Observations and Mechanisms for the Formation of Deep Equatorial and Tropical Circulation

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Key Points:

- Observation of deep Equatorial and Tropical Circulations : a complex system of zonal jets
- Review of theories explaining the generation mechanisms of the Equatorial and Tropical jets
- Remaining gaps and future challenges in our global understanding of Deep Equatorial and Tropical Circulations

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Abstract

The Intermediate and Deep Equatorial and Tropical Circulations (DEC and DTC) consist of a complex system of zonal jets. This paper attempts at unifying existing observations and theories to present our current understanding of this jets system.

Recent in-situ observations suggesting a continuity between DEC and DTC are confronted against the various generation mechanisms that have been proposed in the literature. The key notion to differentiate these previous studies lies in the so-called "cascade of mechanisms", i.e. the energy pathway and equilibration processes chain that lead to the jets from their initial energy source.

Many studies see the Deep Equatorial Intra-seasonal Variability (DEIV) as the initial energy source, highlighting its key role in energizing the DEC and DTC. However, critical gaps remain in this cascade of mechanisms and limit substantially our ability to represent the jets in Ocean Global Circulation Models. This paper aims at identifying such gaps and propose future research directions.

1 Introduction

Understanding and modeling the intermediate and deep circulation in the equatorial oceans remains one of the great challenge in physical oceanography. For long, observations have shown that the deep equatorial circulation consists in several systems of zonal jets with rather complex features [*Luyten and Swallow*, 1976; *Eriksen*, 1981, 1982; *Leetmaa and Spain*, 1981; *Firing*, 1987; *Ponte and Luyten*, 1989, 1990; *Firing et al.*, 1998; *Gouriou et al.*, 1999, 2001; *Rowe et al.*, 2000; *Johnson et al.*, 2002; *Dengler and Quadfasel*, 2002; *Bourlès et al.*, 2003; *Ollitrault et al.*, 2006; *Bunge et al.*, 2008; *Ascani et al.*, 2010, 2015; *Brandt et al.*, 2011; *Cravatte et al.*, 2012; *Qiu et al.*, 2013a; *Ollitrault and Colin de Verdière*, 2014; *Youngs and Johnson*, 2015; *Cravatte et al.*, 2017]. However, our current knowledge of the spatial structure and variability of the jets system is far from being comprehensive, and remains an intense subject of investigation. The very existence of these zonal jets has even been questioned by some studies, claiming that the sparse observations only captured aliased planetary waves signals [e.g. *Jochum and Malanotte-Rizzoli*, 2003]. More recent and comprehensive observational datasets however corroborated the existence of mean zonal currents, on which seasonal Rossby waves currents superimpose. They suggest that the near-equatorial zonal currents should not be considered as isolated pieces [*Cravatte et al.*, 2012, 2017], but rather appear as embedded into a broader meridional system of zonal jets extending to the tropics.

These new observations challenge the many studies that have attempted a theoretical explanation of the presence of the jets in the equatorial region [*Marin et al.*, 2000; *Jochum and Malanotte-Rizzoli*, 2004; *d'Orgeville et al.*, 2007; *Hua et al.*, 2008; *Ménesguen et al.*, 2009a; *Fruman et al.*, 2009; *Ascani et al.*, 2010, 2015] and further away in the tropics [*Marin et al.*, 2000; *Jochum and Malanotte-Rizzoli*, 2004; *d'Orgeville et al.*, 2007; *Hua et al.*, 2008; *Ménesguen et al.*, 2009a; *Fruman et al.*, 2009; *Ascani et al.*, 2010, 2015]. None of them has succeeded in explaining the whole currents system observed from direct measurements highlighting the lack of comprehensive theoretical understanding of Deep Equatorial and Tropical Circulations (DEC and DTC).

This paper is an effort to review our current knowledge of the equatorial and tropical dynamics from subsurface to depth. Its goal is to revisit the existing theories of the mechanisms leading to the Deep Equatorial Circulation (DEC) and Deep Tropical Circulation (DTC) formation, in the light of recent observations. It does not pretend to provide a comprehensive analysis of all the theoretical and modeling studies on zonal jets formation. Rather, it aims to discuss mechanisms specifically relevant for DEC formation, and their possible extension to DTC formation. Indeed, as a tribute to the memory of Lien Hua, we focus on *Hua et al.* [2008]'s theory, and the mechanisms they proposed.

We show that other theories on DEC share a common interpretation of the formation of jets as the results of a cascade of mechanisms. We will also address the question of the applicability of this theory and approach to a broader latitudinal range.

The paper first summarizes, in section 2, the observations of the main features of DEC and DTC, interpreted as a general system of zonal jets. It then discusses, in section 3, the existing theories for the formation of zonal jets, described as a cascade of mechanisms involving key energy sources from which the energy is extracted, and equilibration processes, explaining the formation of stable basin-scale jet structures. In the last section, a discussion on the importance of the key energy sources, focused on the Deep Equatorial Intra-seasonal Variability (DEIV), for the DEC and DTC characteristics is proposed, with open questions toward a possible unifying theory. Gaps in our current knowledge and needs for future model developments to better understand the equatorial and low-latitudes system are also discussed.

2 Observations of the Deep Equatorial and Tropical Circulation (DEC and DTC

One striking characteristic of the tropical oceans is the complex set of zonal currents observed at intermediate and deeper depths (500-2000m) over the whole basin. Cross-equatorial surveys revealed the presence of two patterns of alternating eastward and westward jets schematically represented in Figure 1 : (1) the Equatorial Deep Jets (EDJs), which are equatorially-trapped, vertically-alternating zonal currents with a vertical scale of a few hundreds of meters ; (2) the Extra-Equatorial jets (EEJ) or Equatorial Intermediate Currents (EICs), which are a series of latitudinally-alternating zonal currents with a large vertical extent, and a meridional wavelength of about 3°. This section describes the characteristics of these zonal current systems, their persistence, as well as their similarities and differences in the different basins. Main abbreviations of the different designations of zonal jets are summarized in Table 1 and their main characteristics are given in Table 2.

2.1 Characteristics of the Zonal Equatorial Deep Jets (EDJs)

EDJs are equatorially-trapped (1.5°N-1.5°S) eastward and westward jets stacked over the vertical, with typical wavelengths between 300 and 700 meters [*Youngs and Johnson*, 2015]. They are ubiquitous below the thermocline, extend basin-wide, and are visible in instantaneous profiles, suggesting that they are permanent features of the circulation. They have been observed in all three equatorial basins, though with different properties ; in the Pacific [*Eriksen*, 1981; *Leetmaa and Spain*, 1981; *Firing*, 1987; *Ponte and Luyten*, 1989; *Johnson et al.*, 2002], in the Atlantic [*Eriksen*, 1982; *Gouriou et al.*, 1999, 2001; *Bourlès et al.*, 2003] and in the Indian Oceans [*Luyten and Swallow*, 1976; *Ponte and Luyten*, 1990; *Dengler and Quadfasel*, 2002]. In the Pacific Ocean, they appear to be weaker (10 cm/s) and exhibit a smaller vertical wavelength (300-400 m) than in the Atlantic Ocean (20 cm/s, 500-700m) or the Indian Ocean (15-20 cm/s, 300-600m).

The EDJs zonal continuity, persistence and vertical migrations have been the subject of debate, because of the difficulty to characterize them with sparse data. Multi-years of moored velocity observations and large-scale hydrological profiles showed that EDJs exhibit a slow downward phase propagation (and associated upward energy propagation in the framework of linear wave theory), implying the existence of an energy source at depth, over a 12-30 years period in the Pacific Ocean [*Johnson et al.*, 2002; *Youngs and Johnson*, 2015], and a period of around 4.5 years in the Atlantic Ocean [*Johnson and Zhang*, 2003; *Bunge et al.*, 2008; *Brandt et al.*, 2011; *Claus et al.*, 2016] and in the Indian Ocean [*Youngs and Johnson*, 2015].

2.2 Characteristics of the Extra-Equatorial Jets (EEJs)

The extra-equatorial jets are a system of meridionally-alternating westward and eastward jets, with a large vertical extent, and a 3° meridional wavelength. They are observed at intermediate depths (500-1500m) in both the Atlantic and Pacific Oceans. No evidence of such zonal current systems has been provided yet in the Indian Ocean.

Our knowledge of these zonal jets first came from synoptic cross-equatorial sections ; they were first thought to be confined to the near-equatorial 3°S-3°N band [*Firing*, 1987], with a striking zonal coherence across the Pacific basin [*Firing et al.*, 1998]. In this latitudinal band, this system is composed of the eastward North and South Intermediate Countercurrents (NICC and SICC) located at 1.5°-2°, and the westward North and South Equatorial Intermediate Currents (NEIC and SEIC) found at 3° [*Ascani et al.*, 2010]. Extended meridional sections revealed that similar westward and eastward jets are also present further off-equator [e.g. *Gouriou et al.*, 2006; *Qiu et al.*, 2013a].

The advent of the Argo program allowed estimations of 10-day mean velocity at the floats' parking depth and provided the first basin-wide pictures of zonal currents at 1000m depth in the Atlantic and Pacific Oceans [*Ollitrault et al.*, 2006; *Ascani et al.*, 2010, 2015; *Cravatte et al.*, 2012, 2017; *Ollitrault and Colin de Verdière*, 2014]. These studies confirmed that alternating zonal jets extend in the tropics to at least 18°, with a remarkable zonal coherence in the basin at this depth, and suggested that the NICC, SICC, NEIC and SEIC may be part of a much broader meridionally-alternating current system. In the Pacific and Atlantic, these jets have mean velocities of about 5 to 10 cm/s but instantaneous estimations are much larger (see *Cravatte et al.* [2017]). They are stronger in the Southern Hemisphere, stronger in the western part of the basin, and weaken and disappear toward the eastern part of the basin [*Cravatte et al.*, 2012].

Recently, *Cravatte et al.* [2017] investigated the vertical structure of these EEJs in the Pacific Ocean with a combination of direct velocity measurements and geostrophic estimations. They found a complex vertical structure, consisting in two apparently distinct systems of meridionally-alternating zonal jets equatorward of 10° (Figure 1 ; see *Cravatte et al.* [2017] and Table 1 for the terminology) :
− above 800m, the Low Latitude Subsurface Countercurrents (LLSCs) including the Tsuchiya jets [*Rowe et al.*, 2000], are found just below the thermocline and feel its large-scale slope. These jets deepen and get denser poleward ; they also shoal to lighter density and shift poleward from west to east, thus exhibiting a variable meridional wavelength [*Cravatte et al.*, 2017]. Their mean amplitude of about 20 cm/s or greater close to the equator, weakens to 5 cm/s further poleward.
− below 800m, the Low Latitude Intermediate Currents (LLICs) (including the SICC, NICC, NEIC and SEIC) seem to be a different system of meridionally-alternating zonal jets with a smaller wavelength (3°, Table 2), found on a large vertical extension down to 2000m. Unlike the LLSCs, the latitudinal positions of the LLICs remain constant throughout the basin [*Cravatte et al.*, 2012, 2017].
Both systems of currents merge and are indistinguishable poleward of 10°.

Excepting the Tsuchiya jets which are permanent features of the circulation, the EEJs amplitude and position vary from one instantaneous section to another [e.g. *Gouriou et al.*, 2006; *Cravatte et al.*, 2017]. This is partly explained in the near-equatorial band by the seasonal variability of the currents, related to vertically-propagating annual Rossby waves [*Marin et al.*, 2010]. As the seasonal zonal current anomalies are of similar amplitude or larger than the mean currents on which they superimpose, EEJs may disappear or reverse direction from one section to another. Variability at intra-seasonal and interannual timescales is also observed [*Firing et al.*, 1998; *Cravatte et al.*, 2017], though largely unexplored. These jets and their zonal continuity is thus better revealed on averaged sections, raising the question of the nature of these jets and whether they are latent or permanent features of the circulation. Thus their transport properties have to be assessed. At least, the comparison of 1000m-depth velocities from Argo floats displacements averaged over different periods of 4 years covering the Argo era give similar amplitudes and positions for the LLICs (not shown). This suggests that the LLICs interannual variability, if any, is at periods longer than the Argo era (15 years) ; this also suggests that LLICs are not the result of averaging randomly distributed eddies.

3 Review of proposed mechanisms for the formation and equilibration of the DEC and DTC

3.1 A cascade of mechanisms

Recently, different theories have been proposed to explain some elements of the system of zonal jets described in Figure 1. As in *Hua et al.* [2008] these theories involve a cascade of mechanisms transferring energy from an initial source to equilibrated zonal jet structure. In this cascade of mechanisms, the main ingredients that we can identify are : a key energy source, transfer processes and equilibration processes.

This cascade of mechanisms is schematically represented in Figure 2. Three main energy sources have been proposed in the literature for the generation of various parts of the tropical system of zonal jets :
– the Deep Equatorial Intra-seasonal Variability (DEIV) : originating either from the wind, the instability of deep western boundary currents or from Tropical Instability Waves (that are triggered by the instability of near-surface currents and then partly propagate their energy to depth), this energy source have been invoked for the generation of EDJs and the surrounding SICC and NICC [e.g. *Hua et al.*, 2008; *Ascani et al.*, 2015] or for the observation of Tsuchiya Jets in *Jochum and Malanotte-Rizzoli* [2004].
– Extra-equatorial annual Rossby waves : forced by the seasonal variations of winds, they can lead to the formation of the extra-equatorial zonal jets at latitudes where the LLSC and LLIC have merged (typically poleward of 10°) [e.g. *Qiu et al.*, 2013b].
– Regional permanent features of the ocean circulation or of the wind forcing : upwellings or the equatorial shoaling of the thermocline can explain the formation of the first SSCCs (or Tsuchiya jets) below the thermocline ; small meridional-scale structure of the wind stress curl has also been proposed away from the equator [e.g. *Marin et al.*, 2003; *Taguchi et al.*, 2012].

These different energy sources thus apply to different regions of the equatorial or tropical ocean. The two first energy sources are associated with time-variable forcings (intra-seasonal or seasonal waves), whereas the last one can be mainly seen as a response to permanent ocean or atmosphere forcing. These energy sources are prone to instabilities and mixing, that convert initial energy to other scales. Finally, equilibration mechanisms, involving turbulent processes (e.g. nonlinear rectification) or waves interactions (e.g. basin adjustment or modes, secondary instabilities such as inertial or parametric instabilities), lead to the final formation of the zonal jets (cf Figure 2).

As discussed below, if DEIV is thought to be the main mechanism for the formation of the vertically-alternating EDJs, there is not a unique key forcing for the system of meridionally-alternating zonal jets. Different mechanisms are invoked for various elements of this system, and some have no theory to explain their formation. Because all jets belong to an apparently organized system, we discuss their generation mechanisms following a common formalism. Therefore the following sections follow the schematic diagram of Figure 2 and discuss first the different key energy sources and the associated energy transfer processes, and then the equilibration processes of the zonal jets.

3.2 Key energy sources and energy transfer processes toward zonal jets formation

3.2.1 The Deep Equatorial Intraseasonal Variability (DEIV) and the DEC formation

Most recent studies [*Hua et al.*, 2008; *d'Orgeville et al.*, 2007; *Ménesguen et al.*, 2009a; *Fruman et al.*, 2009; *Ascani et al.*, 2010, 2015] have identified the Deep Equatorial Intraseasonal Variability (DEIV) as the main source of energy for the generation of permanent zonal jets in the vicinity of the equator (red boxes in Figure 2). The DEIV is associated with an equatorial Mixed Rossby-Gravity wave (so-called MRG or Yanai wave) with different characteristics. Figure 4 illustrates in a non-dimensional diagram the wavenumber of MRG waves characteristics for some studies. *Ménesguen et al.* [2009a] or *d'Orgeville et al.* [2007] used short waves as primary energy sources (and long waves were stable in their simulations), while *Ascani et al.* [2015] and specially *Ascani et al.* [2010] considered longer waves that are more divergent and further away from short Rossby waves characteristics. Different processes have thus been invoked for extracting energy from the primary waves.

Hua et al. [2008]; *d'Orgeville et al.* [2007]; *Ménesguen et al.* [2009a]; *Fruman et al.* [2009] considered energy transfers when primary MRG waves are unstable, transferring their energy to smaller scales. Their theory comes from *Lorenz* [1972]'s and *Gill* [1974]'s theory who have studied, in a 2D β-plane configuration, the destabilization of Rossby waves. *Gill* [1974] showed that Rossby waves are always unstable. The destabilization process is a function of the wave intensity, measured by the adimensional number $M = Uk^2/\beta$, where U is the maximum current of the wave, k is its wavenumber and β the planetary vorticity gradient. Strong waves (moderate to large M) are subject to barotropic instability, weaker waves (small M) to triadic instabilities. In both cases, the growth rate of the most unstable mode is proportional to Uk. For strong waves, the most unstable mode has a wavenumber perpendicular to the one of the primary wave with a wavelength of the order of $1/k$. It means that a short zonally-sheared wave will be destabilized into zonal jet-like structures. In the equatorial region, this study has been adapted by *Hua et al.* [2008] to the destabilization, in a stratified ocean, of short equatorial MRG waves (which have characteristics similar to short Rossby waves). Thus, *Hua et al.* [2008] showed that short equatorial MRG waves are destabilized into zonal currents with a meridional scale similar to the (short) zonal wavelength of the primary wave, independently from the primary wave's vertical structure. Because, in the equatorial region, the vertical scale of the baroclinic waves is tied to their meridional scale through the equatorial radius of deformation, short meridional scale is reached through either high meridional wavenumber, either high vertical modes. Thus, the specificity of the equatorial region is that the destabilization of a MRG wave produces both EDJ- and EEJ-like structures : the emerging EDJs are identified as a high baroclinic basin modes (a combination of long Kelvin and Rossby waves) and its vertical scale is proportional to the square of the zonal wavelength of the primary wave, while the emerging EEJs are identified as a low baroclinic basin modes, but with high meridional wavenumber.

In an idealized basin configuration, *Ascani et al.* [2015] [see also *Matthießen et al.*, 2017] have imposed DEIV and achieved an equilibrated simulation reproducing main features of the DEC, like EDJs and EEJs. They analyzed energy transfers between waves and they confirmed that MRG waves (from 30- to 100 day period) participate in the generation of EDJs. They also evidenced energy transfers from short Rossby waves (with periods between 50 and 100 days and large vertical scale) to the EEJ. This is in agreement with *Hua et al.* [2008], but more unexpectedly, they also evidenced a more complex energy transfer from EDJ basin modes to the EEJs.

3.2.2 The extra-equatorial annual Rossby waves and the DTC formation

Focusing on the generation of three subthermocline eastward jets between 9°N and 18°N below the thermocline in the northern Tropical Pacific, *Qiu et al.* [2013b] propose annual Rossby waves as a key energy source (blue boxes in Figures 2). Because of their different characteristics in comparison with DEIV, the processes involved in the energy transfer and equilibration to jets are different from what we discussed above. From both theoretical considerations and numerical models, they conclude that these low baroclinic zonal jets result from two successive processes : (i) the generation of meso-scale eddies through the instability, by wave-triad interactions, of wind-forced annual first-baroclinic Rossby waves in the eastern Pacific and (ii) the subsequent formation of zonal jets through the convergence of potential vorticity (PV) fluxes associated with the eddies. In this framework, the jets appear as latent jets (following the denomination proposed by *Berloff et al.* [2011]), i.e. as weak jets relative to ambient eddies, for which long-term time averages are needed to isolate them.

3.2.3 Permanent forcings for subthermocline DTC formation

The Tsuchiya jets (or primary SCCs as contoured in green in the schematic diagram of Figure 2) were the first observed jet structures off the Equator. Early studies have been argued that they simply result from permanent regional oceanic or atmospheric forcings. Besides TIWs [*Jochum and Malanotte-Rizzoli*, 2004], three permanent energy sources have been proposed to explain the creation of the Tsuchiya jets :

– The Equatorial UnderCurrent (EUC) : *McPhaden* [1984] was the first to analyse the dynamics associated with the Tsuchiya jets, seen as lobes of the EUC on both sides of the equator, resulting from a balance between the poleward diffusion of the cyclonic vorticity associated with the EUC, and the poleward advection of planetary vorticity. Such a balance can only hold in the westernmost part of the equatorial Pacific, and other processes (such as nonlinearities) are required to explain the poleward shift of the Tsuchiya jets (and their progressive separation from the EUC) from West to East.

– The eastern boundary upwelling system : From both analytic and numerical layered models of the tropical Pacific, *McCreary et al.* [2002] suggest that the Tsuchiya jets are mostly driven by upwelling along the South America coast and in the Costa-Rica Dome below the InterTropical Convergence Zone (ITCZ). The Tsuchiya jets are explained as geostrophic currents along arrested fronts, generated by the convergence or intersection of the characteristics of the Rossby waves carrying information about the density structure away from the upwelling regions. Diapycnal fluxes between layers, and the presence of a prescribed Pacific inter-ocean circulation (representing the Indonesian throughflow and a compensating inflow at greater depths from the South), are necessary to create the jets. These results have been further supported in global and regional OGCMs forced by idealized winds [*Furue et al.*, 2007, 2009].

– The equatorward shoaling of the thermocline : *Marin et al.* [2000] propose a mechanism similar to the atmospheric Hadley cells to explain the Tsuchiya jets. In response to the large-scale equatorial shoaling of the ventilated thermocline, meridional ageostrophic and diapycnal cells are created and redistribute angular momentum, creating eastward jets analogue to the atmospheric Jet Streams. In their two-dimensional model, the meridional structure of the thermocline is prescribed and diapycnal fluxes result from an ad hoc relaxation to this background density field. In subsequent papers, *Hua et al.* [2003] and *Marin et al.* [2003] show that this mechanism still holds for a fully three-dimensional Primitive Equations model where the ventilated thermocline is forced by basin-scale Ekman pumping, with strong equatorial recirculations.

The small-scale structure of wind stress curl has also been proposed as a mechanism for the formation of zonal jets below the thermocline, in the presence of islands or in the Southern Pacific Ocean. The local wind stress curl anomalies generated in the lee of an island has been recognized as a forcing for zonal circulation west of this island [e.g. *Belmadani et al.*, 2013]. Following *Kessler and Gourdeau* [2006], *Taguchi et al.* [2012] suggest that the deep oceanic zonal jets observed in the southern tropical ocean are partly forced by coupled ocean/atmosphere processes. In their coupled general circulation model, the deep zonal jets have a signature at the surface, inducing Sea Surface Temperature (SST) anomalies at the same spatial scales through zonal advection. These SST anomalies impact the atmosphere, creating small-scale wind stress curls collocated with the temperature anomalies, which reinforce the zonal jets in accordance with Sverdrup balance.

If these different theories can associate subthermocline zonal jets structure to large-scale or regional permanent forcings, they however can not explain deeper zonal jets. These theories are not incompatible and it is probable that a combination of the different energy transfer processes, involving different key energy sources, are taking place in the equatorial and tropical regions.

3.3 Equilibration processes

After the transfer of energy from key sources to new spatial and temporal scales in equatorial and tropical areas, equilibration processes thus have to take place to form DEC and DTC (cf Figure 2).

3.3.1 DEC equilibration

The equilibration mechanism associated with the equatorial wave instabilities proposed in *Hua et al.* [2008] is a direct structuring of the instability of short MRG waves into zonal currents. The zonal location, the meridional extension of the initial wave, or the presence of boundaries does not modify the tendency to zonal structuring of the resulting circulation but has an influence on the zonal extension of the jets. It has been shown that the MRG wave destabilization produces long high baroclinic Kelvin waves (EDJs-like structures) propagating eastward and long low baroclinic Rossby waves (LLICs-like structures) propagating westward [*Hua et al.*, 2008; *d'Orgeville et al.*, 2007; *Ménesguen et al.*, 2009a; *Fruman et al.*, 2009].

– Inertial instability : short time scale equilibration

This process has been shown to explain the fast homogenization of PV to zero inside the westward jets of the EDJs, while PV exhibits strong gradients on their rim [*Ménesguen et al.*, 2009b]. Since the Coriolis parameter changes its sign at the equator, the equatorial band is particularly propitious to inertial instability [*Hua et al.*, 1997; *d'Orgeville and Hua*, 2005; *Fruman et al.*, 2009]. Inertial instability has only been invoked as a secondary instability and is not necessary to explain the extraction of energy and the equilibration into zonal jets but it influences the final structure of the EDJ.

– Influence of basin modes : long time scale equilibration

As discussed in *Matthießen et al.* [2017], eastern and western boundaries have a strong influence on the equilibration of the zonal jets. In a basin configuration, East and West walls generate reflexion of equatorial waves. The combination of Kelvin waves and their reflection into a first meridional-mode long Rossby wave create a basin mode with a period $T_n = 4L_B/c_n$, with L_B the basin length and c_n the gravity phase speed of the baroclinic mode n. The periodicity for this basin mode is about 4.5 years in the Atlantic for the vertical mode observed for the EDJs [*Johnson and Zhang*, 2003] and about 12-30 years in the Pacific [*Youngs and Johnson*, 2015], with a great uncertainty due to weaker amplitude of EDJs and a broader bandwidth for their vertical mode in the Pacific.

Most of the studies analyzing the influence of basin mode on the structuring -and slow variability- of the DEC concerned EDJ. *d'Orgeville et al.* [2007] have highlighted such low-period oscillations in their simulated solutions of the EDJ. However, they used an idealized forcing over the whole depth which inhibited the upward energy propagation and the vertical dissymmetry observed in the ocean. Longer simulations were performed by *Ascani et al.* [2015] and *Matthießen et al.* [2015, 2017] and their use of a surface intensified forcing term lead to more realistic upward energy propagation for the basin modes, which is however not fully understood. Indeed, in their very long simulations (over 200yrs), *Matthießen et al.* [2017] have in fact found an alternance of downward and upward propagating phase of the basin mode. Observations are not long enough to validate or unvalidate their results and the consequence of the surface-intensified forcing rather than a forcing at depth on the basin mode energy propagation is thus still an open question.

To our knowledge, few studies have analyzed the effect of boundaries and basin modes on the structuring of the LLSC or LLIC. Argo available data do not cover a long enough time period to conclude. A noticeable exception is *Qiu et al.* [2013b]. Even though this is not clearly stated in their paper, their Figure 3 clearly exhibits a slow meridional propagation of the low latitude jets structure (with a time period of 50 years or so that is not discussed). This could be the propagating signature of high meridional basin modes, that could thus be involved in the formation of the whole EEJ structure, at least in numerical simulations.

3.3.2 Nonlinear rectification

Non-linear rectification can reinforce preexisting jets and contribute to their equilibration into permanent circulation features. The influence of eddies through the convergence of PV fluxes is invoked by *Qiu et al.* [2013b] to explain the existence of the low latitude jets. A similar mechanism was found in the tropical Pacific by *Ishida et al.* [2005], where high mesoscale eddy activity is shown to accelerate the SCC (especially for the northern Tsuchiya Jet in the eastern part of basin) and to generate large region of PV homogenization in which strong horizontal recirculations take place.

Besides the permanent forcings that are proposed to explain the Tsuchiya Jets, Tropical Instability Waves have been shown to be an additional contributor to the generation of the Tsuchiya Jets. Using the Transformed Eulerian Mean equations to explore the momentum balance of the southern Tsuchiya Jet in a numerical simulation of the Tropical Atlantic, *Jochum and Malanotte-Rizzoli* [2004] shows that the jet is maintained against dissipation by the convergence of the Eliassen-Palm flux associated with the TIW. The TIWs are also seen as an additional driver for the northern Tsuchiya Jet in *McCreary et al.* [2002], *Hua et al.* [2003] or *Furue et al.* [2009].

The convergence of PV or Eliassen-Palm fluxes invoked by these studies is associated with the tendency of eddies to reinforce the jets. Indeed, zonal jets are associated with zonal PV structures which act as a vortex guide, keeping eddies within some latitudinal bands. Eddies have the tendency to homogenize PV within their vicinity. The combination of both effects leads to a sharpening of the initial PV structure, with homogeneous PV region separated by sharp PV gradients, eventually leading to a staircase structure. In this case, the caracteristics of the eddies determine the final structure and the width of the jets [*Dritschel and McIntyre*, 2008; *Dritschel and Scott*, 2011; *Scott and Dritschel*, 2012].

Besides the equilibration mechanisms reviewed in this section, many other particularities of the ocean may influence the shape of the jets and explain the features we observe [e.g. the large scale zonal slope of the thermocline *Johnson and Moore*, 1997].

4 Discussion

DEIV has thus been invoked in many studies as the key energy source, mainly for the formation of DEC but also possibly of DTC. As discussed above, there is no consensus on the details of the transfer and equilibration processes, from DEIV to zonal jets, which depends on the DEIV structure, location and amplitude.

In the following sections, we therefore address questions concerning DEIV characteristics in previous studies, focusing on their differences, and in observations.

4.1 Influence of the DEIV location on the resulting DEC

As recalled before, in *Hua et al.* [2008], MRG destabilization in the equatorial rail will produce high baroclinic Kelvin waves propagating eastward and low baroclinic Rossby waves propagating westward. In this linear phase, the location of the destabilization is thus important for the extent of the zonal jets, specially in a basin configuration. *Hua et al.* [2008] used an artificial analytical deep forcing to create the primary MRG wave in a channel and in a basin configuration. *d'Orgeville et al.* [2007] and *Ménesguen et al.* [2009a] have used a similar analytical deep forcing restricted to the western boundary of an idealized Atlantic basin, mimicking the deep western boundary variability. In every cases, resulting zonal jets are propagating from the area where the primary wave is destabilized. In basin cases, some energy is lost at eastern or western boundaries. Therefore, the location of the source of the short MRG wave (propagating eastward) in the western boundary is important, principally for EEJ that are propagating west of the destabilization zone.

Ascani et al. [2010] focused on the Pacific basin with the forcing of a vertically propagating MRG wave excited at surface in a restricted equatorial band. They reproduced EEJ structures within and west of the forcing area with an amplitude that remains modest west of the forcing area. In their following study [*Ascani et al.*, 2015], DEIV is still associated with tropical instability waves transferring the surface variability to the deeper ocean. Their idealized configuration of an Atlantic basin reproduces a DEC with EDJ and EEJ like structures in a 6-month mean field. They have noted that the strength of the EEJ rapidly diminishes away from the western boundary and remains weak if realistic coastlines are used.

To summarize, DEIV directly forced at depth or propagating from the surface produce deep jets structures. In a basin configuration, the horizontal location has an impact on the zonal extension of resulting jets and vertical location an impact on the vertical propagation (as discussed in section 3.3.1).

4.2 Influence of the DEIV amplitude and frequency on the resulting DEC and DTC

To be able to differentiate the dynamical regimes applied in previous studies and characteristics of the MRG waves representing DEIV in previous studies, using the vorticity and divergence equations, we derive the general equation governing the dynamics of these waves, including their forcing. As shown in appendix A: , for trapped equatorial waves this equation is :

$$\underset{\text{I}}{\frac{c^3}{c_G^2 c_{sR}}\delta_{xtt}} \quad + \underset{\text{II}}{\frac{c}{c_G}\delta_x} \quad + \underset{\text{III}}{v_x} \quad + \underset{\text{IV}}{\frac{c}{c_{sR}}\nabla^2 \frac{P_{xt}}{\rho_0}} \quad + \underset{\text{V}}{M^*[(\mathbf{u}\cdot\nabla\zeta)_x + ONLT]} \quad = O(M^{*2}) \tag{1}$$

where P is the pressure, $\mathbf{u} = (u,\ v)$ is the horizontal velocity field, $\delta = \partial_x u + \partial_y v$ is the horizontal divergence, $\zeta = \partial_x v - \partial_y u$ the vorticity and $ONLT$ stands for

'other nonlinear terms' whose form is not important for the discussion. The other factors are scaling characteristics associated with the wave types and are explained in appendix A: . The terms (I,II,III,IV) are the classical linear part leading to MRG wave while (V) is the nonlinear term responsible for triadic resonance and barotropic instabilities leading to turbulence and jets. Table 3 gives their order of magnitude for selected previous studies involving DEIV :
– The scaling shows that *Ascani et al.* [2010] force the equatorial ocean more linearly : the nonlinear term is smaller than the linear ones, favoring the gravity part of the MRG wave and departing from the barotropic-like instability (in accordance with Figure 4). Diabatic processes (vertical mixing) thus has to be invoked to transfer energy from this type of waves.
– At higher latitudes, *Qiu et al.* [2013b] force their simulations with $M^* << 1$ which, following *Gill* [1974]'s theory, is a regime of triadic instabilities, as interpreted by the authors.

The length scale needed to destabilize the forced equatorial MRG wave is given by $L_{dest} \simeq 25\frac{c_Y}{kFr_{Hua}}$ [see appendix A: and *Ménesguen et al.*, 2009a]. It is evaluated for the forcing used the different authors in Table 3. For *Qiu et al.* [2013b], we replaced the MRG wave speed by the long Rossby wave speed. When the destabilization length scale is small compared to basin widths, the models have enough time to develop jets (as it is the case for most of the cases of the table 3). But, we confirmed also that the wave forced by *Ascani et al.* [2010] is very stable compared to other studies sampled here.

4.3 DEIV observations

The deep intraseasonal variability can be estimated thanks to the Argo floats, which drift at 1000-m depth during approximately 9 days between two vertical profiles. Averaging drift velocities from thousands of floats has yielded regional and global maps of 1000-m mean velocity [e.g. *Cravatte et al.*, 2012; *Ollitrault and Rannou*, 2013]. Averaging the anomalous drift velocities can provide estimates of eddy kinetic energy (EKE) at 1000m [*Ascani et al.*, 2015].

Here, we compute for each float dive the 10-day mean lagrangian drift velocity anomaly relative to the mean seasonal velocity at the same month and location, and compute for each float dive the 'v-EKE' from the square of the meridional component anomaly only. We then map these EKE values with an optimal interpolation method to obtain an estimate of the deep intraseasonal variability (Figure 3). *Bunge et al.* [2008] found in moored observations that the meridional velocity component at depth in the Atlantic Ocean is dominated by fluctuations at 20-45 days, but that a substantial fraction lies between 10 and 20 days. *Ogata et al.* [2008] also found variability in the deep meridional velocity in the eastern Indian Ocean at 15 days, associated with the propagation of MRG waves. It should be noted that our estimation relies on 10-day lagrangian mean velocities : variability at high frequency (time period less than 20 days) is filtered, and a large part of inertia-gravity and mixed Rossby gravity waves spectrum is missing.

A high v-EKE is observed all along the western boundaries of the three basins, from 20°S to 20°N ; interestingly, it is also observed east of topographic obstacles shallower than 1000m : east of Madagascar Island, of the Mascarene Plateau and of the Chagos-Laccadive ridge in the Indian Ocean, east of the Gilbert Islands or Marianna trench in the Pacific Ocean. All these regions, associated with instabilities of deep western boundary currents, or local generation of meridional oscillations by interaction of currents with topography, are thus potential sources of DEIV radiating away. Meridional EKE is also higher in the equatorial band in the three basins, particularly in the eastern part of the Pacific. As discussed in *Ascani et al.* [2010, 2015], this may reflect the presence of energetic waves at depth, excited either by Tropical Instability Waves producing downward propagating MRG wave beams [*Weisberg et al.*, 1979; *Boebel et al.*, 1999; *Brandt et al.*, 2006; *Bunge et al.*, 2008; *Von Schuckmann et al.*, 2008] or by wind forcing.

Given the previous remarks on the importance of DEIV on the possible impacts on DEC and DTC formations, we understand that its too partial representation in each previously cited study compared to what is observed can be a key factor of misrepresentation of DEC and DTC.

5 Perspective and remaining questions

5.1 Possible avenues for progress in DEC realism in numerical simulations

If theoretical framework and idealized simulations are somewhat convincing in their capability to reproduce some pieces of the DEC, realistic simulations are far from exhibiting a complete DEC. In a realistic Atlantic equatorial configuration, *Eden and Dengler* [2008] are able to reproduce EDJ with the right characteristics, but with an underestimated amplitude compared to observations. Their simulation models the variability of the western boundary current but does not incorporate the wind forcing. The difficulty of approaching realistic configurations is particularly well illustrated in *Ascani et al.* [2015]. In this later study, two solutions are considered : one with a rectangular basin and with a zonally and temporally uniform wind forcing, and one with realistic coastlines and a zonally varying wind forcing annual cycle. Surprisingly, the first idealized solution produces the most realistic deep equatorial circulation. It is not clear whether it is the more realistic wind or the more realistic topography that degrades the realism of the simulation.

Several questions thus remain concerning the DEC formation in a realistic framework. The first issue is about the capability of models to correctly simulate the DEIV. None of the previous studies did try to combine several sources of variability, even in idealized configurations. In the different ocean basins, one can wonder if such a combination of different deep variability sources, with different frequencies, specific zonal or meridional locations, and time-intermittency could improve the simulation of the DEC. More investigations could be done in idealized simulations in combining forcing sources.

As suggested in Figures 3, sources of deep variability in meridional velocity can be located along the western boundary, as well as in the center of the basin, especially in the equatorial Pacific Ocean. How different MRG waves, emanating from the western boundary current variability and propagating downward from the surface, would merge and produce a coherent DEC is worth investigating. Moreover, the question of destabilization of a downward propagating wave compared with the destabilization of a primary wave directly forced at depth is not trivial. Another issue concerns the efficiency of a MRG beam energy propagation through a realistic stratification where the thermocline can be a serious obstacle. The representation of DEIV at depth is therefore dependent on a correct behavior of the model regarding its vertical propagation through a variable stratification.

MRG beams can also be considered as intermittent events forced by strong winds at given times. It would be worth investigating how the DEC may be impacted by time intermittency of the forcing, as it may be by its spatial distribution.

Furthermore, equatorial MRG waves can be excited at different frequencies and wavelengths. In *Hua et al.* [2008]'s theory, short MRG waves are destabilized in a characteristic time inversely proportional to their amplitude and wavenumber. Thus, longer waves or weaker waves can remain stable in a basin configuration (see green, orange and cyan points in Figure 4). It is not known how a DEIV full frequency spectrum would behave in a basin configuration. Which frequencies will be destabilized ? Which interactions between excited waves will emerge ?

To the our knowledge, in the Indian Ocean, existing observations did not reveal low-baroclinic zonal jets at low-latitudes. Mean circulation at 1000m inferred from Argo floats drifts do not show coherent zonal features resembling those found in the Atlantic and Pacific Ocean [*Cravatte et al.*, 2014]. The reason for this is not clear. It might arise because the DEIV in the Indian Ocean is too weak, or do not exhibit propitious frequencies to be destabilized. Alternatively, it might arise because of distinct mean oceanic thermohaline features, or because of a different basin configuration, with oblique coastlines, a northern frontier and a smaller size, inhibiting the development of basin modes for instance. Finally, it may be that some jets exist, but at shallower depths, or are masked by much higher variability. This would definitely require specific observations and investigations, with idealized model studies. Differences with other basin configurations can be a source of a better understanding of the involved mechanisms in the formation of DTC.

5.2 May Deep Tropical Zonal Circulation (20°N-20°S) be explainable with a single theory ?

Equatorial idealized studies have currently focused on DEC within 4°N and 4°S. However, as described in section 2, *Ollitrault et al.* [2006]; *Cravatte et al.* [2012]; *Qiu et al.* [2013a] and other studies have shown that zonal jets are alternating meridionally over a larger range of latitudes in the Pacific and Atlantic oceans. Could the theories explaining DEC be extended to the whole tropical band ?

A first avenue would be to extend the forcing to higher latitudes. As shown in Figure 3, variability is also found at higher latitudes along the western boundary in response to large-scale currents fluctuations. A forcing term similar to *Gill* [1974]'s theory, extending the equatorial response to a DEIV as studied by *Hua et al.* [2008] could be considered.

A second avenue would be to evaluate if DTC could be the result of a high meridional mode equilibration. We can also wonder if *Hua et al.* [2008]'s theory could produce low-baroclinic Rossby waves with high meridional modes. It would therefore be tempting to imagine that the DEC and DTC could be represented as a single structure with an emerging meridional wavelength. The question whether observed Low-Latitude Jets temporality are compatible with a meridional basin mode is still unknown as no observational evidence of a meridional propagation has been provided yet. Another difficulty lies in the complex vertical structure of these jets. They exhibit neither barotropic component nor a clear baroclinic mode and consist in two systems of jets one on the top of the other (cf. Section 2).

Finally, *Qiu et al.* [2013b] proposed a mechanism generating low latitudes zonal jets off equator. Their time-mean solution reproduces a set of such zonal jets, and their observations on two distinct periods encompassing several years suggest that they are at least quasi-stationary features. In their study, the zonal jets meridional scale is fixed by the wind forcing amplitude. The question of how a realistic wind could generate stable meridional scales thus remains open. More generally, the question of the persistence of these jets, their variation or steadiness in latitudinal position and meridional wavelength should be further investigated to see if theories and observations are consistent.

5.3 Is isopycnic turbulence a possible mechanism at low latitudes ?

As an alternative to mechanisms involving DEIV and Rossby waves, some authors have argued that the formation of jets in the ocean can be interpreted following Rhines theory [*Richards et al.*, 2006; *Baldwin et al.*, 2007]. *Rhines* [1975, 1979, 1994] has indeed shown that if the β effect is taken into account, the 2D turbulent cascade to large scales is stopped at some scale where Rossby waves become the driver of the evolution. *Vallis and Maltrud* [1993] [see also *Theiss*, 2004] have stated that the anisotropy leading to the emergence of zonal jets come from the fact that the cascade to larger scale can be pursued much further in the zonal direction, so that turbulence concentrate energy toward large zonal wavelength, eventually leading to the formation of zonal jets [1]. This process is complemented by the jet sharpening effect discussed in 3.3.2 [see also *Dritschel and McIntyre*, 2008; *Dritschel and Scott*, 2011; *Scott and Dritschel*, 2012, for more details].

In fact, any source of turbulence (from small scale noise to vortical structures generated by geostrophic instabilities), follows the anisotropic enstrophy cascade and eventually leads to the formation of zonal jets [*Baldwin et al.*, 2007; *Kamenkovich et al.*, 2009]). Since zonal jets alternating with latitude have first been observed in atmospheric flows of rapidly rotating planets, there exists a very rich literature on this subject in atmospheric and astrophysical sciences (for recent reviews, see for instance *Danilov and Gryanik* [2004]; *Ingersoll et al.* [2004]; *Liu and Schneider* [2010]; *Jougla and Dritschel* [2016], and references therein).

Theiss [2004] revisited Rhines theory and stated that the formation of zonal jets is only possible below a critical latitude, beyond which the turbulence remains dominant and the formation of jets is not possible. Interestingly, this process is consistent with the potential vorticity structures estimated in the tropical Pacific [*Rowe et al.*, 2000; *Cravatte et al.*, 2017; *Delpech et al.*, in prep.]. It is thus also possible to interpret jet-like structures in the ocean as a consequence of the Rhines theory and PV homogenization by isopycnic turbulence.

However Rhines theory has not been applied to realistic configurations and cannot explain all observed jet structures, in particular the vertically alternated EDJ, seen as basin modes, which do not intervene in Rhines theory. In addition, Fig. 3 shows v-EKE is intensified along western boundaries and the equator (note however that EKE is here estimated using 10 days averages). In most studies of the Rhines mechanism, isopycnic turbulence is generated over the whole basin and it is not clear if this is a necessary condition, or if localized turbulence could generate zonal jets extending over the whole basin.

Thus, DEIV direct equilibration and isopycnic turbulence anisotropic rectification are both mechanisms leading to the formation of zonal jets. Their initial source of energy differ : equatorial Rossby-Gravity waves for DEIV and small scale turbulence for isopycnic turbulence. The way energy cascades from large scale forcings is thus crucial to understand which process dominates. Designing experiments that can help distinguish between both processes, maybe differencing the equatorial region and the tropical region, is also an interesting challenge.

6 Conclusion

This study reviewed and discussed the current mechanisms explaining the complex system of zonal jets of the Deep Equatorial and Tropical Circulations. A schematic view of zonal jets structure as observed in the Pacific and Atlantic oceans is proposed in Figure 1. Mechanisms proposed in past studies have been summarized and revisited in terms of a cascade of mechanisms, arising from an initial energy source to the final equilibration of the Deep Circulation. A schematic view (Figure 2) is developed to classify and relate them to the different observed zonal jets systems. A particular stress has been put

1. This comes from the fact that Rossby waves having large zonal wavelength evolve very slowly. Turbulent effects are more rapid and thus continue to be the main driver of the evolution in this case, cascading energy to larger scales in the zonal direction.

on the importance of the DEIV and its ability to be destabilized into zonal jets. However, despite a large amount of theoretical work in the equatorial and tropical region, no consensus to explain the whole circulation has been reached and, above all, no realistic simulation has been able to reproduce convincingly the whole set of zonal jets as observed in the different oceanic basins. Some questions remain open and some new avenues are proposed for further investigations to clarify our global understanding of the Deep Equatorial and Tropical Circulations. We hope that this review will inspire new experiments and stimulate new theoretical work related to this exciting research subject.

Table 1. Main abbreviations found in the literature for the different zonal jets or systems of jets

Abbreviation	Signification
DEC	Deep Equatorial Circulation [a]
DEIV	Deep Equatorial Intraseasonal Variability
EDJ	Equatorial Deep Jets
EEJ	Extra Equatorial Jets
EIC	Equatorial Intermediate Current
SICC	South Intermediate Countercurrent
NICC	North Intermediate Countercurrent
NEIC	North Equatorial Intermediate Current
SEIC	South Equatorial Intermediate Current
LLIC	Low Latitude Intermediate Currents
SCC	Subsurface CounterCurrent
NSCC	North Subsurface CounterCurrent
SSCC	South Subsurface Countercurrent
sSSCC	secondary Southern Subsurface Countercurrent
EUC	Equatorial UnderCurrent
LLSC	Low Latitude Subsurface Countercurrents
SEC	South Equatorial Current
NECC	North Equatorial CounterCurrent
NEC	North Equatorial Current

[a] EDJ + EEJ (=EIC) = DEC

	EDJs		
	Atlantic	Indian	Pacific
Latitude range	1°S-1°N		
Longitude extent	Basin-wide		
Depth	500-3000m		
Intensity of zonal velocity	20cm/s	15-20cm/s (in the west)	10cm/s
Characteristic scale	$\Delta z = 500m$	$\Delta z = 300 - 600m$	$\Delta z = 350m$
Permanent or latent character	permanent		
Low frequency temporal variability	4.5 years	4.5 years	12-30 years

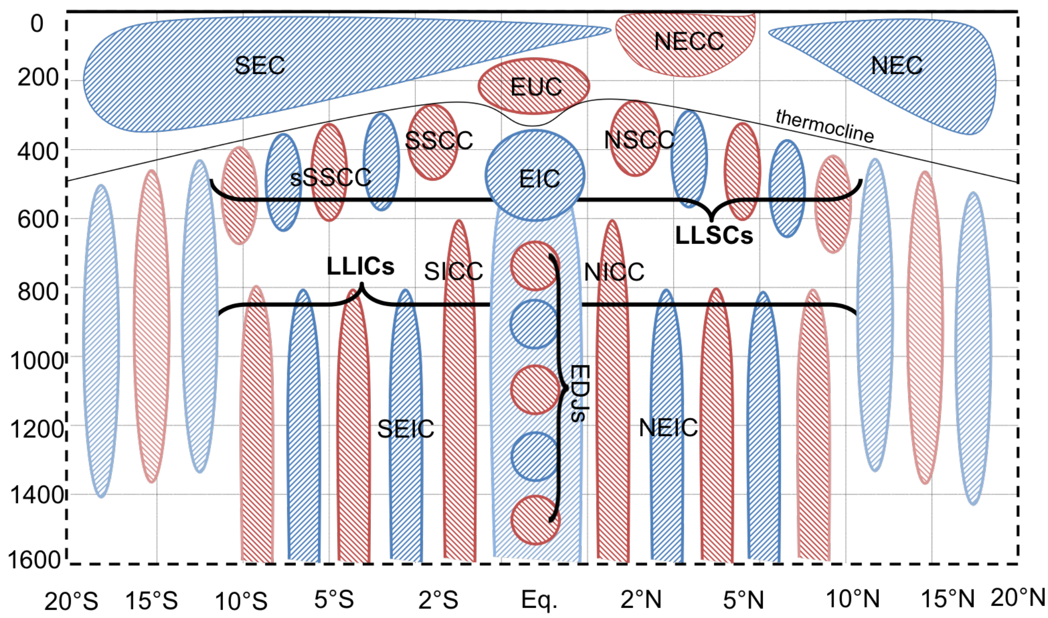
	LLSCs		
	Atlantic	Indian	Pacific
Latitude range			10°S-10°N
Longitude extent	N/A	N/A	Full basin : 8000km
Depth	N/A	N/A	Thermocline-600m
Intensity of zonal velocity	N/A	N/A	5 to 20 cm/s, decreasing poleward
Characteristic scale	N/A	N/A	$\Delta y = 3 - 4°$, increasing eastward
Permanent or latent character	N/A	N/A	latent
Low frequency temporal variability	N/A	N/A	N/A

	LLICs		
	Atlantic	Indian	Pacific
Latitude range	12°S-12°N	N/A	16°S-16°N
Longitude extent	Full basin : 2850km	N/A	Full basin : 8000km, with weaker structures at the east.
Depth	800-1000m	N/A	800-1400m
Intensity of zonal velocity	5 to 15 cm/s, decreasing poleward	N/A	5 to 10cm/s, decreasing poleward and eastward
Characteristic scale	$\Delta y = 2 - 3°$	N/A	$\Delta y = 3°$
Permanent or latent character	N/A	N/A	uncertain
Low frequency temporal variability	N/A	N/A	If any, > 10 years

Table 2. Main characteristics of EDJs, LLSCs and LLICs. N/A stands for not applicable.

	Wave speeds $[ms^{-1}]$ T [days], L [km], N $[s^{-1}]$, H [m], mode m, Fr or M	I	II	III	IV	V (M^*)	destabilization length
***Gill* [1974]**	$c = c_{sR}$	X	X	1	O(1)	$\ll 1$ triads ≥ 1 barotropic instability	
***d'Orgeville et al.* [2007]**	$c = 0.15$, $c_G = 1.6$, $c_{sR} = 0.15$ 40, 540, 2.10^{-3}, 5000, 2, $Fr_{Hua} = 0.2$	0.01	0.1	1	1.5	0.7	15°
***Hua et al.* [2008]**	$c = 0.1$, $c_G = 1.6$, $c_{sR} = 0.1$ 50, 410, 2.10^{-3}, 5000, 2, $Fr_{Hua} = 0.2$	4.10^{-3}	0.06	1	1	0.8	8°
***Fruman et al.* [2009]**	$c = 0.1$, $c_G = 3.2$, $c_{sR} = 0.1$ 50, 400, 2.10^{-3}, 5000, 1, $Fr_{Hua} = 0.15$	1.10^{-3}	0.03	1	1	0.8	10°
***Ménesguen et al.* [2009a]**	$c = 0.1$, $c_G = 1.6$, $c_{sR} = 0.1$ 50, 420, 2.10^{-3}, 5000, 2, $Fr_{Hua} = 0.125$	4.10^{-3}	0.06	1	1	0.5	12°
***Ascani et al.* [2010]**	$c = 0.35$, $c_G = 0.53$, $c_{sR} = 0.5$ 33, 1000, 2.10^{-3}, 5000, 6, $M_{Asc} = 0.2$	0.3	0.6	1	0.7	0.2	62°
***Qiu et al.* [2013b]**	$c = 0.1$, $c_G = 3$, $c_{sR} = 4$, $c_{lR} = 0.1$ 365, 3000, $g' = 0.018m^2s^{-1}$, 500, X, $U = 0.1ms^{-1}$	2.10^{-5}	1	1	0.03	0.03	11°

Table 3. Order of magnitude of the terms (I, II, III, IV, V) present in equation (A.5) for the different models referenced by their associated papers. The parameters of the forcing wave c, the gravity wave speed, the Rossby wave speed (short or long) are also shown. Terms are all compared to the β term in equation (A.5). Nonlinearity (V) is of order O(1) in most of the proposed models which favors barotropic instabilities leading to turbulence and jet formation. In *Qiu et al.* [2013b], this term is small and linear terms dominate leading to triadic resonance generating eddies which averaged exhibit jets. The length of destabilization is also shown in the last column. *Ascani et al.* [2010] shows quite a large length due to a faster Yana wave speed, shorter wave number and smaller nonlinearity. Values used here are crude mean estimates used in the models.



660 **Figure 1.** Schematic representation of the different system of currents found in the tropical oceans. Dashed blue patterns indicate westward flowing currents and dashed red patterns indicate eastward flowing currents. The lower boundary of the thermocline is represented with a solid line. The name of the main currents is indicated (see Table 1). For clarity and readability reasons, the meridional distance between jets is not scaled.

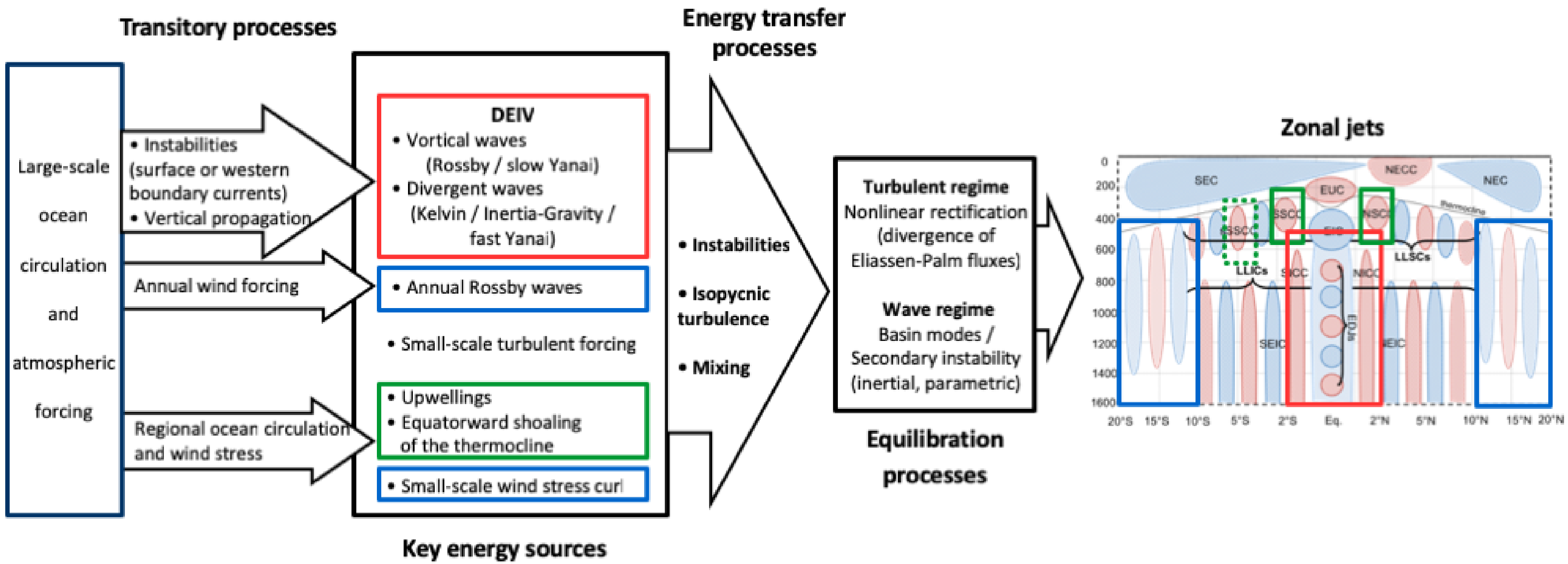


Figure 2. Schematic diagram of the cascade leading to the formation of zonal jets in previous studies. Initial large scale forcing excite key intermediate forcings which transfer energy to zonal structure via processes of instability, mixing or isopycnic turbulence. The permanent circulation is the result of equilibration of these zonal structures with oceanic conditions.

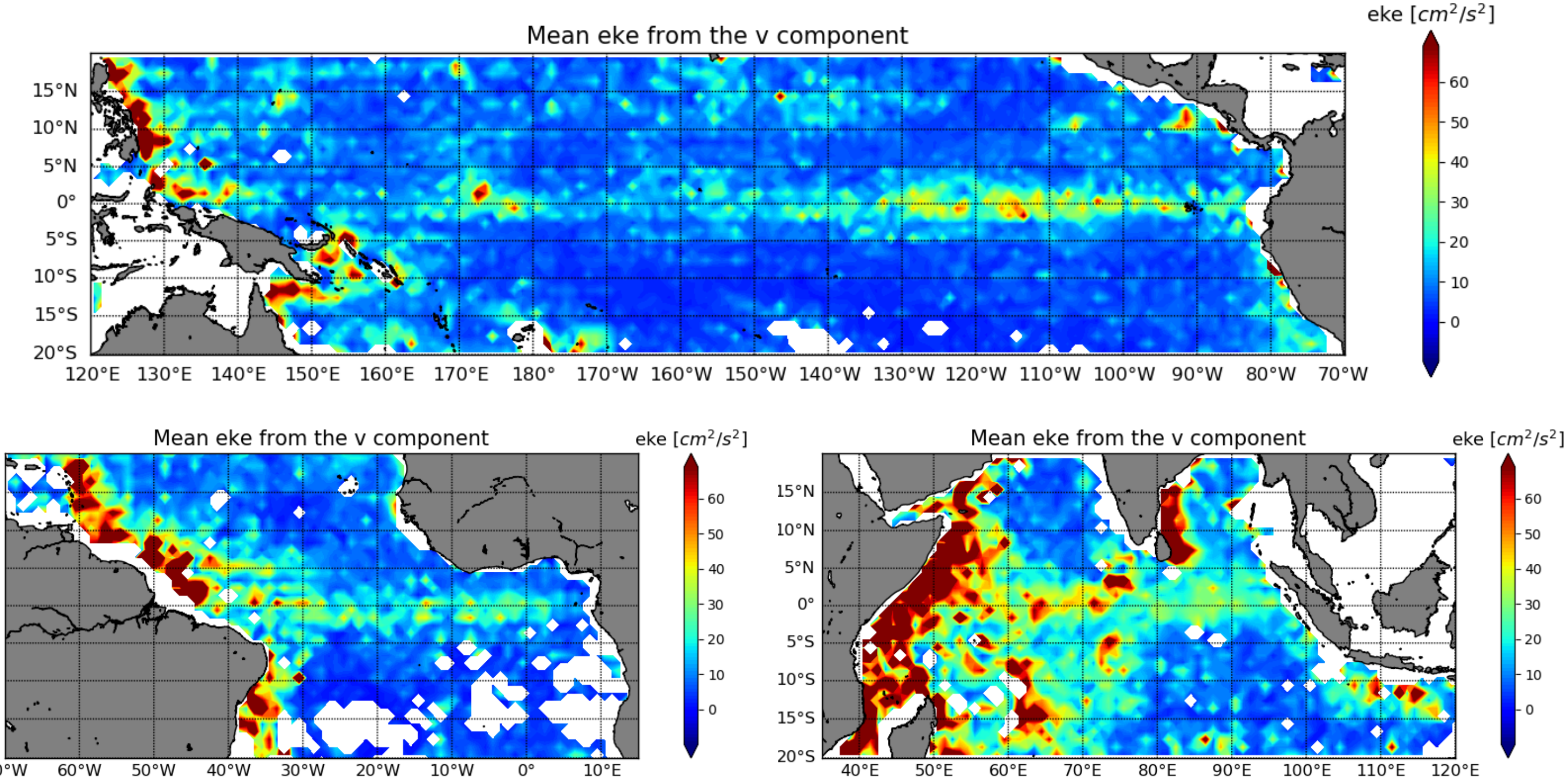


Figure 3. Mean Eddy Kinetic Energy of the v component from deep displacement of Argo floats (ANDRO product, *Ollitrault and Colin de Verdière* [2014]). $(v - \overline{v})^2$ where v is the 10-days average meridional velocity of Argo floats displacements at parking depth between 950 and 1050m and $\overline{v}$ the time average 1995-2016 velocity gridded at a resolution 1°x1°.

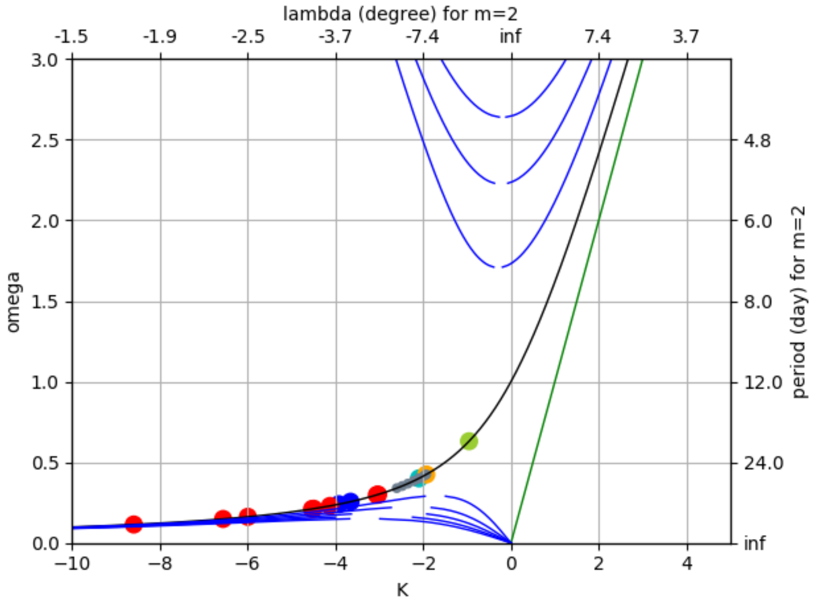


FIGURE 4. Different studies discussed in the paper have reproduced a DEC in idealized simulation, prescribing a primary MRG wave. Characteristic values for the primary wave are reported here in a non-dimensional dispersion diagram. Blue dots are for values found in *Ménesguen et al.* [2009a], with a cyan dot which is for a case where the primary wave is not unstable within their 50°-long basin. Values corresponding to their barotropic forcing are not represented in this diagram because they lie in the [-60 -45] adimensional k range. *d'Orgeville et al.* [2007]'s values are in red and orange when the primary wave is stable within their 50°-long basin. The green dot is a value for the primary wave used in *Ascani et al.* [2010]. Small grey dots stand for an estimation of the spectrum of primary waves forced in *Ascani et al.* [2015]. This figure illustrates how some studies consider primary waves that are closed to short Rossby waves characteristics (blue curves in the low frequency area) and some other depart from these characteristics, approaching a more divergent dynamics.

A: Scaling analysis for equatorial wave dynamics

To evaluate the characteristics of the MRG waves used in different studies, we shall use *Gill* [1974]'s scaling approach, applying it not to the vertical geostrophic vorticity equation (vortical motions) but to the equatorial vorticity and divergence equation (vortical and divergence motions) governing this region.

We consider the vorticity equation

$$\frac{\partial \zeta}{\partial t} + \mathbf{u} \cdot \nabla \zeta + \beta v = -f\delta - \zeta\delta - v_z w_x + u_z w_y \tag{A.1}$$

and the divergence equation

$$\frac{\partial \delta}{\partial t} + \mathbf{u} \cdot \nabla \delta + \delta^2 - 2J(u,v) - f\zeta + \beta u = -\nabla^2 \frac{P}{\rho_0} \tag{A.2}$$

where $\delta = u_x + v_y = -w_z$ is the horizontal divergence.

Combining the x-derivative of the vorticity equation with the x- and t-derivatives of the divergence equation for the equatorial regions and assuming the pressure field in hydrostatic balance leads to

$$\underbrace{\delta_{xtt} + f^2\delta_x + f\beta v_x + \nabla^2 \frac{P_{xt}}{\rho_0}}_{linear} + \underbrace{f(\mathbf{u} \cdot \nabla\zeta)_x + ONLT}_{non-linear} = 0 \tag{A.3}$$

which non dimensionalized gives

$$\frac{\omega^3}{\omega_G^2 \omega_{sR}} \delta_{xtt} + \frac{\omega}{k\beta L_D^2} \delta_x + v_x + \frac{\omega}{\omega_{sR}} \nabla^2 \frac{P_{xt}}{\rho_0} + M^*[(\mathbf{u} \cdot \nabla\zeta)_x + ONLT] = O(M^{*2}) \tag{A.4}$$

where for trapped equatorial waves K=k are zonal wavenumbers, $ONLT$ stands for 'other nonlinear terms' whose form is not important here, and $M^* = \frac{\mathbf{U}^* \cdot \mathbf{K}}{\omega_{sR}} = \frac{U}{c_{sR}}$. The term we keep is the advection of vertical vorticity to be able to harmonize and compare the different authors approaches. Wave number k and frequency ω scale as inverse length and time, U is the zonal velocity magnitude. $\omega_G = \frac{NH}{m\pi} k$ is the Gravity frequency of mode m, N being the Brunt-Väisäla frequency and H the depth of the ocean. $\omega_{sR} = \frac{\beta}{k}$ is the short Rossby wave frequency. For equatorial trapped waves the deformation radius is given by $l_D = \sqrt{\frac{c_G}{\beta}}$.

In terms of wave phase speeds, dividing frequencies by the zonal wave number, the above equation becomes :

$$\underset{\mathrm{I}}{\frac{c^3}{c_G^2 c_{sR}} \delta_{xtt}} \quad + \underset{\mathrm{II}}{\frac{c}{c_G} \delta_x} \quad + \underset{\mathrm{III}}{v_x} \quad + \underset{\mathrm{IV}}{\frac{c}{c_{sR}} \nabla^2 \frac{P_{xt}}{\rho_0}} \quad + \underset{\mathrm{V}}{M^*[(\mathbf{u} \cdot \nabla\zeta)_x + ONLT]} \quad = O(M^{*2}) \tag{A.5}$$

where $c_G = \frac{NH}{m\pi}$ is the Kelvin wave speed.

Gill [1974]'s model is obtained by omitting the I and II linear terms

$$v_x + \frac{c}{c_{sR}} \nabla^2 \frac{P_{xt}}{\rho_0} + M^*[(\mathbf{u} \cdot \nabla\zeta)_x] = 0 \tag{A.6}$$

For mid-latitude dynamics as in *Qiu et al.* [2013b], the deformation radius is given by $\frac{c_G}{f}$ and the term II is scaled as $\frac{c}{\beta L_D^2} = \frac{c}{c_{lR}}$ where c_{lR} is the long Rossby wave phase speed.

To compare the different models, we shall relate the Froude number used by *Hua et al.* [2008] ($Fr_{Hua} = \frac{V}{c_G}$) and the *Ascani et al.* [2015]'s scaling ($M_{Asc} = \frac{U}{c_Y}$) to the more conventional *Gill* [1974]'s $M^* = \frac{U}{c_{sR}}$, scaling the velocity with the short Rossby wave speed $c_{sR} = \frac{\beta}{k^2}$.

The scaling of the nonlinear term is respectively given by

$$M^* = M_{Asc}\frac{c_Y}{c_{sR}} \qquad M^* = Fr_{Hua}\sqrt{\frac{c_G}{c_{sR}}\frac{c_Y}{c_{sR}}} \tag{A.7}$$

Since the meridional velocity is used in the Froude number Fr_{Hua}, for consistency, we have rescaled it to the zonal velocity via the MRG wave (U, V) relation.

The ocean models considered here are forced to excite MRG (also called Yanai) waves $c_Y = c = \frac{\omega}{k}$ while *Qiu et al.* [2013b] forced it with a long Rossby wave $c = \beta L_D^2$ and *Gill* [1974] with a short Rossby wave $c_{sR} = c = \frac{\omega}{k}$.

Acknowledgments

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