Why is there a Front North of the Atlantic Cold Tongue?

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Introduction

- In summer, the equatorial cold tongue is well developed south of the equator with a front across the equator.
  
  - Reynolds’ SST JJA 2006
  - Grad(SST) JJA 2006
  - 2.1°C
  - 20 km
  - 1.1 psu

- In the equatorial front meridional SST gradients reach ~2°C/20 km [Lefèvre et al., 2009]

- This front influences the circulation in the MABL, coastal precipitation and the west African monsoon jump [Thorncroft et al., 2011; Caniaux et al., 2011; Nguyen et al., 2011]
Influence of the CT on the Atmosphere

1. In B, SSTs cool as soon as winds strengthen near 3°S.
2. In B, cooling increased in May-June.
3. Sharp SST gradients between A and B.
4. SST gradients relax in August-September.

1. S.H. winds increase and reach the N.H., never the contrary.
2. As soon as a SST gradient threshold is reached, winds:
   (1) weaken S of the equator;
   (2) strengthen N of the equator up to the continent in July-August.

~2 months 1/2
Mixed-Layer Heat Budget - Seasonal Scale

Giordani et al. (2013)

TURBULENT MIXING

Differential cooling induces a SST front on the Equator?
The Atlantic Cold Tongue - EGEE 2006
Giordani & Caniaux (2011)
Role of the dynamics in frontogenesis?
Frontogenesis
Giordani and Caniaux (2014)

Frontogenetic Function

\[
\frac{1}{2} \frac{d}{dt} (\nabla \theta)^2 = \vec{Q}_h \cdot \nabla \theta + \vec{Q}_d \cdot \nabla \theta
\]

Heat forcing: \(\vec{Q}_h\)

\[
Q_{hx} = \frac{\partial}{\partial x} \left( F_{sol} \frac{\partial I(z)}{\partial z} - \frac{\partial w' \theta'}{\partial z} \right)
\]

\[
Q_{hy} = \frac{\partial}{\partial y} \left( F_{sol} \frac{\partial I(z)}{\partial z} - \frac{\partial w' \theta'}{\partial z} \right)
\]

Dynamic forcing: \(\vec{Q}_d\)

\[
Q_{dx} = -\left( \frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} + \frac{\partial v}{\partial x} \frac{\partial \theta}{\partial y} + \frac{\partial w}{\partial x} \frac{\partial \theta}{\partial z} \right)
\]

\[
Q_{dy} = -\left( \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} + \frac{\partial u}{\partial y} \frac{\partial \theta}{\partial x} + \frac{\partial w}{\partial y} \frac{\partial \theta}{\partial z} \right)
\]
Frontogenesis from a regional PE model

Giordani and Caniaux (2014)

- Frontogenesis in the longitude band [15°W-5°E] and in the latitude band [1°S-1°N]

- Westward of 15°W the frontogenesis vanishes because of weaker SST gradients due to TIW
Heat forcing

\[ Q_{hx} \cdot \frac{\partial \theta}{\partial x} + Q_{hy} \cdot \frac{\partial \theta}{\partial y} \]

Heat fluxes are frontolytic
The dynamic forcing is frontogenetic and mainly supported by the meridional component:

\[ Q \frac{\partial \theta}{\partial x} + Q \frac{\partial \theta}{\partial y} \]
Low-High Frequency Components

The low frequency heat forcing is weakly frontogenetic.

The high frequency component of the dynamical forcing is the strongest frontogenetic term.

The dynamic forcing is mainly supported by the convergence between the nSEC and the GC.
Origin of the dynamical forcing

Wind-stress (shaded) and wind energy flux (black)

SST and dynamic forcing term

Meridional current (shaded) and dynamic forcing term
Frontogenesis

In case of meridional gradients:

\[
\frac{1}{2} \frac{d}{dt} \left( \vec{v} \cdot \theta \right)^2 \approx \frac{\partial}{\partial y} \left( F_{sol} \frac{\partial I (z)}{\partial z} - \frac{\partial w ' T '}{\partial z} \right) \frac{\partial T}{\partial y} \\
- \left( \frac{\partial u}{\partial y} \frac{\partial T}{\partial x} + \frac{\partial v}{\partial y} \frac{\partial T}{\partial y} + \frac{\partial w}{\partial y} \frac{\partial T}{\partial z} \right) \frac{\partial T}{\partial y}
\]

Heat forcing term

Surf. fluxes

Turb. mixing

Frontolysis

Dynamic forcing term

Frontogenesis
$Q_{dy} = -\frac{\partial w}{\partial y} \frac{\partial \theta}{\partial z} < 0$
PV in TIW

\[ q = \left( \zeta + f \right) N^2 - f \left( \frac{\partial \vec{U}_g}{\partial z} \right)^2 \]

\[ q < 0 \quad \text{in frontal/eddy areas where} \quad q_h \ll 0 \]
Destruction of Potential Vorticity

\[ \frac{\partial q}{\partial t} = - f \frac{\partial \vec{U}_g}{\partial z} \frac{\partial \tau}{\partial z} + (\zeta + f) \frac{\partial B}{\partial z} - \vec{U} \nabla q \]

PV destruction:
- Wind oriented down-front
- Buoyancy loss from the ocean to the atmosphere
PV in TIW
\( \bar{J}_{fric} = -f \frac{\partial \bar{U}}{\partial z} \frac{\partial \bar{\tau}}{\partial z} \)

MLD

frontal zone

VORTICITY (\(v \times 10^8\))

Latitude  

V

PREFACE PIRATA CLIVAR TAV Meeting  
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Conclusions

• The heat forcing term associated with fluxes is frontolytic (*i.e.* weakening of the equatorial front)

• Low-frequency heat forcing is frontogenetic and may initiate the equatorial front, which is largely amplified and maintained by the dynamic forcing

• Dynamic forcing is the leading term of frontogenesis: it is driven by the meridional convergence between the Guinea Current and the South Equatorial Current
Conceptual Scheme

WEF

U (SEC)

V (GC)

Convergence SEC and GC

MLD

TKE

Divergence of Turbulent Heat Flux

$d_t \left( \nabla \theta \right)^2 < 0$

$d_t \left( \nabla \theta \right)^2 > 0$

$d_t \left( \nabla \theta \right)^2 > 0$ → Coupling with the Monsoon Flux

Feedback Loop
Conclusions

• Intra-thermocline bolus/eddy of low-PV can modify stratification, circulation, vertical/lateral mixing and ML heat/salt budgets

• How can we document this with data (PIRATA buoys, Gliders, CTDs ...)?
Conceptual Scheme

Baroclinically-low PV

Vortically-low PV

\( f + \zeta \)

\( \nabla b \)

\( \omega_h \)

Transport

North