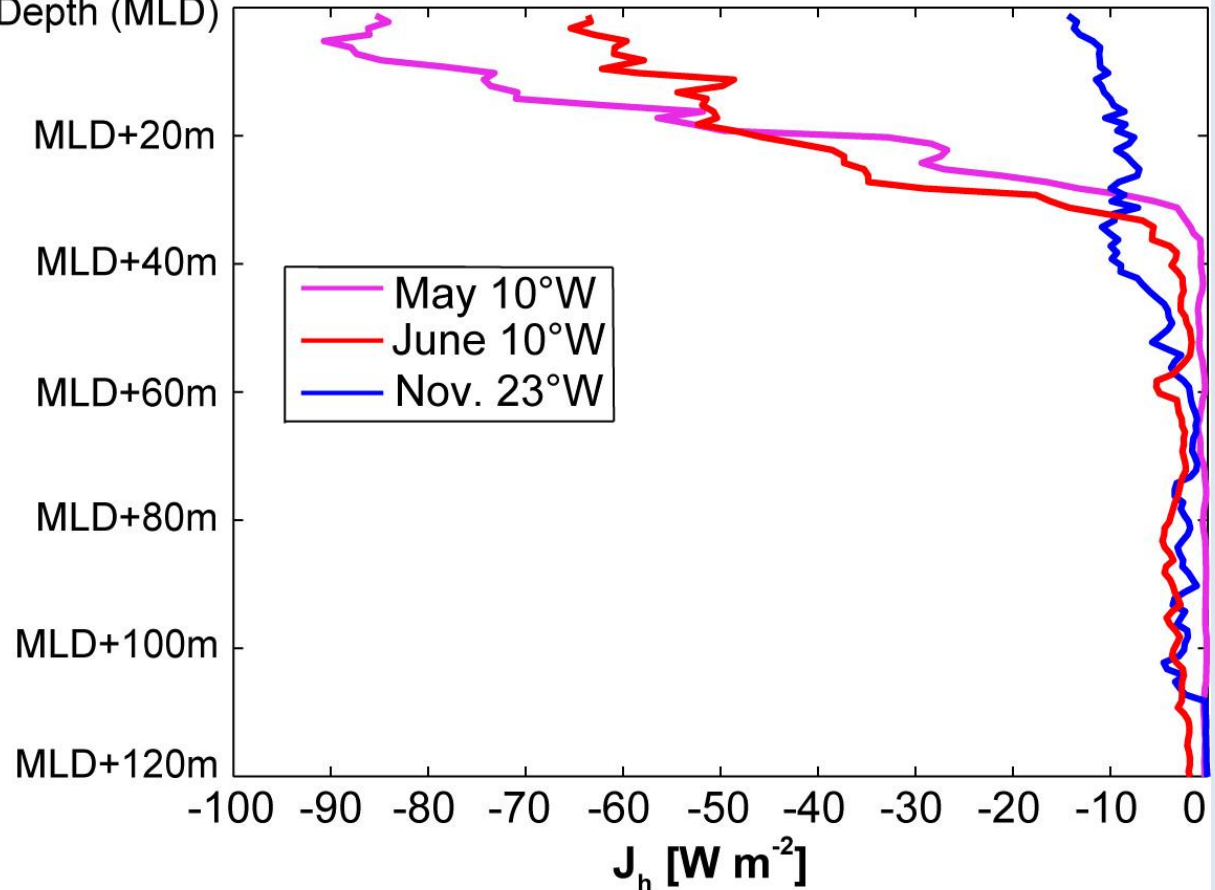
K_ρ - turbulent diffusivity
Γ - mixing efficiency (approx. 0.2)
ε - turbulent dissipation rate

Turbulent heat flux into the deeper ocean at the equator

Pacific

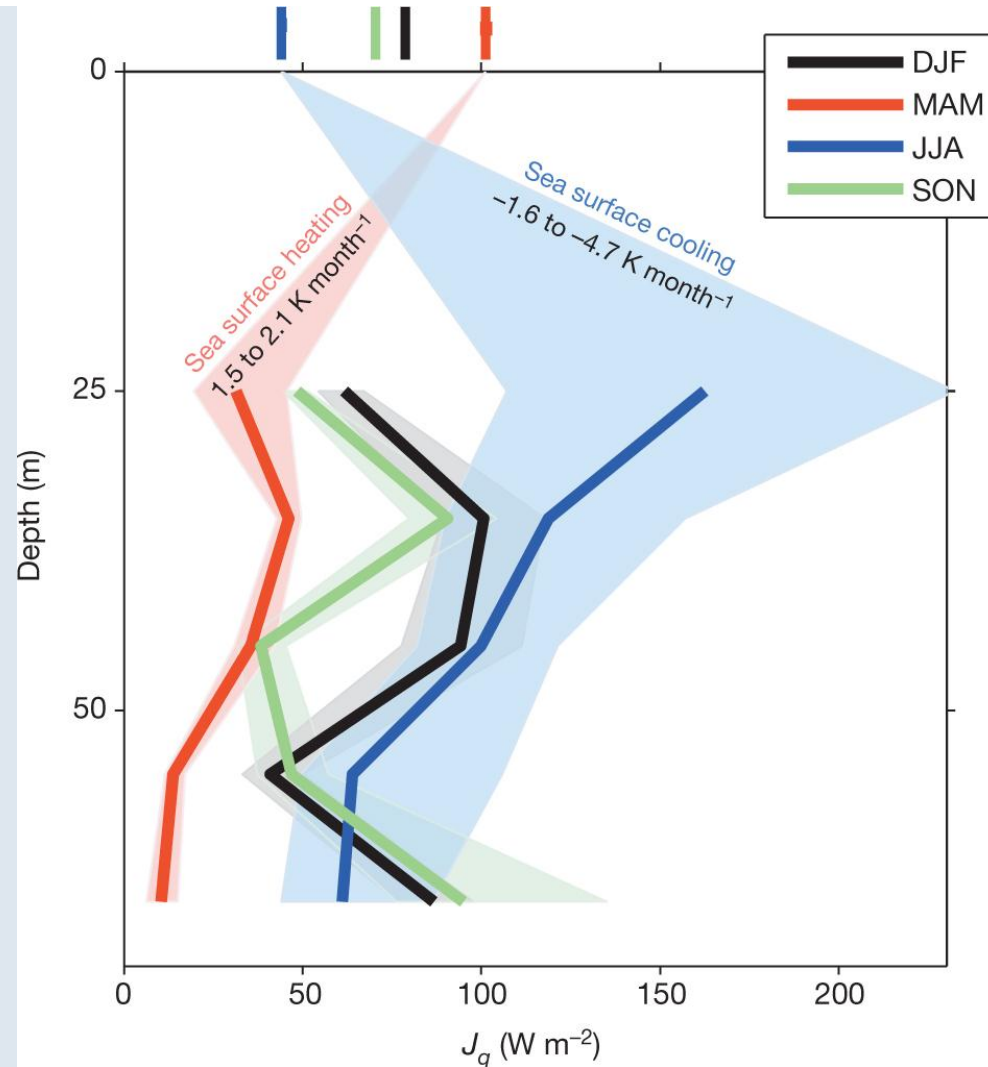▸ Seasonal average heat loss of the mixed layer due to turbulence (from moored observations)

▸ Mixed layer heat is redistributed to the upper 100m of the thermocline

Moun et al., 2013

TIWs and equatorial mixing

- ▸ turbulence float released within a TIW in the Pacific showed elevated average diapycnal heat flux of about 200 Wm^{-2} over a period of 3 weeks
- ▸ model simulations agreed with the observations and attributed the enhanced fluxes to shear instability below the mixed layer
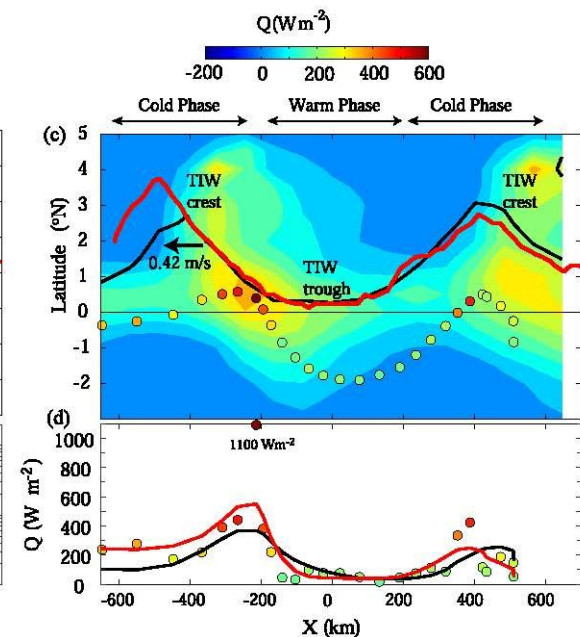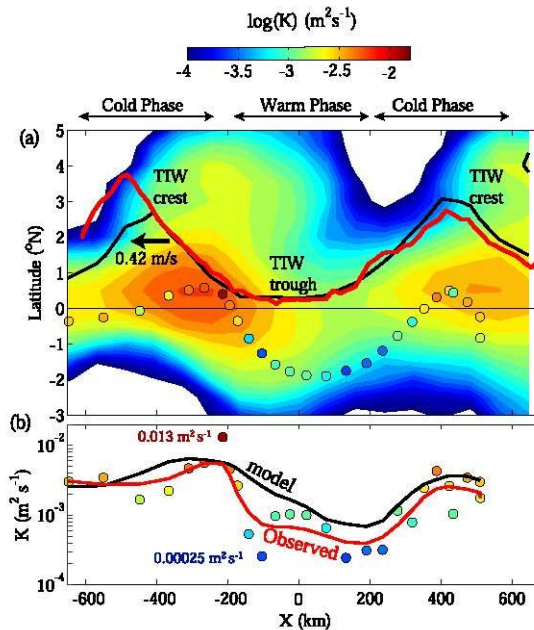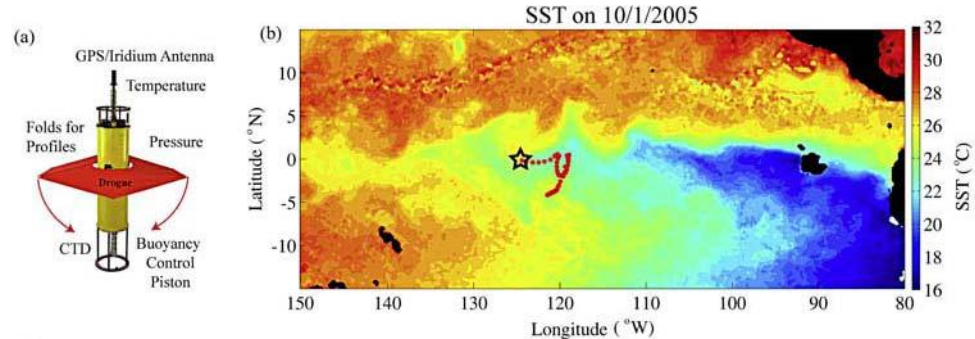
(Lien et al, 2008)

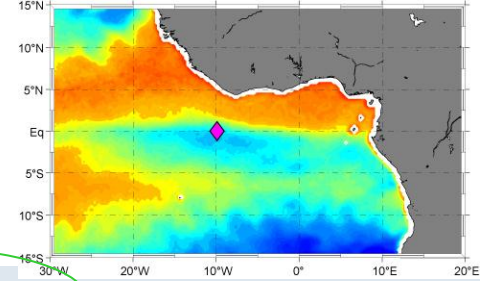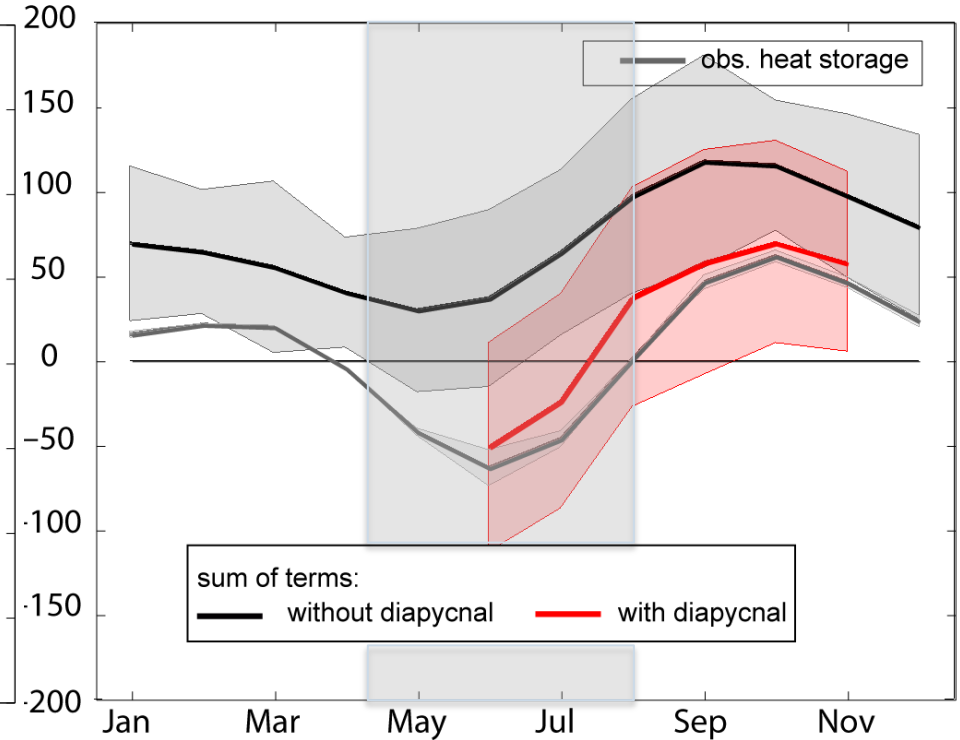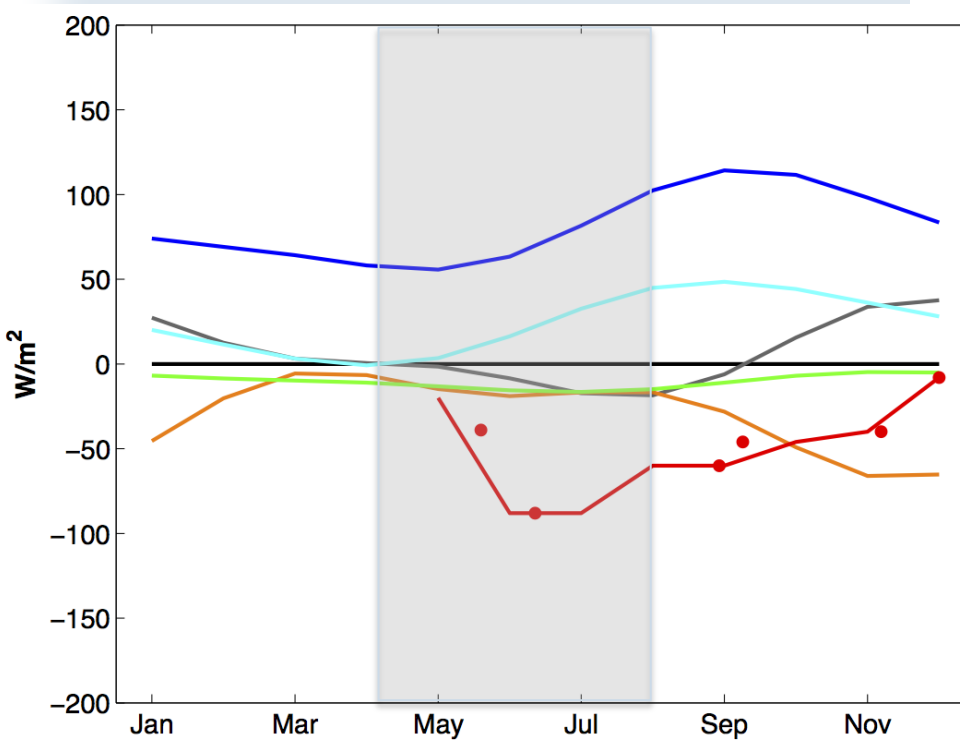# Roadmap: Mixing and the Tropical Oceans

▸ Motivation: Stirring and mixing of a tracer
  - Eddies and stirring
  - Turbulent mixing

▸ Seasonal cycle of the tropical ocean

▸ Eddies in the equatorial oceans

▸ Turbulent mixing within the Equatorial Undercurrent

▸ Mixed layer heat budget

▸ Summary

Mixed layer heat budget

Temporal changes

mean advection

entrainment

$$\rho c_p h \frac{\partial T}{\partial t} = q_{net} - \rho c_p h (\mathbf{u} \cdot \nabla T) - \rho c_p (\overline{\mathbf{u}' \cdot \nabla T'}) - \rho c_p (T - T_{-h}) w_{entrain} - q_{-h}$$

net surface flux

eddy steering

diapycnal mixing

obs. heat storage

sum of terms:
without diapycnal   with diapycnal

(Hummels et al., 2013, 2014)

Mixed layer heat balance in OGCMs

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annual mean trends by different processes (°C/month)
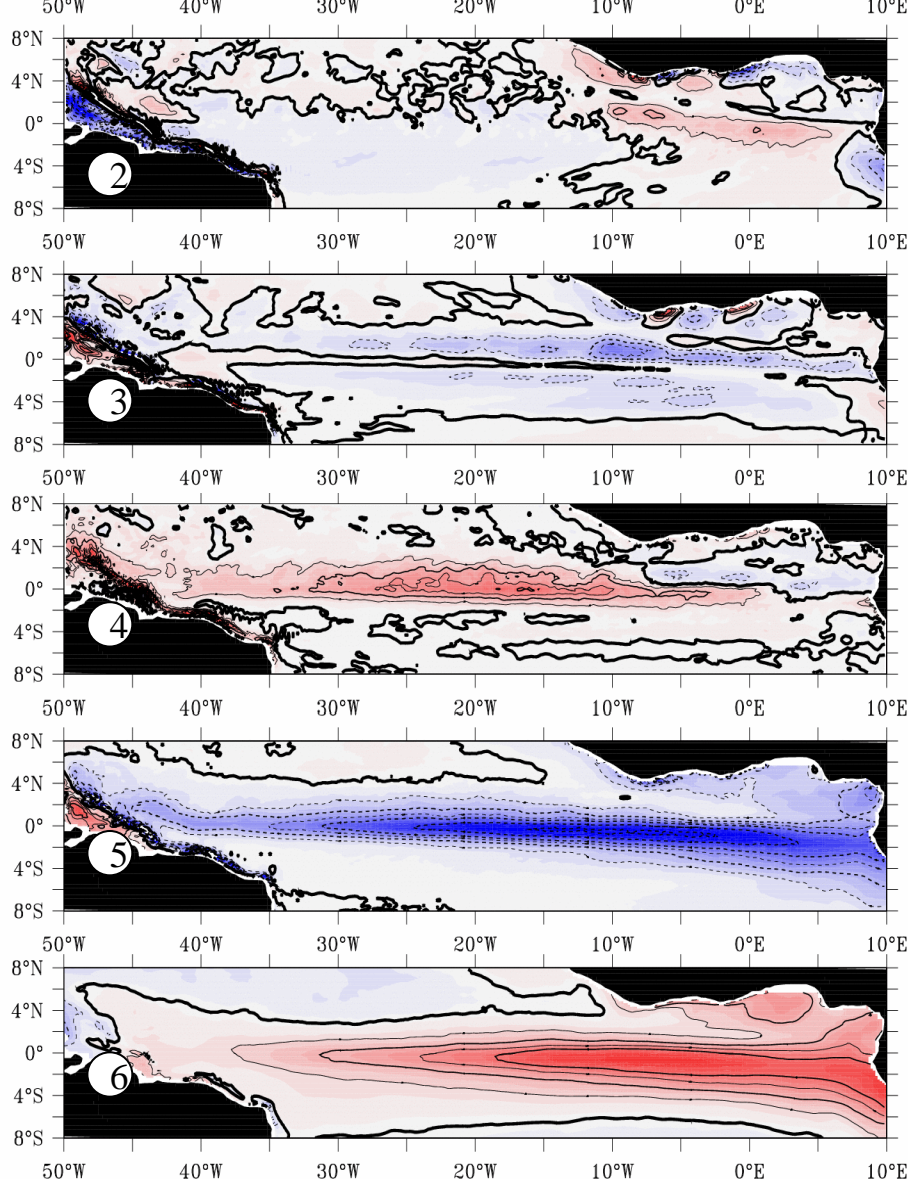
zonal advection by
low frequency currents

$$\langle -\overline{u}\,\partial_x\overline{T}\rangle$$

meridional advection by
low frequency currents

$$\langle -\overline{v}\,\partial_y\overline{T}\rangle$$

high frequencies advection (<35 days)
effects of eddies

$$-\langle\overline{u'\partial_x T'}\rangle - h\langle\overline{v'\partial_y T'}\rangle + h\langle D_l\rangle$$

subsurface mixing (vertical advection,
entrainment and turbulent mixing)

$$-\left(K_z\partial_z T\right)_{(z=h)} - \left(\partial_t h + w_{(z=h)}\right)\!\left(\langle T\rangle - T_{(z=h)}\right)$$

atmospheric
forcing

$$\frac{Q^* + Q_s\left(1 - f_{(z=h)}\right)}{\rho_0 C_p h}$$

(Peter et al., 2006)

Summary

▸ Seasonal SST variability in the equatorial Atlantic and Pacific is controlled by a diapycnal mixing occurring in the stratified layer below the mixed layer

▸ Eddies in the tropical regions warm the could tongue regions during its presents

▸ New autonomous observatories undisturbed measurements of the upper few meters of the ocean. New ideas are expected to result from these data sets

▸ The term "up-mixing region" would be more appropriate for the "equatorial upwelling regions" and perhaps for coastal upwelling regions, too?