Easterly waves over the Eastern Pacific

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Easterly Waves over the Eastern Pacific

by

VICTOR MANUEL TORRES PUENTE

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Submitted to the University at Albany, State University of New York
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Abstract

The research in this thesis explores different processes associated with the life cycle of Easterly Waves over the tropical Eastern Pacific. These include genesis, structural evolution, intensification and a dissipating stage.

First, a review of the mean state was performed considering a PV-θ approach. By using a PV-θ framework, it was found that the distribution of potential vorticity (PV) over the EPAC satisfies the necessary conditions for barotropic instability: the Charney-Stern condition given by $\nabla V \times \nabla Q < 0$, as well as the Fjortoft condition. Together these conditions suggest the potential for barotropic growth of EWs over the EPAC. The sources of positive PV anomalies over the EPAC are associated with stratiform and shallow heating. Also important is the distribution of negative PV anomalies. These are associated with low stability in the lower troposphere that arises in association with dry convection over the western side of the mountain regions of the Sierra Madre.

By using an idealized modeling approach, it was demonstrated that EWs over the EPAC can be triggered by localized forcing (represented by finite amplitude transient heating) in the vicinity of the mid-level jet over the region. In particular, over the Panama bight, heating from a stratiform profile forms an initial trough that develops an EW structure over the following days. The resulting horizontal structure and wave characteristics are in general agreement with results documented in previous studies of EWs in this region. This confirmed that EWs could have an origin in situ breaking the paradigm that they must come from upstream.
The EW storm track over the EPAC was found to be oriented parallel to the coast. Along this, EWs showed a distinctive horizontal tilts against the shear, suggestive of a barotropic growth mechanism, while in the vertical, a tilt at low levels suggested weak baroclinic growth; these characteristics are also agreement with the mean state.

EWs over the EPAC peak in intensity mainly to the west of Mexico over the Western Hemisphere Warm Pool, where they contribute with 20% of the total variance in convective and dynamical metrics. At this region they present a vertical structure closely associated with the transition to tropical cyclones. A rough estimate suggests that this transitions occurs about 13% on average. After reaching a maximum intensification EWs, dissipate.

Within this research, topics associated with the modulation of EWs from convectively coupled equatorial Kelvin Waves and interannual variability of EWs were also explored. Regarding Kelvin Waves, it was found that these can induce EW genesis only about 20% of the time close to the Panama bight. When KWs generate EWs, the convective phase of the MJO is often present.

At longer timescales, changes in EW activity appear to be associated with SST anomalies modulated by low-frequency oscillations. EWs showed a decreased activity during 1980-1997 and increased during the period between 1998-2015. It was found that the EW activity over the EPAC is increased when the Pacific Decadal Oscillation (PDO) is negative and the Atlantic Multidecadal Oscillation (AMO) is positive, as occurred during the period of 1998-2015. The opposite occurred during the years 1980-1997, when a +PDO and a -AMO index occurred. By identifying that SSTs drive the variability of synoptic-scale activity, there exists the potential for predictability of EW activity on long timescales.
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Chapter 1. Introduction

Easterly Waves (EWs) are westward-propagating, synoptic-scale systems found in the tropics, typically observed over Africa and the Atlantic Ocean (Carlson 1969; Reed et al. 1977; Kiladis et al. 2006) and the Pacific Ocean (Serra et al. 2008), originally identified in the 1940’s through the work of Dunn (1940) and Riehl (1945). Since then, analysis of EWs in observations, reanalyses and models has grown into a significant body of literature but focusing particularly on African Easterly Waves (AEWs). Only recently has the interest in the study of EWs over the tropical Americas increased (Serra and Houze 2002; Petersen et al. 2003; Serra et al. 2008; Serra et al. 2010; Rydbeck and Maloney 2014; Rydbeck and Maloney 2015; Rydbeck et al. 2017), often partitioning this region into two areas: the Intra Americas Seas (IAS) and the tropical Eastern Pacific (EPAC). While studies of EWs over the EPAC-IAS region have shown some of their basic characteristics, for example: Structure and energetics (Serra et al. 2008; Serra et al. 2010), intraseasonal variability (Crosbie and Serra 2014; Rydbeck and Maloney 2014), as well as their role on the climate of the tropical Americas (Patricola 2005; Salinas-Prieto 2006), we still lack a complete understanding of the processes that help to explain these basic characteristics, including their presence over a complex region, temporal variability, and genesis. Although the work here covers both regions, the EPAC as well as the IAS region, the overarching goal of this work is to provide a thorough analysis of EWs in the EPAC region. This includes a detailed review of the mean state that supports them; all this through the use of reanalysis and observations as well as idealized modeling. The following subsections will review in further detail the previous research on EWs over the EPAC-IAS region on each of these topics covered in this work.
1.1. Mean state over the EPAC-IAS region

The EPAC-IAS region is characterized by significant variations in topography, Sea Surface Temperatures (SSTs), and regional circulations (see Fig. 1.1). Mountains reach elevations of over 3 km, SSTs within the region are characterized by a pronounced warm pool (Wang 2002), and the atmosphere is characterized by distinctive regional circulations, including the North Atlantic Subtropical High (NASH), as well as local features, such as the Caribbean Low Level Jet (CLLJ - Amador 1998). These interactions between land, ocean and atmosphere set up a complicated mean state in which EWs form, propagate and evolve. In addition to this, it is important to recognize the conditions for dynamic instability within the region. According to Molinari et al. (1997), the sign reversal in the meridional potential vorticity gradient near 700 hPa over the Caribbean, permits either invigoration of EWs or their actual generation. These points raise questions concerning whether the PV sign reversal is enough to invigorate upstream EWs (given the length of the waves compared to the unstable regions), or even if EWs over the EPAC-IAS region can be generated in situ from the given dynamical conditions.

In addition, convection over the EPAC-IAS region plays a fundamental role in the mean state and also in the maintenance and growth of EWs through three main pathways: supporting EWs through establishing mean potential vorticity (PV) gradients (Ferreira and Schubert 1997; Davis et al. 2008), through the direct impact on the EWs themselves (Molinari et al. 1997; Petersen et al. 2003; Serra et al. 2008), and as a potential trigger of EWs (Thorncroft et al. 2008; Rydbeck et al. 2017). These results emphasize the relevance that convection has on the EPAC as well as on EWs. However, the distribution of diabatic heating from convection and its further impact on the creation of PV has not been explored in the same detail over this region as over
the west African region (Janiga and Thorncroft 2013). These points highlight the necessity to clarify the role that convection has on both the mean state and on EWs.

Basic motivating questions with respect to the mean state are: Is the mean state supportive of EWs? What is the role of convection in the establishment of the supportive conditions for EWs? Do the mean conditions allow for the genesis of EWs? Major hypotheses associated with these questions are:

- The presence of jets over the EPAC-IAS region is fundamental for the track and growth of EWs over the region.
- The meridional gradients of PV found over the EPAC-IAS region support growth of EWs through the presence of PV sign reversals.
- Sources of PV are associated with diabatic heating from different convective populations found over the EPAC.

1.2. Variability of EWs

Most of the studies on variability concerned with the EPAC-IAS region have been concerned with the large-scale environment and regional circulations. Examples of these are Karnauskas and Busalacchi (2009) on SSTs, and Giannini et al. (2000) on CLLJ and precipitation. Through its teleconnections, It has also been recognized that El Niño-Southern Oscillation (ENSO) has an important influence on TC activity over the Atlantic (Xie et al. 2005; Kossin et al. 2010), the Caribbean (Amador et al. 2010), and the EPAC (Wang and Fiedler 2006; Wang and Lee 2009; Jin et al. 2014). However, little attention has been given to the variability of weather systems that modulate and produce rainfall on the synoptic timescales, including the EWs. Shelton (2011) addressed the issue regarding EWs over the Caribbean with respect to ENSO, showing that under
El Niño, EW tracks decrease in number. This result can be interpreted as fewer disturbances moving from Africa or a less number of them generated within the region. With only one study covering this topic, it still not clear how ENSO affects EWs over the EPAC.

Even less is known about interdecadal variability of EWs over the EPAC-IAS region. As background, Goldenberg et al. (2001) showed how the Atlantic Multidecadal Oscillation (AMO) modulates the activity of TCs during its positive phase. Shelton (2011) considered the influence of AMO on EWs over the Caribbean, finding that AMO phases do not affect the number of EWs over the IAS region. Instead Shelton (2011) found that these differ only in latitude in association with the location of the NASH. With only a few studies regarding the role of AMO on EWs, it is unclear how this oscillation impacts EWs over the wider EPAC-IAS region. Furthermore, it is not clear if there is a role for other interdecadal oscillations, such as the Pacific Decadal Oscillation (PDO). Motivated by this, the present work also explores the variability of EWs on interannual and interdecadal timescales.

The previous paragraphs raise questions associated with the variability of EWs on interannual and interdecadal timescales such as: How does EW activity vary on a year to year basis? Is there a role of interannual and interdecadal oscillations on the activity of EWs over the EPAC-IAS region? What are the main drivers of EW activity on interannual to interdecadal timescales? For these questions, major hypotheses are:

- EW activity over the EPAC-IAS region is modulated on Interannual and interdecadal timescales
The main driver of interannual and interdecadal variability is SST temperatures over the western hemisphere warm pool, which in turn modulates convection and the meridional gradients of PV.

This enhanced convection over the EPAC-IAS region could also invigorate EWs or even work as trigger of EWs.

### 1.3. Easterly Wave Characteristics

Previous studies on tracking of EWs have provided some insight about their presence over the EPAC-IAS region. Methodologies based on objective tracking of EWs, using relative vorticity, have been conducted for the Atlantic (Thorncroft and Hodges 2001) and for the IAS region (Serra et al. 2010). These studies, in spite of not having their focus on the same region, showed that EW density concentrates mainly over Central America and into the EPAC, where a maximum in wave density is found (see Fig. 1.2.a and 1.2.c). Such results are consistent with those of OLR filtering in the TD band over the EPAC, as Serra et al. (2010) also showed, meaning that a coherent EW track is found west of Central/South America.

Using manual analysis of relative vorticity at mid-levels (700-600 hPa), Kerns et al. (2008) found two wave tracks over the Atlantic Ocean (see Fig. 1.2.b): a northerly one near 20°N, and a southerly one near 10°N. This could be the result of a limited number of years used in their study, not allowing for a general conclusion on the kinematics of EWs over the EPAC-IAS region.

Shelton (2011) used the method of Berry et al. (2007) to track EWs using curvature vorticity and a trough-line algorithm (see Fig. 1.2.d). In comparison with previous results (e.g., compare with Fig. 1.2.c), Shelton (2011) concluded that over the Caribbean, EWs track more along the northern edges of the basin around 20°N, in agreement with Kerns et al (2008).
However, Shelton (2011) also found that waves at the northeastern Caribbean shift towards the central Caribbean and continue westwards into the tropical EPAC (highlighted with pink in Fig. 1.2.d), a feature that had not been observed previously. This could be the result of the merging or elongation of EWs over the Caribbean and the EPAC.

A recent method of Brammer and Thorncroft (2015) showed that over the EPAC-IAS region, EW density is located mostly at about 10°N (see Fig. 1.2.e), over the Caribbean, Central America and the EPAC, but with very low values over the rest of the IAS domain. This may imply that EWs decay as long as they enter the IAS. However, it must be considered that, in order to be tracked, these waves had to have existed over West Africa and therefore a significant number of EWs might be missing over the EPAC-IAS region. This motivates further analysis centered over the EPAC-IAS region.

From all these previous results there is some consensus about the propagation of EWs over the EPAC-IAS region. They all agree on a track over the Atlantic and over the EPAC at about 10°N, but with disagreement on the tracks (where a southern track is found over land in South America, and another over the Caribbean at about 20°N). Therefore, these arguments suggest that results depend on the levels chosen, the algorithm used, the reanalysis, and the number of years, as suggested by Kerns et al. (2008). This complexity has not allowed for a clear idea of EW tracks over the EPAC-IAS region. Hence, it is necessary to make further research to clarify the presence of EWs over the EPAC-IAS region.

Although many of the basic characteristics of EWs have been identified, not much has been explored regarding their life cycle: genesis, development and decay and where this occurs. There are still gaps in our knowledge regarding their origin, vertical evolution, coupling with
convection and intensification processes, as well as their fate. To improve the understanding of EWs over the EPAC, it is necessary to revisit the basic characteristics of EWs, as well as how they evolve in space and time.

Previous paragraphs motivate questions such as: What is the predominant path followed by EWs over the EPAC? To what extent do the EPAC waves originate over the Atlantic? How do EW structures evolve over the EPAC-IAS region? How and where do they intensify? Some of the hypotheses of this topic are:

- EWs track parallel to the coast over the EPAC based on the mean state and PV meridional gradients found over the EPAC.
- EWs over the EPAC can have a link with upstream EWs over the Atlantic.
- The structure and basic characteristics of EWs over the EPAC changes upon the region where they are studied and the conditions where they are embedded.
- EWs intensify over the warm waters found in the EPAC mainly through enhanced convection there.

1.4. **Relationship between EWs and convectively coupled Kelvin waves**

The interaction and influence of convectively coupled equatorial waves on EWs over the EPAC-IAS region has not been addressed previously. It is hypothesized that these could modulate the activity of EWs by influencing the mean state, the convective activity in the EWs, and even act as upstream convective triggers for the genesis of EWs in this region. For example, previous research on AEWs has shown that, during the convective phase of strong convectively coupled Kelvin waves (CCKWs), they can modulate and trigger AEW activity (Ventric and Thorncroft 2013).
Over the Eastern Pacific, previous research has shown the presence of CCKWs over the EPAC (Straub and Kiladis 2002; Roundy and Frank 2004). In particular, Schreck (2015) suggested that CCKWs can significantly amplify EWs leading to tropical cyclogenesis by enhancing cyclonic vorticity at low levels, and upper level outflow through enhanced convection. Therefore, the convection associated with the passage of CCKWs over the EPAC-IAS region could impact the mean state over the EPAC, the development of EWs and even their genesis. Question associated with the presence of CCKWs over the EPAC-IAS region and EWs are: What is the role of CCKWs on convection in the EPAC and how does this influence EWs? How do CCKWs impact the mean state over the EPAC-IAS region? Can they trigger EWs over the EPAC as over Africa? Hypotheses associated with these questions are:

- CCKWs impact directly individual EWs through enhanced convection.
- CCKWs establish PV-meridional gradients during their passage - enhancing instability in the mean state.
- CCKWs are a genesis mechanism, triggering EWs from enhanced convection.

1.5. Origins of EWs

Some mechanisms for EW genesis over the EPAC-IAS region have been proposed previously. These include mainly barotropic instability of the mean flow (Mozer and Zehnder 1996; Molinari et al. 1997; Molinari and Vollaro 2000; Maloney and Hartmann 2000a), and inter-tropical convergence zone (ITCZ) breakdown (Ferreira and Schubert 1997). Overall these mechanisms deal with a barotropic instability mechanism, which extracts kinetic energy from a horizontal sheared flow. Rydbeck et al. (2017) recently suggested another mechanism for the in situ generation of EWs over the EPAC based on the triggering mechanism of Thorncroft et al.
This consists primarily of the forcing of EWs by local convective disturbances resulting from the high terrain over the northern mountain ranges of South America, next to the Panama bight (see Fig. 1.1.c). Thorncroft et al. (2008) showed that EWs over Africa could be triggered by localized forcing in association with latent heat at the entrance of a stable African Easterly Jet. Their results showed structures developed that were similar to observed EWs over Africa, thus providing a viable mechanism for AEW genesis. Motivated by this result, this research explores the triggering hypothesis for EWs over the EPAC, given the convective and dynamical conditions present over the region.

The previous paragraphs pose fundamental questions about the origin of EWs over the EPAC-IAS region, such as: What is the origin of EWs over the EPAC-IAS region? Can EWs be generated in situ? Major hypotheses motivated by these questions are:

- EWs over the EPAC can be generated in situ triggered by convection.
- The genesis of EWs over the EPAC is sensitive to location of the trigger.
- The genesis of EWs over the EPAC is also sensitive to the mean state.

1.6. Overarching scientific questions to be addressed

Each of the topics introduced above will be reviewed in more detail in each chapter. These span from a review of the mean state and the fundamental features of EWs to genesis and further impacts – all this to achieve the overarching goal which is: to improve our knowledge and understanding of the nature of EWs over the EPAC-IAS region. To address this goal and each of these topics, the current dissertation will focus on answering the following overarching questions:
1. What are the characteristics of the mean state over the EPAC-IAS region and how does it support EWs over the EPAC-IAS region?

2. What are the basic characteristics, track and structure of EWs over this region and how does this relate to convection characteristics and mean state?

3. What is the origin of EWs over the IAS-EPAC region: to what extent do EWs form in situ or originate from upstream and what mechanisms are playing a role in this?

4. How do EWs over the EPAC-IAS region vary on interannual and interdecadal timescales?

It is expected that these results will have an impact on our knowledge of the fundamental features of EWs over the EPAC-IAS region from a scientific standpoint, as well as from an operational point of view.

1.7. Outline

This thesis is structured in 7 chapters and it is organized as follows: Chapter 2 presents an overview of the key features of the region through an analysis of the mean state, fundamental for understanding the complexity of the EPAC-IAS region. Then, Chapter 3 will present a thorough analysis of the statistics of EWs as well as their variability, followed by a chapter devoted to its structure and propagation in Chapter 4. The relation between EWs and CCKWs is discussed in Chapter 5, while to explore its genesis, an idealized modeling approach is presented in Chapter 6. This is followed by a summary of results and outline of recommendations for further research in Chapter 7.
Figures

**Figure 1.1.** Main features over the EPAC-IAS region. 

- **a.** Topographic height (in m).
- **b.** Daily climatology of SST (in °C) with the Western Hemisphere Warm Pool enclosed in purple.
- **c.** Outgoing longwave radiation (in Wm\(^{-2}\)), and
- **d.** Daily climatology of wind field and its magnitude (shading, in ms\(^{-1}\)) highlighting the Caribbean Low-level Jet.

Streamfunction (contours in 10\(^6\) m\(^2\)s\(^{-1}\)) also at 950 hPa. For all these plots, the mean climatology is calculated during JJAS for 1980-2015. Data sources: NOAA-ETOPO1 for a; NOAA-Interpolated SST for b; NOAA Interpolated OLR for c; and ERAI for d.
Figure 1.2. Track densities over the EPAC-IAS region from different methodologies. 

Chapter 2. Analysis of the environment over the Eastern Pacific and the Intra-Americas Sea

2.1. Introduction

The Eastern Pacific (EPAC) and the Intra-Americas Sea (IAS) region are characterized by significant variations in topography, Sea Surface Temperatures (SSTs), and regional circulations. In this region, mountains reach elevations over 3 km in height, and their distribution can channel atmospheric flows in some specific locations (Steenburgh et al. 1998). SSTs within the region are characterized by a pronounced warm pool (Wang 2002), which influences atmospheric phenomena that span from a few days (e.g. in tropical cyclones; Xie and Philander 1994; Maloney and Kiehl 2002; Xie et al. 2005) to several years (Philander 1983; Wang and Enfield 2001; Kao and Yu 2009). Regarding the atmosphere, the region is characterized by distinctive regional circulations, including the North Atlantic Subtropical High (NASH), as well as local features, such as the Caribbean Low Level Jet (CLLJ - Amador 1998). Thus, the interactions among land, ocean and atmosphere set up a complicated mean state in this region in which Easterly Waves (EWs) can form, propagate and evolve. The aim of this chapter is to provide a thorough review of the land, ocean and atmospheric features present in the EPAC-IAS region to highlight the potential impact on EWs.

Given that the focus of the present study is on EWs over the EPAC, it is also necessary to explore the conditions for their growth, considering especially the conditions that impact barotropic and baroclinic instability. Dynamically, a necessary condition for the unstable growth of EWs is related to the Charney-Stern criterion for instability of internal jets (Charney and Stern 1962). This requires that the meridional gradient of potential vorticity (PV) has opposite signs in
the interior of the fluid. In addition, the Fjortoft condition (Fjortoft 1950) is a “necessary condition for instability (that) requires that the mean zonal current be positively correlated with PV_y” as Molinari et al. (1997) suggested. To explore these conditions, this chapter presents an analysis of the diagnosed instability associated with the distribution of PV and its horizontal gradients over the region. Furthermore, since previous studies suggest a baroclinic contribution to the growth of EWs (Serra et al. 2008; Serra et al. 2010; Rydbeck and Maloney 2014) the role of baroclinic instability is also explored by considering regional low- and upper-level temperature gradients.

Convection over the EPAC-IAS region plays a fundamental role in the mean state and also in the maintenance of EWs through three main pathways. First: mean convection may help to establish an environment that supports and influences EWs through impacting the PV gradients (Ferreira and Schubert 1997; Davis et al. 2008). Second: convection within the EWs may provide a mechanism for their growth (Molinari et al. 1997; Petersen et al. 2003; Serra et al. 2008), and third: convection may trigger EWs (Thorncroft et al. 2008; Rydbeck et al. 2017). Recognizing that the EPAC-IAS region is characterized by the presence of the Intertropical Convergence Zone (ITCZ), as well as one with the most convective regions of the world (Zipser et al. 2006) over the Panama bight - defined as the region between the Gulf of Panama and Point Santa Elena, Ecuador (Nichols and Murphy 1944) - it is necessary to clarify the role that convection has on the mean state and on EWs over this region. Given this importance, the next paragraphs expand further on the convection characteristics over the region.

Multiple types of convection are found over the EPAC-IAS region. Zuluaga and Houze (2015) studied its distribution over tropical regions considering three types of convection: deep convective cores, wide convective cores, and broad stratiform regions. Figure 2.1.a-c shows their
results illustrating these distributions. Over the EPAC-IAS, deep and wide convective cores (Fig. 2.1.a-b) are found mainly over continental regions, particularly over the west side of Sierra Madre in Mexico and the Panama bight. Broad stratiform regions (Fig. 2.1.c) are found mostly over the ocean within the ITCZ as well as over the Panama bight. Based on the work of Schumacher et al. (2004), Zuluaga and Houze (2015) stated that broad stratiform convection has the highest probability of occurrence over the EPAC because of the unlimited moisture from a warm and humid oceanic boundary layer, thus suggesting this as the dominant convective type over the region. This pattern is also observed over the eastern Atlantic downstream of West Africa, showing similar characteristics.

Huaman and Schumacher (2018) explored the latent heat (LH) vertical profiles over the EPAC ITCZ during JJA using satellite observations from the Tropical Rainfall Measuring Mission (TRMM) Precipitation Radar (PR) and CloudSat, as well as reanalyses (MERRA2, NCEP-NCAR and ERA-Interim, hereafter ERAI). Some of their results are illustrated in Fig. 2.1.d-f for reference. The idealized heating profiles used are characterized by different peaks in the vertical: the deep convective profile by a heating peak at 4 km (600 hPa), the stratiform profile with heating peaks at 8 km (350 hPa), and cooling at 4.5 km (570 hPa), while the shallow convective maximum is centered at 2.5 km, above 750 hPa, these slightly different from Schumacher et al. (2004) based on recent studies associated with peak heights and broad stratiform populations over the EPAC (Houze Jr. et al. 2015; Ahmed et al. 2016). From these profiles, the calculated LH in Fig. 2.1.e shows two strong peaks of LH with maxima at 400 and 700 hPa (~4 K day\(^{-1}\)) with a minimum slightly above 600 hPa (~3 K day\(^{-1}\)). With a LH minimum resulting from the cooling associated with evaporation of stratiform precipitation, this result suggests the dominance of stratiform heating.
over the EPAC at upper levels, in agreement with (Zuluaga and Houze 2015), but also identify shallow convection at lower levels.

Considering that over rainy regions LH is the dominant component in the total diabatic heating $Q_1$ (Houze 1982), Huaman and Schumacher (2018) also calculated $Q_1$ from different reanalyses and compared them with their observed LH. This comparison is illustrated in Fig. 2.1.e-f. Overall, the intensity of $Q_1$ is weaker in all models when compared with observations. This is likely associated with convective parameterizations and their impact on the vertical velocity (Huaman and Schumacher 2018), as well as with model resolution and lack of observations over the EPAC, as several authors have found (Gelaro et al. 2010; Serra et al. 2010; Janiga and Thornicroft 2013; Huaman and Schumacher 2018). However, while MERRA2 (in black) is biased towards upper levels and NCEP (in green) shows a uniform distribution, ERAI (in red) is able to capture two weak maxima slightly above 600 and 850 hPa suggesting that these may be associated with shallow and stratiform heating captured by the model, and providing more confidence in the use of ERAI.

The previous results on convection emphasize the relevance that it has on the EPAC as well as on the distribution of $Q_1$. However, the impact that convection has on the generation of PV, has not been explored in detail in this region. This chapter will also explore the role of convection over the EPAC in the establishment of a basic state impacting the life cycle of EWs.

To explore the previous features associated with EWs over the EPAC-IAS region, this chapter is structured as follows: the data and methodology for this review, as well as the methods employed to study the generation of PV from diabatic heating, are presented in section 2.2. Next, a review of the mean state is presented in section 2.3, followed by an assessment of the instability
mechanisms in section 2.4. Later, the PV generation from diabatic heating is presented in section 2.5, and finally, a summary of results is provided in section 2.6.

2.2. Data and Methodology

Topographic features are identified by using data from NOAA-ETOPO1 (Amante and Eakins 2009), with a horizontal resolution of 1 arc-minute. To explore SSTs, data are retrieved from NOAA-SST interpolated version 2 (Reynolds et al. 2002) on 1° horizontal grid. For convection, outgoing longwave radiation (OLR) from the NOAA-OLR dataset are used (Lee 2014), also on a 1° horizontal grid. To retrieve rainfall, data from the Tropical Rainfall Measurement Mission (TRMM; Huffman et al. 2007) are used from their 3B42 dataset, which uses a combination of precipitation estimates (Adler et al. 1994). This dataset has also been used in previous studies over the region (Kucieńska et al. 2012) showing consistent results over ocean when compared with ground-based rainfall estimates (Wang et al. 2014). Dynamical variables - including PV - as well as derived variables (e.g. \( \theta_e \)), are retrieved from the ERAI dataset (Dee et al. 2011) at 0.5° horizontal resolution and with 28 vertical levels. This spatial grid is appropriate to capture the main features of the mean state. ERAI was selected motivated by the findings of Serra et al. (2010), regarding an improved representation of the tropics over other datasets. For the analysis of stability over the region, we used soundings from the University of Wyoming archived data (University of Wyoming 2020).

To calculate the apparent heat source \( (Q1) \), as well as the PV tendency \( (PVT) \), the approach of Janiga and Thorncroft (2013) is followed. This consists of calculating the diabatic heating as a residual in the thermodynamic equation:

\[
Q1 = \frac{T}{\theta} \left( \frac{\partial \theta}{\partial t} + \mathbf{v} \cdot \nabla_p \theta \right) \tag{2.1}
\]
and then using this to retrieve the PV tendency:

\[ \frac{dp}{dt} \approx -g \nabla_p \cdot [Q1(\Omega + \nabla_p \times \vec{v})] \]  (2.2)

In Eq. (2.1), \( T \) is the temperature, \( \vec{v} = (u, v, w) \) is the vector wind, and \( \nabla_p \) is the gradient operator. The potential temperature is defined as \( \theta = T\left(\frac{p_0}{p}\right)^{R/c_p} \), with \( p_0=1000 \text{ hPa} \) as the reference pressure, \( p \) the pressure, \( R \) the dry air specific gas constant, and \( c_p \) the specific heat of dry air at constant pressure. Here, \( \frac{\partial \theta}{\partial t} \) is the time rate of change of \( \theta \) between analysis times. In Eq. (2.2), \( Q1 \) is retrieved from Eq. (2.1) to represent the diabatic heating, \( g \) is the gravitational acceleration, and \( \Omega = (0, 0, f) \) is the planetary vorticity vector. Equation (2.2) ignores friction, which according, to Krishnamurti et al. (2000), is a valid approximation in the tropics. This is also useful over the EPAC, in regions away from topography, where convection and LH are dominant, as it is for our case.

2.3. Mean state over the Eastern Pacific and the Intra-Americas Sea

The analysis of the mean state over the EPAC-IAS region is made for the rainy season, covering the months of June, July, August and September (JJAS). This will be defined as the extended summer season in this work. As a climatological reference, the years 1980-2015 are considered. Next we provide an overview of the geographic and thermodynamic features, followed by the wind structure.

a. Geography

Figure 2.2.a shows the tropical EPAC and the IAS regions. The tropical EPAC is the oceanic region extending from the Baja California peninsula to the northern coast of Peru, while the IAS region consists of the Caribbean Sea and its islands, the coasts of South America, Central America, and the Gulf of Mexico (Amador 2008). Together, these regions span the deep tropics and the
subtropics. While the EPAC is characterized mostly by a large oceanic region, the IAS region is characterized by complex and diverse topography. This topographic variability is present in its the islands (e.g. Cuba and Hispaniola), as well as over the continental region, from southeastern Mexico to northern South America. In particular, Colón peak (located close to 11°N, 73°W) and the mountain regions north of South America, may have a considerable impact on the dynamical environment by generating potential vorticity through lateral friction (Rodwell and Hoskins 2001). Additionally, these topographic features interact with the large-scale atmospheric circulation, by blocking the flow from upstream (Zehnder 1993; Farfán and Zehnder 1997), as well as shaping the climate of the region (Rasmussen et al. 1989; Smith et al. 2009). Thus, this topographic configuration will have an influence in the thermodynamical environment, as well as the wind structure over the region. These elements are reviewed next.

b. Thermodynamic environment

During the boreal extended summer, SSTs in the EPAC and the IAS region (Fig. 2.2.b) are characterized by values that exceed 26.5°C. High SSTs characterize the northwestern Caribbean and the Gulf of Mexico, while in the EPAC, these are observed west of Mexico and west of Central America. Because of these warm SSTs, this oceanic region has become known as the Western Hemisphere Warm Pool (WHWP), defined as the region of SSTs warmer than 28.5°C, (enclosed by the purple line in Fig. 2.2.b). This corresponds to the second largest warm water body after the Indo-Pacific Warm Pool (Wang and Enfield 2001; Wang 2002; Wang and Enfield 2003). The SSTs in the EPAC-IAS region exceed the suggested thresholds necessary for convection (Zhang 1993; Webster 1994) and tropical cyclogenesis (Palmen 1948). Therefore, SSTs in the EPAC-IAS region are a key thermodynamic element to fuel convection in atmospheric disturbances. There are also notable cooler SSTs over the northern coast of South America. These cooler SSTs impact
the Caribbean Sea by setting up a positive meridional gradient over the Caribbean. Gordon (1967) stated that the source of these cooler SSTs is coastal upwelling in response to the surface wind stress curl present over the Caribbean.

Continuing with thermodynamic variables, the specific humidity at 950 hPa, is shown in Fig. 2.2.c. The maxima in moisture are observed over the western Caribbean and the tropical EPAC, with a local maximum over Central America and the Yucatan peninsula. Notably, the moistest air at this level is not above the warmer SSTs, but mostly over regions where the atmospheric circulations at this level converge (discussed later in the next section). Equivalent potential temperature $\theta_e$ at the same level in Fig. 2.2.d, follows the distribution of humidity. It shows the highest values over Central America indicating an unstable region over the Panama bight.

Regarding convection itself, Fig. 2.2.e shows the distribution of OLR. The first evident feature is the convection associated with the ITCZ over the EPAC, located near 10°N. Convection is also found over continental regions following the topographic distribution, this ranging from northern South America to northwestern Mexico. In particular, the latter is associated with the monsoon of North America (Adams and Comrie 1997; Higgins et al. 1999). Additionally, minima in OLR are found over Central America, in a region of high $\theta_e$ over land, west of the Andes, and also over the ocean in the Panama bight. This region is well known to be associated with deep convection (Zipser et al. 2006), including westward propagating mesoscale convective systems (MCSs) initiated over the land (see Mapes et al. 2003a). In contrast, convection is weak over the central and eastern Caribbean, coincident with cooler SSTs (c.f. Fig. 2.2.b).
Lastly, rainfall is presented in Fig. 2.2.f. The main features of this distribution are clearly seen over the EPAC, again related to the ITCZ, and over the western coast of Mexico. Over the Panama bight, northwest of South America, rainfall coincides with the coldest peak of OLR and partially with elevated $\theta_e$. Over the Caribbean, most of the rainfall is observed over the islands, while the southern Caribbean remains very dry (particularly near northern South America), coincident with the upwelling oceanic region (Fig. 2.2.b).

A detailed consideration of the distribution of $\theta_e$ (Fig. 2.2.d) and rainfall (Fig. 2.2.f) reveals that the relationship between these variables is not necessarily one-to-one. For example, over the Gulf of Mexico, despite finding higher values of $\theta_e$, there are very low rainfall totals. This pattern repeats over Central America at 10°N around 80°W-90°W, where regions of high $\theta_e$ show minima in rainfall, coincident with low SSTs (Fig. 2.2.b) and higher OLR (Fig. 2.2.g). These relationships could be explained by regional overturning circulations within dry environments (Saha et al. 2014). However, to find this, it is necessary to explore the regional and local circulations over the region, described next.

c. Wind structure: Regional and local circulations

The EPAC-IAS region is influenced by both large-scale regional and local circulations. At regional scales, the NASH (in Fig. 2.3) is present mainly over the Caribbean region. This high pressure pattern is characterized by an intensification and a westward extension during July (Hastenrath 1978, 1984; Mapes et al. 2005), which results in local strengthened zonal easterly winds (Curtis and Gamble 2008; Gamble and Curtis 2008; Cook and Vizy 2010). Some studies suggest that these enhanced winds over the Caribbean favor the growth of EWs over the EPAC and the IAS region (Serra et al. 2008; Serra et al. 2010).
Figure 2.4 presents the synoptic-scale atmospheric circulations over the EPAC-IAS, showing the regional winds at different pressure levels. From upper to lower levels, at 200 hPa, Fig. 2.4.a shows an upper-level ridge centered just West of Mexico associated with the tropical easterly jet over the EPAC around 5°N. Over the Atlantic and Caribbean, an upper-level trough encroaches from the east, resulting in northerly flow over the Caribbean. The rest of the IAS region is characterized by relatively weak winds at this level, particularly over the western Caribbean. Around mid-levels, at 600 hPa (Fig. 2.4.b), easterly flow of about 10 ms\(^{-1}\) is observed over the equator, while around 15°N, the wind maximum splits over the southwestern Caribbean and crosses the continent, showing a quasi-continuous wind track. The easterlies over the Caribbean are suggestive of a vertical extension of the CLLJ, which increase in magnitude as pressure increases, as seen at 850 hPa (Fig. 2.4.c) shows. The wind maximum at this lower level splits over the Caribbean, with one branch to the northwest passing over the Yucatan peninsula, and the other traversing through Central America. Figure 2.4.d shows that at the level of 950 hPa, the magnitude of this easterly flow is greater than 10 ms\(^{-1}\), mainly over the Caribbean. This low-level easterly wind maximum is known as the Caribbean Low-Level Jet (Amador 1998; Cook and Vizy 2010).

The Caribbean Low-Level Jet (CLLJ) is a local circulation over the IAS region present throughout the year, and it has been proposed that it can supports EW growth (Molinari et al. 1997; Salinas-Prieto 2006; Serra et al. 2010). It is localized over the Caribbean Sea at about 12.5°N (see Fig. 2.4.d), extending into the EPAC through the Papagayo gap in Central America (Romero-Centeno et al. 2007). During July, the CLLJ reaches a maximum speed close to 14 ms\(^{-1}\) at 950 hPa (Amador 1998), and with a longitudinal extension of about 1000 km. Figure 2.5 presents an
averaged vertical cross-section (5°N-25°N in latitude, and 65°W-80°W in longitude, as illustrated on Fig. 2.4.d) showing that its vertical extension (defined by the isolach of 8 ms⁻¹) reaches around 600 hPa. It is not clear whether the extension of the CLLJ can invigorate upstream waves with wavelengths of ~2500 km (e.g. as AEWs).

In summary, we have briefly introduced an overview of the topographic features, thermodynamical characteristics, as well as the regional circulations over the EPAC-IAS. These are all expected to have an influence on the propagation and growth of EWs over the EPAC. However, these are insufficient in explaining the dynamic conditions associated with their life cycle. For this reason, to identify the potential dynamical mechanisms associated with EW growth, the regions where conditions for instability are satisfied, and how they are going to impact on EWs, a detailed assessment of the dynamical instability within the region is presented next.

### 2.4. Assessment of dynamical Instability

This section assesses the extent to which the environment in the EPAC-IAS region satisfies the necessary conditions for barotropic and baroclinic instability. This analysis is based on consideration of Potential Vorticity (PV), and potential temperature $\theta$ at upper levels at the tropopause and at a lower level close to the ground. The PV-$\theta$ framework is useful for diagnosing fluid instability from a dry dynamics perspective (Drazin and Reid 2004; Hoskins et al. 1985), and may provide insights into possible EW-growth mechanisms.

#### a. Barotropic considerations

The analysis of the three dimensional distribution of potential vorticity is helpful to identify meridional PV gradients that can potentially support EWs. In the horizontal, PV is
explored on isobaric levels, focusing on the mid- to lower troposphere (600, 700, and 850 hPa) given the location of the mid- and low-level jets. Although PV is a conserved quantity on isentropic levels, pressure levels were selected to be consistent with the thermodynamic and wind analysis previously shown. Results were not significantly different from those on isentropic surfaces (not shown) given that over much of the region both coordinate systems are approximately horizontal (which suggests weak baroclinicity). This is followed by an analysis in the vertical, to identify the vertical extent of PV maxima.

i. **Horizontal analysis**

The PV distribution at 600 hPa (Fig. 2.6.a) is characterized by high PV in mid-latitudes and low values in the tropics, consistent with the increase of planetary vorticity with latitude and large-scale shear variations. Over the EPAC around 13°N, a weak PV maximum is observed with a marked positive PV gradient on the equatorward side and a weaker negative gradient on the poleward side. The maximum is formed along the equatorward part of the jet at 600 hPa (c.f. Fig. 2.4.b), and contrasts with the region of weaker PV values poleward of this and along the western side of Mexico, a region of significant topography. This distribution of PV sets up a negative meridional PV gradient from 85°W to 105°W roughly parallel to the coastline. Consequently, the PV distribution over the EPAC defines a region that satisfies the necessary conditions for barotropic instability that, should it be realized, would result in a southeast-to-northwest track for EWs.

At 700 hPa (Fig. 2.6.b), we observe a pattern similar to that seen at 600 hPa, albeit with a weaker PV maximum. The PV maxima at this level are defines a continuous distribution between the Caribbean and the EPAC around 12°N, arising from northern South America and extending
into the EPAC through Central America. This distribution of PV also defines a weak negative meridional PV gradient, close to coastal areas over Mexico and Central America as well as over the Caribbean, revealing a potential track for easterly waves to propagate and grow along.

The distribution of PV at 850 hPa in Fig. 2.6.c, shows richer structure. The PV belt at 10°N shows more connection with the CLLJ (see Fig. 2.4.d). Local maxima are present downstream of the Colón peak and west of Central America (see Fig. 2.2.a), making more evident the meridional PV gradients from the Caribbean and the EPAC.

A particularly interesting feature in the distribution of PV, highlighted above, is the minima observed at 600-700 hPa seen mainly over the mountain regions in the Sierra Madre along the western part of Mexico. We hypothesize that surface heating over this elevated region can result in a reduction of PV through dry convection, analogous to that seen north of the West African Easterly Jet in the Sahara.

To explore the role of surface heating in the distribution of PV, we compared the mean distribution of PV during JJAS (wet season) against that when the surface heating peaks over the region, this is, on May (dry season). This also includes exploring the vertical structure on both regimes using reanalysis and observations. However, given the limitations in data (such as missing soundings), observations were constrained to year 2013, where continuous data were available.

Figure 2.7 shows the PV distribution in May (left column) against the mean JJAS (right column) as obtained from ERAI. During May (Fig. 2.7.a), the PV distribution at 600 hPa is characterized by low PV values centered around 20°N, 100°W. A red circle in this map shows the vertical profiles at Guadalajara station (20.67°N, 103.38°W), where data was available. The mean
vertical structure in Fig. 2.7.b, shows a well-mixed boundary layer characterized by a dry adiabatic environment, vertically reaching 550 hPa. From the definition of PV, we see that the static stability will be close to zero, thus leading to a minimum in PV, as observed in Fig. 2.7.a.

When comparing with the JJAS season (right column in Fig. 2.7), the minima in PV is less pronounced (see Fig. 2.7.c) than that observed in May. The mean sounding in Fig. 2.7.d shows a moister environment with a conditionally unstable profile below 600 hPa and a lifting condensation level close to 750 hPa. It is expected that a minimum in PV would be observed at 750 hPa. (This is indeed observed later in the vertical structure in Fig. 2.9.d).

To compare previous results with observations, Fig. 2.8 shows the vertical profiles during year 2013. For reference, ERAI horizontal maps are also displayed. During May 2013, the horizontal distribution of PV is again characterized by a pronounced minimum over the Mexican Plateau and over the western Sierra Madre (Fig. 2.8a). The mean thermodynamic profile in 2.8.b (in purple) shows a well-mixed adiabatic atmosphere, reaching the 550 hPa. This temperature profile is similar to the long-term mean (Fig.2.7.b), although it is characterized by a drier atmosphere in all the vertical column. To further compare with reanalysis, the ERAI temperature profiles are shown in blue. For May, ERAI resolve adequately the temperature profile, extending adiabatically until 600 hPa, slightly below from observations. However, the dew point temperature is higher, implying a moister lower troposphere over the region. Irrespective of this, we observe a deep well-mixed boundary layer resulting in a PV distribution close to zero.

For the JJAS season during 2013, the right side of Fig. 2.8 shows the horizontal PV distribution and its thermodynamic profile again at the GDL station previously indicated. Consistent with the mean, the horizontal distribution in Fig. 2.8.c is characterized by low PV
values over the western side of Sierra Madre. Observations of the thermodynamic profile in Fig. 2.8.d (in purple) show a temperature with a well-mixed boundary layer reaching 700 hPa. When compared with ERAI (in blue), the temperature profile is very similar. However, the dew point temperature shows some discrepancies above 650 hPa. These profiles, despite some differences, show a well-mixed boundary layer reaching about 700 hPa, leading to a minimum in PV, and a better representation of the atmosphere from the model during the rainy season.

These diagnostics suggest that the minimum of PV found during the rainy season over the EPAC, is associated with low stability in the boundary layer (reaching ~750 hPa) over the western side of Sierra Madre in Mexico. When considering other mountain regions, we observe a similar pattern, characterized by low PV above it and downstream (not shown). Though not dealing directly with PV, this result is in agreement with previous literature over mountainous terrain, showing that convective boundary layers are characterized by dry and well-mixed $\theta$ profiles (Whiteman et al. 2000; De Wekker and Kossmann 2015) that can be advected downstream (Arritt et al. 1992). The minimum in PV from a well-mixed boundary layer is analogous, albeit at a much smaller horizontal scale, to the pronounced minimum observed over Africa (Thorncroft and Blackburn 1999).

Overall, this brief description of PV on isobaric levels lets us identify some important features, potentially significant for the growth and propagation of EWs:

i. A positive PV a strip extending from the IAS and into the EPAC, linked to the CLLJ at low levels, and to the mid-level jet (600 hPa) over the EPAC.
ii. A region of low PV poleward of the observed maximum, linked to surface heating and low stability over the Sierra Madre in Western Mexico, extending from the surface, reaching the level of 750 hPa and decaying aloft.

iii. Consistent with i and ii, the meridional PV gradients are negative at mid-levels of the troposphere, satisfying the Charney-Stern condition for barotropic instability.

ii. Vertical analysis

The latitude-height distribution of PV is presented in Fig. 2.9 at four different longitudes along the PV strip between the IAS and the EPAC (see Fig. 2.6.c for locations). Meridional PV gradients are included (left column, as well as a zoomed-in perspective including zonal winds and isentropes in the lower troposphere where we observe the EWs (right column). All four sections share similar large-scale characteristics such as high PV in midlatitudes and in the stratosphere and low PV in the equatorial region. More notable differences are present in the lower troposphere around 10°N-15°N and these will be discussed below.

At 75°W (Fig. 2.9.a), we see the distinctive CLLJ with easterlies exceeding 10 ms⁻¹ around 14°N. Consistent with this, a pronounced PV maximum is aligned with the cyclonic shear side of the jet. As a consequence, we also note the presence of positive meridional PV gradients on the equatorward side of the jet and negative meridional gradients on the poleward side. Poleward of this maximum, we observe a relative minimum in PV at 18°N, consistent with the anticyclonic shear vorticity of the jet.

At 80°W (Fig. 2.9.b) we still see the CLLJ with peak zonal easterly winds ~14°N exceeding 10 ms⁻¹. The PV maximum south of the jet is much weaker and less distinctive, consistent with weaker cyclonic shear equatorward of the jet maximum. Consequently, we observe a weaker
negative PV meridional gradient poleward. However, a secondary maximum in PV at mid-levels (~800 hPa) starts to be notable, which becomes more apparent further west.

At 90°W - over the Pacific Ocean (Fig. 2.9.c) - the distribution of PV changes slightly. At 10°N, two maxima are present at around 800 hPa and 600 hPa. Both lie on the cyclonic side of the zonal easterly jet, with a minimum between them. At 13°N a very distinctive and deep low-level PV minimum is present, collocated on the anticyclonic side of the easterly jet around 800 hPa, with similarities as over the western side of Sierra Madre, as discussed above. This distribution sets up a negative meridional PV gradient reaching the 400 hPa (see left figure).

Lastly, at 100°W (Fig. 2.9.d), two maxima over the EPAC are again evident: At mid-levels with a maximum at 550 hPa, and a secondary maximum at 750 hPa. Again, these are located on the cyclonic side of the jet, which peaks at 600 hPa. The low PV is again notable on the anticyclonic side of the jet over the western Sierra Madre. This distribution of PV results in a negative meridional PV gradient from the ground and above 400 hPa, and though weak, it satisfies the necessary condition of instability where $\nabla V < 0$.

The previous features in the vertical structure of PV can be summarized as this:

i. The PV maxima over the EPAC-IAS varies in height with longitude, rising westwards (following the easterly jet). While over the Caribbean this is at low levels, over the EPAC its around 600 hPa.

ii. There is a significant region of low PV over the western Sierra Madre that enhances the negative meridional PV gradients in the EPAC. This is in agreement with our findings that this is due to localized dry convection over the high terrain.
iii. There exists a coherent meridional PV gradient along the EPAC-IAS region that satisfies the necessary conditions for barotropic instability.

Figure 2.10 shows a summary of the mean state from a PV-$\theta$ perspective. Along the jet maximum over the EPAC, illustrated in Fig. 2.10.a by the purple dots, we considered the region where it reaches its maximum, enclosed in the box. These points were aligned to obtain the mean-zonal averaged values, shown in Fig. 2.10.b. This figure shows the jet core over the EPAC (in dashed lines), the PV distribution (shading), and isentropes. Hatching illustrates the mountain region of the Sierra Madre at 100.5$^\circ$W.

From this perspective, the averaged jet core reaches 5.5 ms$^{-1}$ at 600 hPa. On its equatorward side, we observe a PV maximum on the cyclonic side of the jet, slightly above it at 500 hPa. A secondary maximum is located at 800 hPa, 2$^\circ$ south relative of the jet maximum, and associated with the low-level jet over the EPAC (see Fig. 2.9.c). A PV minimum is observed over the mountain region of the Sierra Madre, associated with the dry convection and a more well-mixed boundary layer (as hinted by the bowing down isentropes) as previously shown. Together, this PV distribution sets up a negative PV meridional gradient, as illustrated in Fig. 2.10.c. While this PV meridional gradient reaches a maximum at 800 hPa, at this level the winds are relatively shorter in length compared to those above. At 600hPa, both the PV meridional gradient and the jet satisfy the Charney-Stern condition for barotropic instability, as well as the Fjortoft condition, setting up conditions for growth of EWs. After these results, we consider next the potential for the zonal flows to be characterized by the necessary conditions for baroclinic instability.
b. Baroclinic considerations

One of the conditions associated with baroclinic growth requires that the meridional gradient of potential temperature $\theta_y$ at the boundaries (surface or tropopause) has opposite sign to the meridional gradients of potential vorticity in the atmosphere. To explore this, Fig. 2.11 shows the distributions of $\theta$ at two levels: at the upper level near the tropopause where $PV = 2$ PVU, given the strong stratification acting as a lid, and at the low level of 950 hPa close to the surface.

At upper levels, Fig. 2.11.a shows a broad distribution of negative meridional theta distribution all over the EPAC-IAS region - notably at 15°N - stating that at upper levels $\theta_y < 0$. At low levels, Fig. 2.11.b shows a positive theta gradient equatorward over the EPAC. However, over the warm pool, poleward of 10°N, this temperature gradient is weak with a magnitude of 3 K, setting up the condition of $\theta_y > 0$ at low levels. Therefore, the meridional gradients of temperature at the lower boundary, as well as the PV meridional gradients in the interior of the fluid (at 600-700 hPa in Fig. 2.6.a-b and 2.9.c-d), are given as $\theta_y > 0$ and $PV_y < 0$ respectively, satisfying the necessary conditions for baroclinic instability over the region. The relevance of this mechanism on the growth of EWs will be explored in chapters 4 and 6 of this work.

These conditions resemble the environment over Africa, albeit at shorter scales and with weaker intensities. We can observe these differences in the meridional gradients of temperature at the surface, or in the intensity of jet (see e.g. Thorncroft and Hoskins 1994). In particular, we have observed that over the EPAC, the PV maximum is not at the same level with the mid-level jet, as when compared with the AEJ. This result motivated us to consider the role of convection and its associated latent heat on the impact and distribution of PV.
2.5. Diabatic heating over the Eastern Pacific and the Intra-Americas Sea

As stated in the introduction of this chapter, convection has a prominent role in determining the mean state that supports EWs. This section now explores the impact that convection has in terms of diabatic heating, and the role that it has on the generation of PV.

Figure 2.12 shows the apparent heat source (Q1) over the EPAC-IAS region, as calculated from Eq. (2.1) based on ERAI. An important point to make here is that Q1 results from model fields, thus, this variable is determined by the model representation of convection.

Figure 2.12.a shows a vertical section of Q1 averaged over the convective region of the ITCZ between 5°N-12°N (enclosed in Fig. 2.12.b for reference). Local maxima of Q1 are found over the Panama bight at 78°W - west of the Andes - with a peak heating of 7 K·day⁻¹, as well as over 84°W, from 900 to 250 hPa, with peaks close to 5 K·day⁻¹. Additionally, to the west of 85°W, two maxima are observed, around 800 hPa and 500 hPa, with a minimum around 650 hPa. From previous research, we can note that the vertical maximum west of the Andes suggests the existence of predominantly deep convection over the Panama bight (see Fig. 2.2.e), this in agreement with results of Mapes et al. (2003a), Zipser et al. (2006), and Zuluaga and Houze (2015).

The low-level heating maximum around 800 hPa is consistent with the presence of shallow convection (Huaman and Schumacher 2018), resulting from the low-level wind convergence over the ITCZ (Zhang et al. 2004; de Szoeke et al. 2006). The top-heavy maximum around 400 – 550 hPa is consistent with the presence of stratiform heating (Schumacher et al. 2004; Zuluaga and Houze 2015; Huaman and Schumacher 2018). Though the idealized stratiform heating profile is characterized by a cooling region above 600 hPa (see Fig. 2.1.d), this is not
observed here, likely due to the cancellation with the shallow heating. Instead, a minimum is observed close to 600 - 650 hPa. Previous studies have shown that stratiform cooling over oceanic tropical regions is best observed around 4 km in height, because of strong cooling from melting near 0°C (Ahmed et al. 2016; Huaman and Schumacher 2018). Therefore, we can reasonably associate this minimum with the stratiform profile, in agreement with observed results (see Fig. 2.1.e-f). This result also allows us to assume that over this region, ERAI is has shown to reproduce the main convective features over the EPAC.

To explore the horizontal extension of Q1, we considered two regions where maxima are found along the vertical. At upper levels, the layer average between 400 hPa and 600 hPa was considered (see Fig 2.12.a). In this, the level of 600 hPa was also considered given the location of the mid-level jet, and the proximity that it has closer to the upper level maxima. At lower levels, this is composed by averaging between 750 hPa and 800 hPa (see Fig. 2.12.a).

The horizontal distribution of the apparent heat source Q1 is shown in Fig. 2.13. At upper levels, Fig. 2.13.a shows a fairly uniform Q1 maximum over the EPAC. In particular, this maximum is also closer to the coasts over the WHWP, similar to the distribution of broad stratiform regions in Fig. 2.1.c as in Zuluaga and Houze (2015). At lower levels, Fig. 2.13.b shows a maximum more confined to the ITCZ region, highlighting the presence of shallow convection over the EPAC, as found by Huaman and Schumacher (2018). The maximum east of the Panama bight, is associated with deep convection, (Zipser et al. 2006; Zuluaga and Houze 2015), as previously seen in Fig. 2.13.a and consistent with Fig. 2.12.a.

Overall, these horizontal distributions highlight the heating distribution from two different convective profiles found over the region. While at upper levels the stratiform heating
dominates the region (Fig. 2.12.a), at lower levels heating is more characteristic of shallow convection (Fig. 2.12.b), though more confined to the ITCZ. In addition, deep-convective heating is found to the east of the Panama bight. The impact that Q1 has on the PV tendency will be helpful to explain the PV distribution over the EPAC, topic covered next.

Figure 2.14 shows the vertical distribution of the diabatic PV tendency, as calculated from Eq. (2.2). Consistent with the diabatic heating from Fig. 2.12.a, PV production is found clearly at two levels: at mid-levels around 550 hPa, and at low levels around 875 hPa. At mid-levels, the PVT maximum is below the upper-level maximum in Q1 associated with stratiform heating. This is due to strong gradients of diabatic heating (see Fig. 2.12.a) where the term $-\frac{dQ_1}{dp} > 0$ in Eq. 2.2. In the same way, increased PV production is found at low levels around 875 hPa and close to the ground). This is again explained by the increased diabatic heating in the lower troposphere resulting from a prominent contribution from shallow convection over the EPAC. Over the Panama bight region, a maximum is found at low levels, resulting from the same principle, in this case, the vertical gradients in Q1 are even stronger but limited to 800 hPa. At this last point, the environment resembles that over west Africa at the Guinea highlands, where the PVT is not seen above 800 hPa and vertically lies below a destructive region of PVT (Janiga and Thorncroft 2013).

To examine the horizontal distribution of the PV tendency, Fig. 2.15 shows an average of mid- and low- levels as indicated by the dotted lines in Fig. 2.14. These include 550 hPa - 600 hPa, and 850 hPa - 875 hPa levels respectively. Figure 2.15.a shows that at mid-levels the PV tendency is found to the west of Central America and poleward of 10°N. This maximum can be attributed to stratiform heating associated with MCSs in their later stages of development (Zuluaga and Houze 2015; Huaman and Schumacher 2018). Over the lower troposphere, Fig. 2.15.b shows a
defined strip of high PV tendency over the EPAC, following the distribution of Q1 around 750 hPa (see Fig. 2.13.b), in association with the shallow convection from the southern part of the ITCZ (Huaman and Schumacher 2018).

Overall these results show the impact of convection and its associated latent heat release on the distribution of PV. This is helpful to clarify the mean state over the EPAC. At upper levels, LH is associated with stratiform convection, creating a strong PV maximum at mid-levels (550 hPa - 600 hPa). This is in phase with that maximum observed above the cyclonic shear of the mid-level jet in Fig. 2.10.c, suggesting that stratiform convection plays a fundamental role on the creation of PV over the EPAC. Additionally, it is at around the level of 600 hPa, where the Charney-Stern and Fjortoft conditions for barotropic instability are met (see Figs. 2.10 for reference), suggesting that this an optimum level for the growth and propagation of EWs.

At lower levels, the latent heat release is associated with shallow convection. This has a stronger impact on the creation of PV tendency below the 800 hPa, stronger at the ITCZ. While at low levels the distribution of PV sets up the Charney-Stern conditions for barotropic instability, the mean state does not satisfy the Fjortoft condition (see the wind field in Fig. 2.4.c-d), suggesting that the conditions for the growth of EWs are partially satisfied, implying a weakly supportive state for EWs.

If we contrast the results of the distribution of Q1 and PVT over the EPAC with that over West Africa we find key similarities. Over the West African continent, Q1 is centered at 8.5°N, characterized by a vertical distribution at 12.5°W associated with deep convection, and a horizontal layer at 800 hPa associated with shallow convection over the eastern Atlantic. Positive PVT is found below this Q1 maximum (Janiga and Thorncroft 2013). Over the EPAC we have
observed that Q1 is found at two different levels (500 hPa and 800 hPa) with PVT distributions below and tilted in the horizontal. This is also impacted by the presence of the WHWP in the EPAC.

These features highlight the main differences between these two states while showing the complexity over the EPAC. Moreover, if discrepancies arise between models over the region of Africa (c.f. Janiga and Thorncroft 2013), the EPAC pose a bigger challenge regarding the way different models describe it as (Huaman and Schumacher 2018) showed. Further studies are needed regarding this aspect; however, these are left for a future work.

2.6. Summary

This chapter has presented an overview of the main features over the EPAC-IAS and how they may impact the life-cycle of EWs over the region. This was done first by reviewing the basic characteristics of the mean state, followed by an analysis of the conditions for instability using a PV-\(\theta\) framework, and finally reviewing the role of convection within the region. The key results from this analysis include:

- The EPAC-IAS is a region characterized by complex topography, the Western-Hemisphere Warm Pool, the ITCZ at 10\(^\circ\)N, and predominant deep convection over the Panama bight at 9\(^\circ\)N, 78\(^\circ\)W.

- Over the EPAC, the wind structure is characterized by a mid-level jet parallel to the coast, at around 600 hPa. This is characterized by a magnitude of 5.5 ms\(^{-1}\) and a length of 20\(^\circ\).

- The previous point suggests that EWs over the EPAC can track along this jet and intensify from the warm waters over the WHWP.
• A PV strip was identified at low levels that extends from the IAS region into the EPAC, rising from east to west, and linking the CLLJ and the mid-level jet over the EPAC. This distribution of PV is characterized by a maximum located on the cyclonic side of the mid-level jet, and a minimum poleward of this.

• Over the EPAC, at the level of the jet (600hPa), the distribution of PV satisfies the necessary conditions for barotropic instability: the Charney-Stern condition given by $PV_y<0$, as well as the Fjortoft condition, associated with $u<0$ where $PV_y<0$. Together these conditions suggest potential for barotropic growth of EWs over the EPAC region.

• The sources of positive PV anomalies over the EPAC are associated mainly with two different populations of convection: stratiform and shallow (Zuluaga and Houze 2015; Huaman and Schumacher 2018). The stratiform heating is key for the creation of PV at mid-levels (showing similar values to those observed in the mean PV), suggesting that this type of convection is the main driver of PV at 550 hPa - 600 hPa mostly over 10°N – 20°N over the EPAC. At low levels, the PV is created from shallow convection below 800 hPa over the ITCZ between 5°N - 10°N.

• Evidence is provided that suggests that sources of negative PV anomalies may result from dry convection over the western side of the mountain regions of the Sierra Madre. This is evident in May during the dry season, when convection produces a well-mixed boundary layer leading to minimum in PV. For JJAS, the minima are still present, though it is weaker. While dry convection may also have a role, it should be recognized that friction may also contribute. More research is necessary to completely answer this question. This is a key
and novel result that is basic for the setting up of a negative meridional PV gradient and fundamental for the Charney-Stern condition associated with barotropic instability.

- By exploring $\theta$ gradients close to the surface, we found weak meridional gradients of $\theta$ at low levels with a magnitude of 3 K in 12° in latitude, this is $2.27 \times 10^{-3}$ K/km. When compared with West Africa, the meridional $\theta$ gradients are given as 16 K in 12° in latitude, this is $12.12 \times 10^{-3}$ K/km. The much weaker gradients in the EPAC region suggests a more minor role for baroclinic growth of EWs over the EPAC.

The findings in this chapter set up the basis for exploring the characteristics of EWs over the EPAC, as well as their variability and genesis, among other topics covered in this thesis. Additionally, this work can be further extended to explore the differences in the mean state in different reanalysis, in particular to identify the role of convection in these. This can also include climate scales, to identify differences under scenarios of climate change.
Figure 2.1. Convective features over the EPAC from previous works. a. Spatial distribution of the probability of a location under a. Deep convective core, b. wide convective cores, and c. broad stratiform during JJA 1998-2012. Retrieved from Zuluaga and Houze (2015); d. Idealized heating profiles for deep, convective and shallow precipitation; e. LH (K day$^{-1}$) from observations averaged between 130-90°W, and f. Q1 (K day$^{-1}$) vertical profile for JJA comparing models with LH (in continuous black line). Figures d. - f. retrieved and adapted from Huaman and Schumacher (2018).
Figure 2.2. Geographic and thermodynamic features over the EPAC-IAS region. a. Topographic height (in m). Data source from NOAA-ETOPO1. b. Daily climatology of SST (in °C). Enclosed in purple is the Western Hemisphere Warm Pool. Data source from NOAA-Interpolated SST. c. Daily climatology of specific humidity (in g kg\(^{-1}\)), and d. Equivalent potential temperature (in Kelvin) at 950 hPa level. Data source from ERA-Interim. e. Outgoing longwave radiation (in Wm\(^{-2}\)). Data source from NOAA Interpolated OLR, and f. Rainfall (in mm day\(^{-1}\)). Data source from NASA-TRMM using the 3B42 dataset. Mean climatology calculated during JJAS for 1980-2015, except for TRMM data, from 1998-2011.
Figure 2.3. Daily climatology of Mean Sea Level Pressure (in hPa) and winds (in ms\(^{-1}\)) at the level of 1000 hPa for June to September during the years 1980-2015. Data source from ERA-Interim at 0.5° resolution.
Figure 2.4. Panel plot of the daily climatology of wind field and its magnitude (shading, in ms\(^{-1}\)) and streamfunction (contours in 10\(^6\) m\(^2\)s\(^{-1}\)) at a. 200, b. 600, c. 850 and e. 950 -hPa levels for June to September during the years 1980-2015. Wind vector is indicated for reference and units are in ms\(^{-1}\). Red box in d. is the region averaged longitudinally for a cross section in Fig. 2.5. Data source from ERA-Interim at 0.5° resolution.
Figure 2.5. Cross-section of daily climatology of zonal wind (shading, in ms$^{-1}$) and potential temperature $\theta$ (contours, in K) over the Caribbean from 65°W to 80°W and from 5°N to 25°N (as shown in Fig. 2.3.d) during JJAS 1980-2015. Shading at 73°W where topography is most extended. Data source from ERA-Interim at 0.5° resolution.
Figure 2.6. Potential vorticity (in PVU) at the isobaric levels of a. 600 hPa, b. 700 hPa, and c. 850 hPa over the EPAC-IAS region during JJAS 1980-2015. Vertical lines in c. represent the longitudes for cross sections in Fig. 2.9. Data source from ERA-I at 0.5°.
Figure 2.7. Comparison of the horizontal PV distribution and the associated vertical structure at Guadalajara (GDL) station (20.67°N, 103.38°W; enclosed in the pink circle) on two different seasons: maximum dry month (left) against mean JJAS (right), as retrieved from ERAI. a. Mean PV at 600 hPa during May, and b. associated May sounding at 00Z at Guadalajara station. c. PV at 600 hPa during JJAS, and d. associated JJAS sounding at 00Z at Guadalajara station. Data source for horizontal distribution from ERA-I at 0.5° during 1980-2015.
Figure 2.8. Comparison of the horizontal PV distribution and the associated vertical structure at Guadalajara (GDL) station (20.67°N, 103.38°W; enclosed in the pink circle) on two different seasons: maximum dry month (left) against mean JJAS (right) as retrieved from observations (in purple) and ERAI (in blue) during year 2013. a. Mean PV at 600 hPa during May, and b. Associated soundings at 00Z at Guadalajara station. c. PV at 600 hPa during JJAS, and d. associated JJAS soundings at 00Z at Guadalajara station. Data source for horizontal distribution and soundings from ERA-I at 0.5°. Data for observed soundings was obtained from the Department of Atmospheric Science, University of Wyoming.
Figure 2.9. Left: Vertical cross section of potential vorticity in shades (PVU) with meridional PV gradients in contours; and Right: Zoom version of PV in the box indicated on the left, with zonal winds in contours (in ms$^{-1}$) and isentropic levels (in K) at different longitudes over the EPAC: At a. 75°W, b. 80°W, c. 90°W, and d. 100°W during JJAS 1980-2015. For PV gradients, positive meridional gradients are in continuous red line, while negative values dashed. Zero contour is in black. Positive zonal winds are continuous and negative dashed. Data source from ERA-I at 0.5°.
Figure 2.9. Continued.
Figure 2.10. Schematic diagram summarizing the vertical configuration over the EPAC along the jet axis. **a.** Wind speed (ms$^{-1}$) at 600 hPa highlighting the jet core (in purple dots). **b.** Mean vertical structure along the jet core within 106°W to 95°W (as in a.) showing the distribution of PV (PVU), zonal wind (ms$^{-1}$) in black-dashed contours and isentropes (K) in gray contours. **c.** As in b, but black-dashed contours showing negative meridional PV gradients. Latitudes are relative to the center of the jet core. Hatching represents topography at the center of the region at 100.5°W. Data source from ERA-Interim at 0.5° during JJAS 1980-2015.
Figure 2.11. Potential temperature (in K) at upper and lower levels. a. At the level of dynamic tropopause: 2 PVU, and b. at 950 hPa. Data source from ERA-Interim at 0.5° during JJAS 1980-2015.
Figure 2.12. Apparent heat source Q1 (K day$^{-1}$) over the EPAC. a. Average over 5$^\circ$-12$^\circ$N in latitude during June to September 1980-2015. Horizontal lines show the layers used for horizontal analysis. b. Averaged mean precipitation (mm day$^{-1}$) during June to August 1998-2011. The black box shows the latitudes used for a. Data source from ERAI at 0.5$^\circ$ for a., and TRMM-3B42 for b.
Figure 2.13. Horizontal apparent heat source Q1 (K day$^{-1}$) in two different layers (as shown in Fig. 2.12.a). At a. 400-600 hPa layer, and b. 750-800 hPa layer. Data source from ERA-I at 0.5°.
Figure 2.14. Diabatic potential vorticity tendency $Q_1$ PV (x$10^{-1}$ PVU day$^{-1}$) averaged over $5^\circ$-12$^\circ$N in latitude (as in Fig. 2.12.b) during JJAS 1908-2015. Horizontal lines show the selected levels for further exploration. Data source from ERA-I at 0.5°.
Figure 2.15. Horizontal diabatic potential vorticity tendency Q1 PVT ($10^{-1}$ PVU day$^{-1}$) in two different layers as shown in Fig. 2.14. At a. 550-600 hPa layer, and b. 850-875 hPa layer. Data source from ERA-I at 0.5°.
Chapter 3. Analysis of Easterly Waves over the tropical Eastern Pacific and Intra-Americas Sea Region: Relationship with Convection and Variability

3.1. Introduction

Easterly Waves (EWs) are synoptic westward-propagating systems in the tropics observed over Africa and the tropical Atlantic (Reed et al. 1977; Thompson Jr et al. 1979; Kiladis et al. 2006), the tropical western Pacific (Yanai et al. 1968; Reed and Recker 1971; Serra et al. 2008), and over the tropical Eastern Pacific (EPAC) and the Intra-Americas Seas (IAS) region (Tai and Ogura 1987; Petersen et al. 2003; Serra et al. 2008; Serra et al. 2010; Rydbeck and Maloney 2014). Over the tropical Americas, the extent to which EWs over the EPAC are linked with those over the Atlantic including African EWs (AEWs), is still a subject of debate. For example, Frank (1972) found that about 65% of tropical storms over the EPAC could be associated with systems over the Atlantic. Additionally, Frank (1972) stated that “there is little doubt that Atlantic systems play an important role in the east Pacific storm genesis”. While some research supported this idea about the propagation of EWs over the Atlantic to the EPAC (Carlson 1969; Shapiro 1986; Molinari et al. 1997), others tried to elucidate in situ mechanisms for the origin of EWs over the EPAC (Ferreira and Schubert 1997; Serra and Houze 2002; Serra et al. 2008; Rydbeck et al. 2017). However, none of these efforts have provided a consistent explanation for their presence. This may have resulted from the different approaches used to study EWs and TCs in each basin, as well as the lack of a consistent EW metric. One of the first goals of this chapter is to locate the EW storm tracks over the EPAC and to clarify whether the EWs are generated in situ or if they come from upstream using metrics associated with EWs.
Previous studies on EWs over the EPAC based on composite analysis have shown a direct relationship between EWs and convection (Serra et al. 2008; Serra et al. 2010; Rydbeck and Maloney 2014), as well as a preferential location ahead of the trough (Petersen et al. 2003). Despite this coherent association, it has not been well established how EWs modulate convection over the EPAC, including how this varies in space and time. The current knowledge on this topic over the EPAC is limited only to a few papers (Tai and Ogura 1987; Gu and Zhang 2002). While Tai and Ogura (1987) suggested that around 40% of the daily convective variability is associated with EWs, the more recent study of Gu and Zhang (2002) suggested that over the ITCZ, westward propagating synoptic-scale deep convective systems explain only 15%-20% of the total daily OLR variability. The discrepancy between these results may be a consequence of the time period used in these, being only one summer in Tai and Ogura (1987), versus 20 years used by Gu and Zhang (2002). For completeness, this chapter will also explore the percentage of variability explained by convective and dynamical metrics associated with EWs.

In addition, the variability of EWs on interannual and interdecadal timescales will be explored. On interannual timescales, it has been documented that El Niño-Southern Oscillation (ENSO) is the leading mode of interannual variability of SSTs over the tropical Pacific (Wallace et al. 1998) with a periodicity around 5 years (Deser et al. 2010). Through its teleconnections, It has also been recognized that ENSO has an important influence on TC activity over the Atlantic (Xie et al. 2005; Kossin et al. 2010), the Caribbean (Amador et al. 2010), and the EPAC (Wang and Fiedler 2006; Wang and Lee 2009; Jin et al. 2014). However, previous research has not considered the interannual variability of EWs over the EPAC. Only one study (Shelton 2011) addressed the issue regarding EWs over the Caribbean with respect to ENSO, showing that during El Niño, EW
tracks decrease in number. This result can be interpreted as fewer disturbances moving from Africa or a smaller number of them generated within the region. With only one study covering this topic, it is still not clear how ENSO affects EWs over the EPAC.

Even less is known about interdecadal variability of EWs over the EPAC-IAS region. As background, Goldenberg et al. (2001) showed the modulation of TCs over the Atlantic Ocean, including the Main Development Region, by the Atlantic Multidecadal Oscillation (AMO) during its positive phase. Shelton (2011) considered the influence of AMO on EWs over the Caribbean, finding that AMO phases do not affect the number of EWs over the IAS region. Instead Shelton (2011) found that these differ only in latitude in association with the location of the North Atlantic Subtropical High. With only a few studies regarding the role of AMO on EWs, it is unclear how this oscillation impacts EWs over the wider EPAC-IAS region. Furthermore, it is not clear if there is a role for other interdecadal oscillations, such as the Pacific Decadal Oscillation (PDO). Motivated by this, the present work also explores the variability of EWs on interannual and interdecadal timescales.

This chapter focuses on the presence of EWs over the EPAC, their modulation of convection, and variability on interannual to interdecadal timescales, and it is structured as follows: The data and methods used are presented in section 2. Section 3 presents a spectral analysis and an analysis of variance of the dynamical and convective metrics associated with EW activity. This includes a quantitative analysis of the significance of synoptic timescales. In section 4, we study the relationship between synoptic-time convective and dynamical metrics, as well as their year to year variability, which will lead to an exploration of the interannual and interdecadal timescales in section 5.
3.2. Data and Methods

a. Reanalysis dataset

The dynamical fields analyzed in this chapter are based on the ERA-Interim (ERAI) reanalysis (Dee et al. 2011). The horizontal grid spacing used is 0.5°, with a temporal resolution of 6 hours, which is adequate to represent the synoptic scales of interest here. For most of the analysis, the horizontal level of 600 hPa was considered given that, in the mean state, this level is supportive of the growth of EWs over the EPAC (see Chapter 2). All calculations were performed during the extended summer season June to September (JJAS) from 1980-2015.

Previous research on EWs over Africa (Berry et al. 2007; Brammer and Thorncroft 2015) and the IAS region (Shelton 2011) has shown that curvature vorticity is a useful diagnostic to identify and retrieve the trough of an EW from the background shear vorticity. Therefore, curvature vorticity (500 km radially averaged) was calculated from ERAI will be used and defined as the dynamical metric in this research.

b. Satellite datasets

The characteristics of convection are analyzed based on satellite observations of brightness temperature (T_B) and outgoing longwave radiation (OLR). T_B data was obtained from the Cloud Archive User Service (CLAUS) of the European Union (Hodges et al. 2000). It consists of a high-resolution data set in time (3-hourly), and space (1.0° by 1.0° latitude-longitude grid). CLAUS T_B is a product of multiple satellite observations in the 10-12 µm infrared window and the version used spans from July 1983 to June 2005. This dataset has also been used for the analysis of power spectra in a study of AEWs (e.g. Mekonnen et al. 2006).

To cover the whole period of analysis 1980-2015, OLR data from NOAA (Liebmann and Smith 1996) was also used. OLR extends from June 1974 to the present day, and it has a spatial
coverage of $1.0^\circ$ by $1.0^\circ$ latitude-longitude global grid on a daily basis, and it will be used mainly to study the variance of convection. Both, $T_B$ and OLR are utilized as the convective metrics in this study.

c. **Spectral analysis and temporal filtering**

To assess the significance of synoptic timescales in dynamical and convective fields, a spectral analysis was performed on curvature vorticity at 600 hPa and $T_B$. This analysis follows Mekonnen et al. (2006), since it is useful to identify significant frequencies in a simple and objective way. Power spectra were created based on $3^\circ \times 8^\circ$ latitude-longitude boxes within the EPAC-IAS region. For consistency both grids were taken at $1^\circ$, with 36 grid points within each box (including initial points). For each grid point, a spectrum was obtained and then, the mean of the spectrum (MOS) was calculated by averaging over all available years. The MOSs are compared against mean red noise spectra to examine the significance of peak periods (Mekonnen et al. 2006; Slingo et al. 1992). Significant power is defined as a peak in the spectra greater than a corresponding red noise value. To be practical, MOS is only presented for selected regions where convective and dynamical EW-related metrics were the most intense and overlapped.

To isolate the synoptic timescales between 2 and 10 days, a bandpass Lanczos filter was applied to the selected dataset (Duchon 1979). To sharpen the response of the temporal filter, 30-day data points were used on each side of the time series to be filtered. This calculation was performed only on OLR (given that it covers the whole dataset), as well as on the dynamical metric.

To further explore the variability of westward propagating systems (EWs), convective and dynamical metrics are filtered in space and time for the TD-band, removing eastward propagating
systems, such as Kelvin waves. We follow the methodology employed by Wheeler and Kiladis (1999), which filters for westward propagating wavenumbers between 6-20 and temporal scales between 2-10 days.

3.3. Analysis of dynamical and convective metrics

a. Overview

To have a large-scale perspective of the EW activity over the EPAC-IAS region, Fig. 3.1.a-b shows the variance of the dynamical and convective metrics in the 2-10–day band. This spatial analysis is for reference only and it avoids showing the whole region where spectra is not significant. A more detailed spatial analysis, where these figures will be returned to, will be included in Section 3.3.3.

Curvature vorticity variance at 600 hPa in Fig. 3.1.a is dominated by a peak over the EPAC, north of 10°N and west of 100°W, oriented roughly parallel to the Mexican coast. A close inspection of the variance contours suggests a potential origin downstream of or close to Central America. A secondary feature is evident over the Atlantic, characterized by high variance and local maxima. This is closely related to AEW activity originating over West Africa (Thorncroft and Hodges 2001; Brammer and Thorncroft 2015). Interestingly, the curvature vorticity variance shows clearly a minimum over the Caribbean, suggesting from this alone, that the synoptic-scale activity over the tropical EPAC is only weakly related to that over the Atlantic Ocean. Therefore, this result suggests that some EWs in the EPAC may have in situ origins.

OLR variance in Fig. 3.1.b shows the highest activity over the oceanic regions and weaker activity over the land. In particular, a distinctive maximum is evident over the Panama bight, off the coast of Colombia and downstream of there. The increased variance over the ocean is likely
a consequence of the larger availability of moisture and heat over the ocean, in comparison with land. The peak over the Panama bight is associated with deep convection and MCSs (Mapes et al. 2003b; Zipser et al. 2006), and the area has also been hypothesized to be a genesis region for EWs (Serra et al. 2010; Rydbeck et al. 2017). A weaker maximum is observed to the west of the continental region at 15°N, associated with TCs (Renard and Bowman 1976; Zhao and Raga 2015), as well as with EWs (Rydbeck and Maloney 2014). Over the Gulf of Mexico, a local maximum is also observed associated with mesoscale convection and TCs (Jauregui and Zitácuaro 1995). In contrast, the 2-10-day OLR variance shows a minimum over the Caribbean, suggesting a weak contribution of synoptic-scale systems to daily variations in convection there.

A particular characteristic of the 2-10-day filtered OLR distribution is that it looks similar to the OLR distribution, with large values where the mean OLR is also large (see Chapter 2, Fig. 2.2.e for comparison). To help establish the significance of synoptic timescales, a spectrum analysis of OLR was carried out in these key locations in the region.

For brevity, we present only those locations based on the guidance that total variance presents, linking convective and dynamical metrics. This is illustrated in Fig. 3.1.c, where curvature vorticity variance at 600 hPa is shaded and OLR variance is contoured. This arrangement is also collocated in space similar to previous analyses of EWs over the EPAC (Serra et al. 2010; Rydbeck and Maloney 2014). A reference line in purple illustrates the region where curvature vorticity is a maximum. By merging these areas, we selected specific regions linking these metrics, with 14 sectors indicated by the blue boxes in Fig. 3.1.d.
b. Spectral analysis

A spectral analysis of the dynamical and convective metrics associated with EWs is presented in Fig. 3.2. In this figure, shading in green shows significant power, while the red dashed lines the confidence interval. The 2-10-day window is indicated by vertical purple lines, highlighting the period typical of EW timescales.

The dynamic metric (curvature vorticity at 600 hPa in Fig. 3.2 – left) shows significant power within the 2-10-day window for all sectors considered over the EPAC, highlighting the importance of the synoptic-scale systems. These peaks shift power and intensity from 4 days in the south-easternmost sectors, to 6 days on the north-westernmost sector. In addition, there is a notable peak at diurnal timescales, likely associated with the diurnal cycle in convection (discussed next). Though not shown, similar results were obtained when using meridional wind as the dynamical metric.

The convective metric ($T_B$ in Fig. 3.2 – right) shows significant power at EW timescales, though in a more complex way. Power spectra is evident (though weak) in sectors 1-6, peaking between 4 to 5 days. This power diminishes and is not significant in sectors 7 and 8, but it strengthens again after sector 9, where power is significant within the 4-10-day band. In the northernmost sectors, it also must be recognized that there could be a contribution from mid-latitudes systems, however, we will not be focusing on this potential contribution. Additionally, all sectors are characterized by a pronounced diurnal peak, which decreases form east to west. It is clear that these diurnal peaks are more intense over continental regions, in agreement with the diurnal cycle of convection over land (Slingo et al. 1992). Furthermore, the weakening in the diurnal peaks is also likely associated with the drier conditions found poleward over the EPAC in
comparison with more equatorial areas. Significant peaks are also present in sectors 3-10 at timescales longer than 30 days, consistent with the known influence of intraseasonal oscillations in the region, including biweekly oscillations (Serra et al. 2014) and the MJO (Maloney and Hartmann 2000a).

The correspondence between the daily spectral peaks in curvature vorticity and brightness temperature is likely associated with diurnal variations in convection within the EWs, in a process similar to that described in previous research (Gray and Jacobson 1977; Dunion et al. 2014). This topic, however, is out of the scope of this research but deserves further examination.

When considering both spectra for dynamic and convective metrics (Fig. 3.2), we observe that both are significant, except in the sectors 7 and 8. This result could be associated with the life cycle of EWs over the EPAC, characterized by an initial strong coupling that weakens south of the Isthmus of Tehuantepec, reinvigorating later over Western-hemisphere warm pool. However, this result could also be associated with two different tracks of EWs over the EPAC: One zonal track over the ITCZ west of Central America, possibly associated with ITCZ breakdowns (Ferreira and Schubert 1997); and a second track parallel over the EPAC developing over the warm waters, as in Serra et al. (2010) and Rydbeck and Maloney (2014). A more detailed analysis of individual tracks and their characteristics is necessary. These hypotheses will be further explored in chapter 4 when we consider EW structures and their evolution.

When compared with the West African and the eastern Atlantic regions (see Mekonnen et al. 2006), spectral peaks in $T_b$ over the EPAC are weaker. Over West Africa, EWs peak between 2-6 days with higher spectral peaks, indicating a clear and strong periodicity of the waves which
is well known and observed. We hypothesize that these differences arise mainly from the dynamical characteristics in the mean state. Over West Africa, a very favorable environment is found for establishing and supporting EWs, which includes coherent and significant meridional gradients of potential vorticity (PV) at the jet level and deep convection near the entrance of and along the jet. In contrast, over the EPAC, the environment is more complex, with gradients of PV and jets at different levels and an intricate convective distribution. Therefore, differences in the mean state impact the frequency and power of EWs over the EPAC in comparison with West Africa.

To highlight further differences, we explore next the percentage that the metrics associated with EWs explain of the total variance. This will also help to clarify this uncertainty in attribution and to complement the results found in literature (Tai and Ogura 1987; Gu and Zhang 2002).

c. Analysis of spatial variance

To clarify the percentage that EW timescales explain of the total daily variance, we present an analysis of the spatial distribution of variance over the EPAC-IAS. This is performed by obtaining the ratio between the variance associated with timescales of EWs and the total variance. These timescales include first the 2-10–day filtered variance, and second only the westward-moving TD-filtered variance (in this sense excluding eastward propagating systems such as convectively coupled equatorial Kelvin waves) in both metrics.

i. Curvature vorticity

Figure 3.3 presents the spatial variance of the dynamical metric. In this analysis we are omitting values north of 25ºN, associated with mid-latitude weather systems, as observed in Fig.
3.3.a. Figure 3.3.a presents the total variance of curvature vorticity. High variance is located mostly over the tropical EPAC, with a possible link to Central America. Over the IAS region, maxima are observed poleward of 20°N, with low values over the Caribbean. From the synoptic-scale perspective, we observe highest variances located further south in the EPAC than over the IAS, not directly linked with the Caribbean region.

Figure 3.3.b shows the 2-10–filtered curvature vorticity at 600 hPa, which is the same as that in Fig. 3.1.a. Here we just recall that the maximum is over the EPAC and over the Atlantic, with a minimum over the Caribbean, a result that suggests that EWs on both basins are not totally linked at 600 hPa. In addition, a red box illustrates the region where synoptic variance presents a maximum. This has an average value of $0.193 \times 10^{-10} \text{ s}^{-2}$. This value will be used in a later section.

Figure 3.3.c shows the ratio of the 2-10-day filtered variance to the total daily curvature vorticity variance. In this plot we considered only regions over $0.125 \times 10^{-10} \text{ s}^{-2}$ from the total variance to avoid highlighting unimportant large ratios due to small denominators. The ratio explained by the 2-10–day variance in the region of high variances over the EPAC contributes around 45-50% of the total variance highlighting the important role played by synoptic timescales in this region.

When considering exclusively westward-propagating systems, Fig. 3.4 shows the TD-filtered curvature vorticity variance at 600 hPa. The TD-filtered curvature vorticity variance in Fig. 3.4.b is characterized by two maxima: one over the EPAC and another over the Atlantic Ocean, to the north of the Antilles, with a minimum in between. This result shows again, a clear separation between the Atlantic and the EPAC in westward-propagating systems over the tropical Americas. The ratio of the TD variance and the total curvature vorticity variance over the EPAC,
shown in Fig. 3.4.c, highlights the largest percentages of TD-filtered variance representing about a 25-35% of the total daily variance, less than the previous result for 2-10-day filtered variance, which included contributions of eastward-propagating systems. Though not shown, these results are also similar to those using meridional wind as a dynamical metric.

**ii. Convection**

Figure 3.5.a shows the spatial distribution of the total daily OLR variance. This is characterized by a maximum over the Western Hemisphere Warm Pool (WHWP - Wang and Enfield 2001) illustrated by a purple line over the EPAC. Secondary maxima are found over the western Caribbean Sea and the Gulf of Mexico, this last not showing significant peaks in spectral analysis (not shown).

Figure 3.5.b shows the 2-10–day filtered variance of OLR (as previously observed in Fig. 3.1.b), and it is shown only as a reference for the variance ratio. To recall, the main characteristics are a maximum over the Panama bight, a weaker maximum to the west of the continental region at 15°N over the WHWP, associated with TCs and EWs; and a minimum over the Caribbean. Note scaling differences in the colorbar.

Figure 3.5.c shows the percentage ratio of the 2-10–day OLR variance to the total daily OLR variance. For clarity, the area where total variance of OLR is less than 100 W²m⁻⁴ was masked again to avoid highlighting unimportant large ratios. The highest percentages are observed in the Pacific ITCZ region. Ratios greater than 60% are observed over the Panama bight region and downstream in the tropical EPAC. This ratio decreases to 40% over the region adjacent to western Mexico where we see the strongest synoptic scale curvature vorticity variances. These results
highlight the importance of synoptic-scale systems in determining daily variations in convection over the EPAC, in particular, to the west of Central America.

To further compare these results exclusively with westward-propagating systems, the percentage ratios were obtained for TD-filtered OLR, shown in Fig. 3.6. For reference, total OLR variance is displayed again in Fig. 3.6.a. Figure 3.6.b shows OLR-TD filtered variance (note again the scaling differences in the colorbar), which is dominated by a maximum to the west of the Mexican coast. This maximum is very zonal, and it is likely associated with the TD-filtering and the zonal wavenumber decomposition. A much weaker maximum is also observed downstream of the Panama bight, similar to that in the 2-10–day filtered variance. There is also a notable peak over the northern Caribbean. While this is not clear in the 2-10–day variance (see Fig. 3.6.b). It is hypothesized that this results from the TD-filter capturing the signal associated with AEWs.

Figure 3.6.c shows the TD/Total OLR variance ratio. Here, values less than 100 W²m⁻⁴ of the total OLR variance were masked for clarity. Over the EPAC, the highest percentages ratios are located downstream of Central America (close to the coast at 90°W) with values of 20-25%, while over the rest of the EPAC, it ranges between 15-20%. Over the Caribbean Sea, a maximum is found on the eastern side of the basin, with values around 25%. This higher value could be the result from minima in the total variance of OLR and TD-filtering. This result is slightly above that of Gu and Zhang (2002), who reported a 15-20 % of the variability in convection over the EPAC to westward-moving systems.

When compared with the region of West Africa and the eastern Atlantic, the convective metric shows a 30-40% in the 2-6-day variance. This value represents about a 50% more when
compared with the EPAC, highlighting the more frequent presence of EWs over west Africa. This also helps to explain the weak peaks found in previous spectra.

Overall, these results have provided an overview of the presence of EWs over the EPAC and the spatial variance of dynamical and convective metrics associated with them. A main result here is that the EW storm track that is oriented parallel to the coast west of Mexico, suggest an origin west of Central America (~10°N, 85°W), and with little indication of an origin in the Caribbean tracking into the EPAC. This provides a first evidence that many EWs over the EPAC could have their origin “in situ”. Despite Frank (1972) hypothesized that EWs over the Atlantic systems played an important role in the EPAC storm genesis, our results suggest that these may not be totally linked. However, it must be considered that Frank (1972) relied only on satellite imagery for his analysis with results not that evident. We now proceed to investigate how dynamical and convective metrics are related with each other, as well as their variability within the years used.

3.4. Year-to-year variability

Following Mekonnen et al. (2006) we performed a year-to-year comparison of the metrics associated with synoptic-scale systems to further study their variability. This consisted of obtaining the anomalies of the variance of the metrics used and averaging over a selected domain. The anomalies were calculated as deviations from the mean variance during JJAS 1980 - 2015, serving as a simple measure of the annual anomaly of EW activity. The domain was selected as the region where the maximum of TD-filtered variance in OLR was coincident with the highest TD-filtered variance of curvature vorticity at the level of 600 hPa. This region is illustrated in Fig.
3.7.a, enclosed between 11°N-16°N, 80°W-115°W, covering west of Central America and the tropical EPAC.

Figure 3.7.b contrasts the dynamical metric, represented by the 2-10–day filtered curvature vorticity at 600 hPa (in blue), with the convective metric, OLR-TD variance (in red). Considering first the dynamical metric (in blue), two distinctive periods are evident: most of the anomalies are negative between 1980 and 1998, and then positive between 1998 and 2015. Superimposed on this, there are also interannual variations. When considering the convective metric (in red), this follows the same general variation with negative anomalies between 1980-1998, and positives in 1998-2015. Overall, both metrics show a remarkable decadal signal. The relationship between dynamical and convective metrics yields a correlation coefficient of $r=0.304$ (bootstrap significant at 95% with 10000 repetitions). This correlation suggests that a higher synoptic variance in curvature vorticity is associated with higher synoptic variance in convection. The correlation coefficient is considered strong in similar studies (for example, Mekonnen et al. (2006) obtained a $r=0.19$ over West Africa and the tropical eastern Atlantic). Similar results were also obtained using meridional wind as a dynamical metric (not shown).

To explore the nature of the shift from negative to positive anomalies on the synoptic-scale metrics used, we considered the relationship these indexes have with SST anomalies over the same region. Figure 3.7.c shows the relationship between the dynamical metric (in blue) and SST anomalies (in red) over the selected region. There is a clear and evident change in SST anomalies around 1998, shifting from negative to positive values. A correlation coefficient $r=0.454$ (significant at 95%) was obtained, suggesting that changes in temperature in the WHWP drive the synoptic-scale activity over the EPAC. By identifying that SSTs drive the variability of
synoptic-scale activity, there exists the potential for predictability of EW activity on these timescales. The causes for this are explored below.

To clarify if ENSO has an influence on the synoptic-scale metric, Fig. 3.8 shows the detrended dynamical metric in blue, against the Oceanic Nino Index (ONI) in red, which is similar to ENSO34 (Trenberth and NCAR Staff 2020). A correlation coefficient of $r=-0.048$ (significant at 95%) was obtained. This result implies that ENSO has a very weak role on the synoptic-scale systems over the EPAC, particularly over the WHWP. This could be associated with the oceanic response mostly on long timescales; however, this topic deserve its own research and it is left for future work.

Overall, the previous paragraphs have shown the important result that SSTs play a main role as a driver or EW activity on interdecadal timescales. From this, the next section explores how the changes in SSTs on interdecadal variations impact the atmospheric fields (i.e. the mean state), and in turn, how these changes impact on EWs.

3.5. Interdecadal variability

The previous section showed the relevance of interdecadal variability over interannual variability on synoptic systems over the EPAC. This section explores how synoptic-scale systems over the EPAC are modulated on interdecadal timescales. We pay special emphasis to the potential role played by known low-frequency oscillations, such as the Pacific Decadal Oscillation (PDO) and the Atlantic Multidecadal Oscillation (AMO).

As a reminder, the characteristics of PDO and AMO are briefly presented here. A positive PDO index is associated with warmer SSTs in the tropical Pacific Ocean and colder temperatures in the north Pacific Ocean (see Fig. 3.9), similar to an El Niño state but on longer timescales
(Mantua and Hare 2002). The opposite pattern takes place with negative index values. In particular, and of relevance for this study, is that the WHWP is anomalously warm during the negative phase. Regarding the AMO, a positive index is associated with warmer SSTs in the Atlantic Ocean (see Fig. 3.10), with higher SSTs also over the WHWP. A negative index, on the other hand, is associated with colder SSTs (Enfield et al. 2001). These features highlight the impact of SST anomalies in the WHWP, relevant for the mean state over the EPAC and the associated impact on EWs. Thus, we can hypothesize that warm SSTs in the WHWP are expected during the negative phase of PDO and the positive phase of AMO. To investigate this hypothesis, we explore the time series analysis of the dynamical and convective metrics associated with synoptic-scale activity, followed by a composite analysis of thermodynamical and dynamical fields.

a. Time series analysis of convective and dynamical metrics

Figure 3.11 shows the time series of the PDO and AMO in colors (positive in red and negative in blue) compared against the convective and dynamical metrics (in gray, indicated in the upper left side) previously obtained. Each time series shows the interdecadal indices averaged during JJAS of each year. We discuss first the PDO, followed by the AMO.

Figure 3.11.a shows the relationship between the PDO index and 2-10–day filtered curvature vorticity at 600 hPa. We observe that the during the period 1980 – 2015, the PDO signal is mostly positive in the first half of the period and then mostly negative in the second half. When the PDO is positive, we have a weak synoptic variance. The opposite happens when the PDO index is negative, where we see generally stronger synoptic variance. A correlation coefficient of $r=-0.216$ (significant at 95%) was obtained, suggesting that synoptic-scale activity associated with
EWs in the EPAC is generally favored by a negative PDO phase and weakened by a positive PDO phase.

Figure 3.11.b shows the timeseries of PDO against the TD-filtered OLR variance anomaly. In accordance with the dynamical metric, the convective metric shows an inverse relationship with PDO. A correlation coefficient of \( r=-0.412 \) (significant at 95%) suggests that a positive PDO does not favor EWs over the EPAC, whereas a negative PDO index does. This result can be interpreted in terms of the warm SSTs over the WHWP enhancing the convection associated with EWs over the EPAC, whereas cold SSTs inhibit them.

Regarding the AMO index, Fig. 3.11.c shows the relationship between the AMO and the dynamical metric represented by 2-10-day filtered curvature vorticity. An evident distinction between positive and negative periods of AMO is evident, followed closely by the dynamical metric. Enhanced synoptic-scale activity is present with a positive AMO index, and this is reduced with a negative index. These variables have a correlation coefficient \( r=0.494 \) (significant at 95%), suggesting that a positive AMO index favors EWs over the synoptic-scale activity over the EPAC.

Figure 3.11.d shows the relationship between the AMO and the OLR TD-filtered variance anomalies. Here, a correlation coefficient \( r=0.309 \) (significant at 95%) confirms that a positive AMO will favor EWs over the EPAC, whereas a negative will suppress them. With a positive AMO, associated with warmer SSTs over the Atlantic and the WHWP, this result can be interpreted in a similar way to that of the PDO: Warmer SSTs over the WHWP will enhance convection associated with EWs, whereas cold SSTs will inhibit it.

These results together, show a striking anti-correlated interdecadal relationship modulating SSTs over the WHWP. A negative PDO and a positive AMO will result in an increase
of convective activity associated with EWs over the EPAC, which in turn enhances the associated curvature vorticity. In contrast, a positive PDO with a negative AMO, characterized by negative SST anomalies, reduces convection, the activity of EWs over the EPAC, and their associated curvature vorticity.

Variations in EW activity can be interpreted in the changes in the number storms and/or changes in their intensity. For example, higher variability (when PDO is negative and AMO positive), can be interpreted as an increased number of waves with constant intensity, or a reduced number of waves with higher intensity (or both). Similarly, lower variability (when PDO is positive and AMO is negative) can imply a decreased number of waves with constant intensity or as a reduced number of waves with less intensity (or both). This is a point that will be explored in the next section.

b. Composite analysis of thermodynamical and dynamical variables

Our results have shown a change in SSTs in the WHWP on decadal timescales. These changes are modulated by the constructive (or destructive) interference of AMO and PDO signals. Overall, it seems that warm SSTs lead to more EW activity, however, so far, the mechanism for this is not clear. This could be associated with more convection with warmer SSTs impacting the mean state and setting up conditions for EW growth. This section explores the changes in the mean state through the analysis of thermodynamical and dynamical that support EWs.

Figures 3.12 and 3.13 shows the composite averages of anomalous thermodynamic and dynamic fields for the periods 1980-1997 and 1998-2015 during JJAS. Thermodynamic variables used are SST, precipitable water (PW), and OLR; while dynamic variables included winds, streamfunction, potential vorticity, and 2-10–day variance of curvature vorticity, all at 600 hPa.
A purple line (based on the axis of maximum of 2-10-day filtered curvature vorticity in Fig. 3.1.c) is illustrated on each plot and works as a reference for the region of interest. For each field, significant differences from the 1980-2015 mean state were identified using a bootstrap test, which consisted of selecting a random sample of the variable of interest within the period of interest and averaging it, repeating this 1000 times. Anomalies were considered significant at 95% level in a two-sided test if the long-term climatology was not within the bootstrap interval of confidence.

Thermodynamic fields in Fig. 3.12 show evident differences between these two periods. During the period 1980-1997, with a positive PDO (+PDO) and a negative AMO (-AMO) in the left column, the SSTs in the EPAC region (Fig. 3.12.a), are characterized by cold SST anomalies where EWs develop. The cold SST anomalies reduce the moisture availability within the troposphere over the region (Fig. 3.12.b), with negative anomalies in PW closer to continental regions west of Mexico and Central America. Consistent with this, OLR anomalies in Fig. 3.12.c shows weak convection over the WHWP.

In contrast, during the period of 1998-2015, when the mean index signs are given as -PDO and +AMO (right column), Fig. 3.12.d shows positive SST anomalies over the EPAC extending into IAS and the Atlantic. The warmer SST anomalies enhance moisture, as observed in PW in Fig. 3.12.e, west of the Mexican coast and Central America, as well as over the Atlantic. This distribution is followed closely by enhanced convection west of Mexico and Central America, as shown in Fig. 3.12.f

Dynamic variables associated with EWs are now shown in Fig. 3.13. During the period 1980-1997, when interdecadal indexes are given as +PDO and -AMO (left column), the wind field
at 600 hPa in Fig. 3.13.a shows a westerly anomaly of ~1 m s⁻¹ west of Central America, with a trough over the Isthmus of Tehuantepec and a ridge to the west. This distribution weakens the mid-level jet over the EPAC, key for establishing the dynamical conditions associated with the growth of EWs. The total PV distribution at 600 hPa (Fig. 3.13.b) shows negative meridional gradients between 95°W-105°W (see contours), intensifying west of this. The negative meridional gradient between 5°N and 20°N (illustrated by the vertical black line) is -1.017x10⁻⁴ PVU·km⁻¹, which is higher than the mean value during 1980-2015 of -1.176x10⁻⁴ PVU·km⁻¹. These two dynamical conditions in this mean state (weaker mid-level winds, and weaker negative meridional PV gradients) help to explain the distribution of variance of 2-10-day curvature vorticity shown in Fig. 3.13.c, reaching a maximum around 20°N, 110°W and characterized by a mean variance value (within the red box) of 0.165x10⁻¹⁰ s⁻², showing diminished synoptic activity when compared against the variance during the whole period 1980-2015 of 0.193x10⁻¹⁰ s⁻² (see Fig. 3.3.b).

When considering the period -PDO and +AMO during 1998-2015, Fig. 3.13 (right column), the opposite pattern is observed. The wind field in Fig. 3.13.d is characterized by easterly anomalies west of Central America, with a ridge over Yucatan and a trough over the EPAC west of the Mexican coast. This wind anomaly results in a stronger mid-level jet and associated with this, a longer and more negative meridional PV strip, as observed in Fig. 3.14.e. The magnitude of the negative meridional gradient (as in Fig. 3.13.b) is -1.334x10⁻⁴ PVU·km⁻¹, more negative when compared with the mean of value of -1.176x10⁻⁴ PVU·km⁻¹. These dynamical aspects help to explain the increased variability of EWs, as shown in Fig. 3.14.f by the 2-10-day filtered curvature vorticity. This last shows a mean variance (within the red box) of 0.221x10⁻¹⁰ s⁻², larger
than the mean during 1980-2015 of $0.193 \times 10^{-10} \text{ s}^{-2}$. Additionally, EW activity is also increased over the Gulf of Mexico and the Atlantic Ocean, consistent with higher SSTs and convection there, as observed in Fig. 3.12.c and f. Despite the increase in synoptic activity, Fig. 3.13.f shows a weak link with EWs from upstream.

Figure 3.14 shows the tracks of EWs and their intensity (in color) during the different periods (details of the tracking methodology will be described in Chapter 4, section 2.d). Considering a box within 13°N to 16°N and 101°W to 105°W (as illustrated in the figure by the blue box) it was found that during 1980-1997, a total of 121 systems were counted. Red color shows the more intense systems, with most of them west of 100°W. When considering the period 1980-2015, a slightly increase in number of systems was found, with 137 in total. For this period most of the systems present an intensification west of 90°W. Thus, these results show that changes in variability of EWs over the EPAC are observed mostly on intensity with a slight difference in number. In addition, over the Atlantic Ocean, it is observed an increase in number of EWs. This is in agreement with the results of Goldenberg et al. (2001), linking the increase of tropical cyclones in association with a warmer SST during the positive phase of the +AMO. Overall, these results also follow those presented in Figs. 3.12-13.

3.6. Summary and conclusions

This chapter has provided insight into the characteristics of EWs over the EPAC mainly through the statistical analysis of EW activity and an analysis of its variability. From this, it was possible to identify their presence over the EPAC-IAS region, the role that they play in the modulation of convection, as well as their variability.
From the statistical analysis, it was found that over the EPAC, the EW storm track is oriented parallel to the coast, peaking to the west of Mexico. Our diagnostics suggest an origin close to Central America and a weak link with systems over the Atlantic. In particular, there is little indication that EWs could have an origin in the Caribbean and tracking into the EPAC. This provides evidence that most EWs over the EPAC are generated “in situ”. While Frank (1972) hypothesized that tropical storms over the Atlantic had a prominent role on the EPAC, this seems to be inaccurate, however, his work was based only on satellite imagery and a few systems crossing Central America. The EW track over the EPAC is in agreement with recent previous literature (e.g. Serra et al. 2010; Rydbeck and Maloney 2014), and it also captures those findings from Aiyyer and Molinari (2008) and Rydbeck and Maloney (2014) associated with a most zonal track over the ITCZ west of the Panama bight.

Spectral analysis showed that dynamical and convective metrics present significant spectral peaks in the 4-6 days range. In particular, the convective metric (T_B) showed spectral peaks over two regions: the first characterized by peaks in the 4-5-days regime between the Panama bight and the Gulf of Tehuantepec, and the second between 4-10-days after the Isthmus of Tehuantepec heading northwestwards. This result can be interpreted in terms of the life cycle of EWs over the EPAC characterized by genesis, weakening and reinvigoration. Also, this result can be interpreted as two different tracks of EWs. These results serve as a basis to further explore this based on a tracking approach (in chapter 4).

Considering the modulation of convection over the EPAC, synoptic-scale systems contribute at least to 40% of the total convective variance, while westward-propagating synoptic-scale systems (EWs), were found to contribute 20% of the total variance in convective and
dynamical metrics. These results are in agreement with those of Gu and Zhang (2002). A summary of results on dynamic and convective metrics is shown in Table 3.1. Additionally, from spectral analysis, it was found that convection over the EPAC is also associated with intraseasonal timescales (peaking around 30 days) to the west of the Mexican coast. This in agreement with previous research on the presence of MJO over the EPAC (Maloney and Hartmann 2000b; Rydbeck and Maloney 2014).

Convective and dynamical metrics associated with EWs, suggest a strong relationship between them. Statistically, this is supported by a correlation coefficient $r=0.304$ (significant at 95%) which is considered strong in similar studies (Mekonnen et al. 2006). Moreover, when exploring their variability, a change in activity was also identified, shifting from reduced synoptic-scale activity in the period 1980-1997, to increased activity during 1998-2015.

Changes in EW activity on interdecadal timescales appear to be associated with SST anomalies modulated by low-frequency oscillations, such as the PDO and the AMO. It was found that EW activity over the EPAC is increased when there is a -PDO index and a +AMO index, as occurred during the period of 1998-2015. The opposite occurred during the years 1980-1997, when a +PDO and a -AMO index occurred. The different correlations obtained between dynamical and convective metrics with the oscillation indexes showed a striking interdecadal relationship.

The physical mechanism for this is that during -PDO and +AMO (1998-2015), positive SST anomalies over the WHWP enhanced convection over the region and the associated curvature vorticity. A detailed analysis showed that while thermodynamic fields were characterized by higher SSTs, with increased moisture and convection, the dynamic fields were characterized by
an enhanced mid-level jet over the EPAC, larger PV meridional gradients and evidently, increased synoptic-scale variability in the synoptic-scale activity, particularly on intensity, as reflected by EW track analysis.

The opposite occurred during a +PDO and a -AMO (1980-1997), when this period was characterized by cold SSTs anomalies over the WHWP, diminished moisture and convection. The dynamic fields showed a diminished mid-level jet, a westward shift of a weakened meridional PV gradients and less intense synoptic-scale EWs, as the tracking showed.

As previously stated, by identifying that SSTs drive the variability of synoptic-scale activity, there exists the potential for predictability of EW activity on long timescales. These results provide a first step to understand the long-term variability of EWs over the EPAC and it can be helpful to discern changes on longer timescales, for example, those associated with climate change.
Table 3.1. Summary of explained variance upon metric and comparison with previous results.

<table>
<thead>
<tr>
<th>Metric</th>
<th>OLR</th>
<th>Curv. vort.</th>
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<tbody>
<tr>
<td>2-10–day filtered</td>
<td>40-60%</td>
<td>45-50%</td>
</tr>
<tr>
<td>TD-filtered</td>
<td><strong>20-25%</strong></td>
<td><strong>25-35%</strong></td>
</tr>
<tr>
<td>Gu and Zhang</td>
<td>15-20%</td>
<td>-</td>
</tr>
<tr>
<td>(2002)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>-</td>
<td></td>
<td>Not reported.</td>
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</tbody>
</table>
**Figure 3.1.** Horizontal variance of dynamical and convective metrics showing regions where they overlap and target regions.  

- **a.** 600 hPa 2-10-day filtered curvature vorticity variance ($10^{-11}$ s$^{-2}$).  
- **b.** 2-10-day filtered OLR (W$^2$ m$^{-4}$).  
- **c.** Overlapping signals showing regions in common. Purple line along the curvature vorticity maxima, illustrating the link with the region west of Central America.  
- **d.** Regions considered to perform spectral analysis as result c.  

Data obtained from ERA-I at 0.5°, and from NCDC-OLR for JJAS during 1980-2015.
Figure 3.2. Power spectra of curvature vorticity at 600 hPa (left) and brightness temperature (right) for the regions over the EPAC, as in Fig. 3.1.d for June to September during the years 1980-2015 and 1984-2004, respectively. Significant power (greater than or equal to the red noise) is in green. Vertical purple lines delineate the power for the periods of 2-10–days. Power on the ordinate (in s$^2$ and K$^2$). Frequencies on the horizontal axis are in cycles per day, and days, as can be observed in the upper horizontal axis. $T_B$ data obtained from CLAUS dataset and curvature vorticity from ERAI at 0.5°.
Figure 3.3. Daily climatology of a. the total variance of curvature vorticity (s$^{-1}$), b. the 2-10–day filtered curvature vorticity variance (s$^{-1}$), and c. Percentage ratio of the 2-10–day filtered curvature vorticity variance to the total curvature vorticity variance (in %), for JJAS during 1980–2015. The red box illustrates the region with maximum 2-10–day filtered variance, used to calculate a mean variance within the region. Data calculated from ERA-I at 0.5°.
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Chapter 4. Easterly Wave Characteristics over the tropical Eastern Pacific and Intra-Americas Sea region

4.1. Introduction

Easterly Waves (EWs) are westward propagating systems found over tropical regions, where they are often associated with deep convection over the ITCZ (Tai and Ogura 1987; Gu and Zhang 2002). Such waves can also serve as seeds for tropical cyclones (TCs) over the Atlantic (Kiladis et al. 2006), as well as over the Pacific (Avila and Guiney 2000; Avila et al. 2003; Pasch et al. 2009). Aside from their convective and synoptic relevance, it is necessary to study their wave characteristics - as well as their structure - to improve the understanding of the processes leading to intensification and possible evolution into TCs. With only a few studies on this topic over the Eastern Pacific (EPAC), a region where the observations are limited, the goal of this chapter is to retrieve the fundamental features of EWs over this basin from a statistical and an objective tracking approach (Brammer et al. 2018). The results of this chapter will build on those previously obtained in Chapter 3, regarding regions where EWs are present over the EPAC.

Knowledge of the fundamental features of EWs are key for helping us understand their life cycle, growth mechanisms, and genesis (a topic that will be discussed later in Chapter 6). Over the tropical EPAC and the Intra-Americas Seas (IAS), it has been found that EWs are generally characterized by a phase speed ($c$) of 5-11 ms$^{-1}$, a period ($T$) of 3-5 days, and a wavelength ($\lambda$) of 2200 km on average (Tai and Ogura 1987; Petersen et al. 2003; Kerns et al. 2008; Serra et al. 2008; Serra et al. 2010). They tend to be characterized by a tilt against the horizontal wind shear, being oriented from northeast to southwest over the EPAC (Serra et al. 2008; Rydbeck and Maloney 2014; Rydbeck and Maloney 2015), suggesting barotropic growth. However, in the
vertical, some differences arise between observations showing a vertical structure (Raymond et al. 1998; Petersen et al. 2003), against reanalysis, which show a tilt in the vertical (Serra et al. 2008; Serra et al. 2010). The vertical structure is also important because it indicates the growing mechanism, this associated with baroclinic development. Although many of their basic characteristics have been identified previously not much has been explored regarding their life cycle, which could possibly be ending up as tropical cyclones (TCs). Therefore, to improve our understanding of EWs over the EPAC this chapter revisits and retrieves the fundamental features of EWs over the EPAC and how they evolve in space and time.

Different methodologies have been used in the past to study EW structures. Overall, these can be grouped into two groups: those using kinematic indices, composites and regressions (or Eulerian) - and those using a tracking methodology, consisting of following a specific feature within the fluid (or Lagrangian). Under the Eulerian framework, the study of EWs has offered a clear and smooth synoptic view of their characteristics, as several studies have shown over Africa (e.g. Reed and Recker 1971; Burpee 1972; Kiladis et al. 2006; Mekonnen et al. 2008; Ventrice and Thorncroft 2013), as well as over the EPAC (e.g. Serra et al. 2008; Serra et al. 2010; Rydbeck and Maloney 2014).

Following-storm (or Lagrangian) approaches have also been used in the study of EWs. Over Africa, Thorncroft and Hodges (2001) used it to have a better perspective of AEWs statistics and basic information of the storm track. More recently, Dunkerton et al. (2009) - through the “pouch” paradigm - highlighted EWs characteristics in the co-moving frame of reference and their link with TC genesis. In this sense, tracking methodologies have also given additional insight into EWs regarding structure and thermodynamical characteristics (e.g. Kerns et al. 2008), as well as
their association with TC genesis (Brammer and Thorncroft 2015). Therefore, this approach could potentially be useful to have a complementary and deeper understanding in the analysis of EWs over the EPAC-IAS region.

A summary of the methods used in the analysis of EWs over the EPAC and some of their fundamental features is presented in Tables 4.1 and 4.2 based upon metric used for reference. We note that only 4 of these works use a tracking methodology and their results are consistent in finding a dense region of EWs over the EPAC. However, they present mixed results about their origin, and only one (Shelton 2011) provided results regarding their structure, with all others leaving aside the basic features of EWs.

The aim of this chapter is to retrieve fundamental features of EWs over the EPAC-IAS region using both regression analysis and a complementary tracking methodology. This will allow us to increase our understanding of the structure, evolution and possible intensification into TCs over the EPAC. This will also involve retrieving information about their genesis within the region. For these goals, this chapter is structured as follows: The data and the tracking methodology used in this work is presented in section 4.2. Then, section 4.3 presents an overview of the storm tracks and statistics based on the tracking methodology. This is followed by the analysis of EWs over the EPAC-track, initiating over the region of Central America in section 4.4 and their subsequent intensification over the warm waters of the EPAC in section 4.5. Finally, a summary of results is presented in section 4.6.
4.2. Data and methodology

a. Key Datasets
   ERA-Interim (Dee et al. 2011) was used in this chapter to obtain dynamical and thermodynamical fields for different analysis. The horizontal grid spacing of the ERA-Interim used is 0.5°x0.5° at 28 vertical levels, with a temporal resolution of 6 hours. As in previous chapters, calculations were performed during the extended summer season JJAS from 1980-2015.

   Convective fields were retrieved from NOAA-OLR spatial coverage of 1.0° by 1.0° latitude-longitude global grid (Liebmann and Smith 1996) on a daily basis. This dataset was interpolated also to a 0.5° grid spacing to be consistent with reanalysis (and results of Chapter 3).

b. Filtering Methodology
   The associated EW-cloudiness locations over the EPAC-IAS were isolated through space-time filtering of OLR dataset following Wheeler and Kiladis (1999), using the spectral peaks in OLR, corresponding to EWs referred to as tropical disturbance band (TD-band). The space-time filtered region includes 2-10-day fluctuations in westward-moving zonal wavenumbers 6-20.

c. Regression Analyses
   Winds, temperature, and streamfunction from ERA-I were linearly regressed onto the TD-filtered OLR time series at selected base points to obtain a statistical representation of wave structure. The linear regressions were computed only for those intervals when OLR is chosen to be below -1.5 standard deviations of the windowed and filtered OLR, yielding magnitudes that are typical of individual wave events based on case studies (as in Wheeler et al. 2000). Statistical significance of the local linear relationship in wind was retrieved using a bootstrap test (Efron and Gong 1983).
d. Tracking

The objective tracking approach used here was similar to Brammer et al. (2018), based on Marchok (2002). This methodology consists of analyzing mass-weighted centers on three levels (600, 700 and 850 hPa). While Brammer et al. (2018) used multiple fields, here we only tracked positive curvature vorticity. This methodology finds a single center from the average of these centers that fall within 300-km tolerance from an estimated center. Then this leads to find a smooth and continuous track for each EW trough to capture well-defined tropical disturbances.

Figure 4.1 shows an example of the tracks during the year 2015 to illustrate this methodology. First, a single track is shown in Fig. 4.1.a of Hurricane Hilary (in gray). Overall, the track of the tropical cyclone (in blue) is well represented by the tracking methodology employed, showing a nice agreement with observations. It is also important to highlight that the tracking algorithm goes further east than the initial point indicated by the National Hurricane Center (NHC) showing the association of this TC with the precursor EW over the region. Figure 4.1.b also shows the tracks obtained for the JJAS season in 2015, including some from the Atlantic – which are similar to observed tracks from previous works over the region (e.g. Kerns et al. 2008). It is important to mention that the results from this methodology offer a complimentary view to results using only statistical analysis (e.g. Serra et al. 2008; Serra et al. 2010).

4.3. Storm track statistics and characteristics

In this section we present the EW activity over the EPAC-IAS region. Figure 4.2 shows EW density, mean intensity, and genesis regions. These statistics are similar to those in previous works (e.g. Thorncroft and Hodges 2001; Kerns et al. 2008; Brammer and Thorncroft 2015; Belanger et al. 2016; Brammer et al. 2018), and they will serve as a basis to understand where EW activity originates, tracks, and where they intensify the most, all this based on a dynamical
approach. EWs considered here were required to exceed a vorticity threshold of $0.1 \times 10^{-5} \text{ s}^{-1}$, and last longer than 4 days.

a. Statistics

The track density in Fig. 4.2.a shows two track density maxima in the region: the largest over the EPAC, and a much shorter region over the southwestern Caribbean. The track pattern over the EPAC starts narrow and dense west of Central America, widening to the west while getting less dense. This feature suggests the spread of EW tracks as they propagate over the EPAC, recurving at different longitudes. The region north of the Panama bight, though shorter, is denser than that over the EPAC. This peak is linked to weak storm systems originating downstream of the topography over northern Colombia (see Chapter 2). Also, this maximum could be associated with persisting convection over the Panama bight, and with EWs genesis over this region, as some previous studies have suggested (Serra et al. 2010; Rydbeck et al. 2017). In addition to these two maxima, low density values are observed over the Atlantic, which are associated with AEWs to the west of the lesser Antilles. The shape of the track density is similar to previous results seen in Kerns et al. (2008), Brammer and Thorncroft (2015), and Belanger et al. (2016).

Figure 4.2.b shows the mean intensity of the EWs. Over the EPAC, the spatial distribution shows a maximum west of the Mexican coast, highlighting intensification regions of EWs, and possible transition into tropical cyclones. Over the Atlantic, we observe high intensities in association with TCs over the Atlantic, with an evident recurving pattern. In opposition to these high intensities, a notable region with low mean intensities is found over the Caribbean and to
the west of Central America. When considering Fig. 4.2.a we observe that these last regions are characterized by a large number of waves with very weak intensities.

Lastly, Fig. 4.2.c shows genesis regions as obtained from the tracking methodology. The distribution shows that genesis can happen over a broad region of the EPAC. However, preferential regions can be identified around 15°N, 100°W off the coast of Mexico, west of Central America at 10°N, 87°W; and to the north of the Panama bight, around 10°N, 80°W. If EWs can also be generated over this last region, it is a hypothesis that will be further explored in Chapter 6.

b. Tracks

This section explores the tracks of the EWs over the EPAC over two regions. The first one is where EWs are expected to be generated, over the EPAC west of Central America, (see Fig. 4.2.c). This sector will be called the upstream region. The second one is where they are more intense over the EPAC, off the coast of Mexico (as in Fig. 4.2.b), and it will be referred to as the mature EW region.

Figure 4.3 shows all the EW tracks (in light blue) passing through the mentioned regions limited by boxes, as well as the genesis points for each of these tracks (illustrated by the darker blue dots). As a reminder, a genesis point is defined as the first occurrence of an EW over the thresholds and constraints defined in the methodology. To simplify the analysis, the storms recurving into the continent were not considered to have a more coherent signal when compositing (in later sections).

The upstream region is illustrated by the red box in Fig. 4.3.a, west of Central America, and defined by the region bounded by the box from 9°N-13°N, 87°W-92°W. It is observed that
EW-systems have origin both to the north of the Panama bight as well as west of Central America. From the tracking analysis, 336 cases were retrieved, from which 145 cases (43%) had an origin over the EPAC, while the rest 191 cases (57%) were generated north of the Panama bight and eastwards. Though most of the genesis points are generated north of the Panama bight, another important number of them is generated to the west of Central America, in agreement with Fig. 4.2.c and possibly associated with observations of Frank (1972). Based on their genesis, we considered this region first to explore the initial stages of the life cycle of EWs over the EPAC.

The mature EW region is illustrated in Fig. 4.3.b and defined as the sector limited by the box within 13°N-16°N, 101°W-105°W - where EWs are more intense. With similar characteristics to the previous region (regarding tracks and genesis points), a total of 258 cases were obtained. As can be observed, the distribution of genesis points is spread all over the tropical EPAC, most of them west of Central America (199 cases or 77%), and with the rest of them (59 cases or 23%) to the east of the Panama bight and the Caribbean Sea. This result reinforces the idea that EWs over the EPAC can have a genesis over the region.

From this analysis we proceed to study the evolution and characteristics of EWs over the EPAC track first exploring the upstream region followed by the mature EW region.

4.4. Upstream region

The previous results of the storm tracks suggest that EWs originate north of the Panama bight and west of Central America around 10°N. We hypothesize that EWs can have an origin or *genesis* associated with the convection found over the Panama bight (see Chapter 2 - Fig. 2.2.e), in an analogous way to that proposed by Thorncroft et al. (2008). In particular, over the EPAC,
this hypothesis has also been raised by Serra et al. (2010) and Rydbeck et al. (2017). Therefore, we choose to explore the nature of EWs over this “upstream region”.

We first begin by analyzing EWs from the regression approach to have a large-scale synoptic view of these. This is followed by an analysis based on the tracking approach to have a complementary analysis about their evolution.

a. Regression analysis

The composited structure over the Panama bight is shown in Fig. 4.4 at the regression point at 11°N, 89.5°W. The horizontal composite shows OLR (shading), and 600 hPa - streamfunction (contours) and regressed winds, while the vertical section shows meridional wind. The point of reference is indicated by a red dot on the horizontal map and with a vertical line in the cross section.

The EW structure depicted in Fig. 4.4.a highlights a synoptic ridge-trough-ridge pattern. The wind field shows a pronounced northwest-to-southeast tilt to the north of the regressed point, and a northeast-to-southwest tilt south of this. By examining the zonal wind shear at the same level in Fig. 4.5, we observe that the streamfunction is tilted against this zonal wind shear, suggesting a barotropic growth configuration, which is in agreement with the existing literature (e.g. Serra et al. 2010). Considering the length from ridge-to-ridge the EWs over this region have a wavelength around 2900 km, with a pattern suggestive of an upstream origin. The strongest convective center is located in the trough, near 11°N, 90°W (as expected). Convection is also spread over land, suggesting that these EWs can be impact the weather over Mexico and Central America.
The vertical structure at 11°N is represented by meridional wind in Fig. 4.4.b, and with the convective signal above it (Fig. 4.4.c). A meridional wind maximum is observed at mid-levels, between 400 – 650 hPa. In particular, this is characterized by a vertical structure tilted with height to the east, with an approximate slope of 100 hPa·deg⁻¹, possibly associated with baroclinic growth (as discussed in Chapter 2). In addition, the vertical structure at this point is characterized with convection within the trough, as Fig. 4.4.c shows.

To study the evolution of EWs at this regressed point, Fig. 4.6 shows the horizontal and vertical evolution of EWs from -6 to +4 days every two days. Similar to the regression at lag 0, this figure shows the EW structure horizontally (left column), and vertically (right column). Again, the point of regression is indicated by a red dot on the horizontal maps, and by a vertical black line in the vertical plots. Overall, this evolution presents an EW structure with a propagation speed of ~5.7 ms⁻¹, a period of 6 days, and a vertical structure reaching the 400 hPa.

In the horizontal (left side of Fig. 4.6), the evolution of the regressed EWs shows the presence of diminished convection at lag-4 (Fig. 4.6.b), followed by incipient convection by day -2 north of the Panama bight (Fig. 4.6.c). After two more days, the structure is characterized by a well-defined EW trough with convection within it. The structure then propagates towards the northwest, characterized by a northeast-southwest horizontal tilt, with convection in phase with the trough (Fig. 4.6.e). This is in agreement with the structure of EWs over the EPAC, as reported by previous research (e.g. Serra et al. 2008; Serra et al. 2010; Rydbeck and Maloney 2014). The convection and its associated latent heat release in the trough persist during the next two days when they start weakening (see Fig. 4.6.f).
The vertical structure of EWs at 11°N (right side of Fig. 4.6) shows the progression of westward-moving trough with maximum intensification at lag 0 (Fig. 4.6.d-right), particularly at mid-levels and below. This is characterized by two meridional wind maxima at 550 hPa and 200 hPa. After this, the vertical structure presents a tilt mostly below 400 hPa (as previously indicated) and above, an out-of-phase westward tilting with height. This last results from the convective outflow aloft. It is also possible to observe that convection is within the trough all times, except at lag 4 (Fig. 4.6.f) when EWs have recurved. This structure is similar to that previously reported by Serra et al. (2010)

As a summary, the previous paragraphs showed that the main features of EWs over the upstream region, west of Central America are:

i. EWs west over the upstream region (west of Central America) are characterized by a propagation speed $c \approx 5.7 \text{ ms}^{-1}$, a wavelength $\lambda \approx 2900 \text{ km}$, and a period $T \approx 6$ days.

ii. Their horizontal structure is characterized by a southwest to northeast tilt against the mean zonal wind shear, suggesting barotropic growth.

iii. Convection is found within the trough.

iv. The vertical structure reaches the 400 hPa and it is characterized by vertical tilt at lower levels associated with baroclinic growth.

v. The evolution of EWs suggest an origin north of the Panama bight, intensifying over the EPAC.
b. Tracking analysis

Figure 4.7 shows the horizontal and vertical composite structure of EWs now based on the tracking methodology. This EW composite consists of all the storms centered within the red box in Fig. 4.3.a, here represented by a black box. The horizontal plot in Fig. 4.7.a shows curvature vorticity (shading) at 600 hPa, while streamfunction is in contours. The green box depicts the region used to average the meridional wind for the vertical structure in Fig. 4.7.b. The shape of this box is based on the tracks presented in Fig. 4.3.a. We can see that these tracks are zonal up to 90°W, and then they move parallel to the coast. This will be used to compare with structures obtained from regressions.

Figure 4.7.a shows the horizontal structure characterized with a wavelength around 1700 km (shorter when compared against the regressed structure). The center of maximum curvature vorticity is located west of Central America within a ridge-trough-ridge structure along the EPAC. The structure is also characterized by a northwest to southeast tilt, consistent with barotropic growth.

The vertical structure, as illustrated by the meridional wind on the right side, shows an intense wave structure at mid-levels but weaker below. This extends vertically reaching the 300 hPa, characterized by weak convection within the trough and no tilt in the vertical.

The evolution of the EWs along the tracks is presented in Fig. 4.8, displayed in the same way that the composite at day 0. Overall, EWs over this region are characterized by propagation speeds $c \approx 3.5 \text{ ms}^{-1}$, a period $T \approx 5$ days, and a vertical height reaching 300 hPa, on average.

In the horizontal, the evolution does not show positive curvature vorticity until lag-2 (Fig. 4.8.c), when a ridge-trough signal is established over the Panama bight. This is characterized by
a northeast to southwest tilt, starting to define a wavy pattern over the EPAC hardly seen 2 days before. Only after this lag, the EW structure propagates keeping its horizontal tilt and reaching the maximum amplitude to the west of Central America at day 0 (Fig. 4.8.d). After this, the EW structure quickly decays (Fig. 4.8.e), diminished by day 4 (Fig. 4.8.f). This quick decay may be the result of the interference of all the points used for the composite.

The vertical evolution of the meridional wind shows a mid-level trough development with a weak convective signal two days before the target region (Fig. 4.8.c-right). This structure reaches its maximum vertical extension at day 0 (Fig. 4.8.d), from surface to 300 hPa and mostly vertical, but decaying rapidly in association with the weak convection, this in pace with the horizontal structure.

The tracking methodology allowed consideration of snapshots at certain times during the evolution of EWs over the EPAC. However, the tracking approach allows us to retrieve key variables following the center of the curvature vorticity every 6 hours. Therefore, we can see more clearly the evolution of their structure along the track, including growth and decay. For this end, variables such as relative vorticity, vertical velocity omega, specific humidity, temperature, diabatic heating Q1, and potential vorticity tendency (PVT) from Q1, were averaged over a radius of 500 km around each trough center and plotted in a longitude-pressure diagram, which can also be interpreted as an time-pressure diagram. This will help to understand the vertical evolution of EWs over the EPAC.

The vertical evolution of variables associated with EWs over the EPAC is presented in Fig. 4.9, together with a reference map (above). A gray dashed line is used as a reference for the easternmost side of the red box (region through which all EWs passed), in this case at 87°W. For
the vertical evolution, all tracks were put into a single track in space with a reference point at 87°W, this location being 0° in the vertical maps; +5° represents then five degrees in the evolution of the individual track, and so on.

Beginning with the dynamic variables, relative vorticity anomalies in Fig. 4.9.a show an intensification at mid-levels (550 hPa - 600 hPa) starting at -7.5° (this is around 79.5°W), being more evident at 0°. This evolution intensifies later from mid- to low-levels exceeding 8.0x10^-5 s^-1 after +17°, this is about 104°W. The vertical extent reaches 200 hPa and its evolution suggests a sudden and vigorous intensification at this point, which persists for the next 3° weakening after this point.

To diagnose the potential convective role, Fig. 4.9.b shows vertical velocity (omega). Initially at 0°, increased vertical motion is associated with the positive anomalous vorticity at mid-levels, showing larger values at +7° (94°W). At this location, the signal extends from low levels to 200 hPa right over the mature EW region.

The moisture field in Fig. 4.9.c shows initially a positive anomaly around 500-700 hPa (at -3°), moistening throughout the lower levels over oceanic regions and exceeding 2.0 g kg^-1 after +7°W (94°W) and reaching the 300 hPa. This evolution suggests moistening of the atmospheric column within the EW, with an increased moisture at low levels associated with rainfall (Serra et al. 2010).

Thermal anomalies in Fig. 4.9.d show a signal dominated by warming at mid- to upper-levels (above 500 hPa and up to 200 hPa), increasing in magnitude at +10° (97°W). Cooling at low levels is also evident, this associated with rainfall in the boundary layer. This result is also associated with a peak in vorticity at mid-levels and it is in agreement with those of Serra et al.
(2010) regarding the classic Riehl structure (Riehl 1979), showing warming above and cooling below mid-levels, respectively.

The above results about enhanced cooling below 700 hPa and warming above 500 hPa, led us to explore the evolution of the apparent heat source (Q1), shown in Fig. 4.9.e. A starting signal is observed at -5° (82°W), later exceeding 7 K·day⁻¹ in the mid-to-upper troposphere after 0°, likely associated with stratiform heating (as seen in Chapter 2). In particular, a maximum is observed at +7°, from 850 hPa to 300 hPa, suggesting the presence of deep convection and the associated latent heat in the growth of EWs. This can be explained by considering that in regions of high precipitation, latent heat is the most important term in the apparent heat source Q1 (Janiga and Thorncroft 2013). The distribution of Q1 highlights the importance that diabatic heating (specifically stratiform) has on the evolution and intensification of EWs, necessary for their invigoration over the EPAC.

It can be expected that the distribution of Q1 will impact in the creation of PV. To explore this, Fig. 4.9.f shows the evolution of potential vorticity tendency (PVT). It is observed that PVT is characterized by a maximum at 600 hPa, which is a key level for their life-cycle (as discussed in Chapter 2). This PVT anomaly starts developing at -7°, exceeding the 0.12 PVU after 0°.

Overall, the vertical evolution of EWs exemplifies the life cycle of EWs over the EPAC from the upstream region. This goes from a nascent vortex at mid-levels around 79.5°W, that intensifies over 104°W. It is over this region that several factors coincide: a maximum mid-level jet, warmer SSTs, and the presence of latent heat release from deep convection (as seen in
Chapter 2). After crossing this region, EWs seem to decay around 107°W. Next we explore briefly the possible link with TCs from this upstream region.

c. **Link to tropical cyclones**

The intensification of EWs at low-levels can lead to favorable seedlings for TC genesis, as observed in other basins. After the structural changes observed, here we just provide an estimate of relationship of EWs with TCs over the EPAC. This is simple a ratio based on the number of genesis points retrieved from the tracking methodology in the upstream region, against TCs retrieved from the IBTrACS dataset (Knapp et al. 2010) within the same boxed region. When considering the relationship between EWs and TCs over the upstream region west of Central America, TCs from the IBTrACS dataset showed only 8 cases, this representing only a 2% of the cases (see Fig. 4.10). This result shows only that stronger EWs are necessary to develop into TCs, and that may be achieved at the mature EW region, which will be explored next. The relationship between EWs and TCs can be further explored by considering all EWs passing through this box and developing into TCs, and studying those that develop from those that not develop. However, this is a topic left for future research.

As a summary, the tracking of EWs over the upstream region shows:

i. EWs over the EPAC (from the upstream region) are characterized by a propagation speed $c \approx 3.5 \text{ ms}^{-1}$, a wavelength of 1700 km, a period $T = 5$ days, and a vertical extension reaching the 300 hPa on average. Their horizontal suggests barotropic growth.
ii. EWs over the upstream region are characterized by a mid-level vortex on the initial stage and developing to lower levels. This characteristic is similar to that over west Africa over land.

iii. The transition to TCs over from this location is marginal, of only 2% of them transitioning.

4.5. Mature EW region

In this section we present an analysis of the evolution and growth of EWs over the mature EW region. As in the upstream region, we first present the analysis of EWs from the regression approach followed by the analysis based on the tracking approach for the complementary analysis about their evolution.

a. Regression analysis

Figure 4.11 presents the structure of EWs over the EPAC with a base point at 14°N, 99°W, as well as their coupling with convection, based on a regression approach. Again, the horizontal composite shows OLR (shading), and 600 hPa - streamfunction (contours) and regressed winds, while the vertical section shows meridional wind. The point of reference is indicated by a red dot on the horizontal map and with a vertical line in the cross section.

The EW structure depicted in Fig. 4.11.a highlights a synoptic ridge-trough-ridge pattern. The wind field shows a pronounced northwest-to-southeast tilt to the north of the regressed point, and a northeast-to-southwest tilt south of this. By examining the zonal wind shear at the same level in Fig. 4.12, we observe that the streamfunction is tilted against this zonal wind shear, suggesting a barotropic growth configuration. This is also in agreement with the existing literature (Serra et al. 2008; Rydbeck and Maloney 2014). Considering the length from ridge-to-
ridge (in Fig. 4.11) the EWs over this region have a wavelength around 2900 km, with a pattern suggestive of an upstream origin. The strongest convective center is located in the trough, near 15°N, 100°W (as expected). In this case, convection is also spread over land, suggesting that these EWs can be impact the weather over Mexico.

The vertical structure of EWs at 14°N is shown in Fig. 4.11.b with the convective signal included above it (Fig. 4.11.c). Overall, the vertical structure is characterized by two wind maxima; one at 600 hPa and another at 200 hPa. In the lower troposphere the EW structure does not show a significant tilt in the vertical, reaching as high as 300 hPa. Above this, there is an out-of-phase westward tilting with height, resulting from the convective outflow aloft. The vertical structure also shows convection aligned with the tropospheric trough, as observed in Fig. 4.11.c. This structure is similar to that found in EWs at 10°N over Africa (Kiladis et al. 2006).

The daily evolution of EWs (based on the regression analysis from lag -6 to lag +4) is presented in Fig. 4.13. Similar to the previous figure, this shows the EW structure horizontally (left column), and vertically (right column). Again, the point of regression is indicated by a red dot on the horizontal maps, and by a vertical black line in the vertical plots.

Overall, the EWs propagate westwards with a phase speed of about 5 ms⁻¹ and with a period of about 6 days, consistent with results in Chapter 3. The evolution shows that the trough at day 0 appears to have its origins further east at around 80°W four days before (Fig. 4.13.b). During its passage westwards, the trough is convectively active and the ridge convectively suppressed (Fig.4.13.c-f).

Winds exhibit a northwest-to-southeast tilt to the north of the circulation and a northeast-to-southwest tilt to the south. This pattern continues until lag+4 (Fig. 4.13.f) when the
cyclonic center reaches about 120°W in longitude with a weakening convection signal, and recures parallel to the peninsula of Baja California.

The EWs exhibit large scales of about 30° in longitude and 25° in latitude, with cross-equatorial flow to the south (Fig. 4.13.e-f). An interesting aspect to highlight in this evolution is that EWs initially move zonally and later their orientation shifts NW-SE. This is a key aspect that has to be considered in any tracking and modeling study.

In the vertical, the evolution shows a trough intensification from lag-4 to lag+2, mainly under 300 hPa, with outflow above from convection tilted westward. This vertical structure is aligned with convection, as OLR anomalies are in phase with the trough.

This analysis of the structure and evolution of EWs on the mature EW region can be summarized as follows:

i. EWs over the EPAC are characterized by a propagation speed of about 5 ms⁻¹, a wavelength of about 2900 km, and a period of about 6 days. These results agree with those in previous literature (Serra et al. 2008; Rydbeck and Maloney 2014).

ii. Their horizontal structure shows tilting against the zonal wind shear, indicative of barotropic growth.

iii. Convection is found within the trough.

iv. The vertical structure reaches the 300 hPa, consistent with the latent heat release sustaining their growth given the phasing, with a westward tilt above it. This associated with the output flow from convection.
v. The evolution of EWs at this point suggests a frequent origin over the western Caribbean, moving into the EPAC and intensifying in this basin. This evolution is consistent with the variance results presented in Chapter 3.

b. Tracking analysis

Figure 4.14 shows the horizontal and vertical structure of EWs with the same variables as those used in the previous composite in Fig. 4.7, now over the EW mature region, as illustrated by the black box.

In the horizontal, Fig. 4.14.a shows that the EW compositied structure is characterized by a maximum of curvature vorticity within the region of interest, with a clear EW track tilted over the EPAC. This extends down to the west of Central America and is characterized by a wavelength of ~2000 km.

The vertical structure in Fig. 4.14.b, represented by the meridional wind, shows an upright vertical structure that reaches 300 hPa, with a westward tilt with height above it. Convection within the trough, highlights a strong role of deep convection in the vertical structure. Overall, the horizontal and vertical structure is similar to that retrieved from the regression approach (see Fig. 4.11). Differences on scale and tilt result from the zonal decomposition from the space-time filtering and the zonal decomposition.

The evolution of the EWs along the tracks is presented in Fig. 4.15, displayed in the same way that the composite at day 0. Together, the horizontal and vertical evolution of EWs show that they are characterized by a propagation speed $c \approx 3 \text{ ms}^{-1}$, a period $T=6$ days, and a vertical extension reaching the 300 hPa on average, with similar structures to those in the regressions.
The evolution in the horizontal shows that the positive curvature vorticity associated with the main EW at lag 0 can be identified 4 days before (Fig. 4.15.b). This is characterized by an elongated maximum of positive curvature vorticity tilted northeast to southwest. The trough intensifies as it moves westwards, showing a maximum at day 0 with a clear trough-ridge-trough structure, and inducing a new positive curvature vorticity maximum to the west of Central America. This remains intense for 2 more days (Fig. 4.15.e), and then it recurves west (Fig. 4.15.f) north of 20°N. In contrast with regressions, the zonal extent is shorter, in the order of 16.5° in longitude and 10° in latitude.

The evolution of the vertical structure, represented by the meridional wind on the right side of Fig. 4.15, shows a vertical intensification between day -4 and day 0 (Fig. 4.15.b-d) mainly between the surface and the level of 300 hPa. This intensification goes in phase with convection (as indicated above the meridional wind), which produces an outflow aloft, characterized by westward tilt. Convection within the trough highlights the role of the latent heat release for their intensification. Convection diminishes considerably by day+2 (Fig. 4.8.e), likely associated with cooler SSTs over the EPAC (see line in blue defining the warm pool as in Chapter 2, Fig. 2.2.b), suggesting an uncoupling with the trough, and a slow weakening of the trough EWs at later times.

To explore more in depth the life cycle of EWs in the vertical, and to be consistent with the previous location (upstream region), the vertical evolution of variables associated with EWs is presented in Fig. 4.16. A reference map is shown in the upper part, where the gray dashed line indicates the easternmost side of the red box west of Central America, in this case, at 101°W.

Relative vorticity anomalies (Fig. 4.16.a) show an initial development at mid-levels at -15°, this is about 86°W. A stronger and clearer signal is observed at -10° (91°W), exceeding the
5.0x10^{-5} \text{s}^{-1}. At the target region (0°) the relative vorticity anomalies have developed from the surface to 500 hPa and a later reaching up to 300 hPa at +8° (109°W) over the warm SSTs of the western hemisphere warm pool (WHWP).

Following this, rising motions by the vertical velocity omega (Fig. 4.16.b) start showing up at mid-levels at -10°, strengthened and present at lower levels at -5° and clearly extended from the surface to 200 hPa at 0°. After this location, the vertical velocity decays within the next 10° (111°W).

The evolution of variables associated with the thermodynamical structure show moisture (Fig. 4.16.c) initially at mid-levels, and extending below at -5°, with temperature anomalies (Fig. 4.16.d) showing warming aloft (between 500 hPa - 200 hPa) and cold below, extending vertically after 0°. The evolution of these variables is similar to that in Fig. 4.9, though more intense.

Diabatic heating Q1 and PVT highlight again the role of convection for the maintenance of EWs. For example, Q1 in Fig. 4.16.e shows an intensification at +3° (104°W) over warm waters of the WHWP, emphasizing an increased availability of latent heat. In agreement with this, PV from Q1 in Fig. 4.16.f shows creation of PV at 600 hPa following a more continuous distribution after -5° and decaying at +10°.

Overall, the evolution of EWs at the mature stage show similar results as that over the upstream region. However, these are more intense (as expected) given the target region is characterized by a supporting mean state (mid-level winds, and SSTs) favors their intensification and confirming the life cycle of EWs over the EPAC. With more intense characteristics, it is expected to have more TCs, topic briefly discussed next.
c. Link to tropical cyclones

The relationship between EWs and TCs over the mature EW region showed 33 cases, which represents about a 13% of the total EWs (see Fig. 4.17). Again, it should be considered that this result is restricted only to a small region where EWs intensify, and to storms not recurving into land. This result just gives a rough estimate about the relationship between EWs and TCs and it is left for future work.

Overall, the results of the structure and evolution of EWs over the EPAC from the tracking approach can be summarized as follows:

i. EWs over the EPAC (in the intensification region) are characterized by a propagation speed $c \approx 3 \text{ ms}^{-1}$, a wavelength of 2000 km, a period $T=6$ days, and a vertical extension reaching the 300 hPa on average. Their horizontal and vertical structure suggests barotropic growth.

ii. The vertical evolution shows the structure of EWs changes from a mid-level vortex to one dominated more by low-level vorticity over the mature EW region. This result suggests that intensification of EWs over the EPAC is due to interactions from two components: the first linked to a mid-level vortex, and the second associated with latent heat release from deep convection over the high SSTs from the WHWP.

iii. The relationship between EWs to TCs over this region is about 13% of the cases.
4.6. Summary

This chapter has presented the basic features of EWs over the EPAC—such as wavelength, propagation speed, period and vertical extent—as well as their structure, evolution, coupling with convection and genesis regions. This was performed using two complementary methodologies, the first based on a linear regression, and the second based on a Lagrangian approach (tracking). The main results can be summarized as follows:

- From the regressions we could retrieve the large-scale features of EWs over the EPAC. These are characterized by a propagation speed of about 5.0 ms\(^{-1}\), a wavelength of about 2900 km, and a period of about 6 days. Their distinctive horizontal tilts against the shear, is suggestive of a barotropic growth mechanism. In the vertical, their structure reaches 300 hPa. At low levels, they are characterized by an eastward tilt of 100 hPa·deg\(^{-1}\) likely associated with baroclinic growth. At upper levels, EWs are characterized with a westward tilt resulting from convective outflow.

- Based on the tracking, we identified an EW storm track over the EPAC oriented parallel to the coast, reaching maximum intensities around 15°N, 100°W (mature region). The NW-SE orientation of the storm track hinted at a genesis region to the west of Central America (upstream region).

- While regressions present EWs coming from the Caribbean, tracking showed EWs coming from Central America. This results mainly from the zonal wavenumber decomposition used in the regressions, whereas tracking follows their dynamic center allowing for a better trough identification.
• The tracking methodology also showed similar EW characteristics regarding structure and evolution. However, differences were found for propagation speed (3.0 ms\(^{-1}\)) and wavelength (of about 2000 km). This is also a consequence of the trough identification as indicated in the previous bullet. Differences in EW characteristics are summarized in Table 4.3.

• The vertical evolution of EWs over the EPAC is characterized by a mid-level vortex at early stages, developing to lower levels in association with deep convection over the WHWP, where they present further intensification at a mature stage, and dissipating later. The evolution considered at two different regions showed common results, this allowing to identify the life cycle of EWs over the EPAC.

• The transition to TCs over the upstream region is marginal, representing only about 2% of the cases within the region west of Central America.

• The previous result changed over the mature EW region suggesting that this EW-TC transition occurs about 13% on average. This ratio could be larger when considering all EWs over the EPAC, for example, considering recurving systems into continental regions and considering not a limited domain. However, this is left as a future work.

• Based on the tracks and genesis regions, 77% of EWs are generated over the EPAC, increasing the confidence to propose a genesis mechanism in situ. This result does not mean that all EWs are generated over the EPAC and some of them could be linked with upstream systems associated with African Easterly Waves, for example.
The next chapter explores how the variability of convection over this region is impacted by convectively coupled Kelvin waves, followed by a more general approach from a model perspective.
### Tables

**Table 4.1.** Tracking methodologies based on relative vorticity.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Data</th>
<th>Period</th>
<th>Region</th>
<th>Method</th>
</tr>
</thead>
<tbody>
<tr>
<td>Serra et al.</td>
<td>ERAI</td>
<td>(JJASON)</td>
<td>5°-25°N, 120°W-60°W</td>
<td>Automated tracking (partial)</td>
</tr>
<tr>
<td>Serra et al.</td>
<td>NCEP-NCAR</td>
<td>(JJASON)</td>
<td>24°N-24°S, 80°W-180°W</td>
<td>-</td>
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<tr>
<td>(2008)</td>
<td>1979-2002</td>
<td>700 hPa</td>
<td></td>
<td></td>
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<tr>
<td>Kerns et al.</td>
<td>ERA40</td>
<td>(JJASO)</td>
<td>0-35°N, 140°W-10°W</td>
<td>Manual analysis</td>
</tr>
<tr>
<td>Petersen</td>
<td>Observation</td>
<td>(SO)</td>
<td>East Pacific</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>EPIC-2001</td>
<td></td>
<td>(10°N, 95°W)</td>
<td></td>
</tr>
<tr>
<td>Tai &amp; Ogura</td>
<td>Observation/Model</td>
<td>(MJJAS)</td>
<td>Eastern Pacific</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>1979</td>
<td></td>
<td></td>
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</table>
Table 4.2. Tracking methodologies based on curvature vorticity.

<table>
<thead>
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<th>Region</th>
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<td>(JIAS)</td>
<td>-10°-35°N, 40°-140°W</td>
<td>Automated tracking</td>
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<tr>
<td>Belanger et al.</td>
<td>Multiple</td>
<td>Multiple</td>
<td>Globe</td>
<td>Automated tracking</td>
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<tr>
<td>(2016)</td>
<td>ERAI</td>
<td>(JASO)</td>
<td>5°-25°N, 110°W-40°W</td>
<td>Automated tracking</td>
</tr>
<tr>
<td>Shelton</td>
<td>ERAI</td>
<td>1979-2009</td>
<td>600 hPa</td>
<td>Automated tracking</td>
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Table 4.3. Differences between EW characteristics from Regression and Lagrangian approaches.

<table>
<thead>
<tr>
<th>Method</th>
<th>Region</th>
<th>$c$ (ms$^{-1}$)</th>
<th>$\lambda$ (km)</th>
<th>$T$ (days)</th>
<th>$h^*$ (hPa)</th>
<th>$k$ ($\theta=15^\circ$N)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Regressions</td>
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<td>2900</td>
<td>6</td>
<td>400</td>
<td>13</td>
</tr>
<tr>
<td></td>
<td>Mature</td>
<td>5.0</td>
<td>2900</td>
<td>6</td>
<td>300</td>
<td>13</td>
</tr>
<tr>
<td>Tracking</td>
<td>Upstream</td>
<td>3.5</td>
<td>1700</td>
<td>4</td>
<td>300</td>
<td>22</td>
</tr>
<tr>
<td></td>
<td>Mature</td>
<td>3.0</td>
<td>2000</td>
<td>6</td>
<td>300</td>
<td>19</td>
</tr>
</tbody>
</table>

Numbers in white represent results from the regression methodology, while numbers in purple represent values from the Lagrangian approach. $h^*$ is for height, not equivalent depth. Wavenumbers are for reference, not used in the tracking methodology.
Figures

Figure 4.1. Example of the tracks obtained over the EPAC – IAS region during year 2005 using the described methodology. **a.** Hurricane Hilary as retrieved form the tracking methodology (in blue) against the best track (in grey) from National Hurricane Center (NHC). Dots indicate best track every 6-hours. **b.** All tracks for year 2005 during JJAS. Data retrieved from ERAI at 0.5°.
Figure 4.2. Statistics associated with EWs from the tracking methodology. 

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b. Mean curvature vorticity intensity over the EPAC - IAS region.

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Figure 4.4. a. Regressions of OLR (shading), 600-hPa – winds, and streamfunction (contours) onto TD-filtered OLR at 11°N, 89.5°W; b. Vertical regressions of meridional wind (in ms⁻¹) along 8°N; and c. OLR (in Wm⁻²) - all of them at lag 0 during JJAS 1980-2015. For a: every other 95% significant wind vector is shown. Streamfunction contours are every 0.1x10⁻⁶ m²s⁻¹; negative contours are dashed. For b: Vertical line indicates the reference at 89.5°W.
Figure 4.5 Mean zonal wind shear (in s^{-1}) and contours of streamfunction (10^{-6} m^2 s^{-1}) of EWs at 600 hPa over 11^\circ N, 89.5^\circ W (indicated by the red dot) at lag 0 from the regression approach. It can be observed the tilted structure of the EW against the zonal wind shear of the mid-level jet. Data retrieved from ERAI at 0.5^\circ.
Figure 4.6. Left. Regressions of OLR (shading), 600hPa-winds and streamfunction (contours), onto TD-filtered OLR at 11°N,89.5°W during JJAS 1980-2015 for a. lag -6, b. lag -4, c. lag -2, d. lag 0, e. lag 2, and f. lag 4. Every other 95% significant wind vector is shown. Streamfunction contours are every 1x10^5 m^2 s^-1; negative contours are dashed. Right. Vertical regressions of meridional wind (in ms^-1) along 8°N and OLR (in Wm^-2) at a. lag -6, b. lag -4, c. lag -2, d. lag 0, e. lag 2, and f. lag 4. Vertical line indicates the reference at 89.5°W.
Figure 4.6. Continued.
Figure 4.7. Horizontal structure of EWs over the EPAC from the Lagrangian approach (EW-tracking). a. EW composite on the region limited by the black box, with curvature vorticity in colors ($10^{-5}$ s$^{-1}$) and streamfunction in contours ($10^{-6}$ m$^2$s$^{-1}$) at 0 hrs. b. Meridional wind composite (ms$^{-1}$) within the green sector depicted on the horizontal, with OLR (Wm$^{-2}$) above it. Results obtained for JJAS 1980-2015 with data from ERAI and NOAA-OLR.
**Figure 4.8. Left:** Horizontal evolution of curvature vorticity \((10^5 \text{s}^{-1})\) from -6 to +2 days (every 2 days) associated with EWs over the tropical Eastern Pacific. The green sector depicted in each figure was used to average the meridional winds. **Right:** Vertical evolution of meridional wind \((\text{ms}^{-1})\) along the green sector in each figure. Black horizontal line indicates the level of 600 hPa, while the vertical line indicates 70°W, the easternmost latitude from the green sector. The vertical dashed line indicates the latitude where the averaged sector tilts, while the vertical bold line indicates the reference region at 87°W. Data from June to September during the years 1980-2015. Data source from ERA-Interim at 0.5°.
Figure 4.8. Continued.
Figure 4.9. Vertical evolution of trough variables associated with the evolution of EWs over the upstream region west of Central America (as illustrated in the map above). This shows anomalies of a. Relative vorticity (10⁻⁵ s⁻¹), b. Omega vertical velocity (Pa s⁻¹), c. Specific humidity (g kg⁻¹), d. Temperature (°C); and total fields for: e. Apparent heat source Q1 (K day⁻¹), and f. Potential vorticity tendency PVT (PVU). The gray dashed line represents the approximate longitude of the easternmost side of the red box area.
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Figure 4.12. Mean zonal wind shear (s⁻¹) and contours of streamfunction (10⁻⁶ m²s⁻¹) of EWs at 600 hPa over 14°N, 99°W (indicated by the red dot) at lag 0 from the regression approach. It can be observed the tilted structure of the EW against the zonal wind shear of the mid-level jet. Data retrieved from ERAI at 0.5°.
Figure 4.13. **Left.** Regressions of OLR (Wm$^{-2}$), 600-hPa winds and streamfunction (contours), onto TD-filtered OLR at 14°N,99°W during JJAS 1980-2015 for a. lag -6, b. lag -4, c. lag -2, d. lag 0, e. lag 2, and f. lag 4. Every other 95% significant wind vector is shown. Streamfunction contours are every $10^5$ m$^2$s$^{-1}$; negative contours are dashed. **Right.** Vertical regressions of meridional wind (in ms$^{-1}$) along 14°N and OLR (in Wm$^{-2}$) at a. lag -6, b. lag -4, c. lag -2, d. lag 0, and e. lag 2, and f. lag 4. Vertical line indicates the reference at 99°W.
Figure 4.13. Continued.
Figure 4.14. Horizontal structure of EWs over the EPAC from the Lagrangian approach (EW-tracking). a. EW composite on the region limited by the black box, with curvature vorticity in colors \(10^5 \text{s}^{-1}\) and streamfunction in contours \(10^6 \text{m}^2\text{s}^{-1}\) at 0 hrs. b. Meridional wind composite (ms\(^{-1}\)) within the green sector depicted on the horizontal, with OLR (Wm\(^{-2}\)) above it. Results obtained for JJAS 1980-2015 with data from ERAI and NOAA-OLR.
Figure 4.15. **Left:** Horizontal evolution of curvature vorticity \((10^{-5} \text{s}^{-1})\) from day -6 to day +4 (every 2 days) associated with EWs over the tropical Eastern Pacific. The green sector depicted in each figure was used to average the meridional winds. **Right:** Vertical evolution of meridional wind \((\text{ms}^{-1})\) along the green sector in each figure. Black horizontal line indicates the level of 600 hPa, while the vertical line indicates 70°W, the easternmost latitude from the green sector. The vertical dashed line indicates the latitude where the averaged sector tilts, while the vertical bold line indicates the reference region at 101°W. Data from June to September during the years 1980-2015. Data source from ERA-Interim at 0.5°.
Figure 4.15. Continued.
Figure 4.16. Vertical evolution of trough variables associated with the evolution of EWs over the mature EW region (as illustrated in the map above). This shows anomalies of a. Relative vorticity \((10^{-5} \text{ s}^{-1})\), b. Omega vertical velocity \((\text{Pa s}^{-1})\), c. Specific humidity \((\text{g kg}^{-1})\), and d. Temperature \((^\circ\text{C})\); and total fields for: e. Apparent heat source \(Q_1\) \((\text{K day}^{-1})\), and f. Potential vorticity tendency \((\text{PVU})\). The gray line represents the approximate longitude of the easternmost side of the red box area.
Figure 4.17. TC genesis and their best tracks over the boxed region used during the same period of time JJAS 1980-2015. Data retrieved from IBTrACS dataset.
Chapter 5. The role of convectively coupled equatorial Kelvin waves on Easterly Wave activity over the tropical Eastern Pacific

5.1. Introduction

Easterly Waves (EWs) are westward propagating disturbances found in the summer season over the tropical band of 15°N-20°N across the world. These are observed in West Africa and the Atlantic (e.g. Kiladis et al. 2006, and references therein), Australia (e.g. Dickinson and Molinari 2000), the west Pacific Ocean (e.g. Serra et al. 2008 and references therein), and over the tropical Eastern Pacific (e.g. Molinari et al. 1997; Serra et al. 2008; Rydbeck and Maloney 2014; Rydbeck and Maloney 2015). The relationship between EWs and eastward-propagating synoptic-scale systems has been studied mostly over Africa (e.g. Diaz and Aiyyer 2013) associated with the Madden-Julian Oscillation (Ventrice et al. 2011), as well as with convectively coupled equatorial Kelvin Waves (Mekonnen et al. 2008; Ventrice et al. 2012; Ventrice and Thorncroft 2013), hereafter called KWs. These results have demonstrated that KWs impact the modulation of convection and can act as triggers of African Easterly Waves (AEWs), particularly over mountain regions. This chapter presents an analysis of the impacts that KWs have on the EW activity over the tropical Eastern Pacific (EPAC).

Over the EPAC, the relationship between westward- and eastward-propagating synoptic-scale systems has focused more on the MJO, given their modulation of mean state convection and tropical cyclones (Maloney and Hartmann 2000a, 2001), as well as on EW activity (e.g. Rydbeck and Maloney 2014). It is still unclear if there exists a relationship between KWs and EWs in this basin. With only one study dealing with the interaction between KWs and TCs (Schreck 2015) linking EWs, there are still gaps in our knowledge regarding how KWs can modify.
convection over the EPAC, the impacts of convection within EWs and importantly, if they can trigger EWs over the EPAC.

Over Africa it has been demonstrated that KWs modulate the activity of AEWs. For example, by analyzing the structure and variability of KWs over Africa, Mekonnen et al. (2008) observed that a strong KW was associated with a series of EW initiations over Africa. This result was further explored by Ventrice and Thorncroft (2013), who found that AEW activity increases during and after the passage of the convective phase of strong KWs. KWs enhance the convective environment favorable for AEWs and the meridional gradients in potential vorticity. It is expected that these mechanisms could also enhance the activity of EWs over the EPAC given the similarities in the mean state.

The presence of KWs over the EPAC has been reported in previous studies from reanalysis (Roundy and Frank 2004) and observations (Straub and Kiladis 2002). Straub and Kiladis (2002) showed the observed characteristics of the passage of a KW over the EPAC during the 1997. From their results, KWs over the EPAC are characterized by a zonal scale of 1000-2000 km and an eastward-propagation of 15 ms⁻¹. The convective envelope consists of westward moving disturbances with zonal scales of 100-500 km. Synoptic features include a dynamical structure nearly symmetric with respect to the equator. However, the convective signal associated with the dynamical perturbation is located north of the equator, in the region of warm sea surface temperatures, and along the climatological intertropical convergence zone (ITCZ), in agreement with several previous studies (Takayabu and Nitta 1993; Wheeler and Kiladis 1999; Wheeler et al. 2000). Of relevance to this study, was the result that KWs enhance convection over Central and South America, also later confirmed by Liebmann et al. (2009). In particular observations
from the KW passage reported that stratiform convective regions become predominant within
the next 24 hours.

A more recent study on the role that KWs have on tropical cyclogenesis globally, was by
Schreck (2015) who included analysis of the EPAC. From his results, Schreck (2015) found that for
the process of tropical cyclogenesis, an interaction between KWs and EWs is necessary, this is,
there must be present a pre-existing EW over the region. His results suggested also that EWs
transitioned to TCs 3.5 days after the KW passage. However, the mechanisms and processes that
resulted in genesis from EWs over the EPAC were not studied.

This chapter is motivated by similar hypotheses of Ventrice and Thorncroft (2013), namely
that KWs can enhance convection over the Panama bight and the mountainous regions of the
Andes, acting as convective triggers for EWs over the EPAC. This will extend the work of Schreck
(2015) on interactions of KWs and TCs, filling the gap on our understanding of KWs over the EPAC.

This chapter is structured as follows. Section 2 discusses the datasets and methodology
used to explore the presence of KWs over the EPAC and their relationship with EWs. Then, section
3 presents an overview of the presence of KWs over the EPAC. This is followed by the analysis of
the relationship between KWs and EWs in section 4, with a summary and conclusions in provided
in Section 5.

5.2. Datasets and Methodology

a. Key Datasets

ERA-Interim (Dee et al. 2011) was used in this chapter to obtain dynamical and
thermodynamical fields for different analysis. The horizontal grid spacing of the ERA-Interim
(ERAI) used is 0.5° x 0.5° at 28 vertical levels, with a temporal resolution of 6 hours. As in previous
chapters, calculations were performed during the extended summer season June to September (JJAS) from 1980-2015.

Convective fields were retrieved from NOAA-OLR spatial coverage of 1.0° by 1.0° latitude-longitude global grid (Liebmann and Smith 1996) on a daily basis. This dataset was interpolated also to a 0.5° grid spacing to be consistent with reanalysis.

To illustrate examples of Kelvin waves, brightness temperature ($T_B$) was used given its higher temporal resolution. $T_B$ data was obtained from the GridSat dataset (Knapp et al. 2011). It consists of a high-resolution data set in time (3-hourly), and space (about 8 km in latitude-longitude grid at the equator). GridSat is a product of multiple satellite observations in the 11 µm infrared window.

To retrieve rainfall, data from the Tropical Rainfall Measurement Mission (TRMM; Huffman et al. 2007) are used from their 3B42 dataset, which uses a combination of precipitation estimates (Adler et al. 1994). This dataset has also been used in previous studies over the region (Kucieńska et al. 2012) showing consistent results over ocean when compared with ground-based rainfall estimates (Wang et al. 2014). This dataset was used for JJAS 1989-2009.

Anomalies for dynamical and thermodynamical variables were constructed by subtracting the long-term mean and the first three harmonics of the seasonal cycle.

b. Filtering

The associated KW-cloudiness over the EPAC-IAS were isolated through space-time filtering of OLR and $T_B$ dataset following Wheeler and Kiladis (1999), using the spectral peaks in OLR, corresponding to the KW-band. The space-time filtered region includes eastward
propagating wavenumbers 1-14, periods of 2.5 to 20 days, and dispersion curves for equivalent depths of 8-90 meters.

c. Regression Analyses

Winds, temperature, and streamfunction from ERAI were linearly regressed onto the TD-filtered OLR time series at selected base points to obtain a statistical representation of wave structure. The linear regressions were computed only for those intervals when OLR is chosen to be below -1.5 standard deviations of the windowed and filtered OLR, yielding magnitudes that are typical of individual wave events based on case studies (as in Wheeler et al. 2000). Statistical significance of the local linear relationship in wind was retrieved using a bootstrap test (Efron and Gong 1983).

d. Kelvin Wave – Easterly Wave relationship

To find the relationship between KWs and EWs, we use a Hovmoller methodology, associating KW-filtered OLR with curvature vorticity from EWs. Basically, this methodology consists of finding the dates when EWs had genesis (details below), while the convective phase of a KW was present over the region of maximum KW-variance. The OLR associated with KWs was constrained to the band of 3°-12°N, while for EWs this was between 6°-13°N (see Fig. 5.1.a). These latitudinal bands were considered from previous results from KW convection (Straub and Kiladis 2002; Roundy and Frank 2004) and EWs over the EPAC (as in Chapter 4). Initial base points for EW genesis were selected at 84°W ± 2o in longitude, and at each of these points 1° of tolerance was given without overlapping with the central location (this is: intervals from 81°W to 83°W; from 83°W to 85°W and 85°W to 87°W), as illustrated in Figure 5.1.b. To exemplify this, Fig. 5.2 shows an example of a location within a Hovmoller diagram. In this plot, the convective envelope
of a KW generates an EW on July 28 of 1997 (orange line) at 86°W, as highlighted by the red point. This date then is selected as an EW genesis point from a KW.

The convective signal associated with KWs was selected when their anomalies were under -0.5 standard deviation. A genesis point was defined as the first point of a westward-propagating system lasting at least 4 days (based on the 500 km radially averaged curvature vorticity, associated with EWs) and exceeding $0.1 \times 10^{-5} \text{s}^{-1}$, following the methodology of Brammer and Thorncroft (2015) to identify EWs.

5.3. Kelvin Waves over the tropical EPAC

This section presents a brief review of the characteristics of KWs over the EPAC. This will be helpful to identify the impacts of KWs on dynamical and thermodynamical fields and to explore the possible relationship with EWs.

a. Variance of Kelvin Waves over the tropical EPAC

The geographical distribution of variance associated with Kelvin-wave filtered OLR during the months of July to September over the EPAC is shown in Fig. 5.3.a. Overall, convective activity associated with KWs distributes uniformly around 8°N across the EPAC. Large values are found around 120°W, weakening towards South America. This distribution is located north of the equator occurring between 5°N and 10°N. This is related to the climatological position of the ITCZ. Over the equator, cold SSTs prevent deep convection from occurring, as Straub and Kiladis (2002) pointed out.

The variance of rainfall associated with KWs over the EPAC is shown in Fig. 5.3.b. This closely follows the convective variance with a continuous distribution around 8°N. Both variables show impacts over Central America and the Panama bight, potential regions for EW genesis.
These results are in agreement with previous studies (Wheeler et al. 2000; Straub and Kiladis 2002; Roundy and Frank 2004).

Assuming that expected impacts of KWs related to EWs over the EPAC are over the Central America region, we selected a point where the maximum in KW-filtered OLR and rainfall variance were coincident. This point was considered at 9°N, 84°W and illustrated in Fig. 5.3.a. To identify the periods when KWs are present over the EPAC around this point, a 30-day running averaged variance of the Kelvin-filtered OLR was calculated over a boxed region with ±3° in longitude, and ±2° in latitude (see Fig.5.3.a). Figure 5.3.c shows that KWs occur in this region mostly between the months of April and August, peaking during May and August. When considering that EWs over the EPAC are mostly active during June to September, this result suggests that our results on KWs will not consider those EWs occurring during September; however, this will not impact the results on the KW-EW relationship.

To have an overview of KWs over the EPAC, Fig. 5.4 shows a time-longitude diagram of OLR anomalies averaged between 3°N-12°N (used as convective signature) and velocity potential at 200 hPa (used as a dynamical signature) during JJAS 1980-2015. Overall, an envelope of enhanced convection propagates eastward at about 14 ms⁻¹ from 130°W (though originating outside of the domain), with convection peaking around 80°W. Following the convective signal, negative velocity potential anomalies, representing upper-level divergence, are in phase along this convective trace. These features are in general agreement with the previous literature (Wheeler et al. 2000; Straub and Kiladis 2002; e.g. Roundy and Frank 2004). Despite these features, however, an important aspect should be pointed out. After the passage of the convective signature associated with the KW around 80°W, a westward propagating positive OLR
anomaly is observed. This feature suggests that after the passage of a KW, a dry ridge propagates westward, an unexpected result. This will be considered further on the development of westward perturbations over the EPAC. To have a deeper knowledge of KWs and to analyze this feature, the KW structure and evolution is presented next.

b. Structure of Kelvin Waves over the tropical EPAC

Figure 5.5 shows the horizontal and vertical composited structure of EWs over the EPAC together with their convective signal. In the horizontal, Fig. 5.5.a shows OLR anomalies (shading), and 850 hPa – streamfunction (contours) and winds (significant at 95%). At lag 0, enhanced convection (north of the equator) is observed over EPAC and the Central America region, with significant westerlies impacting the mountain regions of Central and South America, all these ahead of positive streamfunction over the equator.

For the vertical analysis, averaged OLR anomalies between 7°N-11°N in Fig. 5.5.b are shown for reference. The vertical structure of relative vorticity in Fig. 5.5.c shows a positive dominant signal at low levels, with a vertical extent reaching the 300 hPa, over the EPAC to the west of the Andes mountains. This is highest in the lower troposphere below 500 hPa, in the wake of KW convection, in agreement with previous results (Roundy 2008; Ventrice and Thornicroft 2013). Vertical velocity in Fig. 5.5.d shows two peaks, one at 84°W (central point), and another at 76°W close to the Andes, showing the enhanced convection from westerlies impinging on the topography, similar to results of Ventrice and Thorncroft (2013) in the Guinea highlands region. Also notable are the regions of descent west of 95°W below 500 hPa, consistent with subsidence behind the convective envelope.
Figure 5.6 shows the horizontal and vertical evolution of KWs over the EPAC from lag-4 to lag+4. Dynamical and convective fields are the same that were used at lag 0, as shown above. Overall, the evolution of KWs over the EPAC is characterized by an eastward propagating convective envelope of about 30° in longitude, a phase speed of about 12.5 ms⁻¹, and a period of 8 days, extending vertically to about 300 hPa.

Figure 5.6.a shows that at lag-4 the EPAC is characterized by a suppressed convective environment, and negative relative vorticity anomalies close to Central America. After two days, Fig. 5.6.b (lag-2) shows a convective envelope over the EPAC of 30° in longitude. This is characterized by positive relative vorticity anomalies ahead of the convective envelope and below 500 hPa. After crossing the point of regression at lag 0, Fig. 5.6.d shows a weaker convective envelope over the Caribbean Sea, followed by an incipient negative anomaly of relative vorticity, which suggest a westwards propagation (see relative vorticity in Fig. 5.6.e).

In summary, the evolution of KWs over the EPAC shows that during their passage over the region of Central America and the Panama bight, the environment is characterized by suppressed convection and positive relative vorticity anomalies. Regarding EWs, this result suggests that the passage of KWs over the EPAC may not provide the necessary convective and dynamical elements for their growth considering that convection can trigger EWs. To explore this, the next section considers an analysis of metrics associated with EWs during the passage of KWs over the region.
5.4. Relationship between Kelvin Waves and Easterly waves over the tropical EPAC

a. Hovmoller analysis

To study the potential effect of KWs on EWs, eddy kinetic energy (EKE) and curvature vorticity at 850 hPa were first used, motivated by previous studies of EWs in the West African region (e.g. Berry and Thorncroft 2005; Leroux et al. 2010; Ventrice and Thorncroft 2013). Figure 5.7 shows Hovmoller diagrams of 2-10-day filtered EKE and curvature vorticity at 850hPa (shaded) overlaid by KW-filtered OLR anomalies (contours).

Figure 5.7.a shows 850 hPa EKE with three maxima. One maximum lie along the KW convective envelope associated with the vorticity in the convective phase of the KW. Two maxima (highlighted by the thick black lines) show westward propagating anomalies, initiated around day-1 at 105°W and at day+1 at 75°W. The first preferred locations of enhanced EKE is likely associated with convection over the warm pool, a region where EWs intensify and genesis also occurs (see Chapter 4, Fig. 4.2.b-c). The second location is likely associated with the westward propagation of a ridge given the presence of a positive convective anomaly and in agreement with negative anomalies of relative vorticity as observed in the evolution of KWs over the region.

Figure 5.7.b which shows 850 hPa 2-10-day filtered curvature vorticity, highlights positive anomalies from day -4 to day 0, moving along with the convective KW envelope, similar to EKE. However, at lag+1 at 80°W, a westward propagation of negative curvature vorticity is seen again, along with positive OLR anomalies.

Overall, the results on metrics associated with EWs do not show a clear relationship between KWs and EWs over the EPAC as evident as that observed over West Africa. Despite this
result, there are cases that show the opposite, this is, that EWs can be generated from KWs. An example of this is presented next.

b. Influence of a Kelvin Wave on initiation of a tropical depression

Figures 5.8 and 5.9 show the sequence for a case when a KW generated an EW with a later development into a tropical depression, which took place on August 1, 1997. Figure 5.8 illustrates this case using brightness temperature ($T_B$) from GridSat in shading and its Kelvin band filtered anomalies in contours, while Fig. 5.8 shows again $T_B$ in shading and 2-10-day filtered curvature vorticity at 850 hPa in contours in association with EWs.

The succession of events starts on July 25 (Fig. 5.8.a), when the convective envelope of a KW and its associated convection is located around 120°W. This KW propagated eastwards impacting the Central America region on July 28 (Fig. 5.8.d). On this same day, Fig. 5.9.d allows us to see the generation of positive curvature vorticity in association with the convective envelope of the KW at 85°W, west of Central America (compare with Fig. 5.8.d). After this convective envelope passes 80°W on July 29 (Fig. 5.8.e), a trace of convection and its associated curvature vorticity is observed around 10°N (Fig. 5.9.e), west of Central America. This feature moves northwestward, showing organized convection on July 31 (Fig. 5.8.g). By August 1, this storm has a very well-defined circulation consistent with a tropical depression (Fig. 5.9.h), where a ridge-trough-ridge wave train can be observed close to the coast of Mexico.

The previous example suggests that KWs can have an influence on the EW activity over the EPAC. Even though our first result suggested that over the EPAC, KWs are generally followed by suppressed convection and an associated ridge. This example suggests that under special circumstances, EWs can be generated from KWs over the EPAC. Therefore, it is necessary to
identify such conditions associated with these events, which in turn, can be used to detect a favorable environment for genesis of EW genesis over the EPAC.

c. Easterly Waves generated by Kelvin Waves

To explore more generally how EW systems may develop from KWS, we applied the KW-EW methodology described in section 5.2.e, consisting basically on finding dates when the convective envelope associated with KWS and westward propagating associated with EWs coincide.

The results from this methodology are shown in Fig. 5.10, where a Hovmoller composite of all retrieved cases detected is presented. Shading presents 2-10 day filtered curvature vorticity at 850 hPa, and contours the convective envelope associated with KWS. Day 0 represents the minimum OLR associated with KWS around 84°W, 10°N; this point highlighted by the vertical black line in the Hovmoller diagram.

Figure 5.10.a shows the composite of all the cases identified (40 cases), representing only 20% of all KWS found over the EPAC. This composite shows westward-propagating systems that can last 7 days reaching longitudes as far as 115°W where they recurve. Interestingly, this EW is not the only one generated at day 0. Seven days before, a weak KW also invigorates an upstreaming westward propagating system, possibly from Africa, and followed by a westward propagating ridge, as in the previous result. As can be seen, this methodology has allowed the retrieval of cases of EWs genesis within the Central America region and the Panama bight, given that the curvature vorticity signal does not start further east.

To compare against a regular composite, Fig. 5.10.b shows 40 cases randomly selected and not following this methodology. As can be seen, while KWS do generate curvature vorticity
at the point of interest (84°W at day 0), this does not lead to EW propagation westward, followed by negative curvature vorticity, as in the general case. Even though there are some traces of westward propagating systems, these are weak, and they could be related to individual cases projecting into the composite.

In summary, the previous results allowed us to observe that most of the time, KWs do not generate EWs over EPAC. However, under some circumstances, KWs can generate westward propagating EWs, similar to results Africa (Ventrice and Thorncroft 2013). However, the immediate question related to the particular conditions under which EWs are generated from KWs are still unanswered. After exploring composites of variables associated with EWs, such as convection, potential vorticity, and SSTs, none of these led into a significant difference between developing versus non-developing cases (not shown). This was associated with the few number of cases found, resulting in noisy averages.

In light of this result the considered approach consisted in exploring the phases of the Madden-Julian Oscillation given that it has been observed that it modulates the EW activity over Africa (Ventrice et al. 2011; Ventrice et al. 2012) as well as over the EPAC (Maloney and Hartmann 2000a; Rydbeck and Maloney 2014) and results of (Schreck 2015) considered its role during the presence of KWs. Therefore, it is hypothesized that the MJO provides an enhanced environment where EWs can develop from KWs, closely associated with meridional PV gradients. This topic is explored next.
d. The role of the Madden-Julian on the relationship between Kelvin Waves and Easterly Waves over the tropical EPAC

This section presents first a quick overview of the presence of the MJO over EPAC, in particular, its influence on the horizontal distribution of PV and its associated negative meridional gradients (as discussed in Chapter 2). Figure 5.11 shows the total PV at the isentropic level of 320 K, which is near 600 hPa level (see Fig. 2.10) during JJAS 1980-2015. The convective and non-convective phases of the MJO were retrieved by considering the RMM sectors 8-1-2 for the convective phase and 4-5-6 for the non-convective phase.

Figure 5.11.a shows the PV distribution for the mean state during JJAS. Over the EPAC, a value of 0.23 PVU units (PVU; $1\text{PVU} = 1 \times 10^{-6} \text{K m}^2 \text{kg}^{-1} \text{s}^{-1}$) is observed at 100°W, 13°N (indicated by the white spot, taken as a reference). The distribution is parallel to the coast downstream from western Central America, linking the western Caribbean, associated with convection over the warm pool in the Pacific (as observed in Chapter 2). This PV distribution set up a negative meridional PV gradient along the coast, satisfying the Charney-Stern condition for dynamical instability (Charney and Stern 1962). Therefore, these conditions allow for the propagation of EWs over the region as discussed in chapter 2.

Figure 5.11.b shows the mean PV distribution during the convective phase of the MJO over the EPAC (RMM phases 8, 1 and 2). Overall, the distribution of total PV is similar to that of Fig. 5.11.a, except that the PV maximum is higher, about 0.26 PVU at the reference point (white spot), associated with enhanced convection (Maloney and Hartmann 2000a), thus increasing the meridional PV gradients at this level.
During non-convective MJO events (RMM phases 4, 5 and 6), Fig. 5.11.c shows a weaker PV distribution, of about 0.20 PVU at the reference point, thus diminishing the meridional PV gradients in comparison with the mean state.

By identifying the impacts that MJO sets up on the distribution of PV, and to observe the impact of KWs on EWs during the convective phase of the MJO over the EPAC, the identified events were partitioned in those occurring during the convective phase of the MJO and those occurring during the non-convective phase of the MJO. Figure 5.12 shows these Hovmoller composites where shading represents 2-10-day filtered curvature vorticity, purple-dashed contours negative OLR associated with KWs, while red dashed contours negative OLR anomalies associated with the convective phase of the MJO.

Figure 5.12.a shows the identified events during the convective phase of the MJO (18 cases). From this subset, it is observed that the convective envelope of the MJO intensify EWs propagating from upstream (possibly related to AEWs), with this invigoration lasting for the next 6 days. However, it is only when minimum OLR associated with the KWs is present, that EWs are generated in situ over the EPAC, without any perturbation from upstream. The westward propagation associated to these EWs last also 6 days followed by a negative curvature vorticity. A possible weakening may be related to the Hovmoller band used for these diagrams.

When compared with non-convective MJO episodes (12 events), Fig. 5.12.b shows that KWs also initiate curvature vorticity associated to EWs, but these systems are characterized by a shorter duration, being weaker and propagating only during about 4 days. This shorter duration could be associated with a recurving path or the interaction with a secondary KW and its convective suppressed phase, as seen at day+4.
Overall, these results indicate that the convective environment of the MJO is more favorable for the initiation of EWs from KWs over the EPAC. The environment of the MJO intensifies or diminishes convection and the meridional PV gradients necessary for EWs, allowing them to develop from KW events. However, there are cases when EWs can be generated from KWs, for example, the one presented at the end of July of 1997, when the RMM index was in phase 6.

5.5. Summary and conclusions

This chapter has explored the relationship between Kelvin waves and Easterly Waves over the EPAC during the extended summer season (June-September) during 1980-2015. Despite previous research has suggested that KWs modulate positively the activity of EWs and tropical cyclogenesis, over the EPAC, our analysis highlighted that KWs often generate a westward propagating ridge which is the opposite to what has been reported in other regions. Though not clear, it is hypothesized that this could be related to the reflected component of the Kelvin waves impacting topography, as previous studies suggest (Kleeman 1988; Gandu and Geisler 1991).

However, there are cases where KWs can indeed act as a trigger for EWs. Here we provided an example that showed this, suggesting that KWs can generate EWs. The methodology developed here, though simple in principle, allowed us to find cases where EWs are actually generated from KWs in the EPAC. From this, it was found that the total number of cases represent only about 20% of the total cases found in the period between 1980-2015, this consistent with no signal seen in the mean.

The results here suggest that the specific conditions present when EWs can be generated from KWs are associated with the presence of the convective phase of the Madden-Julian
Oscillation. During the convective phase, the MJO over the EPAC enhances convection, and increases relative vorticity in the mean state (Maloney and Hartmann 2000), but more important for EWs is the enhancement of negative meridional PV gradients, fundamental for the development of EWs and their further intensification downstream. Under this enhanced convective environment, EWs are generated from KWs in situ over the EPAC, without any precursor from upstream. This can be associated with triggering from convection (particularly from stratiform heating), enhancement of negative meridional PV gradients or possibly convection within the EWs. When this occurs, EWs can last 6 days on average. The origin of EWs from KWs on non-convective phases of the MJO is also possible, but this only happens a few times. These EW systems are weak and short-lived, persisting only for 4 days in average.

Figure 5.13 shows two schematic diagrams illustrating the presence of KWs and their relationship with EWs. Figure 5.13.a shows the mean effect of KWs together with a schematic Hovmoller representing their effect over the EPAC. The eastward progression of KWs and the associated westerlies generates deep convection over Central America, followed by downward vertical motion (as a direct circulation), creating a ridge in the lower troposphere. The Hovmoller shows that after the KW passage (in purple), a westward propagating ridge (in yellow) is initiated and lasts for about 4 days.

When KWs generate EWs, the convective phase of the MJO is often present. Figure 5.15.b illustrates this interaction, by highlighting the enhanced convection as the influence of the MJO over the EPAC. This convection in turn enhances vorticity over the region and in the nascent storms. The Hovmoller depicts the presence of the MJO (in gray), and when the KWs are present.
under this environment, the westward-propagating trough (in green) arises and propagate westward, lasting for about 6 days on average.

More research is needed, with a different approach, to clarify the synoptic-scale environment associated with the genesis of EWs from KWs. One hypothesis is on the convective forcing that KWs produce. For example, while over Africa deep convection generates vorticity at mid-levels that grows taking advantage from PV gradients, over the EPAC it may be necessary that the stratiform heating left behind the KWs be more persistent. That in turn can produce vorticity at around 600 hPa and grow downstream over the EPAC from the PV gradients. This will also explain why under the convective phase of the MJO the KWs generates more often EWs. A second hypothesis would be to recognize if changes on the phase speed of KWs also enhance the stratiform heating. For example, Fig. 5.12, shows differences in phase speed. When low phase speeds are present in KWs these develop longer EWs. Exploring these hypotheses can help to clarify this relationship and these topics left for future work.
Figures

![Map of Easterly wave track density](image)

**Figure 5.1.** a. Easterly wave track density (number of counts per 0.5°) during JJAS 1980-2015. Green lines (3°N-12°N) show the Hovmoller region used to retrieve convectively coupled equatorial Kelvin waves, while black lines (5°N-15°N) show the Hovmoller region to track EWs. Track density obtained from data from ERAI. b. Schematic showing the longitudes used for the interaction KWS-EWs.
Figure 5.2. Example of the Hovmoller methodology used at 86°W during 1997. Shades represent unfiltered OLR anomalies between 6°N-13°N. Green contours represent KW-filtered OLR as in Fig. 5.1a. Purple lines represent TD-filtered OLR (between the 5°N-15°N in Fig. 5.1) while black lines represent EWs retrieved from the Hovmoller methodology similar to Brammer and Thorncroft (2015). Red spots show EWs genesis during a KW passage. Bold vertical line indicates the coast of South America (at 78°W), and the longitude where EW genesis is expected, in this case at 86°W (thin line).
Figure 5.3. Distribution of the mean variance during the rainy season (June-September) of Kelvin-band filtered for a. OLR \( (W^2 m^{-2}) \), and b. rainfall \( (mm^2 day^{-2}) \). Kelvin wave activity was determined by filtering each variable in the Kelvin-wave band with periods of 2.5-20 days, wavenumbers 1-14, and depths of 8-90 m. c. Annual variability \( (W^2 m^{-2}) \) of Kelvin wave around the boxed area on a. Blue line represents the mean annual variance while the red line represents the mean variance plus 1.5 standard deviation. Data from NOAA daily OLR at 0.5° during JJAS 1980-2015 and from TRMM 3B42 during JJAS 1989-2009.
Figure 5.4. Time-longitude composite of daily averaged OLR anomalies (in Wm$^{-2}$) of Kelvin wave filtered in shading and velocity potential (m$^2$s$^{-1}$) in contours averaged over each CCKW lag between 3°N-12°N. Positive (negative) OLR anomalies statistically different than zero at the 90% confidence level are dotted. Shading interval is 0.1 Wm$^{-2}$. Contour intervals are 0.5m$^2$s$^{-2}$. Data from NOAA Daily OLR at 0.5° resolution, and ERAI at 0.5° resolution during JJAS 1980-2015.
Figure 5.5. Horizontal and vertical structure of Kelvin waves at 9°N, 84°W (indicated by the red dot in a) at lag 0. 

a. OLR (Wm⁻²) and 850 hPa - streamfunction (10⁻⁵ m²s⁻¹) and winds (ms⁻¹). Every other 95% significant wind vector is shown. Streamfunction contours are every 0.1x10⁵ m²s⁻¹, with negative contours dashed. b. OLR anomalies (Wm⁻²) averaged between 7°N-11°N. c. Relative vorticity anomalies (10⁻⁶ s⁻¹) along 9°N, with the vertical black dashed line indicating the reference point at 84°W, while the brown continues one illustrating the coast; and d. Omega vertical velocity anomalies (Pa s⁻¹) along 9°N. Data source from NOAA Daily OLR at 1° resolution, and ERAI at 0.5° grid space during JJAS 1980-2015.
Figure 5.6. Kelvin wave passage over the EPAC from lagged regressions at 9°N, 84°W (indicated by the red point). a. lag-4, b. lag-2, c. lag 0, d. lag+2, e. lag+4. OLR (Wm⁻²) in shading, 850 hPa streamfunction (m²s⁻¹) in contours, and winds (ms⁻¹) indicated by arrows. Every other 95% significant wind vector is shown. Streamfunction contours are every 1x10⁵ m²s⁻¹; with negative contours dashed. Data from NOAA Daily OLR at 0.5° resolution, and ERA-I at 0.5° resolution during JJAS 1980-2015.
Figure 5.6. Continued.
Figure 5.7. Time-longitude composites of metrics associated with EWs during the convective envelope of the passage of KWs. Shading represents: a. Daily averaged 850 hPa eddy kinetic energy (in m$^2$s$^{-2}$) of 2-10-day filtered 850 hPa winds; and b. 850 hPa 2-10-day filtered curvature vorticity. Contours represent the convective Kelvin-filtered OLR anomalies (Wm$^{-2}$), with positive values in a continuous line and negative dashed line. EW metrics were averaged for each KW between 6°N-13°N while Kelvin-filtered OLR anomalies were averaged between 3°N-12°N. Contour intervals are 5 Wm$^{-2}$. Dashed line represents the target longitude at 84°W, while the brown continuous line the continental coast. Data retrieved from NOAA Daily OLR at 0.5° resolution, and ERA-I at 0.5° resolution during JJAS 1980-2015.
Figure 5.8. Daily sequence for a case of a Kelvin Wave and its interaction with an Easterly Wave over the tropical EPAC from July 25 to August 1, 1997. Unfiltered brightness temperature $T_b$ (in °C) in shades, while $T_b$ Kelvin filtered in contours. Data retrieved from GridSat at 8 km.
Figure 5.9. Daily sequence for a case of a Kelvin Wave and its interaction with an Easterly Wave over the tropical Eastern Pacific from July 25 to August 1, 1997. Unfiltered brightness temperature $T_B$ (in °C) is in shades, with 2-10-day filtered curvature vorticity at 850 hPa in contours. Negative values are dashed. $T_B$ data retrieved from GridSat at 0.5°. Curvature vorticity calculated from ERAI at 0.5°.
Figure 5.10. Time-longitude composite of daily averaged 850 hPa 2-10-day filtered curvature vorticity (shaded) averaged over each KW between 6°N-13°N, and Kelvin filtered OLR anomalies are averaged between 3°N-12°N a. For EWs developing from KWs following the methodology, and b. For 40 random cases. Negative Kelvin filtered OLR anomalies are dashed, while positive OLR anomalies are continuous. Shade interval is 5x10^{-5} s^{-1}; contour intervals are 5 Wm^{-2}. Vertical line indicates reference point at 84°W. Data retrieved from NOAA Daily OLR at 0.5° resolution, and ERA-I at 0.5° resolution during JJAS 1980-2015.
Figure 5.11. Mean potential vorticity (PV) at 320 K (~600 hPa) during MJO phases during JJAS 1980-2015. a. Mean conditions, b. MJO Westerly (or convective) phase, and c. MJO Easterly (or non-convective) phase. White circle denotes is used for reference values. Data obtained from ERAI at 0.5°.
Figure 5.12. Time-longitude composite of daily 850 hPa 2-10-day filtered curvature vorticity (shaded) averaged between 6°N-13°N, and Kelvin-filtered OLR anomalies averaged between 3°N-12°N for convective and non-convective MJO phases. **a.** For cases during MJO Westerly (phases 8-1-2), and **b.** For cases during MJO Easterly phases (4-5-6). Negative Kelvin-filtered OLR anomalies are dashed, while positive OLR anomalies are continuous. Shading interval is 5x10^{-5} s^{-1} while contour intervals are 5 W m^{-2}. Vertical line indicates reference point at 84°W. These figures can be compared with Fig. 5.10.**a** for all EWs developing from KWs. Data retrieved from NOAA Daily OLR at 0.5° resolution, and ERA-I at 0.5° resolution during JJAS 1980-2015.
Figure 5.13. Schematic representation of the interaction between convectively coupled equatorial Kelvin waves and Easterly Waves over the tropical Eastern Pacific. **a.** On regular cases, when Kelvin waves generate a ridge (in yellow) after their passage, and **b.** When Kelvin waves generate an Easterly wave (in green) when the convective phase of the MJO is present.
Chapter 6. Genesis of Easterly Waves over the tropical Eastern Pacific
and the Intra Americas Seas

6.1. Introduction

Easterly waves (EWs) are westward propagating disturbances found in the summer season over the tropical band of 15°N-20°N across the world. These are observed in West Africa and the Atlantic (e.g. Kiladis et al. 2006, and references therein), Australia (e.g. Dickinson and Molinari 2000), the west Pacific Ocean (e.g. Serra et al. 2008 and references therein), and over the tropical Eastern Pacific (e.g. Molinari et al. 1997; Serra et al. 2008; Rydbeck and Maloney 2014; Rydbeck and Maloney 2015). A general consensus exists regarding their genesis mechanism which is often viewed to be associated with a mixed barotropic-baroclinic instability mechanism. Recently, Thorncroft et al. (2008) questioned this paradigm by showing that EWs over Africa could be triggered by localized forcing in association with latent heat at the entrance of a stable African Easterly Jet. Their results showed structures similar to observed EWs over Africa, thus providing a viable mechanism for EW genesis. Motivated by this result, this research explores the triggering hypothesis for EWs over the tropical Eastern Pacific (hereafter EPAC), given that similar convective and dynamical conditions are also met over the region (see chapter 2). Therefore, the aim of this chapter is to test if the triggering hypothesis of EWs could have validity over the EPAC, a region where the mechanisms for genesis of EWs remains uncertain.

The characteristics of EWs over the EPAC and the Intra Americas Seas (IAS) have recently been documented in several studies (Tai and Ogura 1987; Petersen et al. 2003; Kerns et al. 2008; Serra et al. 2008; Serra et al. 2010), as well as in this work. Previous studies have found that, on average, EWs are characterized by a phase speed of 5-11 ms⁻¹, a period of 3-5 days, and a
wavelength of around 2200 km. It has also been found that EWs can grow from barotropic energy conversions (Serra et al. 2008; Maloney and Hartmann 2000a; Serra et al. 2010), wave accumulation (Webster and Chang 1988; Shelton 2011), and growth of random vorticity noise by barotropic energy conversions (Hartmann and Maloney 2001; Maloney and Hartmann 2001). Additionally, recent work finds that they can also grow from baroclinic energy conversions over the EPAC (Serra et al. 2010; Rydbeck and Maloney 2014) with mixed results about their relative importance (Serra et al. 2008; Serra et al. 2010; Rydbeck and Maloney 2014). However, these results focus on growing mechanisms assuming that EWs pre-existed, and do not properly address the genesis mechanism.

Some mechanisms about EW genesis over the EPAC have been proposed. These include barotropic instability of the mean flow (Mozer and Zehnder 1996; Molinari et al. 1997; Molinari and Vollaro 2000; Maloney and Hartmann 2000a); inertial instabilities from cross-equatorial pressure gradients (Toma and Webster 2010a; Toma and Webster 2010b); and the inter-tropical convergence zone (ITCZ) breakdown (Ferreira and Schubert 1997). Overall these mechanisms deal with a barotropic instability mechanism, which extracts kinetic energy from a horizontal sheared flow. Rydbeck et al. (2017) recently suggested another mechanism for the in situ generation of EWs over the EPAC. This consists primarily of the forcing of EWs by local convective disturbances resulting from the high terrain over the northern mountain ranges of South America, next to the Panama bight. However, the result of Rydbeck et al. (2017) only shows the suppression of EW variability from removing topography and does not explicitly demonstrate the generation of an EW by a convective forcing.
Observational results also suggest that EWs can be triggered by convection. For example, Berry and Thorncroft (2005) establish that convection was the precursor of a strong African Easterly Wave (AEW) in 2000. This convection was initially composed of several mesoscale convective systems (MCS) located over the Darfur mountains, close to the entrance of the African Easterly Jet. In comparison, over the EPAC, there exits the potential for triggering of EWs from convection over the mountain regions near the Panama bight, close to mid-level jet over the EPAC (see Chapter 2). Additionally, convection over these locations can also be associated with MCS forced by orographic lifting (Mapes et al. 2003a). Thus, the convective forcing present over the EPAC can be considered as a potential trigger for EWs and should be explored.

Recently, the triggering hypothesis of AEWs by convection was tested via a modeling approach (Thorncroft et al. 2008). This approach consists of two steps: the first one sets up a dry mean state, and the second provides a localized transient heating region to simulate the impact of convection. Overall, the results of these experiments showed that the simulated response, after several days of integration, resembles the observed structures of AEWs. Motivated by this result, here we propose that EWs over the EPAC can be triggered by a localized convective forcing associated with latent heating close to or along the mid-level jet over the EPAC (see Chapter 2, Fig. 2.4.b and Fig. 2.10). To achieve this objective, we perform a modeling study, similar to Thorncroft et al. (2008), considering different heating profiles that are representative of the different convective population over the EPAC-IAS region (Huaman and Schumacher 2018; Zuluaga and Houze 2015).

For this goal, this chapter is organized as follows: Section 2 provides a background about the main features of the mean state over the EPAC-IAS. In section 3, we present the data, model,
and approach used in this study. Section 4 analyses the simulated EWs over the EPAC-IAS region in response to heating triggers based on a stratiform convective profile, considered as a reference. After this, section 5 presents an energetics analysis and a comparison with AEWs to observe differences between them. Later, a sensitivity analysis to heating profile, genesis location, as well as varying mean state is presented in section 6. This is followed by a summary and some discussion on the significance of the present findings in section 7.

6.2. Background: Main features over the EPAC-IAS region

The EPAC-IAS region is characterized by a complex atmospheric mean state in a region of significant variations in topography. To explore how the mean state supports EWs, as well as possible locations for a triggering mechanism over this region, this section presents a quick overview of the atmospheric state during the extended summer season (June to September, hereafter JJAS) from 1980-2015 using data from ERA-I (Dee et al. 2011). The key atmospheric features are presented in Fig. 6.1. The wind distribution at 600 hPa in Fig. 6.1.a (considered from dynamical conditions discussed later) shows a distinctive easterly flow over the southwestern Caribbean Sea and into the EPAC. In particular, in the latter region, the jet is characterized by magnitudes ranging from 6-7 ms⁻¹. This flow is important for the evolution of EWs, given that most of the atmospheric perturbations, such as EWs and TCs, track along this jet (Serra et al. 2010; Brammer and Thornicroft 2015). At lower levels (Fig. 6.1.b), the Caribbean Low-level jet (CLLJ: Amador 1998; Amador et al. 2010; Cook and Vizy 2010), is a dominant local climatic feature over the IAS region that is present during the boreal summer. Previous research suggests that this low-level jet can support localized EW growth through a barotropic growth mechanism (Molinari et al. 1997; Molinari et al. 2000; Serra et al. 2010). However, we will focus on the mid-
level jet, given that its extension is larger (about 20° in longitude over the EPAC plus 10° more over the western Caribbean) and it is linked with the CLLJ thus arguably providing a more favorable dynamical environment for EW growth.

The dynamic instability of the mid-level easterly jet over the EPAC-IAS region is assessed using a potential vorticity (PV) perspective, consistent with previous analyses of other tropical easterly jets (Burpee 1972; Thorncroft and Hoskins 1994; Molinari et al. 1997). At the level of 600 hPa, Fig. 6.1.c shows the mean JJAS horizontal distribution of PV. Over the EPAC, relatively high PV is present south of 20-25°N (generated by latent heat, see chapter 2) is observed over the oceanic regions, while low PV is observed over the Sierra Madre (linked to low stability and dry convection as proposed in Chapter 2). This distribution of PV results in a local negative meridional gradient in PV roughly parallel to the continent. This distribution then satisfies the Charney-Stern condition for instability of internal jets (Charney and Stern 1962), as well as the Fjortoft condition (Fjortoft 1950), which requires that the mean zonal easterlies are positively correlated with the negative meridional PV gradient (see Fig. 6.1.a and Fig. 6.1.c). The fulfillment of these conditions allows for the growth of EWs over this region.

The mean JJAS Outgoing Longwave Radiation (OLR), in Fig. 6.1.d, highlights the most convectively active regions that might provide potential triggering locations for EWs. Notably, these are located equatorward of the mid-level jet over the EPAC. A striking feature is the convectively active region over the Panama Bight. From its associated diabatic heating (Zuluaga and Houze 2015), it could be a likely triggering location for EWs, as suggested in previous studies (Serra et al. 2010; Rydbeck et al. 2017). The distribution of OLR also shows the ITCZ over the EPAC, characterized by a minimum over the convergence zone, in agreement with de Szoeke et
al. (2006). In contrast, over the Caribbean Sea, most of this basin remains without convection, particularly the southern region. This is coincident with the CLLJ (see Fig. 6.1.b) and the upwelling oceanic area associated with cold Sea Surface Temperatures (SSTs) as Gordon (1967) noted.

In summary, the atmospheric mean state over the EPAC-IAS region during JJAS is characterized by a jet-like steering flow at mid-levels (600 hPa), and a low-level jet at 950 hPa over the Caribbean Sea (Fig. 6.1.c). The mid-level flow at 600 hPa meets the Charney-Stern and the Fjortoft necessary conditions for instability of atmospheric flows and supporting the growth of EWs. We could expect that the EWs over the EPAC may be weaker in comparison with those over Africa given the magnitudes of the mid-level jet and the PV gradients. While the intensities of these over the EPAC reach a maximum of 6.0 ms\(^{-1}\) and 0.3 PVU, over Africa these are 13 ms\(^{-1}\) and 0.4 PVU respectively. The OLR distribution over the EPAC-IAS region suggests that triggering for EWs via diabatic heating may most likely occur over the Panama bight region, equatorward of the mid-level jet. We hypothesize that these features, together, set up a suitable environment for triggering the development of EWs over the EPAC. Having identified the main features or the mean state over the EPAC, we provide in the next section the characteristics of the model used, as well as an assessment of how well the features here introduced are captured by this model given its resolution.

6.3. Data and Modeling Approach

The overall approach used in this research is similar to that in Thorncroft et al. (2008), which consists of exploring the dry adiabatic response to finite-amplitude, transient heating perturbations, to explore the role of dry dynamics on the establishment of EWs. The characteristics of the model used for these simulations are described next.
a. Model

The model and setup used for this research are similar to Thorncroft et al. (2008) but with some differences, following Hall et al. (2019). The model is a dry global spectral primitive equation model with truncation at wavenumber number 42 (T42), corresponding to a horizontal grid of about 2.8° at the equator and 15 vertical levels in a sigma coordinate system. This is a higher resolution in the horizontal and vertical than that used by Thorncroft et al. (2008) who used T31 and 10 vertical levels. Integrations of the full nonlinear equations for vorticity, divergence, temperature, and surface pressure are carried out using a semi-implicit 22.5-minute time step. As in similar studies using this model (Hall 2000; Hall et al. 2006; Thorncroft et al. 2008) the basic state is maintained by adding a forcing term that represents the combined effects of diabatic heating and transients. The perturbation approach requires us to constrain the experiments to be linear by imposing a very small initial forcing perturbation and subsequently rescaling the response. Consistent with previous work with this model, a 12h-$\nabla^6$ diffusion is applied to the momentum and temperature equations. Low-level damping (representative of turbulent transfer of momentum and heat with the surface) is also included in all these simulations as described in Hall et al. (2006). Damping rates used in this model version are 16 hours for momentum and temperature near the surface with damping coefficients decreasing linearly from the surface $\sigma = 1$ to $\sigma = 0.79$. Above $\sigma = 0.79$ and in the free atmosphere, damping time scales for momentum and temperature are on a 20-day timescale. In addition, a radiative-convective restoration rate of 12 days is applied to on temperature only and is independent of height.
A basic run involves prescribing the heating in a particular location for a period of one day and then switching it off. The model is integrated for several days after this to see the adiabatic response to this transient heating. After the first day, no heating is included within the developing EW, and so it is expected that EW amplitudes will be weaker than those observed or seen in models with full physics (Serra et al. 2008; Serra et al. 2010; e.g. Crosbie and Serra 2014; Rydbeck and Maloney 2014; Rydbeck and Maloney 2015). The main objective of this work is to identify whether upstream heating can lead to realistic EW activity downstream, and to investigate the impact of the heating on the nature of the EW developments.

b. Data

The data used for this model was obtained from the ERA-Interim (ERAI) reanalysis (Dee et al. 2011) and ingested into the model at T42 resolution. A brief comparison of the main features of the mean state over the EPAC-IAS from the T42L15 model is shown in Fig. 6.2.

In the horizontal, Fig. 6.2.a shows the level of $\sigma = 0.60$, where easterly flow over the Caribbean and EPAC reaches a magnitude of about 7 ms$^{-1}$, closely reproducing those values from ERAI reanalysis (c.f. Fig. 6.1.b). The distribution of mean PV at the same level in Fig. 6.2.b from the model shows weaker PV anomalies and thus a weaker negative meridional gradient over the region, consistent with that from ERAI reanalysis (see Fig. 6.1.c).

The vertical structure of the zonal wind in the model is shown in Fig. 6.3 (a-c, inverted order in the upper row) at different longitudes (red lines in Fig. 6.2.a). Focusing on the region of interest (between 10$^\circ$N-20$^\circ$N), at 80$^\circ$W the CLLJ is resolved around 15$^\circ$N with a vertical extension up to 500 hPa. At 90$^\circ$W we observe the continuation of the jet at 15$^\circ$N. At 100$^\circ$W, the mid-level jet is at 15$^\circ$N between 500-600 hPa. Also, a prominent signal from the tropical easterly jet at
upper levels is present. Overall, all these features suggest that the T42L15 model is able to resolve the key mean wind features over the region, as can be compared with ERAI (Fig. 6.3.d-f illustrated in the lower row).

As previously mentioned, the distribution of PV is likely important for the growth of EWs. Figure 6.4.a-c shows the vertical cross-sections of PV at the same latitudes previously illustrated. Overall, the model resolves the vertical PV distribution between 12.5°N-15°N though weaker (compare with Fig. 6.4.d-f). PV minima are well represented around 15°N with negative meridional PV gradients at 80°W (Fig. 6.4.a), and weaker at 90°W and 100°W (Figs. 6.4.b-c, respectively). This is expected given the coarse resolution of the model, in particular, in the vertical. This distribution can provide the necessary elements for the growth of EWs over the region at this resolution. It must be recognized also the presence of local PV maxima observed at 850 hPa in all longitudes, however, these may result from the calculation of PV in the vertical within the model.

c. Heating profile

The initial localized heating is prescribed in the model through the thermodynamic equation, following Thorncroft et al. (2008). In the horizontal, the heating function is given by the form:

\[
H(r) = \begin{cases} 
H_0 \cos^2 \frac{\pi}{2} \left( \frac{r}{r_0} \right), & r \leq r_0 \\
0, & r > r_0 
\end{cases}
\]

where \(H_0\) is the peak heating rate at the center, \(r\) is the distance from the center, and \(r_0\) determines the horizontal scale of the heating. For the basic heating run, \(r_0\) is set to 7.5° and the heating is centered at 7°N and 78°W (illustrated in Fig. 6.5.a), downstream of the Colombian
Andes and over the Panama bight. This point was chosen because we hypothesize that convection can trigger EWs over the EPAC, and this location represents a region of heating associated with enhanced mean convection over the region of the Gulf of Panama (Zipser et al. 2006; Zuluaga and Houze 2015).

In the vertical, the heating function initially considered is characterized by a stratiform convective profile with the analytic form given by:

\[
H_0 = \frac{75\pi}{(71 - 18e^{-2\pi})}[e^{-2\pi(1-\sigma)} - 1] \cos \frac{3\pi}{2} (1 - \sigma)
\]

where \(\sigma\) is the model vertical coordinate with normalized pressure (see gray line in Fig. 6.5.b). This profile was first considered given that the EPAC-IAS region is dominated by stratiform heating including the Panama bight (Zuluaga and Houze 2015), and according to Hertenstein and Schubert (1991), such a profile is particularly efficient in generating cyclonic PV anomalies (and their associated relative vorticity) in the mid-troposphere. This vorticity is caused by strong heating gradients in the vertical at such levels. The peak heating rate for this profile is 3.12 K·day\(^{-1}\) at \(\sigma = 0.31\), and a cooling rate of -1.47 K·day\(^{-1}\) at \(\sigma = 0.85\). This convective profile integrates to unity between \(\sigma = 0\) and \(\sigma = 1\) (as well as the others illustrated, hence they all account for the same total heating to the system). For our purposes the stratiform convective profile is of particular interest given that the present study will focus on \(\sigma = 0.60\) from the dynamical conditions found over the mean state.
6.4. Forced EWs over the EPAC-IAS region

EWs forced by localized stratiform heating

The atmospheric response to stratiform heating is shown in Fig. 6.6 at σ=0.60. After the initial adjustment to the heating by day 1, the evolution in the relative vorticity field takes the shape of a coherent EW wave train downstream of the heating location. The initial heating directly spins up a trough at 600 hPa (Fig. 6.6.a) similarly to that seen in Thorncroft et al. (2008). By day 4, however, a clear and well-defined EW structure with troughs and ridges, is observed, very similar to observed EWs over the EPAC (Serra et al. 2008; Serra et al. 2010; Crosbie and Serra 2014; Rydbeck and Maloney 2014; Rydbeck and Maloney 2015; Rydbeck et al. 2017), as well as with our results in Chapter 4 (see Figs. 4.7 and 4.14). By day 6 the EW structure has lost its coherence and appears to be dissipating (Fig. 6.6.f). The initial vortex is also weakening as it reaches 105°W, though according to our results in Chapter 4, this vortex possibly would continue developing further.

The simulated EWs track parallel to the coast following the negative PV gradient observed at mid-levels, as well as the zonal winds (see Fig. 6.2). This conditions will support their growth and propagation, as stated by dynamical conditions (Fjortoft 1950; Molinari et al. 1997; Thornicroft et al. 2008). The simulated EWs have a wavelength of about 2000 km, and horizontal tilts characteristic of barotropic growth. They propagate north-westwards with a speed of about 4.6 ms⁻¹. Together, these characteristics resemble the EWs seen from regressions in previous studies (Serra et al. 2008; Serra et al. 2010; Rydbeck et al. 2017). Regarding the amplitudes of these simulated EWs, of about 4x10⁻⁶ s⁻¹, we hypothesize that convection and its associated latent
heat enhance them and compare with observations. These results provide strong evidence that supports the hypothesis that East Pacific EWs can be triggered by upstream localized heating.

To explore the structure of these simulated waves in the vertical, Fig. 6.7 shows the cross section of the meridional wind along the purple line included in Fig. 6.6.d. The vertical structure (left column in Fig. 6.7) during the first day (Fig. 6.7.a) shows a vertical development consistent with the direct response to the stratiform heating profile, characterized by a mid-level trough centered around 600 hPa. There is also a response at lower and upper levels (900 hPa and 150 hPa respectively) that is consistent with the day 0 results in Thorncroft et al. (2008). In subsequent days the mid-level trough can be seen to propagate westward with the peak moving to the 500 hPa level. By day 4 (Fig. 6.7.d) the wave has lost its low-level signal while at upper-levels this has also weakened.

Throughout the evolution the meridional wind indicates a vertical structure with a tilt with height (illustrated in Fig. 6.7.d) of about 1.8 hPa km$^{-1}$, suggesting a role for baroclinic growth processes. In addition, by day 4 (Fig. 6.7.d), the meridional wind shows a maximum southerly above the jet level ($\sigma = 0.50$) of 0.8 m$^{-1}$ behind the trough. This value is similar to Serra et al. (2008), who found a value of 0.9 m$^{-1}$ from composited EWs from reanalysis. Previous results from regressions (in Chapter 4) show that meridional wind values can reach 1.5 m$^{-1}$ however, the weaker values in the dry model are likely associated with the lack of convection.

An interesting aspect regarding predictability arises when considering the time that it takes for the initial trough, triggered by convection, to reach the mature EW region (as in Chapter 4). According to our results, it takes about 5 days to reach such region (see Fig. 6.6.e). This implies
that knowing that convection can trigger an EW, it takes this time to reach the warm pool, thus being a predictable feature.

The simulated EWs over the EPAC, when compared with those over Africa (Thorncroft et al. 2008), show weaker amplitudes (around 2.5 ms\(^{-1}\) for stratiform convective regions over Africa) and a shorter life cycle with EWs breaking apart after day 4. However, we must consider that the mean state configuration is very different. While the EPAC-IAS region presents a mid-level jet of about 2000 km, the African easterly jet is characterized by a length of about 5000 km. Despite the differences, these results support the initial hypothesis that EWs over the EPAC can be forced by a finite-amplitude, transient heating, through an adiabatic response. Further comparisons with African Easterly Waves are included in the next section which explores eddy fluxes and energetics of this control run.

6.5. Eddy fluxes and energetics

To shed light on the possible EW growth mechanisms over the EPAC, this section briefly explores the EW energetics. This will be helpful to clarify the contributions associated with barotropic and baroclinic growth. To achieve this, we considered the integration of covariances from the perturbations terms, similar to the approach of Thorncroft and Hoskins (1994). However, the present section does not include full energetics with conversion terms. For the present case, this analysis considers the dry run of the mean JJAS state with a stratiform convective profile averaged longitudinally between 110\(^\circ\)W and 80\(^\circ\)W (see black box in Fig 6.6.d) from day 2 to day 6 of the control run.

Figure 6.8 shows different meridional cross sections for the four second-momentum terms, zonally averaged. For these calculations, the overbar denotes a longitudinal mean, and
primes deviations from that mean, averaged over the longitudes previously indicated. For brevity, this analysis will be focused for EWs over the EPAC around 15°N.

The zonal-averaged eddy kinetic energy (EKE) is presented in Fig. 6.8.a (shading) together with zonal winds (in contours), highlighting the mid-level jet around 15°N. The EKE structure is dominated by a maximum equatorward and just above the jet core between about 400 hPa and 600 hPa. This maximum reaches a value of 0.8 m²s⁻² and it is collocated with the mean winds (see Fig. 6.3.a-c for reference). This maximum result from vorticity generated by the stratiform heating profile below the maximum of heating (see grey line in Fig. 6.5).

The horizontal eddy momentum flux in Fig. 6.8.b, shows a maximum at 400 hPa coincident in location with EKE maxima. A minimum is located poleward of the jet maximum (in contours) at 500 hPa and extending from 400 – 600 hPa on the southern flank of the mid-level jet, in the region of meridional shear. At 600 hPa, a positive-negative meridional structure shows the flux of easterly momentum away from the jet, associated with barotropic growth of systems tilted against the meridional shear, implying a weakening of the mean jet and a strengthening of the eddies. This is a very coherent pattern suggesting a significant contribution to EW growth from barotropic processes, consistent with the horizontal tilts seen in Fig. 6.6. Eddy momentum flux maxima on the equatorward and poleward side of the jet are +0.29 m²s⁻² and -0.04 m²s⁻², respectively. The structure on the southern flank of the mid-level jet is consistent with the EKE field, associated with the zonally averaged values on a horizontal tilted structure. Overall, we can observe that these perturbations (EWs) are really taking advantage of the PV maximum observed in the model between 400 hPa and 600 hPa.
Large-scale heat fluxes in Fig. 6.8.c are found above and below the jet (15°N). Negative values under the jet level are about -0.012 K·ms\(^{-1}\), while positive values above it are of about 0.020 K·ms\(^{-1}\). This is consistent with cold air moving poleward below the mid-level jet and warm air equatorward above it. This is actually observed from values of potential temperature at 950 hPa and at 2 PVU (see Chapter 2, Fig. 2.11). Consistent with this, the vertical heat flux in Fig 6.8.d shows an upward heat flux around 15°N. Together, the horizontal and vertical heat fluxes show very small values suggesting a weak baroclinic growth over the region consistent with the vertical tilts observed in Fig. 6.7.d.

To compare the zonal-mean diagnostics of EWs over the EPAC with AEWs over West Africa, Fig. 6.9 shows the same diagnostics for EWs over Africa generated from a simulation with the same model previously described and with the same characteristics regarding convection and time. The zonal average was done from 31°W to 10°E, as in Thorncroft and Hoskins (1994). In short, the zonal mean EKE in Fig. 6.9.a shows a mid-level maximum equatorward of the jet, with a deeper reach aloft greatly exceeding 1.0 m\(^2\)s\(^{-2}\). The EKE values over the EPAC of 0.8 m\(^2\)s\(^{-2}\) at most, showing a weaker intensity.

The horizontal eddy momentum flux over Africa (Fig. 6.9.b) shows a distribution similar to the EPAC characterized by the flux of momentum away from the jet consistent with barotropic energy conversions. Fluxes over Africa have a greater amplitude than those over the EPAC, exceeding +0.5 m\(^2\)s\(^{-2}\) compared against +0.29 m\(^2\)s\(^{-2}\). This comparison illustrates the strength of the African Easterly Jet and the PV-\(\theta\) structure over Africa, as compared to the EPAC.

The horizontal eddy flux over the African region (Fig 6.9.c) shows a very clear dipole above and below the African easterly jet (located about 700 hPa), with negative values of -0.144 K·ms\(^{-1}\)
and positive values of +0.121 K·ms\(^{-1}\) respectively, this implying cold air moving poleward at low levels and warm air moving equatorward, clearly showing baroclinic growth. From this, it is evident that heat fluxes over the EPAC (in Fig. 6.8.c) with values of -0.012 K·ms\(^{-1}\) and +0.020 K·ms\(^{-1}\) are weaker when compared with these over Africa, and though the structures are similar, this result suggests a weak contribution to the growth of EWs over the EPAC from baroclinic growth.

Regarding the vertical heat fluxes over the African region, Fig. 6.9.d shows a clear upward heat flux below the jet together with cold air sinking around 400 hPa. The magnitudes of these fluxes are -0.13 K hPa hr\(^{-1}\) and 0.009 K hPa hr\(^{-1}\), and when compared with those over the EPAC of -0.01 K hPa hr\(^{-1}\) and 0.008 K hPa hr\(^{-1}\) (both below and above the jet, respectively), it is evident that those over the EPAC are very weak because of differences in temperature gradients in the mean state over Africa.

This comparison shows that over the EPAC, EWs grow from the barotropic and baroclinic contributions though these are both much weaker when compared to AEWs. In particular, the values associated with barotropic processes over the African region exceed those over the EPAC by about a 5:1 ratio. This highlights the differences between the characteristics of the mean state regarding jet length, and PV-\(\Theta\) structure in comparison with the mean state over the EPAC.

6.6. Sensitivity analysis

a. Sensitivity to heating profile

Given that the EPAC-IAS region is impacted by other types of convection (Schumacher et al. 2004; Zuluaga and Houze 2015; Huaman and Schumacher 2018), the previous simulation was repeated with two additional convective profiles: shallow and deep following Thorncroft et al (2008). These are illustrated in Fig. 6.5.b.. The results will also serve for comparison with the
findings of Thornicroft et al. (2008). Heating rates of the shallow and deep convective profiles are 2.71 K·day\(^{-1}\) at \(\sigma = 0.79\), and 2.09 K·day\(^{-1}\) at \(\sigma = 0.50\), respectively. The functional form of these profiles is given by:

\[
H_0 = \left(\frac{1}{\pi} - \frac{12}{\pi^3} + \frac{48}{\pi^5}\right)^{-1} \sigma^4 \sin (\pi\sigma)
\]

for the shallow heating profile, and:

\[
H_0 = \frac{\pi}{2} \sin (\pi\sigma)
\]

for deep convective profile. The atmospheric response to these profiles is presented next.

\(\text{i. Shallow convective heating profile}\)

Simulation results for the shallow heating profile are shown in Fig. 6.10. The horizontal evolution of relative vorticity is considered at the level of \(\sigma = 0.60\) to compare with the previous case, despite the forcing being initially more intense at lower levels (c.f. Fig 6.7.e). Analogous to the previous run, an adjustment takes place by day 1, but in this case, the horizontal structure is dominated by a mid-level ridge resulting from the maximum in diabatic heating below this level. On the following days, the propagation of this ridge sets up a ridge-trough configuration, showing a weak pattern by day 4, and by day 6, the horizontal structure is completely unorganized.

The vertical structure (central column of Fig. 6.7) shows a weaker development at the jet level than for the previous case. The shallow heating triggers a strong trough at low levels (due to the initial heating profile) together with a mid-level ridge, also characterized by a tilt in the vertical (see Fig. 6.7.h). However, the ridge and the trough quickly decay despite having the same column-averaged heating. The meridional wind in the stratiform run, reaches an amplitude of 1.20 ms\(^{-1}\) at 600 hPa, while the shallow profile barely reaches 0.40 ms\(^{-1}\) at this same time and
level. This results from the initial wind response to the heating imposed, resulting in different relationships with the mid-level jet over the EPAC (see Fig. 6.3.a-b for reference). Examples of these response are observed in the vertical structures on day 1 in Fig. 6.7.a and e. While at 600 hPa the stratiform profile produces a meridional wind of $3.2 \text{ ms}^{-1}$, the shallow convective produces $1.2 \text{ ms}^{-1}$, suggesting a weaker interaction at the level of the jet.

Overall, a shallow heating perturbation results in a weaker EW response along the mid-level jet in comparison with the stratiform heating profile. The perturbation structures resemble weak EWs over the EPAC, but their quick decay suggests that shallow convection is less efficient at triggering EWs. This is due to the less impact of the atmospheric response (and associated circulations) at the jet level to the initial heating.

**ii. Deep convective heating profile**

For completeness, experimental runs initialized with a deep convective profile were also performed. As in previous cases, the results of the perturbation runs are shown at the level of $\sigma = 0.60$, which, in this case, is below the level of maximum heating (at $\sigma = 0.50$). The horizontal evolution of relative vorticity is shown in Fig. 6.11. This is very similar to that seen with the stratiform convective profile but weaker (c.f. Fig. 6.6). After the first day of adjustment, the EW system generated by this heating profile is smaller and weaker when compared with the stratiform profile (see Fig. 6.6). By day 4, the system is weak and still coherent, but by day 6 this is weaker. We infer that the growth of EWs is weak and limited, again given by the atmospheric response at the jet level from the initial heating profile, similar to the previous case. This is clearly observed on the vertical structure (right column of Fig. 6.7), where maximum amplitudes in meridional wind are at 700 hPa below the jet level (about 600 hPa), thus showing less interaction.
In summary, model runs with different heating profiles over the EPAC-IAS region show EWs are also triggered, but the type of heating profile determines the general structure and evolution of the EWs in the region. The different heating profiles drive an initial trough (except the shallow heating), that impact the atmospheric response, particularly at the jet level. Therefore, the general structure of EWs is sensitive to the response to the heating profile. When compared with results of Thorncroft et al. (2008), the sensitivity impacts mostly on the amplitude of EWs, whereas over the EPAC it also impacts structure. These differences are explained mostly by differences in the mean state, in particular by the wind field and PV distributions.

We now proceed to explore where is the most likely region where EWs can be triggered, by performing sensitivity analysis to location of the heating, again following the approach of Thorncroft et al (2008).

b. Sensitivity to location of heating

Previous sections described the effect of convective heating on a single point over the EPAC-IAS region (specifically in the Panama bight region at 7°N, 78°W) based on the suggestion that convective forcing could be a triggering mechanism for EWs (see Zuluaga and Houze (2015), and Rydbeck et al. (2017)). However, it should be noted that convection and the associated heating occur not only over the Panama bight, but also over a wide region covering oceanic and land regions over the tropical EPAC and IAS. (see Fig. 6.1.d, and also Zuluaga and Houze (2015) in their Fig. 3). Motivated by this, the impact of heating in different locations on the EW response over the EPAC-IAS region was investigated. This will help identify the most sensitive location to force EWs over the EPAC-IAS region on the basic state. For this, a large number of simulations were performed over the region enclosed between 5°S and 35°N, and from 50°W to 120°W at
the spaced interval of about 3° in latitude and longitude. This region can be observed in Fig. 6.12, enclosed by the green box.

Figure 6.12 shows the results of these simulations by using an influence function for the three different profiles. We defined this influence function as the variance of relative vorticity at \( \sigma = 0.60 \) on day 4 of the simulation over a target area of 0°N to 20°N, and 75°W to 105°W (shown in red in the same figure). Additionally, this figure shows the zonal winds (in ms\(^{-1}\)) in contours, with warm colors indicating westerlies and cold colors easterlies.

Despite perturbing a wide region, we observe a peak region around 12.5°N, 70°W extending from the tropical EPAC trough Central America and into the Western Caribbean. This is located along the mid-level jet over the northern region of South America over the Caribbean Sea, ranging in latitude from 5°N to 20°N. This suggests that the best (or most sensitive) location to force EWs over the EPAC-IAS region (under a dry basic state) would be around 12.5°N, 70°W. This region is located northeast of the initial point in the basic runs, as observed from the zonal wind contours. Also, we observe that the influence function is much stronger for the stratiform heating profile (Fig. 6.12.a) when compared with shallow (Fig. 6.12.b) and deep (Fig. 6.12.c) convection. Basically, this would be explained by differences in the heating profile and the stronger interaction with the mid-level jet, as previously discussed.

While this result suggests that the best location to initiate EWs in the EPAC is along the mid-level jet, in the Western Caribbean actual observations do not show frequent convection in this location. This is possibly related to the lack of convection in the mean state (see Fig. 6.1.d) and cool SSTs. It is worth mentioning that close to this point, the only region with convection is
the that over the Maracaibo lake, to the east of Colón peak over northern South America (Molinari et al. 1997; Zuluaga and Houze 2015) where convection often develops.

To illustrate differences and similarities in the atmospheric response to different triggering locations, we explored the atmospheric response (trough relative vorticity) from the stratiform convective heating profile (as reference) at $\sigma = 0.60$ for three different points. The first location was chosen based on the most efficient region at 12.5°N, 70°W; the second based on TD-filtered OLR over the Atlantic at 10°N, 55°W (not shown), and the third far from the latitude of the mid-level jet at 0°N, 78°W. Figure 6.13 shows relative vorticity at day 4 (left column), considered as reference to compare with previous results in Fig. 6.6 at the same time. In addition, to compare structures over the EPAC, these were explored at later times (right column) on day 6 or 8 upon location.

Figure 6.13.a shows the response to heating at the location of most effective forcing at 12.5°N, 70°W. After four days (Fig. 6.13.a - left), the horizontal structure shows a trough over the Gulf of Mexico with tilting structures over the western Caribbean associated with barotropic growth. On this day, a second trough is initiated over the Panama bight, with a well-defined wave train. These characteristics show similarities with the EWs forced by heating at the reference location at 7°N, 78°W in Fig. 6.6. After six days (Fig. 6.13.a - right) a quite similar structure as in Fig. 6.6 is found over the EPAC. This confirms that as long as the convective forcing is along the jet, EW growth will take place, similar to the result of Thorncroft et al. (2008).

Figure 6.13.b shows the response to heating at 10°N, 55°W located over the Atlantic Ocean at day 4. This point was selected because of a maximum in TD-filtered OLR region (not shown), and to observe the evolution of heating associated with upstream Easterly Waves (e.g.
AEWs) over the EPAC-IAS region. The forced response moves westwards and the initial trough is maintained, forming a tilted structure over the Caribbean Sea. By day 8 (Fig. 6.13.b - right), the leading trough is over the continent with a structure similar to that obtained with forcing over the tropical EPAC in Fig. 6.6.d (on day 4). Heating over the Atlantic suggests that upstreaming systems can induce EWs over the EPAC-IAS region, as long as they get placed along the axis of the mid-level jet.

Lastly, a location south of the mid-level jet at 0°N, 85°W (Fig. 6.10.c - left) was considered to study the effects of a localized forcing far from the jet region. This figure shows that there is not a clear EW-development forced from this point, with an atmospheric response quickly decaying by day 4. By day 6, all the relative vorticity anomalies have already dispersed without any further influence over the EPAC highlighting the importance of the jet.

We can summarize this section by stating that the most sensitive location for EW initiation from finite amplitude transient heating is located along the mid-level jet at 12.5°N, 72°W. This is located over the northern part of South America and extending to the tropical Eastern Pacific trough Central America. Triggering along the mid-level jet shows an initial trough followed by a wave train over the EPAC, similar to the reference run in Fig. 6.6. The closer the triggering is to the jet, the bigger the response, particularly over the EPAC. If the triggering is at locations far from the mid-level jet, there is only limited EW growth. This is related to the extent to which the initial atmospheric response to the heating overlaps with the mid-level jet. These results also show that under idealized conditions, EWs over the EPAC can also be triggered by upstream EWs.
After this analysis on sensitivity to location, we proceed now to explore the sensitivity to the mean state.

c. Sensitivity to the mean state

In this section, we study the impact of different base states on the convective response. For this purpose, we selected the Madden-Julian Oscillation (MJO) in its easterly and westerly phases as base states, to show realistic contrasting conditions. Previous research has shown that the MJO modulates the TC and EW activity over the EPAC-IAS region (Molinari et al. 1997; Aiyyer and Molinari 2008; Crosbie and Serra 2014; Rydbeck and Maloney 2014; Rydbeck and Maloney 2015). During the westerly (or convectively active) phase of the MJO (hereafter MJOW), low-level positive relative vorticity and convergence are enhanced over the tropical EPAC, with EWs close to coastal areas. During easterly (or convectively suppressed) phase of the MJO (hereafter MJOE), these characteristics are reversed. Low-level positive relative vorticity anomalies and divergence are found over the EPAC. According to Aiyyer and Molinari (2008), zonally over the ITCZ. Therefore, this section addresses the response to transient heating introduced into these MJO states and compares the responses with previous literature. We hypothesize that during the MJOE phase, a stronger mid-level jet is located over the EPAC, leading to stronger dry response to triggering. During MJOW events, a weaker mid-level jet could result in weaker EWs. This is only considering that the model simulations are dry.

Figure 6.14 shows the distribution of PV (shaded) and the zonal winds at 600 hPa for the different MJO states in the T42L15 model. As a reference, Fig. 6.14.a shows the JJAS mean, previously described in Section 6.3. The distribution of PV and zonal wind on the MJOW phase is shown in Fig. 6.14.b. This shows enhanced PV west of 100°W over the EPAC when compared with
the JJAS mean. Also, positive zonal winds (absolute westerlies) are found at 10°N, with negative zonal winds at 20°N, 105°W closer to continental areas and parallel to the coast. During the MJOE phase, Fig. 6.14.c shows enhanced easterlies compared to the mean JJAS state, and a weaker and more zonal PV distribution at 13°N over the EPAC. Despite differences in direction, the amplitudes of the zonal winds exceed the 5 ms⁻¹ around 13°N, 100°W in all of these mean states.

Of importance for the growth and intensification of EWs over the EPAC are the meridional PV gradients and the associated PV sign reversals. However, these are not clearly defined from the horizontal distributions. For a better analysis, Fig. 6.15 shows the vertical cross-sections PV at 80°W, 90°W, and 100°W (as illustrated in Fig. 6.2) for the mean, MJOW, and MJOE – states. While the shading presents the distribution of PV, the dashed contours in black show the negative meridional PV gradients. The mean state is just shown as a reference in comparisons.

The MJOW as captured by T42L15 model (Fig. 6.15 - middle row) shows a more intense distribution of PV (this more evident at mid-levels at 100°W), with a distribution of negative meridional PV gradients similar to the mean state. For the MJOE state (Fig. 6.15-bottom-row), the negative meridional PV gradients are similar in shape and location, but stronger in magnitude. This arises basically from the intensification of the easterlies and the associated shear during the MJOE phase.

To explore the response to transient heating for these different MJO states in the T42L15 model, perturbation runs are run with heating at 7°N, 78°W (the same location used initially) with a stratiform convective heating profile. Results of these experiments are shown in Fig. 6.16 for day 4 on the same level of σ = 0.60, together with the zonal winds for each basic state. Again, for reference Fig. 6.16.a shows the JJAS mean state run. Figure 6.16.b shows the results of the
perturbation run with a MJOW state. It is important to mention that the convective anomaly associated with the presence of the MJOW is not included in these runs. The dry response to the prescribed heating shows a horizontal structure of relative vorticity similar to the basic reference state but shifted north (over land), consistent with the jet shifted north and with the centers of vorticity elongated (compare against Fig. 6.16.a). This shift is caused mainly by the presence of westerlies over the EPAC. The structure is also characterized by horizontal tilts associated with barotropic growth.

Figure 6.16.c shows the response to the prescribed heating on the MJOE state. The response forms more coherent, continuous, and more intense EW structures when compared with the basic state run. The marked horizontal tilts equatorward of the mid-level jet suggests barotropic growth resulting from the horizontal shear of the mean wind in this state. The horizontal structures on these three mean states present similar characteristics associated with their growth, however they differ regarding amplitudes and geographic location.

These simulations also show several differences on EWs between MJO states. The MJOE state supports more intense EWs as compared to MJOW, in association with stronger zonal winds and stronger sign reversals too, as observed in Fig. 6.15 (bottom row). Regarding location, during MJOW states the EWs are located further north, possibly associated with the a more poleward jet over the EPAC. The tilting and shape of the EWs over the EPAC are similar in the different mean states, highlighting again that dry dynamics is key for the growth of EWs over the EPAC-IAS region. However, differences in structures are also modified by latent heat within the waves, which is not included in these perturbation runs.
6.7. Summary and discussion

In this chapter, we have demonstrated that Easterly Waves can be triggered by localized forcing (represented by finite amplitude transient heating) in the vicinity of the mid-level jet over the EPAC-IAS region. In particular, over the Panama bight, heating from a stratiform profile forms an initial trough that develops an EW structure in 4 days, with a wavelength and propagation speed of 2000 km and 4.6 ms$^{-1}$, respectively. The resulting horizontal structure and wave characteristics are in general agreement with results documented in previous studies of EWs in this region (Serra et al. 2008; Serra et al. 2010; Rydbeck and Maloney 2014; Rydbeck et al. 2017).

These results reinforce the idea that over the EPAC, EWs can have genesis in situ (Serra et al. 2010; Rydbeck and Maloney 2017), by explicitly showing that the EWs can be triggered by upstream heating. This approach suggests a new paradigm for studying the nature of EWs over the EPAC, as well as their variability on longer timescales.

Zonal mean energy fluxes showed that over the EPAC, the horizontal eddy momentum fluxes associated with barotropic processes are the main source for EW growth, in agreement with eddy kinetic energy calculations. In comparison, vertical and horizontal heat fluxes show weak values, suggesting that baroclinic processes are secondary over the EPAC. When compared with momentum fluxes over Africa, values over the EPAC are weaker, in particular, the values associated with barotropic processes over the African region exceed those over the EPAC by about a 5:1 ratio.

In this study, convection is only considered as a triggering mechanism, and the interactions and feedbacks between EWs and convection (e.g. Janiga and Thorncroft 2016) are not considered here. Therefore, these simulations show amplitudes below those reported from
studies using reanalysis. This is expected given that the contribution from convection and the associated latent heat is absent, as Serra et al. (2010) and Rydbeck and Maloney (2015) have suggested.

Sensitivity tests showed that the EW evolution and amplitudes are highly impacted by the heating profile. The stratiform profile showed the strongest response while the shallow profile showed the weakest response. This is expected given the vertical gradients in the heating profiles (Hoskins et al. 1985), which drive the intensity of the initial trough.

Sensitivity analysis to the location of the initial heating shows that the most sensitive location for EW initiation is found along the mid-level jet, particularly at 12.5°N, 72°W. This region is found over the northern part of South America and extends into the EPAC trough Central America. This preference is because the region is located along the mid-level jet, where the dynamical conditions for instability, satisfied. EWs generated along the mid-level jet can impact continental regions during their development stage within 4-6 days (depending on location) and then potentially initiate an EW outbreak over the EPAC. At locations where dynamical conditions are not met, the transient heating cannot promote the growth of EWs, as observed in an example location south of the mid-level jet.

By varying the mean state, particularly focusing on the MJO in its westerly/easterly phases, results showed a similar response to heating. This is characterized by a wave train of the same length, tilt, and shape over the EPAC when compared against the mean state. Subtle differences were found in their intensity and location, associated with each particular phase. While the structure of EWs is more intense when the MJO is on its easterly phase, EWs are located more northerly during the MJO westerly phase. It is necessary to remark that the
amplitudes retrieved are inconsistent with observations due to the lack of coupling with convection.

Though previous studies suggest that EWs over the EPAC can be initiated \textit{in situ}, we have proven the hypothesis that triggering of EWs by transient heating is possible. The model confirms that EWs can be triggered by finite-amplitude heating along the mid-level jet over the EPAC-IAS region. EW growth is best achieved by the stratiform heating profile, which produces vorticity at mid-levels due to the vertical gradients in the heating profile. Though several other mechanisms for the genesis of EWs have been proposed, dealing mostly with the barotropic instability within the region, it is necessary to say that these are not mutually exclusive and each of them may be important given different conditions. Also it must be recognized that this is only a pathway through which EWs can be generated over this complex as it is the EPAC.
Figure 6.1. Climatology of the mean state over the EPAC-IAS region. Wind field and its magnitude (shading, in ms$^{-1}$), and streamfunction (contours in 10$^6$ m$^2$ s$^{-1}$) at a. 600 hPa, and b. 950 hPa. Wind vector is indicated for reference. c. Potential vorticity (in PVU) at 600 hPa. Mean calculated for JJAS during the years 1980-2015. Data source from ERA-Interim at 0.5° resolution. d. OLR (Wm$^{-2}$) for JJAS 1980-2015. Data source from NOAA.
Figure 6.2. Horizontal maps at 600 hPa (σ=0.60) of a. Wind field magnitude (shades, in ms⁻¹), direction (arrows), and streamfunction (in contours, 10⁶ m²s⁻¹); and b. Potential vorticity (in PVU) over the Eastern Pacific - Intra Americas Seas from the T41L15 model using ERAI reanalysis during JJAS 1980-2015. Vertical purple lines in a. show cross sections in Figs. 6.3 and 6.4.
Figure 6.3. Comparison of the mean state winds between the model T42L15 (above) and ERAI reanalysis (below). Cross-sections show zonal wind (shades, in ms$^{-1}$) and meridional wind (contours, in ms$^{-1}$) from east (right) to west (left) at: a, d. 80°W; b, e. 90°W; and c, f. 100°W. Data source from ERAI. Mean calculated for JJAS 1980-2015.
Figure 6.4. Potential vorticity (PV) comparison of the mean state between the model T42L15 (above) and ERAI reanalysis (below). Cross-sections show total PV (shades, in $10^{-1}$ PVU) and meridional PV gradients (contours) from east (right) to west (left) at: a, d. 80°W; b, e. 90°W; and c, f. 100°W. Negative meridional PV gradients are in black (dashed), and positive in pink (continuous). Data source from ERAI. Mean calculated for JJAS 1980-2015.
Figure 6.5. Triggering location and heating profiles used. a. Center point (7ºN, 78ºW) where these heating profiles were used (in black) and radius of the heating (7.5º). b. Stratiform convective (in grey), deep convective (in green), and shallow convective (in blue), analogous to Thorncroft et al. (2008). The vertical coordinate is $\sigma \times 1000$, approximately the same value of pressure in hPa. Horizontal axis shows normalized magnitude, with each profile integrating to 1 in the vertical.
Figure 6.6. Relative vorticity (shades, in $10^{-6}\text{ s}^{-1}$) at $\sigma = 0.60$ for days 1 - 6 for the response to a *stratiform convective* anomaly at $7^\circ\text{N}, 78^\circ\text{W}$. Contours show the zonal wind (in ms$^{-1}$) at the same level with warm colors referring to westerlies and cold colors to easterlies. Purple tilted line on Day 4 is a reference for vertical cross sections in Fig. 6.7, while the black box shows the area used for flux calculations in Fig. 6.8.
Figure 6.7. Vertical cross section of meridional wind (in ms\(^{-1}\)) across the line in Fig. 6.6.d for days 1 – 4 for the three different heating profiles. Contours are every 0.20 ms\(^{-1}\) for stratiform and shallow profile, and 0.10 ms\(^{-1}\) for deep convective, with negative contours dashed in blue. The vertical coordinate is \(\sigma \times 1000\), which is approximately the same value of pressure in hPa. Dashed line in d. is to illustrate the tilting.
Figure 6.8. Zonal-mean eddy diagnostics for EWs over the EPAC calculated from day 1 to day 6 which grows on the basic state from Fig. 6.1 within the region 110°W to 80°W.  

- **a.** Eddy kinetic energy $\frac{1}{2}(\bar{u}'u' + \bar{v}'v')$ (in shades, m$^2$s$^{-2}$), and mean zonal wind (contours, in ms$^{-1}$).  
- **b.** Horizontal momentum flux $u'v'$ (shades, m$^2$s$^{-2}$) with mean zonal wind (contours, in ms$^{-1}$).  
- **c.** Horizontal eddy head heat flux $v'T'$ (shades, in K ms$^{-1}$) and mean meridional wind (contours, in ms$^{-1}$), and  
- **d.** Vertical eddy heat flux $\omega'T'$ (shades, in K hPa hr$^{-1}$) and mean vertical velocity omega (contours; Pa s$^{-1}$). The overbar denotes a zonal mean, while primes denote deviation from the zonal mean. Negative contours values are dashed with zero contour bold.
**Figure 6.9.** Zonal-mean eddy diagnostics for EWs over **Africa** calculated from day 1 to day 6 which grows on the basic state within the region 30°W to 10°E. 

- **a.** Eddy kinetic energy $\frac{1}{2}(u' u' + v' v')$ (in shades, m$^2$s$^{-2}$), and mean zonal wind (contours, in ms$^{-1}$).  
- **b.** Horizontal momentum flux $u' v'$ (shades, m$^2$s$^{-2}$) with mean zonal wind (contours, in ms$^{-1}$).  
- **c.** Horizontal eddy head heat flux $v' T'$ (shades, in K ms$^{-1}$) and mean meridional wind (contours, in ms$^{-1}$), and 
- **d.** Vertical eddy heat flux $\omega T'$ (shades, in K hPa hr$^{-1}$) and mean vertical velocity omega (contours; Pa s$^{-1}$). The overbar denotes a zonal mean, while primes denote deviation from the zonal mean. Negative contours values are dashed with zero contour bold.
Figure 6.10. Relative vorticity (shades, in $10^{-6} \text{s}^{-1}$) at $\sigma = 0.60$ for days 1 - 6 for the response to a shallow convective anomaly at $7^\circ \text{N}, 78^\circ \text{W}$. Contours show the zonal wind (in $\text{ms}^{-1}$) at the same level with warm colors referring to westerlies and cold colors to easterlies.
Figure 6.11. Relative vorticity (shades, in $10^{-6}$ s$^{-1}$) at $\sigma = 0.60$ for days 1 - 6 for the response to a deep convective anomaly at 7°N, 78°W. Contours show the zonal wind (in ms$^{-1}$) at the same level with warm colors referring to westerlies and cold colors to easterlies.
Figure 6.12. Influence function for a. stratiform, b. shallow, and c. deep – convective anomalies at 600 hPa, showing the most sensitive location for the initiation of EWs over the EPAC-IAS region that have high amplitude on day 4 in the red box. Contours show the zonal wind (in ms\(^{-1}\)) at each level with warm colors referring to westerlies and cold colors to easterlies. Green box shows the “bombed” region.
Figure 6.13. Examples of $\sigma = 0.600$ relative vorticity response (shades, in $10^6 \text{s}^{-1}$) for different initial points. Heating is convective stratiform centered on a. 12.5°N, 70°W; b. 10°N, 55°W; and c. 0°N, 78°W. Contours show the zonal wind (in ms$^{-1}$) at each level, with warm colors referring to westerlies and cold colors to easterlies.
Figure 6.14. Potential vorticity (shades, in $10^{-1} \text{ PVU}$), and zonal wind (contours, ms$^{-1}$) at 600 hPa for different mean states: a. JJAS, b. MJOW, and c. MJOE - from the T42L15 model.
Figure 6.15. Vertical cross-section of PV (shades, in $10^{-1}$ PVU), and its meridional gradient (contours) at: 80°W(right column), b. 90°W(center), and c. 100°W (left column) - as marked in Fig. 6.2.a - for different states: mean state (upper row); MJOW state (middle row); and MJOE state (bottom row). PV retrieved from the T42L15 model using ERAI reanalysis. Mean calculated for JJAS 1980-2015.
**Figure 6.16.** Relative vorticity (shades in $10^6 \, \text{s}^{-1}$) at $\sigma = 0.60$ for day 4 (Iter. 16) for the response to a stratiform convective anomaly at $7^o\text{N}, 78^o\text{W}$ with different basic states: **a.** Basic state JJAS (for reference), **b.** MJO West, and **c.** MJO East. Contours show the zonal wind (in ms$^{-1}$) at the same level with warm colors referring to westerlies and cold colors to easterlies.
Chapter 7. Summary and future work

7.1. Introduction

This dissertation has addressed several aspects regarding Easterly Waves (EWs) over the tropical Eastern Pacific (EPAC) and the Intra Americas Sea (IAS) region. This included a review of the mean state over the EPAC, their track, structure, evolution, and genesis; as well as adjacent topics associated such as the role of Kelvin Waves and interdecadal variability. Together, these features provided a thorough analysis of EWs in this region, and allowed the description of a proposed life-cycle for EWs over the EPAC in four different phases:

1. Genesis
2. Early development
3. Intensification
4. Weakening

Figure 7.1 shows a schematic diagram of the approximate location where each of these phases occurs. Before addressing each of them here, we first present a summary of the mean state that was discussed in section 2. This will be followed by a summary of the EW phases in section 3. Then, adjacent topics in this research are presented in section 4. Lastly, avenues for future work presented in section 5.

7.2. Summary of the mean state

The mean state showed a PV strip at low levels that extends from the IAS region into the EPAC, rising from east to west, and linking the CLLJ and the mid-level jet over the EPAC. This distribution of PV is characterized by a maximum located on the cyclonic side of the mid-level jet, and a minimum poleward of this.
Over the EPAC, at the level of the jet (600 hPa), the distribution of PV satisfies the necessary conditions for barotropic instability: the Charney-Stern condition given by $PV_y < 0$, as well as the Fjortoft condition, associated with $u < 0$ where $PV_y < 0$. Together these conditions suggested potential for barotropic growth of EWs over the EPAC region. By exploring $\theta$ gradients close to the surface, we found weak meridional gradients of $\theta$ at low levels, with a magnitude of 3 K in 12° in latitude, this is $2.27 \times 10^{-3}$ K/km. When compared with West Africa, the meridional $\theta$ gradients are given as 16 K in 12° in latitude, this is $12.12 \times 10^{-3}$ K/km. The much weaker gradients in the EPAC region suggested a more minor role for baroclinic growth of EWs over the EPAC.

Important in the setting up of the dynamical conditions associated with the Charney-Stern conditions, are the sources of positive and negative PV anomalies. The sources of positive PV anomalies over the EPAC are associated mainly with two different populations of convection: stratiform and shallow (Zuluaga and Houze 2015; Huaman and Schumacher 2018). The stratiform heating is key for the creation of PV at mid-levels (showing similar values to those observed in the mean PV), suggesting that this type of convection is the main driver of PV at 550 hPa - 600 hPa mostly over 10°N – 20°N over the EPAC, while at low levels, PV is created from shallow convection below 800 hPa over the ITCZ between 5°N - 10°N.

We provided evidence that suggests that sources of negative PV anomalies may result from dry convection over the western side of the mountain regions of the Sierra Madre. This is particularly evident in May during the dry season, when convection produces a well-mixed boundary layer leading to minimum in PV. For JJAS, the minima are still present, though it is weaker. While dry convection may also have a role, it should be recognized that friction may also contribute over mountain regions leading to the destruction of PV. This last point also leads to a
research topic to consider in future work. Arguably, the processes over land may impact the development of EWs over the ocean. This may pose a predictability issue on the land-ocean interactions and their impact on weather systems over the EPAC.

7.3. Summary of EW phases

**Phase 1: Easterly Wave Genesis**

Evidence has been provided to suggest that convection over the region of Central America and the Panama bight, close to the mid-level jet, can trigger EWs in the EPAC. A simple approach considering variance of synoptic-timescales associated with EWs, suggested an origin close to Central America and a weak link with systems over the Atlantic. In particular, there is little indication that EWs could have an origin in the Caribbean and tracking into the EPAC. This provided a first evidence that most EWs over the EPAC are generated “in situ”. While Frank (1972) hypothesized that tropical storms over the Atlantic had a prominent role on the EPAC, this seems to be inaccurate.

By using an idealized model, it was demonstrated that EWs over the EPAC can be triggered by localized forcing, this represented by finite amplitude transient heating, in the vicinity of the mid-level jet over the region. A control run with transient heating over the Panama bight showed that heating characterized by a stratiform profile forms an initial mid-level trough that develops an EW structure downstream within 4 days, with a wavelength of about 2000 km and propagation speed of about $4.6 \text{ ms}^{-1}$. The wave characteristics are in general agreement with results documented in previous studies of EWs in this region although the wavelength is on the short side of the spectrum.
These results reinforce the idea that over the EPAC, EWs can have genesis in situ (Serra et al. 2010; Rydbeck et al. 2017), by explicitly showing that the EWs can be triggered by upstream heating. This approach suggests a new paradigm for studying the nature of EWs over the EPAC, as well as their variability on longer timescales.

Sensitivity tests showed that the EW evolution and amplitudes are highly impacted by the nature of the heating profile. The stratiform profile showed the strongest response while the shallow profile showed the weakest response. This is expected given the vertical gradients in the heating profiles (Hoskins et al. 1985), which drive the intensity and structure of the initial trough. The more this overlaps with the mid-level jet and associated PV gradients, the greater the downstream response.

Sensitivity analysis to the location of the initial heating shows that the most sensitive location for EW initiation is found along the mid-level jet, particularly around 12.5°N, 72°W on the southern-central region of the Caribbean Sea. This region is found over the northern part of South America and extends into the EPAC trough Central America. This preference is because the region is located along the mid-level jet, where the dynamical conditions for instability are satisfied.

By varying the mean state, particularly focusing on the MJO in its westerly/easterly phases, results showed a similar response to heating. This is characterized by a wave train of the same length, tilt, and shape over the EPAC when compared against the mean state. Subtle differences were found in their intensity and location, associated with each phase. While the structure of EWs is more intense when the MJO is in its easterly phase, EWs are located more north during the MJO westerly phase. The PV gradients are weaker in the easterly phase of the
MJO; however, it is the more intense zonal winds and their associated meridional shear what drives this intensity. Along the same line, it is necessary to remark that the amplitudes retrieved are inconsistent with observations due to the lack of coupling with convection in the dry model.

**Phase 2: Early development of Easterly Waves**

EWs propagate following the negative meridional PV gradients, oriented parallel to the western coast of Mexico and Central America. When considering the structure of EWs along this path, regression analysis in chapter 3 shows that EWs are seen to have structures characterized by a propagation speed of about 5.0 ms\(^{-1}\), a wavelength of about 2900 km, and a period of about 6 days. Their distinctive horizontal tilts against the shear, is suggestive of a barotropic growth mechanism. In the vertical, their structure reaches 300 hPa. At low levels, they are characterized by a weak eastward tilt with height likely associated with weak baroclinic growth.

The relative contributions of barotropic and baroclinic development were confirmed from the idealized modeling. When considering zonal mean energy fluxes, the horizontal eddy momentum fluxes associated with barotropic processes are seen to be the main source for growth of EWs. This was also in agreement with eddy kinetic energy calculations performed. In comparison, vertical and horizontal heat fluxes showed a weaker contribution, suggesting that baroclinic processes are secondary over the EPAC. When compared with momentum fluxes over Africa, values over the EPAC are weaker, in particular, the values associated with barotropic processes over the African region exceed those over the EPAC by about a 5:1 ratio. Considering values associated with baroclinic processes, these are in the order of about 10:1. Regarding amplitude (at the level of maximum heating response) these are about ~1.7:1.
From regressions, the evolution of EWs showed in the horizontal, a structure characterized by the presence of incipient convection north of the Panama bight consistent with genesis over that region. After two more days, the EW structure is characterized by a well-defined EW trough with convection within it. This then propagates towards the northwest, is characterized by a northeast-southwest horizontal tilt (as previously mentioned), with convection in phase with the trough. In the vertical, EWs present a tilt mostly below 400 hPa, and above, an out-of-phase westward tilting with height. This last resulting from the convective outflow aloft.

A more detailed analysis in the vertical, showed relative vorticity anomalies with a maximum at mid-levels (550 hPa - 600 hPa). This started at around 79.5ºW, consistent with the triggering hypothesis. During the westward propagation of EWs, the thermodynamical structure was dominated by warming at mid- to upper- levels (above 500 hPa and up to 200 hPa), increasing in magnitude as they propagated westwards (97ºW). Cooling at low levels was also evident, associated with evaporation of rain in the boundary layer. This result is also linked with peak vorticity at mid-levels and it is in agreement with previous results of Serra et al. (2010) regarding the classic Riehl structure (Riehl 1979), showing warming above and cooling below mid-levels, respectively.

**Phase 3: Easterly Wave intensification**

The further evolution of EWs showed the intensification of the mid-level vortex over the warm waters of the EPAC. At this stage, the horizontal structure kept showing a tilt against the zonal wind shear, continuing with the barotropic configuration. It was observed that over the Western Hemisphere Warm Pool (WHWP), low-level vorticity developed beneath the EW trough,
in association with deep convection. This led to a vorticity structure extending vertically through the troposphere.

The previous result suggested EW intensification could lead to a possible EW-TC transition, this being of about 13% on average. This ratio could be larger when considering all EWs over the EPAC transitioning to TCs.

**Phase 4: Weakening of Easterly Waves**

After the intensification phase, EWs show a weakening in convection and relative vorticity. This is likely associated with the lack of a supporting environment (such as meridional PV gradients, convection and moisture) as observed in the characteristics of the mean state. At this stage EWs can dissipate or possibly recurve into the continent (These EW-systems were not considered in the present work). If EWs recurve, they can possibly merge with midlatitude systems as noted by Corbosiero et al. (2009), or interact with the monsoon system over North America, being a possible source of moisture for the region (Adams and Comrie 1997; Pascale et al. 2019). Though it was not shown in this work, the influence of EWs on the monsoon of North America was also explored, but it will be a topic to present in future research.

### 7.4. Summary of adjacent topics in the research of EWs over the EPAC

**Kelvin Waves**

Convectively coupled equatorial Kelvin Waves (KWs) were considered to study their influence on the mean state over the EPAC, the modulation of EW activity, and as a particular case of the triggering hypothesis.

Over the EPAC KWs impact Central America and the Panama bight region with enhanced convection. Our analysis highlighted that KWs often generate a westward propagating ridge in
their wake, which is the opposite to what has been reported in other regions. Though not clear, it is hypothesized that this could be related to the reflected component of the Kelvin waves impacting topography, as previous studies suggest (Kleeman 1988; Gandu and Geisler 1991).

As a genesis mechanism, it was demonstrated that KWs can trigger EWs in only about 20% of the total cases. In particular, this is favored during the convective phase of the Madden-Julian Oscillation (MJO). During the convective phase, the MJO over the EPAC enhances convection in the mean state and increases relative vorticity in the environment (Maloney and Hartmann 2000). More important for EWs, is the enhancement of negative meridional PV gradients, necessary for their development and further intensification downstream. Under this enhanced convective environment, EWs can be generated from KWs in situ over the EPAC, without any precursor from upstream. When this occurs, EWs last about 6 days on average.

The main mechanism through which KWs could induce EW genesis during the background environment of the MJO, may result from the stratiform heating left behind after the KW passage. This will enhance the stratiform heating from the MJO background over the EPAC, particularly over Central America and the Panama bight. Then, with this enhanced stratiform heating, the triggering mechanism observed in the modeling approach would follow, leading to EW genesis. As this scenario is difficult to achieve in reality, that explains why only a few cases were found. Therefore, the stratiform heating from KWs under MJO would create a more unstable basic state acting as a trigger of EWs.
Interdecadal variability

When considering the relationship between dynamical and convective metrics associated with EWs over the EPAC, a strong relationship between them was found, with a correlation coefficient of $r=0.304$ (significant at 95%), which is considered strong in similar studies (Mekonnen et al. 2006).

EW activity in this region was found to be characterized by marked decadal variations. Reduced synoptic-scale activity was present during the period 1980-1997, and increased activity was present during 1998-2015. Changes in EW activity on interdecadal timescales appear to be associated with SST anomalies modulated by low-frequency oscillations, including most notably the Pacific Decadal Oscillation (PDO) and the Atlantic Multidecadal Oscillation (AMO). It was found that EW activity over the EPAC during the period 1980-2015 was increased when there is a -PDO index and a +AMO index, as occurred during the period of 1998-2015. The opposite occurred during the years 1980-1997, when a +PDO and a -AMO index occurred. The different correlations obtained between dynamical and convective metrics with the oscillation indexes showed a striking interdecadal relationship.

The physical mechanism for this is that during -PDO and +AMO (1998-2015), positive SST anomalies over the WHWP enhanced convection from the mean state over the region and this, in turn, invigorating EWs and the associated curvature vorticity. A detailed analysis showed that the thermodynamic fields were characterized by higher SSTs, increased moisture and more convection. The dynamic fields were characterized by an enhanced mid-level jet over the EPAC, larger PV meridional gradients and evidently, increased synoptic-scale variability. In particular, EW track analysis showed an impact on intensity. The opposite occurred during a +PDO and a
AMO (1980-1997), when this period was characterized by cold SSTs anomalies over the WHWP, diminished moisture and convection. The dynamic fields showed a diminished mid-level jet, a westward shift of a weakened meridional PV gradients and less intense synoptic-scale EWs, as the tracking showed.

By identifying that SSTs drive the variability of synoptic-scale activity, there exists the potential for predictability of EW activity on long timescales. These results provide a first step to understand the long-term variability of EWs over the EPAC. This can be helpful to discern changes on longer timescales, for example, those associated with climate change.

Current indices indicate a +AMO index and a -PDO index, suggesting that we may continue to observe enhanced EW activity over the EPAC.

7.5. Future work

Previous results open research questions associated with the different topics associated with the life cycle of EWs, as well as the adjacent topics. Some of these lines of research are presented next.

**For the mean state**

- A detailed analysis of the sources of negative PV during the summer over the Sierra Madre. This can be achieved from a modeling approach as that considered by Thorpe et al. (1993). By identifying how the distribution of PV can be affected, in particular the negative PV anomalies, can have an impact on variability of EWs, as well as consequences on predictability. Additionally, it is necessary to have more observations over the region to recognize and study in detail the sources of negative PV anomalies during the summer. This knowledge will improve our understanding of the mean state over the EPAC.
After the analysis of the mean state, it was recognized a region of enhanced heat over the Sierra Madre. It can be hypothesized that the origin of the mid-level jet over the EPAC is due to moist convection over the ITCZ and dry convection over the Sierra Madre, similar to that over Africa as demonstrated by Thorncroft and Blackburn (1999). In addition, the variability of the EPAC mid-level jet should be explored to recognize the impacts of this on EWs, similar to Leroux and Hall (2009).

For the different phases of EWs

- Given the similar downstream EW response for every forcing applied, we suggest that an analysis of the normal modes should be carried out to study the structures that the mean state supports, similar to Hall et al. (2006). It is expected that this analysis can further support the need of an initiation mechanism for EWs over the EPAC.
- A comprehensive analysis on the contribution from EWs to TCs over the EPAC. Future work consists in tracking EWs that develop in TCs to identify the EW characteristics and the environment influence developing versus non-developing EWs into TCs.
- Investigate the possible relationship of EWs with mid-latitude systems as well as their role as a source of moisture for the North American monsoon.

For adjacent topics

- For the relationship between Kelvin waves and EWs, it is necessary to investigate, in more detail, the impacts that stratiform heating has on the mean environment over the EPAC, as well as considering this response as a function of the propagation speed. For this last point, it is hypothesized that the KWs with slower propagation speeds could induce a more unstable mean stable.
- Results on interdecadal variability were constrained to the timespan from the reanalysis. Therefore, there is opportunity to explore EW activity further back in the 20\textsuperscript{th} century from longer datasets (e.g. Kaplan et al. 1998) to explore the climate variability of variables associated with EWs over the EPAC. This will be useful to understand past and future EW activity. Some key variables include SSTs, convection, and PV meridional gradients.

- This analysis could be extended to climatic scales explore how the environmental factors associated with EWs over the EPAC will be modified under different climate change scenarios.

This dissertation has explored various aspects of EWs during the rainy season over the EPAC. It has provided a new framework to understand the mean state, an analysis of the life cycle of EWs in different phases from a following-storm approach, as well as breaking paradigms about their genesis and link with upstreaming EW systems.

These findings can help to improve the forecast and formation of EWs. In addition, it shed light on the process associated with tropical cyclogenesis over the EPAC. By considering this knowledge on EWs and the effect of KWs, the predictability of severe weather over the tropical Americas can be enhanced from eastward and westward propagating systems.

Lastly, but nevertheless important, is the modulation of EWs on interdecadal timescales. This work has also provided new knowledge on longer timescales as well as predictability based on low-frequency oscillations.
Thus, the work achieves the overarching goal of *improving our knowledge and understanding of the nature of EWs over the EPAC-IAS region*, opening new avenues for future work necessary to increase our understanding of the physical process over the EPAC.
**Figures**

*Figure 7.1.* Schematic plot illustrating the different phases of EWs over the EPAC: 1. Genesis. 2. Early development. 3. Intensification, and 4. Weakening. In phase 1 (genesis), the cloud represents the localized forcing. In phase 2 (Early development) the tilted lines illustrate the barotropic growth, following the track over the EPAC (illustrated by the transparent purple line parallel to the coast). In phase 3 (Intensification) the circle represents EW intensification and possible transition to tropical cyclones over warm waters of the EPAC (illustrated by the thin blue line). In the weakening phase (4), the dotted lines illustrate their dissipation and possible interaction with other synoptic-scale systems.
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