The Mountain Gap Wind of the Isthmus of Tehuantepec and Its Impact on Tropical Cyclogenesis in the Eastern Pacific

Adolfo Lugo Rios

University at Albany, State University of New York, alugorios@albany.edu

The University at Albany community has made this article openly available. Please share how this access benefits you.

Follow this and additional works at: https://scholarsarchive.library.albany.edu/cas_daes_etd

Part of the Atmospheric Sciences Commons, and the Meteorology Commons

Recommended Citation
https://scholarsarchive.library.albany.edu/cas_daes_etd/2

License
This Dissertation is brought to you for free and open access by the Atmospheric and Environmental Sciences at Scholars Archive. It has been accepted for inclusion in Atmospheric and Environmental Sciences Theses and Dissertations by an authorized administrator of Scholars Archive. Please see Terms of Use. For more information, please contact scholarsarchive@albany.edu.
THE MOUNTAIN GAP WIND OF THE ISTHMUS OF TEHUANTEPEC AND ITS IMPACT ON TROPICAL CYCLOGENESIS IN THE EASTERN PACIFIC

by

Adolfo Lugo Rios

A Dissertation
Submitted to the University at Albany, State University of New York
In Partial Fulfillment of
the Requirements for the Degree of
Doctor of Philosophy

College of Arts and Sciences
Department of Atmospheric and Environmental Sciences

2022
ABSTRACT

In this dissertation we research two aspects of the mountain gap wind northerlies of the Isthmus of Tehuantepec in Southern Mexico. The causes of extreme gap wind northerlies, emphasizing the causes by tropical weather systems, like easterly waves troughs or hurricanes, and the impacts of an extreme gap wind events on tropical cyclogenesis.

Through the use of an index to diagnose the mountain gap wind strength, its evolution through the year and the mechanisms leading to extreme mountain gap winds was highlighted. The index revealed that the gap wind is strongly related to high surface pressure anomalies over the GoM throughout the year. We also observed that between June and August (summer), low surface pressure anomalies over the EPac also impact the gap wind, albeit to a lesser degree. Also, we detected that the gap wind has a significant diurnal cycle in summer.

The index diagnosing the mountain gap also allows the identification of the extreme gap northerlies. We found that extreme northerlies are strongest between November and January (winter) and weaken between June and August (summer). Nevertheless, throughout the year, most of the extreme gap northerlies were associated with fluctuations with periods between 2 and 10 days and a surface pressure gradient across the mountain gap established in association with high surface pressures over the Gulf of Mexico. The high pressures in the Gulf of Mexico were related to surface ridges of midlatitude origin.

In summer, when the midlatitude ridges retreat poleward and their pressure gradient diminishes over the Gulf of Mexico, other factors may trigger extreme gap northerlies. These factors consist of easterly wave troughs and tropical cyclones. When these tropical systems are located over the EPac, their low surface pressure establishes a pressure gradient that enhances the gap northerlies. When these systems are located over the GoM and their circulations are strong,
their associated northerlies may enhance the gap northerlies as well. Additionally, we found that the gap outflow in the Eastern Pacific merges with the circulation of the easterly wave and the circulation intensifies.

The case study of hurricane Patricia (2015) is used to explore the potential impacts of the gap northerlies on tropical cyclogensis. Hurricane Patricia was formed from a low-level vortex, at around 950 hPa, underneath an easterly wave trough over the Eastern Pacific on 0600 20th of October, 2015. Leading up to cyclogenesis, three distinctive features were involved: an EW trough, the gap outflow vorticity and the Intertropical Convergence Zone (ITCZ) vorticity anomalies. The latter two were more evident at low levels. The EW trough stalled over the Eastern Pacific and the Yucatan peninsula for two days from the 18th until cyclogenesis occurred on the 20th of October. While stalled, the trough southerlies advected ITCZ vorticity anomalies and moisture underneath the trough. At the same time the trough northerlies as well as a midlatitude ridge enhanced the gap wind. Eventually, the two low-level vorticity anomalies merged underneath the trough and formed a vortex which leads to cyclogenesis. Once the vortex was formed, the ITCZ vorticity anomalies kept merging with the vortex for about 48 hours, between 20th and 22nd. These vorticity anomalies were associated with convection in the trough, also released latent heat by condensation helping to intensify the vortex. The vortex achieved hurricane strength winds on the 22nd of October and kept intensifying for the next two days.

Through a numerical experiment consisting of filling the mountain gap of the Isthmus of Tehuantepec the role of the gap outflow on the cyclogensis of hurricane Patricia was studied. The experiment revealed that when the mountain gap is filled, the gap wind is suppressed, the vorticity underneath the EW trough is weaker, and cyclogenesis does not occur. We found that the gap outflow vorticity helps to form a low-level vortex underneath the trough. The main consequence
of the absence of the gap wind and its outflow is that the vorticity in the EPac at low levels is decreased and only the ITCZ vorticity anomalies remain in the EPac. In this situation, the ITCZ vorticity anomalies remain scattered over a broad vortex underneath the trough failing to merge around a sole vorticity maximum characteristic of TC vortices. Thus, the additional vorticity associated with the gap outflow contributes to establish a vorticity maximum for the ITCZ vorticity anomalies to merge around, leading to a developing vortex. Therefore, the vorticity of the gap outflow acts as a catalyst to trigger cyclogenesis for hurricane Patricia.
ACKNOWLEDGMENTS

I am very grateful to Professor Christopher Thorncroft, first for allowing me to give the opportunity to pursue a Ph.D in Albany, and second, for his infinite patience, dedication, and understanding through the course of this research.

I am also thankful to my committee members Professors Lance Bosart, Robert Fovell, and Ryan Torn for their comments that enriched this research. I want to thank all the faculty in the Department of Atmospheric and Environmental, Mark Beauharnois and Hector Martinez for their technical support to use the cluster KRATOS at the ASRC.

Also special thanks to Caro for her moral and unconditional support during all these years. To my parents, Adolfo and Carlota. To my aunt Sandra, my uncle Max and my cousins Diego and Andres for the additional support in the last months prior submission.

I am very thankful to Graciela Raga and Luis M. Farfan for being my mentors in Mexico. They introduce me to the atmospheric sciences

This research received generous funding from the National Council of Science and Technology (CONACyT) in Mexico (grant 280338/4408732 for graduate studies), the Fulbright-Garcia Robles (grant 15151005 for graduate studies), and from the Department of Atmospheric and Environmental Sciences (several teaching assistance positions).
TABLE OF CONTENTS

ABSTRACT ........................................................................................................................................... ii

ACKNOWLEDGMENTS ............................................................................................................................ v

TABLE OF CONTENTS ............................................................................................................................ vi

1. INTRODUCTION ................................................................................................................................. 1

   1.1 The Gap Wind of the Isthmus of Tehuantepec .............................................................................. 3
       1.1.1 Structure of the Isthmus of Tehuantepec Mountain Gap Wind ........................................... 3
   1.2 Tropical Cyclogenesis in the Eastern Pacific .................................................................................. 6
       1.2.1 Cyclogenesis of Hurricane Guillermo (1991) ...................................................................... 8
       1.2.3 Cyclogenesis of Hurricane Hernan (1996) ......................................................................... 9
       1.2.4 Cyclogenesis of Hurricane Patricia (2015) ......................................................................... 9
   1.3 Overarching Objectives ............................................................................................................... 10

Figures ..................................................................................................................................................... 11

2. HYPOTHESES AND APPROACH ....................................................................................................... 18

   2.1 Climatology of the Gap Flow During Summer ........................................................................... 18
   2.2 Impacts of the Gap Flow on the Formation of TCs ................................................................... 18
   2.3 General Approach ..................................................................................................................... 19

Figures ..................................................................................................................................................... 21

3. THE INFLUENCE OF SYNOPTIC-SCALE TROPICAL WEATHER SYSTEMS ON
   THE ISTHMUS OF TEHUANTEPEC GAP WIND IN SUMMER ..................................................... 22

   3.1 Methodology and Data ................................................................................................................. 23
       3.1.1 Representation of the Isthmus of Tehuantepec Gap Wind ................................................... 23
   3.2 Climatology of the Isthmus of Tehuantepec Gap wind .............................................................. 24
3.2.1 Annual cycle of the gap wind ................................................................. 25
3.2.2 Wavelet Power Spectrum ......................................................................... 26
3.2.3 Duration of Strong Northerlies ................................................................. 27
3.3 Role of Mean Sea Level Pressure Anomalies on the Tehuantepec Gap Wind .......... 28
3.4 Composite of Atmospheric Patterns During Gap Wind Events ......................... 30
  3.4.1 Spring Gap Wind Events ......................................................................... 31
  3.4.2 Summer Gap Wind Events ..................................................................... 32
  3.4.3 MSLP in the GoM and the Outflow Length During Gap Wind Events .......... 34
  3.4.4 Summary ............................................................................................... 35
3.5 Influence of the Diurnal Cycle on Summer Gap Wind Events ............................. 35
3.6 Influence of Synoptic-Scale Tropical Systems on Gap Wind Events ................... 37
  3.6.1 Examples of Tropical Systems Driving Gap Wind Events ......................... 38
  3.6.2 Composites of Tropical Systems at the Time of Gap Wind Events ............... 40
3.7 Summary and Concluding Remarks ................................................................ 45
Figures ............................................................................................................. 47

4. SYNOPTIC-SCALE ASPECTS OF PATRICIA CYCLOGENESIS ....................... 58
4.1 Background State of the Atmosphere between 8th and 20th October. .................... 59
4.2 Evolution of the Synoptic-scale elements. ....................................................... 61
  4.2.1 Formation of the Developing EW Troughs for Olaf and Patricia .................. 61
  4.2.2 Evolution of the developing trough for Patricia ........................................... 64
4.3 Summary and Concluding Remarks ................................................................ 66
Figures ............................................................................................................. 68
5. ANALYSIS OF THE DEVELOPING EASTERLY WAVE TROUGH FOR PATRICIA

5.1 Evolution of the Developing Trough for Patricia

5.2 The Low-Level Circulation Leading to Cyclogenesis

5.2.1 Horizontal Structure and Development of the Low-Level Circulation Leading to Cyclogenesis

5.2.2 Contributions of the ITCZ Vorticity to Cyclogenesis and Vertical Development of the Vortex

5.3 Summary

Figures

6. IMPACTS OF THE TEHUANTEPEC GAP WIND ON THE CYCLOGENESIS OF HURRICANE PATRICIA

6.1 The Gap Wind

6.1.1 Evolution of the Gap Wind

6.1.2 Spatial Structure of the Gap Wind

6.1.3 Spatial Extension of the Gap Flow and Interactions with the Vortex Leading to Cyclogenesis

6.1.4 Summary

6.2 Potential Effects of the Tehuantepec Gap Wind on Cyclogenesis

6.2.1 Impacts of the Gap Outflow Vorticity on the Cyclogenesis Region

6.2.2 Impacts of the Gap Outflow $\theta e$ on the Cyclogenesis Region
6.2.3 Evolution of the Mountain-Gap Parcel’s Vorticity and \( \theta_e \) Flowing into the Cyclogenesis Region

6.2.3 Summary

6.3 Summary and Discussion

Figures

7. **OPERATIONAL FORECASTS FOR THE GENESIS OF HURRICANE PATRICIA**

7.1 Operational Forecasts of Patricia Cyclogenesis

7.2 The Cyclogenesis of Patricia in the GFS and ECMWF Operational Forecasts

7.2.1 The GFS Forecast for Cyclogenesis of Hurricane Patricia

7.2.2 ECMWF Forecast for Hurricane Patricia

7.2.3 Discussion

7.3 Summary

Figures

8. **A NUMERICAL SIMULATION OF PATRICIA AND THE GAP FLOW**

8.1 Experimental Setup

8.2 Retrieval from the Control Run of the ERA 5 Atmospheric Features and Best-Track Characteristics

8.2.1 The Gap Wind and Hurricane Patricia, First Insights

8.2.2 The Structure of the Gap Wind

8.2.3 The Developing Trough, and the ITCZ Vorticity Strip

8.3 Hurricane-like Characteristics of the simulated TC

8.3.1 Vertical Structure of the Simulated TC
8.4 Summary ........................................................................................................................................... 130
Figures ....................................................................................................................................................... 131

9. THE SENSITIVITY OF PATRICIA’S CYCLOGENESIS TO THE GAP WIND OF
THE ISTMUS OF TEHUANTEPEC ........................................................................................................ 139

9.1 Model Setup for the Blocked Run ........................................................................................................ 139
9.2 Overview of the Control and Blocked Runs ......................................................................................... 140
9.3 Synoptic-Scale Aspects of the Control and Blocked Runs .................................................................... 142
  9.3.1 Synoptic-Scale Evolution of the Control Run ............................................................................... 142
  9.3.2 Synoptic-Scale Evolution of the Blocked Run ............................................................................. 143
  9.3.3 General Considerations .................................................................................................................. 145
9.4 Evolution of the Developing Vortex and of the Non-Developing Vortex ........................................... 146
  9.4.1 The Spatial Evolution of the Developing Vortex and the Non-Developing Vortex ..................... 148
  9.4.2 The Contributions of the Gap Wind to the Vortex Development ............................................... 151
9.5 Discussion and Conclusions ............................................................................................................... 155
Figures ....................................................................................................................................................... 156

10. CONCLUSIONS AND PROPOSED FUTURE WORK ....................................................................... 168

10.1 The Impacts of Tropical Weather Systems on the Gap Wind ............................................................ 168
10.2 The Cyclogenesis of Hurricane Patricia .............................................................................................. 169
  10.2.1 The role of the Gap Wind during the Cyclogenesis of Hurricane Patricia ................................... 172
  10.2.2 The importance of the ITCZ Vorticity Anomalies during the Cyclogenesis of Hurricane Patricia ..... 174
10.3 Future Work ........................................................................................................................................ 175
Figures ....................................................................................................................................................... 177
REFERENCES

180
1. INTRODUCTION

The low-level wind flow in the Isthmus of Tehuantepec in southern Mexico, between the Gulf of Mexico and the Gulf of Tehuantepec is strongly influenced by the mountain gap there (Figure 1.1). This narrow pass, 40 km wide and about 1.8 km depth channels the wind and frequently forms low-level flow with jet like characteristics that extends several hundreds of km into the Eastern Pacific (EPac). The intensity of the mountain gap wind experiences abrupt fluctuations that last from a couple of hours to several days. During winter the gap wind is due to cold air masses that form in the Great Plains extending over the Gulf of Mexico (GoM), reach the Isthmus of Tehuantepec and flush into the Gulf of Tehuantepec (Shultz, et al, 1998). In the Mexican region, these cold surges are called Tehuanos. During Summer, features that enhance the wind include easterly waves (EWs) (Farfán & Zehnder, 1997) and the position and intensity of the Azores-Bermuda high (ABH) (Romero-Centeno, et al, 2007). Also, the gap flow extends into the tropical cyclogenesis region in the Eastern Pacific (Romero-Vadillo, et al, 2007) where it may affect tropical cyclogenesis (Bosart, et al., 2016; Holbach & Bourassa, 2014 and, Zehnder, et al, 1999).

There are a number of potential impacts that the gap wind can have depending on its intensity and the time of year that motivate research on this flow, including:

- Formation of tropical cyclones (TC) (Farfán & Zehnder, 1997 and Zehnder, et al, 1999). This has been the subject of several recent studies (e.g. Holbach & Bourassa, 2014 and Bosart, et al, 2016), and is paramount to tropical cyclogenesis forecasting in the region;

- Modulation of convection in the region, which can be as far-reaching as the intertropical convergence zone (ITCZ) (Romero-Centeno, et al, 2007);

- Intense gap winds are a threat to passing ships in the Gulf of Tehuantepec (e.g. Hurd, 1929).
• Advection of pollutants from the oil refinery in the coastal town of Salina Cruz into the ITCZ modifies the microphysics processes in the clouds there (Baumgardner, et al, 2005);

• Abrupt changes in sea surface temperature (SST) (Trasviña, et al, 1995) and Ekman transport in the shallow layer of the sea close to the coast (McCreary, et al, 1989);

• Biological impacts such as, increase of phytoplankton production and fisheries due to Ekman transport (Liang, et al, 2009);

• The strong low-level flow provides energy to the wind farms near the coast of the Gulf of Tehuantepec (AMDEE, 2014).

Of all these points; the impact of the gap wind on the formation of tropical cyclones creates a unique configuration in which topography driven flows may contribute to cyclogenesis. Therefore, the dissertation aims, in order of importance to 1) understand the impacts of the Tehuantepec gap wind on tropical cyclogenesis considering a study case and 2) elucidate if tropical systems, e.g., EWs or troughs and hurricanes enhance the gap wind. These aims achieve to understand whether tropical disturbances enhance gap winds and whether tropical disturbances develop by enhanced gap winds. This will lead to a better understanding of the interactions between the gap flow and tropical disturbances benefiting the forecasting of both tropical disturbances (specifically cyclogenesis) and the gap wind itself.

Therefore, the overarching goal of this work is to understand the nature of the gap flow, its causes and the variability in Summer, and its impact on tropical cyclogenesis. The next sections present a literature review about the characteristics and structure of the gap wind and about the current studies of the impacts of the gap wind in cyclogenesis.
1.1 The Gap Wind of the Isthmus of Tehuantepec

The wind measured by weather stations at the town of La Venta (Fig. 1.1) are representative of the gap wind (Romero-Centeno, et al; 2003). The intensity of these surface winds has a strong seasonal variation with maximum intensity in December and January, a minimum in May and June and a secondary maximum in July (Figure 1.2 (a)). The gap wind and the mean sea level pressure (MSLP) gradient between the GoM and the EPac are closely related year-around. As the gradient increases, i.e., the pressure in the GoM is higher than in the GoT, the wind intensity increases and becomes predominantly northerly. When the pressure gradient diminishes, the wind intensity decreases and is less likely to be northerly (Romero-Centeno, et al, 2003). Therefore, strong gap wind are northerlies and weak gap wind have different direction (Fig. 1.2 (b)).

In Summer, despite the gap wind weakens and the direction of the gap wind is more variable (Romero-Centeno, et al, 2003) the northerlies are still frequent and account for ~50% of the time (Holbach & Bourassa, 2014). Also, the weaker gap winds in Summer, has a low-level outflow similar in shape and structure to that of the strong northerlies of December and January, but with diminished values (Holbach and Bourassa, 2014).

1.1.1 Structure of the Isthmus of Tehuantepec Mountain Gap Wind

The structure of the gap outflow has been widely described in the literature (e.g., Bourassa, et al, 1999, Chelton, et al, 2000a and 2000b, Holbach & Bourassa, 2014, Shultz, et al, 1998 and Steenburgh, et al, 1998) and its summarized next, based on one representative case. The maximum wind speed locates over the EPac and close to the shore (Fig. 1.3 (a)). The outflow turns anticyclonically and extends hundreds of kilometers into the EPac; although in the presence of low-pressure disturbances to the east of the mountain gap, the outflow may turn cyclonically (Bourassa, et al, 1999). The outflow is generally a divergent flow, flanked by small convergent
regions (Fig. 1.3 (b)). During summer, the low-level divergence associated with the gap wind tends to be associated with suppressed convection (Romero-Centeno, et al, 2007). Also, the outflow forms two vorticity patches flanking the maximum winds, an anticyclonic patch to its west and a cyclonic patch at its east. The negative patch is generally stronger than the positive one (Fig. 1.3 (c)), about double that of the positive patch, this asymmetry is caused by the orientation of the surface ridge triggering the gap wind over the Gulf of Tehuantepec. Thus, the structure of the gap outflow is consistent with that of a northerly jet.

The vertical structure of the gap wind is illustrated by one representative case on 20th October 2015 (Figure 1.4). The maximum wind in the mountain gap peaks at the 900 hPa level, with strong winds typically spanning up to the 800 hPa level. At times it extends up to the 700 hPa level (Steenburgh, et al, 1998) and is limited longitudinally by the mountain pass. Depending on the orientation of the gap inflow in the Gulf of Mexico, a northerly flow over the Sierra Madre is also present (Steenburgh, et al, 1998).

The time scale of the mountain gap northerlies varies from hours to days, indicating that synoptic structures impacting the surface pressure in the GoM or even the EPac influence the gap flow. Literature points to midlatitude systems (e.g. surface ridges), as the primary drivers of the gap wind at the synoptic scale (Steenburgh et al, 1998). Also, potentially EWS or other high-pressure systems in the GoM may drive the gap wind but this needs to be explored. The typical atmospheric configuration for a gap wind event triggered by a midlatitude system is exemplified by a case occurred on 13th March, 1993 and shown in Fig. 1.5.

The atmospheric configuration during this gap wind event consists of a surface ridge indicated by positive surface pressure anomalies (~12 hPa) over much of the GoM, blocked by the Sierra Madre (Fig. 1.5). Strong surface winds (15 ms⁻¹) occur along the pressure gradient over the
Isthmus of Tehuantepec. In turn the strong winds cause upwelling in the EPac and the SST decreases on the EPac (blue contours, Fig. 1.5). Note that the positive pressure anomalies are aligned with the topography indicate a strong influence of the topography not only on the gap wind itself but also on the surface ridge. The mechanism triggering the gap wind presented here is dominant from October to March when midlatitude systems reach the GoM, in Summer they retreat poleward. Hence other mechanisms different from midlatitude systems have to modulate the gap wind in the synoptic timescale in Summer. However, these mechanisms have not been studied in detail, and we propose to study these potential mechanisms.

Despite a comprehensive understanding of the synoptic-scale modulations of the gap wind in Summer is yet to be presented. It has been recognized that factors modulating the gap wind at synoptic timescales in the summer include the passage of EWs north of the Isthmus of Tehuantepec (Farfán & Zehnder, 1997, Zehnder, et al, 1999 and Molinari & Vollaro, 2000), and low-pressure anomalies to the south of the mountain gap, in the EPac (Bourassa, et al, 1999). Further evidence of the interaction of EWs with the gap wind was found by Serra, et al (2010), who showed that the wind across the Isthmus of Tehuantepec has significant spectral energy in the 3–6-day band, the same as EWs. However, the gap wind caused by these features are not as strong as the winter ones.

Despite the lack of a comprehensive understanding of synoptic variation of the gap wind it is understood why the gap wind in Summer weakens (Fig. 1.2 (a)). The weakening is due to the diminished pressure gradient between the GoM and the EPac when compared to Winter (Romero-Centeno, 2003). Nevertheless, the gap wind experiences some intensification during July due to an increase in the pressure gradient between the GoM and the EPac caused by the intensification and westward extension of the ABH (Romero-Centeno, et al, 2003, and Fig. 1.6).
In summary, at seasonal scales, the gap wind is modulated by the seasonal evolution of the meridional pressure gradient between the GoM and the EPac. At synoptic scales, the gap wind is primarily dominated by midlatitude surface ridges and potentially by weaker disturbances, like EWs close to the mountain gap.

1.2 Tropical Cyclogenesis in the Eastern Pacific

In this section we discuss in general terms the tropical cyclogenesis in the EPac and what is the current understanding of the impacts of the gap wind on cyclogenesis. Cyclogenesis in EPac occur close to the coast of Mexico becoming less frequent westward (Fig. 1.7; Allard (1984); and Gray (1968)). Climatologically, the region is characterized by low wind shear values in the 800-200 hPa layer (Fig. 1.8 (a) Maloney & Hartmann, 2000), has frequent convective activity due to the ITCZ (Molinari and Vollaro, 2000) and has SST greater than 28°C. All these characteristics favor the formation of TCs and make of the EPac a prolific hurricane basin.

Operational reports from the NHC (Avila, 1991, 1995, 2000, 2006 and 2009; Beven, 2004 and 2005; Blake, 2010 and 2013; Brennan, 2009; Franklin, 2002, Kimberlain, 2011; Knabb, 2008; Lawrence, 1994, 1999 and 2001; Mayfield, 1998; Pasch, 1996 and 2009; Rappaport, 1992 and 1998; Stewart, 2012) indicate that most of the TCs in the EPac originate from African EWs traversing Central America and developing in the EPac. Less frequently, reports indicate that weather disturbances in the Gulf of Tehuantepec evolve into TCs. Rare causes (in the sense that these causes are explicitly mentioned once or twice in these reports) point to local TC genesis, in which Kelvin waves or the MJO interacts with the ITCZ, monsoon troughs over Central America, or mesoscale complex systems close to the coast may trigger tropical cyclones.

At sub-seasonal scale, the MJO strongly influences the EPac environment for EWs to develop. Maloney & Hartmann (2000) observed that during the westerly phase of the MJO (phase-
2) convective activity occurs in the easternmost EPac, wind shear decreases and the probability of cyclogenesis increases (Fig. 1.8 (b,c)). This leads to prolific periods of TC activity when several TCs may form simultaneously (Fig. 1.9). During this phase of the MJO, incoming EWs intensify and develop in the EPac (Molinari and Vollaro, 2000a).

Additionally, it has been recognized that the orography in Central America and south Mexico helps to create flows favoring TC genesis. In this regard, it has been found that the Papagayo gap wind in Central America creates a negative meridional potential vorticity (PV) meridional gradient, satisfying the Charney-Stern instability criterion (Molinari, et al, 1997). This mechanism partially explains how an incoming EW can develop. Also, it has been suggested that this negative PV gradient is able to form an EW (Whitaker, 2020). Additionally, Zehnder (1991a and 1999) proposed that Equatorial perturbations interacting with the orography of Central America may spawn vortices on the lee side, which are hypothetical seedlings for TCs. Moreover, several study cases of the formation of individual TCs (e.g., Farfán and Zehnder (1997) and Bosart, et al (2016)) have recognized that the gap wind of the Isthmus of Tehuantepec may contribute to cyclogenesis. Thus, then, the orography limiting the EPac and the flows associated to it may enhance weather disturbances leading to cyclogenesis (e.g. Zehnder, 1991a and 1991b and 1999). Hence, is important to increase our understanding on the underlying mechanisms causing topography flows that impact positively on cyclogenesis to have better cyclogenesis forecasts in the basin. In the case of this research focus is given to the impacts of the gap wind of the Isthmus of Tehuantepec on cyclogenesis. The next section provides an insight of the effects of gap wind of the Isthmus of Tehuantepec on cyclogenesis based on induvial study cases.
1.2.1 Cyclogenesis of Hurricane Guillermo (1991)

The cyclogenesis of Guillermo, studied by Farfán and Zehnder (1997), is the first study case addressing the potential positive impact of the gap wind on tropical cyclogenesis. Guillermo was formed from a cyclonic circulation (cyan circle in Fig. 1.10) at 900 hPa to the south of the Isthmus of Tehuantepec on the leading edge of trough by 0600 2nd August (solid black line in Fig. 1.10).

They found that the low-level circulation leading to cyclogenesis was influenced by three low-level flows: 1) a southerly flow from the ITCZ (green arrow, Fig. 1.10 (c)), 2) a flow parallel to Central America from the Papagayo mountain pass gap (cyan arrow, Fig. 1.10 (b)) and 3) the Isthmus of Tehuantepec gap outflow (red arrow, Fig. 1.10 (c)) that intensified as the EW moved north of Central America (Fig. 1.10 (c)).

Through a numerical experiment it was concluded that this circulation does not form, and by extension cyclogenesis fails to occur if the gap wind is suppressed. Thus, they concluded that the orographic flows had a substantial role in the cyclogenesis of Guillermo.

1.2.2 Cyclogenesis of Hurricane Henriette (1995)

Cyclogenesis of Henriette in 1995 was similar to that of Guillermo, a low-level circulation to the south of the Isthmus led to cyclogenesis, but in this case, an incoming EW crossed over Central America in to the EPac (Zehnder, 1999). When the EW crossed over Central America, the gap wind intensified. Simultaneously, deep convection formed over the low-level circulation leading to cyclogenesis over the EPac. They speculated that enhanced surface heat fluxes associated with the Isthmus of Tehuantepec gap outflow were fundamental to develop this convection which in turn seemed to be important to form the TC. However, they did not study the relative importance of the gap outflow in the formation of Henriette.
1.2.3 Cyclogenesis of Hurricane Hernan (1996)

Molina, et al (2000b) described the cyclogenesis of hurricane Hernan. Its cyclogenesis was linked to an African EW (tracked at 700 hPa) that crossed Central America at 13°N. The circulation leading to Hernan was formed after the EW moved into the Pacific, and similar to the previous cases, an enhanced gap wind was observed.

Furthermore, Molinari, et al (2000b) conjectured that the gap wind may have contributed to create cyclonic vorticity close to the surface which then aligned with the mid-level vorticity of the incoming EW as the ultimate cause of cyclogenesis. At the same time, they considered also possible that the gap wind is a consequence of the developing of the trough indicating that both the gap wind and the EW are inexorably dependent on each other. This means that a gap wind can be triggered by the circulation of an incoming EW, in turn the gap wind vorticity may further increase the low-level vorticity of the EW favoring its development. Nevertheless, they considered that the gap wind contributes to increase the cyclonic vorticity underneath the trough regardless of whether the gap wind is caused by the trough or by other exogenous factors.

1.2.4 Cyclogenesis of Hurricane Patricia (2015)

The cyclogenesis of hurricane Patricia was thoroughly discussed by Bosart, et al (2016). They observed that it consisted of a complex interaction involving an EW and the gap wind. The EW formed a broad circulation over the Yucatan peninsula extending as far as the EPac (yellow contours in Fig. 1.13).

Cyclogenesis occurred in the center of a low-level (925 hPa) circulation formed by the gap wind outflow and the trough southerlies (inset in Fig. 1.11). Contrasting with the study cases discussed earlier, the gap wind was caused by the surface ridge associated with a midlatitude system in the GoM (blue contours in Fig. 1.11). Also, the enhanced latent heat fluxes associated
with the gap outflow could have increased the convection within the trough further contributing to
cyclogenesis.

These case studies strongly suggest that the gap wind contributed to form the low-level
circulations leading to cyclogenesis. However, with the exception of Farfán and Zehnder (1997) it
is unknown whether the gap wind is needed in these cyclogenesis cases or if it were to occur
regardless the gap wind. Nevertheless, it has been illustrated the importance of the gap wind on
cyclogenesis, but more work is needed to understand its role.

These study cases serve as a motivation to understand the role of the Isthmus of
Tehuantepec gap wind on cyclogenesis. Therefore, an overarching objective of this dissertation is
to understand the role of the gap wind on tropical cyclogenesis. To achieve this, we will study the
cyclogenesis of hurricane Patricia in detail to reveal what is the underlying mechanism of the gap
outflow during cyclogenesis. The overarching objectives are discussed next.

1.3 Overarching Objectives

Based on the presented individual study cases of cyclogenesis, we want to elucidate the
role of gap outflow of the Isthmus of Tehuantepec on the development of the low-level circulations
leading to cyclogenesis based on the case of hurricane Patricia. Also, we want to explore what
other synoptic factors different from midlatitude systems modulate the Isthmus of Tehuantepec
gap wind. Hence, the proposed overarching objectives of this study are:

1. To improve our knowledge and understanding of the nature of the gap flow and its
   modulation by synoptic systems of tropical origin during Summer.

2. To elucidate the extent to which the Tehuantepec gap flow impacts tropical
cyclogenesis in the Eastern Pacific and about the underlying mechanism based on the cyclogenesis
of hurricane Patricia.
By accomplishing these objectives, it will be clear what elements should be adequately resolved in operational models and what elements should be tracked down by forecasting agencies in the region specially for tropical cyclogenesis in the EPac.

This dissertation has the following structure. Chapter 2 presents the hypotheses and sub-hypothesis to be tested in the present study. Chapter 3 presents a climatology of the gap wind emphasizing the impacts of synoptic-scale elements on the gap wind including summer. Chapters 4 to 9 provide analysis of the cyclogenesis of hurricane Patricia and the role of the gap wind in it. Chapter 10 presents the conclusions and proposed future work.

The afore mentioned study of the cyclogenesis of Patricia is as follows, chapters 4 to 6 analyze the cyclogenesis of Patricia, including the potential impacts of the gap wind on it based on the ERA-5. Chapter 7 studies the potential impacts of the ITCZ vorticity anomalies on the cyclogenesis of Patricia based on operational forecasts. Chapters 8 and 9 explore the importance of the gap wind on cyclogenesis through a numerical sensitivity experiment.

**Figures**

![Figure 1.1. Topography. Salina Cruz (SCZ), La Venta (LV) and wind farms (crosses). Sierra Madre (SM) and Central America Mountains](image-url)
Figure 1.2 a) Monthly mean wind speeds at La Venta, error bars represent the $\pm 1\sigma$ from the mean. b) Frequency distribution of wind speed at la Venta for all winds (circles), only northerly winds (triangles) and for the remaining winds (squares). From Romero-Centeno, et al (2003). © American Meteorological Society. Used with permission.

Figure 1.3. Composite averages of a) the NSCAT 10-m wind field, b) wind divergence and c) relative vorticity over the Gulf of Tehuantepec for the 2-day period 19–20 Dec. 1996. The heavy line in the upper-left panel is the inertial path originating at the gap opening with the observed wind at 15.25ºN, 95ºW. From Chelton (2000a). © American Meteorological Society. Used with permission.
Figure 1.4. Vertical cross section at 17°N. Horizontal wind (vectors), wind speed (in kt, shading) and temperature (contour). Inset shows the horizontal wind at 850hPa (vectors) and windspeed. From Bosart, et al (2016).

Figure 1.5. Surface pressure anomalies with respect to the monthly climatology (every 4 hPa starting ±4hPa, solid contours for positive values and dashed for negative values), SST anomalies (every °C, starting at -1°C, blue contours) at surface in heavy solid (dashed) lines every 4 hPa starting at ±4hPa. Letters H and L denote the maximum and minimum surface pressure anomaly values.
Figure 1.6. Monthly climatology values of the Pressure at mean sea level pressure from NCEP reanalysis for 1948–2001 (Contours every 1 hPa). From (Romero-Centeno, et al, 2003) in hPa. © American Meteorological Society. Used with permission.

Figure 1.8 a) The 200-mb zonal wind minus 850-mb zonal wind during May–Nov 1979–95 for phase 2. Contours are every 5.0 m s\(^{-1}\). Negative contours in dashed. Positive values mean that 200-hPa winds are more westerly than 850-hPa winds. b) Number of hurricanes as a function of MJO phase for the EPac region during May–Nov 1979–95. Error bars represent 95% confidence limits. c) Composites of bandpassed 850-hPa wind anomalies and MSU precipitation anomalies during May–Nov 1979–95 for phase 2. Maximum vectors are 3.0 m s\(^{-1}\). Precipitation contours are at intervals of 0.6 mm day\(^{-1}\) starting at 0.3 mm day\(^{-1}\). Negative contours are dashed. Precipitation is spatially smoothed by a 1–2–1 filter. From Maloney & Hartmann (2000). © American Meteorological Society. Used with permission.
Figure 1.9 GOES visible image at 1830 UTC 19 August 1993 showing hurricanes Greg, Hillary and tropical storm Irwin in its formative stage. From Avila (1995). © American Meteorological Society. Used with permission.

Figure 1.10. 950 hPa Wind (vectors) and its relative vorticity (>0 s⁻¹ shading) for 1200 2nd, 0000 3rd and August (b), and 1200 3rd August (c). The thick line indicates the axis of the trough. Blank areas indicate topography. The blue circle indicates the cyclonic circulation lading to Guillermo. Red (green, cyan) arrow indicates the gap wind (southwesterlies, easterlies crossing Central America). Adapted from Farfán and Zehnder (1997). © American Meteorological Society. Used with permission.
Figure 1.11. Scheme of elements present during Patricia cyclogenesis. From Bosart et al (2016).
2. HYPOTHESES AND APPROACH

This dissertation covers two aspects of the gap wind, one is about what factors, in addition to surface ridges in the Gulf of Mexico modulate the gap wind. The other is about the impact of the gap wind on tropical cyclogenesis. Each of these aspects are discussed next together with testable hypotheses and an approach.

2.1 Climatology of the Gap Flow During Summer

The underlying mechanism enhancing the gap wind in the synoptic time-scale is the establishment of a relative steep pressure gradient between the Gulf of Mexico and the East Pacific (EPac). At synoptic scales and when surface ridges are absent in the GoM, like in summer we hypothesize that *the synoptic variability of the gap flow in summer is primarily modulated by the pressure gradient established by incoming tropical disturbances like EWs or hurricanes*, hypothesis I.

2.2 Impacts of the Gap Flow on the Formation of TCs

It is hypothesized that the gap wind can favorably trigger cyclogenesis in the following scenario represented by Fig. 2.1. The elements considered are:

1) A low-level (~950 hPa) circulation (indicated by a negative stream function anomaly, red contours) underneath a trough (not shown).

2) A gap wind outflow (vectors) vorticity (vertical hatching) overlaps with the circulation;

3) An upstream moisture or vorticity anomalies in the GoM (represented by a hatched oval).

We propose the following mechanisms for the gap outflow to promote cyclogenesis considering these elements:

- Dynamical mechanism: The positive vorticity (vertical hatching) of the gap wind outflow merges with the vorticity of the low-level circulation resulting in a vortex. The vortex helps
convective-scale vorticity anomalies to aggregate and form a nascent TC as described in Kilroy et al (2016), hypothesis II.

- Thermodynamical mechanism: the enhanced surface latent heat fluxes associated with the gap wind outflow contribute to increase the equivalent potential temperature close to the surface and help to develop convection increasing the low-level vorticity and to increase the equivalent potential temperature within the precursor circulation, hypothesis III.

- Transport mechanism: the gap wind outflow and the circulation are materially connected, in this situation the gap outflow may transport upstream moisture or vorticity anomalies into the circulation, hypothesis IV.

Regarding hypothesis IV, it has not been documented as part of the gap wind contributions to cyclogenesis; nevertheless, it has been recognized that the ingestion or entrainment of environmental air into precursor circulations is important to their development or lack of. For example, Dunkerton, et al (2009) highlighted that it is important to “isolate” the precursor circulation from incoming dry and/or cold air. Also, Brammer and Thorncroft (2017) discussed the importance of the advection of moist or dry air beneath the midlevel trough to determine the growth of the precursor.

2.3 General Approach

The first part of the dissertation presents a climatology of the mountain gap wind of the Isthmus of Tehuantepec in Summer and its causes. The second part provides an in-depth analysis of the impacts of the gap wind on tropical cyclogenesis in the EPac through the study case of hurricane Patricia including modeling. Each part is now introduced.
The climatology consists of studying the synoptic-scale variations of the gap wind in Summer when surface ridges are absent using the CFS reanalysis. These may be related to synoptic-scale tropical systems, like EW or TCs. This part addresses hypothesis of section 2.1.

The second part addresses the hypotheses II, III and IV. In this part we study thoroughly the cyclogenesis of hurricane Patricia focusing on the potential impacts of the gap wind using the ERA-5. The study covers the period from the formation of the developing trough up to cyclogenesis. The purpose of studying such period in detail is to fully understand the environment surrounding the developing trough, to identify all the vorticity anomalies contributing to form the hurricane, and to assess their relative importance during the cyclogenesis process, emphasizing the gap outflow vorticity. The impacts of the gap wind on the cyclogenesis of hurricane Patricia include a numerical experiment consisting of suppressing the gap wind to explore a scenario of the evolution of the trough in absence of the gap wind. The outcomes of the numerical experiment are compared with the conceptual model presented in section 2.2.
Figures

Figure 2.1. Low level atmosphere. A TC precursor in the Eastern Pacific is represented by the stream function anomaly ($\psi'$). Northerly flow, vorticity ($\xi$), shear vorticity ($\xi_S$) and a generic anomaly are also shown. See legend for further details.
3. THE INFLUENCE OF SYNOPTIC-SCALE TROPICAL WEATHER SYSTEMS ON THE Isthmus of TEHUANTEPEC GAP WIND IN SUMMER

Mountain gap winds are caused by pressure gradient forces along the mountain gap. In the case of the Isthmus of Tehuantepec, the pressure gradient force underlying the gap wind in winter is mainly due to MSLP anomalies over the GoM leading to gap winds. These anomalies are of midlatitude origin. It is also possible for MSLP anomalies to the south of the Isthmus and upstream circulations like EWs to establish the pressure gradient force leading to a gap wind. However, the potential impacts on the gap wind by MSLP anomalies downstream of the mountain gap in the EPac and by circulations upstream of the mountain gap remain vastly underexplored. In this chapter we will explore these potential impacts.

The most notable impact of tropical systems on the gap wind consists of a change in the gap outflow shape (Bourassa, 1999). This means that the gap outflow which usually has an anticyclonic curvature, curves cyclonically in presence of tropical cyclones to the southeast of the Isthmus (Bourassa, 1999). This curvature change may lead the outflow to wrap around circulations helping to develop them (Farfán and Zehnder, 1997 and Molinari, 1999). This chapter aims to better understand the interactions between the gap wind and tropical systems considering the potential role of the MSLP anomalies associated to these systems and their circulation.

Although the main topic of this chapter is to understand the impacts of tropical systems on the gap wind in summer, the impacts of midlatitude systems on the gap wind will also be revisited. This additional topic will help to contextualize the significance of the impacts of tropical systems on the gap wind next to the midlatitude systems. This chapter comprises the following sections: Section 3.1 presents the data used; section 3.2 presents a climatology of the gap wind; section 3.3 studies the modulation by MSLP anomalies on the gap wind; section 3.4 analyzes the
wind at the time of strong northerlies for different seasons of the year; section 3.5 discusses the influence of diurnal circulation on the gap wind; section 3.6 describes the influence of tropical systems on the gap wind; and section 3.7 presents the summary and conclusions.

3.1 Methodology and Data

The Climate Forecast System (CFS) Reanalysis is used to examine the gap wind and the environment surrounding it. The CFS reanalysis has a global domain and spans in the vertical between 1000 hPa and 30 hPa covering the period from 1st January of 1979 to the present. In this chapter, the period of study ranges from 1979 to 2019. The CFS has a horizontal resolution of 0.5°x0.5° and a vertical resolution of 25 hPa close to the surface and at the top of the atmosphere, but in the middle atmosphere is 50 hPa and has temporal resolution of 6 hours. The spatial and temporal resolution of the CFS reanalysis makes it suitable for analyzing synoptic-scale weather patterns, including EWs, and midlatitude systems.

To highlight the weather patterns, the anomalous wind and MSLP are used instead of total fields. The anomaly for each variable is with respect to the monthly climatology values of the wind and the MSLP. In turn, the monthly climatology value for any time of the year is given by the climatology value within a 30-days time-window centered at any time of the year. This choice allows for continuous climatology variables.

3.1.1 Representation of the Isthmus of Tehuantepec Gap Wind

The standardized anomaly of each wind component averaged within the interval [96.5°W,92.5°W]x[16°N,18°N]x[Surface,800 hPa] encompassing the Isthmus of Tehuantepec diagnoses the gap wind every six hours (Fig. 3.1). The value for the meridional component is referred as $I_T$ and is the foundation to the climatology presented in this chapter to represent the
gap wind. Composites of the wind and MSLP for anomalies different values of $I_T$ representing strong northerlies are analyzed to understand the state of the atmosphere at these times.

### 3.2 Climatology of the Isthmus of Tehuantepec Gap wind

The mean wind at 900 hPa between 1979 and 2019 (Fig. 3.2 (a)), shows the Tehuantepec gap wind as a northerly flow connecting the GoM with the EPac. Its inflow is a southward deviation of the anticyclonic flow over the GoM, and its outflow turns westward and merges with the easterlies over the EPac. The gap wind is associated with a pressure gradient force between the GoM and the EPac. The anticyclonic curvature of the gap outflow results from an adjustment to the easterlies in the EPac.

Figure 3.2 (b) shows a vertical cross section of the mean meridional wind at 17°N between 1979 and 2019. The mean gap wind is the northerly wind constrained by the mountain gap spanning from the surface to about 600 hPa. The wind is greater than 4m/s between the surface and 800 hPa and maximizes at 900 hPa. The 800 hPa level will be considered as the upper limit to represent the gap wind by $I_T$ because it accounts for the strongest wind.

While the northerlies within the mountain gap are strongest at 900 hPa, the outflow over the EPac is strongest at 950 hPa (cross section not shown), indicating that the gap outflow has a downward motion. This is consistent with the studies of Gaberšek and Durran (2004 and 2006) where they found that gap outflows frequently develop downward motions. The downward motion has been associated to the formation of lee-waves similar to hydraulic jumps. The subsequent sections and chapters will consider the 950 hPa pressure level to focus on the outflow to highlights the downstream flow associated with the gap wind.

The next section presents a study of the annual cycle of the gap wind and its fluctuations thorough the year, including its periodicity and amplitude.
3.2.1 Annual cycle of the gap wind

Figure 3.3 shows the monthly climatology values (black line), interquartile range (IQR) and 5th to 95th percentile range of the zonal and meridional wind vertically averaged within the Isthmus of Tehuantepec (red box in Fig. 3.1). The zonal wind is easterly and roughly constant throughout the year (Fig. 3.3 (a)). Its median value is roughly, $u = -0.9 \text{ ms}^{-1}$, and its standard deviation is $\sigma = 1.7 \text{ ms}^{-1}$. For more than 90% of the time, the wind speed is less than 4 ms$^{-1}$ as indicated by the monthly 5th to 95th percentile range. In general, the zonal wind is weak consistent with the mean zonal wind shown in Fig. 3.2 (a).

The meridional wind (Fig. 3.3 (b)) is predominantly northerly (about 75% of the time) and has a pronounced seasonal cycle. The median northerlies are strongest in November, $v = -8.4 \text{ ms}^{-1}$, and are weakest during April and May, $v = -2.2 \text{ ms}^{-1}$. Additionally, a secondary peak occurs in July, $v = -6.9 \text{ ms}^{-1}$. The amplitude of the monthly 5th to 95th percentile range also has an annual cycle. It maximizes during February, March (~3.7 $\sigma$), and April and minimizes during July and August (~1.85 $\sigma$). The extreme northerlies, represented by the 5th percentile, follow a seasonal cycle, are strongest between November and January and weaken between June and August. These variations of the meridional wind through the year suggest that different mechanisms drive the gap wind at different times of the year. These mechanisms could be due to variations in the strength of a unique factor or to several factors underlying the gap wind.

In general terms, the meridional wind is much stronger than the zonal wind. More specifically, when the northerlies are stronger than 4 ms$^{-1}$, which occurs for about 66% of the time, the gap wind is mostly meridional. Thus, hereafter it is only considered the meridional component of the gap wind, represented by $I_T$ since we are interested in the underlying mechanisms behind strong gap winds.
Based on the monthly climatology values of $I_T$, diagnosing the gap wind and the amplitude of its 5<sup>th</sup> to 95<sup>th</sup> percentile range, the evolution thorough the year of $I_T$ is represented by three phases:

- The first phase goes from October to January and is characterized by the strongest mean and extreme northerlies (i.e., when the monthly 5% percentile achieves its minimum), and considerable variability. This phase is referred to as fall-winter.
- The second one goes from February to May and is characterized by the weakest mean northerlies, but they have the largest variability. This season is referred to as spring since it mainly includes the spring months.
- The third phase goes from June to September is characterized by minimum variability and a secondary peak in the mean northerlies; it will be referred to as summer.

An excerpt of the gap wind for 1996 (Fig. 3.3 black line) shows that the gap wind is characterized by fluctuations which amplitudes span across the monthly IQR and even the monthly 5<sup>th</sup> to 95<sup>th</sup> percentile range (Fig. 3.3 red lines) every few days (about 2-6 days). The periodicity of the fluctuations suggests a strong modulation by synoptic weather systems on the gap wind. Note that other year present a similar behavior (not shown).

### 3.2.2 Wavelet Power Spectrum

The monthly mean wavelet power spectrum for $I_T$ (Fig. 3.5 (b)) provides insight into the characteristic timescales associated with its fluctuations at different times of the year. Four features characterize it:

- The annual cycle has a large power, consistent with the seasonal cycle shown in Fig. 3.3.
- About 40% of the variability during Spring and Fall-Winter (Fig. 3.4 (a)) and about 20% during Summer is attributed to the synoptic scales (2-10 days). The contribution of the synoptic
band (2-10 days periods) to the gap wind variability is estimated as the ratio of the accumulated power in this band to all the power.

- The variability attributed to the synoptic band follows the seasonal cycle; it maximizes in February-March and minimizes during July and August, similar to the seasonal cycle of the extreme northerlies.
- The diurnal cycle forms a relative maximum from May through September, suggesting that gap wind has strong diurnal circulations. Then it weakens from October to December.

In general, the synoptic band contributes significantly to the wavelet power spectrum from October to April (fall-winter and spring phases) and minimizes between June and September (summer). In summer, the power is roughly constant for periods greater than three days suggesting that different weather phenomena equally impact the gap wind in a broad period band.

The weak power in the synoptic band during the summer phase leads us to speculate that synoptic systems have a weaker influence on the gap wind fluctuations then. The impacts of synoptic systems at this time of the year will be addressed in section 3.4.

3.2.3 Duration of Strong Northerlies

As mentioned earlier, the fluctuations of the gap wind in general are dominated by the synoptic band in Fall-Winter, but in Summer that is not the case. To know to what extent the synoptic timescales modulated the strong northerlies in Summer we will study the distributions of the durations of strong gap winds for each phase of the year. Northerlies less than -5 ms\(^{-1}\) are considered strong, this value is slightly greater than the mean between 1979 and 2019. The boxplots for the distributions of the duration of strong northerlies (Fig. 3.6) show a median duration for Spring and Summer of 1.75 and 1.5 days, respectively, and larger for Fall-Winter, 2.5 days. In general, the duration of strong northerlies fluctuates between 0.5 days up to 10 days regardless of
the phase of the year. This suggests that strong gap wind northerlies are modulated by synoptic-scale factors regardless of the phase of the year.

In summary, the Isthmus of Tehuantepec gap wind is characterized by fluctuations of the northerlies and the monthly mean northerlies follows a marked annual cycle which can be divided into three phases: fall-winter, spring, and summer. During fall-winter the mean northerlies are strongest, in spring are weakest and in summer experience a secondary enhancement. Regarding the fluctuations of the northerlies, it was found that their amplitude maximizes during Spring and minimizes in Summer. Their periodicity, in general is dominated by the synoptic period (2-10 days) during Fall-Winter and Spring but not in Summer. Nevertheless, the periodicity of fluctuations with strong northerlies (<5m/s) are dominated by the synoptic period regardless of the time of the year.

3.3 Role of Mean Sea Level Pressure Anomalies on the Tehuantepec Gap Wind

Pressure gradient forces produce mountain gap winds (Chelton, 2000; Gaberšek and Durran, 2004; and references within). In the case of the Tehuantepec gap wind, the pressure gradient force is often due to MSLP anomalies in the GoM, such that positive MSLP anomalies related to the cold sector (cold air masses behind cold fronts) of midlatitude systems lead to strong northerlies. This section explores the modulation by MSLP anomalies on the gap wind during Spring and Summer regardless of the source. The fall-winter phase is not discussed because it is similar to spring.

The modulation by MSLP anomalies on the gap wind is analyzed through maps of the zero-day and one-day lag correlation coefficients between the $l_T$ time series and the MSLP time series at every grid point of the CFS reanalysis, (i.e, every $0.5^\circ \times 0.5^\circ$). The following conventions facilitate the interpretation of the correlation maps: Positive values are referred to as positive
correlations and are interpreted as northerlies correlate with negative MSLP anomalies (lows), while negative values are referred to as negative correlations and are interpreted as northerlies correlate with positive MSLP anomalies (highs).

The zero-day lag MSLP correlation map for Spring (Fig. 3.7 (a)) shows strong negative correlations in the GoM; the negative correlations are strongest over the western sector of the GoM (R=-0.8 isopleth). This is consistent with high MSLP in the GoM establishing a pressure gradient between the GOM and EPac leading to gap winds. The one-day lag correlation map (Fig. 3.6 (b)) shows the negative correlation is strongest (-0.7 isopleth) over North America, leeward the Rocky Mountains. This indicates that southward high MSLP anomalies strongly modulate the gap wind northerlies, in agreement with the literature (Shultz et al., 1997 and 1998 and Fig. 1.5).

During Summer, the negative correlation over the GoM weakens (Fig. 3.6 (c,d)), and now there is a positive correlation over the EPac, indicating that the gap wind in Summer is driven by factors different from those in Spring. In terms of the northerlies, the weaker negative correlation in the GoM indicates that positive MSLP anomalies have a diminished impact on them. Meanwhile the positive correlation in the Panama Bight suggests that negative MSLP anomalies impact the northerlies.

The one-day lag correlation map for Summer (Fig. 3.6 (d)) is similar to the zero-day lag map in the GoM, but the positive correlation in the Panama Bight increases. The steady location of the strongest anticorrelation values on the GoM for the zero-day and one-day lag maps indicates that stationary or slowly moving MSLP anomalies modulate the gap wind in Summer, in contrast with the southward positive MSLP anomalies in Spring. Meanwhile, the increased correlation with the MSLP over the Panama Bight, a genesis region for EWs (Whitaker; 2020, and Torres et al;
2021), hints at a potential role for downstream negative MSLP anomalies associated with tropical systems as factors modulating the gap northerlies.

Overall, the correlation maps show that gap winds northerlies correlate with positive MSLP anomalies in the GoM in Spring as expected. In Summer the correlation with MSLP anomalies in the GoM is weaker. Interestingly, in Summer MSLP in the EPac correlates with the Summer, suggesting that low pressure systems in the EPac, particularly those formed over the Panama Bight may have an impact in gap wind northerlies.

3.4 Composite of Atmospheric Patterns During Gap Wind Events

This section analyzes the composites of the wind and MSLP anomalies at the time of gap wind events (defined in next paragraph) to understand the common weather patterns behind these events for the Spring and Summer phases. The MSLP anomalies over the GoM are also analyzed to assess their association with gap wind events in spring an in summer. The composite for the fall-winter phase is very similar to that of Spring and it is not shown.

A gap wind event is defined as a relative minimum of the time series of $I_T$ below the monthly 5th percentile within a one-day window, i.e., extreme northerlies relative to the monthly climatology. This threshold is chosen because extreme northerlies should be due to discernable weather patterns than weak northerlies. The analysis here focuses on revealing patterns associated with tropical factors, yet the discussion for Spring is provide for a comprehensive analysis of the gap wind.

The mean anomalous MSLP value for each event is calculated over the interval $[100^\circ W, 90^\circ W] \times [18^\circ N, 28^\circ N]$; this region represents the part of the GoM where correlation is strong for spring and summer. Finally, the length of the gap wind outflow is given by the length of the shear vorticity isopleth equal to zero s$^{-1}$ at 950 hPa such that the angle with the wind is less
than 45° and the wind speed is above 5 ms$^{-1}$. The criteria to identify the outflow are similar those used in Hewson (1998) and Berry and Thornicroft (2007) to identify the axis of jets-like flows. The distributions of these characteristics for the summer and spring gap events are analyzed to reveal further differences for each phase of the year.

### 3.4.1 Spring Gap Wind Events

The vertical cross section of the composite of the horizontal wind for spring gap wind events is shown in Fig. 3.8 (a). The vertical cross section shows northerly winds spanning thorough all the pressure levels and to the east of the mountain gap. Note that the strongest wind is contained between the surface and 800 hPa (orange shading) similar to the climatological meridional wind (Fig. 3.2 (b)).

Figure 3.9 (a,c) shows the composite of the wind anomalies and temperature anomalies at 700 (panel a) and 950 hPa (panel c) and of the MSLP (panel a) for the gap wind Spring events. The anomalous wind at 700 hPa (Fig. 3.9 (a)) forms an anticyclonic flow to the west of a cyclonic flow over North America north of 30°N characteristic of a midlatitude ridge-trough pattern. The negative temperature anomalies overlap the anticyclonic flow. The MSLP anomalies (Fig. 3.9 (c)) consist of a positive anomaly over the GoM limited to the west by the Sierra Madre and a negative anomaly to the west over the Atlantic. This configuration of the wind, temperature and MSLP clearly show a cold front over the GoM and evidence that high pressure anomalies associated with cold fronts drive the gap wind events. This weather pattern is consistent with the literature (Steenburg, et al, 1998) where gap winds are driven by cold fronts associated with midlatitude systems.

At 950 hPa, the gap wind consists of the equatorward penetration of the northerly flow over the GoM channeled by the Sierra Madre into the EPac through the Isthmus of Tehuantepec.
(Fig. 3.9 (c)). Note that negative temperature anomalies penetrate in the EPac through the mountain gap as well, indicating advection of cold temperature in the EPac by the gap wind. The composite shows that the gap wind northerlies are parallel to the pressure gradient force of the anomalous MSLP along the mountain gap, indicating that they are driven by a pressure gradient.

In summary, midlatitudes systems drive the Spring gap wind events by establishing a meridional pressure gradient between the GoM and the EPac of about 3 hPa per 100 km. Additionally, the Sierra Madre constrains the westward extension of the midlatitude systems, by blocking the high pressures (positive MSLP anomalies) and the flow leading the northerlies to propagate equatorward. Overall, the scenario is consistent with high pressure cold surges impinging the Isthmus of Tehuantepec and penetrating the EPac modulating the gap wind (Steenburg et al., 1998).

3.4.2 Summer Gap Wind Events

Vertical cross section of the composite of the horizontal wind for Summer gap wind events is shown in Fig. 3.8 (b). There, it is shown that the wind is easterly except in the mountain gap itself where it is northerly. The gap wind northerlies are about half of the Spring values, are not as depth as in Spring, and the wind weakens faster from its maximum. This indicates a narrower and weaker gap wind in Summer, suggesting that the Summer gap winds are driven by different mechanisms than those in Spring.

The anomalous wind at 700 hPa (Fig. 3.9 (b)) is dominated by an elongated and weak anticyclonic flow over the western GoM spanning to the EPac. Beneath the anticyclonic flow the MSLP anomalies are positive and weak (Fig. 3.9 (d)). This suggest some influence of midlatitude systems on the gap wind even in summer, albeit there are no significant temperature anomalies at any pressure level when compared with spring.
The composite of the anomalous wind at 950 hPa (Fig. 3.9 (d)) for summer gap wind events shows the gap wind as the southernmost extension of the GoM northerlies. On the EPac, the gap outflow adjusts to a weak easterly flow in the EPac. Over the Isthmus of Tehuantepec, the gap wind is parallel to the pressure gradient, indicating that the gap wind is still driven by the pressure gradient force; however, the gradient in summer is weaker than in spring (1 hPa per 100 km in summer vs 3 hPa per 100 km in spring). Therefore, considering the positive MSLP anomalies over the GoM and the anticyclonic flow at 700 hPa it appears that midlatitude systems still have an impact on the summer gap wind.

Is notable that the MSLP gradient over the mountain gap is due to a positive anomaly over the western GoM and a negative MSLP anomaly over the EPac. The positive MSLP anomaly over the GoM is consistent with the MSLP correlation maps and suggests some influence of midlatitude systems on gap wind events. The negative MSLP anomaly over the EPac shows that negative MSLP anomalies over the EPac contribute to enhance the gap wind, indicating that summer gap winds events may be driven by factors different from those in Spring.

Also, is notable that the wind over the GoM in Summer has an easterly component but in Spring the wind has a westward component. This difference in the sign of the zonal component of the wind lead us to speculate that these northerlies could be part of the cyclonic circulations upstream the mountain gap, however this needs further examined (see section 3.6).

Overall, the state of the atmosphere for summer gap wind events (Fig. 3.9 (b,d)) indicates that midlatitude systems still may trigger gap wind events. However, gap winds events are also influenced by negative MSLP in the EPac suggesting that low-pressure systems in the EPac may also contribute to gap wind events. Furthermore, we speculate that upstream cyclonic circulation, like those associated with EWs or TCs could also have some impact on the gap wind.
3.4.3 MSLP in the GoM and the Outflow Length During Gap Wind Events

The distributions of the gap wind outflow length and of the average MSLP anomalies over the GoM for the Summer and Spring gap wind events are discussed to further evidence that gap wind events in Summer are different from those in Spring (Fig. 3.10).

The distribution of the outflow length (Fig. 3.10 (a)) for Spring events ranges between 400 and 800 km, as indicated by the IQR, and at times exceeding 1000 km. While the average MSLP anomaly value (Fig. 3.9 (b)) associated with gap wind events, ranges between 4 and 8 hPa, as indicated by the IQR but less than 12 hPa for 95% of the events. Unsurprisingly, the positive mean MSLP during gap wind events are consistent with surface cold fronts as the mechanism behind the gap wind.

The corresponding distributions for summer events show shorter outflow lengths (Fig. 3.10 (a)) and weaker MSLP anomalies than in Spring. The outflow lengths range between 125 and 400 km for about 50% of the events, and the MSLP anomalies (Fig. 3.10 (b)) range between 0 and 3 hPa for 50% of the events consistent with the weak MSLP gradient between the EPac and the GoM shown before (Fig. 3.10 (d)). Surprisingly, about 25% of the summer events occur in the presence of negative MSLP anomalies over the GoM versus only 5% of the Spring events. These differences in the outflow length and anomalous MSLP distributions for the summer and spring events further illustrate that summer gap wind events are different from those in spring.

We found that for the case of the Spring gap wind events, their intensity, estimated by $I_T$ was strongly correlated with the outflow length and the MSLP anomalies over the GoM. This means that stronger MSLP anomalies led to stronger gap wind events and longer outflows. However, this is not the case for Summer gap wind events. That is, strong events occur regardless
the sign of the MSLP anomalies and may have short or long outflows. The weak correlation between these variables is interpreted as diverse factors triggering gap wind events in Summer.

3.4.4 Summary

In summary, the composites of the MSLP, and wind anomalies for gap wind events, show that Spring gap wind events, and to a lesser degree, the summer events are both modulated by positive MSLP anomalies over the GoM. However, there are substantial differences between the summer and spring gap wind events, these are: 1) Negative MSLP anomalies over the EPac to summer events indicating a potential impact of downstream EWs on the gap wind; 2) summer gap wind events may occur even in presence of negative MSLP anomalies over the GoM and; 3) summer gap wind events may have short outflows being as short as 100 km, while in Spring, these are long, about 400 km. Therefore, is very likely that summer gap wind events can be triggered by other mechanisms in addition to positive MSLP associated with midlatitude systems. Likely by downstream EWs and probably upstream EWs, these other triggering mechanisms will be explored in section 3.6.

3.5 Influence of the Diurnal Cycle on Summer Gap Wind Events

Before studying mechanisms other than midlatitude systems triggering gap wind events in Summer. We will first explore the diurnal cycle of the gap wind during summer, because as per the wavelet power spectrum, the diurnal cycle has power similar to other timescales (Fig. 3.5 (b)) suggesting a strong contribution to the gap wind by the diurnal cycle.

Evidence of the importance of the diurnal cycle on the gap wind is highlighted by the histograms of the frequencies of the time of the day when gap wind events occur (Fig. 3.10). The histograms show that peak northerlies during gap wind events in spring (Fig. 3.11 (a)) and to a greater degree in Summer (Fig. 3.11 (b)) frequently occur at 0000 UTC, corresponding to 1800
local time. Considering the time of the peak northerlies, composites of the daily wind at 0000 and 1200 UTC will be analyzed for the peak northerlies during gap wind events. The daily wind is represented by the 1-day high pass filtered wind.

It was found that the meridional daily wind over the Isthmus of Tehuantepec is relatively intense (5m/s) at 0000 and 1200 UTC above 900 hPa. Northerly at 0000 and southerly at 1200. At 0600 and 1800 the wind weakens and is mainly zonal (not shown). Thus, the northerly wind above 900 hPa is responsible of enhancing the gap northerlies at 0000.

The meridional flows over the Isthmus are represented by the streamlines of the composited daily meridional and vertical wind for the summer gap wind events at the meridian 94.5°W, primarily because the zonal wind is weaker. These are shown at 0000 and 1200 UTC when the meridional wind is strong (Fig. 3.12) and are representative of the diurnal circulations surrounding the Isthmus of Tehuantepec. These streamlines reveal a meridional circulation over the Isthmus of Tehuantepec between 15°N and 17°N and between 950 and 875 hPa. Also, a meridional flow below 950 hPa converges/diverges in the Isthmus of Tehuantepec.

The streamlines of the daily wind at 0000 UTC (Fig. 3.12 (a)) clearly shows a convective cell over the Isthmus of Tehuantepec (black shading). The convective cell consists of ascending air over land which then flows southward above 875 hPa and sinks over the EPac. Twelve hours later, at 1200 UTC the convective cell has an opposite circulation, the wind above 875 hPa is southerly and the air sinks over land (Fig. 3.12 (b)). This diurnal circulation over the Isthmus of Tehuantepec is consistent with a land-sea breeze caused basically due to the temperature contrast between land and sea. At 0000 UTC, 1740 local time, the land is warmer than the EPac leading the ascending motion over the EPac (Fig 3.13 (a)). And at 1200 UTC, 0540 local time, the sea is warmer than land (Fig 3.13 (a)).
Thus, the meridional wind over the Isthmus of Tehuantepec is mainly caused by a land-sea breeze circulation. Above 875 hPa, the daily wind is mainly meridional, northerly at 0000 UTC and southerly at 1200 UTC. These daily northerlies help to explain why summer gap wind events frequently peak at 0000 UTC.

However, a more detailed analysis of the diurnal cycle of the gap wind is left for future work, including the contributions of surface mixing that are hard to visualize in the CFS due to its coarse resolution (every 0.5°x0.5° and 6 hours). Particularly over land where the temperature changes are greater than over the ocean resulting in a more significant role of diurnal surface mixing (Fig. 3.13).

Finally, it is possible that the interaction between the gap wind and EW-troughs may impacted by the diurnal cycle. This means that the diurnal cycle could be important when forecasting the interactions between EW troughs and the gap wind, when the northerlies are reinforced at 0000 UTC.

### 3.6 Influence of Synoptic-Scale Tropical Systems on Gap Wind Events

So far, convincing evidence of the influence of tropical systems on the gap wind remains elusive. This section attempts to evaluate the influence of tropical systems, specifically of EWs and hurricanes during summer gap wind events. First, two individual events demonstrate different pathways for EWs and hurricanes to trigger gap wind events. Then, the summer gap wind events in the presence of these tropical systems are classified considering their location and the length of outflow to better understand their role in gap wind events.
3.6.1 Examples of Tropical Systems Driving Gap Wind Events

Excerpts for the gap wind index ($I_T$) of September 1988 and July 1996 show two different gap wind events (Fig 3.14). The 1988 event peaked at 1800 14th September and the 1996 event, peaked at 0000 29th July. Each event is described briefly.

The 1988 Event: Example of an Upstream Hurricane Influence

The $I_T$ for the 1988 gap wind event (Fig. 3.14, left panel) shows a steady intensification of the northerlies for six consecutive days starting on September 9th, followed by a sudden weakening of the northerlies and reversal of the wind direction on the 15th. Note the rapid decline and reversal of the meridional wind of about $-24 \text{ ms}^{-1}$ in less than a day after 0000 15th illustrates the abrupt changes the gap wind may experience (Fig. 3.14, (a)).

The atmosphere at the time of the peak of the event is dominated by hurricane Gilbert. A cyclonic flow over the Yucatan peninsula at 700 hPa and 950 hPa, this flow is accompanied by negative MSLP anomaly (Fig. 3.15 (a,c)). It is revealing that this event lacks an MSLP gradient along the mountain gap and the hurricane cyclonic flow extends over the mountain gap. As the hurricane moved westward approaching the Yucatan peninsula the gap wind northerlies slowly ramp up. After, the hurricane traverses the GoM and makes landfall on Mexico on the 16th its southerlies covered much of the GoM causing a southerly gap wind (not shown).

At 950 hPa (Fig. 3.15 (c)), the hurricane circulation is blocked by the topography except over the mountain gap of the Isthmus of Tehuantepec where the wind is enhanced triggering the gap wind event. Downstream, in the EPac, the hurricane circulation intensifies again; with maximum wind speed of about 300 km to the south of the mountain gap. The gap outflow (green line), which also forms part of the hurricane circulation penetrates about 600 km into the EPac. This event shows that gap wind may be enhanced by the circulations of tropical systems without
the need of a significant MSLP gradient along the mountain gap. Furthermore, these events would be impossible to discern from the composites illustrating the potential diversity for mechanisms causing gap wind events in Summer. This mechanism in absence of a pressure gap wind is referred the kinematic mechanism for a gap wind, because in absence of pressure gradients, the strong wind downstream the gap wind is due to vertical momentum fluxes (Gaberšek and Durran, 2004, eq. 10 and Fig. 11 within). This may be caused by subsidence associated with the hurricane circulation over the region.

**The 1996 Event: Example of a gap wind triggered by a downstream EW influence**

For the 1996 event (Fig. 3.1), $I_T$ shows a brief fluctuation of the mountain gap northerlies between the 27th and 30th of July, achieving maximum intensity on 0000 29th of July. For this event, the evolution of $I_T$ is very different from that of the Gilbert event evidencing that summer gap winds events can be triggered by diverse elements.

This event shows a short trough over the EPac at 700 hPa characterized by its cyclonic circulation (Fig. 3.15 (b)). The trough has a negative pressure anomaly to the south of the Isthmus underneath the trough (Fig. 3.15 (d)). The negative MSLP anomaly creates a pressure gradient downstream the gap wind indicating that the gap wind is due to the negative MSLP anomaly associated with the trough.

At 950 hPa (Fig. 3.12 (d)), the wind forms a broad cyclonic flow centered on the minimum MSLP anomaly. The wind over the EPac maximizes along the gap wind outflow (green line), wraps the negative MSLP anomaly, and merges with the trough circulation. Over the U.S., a surface ridge (red shading) is found, but too far from the Isthmus to be considered as the cause of the gap wind. Therefore, the gap wind at 950 hPa is highly likely due to the trough circulation and is enhanced by the pressure gradient over to the south of the Isthmus. As in the case of Gilbert, the
circulation of the trough channels the gap wind outflow suggesting that the gap outflow is related to the EW circulation.

**3.6.2 Composites of Tropical Systems at the Time of Gap Wind Events**

The 1988 event suggests that circulations associated with tropical systems can trigger gap wind events, suggesting a kinematic pathway to gap wind events; additionally, negative MSLP anomalies to the south of the gap wind can further enhance the gap wind, like in the 1996 event. Additionally, both events revealed that the gap outflow is controlled by the circulation of a hurricane or an EW. Also, by considering the complexity of the composite of the summer gap wind events, it is suggested that several different weather patterns may trigger gap wind events. Thus, to understand the impacts of tropical systems on summer gap wind events, these events are grouped considering their outflow length and the location of tropical systems at the time of each event. The outflow length is considered because it results surprising that some gap wind events, which account for the strongest events relative to the seasonal cycle, have develop very short outflows (e.g. Fig. 3.10 (a)). Additionally, the length of the gap outflow and the formation of strong winds downstream the mountain gap are indicative of different mechanism triggering the gap wind (Gaberšek and Durran, 2004).

**Identification of Tropical Systems**

The tropical systems are identified based on Gilbert and the trough structure of MSLP anomalies and positive curvature vorticity at 950 and 700 hPa (red isoliths in Fig. 3.15). We focus on tropical systems satisfying the following criteria that are representative of lower strong tropospheric vortices: 1) locate the MSLP anomaly minima less than -2 hPa and the curvature vorticity maxima at 950 and 700 hPa larger than $2 \times 10^{-5}$ s$^{-1}$. 2) Identify triads consisting of a MSLP minimum, vorticity maxima at 950 hPa and at 700 hPa such that the vorticity maxima are
within 150 km from each other and within 250 km from the MSLP minimum. 3) The MSLP minimum indicates the system location.

The chosen threshold values identify discernable strong vortices in the lower troposphere. This is important because we have observed that in some events there is no apparent weather system at the time of the gap wind event and their interpretation is complicated. Thus, we only consider summer gap wind events where is easy to discern a weather system at the peak of the event.

Once the tropical systems satisfying the criteria at the time of summer gap winds are identified, gap wind events with tropical systems within 1200 km from the Isthmus (circumference in Fig. 3.12 (d)) are clustered according to the location of the systems and the outflow length. About 40% of summer events have tropical systems. The k-means technique is used to identify the centroids of four clusters based on the tropical system’s locations. The resulting clusters spread over the quadrants around the Isthmus of Tehuantepec (Figure 3.13) and are: the upstream, the easternmost EPac (EEPac), the EPac, and the Mexico (Mx) clusters. The outflow length is categorized as short (<250 km) or long (≥250 km). The Mexico cluster is not discussed since it consists of systems unrelated to the gap wind. The composite for each of the 6 clusters is analyzed next.

A) Upstream Cluster

Is surprising that upstream tropical systems are related with the gap wind, since the primary mechanism for the gap wind is a pressure gradient along the mountain gap and the identified tropical systems have low MSLP. The composite for the upstream cluster (northeast quadrant) with long outflows (Fig. 3.17 (a)) is dominated by a broad negative MSLP anomaly and a circulation spanning from the Yucatan peninsula to the EPac. The MSLP anomaly over the EPac
forms a gradient to the south of the Isthmus enhancing the gap wind; the gap outflow is strongest 100 km to the south of the Isthmus. Thus, circulations of tropical systems upstream the gap wind may lead to gap wind events, while the MSLP gradient downstream the gap wind enhances its outflow.

It's worth asking if the MSLP anomaly pattern correspond to different systems, one over the Yucatan peninsula and other over the EPac, or to a broader system all over the domain. It turns out that some individual events (not shown) are due to single and broad trough spanning over much of the domain, while others consist of two different systems. For the events consisting of a broad trough, their circulation at 950 hPa is split by the topography in Central America, forming two circulations, one over the EPac and other over the Yucatan peninsula forming one broad circulation at 700 hPa. In the case of the events with two systems it was found that the system over the Yucatan peninsula is stronger than the one over the EPac and suggests that the gap wind is mostly due to the circulation of the upstream system.

The composite for the upstream cluster with short outflows (Fig. 3.17 (b)) forms similar patterns than the one for long outflow (Fig. 3.17 (a)) but lacks the MSLP gradient over the EPac. This suggests that the circulation of tropical systems upstream the gap wind triggers the gap wind, however if there is no pressure gradient along the mountain gap, then the resulting outflow is short. This is not the case if the circulation of the system is strong enough, as in the Gilbert event; however, these cases are rare.

Both composites show that the circulation of tropical systems upstream the Isthmus of Tehuantepec can lead to gap winds; and a long outflow is formed if a pressure gradient along the mountain gap is present.
B) Easternmost Eastern Pacific (EEPac) Cluster

The composite for the EEPac cluster, southwestern quadrant, with long outflows (Fig. 3.18 (a)) shows a trough over the EPac and a northerly flow over the GoM. The anomalous MSLP associated with the trough forms a gradient along the mountain gap leading to a gap wind and its outflow over the EPac. However, there is no positive MSLP anomaly over the GoM signaling surface ridges. Regarding the outflow, it forms the strongest winds over the EPac 100 km offshore, and merges with the trough circulation. Overall, the composite shows that negative MSLP anomalies associated with troughs can enhance northerly flows and the outflow is channeled by the trough circulation, indicative of the kinematic effect on the gap outflow by the trough.

The composite for events with short outflows (Fig. 3.18 (b)) shows a short trough over the EPac and a weaker MSLP gradient over the Isthmus. The northerlies over the GoM does not extend poleward up to the U.S. indicating even weaker surface ridges. A westerly flow to the west wraps around the negative MSLP anomaly. The westerlies suggest enhanced convection over the EPac, e.g., the phase 2 of the MJO. This suggests that troughs formed when convection over the EPac is widespread help to enhance the gap wind despite. The main difference between the long outflow and short outflow composite is the steeper MSLP gradient found in the long outflow composites suggesting that the short troughs develop are weaker. However, both composites reveal that EPac troughs may enhance the gap wind. This strengthens the idea that MSLP gradient due to negative MSLP anomalies downstream the gap wind may as well trigger gap wind events, and steeper MSLP gradients may lead to longer outflows.

C) EPac Cluster

The composite for the EPac cluster, tropical systems in the southwest quadrant, with long outflows (Fig. 3.19 (a)) shows a MSLP gradient caused by a positive MSLP anomaly in the GoM
and a negative MSLP anomaly in the EPac. The gradient maximizes along the Isthmus of Tehuantepec leading to a gap wind. The gap outflow in the EPac merges with the trough’s circulation, however its MSLP anomaly is weaker than in the EEPac composite. As in the previous composites the outflow deflects toward the trough and merges with its circulation. Thus, the negative MSLP anomalies of EPac troughs contribute to establishing a MSLP gradient along the mountain gap and the trough circulation channels the outflow.

The composite for events with short outflows (Fig. 3.19 (b)) show weak northerlies limited to the western GoM, but do not have a MSLP gradient along the mountain gap. Despite the minimum MSLP anomaly is stronger than in the long outflow composite (Fig. 3.19 (a)), the MSLP gradient along the mountain gap is weaker. Overall, the cases conforming this cluster are the weakest and have the shortest outflow. The gap wind enhancement in these cases results mainly from the topography constraining the upstream flow, like in a typical Venturi tube. In the Venturi tube case, the flow is only enhanced in the constriction itself and has the same speed at the entrance and exit of the constriction (this occurs here as well).

In general, the EPac cluster shows that EPac troughs to the southwest of the Isthmus of Tehuantepec help to trigger gap wind events. When the troughs MSLP anomaly occur simultaneously with a surface ridge a long gap outflow form and the outflow is channeled by the trough circulation. On the contrary in absence of a surface ridge over the GoM the gap outflow is very short.

D) Summary of K-Means analysis

Two gap wind events exemplified the influence of tropical systems on gap wind events. The examples show that tropical systems can establish the MSLP gradient along the mountain gap leading to a gap wind event; or their circulation channeled through the Isthmus may as well lead
to a gap wind event (the kinematic mechanism). Composites of Summer gap wind events based on the location of tropical systems and the outflow length revealed a diversity of configurations favorable to gap wind events. In general terms, the composites revealed that tropical systems over the EPac can trigger a gap wind by establishing a MSLP gradient along the mountain gap or by having their circulation flowing through the mountain gap (the kinematic mechanism).

Additionally, it was observed that the circulation of downstream EWs channels the gap outflow. The length of the outflow is related with the steepness of the MSLP gradient. If the MSLP gradient is steep the gap outflow is long, if not, is short.

In conclusion, the summer gap wind events have a large variability of causes. This means that the gap wind is equally impacted by different locations and MSLP anomalies associated with tropical systems.

3.7 Summary and Concluding Remarks

Fluctuations of the meridional wind characterize the Isthmus of Tehuantepec gap wind and are diagnosed by the time series of the Tehuantepec gap wind index ($I_T$). $I_T$ follows a distinctive seasonal cycle such that the mean northerlies peak between November and January and are weakest in May. A secondary maximum occurs in August. The fluctuations of $I_T$ last a few days and their amplitude is maximum in winter and spring but is minimum in summer. This means that the fluctuations in summer are also weaker than in the rest of the year. These two differences of the summer fluctuations with respect those in the rest of the year indicate that the gap wind is modulated by other factors in summer.

The gap wind events in spring and winter are primarily modulated by positive MSLP anomalies associated with midlatitude systems. However, in summer their influence is diminished
because they retreat poleward yet are still important. During Summer it was found that other factors also modulate the gap wind, these are:

- **Upstream tropical systems, e.g., hurricanes or EWs over the Yucatan peninsula.** The circulation of these systems when channeled by the mountain gap results enhanced, leading to a kinematic mechanism for gap wind events. Notably these systems do not necessarily establish a meridional MSLP gradient across the mountain gap.

- **Downstream tropical systems in the EPac.** The negative MSLP anomalies associated with troughs or hurricanes in the EPac may establish a meridional MSLP gradient leading to gap wind events. The MSLP gradient can be further enhanced by positive MSLP anomalies over the GoM typically caused by midlatitude systems.

- **The diurnal cycle.** Most of the gap wind events occur at 1800 local time (0600 UTC), meaning that the gap wind has a strong response to the diurnal cycle due to the coastal breeze in the Isthmus. However, we recognize the potential role of diurnal mixing to contribute to the meridional wind, but the CFS is not adequate to study this mixing and is left for future studies.

  Additionally, it was found that the length of the gap outflow during summer gap wind events has broad variability ranging from less than 100 km to 400 km. In general, the outflow length in summer is substantially shorter less than in spring events. In spring, the gap outflow typically has outflows longer than 400 km. This is a surprising result because a gap wind event is given by the relatively strongest northerlies, i.e., minima below the monthly 5th percentile. In the case of the summer gap wind events, the length of the outflow is related to the minimum MSLP of the tropical systems (troughs or hurricanes) in the vicinity of the Isthmus of Tehuanantepec. If the MSLP of these systems is such that establishes a steep gradient along the mountain gap, then the outflow is long, on the contrary if the gradient is weak, the outflow is short. Notably the gap
outflow merges with the circulation of the tropical systems. This leads us to speculate that longer outflows may enhance the circulation of the triggering tropical systems, resulting in a positive feedback between them. First, the gap wind is triggered by the system’s negative MSLP anomaly, then the gap outflow may enhance its circulation favoring the intensification of its MSLP anomaly further. In turn the MSLP anomaly intensifies further enhancing the gap wind.

In conclusion, the Isthmus of Tehuantepec gap wind is primarily modulated by surface ridges (positive MSLP anomalies) thorough the year. However, in summer, tropical systems like EWs troughs or hurricanes, may also trigger gap winds events. If these tropical systems are located downstream, they may establish a MSLP gradient favorable to a gap wind; if located upstream, the gap wind may be enhanced by their circulation. Additionally, the gap outflow is channeled by the circulation of the triggering tropical systems and merging with the circulation. Therefore, tropical systems may provide different pathways leading to gap wind events in summer.

**Figures**

Figure 3.1. A) Topography (shading), region where the wind is averaged (red box), and parallel 17°N. B) Vertical cross section of the topography at 17°N (black) and region where the wind is averaged (red box).
Figure 3.2. Monthly climatology value of the wind components in black line. Interquartile range (IQR) in dark gray shading and monthly 5th-95th percentile range in light shading of A) the zonal wind (upper panel) and of B) the meridional wind (lower panel) in terms of m/s (left axis) and standardized anomalies (right axis, red ticks). Horizontal lines in B) are at ±4 m/s. Labels in the horizontal axis are located at the first day of the month.

Figure 3.3. Mean wind at 900 hPa at hPa (A) and vertical cross section of the meridional wind at 17°N (isopleths every m/s). Horizontal line at 17°N in panel A indicates the latitude of the vertical cross section of B. Black shading in B indicates the orography.
Figure 3.4. As figure 3.2 for the meridional wind (left axis) and $I_T$ (right axis) with an excerpt of the 1996 year. The triangles mark gap wind events. Individual fluctuations are highlighted in red and are discussed in the text.
Figure 3.5. A) Percentage of the power of the synoptic band (2-10 days). B) Monthly climatology of the wavelet power spectrum of the gap wind index. Horizontal lines marked at selected periods.

Figure 3.6. Boxplot of the number of consecutive days with meridional winds below -5 m s\(^{-1}\).
Figure 3.7. Zero-day lag correlation coefficients between the MSLP and the gap wind index (left column) and of the one-day lagged correlation coefficient (right column) for Spring (A,B) and Summer (C,D). Labels every ±0.2 units and isopleths every ±0.1 units.

Figure 3.8. Vertical Cross sections of the horizontal wind (barbs) and its intensity (shading) at the parallel 17°N.
Figure 3.9. Composite of the wind anomaly (barbs), and temperature anomaly (every $\pm 1$K regions in Spring and $\pm 0.5$K regions in Summer; blue patches for negative, and red for positive, in K) at 700 (A and B) and 950 hPa (C and D) for Spring (A and C). Also is shown the corresponding composite of the MSLP anomalies (every $\pm 1$ hPa (green contours)) in panels C and D.
Figure 3.10. Boxplots of the Jet Length (A), and mean MSLP anomalies over the GoM (B) distributions for Spring and Summer gap wind events.

Figure 3.11. Histograms of absolute frequencies for the time of the day when Spring (A) and Summer (B) gap winds events occur.
Figure 3.1. Excerpts of the gap wind index for September 1988 and July 1996. Gray envelopes indicate the 5th to 95th monthly percentile range and the monthly interquartile range. The heavy solid line is the monthly median. Red crosses indicate the time of gap wind events.

Figure 3.12. Vertical cross sections at the meridian 94.5°W of the composite of the streamlines of the daily meridional and vertical wind for Summer gap wind events. Panel A at 0000 UTC and B at 1200 UTC. Horizontal axis in degrees north. Gray shading indicates the topography and heavy solid line indicates land.

Figure 3.13. Composite of the temperature at 1000 hPa for the Summer Gap Wind events occurring at 0000 (A) and 1200 UTC (B).

Figure 3.14. Excerpts of the gap wind index for September 1988 and July 1996. Gray envelopes indicate the 5th to 95th monthly percentile range and the monthly interquartile range. The heavy solid line is the monthly median. Red crosses indicate the time of gap wind events.
Figure 3.15. 950 (A,B) and 700 hPa (C,D) anomaly maps for the gap wind events of 1800 1988-09-14 (A,C) and 0000 1996-07-29 (B,D). Anomalous wind (barbs) and positive curvature vorticity \((10^{-5} \text{s}^{-1})\), dotted red isopleths), mslp (in hPa dashed blue isopleths for negative anomalies and red isopleths for positive anomalies in c, d), and gap outflow (green line in c, d). Isopleths are every \pm 4 units. Color shading corresponds to \pm 4 units isopleths, blue for negative and red for positive anomalies. Also are indicated MSLP relative minima by blue circles (\leq -4 hPa), curvature vorticity relative maxima (\geq 4 \times 10^{-5} \text{s}^{-1}) at 950 hPa by crosses and at 700 hPa by plus signs, concentric circumferences every 100 km from the Isthmus of Tehuantepec and at 1250 km are also shown.
Figure 3.16. Locations of MSLP minima corresponding to tropical systems within 1200 km from the Isthmus of Tehuantepec at the time of gap wind events. Colors indicate their respective cluster.

Figure 3.17. 950 hPa anomaly composites for the upstream cluster with short (A) and long (B) gap wind outflows length. Anomalous wind (barbs), MSLP (in hPa, every ±1 hPa, dashed and solid isopleths for negative and positive anomalies, respectively), and gap outflow (green line in A). Color shading corresponds to −2 hPa isopleths.
Figure 3.18. 950 hPa anomaly composites for the EEPac cluster with short (A) and long (B) gap wind outflows. Anomalous wind (barbs), MSLP (in hPa, every ±1 hPa, dashed and solid isopleths for negative and positive anomalies, respectively), and gap outflow (green line in A). Color shading corresponds to −2 hPa isopleths.

Figure 3.19. 950 hPa anomaly composites for the EPac cluster with short (A) and long (B) gap wind outflows. Anomalous wind (barbs), mslp (in hPa, every ±1 hPa, dashed and solid isopleths for negative and positive anomalies, respectively), and gap outflow (green line in A). Color shading corresponds to −2 hPa isopleths.
4. **SYNOPTIC-SCALE ASPECTS OF PATRICIA CYCLOGENESIS**

Hurricane Patricia was a short-lived TC (20-24 October, 2015) category 5 hurricane that formed over the EPac to the south of the Isthmus of Tehuantepec, moved parallel to the coast and then recurved to make landfall on 0600 23rd (Fig. 4.1). Patricia is the most intense TC ever recorded in the western hemisphere since records began in 1979 (Rogers et al., 2017). Patricia has been the subject of numerous studies (see Bosart et al., 2016; Kimberlain, 2016; Rogers et al., 2017; Qin and Zhang, 2018; Fox and Judt, 2018; Martinez, 2019). The vast majority of them focus on its unprecedented rapid intensification ending in its record-breaking intensity that lasted two days beginning on the 21st of October short after reaching tropical storm category. Notably, except for Bosart et al. (2016) and the NHC report (Kimberlain, 2016), none of these studies focused on the complex processes leading up to cyclogenesis. This process involved the interaction between several weather systems (see Fig. 1.15). This complexity was further manifested in poor operational forecasts that formed the TC two days earlier than the best track record e.g., GFS, and is discussed in chapter 7. Motivated by this, this chapter tries to shed light on the synoptic evolution leading up to cyclogenesis using the state-of-the-art ERA-5 reanalysis.

The objective of the present chapter is to investigate and document the series of events in the synoptic scale between the 8th and the 20th of October of 2015 that culminate in the formation of Patricia on 0600 20th October. Focus is given to the structures and spatial patterns of the wind and its vorticity at 950 hPa and 700 hPa, and to a lesser extent at 500 hPa. The 700 hPa level has been used in prior studies to highlight easterly wave (EW) circulations (Serra et al. 2008, 2010 and Rydbeck and Maloney, 2014) and the 950 hPa level captures the interactions between the developing trough with the low-level mountain gap flows caused by the Papagayo and the
Tehuantepec mountain gaps. The 500 hPa is used to highlight the formation of EWs over the Panama Bight.

4.1 Background State of the Atmosphere between 8th and 20th October.

Before discussing the synoptic-scale evolution, the background state of the atmosphere for the period between the 8th and the 20th of October is analyzed to understand what conditions may be present in terms of the vorticity fields at 950 and 700 hPa levels. Additionally, the following parameters are also considered after Gray (1968): sea surface temperatures (SSTs), environmental vertical wind shear between 800 and 200 hPa, the total precipitable water to evaluate if the EPac in that period is favorable to TC genesis.

The background state winds at 950hPa (see Fig. 4.2(a)) are dominated by an easterly flow over the Caribbean and the GoM, southwesterlies in the southern boundary of the domain (highlighted by the green line) and the mountain gap wind of the Isthmus of Tehuantepec. The easterlies in the Caribbean form the CLLJ, flanked to the north and to the south by negative and positive vorticity patches, respectively. The extension of the CLLJ towards Central America is split by the orography in two branches. The northern branch flows into the Yucatan peninsula (heavy red line) and the southern branch flows through the Papagayo mountain gap (black line on top of the red line). Over the GoM, the easterlies form part of an anticyclonic circulation centered over North America that impinges the Sierra and is channeled to the Isthmus of Tehuantepec forming a gap wind. This anticyclonic flow is indicative of the presence of surface ridges over North America in this period. The southwesterlies in the southern East Pacific are associated with a cross-equatorial flow deflected by the Coriolis force that advects moisture into the ITCZ region (Gallego, et al, 2019) at ~10°N and frequently into South America (Poveda and Mesa, 1999). The southwesterlies may combine with the Papagayo mountain gap easterlies to their north forming a
monsoon trough with high vorticity values favorable for the development of disturbances (Holbach and Bourassa, 2014).

At 700 hPa (Fig. 4.2(b)), the background mean wind is characterized again by an easterly flow from the Caribbean into the EPac. The mean wind in the Caribbean (red streamlines) has a southerly component that may steer poleward any incoming EWs from the east. Meanwhile in the EPac, the mean wind is roughly aligned with the anticyclonic flow over North America. In turn, the anticyclonic flow (blue streamline), west of a trough and leeward of the Rocky Mountains indicates the presence of midlatitude systems on this two-week period.

The background vorticity is large ($\geq 1 \times 10^{-5} \text{ s}^{-1}$) to the south of the Isthmus of Tehuantepec and along a strip of vorticity at 9°N at 950 and at 700 hPa (Fig. 4.2, dotted contours). The vorticity strip is associated with the ITCZ and is due to the release of latent heat by the convection causing much of the vorticity. It extends eastward up to 90°W, between the CLLJ southern branch and the southwesterlies coinciding with a region favorable for the development of synoptic disturbances due to a PV gradient sign reversal (Ferreria and Schubert, 1997; Holbach and Bourassa, 2014 and Molinari, 1997).

Figure 4.3 shows the background SSTs, environmental vertical shear between 200 and 800 hPa (averaged within 800 km from each grid point) and precipitable water (pw) of the total column (pw). The figure shows that for this period the EPac had mean SSTs above 29°C over a broad region and weak vertical shear of less than 5ms$^{-1}$ between 5 and 15°N over the EPac. The background pw over the ITCZ is indicative of convection (Hagos et al, 2021), where the pw is larger than 55 mm m$^{-2}$. Overall, the background mean state over the EPac is favorable to develop weather disturbances, like EWs or tropical cyclones, particularly close to Central America. Finally, it is worth noting the region where the vorticity maximizes over the ITCZ, the wind shear is low,
the pw is greater than 55 mm m$^{-2}$, and the SST is about 27°C. This region is favorable to convection and weather disturbances.

Therefore, the EPac during the period of study forms a region favorable to tropical cyclogenesis during this period, with high precipitable water, high SSTs, and low wind shear in agreement with the observations of Bosart et al (2016) and Kimberlain (2016).

4.2 Evolution of the Synoptic-scale elements.

The works of Bosart et al. (2016), summarized in Fig. 1.15, and of Kimberlain (2016) already provide some background about the key synoptic features that contributed to the cyclogenesis of Patricia. These include: 1) a southwesterly flow in the EPac from the ITCZ region, 2) an incoming EW from the Caribbean, and 3) an Isthmus of Tehuantepec gap wind event. The analysis presented here will consider these elements to provide an interpretation of the events leading up to the cyclogenesis of Patricia in terms of the anomalous wind and its vorticity. The wind anomalies are considered with respect to the background state.

In addition to analyzing the developing EWs trough for Patricia, the corresponding one for Olaf that happened a few days earlier is also discussed. This is because the developing EW troughs for both TCs share a common origin that can be traced back to weather disturbances formed by the Panama Bight that experience some development when they move over the region of high vorticity in the ITCZ (dotted region over the EPac in Fig. 4.2) yet they have different fates. The disturbance that yielded Olaf moved westward while the one for Patricia moved northwestward and led to a different cyclogenesis pathway.

4.2.1 Formation of the Developing EW Troughs for Olaf and Patricia

Figures 4.4, 4.5 and 4.6 show the evolution of the anomalous winds and associated vorticity every two days between the 8th and 20th of October at 950, 700 and 500 hPa. The
discussion is initially dedicated to the formation of two weather disturbances formed close to the region of the ITCZ where the vorticity is large and occurred in presence of a Papagayo wind that led to Olaf and Patricia. These are called the developing troughs or simply the troughs. Then, the discussion shifts to the evolution of the trough for Patricia from 14th to 20th of October until cyclogenesis.

On the 8th October, the 700hPa and 950hPa pressure levels are both characterized by a NE-SW strip of positive vorticity between the North Atlantic and the GoM (Fig. 4.4 (a, b)), consistent with the passage of a midlatitude system which also triggered the Tehuantepec gap wind event. The westerlies over the GoM at 950 hPa are due to a midlatitude system, and curve anticyclonically over Central America and enhance the Papagayo gap wind. Both gap winds are associated with positive and negative vorticity anomalies either side of the peak winds (Fig. 4.4 (a)). In the tropics, at 700 hPa (Fig. 4.4 (b)) the wind consists of a northeasterly flow resembling the leading edge of an EW that spans from the Papagayo gap to the EPac which may also have contributed to enhance the Papagayo at low levels.

Also, at this time at 500 hPa (Fig. 4.6 (a)) the wind over Central America is also northeasterly, similar to that at 700 hPa. To the south of this northeasterly flow is found a vorticity maximum. It appears that this vorticity was formed there since it cannot be traced back. Thus, the northeasterly anomalous flow west of Central America spanning from the 950 to 500 hPa suggest the formation of a trough that formed in the Panama Bight. The formation of vorticity anomalies at these pressure levels has been associated to the genesis of EWs (Whitaker, 2020 and Torres, 2021).

By the 10th October, some small-scale vorticity anomalies can be seen at around 5°N at 950 hPa and 700 hPa (Fig. 4.4 (c,d)) consistent with convection in the ITCZ region. These vorticity
anomalies are underneath the 500 hPa vorticity patch (Fig. 4.6 (b) dashed line), and together form the trough that was associated with the development of Olaf and can be tracked for several days after (diagonal lines connecting the panels in Fig. 4.4 and dashed lines in Fig. 4.6).

Additionally, on the 10th of October a new weak vorticity patch is seen close to the Panama Bight at 500 hPa (Fig. 4.6 heavy line). This new vorticity patch indicates the genesis of the trough that was associated with the genesis of Patricia and will be discussed in section 4.2.2.

By the 12th of October (Fig. 4.4 (e, f)), the developing trough for Olaf has moved westward and intensified at 950 and 700 hPa levels with its vorticity maximum at 700 hPa (Fig. 4.4 (f)). From this time onward, the developing trough continues to intensify its circulation and eventually forms Olaf on the 18th.

On the 12th, the trough that formed two days (10th) earlier close to the Panama Bight at 500 hPa, has moved westward (Fig. 4.6, heavy solid line). Underneath the 500 hPa EW trough some positive vorticity structure both at 700 hPa (Fig. 4.4 (f), heavy solid line) and 950 hPa is seen. At 950 hPa the trough vorticity overlaps with the positive vorticity patch of the Papagayo gap wind (Fig. 4.4 (e)). Is interesting that the trough vorticity at 950 hPa coincides with a Papagayo gap wind suggesting that the Papagayo gap wind influences positively on the development of the EW trough. This trough is the developing trough for Patricia.

Interestingly, it seems that both developing troughs became evident at 700 hPa after a Papagayo wind event, suggesting that this wind is beneficial to intensify EW troughs. The literature indicates that the Papagayo wind enhances the formation and development of disturbances into EWs (Holbach and Bourassa, 2014). This mechanism consists of energy conversions from the mean kinetic energy associated with the meridional shear of the zonal wind to eddy kinetic energy that occur when the Papagayo wind becomes unstable and intensifies the disturbances (Serra,
However, the role of the Papagayo wind to the development of these two troughs is out of the scope of this research and is left as future work.

### 4.2.2 Evolution of the developing trough for Patricia

The next days (Figs. 4.4 (g, h) and Fig. 4.5) show the northwestward motion of the developing trough for Patricia parallel to Central America contrasting with the westward motion of the developing trough for Olaf.

By the 14th of October (Fig. 4.4 (g, h)), the wind and its vorticity form complex patterns in the easternmost EPac. The trough vorticity has weakened by this day. At 950 hPa only a small patch of vorticity at the southern end of a northerly flow offshore Central America remains (Fig. 4.4 (g)). At 700 hPa, the continuous vorticity band has yielded a number of weak, positive, and small-scale vorticity anomalies dispersed along a SW-NE oriented axis. The axis straddles the continent (heavy line in Fig 4.4(h)) and spans from the EPac to the Caribbean (Fig. 4.4 (h)). The extension of the trough in the Caribbean, coincides with the passage of an incoming EW and eventually both systems merge. This merging is more evident at 500 hPa (Fig 4.6 (c)). The system resulting from the merging will be referred as the developing trough. It is worth noting that the upstream EW can only be traced back at 500 hPa to the 12th (Fig. 4.6, dashed line) suggesting that it formed in the Caribbean. This is the EW referenced by Bosart et al (2016); as both systems merge, the EPac trough drifts northwestward in contrast to the westward motion of the trough that led to Olaf.

In simpler terms, the developing trough in the EPac observed on the 12th weakens and is approached by an incoming Caribbean EW. This results in a larger SW-NE oriented trough that spans from the EPac to the Caribbean and moves northwestward.
By the 16\textsuperscript{th} of October (Fig. 4.5 (a, b)), the developing trough in the easternmost EPac strengthens and forms again a continuous vorticity patch. At 950 hPa (Fig. 4.5 (a)), the positive vorticity patch in the EPac is limited to the north by the mountains and is accompanied by a southerly flow. At 700 hPa (Fig. 4.5 (b)), the trough (indicated by a schematic trough black line) retains its SW-NE orientation, similar to an EPac EW with maximum vorticity over Central America extends over the EPac. This time coincides with the onset of the Central America Gyre (CAG) discussed by Bosart et al (2016) that helped to develop a circulation over the EPac. Also, the Tehuantepec gap wind and its associated vorticity dipole is clearly seen at this time. This gap wind is triggered by a midlatitude system.

By the 18\textsuperscript{th} of October (Fig. 4.5 (c,d)) the trough occupies much of the Yucatan peninsula stalling for the next two days. Its southerlies at 950 and 700 hPa experience an abrupt enhancement and advect vorticity from the ITCZ poleward, also the the vorticity over the EPac along the southerlies increases. At 950 hPa (Fig. 4.5 (c)) the trough vorticity is close to the positive vorticity patch associated with the Tehuantepec gap wind and start to merge. The merging of the gap wind vorticity with that of the trough forms a broad vorticity patch, this patch leads to cyclogenesis.

By 19\textsuperscript{th} of October (Fig. 4.5 (e, f)), the day before cyclogenesis, the positive vorticity patch associated with the Tehuantepec gap wind has fully merged with the vorticity of the trough at 950 hPa (Fig. 4.5 (e)); i.e., they become undistinguishable one from each other. At 700 hPa, the trough vorticity connects with the midlatitude system vorticity (Fig. 4.5 (f)); note the filament of vorticity going from the Yucatan peninsula to the Atlantic going through the GoM. This leads to a weakening of the trough over the Yucatan peninsula, however the circulation intensifies over the EPac.
On the 19\textsuperscript{th}, infrared imagery reveals a curved band of clouds originating in the ITCZ that surrounds the maximum of vorticity at 950 hPa (Fig. 4.7 (a), cyan cross). This curved band is indicative of the formation of a TC (Dvorak, 1984 and Velden 2006). Over this curved band, deep convection develops as indicated by the location of individual lightning strikes. Note that deep convection occurring close to the vorticity maximum has been associated with strengthening vortices (Wang, 2018) further indicating the imminent cyclogenesis.

Finally, by 20\textsuperscript{th} of October (Fig. 4.5 (g,h)) Patricia has formed, as is indicated by the stacked vorticity maxima in the EPac at 950 and 700 hPa. By this time, the Tehuantepec wind achieves its maximum intensity and wraps around the southern sector of the nascent TC. Strongly suggesting an interaction between the nascent TC and the gap wind. The interaction between the gap wind and the TC is further explored in chapter 6. Additionally, the curved band of clouds remains convectively active and is better defined when compared with the day earlier consistent with the overall increase of vorticity (Fig. 4.7 (b)).

4.3 Summary and Concluding Remarks

The evolution of the developing trough for Patricia involved complex processes where several systems impacted its evolution. These can be summarized in terms of four key stages (Fig. 4.8). These are: I) Formation of the developing EW trough in the EPac, II) northwestward motion and weakening of the trough, III) merging with a Caribbean EW and IV) interaction between the developing trough and the gap wind leading to cyclogenesis.

\textbf{Stage I:} (Fig. 4.8 (a)) The developing trough is first detected as a vorticity anomaly on the 10\textsuperscript{th} of October over the Panama Bight at 500 hPa where heating there may trigger disturbances (Torres et al, 2021). Then it moved westward, in a fashion similar to an EPac EW described in Whitaker (2018). By the 12\textsuperscript{th}, its signature was evident at 950 hPa and coincided with a Papagayo
gap wind. It has been observed that instabilities of the vorticity strip associated with the Papagayo wind contribute to develop disturbances (Serra, 2010 and Holbach and Bourassa, 2014).

**Stage II:** (Fig. 4.8 (b)) On the 14\textsuperscript{th}, the trough is approached by an incoming Caribbean EW that was propagating along the CLLJ (Fig. 4.1 (b)). Both the trough and the Caribbean EW move northwestward, and its vorticity weakens. At 700 hPa the of both systems vorticity forms an SW-NE axis made of small patches (red shading).

**Stage III:** (Fig. 4.8 (c)) Two days later, on the 16\textsuperscript{th}, both systems, the incoming Caribbean EW and the EPac EW merge resulting in a broader trough and forming a circulation at 700 hPa over Central America. When the trough resulting from the merging reaches the Yucatan peninsula on the 18\textsuperscript{th} it stalls, and its vorticity and southerlies strengthen over the EPac (red shading and green wind barbs). While stalled, the trough vorticity increases at 950 hPa, in part aided by merging with the positive vorticity patch associated with the Tehuantepec gap wind. At this point the vorticity within the trough maximizes over the Yucatan Peninsula at 700 hPa.

**Stage IV:** (Fig. 4.8 (a)) On the 19\textsuperscript{th}, most of the changes within the trough occur at 950 hPa; the circulation intensifies, and the vorticity builds up in the southwestern part of the trough over the EPac. The increase in vorticity is mainly a result of the merging of the trough vorticity with the gap outflow vorticity trough. The merging was accompanied by the deep convection occurring close to the vorticity maximum at 950 hPa further increasing the vorticity there (Fig. 4.8 (c) black dots). As the circulation develops in the EPac, the cyclonic flow over the Yucatan peninsula weakens. This leads to an apparent split of the trough by forming two circulations, one leading to Patricia on 0600 UTC 20\textsuperscript{th} October (Fig. 4.8 (d) crosshairs) over the EPac and other over the Yucatan peninsula (barbs in Fig. 4.8 (d)) that dissipated as the midlatitude system passes. The merging of the gap wind vorticity with the vorticity of the developing trough and subsequent
increase of vorticity at low levels strongly suggests a positive net effect of the gap wind to
cyclogenesis. This is studied further in chapter 6.

In general, Stage IV depicts a complex scenario for cyclogenesis when the developing
trough stalled for two days and was characterized by the intensification of the circulation and
vorticity increase in the EPac at 950 hPa. The vorticity increased mainly by the merging of the
trough vorticity with that of the gap wind between the 18th and the 20th.

The next chapter studies the inner structure of the trough and the mesoscale-scale processes occurring within the trough over the EPac leading cyclogenesis during the two days before cyclogenesis, 18th to 20th.

Figures

![Figure 4.1. Best track of hurricane Patricia and storm intensity starting south of the Isthmus in the EPac. Horizontal axis in right panel are in days of October. Heavy dots placed every 24 hours from cyclogenesis.](image)
Figure 4.2. Mean wind field and mean vorticity in color shading from the 8th to 20th of October 2015 at 700 hPa and 950 hPa, red for positive and blue for negative vorticity values. Wind (barbs, color coded according to their intensity as shown below panel A). A full pennant is placed every 10 knots and a half pennant every 5 knots. Heavy lines are selected streamlines to represent the anticyclone and the trough of the midlatitude system in blue. The red line is selected streamline to illustrate the CLLJ. The thin black line is a selected streamline to illustrate the southern branch of the CLLJ. The green line is a selected streamline to illustrate the southerlies. Thin black contour indicates a region in the EPac with vorticity greater than $10^{-5}$ s$^{-1}$ at 950 and at 700 hPa simultaneously.

Figure 4.3. Mean SST in color shading; black isopleth for SST at 29°C. Mean precipitable water isopleths at 55 and 60 mm m$^{-2}$ and mean environmental wind shear are also shown. The sources of vorticity in the EPac are also shown.
Figure 4.4. Anomalous wind and its vorticity in color shading at 950 hPa (A, C, E and G) and 700 hPa (B, D, F and H) for the 8th (A, B), 10th (C, D), 12th (E, F) and 14th (G, H) at 0600 UTC of October. The tilted lines is the manually analyzed location of the precursor for Olaf and for Patricia in F and H. Bold wind barbs (D) indicate a weak northerly flow.
Figure 4.5. Greater than 2 days filtered in time wind and its vorticity in color shading at 950 (A, C, E and G) and 700 hPa (B, D, F and H) for the 16th (A, B), 18th (C, D), 19th (E, F) and 20th (G, H) at 0600 UTC of October. The tiled black line in panels B, D, F and I represents the manually analyzed axis of the precursor of Patricia.
Figure 4.6. Anomalous wind and its vorticity in color shading at 500 hPa at the 0600 UTC of the 8\textsuperscript{th} (A), 10\textsuperscript{th} (B) and 12\textsuperscript{th} (C) and 14\textsuperscript{th} (D) of October. The tilted black lines represent the manually analyzed axis of the incoming EW.
Figure 4.7 Infrared satellite imagery at 0600 of 19th and 20th of October. Red dots indicate the locations of individuals lightings within 30 minutes of each image. Lightning data is from the Worldwide Lightning Location Network (https://wwlln.net). The cyan cross marks the location of the maximum in vorticity at 950 hPa in the cyclogenesis region.
Figure 4.8. Key elements of the evolution of the precursor for Patricia. $\geq 2 \times 10^{-5}$ s$^{-1}$ vorticity regions in color at 700 hPa for stages I, II and III and at 950 hPa for stage IV. Dates representative of each phase in red. Also shown, the Papagayo wind at 950 hPa in stage I, the anomalous wind at 700 hPa for stages II and III, and the anomalous wind at 950 hPa for the 20$^{th}$ in stage IV. The track of the incoming EW at 500 hPa is indicated in the dashed line. Its location in heavy dots corresponds to the color of each date. The cyclogenesis location on the 20$^{th}$ and lightning locations for the same time are shown in stage IV.
5. **ANALYSIS OF THE DEVELOPING EASTERLY WAVE TROUGH FOR PATRICIA**

Chapter 4 presented the synoptic evolution leading up to the cyclogenesis of Patricia. There, it was found that an EW trough over the EPac provided a synoptic environment that led to genesis. The EW trough was influenced, by an incoming Caribbean EW and subsequently by a Tehuantepec gap wind event. The developing EW trough was formed in the Panama Bight (Phase I) and weakened shortly after its formation (Phase II). It was then approached by an incoming EW which merged with it (Phase III). Then a Tehuantepec gap flow appeared to contribute to increase the vorticity and to form a low-level circulation (950 hPa) in the EPac (Phase IV). This chapter examines the elements and mesoscale structures within the trough favorable to cyclogenesis.

Known elements that favor cyclogenesis include high environmental moisture content, high equivalent potential temperature ($\theta_e$), high low-level vorticity. Within the trough, characteristics favorable to cyclogenesis include the formation of convectively active circulations (Wang and Hankes, 2016). Closed circulations within the trough allow the air, isolated from the environment, may moisten, saturate and precipitate releasing latent heat by condensation favorable for TC genesis (Dunkerton, Montgomery and Wang, 2009). Meanwhile the trough’s vertical development involves the formation of a coherent vortex spreading in the vertical.

This chapter is organized as follows: section 5.1 presents the trough track and discusses the global evolution of the trough characteristics that may influence the probability of cyclogenesis; section 5.2 analyzes the air in the closed circulations within the trough and section 5.3 provides a summary of the chapter.
5.1 Evolution of the Developing Trough for Patricia

To recall, the developing trough was first detected at 500 hPa over the Panama Bight region (Fig. 4.6 (b)) as a positive vorticity anomaly. The location of the maximum vorticity value of this anomaly is used to estimate the trough track (Fig. 5.1 red line). Its track shows that the trough originates over the Panama Bight region, then moves parallel to Central America until the 16th. Afterward it slowly moves westward until the 20th, cyclogenesis time and then the nascent TC follows closely the best track (Fig. 5.1 black line).

Additionally, when the curvature vorticity at 950 hPa is spatially averaged within a 300 km radius a separate curvature vorticity anomaly was also found to be following the TC track (Fig. 5.1 blue line). This vorticity anomaly originated in the ITCZ by 1700 17th of October (map of curvature vorticity not shown). Initially, it was advected northeastward toward the cyclogenesis region by the southerlies of the trough. After cyclogenesis, this vorticity anomaly follows the TC track. Interestingly, this vorticity anomaly moved parallel to the southerlies of the trough (Fig. 4.4 (c,e)), this suggests that the ITCZ vorticity advected by the southerlies contributed to cyclogenesis by merging with the gap outflow vorticity.

5.1.1 Evolution of the Variables Within the Trough Favoring Cyclogenesis

The evolution of the variables that favor TC formation is explored through timeseries centered on the trough location (Fig. 5.2). These are: the vorticity (averaged within a 300 km radius at 950 hPa, and between 850 and 650 hPa); and the convection proxied by the infrared brightness temperature coverage under 240°K, and by the number of individual lightning strokes (within a 300 km radius disk). Also, we consider the environmental moisture to determine if the passing midlatitude system caused some entrainment in the vicinity of the trough (the specific humidity averaged within a 300–750 km annulus at 850 hPa). These variables are the same analyzed by
Brammer and Thorncroft, (2018) for a non-developing trough. The evolution of these variables reveals the following:

1) **Vorticity**: The vorticity between 850 and 650 hPa (red line Fig. 5.2) was characterized by a general increase after the 14\textsuperscript{th}, when the EW trough was approached by an incoming EW and both merged (Fig. 4.4 (e-h), stage III (Fig 4.8 (c)). On the 18\textsuperscript{th}, the vorticity weakens; the weakening is attributed to the midlatitude system passing over the GoM. Then after the 19\textsuperscript{th} (stage IV), the vorticity increases again. Finally, on the 20\textsuperscript{th} there is a small weakening in the vorticity, this is due to closeness of trough center with the negative vorticity patch of the gap wind. Despite the weakening on the 18\textsuperscript{th} and 20\textsuperscript{th}, the global trend is positive on these days, coinciding with the merging of the gap wind vorticity with the trough vorticity (Fig. 4.5 (e-h)). In general, the vorticity within the trough increased first when it merged with the EW, and then, when the trough merged with the vorticity of the gap wind.

2) **Vorticity at 950 hPa**: The vorticity at 950 hPa (red line Fig. 5.3) has a similar trend to that between 850 and 650 hPa. However, on the 19\textsuperscript{th}, the vorticity at 950 hPa increases substantially. At this time, the gap wind intensifies. This suggests that the gap outflow vorticity contributes to establish the low-level vorticity patch leading to cyclogenesis.

3) **Convection**: The convective activity represented by the cool IR temperature coverage (area of IR cloud temperatures less than 240°K) and the number of lightning strikes exhibits a pronounced diurnal cycle until the 16\textsuperscript{th}. Between the 11\textsuperscript{th} and the 16\textsuperscript{th} the trough center moves over land. Both, the lightning and the cool IR coverage signals peak at about 1800 UTC and correspond to a land-based diurnal cycle (Minobe et al, 2020). After the 16\textsuperscript{th}, when the trough’s center moves over the EPac, the diurnal cycle becomes less obvious. At this time, the cool IR coverage increases dramatically, and the number of strokes occurs continuously for the next two days indicating that
the EW trough is convectively active. This coincides with the enhancement of the trough southerlies which advected ITCZ vorticity. The convection then decreases on the 18th and increases again after the 19th. This abrupt decrease coincides with the passage of the midlatitude system over the GoM (Fig. 4.5 (d,f,h)), suggesting that the midlatitude system weakened the trough by shearing the environment.

4) **Humidity**: The environmental moisture (black solid line) remained roughly constant through the trough’s life, and after cyclogenesis. The moisture values show that the trough was embedded in a moist atmosphere. Nevertheless, is notable that the moisture remains unchanged despite the trough moves initially over land. However, it is recalled that during this time of the year, the atmosphere is moist over Central America and convection is frequent (Magana et al, 1999).

Thus, the vorticity of the trough increased continuously after it interacted with the incoming EW likely due to the vorticity merging on the 14th between the EPac trough with the incoming Caribbean EW (Stage II). Then, on the 16th, the center of the trough moved over the EPac, and it became convectively active, and the vorticity increase on the 17th. On the 18th (Stage III) the convection and vorticity weakened, suggesting that the midlatitude system over the GoM weakened the trough, nevertheless this weakening was transient, and vorticity and convection increased again on the 19th and cyclogenesis occurred on 0600 20th.

**5.2 The Low-Level Circulation Leading to Cyclogenesis**

This section explores the development of the vortex leading to Patricia in terms of the wind, the precipitation and of the equivalent potential temperature ($\theta_e$). The vortex is characterized by stacked closed streamlines of the relative wind at 950 and 700 hPa. Within the closed streamlines $\theta_e$ maximizes and precipitation occurs. The discussion is similar to that of Brammer
and Thorncroft, (2018). Additionally, we explore the potential contributions of the curvature vorticity anomalies originated in the ITCZ to the vortex leading to cyclogenesis (see section 5.1).

It should be noted that closed streamlines are not actual material walls that compartmentalize the fluid yet can be used as proxies for them provided the wind varies slowly. However, they are indicative of regions favorable to cyclogenesis provided air within them is moist. Closed streamlines persistent in time and vertically stacked diagnose a favorable environment to develop the trough, because they signal a coherent vortex which minimizes entrainment (Tang and Emmanuel, 2010). On the contrary if the circulations are tilted they are less resilient (Brammer and Thorncroft, 2018).

5.2.1 Horizontal Structure and Development of the Low-Level Circulation Leading to Cyclogenesis

Figure 5.4 shows the relative wind, closed streamlines and $\theta_e$ at 950 and 700 hPa, and the precipitation for the 1800 18th, 1200 19th and 0600 20th. In general terms, the sequence of time frames shows a set of vertically stacked closed streamlines, two at 700 (Fig. 5.5 (a,b,c)) and one at 950 hPa (Fig. 5.5 (d,e,f)). This suggests that low-level air may recirculate within the closed streamlines, moistens, then ascends and precipitates leading to an overall increase vorticity and temperature $\theta_e$ that favor cyclogenesis.

The relative wind at 700 hPa forms two concentric closed streamlines. The outer one initially spans from the EPac to the Yucatan peninsula (the SW-NE oriented closed streamline in Fig. 5.4 (a)) and is representative of the developing EW trough. The inner one locates over the EPac but shrinks by cyclogenesis time (Fig. 5.4 (f)). The time frames show that the outer streamline retraces from the Yucatan peninsula to the EPac. Its shrinking coincides with the passage of the midlatitude system which may have sheared the EW trough over the Yucatan peninsula. The inner
streamline, before cyclogenesis is collocated with the closed streamline at 950 hPa forming coherent vortex, however the inner streamline shrinks by cyclogenesis time (Fig. 5.4 (c)). Thus, as the trough weakened over the Yucatan peninsula, the coherent vortex over the EPac remains intact despite the inner closed streamline at 700 hPa shrunk yet the outer remains.

Regarding the closed streamline at 950 hPa, it is located between the gap northerlies and the trough southerlies. This strongly suggests that the formation of this closed streamline was impacted by the gap wind indicating an influence of the gap wind on the circulation leading to cyclogenesis.

\( \theta_e \) maximizes, at 950 hPa within the closed streamline at 950 hPa, and increases between the 1800 18\(^{th}\) to 0600 20\(^{th}\) from 348°K to 355°K. At 700 hPa, \( \theta_e \) maximizes along an arc within the outer closed streamline this arc is indicative of the upward transport of moisture and heat within the outer closed streamline favoring cyclogenesis. Overall, \( \theta_e \) has a structure favorable to cyclogenesis.

In addition to the relatively high \( \theta_e \) within the closed streamlines, two tongues of low \( \theta_e \) are seen, one coming through the mountain gap of the Isthmus of Tehuantepec at 950 hPa (Fig. 5.4 (c)) and another from the Equator at 700 hPa (Fig. 5.4 (f)). The tongue over the Isthmus of Tehuantepec is due the midlatitude system which transports low \( \theta_e \) into the tropics through the mountain gap. Even more, the tongue of low \( \theta_e \) at 950 hPa illustrates that air associated with the midlatitude system over the GoM penetrates into the EPac Regarding the Equatorial tongue at 700 hPa is due to subsidence associated with convection, as will be discussed in the next paragraphs.

Regarding the tongue of low \( \theta_e \) at 950 hPa, is worth noting that is close to the closed streamline at 950 hPa and the maximum \( \theta_e \) tongue at 950 hPa. Their closeness suggests that the gap wind could have transported air with low \( \theta_e \) which would be detrimental for cyclogenesis.
Nevertheless, this effect, if occurred at all should be negligible since cyclogenesis took place. This potential will be explored in the next chapter.

The precipitation forms a curved band on close to the periphery of the outermost streamline at 700 hPa, being maximum within it at (93°W, 14°N) by genesis time where deep convection occurs (Fig. 5.5 (i)). This curved band of precipitation coincides with the curved band of clouds in satellite imagery (Fig. 4.7 (b)). This means that within the outermost closed streamline: 1) air moistens, 2) that vorticity is increasing by convection, and 3) the release of latent heat by condensation helps to develop the wind. Leading to develop favorable environment to cyclogenesis.

Overall, this configuration consisting of precipitation associated with deep convection and high $\theta_e$ within the closed streamlines has been associated with an overall strengthening of developing troughs during the pre-genesis stages (Dunkerton et al, 2009; and Wang and Hankes, 2016), and portrays a structure favorable for cyclogenesis.

5.2.2 Contributions of the ITCZ Vorticity to Cyclogenesis and Vertical Development of the Vortex

The contributions of the ITCZ vorticity to cyclogenesis are illustrated by the averaged curvature vorticity (950 hPa) within a radius of 300 km between the 0600 18th and 0600 20th. The vertical development of the trough vorticity and $\theta_e$ over the EPac is illustrated through a sequence of vertical cross sections across a SW-NE rhumb line ($S_1$, dashed line in fig. 5.1).

The 950 hPa curvature vorticity shows a curvature vorticity anomaly at (96°W,10°N), the ITCZ vorticity anomaly (cross hairs in Figure 5.5 (a)) and other at (91°W,13°N), the EW trough vorticity on 0600 18th. The track of the ITCZ vorticity anomaly is indicated in blue (Fig. 5.5). On the next day, the ITCZ vorticity anomaly merges with the EW trough vorticity and becomes
undistinguishable from each other (Fig. 5.5 (b)). After merging, the vorticity increases for the next day, and on 0600 20\textsuperscript{th} the vorticity led to cyclogenesis (Fig. 5.5 (c)). In general terms, the trough vorticity in the EPac increases after an ITCZ vorticity has merged with it. Thus, the ITCZ vorticity is a source of vorticity to the EW trough prior cyclogenesis.

The vertical cross sections show the evolution of the spatially averaged $\theta_e$ and curvature vorticity within a 300 km radius (Fig 5.6). Also, they show the potential contributions of the curvature vorticity anomaly originating in the ITCZ to the developing EW trough over the EPac. $S_1$ results from the linear fit of the ITCZ curvature vorticity anomaly track between the 18\textsuperscript{th} and 20\textsuperscript{th} (Fig. 5.1). The vertical cross sections span over the EPac (left part), Central America (central part, black shading) and the Caribbean (right part).

The curvature vorticity anomalies show the evolution of a low-level anomaly (vertical dashed line) over the EPac moving northeastward to Central America and approaching to another curvature vorticity anomaly there (Fig. 5.6). These anomalies are embedded in a high $\theta_e$ environment (> 340°K) surrounding Central America. When the ITCZ curvature vorticity anomaly approaches the anomaly located to the southwest of Central America on the 0600 19\textsuperscript{th}, the curvature vorticity in the EPac increases and spans along the vertical (Fig. 5.6 (b-c)). The resulting column indicates the nascent TC at cyclogenesis time (Fig. 5.5 (c)). Simultaneously, the vorticity in the Caribbean weakens, and $\theta_e$ at low levels (below 700 hPa) increases over the EPac.

The arrival of the ITCZ vorticity anomalies to Central America coincide with the increase and vertical development of the vorticity there. It suggests that the ITCZ vorticity anomalies merge with the vorticity in the cyclogenesis region contributing to develop the vortex leading to cyclogenesis. The merging of the ITCZ vorticity with the vorticity southwest of Central America may lead to an increase of the low-level wind, increasing the surface heat fluxes causing $\theta_e$ to
increase. However, to what extent these vorticity anomalies are important to cyclogenesis will be further explored in chapter 7.

5.3 Summary

The developing EW trough for Patricia had a steady increase of vorticity between 850 and 650 hPa after it merged with the incoming Caribbean EW on the 14th of October (stage II). By the 16th of October the trough became convectively active when its vorticity started to increase over the EPac, and its center was displaced to the EPac due to a weakening or removal of the Yucatan component. After the 18th several elements associated with cyclogenesis of Patricia were detected (Phases III and IV)

- **Low-Level circulation leading to cyclogenesis**: By the 1800 18th, closed streamlines at 950 and at 700 hPa formed within the trough over the EPac. within the closed streamlines, convection, precipitation, as well as high \( \theta_e \) at 950 and 700 hPa were present providing a favorable region for cyclogenesis which occurred about a day later.

- **ITCZ vorticity contribution**: A curvature vorticity anomaly originated in the ITCZ and was transported by the southerlies, and potentially contributing to increase the vorticity in the cyclogenesis region. This vorticity anomaly was initially shallow, but later stretched vertically and merged with the vorticity in the cyclogenesis region over the EPac. By cyclogenesis time, the anomaly was part of a vorticity column associated with the nascent TC. Therefore, vorticity from the ITCZ appears to increase the vorticity in the cyclogenesis region.

- **Interaction between the circulation leading to cyclogenesis and the gap wind**: The circulation leading to cyclogenesis at 950 hPa formed close to the gap wind outflow. We, therefore, expect some interaction between the gap wind outflow and the circulation. The outflow could have contributed to increase the vorticity within the circulation, but simultaneously could have
ventilated the circulation since it was associated with low $\theta_e$ values. This potential interaction is the matter of the next chapter.

**Figures**

**Figure 5.1.** Track of the precursor trough every hour given by the maximum vorticity at 500 hPa in red. Track of the small blob smoothened maximum curvature vorticity at 950 hPa every hour in blue. Heavy dots indicate the locations two days since the 0600 UTC 10th October (red or blue numbers). Best track of Patricia in black. Vertical cross sections are plotted along SW-NE line S1.

**Figure 5.2.** Time series of analysis and observed variables. GridSat IR areal coverage (dashed black and red) and IMERG mean precipitation rate (solid blue) within 300-km radius. ERA-5 specific humidity at 850 hPa averaged between 300-750 km (solid black line). ERA-5 850-600 hPa mean vorticity within 300 km radius (solid red line). ERA-5 950 hPa vorticity within 300 km radius (solid green line). Horizontal axis in days of October.
Figure 5.3. Time series of analysis and observed variables. GridSat IR areal coverage (dashed black and red) and IMERG mean precipitation rate (solid blue) within 300-km radius. ERA-5 specific humidity at 850 hPa averaged between 300-750 km (solid black line). ERA-5 850-600 hPa mean vorticity within 300 km radius (solid red line). ERA-5 950 hPa vorticity within 300 km radius (solid green line). Horizontal axis in days of October.
Figure 5.4. Selected streamlines in red (700 hPa) and orange (950 hPa), $\theta_e$ in color shading, relative to the developing trough wind at 700 (A, B and C) and 950 hPa (D, E and F), for the 1800 UTC 18th (A, D, G), 1200 UTC 19th (B, E, H) and 0600 UTC 20th (C, F, I). Panels G, H and I show the IMERG precipitation in color shading and lightning locations in red for the same dates. The panels are centered at the precursor’s location red crosshairs. Criss-cross hatching in d, e and f indicates the orography at 950 hPa.
Figure 5.5. Map of the curvature vorticity ($\times 10^{-5}$s$^{-1}$, shading) at 950 hPa for the 0600 UTC, 18th, 19th and 20th of October in A, B and C respectively. Also shown the track of the maximum curvature vorticity value originated in the ITCZ and its location at each time (cross hair).
Figure 5.6. Vertical cross sections along the rhumb line $S_1$ of the vorticity, in color shading and $\theta_e$, in °K for the 0600 UTC, 18th, 19th and 20th of October in A, B and C respectively. Vertical line marks the location of the small curvature vorticity anomaly. Black shading indicates the topography in Central America, left of it lies the EPac and right of it the Caribbean. Each horizontal tick is placed about 100 km.
6. IMPACTS OF THE TEHUANTEPEC GAP WIND ON THE CYCLOGENESIS OF HURRICANE PATRICIA

Chapter 4 highlighted the presence of a Tehuantepec gap wind event during the cyclogenesis of Patricia. It was found that the vorticity patch associated with the gap outflow merged with the vorticity associated with the vorticity anomalies underneath the developing EW trough. Chapter 5 showed that within the cyclogenesis region, the circulation leading to cyclogenesis formed close to the gap wind outflow such that the gap wind northerlies are indistinguishable from those of the circulation. These observations motivate us to consider the possible potential positive influence of the gap wind on the genesis of Patricia.

Additionally, chapter 5 showed that the gap outflow at cyclogenesis time was characterized by low $\theta_e$. This suggests that the gap wind outflow could have ventilated the low-level of the cyclogenesis region having a negative effect on cyclogenesis. Nevertheless, it is expected that this possible negative effect to be insignificant given that cyclogenesis took place.

This chapter analyses the interactions between the gap wind and the developing trough on the days prior to cyclogenesis and consists of three sections; 6.1) the gap wind intensity and its structure prior to cyclogenesis, 6.2) the gap outflow contributions of the vorticity and $\theta_e$ to the circulation leading to genesis of Patricia, and 6.3) a general discussion and summary.

6.1 The Gap Wind

The gap wind is analyzed first, in terms of $I_T$ and then in terms of its spatial structure. The $I_T$ for these days is compared with the climatology to assess its significance. The structure of the gap flow is explored through vertical cross sections of the wind and through a Lagrangian analysis of the air parcels flowing through the mountain gap. The vertical cross section will show the structure of the gap wind prior cyclogenesis. The Lagrangian analysis will reveal the origins
of the gap wind and an assessment of the extent to which the gap wind air reaches the low-level circulation that led cyclogenesis.

6.1.1 Evolution of the Gap Wind

Figure 6.1 shows $I_T$ and the monthly climatology including the 5th to 95th percentile range. The figure shows the gap wind northerlies ramp up between the 13th and the 20th of October. They peaked at -17 ms$^{-1}$ around 0000 20th, which is about 2 standard deviations and clearly is a gap wind event. This peak occurs about 7 hours before cyclogenesis. After 1800 20th, the northerlies weaken drastically, and the gap wind turned southerly and became extreme (9 ms$^{-1}$) on the 24th by the time of Patricia cyclolysis (see crosshairs). The change in the direction of the gap wind during Patricia’s life, indicates that both Patricia and the gap wind were related, i.e., at cyclogenesis the gap wind was northerly and at cyclolysis the gap wind was southerly.

Between the 20th and the 24th, the anticyclonic flow associated with the passing midlatitude system dominates Mexico and the GoM (Fig. 6.2). This flow modulated the direction of the gap wind and the track of Patricia. At cyclogenesis time (0600 20th), the gap northerlies are due to the anticyclonic flow that extends to the west of the developing trough over the Yucatan peninsula and merge with the nascent TC (Fig. 6.2 (a)). Then, at cyclolysis time (0600 24th) the gap southerlies are due to the Patricia’s southerlies as it made landfall (Fig. 6.2 (b)). In turn the anticyclonic flow associated with the midlatitude system, make the TC to recurve and to landfall (Roger et al, 2017). Thus, the midlatitude system modulates both the gap wind and Patricia track.

6.1.2 Spatial Structure of the Gap Wind

The mountain gap of the Isthmus of Tehuantepec imposes a physical constraint on the wind flowing toward the cyclogenesis region. The northerlies flowing through it may influence
the transport of vorticity and $\theta_e$ into the cyclogenesis region. These northerlies, as well as the vorticity and $\theta_e$ are studied in vertical cross sections at constant latitude (17°N).

Figure 6.3 shows the vertical west-to-east cross section of the horizontal wind, together with its vorticity and $\theta_e$ for 0600 18th, 0600 19th and 0600 20th of October. These time frames are representative of the period between the time when the developing trough stalls over the Yucatan peninsula (Phase III) until cyclogenesis (Phase IV). In general terms, the vertical cross sections show northerlies within and over the mountain gap, intense vorticity along the mountain slopes within the gap. Also, $\theta_e$ fluctuates in time along the western slope of the gap but increases in general over its eastern slope.

The vertical cross sections show a steady intensification of the northerlies within the mountain gap between the 18th and the 20th (Fig. 6.3 (b,c,d)). This is consistent with the intensification of $I_T$ in the same period (Fig. 6.1). As the northerlies intensify, the vorticity patches at the bottom of the mountain gap (below 900 hPa) broaden and intensify. The vorticity generated by the gap wind at low-levels (below 900 hPa) may be an important vorticity source in the EPac that helps to form the low-level vortex leading to cyclogenesis.

The vertical sections between the 16th and the 20th also show $\theta_e$ (color shading, Fig. 6.3). $\theta_e$ is lower in the western slope of the mountain gap than in the eastern slope, also $\theta_e$ over each slope has a characteristic behavior. Over the western slope, $\theta_e$ fluctuates and over the eastern slope increases steadily. The fluctuations over the western slope are indicative of midlatitude air masses flowing through the mountain gap (Fig. 6.3 (a,d)). This is consistent with the tongue of low $\theta_e$ at the 20th (Fig. 5.4 (f)) and with the presence of midlatitude systems over the GoM. The increase over the eastern slope of the gap coincides with the arrival of the developing trough to the Yucatan peninsula on the 18th. (Fig. 4.5 (d)). This means that the trough northerlies transport high $\theta_e$ toward
the EPac. The arrival of the developing trough over the mountain gap is further evidenced in the cross sections by the shift on the east of the mountain gap of the wind direction from easterly, on the 16th (Fig. 6.3 (a)) to southerly between the 18th and the 20th (Fig. 6.3 (b,c,d)). Thus, then, air masses of midlatitude (associated with the midlatitude system) and tropical origin (associated with the developing trough) flow through the mountain gap. It is of interest here to assess any potential negative impact of air with low $\theta_e$, of midlatitude origin, on cyclogenesis.

In summary, the gap wind northerlies are associated with the southward transport of positive vorticity between the surface and 900 hPa and with the transport of low and high $\theta_e$ values between the surface and 800 hPa prior cyclogenesis. The positive vorticity potentially contributes to develop the low-level vortex leading to cyclogenesis. But to what extent the potential negative impacts of low $\theta_e$ air flowing through the mountain gap on cyclogenesis, if any, are yet to be understood.

6.1.3 Spatial Extension of the Gap Flow and Interactions with the Vortex Leading to Cyclogenesis

To study the transport of gap-flow vorticity and $\theta_e$ toward the cyclogenesis region in the EPac, a Lagrangian approach is taken. Moreover, it allows to reveal the origins of the gap flow, and its fate as it moves into the EPac. Also, it indicates the spatial extension of the gap flow, which is useful to determine to what extent the gap flow interacted with the vortex leading to genesis.

Lagrangian Methodology

The Lagrangian representation of the gap flow consists of a set of parcel trajectories that are chosen to pass through the mountain gap. These parcels pass at every grid point of the ERA-5 within the mountain gap, and every hour between the 14th to the 22nd of October. In other words, the parcels located every quarter degree, eta level, and hour within the interval
[96°W,93°W]×[16.5°N,17.5°N]×[Surface,800 hPa] conform the gap flow (Fig. 6.4). These parcels are referred to as the mountain-gap parcels and their trajectories are calculated backward and forward in time for two days.

The parcel locations at a given time indicate the extension of the gap flow. To avoid considering parcels that move too far from others, only those neighboring other 75 parcels within a 10,000 km² vicinity (100x100 km² or ~1°x1° at the Equator) in a vertical layer with thickness of 50 hPa are considered. This density threshold is representative of the grid points density in the ERA-5.

**Spatial Extension of the Gap Flow**

Figures 6.5 and 6.6 show the spatial extension of the gap flow, between the surface and 950 hPa on the 16th, 18th, 19th, and 20th. This layer is referred to as the 950 hPa layer. We choose this layer to study the interactions of the gap flow with the trough at low levels. The figures also show the isochrones for departure times from the gap exit, in hours relative to the time atop each panel. The departure times are proxies to the parcel motion. Positive isochrones indicate the elapsed time since a parcel departed the mountain gap and correspond to the outflow. Negative isochrones indicate the time it takes for a parcel to arrive to the mountain gap (locations shown in Fig. 6.1) and correspond to the inflow. For example, the +24 hours isochrone indicates the parcels that have departed the mountain gap 24 hours ago; the -24 hours isochrone indicates the locations for parcels that will arrive to the gap in 24 hours.

The gap inflow at the 950 hPa layer between the 16th and the 20th extends over a wide area in the GoM, the Yucatan peninsula and the Caribbean (Fig. 6.5). The fraction of the inflow originated in the GoM flows anticyclonically and is associated with the midlatitude system, i.e. is of midlatitude origin. The fraction originated over the Yucatan peninsula and the Caribbean is
tropical air associated with the developing EW trough. Thus, two different air masses, tropical and midlatitude flow through the mountain gap. The midlatitude air has low $\theta_e$ relative to the tropical air. Their different $\theta_e$ potentially impact the thermodynamical properties of the low-level vortex leading to cyclogenesis. In particular is of interest to understand the role of the midlatitude air, which may have detrimental impacts on its formation.

Regarding the gap outflow, it forms a narrow strip extending southward over the EPac and has a more complicated evolution (Fig. 6.5). Its evolution consists in a curvature change between the 16$^{th}$ and the 18$^{th}$. The outflow has an anticyclonic curvature on the 16$^{th}$, as indicated by the clockwise increase of the departure times. Then, on the 18$^{th}$ most of the outflow has a cyclonic curvature, indicated by the counterclockwise increase of the departure times. The shift from anticyclonic to cyclonic curvature in most of the outflow coincides with the establishment of the cyclonic flow in the cyclogenesis region underneath the trough (see oval in Fig. 5.4 (c)). The change in the curvature of the outflow is consistent with the observations of Bourassa et al (1999). Their observations consist of a change of the gap outflow curvature due to the presence of low pressures to the east, or southeast of the mountain gap typically associated with hurricanes. Thus, the change of the outflow curvature from anticyclonic to cyclonic indicates that the gap wind outflow was influenced by the low pressure associated with the developing trough over the EPac.

It is notable that some of the mountain gap parcels flowing cyclonically over the EPac penetrate a closed streamline on the 19$^{th}$ and the 20$^{th}$ (magenta ovals, Fig. 6.5 (c,d)). The closed streamlines are with respect the storm-relative wind and delimit the low-level vortex at 950 hPa that led to cyclogenesis. We want to recall that the closed streamlines are not actual material walls moreover when there are vertical motions, however their persistence in time are signals of vortices. This penetration clearly indicates a material connection between the gap outflow and the nascent
vortex, meaning that the gap outflow transports material properties like vorticity and $\theta_e$ into the nascent vortex.

6.1.4 Summary

In summary, the gap wind northerlies before and at cyclogenesis time were extreme. Two different air masses conform the gap flow, one of tropical origin (high $\theta_e$) and other of midlatitude origin (low $\theta_e$). A fraction of the low-level outflow (below 950 hPa) penetrated the closed streamlines in the EPac. The closed streamlines are representative of the vortex leading to Patricia at low levels. The penetration of the outflow into the region of closed streamlines shows a material connection between the gap wind and the nascent vortex. Through this connection the gap outflow transports vorticity and $\theta_e$ toward the vortex potentially influencing cyclogenesis. These potential influences are discussed next.

6.2 Potential Effects of the Tehuantepec Gap Wind on Cyclogenesis

Mountain-gap parcels flowing into the region where cyclogenesis is taking place (e.g. by the closed streamlines in Fig. 6.5) bring vorticity and $\theta_e$ to the region influencing the cyclogenesis process. If these mountain-gap parcels bring high vorticity and high $\theta_e$, then the effect of the gap outflow will favor cyclogenesis. On the contrary if they transport low vorticity or low $\theta_e$ the gap outflow will have a detrimental influence on cyclogenesis.

To reveal if the gap outflow transports high vorticity and high $\theta_e$ relative to the vorticity and $\theta_e$ in the cyclogenesis region we will compare the vorticity and $\theta_e$ values of the mountain-gap parcels flowing into the cyclogenesis region against the vorticity and $\theta_e$ within the cyclogenesis region. The cyclogenesis region is defined by a fixed in space circle with a 200 km radius centered on the cyclogenesis location (cross in Fig. 6.6 (a)). The circle covers the closed streamlines of the storm-relative wind at low-levels associated with the vortex leading to cyclogenesis between the
18th and the 20th (e.g., Fig. 5.4 (d,e,f) and Fig. 6.4 (c,d)). Moreover, the closed streamlines for the first hours after cyclogenesis, are also covered by this circle.

The upcoming analysis focuses on the parcels in the 950 hPa layer because it is where the vorticity is the most intense in the mountain gap (see Fig. 6.2). Also, because is in this layer where heat and moisture fluxes influence $\theta_e$ potentially modify the thermodynamics of the gap wind air flowing over the into the cyclogenesis region.

6.2.1 Impacts of the Gap Outflow Vorticity on the Cyclogenesis Region

The timeseries of the median vorticity between the surface and 950 hPa within the cyclogenesis region is shown in Fig. 6.6 (b). The timeseries show that in the cyclogenesis region the vorticity increases between the 18th and the 20th, consistent with the formation of a vortex in the region on those days. Before the 18th the median vorticity remains roughly constant. When comparing the gap outflow vorticity with that of the cyclogenesis region it is seen that its median vorticity values correspond approximately to the 75th percentile value within the cyclogenesis region between the 18th and the 20th (gray envelope in Fig. 6.6 (b)). This means that the difference between the median vorticity of the gap outflow and the median vorticity in the cyclogenesis region is statistically significant at the 95% level. Additionally, the vorticity increase is coincident with the arrival of enhanced gap flow. The relative high vorticity values of the gap outflow are interpreted as an enhancement to the vortex vorticity. The contributions to the gap outflow vorticity to enhance the vortex will be further discussed in chapter 9.

6.2.2 Impacts of the Gap Outflow $\theta_e$ on the Cyclogenesis Region

Regarding $\theta_e$ in the cyclogenesis region, it is seen that its median $\theta_e$ increases from 352K to 355K between the 17th and the 20th (Fig. 6.6 (c)) and indicates that $\theta_e$ in the cyclogenesis region does not change much prior cyclogenesis. When the median $\theta_e$ of the gap wind outflow is
compared with the median $\theta_e$ of the cyclogenesis region, it results that the median $\theta_e$ of the gap outflow results slightly higher than the median $\theta_e$ of the cyclogenesis region. Their difference is not statistically significant. This indicates that the gap outflow does not contribute to increase $\theta_e$ in the cyclogenesis region, meaning that $\theta_e$ increases locally instead of being transported from the gap outflow. However, considering that through the mountain gap, air of midlatitude origins flows, it means that the mountain-gap parcels have their $\theta_e$ quickly modified over the ocean increasing their $\theta_e$ and their detrimental effect to cyclogenesis is nullified.

### 6.2.3 Evolution of the Mountain-Gap Parcel’s Vorticity and $\theta_e$ Flowing into the Cyclogenesis Region

The gap outflow transports high positive vorticity into the cyclogenesis region and should help to increase the overall vorticity of the vortex leading to cyclogenesis (Fig. 6.6 (b)). In the case of $\theta_e$, the gap outflow transports air with similar $\theta_e$ to the cyclogenesis region, meaning that $\theta_e$ in the cyclogenesis region is not influenced by the gap outflow. Thus, we want to understand how the vorticity and $\theta_e$ of the mountain-gap parcels evolve before penetrating the cyclogenesis region to better understand the dynamics of the gap outflow. Considering that the vorticity (Fig. 4.5) and $\theta_e$ (Fig. 5.4) of the gap inflow is lower over the GoM than in the EPac, it is worth exploring the evolution of the gap outflow vorticity and $\theta_e$ into the cyclogenesis region.

The evolution of the mountain-gap parcels vorticity and $\theta_e$ that penetrate the low-level closed streamline under 950 hPa by cyclogenesis time, 0600 20th is explored next (parcels inside the magenta oval in Fig. 6.5 (d)).

**Evolution of the Vorticity of the Parcels that Penetrate the Vortex leading to Patricia**

The evolution of the vorticity of the mountain-gap parcels penetrating the low-level closed streamline at cyclogenesis time is shown in Fig. 6.7 (a). These parcels are those inside the
magenta oval in Fig. 6.5 (d). Their median vorticity is negative on 0600 18\(^{th}\), and the IQR shows that most of the parcels have negative vorticity as well (Fig. 6.7 (a)). Then, on 0000 19\(^{th}\) their vorticity increases, and half of the parcels have positive vorticity. On 1200 19\(^{th}\) their vorticity increases substantially until 0600 20\(^{th}\), when cyclogenesis takes place. These times, 0600 18\(^{th}\), 0000 19\(^{th}\), and 1200 19\(^{th}\), are key for the evolution of their vorticity, thus it is important to know what causes the vorticity to change at these times.

Figure 6.7 (b) shows the location of the parcels every hour (points), and the vorticity at those locations and times (shading). The contours delimit regions that encompass about 75% of these mountain-gap parcels at each key times (magenta for 0600 18\(^{th}\), green for 0000 19\(^{th}\), gray for 1200 19\(^{th}\) and black for 0600 20\(^{th}\)) and are representative of the parcel’s location. Figure 6.7 (b) reveals that most of the parcels on the 0600 18\(^{th}\) (magenta contour) flow southward and very close to the Sierra Madre. On 0000 19\(^{th}\) the parcels flow across the mountain gap (green contour). By 1200 19\(^{th}\) they flow over the EPac (grey contour) and on 0600 20\(^{th}\) they locate over a hook-like region within the vortex (black contour).

To understand the sources of vorticity for the parcels we have considered their vorticity budget:

\[
\frac{d\zeta}{dt} = -(\zeta + f)\nabla \cdot \bar{u} - \nu \frac{\partial f}{\partial y} - \hat{k} \cdot \left( \nabla \omega \times \frac{\partial \bar{u}}{\partial p} \right) + \text{Res}
\]

Where \(\zeta = \partial v / \partial x - \partial u / \partial y\) is the vorticity, \(d\zeta / dt\) is the total derivative of the vorticity, \(f\) is the Coriolis parameter, \(\nabla = (\partial / \partial x, \partial / \partial y, \partial / \partial p)\) is the gradient, \(\bar{u} = (u, v, 0)\) is the horizontal velocity vector, \(\omega\) is the vertical velocity (Pa/s), \(p\) is the pressure, \(\hat{k} = (0, 0, 1)\) and Res is a residual term accounting for other sources. The first term of the right-hand side in the vorticity equation is the stretching, the second the advection of Coriolis, and the third is the tilting.
The vorticity budget of the mountain gap parcels close to the surface (under 950 hPa and very close to the Sierra Madre) before flowing through the mountain revealed that the residual in the vorticity equation is about 50% of the vorticity budget. We speculate that their vorticity is likely due to shear caused by the surface friction, but this should be further explored. Afterward, when the mountain-gap parcels flow southward over the EPac the residual term in the vorticity equation decreases. Between 1200 19th (gray shading) and 0600 20th (black shading) when these parcels start to flow cyclonically, the vorticity budget is dominated by the stretching terms, which accounts for about 80% of the vorticity budget (Fig. 6.6 (c)). Thus, the vorticity of the mountain-gap parcels is likely driven by the surface friction associated with the Sierra Madre when they are upstream of the gap. Then, their vorticity increases due to stretching when they flow cyclonically over the EPac.

**Evolution of the Mountain-Gap Parcel’s \( \theta_e \) Flowing into the Vortex leading to Patricia**

The \( \theta_e \) along the trajectories of the mountain-gap parcels that flow into the cyclogenesis region is shown in Fig. 6.8 (a). The figure shows a southward increase of \( \theta_e \) over the EPac and when the parcels flow cyclonically (gray and black contours). The timeseries of \( \theta_e \), indicated by the median and IQR (Fig. 6.8 (c)) show a steady increase in time and a spread that narrows in time. It is notable that their \( \theta_e \) increases dramatically between 0000 19th and 1200 19th, after the parcels cross the mountain gap. Afterward, \( \theta_e \) increases at a slower rate yet its spread keeps narrowing until cyclogenesis in 0600 20th.

The sudden \( \theta_e \) increase after the parcels cross the mountain gap by 0000 19th is due to the heating and moistening by the temperature and moisture differences between the sea and the atmosphere. As they travel southward, the heating and moistening diminishes and \( \theta_e \) stops increasing and their spread narrows, meaning that they are in equilibrium with their surroundings.
Additionally, the narrowing and slow $\theta_e$ increase afterward 0000 19th, indicates that parcels have achieved a similar $\theta_e$ of the cyclogenesis region (Fig. 6.4 (e)). This means that the $\theta_e$ of most of the mountain-gap parcels only achieve the $\theta_e$ of the whole cyclogenesis region and do not contribute to increase the $\theta_e$ in the cyclogenesis region.

It is notable that when the parcels enter the cyclogenesis region, after the 1200 19th, they are close to saturation, with relative humidity values greater than 80% (Fig. 6.8 (d)) and just like in the case of $\theta_e$, the parcels RH values are similar to those in the cyclogenesis region (not shown). Nevertheless, these RH values are indicative of environments favorable to cyclogenesis (Holland, 1996), thus the potential negative effects of the gap outflow on the development of the vortex fail to materialize.

6.2.3 Summary

In summary, the low-level (below 950 hPa) gap outflow transports relatively high vorticity into the cyclogenesis region, i.e., the low-level vortex proxied by the closed streamlines, favoring cyclogenesis. Regarding the $\theta_e$ of the gap outflow; its $\theta_e$ appears to not have any impact on the $\theta_e$ of the vortex leading to cyclogenesis since they have similar values to the environment. Thus, the gap outflow does not ventilate the low-level vortex considering that air with low $\theta_e$ flow through the mountain gap and then penetrates the cyclogenesis region.

Additionally, it was observed that the mountain-gap parcels have their vorticity upstream due to friction as they flow close to the mountain. Over the EPac approaching the cyclogenesis region, their vorticity further increases mainly by stretching.

6.3 Summary and Discussion

The Tehuantepec gap northerlies were extreme prior to cyclogenesis of hurricane Patricia (Fig. 6.1), around -17 ms$^{-1}$. Initially, the gap outflow interacted with the developing trough by
becoming cyclonically curved. Subsequently, the mountain-gap parcels flowed into the vortex leading to cyclogenesis (represented by closed streamlines at 950 hPa) and a material connection between the gap outflow and the vortex was formed. Through this connection, the gap outflow vorticity and $\theta_e$ were transported into the cyclogenesis region impacting its development.

With respect the material connection between the gap outflow and the low-level vortex, the following was found:

- High values of vorticity within the cyclogenesis region were associated with air from the mountain gap. This means that the gap outflow contributed with high vorticity to the cyclogenesis region favoring its development.

- The mountain gap parcels flowing into the cyclogenesis region had similar $\theta_e$ values to those in the cyclogenesis region, suggesting that the gap outflow does not have a significant impact on the $\theta_e$ in the cyclogenesis region. Also, it was found that mountain gap parcels with initially low $\theta_e$ are moistened and heated until they reach equilibrium with the environmental $\theta_e$.

The structure of the flow crossing the mountain gap of the Isthmus of Tehuantepec revealed the following:

- The highest values of positive vorticity were limited to the lower (surface to 950 hPa) atmosphere on the eastern flank of the mountain gap. The vorticity generated in the mountain gap is due to shear caused by the air flowing close to the mountain slopes. Then, the gap outflow vorticity increases by stretching once it reaches the cyclogenesis region.

- Through the mountain gap, air with different values of $\theta_e$ ranging from 330 to 350°K flow. This spread of $\theta_e$ is attributed to different air sources flowing through the mountain gap.
The gap outflow and the vortex that led to genesis were materially connected. This means that the air flowing through the mountain gap transported vorticity and $\theta_e$ to the low-level vortex impacting its evolution.

Therefore, the impacts of the gap outflow on cyclogenesis appear to be more dynamic than thermodynamic, i.e. the gap outflow contributes to increase the vorticity in the cyclogenesis region. However, it is yet to be determined whether this vorticity contribution is necessary for cyclogenesis. To explore if the gap outflow vorticity contribution is necessary or not, the gap outflow will be suppressed through a numerical experiment, which will be described in chapters 8 and 9.

**Figures**

![Figure 6.1. Mountain gap wind index for October 2015 in black. Monthly 5th-95th percentile range in gray. Cyclogenesis (P15) and cyclolisis times are indicated by red-crossed circles. Horizontal axis is time in days of October.](image-url)
Figure 6.2. Horizontal wind at 850 hPa for 0600 20 (A) and 0600 24 (B). Also is shown the track of Patricia in red and its position for each date (heavy red dot).

Figure 6.3. Vertical cross sections of the horizontal wind (barbs), $\theta_e$ (color shading) and vorticity of the horizontal wind (isopleths every $10^{-4} s^{-1}$, red and blue for positive and negative values, respectively) at 17°N for the 0600 16th (A), 0600 18th (B), 0600 19th (C), and 0600 20th (D).
Figure 6.4. A) Location of mountain gap parcels inside the mountain gap (heavy dots) and topography (every 50 hPa starting on 950 hPa). B) Parcel locations (dots) within the mountain gap at 17°N in eta coordinates.

Figure 6.5. Location of individual parcels between the surface and 950 hPa and their $\theta_e$ (shading). For the 0600 16th (A), 0600 18th (B), 0600 19th (C) and 0600 20th (D). Mean Isochrones (in hours relative to the time after departing the mountain gap, black contours). Closed streamlines of the relative wind at 950 hPa (Magenta ovals). Note the different axis scale of panel D.
Figure 6.6. A) Cyclogenesis region. Disk of 200 km radius centered on the cyclogenesis location for Patricia. B) Timeseries of the median vorticity value of the mountain-gap parcels within the cyclogenesis region and between the surface and 950 hPa. Median vorticity value and IQR within the cyclogenesis region between the surface and 950 hPa. Panel C) as B) but for the $\Theta_e$. Time in horizontal axes is in days of October.
Figure 6.7. A) Timeseries of the median (black line) and IQR (gray envelope) vorticity of the mountain-gap parcels entraining the closed streamline at cyclogenesis time (Fig. 6.5 (d)). Horizontal axis in days of October. B) Vorticity of the mountain-gap parcels (shading). Stretching term of the vorticity equation along the trajectory of the parcels. Vertical lines in A correspond to the locations of the parcels in contours in panels B and C.
Figure 6.8. A) Timeseries of the median (black line) and IQR (gray envelope) vorticity of the mountain-gap parcels entraining the closed streamline at cyclogenesis time (Fig. 6.5 (d)). Horizontal axis in days of October. B) Vorticity of the mountain-gap parcels (shading). Stretching term of the vorticity equation along the trajectory of the parcels. Vertical lines in A correspond to the locations of the parcels in contours in panels B and C.
7. OPERATIONAL FORECASTS FOR THE GENESIS OF HURRICANE PATRICIA

Chapters 4 through 6 described the synoptic environment leading to Patricia's cyclogenesis, the structure of the developing trough, and the potential interactions of the trough with the mountain gap wind. The evolution of the developing trough and the vortex leading to cyclogenesis within it were complex because it involved numerous factors occurring at different spatial scales. This complexity may have contributed to poor cyclogenesis forecasts during the earliest forecasts predicting cyclogenesis. For example, the first GFS forecast to predict a TC, developed a tropical storm 30 hours before Patricia was declared a tropical storm. In contrast the first ECMWF cyclogenesis forecast only developed a tropical storm twelve hours before Patricia was declared a tropical storm, but it dissipated soon after.

This chapter aims to explore why the first GFS and ECMWF forecasts predicting cyclogenesis take diverging pathways. Focus is given to identify the vorticity anomalies that were associated with cyclogenesis. This will help us to understand what factors may have impacted Patricia cyclogenesis. The forecasts studied here are the GFS initialized on the 1800 17th, and the ECMWF initialized 6 hours later, on the 0000 18th. These are the first initialization times that forecasted cyclogenesis for each respective model. The forecasted TC of these forecasts are discussed and compared with ERA-5.

The GFS forecast data is obtained from the NCEI (National Centers for Environmental Information) with initialization times every 6 hours, 0000 0600 1200 and 1800 UTC. The ECMWF forecast data is obtained from the TIGGE (The International Grand Global Ensemble; Bougeault et al, 2010) database with available initialization times every 12 hours, 0000 and 1200 UTC. Both retrieved datasets have a horizontal resolution of 0.5°×0.5°. The GFS model has 26 vertical levels
up to 70 hPa, and the ECMWF stored in TIGGE has eight vertical levels up to 200 hPa, but the operational model has 137 vertical levels.

This chapter consists of 3 sections; section 7.1 briefly describes the track and intensity of the first GFS and ECMWF operational forecasts for the cyclogenesis of Patricia. Section 7.2 discusses the synoptic conditions leading to cyclogenesis as seen by the GFS and ECMWF operational forecasts. Section 7.3 provides a summary of the chapter.

7.1 Operational Forecasts of Patricia Cyclogenesis

Several operational models initialized between 1800 17\textsuperscript{th} and 0000 18\textsuperscript{th} were the first to predict a TC. Example of these models are the GFS, BAMD and SHIP initialized on the 1800 17\textsuperscript{th}, and the HWRF GFDL, SHIP, COAMP, and ECMWF initialized on 0000 18\textsuperscript{th}. The predicted track and intensity of these forecasts are shown in Fig. 7.1. Except for the ECMWF, the forecasted tracks and intensities are obtained from the ATCF (Automated Tropical Cyclone Forecasting). In the case of the ECMWF, the forecasted track and intensity are given by the track of the maximum vorticity at 950 hPa, and by the maximum 10-m wind speed within 150 km from the maximum vorticity.

The mentioned forecasts predict genesis close to the Best-Track genesis location (Fig. 7.1 (a)), but cyclogenesis is premature in these forecasts except for the ECMWF. In all of them, except for the ECMWF, cyclogenesis occurs about two days before the Best-Track genesis, on 1800 17\textsuperscript{th} (GFS, BAMD, and SHIP) on or 0000 18\textsuperscript{th} (HWRF, GFDL, and COAMPS). Regarding the ECMWF, it forecasts cyclogenesis until 1200 20\textsuperscript{th}, similar to the Best-Track. Additionally, the predicted tracks have a right-of-track bias when compared to the Best-Track, being the ECMWF the one with the largest bias. These observations are indicative the both the synoptic flow and the cyclogenesis process in the forecasts are very different from in the ERA-5.
Regarding the intensity forecasts, these models, except for the ECMWF predict a TC with tropical storm winds on 0000 19\textsuperscript{th} (SHIP) or 1800 19\textsuperscript{th} (GFDL, COAMP, HWRF). However, Patricia developed tropical storm winds until 0000 21\textsuperscript{th} (Fig. 7.1 (b)). This means that these models predict a premature intensification. In contrast, the ECMWF predicted a weak TC. Also, neither model predicted the peak intensity of the Best-Track. Overall, the poor intensity forecasts for Patricia have been documented and it has been concluded that it is mostly due to the coarse resolution of the operative models (Qin and Zhang, 2018; and Fox and Judt, 2018).

In general terms, these forecasts show two behaviors regarding the cyclogenesis of Patricia. The GFS, HWRF, GFDL, COAMP and SHIP forecasts predicted genesis prematurely. The ECMWF forecast predicted genesis at the correct time, but the predicted TC was very weak. Neither model adequately forecast the track nor the peak intensity of Patricia. Considering that part of the scope of this work is to understand the cyclogenesis of Patricia we will further analyze the GFS forecast initialized on 1800 17\textsuperscript{th} and the ECMWF initialized on 0000 18\textsuperscript{th} to reveal what factors led to poor cyclogenesis forecasts.

7.2 The Cyclogenesis of Patricia in the GFS and ECMWF Operational Forecasts

The GFS forecast initialized on 1800 17\textsuperscript{th}, hereafter GFS\textsubscript{1800-17}, and the ECMWF initialized on 0000 18\textsuperscript{th}, hereafter ECMWF\textsubscript{0000-18}, referred before show two different biases in the genesis forecasts for Patricia. This section analyzes the predicted synoptic patterns and are compared with the ERA-5 to elucidate what factors lead the forecasts to diverge from the ERA-5. The analysis is based on maps at 950 and 700 hPa of the wind and its associated vorticity for each forecast and for the ERA-5. We also present an analysis of the forecasted TC by the GFS and the ECMWF which are also compared with the ERA-5 TC.
7.2.1 The GFS Forecast for Cyclogenesis of Hurricane Patricia

The GFS\textsubscript{1800-17} forecast for the 950 hPa and 700 hPa wind and its vorticity are shown in the left columns of Figs. 7.2 and 7.3, respectively. The forecasts are valid for the 0600 18\textsuperscript{th}, 0600 19\textsuperscript{th}, 0600 20\textsuperscript{th} and 0600 21\textsuperscript{st} of October. Recall that the GFS\textsubscript{1800-17} forecast is initialized on 1800 17\textsuperscript{th}.

The 0600 18\textsuperscript{th} of October shows a vortex to the south of the gap outflow at (13\textdegree N, 96\textdegree W), that dominates the vorticity field over the EPac at 950 and 700 hPa (Fig. 7.2 (a) and Fig. 7.3 (a)); this vortex represents Patricia in the forecast. The vortex is embedded within the EW trough over the Yucatan peninsula at 700 hPa (Fig. 7.3 (a)), which is similar to that in the ERA-5 including the enhanced southerlies. Considering its coherent structure between 950 and 700 hPa and its vorticity intensity at 950 hPa, much stronger than that of the gap outflow it seems that it developed independently of the gap outflow vorticity.

During the next days, 0600 19\textsuperscript{th} and 0600 20\textsuperscript{th} the vortex keeps intensifying at both 950 and 700 hPa levels, being the 950 hPa southerlies the strongest wind surrounding the vortex (Figs. 7.2 (c,e) and 7.3 (c,e)). On the 0600 20\textsuperscript{th}, the GFS\textsubscript{1800-17} predicts a vortex with hurricane winds at 950 hPa but the ERA-5 only develops winds of 17 m/s (grey line, Fig. 7.1 (b)), these winds are related to the mountain gap (Fig. 5.4 (f)). This means that when the ERA-5 was forming Patricia on the 0600 20\textsuperscript{th}, the GFS\textsubscript{1800-17} already forecasted an intensifying TC. The cause of the premature cyclogenesis is attributed to the intensification of a vorticity anomaly on the ITCZ.

This vorticity anomaly originated in the ITCZ, about 1500 km from the Isthmus of Tehuantepec, and was advected by the trough southerlies (vorticity map at 0000 18\textsuperscript{th} not shown). As it was transported northward it intensified, such that by 0600 18\textsuperscript{th}, or 12 hours into the forecast its vorticity is easily recognized at both 950 and 700 hPa. Subsequent times show that this vortex
keep intensifying. Therefore, the GFS$_{1800-17}$ models the cycogenesis of Patricia as the intensification of a sole ITCZ vorticity anomaly likely independent of the gap outflow.

Regarding the vorticity anomaly leading to cyclogenesis in the GFS$_{1800-17}$, is surprising that its location (95°W,11°N; see Fig. 7.2 (a)) on 0600 18th is very close to the corresponding location (96°W,10°N; see Fig. 5.1) of the curvature vorticity anomaly discussed in chapter 5. This would be consistent with the discussion of the contributions of the ITCZ vorticity to the cyclogenesis of Patricia; but in the case of the GFS$_{1800-17}$ these anomalies are stronger and appear to be the main factors underlying cyclogenesis.

7.2.2 ECMWF Forecast for Hurricane Patricia

The ECMWF$_{0000-18}$ forecast for the 950 hPa and 700 hPa wind and its vorticity are shown in the right columns of Figs. 7.2 and 7.3, respectively. The forecasts are valid for the 0600 18th, 0600 19th, 0600 20th, of October. Recall that the ECMWF$_{0000-18}$ forecast is initialized on 0000 18th.

On 0600 18th, a broad cyclonic flow over the Yucatan peninsula and the EPac dominates the vorticity at 700 hPa (Fig. 7.2 (b)). At 950 hPa, another cyclonic flow underneath the trough to the south of the Isthmus of Tehuantepec and limited to the north by the topography is forecasted (Fig. 7.2 (a)). This cyclonic flow underneath the trough is limited between the gap outflow and the trough southerlies. Also, several vorticity anomalies to the north of the trough southerlies are also forecasted; these are similar to the ITCZ vorticity anomalies of the ERA-5. Overall, this configuration of the wind and its vorticity at 950 hPa is very similar to that on the ERA-5 when the trough vorticity merged with that of the gap outflow forming a large vortex; see chapter 4 (Fig. 4.5 (c,d)).

On the next days the vorticity at 950 hPa slowly intensifies and becomes more compact yet the wind remains roughly unchanged. The wind and its vorticity form a vortex close to the gap
outflow at 950 and 700 hPa located (96°W, 14°N) on 0600 19th (Fig. 7.2 (d) and 7.3 (d)) and by (97°W, 15°N) on 0600 20th (Fig. 7.2 (d) and 7.3 (d)). This vortex is the precursor for Patricia.

Despite the similar wind and vorticity patterns of the ECMWF0000-18 with the ERA-5 until the 20th, the ECMWF0000-18 diverges from the ERA-5 on the 21st (Fig. 7.4). The ECMWF0000-18 predicts weaker 950 hPa winds than the ERA-5, including the gap wind. Additionally, the vorticity to the south of the Isthmus in the ECMWF0000-18 is weaker and more spread than in the ERA-5. This indicates that the ECMWF0000-18 fails to develop a compact low-level vortex, leading to a weak system. We speculate that the ECMWF0000-18 fails to develop the low-level vortex because the weaker gap outflow produces less vorticity than the ERA-5 and it is the extra vorticity of the gap outflow that helps to consolidate the vortex. However, a more detailed explanation of why this specific forecast fails to develop the vortex is out of the scope of this research.

In general terms, the vorticity patterns of the ECMWF0000-18 are similar to those of the ERA-5 (Fig. 4.4 (e,f)) until the 20th, and very different from those of the GFS1800-17. That is, the ECMWF0000-18 forecasts a broad cyclonic flow between the gap outflow northerlies and the trough southerlies. However, the vorticity maximum is not as prominent nor as compact as in the ERA-5 resulting in a weak TC. We speculate that the predicted weak TC is partially related by the weaker ECMWF0000-18 winds, specifically the gap outflow, resulting in a weaker vorticity. In turn the weaker vorticity makes the vortex to develop slowly than the ERA-5 vortex.

7.2.3 Discussion

Based on the GFS1800-17 forecast, it can be said that the intensity of the low-level ITCZ vorticity transported by the trough southerlies is important for cyclogenesis. If the ITCZ develops strong vorticity anomalies, these may develop on their own, regardless of the gap outflow vorticity, triggering cyclogenesis. Therefore, the ITCZ vorticity anomalies need to be adequately modeled
in cyclogenesis forecasts. In turn, considering their horizontal scale, less than 50 km (the operative GFS forecast has a resolution of ~25km), their development is strongly dependent on the modeling of the convection, planetary boundary layer, and other small-scale processes modeled in the forecasts. Ultimately, the cyclogenesis forecast is dependent on the representation of these small-scale processes.

The ECMWF\textsubscript{0000-18} forecast suggests that if the gap outflow is weaker, and the ITCZ vorticity is “weak.” Then, the formation of the low-level vortex leading to cyclogenesis would take more time and would be weaker, suggesting that the gap outflow and its vorticity has a strong impact in cyclogenesis. We recognize the speculative aspect of this statement requiring further investigation, including what could be a threshold intensity for the ITCZ vorticity to trigger cyclogenesis on its own. However, in the next chapters we will study the role of the gap wind intensity in cyclogenesis, to elucidate the validity of this statement.

7.3 Summary

The first operational GFS and ECMWF forecasts for the cyclogenesis of Patricia were analyzed. The GFS predicted cyclogenesis too early, while the ECMWF predicted cyclogenesis in the correct time and location albeit the forecast TC was too weak. The chapter explores what caused these biases in the forecasts.

The GFS forecast initialized on the 1800 17\textsuperscript{th} (GFS\textsubscript{1800-17}) was the earliest GFS forecast to predict a hurricane. It predicted a tropical storm (TS) about 30 hours earlier than the Best-track, and genesis occurred 200 km from the genesis location in the Best-Track. Genesis in the GFS\textsubscript{1800-17} was attributed to a low-level vorticity anomaly that originated by the ITCZ transported by the trough southerlies. This anomaly developed fast, and likely independent of the gap outflow
vorticity because its vorticity intensity was much larger than the gap outflow vorticity; also because it formed far from the gap outflow.

The ECMWF forecast initialized on 0000 18\textsuperscript{th} (ECMWF\textsubscript{0000-18}) was the first ECMWF forecast for cyclogenesis. It predicted genesis close to the Best-track location and developed tropical storm winds about 12 hours later than the Best-track, however, the predicted TC was very weak. Cyclogenesis was attributed to a broad vortex similar to the ERA-5, comprised between the gap outflow vorticity and the trough southerlies underneath the developing trough. However, the vortex developed at a slower pace than in the ERA-5 resulting in the weaker TC. One potential cause for the slow development of the vortex could be a weak gap wind with weak vorticity. Ultimately this translates in weaker low-level vorticity underneath the trough leading to a weaker TC vortex.

In summary, by analyzing GFS\textsubscript{1800-17} and ECMWF\textsubscript{1800-17} forecasts it was revealed that both the ITCZ vorticity and the gap outflow vorticity are factors that may have a significant impact on cyclogenesis. Thus, it is expected that a correct cyclogenesis forecast would model properly these two origins of vorticity in space (the location of the ITCZ vorticity anomalies), time, and intensity. Not to mention the EW trough itself and its vorticity. The next chapter will explore the impacts of the gap outflow on cyclogenesis by suppressing it. The potential role of the ITCZ it is proposed for future studies.
Figure 7.1. A) Track forecasts of GFS (green), SHIP (magenta) and BAMD (blue) operational models initialized on the 1800 17th, and of GFDL (brown), COAMP (navy blue), and ECMWF (cyan) forecast models initialized on the 0000 18th; The Best Track (black) and ERA-5 (gray) TC tracks are also shown. B) As in A) but for the intensity forecast. The horizontal axis in B) in days of since 1800 17th of October. Heavy dots are located every day starting on 1800 17th.
Figure 7.2. GFS1800-17 (left column) and ECMWF0000-18 (right column) wind (barbs) and vorticity (shading) at 950 hPa forecasts valid for the 0600 18th (top row), 0600 19th (middle row), 0600 20th (bottom row).
Figure 7.3. GFS$_{1800-17}$ (A, C, E) and ECMWF$_{0000-18}$ (B, D, F) wind (barbs) and vorticity (shading) at 700 hPa forecasts valid for the 0600 $18^{\text{th}}$ (A, B), 0600 $19^{\text{th}}$ (C, D), 0600 $20^{\text{th}}$ (E, F).
Figure 7.4. Wind and vorticity for the ECMWF 0000-18 at 0600 21st.
8. A NUMERICAL SIMULATION OF PATRICIA AND THE GAP FLOW

The key elements for understanding the cyclogenesis of hurricane Patricia were presented in chapters 4, 5 and 6. There it was found that three vorticity anomalies likely contributed to form the vortex leading to cyclogenesis. These vorticity anomalies are the vorticity associated with the developing EW trough (Fig. 8.1), the ITCZ vorticity anomalies and advected northward by the trough southerlies, and the positive shear vorticity associated with the gap wind.

The developing trough is necessary for cyclogenesis as it provides a favorable synoptic environment for the formation of a TC vortex. Within it, air may recirculate isolated from the environment developing sustained convection, and coherent vortices may form, the so-called pouch theory (Dunkerton and Montgomery, 2009). Additionally, ITCZ vorticity anomalies were transported northward by the EW trough southerlies. This indicates that ITCZ vorticity anomalies contribute to cyclogenesis.

Regarding the Isthmus of Tehuantepec gap wind. It was triggered by a passing midlatitude system and was materially connected with the trough. Through this material connection vorticity was transported from the gap outflow into the cyclogenesis region, likely contributing to favor cyclogenesis. However, it is unclear if this contribution of gap-flow vorticity actually affected the favorability for cyclogenesis, and to what degree. Nevertheless, a study on the formation of Guillermo (1993; Farfán and Zehnder, 1997) found in numerical simulations that the contributions of the gap wind flow are needed for cyclogenesis.

Therefore, to solely evaluate the role of the gap wind on cyclogenesis, a numerical experiment is done. The experiment consists of two WRF runs; the first one, the Control Run described in this chapter, models the cyclogenesis of Patricia including the gap flow. The second, the “Blocked Run,” is similar to the control run but with the gap wind subtracted by filling the
mountain gap of the Isthmus of Tehuantepec. The blocked run will be discussed in the next chapter. The purpose of this chapter is only to evaluate the control run. The next chapter studies the effects of the gap wind by comparing the outputs of both runs.

This chapter is organized as follows. Section 8.1 describes the experimental setup. Section 8.2 evaluates the vorticity sources related with cyclogenesis on Control Run. Section 8.3 describes the simulated TC and Section 8.5 summarizes the control run.

8.1 Experimental Setup

The control run uses a triple nested-grid version of the ARW-WRF (version 4.0; Skamarock et al. 2008) with a finest resolution of ¾ km for the inner domain. Figure 8.2 shows the triple nested domains with horizontal grid resolutions of 6, 2 and ¾ km, and 452 × 299, 907 × 490, and 1401 × 714 grid points, respectively. The exterior domain, D₁ is centered at (94°W,16.4°N) and uses a Mercator projection. All the domains use the same 55 vertical levels used by Chen et al. (2011) for the simulation of hurricane Wilma (2005), with the top pressure level at 30 hPa. Domains D₁ and D₂ are initialized on 0000 13th October. The innermost domain D₃ is initialized at 1800 on the 18th. Domain D₃ captures the formation of Patricia, which is indicated by its minimum surface pressure and its vortex, as well as its subsequent northwestward motion.

The reason to opt for a long integration time for the experiment, initiated seven days before cyclogenesis is to completely model the ramp up of the gap northerlies since the 0000 13th and its interaction with the developing EW trough. Recall that the developing EW trough results from the merging of an EW formed in the EPac with an incoming Caribbean EW (chapter 4).
The initial conditions and boundary conditions are provided by the ERA-5 reanalysis with a horizontal resolution of $0.25^\circ \times 0.25^\circ$ every 3 hours. The SST is prescribed by the ERA-5. All the inner domains derive lateral boundary conditions from their corresponding parent domains.

The model physics, except for the topographic effects are chosen as in Chen et al. (2011) for the simulation of hurricane Wilma and are as follows: i) the Thompson scheme for microphysics (Thompson et al; 2008); ii) the Yonsei University PBL scheme (Hong et al, 2006); iii) the Rapid Radiative Transfer Model (RRTM) for longwave and shortwave radiation (Mlawer et al, 1997); iv) the topographic drag adjustment to the surface winds (Jimenez and Dudhia, 2012); v) the shadow effects of the topography on the shortwave radiation (Arthur, 2018); vi) the drag coefficient for the horizontal momentum fluxes over warm water of Donelan et al. (2004); vii) the exchange coefficient for temperature and moisture over warm water of Garrat (1992) and, viii) the unified Noah land-surface model (Tewari et al, 2004). Also, the model is run with no cumulus parameterization in any domain.

8.2 Retrieval from the Control Run of the ERA 5 Atmospheric Features and Best-Track Characteristics

The Control Run is evaluated against the ERA-5 and against the Best-Track by comparing the key features related to cyclogenesis found in the ERA-5 and in the Best-Track. These features are, the developing trough, the transport of vorticity from the ITCZ to the cyclogenesis region, the gap wind, the TC track and its intensity. It must be kept in mind that it is not intended for the control run to be a higher resolution exact copy of the ERA-5 nor a better forecast of the early GFS or ECMWF forecasts. On the contrary, the control run is the foundation for an experiment designed to assess the impacts of the gap wind on cyclogenesis and, since cyclogenesis takes place about 7 days after the initial time some differences between the Control Run and the ERA-5 and the Best-
Track are expected. So, the control run is considered to be useful if it reproduces the gap wind and forms the TC in approximately the correct time and place, including the interaction between the gap outflow and the developing trough.

### 8.2.1 The Gap Wind and Hurricane Patricia, First Insights.

The gap wind in the Control Run is compared with the gap wind in the ERA-5. The gap wind is computed as the averaged meridional wind within the interval $[95.25^\circ W, 94.75^\circ W] \times [16.75^\circ N, 17.25^\circ N] \times [1000 \, hPa, 800 \, hPa]$, this is the same region used to define the mountain gap wind index in chapter 3. The gap wind of the control run (red line) and of the ERA-5 (black line) is shown in Fig. 8.3 (a). Clearly, the Control Run successfully reproduces the ERA-5 gap wind. However, the minimum on the 20th and the peak on the 22nd are slightly larger than those in the ERA-5 indicating that the extreme gap wind in the Control Run are stronger than in the ERA-5. This is possibly due to the higher resolution of the simulation, but in general they are similar.

Figure 8.3 (b) shows the track of the Best-Track and of the simulated TC. Its track is given the MSLP minimum over the EPac. The simulated TC is first detected at 1200 19th (Fig. 8.3 (c)) and to the remains south of the mountain gap for about 12 hours. By 0600 20th, the simulated TC moves westward following closely the Patricia’s track until 0600 21st. Afterward, the tracks diverge, and the simulated TC deviates to the northwest of the Best-track, similar to the right-of-track biased ECMWF and GFS forecasts.

Figure 8.3 (c) shows the maximum 10-m windspeed within 100 km from the location of the simulated TC, its intensity, and the intensity of the Best-Track. The intensities are very similar between the 0600 20th and the 22nd. Afterward the simulated TC stops intensifying as it approaches land. Thus, the control run, except for the right-of-track bias, models Patricia and its genesis.
location and time satisfactorily. More details on the hurricane-like structure of the simulated TC and of the are given in section 8.3.

An explanation for the right-of-track bias of the simulated TC is suggested by the diagnosed differences between the steering flow for the TC in the Control Run and that in ERA5. The steering flow is usually defined as the residual wind after subtracting the wind vortex (Holland, 1983) and has been estimated by the 500 hPa wind (Brand et al, 1981). Here we estimate the mean steering flow as the 500 hPa wind averaged between the 0600 20th and the 2100 22nd and is shown in Fig. 8.4 for the ERA-5 (Fig. 8.4 (a)) and for the Control Run (Fig. 8.4 (b)). The figure reveals that in both cases, the Best-track and the simulated TC moved around the periphery of an anticyclonic steering flow over the EPac. The ERA-5 wind is a stronger (about 4 m/s stronger) and has a stronger zonal component than the Control Run. This causes the right of track bias of the simulated TC track compared with the Best-Track.

Overall, the Control Run models the cyclogenesis of Patricia in a place and time close to the Best-Track. However, the resulting simulated TC has a right-of-track bias attributed to a weaker but northward steering flow at 500 hPa associated with a weaker anticyclonic flow associated with the midlatitude system over the EPac.

As a side topic, is important to properly forecast the steering flow in which the TC is embedded. In the case of hurricane Patricia and generally of recurving TCs, the anticyclonic circulation associated with midlatitude systems gives the steering flow. However, the predictability of these midlatitude circulations, which may cause the right-of-track bias is out of the scope of this work.
8.2.2 The Structure of the Gap Wind

A comparison of the gap wind between the Control Run and the ERA-5 is presented next. It considers the study of vertical cross section at 17°N of the meridional wind and the vorticity between the 0600 18th and the 0600 20th for both the ERA-5 and the Control Run.

The vertical cross sections of the meridional wind between the 18th and the 20th for the ERA-5 and the Control Run show a northerly flow in the mountain gap from the surface up to 500 hPa (Fig. 8.5). For both the ERA-5 and the Control Run, the northerlies are stronger along the west flank of the mountain gap and have similar intensity. On the eastern flank, however, the northerlies close to the surface are stronger in the Control Run than in the ERA-5 by about 5 m s⁻¹. This is most notable on the 0600 20th (Fig. 8.5 (c,d)).

Regarding the vorticity associated with the gap wind; both the ERA-5 and the Control Run develops two vorticity patches in the mountain gap (95°W) with different signs (Fig. 8.6 (a,c)). The negative patch locates over the western flank of the mountain gap (west of 95°W), and the positive patch over its eastern flank (east of 95°W). The positive patch of the ERA-5 is smaller and weaker. In contrast the Control Run develops a stronger and wider positive vorticity patch that dominates the eastern flank of the mountain gap. This suggests that the Control Run may transport higher vorticity toward the cyclogenesis region than the ERA-5, which in any case favors cyclogenesis.

The vorticity of the Control Run is compared with that of the ERA-5 by averaging to the same resolution as the ERA-5 (25 km). The resulting averaged vorticity within the mountain gap in the Control Run is higher than the vorticity in the ERA-5. This means that the Control Run predicts a gap wind with stronger vorticity than that in the ERA-5, potentially having a larger
impact on cyclogenesis. This suggests that the predictability of cyclogenesis as a function of the horizontal resolution should be further explored.

8.2.3 The Developing Trough, and the ITCZ Vorticity Strip

The developing trough, the ITCZ vorticity the advected by the trough southerlies, and the gap outflow are modeled by the Control Run and are similar to their representations in the ERA-5. These three features are modeled by the Control Run and are shown in Fig. 8.7 by the wind and its vorticity at 950 and 700 hPa for the 0600 18\textsuperscript{th}, and the 0600 20\textsuperscript{th}.

The ITCZ vorticity anomalies transported northward by the trough southerlies is clearly seen at 950 hPa before and on cyclogenesis in the Control Run (Fig. 8.7 (a,c)). On the 18\textsuperscript{th} they start to merge with the vorticity patch associated with the gap outflow (Fig. 8.7 (a)). By the 20\textsuperscript{th} (Fig. 8.7 (c)) the ITCZ vorticity anomalies have merged with the gap outflow vorticity forming a broad patch characterized by several small-scale maxima. This merging of the vorticity in the EPac prior cyclogenesis in the Control Run is similar to that seen in the ERA-5 (see Fig. 8.8 (a,c) and thoroughly discussed in chapter 4). However, the control run’s higher resolution resolves processes and features not resolved by the ERA-5 (2 km vs 25 km). This is seen by comparing the complex vorticity pattern in the EPac of the Control Run with the “simpler” vorticity field in the ERA-5 (Fig. 8.7 (c) vs Fig. 8.8 (c)). Thus, at large scales, the EPac vorticity in the Control Run and in the ERA-5 are similar, but at small scales (less than 10 km) some differences arise, mainly the several vorticity maxima present in the Control Run. These differences are mostly due to how the smaller-scale processes are modeled in the Control Run (microphysics, Planetary Boundary Layer, heat and momentum fluxes), which are not equally resolved by the ERA-5.

At 700 hPa, the developing EW trough in the Control Run is indicated by a broad circulation extending from the Yucatan peninsula to the EPac. The EW trough is similar to the
trough in the ERA-5 (compare Fig. 8.7 (b) with Fig. 8.8 (b)). The evolution of the trough, not shown for earlier times is similar to that of the ERA-5 where an EPac trough merged with an incoming Caribbean EW leading to the elongated trough straddling the Yucatan peninsula (see sections 4.3 and 4.4). The trough circulation weakens over Yucatan and a circulation develops over the EPac between the 18th and the 20th (Fig. 8.7 (b,d)); this agrees with the ERA-5 (Fig. 8.8 (b,d)). Additionally, the shape of the developing EW trough is similar both in the control run and in the ERA-5. However, in the Control Run, the circulation in the EPac is about 3 ms\(^{-1}\) stronger than in the ERA-5.

Therefore, the control run reproduces the atmospheric elements involved in the cyclogenesis of Patricia features, the gap wind, the developing EW trough and the ITCZ vorticity anomalies. Additionally, the vorticity patterns at 950 and 700 hPa on 0600 20th between the ERA-5 and the Control Run are similar in intensity and shape, indicating that the merging of the gap outflow vorticity and the trough vorticity also occurs in the Control Run (Figs. 8.7 (c,d) and 8.8 (c,d)). The only notable difference in the times leading to cyclogenesis is the complex vorticity structure within the vorticity strip, however this difference is mainly due to the coarser resolution of the ERA-5, so are expected. The next section provides evidence of the hurricane-like structure of the simulated TC.

8.3 Hurricane-like Characteristics of the simulated TC

The hurricane-like structure of the simulated TC is evidenced by the wind and its vorticity at 950 and 700 hPa (Fig. 8.9), and the vertical cross sections of the azimuthally averaged equivalent potential temperature, vorticity, and tangential and radial winds (Fig. 8.10). These fields are shown for the 0600 22nd of October corresponding to the time when the simulated TC achieved maximum
intensity. Additionally, the Control Run reveals some high-resolution characteristics not modeled by the ERA-5 including the eye.

At 0600 22\textsuperscript{nd}, the simulated TC forms a coherent vortex located at (99.9\textdegree W, 15.8\textdegree N) whose diameter is about 50 km at 950 (Fig. 8.9 (a)) and 700 hPa (Fig. 8.9 (b)). Within the vortex, the vorticity maximizes along an arc on the eastern sector and inside the radius of maximum winds. This arc is aligned with the radius of maximum winds at both pressure levels. Subsequent time frames (not shown) indicate that the arc tries to close in upon itself, suggesting that the Control Run attempts to develop an axisymmetric eyewall yet never becomes axisymmetric in the integration time. Observational studies have detected arcs of maximum vorticity within the radius of maximum winds in intensifying storms (Martinez et al; 2017). Thus, the vortex associated with the simulated TC is similar to that of an intensifying TC.

The TC vortex for Patricia is accompanied by small-scale vorticity anomalies distributed along a spiral band to its south at 950 and 700 hPa. The spiral band is parallel to the southerlies. These vorticity anomalies are associated with convective cells in the control run, since they are associated with precipitation (precipitation not shown). However, the ERA-5 does not resolve individual convective cells but forms a long strip of positive vorticity similar to this spiral band (Fig. 7.2 (a,c,e)). Nevertheless, both the ERA-5 and the Control Run develop this band parallel to the southerlies indicating that the control run reproduces reasonably the vorticity surrounding the TC vortex.

Outside the radius of maximum winds, the flow is asymmetric, its weaker in the western sector and stronger in the eastern sector, being the southerlies the strongest winds. This is consistent with the asymmetries of the ERA-5 and of the Best-Track winds. The radius of maximum winds of the TC vortex (25km) is similar to the radius of maximum winds of the
Extended Best-Track by 22\textsuperscript{nd} of October (Demuth, DeMaria and Knaff, 2016). The Extended Best-Track dataset complements the Best-Track with additional information about the radius of maximus winds, the diameter of the storm eye, and the radius and pressure of the outer closed isobar. Thus then, the Control Run models a developing vortex with a vorticity maximum within the radius (25 km) of maximum winds at its time of maximum intensity. Both the radius of maximum winds and the maximum intensity on 0600 22\textsuperscript{nd} are similar to the Best-Track and the extended Best-Track. This vortex has characteristics of actual TCs.

\textbf{8.3.1 Vertical Structure of the Simulated TC}

The vertical cross sections of the azimuthally averaged $\theta_e$, vorticity, tangential winds and radial winds (Fig. 8.10) provide an axisymmetric perspective of the TC vortex. However, it is recalled that the simulated TC is not completely axisymmetric. Figure 8.7 reveals that the TC vortex of the Control Run has a similar shape to that of ERA-5 (Fig. 7.6 (c)), albeit its horizontal size is smaller. This is a consequence of the higher resolution of the Control Run; this allows to resolves the complex structures within the TC radius of maximum winds (less than 25 km from the TC center, which is the resolution of the ERA-5).

Additionally, the following can be said about the vertical cross sections:

1) The TC developed a warm core, greater than 350K, as evidenced by the highest values of $\theta_e$ close to the TC center.

2) The mean vorticity forms a column up to about 350 hPa and forms a ring aloft up to 100 hPa. These rings have been observed in PV structures of hurricanes (Hendricks, et al; 2009 and Yau, 2004). The vorticity maximum locates close to the center of the TC vortex, between the surface and 700 hPa inside the radius of maximum winds.
3) The mean tangential wind maximizes at about 25 km from the TC center (dashed lines), closer to the TC center, the vorticity increases substantially. The maximum tangential winds are between 1000 and 600 hPa in agreement with vertical wind profiles in tropical cyclones. Being the most intense winds close to the surface.

4) The mean radial wind forms an inflow up to 50 km form the storm center; aloft, at 200 hPa. An upper-level outflow formed on top of the inflow characteristic of TC is seen. Therefore, the Control Run simulates a realistic TC with a structure similar to that of the ERA-5.

8.4 Summary

The control run captures all the atmospheric and key vorticity features involved in the cyclogenesis of hurricane Patricia; the developing EW trough, the ITCZ vorticity anomalies advected by the southerlies, and the mountain gap wind. The control run developed a developing vortex characterized by a surface low-pressure. This developing vortex leads to Patricia and has hurricane-like characteristics: The developing vortex formed close to the place and the time of Patricia’s genesis. It developed maximum 10-m winds similar to the intensity of Patricia until the 0000 22\textsuperscript{nd}. However, its track drifted to the right of Patricia’s Best-Track. This right-of-track bias was attributed to the weaker than the ERA-5 flow at 500 hPa; nevertheless, it occurs 7 days into the simulation and some biases between the Control Run and the ERA-5 should be expected.

Therefore, the control run is considered useful enough to be a basis to explore the sensitivity of the formation of Patricia to the Tehuantepec gap wind. The next chapter explores a scenario where the gap wind is suppressed by blocking the mountain gap of the Isthmus of Tehuantepec. This scenario allows to determine the importance of the Tehuantepec gap wind during cyclogenesis.
Figures

Figure 8.1. Schematic of the vorticity associated with the developing EW trough (blue hatching) at 700 hPa (A). Schematic of the gap outflow positive vorticity (black hatching) and the ITCZ vorticity anomalies (red hatching) at 950 hPa (B). Topography (Shading).
Figure 8.2. Model configuration for the control run. Nested domains D₁ (heavy solid line), D₂ and D₃ have 6, 2 and ⅔ km horizontal grid lengths respectively.

Figure 8.3. A) Tehuantepec gap wind based on the ERA-5 (black line) and control run (red line). B) Best-Track (black) and control run (red) TC track. C) Intensity of the Best-Track (black) and of the control run (red) TC. Heavy dots, black (for Best-Track) and red (for control run) placed on the 0600 20th, 0600 21st and 0600 22nd of October. Gray shading in C) indicates land. Horizontal axis in A) and B) are in days of the simulation initial time (upper row) and the date (days of October, lower row).
Figure 8.4. Wind at 500 hPa averaged between the 0000 20th and 2100 22nd of October for the ERA-5 and the Control Run in panels A and B, respectively. Also shown the Best-Track and the simulated track of the control run for those days.
Figure 8.5. Vertical cross sections at 17°N of the meridional wind (color shading contours every ±5 m/s) on 0600 18th (top row), 0600 19th (middle row), and 0600 20th October (bottom row) for the ERA-5 (left column) and the control run (right column). Vertical coordinate in hPa. Black shading indicates the topography.
Figure 8.6. Vertical cross section of the wind (barbs) and its vorticity (shading) at 17°N on 0600 18th and 0600 20th October for the ERA-5 (A,C) and the control run (B,D). Vertical coordinate in hPa.
Figure 8.7. Control Run wind (barbs) and its vorticity (color shading) at 950 and 700 hPa in the left and right column, respectively, for the 0600 18th and 20th of October in the upper and lower rows, respectively. Upper right box in each panel indicates the simulation time in days and hours and in the second line the date.
Figure 8.8. ERA-5 wind (barbs) and its vorticity (color shading) at 950 (left column) and 700 hPa (right column) for the 0600 18th (upper row) and 20th of October (lower row).

Figure 8.9. Wind (barbs) and its vorticity (shading) at 950 and 700 hPa for the Control Run on the 0600 22nd.
Figure 8.10. Vertical cross section of the radially averaged $\theta_e$ (color shading), tangential wind every 5 m s$^{-1}$ (black line), radial wind every -3 m s$^{-1}$ (dashed blue line) and 3 m s$^{-1}$ (dashed blue line), and vorticity in $10^{-5}$ s$^{-1}$ units and indicated in the isopleths (green line) for Control Run on the 0600 22nd. Also shown the radius of maximum winds as a function of the pressure.
9. THE SENSITIVITY OF PATRICIA’S CYCLOGENESIS TO THE GAP WIND OF THE ISTMUS OF TEHUANTEPEC

The Control Run simulates the evolution of a hurricane-like vortex similar to hurricane Patricia and is the basis to explore the sensitivity of the formation of Patricia to the Isthmus of Tehuantepec gap outflow. The sensitivity of the cyclogenesis process to the gap outflow is explored by comparing the Control Run with the Blocked Run. In the Blocked Run the mountain gap is filled creating a scenario with no gap wind.

Most of the discussion consists of presenting and understanding the differences between the Control Run and the Blocked Run weather patterns over the EPac, focusing on the interaction between the gap wind and the trough. The differences between the runs are attributed to the gap wind.

This chapter consists of five sections. Section 9.1 describes the blocked run setup; section 9.2 presents some overarching differences between the Blocked and Control Runs. Section 9.3 includes analyses of the evolution of the synoptic-scale weather patterns in the EPac for each run. Section 9.4 shows analyses of the sub-synoptic scale aspects of both runs focusing on the impacts of the gap wind on cyclogenesis. Finally, section 9.5 presents the summary and conclusions.

9.1 Model Setup for the Blocked Run

The blocked run has the same configuration as the control run except with the mountain gap of the Isthmus of Tehuantepec filled (Fig. 9.1 (a)). The mountain gap is filled with two superposed meridional paraboloids with a height of 0 km over the sea and a maximum total height of two km on top of the mountain gap (red line, Fig. 9.1 (b)). It extends zonally for about 600 km covering the mountain gap of the Isthmus of Tehuantepec such that the resulting topography has an average height typical of that of the surrounding mountains (grey shading, Fig. 9.1 (b)).
For this study we focus on the period between the 19th (stage IV, when the trough stalls over the Yucatan peninsula) and the 22nd (two days after cyclogenesis on the 20th). Remembering that the model is initiated on the 13th of October, the period emphasized here will thus be between days 5 and 7 of the forecast. During this period in the control run, cyclogenesis occurs, and the TC moves over the EPac. Also, this period covers the interaction between the gap wind and when the developing EW trough stalled over the Yucatan peninsula prior to cyclogenesis. The Blocked Run allows us to carry out an analysis of the evolution of the EW trough over the EPac in absence of the gap wind.

9.2 Overview of the Control and Blocked Runs

Through this work it has been argued that the gap wind, which is mainly meridional, played a favorable role in the cyclogenesis of hurricane Patricia. Hence, first we want to show the impacts on the meridional gap wind in Isthmus of Tehuantepec when the mountain gap is filled. Second, then we show what happens to the flows on the EPac when the mountain gap is filled. This is achieved by comparing the Control Run with the Blocked Run.

The impacts on the wind in the Isthmus of Tehuantepec when the mountain gap is filled are illustrated by the time series of mean meridional wind for the Control Run and for the Blocked Run. Recall that the Isthmus consists of the interval [96.5°W,92.5°W]×[16°WN,18°N]×[Surface,800hPa]; which is the same used in chapter 8 to diagnose the mountain gap wind in both the ERA-5 and the Control Run. Fig. 9.2 (a) shows the timeseries of the mean meridional wind in the Isthmus of Tehuantepec for the Control Run (red), the Blocked Run (blue) and the ERA-5 (black). As discussed in chapter 8 and recalled here, the gap wind in the Control Run is very similar to that of the ERA-5 and cyclogenesis occurs when
the northerlies are very strong on 0600 20th. Then the gap wind becomes southerly at cyclogenesis time. In contrast, in the Blocked Run the mean meridional wind in the Isthmus of Tehuantepec is weaker. Contrast the peak northerlies of the Control Run, -20m/s vs the peak northerlies in the Blocked Run, -5m/s. This shows that the Blocked Run is representative of a scenario where the gap wind and its outflow is almost suppressed.

Additional effects of filling the mountain gap are shown by the wind and its intensity at 950 hPa on 0600 20th of October (cyclogenesis time) for the Blocked and Control Runs (Fig. 9.3). The 950 hPa wind in the Control Run has a strong gap outflow and has a broad cyclonic flow over the EPac (Fig. 9.3 (a)). In contrast, the Blocked Run has very weak northerlies and a weak and elongated cyclonic flow forms over the EPac at (93°W,13°N; Fig. 9.3 (b)). In the Control Run this cyclonic flow develops in hurricane Patricia (chapter 8).

Figure 9.2 (b) shows the track of the minimum MSLP associated with each cyclonic flow shown in Fig. 9.3 for each run over the EPac. Fig. 9.2 (c) shows their respective intensities of each cyclonic flow, defined as the maximum 10-m wind speed within 100 km from their corresponding minimum MSLP location. Clearly the cyclonic flow in the Control Run (red) develops and intensifies similar to the Best-Track (black), but in the case of the Blocked Run the cyclonic flow fails to develop. Each cyclonic flow in each run forms a vortex, as will be discussed in section 9.3. In the case of the Control Run, the vortex is referred as the developing vortex, in chapter 8 is referred as the simulated TC. In the case of the Blocked Run, the vortex is referred as the non-developing vortex.

Even though the vortex in the Control Run develops and the vortex in the Blocked run does not, both nevertheless have similar tracks. They originate close to the Best-Track cyclogenesis location and follow the Best-Track moving roughly parallel to the coast. Their similar tracks
suggests that the synoptic-scale flow is similar in both runs. Is worth to note that both the developing and non-developing vortices have complex tracks prior the 20th. This is because in both runs, the minimum MSLP is establishing and their corresponding MSLP fields have fluctuations in space, hence the noisy tracks at the beginning.

Overall, these results suggest that the presence of the gap wind in the Control Run favors tropical cyclogenesis and its absence in the Blocked Run results in a lack of genesis. In the next section we analyze the synoptic-scale flow for each run to reveal the common aspects for each run and what are the largest differences between the runs.

9.3 Synoptic-Scale Aspects of the Control and Blocked Runs

Before considering the role of the gap wind on the cyclogenesis of Patricia, we briefly discuss the evolution of the synoptic-scale flow between the 19th and the 22nd for both runs. The synoptic scale flow of each run is represented by the wind and the geopotential height anomaly at 700 hPa and 950 hPa (Figs. 9.4 and 9.5). The geopotential height anomalies are with respect to their mean value between the 13th and the 22nd for each pressure level rounded to the closest decameter. The geopotential height is useful for highlighting both the midlatitude system and the EW trough in the simulations.

9.3.1 Synoptic-Scale Evolution of the Control Run

The evolution of the synoptic-scale weather patterns in the Control Run is as follows. On the 19th, a broad cyclonic flow accompanied by a negative geopotential heights anomaly dominates the domain (Fig. 9.4 (a,b)). This structure is the developing EW trough for Patricia (chapters 4 and 8). At 700 hPa, the EW trough is seen as a broad cyclonic circulation extending from the Yucatan peninsula to the EPac (Fig. 9.4 (b)). Its associated geopotential height anomaly finds its minimum in a sub-synoptic vortex over the EPac at (93.5°W,15°N). Underneath the trough, at 950 hPa, the
mountains over Central America block the EW trough’s cyclonic flow except on the Isthmus of Tehuantepec (Fig. 9.4 (a)). The topography allows the trough northerlies to flow over the EPac but blocks the southerlies from reaching the Yucatan peninsula. This leads to a cyclonic flow over the EPac separated from the wind over Yucatan.

It is worth noting that although the sub-synoptic vortex achieves tropical storm winds, it is a transient feature unrelated to the cyclogenesis of Patricia (Fig 9.3 (a,b)). This vortex originates as a vorticity anomaly in the ITCZ and is advected by the trough southerlies. As it moves northward it intensifies but dissipates once it made landfall in the following hours. A detailed description of its evolution is not provided. However, its presence suggests that ITCZ vortices contribute to enhance the EW trough over the EPac.

The 20th shows a general intensification of the EW trough at 950 and 700 hPa evidenced by the intensification of the negative geopotential height anomalies (Fig. 9.4 (c,d)). Additionally, the 950 hPa wind over the EPac strengthens with respect to the 19th and forms a broad cyclonic vortex underneath the EW trough (Fig. 9.5 (c)). This vortex is the developing vortex mentioned in section 9.2.

The subsequent days show the westward propagation of the EW trough parallel to the coast (Fig. 9.4 (e-h)). At 700 hPa, it is seen that the EW trough over the EPac remains but its part over the GoM or the Yucatan peninsula dissipates as the ridge of the midlatitude system displaces the EW trough (Fig. 9.4 (f,h)). Additionally, the developing vortex within the EW trough keeps intensifying, forming the vortex of a developing TC.

### 9.3.2 Synoptic-Scale Evolution of the Blocked Run

The Blocked Run shows a negative geopotential height anomaly over southern Mexico accompanied by a broad cyclonic flow spanning from the Yucatan peninsula to the EPac at 700
hPa at 0600 19th (Fig. 9.5 (b)). This pattern is similar to that of the developing EW trough in the Control Run, albeit the negative geopotential anomalies are weaker in the Blocked Run. At 950 hPa, the mountains block the geopotential height anomalies and the wind underneath the EW trough (Fig. 9.5 (a)). This yields the wind to form two disconnected cyclonic flows, each one associated with a negative geopotential height anomaly; one located over the EPac, and the other over Central America limited to the south by the mountains.

On the 20th and the 21st, the EW trough keeps developing, as evidenced by the intensification of the negative geopotential height anomalies mostly over the EPac at 700 hPa (Fig. 9.5 (d)). At 950 hPa, the negative geopotential height anomalies of the two disconnected cyclonic flows underneath the trough have moved westward and keep intensifying (Fig. 9.5 (c)). The flow over the EPac is broader and its associated geopotential height anomaly is more intense than the one over Central America. This is consistent with the fact that the ocean is a more favorable environment than land for developing cyclonic flows. On the 21st, the cyclonic circulation over the EPac forms a broad vortex with center at (95.5°W,13°N) characterized by the minimum geopotential anomalies (Fig. 9.5 (e)). Regarding the cyclonic flow formed over Central America, it has yielded to a compact vortex over the GoM at (93.5°W,19°N).

On the 22nd, at 700 hPa, the geopotential height anomalies become positive over the GoM (Fig. 9.5 (h)). Over the EPac, the EW trough moves westward and remains roughly of the same intensity (Fig. 9.5 (h)). The positive vorticity anomaly over the GoM is due to the passage of a midlatitude ridge that has displaced most of the EW trough there, limiting the EW trough to the EPac. At 950 hPa, the vortex over the EPac underneath the trough weakens and becomes more compact, but the vortex over the GoM keeps developing (red circles in Fig. 9.5 (g)); however, it dissipates in the subsequent hours. The shrinking of the vortex over the EPac suggests that the
Blocked Run attempts to develop this vortex but fails. Thus then, in the Blocked Run as in the Control Run, the EW trough still harbors a sub-synoptic vortex, but it fails to develop. Surprisingly the blocked run forms a strong and shallow vortex form over the GoM. The evolution of this vortex is out of the scope of this work yet is an interesting feature arising when the mountain gap is filled.

**9.3.3 General Considerations**

In general terms, the synoptic-scale patterns are similar for both the Control and Blocked Runs. That is, an EW trough spanning from the Yucatan peninsula to the EPac on the 19th moves westward and retraces to the EPac because of a midlatitude ridge over the GoM displaces it. This weather system has been referred as the developing EW trough in chapters 4 and 5 and results from a Caribbean EW trough with an EPac EW trough that approach each other (see figure 4.6 (c,d)). In both runs, the EW trough forms a sub-synoptic vortex over the EPac that experiences some development between the 19th and the 21st. Afterward the evolutions of the sub-synoptic vortices diverge. In the Control Run this vortex develops (and has TC-like properties, see chapter 8), but in the Blocked Run the vortex fails to develop. The developing vortex locates at (100°W,16°N) in the Control Run and the non-developing vortex at (101°W,15°N) in the Blocked Run for the 0600 22nd. Each vortex is indicated by a red circle over the EPac in Figs. 9.4 (g) and 9.5 (g), for the Control Run and the Blocked Run respectively.

Despite both runs model the EW trough, there are perceivable difference in its structure before the 21st. The negative geopotential height anomaly associated with the EW trough over the EPac in the Control Run is more intense than in the Blocked Run (lower geopotential height anomalies). This suggests that the gap wind is favorable to the development of the vortex within the EW trough over the EPac. The gap wind could provide additional vorticity during the early stages of the vortex that help to develop the vortex.
The next section discusses in detail the evolution of the vortex of each run to reveal the mechanism to explain how the gap wind favors the developing vortex in the Control Run.

9.4 Evolution of the Developing Vortex and of the Non-Developing Vortex

As was discussed in section 9.3, after the 20th, a vortex within the EW trough develops in the Control Run. The Blocked Run forms also a vortex but fails to develop. Moreover, the developing vortex in the Control Run has a TC-like structure (see chapter 8). In this section we study the respective evolution of the developing and of the non-developing vortex, as well as the impacts of the gap wind on developing vortex by comparing them.

We first compare the timeseries of the mean vorticity and mean $\theta_c$, to diagnose the developing vortex and the non-developing vortex considering a reference frame centered on the location of their respective minimum MSLP. This will reveal us the time when the vortices become substantially different $\theta_c$ of each vortex is considered within 30 km from its respective center in a layer between 800 and 600 hPa. The 950 to 800 hPa layer is indicative of the potential impacts of the gap wind on the developing vortex at low levels. The 800-600 hPa layer diagnoses the development of a warm core, characteristic of TCs above the planetary boundary layer $\theta_c$ to increase in the developing vortex but not in the non-developing vortex. We consider a radius of 30 km because it is about the size of the developing vortex when it achieves maximum intensity (Fig. 8.9). However, we recognize that the developing vortex is larger during its initial stages, about 200 km (Fig. 9.3 (a)), and then it shrinks to little less than 25 km, the size of the vorticity patch in the mature stages, e.g. Fig. (8.9 (a)). Thus, 30 km is considered as a mean value to represent the later stages of the developing and the non-developing vortices. We note that when larger radii values (~100 km) are considered the resulting timeseries for both variables are similar for both vortices and are representative of the evolution of the EW trough (not shown). Moreover, for radii between
10 km and 50 km, the resulting timeseries yield to results similar to those presented in the next paragraphs. By analyzing the evolution of these variables for each vortex we can identify at what time the vortices become different and what processes lead them to take different pathways.

Figure 9.6 (red lines) shows the timeseries of the mean vorticity and the mean $\theta_e$ for the developing vortex. The vorticity increases in time, first slowly, and then faster after the 21st (Fig. 9.6 (a)). Its mean $\theta_e$, also experiences a significant increase after the 21st, suggesting that on the 21st the developing vortex develops a warm core (Fig. 9.6 (b)). In contrast, the corresponding timeseries of the mean vorticity and mean $\theta_e$ for the non-developing vortex remain roughly constant, except for some fluctuations of the mean $\theta_e$ (blue line Fig. 9.6 (a,b)). These mean $\theta_e$ fluctuations are associated with some individual convective cells occurring close to its center (not shown, but at this time precipitation occurred close to the center and columns with high $\theta_e$ were observed close to its center). In a similar fashion to the evolution of their intensity (Fig. 9.2 (b)), the mean vorticity, and mean $\theta_e$ of the two vortices are similar prior to the 21st. Then on the 21st they differ; the mean vorticity and mean $\theta_e$ of the developing vortex increase but in the case of the non-developing they do not increase. This means that the effects of the gap wind on the developing vortex takes about to two days to become evident, i.e., the gap wind peaks at 2100 19th, but the vortex develops until the 21st (its vorticity and wind intensify).

The evolution of the vertical profiles of the mean vorticity and mean $\theta_e$ within 30 km from the respective location of each vortex is analyzed next (Fig. 9.7). The analysis will reveal the vertical levels where the differences between the vortices occur first.

The vorticity of the developing vortex has values larger than $1 \times 10^{-4} s^{-1}$ in a shallow layer below 900 hPa before 1800 21st (black isopleths Fig. 9.7 (a)). Afterward, the mean vorticity deepens and intensifies steadily; this deepening and growth is characteristic of developing TCs. In
contrast the vorticity of the non-developing vortex (black isopleths Fig. 9.7 (b)) increases is characterized by periods where the vorticity increases followed by periods of weakening. For example, the vorticity below 900 hPa increases between 0600 20th and 1200 20th, and along the vertical between the 1200 21st and 0000 22nd. Thus, the non-developing vortex attempts to increase its vorticity but it fails. Therefore, considering that on the 20th both vortices are to the south of the Isthmus of Tehuantepec, and that in the Control Run the gap outflow is close to the developing vortex. Then, the gap outflow has to contribute to sustain the low-level (~900 hPa) vorticity of the developing vortex on the 20th, and the subsequent differences are likely due to the different pathways each vortex takes.

The vertical profiles of $\theta_e$ are similar for both vortices before 1800 20st (color shading Fig. 9.7). Afterward, the $\theta_e$ of the developing vortex increases substantially along the vertical, indicating the formation of a warm core (Fig. 9.7 (a)). In contrast, the $\theta_e$ of the non-developing trough remains roughly constant through all the period (Fig. 9.7 (b)). This agrees with the previously shown timeseries of $\theta_e$, similar for both vortices prior the 21st. Afterward, they diverge (Fig. 9.6 (b)). Thus, the $\theta_e$ vertical profiles suggests that the gap wind has little impact on $\theta_e$ at low levels.

Considering that before the 21st, $\theta_e$ close to the surface is similar for both vortices, but the vorticity close to the surface of the developing vortex is greater than that of the non-developing vortex. Then, the main impact of the gap outflow on the developing vortex is to increase the vorticity, having little impact on the $\theta_e$ (in agreement with the discussion of chapter 6). Furthermore, the differences in $\theta_e$ between both vortices after the 21st likely arise due to the intensification of the development vortex (cyclogenesis) and are not as a direct effect of the gap flow on the vortex.
9.4.1 The Spatial Evolution of the Developing Vortex and the Non-Developing Vortex

We showed that the developing vortex and the non-developing vortex are similar prior the 21st. Afterward they take different pathways. An analysis of the evolution of their environment will reveal when and where the gap wind influences the developing vortex. Their surroundings are represented by Hovmoller diagrams of azimuthal averages for different variables centered on the location of the minimum MSLP. These variables are the 950 hPa tangential wind, and the vorticity and $\theta_e$ at 950 and 700 hPa. As said earlier, the differences in their evolution are attributed to the effects of the gap wind on the trough.

Figure 9.8 shows the tangential wind at 950 hPa for the developing vortex and the non-developing vortex. For the case of the developing vortex the tangential wind shows that on the 20th, the wind close to the center (~50 km) is weak (~5 m/s) but is stronger far from the center (~12 m/s). The strong wind far from the vortex center is the gap outflow. Then on the 21st, the tangential wind increases close to the vortex center reaching intensities of 25 m/s on the 22nd but remains unchanged far from it (Fig. 9.8 (a)). In contrast, the Hovmoller diagram for the non-developing vortex shows that the wind remains roughly constant throughout the simulation (Fig. 9.8 (b)). However, it can be recognized that there is a transient intensification of the wind on the 1200 21st lasting a few hours. This transient intensification of the non-developing vortex wind coincides with a slight increase of the vorticity and $\theta_e$ (see their respective vertical profiles at 1200 21st (Fig. 9.8 (b))). By this time, several vorticity anomalies are located close to the minimum MSLP (see Fig. 9.14 (f)). However, these vorticity anomalies instead of aggregating around a sole vorticity patch, like during a cyclogenesis process, are spread over a broad area, and the wind intensification fails to occur.
The Hovmoller diagram of the vorticity for the Control Run shows a substantial increase at 700 hPa and at 950 hPa within 25 km from the center of the developing vortex after 0000 21st (Fig. 9.9 (a) and 9.10 (a)). This indicates that the developing vortex spun up one day after the minimum MSLP is formed (1800 19th). In contrast, the vorticity of the Blocked Run shows some sporadic maxima close to the non-developing vortex (~50 km) by 1200 21st, more evident at 950 hPa than at 700 hPa (Fig. 9.9 (b) and Fig. 9.10 (b)). These maxima are related to the aggregation of some convectively active vorticity anomalies suggesting that the non-developing vortex attempts to form a strong vortex but fails to complete the cyclogenesis process. The vorticity beyond 100 km the minimum MSLP is similar for both vortices indicating that they have similar surroundings, consisting of the EW trough in the EPac.

Both Hovmoller diagrams for $\theta_e$ are roughly similar for the developing vortex and the non-developing vortex before the 21st (Figs. 9.11 and 9.12). Afterward, the developing vortex increases its $\theta_e$ and by the 22nd there is a distinguishable maximum at 950 and at 700 hPa close to the minimum MSLP. This maximum occurs simultaneously with the increase of the vorticity (Figs. 9.9 (a) and 9.10 (a)). Meanwhile, the $\theta_e$ of the non-developing vortex fails to develop a persistent warm core, yet a transient increase $\theta_e$ occurs on 1200 21st at 700 hPa (Fig. 9.12 (b)). This increase of $\theta_e$ occurred simultaneously with a vorticity (Fig. 9.10 (b)) and wind intensity (Fig. 9.8 (b)) increase, suggesting that the non-developing vortex “attempted” to develop a TC-like vortex (high vorticity and high $\theta_e$). The $\theta_e$ beyond 100 km from the minimum MSLP, is similar for both vortices indicating that both systems share a similar environment.

In summary, by considering a reference frame centered on the respective locations of the MSLP for each vortex it was found that between the 20th and the 21st, both vortices are similar in terms of their wind intensity and $\theta_e$, but their vorticity below 900 hPa is different. The developing
vortex has stronger vorticity than the vorticity of the non-developing vortex. Afterward the developing vortex intensifies but the non-developing vortex does not. Therefore, the primary impact of the gap wind on the formation of the developing vortex is to increase the low-level vorticity and is consistent with the discussion in chapter 6. In chapter 6 it was shown that the large vorticity values within the cyclogenesis region were attributed to the gap outflow. The gap wind contributions to the development of the vortex in the Control Run is explored next.

9.4.2 The Contributions of the Gap Wind to the Vortex Development

In this subsection we discuss the underlaying mechanism responsible for the formation of the developing vortex. This vortex was first detected close to the surface and is barely detected at mid-levels. Hence, we focus on the evolution of the wind and its vorticity at 950 hPa between the 19th and the 22nd for both the Control and Blocked Runs (Figs. 9.13 to 9.15). At 700 hPa, the mid-levels, there are not perceivable effects of the gap wind during this its development in this period (compare the vorticity, $\theta_e$ between both runs in Fig. 9.7). The precipitation is also considered to indicate regions where latent heat is released by condensation. The period analyzed here covers a little after the minimum MSLP is detected in the Control Run and when the developing vortex has formed a hurricane at 0600 22nd.

Figure 9.13 shows that both runs simulate the trough southerlies and the ITCZ vorticity anomalies along them. The ITCZ vorticity anomalies consist of a SW-NE strip of convective scale (~10km) vorticity anomalies accompanied by rainfall. However, the Control Run (Fig. 9.13 (a,c,e,g)) has stronger northerlies than the Blocked Run (Fig. 9.13 (b,d,f,h)). The gap outflow vorticity in the Control Run consists of a S-N strip of vorticity to the east of the strongest northerlies. A consequence of the weak northerlies and its associated vorticity in the Blocked Run is the weaker vorticity between 13°N and 15°N and the absence of a cyclonic flow between these
latitudes (Fig. 9.13 (b,d,f,h)). Nevertheless, each run develops a minimum MSLP (green circles) surrounded by a cyclonic flow, this pattern corresponds to the vortex in each run. In the Control Run the minimum MSLP locates close to the gap outflow, and in the Blocked Run locates close to the ITCZ vorticity anomalies. The vortex in the Control Run is characterized by strong winds and by a large vorticity patch. In contrast the vortex in the Blocked Run has weaker winds and its vorticity consists of convective scale vorticity anomalies spread over a large area. Moreover, at 0600 20th, the minimum MSLP in the Blocked Run separated from the vorticity anomalies. Therefore, the gap outflow helps to develop a stronger vortex.

The evolution between the 2100 19th and the 0600 20th of the vorticity for each vortex at 950 hPa is substantially different. In the case of the Control Run, the gap outflow vorticity merges with the ITCZ vorticity anomalies (Fig. 9.13, left column). Whereas in the Blocked Run, the vortex vorticity is the ITCZ vorticity which roughly remains unchanged over the SW-NE strip surrounded by a weak cyclonic flow (Fig. 9.13, right column). This evidence that the gap outflow vorticity assists to form a vortex to its east.

The merging of the gap outflow vorticity with the ITCZ vorticity anomalies in the Control Run is as follows. The ITCZ vorticity anomalies along the southerlies, propagate along an arc (hand drawn in green; Fig. 9.13, left column) that increases in length until it merges with the gap outflow vorticity. On 0600 20th, the ITCZ vorticity is undistinguishable from the gap outflow vorticity and the vorticity within the vortex consists of a spiral centered on a vorticity patch at (94°W,13.5°N).

It is worth noting that the gap outflow usually has an anticyclonic curvature, but if close to a low MSLP it acquires a cyclonic curvature (Bourassa, 1999). In this case, the EW trough is responsible for establishing a low MSLP in the EPac (inferred by its low geopotential heights, Fig
9.4 (left column)), and the gap outflow turns cyclonic in response to the low MSLP. Then, the ITCZ vorticity anomalies advected by the trough southerlies merge with the outflow vorticity forming a “broad” positive vorticity patch (Fig. 9.13 (g), center of the spiral).

When the vorticity merges between the 19th and the 20th, it is seen that the low-level wind remains roughly constant. For example; the intensity of the 10-m wind close to the vortex (Fig. 9.2 (c)) and the intensity of the 950hPa wind (Fig. 9.14 (a,c,e,g)) remain roughly unchanged. This is characteristic of a barotropic merging. During a barotropic process two vorticity patches close enough together will undergo an irreversible merging (Melander et al, 1987, and Dritschel et al, 1992), and the wind does not intensify (Kilroy, et al, 2016). Regarding the evolution of the developing vortex, Kilroy, et al (2016) found that prior cyclogenesis smaller vorticity anomalies merge through a barotropic process prior to forming a broader vorticity patch. Thus, the vorticity merging in the blocked run is supportive of cyclogenesis process between 1800 19th and 0600 20th.

Despite the fact that a vortex had already formed in the Control Run by 0600 20th, it does not experience any significant intensification of the wind and its vorticity in the next 24 hours, between the 0600 20th and the 0600 21st (Fig. 9.13 (g) and 9.14 (a,c); also Fig. 9.6 (a) and Fig. 9.7 (a)). During this period, precipitation is absent within the vortex yet it occurs in the convective-scale (10km) vorticity anomalies surrounding it (Fig. 9.13 (g) and 9.14 (a,c)). The absence of precipitation within the vortex indicates that the release of latent heat by condensation is not yet a dominant mechanism within vortex prior 0600 21st.

Finally, at 1800 21st precipitation occurs within the developing vortex and becomes convectively active and the wind intensifies. At this time the developing vortex has formed a smaller, yet strong vorticity maximum surrounded by convectively active spiral bands and strong winds (~35 m/s) characteristic of TCs (9.14 (e)). Additionally, between 0600 and 1800 21st, a
surface radial inflow develops around the vorticity maximum and the surface latent heat fluxes (not shown). This in turn enhances convection and subsequently the vorticity. The formation of the surface radial wind and merging of convectively active vorticity anomalies, resembles the so-called tipping point leading to cyclogenesis or to TC intensification described in Kilroy et al (2016). The tipping point is when convection occurs close to the vorticity maximum (the developing vortex) leading to an intensification of the wind and the vorticity due to the release of latent heat by convection. Also, the vorticity builds up and $\theta_e$ increases by 0600 21st (Fig. 9.6). We want to comment that the mechanism presented by Kilroy (2016) is consistent with the WISHE mechanism in the sense that both are dependent on surface heat fluxes and radial inflows as a mechanism to sustain the convection around the TC vortex.

Overall, the cyclogenesis of Patricia on 1800 19th and its early stages on the 21st follows the cyclogenesis process described by Kilroy et al (2016). That is, cyclogenesis is due to vorticity merging or aggregating at low-levels. Subsequently, the small-scale convectively active vorticity anomalies keep merging around the developing vortex helping to further increases the vorticity, the inward radial flow, and the surface heat fluxes. The net result is the intensification and vertical extension of the vorticity in phase with the intensification of the wind.

In contrast, the non-developing vortex of the Blocked Run shows that convectively active vorticity anomalies (precipitation) aggregate slowly around a broad vorticity patch, yet it lacks the vorticity band over the northwest quadrant (Fig. 9.14 (b,d,f)). Also, the wind slightly intensifies until the 1800 21st but forms an elongated circulation that is dissimilar to a TC-like circulation (Fig. 9.14 (f)). However, the wind weakens and precipitation ceases in the next hours (Fig. 9.15 (b)) whereas the developing vortex keeps intensifying (Fig. 9.15 (a)). Thus, considering that the ITCZ vorticity anomalies underneath the trough attempt to form a vorticity maximum but fail to
trigger cyclogenesis, it can be said that the gap outflow vorticity acts as a “catalyst” to develop the vortex and to trigger cyclogenesis.

9.5 Discussion and Conclusions

The formation of the developing vortex (which leads to cyclogenesis) results from the merging of the gap outflow positive vorticity with ITCZ vorticity anomalies between the 19th and 20th. The Blocked Run showed that in absence of the gap wind the vortex does not develop; and cyclogenesis fails.

Once the low-level vortex is formed, convective-scale (~10 km) vorticity anomalies keep aggregating around it for one day between the 20th and the 21st and the vortex becomes more compact. By 1800 21st, the vortex becomes more compact, and convection associated with the vorticity anomalies occurs around the vorticity maximum. This convection close to the vorticity maximum helps to increase the vorticity and to further intensify the vortex wind, and the vortex is similar to a mature TC-like vortex reaching a hurricane category 1 status on 0300 22nd. In general terms, the gap wind acts as a catalyst to increase the low-level vorticity and to help intensify a vortex underneath the trough. After that the vortex develops via the additional convection occurring within the vortex.

In contrast, the Blocked Run shows that in absence of the gap outflow, cyclogenesis fails. Nevertheless, a non-developing low-level vortex lingers within the trough over the EPac. This non-developing vortex follows a similar path to that of the developing vortex but is weaker and much broader than the developing vortex of the Control Run. Also, it fails to intensify at least prior to when the simulation stops on 2100 22nd. We suspect that there won’t be additional attempts since developing vortex approaches land and the environmental shear increases by the 23rd in the ERA-
5 (not shown). However, there appear to be attempts to develop the vortex since vorticity anomalies associated with convection merge around the vortex.

In conclusion, the gap wind acts as a *catalyst* to develop the vortex leading to the cyclogenesis of Patricia. The gap outflow increases the low-level vorticity underneath the trough, and merges with the ITCZ vorticity to develop a vortex. In absence of the gap wind the vortex fails to form.

**Figures**

![Figure 9.1. A) Model domains for the blocked run. Nested domains D₁ (heavy solid line), D₂ and D₃ have 6, 2 and ⅔ km horizontal grid lengths respectively. B) topography profiles at 17°N of the original topography (black), synthetic topography anomaly (red) and resulting topography (grey). Horizontal scale in degrees west and vertical scale in km.](image-url)
Figure 9.2. Mean meridional wind in [96.5°W,92.5°W]x[16°WN,18°N]x[Surface,800hPa] for the ERA-5 (black), Control Run (red) and Blocked Run (blue) in upper panel. Best-Track (black), developing vortex(red) and non-developing vortex (blue) tracks and intensity in middle and bottom panels, respectively. Heavy dots, black, red or blue placed on the 0600 20th, 0600 21st and 0600 22nd of October. Gray shading in B) indicates land. Horizontal axis in A) and C) are in days after the initial time (upper row) and the date (lower row).
Figure 9.3. 950 hPa wind (barbs) at 0600 19th for the Control (A) and Blocked Runs (B). Black barbs for wind intensities less than 10 m/s, green for wind intensities between 10 and less than 15 m/s, orange for wind intensities between 15 and less than 20 m/s, red for wind intensities between 20 and less than 25 m/s, and magenta for wind intensities larger than 25 m/s.
Figure 9.4. Wind (barbs), geopotential height (color shading) at 700 hPa (right column) and 950 hPa (left column) and topography (hatching) at 0600 19th (upper row), 0600 20th (middle-top row), 0600 21st (middle-bottom row), 0600 22nd (bottom row) for the Control Run. The geopotential height is expressed as the deviation from the mean geopotential value between the 13th and the 22nd to the closest dam (550-m for the 950 hPa level and 3150-m for the 700 hPa level).
Figure 9.5. Wind (barbs), geopotential height (color shading) at 700 hPa (right column) and 950 hPa (left column) and topography (hatching) at 0600 19th (upper row), 0600 20th (middle-top row), 0600 21st (middle-bottom row), 0600 22nd (bottom row) for the Blocked Run. The geopotential height is expressed as the deviation from the mean geopotential value between the 13th and the 22nd to the closest dam (550-m for the 950 hPa level and 3150-m for the 700 hPa level).
Figure 9.6. Timeseries centered in the locations of the simulated TC (red) and of the remnants of the trough (blue) of A) The mean vorticity between 950 and 800 hPa. B) The mean $\theta_e$ at 700 hPa Each variable is averaged within 30 km from the location of the system. Horizontal axis in days into the simulation (upper row) and in days of October (lower row). Circles located every day starting on 0600 20th.
Figure 9.7. Evolution of the vertical profiles of the $\theta_e$ (shading) and of the vorticity (isopleths, $\times 10^{-4}$ s$^{-1}$) for the developing vortex in the control run (A) and for non-developing vortex in the blocked run (B).

Figure 9.8. Hovmoller diagram of the tangential wind at 950 hPa for the Control Run (A) and the Blocked Run (B). The horizontal axis is in km from the center of the TC vortex and from the (A) non-developing vortex (B).
Figure 9.9. Hovmoller diagram of the vorticity at 700 hPa along the track of the developing vortex (A) and along the non-developing vortex (B).

Figure 9.10. Hovmoller diagram of the vorticity at 950 hPa along the track of the developing vortex (left) and along the non-developing vortex (right).
Figure 9.11. Hovmoller diagram of $\theta_e$ at 700 hPa along the track of the developing vortex (A) and along the non-developing vortex (B).

Figure 9.12. Hovmoller diagram of $\theta_e$ at 700 hPa along the track of the developing vortex (A) and along the non-developing vortex (B).
Figure 9.13. 950 hPa horizontal wind (barbs), 950 hPa vorticity (shading), precipitation (gray shading >5mm hr⁻¹), and topography (hatching) at 2100 19th (A,B), 0000 20th (C,D), 0300 20th (E,F) and 0600 20th (G,H) for the Control Run (left column) and the Blocked Run (right column). Green arc indicates the ITCZ vorticity anomalies in the Control Run. Green circles (30 km radius), correspond to the location of the minimum PMSL of figure 9.2 (b).
Figure 9.14. 950 hPa horizontal wind (barbs), 950 hPa vorticity (shading), precipitation (gray shading >5mm hr\(^{-1}\)), and topography (hatching) at 1800 20\(^{th}\) (A,B), 0600 21\(^{st}\) (C,D), and 1800 21\(^{st}\) (E,F) Control Run (left column) and the Blocked Run (right column). Green circles with 30km radius located at the minimum MSLP location.
Figure 9.15. 950 hPa horizontal wind (barbs), 950 hPa vorticity (shading), precipitation (gray shading >5mm hr\(^{-1}\)) at 0600 22\(^{nd}\) for the Control Run (A) and the Blocked Run (B). Green circles with 30km radius located at the minimum MSLP location. Note the different horizontal scales between panel A and B.
10. CONCLUSIONS AND PROPOSED FUTURE WORK

Here we present a summary and concluding remarks of the impacts of tropical weather systems on the gap wind, including its climatology in section 10.1. The cyclogenesis of hurricane Patricia (occurred in October of 2015) in section 10.2. The sensitivity of the cyclogenesis of hurricane Patricia to the intensity of the gap wind in section 10.3. And we discuss possible avenues for future work in section 10.4.

10.1 The Impacts of Tropical Weather Systems on the Gap Wind

The mountain gap wind of the Isthmus of Tehuantepec follows the seasonal cycle and is characterized by fluctuations of the meridional wind. The mean northerlies are stronger in winter than in summer. The fluctuations last a few days, and their amplitude maximizes in winter and spring, and minimizes in summer.

The wavelet power spectrum of the Isthmus of Tehuantepec gap wind was used to identify the dominant periods throughout the year. In winter and spring, the period band between 2-10 days strongly dominates the fluctuations, indicating a strong influence of synoptic weather on the gap wind. In summer, the period band broadens, indicating that weather systems in different time scales equally impact the fluctuations of the gap wind. These periods range from the diurnal cycle to sub-seasonal timescales. However, when summer fluctuations associated with northerlies are stronger than the monthly climatology, their dominant periods fall back to the synoptic scale period band. Therefore, the gap wind fluctuations associated with extreme gap northerlies are mostly modulated by synoptic scale weather phenomena in summer.

The primary factors modulating the extreme gap northerlies fluctuations all year are surface ridges in the Gulf of Mexico associated with midlatitude systems that establish a MSLP gradient along the mountain gap. However, their influence on the gap wind weakens in Summer,
when the midlatitude systems retreat poleward, leading to other meteorological features that modulate the gap wind. These features include tropical systems (e.g., EW troughs and TCs), and the diurnal cycle. Regarding the tropical systems, they impact the gap wind in two different ways, one is through their associated negative PMSL anomalies on the EPac establishing a pressure gradient along the mountain gap. The second is a kinematic effect consisting in their low-level (950 hPa) circulation channeling the gap wind outflow in the EPac. In essence, the gap outflow merges with their circulation, enhancing part of the trough circulation, where the gap outflow merges with the trough circulation.

10.2 The Cyclogenesis of Hurricane Patricia

The environment over the EPac in the week before cyclogenesis (between the 8th and the 20th of October) was favorable for tropical cyclogenesis (Gray, 1968). It consisted of low vertical wind shear, large precipitable water values (larger than 55 mm), and SSTs greater than 26°C. Additionally a mean northerly mountain gap flow in the Isthmus of Tehuantepec was also present.

Based on the results shown in this thesis, the sequence of events leading to the genesis of Patricia, and its evolution prior to its rapid intensification can be characterized by three stages. The stages consist of the formation of an EW trough over the EPac on the 10th of October of 2015 and by the approaching of a Caribbean EW trough (stage I). The merging of the EW troughs (stage II) and the formation of a low-level vortex over the EPac involving the gap outflow vorticity and the ITCZ vorticity (stage III). Then, the vortex intensification and hurricane strength winds on the 22nd of October (stage IV).

These stages are illustrated by a schematic of the evolution of the vorticity anomalies and their associated flows involved in the cyclogenesis process at 950 and 700 hPa between the 10th and the 22nd of October. Red dots indicate regions of positive relative vorticity associated either
with the ITCZ (east-to-west strip) or the EW troughs or vortices (ovals). The blue dots indicate the regions of negative relative vorticity associated with the midlatitude ridge. The red lines indicate the positive vorticity patch associated with the gap outflow. The blue curved arrows represent the associated flows.

**Stage I) Formation of the EPac EW trough (Fig. 10.1 (a,b)).** The EPac EW trough is depicted by its flows (blue arrows) and its vorticity (red oval with dots) at 700 hPa over the EPac (Fig. 10.1 (b)). The trough was formed over the Panama Bight region on the eastern limits of the ITCZ (red dots delimited by the blacked dashed line) on the 10th of October and moved northwestward over the EPac parallel to Central America and overlapped with the ITCZ at 950 and 700 hPa. The Panama Bight is a region that is prolific for EW genesis (e.g., Torres and Thorncroft, 2021). Simultaneously, over the Caribbean another weak EW trough, only discernable at 700 hPa, moves toward Central America and approached the EPac EW (red dots over the Caribbean, Fig. 10.1 (b)). Also, a persistent gap wind in the Isthmus of Tehuantepec is present at this time which forms in association with a midlatitude ridge and its anticyclonic flow. This gap wind will impact the vorticity underneath the EW trough in the subsequent stages.

**Stage II) Merging of the EW troughs (Fig. 10.1 (c,d)).** The EPac and the Caribbean EW troughs at 700hPa merged and formed a broader trough straddling Central America (Fig. 10.1 (d)). The resulting system stalled over the Yucatan peninsula and spanned to the ITCZ between the 18th and the 20th of October (red dots, Figure 10.1 (d)). The southerlies underneath the broader trough at 950 hPa advect ITCZ vorticity anomalies and moisture toward the continent. The vorticity anomalies are associated with convection helping to increase the vorticity at low levels. Additionally, the trough northerlies together with the midlatitude ridge over the GoM further enhanced the gap wind as well as the low-level shear positive vorticity (Fig. 10.1 (c) red straight
lines indicate the positive vorticity). Therefore, two vorticity sources are found over the EPac underneath the trough which are key points to the next stage.

**Stage III) Formation of the low-level TC vortex over the EPac (Fig. 10.1 (e,f)).** At 950 hPa, the vorticity increased, underneath the EW trough over the ocean in the EPac (Figure 10.1 (e)). The increase was due to the merging of 1) the vorticity associated with the enhanced mountain gap outflow (straight lines) and 2) the ITCZ vorticity anomalies. The increase of vorticity leads to a broad low-level vortex (red circle) underneath the EW trough. The air originated in the ITCZ is advected by the southerlies, enters the vortex, recirculates and moistens it. Also, the convection associated with the ITCZ vorticity anomalies contribute to release latent heat by condensation along the southerlies and within the vortex helping to increase the vortex vorticity. Overall, the moist and elevated vorticity at low-levels creates a favorable precursor vortex for cyclogenesis.

At 700 hPa, the EW trough (red oval), does not intensify over land but it does over the EPac (red circle) atop the low-level vortex. The intensification over the EPac is due to the stretching of the low-level vortex.

**Stage IV) Intensification of the TC vortex (Fig. 10.1 (g,h)).** By late on the 20th and early on the 21st, a radial inflow at 950 hPa centered on the vortex develops and is surrounded by convection. The radial inflow helps to transport moisture to the convective cells and helps to increase the $\theta_e$ at low levels. Together the convection and the release of latent heat by condensation helps to increase the vortex’s vorticity and consequently its winds. The vortex reaches hurricane winds on the 22nd.

Overall, the mechanism for the intensification of the vortex for Patricia, in stage IV is similar to that proposed by Kilroy et al (2016). The intensification mechanism is characterized by the so-called tipping point. The tipping point occurs after the merging of small-scale vorticity.
anomalies, typically associated with convection around a larger vorticity maximum within a vortex. It is worth noting that Kilroy et al. (2016) did not offer a threshold vorticity, or spatial of the vorticity maximum. When persistent convection leads to convergence at low levels, forming a persistent radial inflow. The convergence associated with the radial inflow allows for the convergence of moisture and surface latent heat that sustains the convection, which in turns increases the vorticity developing the vortex. In the case of hurricane Patricia, the ITCZ vorticity anomalies merge with the gap outflow vorticity to form the vorticity maximum. Also, the ITCZ vorticity anomalies contribute with most of the convection in the early stage of the TC vortex. The role of the gap outflow vorticity in the formation of the vorticity maximum will be discussed in the next section.

In general, the vorticity of the gap northerlies and the ITCZ vorticity anomalies are important factors for the cyclogenesis of Patricia. The role of each vorticity anomaly in the cyclogenesis is summarized in the next subsections.

10.2.1 The role of the Gap Wind during the Cyclogenesis of Hurricane Patricia

The role of the gap wind on the cyclogenesis of hurricane Patricia is studied through the ERA-5 and through a numerical experiment using the WRF model. The numerical experiment consists of comparing two realistic simulations. One simulation is the Control Run and reproduces the cyclogenesis of hurricane Patricia. The second one is the Blocked Run, is identical to the Control Run but with the mountain gap filled. By comparing both simulations it is possible to determine the impacts of the mountain gap wind.

The ERA-5 revealed that the gap outflow was materially connected with the low-level cyclonic flow associated with the vortex leading to cyclogenesis. This means that air parcels flowing through the mountain gap flowed into the vortex leading to cyclogenesis. This connection
allows the mountain gap parcels to transport vorticity and $\theta_e$ toward the vortex leading to cyclogenesis. We found that the vorticity of these parcels was relatively high compared with the vorticity in the low-level cyclonic flow over the EPac. But their $\theta_e$ was similar to the mean value in this cyclonic flow. Therefore, the gap outflow contributes to increase the vorticity within the cyclonic flow but is not the case of the $\theta_e$.

The numerical experiment shows two consequences of filling the mountain gap. 1) unsurprisingly, the gap wind is suppressed; 2) cyclogenesis does not occur, and 3) the EW trough over the EPac is weaker between the 19th and the 20th. Nevertheless, there is a failed attempt of the EW trough over the EPac to trigger cyclogenesis. In general, the experiment reveals that the gap wind favors the EW trough development and is a determining factor for the cyclogenesis of hurricane Patricia.

In the Blocked Run, the ITCZ vorticity anomalies at 950 hPa failed to merge around a distinguishable vorticity maximum and remain spread along the southerlies, nevertheless the EPac EW retains a cyclonic flow at 950 hPa. The consequences of suppressing the gap wind are further illustrated by highlighting the missing elements of the schematic of Fig. 10.1 in the Blocked Run. First, the vorticity of the cyclonic flow over the EPac is weaker because the gap northerlies and its vorticity are absent (Fig. 10.1 (c,e)). Second, as the EW trough in the Blocked Run moves over the EPac, the easterlies (represented by the gap outflow) and its vorticity associated are much weaker (Fig. 10.1 (e)). Also, the convection is only present along the southerlies on the western quadrant and is associated with the ITCZ vorticity anomalies.

In terms of the hypotheses presented in chapter 2 we found: 1) the gap outflow vorticity contributes to increase the vorticity of the vortex leading to cyclogenesis; moreover, this additional vorticity is needed to trigger cyclogenesis. 2) The gap outflow does not contribute to increase the
\(\theta_e\) of said vortex. Both the Control Run and the Blocked Run have similar \(\theta_e\) values. 3) There was no evidence of upstream anomalies over the GoM being transported by the mountain gap northerlies into the cyclogenesis region. Therefore, the dynamic mechanism (hypothesis II) is the principal contribution of the gap wind to cyclogenesis. In contrast the thermodynamic mechanism (hypothesis III) and the transport mechanism are not present (hypothesis IV). Nevertheless, we should not rule out that for other cyclogenesis cases, upstream anomalies, e.g., vorticity or moisture anomalies over the GoM may be transported by the gap outflow into the cyclogenesis region.

10.2.2 The importance of the ITCZ Vorticity Anomalies during the Cyclogenesis of Hurricane Patricia

The importance of the ITCZ vorticity anomalies during the cyclogenesis of Patricia was explored through an analysis of the first GFS and ECMWF operational forecasts that predicted cyclogenesis. The GFS and ECMWF operational forecasts initialized on the 1800 17\(^{th}\) of October and 0000 18\(^{th}\) of October, respectively were the first to capture the cyclogenesis of Patricia. Interestingly cyclogenesis in each of these forecasts was triggered due to different mechanisms. Nevertheless, their predictions for the gap wind at low levels (950 hPa) and the structure of the EW trough were roughly similar. This led us to investigate what could have impacted cyclogenesis in these forecasts differently.

By comparing both forecasts, it was found that cyclogenesis is modulated by the intensity of the ITCZ vorticity anomalies. If these vorticity anomalies are too strong, as is the case in the GFS forecast, they appear to trigger cyclogenesis by themselves prematurely, regardless of the gap wind. If they are weaker, the ITCZ vorticity merges with the gap outflow vorticity forming the vortex leading to cyclogenesis. Ultimately the ITCZ vorticity anomalies need to be properly forecasted for an adequate cyclogenesis forecast. In turn their spatial scale of about 10 km means
that the cyclogenesis predictability is limited by the predictability of the convective scale processes.

10.3 Future Work

We propose several additional avenues for future work to further understand the dynamics of the mountain gap winds and their interactions with EW troughs and hurricanes.

- **The Importance of the Diurnal Cycle on the Mountain Gap Wind of the Isthmus of Tehuantepec.** Future work should consider a more detailed analysis of the diurnal land/sea breeze in the Isthmus of Tehuantepec to understand the wind patterns enhancing the northerlies in the mountain gap at 0000 UTC. This can be done by analyzing the diurnal wind and their associated circulation for different periods of the year in datasets with higher temporal and spatial resolutions than the CFS, e.g., ERA-5.

- **Sensitivity of the Cyclogenesis of Hurricane Patricia to the Intensity of the Low-Level ITCZ Vorticity Anomalies.** The analysis of the GFS forecast revealed that cyclogenesis of hurricane Patricia is sensitive to the amplitude of the ITCZ vorticity anomalies. These vorticity anomalies have sizes of about 10 km, meaning that they are dependent on convective scale processes. If these vorticity anomalies are strong enough, they appear to be able to trigger cyclogenesis regardless of the gap wind. Thus, we propose to perform additional numerical experiments aimed to strengthen or weaken convection, e.g., the microphysics, and the convection and planetary boundary layer schemes. These simulations will reveal whether strong vorticity anomalies are capable of triggering cyclogenesis on their own or if the gap wind is still needed for their development. On the contrary if we weaken these vorticity anomalies it would be possible to address the question of whether the gap wind is enough to develop the vortex leading to cyclogenesis.
In this regard we have done one preliminary run identical to the Control Run and other identical to the Blocked Run but with the southern boundary of the outermost domain extended southward to 5°N in each run (figures not shown). This southward extension of the outermost domain covers the ITCZ and allows the model to fully model the ITCZ convection. Both preliminary runs produced strong ITCZ vorticity anomalies that develop strong vortices that appear to be TC vortices. Therefore, we are confident that ITCZ vorticity anomalies are enough to trigger cyclogenesis provided they are strong enough. Nevertheless, a more detailed analysis of this run is needed to fully validate this assertion.

- **The Importance of Gap Winds on the Trough Development.** Considering that the Isthmus of Tehuantepec gap wind helps to strengthen the EW trough over the EPac it is likely that the Papagayo gap wind in Central America may also strengthen EW troughs. In this regard we studied in chapter 4 the early stages of the developing EW troughs for Patricia and for Olaf. There, it was observed that their low-level wind experienced some enhancement due to the presence of a Papagayo gap wind. Therefore, we recommend that a numerical experiment should be carried out that is similar to the one carried out here but applied to the case of the developing EW trough for Olaf to assess the importance of the Papagayo gap wind on its evolution.

10.4 Final Comments

There is a close relationship between the mountain gap wind of the Isthmus of Tehuantepec and tropical systems close to said mountain gap. It has been shown that tropical systems, like EW troughs or hurricanes can enhance the gap wind but they also can be enhanced themselves by the gap outflow. More specifically, it has been shown that regardless of its cause, the gap wind can develop EWs in the EPac and trigger cyclogenesis by increasing the low-level vorticity. This interaction forms a complex environment for EW troughs in the EPac where two
aspects need to be considered in the forecast for EWs troughs in the EPac. These aspects include, the location, extension and timing of the EW trough, and simultaneously, the timing of the weather systems that impact the gap wind, whether they are of midlatitude or tropical origin. The proper modeling of these two aspects will improve our forecasts for EWs troughs in presence of gap winds.

**Figures**
Fig. 10.1. Schematic of the events involved in the cyclogenesis of hurricane Patricia and its early stages. Flow direction (blue arrows), regions of positive vorticity (red dots), negative vorticity (blue dots) positive vorticity associated with the northerlies (red straight lines), ITCZ vorticity (hatched with black contour) and topography at 950 hPa (crisscross). The intensity of the vorticity is illustrated by the dot density.
Fig. 10.1. Continuation.
REFERENCES


