

Dynamical Origin Of Coastal Upwelling At Cape Sainte-Marie, Southern Madagascar

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Abstract: This study investigates the origin of coastal upwelling at Cape Sainte-Marie in southern Madagascar, using a one-year CROCO simulation representative of 2010, referred to as Core 2 by Ramanantsoa et al. [1]. The objective is to characterize the life cycle of this cell, from its formation at the cape to its export toward the southern gyre and to clarify the respective roles of ocean dynamics and wind forcing in its onset. Joint analysis of surface currents, normalized relative vorticity, sea level, kinetic energy, thermohaline structure, and instability diagnostics shows that Core 2 is associated with current-separation upwelling. Between days 40 and 90 (mid-February to early April), a pre-existing boundary flow intensifies and separates at Cape Sainte-Marie, generating a cyclonic cell in geostrophic balance that uplifts the isopycnals and produces a cold, saline anomaly. At the cape, the eddy is primarily energized by baroclinic instability, which clearly dominates the other energy sources and for which the necessary conditions for development are satisfied. The cell subsequently detaches from the cape and migrates westward toward the southern gyre by mid-April. The process is recurrent, with a second eddy generation developing in early September. Core 2 therefore emerges not simply as wind-driven upwelling, but as a current-separation upwelling system: wind modulates its intensity, whereas its origin and evolution are controlled by current separation at Cape Sainte-Marie, baroclinic amplification, and export toward the southern gyre.

Keywords: coastal upwelling; current separation; baroclinic instability; Cape Sainte-Marie; CROCO.

1. INTRODUCTION

Southern Madagascar is a highly dynamic oceanic region where winds, bathymetry, boundary currents, and mesoscale eddy activity interact. Previous studies have distinguished two upwelling cells in this region [1]: Core 1, southeast of Madagascar, associated with separation of the East Madagascar Current, and Core 2, west of Cape Sainte-Marie. Core 1 is now well established as a current-separation upwelling system, whereas Core 2 has been interpreted as predominantly wind-driven, while also being influenced by exchanges with the Mozambique Channel. However, the maximum offshore Ekman transport does not coincide with the maximum upwelling signal [1], indicating that local wind forcing alone is insufficient to explain Core 2 variability.

The specific mechanism origin of Core 2 therefore remains unresolved. A CUTI-based analysis [2] confirmed that part of the transport associated with the Core 2-Cape Sainte-Marie-southern gyre system cannot be attributed to local wind forcing and highlighted the open character of the system, without identifying the physical mechanism responsible for upwelling.

The present study addresses this gap using a CROCO simulation dedicated to this system. It aims to identify the origin mechanism of Cape Sainte-Marie upwelling, determine whether it results from current separation rather than direct wind forcing, and characterize the roles of baroclinic instability and export toward the southern gyre. The 2010 simulation is analyzed using the dynamic, thermohaline, and energetic diagnostics described in Section 2.

2. METHODS

2.1. Model configuration and study domain

The simulation is performed with the CROCO model [3], [4], inherited from ROMS, over the domain 42°E-46°E / 27°S-25°S. The horizontal grid contains 145×82 points, corresponding to a mean resolution of approximately 2.7 km, with 40 sigma vertical levels. The simulation covers a 360-day numerical annual cycle representative of atmospheric conditions in 2010. It is forced by ERA5 [5] using COARE-type bulk formulae [6], with KPP turbulent closure [7]. Daily averaged outputs are used for the diagnostics. The analyzed variables are horizontal currents, sea level, temperature, salinity, relative vorticity, and energy diagnostics.

2.2. Analysis boxes and origin window

The analysis is based on four boxes used consistently with the CUTI study [2]: B1 western upstream, B2 coastal Core 2, B3 Cape Sainte-Marie, and B4 export toward the southern gyre. This configuration distinguishes the westward-advected response area, the cold Core 2 cell, the dynamic source region at the cape, and the export zone toward the southern gyre.

The main analysis window corresponds to days 40-90, during which the boundary flow intensifies and separates at the cape. Simulation days are referenced to the year 2010, with day 0 corresponding to 1st January. This convention places day 40 around mid-February and day 90 around early April.

Table 1. Analysis boxes used to distinguish the upstream area, Core 2, Cape Sainte-Marie, and export toward the southern gyre.

Box	Longitude	Latitude	Dynamic role
B1 western upstream	43.0-44.0°E	26.8-25.8°S	Residence and advection zone of the detached eddy
B2 coastal Core 2	44.0-45.0°E	26.6-25.4°S	Cold cell and primary thermohaline response
B3 Cape Sainte-Marie	45.0-46.0°E	26.6-25.6°S	Current-separation source region and energy generation
B4 export toward southern gyre	44.0-46.0°E	27.0-26.6°S	Export zone toward the southern gyre

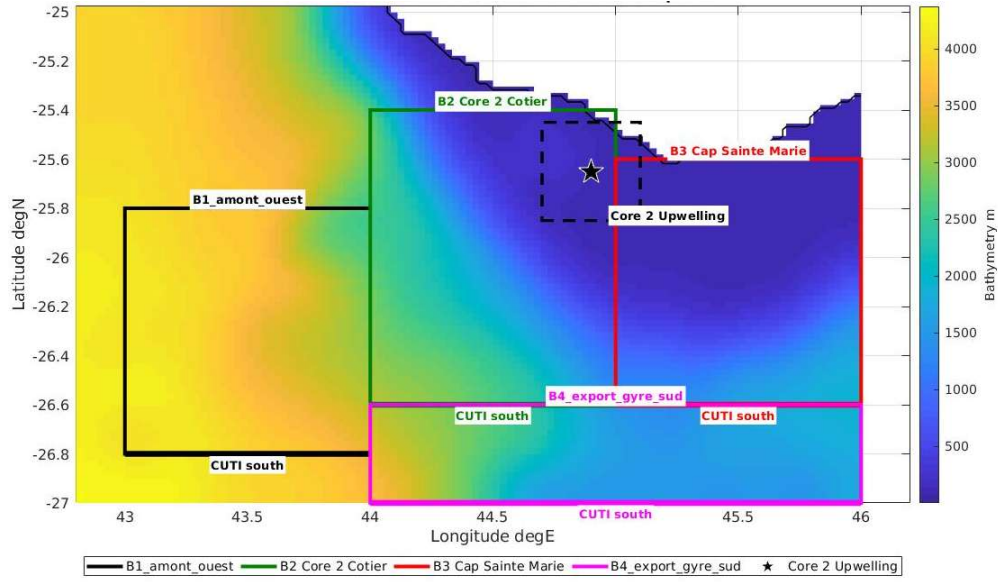


Figure 1. Location of the four analysis boxes over the domain bathymetry. B2 corresponds to the coastal Core 2 cell and B3 to the Cape Sainte-Marie sector

2.3. Dynamic diagnostics

The propagation of the structures is tracked using surface relative vorticity normalized by the Coriolis parameter, defined as:

$$\frac{\zeta}{f} = \left(\frac{1}{f}\right) \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right) \quad (1)$$

- where $f = 2\Omega \sin(\phi)$ and ζ_r denotes relative vorticity.

In the Southern Hemisphere, positive values of ζ/f indicate cyclonic circulation. A Hovmöller diagram of ζ/f is constructed along an axis connecting Cape Sainte-Marie to the southern gyre to track the formation and subsequent propagation of eddy structures.

The Charney-Stern criterion [8] is evaluated from the sign of the meridional potential-vorticity gradient. Baroclinic instability is possible when $\frac{\partial q}{\partial y}$ changes sign in the vertical:

$$\left(\frac{\partial y}{\partial x}\right) z_1 \times \left(\frac{\partial q}{\partial y}\right) z_2 < 0 \quad (2)$$

This criterion is necessary but not sufficient. It indicates that the flow has a vertical structure capable of supporting instability, while the nature of that instability is subsequently assessed from the energy-conversion balance.

The growth rate of baroclinic instability is estimated using the Eady formulation [9]:

$$\sigma E = 0.31 \left(\frac{|f|}{N}\right) \left|\frac{\partial U}{\partial z}\right| \quad (3)$$

- where N is the Brunt-Väisälä frequency and $\left|\frac{\partial U}{\partial z}\right|$ is the vertical shear of horizontal velocity. The absolute value of f is used to obtain a positive growth rate.

Baroclinic conversion measures the transfer of available potential energy to eddy kinetic energy [10]. It is defined here as:

$$BC = -\left(\frac{g}{\rho_0}\right) \langle w' \rho' \rangle \quad (4)$$

A positive BC value indicates that isopycnal tilting feeds eddy growth. Density fluctuations ρ' are reconstructed from temperature, salinity, and pressure using the seawater equation of state.

Barotropic conversion measures the transfer of energy from the mean current to eddy fluctuations through the horizontal shear of the mean flow [10]:

$$BT = -\langle u'u' \rangle \frac{\partial U}{\partial y} - \langle u'v' \rangle \frac{\partial v}{\partial x} - \langle v'v' \rangle \frac{\partial v}{\partial y} \quad (5)$$

In both expressions, primes denote departures from the temporal mean over the origin window and angle brackets denote the averaging operator. A positive baroclinic conversion whose magnitude exceeds that of barotropic conversion indicates that eddy amplification is primarily baroclinic.

3. RESULTS

3.1. Initial state and intensification of the boundary flow

At the initial state, in early January, Figure 2, a coherent boundary flow is already present along southern Madagascar toward the west and southwest. Its surface speed is approximately $0.3 \text{ m}\cdot\text{s}^{-1}$, while kinetic energy remains low and relative vorticity is weakly structured. This pre-existing flow constitutes the boundary current likely to separate at Cape Sainte-Marie. Between days 40 and 90, from mid-February to early April, Figure 3, this boundary flow intensifies markedly. Velocities increase in the Core 2 sector and in the region where separation subsequently develops, while surface kinetic energy becomes concentrated near the cape. This phase precedes and conditions current separation at the cape

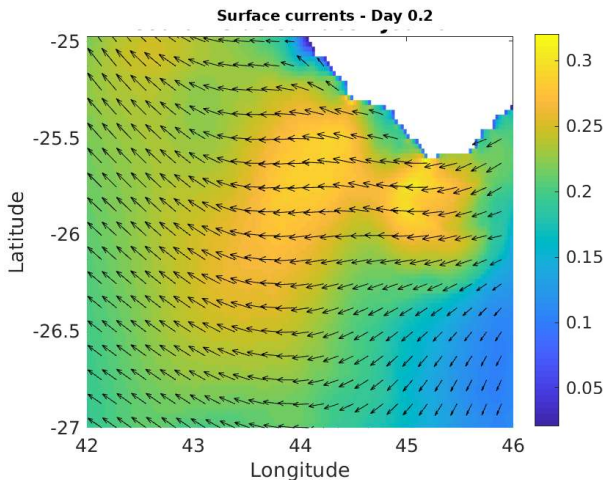


Figure 2. Day 0: pre-existing boundary flow, speed $\approx 0.3 \text{ m}\cdot\text{s}^{-1}$.

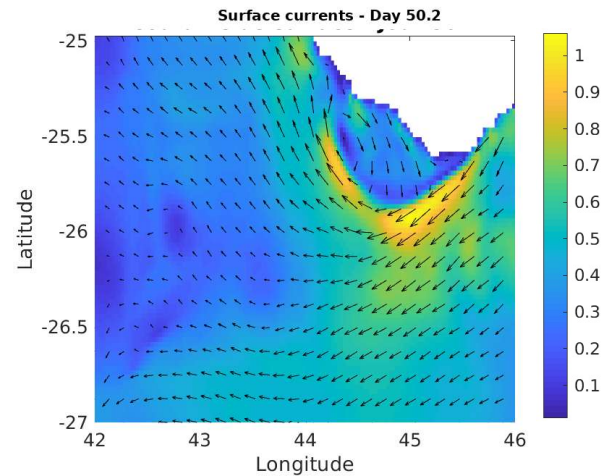


Figure 3. Day 50: current intensification and onset of separation at Cape Sainte-Marie (speed up to $\approx 1 \text{ m}\cdot\text{s}^{-1}$).

3.2. Current separation and formation of a cyclonic cell

Around day 50, at the end of February, Figure 4, the flow separates at Cape Sainte-Marie. Normalized vorticity then reveals a cyclonic ribbon bordered on the coastal side by an anticyclonic signal. This dipolar structure is characteristic of a current separating from a cape and generating a cyclonic separation cell.

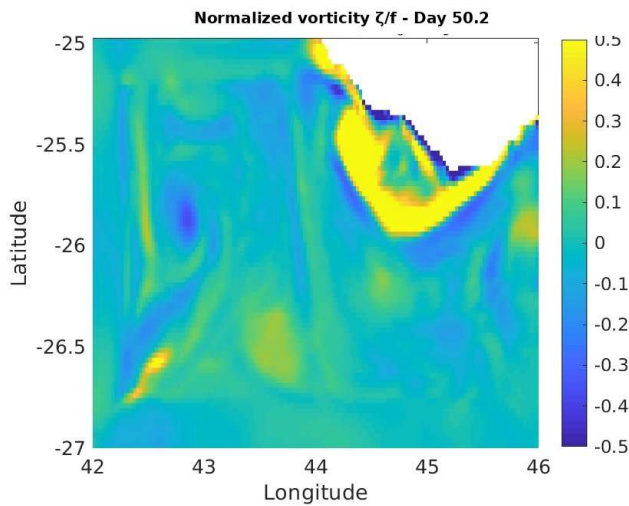


Figure 4. Day 50: appearance of the cyclonic ribbon at the separation point.

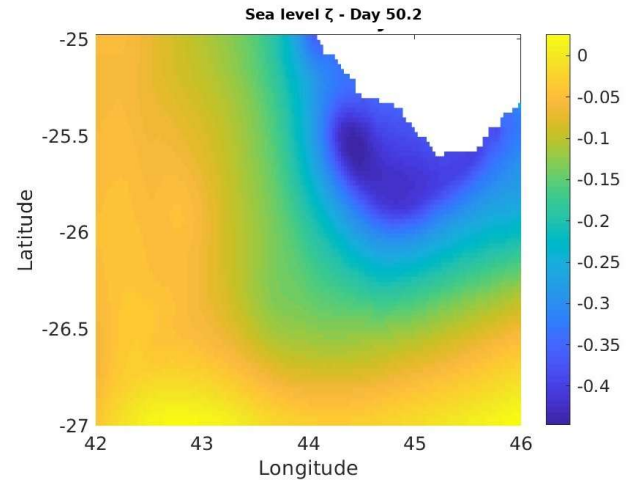


Figure 5. Day 50: incipient geostrophic depression.

At the same time, sea level exhibits a localized depression over the cell, coincident with the cyclonic ribbon, Figure 5. This co-location of the cyclonic vorticity signal and the sea-level trough indicates that the cell is in geostrophic balance.

3.3. Baroclinic amplification at the separation source region

The meridional potential-vorticity gradient averaged over the Core 2-Cape Sainte-Marie sector changes sign in the vertical. This sign reversal satisfies the necessary Charney-Stern criterion and indicates that the vertical structure of the flow is favorable to the development of baroclinic instability.

The vertical structure of the current also exhibits marked shear in the upper layer. At Core 2, velocity decreases from $0.35 \text{ m}\cdot\text{s}^{-1}$ at 5 m depth to $0.22 \text{ m}\cdot\text{s}^{-1}$ at 50 m depth, corresponding to a vertical shear of approximately $2.8 \times 10^{-3} \text{ s}^{-1}$, Figure 6.

This shear develops within a stably stratified water column, with surface temperature clearly higher than that observed near 100 m. The estimated Eady growth rate indicates rapidly growing instability, consistent with the eddy closure observed around day 100

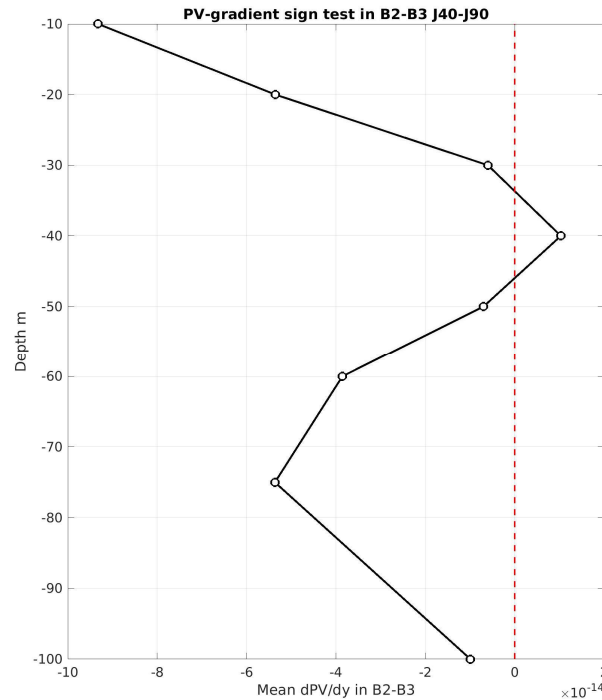


Figure 6. Meridional potential-vorticity gradient averaged over the Core 2-Cape Sainte-Marie sector.

The energy-conversion balance confirms the baroclinic nature of the amplification. Baroclinic conversion is maximum and positive at Cape Sainte-Marie (B3), where it dominates barotropic conversion ($|BC|/|BT| = 1.74$), Table2. By contrast, it becomes negative upstream (B1) and in the export zone (B4): eddy-energy generation is therefore not distributed throughout the domain, but is concentrated at the current-separation point, Figure 8 and Figure 9.

Table 2. Baroclinic (BC) and barotropic (BT) conversions averaged by box over days 40-90.

Box	BC ($\text{m}^2 \cdot \text{s}^{-3}$)	BT ($\text{m}^2 \cdot \text{s}^{-3}$)	$ BC / BT $	Interpretation
B1 western upstream	-9.3×10^{-9}	$+8.0 \times 10^{-9}$	1.15	No local origin; residence/advection zone
B2 coastal Core 2	$+3.6 \times 10^{-9}$	-6.2×10^{-9}	0.57	Secondary baroclinic source; thermohaline response
B3 Cape Sainte-Marie	$+2.5 \times 10^{-8}$	$+1.4 \times 10^{-8}$	1.74	Generation source region; baroclinic conversion dominant
B4 export toward southern gyre	-5.4×10^{-9}	$+5.2 \times 10^{-9}$	1.03	No origin; export/dissipation zone

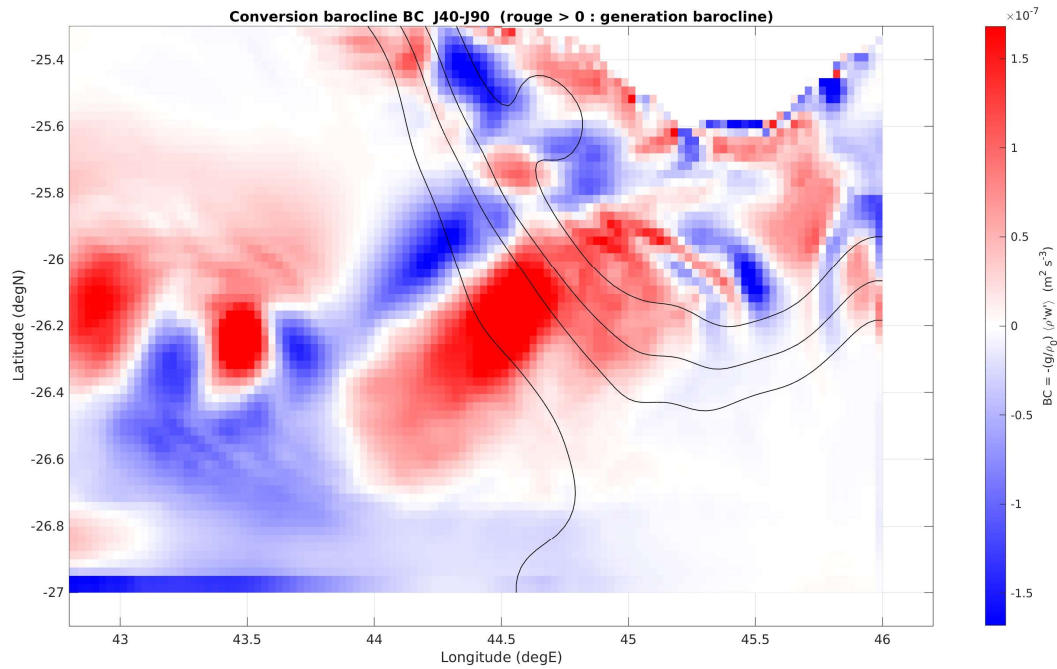


Figure 7. Baroclinic conversion, with positive cores along the slope at the cape.

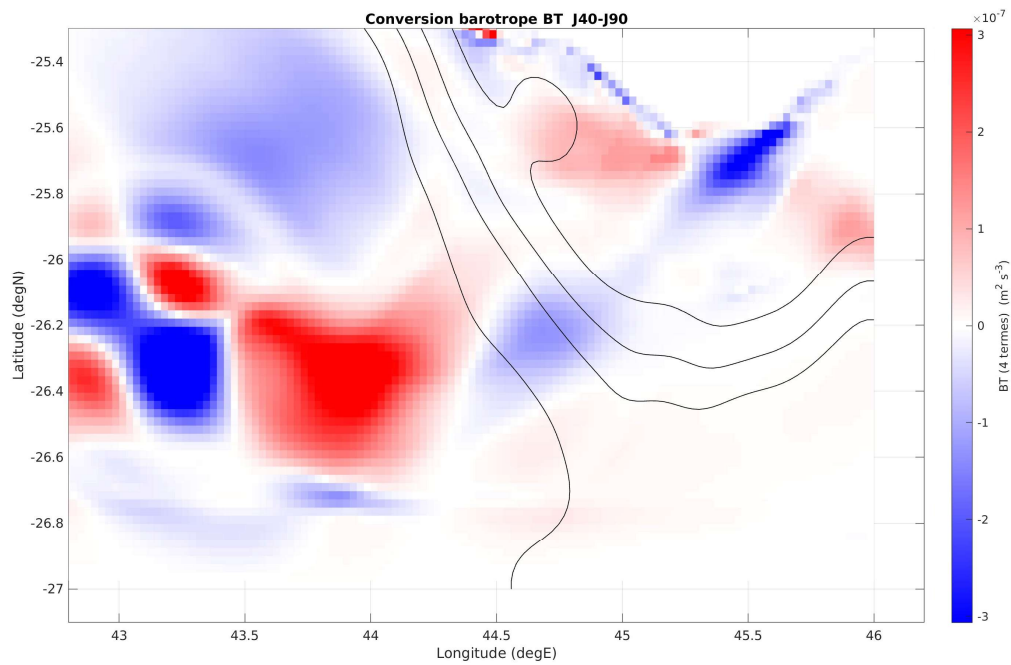


Figure 8. Barotropic conversion, organized as a dipole without coherent net generation.

To place this structure in the context of the characteristic dynamical scale of the region, the baroclinic Rossby deformation radius was estimated for each box over the origin period (J40-J90). It is defined as the ratio of the first baroclinic-mode phase speed to the Coriolis parameter:

$$Rd = \frac{c}{|f|}, = \left(\frac{1}{\pi}\right) \int_{-H}^0 N(z) dz \quad (6)$$

- where c is the baroclinic-wave phase speed, f is the Coriolis parameter, $N(z)$ is the Brunt-Väisälä frequency, and H is the integration depth.

Calculated over the surface layer (0-100 m), the mean deformation radius is 5.2 km in B2 coastal Core 2, 5.3 km in B3 Cape Sainte-Marie, and 4.4 km in B4 export toward the southern gyre. When calculated over the full water column, the corresponding values are 26.2 km, 14.0 km, and 36.4 km. The surface deformation scale of the Core 2 cell (approximately 5 km) is substantially smaller than that of the large deep eddies of the Mozambique Channel (40-60 km) [10]. This difference supports the interpretation of a surface-intensified baroclinic instability distinct from the deeper dynamics described in this region. The reduced full-column radius at Cape Sainte-Marie (B3; 14.0 km, compared with 26-36 km in neighboring boxes) reflects the topographic constraint of the shelf at the cape. This constraint is consistent with current separation at this location. Calculated over the surface layer (0-100 m), the mean deformation radius is 5.2 km in B2 coastal Core 2, 5.3 km in B3 Cape Sainte-Marie, and 4.4 km in B4 export toward the southern gyre.

3.4. Detachment, migration, and export toward the southern gyre

By day 100, in mid-April, the cyclonic ribbon has closed into a coherent, nearly circular eddy core that is detached from the cape and advected westward (Figure 9). The main core is centered near 43°E/26.1°S, with a secondary core near 44.3°E/26.5°S. The eddy has therefore migrated approximately 200 km from its formation region at the cape, quantifying the export of upwelled waters toward the southern gyre (Figure 10). Kinetic energy, initially concentrated at the cape during the separation phase (Figure 11), is subsequently transported within the detached core. By day 100, it forms an energetic ring around a comparatively quiescent center, a classical signature of a mature eddy structure (Figure 12).

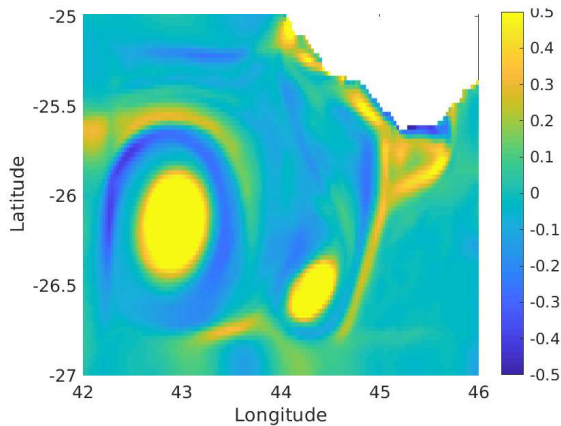


Figure 9. Vorticity ζ/f on day 100

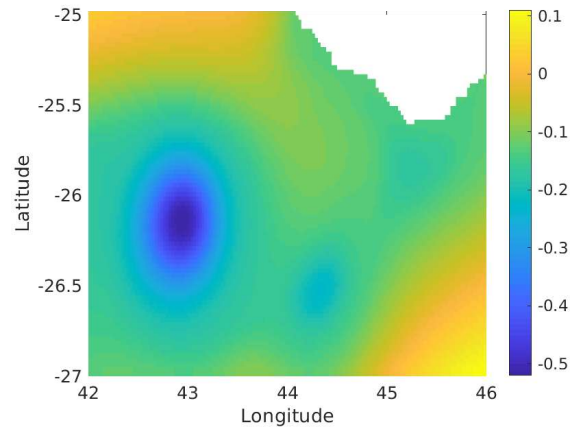


Figure 10. Sea level on day 100.

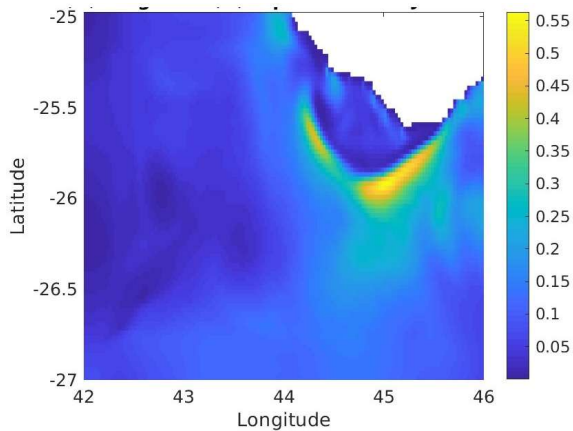


Figure 11. Kinetic energy concentrated at the cape on day 50.

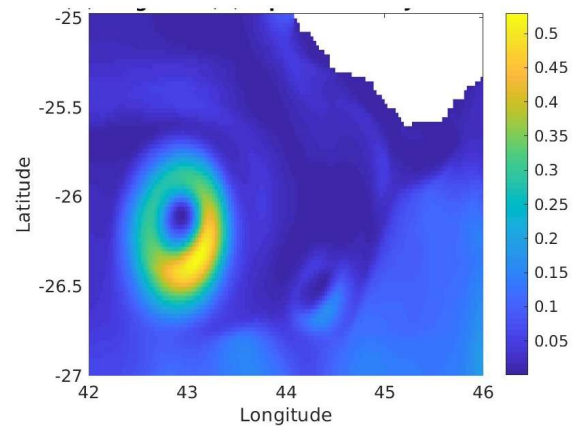


Figure 12. Kinetic energy transported within the detached eddy on day 100.

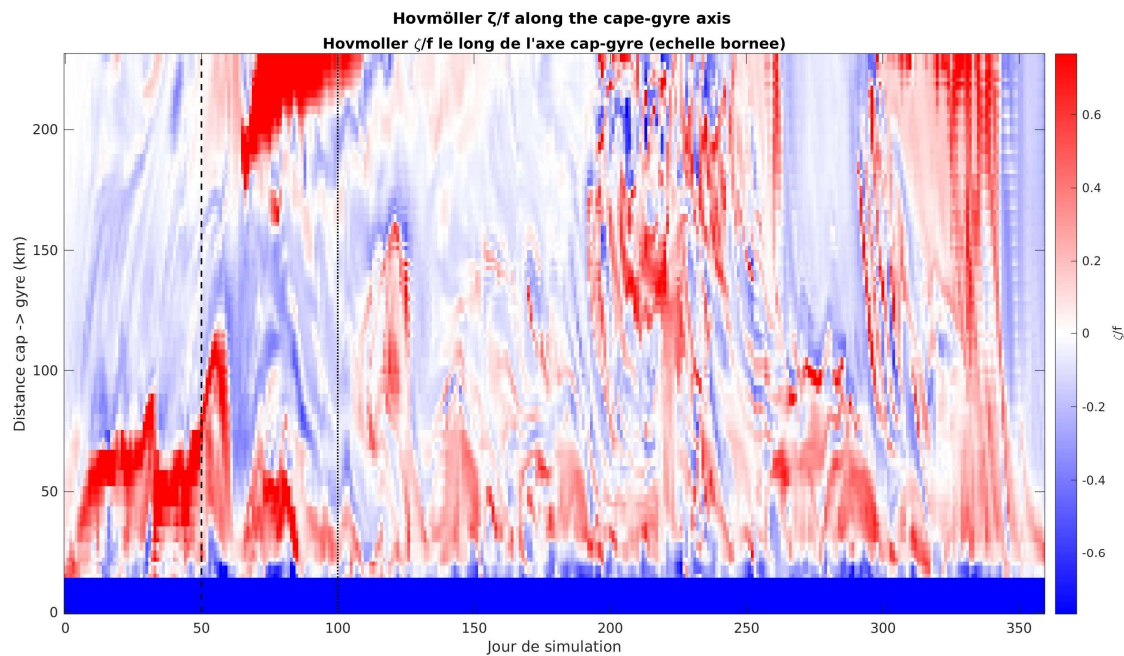


Figure 13. Hovmöller diagram of ζ/f along the cape-gyre axis.

3.5. Chronology of dynamic and wind-forcing peaks

Comparison of the peak dates of the different diagnostics at Cape Sainte-Marie over the origin window (day 40- day 90) clarifies the order in which the signals emerge. Dynamic indicators - relative vorticity, the sea-level trough, surface velocity, and surface kinetic energy - peak first, around days 57-60 (mean $\approx$ J59). Transport diagnosed by CUTI follows with a lag of 4-6 days, whereas Bakun wind forcing [13] reaches its maximum only about eleven days later (Table 3). The dynamic separation signal therefore precedes the maximum wind forcing, consistent with an origin controlled by ocean dynamics rather than by wind. Wind thus appears to strengthen the upwelling without acting as its initial driver.

Table 3. Peak day of each diagnostic at Cape Sainte-Marie over the J40-J90 window and lag relative to the mean dynamic peak ($\approx$ J59). The lag is positive when the indicator follows the mean peak; transport responds after a few days, whereas wind forcing peaks last, approximately eleven days after the dynamic signal.

Diagnostic	Peak day	Lag (d)	Family
Relative vorticity ζ/f	57	-2	Dynamic
Surface velocity	57-60	-1	Dynamic
Surface kinetic energy	60	+1	Dynamic
CUTI _{box} (balance)	63	+4	Transport
CUTI _{south} (export)	65	+6	Transport
Bakun forcing (wind)	70	+11	Wind

3.6. Thermohaline response

The cyclonic eddy formed at the cape uplifts the isopycnals and produces a cold, saline signature. From day 50, at the end of February, a cold anomaly develops near the cape and Core 2, accompanied by simultaneous salinization. This cooling-salinization association indicates a contribution from colder, saltier subsurface waters.

Depth-time profiles show a progressive decrease in temperature throughout the upper water column, with a pronounced signal between 25 and 75 m. The dynamic maximum observed during the separation and intensification phase precedes the surface thermal minimum. This lag indicates that the thermohaline response develops more slowly than the eddy dynamics (Figure 14).

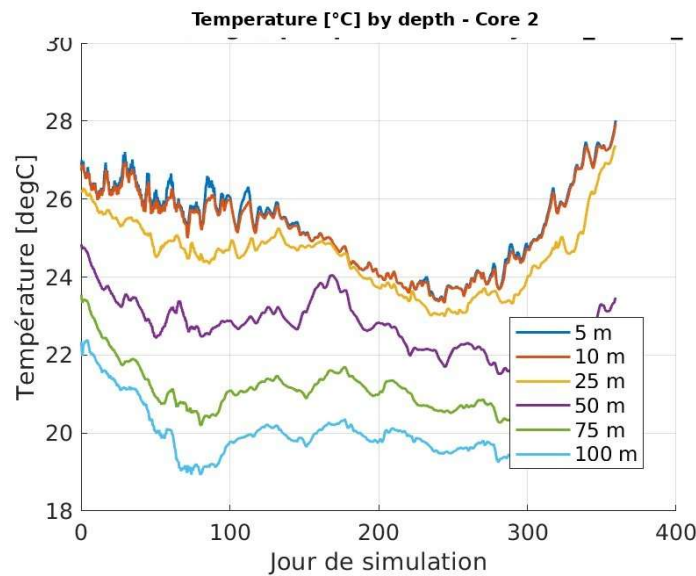


Figure 14. Depth-time evolution of temperature at Core 2.

3.7. Recurrence of the mechanism

After detachment of the first eddy, surface kinetic energy decreases and is subsequently redistributed into filaments. Between days 150 and 280 (June-October), cooling reaches its maximum spatial extent, with local dynamic upwelling superimposed on the regional seasonal cooling of the austral winter. Around day 250 (early September), a second cyclonic cell forms, visible in both vorticity and sea level. This second generation indicates that the separation process at Cape Sainte-Marie can recur during the simulated year.

4. DISCUSSION

The diagnostics converge on a two-stage mechanism. The first stage corresponds to the separation of a boundary flow at Cape Sainte-Marie. This separation generates a cyclonic cell, identified by the vorticity dipole at the cape and the associated sea-level depression. The second stage corresponds to baroclinic amplification of this cell, evidenced by positive and dominant baroclinic conversion at the cape, the Charney-Stern criterion, and the vertically sheared current structure. This mechanism is comparable to that described in other eastern-boundary upwelling systems. In such systems, current separation at a cape generates a cyclonic cell that uplifts and advects dense waters. Off central Chile, a cyclonic wake cell enhances upwelling at Cape Punta Lavapié [11]. South of Taiwan, Kuroshio separation at a cape generates cyclonic eddies through recirculation [12]. The present study extends this mechanism to Cape Sainte-Marie and identifies it as the origine driver of Core 2.

This distinction is important. Baroclinic instability is not necessarily the initial trigger of the eddy; rather, it acts primarily as an amplification and maintenance mechanism after current separation. The onset is linked to cape geometry, bathymetry, and the presence of a pre-existing boundary flow. The resulting eddy uplifts the isopycnals, produces a cold, saline anomaly, and subsequently exports upwelled waters toward the southern gyre.

The role of wind must therefore be interpreted with care. Results from the CUTI study [2] show coherence between the Bakun index and CUTI in some sectors, confirming the existence of a wind-driven component of the transport. However, the diagnostics presented here show that Core 2 origin cannot be reduced to a direct response to local wind forcing. Wind acts as a strengthening or modulation factor for the upwelling signal, whereas the formation mechanism is based on current separation at the cape and subsequent baroclinic amplification.

The localization of energy generation at Cape Sainte-Marie provides a central line of evidence. If the upwelling were exclusively wind-driven, the dynamic response would be more broadly distributed along the coast, following the spatial structure of upwelling-favorable winds. Here, the cores of baroclinic conversion, the emergence of vorticity bands, and the sea-level depression are concentrated at the cape. This spatial localization is characteristic of current-separation upwelling.

These results complement and refine the conclusions of Ramanantsoa et al. [1]. Those authors distinguished a dynamic Core 1 associated with separation of the East Madagascar Current and a Core 2 mainly related to wind forcing.

The diagnostics presented here show that Core 2 also exhibits current-separation dynamics, but associated with the Cape Sainte-Marie sector and export toward the southern gyre. Core 2 may therefore be interpreted as an analogue of Core 1 while retaining its own regional dynamics and interaction with exchanges originating from the Mozambique Channel.

5. CONCLUSION

Core 2 upwelling south of Cape Sainte-Marie is interpreted here as a current-separation upwelling system. Separation of a boundary flow at the cape triggers the formation of a cyclonic cell that uplifts the isopycnals and produces a cold, saline anomaly. The cell subsequently detaches from the cape, migrates westward, and exports upwelled waters toward the southern gyre.

Baroclinic instability contributes to eddy amplification and persistence. It is evidenced by positive and dominant baroclinic conversion at the cape, a $|BC|/|BT|$ ratio greater than 1 in the origin source region, and a sign reversal of the potential-vorticity gradient in the vertical. The process recurs during the simulated year, with a second eddy generation around day 250.

These results show that Core 2 should not be interpreted solely as wind-driven upwelling. Wind strengthens the signal, but the origin of the process is linked to current separation at the cape and baroclinic eddy amplification. This mechanism brings Core 2 closer to Core 1 described by Ramanantsoa et al. [1], while highlighting the specific dynamics of the southwestern Cape Sainte-Marie sector. This conclusion is based on a one-year simulation; the climatological representativeness of the mechanism should be assessed in a future multiyear study.

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