Stratospheric Intrusions and Transient Convection in the Eastern Tropical Pacific

by

Beatriz M. Funatsu

A dissertation submitted to The Johns Hopkins University in conformity with the requirements for the degree of Doctor of Philosophy.

Baltimore, Maryland
August, 2005

© Beatriz M. Funatsu 2005
All rights reserved
Abstract

Deep convection is a key aspect of the tropical atmosphere. Local factors control its existence and strength, and these factors are in turn affected by large-scale processes. One example is found in the eastern tropical Pacific, where observations have suggested a link between deep convection and stratospheric intrusions into the tropical upper troposphere. Here we perform an analysis of the connection between these stratospheric intrusions and convection in this region, using a combination of data analysis, potential vorticity (PV) inversion diagnostics, and numerical simulations.

First a climatological study of intrusions with respect to its three-dimensional structure and accompanying outgoing longwave radiation (used as proxy for convection) is performed. This analysis shows that intrusion events have remarkably similar evolution and structure, and that transient convection nearly always occurs at the leading edge of the PV tongue. This confirms previous studies that have shown a close link between Rossby wave activity and transient convection.

PV inversion is then used to quantitatively assess the contribution of the upper level PV anomaly associated with the intrusion to changes in the dynamical and thermodynamical structure of the atmosphere. This establishes a formal link between the PV anomaly and quantities that characterize convection (convective available potential energy CAPE, vertical velocity, static stability). We verified that there is lower static stability and upward vertical motion at or around the leading edge of the intrusion, coincident with the area of
convection and of positive advection of PV, and consistent with proposed theory. Moreover, we determined that the upper-level PV intrusion has the dominant contribution to promote atmospheric destabilization, CAPE build-up and vertical velocity.

Finally, numerical model simulation are used to corroborate the above results (mechanisms), and also to examine the role of convection in the evolution of the PV anomaly. These simulations confirm that the PV intrusion is a key element for convection to occur. Also, analysis of simulations showed that here is a qualitatively good agreement with results from PV inversion diagnostics. Simulation results also show that the diabatic source of PV due to latent heat release plays a significant role in determining the shape of the intrusion.

Advisor: Prof. Darryn W. Waugh
Reader: Prof. Thomas W. N. Haine
Acknowledgements

My first thanks are to Prof. Darryn W. Waugh, my advisor. I benefited from his extreme generosity: He shared his expertise without constraints, dedicated time, invested in my training, and at the same time that he was critical he was also tactful, fair, objective. I thank him for giving me the opportunity to develop scientific skills and for doing everything possible to make my research go as smooth as possible. I thank him also for letting me ‘have a life’ while I pursued this Ph.D. degree. Prof. Waugh is certainly one of the people that I admire the most, both as a scientist, as well as a human being.

Prof. Thomas Haine is also a huge source of inspiration. I enjoyed very much his teaching, his criticism and objectivity. Like Prof. Waugh, he did not let me get away with anything. I thank him very much for the careful reading of this manuscript, for discussions on issues related and unrelated to this document, and for always being accessible regardless of the question I had when I knocked at the door of his office.

I acknowledge the National Center for Atmospheric Research (NCAR) for providing me with data and computational resources to conduct my research. I thank Mr. Bob Dat-tore (MSS), Ms. Cindy Bruyere (MM5), and SCD staff, for their kind help in very many occasions.

I thank Dr. Paul Newman (NASA) for the use of NCEP data from NASA, and Dr. Eric Ray for generously providing me a suite of programmes that helped me in this project.

I thank the Professors, Staff and Colleagues of the Johns Hopkins Dept. of Earth and Planetary Sciences. Being part of this department was an enriching experience in many levels. Many thanks to Profs. George Fisher and Peter Olson, and Drs. Edward Urban
and Katalin Szlavecz for the engaging conversations at lunch time, about scientific and non-scientific issues alike.

I also thank Dr. Manuel Alonso Gan (Brazilian National Institute for Space Research), Prof. Tercio Ambrizzi (University of São Paulo), and Prof. Amauri Oliveira (University of São Paulo), for their kind support.

My friends from near and from far were (and are) a huge source of comfort and sanity... Special thanks to Sarah Carmichael, Taber Hersum and Eyal Shalev (formerly Stanislavsky), and my lab-mates Hong Zhang, Ping-Ping Rong and Andrea Modod: They helped me much more than they can ever imagine!

Thanks to Elícia Inazawa, Rita Ynoue, Fernanda Ide, Katia Fernandes and Rodrigo Souza - although walking different paths I got strength from their friendship and common experiences.

Last, but not least... To my dear family: Hagay Amit, my best friend and better half, thanks from my heart for taking all the waves (my ups and downs) holding my hand tight, and for making me laugh. I thank my parents, my big sister Leila, my brother Sho, and my sweet nephew Daniel Jun, for their endless support, without complaints, without pressures, only cheering.

This work was supported by NSF.
Contents

Abstract ii
Acknowledgements iv
List of Figures viii
List of Tables xiv

1 Introduction 1

2 Three-dimensional structure and connection of PV intrusion with deep convection 5
   2.1 Introduction ................................ 5
   2.2 Data ...................................... 6
   2.3 Illustrative Examples ....................... 7
   2.4 Potential Vorticity .......................... 11
   2.5 Outgoing Longwave Radiation ............... 15
   2.6 Ozone .................................... 21
   2.7 Concluding Remarks ......................... 25

3 Potential Vorticity Inversion 28
   3.1 Introduction ................................ 28
   3.2 PV inversion theory and Attribution: Overview ........ 31
   3.3 Idealized PV anomaly ....................... 36
   3.4 Case Study 1: 12-17 January, 1987 ............. 41
       3.4.1 PV inversion ........................ 44
   3.5 Other case studies ......................... 53
   3.6 Summary .................................. 64

4 Numerical Model Simulations 66
   4.1 Introduction ................................ 66
   4.2 Model description ........................... 67
<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.3</td>
<td>Methodology</td>
<td>69</td>
</tr>
<tr>
<td>4.4</td>
<td>Control run: 13-17 Jan 1987</td>
<td>70</td>
</tr>
<tr>
<td>4.5</td>
<td>Sensitivity analysis</td>
<td>74</td>
</tr>
<tr>
<td>4.5.1</td>
<td>Initial and Boundary conditions</td>
<td>74</td>
</tr>
<tr>
<td>4.5.2</td>
<td>Initial condition time</td>
<td>74</td>
</tr>
<tr>
<td>4.5.3</td>
<td>Gridsize</td>
<td>75</td>
</tr>
<tr>
<td>4.5.4</td>
<td>Choice of cumulus parameterization</td>
<td>75</td>
</tr>
<tr>
<td>4.5.5</td>
<td>Summary</td>
<td>77</td>
</tr>
<tr>
<td>4.6</td>
<td>Factor Separation Analysis</td>
<td>78</td>
</tr>
<tr>
<td>4.7</td>
<td>Additional Cases</td>
<td>88</td>
</tr>
<tr>
<td>4.8</td>
<td>Latent Heat and PV evolution</td>
<td>93</td>
</tr>
<tr>
<td>4.8.1</td>
<td>Background</td>
<td>93</td>
</tr>
<tr>
<td>4.8.2</td>
<td>Results</td>
<td>96</td>
</tr>
<tr>
<td>4.9</td>
<td>Summary</td>
<td>107</td>
</tr>
<tr>
<td>5</td>
<td>Summary and Conclusions</td>
<td>110</td>
</tr>
<tr>
<td>References</td>
<td></td>
<td>115</td>
</tr>
<tr>
<td>A</td>
<td>PV inversion equations</td>
<td>123</td>
</tr>
<tr>
<td>B</td>
<td>Overview of the PSU/NCAR Mesoscale Modeling System MM5</td>
<td>126</td>
</tr>
<tr>
<td>Vita</td>
<td></td>
<td>131</td>
</tr>
</tbody>
</table>
List of Figures

1.1 20-year December-January-February climatology (1980-1999) of zonal wind ($\bar{u}$, in m.s$^{-1}$) at 350 K isentropic surface. Light shading for regions with $\bar{u} < 0$ (i.e., easterlies) and dark shading for $\bar{u} \geq 40$ m.s$^{-1}$ (showing the climatological position of the subtropical jet). .................. 2

1.2 The 6-14 day anomalies in OLR, 200-hPa streamfunction, and locally significant 200-hPa wind vector for a -1 standard deviation in OLR at 5-15°N, 140-130°W (shown by open box). The period used is December-February 1983/84-1987/88. The contour interval is $10 \times 10^5$, with zero contour omitted. Shading outlines regions of OLR anomaly less than -6 W.m$^{-2}$, hatching regions of OLR anomaly greater than +6 W.m$^{-2}$ (from Kiladis and Weickmann 1992a). ........................................ 3

2.1 Maps of PV on the 350 K surface (contours) and OLR (dark shading) over the northern tropical Pacific between 13 and 18 January 1987. PV contours are 1, 2, 4, 6, and 8 PVU, while shaded region shows OLR (south of 30°N) less than 180 W/m$^2$. Solid circle is Hilo, Hawaii. .................. 8

2.2 PV and O$_3$ distributions on 15 January 1987. (a) to (c) PV on the 330 K, 380 K, and 410 K surfaces (contour interval is 1, 1, and 2 PVU, with PV larger than 2, 4, and 8 PVU shaded, respectively). (d) Total ozone from TOMS (contour interval 25 DU, greater than 275 DU shaded). The solid curves in (e) and (f) show vertical profiles of PV and O$_3$ at Hilo (the dashed curves are the November-March mean values and the shaded regions the mean plus and minus one standard deviation). .................. 10

2.3 Maps of phase-shifted composite mean PV at (a) 330 K, (b) 350 K, (c) 370 K, and (d) 410 K at day 0 of intrusion events. Contour intervals are 0.5 PVU in (a) and 1 PVU in (b) to (d), with shading for PV greater than (a) 1, (b) 2, (c) 3, and (d) 7 PVU. Crosses mark the locations of the distributions shown in Fig. 2.4. ........................................ 13

2.4 Distributions of PV at 330 K (left), 350 K (middle), and 410 K (right) at 20°, 15°, and 10°N and reference longitude, for day 0 of intrusion events. Vertical dotted lines are mean values and horizontal solid lines are mean plus and minus one standard deviation. ............... 14
2.5 Shifted composites of PV on 350K, for the intrusions in (a) North Pacific, (b) North Atlantic, (c) South Pacific and (d) South Atlantic, based on the climatology of Waugh and Polvani (2000). ........................................ 16

2.6 Maps of phase-shifted composite mean 350K PV (thick contours) and OLR south of 30°N (light contours and shading) for days -3 to +2 of North Pacific events. Contour interval for PV is 1 PVU, and 20 W/m² for OLR (shaded region shows OLR less than 240 W/m²). All events have been shifted so the reference longitude is at 140°W. ......................... 17

2.7 Distribution of (a) minimum OLR and (b) longitude of minimum relative to reference longitude for intrusion events. ................................. 19

2.8 Longitude-time contour plot of OLR at 15°N for December 1986 through February 1987. Contour interval is 30 W/m² with values less than 210 W/m² shaded. The dashed lines show the region used to identify “OLR events”. ....................................................... 20

2.9 Maps of phase-shifted composite mean 350K PV (thick contours, contour interval is 1 PVU) and OLR (shaded regions) for days -3 to +2 of OLR events in 10°-20°N, 220°-230°E. Only OLR south of 30°N and less than 230 W/m² is shown. ............................................................... 22

2.10 Maps of phase-shifted composite mean total ozone (contour interval 10 DU, with shading for values larger than 260 PVU) and PV = 2 PVU at 350 K (thick contour) for days -3 to 2 of intrusion events. .......................... 24

3.1 Schematic representation of the proposed mechanism for the occurrence of convection, based on Dixon et al. (2003). The presence of an upper level PV trough causes a parcel of air on an isentropic surface to move northwestward and upward on its downstream side, providing conditions favorable to trigger convection. ................................. 30

3.2 (a) Vertical cross-section at 30°N of hypothetical PV distribution. PV anomaly with a maximum of \( q_0 = 9 \) PVU at a height of 5.5km. (b)-(d) shows the results by performing PV inversion using Eqs. (3.1) and (3.2). (b) Geopotential perturbation \( \Phi' \) at 900hPa. (c) Vertical cross-section of meridional velocity \( v' \) (m/s, solid) and potential temperature \( \theta \) (K, dotted). (d) Zonal (solid) and meridional (dotted) components of wind at 900hPa. . . . . . . . . 37

3.3 Variation of (a) maximum horizontal wind speed (m.s\(^{-1}\)) at 900hPa, (b) ratio between maximum horizontal wind speed at 900hPa and maximum horizontal wind speed at the level of the anomaly, (c) ratio between height where the maximum velocity is equal to 1/e of the maximum velocity at the level of the anomaly and the height of the (center of) anomaly, and (d) “static stability” in the 1000-850 hPa layer (deg/km), with strength of anomaly \( q_a \), at 30°N, 215°E. Solid line show variation for anomaly at \( z_c = 4\) km, dotted line for \( z_c = 5.5\) km, and dashed line \( z_c = 9\) km. Thick dark line in (d) represents the the background \( d\theta/dz \) in the 100-850 hPa layer. . . . . 39
3.4 Vertical profile of virtual temperature of “environment” (solid) and parcel (dashed), for (a) background; and (b) background+perturbation, for the conditions described in Fig. 3.2. Both profiles are at 30°N 215°E.

3.5 Potential vorticity (PVU, solid contours) at 200hPa, OLR (< 180 W.m\(^{-2}\), shaded), and vertical velocity \(\omega\) (Pa/s, > 0 thick black solid contours, < 0 thick dot-dashed contours), for the period 12-17 January 1987 12UTC. Contours of \(\omega\) were cropped at 30°N.

3.6 Vertical cross-section of \(\theta\) (K, thin lines), PV (PVU, thick lines) and vertical velocity \(\omega\) (Pa.s\(^{-1}\), > 0 thick black solid, < 0 thick dot-dashed) at (a) 25N and (b) 215E, on 15 January 1987 12UTC. Vertical coordinate is pressure (Pa), in log scale.

3.7 Vertical profile of virtual temperature of “environment” (NCEP data, solid) and parcel (dashed), on (a) 13 and (b) 15 January 1987. Both profiles are at 20°N 215°E.

3.8 Distribution of CAPE (thin) and 2PVU (thick) for 14-17 January 1987 12UTC. Parcels are lifted from 950 hPa and assumed to undergo pseudo-adiabatic ascent. CAPE contour intervals are 200 J.kg\(^{-1}\).

3.9 (a) NCEP reanalysis geopotential height (gpm, solid contours) and balanced geopotential field obtained by inverting PV (gpm, dotted contours) on 200hPa, (b) Same as (a) but for 850hPa.

3.10 Vertical velocity \(\omega\) at 500 hPa (Pa/s) given by: (a) Reanalysis data; (b) solving the Q-vector form of \(\omega\)-equation (Eq. (3.5), where \(Q\) is calculated using \((u, v, T)\) from reanalysis data; (c) idem to (b), except \(Q\) is calculated using balanced fields \((u, v, T)\) from PV inversion. Positive values (downward motion) in solid contours, negative values (upward motion) in dash-dot contours. Contour values ±[0.1, 0.2, 0.3, 0.5].

3.11 2PVU contour (thick) on 200 hPa and perturbation geopotential fields (in gpm) on 15 January 1987 obtained by inversion of total \(PV'\) (left column), and inversion of \(PV'\) in the upper atmosphere (400-100 hPa) (center column). Positive values are shown in solid contours, negative in dotted contours. Right column shows difference between the left and center columns. Pressure level where the perturbations are being displayed is shown in the left side of each row.

3.12 Time series of dry static stability (=\(-g\partial\theta/\partial p\), in K.m\(^2\).kg\(^{-1}\)) for the period 13-17 January 1987, for (a) 20°N 205°E and (b) 20°N 212.5°E, calculated using results from PV inversion. Solid line for total field, dashed line represents the contribution of \(q'_U\) (i.e., \(\overline{S} + S'_U\)).

3.13 Time series of CAPE for the period 13-17 January 1987, at 20°N 212.5°E. The thick solid curve is total CAPE, the thin dot curve corresponds to CAPE calculated using 5-day averaged mixing ratio profile, thin dash-dot curve is CAPE calculated using 5-day averaged temperature profile, and thick dashed curve is CAPE calculated using temperature associated with \(q'_U\).
3.14 Vertical velocity obtained by solving Eq. (3.5) using balanced \( \mathbf{v} \), \( T \) from PV inversion to calculate \( \mathbf{Q} \), left column; and \( \mathbf{\omega} \) associated with \( q'_U \). See section 3.2 for details of calculation. .......................................................... 52

3.15 Same as Fig. 3.5, except for the dates shown above each map. ........................................ 54

3.16 Vertical cross-section of \( \mathbf{\theta} \) (K, thin lines), PV (PVU, thick lines) and vertical velocity \( \mathbf{\omega} \) (Pa.s\(^{-1}\), > 0 thick black solid, < 0 thick dot-dashed for the dates and latitudes indicated on the top of each cross section .................................................. 55

3.17 2PVU contour (thick) and CAPE (thin) for parcels lifted from 950 hPa, for the dates indicated in the top of each map. Contour intervals 200 J.kg\(^{-1}\)........................................ 56

3.18 Same as in Fig. 3.12, but for the dates and gridpoints shown on the top of each plot. Notice that the y-scale for the first panel (February 1991) is lower than for the others. ................................................................. 58

3.19 CAPE (solid) and CAPE\(_U'\) (dashed) for the dates and gridpoints shown on the top of each plot .......................................................... 61

4.1 Potential vorticity (PV contours of 1, 2, 4 and 8, in units of \( 10^{-6} \text{m}^2 \text{K.kg}^{-1} \text{s}^{-1} \)) on 200 hPa, and OLR (< 210 W.m\(^{-2}\) shaded), for control run, for 14-17 January 1987 12UTC. ................................................................. 72

4.2 Potential vorticity (PV contours of 1, 2, 4 and 8, in units of \( 10^{-6} \text{m}^2 \text{Kkg}^{-1} \text{s}^{-1} \)) on 200 hPa, and OLR (shaded < 210 W.m\(^{-2}\) (left column), PV = 2PVU contour and CAPE (J.kg\(^{-1}\)), in intervals of 500 J.kg\(^{-1}\). Lowest value is 500 J.kg\(^{-1}\) (central column), and cross section at 20\(^\circ\)N of PV, potential temperature \( \mathbf{\theta} \) (K, intervals of 5K) and vertical velocity \( \mathbf{w} \) (cm.s\(^{-1}\), values greater than 2cm.s\(^{-1}\) shaded (right column), for (a) MM5 control run simulation, (b) NCEP/NOAA data, and (c) MM5 simulation with initial and boundary conditions using NCEP data. On 15 January 1987 12UTC. 73

4.3 PV at 200 hPa (1,2,4,8 PVU) and OLR (< 210W.m\(^{-2}\)shaded) for simulations (a) beginning at 12 January 1987 00 UTC, and (b) with twice horizontal resolution, i.e., 25 km. ................................................................. 75

4.4 PV at 200 hPa (1,2,4, 8 PVU) and OLR (< 210W.m\(^{-2}\)shaded) on 15 January 1987 12UTC (top row), and cross section at 20\(^\circ\)N of PV (1, 2, 4 PVU), \( \mathbf{\theta} \) (intervals of 5K) and vertical velocity \( \mathbf{w} \) (±2 cm.s\(^{-1}\), positive values shaded), for simulations identical to the control run, except using (a) Betts-Miller, (b) Anthes-Kuo and (c) Kain-Fritsch cumulus parameterizations instead of none. ................................................................. 77

4.5 Same as Fig. 4.4, except for runs UPV, LH and OTHR. ............................................ 79

4.6 Time sequence of area-averaged (a) OLR (W.m\(^{-2}\)), (b) dry static stability \( S \) (=g d\(\theta\)/dp, in K.m\(^2\)kg\(^{-1}\)), (c) CAPE (J.kg\(^{-1}\)), (d) vertical velocity \( w \) (=dz/dt, cm.s\(^{-1}\)), (e) latent heat flux (W.m\(^{-2}\)), and (f) \( \mathbf{\theta}_e \) (K). Results from MM5 simulation, including \( q'_U \) and latent heat release (solid line), including \( q'_U \) but no latent heat (dashed line), including latent heat but no \( q'_U \) (dotted line), and removing both \( q'_U \) and latent heat (dot-dashed line). ........... 81
4.7 (a) PV (1, 2, 4, 8 PVU) at 200hPa, and OLR (< 210 W.m⁻² shaded), for the dates indicates on the top of each panel. (b) 2 PVU contour and vertical velocity w (=dz/dt, 5, 10, 15 cm.s⁻¹, positive values shaded). (c) 2 PVU contour and CAPE (500, 1000, 1500 J.kg⁻¹), and (d) cross sections of PV (1, 2, 4 PVU) (thick solid), potential temperature (5K intervals) and relative humidity (≤ 80 % dark gray, ≤ 40% light gray), for dates: (1) 12 February 1991 12UTC, cross section at 17.5°N, (2) 23 January 1999 12UTC, cross section at 17.5°N, (3) 16 January 1997 12UTC, cross section at 17.5°N and (4) 28 January 2003 12UTC, cross section at 15°N.

4.8 Time sequence of area-averaged (a) OLR (W.m⁻²), (b) dry static stability $S (= g d\theta/dp, \text{in K.m}^2\text{.kg}^{-1})$, (c) CAPE (J.kg⁻¹), (d) vertical velocity w (=dz/dt, cm.s⁻¹), (e) latent heat flux (W.m⁻²), and (f) $\theta_e$ (K), for cases (i) February 1991, (ii) January 1999, (iii) January 1997, and (iv) January 2003. Results from MM5 simulation, including $q'_U$ and latent heat release (solid line), including $q'_U$ but no latent heat (dashed line), including latent heat but no $q'_U$ (dotted line), and removing both $q'_U$ and latent heat (dot-dashed line).

4.8 Continued.

4.8 Continued.

4.9 Potential Vorticity (1, 2, 4 and 8 PVU) at 200hPa and OLR (<210 W.m⁻²) given by control run (first row), and run with $q'_U$ only (no latent heat, second row), on (a) 14 January 1987 12UTC, (b) 16 January 1987 12UTC, and (c) 18 January 1987 00UTC.

4.10 Potential Vorticity at 200hPa: contribution of LH only to total PV (first row), and contribution of interaction of LH with PV (second row), for (a) 14 January 1987 12UTC, (b) 16 January 1987 12UTC, (c) 18 January 1987 00UTC. PV contours are ±0.1, 0.2, 0.4, 0.8 PVU for LH, and 1, 2, 4 and 8 PVU for INTR.

4.11 Cross-section at 20°N of potential vorticity (1, 2, 4 and 8 PVU) and relative humidity (40% light gray, and 80% dark gray) given by control run (first row), run with $q'_U$ only (no latent heat, second row). (a), (b) and (c) the same dates as in Fig. 4.9.

4.12 Cross-section at 20°N of contribution of LH to total PV (top row) and contribution of INTR (bottom row). PV contours ±0.1, 0.2, 0.4, 0.8 PVU for LH and ±1, 2, 4 and 8 PVU for INTR. (a), (b) and (c) the same dates as in Fig. 4.9.

4.13 Potential Vorticity (1, 2, 4 and 8 PVU) at 200hPa and OLR (<210 W.m⁻²) given by control run (first column), run $\nabla$PV only (no latent heat, second column), and the contribution of the interaction term to PV (third column, negative values shaded), on (a) 14 February 1991 00UTC, (b) 17 January 1997 12UTC, (c) 25 January 1999 00 UTC, and (d) 30 January 2003 00UTC.
4.14 Cross-section of potential vorticity (1, 2, 4 and 8 PVU) and relative humidity (40% light gray, and 80% dark gray) given by runs CNTL (first column) UPV (no latent heat, second column), and contribution of INTR to PV (third column, negative values shaded). Dates in (a)-(d) are the same as in Fig. 4.13, and the cross-sections are at: (a) 20°N, (b) 20°N, (c) 15°N, (d) 12.5°N. 106

B.1 Schematics of MM5 modeling system flow. . . . . . . . . . . . . . . . . . . . 127
B.2 Schematics of the direct interaction of parameterizations in MM5. . . . . . . 129
## List of Tables

<table>
<thead>
<tr>
<th>Section</th>
<th>Table Title</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1</td>
<td>Time variation of perturbation static stability $S'$ (in units of $10^{-5}$ K.m².Kg⁻¹.h⁻¹) averaged over 48h before 12UTC of the date indicated. $\partial S' / \partial t$ is the change associated with total anomaly $q'$, and $\partial S'_U / \partial t$ is the change attributable to $q'_U$.</td>
<td>59</td>
<td></td>
</tr>
<tr>
<td>3.2</td>
<td>Relative contribution of CAPE'$_U$ to total CAPE (in %) for the dates and locations indicated (same as in Fig. 3.19). (i) max[CAPE'$_U$]/CAPE, represents the ratio of the maximum CAPE'$_U$ in the time period ranging from two days prior to two days after the date indicated, to the total CAPE for this particular day. (ii) CAPE'$_U$/maxCAPE, is the contribution of CAPE'$_U$ when CAPE is maximum in this same time period. (iii) shows CAPE'$_U$/CAPE averaged over 36h prior to the date shown. (iv) CAPE'$_U$/CAPE for the date and location given. (v) shows CAPE'$_U$/CAPE averaged over the 9 nearest gridpoints (to the location given). For this last case, only values greater than zero were included in the calculation. The values in brackets are the standard deviation of the average.</td>
<td>60</td>
<td></td>
</tr>
<tr>
<td>3.3</td>
<td>Area-averaged $\omega'$ and $\omega'_U$ obtained by solving Eq. (3.5); see text for details. % indicates the relative contribution of $\omega'_U$ to $\omega'$ (the total vertical velocity perturbation). Since both $\omega'$ and $\omega'_U$ can be positive or negative, the relative contribution can exceed 100% or be negative. In the first case, it means that $\omega'_U$ actually contributes to the whole of the total vertical velocity in the same direction; in the second case, it acts in the opposite direction.</td>
<td>63</td>
<td></td>
</tr>
<tr>
<td>4.1</td>
<td>Absolute (second column) and relative contributions of factors and their interactions to OLR. OLR$<em>{cs}$ refers to the clear-sky value, and OLR$</em>{fac}$ the contribution due to each factor. Values are averages over the area 17.5-22.5°N, 215-220°E, on 15 January 1987 12 UTC.</td>
<td>83</td>
<td></td>
</tr>
<tr>
<td>4.2</td>
<td>Time variation of static stability $S$ (in units of $10^{-5}$ K.m².kg⁻¹.h⁻¹), averaged over the area 17.5-22.5°N, 210-215°E. The first column is the average contribution over 24h ending at 15 January 1987 00UTC, and the second (rightmost) column is the averaged contribution on 15 January 1987 for 00-12UTC.</td>
<td>84</td>
<td></td>
</tr>
</tbody>
</table>
4.3 Area-averaged (15-20°N, 207.5-212.5°E) contributions of factors to total (CNTL) CAPE, in J.kg\(^{-1}\). First column (maxCAPE\(_{fac}\)) shows the maximum CAPE attained by each factor between the period 13 January 00UTC to 17 January 00 UTC. The second column shows CAPE\(_{CNTL}\) when maximum CAPE due to each factor is found. Third column shows the contributions for the maximum CAPE\(_{CNTL}\) and the last column, the average of first 36 hours.

4.4 Vertical velocity for control run and due to \(q'_U\), latent heat, their interaction and to neither of those, averaged over the area 15-20°N, 210-215°E, on 15 January 1987, 12UTC.

4.5 Control values and relative contributions of upper level PV (\(q'_U\)), latent heat (LH), their interaction (INTR), and factors unrelated to any of them (OTHR) for OLR (W.m\(^{-2}\)), tendency of \(S\) (in units of 10\(^{-5}\) K.m\(^2\).kg\(^{-1}\).h\(^{-1}\)) for the period of 48 hours prior to the date on top, CAPE (J.kg\(^{-1}\)), and \(w\) (cm.s\(^{-1}\)) at 500 hPa. OLR (W.m\(^{-2}\)), CAPE (J.kg\(^{-1}\)) and \(w\) (cm.s\(^{-1}\)) are values on the dates indicated in top row.

4.6 Relative contribution (in %) of \(q'_U\) to the static stability tendency for 48h prior to convection, CAPE, and vertical velocity (=dz/dt) at 500hPa, for the dates indicated in the left column. The values between brackets are the time (UTC) for PV inversion and numerical simulations, respectively. Values marked with * correspond to the percentage relative to the total negative contributions.
To my siblings Leila M. F. Brambilla and Sidney Sho Funatsu,

for always taking my side.
Chapter 1

Introduction

Deep convection is a key aspect of tropical atmosphere. Latent heat release in deep convective regions in the tropics is a very important source of energy for the general circulation of the atmosphere (Hoerling 1992). Also, waves generated from extensive areas of convection can travel around the globe, causing changes in weather away from the source region. Locally, deep convection may be important in cross-tropopause mass exchange (e.g. Lamarque and Hess 1994), and in redistributing water vapor, ozone and other atmospheric constituents (e.g., Waugh 2005 and references therein).

Local factors such as temperature, humidity and wind profiles exerts primary control on the existence and strength of deep convection. Those factors are in turn modified by large-scale processes, such as large scale low-level convergence, or destabilization through quasi-geostrophic motion. In the midlatitudes, this is generally the case, with deep convection being “forced” by quasi-geostrophic motions (Raymond 2001). In the deep tropics, convection is basically controlled by changes in heat and moisture surface fluxes in the boundary layer, which affects the convective available potential energy (CAPE). In the subtropics, the picture is more blurred, with both mid-latitude and tropical effects playing a role.
Figure 1.1: 20-year December-January-February climatology (1980-1999) of zonal wind ($\bar{u}$, in m.s$^{-1}$) at 350 K isentropic surface. Light shading for regions with $\bar{u} < 0$ (i.e., easterlies) and dark shading for $\bar{u} \geq 40$ m.s$^{-1}$ (showing the climatological position of the subtropical jet).

An example of a region where tropical convection may be affected by both mid-latitude and tropical effects is the eastern tropical Pacific. In a series of papers, Kiladis and Weickman (1992a,b, 1997) Tomas and Webster (1994), Kiladis (1998) and others showed that Rossby wave-type of disturbances with periods between 6-30 days propagate from the extratropics to the tropics in this region, through a window of westerlies within the equatorial easterlies during the boreal winter (Fig. 1.1). These equatorward propagating waves have a significant contribution to the momentum balance in the tropical eastern Pacific region (Kiladis and Feldstein 1994, Kiladis 1998), acting to slow the equatorial westerlies (Kiladis 1998).

The above studies showed that the path, velocity, and energy dispersion of these waves
Figure 1.2: The 6-14 day anomalies in OLR, 200-hPa streamfunction, and locally significant 200-hPa wind vector for a -1 standard deviation in OLR at 5-15°N, 140-130°W (shown by open box). The period used is December-February 1983/84-1987/88. The contour interval is $10 \times 10^5$, with zero contour omitted. Shading outlines regions of OLR anomaly less than -6 W.m$^{-2}$, hatching regions of OLR anomaly greater than +6W.m$^{-2}$ (from Kiladis and Weickmann 1992a).

agree well with those predicted by the linear theory for Rossby wave propagation (Webster and Holton 1982, Hoskins and Ambrizzi 1993). If these Rossby waves are of sufficient large amplitude “wave breaking” can occur, influencing the distribution of tracer constituents in the upper troposphere (e.g., Scott et al. 2001). Moreover, circulation anomalies associated with these disturbances were found to correlate well with outgoing longwave radiation (OLR) anomalies centered in the central and eastern Pacific, e.g., Fig. 1.2. Kiladis and Weickmann (1992b) and Kiladis (1998) hypothesized that the link between intrusion and deep convection happens through advection of positive potential vorticity (PV) downstream, coincident with areas of decreased static stability and enhanced vertical motion.

Waugh and Polvani (2000) formed a climatology of these Rossby-wave type of intrusions events based on PV evolution on 350K insentropic surface and showed that these events are more frequent during the boreal winter. Also, they found that there is large interannual variability in the occurrence of intrusions. This variability is highly correlated
with the phase of ENSO, with fewer intrusions in the warm phase (El Niño). For example, in the northern hemisphere eastern tropical Pacific, intrusion events occur predominantly within the westerly ducts, and their frequency is much lesser in years of El Niño, when the westerlies decrease from about 20 m.s\(^{-1}\) to 5 m.s\(^{-1}\) or less.

The above studies have shown a strong relationship between Rossby wave activity propagating into the tropical Pacific and deep convection, and have presented examples where the convection occurs ahead of an intruding tongue of high PV. There are however a range of issues that need to be addressed for a better understanding of these systems and their impact on transport and deep convection. For example, what is the exact link between PV intrusions and deep convection? Are all convective events linked to intrusions and vice versa. If a link exist, what is the “cause and effect”? Finally, is there any feedback between convection and intrusions? We will address the above issues here.

The layout of this document is as follow: In Chapter 2 we investigate the climatological three-dimensional structure of the intrusions, and the accompanying OLR and O\(_3\) structures. The quantitative assessment of the contribution of the upper-level PV intrusion to key quantities related to convection (CAPE, static stability and vertical velocity) is addressed in Chapter 3, using PV invertibility concepts and diagnostics. In Chapter 4 we use a mesoscale model to confirm the results obtained by PV inversion and also to examine the impact of convection to the evolution of the intrusion through latent heat release. A summary and discussion of the results, as well as unresolved issues to be addressed in the future, are presented in Chapter 5.
Chapter 2

Three-dimensional structure and connection of PV intrusion with deep convection

2.1 Introduction

In this Chapter we reproduce part of the article published in the Journal of Atmospheric Sciences (Waugh and Funatsu, 2003). In this paper, we extended the study of Waugh and Polvani (2000; hereafter WP2000), who performed a climatological analysis of deep tropical intrusions. They found that these intrusions occur more frequently in the northern hemisphere winter, in both northern and southern Pacific and Atlantic Oceans. Moreover, they found that the events in the Pacific occur almost exclusively within the westerly ducts and there is strong dependency on the background flow. Consequently, there is strong interannual variability, with events more frequent in La Niña years than in El Niño years. Here, we examine the spatial structure of the intrusions over the northern Pacific Ocean in terms of both PV and ozone (O$_3$), and also the relationship of intrusions with transient tropical
convection as diagnosed by outgoing longwave radiation (OLR) measurements. In the next section we describe the various data sets used in our analysis. Example intrusion events that illustrate the three-dimensional PV structure of an intrusion and the associated changes in convection and ozone are presented in Section 2.3. The climatological PV structure is examined in Section 2.4, while the accompanying OLR field is examined in Section 2.5. The spatial distribution of total column ozone, as observed by the Total Ozone Mapping Spectrometer (TOMS) during intrusion events is examined in Section 2.6. Concluding remarks are in Section 2.7.

2.2 Data

Potential vorticity (PV) calculated from National Center for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) reanalysis data (Kalnay et al. 1996) was used to examine the structure of the intrusions. These data have 2.5° latitude by 2.5° longitude resolution and are available 6-hourly. In our analysis we focus on the PV distribution on isentropic surfaces between 330 K (≈ 400 hPa in the tropics) to 410 K (≈ 80 hPa), and use 12Z data from 1980 to 1999.

These data are the same as used by WP2000 to form their climatology of the occurrence of intrusions, and we use the events defined in the WP2000 climatology to examine the mean structure of the intrusions. WP2000 defined their climatology by first identifying the occurrence of high PV (> 2 PVU; 1 PVU = 10^{-6} K s^2/kg) at 10°N or 10°S. Then occurrences within 10° longitude or within 6 days were grouped into single intrusion events. Below, the middle day when PV at 10°N or S exceeded the critical value is defined as “day 0” of the intrusion event (and “day -2” corresponds to 2 days before this day).

Another check on the reality of features in the analysed PV is comparisons with O₃ observations. A high correlation between PV and O₃ is expected in the upper troposphere
and lower stratosphere, and similar features should be seen in PV and O₃ fields, i.e., high O₃ when there is high PV. Unfortunately three dimensional O₃ data are not available for the region of interest. However, near-global measurements of the total column ozone are made by the Total Ozone Mapping Spectrometer (TOMS) satellite instruments (http://toms.gsfc.nasa.gov). Three different TOMS instruments were operational at different periods during the 20 years of interest. In our analysis we use total ozone data from Nimbus 7 (1 January 1980 to 6 May 1993), Meteor 3 (7 May 1993 to 24 November 1994) and Earth Probe TOMS (25 July 1996 to 31 December 1999). There is no TOMS data between 24 November 1994 and 25 July 1996. The TOMS data used has 1° latitude by 1.25° longitude resolution.

The occurrence of convection near the intrusions is examined using outgoing long-wave radiation (OLR) as a proxy for tropical convection. We use the NOAA spatially and temporally interpolated OLR (Liebmann and Smith, 1996), which were obtained from the NOAA-CIRES Climate Diagnostics Center (http://www.cdc.noaa.gov/). These data are available daily and on the same horizontal grid as the NCAR/NCEP reanalysis data, and we use the same 20-year period as the PV data (1980 to 1999).

### 2.3 Illustrative Examples

Before considering the climatological structure of the intrusions we first consider two illustrative examples. The first is an intrusion event that occurred over the northern Pacific in the middle of January 1987. This event is actually not counted in the WP2000 climatology as the PV at 10°N never exceeds 2 PVU during the event. However, the occurrence of tropical convection and relationship with meteorological fields during this event was examined by Kiladis and Weickmann (1992b) and ozone measurements from Hilo were made on January 15 when the high-PV tongue was overhead. The detailed analysis of Kiladis and
Weickmann (1992b) and the availability of ozone data means that this event can be used not only to illustrate the PV structure but also the connection with convection and ozone.

Figure 2.1 shows the evolution of the PV on the 350 K surface and (low) OLR over the northern tropical Pacific between 13 to 18 January 1987. These maps show that there were undulations in the PV contours within the region of steep meridional PV gradients (the tropopause) which amplified over the 13 and 14 January. As they amplified, these

Figure 2.1: Maps of PV on the 350 K surface (contours) and OLR (dark shading) over the northern tropical Pacific between 13 and 18 January 1987. PV contours are 1, 2, 4, 6, and 8 PVU, while shaded region shows OLR (south of 30°N) less than 180 W/m². Solid circle is Hilo, Hawaii.
disturbances (Rossby waves) propagated eastward and produced a tongue of high PV which intruded into the tropics (15 and 16 January). This tongue is relatively narrow and has an almost north-south orientation (there is only a slight SW-NE tilt). Over the next two days the tongue decayed, and the whole lifecycle (amplification and decay) of the tongue lasted around 6 days. As shown below, the above sequence of events is typical of the intrusions examined, and the above features can be seen in PV maps for other events, e.g., figures in Hsu et al. (1990), Tomas and Webster (1994), Numaguti (1995), and WP2000.

The maps in Fig. 2.1 also show changes in the OLR during this period. In particular, a region of low OLR, and presumably deep convection, occurred just ahead of the high-PV tongue. The region of low subtropical OLR appears around 15 January and lasts around 4 days. As mentioned above, Kiladis and Weickmann (1992b) performed a detailed analysis of the meteorology and OLR during this event. They showed that the region of low OLR ahead of the tongue is co-located with a region of upward motion at 500 hPa, and argued that this ascent (and reduced static stability) could be attributed to the anomalous PV within the intruded tongue. The link between low OLR and intrusions is examined further in Section 5.

The vertical structure of the intrusion on 15 January is shown in Fig. 2.2. Figure 2.2(a)-(c) show that a tongue of high PV intrudes into the tropics at all levels from 330 K (middle troposphere) to 410 K (lower stratosphere). The width of the tongue increases slightly with height, but the location and NW-SE tilt varies little between 330 and 410 K. The anomalous nature of the PV in the tongue can be seen by comparing the vertical profile of PV over Hilo (solid circle in panels (a)-(d)) on 15 January with the climatological winter values, see Fig. 2.2(e) (the shaded region shows the climatological mean plus and minus the standard deviation of daily values). The values on 15 January are around or larger than the climatological mean plus one standard deviation from 700 hPa to over 70 hPa, with very large values between 300 and 200 hPa (340-350 K).
Figure 2.2: PV and O₃ distributions on 15 January 1987. (a) to (c) PV on the 330 K, 380 K, and 410 K surfaces (contour interval is 1, 1, and 2 PVU, with PV larger than 2, 4, and 8 PVU shaded, respectively). (d) Total ozone from TOMS (contour interval 25 DU, greater than 275 DU shaded). The solid curves in (e) and (f) show vertical profiles of PV and O₃ at Hilo (the dashed curves are the November-March mean values and the shaded regions the mean plus and minus one standard deviation).
The impact of this intrusion event on the ozone distribution is shown in Fig. 2.2(d) and (f), which show total column ozone from TOMS and the vertical profile of ozone above Hilo, respectively. There is a strong signature in the both the horizontal and vertical distributions of ozone. The ozone above Hilo is larger than the mean throughout the troposphere, with anomalously large values from below 200 hPa to above 100 hPa, while the total column ozone in the tongue near Hilo is over 300 DU (compared with the climatological mean January value of around 255 DU). The similarities in the structures in the O₃ and PV fields, and in particular the near north-south orientation in both the tongues of total column ozone and PV, gives us confidence in the reality of the analysed PV.

The above examples show that intrusion events can produce tongues of high PV and ozone which extend from the middle troposphere to lower stratosphere, and simultaneously deep convection (as diagnosed by low OLR) can occur ahead of the high PV/ozone tongue. In the following sections we examine whether these are climatological features of the intrusion events.

### 2.4 Potential Vorticity

We now examine the climatological PV structure of the intrusion events in the North Pacific region. All days when there is an intrusion, as defined by WP2000 (see Section 2), in the North Pacific are grouped together, and the mean PV field (“composite-mean” PV) is calculated. The temporal evolution of the composite-mean PV is examined by repeating this analysis for days -3 to 3, i.e., 3 days prior to and 3 days after the peak of the intrusion. Although we formed composites for just north Pacific events there is still a large variation in the longitude of the high-PV tongues, and the averaging of all events removed a lot of the structure. We therefore formed composites in which the PV fields were shifted in longitude so that the PV tongue at day 0 occurred at the same notional longitude (140°W), i.e., the
PV fields were shifted so that events are “in phase”. All results presented below are from these “phase-shifted” composites.

Figure 2.6 shows maps of 350K PV for the phase-shifted composite of events in the North Pacific, for day -3 to day 2. Also shown is the composite OLR, which will be discussed in the next Section. The evolution and structure of the composite PV is very similar to that of the PV in the individual event shown in Fig. 2.1. A disturbance to PV contours near the tropopause propagates eastward and amplifies, producing a tongue of high PV that extends south of 10°N. As in the January 1987 event the high-PV tongue in the composite-mean has almost north-south orientation and lasts around 3 days. The gradients of PV and maximum PV in the tongue are weaker than in the individual event shown in Fig. 2.1. However, considering the composite mean is the average of 103 events, the similarity between the composite-mean and individual events suggests that the vast majority of the intrusions have very similar evolution and structure. The variability between events is examined below.

The mean vertical structure of the intrusions is examined by forming composite PV fields on several other isentropic surfaces, for the same days used to form the 350 K composites. Figure 2.3 shows the composite PV at day 0 for levels between 330 to 410 K. At all levels there is a tongue of high PV reaching to lower latitudes at roughly the same longitude. The strength of the tongue weakens as you move up or down from 350-370 K, and is relatively weak at the lowest and highest levels shown. For example, at 350 K the PV contours which are normally around 30°N (PV ≈ 2 PVU) reach as far south as 10°N, whereas at 330 and 410 K the southward movement of contours around 30°N is only a few degrees latitude.

To examine the variability between different individual events we examine the distribution at different locations of the PV at day 0. Figure 2.4 shows the distributions of PV at 330 K, 350 K and 410 K for the reference longitude and three different latitudes (10°, 15°, 12°...
and 20°N), see crosses in Fig. 2.3.

Consider first the distributions at 350 K (middle column in Fig. 2.4). At 10°N all the PV values are equal to or greater than 2 PVU (as required to be included in the WP2000 climatology), and the distribution is very narrow with PV between 2 and 2.5 PVU in all but a few cases. More symmetric distributions are found at more northerly points, and also east and west of the reference longitude (not shown). However the spread at 15°N is still relatively small. This indicates that there are rather small variations in PV between different intrusion events, i.e., intrusions have very similar spatial structure. Note that the mean value at 15°N is actually greater than at 20°N. This is because of the NE-SW tilt of the intrusions (see Fig. 2.3), and at 20°N the largest PV values are east of the reference longitude (the mean PV 5° east of the reference longitude is larger by 1 PVU).

Figure 2.3: Maps of phase-shifted composite mean PV at (a) 330 K, (b) 350 K, (c) 370 K, and (d) 410 K at day 0 of intrusion events. Contour intervals are 0.5 PVU in (a) and 1 PVU in (b) to (d), with shading for PV greater than (a) 1, (b) 2, (c) 3, and (d) 7 PVU. Crosses mark the locations of the distributions shown in Fig. 2.4.
Figure 2.4: Distributions of PV at 330 K (left), 350 K (middle), and 410 K (right) at 20°, 15°, and 10°N and reference longitude, for day 0 of intrusion events. Vertical dotted lines are mean values and horizontal solid lines are mean plus and minus one standard deviation.

The distributions at 410 K (right column in Fig. 2.4) are also symmetric with a relatively small spread about the mean value. The mean values increases from around 5 to 8 PVU from 10°N to 20°N (compared to the DJF mean values of 3 to 6 PVU for 10°N to 20°N at 220°E), while the standard deviation is around 1.5 PVU for all latitudes. This shows there are only small variations in 410 K PV between different events, and that most events having higher than normal PV at 410 K.

At 330 K (left column in Fig. 2.4) the distributions are somewhat different. In nearly
all cases the PV is less than 1 PVU, and only in a few cases is the PV greater than 1.5 PVU at, or south of, 20°N. This indicates that in only a few of the intrusions is there a deep downward signal in the PV. In other words, the strong signal in 330 K PV seen in the January 1987 event (Fig. 2.2) is not seen in most intrusion events. Note that even though there is not a strong signal in the analyzed PV there may be finer scale features, as seen in high-resolution trajectory calculations (e.g., Scott et al., 2001), that are not resolved in the analyzed PV. Visual inspection of the PV for events with strong signature at 330 K (i.e., PV ≥ 2 PVU at 20°N) shows that for these events the tongue at 350 K is narrower, with stronger gradients, and more north-south orientation than the tongue in the composite-mean field.

Examination of phase-shifted composites for events in the South Pacific and the North and South Atlantic regions show very similar features to the North Pacific region (Fig. 2.5). The climatological mean PV fields show a well defined high-PV tongue with almost north-south orientation, and there is only small event-to-event variability. However, the southern hemisphere events are noticeably weaker than the northern hemisphere events.

2.5 Outgoing Longwave Radiation

As discussed in the Introduction, several studies have shown that there is a link between Rossby waves propagating into the tropics and transient deep convection in the eastern tropical Pacific (e.g., Kiladis and Weickmann 1992a,b, Kiladis 1998, Slingo 1998, Matthews and Kiladis 1999). Kiladis and Weickmann (1992b) examined the January 1987 event discussed in Section 3 and showed that convection, as diagnosed by OLR, occurred ahead of an intruding tongue of high PV. They further suggested that the ascent and reduced static stability ahead the PV tongue initiated/supported the convection (see also Kiladis, 1998). However, it is not clear whether all convective events are linked with tongues of PV, and
whether there is convection ahead of all tongues of PV. We examine these issues by examining the OLR for the intrusion events examined in the previous section.

We form phase-shifted composites of the OLR in exactly the same manner as the PV composites in the last section. Fig. 2.6 shows the composite-mean OLR for north Pacific intrusions events, with values less than 240 W/m² and south of 30°N shaded. As in Figure 1, the region of low OLR occurs at the leading edge of the tongue of high PV, and appears around the peak of the intrusion event. This signature of low OLR in the composite fields supports the hypothesis of Kiladis and Weickmann (1992b) and Kiladis (1998) that the
ascent and reduced static stability ahead the PV tongue will initiate/support convection.

The value in the composite OLR is not as low as in the January 1987 example shown in Fig. 2.1 (the minimum value in the composite is around 220 W/m² compared to less than 160 W/m² on 15 January 1987). However, examination of individual events shows that there is generally low OLR around the leading edge of the tongue but that there is a lot of variability in the exact location of the low OLR. Because of this spatial variability
the minimum value in the composite mean is significantly higher than the minimum in individual events.

To examine the variations in low OLR we determine, for each intrusion event, the minimum OLR in the vicinity of the PV tongue. Figure 2.7 shows the distributions of minimum OLR and the longitude of this minimum relative to the reference longitude. These plots show that nearly all events have a region of OLR much lower than in the composite mean (70% of the events have a region of OLR less than 170 W/m²), and that the location of low OLR is nearly always ahead of the tongue (longitude is around 10° to 12.5° ahead of the reference point). We have repeated this analysis using the average OLR in 5° latitude by longitude boxes, and the distributions of minimum OLR and location are similar.

The above indicates that when there is an intrusion (tongue) of high PV there is nearly always low OLR, and hence deep convection. But does similar convection occur when there are no such PV events? To examine this we define events where the OLR is anomalously low (“low-OLR” events), and form phase-shifted composites of both the OLR and PV for these events.

To define low OLR events we first examine the variability of OLR in the eastern tropical Pacific. Figure 2.8 shows a longitude-time plot OLR at 15°N from 1 December 1986 to 28 February 1987, with values less than 210 W/m² shaded. This shows that there were several brief periods during this winter when there was very low OLR (e.g., below 210 W/m²) and that these occurred between 100° and 150°W. Note that one of these periods (15-17 January) is associated with the intrusion event shown in Fig. 2.1. Plots for other years show similar temporal variability, with brief periods with OLR much lower than the winter mean values (see also figure 6 of Matthews and Kiladis (1999) for OLR time series at 130°-140°W for several winters). For our analysis we define “OLR events” as periods when the average OLR within the region of interest is less than 210 W/m² (as in the PV intrusion climatology, consecutive days are grouped together as one single event). In the analysis
Figure 2.7: Distribution of (a) minimum OLR and (b) longitude of minimum relative to reference longitude for intrusion events.
below we focus on OLR events in the region 10°-20°N and 130°-140°W. For December 1986 to February 1987 there were 6 events in the region (see dashed lines in Fig. 2.8), but there is large interannual variability in the number events, with the number varying between 0 and over 10 (with fewer events during the warm phase of ENSO). See Matthews and Kiladis (1999) for further discussion on the interannual variability of convection in the eastern tropical Pacific.

Figure 2.8: Longitude-time contour plot of OLR at 15°N for December 1986 through February 1987. Contour interval is 30 W/m² with values less than 210 W/m² shaded. The dashed lines show the region used to identify “OLR events”.

Using the above criteria applied to 20 years of DJF data we form composite-mean OLR and PV fields as in Section 3, except here the events are defined on low OLR rather than high PV. Figure 2.9 shows maps of the composite OLR and 350K PV for days -3 to 2 of the low OLR events in the region 10°-20°N, 130°-140°W. (Similar maps for composites of low OLR events in different, nearby, regions show very similar features.) These maps
show a similar evolution of PV and OLR as the composite maps for PV events (Fig. 2.6). A trough in the PV contours forms to the north west of the reference region (with low OLR) around 2 to 3 days before the reference date; the minimum OLR occurs around the same time as the maximum perturbation to the PV contours; and the low OLR is ahead (east) of the PV trough. As might be expected, the minimum in OLR is better defined (lower) and the tongue of high PV less defined (weaker) in the composite based on OLR events than the composites for the PV events. However, both composites indicate a strong connection between high PV at 350K and low OLR.

As for the case of the composite-mean OLR for the high PV events, the structure and magnitude of the composite-mean PV for low OLR events is reduced because of the averaging over a large number of events. Analysis of the distribution of maximum PV around the reference region shows that in nearly all of the low OLR events there is a tongue of high PV west (upstream) of the region of low OLR, and, consistent with the analysis of OLR associated with PV events, the mean shift is around 10° longitude. Note that in the majority of these tongues the PV = 2 PVU contour does not reach as far south as 10°N, and hence these days are not included in the WP2000 intrusion climatology (and the above analysis of PV events). However, this analysis still shows that low OLR, in the northeastern tropical Pacific, usually occurs in the vicinity of a tongue of anomalously high PV in the UT.

2.6 Ozone

As discussed in Section 2 the impact of intrusions on ozone is examined using ozonesonde measurements at Hilo (19°N, 205°E) and total ozone from the TOMS satellite instrument. Changes in the vertical structure of ozone over Hilo are discussed in detail in Waugh and Funatsu (2003). They showed that there is large variability in O₃ in UT/LS over Hilo, and that the variations are well correlated with variations in the analyzed PV. Large values
Figure 2.9: Maps of phase-shifted composite mean 350K PV (thick contours, contour interval is 1 PVU) and OLR (shaded regions) for days -3 to +2 of OLR events in 10°-20°N, 220°-230°E. Only OLR south of 30°N and less than 230 W/m² is shown.
of O₃ (and PV) above the mean winter values are associated with intrusion events, and the occurrence of a tongue of high PV over Hilo. Also, very low values of O₃ can be associated with intrusion events: the air ahead and behind the high-PV tongues can come from equatorial regions, and have low ozone (see e.g. their Fig. 12).

In this section we discuss only the spatial and temporal variations in total column ozone, using TOMS (Total Ozone Measurement Sonde) data. The ozonesonde data provides high vertical resolution, but has limited temporal resolution and are available from only a few locations. It is therefore not possible to examine the daily variability of ozone or its horizontal structure from these data. However, information on daily and spatial structure of total ozone is available from TOMS (and other satellite) instruments.

We form phase-shifted composites of TOMS total ozone for days in the WP2000 intrusion climatology in the same manner as the PV and OLR composites discussed above. Figure 2.10 shows the phase-shifted composite-mean total ozone for days -3 to 2 of north Pacific events (also shown is the composite-mean PV = 2 PVU contour at 350 K). There is some signature in the total ozone composite (i.e., southward undulation of the 260 and 270 DU contours on days -1 and 0), but it is weak. Although the weak signal in total ozone could be due to averaging between events with a strong signal but differing spatial structure, visual inspection of total ozone maps shows that most intrusion events have only a weak signal in total ozone.

Although most events have a weak signal, the analysis of the 15 January 1987 event (Fig. 2.2) showed a strong signature in total ozone can occur in individual events (with total ozone as large as 300 DU at 20°N). One difference between the 15 January 1987 event and most intrusion events is the deep downward extent of this event (i.e., as discussed in Section 4 most events don’t have a strong PV signal at 330 K). This suggests that a large change in total ozone may occur only in “deep” intrusion events with large PV at 330 K. This is supported by examination of the total ozone for the events with a strong signature
Figure 2.10: Maps of phase-shifted composite mean total ozone (contour interval 10 DU, with shading for values larger than 260 PVU) and PV = 2 PVU at 350 K (thick contour) for days -3 to 2 of intrusion events.
in PV at 330 K (PV ≥ 2 PVU at 20°N) which shows a tongue of high total ozone for each event (not shown).

2.7 Concluding Remarks

The analysis of 20 years of NCEP/NCAR potential vorticity (PV) data shows that intrusion events over the northern tropical Pacific have remarkably similar evolution and structure between 340 and 370 K, with all events producing narrow tongues of high PV that have an almost north-south orientation and last 2 to 3 days. This can be contrasted with tropopause folds occurring in middle latitudes where there is large variability between events and the tongues of high PV are seldom aligned north to south. The reason why the subtropical intrusions have near north-south alignment is unclear, but may be related to the very weak meridional gradients in the zonal wind within the westerly ducts.

The intrusions can extend from the middle troposphere (330K) to lower stratosphere (410K). The upward extension into lower stratosphere is a robust feature and occurs in the vast majority of events, but deep downward penetration occurs only in a small percentage of the events. The extension of the events into the lower stratosphere may play an important role in determining the chemical composition in the tropical lower stratosphere. Analysis of trace gas observations indicates that while there is a subtropical barrier to transport into the tropical stratosphere it is not a perfect barrier and there some mixing into the tropics (e.g., Avallone and Prather 1996, Hall and Waugh 1997, Minschwaner et al. 1996, Volk et al. 1996). The intrusion events examined here appear to be one processes by which this transport occurs (see Horinouchi et al. (2000) for examination of this stratospheric transport in a general circulation model).

Analysis of the relationship between OLR (as a proxy for deep convection) and intrusions shows that transient convection and tongues of high PV in the UT nearly always occur
together. This is consistent with the previous studies that have shown a close link between Rossby wave activity and transient convection (e.g., Kiladis and Weickmann 1992a,b, Kiladis 1998, Slingo 1998, Matthews and Kiladis 1999). It further shows that the relevant Rossby waves are of large amplitude and produce “anomalous” tongues of PV. This tight connection between convection and tongues of PV, with the convection at the leading edge of the PV tongue, adds support to the hypothesis of Kiladis and Weickmann (1992b) that the anomalous PV within the intruded tongue produces a region of ascent and reduced static stability that initiates the convection.

Although the analysis presented here has provided useful information of the structure of PV, O₃ and OLR during intrusion events, there are still several areas that need further investigation. The analysis performed here has confirmed that close link between intrusions and deep convection, but it does not address “cause and effect”. This issue is addressed in the following two chapters.

Also, our analysis has shown a link between the PV and O₃ distributions during intrusions but has not addressed the issue of the amount of irreversible transport of O₃ caused by the intrusion events. The lack of high-resolution three-dimensional data means that this issue will need to be examined using high resolution models (e.g., Scott et al., 2001; Scott and Cammas 2002), in combination with available data. Also the impact of intrusion events on other trace constituents, e.g., water vapor, needs to be examined. As the frequency of the intrusions varies with the phase of ENSO it is possible that the intrusions play a role in the observed ENSO-related variations in UT humidity (e.g., Newell et al., 1997, Bates et al. 2001).

It is interesting to note that there may be some coupling between the convection and irreversible transport of trace constituents. The diabatic heating associated with the convection may play a role in the weakening of the PV anomaly (as noted by Kiladis (1998)), and the mixing of stratospheric and tropospheric air. So it is possible that the PV tongue
may induce convection that leads to its own destruction and mixing of different air masses. Therefore further investigations of transport and convection connected with intrusion events should be coupled together.
Chapter 3

Potential Vorticity Inversion

3.1 Introduction

In the previous Chapter it was shown that the connection of transient convection and intrusions of high-PV tongues into the tropics is a robust feature, with the convection occurring at the leading edge of the PV intrusions. However, this connection between PV and convection says nothing about causality.

To examine the causality we now make a more detailed assessment of the evolution of various variables such as static stability and the convective available potential energy (CAPE) during the intrusion event. Causality can only be definitely established in linear systems. Atmospheric systems are chaotic and therefore allow multiple stable equilibria. However, in some cases, for relatively short-lived phenomena, its evolution may be weakly non-linear, in which case the issue of causality may be addressed. Intrusion events fall in this category. Although the evolution of the system is not purely linear (Rossby wave breaking and irreversible mixing may occur), the non-linearity is weak, especially prior to convection.

The existing theory proposes that “decreased static stability and enhanced upward
motion occur in the area of positive potential vorticity advection, ahead of the intrusive trough” (Hoskins et al. 1985 (HMR), Kiladis and Weickmann 1992b). HMR showed that the effect of a PV disturbance reflects not only locally in the thermal and dynamical fields, but throughout a surrounding area. Changes in the vertical velocity arise in response to the adjustment of the atmosphere to the presence of the PV anomaly. A helpful analogy of this effect is the particle-based theory of electrostatics, in which the stream function, vertical velocity, etc. are “attributed” to a particular PV anomaly in the same way a charged particle affects the electrical field by “action-at-a-distance” (HMR, Bishop and Thorpe 1994, Clough et al. 1996).

HMR and Thorpe (1995) showed that a positive (cyclonic) upper-level PV has a less stable potential temperature distribution within and immediately below the anomaly, while it is more stable in the layers above. The distortion of the temperature field as it adjusts to the cyclonic anomaly and the translational motion of the anomaly itself results in vertical motion in low levels. Physically, the vertical velocity can be understood as a sum of two effects. First, there is a contribution proportional to the local tendency of buoyancy, as a particle moves up or down as the isentropes adjust to the presence of the anomaly. This is referred by Hoskins et al. (2003) as the isentropic displacement. Second, if we consider now a frame of reference moving with the system, there will be a contribution from the “advection” of buoyancy, due to the horizontal motion of the particle moving along an isentropic surface (called isentropic upglide). The isentropic upglide will cause a particle moving along an isentropic surface to go upwards, northwards and westward, in a reference frame where the isentropic displacement can be discarded (Dixon et al., 2003). The contribution of vertical velocity of the basic state will lead to an upward and northward movement ahead of the intrusion (due to the inclination of the basic state isentrope) and the perturbation vertical velocity will lead to an upward movement in the E-W direction due to the distortion of the isentropic surface caused by the presence of the anomaly. This mecha-
Figure 3.1: Schematic representation of the proposed mechanism for the occurrence of convection, based on Dixon et al. (2003). The presence of an upper level PV trough causes a parcel of air on an isentropic surface to move northwestward and upward on its downstream side, providing conditions favorable to trigger convection.

A commonly used diagnostics to quantify the potential for convection is CAPE, which
is a measure of the energy due to buoyancy, available for convection. Juckes and Smith (2000) showed using a theoretical model that upper-level troughs in the tropical region, as well as in mid-latitudes, cause an increase in CAPE, with the increase being larger for stronger and/or broader troughs. Therefore, it is expected that there should be enough CAPE accumulation prior to the outbreak of convection, and gradual dissipation as this energy is consumed by convection itself. Vertical ascent must be present for this energy to be released and for convection to be able to persist beyond the timescale of a cumulonimbus cloud (Trier 2003; Raymond et al. 2003).

Therefore, the combination of changes in the dynamical (upward vertical velocity) and thermodynamical (decreased static stability) structure of the atmosphere due to the cyclonic anomaly in the upper troposphere and CAPE build-up provide the frame for the occurrence of convection. The proposed theory has not been, to our knowledge, verified for intrusion-like systems within the tropical/subtropical region. In this work, we examine whether the available data will be compatible with the existing theory. In an effort to determine the role the intrusions play in forcing the ascent (convection) we shall perform PV inversion calculations (e.g. Davis and Emanuel 1991; hereafter DE91) to determine the vertical motion attributable to the anomalous PV within the intrusion.

The structure of the chapter is as follow: Section 3.2 provides an overview of PV inversion and section 3.3 show its application to an idealized case. Sections 3.4 and 3.5 describe results for selected intrusion cases and finally, summary is presented in section 3.6.

### 3.2 PV inversion theory and Attribution: Overview

“PV inversion” refers to the mathematical procedure in which given a distribution of PV it is possible to recover the dynamical fields associated with this distribution by “inverting” the operator relating those quantities (HMR). The inversion is constrained by the choice of
balance condition between mass and wind fields, e.g., quasi-geostrophy, gradient balance, etc. Also, proper boundary conditions must be defined. For example, DE91 developed a technique to perform PV inversion using Charney balance condition, which is weakly non-linear. The geopotential $\Phi$ and streamfunction $\psi$ are related by neglecting the divergent and vertical components of the wind. This relation can be expressed symbolically as

$$\nabla^2 \Phi = M(\psi)$$  \hspace{1cm} (3.1)

where $M$ is a non-linear operator and $\nabla^2$ is the horizontal Laplacian operator. A second equation relates PV to $\Phi$ and $\psi$, using the definition of Ertel’s PV and the hydrostatic approximation, i.e.,

$$q = \mathcal{N}(\Phi, \psi)$$  \hspace{1cm} (3.2)

where $q$ is the potential vorticity and $\mathcal{N}$ is another non-linear operator. Equations (3.1) and (3.2) form a coupled system of non-linear equations to be solved for $\Phi$ and $\psi$, given a distribution of PV (the explicit set of equation is presented in Appendix A).

The applicability and usefulness of so-called “PV thinking” and “PV invertibility principle” (HMR) are abundant in the literature. For example, it has been used to diagnose cyclonic development (DE91, Stoelinga 1996, Martin and Marsili 2002; Evans and Prater-Mayer 2004; Augusti-Paneda et al. 2004; Ahmadi-Givi et al. 2004), frontal development (Thorpe 1997; Morgan 1999; Chen et al. 2003), and tropical dynamics (Moller and Jones 1998; Jones et al. 2003; Lakmann and Yablonsky 2004).

The “PV invertibility” also allows to address the issue of “attribution”, in the sense proposed by Thorpe (1997). As explained by Thorpe (1997), in many meteorological systems it is not clear to distinguish the cause-and-effect of dynamical features. However, by using PV inversion, it is possible to verify whether a certain feature “B” is attributable to another feature “A” (the association is ‘weaker’ than in a cause-and-effect statement). These ideas
have also been explored in the literature (e.g., DE91, Davis 1992b, Morgan 1999, Martin and Marsili 2002). Usually, to explore the issue of attribution, PV is split into a mean part and a perturbation relative to this mean,

$$q = \bar{q} + q'. \quad (3.3)$$

The perturbation $q'$, in turn, may be split up into several non-overlapping components, which would represent specific aspects of the anomaly. Commonly, $q'$ gets split into three parts: (a) upper level anomaly, representing disturbances near and below the tropopause, (b) interior PV anomaly, due to latent heat release, and (c) boundary temperature anomalies, also incorporating the effects of surface heating to the layers immediately above. Equation (3.3) can then be expressed as

$$q = \bar{q} + q'_U + q'_I + q'_L, \quad (3.4)$$

the last three terms representing contributions by (a), (b) and (c), respectively.

For the non-linear system given by Eqs. (3.2) and (3.1), the attribution of $\Phi'$ and $\psi'$ to a particular $q'_i$ is not straightforward as it is in the linear case. There is not a unique way to attribute dynamical fields to a particular $q'_i$. Direct piecewise inversion can only be accomplished by linearizing the interactions between the perturbation fields and the perturbations with the basic state in the non-linear system. In DE91 piecewise technique, non-linear terms are “hidden” in non-constant coefficients of the linear differential operator, and the sum of the perturbations sum up to the total perturbation, i.e., $\sum_i q'_i = q'$, $\sum_i \Phi'_i = \Phi'$, and $\sum_i \psi'_i = \psi'$. An alternative procedure to evaluate the dynamical quantities attributable to a certain $q'_i$ consists in removing (only) the perturbation $q'_i$ from the total flow, inverting for this new, altered distribution of $q$, and finally, removing the resulting fields from the fields obtained by inverting “total” $q$. This methodology is meaningful only when the non-linearity of the system is weak, and in this case, the results using this procedure and
the DE91 piecewise inversion technique do not yield significant differences (e.g., Davis 1992a). We performed PV inversion using these two approaches and verified that the differences between the results are negligible (not shown). In this study we opted to use the second approach for the attribution problem.

The issue of attribution of vertical velocity to a particular $q'_i$ is somewhat difficult. First, we need to relate the vertical motion to the balanced dynamical fields from the PV inversion, i.e., we need to find the ‘balanced vertical motion field’, consistent with the balanced horizontal winds. In this study we opted to work with the Q-vector form of the $\omega$ equation (e.g., Morgan 1999) and included the $\beta$-effect (Bluestein, 1992):

$$L(\omega) = -2\nabla \cdot Q - \beta \frac{\partial \theta}{\partial x},$$

where

$$L = S \nabla^2 + \frac{f_o^2}{\gamma} \frac{\partial^2}{\partial p^2}, \quad S = -\frac{T}{\theta} \frac{\partial \theta}{\partial p}, \quad \gamma = \frac{R}{p} \left( \frac{p}{p_0} \right)^\kappa,$$

and

$$Q = \left[-\frac{dv}{dx} \cdot \nabla \theta, -\frac{dv}{dy} \cdot \nabla \theta \right].$$

$R = 287 \text{ J.kg}^{-1}\text{K}^{-1}$, $\kappa = R/c_p$, $c_p = 1004.5 \text{ J.kg}^{-1}\text{K}^{-1}$, $\beta = df/dy = 2\Omega \cos \phi/a$, $a$ is the average radius of the Earth, $\Omega$ is the Earth’s rate of rotation, and $v = (u,v)$. Notice that $T/\theta = (p/1000)^\kappa$, and therefore, $L$ is linear despite the $\theta$ dependency on the right hand side. The Q-vector is calculated using the balanced fields obtained by inverting PV. In the context of the omega- Q-vector formulation, there is no diabatic or frictional forcing for the vertical motion, which means that the flow is adiabatic, i.e., a particle initially on an isentropic surface will remain in this surface. For the intrusion events we are studying, this assumption does not hold for the entire time span of the system evolution. However, it can give a quantitative estimate of the vertical motion as the system evolves (see Section 3.4.1).
The operator $\mathcal{L}$ is linear, and therefore, $\mathcal{L}(\omega) = \mathcal{L}(\bar{\omega}) + \mathcal{L}(\omega')$. However, since $\mathbf{v} = \mathbf{v} + \mathbf{v}'$ and $T = T + T'$, the Q-vector forcing will have cross-terms representing the interaction of the basic state and the perturbation. Therefore, again the issue of attribution is not straightforward. The approach taken here is analogous to that described earlier this section (e.g., Davis 1992b): First, we calculate $\omega = \mathcal{L}^{-1}(Q\text{-forc.})$, then remove $q'_{U}$ from $q$ obtaining, say, a distribution $q^*$. Using $q^*$ we find $\mathbf{v}^*, T^*$ associated with this distribution, calculate $Q^*$ and from that, get $\omega^* = \mathcal{L}^{-1}(Q^*\text{-forc.})$. Finally, $\omega'_{U} = \omega - \omega^*$.

We wish to also attribute changes to CAPE due to $q_i$. CAPE is defined as

$$CAPE = \int_{LFC}^{LNB} B dz$$

where the integration limits LFC and LNB stands for “level of free convection” and “level of neutral buoyancy”, respectively, and $B$, the parcel buoyancy, is defined as

$$B = g \left[ \frac{\theta_{vp}(z) - \theta_v(z)}{\theta_v(z)} \right]$$

where $\theta_{vp}(z)$ is the virtual potential temperature of the air parcel and $\theta_v(z)$ is the virtual potential temperature of the environment. As discussed CAPE is often used as a diagnostic of the potential for convection. It relies on the assumption of the parcel theory, i.e., that a parcel of air being displaced vertically will not exchange mass or heat with its surroundings and will adjust immediately to the local pressure to where it has been displaced. The attribution of changes in CAPE to a given $q'_i$ is more complicated because of the dependency of CAPE on both temperature and moisture distribution. If the dependency of CAPE on moisture is small enough such that it can be neglected, then CAPE due to $q'_i$ can be obtained by calculating the temperature attributable to $q'_i$. This issue will be discussed further in Section 3.4.1.

In summary, given a distribution of PV, it is possible to find the geopotential and wind distribution satisfying the determined balance (in this case, Charney balance), by solving
Eqs. (3.1) and (3.2). Then the invertibility principle allows the issue of attribution to be addressed, i.e., to find what part of the geopotential and wind fields are attributable to a certain $q_i'$. Once the balanced geopotential field is obtained, it is possible to calculate the temperature distribution by using e.g., the hydrostatic approximation. In this case, the dependency of temperature on the geopotential is linear, it is also possible from the geopotential field to obtain the temperature distribution and change in CAPE related to a particular $q_i'$.

### 3.3 Idealized PV anomaly

Before considering realistic PV intrusions we examine the flow due to an idealized PV anomaly. We consider a hypothetical background state such that $\overline{T} = \overline{T}(z)$ only, decreasing monotonically with height. This implies that the background flow $\mathbf{v} = (u,v) = 0$ everywhere, in order to satisfy the thermal wind relationship. Moreover, the geopotential and streamfunction fields are constant on isobaric surfaces. A PV anomaly distribution following $q' = q_0[1 - (r/R_0)^2]$ ($q_0$ the maximum strength and $R_0$ the radius of the anomaly, and $r$ the distance from the center of the anomaly) was superimposed to this background state. $R_0$ was set to 300km in the horizontal direction and 1.7km in the vertical, i.e., the vertical and horizontal scales were decoupled, such that $(r/R_0)^2 = r_h^2/R_{0h}^2 + r_v^2/R_{0v}^2$ ($h$ subscript refers to horizontal and $v$ to vertical). The total size of the grid was 71 × 51 gridpoints in the (x,y) domain, with spacing of 1°, corresponding to an area span of 5-55°N, 180-250°E. In the vertical it was used 20 levels, from 1000 to 50 hPa, with spacing of 50hPa.

Figure 3.2a shows the PV anomaly distribution for an anomaly centered at 5.5km height and with central (maximum) strength of $q_0 = 9$PVU. The dynamical fields associated with this PV anomaly are shown in Fig. 3.2b-d (note that in this case, since the background state is at rest, the winds obtained by using Eqs. (3.1) and (3.2) are actually “perturbation”
Figure 3.2: (a) Vertical cross-section at 30°N of hypothetical PV distribution. PV anomaly with a maximum of $q_0 = 9$ PVU at a height of 5.5km. (b)-(d) shows the results by performing PV inversion using Eqs. (3.1) and (3.2). (b) Geopotential perturbation $\Phi'$ at 900hPa. (c) Vertical cross-section of meridional velocity $v'$ (m/s, solid) and potential temperature $\theta$ (K, dotted). (d) Zonal (solid) and meridional (dotted) components of wind at 900hPa.
winds). The temperature field is distorted due to the presence of the anomaly; it is more stable “inside” the anomaly and more unstable below and ahead of the anomaly. There is cyclonic circulation associated with this positive PV anomaly and the geopotential field shows a low due to this circulation. These results are in accordance with the theoretical results obtained by Thorpe (1985). HMR showed that this wind and temperature distributions represent qualitatively well the observed structure of mid-latitude cyclonic systems (e.g. see their Fig 8).

Similar PV inversions were performed for anomalies with strength \( q_0 = 1, 2, 3, 4, 6, 12, 18 \) PVU, placed at heights of \( z_c = 4, 5.5, 9 \) km (corresponding pressure levels are 612.13, 504.13, 310.55 hPa). Figure 3.3 shows a few diagnostics based on these inversions. The behavior of the maximum horizontal velocity near the surface is, as expected, larger when the anomaly intensity is larger or the anomaly is closer to the ground (Fig. 3.3a). Figure 3.3b shows the ratio of the maximum velocity near the ground to the maximum velocity at the level of the anomaly. This is the diagnostic used by Davis (1992a) to measure the vertical penetration. The results presented here are in good agreement with those obtained by Davis (1992a). It is not possible to find a generalized vertical penetration scale due to the non-linearity inherent in the mathematical problem. As an alternative measure of the vertical penetration, Fig. 3.3c shows the ratio between the height were the maximum velocity drops to \( e^{-1} \) of the maximum velocity at \( z_c \), and \( z_c \) itself. For the anomalies placed at \( z_c = 4 \)km, the winds are greater than \( 1/e \) of the maximum wind speed when the anomaly strength reaches 6 PVU. For the anomalies placed near the top of the troposphere (\( z_c = 9 \)km), the e-folding scale-height increases almost linearly with its strength. The anomalies placed at \( z_c = 5.5 \) km present a mixed behavior; when the anomalies are relatively small, the behavior is almost linear. After some threshold, there is a change in slope, for the increase of the scale-height.

Finally, the near-surface static stability (the slope of the potential temperature in the
Figure 3.3: Variation of (a) maximum horizontal wind speed (m.s\(^{-1}\)) at 900hPa, (b) ratio between maximum horizontal wind speed at 900hPa and maximum horizontal wind speed at the level of the anomaly, (c) ratio between height where the maximum velocity is equal to \(1/e\) of the maximum velocity at the level of the anomaly and the height of the (center of) anomaly, and (d) "static stability" in the 1000-850 hPa layer (deg/km), with strength of anomaly \(q_a\), at 30°N, 215°E. Solid line show variation for anomaly at \(z_c = 4\)km, dotted line for \(z_c = 5.5\)km, and dashed line \(z_c = 9\)km. Thick dark line in (d) represents the the background \(d\theta/dz\) in the 100-850 hPa layer.
1000-850hPa layer) immediately below the anomaly center shows that there is a substantial
decrease in stability - i.e., this layer becomes more unstable - due to the presence of the
anomaly. For the anomalies placed in the mid-atmosphere and closer to the ground, it seems
that there is a ‘saturation limit’, in which $\frac{d\theta}{dz}$ seems to converge towards an asymptote.
For $z_c = 9$km, such behavior can not be deduced. However, it is clear that even for the
higher and weaker anomaly ($z_c = 9$km, $q_0 = 1$PVU), there is a signal of destabilization of
the atmosphere at low levels.

Figures 3.4a,b show the change in $\theta_v$ vertical profiles of the parcel and environment due
to the presence of the anomaly. Figure 3.4a show the profile for the stable conditions of the
background state, and it is clear that there is no CAPE at this point. However, the positive
PV anomaly changes the $\theta_v$ profile (Fig. 3.4b) so that energy for convection is available.

Those calculations were repeated for the same conditions, but for the latitude displaced
from 5-55$^\circ$N to 20-70$^\circ$N. The results were just slightly stronger than for the ‘subtropical’
case (not shown).

![Figure 3.4](image)

Figure 3.4: Vertical profile of virtual temperature of “environment” (solid) and parcel
(dashed), for (a) background; and (b) background+perturbation, for the conditions de-
dscribed in Fig. 3.2. Both profiles are at 30$^\circ$N 215$^\circ$E.
3.4 Case Study 1: 12-17 January, 1987.

We now examine whether the data for a particular intrusion event are qualitatively consistent with the theory through the analysis of pertinent variables ($\theta$, PV, vertical velocity $\omega$, OLR, CAPE). We also use the PV inversion technique of DE91 to address the issue of attribution, i.e., to quantify the contribution of the upper level PV anomaly to the total fields of the above variables. The complete set of equations and description of the method can be found in DE91 and Davis (1992b), and is also presented in the Appendix. In our calculations we use Neumann condition (hydrostatic approximation $\partial \Phi / \partial \pi = -\theta$) for horizontal boundary conditions, and Dirichlet (prescribed $\Phi$ and $\psi$) for lateral boundary conditions.

The intrusion event we examine occurred between 12-17 of January, 1987. This case has been previously investigated in Chapter 2 (and in Kiladis and Weickmann 1992b), and therefore, is a natural choice for further investigation. Figure 3.5 shows the evolution of PV, OLR and vertical velocity $\omega$ for this period. The PV = 2 PVU (1PVU = $10^{-6}$ K s$^2$ kg$^{-1}$) contour shows the position of the dynamical tropopause and it is seen to undulate and amplify in the early stages, turning into a sharp trough by 15 January and decaying as it moves eastwards. A large area of deep convection (low OLR) is observed ahead of the intruding PV trough from the 15th on. Upward vertical velocity is found in areas of low OLR (cloudiness). In particular, there is strong upward motion nearly coincident with the area of low OLR on the 15th. All these elements are consistent with the theoretical description discussed earlier.

Figure 3.6 show cross sections of $\theta$ and PV on 15 January 1987. Figure 3.6a shows a latitudinal cross section at 25°N, and Fig. 3.6b a longitudinal cross section at 215°W, which cuts through ahead of the intrusive tongue. It is clear from Fig. 3.6a that there is decreased static stability downstream of the intrusion at low levels (between 1000-600 hPa, and 205-220°E). The tilt of the isentropic surface and translation of the system to the east
Figure 3.5: Potential vorticity (PVU, solid contours) at 200hPa, OLR (< 180 W.m$^{-2}$, shaded), and vertical velocity $\omega$ (Pa/s, > 0 thick black solid contours, < 0 thick dot-dashed contours), for the period 12-17 January 1987 12UTC. Contours of $\omega$ were cropped at 30°N.
indicate that in a frame of reference moving with the intrusion a particle will move upwards and westward in the region downstream of the upper level trough. Also, we see from Fig. 3.6b that the isentropes slope upward towards the north pole “ahead” of the intrusion. In a 3-D perspective, the particle is moving westwards, northwards and upwards in the region ahead of the intrusion, i.e., it acquires cyclonic circulation.

Figure 3.6: Vertical cross-section of \( \theta \) (K, thin lines), PV (PVU, thick lines) and vertical velocity \( \omega \) (Pa s\(^{-1}\), > 0 thick black solid, < 0 thick dot-dashed) at (a) 25N and (b) 215E, on 15 January 1987 12UTC. Vertical coordinate is pressure (Pa), in log scale.

Figure 3.7: Vertical profile of virtual temperature of “environment” (NCEP data, solid) and parcel (dashed), on (a) 13 and (b) 15 January 1987. Both profiles are at 20°N 215°E.
Figures 3.7 and 3.8 address time changes in CAPE as the intrusion evolves. Figure 3.7 shows vertical profiles of $\theta_v$ and $\theta_{vp}$ for the 13th and 15th, both at 20°N 215°E. CAPE is proportional to the area between the curve of $\theta_v$ and $\theta_{vp}$. It is clear that there is a greater value of CAPE on the 15th, while on 13th CAPE is not significant. Figure 3.8 shows the time evolution of CAPE (at 12 UTC) from 13 to 17 January, along with 2 PVU contour to locate the position of the intruding trough. There is a clear build-up of CAPE downstream of the trough, and decay of CAPE as the intrusion moves eastwards. The potential energy for convection can be released by the vertical upward movement in the lower levels by isentropic upglide and isentropic displacement due to the anomaly.

![Figure 3.8: Distribution of CAPE (thin) and 2PVU (thick) for 14-17 January 1987 12UTC. Parcels are lifted from 950 hPa and assumed to undergo pseudo-adiabatic ascent. CAPE contour intervals are 200 J.kg$^{-1}$.](image)

The analysis above shows that there is good qualitative agreement between the data to the proposed theory, e.g. compare with Fig. 3.1. We now examine whether there is quantitative agreement.

### 3.4.1 PV inversion

The main purpose of this section is to quantify the changes in $\theta$, $\omega$ and CAPE that are attributable to the upper level PV anomaly due to the intrusion. We first perform the inversion of the PV field using DE91 technique, and verify its accuracy for the type of
events that we are studying. In a second step, we address the attribution issue itself, i.e.,
isolate the upper level anomaly and verify its contribution to the mentioned fields.

**Charney balance**

The Charney balance (Charney 1955) condition is suitable for mid-latitude large-scale
flows and several authors have used the DE91 technique for studying mesoscale structure
and dynamics, such as tropopause folds and frontogenesis in the midlatitudes (e.g., Thorpe
1997, Morgan 1999, Griffiths et al. 2000). However this method has not been applied
to subtropical structures we are interested in, which are quasi-horizontal and have a very
narrow tongue-like shape. In the subtropics, the Rossby number is larger than in the extra-
tropics, and the irrotational component of the wind may be significant. With this in mind,
the question arises of how different the fields obtained by PV inversion using this balance
condition will be with respect to the observed field, and whether this balance condition is
still suitable for the type of events we want to study.

![Figure 3.9](image_url)

Figure 3.9: (a) NCEP reanalysis geopotential height (gpm, solid contours) and balanced
geopotential field obtained by inverting PV (gpm, dotted contours) on 200hPa, (b) Same as
(a) but for 850hPa.
We examine this by applying the DE91 inversion method to the PV for 15 January 1987. Figure 3.9a,b compares the geopotential field from NCEP reanalysis (solid contour) with that from the balanced inversion (dashed). There is in remarkably good agreement with the NCEP reanalysis data at all pressure levels. The discrepancies are larger at 1000hPa (up to \( \sim 30\% \)), but at and above 925hPa the errors do not exceed 4.5\%. In the wind field, the differences between reanalysis data and inverted fields are larger. These differences stem from the fact that the irrotational part of the flow was neglected in the Charney balance, and discrepancies in the magnitude of the wind were found especially downstream of the upper-level trough (not shown). At 200 hPa, the maximum difference was of \( \sim 25 \text{ m.s}^{-1} \) in the zonal direction, and \( \sim 15 \text{ m.s}^{-1} \) in the meridional wind field. These discrepancies are of the same magnitude of those found by Agusti-Panareda et al. (2004) for upper-level troughs. At 850hPa, the differences were mainly to the order of 3 m.s\(^{-1}\), which can be quite significant since the wind speed at these levels are of \( \sim 10 \text{ m.s}^{-1} \). However, the main pattern and position of centers of maxima and minima of the wind were very well captured by the inverted fields at all pressure levels. These calculations show that flow satisfies the Charney balance to a good approximation, even though the divergent part of the wind cannot be captured by using this balance. We performed the same technique for a narrow trough approaching the west coast of North America on 8 December 1996, but with the trough located well within the midlatitudes (trough axis between 150-135W, and covering \( \sim 25-55^\circ\text{N} \)), and the results are similar, although with less pronounced differences in both geopotential and wind magnitudes, as expected (not shown).

The vertical velocity \( \omega \) was calculated using the Q-vector formulation (see Section 3.2). Figure 3.10 shows a comparison between the vertical velocity \( \omega \) at 500 hPa from (a) reanalysis data, (b) inverting Eq. (3.5) using \( v \) and \( T \) from the reanalysis to calculate \( Q \) and (c) inverting Eq. (3.5) using balanced \( v \) and \( T \) from PV inversion. There are noticeable differences between \( \omega_{\text{rean}} \) and inverted \( \omega \) (obtained solving Eq. (3.5)): the upward velocity
(dash-dotted line) is weaker and the downward velocity (solid) is stronger than the reanalysis data. Differences are expected as the $\omega$-equation is based on the quasi-geostrophic theory and represents the instantaneous vertical velocity of the system as it adjusts itself to be in quasi-geostrophic balance and therefore is not accurate quantitatively. However, it qualitatively provides the correct position and pattern of upward and downward motions, and the correct order of magnitude of the velocity. Figure 3.10(c) shows the solution of Eq. (3.5) but using the balanced horizontal wind and temperature fields to calculate the forcing term. This result is very similar to $\omega$ calculated using reanalysis data for the forcing. We will use values calculated from the $\omega$-equation to obtain a rough estimate of the contribution of intrusion on the vertical field relative to the total field. The proposed hypothesis (see Chapter 1) is that intrusion initiates and support convection by destabilizing the atmosphere in the regions below and ahead. Once convection is triggered, it is “self-sustained” and the contribution of the intrusion to the vertical velocity may decrease.

Figure 3.10: Vertical velocity $\omega$ at 500 hPa (Pa/s) given by: (a) Reanalysis data; (b) solving the Q-vector form of $\omega$-equation (Eq. (3.5), where $Q$ is calculated using $(u, v, T)$ from reanalysis data; (c) idem to (b), except $Q$ is calculated using balanced fields $(u, v, T)$ from PV inversion. Positive values (downward motion) in solid contours, negative values (upward motion) in dash-dot contours. Contour values $\pm[0.1, 0.2, 0.3, 0.5]$. 
**Attribution**

Having verified that this technique is appropriate to recover the dynamical fields for this system, we now address the issue of attribution. The quantitative definition of a perturbation depends on the choice of basic state. For this case, we used a 5-day mean as a basic state, which represents the period in which the intrusion grew and decayed. The perturbations were defined as follow: $q'_U$ is the perturbation in upper levels (400-100 hPa), $q'_I$ is the perturbation in intermediate levels (700-500 hPa), and $q'_L$ encompasses the contributions of the lower atmosphere and the surface (potential) temperature anomalies (e.g., Davis et al. 1996). The choice for $q'_U$ was based on the map of PV minus the time mean PV ($\bar{PV}$). Latitudinal cross sections show that the 1PVU contour was close but did not go below 400 hPa (not shown).

Figure 3.11 show the geopotential anomaly associated with total perturbation PV (left column) and perturbation PV in the layer 400-100 hPa ($q'_U$, middle column). Comparing with the total perturbation fields, it is noticeable that the effect of $q'_U$ is dominant throughout the troposphere. The contributions of the geopotential anomalies attributable to mid-lower atmosphere ($q'_I$) and boundary $\theta$ perturbations ($q'_L$) are small, the joint contribution being at most 50% in the region near surface, in the area immediately below the intrusion.

The impact of the temperature rearrangement due to the intrusion can be also accessed by PV inversion. Figure 3.12 shows the temporal evolution of dry static stability ($S = -g \partial \theta / \partial p$) for two gridpoints, one near Hilo (19.7°N 204.9°E) and one gridpoint within the area of ascent ahead of the intrusion on 15 January. The solid line is the static stability $S$ calculated using the temperature from PV inversion of total PV field, while the dashed line is the contribution of $q'_U$ only. The evolution of the static stability at the point near Hilo clearly shows that $q'_U$ dictates the decrease of static stability as the intrusion hits Hilo. For the other location the contribution of $q'_U$ is smaller, however there is a steady decrease of
Figure 3.11: 2PVU contour (thick) on 200 hPa and perturbation geopotential fields (in gpm) on 15 January 1987 obtained by inversion of total $PV'$ (left column), and inversion of $PV'$ in the upper atmosphere (400-100 hPa) (center column). Positive values are shown in solid contours, negative in dotted contours. Right column shows difference between the left and center columns. Pressure level where the perturbations are being displayed is shown in the left side of each row.
Figure 3.12: Time series of dry static stability \((-\frac{g \partial \theta}{\partial p}, \text{in K.m}^2.\text{kg}^{-1}\) for the period 13-17 January 1987, for (a) 20°N 205°E and (b) 20°N 212.5°E, calculated using results from PV inversion. Solid line for total field, dashed line represents the contribution of \(q'U\) (i.e., \(\overline{S} + S'U\)).

Static stability with time attributable to \(q'U\). Calculations show that the contribution of \(q'U\) to this decrease is of 100% for the period before the onset of convection (see Table 3.1), ahead of the intrusion.

Figure 3.13 shows the time evolution of CAPE (dark solid) on 20°N 212.5°E which coincides with the maximum in the CAPE field by 15 January (see Fig. 3.8). CAPE values are dependent on temperature and moisture fields. Moisture effect is incorporated in the calculation of virtual potential temperature and also determines, together with the temperature, the lower limit of CAPE integration (lifting condensation level, LCL). PV inversion cannot determine the amount of moisture directly associated with a particular distribution of PV. Also, the relationship between CAPE and CAPE due to \(q'U\) (CAPE\(_U'\)) is not exactly linear because of CAPE dependency on both temperature and moisture. Therefore, before quantifying the contribution of \(q'U\) to CAPE, first we need to access the sensitivity of CAPE to both moisture and temperature changes. The sensitivity test was performed in the following manner: CAPE was calculated using average water vapor mixing ratio profiles (for the period 13-17 January 1987) but the observed variation in temperature, and then, using
averaged temperature and varying mixing ratio. The results showed that for fixed temperature the signal of CAPE is strong in the early part of the event, but does not show any build-up prior to the onset of convection (Fig. 3.13). For fixed mixing ratio, however, there is a very strong signal of CAPE, with the pattern and intensity resembling that of CAPE calculated using the actual profiles of $T$ and mixing ratio (Fig. 3.13). Therefore, CAPE is more sensitive to the temperature changes than to the moisture changes, at least for this event, and we chose to calculate $\text{CAPE}'_U$ using the actual value of mixing ratio.

The contribution of $q'_U$ to CAPE is shown as dark dashed curve in Fig. 3.13. It has the same pattern of “total” CAPE, and in the period of strong convection (between 15-16 January 00UTC for this point), the contribution of $T'_U$ is about 80% of the total CAPE. It is important to point out, once more, that $\text{CAPE}'_U$ is actually CAPE that the atmosphere

![Figure 3.13: Time series of CAPE for the period 13-17 January 1987, at 20°N 212.5°E. The thick solid curve is total CAPE, the thin dot curve corresponds to CAPE calculated using 5-day averaged mixing ratio profile, thin dash-dot curve is CAPE calculated using 5-day averaged temperature profile, thin dash-dot curve is CAPE calculated using temperature associated with $q'_U$.](image)

51
The last dynamical variable we want to examine is the vertical velocity $\omega$. Figure 3.14 shows the contribution of $q'_U$ to $\omega$ for the levels of 500 hPa and 925hPa, calculated as explained in Section 3.2. The left panel shows $\omega$ obtained by inverting the total field, and the right panel, $\omega'_U$. It shows that $q'_U$ contributes to virtually all of the total upward motion field ahead of the intrusion in the lower and mid-levels of the atmosphere (see also Table 3.5). Therefore, the vertical motion is attributable almost exclusively to the PV intrusion.

The results above show that the PV intrusion has a strong signal in key dynamical
and thermodynamical diagnostics that are related to convection. Decreased static stability, 
CAPE accumulation and vertical ascent are dominated by contribution due to \( q'U \). We now 
examine whether these signals are consistently found for cases that penetrate deeper in the 
tropical region.

### 3.5 Other case studies

The above methodology was applied to 6 other intrusions: 12 February 1991, 16 January 1997, 14 February 1997, 14 January 1999, 23 January 1999 and 28 January 2003. The first five are from the climatology provided by WP2000; the case of January 2003 is outside of the period WP2000 used in their climatology and was chosen because it was a recent event. The cases of February 1992, January 1997 and early January 1999 were chosen because they are events in which the 2PVU line was detected on the 330K at 20°N (most of the intrusion events did not have deep vertical penetration). Figure 3.15 shows the PV, OLR and vertical velocity for the 6 cases on the day that an “intrusion” as defined by WP2000 was identified. In all of them there is a pronounced trough, upward vertical velocity ahead of the intrusion and low OLR downstream of the trough.

Although there is similarity on the shape and pattern of convection in all these events, the PV anomaly for all except the February 1991 event does not have a signal as deep in the atmosphere as in the case of January 1987 (Fig. 3.16). For the cases of January 1987 and February 1991, 1 PVU anomaly reaches 400hPa, while in the other cases it reaches only 300 hPa (not shown). Recall from section 3.3 that the vertical radius of influence varies with the strength of the PV anomaly and its position relative to the ground. This should be kept in mind in the analysis of \( \omega \), CAPE, and static stability.

The first thing to verify is whether the reanalysis data shows any destabilization of the atmosphere and upward vertical velocity downstream of the intrusion. Latitudinal and
Figure 3.15: Same as Fig. 3.5, except for the dates shown above each map.
Figure 3.16: Vertical cross-section of θ (K, thin lines), PV (PVU, thick lines) and vertical velocity ω (Pa.s\(^{-1}\), > 0 thick black solid, < 0 thick dot-dashed for the dates and latitudes indicated on the top of each cross section

Longitudinal (downstream of the intrusion) cross sections are shown in Fig. 3.16. Upward vertical velocity (dot-dashed lines) is present ahead of the intruding PV tongue, which is evident in both latitudinal and longitudinal cross sections. This upward vertical velocity is fairly coincident with regions of lowered static stability, at least qualitatively. Figure 3.17 shows sequences of 4-day CAPE evolution for each of the cases. Except for the cases of January 1997, there is CAPE accumulation as the PV intrusion evolves.

The next point to be verified is how adequate is Charney balance for the intrusion events
that come deep into the tropical region. PV invertibility itself is independent of domain; however, it is physically constrained by the balance between the mass and wind fields. The results of the inversion showed a remarkable agreement between the balanced fields and the NCEP data, again, especially for the geopotential field. The wind fields are weaker, as was also found previously for the case of January 1987, but maxima and minima patterns and position were relatively well represented (not shown).

Having verified that in these events there is destabilization ahead of the intrusion and that Charney balance is suitable we now, as in the previous section, use PV inversion to quantify the contribution of upper level PV anomalies.

a. Static Stability

Figure 3.18 shows the average static stability $S$ between the 850-500hPa layers. For all cases, there is a clear drop in the static stability prior to the outbreak of convection
The decrease in $S$ related to $q'U$ (dashed lines) shows that for most cases, the overall decrease in static stability can be attributed to the intrusion. Even for the case of February 1997, where the static stability increases before 13 February 12UTC, there is a drop in static stability 12 hours before the intrusion reaches its maximum penetration into the tropical region and its slope is nearly identical to the total decrease in the static stability. Notice that more important than the value of the static stability itself is its time variation. Therefore, it is clear from these results that the decrease in the static stability in the lower levels of the atmosphere before the outbreak of convection is to a very large extent attributable to $q'U$.

The above findings are summarized in Table 3.1. The first 2 columns give the time variation of the total $S'$ perturbation and rates of variation of $S'$ due to $q'U$ at the gridpoint indicated, for the 48h before 12UTC of the date indicated (when deep intrusion was detected). Results show that in 4 out of 7 cases, $q'U$ contributes with 88% or more to the decrease in static stability, and in 2 out of 7, with about 50%. When we make an area average around the gridpoint (shown in the last 2 columns), the contributions of $q'U$ are even larger ($\sim 100\%$ of total decrease in 6 out of 7 cases). Even for the case of 14 February 1997, on average $q'U$ contributes to decrease the static stability. Therefore, it is clear from the above results that the upper level intrusion has a great impact in destabilizing the atmosphere prior to convection.

b. CAPE

CAPE and CAPE$_U'$. are shown in Fig. 3.19 for the dates and gridpoints shown on the top of each panel. These gridpoints are located close to the area of large CAPE (Fig. 3.17). CAPE of the background (mean) state is zero, and therefore in this case CAPE itself is a ‘perturbation’ value and represents the total contribution of all anomalies.

For the events of February 1991, January 1999 and January 2003, it is clear that the
Figure 3.18: Same as in Fig. 3.12, but for the dates and gridpoints shown on the top of each plot. Notice that the y-scale for the first panel (February 1991) is lower than for the others.
Table 3.1: Time variation of perturbation static stability $S'\,$ (in units of $10^{-5}$ K.m$^2$.Kg$^{-1}$.h$^{-1}$) averaged over 48h before 12UTC of the date indicated. $\frac{\partial S'}{\partial t}$ is the change associated with total anomaly $q'$, and $\frac{\partial S'_{U}}{\partial t}$ is the change attributable to $q'_{U}$.

<table>
<thead>
<tr>
<th>DATE</th>
<th>Gridpoint</th>
<th>$\frac{\partial S'}{\partial t}$</th>
<th>$\frac{\partial S'_{U}}{\partial t}$</th>
<th>Gridpt. ave.</th>
<th>$\frac{\partial S'}{\partial t}$</th>
<th>$\frac{\partial S'_{U}}{\partial t}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>15 Jan 1987</td>
<td>20°N 212.5°E</td>
<td>-3.25</td>
<td>-1.78</td>
<td>17.5–22.5°N 210–215°E</td>
<td>-1.88</td>
<td>-3.22</td>
</tr>
<tr>
<td>12 Feb 1991</td>
<td>15°N 235°E</td>
<td>-2.57</td>
<td>-2.25</td>
<td>12.5–17.5°N 232.5–237.5°E</td>
<td>-2.00</td>
<td>-2.24</td>
</tr>
<tr>
<td>16 Jan 1997</td>
<td>12.5°N 245°E</td>
<td>-0.57</td>
<td>-0.62</td>
<td>10–15°N 242.5–247.5°E</td>
<td>-0.69</td>
<td>-0.64</td>
</tr>
<tr>
<td>14 Feb 1997</td>
<td>15°N 215°E</td>
<td>-0.96</td>
<td>0.52</td>
<td>12.5–17.5°N 212.5–217.5°E</td>
<td>0.57</td>
<td>-0.82</td>
</tr>
<tr>
<td>14 Jan 1999</td>
<td>15°N 220°E</td>
<td>-2.82</td>
<td>-2.57</td>
<td>12.5–17.5°N 217.5–222.5°E</td>
<td>-2.21</td>
<td>-2.30</td>
</tr>
<tr>
<td>23 Jan 1999</td>
<td>15°N 202.5°E</td>
<td>-1.55</td>
<td>-1.50</td>
<td>12.5–17.5°N 200–205°E</td>
<td>-1.46</td>
<td>-1.43</td>
</tr>
<tr>
<td>28 Jan 2003</td>
<td>12.5°N 235°E</td>
<td>-1.46</td>
<td>-0.71</td>
<td>10–15°N 232.5–237.5°E</td>
<td>-0.72</td>
<td>-1.37</td>
</tr>
</tbody>
</table>

upper-level trough had a large contribution to (total) CAPE build-up, this contribution being higher than 50% of the total value. For the cases in 1997 this contribution is smaller, but $q'_{U}$ still has a strong signal in CAPE.

Table 3.2 show values of the relative contribution of $q'_{U}$ for CAPE. The maximum value of CAPE$'_{U}$ is generally found prior to or at the same time of the maximum in the total CAPE (Fig. 3.19). The contribution of maximum CAPE$'_{U}$ to the total CAPE is about 60% or more in all but the case of January 1997 (column (i)). Column (ii) shows the contribution of CAPE$'_{U}$ to the maximum CAPE found throughout the event (for the given gridpoint). Note that since the maximum of CAPE$'_{U}$ does not always coincide with the maximum of CAPE, its relative contribution have some relatively low values (e.g., 34% for the case of January 1999). The contribution of $q'_{U}$ is still significant, though, being of about 60% and higher for 5 out of 7 cases. Column (iii) shows the averaged contribution of CAPE$'_{U}$ for the build-up of CAPE 36h prior to the maximum intrusion. Values of zero CAPE$'_{U}$ were not taken into
Table 3.2: Relative contribution of $\text{CAPE}_U'$ to total CAPE (in %) for the dates and locations indicated (same as in Fig. 3.19). (i) $\text{max}[\text{CAPE}_U']/\text{CAPE}$, represents the ratio of the maximum $\text{CAPE}_U'$ in the time period ranging from two days prior to two days after the date indicated, to the total CAPE for this particular day. (ii) $\text{CAPE}_U'/\text{maxCAPE}$, is the contribution of $\text{CAPE}_U'$ when CAPE is maximum in this same time period. (iii) shows $\text{CAPE}_U'/\text{CAPE}$ averaged over 36h prior to the date shown. (iv) $\text{CAPE}_U'/\text{CAPE}$ for the date and location given. (v) shows $\text{CAPE}_U'/\text{CAPE}$ averaged over the 9 nearest gridpoints (to the location given). For this last case, only values greater than zero were included in the calculation. The values in brackets are the standard deviation of the average.

<table>
<thead>
<tr>
<th>DATE</th>
<th>location</th>
<th>(i)</th>
<th>(ii)</th>
<th>(iii)</th>
<th>(iv)</th>
<th>(v)</th>
</tr>
</thead>
<tbody>
<tr>
<td>15 Jan 1987</td>
<td>20°N 212.5°E</td>
<td>84.8</td>
<td>67.3</td>
<td>56.3</td>
<td>67.3</td>
<td>64.9 (11.2)</td>
</tr>
<tr>
<td>12 Feb 1991</td>
<td>17.5°N 235°E</td>
<td>74.4</td>
<td>74.4</td>
<td>68.7</td>
<td>64.7</td>
<td>64.5 (27.3)</td>
</tr>
<tr>
<td>16 Jan 1997</td>
<td>12.5°N 245°E</td>
<td>42.6</td>
<td>42.6</td>
<td>22.3</td>
<td>0</td>
<td>28.3 (10.2)</td>
</tr>
<tr>
<td>14 Feb 1997</td>
<td>15°N 215°E</td>
<td>59.4</td>
<td>59.4</td>
<td>46.4</td>
<td>37.9</td>
<td>18.9 (12.4)</td>
</tr>
<tr>
<td>14 Jan 1999</td>
<td>12.5°N 220°E</td>
<td>92.8</td>
<td>34.0</td>
<td>76.6</td>
<td>49.8</td>
<td>43.1 (16.6)</td>
</tr>
<tr>
<td>23 Jan 1999</td>
<td>15°N 202.5°E</td>
<td>99.9</td>
<td>75.6</td>
<td>89.4</td>
<td>80.1</td>
<td>80.1 (31.1)</td>
</tr>
<tr>
<td>28 Jan 2003</td>
<td>20°N 212.5°E</td>
<td>58.3</td>
<td>58.3</td>
<td>63.3</td>
<td>76.3</td>
<td>59.9 (32.9)</td>
</tr>
</tbody>
</table>

account only when the total CAPE was also zero. The average contribution of all 7 events is about 58%, and this value increases to about 64% when the weakest event is not included. Column (iv) shows the ratio of $\text{CAPE}_U'$ to total CAPE at the date and location given, and column (iv) gives the averaged value for the closest neighbouring gridpoints (values of zero CAPE and/or $\text{CAPE}_U'$ were excluded from the average), along with the standard deviation. There is quite a lot of variability in the contribution of $\text{CAPE}_U'$ to the total CAPE, at least for those specific dates. For example, for the case of January 1997 (which has the smallest standard deviation), the contribution of $\text{CAPE}_U'$ varied roughly from 20 to 40%, while for the case of January 2003 (highest standard deviation), the contribution of the upper level anomaly could range from 17% to $\sim$93%.

These results show that $\text{CAPE}_U'$ has a large contribution to the total CAPE in the stages previous to the onset of convection. The maximum contribution occurs before or at the same time of the maximum CAPE build-up, and the average contribution is about 60%.

60
Figure 3.19: CAPE (solid) and CAPE\textsubscript{U} (dashed) for the dates and gridpoints shown on the top of each plot
c. Vertical velocity

For the cases studied in this section, the vertical velocity calculation was more complicated due to proximity to the equator and the quite erratic signal of vertical velocity (e.g., see Fig. 3.15). The proximity to the equator has a twofold influence. On one hand it makes it complicated to handle computational issues related to stability (in the PV inversion) due to the change in PV signal from positive to negative as one crosses to the southern hemisphere. For that reason, all inversions were performed with the equator as the southernmost boundary. This limitation is reflected later when calculating $\omega$ using Eq. (3.5). On the other hand, the quasi-geostrophic theory itself loses its strength as it goes away from the mid-latitudes. Therefore, the results of this section should be analysed with extra caution.

As a first step we compare the results of solutions of Eq. (3.5) using reanalysis winds and temperature fields with the reanalysis $\omega$ field. The inverted $\omega$ does give a qualitatively good representation of the vertical velocity pattern, especially in the mid-troposphere. Closer to the ground ($\sim$925hPa) and at upper levels, the patterns of inverted $\omega$ get more fuzzy, as the values of reanalysis $\omega$ gets weaker.

Table 3.3 shows the area average of $\omega'$ and $\omega' U$ obtained by solving Eq. (3.5) and the relative contribution of $q'_U$ to the total vertical velocity field. In three out of the 7 cases there is a strong contribution of $q'_U$ in the mid-troposphere; two other show some contribution but not as strong. It is also interesting to note that there is significant contribution of $q'_U$ at low levels in 5 out of 7 cases, and a mixed result for the 700 hPa level. The cases of January 1987, February 1991 and January 2003 show a very strong contribution of $q'_U$ to the upward vertical velocity in all levels; they are also the cases that had strongest $q'_U$ (see Fig. 3.16). The cases of January 1997 and early January 1999 had a relatively weaker anomaly, and in those cases, $q'_U$ contributes to ascent in the lower and mid-tropospheric layers, but at 700hPa the contribution was very small and even negative (January 1999).
<table>
<thead>
<tr>
<th>DATE</th>
<th>Averaged Area</th>
<th>Level (hPa)</th>
<th>$\omega^\prime$</th>
<th>$\omega_U^\prime$</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td>15 Jan 1987</td>
<td>15-20°N 212.5-217.5°E</td>
<td>925</td>
<td>−0.054</td>
<td>−0.076</td>
<td>&gt; 100</td>
</tr>
<tr>
<td></td>
<td></td>
<td>700</td>
<td>−0.007</td>
<td>−0.023</td>
<td>&gt; 100</td>
</tr>
<tr>
<td></td>
<td></td>
<td>500</td>
<td>−0.076</td>
<td>−0.115</td>
<td>&gt; 100</td>
</tr>
<tr>
<td>12 Feb 1991</td>
<td>10-15°N 235-240°E</td>
<td>925</td>
<td>−0.111</td>
<td>−0.084</td>
<td>75.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>700</td>
<td>−0.153</td>
<td>−0.069</td>
<td>45.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>500</td>
<td>−0.323</td>
<td>−0.197</td>
<td>60.9</td>
</tr>
<tr>
<td>16 Jan 1997</td>
<td>7.5-12.5°N 240-245°E</td>
<td>925</td>
<td>−0.075</td>
<td>−0.036</td>
<td>47.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>700</td>
<td>−0.087</td>
<td>−0.015</td>
<td>17.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td>500</td>
<td>−0.160</td>
<td>−0.060</td>
<td>37.8</td>
</tr>
<tr>
<td>14 Feb 1997</td>
<td>7.5-12.5°N 210-215°E</td>
<td>925</td>
<td>−0.050</td>
<td>−0.007</td>
<td>13.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>700</td>
<td>−0.065</td>
<td>−0.001</td>
<td>1.9</td>
</tr>
<tr>
<td></td>
<td></td>
<td>500</td>
<td>−0.103</td>
<td>−0.009</td>
<td>9.0</td>
</tr>
<tr>
<td>14 Jan 1999</td>
<td>7.5-12.5°N 220-225°E</td>
<td>925</td>
<td>−0.057</td>
<td>−0.008</td>
<td>13.9</td>
</tr>
<tr>
<td></td>
<td></td>
<td>700</td>
<td>−0.058</td>
<td>0.024</td>
<td>&lt; 0</td>
</tr>
<tr>
<td></td>
<td></td>
<td>500</td>
<td>−0.156</td>
<td>−0.007</td>
<td>4.4</td>
</tr>
<tr>
<td>23 Jan 1999</td>
<td>12.5-17.5°N 220-225°E</td>
<td>925</td>
<td>−0.033</td>
<td>−0.017</td>
<td>52.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>700</td>
<td>−0.043</td>
<td>−0.028</td>
<td>64.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>500</td>
<td>−0.015</td>
<td>−0.020</td>
<td>&gt; 100</td>
</tr>
<tr>
<td>28 Jan 2003</td>
<td>10-15°N 235-240°E</td>
<td>925</td>
<td>−0.032</td>
<td>−0.043</td>
<td>&gt; 100</td>
</tr>
<tr>
<td></td>
<td></td>
<td>700</td>
<td>−0.012</td>
<td>−0.021</td>
<td>&gt; 100</td>
</tr>
<tr>
<td></td>
<td></td>
<td>500</td>
<td>−0.036</td>
<td>−0.035</td>
<td>98.9</td>
</tr>
</tbody>
</table>

Table 3.3: Area-averaged $\omega^\prime$ and $\omega_U^\prime$ obtained by solving Eq. (3.5); see text for details. % indicates the relative contribution of $\omega_U^\prime$ to $\omega^\prime$ (the total vertical velocity perturbation). Since both $\omega^\prime$ and $\omega_U^\prime$ can be positive or negative, the relative contribution can exceed 100% or be negative. In the first case, it means that $\omega_U^\prime$ actually contributes to the whole of the total vertical velocity in the same direction; in the second case, it acts in the opposite direction.

It is interesting to notice, though, that although the perturbation velocities in the case of 23 January 1999 (weakest anomaly of all cases) were very weak, the contribution of $q_U^\prime$ was quite strong. Overall, the effect of $q_U^\prime$ was to provide a negative perturbation vertical velocity in all levels, i.e., to promote vertical ascent.
3.6 Summary

In this chapter we have shown, using NCEP reanalysis, that the dynamical and thermodynamic structure of the atmosphere in intrusion events are consistent with theoretical expectations, even for intrusions that penetrate deep into the tropical region. We verified that Charney balance yields a remarkably good description of the balanced nature of the flow, even though the intrusions have a considerable amount of divergence (downstream of the trough) and are so deep in the tropical region. Furthermore, using PV inversion we have established formally a link between dynamical quantities that characterize convection (CAPE, $\omega$, static stability) and the PV anomaly due to the intrusion. There is a consistent pattern of CAPE accumulation and decreasing static stability due mostly to the upper-level PV anomaly prior to the outbreak of convection. The contribution of $q'U$ to CAPE through changes in the temperature before the onset of convection is larger than 50%. The rate of destabilization was found to be higher than 90% for the 48h before the deep penetration of the intrusion into the tropics. Estimates of $\omega'U$ were obtained by using the Q-vector form of the so-called $\omega$ equation, which is based in the quasi-geostrophic theory. The values obtained suggest that when the convection is fully developed there is a strong component of the upward vertical velocity due to the PV anomaly, whenever the anomaly is relatively strong. As the main area of investigation lies close to the equator (around 10°N), these results should be taken with caution. As the intensity of the anomaly decreases, it is not clear whether the low level ascent associated with $q'U$ would be enough for the release of CAPE and activation of convection. The above results support the hypothesis (put forward by Kiladis 1998) that the upper level PV initiates and supports convection in the tropical eastern Pacific by destabilizing the lower troposphere and causing upward motion ahead of the tongues.

There are, however, some unresolved issues. We have not distinguished the contribution
of differential advection (of moisture and temperature) to CAPE, static stability and $\omega$, nor the possible interactions between the different anomalies. Also, we have not investigated the role of convection on the evolution of PV. It is expected that the latent heat release and turbulent mixing due to the strong convection will affect the PV budget. In the next chapter we investigate some of these issues using the Fifth-generation Penn State/NCAR mesoscale model (MM5). In this model, the effects of diabatic heating can be included and also removed. In addition, by using PV inversion, it is possible to isolate the contribution of different anomalies and modify boundary and initial conditions in the model simulations, and to verify the impact of those anomalies in the evolution of the system.
Chapter 4

Numerical Model Simulations

4.1 Introduction

In Chapter 3 we presented a quantitative estimate of the effect of the upper level PV anomaly associated with an intrusion on dynamical and thermodynamical fields characterizing convection. These estimates were obtained using PV inversion diagnostics and they showed that intrusions play a key role in destabilizing the lower atmosphere, building-up CAPE and promoting vertical ascent.

In this Chapter, we use a mesoscale model to confirm the results obtained previously and also to address some of the issues that could not be resolved using PV inversion diagnostics. The PV inversion approach is only meaningful if we assume that the system evolves with weak non-linearity. This is an important assumption, since atmospheric dynamics is chaotic and therefore allow for multiple equilibria state, turning the causality issue a difficult task. For the intrusion evolution, non-linearity is weak specially prior to the outbreak of convection, however as the system evolves the non-linearities grow and wave breaking can occur. As convection evolves and matures, PV is no longer conserved due to diabatic and frictional effects. At this point the use of a numerical model is an appro-
appropriate and efficient approach to study the impact of the diabatic effects on the PV evolution and their interaction.

There are two aims of this chapter: (1) to investigate the role of key quantities (PV, latent heat) in the development of convection; (2) to investigate the role of latent heat in the evolution of PV. We use the “factor separation method” to evaluate the contributions of the upper level PV and latent heat to CAPE, $\omega$ and $S$, and compare these estimates with those obtained by PV inversion in the previous chapter. This method also allows to examine the changes in PV evolution due to latent heat release.

The chapter is divided as follow: Section 4.2 provides a brief description of the mesoscale model we use, and Section 4.3 presents the methodology, and Section 4.4 describes sensitivity analysis results relative to model configuration setup, i.e., sensitivity to the choices of cumulus parameterization, initial and boundary conditions, and gridsize. Section 4.6 and 4.7 present the results using the methodology proposed and discussion for selected intrusion cases. Section 4.8 discusses feedback aspects of latent heat and PV interaction and finally, Section 4.9 presents a summary of the chapter.

### 4.2 Model description

We use version 3 of the Fifth-Generation NCAR/Penn State Mesoscale Model (MM5; Dudhia 1993, Grell et al. 1995) to address the goals described in the Introduction. MM5 solves the equations of momentum, heat and continuity for a fully compressible, non-hydrostatic atmosphere. The vertical coordinate is a sigma-type terrain-following coordinate based on reference pressure, and there is the option of polar stereographic, Lambert or Mercator conformal map projections, each of which uses a mapscale factor that depends only on latitude. MM5 also has multiple nesting capability, in which nested domains are forced by the parent domain at their boundaries. This limited area mesoscale model has
been used for a variety of purposes that include theoretical and predictive simulations of atmospheric phenomena ranging cyclones and hurricanes to meso-beta and meso-gamma systems (2-200km, e.g., mesoscale convective systems, fronts, urban heat islands).

In our simulations we use a Mercator projection, true at 24°N and with central longitude varying depending on the location of the intrusion. A single domain with 167×111 gridpoints in the W-E × N-S directions, and resolution of 50km is used. This domains covers a region of fixed latitude of 0.22°-44.10°N, while the longitude is variable depending on the intrusion location (resolution is approximately of 0.45 degrees in the E-W direction).

MM5 requires the input of geopotential height, horizontal wind, temperature, relative humidity, sea-level pressure, sea-surface temperature, and snow-cover data for the initial and boundary conditions. We use NCEP reanalysis data, except for the two first variables. The geopotential and wind fields for both initial and boundary conditions were replaced by the balanced fields obtained by PV inversion of the NCEP reanalysis data as described in Section 3.2. The area used in the PV inversion spanned 45x25 gridpoints, with longitude range varying according to the intrusion case, and latitude span of 0-60°N. This was done to eliminate undesired high-frequency waves, such as gravity waves, from the initial condition. For the vertical resolution, we set the top at 50hPa and defined 31 unevenly spaced σ-levels. The spacing was such that it sampled slightly more the boundary layer and the upper troposphere.

Several physical parameterizations can be chosen in MM5: Cumulus parameterizations, planetary boundary layer (PBL), explicit moisture, radiation, and surface schemes. For the most of our simulations, we choose to use simple ice microphysics scheme (Dudhia 1993), cloud-radiation scheme for radiation (Dudhia 1993), five-layer soil model for surface scheme, MRF PBL scheme (Hong and Pan, 1996), and variable cumulus parameterizations. Those schemes interact with each other via cloud processes, radiative fluxes, surface physical properties (e.g., albedo) and dynamical quantities (e.g., temperature, wind,
etc.). An important feature of MM5 is that “no cumulus parameterization” can be chosen, and also, water vapor can be set to be passively advected. Those features are important for the sensitivity studies that we want to perform.

### 4.3 Methodology

Numerical models can be used to isolate the pure contribution of given “factors” - e.g., topography, surface heat fluxes and latent heat release, as well as the contribution of their interaction, to chosen atmospheric fields. In the case we are studying, the main purpose of the experiments is to investigate the impact of the PV anomaly on the development of the convection. Since latent heat release modifies the PV distribution, it is relevant to also investigate the interaction between latent heat release and PV.

Davis et al. (1993) argue that numerical model studies that compare simulations with and without latent heat are shortcoming because of artificial removal of this diabatic effect can affect all aspects of model integration. Stein and Alpert (1993) suggested a new approach to compare model results and isolate contributions of distinct “factors” to a certain field. The method, called “factor separation” (FSM) is based on the idea that a certain field is build by contributions of factors and their interactions. Here we apply FSM for the problem we want to investigate.

Consider the case of a predicted field $f$ that depends on factors $\alpha_1 = q_U'$, and $\alpha_2 = \text{latent heat (LH)}$. If each of these factors are multiplied by a changing coefficient $c_i$, $i = 1,2$, then $f$ is a continuous function of those coefficients, such that $f = f(c_1, c_2)$. The function $f$ can be decomposed as

$$f(c_1, c_2) = \hat{f}_0 + \hat{f}_1(c_1) + \hat{f}_2(c_2) + \hat{f}_{12}(c_1, c_2)$$  \hspace{1cm} (4.1)

The term $\hat{f}_0$ is a constant part, which is independent of $c$, and all the other terms are
c-dependent. Each variable $c_i$ can be any value between 0 and 1, and in this method, it is arbitrarily set to be either 1 or 0, meaning that the factor is ‘switched on’ or ‘off’, respectively. In the present case (2 factors) Eq. (4.1) yields the combinations: $f_0 = f(0,0) = \hat{f}_0$, $f_1 = f(1,0) = \hat{f}_1 + \hat{f}_0$, $f_2 = f(0,1) = \hat{f}_2 + \hat{f}_0$, and $f_{12} = f(1,1) = \hat{f}_{12} + \hat{f}_1 + \hat{f}_2 + \hat{f}_0$. From these, the solution obtained for each $\hat{f}$ is:

$$\hat{f}_0 = f_0, \quad \text{(4.2)}$$
$$\hat{f}_1 = f_1 - f_0, \quad \text{(4.3)}$$
$$\hat{f}_2 = f_2 - f_0, \quad \text{(4.4)}$$
$$\hat{f}_{12} = f_{12} - (f_1 + f_2) + f_0. \quad \text{(4.5)}$$

It is important to understand the meaning of the terms $\hat{f}$. The term $\hat{f}_0$ represents the part of the predicted field that is not dependent on any of the factors. The term $\hat{f}_i$ represents the fraction of $f$ that is induced by the factor $\alpha_i$, i.e., $\hat{f}_1$ is the part of $f$ solely induced by $q'_U$, and $\hat{f}_2$ due solely to $\alpha_2 = \text{latent heat}$. The factor $\hat{f}_{12}$ is the contribution due to the pure (non-linear) interaction between the upper level PV anomaly and latent heat release. In this method, it is assumed that there is no interaction with the background field\(^1\). This was also assumed in the analyses performed in Chapter 3, therefore the results obtained from this method may be compared to the PV inversion analyses.

### 4.4 Control run: 13-17 Jan 1987

The first step of the model analysis is to verify that the MM5 model can simulate intrusion events and to examine the sensitivity to the choice of parameters. For this purpose, we

---

\(^1\)A modification of this method to evaluate the potential nonlinearity of the basic system and the response of the model to the fractional effect of a factor was presented by Krichak and Alpert (2002), but was not explored here.
conducted a series of simulations for the intrusion case of 13-17 January 1987. This case was examined in detail in Chapter 3 and we wish to compare model results with analyses (discussed in Chapter 3). A “control simulation” (or control run) was performed using the setup described in Section 4.2 with no convective parameterization and balanced geopotential and wind fields as the initial and boundary conditions. The initial condition was set on 13 January 1987 00UTC, and the model was allowed to run for 120 hours, i.e., until 18 Jan 1987 00UTC. The option to use no convective parameterization is, in general, meaningful for scales that may resolve updrafts and downdrafts (typically gridsizes of \( \sim 10 \text{km} \)). It will be shown in Section 4.5 that the choice of convective parameterization does not play a critical role in producing convection, because the convection occurs in resolved scale.

Figure 4.1 shows the time sequence of PV and OLR for the period of 14-17 January 1987 12UTC for the control run. Comparison with analysis (Fig. 3.5) shows that PV evolution is fairly well simulated, with high-PV intrusion and deep convection occurring downstream of the PV tongue. In the PV field, perhaps the major disagreement with NCEP reanalysis data is the downstream side of the trough in the lat\( \times \)lon plane, in which the simulations cannot reproduce the sharper ‘kink’ that can be seen in PV from reanalysis. By the 15th, the simulated convection had a much smaller extent than the OLR data (compare Fig. 4.1 to Fig. 3.5). However, by the 16th, the cloud coverage was more similar to the data. The simulated OLR and vertical velocity are consistent: Rather than having a large, continuous area of upward motion \( \text{w} = dz/dt \) (shaded in Fig 4.2a, far left) model results yielded a very fragmented \( \text{w} \) field, which covered a smaller area than that of NCEP reanalysis (Fig. 4.2b, far left). The relative humidity output from the model represented fairly well the dry region (\( \text{RH} < 40\% \)) at the northwest and northeast of the intrusion, but for all remaining areas, the model shows an excessively moist atmosphere at 900hPa (not shown). Cross-sections of RH show that this moisture is concentrated in the lower troposphere, and there is a “column” of high RH below and slightly ahead of the intrusion.
Figure 4.1: Potential vorticity (PV contours of 1, 2, 4 and 8, in units of $10^{-6}$ m$^2$.K.kg$^{-1}$.s$^{-1}$) on 200 hPa, and OLR (< 210 W.m$^{-2}$ shaded), for control run, for 14-17 January 1987 12UTC.

but this high RH is concentrated in the lower troposphere, while the NCEP data shows a large area of high RH in the mid-troposphere mostly ahead of the PV tongue. The evolution of CAPE given by the simulation showed a high amount in the vicinity where convection takes place, but the maximum CAPE is located more to the south and well within the high PV tongue (more to the west compared to values calculated using NCEP data, see Fig. 4.2a, b central column).

The results above show that the control run is able to simulate the general characteristics of convection, CAPE accumulation, vertical ascent and destabilization of the atmosphere, which are the main aspects we have been focusing so far to diagnose convection. There is, of course, a large number of physical parameterization options and their combinations, and an even larger number of possible initial and boundary conditions. In the next section,
we attempt to compromise some of the possible choices, and compare with results of the control run to test the sensitivity of the model.

Figure 4.2: Potential vorticity (PV contours of 1, 2, 4 and 8, in units of $10^{-6} \text{m}^2\text{K kg}^{-1}\text{s}^{-1}$) on 200 hPa, and OLR (shaded < 210 W.m$^{-2}$) (left column), PV = 2PVU contour and CAPE (J.kg$^{-1}$), in intervals of 500 J.kg$^{-1}$. Lowest value is 500 J.kg$^{-1}$) (central column), and cross section at 20°N of PV, potential temperature $\theta$ (K, intervals of 5K) and vertical velocity $w$ (cm.s$^{-1}$, values greater than 2cm.s$^{-1}$ shaded (right column), for (a) MM5 control run simulation, (b) NCEP/NOAA data, and (c) MM5 simulation with initial and boundary conditions using NCEP data. On 15 January 1987 12UTC.
4.5  Sensitivity analysis

4.5.1  Initial and Boundary conditions

We compared a simulation using NCEP reanalysis data for geopotential and wind fields in the initial and boundary conditions with the control one. Using NCEP data for all fields showed that simulated OLR ahead of the high-PV tongue resembles better with the data (Fig. 4.2c). Since the wind fields provided by the PV inversion are weaker up to \(\sim 20\%\) (see section 3.4.1), it is likely that these differences are being reflected in the triggering mechanisms of the convection, as this is extremely sensitive to subtle differences in the thermal and dynamical structure of the atmosphere. Once convection is triggered, though, the control simulation also produces vigorous convection, covering a large area ahead of the intrusion (e.g., Fig. 4.1 for 16 January 1987 12UTC).

There are not substantial differences in \(\theta, w\) (e.g., Fig. 4.2a vs. 4.2c, far left), \(u, v\) and relative humidity fields (not shown) for runs initialized with NCEP and balanced fields. Zonal velocity fields were the most different from each other, but still not strikingly so. CAPE for the run with NCEP geopotential and wind fields in initial and boundary conditions is also not larger (in area and/or intensity) than equivalent with balanced fields (Fig. 4.2a,c central column).

4.5.2  Initial condition time

We also investigated the impact of change in the time of initialization for the simulation. We kept the setup as described in Section 2, with no cumulus parameterization, but the initial time is set to 12 January 1987 00UTC. There is a slightly larger area of convection for this case than for that started 24 hours later (Fig. 4.3a), however, differences in any other fields are not major. These results were also confirmed by the cross-sections of PV,
Figure 4.3: PV at 200 hPa (1,2,4,8 PVU) and OLR (< 210W.m\(^{-2}\) shaded) for simulations (a) beginning at 12 January 1987 00 UTC, and (b) with twice horizontal resolution, i.e., 25 km.

RH, θ, etc. (figures not shown).

4.5.3 Gridsize

Another experiment similar to the control run was conducted, with all parameters kept the same except with resolution of 25km, i.e., double of original resolution. PV, RH and w fields in this run show more fine scale structures, especially in the vertical velocity field and in the vertical structure of PV. In the latter, there is much more indenting in the PV tongue, and small blobs of PV in low levels, likely due to latent heat release. There is some improvement in the simulation of OLR compared to control run (slightly larger area of deep convection, see Fig. 4.3b, for 15 January 1987 12UTC), however, no major differences are detected.

4.5.4 Choice of cumulus parameterization

We compared the results of the control simulation with those using the following parameterizations: Betts-Miller (BM), Anthes-Kuo (AKuo) and Kain-Fritsch (KF). The PV evolution is almost identical for all MM5 simulations. The OLR evolution is very similar
among different schemes: All of them yield a smaller area of low OLR compared to NOAA data for 15 January 1987 12UTC (Fig. 4.4 top row), and all simulate vigorous convection thereafter. These are consistent with the CAPE distribution for each run, which were also very similar (not shown). A fairly high amount of CAPE is present in the vicinity of low OLR, with slight variations in the place and intensity with varying scheme. For example, KF was the scheme that yielded the largest area of deep convection on the 15th, and consistently, was the experiment that showed the largest CAPE values in that area. The OLR pattern is also consistent with the vertical velocity fields given by the simulations, which showed that, similar to the control run, rather than having a large, continuous area of upward motion all simulations yielded very fragmented w field which covered a smaller area than that given by NCEP reanalysis (e.g., Fig. 4.4 bottom row). Cross-sections of PV show that PV = 2 PVU does penetrate only until ∼300 hPa, and there is quite a lot of distortion in the 1 PVU line, much like the control run. There are some differences, though: For example, both AKuo and control runs yield thin PV filaments in the regions of or bordering high-RH values, but those filaments are not observed in KF or BM (Fig. 4.4 bottom row).

Another set of runs were performed, one without cumulus parameterization and another modified Kain-Fritsch, in which shallow convection was enabled (in all the previous runs, shallow convection option was turned off). The results showed that there is not any substantial difference in the amount of convection produced, and even a slight decrease in deep convection, for both cases. The general results were not significantly distinct from the previous runs.

The results above show that there is not a strong departure in the simulations from one scheme relative to another. Slingo (1998) showed that Betts-Miller scheme worked better to simulate convection in the eastern (tropical) Pacific as compared to Kuo scheme (using UK Universities Global Atmospheric Modeling Programme (UGAMP) General Circulation Model (UGCM)). Also, Emanuel (1993, 1994) have very strong criticism towards the
Figure 4.4: PV at 200 hPa (1,2,4, 8 PVU) and OLR (< 210W.m$^{-2}$ shaded) on 15 January 1987 12UTC (top row), and cross section at 20°N of PV (1, 2, 4 PVU), $\theta$ (intervals of 5K) and vertical velocity w (±2 cm.s$^{-1}$, positive values shaded), for simulations identical to the control run, except using (a) Betts-Miller, (b) Anthes-Kuo and (c) Kain-Fritsch cumulus parameterizations instead of none.

physics underlying Kuo parameterization. It was shown here that since the convection is resolved, the choice of cumulus parameterization does not cause significant differences in simulation results, with all schemes producing similar features for most atmospheric variables, and in particular, OLR.

4.5.5 Summary

In summary, all the above results show that convection is more sensitive to initial and boundary conditions (NCEP vs. balanced fields) than to the physical parameterizations or gridsize choices. The control simulation is able to reproduce the evolution of PV and OLR and yields consistent fields of CAPE, w, $\theta$. Although the run using NCEP reanalysis as initial and boundary conditions reproduces the analysis/observations better than runs using the balanced fields, we use balanced fields to initialize the runs because it will allow for a
balanced dynamical field without spurious waves even when $q'U$ is removed, and therefore, results will be self-consistent and comparison between different runs will be meaningful.

### 4.6 Factor Separation Analysis

Following the methodology proposed in Section 4.3, four simulations were performed in order to investigate the impact of $q'U$, latent heat release (LH) and their interaction in the development of convection. All simulations were initialized on 13 January 1987 00UTC, with balanced geopotential and horizontal velocity fields in both initial and boundary conditions. The physical parameterization options were kept the same for all runs (no cumulus parameterization, simple-ice microphysics scheme, cloud-radiation scheme for radiation, five-layer soil model for surface scheme, MRF PBL scheme, see Section 4.2 for references). The simulations are as follow:

1. control simulation (**CNTL**, $f_{12}$ in Stein and Alpert 1993 nomenclature), which is the same as defined in Section 4.4
2. simulation with $q'U$ only (no latent heat) (**UPV**, $f_1$)
3. simulation with latent heat only (no $q'U$) (**LH**, $f_2$)
4. simulation without $q'U$ and without latent heat (**OTHR**, $f_0$)

In the runs **LH** and **OTHR**, $q'U$ was “removed” from the initial and boundary conditions in the following manner: First, a time-averaged “basic state” was defined (for the present case we used the 5-day mean as in Chapter 3). Then, we select the levels in which a PV anomaly equal or greater than 1 PVU is found for the day of maximum (deepest) vertical intrusion (400-100hPa for this case). Next, we replace the geopotential and wind fields by the mean field. Finally, we perform PV inversion of this “altered” distribution (mean
field replacing the PV anomaly) and use the resulting balanced geopotential, wind and temperature fields as the initial and boundary conditions for those runs. Also, in these runs upper-air and surface data were not incorporated, to avoid inconsistencies between the observed and idealized fields. Using the balanced fields was a necessary measure to minimize spurious results that could arise due to sharp gradients in the fields where the anomaly was removed and the level immediately below or above.

Figure 4.5 top row shows the PV and OLR distributions on 15 January 1987 12 UTC for runs UPV, LH and OTHR. It is clear that the convection must be related to the presence of $q'_U$, since there is not a signal of OLR when $q'_U$ is removed. The impact of intrusion on the thermal and vertical velocity fields can be seen in Fig. 4.5 bottom row: When the intrusion is removed, the thermal structure is nearly undisturbed throughout the depth of the troposphere.

This result can be seen also in Fig. 4.6, which shows the time evolution of OLR, dry static stability $S$, CAPE and vertical velocity $w$ ($=dz/dt$), for the four simulations. The
values are area-averaged because, for most of these parameters, there is a lot of fine-scale structure. Therefore, an area average should be more representative than a point value. For each parameter a 5°x5° area was chosen, as indicated at the top of each plot. These regions differ slightly for each variable and were chosen based on visual inspection of individual maps for each quantity, such that it would include the larger area of maxima (or minima, depending on the parameter) that the box could contain.

The time sequence of OLR (Fig. 4.6a) for run CNTL shows that its value begin to decline substantially after the 15th, reaching a deep convection threshold (around 205 W.m⁻², e.g., Gu and Zhang, 2002) at around 12 UTC this day. This plot also shows that the OLR sequence from the run UPV matches closely the control run until it reaches the deep convection threshold, while for the remaining runs, OLR values remain very high (i.e., no deep convection present) throughout the whole simulation period. This behavior is similarly observed in the sequence of $S$ (Fig. 4.6b). The variation of the static stability in the run UPV follows the curve of CNTL prior to the onset of convection. For the other two runs, the decrease of static stability prior to the 15th is weaker compared to run UPV.

The time sequence of CAPE (Fig. 4.6c) shows a clear CAPE build-up prior to the onset of convection for runs CNTL and UPV. The run with only latent heat (LH) shows an early CAPE build-up, but a decrease subsequently, while UPV shows a sharp increase immediately before the convection occurs. This suggests that CAPE is built up in response to thermodynamical changes due both to latent heat release and the presence of the PV anomaly, with each contributing more or less depending on the stage of evolution of the system. It is interesting to see that there is also a small amount of CAPE unrelated to either $q'_U$ and water vapor content. It may be related to changes in the thermodynamical structure due to turbulence and/or surface heat fluxes.

Figure 4.6d shows the time sequence for the vertical velocity $w$. Until about 24h before the onset of convection there was, on average, weak subsidence in this area, followed by
Figure 4.6: Time sequence of area-averaged (a) OLR (W.m\(^{-2}\)), (b) dry static stability \( S \) (=g \( \frac{d\theta}{dp} \), in K.m\(^{-2}\).kg\(^{-1}\)), (c) CAPE (J.kg\(^{-1}\)), (d) vertical velocity \( w \) (=dz/dt, cm.s\(^{-1}\)), (e) latent heat flux (W.m\(^{-2}\)), and (f) \( \theta_e \) (K). Results from MM5 simulation, including \( q_U' \) and latent heat release (solid line), including \( q_U' \) but no latent heat (dashed line), including latent heat but no \( q_U' \) (dotted line), and removing both \( q_U' \) and latent heat (dot-dashed line).
a relatively strong ascent, in the control simulation. Until the 14th 12UTC $w_{UPV}$ is always less negative than the other simulations. Actually, run $UPV$ is the only one other than $CNTL$ that showed positive $w$ (i.e., ascent), even after the 15th.

The next step is to perform analysis of the results in a similar manner as done in the previous chapter, i.e., to quantify the relative contributions of each factor to OLR, CAPE, $w$ and $S$. The results from this analysis will allow a direct comparison with results obtained in the previous chapter, and further reinforce the proposed hypothesis of convection being triggered by upper level high-PV anomalies.

Equations (4.2)-(4.5) were solved to obtain the contribution of each factor to a given field. In the following, INTR will refer to the interaction of $q'U$ and latent heat ($\hat{f}_{12}$) and the contribution unrelated to either $q'U$ or latent heat is referred to as OTHR ($\hat{f}_0 = f_0$).

a. OLR

The contribution of each factor to the total OLR depends on the value of “clear sky” OLR ($OLR_{cs}$). There is not a unique value, since $OLR_{cs}$ depends on the surface temperature (see, e.g., Gu and Zhang 2002). One possible value would be the value of OLR in the run OTHR, since in this run no convection developed, and it has quite a high value of OLR (time mean of the area average of OLR in run OTHR was $\sim 265\, \text{W.m}^{-2}$). Following Gu and Zhang (2002), estimates of $OLR_{cs}$ using surface temperatures of 295K and 300K yields clear-sky OLR values of 275.12 W.m$^{-2}$ and 284.14 W.m$^{-2}$ respectively. Results relative to each of these choices as shown in Table 4.1.

A negative contribution of a factor means that this factor acts to decrease the OLR, i.e., contributes towards the deep convection threshold. For example, if we consider that clear-sky OLR is 265 W.m$^{-2}$, then the contribution of $q'U$ acts to lower the OLR to $\sim 201\, \text{W.m}^{-2}$, meaning that there deep convection would be present. From the relative contribution analysis it is not possible to determine whether deep convection is achieved, however, it clearly
Table 4.1: Absolute (second column) and relative contributions of factors and their interactions to OLR. OLR\textsubscript{cs} refers to the clear-sky value, and OLR\textsubscript{fac} the contribution due to each factor. Values are averages over the area 17.5-22.5\textdegree N, 215-220\textdegree E, on 15 January 1987 12 UTC.

<table>
<thead>
<tr>
<th>run/contr.</th>
<th>abs.value</th>
<th>OLR\textsubscript{cs} (W.m\textsuperscript{-2})</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>285</td>
</tr>
<tr>
<td>CNTL</td>
<td>195.29</td>
<td>89.7</td>
</tr>
<tr>
<td>q'\textsubscript{U}</td>
<td>-61.14</td>
<td>68.1</td>
</tr>
<tr>
<td>LH</td>
<td>-0.88</td>
<td>0.9</td>
</tr>
<tr>
<td>INTR</td>
<td>-5.57</td>
<td>6.2</td>
</tr>
<tr>
<td>OTHR</td>
<td>262.9</td>
<td>-</td>
</tr>
</tbody>
</table>

shows that the upper level PV anomaly has by far the most significant contribution acting to lower OLR.

b. Static Stability

The contribution of each factor and their interaction to the time variation of static stability are shown in Tables 4.2 and 4.5 (first column, second row). The static stability is an average of the values between 850-500 hPa, and over the area 17.5-22.5\textdegree N, 210-215\textdegree E. There is a steady decrease in static stability in the control run before the 14th at 12 UTC (see Fig. 4.6), with q'\textsubscript{U} contributing with \sim 88\% to the total destabilization until this point (later it still contributes with \sim 83\% of the total negative trend). Both LH and q'\textsubscript{U} help destabilize the atmosphere, but INTR does not contribute to destabilization, while factors unrelated to both LH and q'\textsubscript{U} act overall to favor destabilization.

c. CAPE

CAPE associated to each factor are given in Table 4.3. The first column shows the maximum value attained by each factor in the time period of 13 January 1987 00UTC to 17 January 00UTC. Note that these values do not coincide in time for each factor. For example, the maximum CAPE due to q'\textsubscript{U} is found at 15 January 12 UTC, while for LH,
The maxima in CAPE is found much earlier. The values given show that the maximum $q_U'$ contribution is at least three times more than the next larger contribution (LH). The second column shows the value of CAPE in the control run when the CAPE associated with a particular factor was maximum. It shows that, for example, CAPE due to $q_U'$ accounts for the whole CAPE at that time, while even when LH is maximum, it accounts for less than 50% of the total CAPE at that time.

The third and fourth columns of Table 4.3 show the contributions of the factors to CAPE when it is maximum (15 January 06UTC; 555.31 J.kg$^{-1}$, shown in the last row), and the average of 36h prior to convection, i.e., from 14 January 00UTC until 15 January 12UTC. The averaged values indicate that although $q_U'$ has the most important contribution ($\sim 53\%$). Combined with the results given in the first two columns, they show that the relative contribution of factors throughout the period preceding the convection varied, and contributions from LH and OTHR also played a role.

Negative values were found for the contribution of INTR. They arise because of the mathematical formulation of the analysis methodology and do not have a physical interpretation, except that INTR does not contribute at all to total CAPE. It should not be interpreted as consumption of CAPE because convection was not present before the 15th, and
Table 4.3: Area-averaged (15-20°N, 207.5-212.5°E) contributions of factors to total (CNTL) CAPE, in J kg\(^{-1}\). First column (maxCAPE\(_{fac}\)) shows the maximum CAPE attained by each factor between the period 13 January 00UTC to 17 January 00 UTC. The second column shows CAPE\(_{CNTL}\) when maximum CAPE due to each factor is found. Third column shows the contributions for the maximum CAPE\(_{CNTL}\) and the last column, the average of first 36 hours.

<table>
<thead>
<tr>
<th>factor</th>
<th>maxCAPE(_{fac})</th>
<th>CAPE(_{CNTL})</th>
<th>CAPE(_{fac})</th>
<th>aveCAPE(_{fac})</th>
</tr>
</thead>
<tbody>
<tr>
<td>(q'_U)</td>
<td>644.73</td>
<td>521.73</td>
<td>541.77</td>
<td>228.84</td>
</tr>
<tr>
<td>LH</td>
<td>126.31</td>
<td>539.43</td>
<td>79.73</td>
<td>75.06</td>
</tr>
<tr>
<td>INTR</td>
<td>86.62</td>
<td>309.16</td>
<td>−172.99</td>
<td>−30.96</td>
</tr>
<tr>
<td>OTHR</td>
<td>118.83</td>
<td>408.47</td>
<td>104.79</td>
<td>95.41</td>
</tr>
<tr>
<td>CNTL</td>
<td>553.31</td>
<td>553.31</td>
<td>553.31</td>
<td>428.35</td>
</tr>
</tbody>
</table>

therefore, there were no mechanism to burn up the available potential energy.

d. Vertical velocity

Vertical velocity values for control run, and contributions of \(q'_U\), LH, INTR and OTHR on 15 January 1987 12UTC are given in Table 4.4. In the mid-troposphere (500hPa) the most important contribution to the vertical velocity is \(q'_U\), followed by the interaction of it with latent heat release. At 700 hPa, \(q'_U\) contribution drops to \(~21\%\) and is surpassed by the contribution of the interaction term. At low levels, INTR is the only term that contributes in the same direction of the total vertical velocity (control run).

e. Surface fluxes

To further ascertain that the upper level PV anomaly was the crucial element to the development of convection rather than surface conditions, we show in Fig. 4.6e the time sequence of the latent heat flux for the same area and period of that for CAPE (15-20°N, 207.5-212.5°E). It is expected that latent heat flux would be enhanced before convection is activated (e.g., Zhang 2003, Raymond 2001). We see indeed that there is an increase in the LH flux in the 24 hours preceding the convection (14 to 15 January 12 UTC) for run CNTL.
Table 4.4: Vertical velocity for control run and due to $q'_U$, latent heat, their interaction and to neither of those, averaged over the area 15-20°N, 210-215°E, on 15 January 1987, 12UTC.

<table>
<thead>
<tr>
<th>factor</th>
<th>500 hPa</th>
<th>700 hPa</th>
<th>900 hPa</th>
</tr>
</thead>
<tbody>
<tr>
<td>CNTL</td>
<td>3.99</td>
<td>2.26</td>
<td>0.92</td>
</tr>
<tr>
<td>$q'_U$</td>
<td>2.44</td>
<td>0.57</td>
<td>-0.19</td>
</tr>
<tr>
<td>LH</td>
<td>-0.21</td>
<td>-0.34</td>
<td>-0.03</td>
</tr>
<tr>
<td>INTR</td>
<td>2.27</td>
<td>2.72</td>
<td>1.27</td>
</tr>
<tr>
<td>OTHR</td>
<td>-0.52</td>
<td>-0.68</td>
<td>-0.12</td>
</tr>
</tbody>
</table>

however OTHR has the higher latent heat flux and still was unable to trigger convection. Run LH had values of LH flux very similar to CNTL and also failed to produce convection.

These results are corroborated by the evolution of the equivalent potential temperature ($\Theta_e$) at 2m for the same area and period (Figure 4.6f). Despite the fact that $\Theta_e$ evolution is very similar in runs CNTL and LH, the latter did not present any convection, reinforcing the hypothesis that surface processes by themselves do not have a sufficient impact in triggering convection, and that the presence of $q'_U$ is of fundamental importance for that, at least for this case.

\textit{f. Comparison with results from PV inversion analysis.}

The model results obtained in this Section allow a direct comparison with results from analyses presented in Chapter 3. The gridpoints/areas generally do not coincide, but the comparison is still justified because in both cases, the gridpoint/area used is representative of the region of maximum vertical ascent, maximum cape, minimum OLR, etc. Note that this comparison is only possible for values relative to $q'_U$, since this is the only contribution that was directly inferred using the PV inversion technique.

In some aspects there is qualitatively good agreement from the results obtained here and in the previous chapter. For example, the contribution of $q'_U$ to CAPE increases with time, and is dominant in the 24 hours preceding the onset of convection. Also, it causes a
decrease in static stability prior to convection, and vertical ascent in the mid-troposphere at the leading edge of the PV intrusion. In a qualitative fashion, therefore, there is a good agreement between results obtained in the previous chapter with those presented here.

In terms of absolute values, there are differences between the methods results. For example, for the static stability local tendency parameter, there are discrepancies relative to the PV inversion analyses. The decrease in static stability for the control run is much smaller (by at least a factor of 3) than that given by the analyses. Careful inspection of $\theta$ distribution (e.g., from Fig. 4.1) shows that there is indeed much more distortion of $\theta$ in the NCEP data than given by the simulations. Also, since the vertical penetration of PV in the simulation is weaker, the destabilization in the lower levels of the atmosphere is weaker, as expected. The results using PV inversion technique showed that $q_U'$ contributed 76% of the static stability decrease in the period 14-15 January 00UTC, and to 100% in the 48 hours preceding the convection (Table 3.1). Results given by the simulations were in very close agreement, showing that the contribution of $q_U'$ in the period 13-15 January 12UTC was of 73%.

The analyses showed that the maximum value of CAPE due to $q_U'$ was $\sim$85% of the CAPE at the same time (Table 3.2), compared to $>100\%$ from model results (Table 4.3). When total CAPE was maximum, the contribution of $q_U'$ was about 98% (PV inversion showed that the maximum of CAPE due to $q_U'$ did not coincide in time with the maximum of total CAPE). The averaged contribution for the 36 hours prior to the convection was 62% for the model result, and 56% using PV inversion method.

With respect to the vertical velocity, $q_U'$ is dominant in the mid-troposphere (500hPa), with values of 61% and 100% from the simulations (Table 4.4) and analyses (Table 3.3), respectively. For lower levels in the troposphere, PV inversion analyses still yields full contribution of $q_U'$ for the vertical ascent (i.e., $>100\%$), while the simulations give about 25% at 700hPa and no contribution at 900hPa. Again, this may stem from the fact that the
vertical penetration of PV tongue was not as strong in the simulation results, resulting in lesser influence in the lower levels.

Results using MM5 simulations provide further insight in the role of other factors in convection and the PV evolution. For example, it shows that latent heat alone makes a small contribution in providing energy for convection in the early period and it hinders vertical ascent. The role of the interaction of $q' U$ and LH also has mixed results: it does not contribute at all to CAPE build-up (Table 4.3), but has as strong or stronger contribution to the upward vertical motion than $q' U$ alone (57% at 500 hPa and > 100% at 700 and 900 hPa). Factors unrelated to either $q' U$ and LH have non-negligible contributions for CAPE prior to convection (~22%), at the same time that causes relative stabilization of the atmosphere and acts to hinder vertical ascent.

### 4.7 Additional Cases

We next perform the same diagnostics to other selected cases, also shown in Chapter 3: 12 February 1991, 16 January 1997, 23 January 1999, and 28 January 1993. A series of experiments similar to that performed for the case of January 1987 were performed for these cases, i.e., control simulation (CNTL, with upper level PV anomaly and latent heat included), $q' U$ only (UPV, no latent heat), no $q' U$ (LH), and $q' U$ and latent heat excluded (OTHR). All simulations have 167x111 horizontal gridpoints with 32 vertical (sigma) levels, no convective parameterization and all other physical parameterization choices the same as before (see Section 4.2). One difference was that the time average mean field for these cases to feed the initial and boundary conditions for runs LH and OTHR was a one-month average, rather than a 5-day mean. Also, for the case of February 1991, the upper level PV anomaly was removed between 400-100 hPa for runs UPV and OTHR, while for the other 3 cases, the anomaly did not penetrate very deep into the troposphere, and PV was removed.
between 300-100 hPa.

The control runs for these events showed similar general patterns: The PV tongue is fairly well simulated, with convection developing downstream of the intrusion. Figure 4.7a) shows PV at 200 hPa and OLR, for each case’s CNTL (compare with Fig. 3.15). Similar to case of January 1987, the vertical velocity has very fragmented areas of upward and downward motion (Fig. 4.7b)). As convection organizes and develops, there is a corresponding increase in CAPE nearly coincident with low OLR area, except for the case of January 1999, which show a quite unexpected CAPE accumulation to the west of the intrusion, and not coinciding with the convective area (Fig. 4.7d).

Figure 4.8 show the time evolution of OLR, static stability, CAPE, vertical velocity, surface latent flux and surface equivalent potential temperature for the four runs (CNTL, UPV, LH, and OTHR), for each case. The most striking result shown here is the behavior of OLR for run UPV for cases of January 1997 and January 2003, which remains high throughout the simulation period (Fig. 4.8a). We calculated the relative contributions of $q'U$, latent heat and their interactions to the total fields, and Table 4.5 summarizes the results obtained for these cases.

We notice that, for the cases where the intrusion reaches deeper vertically, i.e., for January 1987 and February 1991, the results are more uniform. For these deep events, there is a consistent pattern of decrease in OLR mainly due to $q'_U$ (90% and 100%, respectively), decrease in static stability (100% contribution of $q'_U$ for destabilization) for some time before the convection is activated, CAPE build up (62% and 100%, for the period of 36h from beginning of simulation) and upward vertical velocity (61% and 55%). Therefore, $q'_U$ is clearly the dominant factor in these events.

For the shallower intrusion cases (January 1999, January 1997 and January 2003), the results obtained here reveal a mix of dominant influences, mainly of $q'_U$ and its interaction
Figure 4.7: (a) PV (1, 2, 4, 8 PVU) at 200hPa, and OLR (≤ 210W.m\(^{-2}\) shaded), for the dates indicates on the top of each panel. (b) 2 PVU contour and vertical velocity \(w (=dz/dt, 5, 10, 15 \text{ cm.s}^{-1}, \text{ positive values shaded})\). (c) 2 PVU contour and CAPE (500, 1000, 1500 J.kg\(^{-1}\)), and (d) cross sections of PV (1, 2, 4 PVU) (thick solid), potential temperature (5K intervals) and relative humidity (≤ 80 % dark gray, ≤ 40% light gray), for dates: (1) 12 February 1991 12UTC, cross section at 17.5°N, (2) 23 January 1999 12UTC, cross section at 17.5°N, (3) 16 January 1997 12UTC, cross section at 17.5°N and (4) 28 January 2003 12UTC, cross section at 15°N.
Table 4.5: Control values and relative contributions of upper level PV ($q_U$), latent heat (LH), their interaction (INTR), and factors unrelated to any of them (OTHR) for OLR (W.m$^{-2}$), tendency of $S$ (in units of 10$^{-5}$ K.m$^2$.kg$^{-1}$.h$^{-1}$) for the period of 48 hours prior to the date on top, CAPE (J.kg$^{-1}$), and $w$ (cm.s$^{-1}$) at 500 hPa. OLR (W.m$^{-2}$), CAPE (J.kg$^{-1}$) and $w$ (cm.s$^{-1}$) are values on the dates indicated in top row.
with latent heat. For example, for the case of January 1999 $q'_U$ contribution was again far dominant over the other components, except for CAPE. However, for the cases of January 1997 and January 2003, there are mixed results. The most striking result is that for these cases, the simulation with $q'_U$ only does not yield convection (though LH and OTHR also did not; see Fig. 4.8). The analysis shows that for these events, the interaction of $q'_U$ and latent heat was of crucial importance for lowering OLR (see OLR in Table 4.5). CAPE build-up is mainly due to $q'_U$, and $q'_U$ contributes to destabilize the lower-mid atmosphere in all cases, even when static stability in control run was increasing (January 1997). The vertical velocity again show mixed results, with the main upward contribution being in most cases from $q'_U$ and on January 1997 case, due to its interaction to latent heat.

Results obtained here can again be compared with results obtained in Chapter 3, for the contribution of $q'_U$. According to the PV inversion, changes in static stability were negative for all cases and the contribution of $q'_U$ was nearly 100% for decrease. The contribution of $q'_U$ to CAPE were about or greater than 60% except for January 1997, which was very low (about 28%). The present analysis show that $q'_U$ also promoted destabilization in all cases; the contributions ranged from 34 to 100% relative to the total negative contributions (notice that sometimes the control run had increase in $S$, while individual contributions could be negative). The upper level PV anomaly had an even larger contribution to total CAPE than estimated before ($>70\%$) in all cases but for January 1999, which had a low contribution (about 19.5%). The vertical velocity fields also showed a good agreement between the results here and from PV inversion diagnostics. Contribution to $w$ at 500 hPa for cases of 1991, 1997, 1999 and 2003 were respectively 60.9%, 37.8%, 100%, and 98.9% for the latter case. The FSM analysis yields contributions of 55.8%, 0 (zero), 100% and 86.2%, comparatively (the last value is actually relative to the total positive velocity, since the CNTL value was negative).

With respect to surface processes, it seems that there is not a clear pattern of either LH
flux or $\theta_e$ that is consistent in one run compared to another. The case of February 1991 is very similar to January 1987 case: UPV has the lowest LH flux however is able to trigger convection, whereas LH and OTHR have larger LH flux however does not yield convection. In the case of January 1997 there isn’t convection in any run except CNTL, despite the fact that LH flux is higher in all other runs after the 16th (Fig. 4.8fII). However, the same conclusions cannot be drawn for the other 2 cases. From these analyses it is not clear what is the role of the surface processes in triggering the convection.

The results from the above cases provide strong evidence that the presence of the PV intrusion (i.e., $q'U$) is of fundamental importance for convection to occur. These numerical model results show however, that sometimes latent heat and/or the interaction of $q'U$ with latent heat have a stronger effect on CAPE or $w$ than the effect of $q'U$. This implies that $q'U$ is a necessary - though not always sufficient - condition for the convection to develop. When the intrusion is vertically deep enough the sufficient condition seems to apply, whereas when the intrusion is ‘shallow’, it may not be sufficient to promote convection without the interaction with latent heat.

### 4.8 Latent Heat and PV evolution

#### 4.8.1 Background

Potential vorticity is conserved in adiabatic and frictionless conditions. Although those processes may be neglected in some circumstances rendering PV as a tracer, this is not true for the case studied, specially during and after the onset of convection, where turbulent mixing, latent heat release and cloud effects may cause material changes in PV. The most important sources and sinks of PV due to diabatic effects are convective regions with precipitation and regions with large population of shallow stratocumulus clouds (Krishna-
murti et al. 2000, Brunet et al. 1995). Globally, diabatic sources and sinks of PV have an important impact in maintaining the general circulation (Hoerling 1992). Locally, they are important in cross-tropopause mass-exchange (e.g., Lamarque and Hess, 1994), as the latent heat in the cloud interior and radiative cooling from cloud top result in a transport of PV from the tropopause level to middle and lower tropospheric levels (HMR).

The Eulerian change (or, tendency) of PV can be expressed as:

\[ \frac{d(PV)}{dt} = \frac{\partial PV}{\partial t} + v \cdot \nabla PV = \frac{1}{\rho} \dot{\zeta}_a \cdot \nabla \dot{\theta} + \frac{1}{\rho} K \cdot \nabla \theta \]  (4.6)

where \( v \) is the three-dimensional wind velocity field, \( \dot{\zeta}_a \) is the absolute vorticity, \( \dot{\theta} \) is the diabatic heating term, and \( K \) is the curl of the frictional force per unit mass. The first term in the r.h.s. represents diabatic processes, which include effects of radiative heating/cooling, latent heat release and heat conduction, which is usually neglected. The last term refers to the diffusive terms, i.e., molecular dissipation (in most cases neglected) and turbulent mixing.

Let us consider a flow under small Rossby number and large Richardson number conditions. Then the contributions of the vertical component of the dot products in Eq. (4.6) dominate over the horizontal ones, and

\[ \frac{d(PV)}{dt} \simeq \frac{1}{\rho} k \cdot \dot{\zeta}_a \left( \frac{\partial \dot{\theta}}{\partial z} \right) + \frac{1}{\rho} k \cdot K \left( \frac{\partial \theta}{\partial z} \right) \]  (4.7)

The diabatic term shows that when there is latent heat release, i.e., heating, there will be production of PV at lower levels (where \( \partial \dot{\theta} / \partial z \) is positive) and destruction of PV at the top (\( \partial \dot{\theta} / \partial z < 0 \)). Conversely, when there is cooling, the reverse happens. Therefore, the contribution of the diabatic heating on PV changes depends on the strength of latent heat release (heating) compared to the radiative cooling. For the friction term, the production or destruction of PV will depend on whether \( K \) is “cyclonic” or “anticyclonic”.
If we consider a material volume $\Lambda$ with surface $S$, then the integration of Eq. (4.6) over $\Lambda$ yields the integral statement (using the relations $\nabla \cdot \zeta_a = \nabla \cdot K = 0$, $\rho PV = \nabla \cdot (\zeta_a \theta)$, etc.)

$$\frac{d}{dt} \int_\Lambda PV \rho d\Lambda = \int_S (\dot{\theta} \zeta_a + \theta K) \cdot \mathbf{n} dS$$  \hspace{1cm} (4.8)

The interpretation of Eq. (4.8) is that diabatic and frictional forces inside $\Lambda$ can only redistribute $PV$, and $PV$ changes in the interior of the domain occurs only when there are non-zero values of $\dot{\theta}$ or $K$ over the domain $S$. Therefore, when latent heat release heating dominates over radiative cooling, the net effect of convection is to (locally) reduce the strength of the upper level PV anomaly and “move it” to lower levels. HMR points out that the “most dramatic effects” of $\dot{\theta}$ occur near the tropopause, because in the lower stratosphere PV values are too large and the diabatic term is comparatively too small for any reasonable estimate of $\dot{\theta}$. If there is a vertical redistribution of PV due to diabatic effects, and destruction of PV by surface friction, then the upper level anomaly can be significantly weakened. In fact, estimates given by HMR suggest that in midlatitudes, the combined effect of diabatic heating and surface friction have a very large impact on the timescale of the upper-level PV anomalies.

The above discussion describe the contribution of condensation to the PV distribution as a localized source due to vertical variations in heating. Another contribution arises from the advective effect by the irrotational flow, which is in part related to heating (Hoerling, 1992). Moreover, the redistribution of PV due to diabatic processes can affect the evolution of a system by a number of ways, e.g., (i) advective feedback from the surface thermal anomaly, (ii) direct reduction of PV aloft by latent heat, (iii) latent heat influence on the surface potential temperature, which in turn alters its induced effects at upper levels, and (iv) enhancement of upper-level cyclonic wrapping of PV anomalies (Ahmadi-Givi et al. 2004, Davis 1992b).

Sensitivity studies using numerical simulations show consistently that the effect of con-
densation on the evolution of atmospheric systems is to decrease the effective stability of ascending air and to increase the growth rate (Davis et al 1993, and references therein). However, these studies also show that there is a large case-to-case variability based on the location and strength of the latent heating. (e.g. Davis 1992b, Davis et al. 1993, Ahmadi-Givi et al. 2004). For example, Davis (1992b) and Davis et al. (1993) propose that the effect of condensational heating on a system development can be addressed in terms of superposition of PV anomalies of different sources. Using this approach, they showed that the primary effect of condensational heating at low levels is to superpose a positive PV anomaly above the low level baroclinic zone causing its intensification. Ahmadi-Givi et al. (2004) found that low-level PV anomalies associated with latent heat release had as strong contribution to the intensification of a surface cyclone as the upper level PV anomaly, although the upper level feature was identified as crucial for the initiation of the system.

A more complicated question is the impact of non-linear interactions of PV and latent heat. Results are case-dependent, with some showing that mutual intensification between upper and lower (diabatically produced) PV perturbations was important (Davis and Emanuel 1991, Stoelinga 1996), while in others the upper level PV anomaly controlled the entire development (Davis 1992b, Davis et al. 1993). In the case study shown by Ahmadi-Givi et al. (2004), the latent heat inhibited downward penetration of the upper-level PV anomaly and decreased the horizontal extent of the tropopause fold. Therefore, there seems to be a dual role of the interaction of PV and latent heat, first by intensifying the surface disturbance but ultimately contributing to the decay of the upper-level PV anomaly and subsequent cycloysis.

### 4.8.2 Results

The numerical simulations results presented in previous sections provide strong evidence that the presence of the upper level PV anomaly leads to atmospheric destabilization,
CAPE build-up and vertical motion, which ultimately cause the outbreak of convection. The next step is to investigate the feedback effect of the convection on the PV evolution itself. Once convection is activated, PV is no longer conserved because the diabatic effects due to latent heat and radiative cooling/heating cannot be neglected.

In this section, only the impact of latent heat release on the PV evolution is examined. We first compare the evolution of PV in runs CNTL and UPV for the case of January 1987. Figure 4.9 shows snapshots of PV and OLR for these runs for three stages of evolution of the intrusion. When convection is not present yet (14 January 12UTC) the two runs yield very similar PV and OLR fields. However, as the convection develops (16 January 12UTC), there is a clear signal of low OLR in both runs with difference in the position and extent of convection. Also, the intrusion is more tilted and narrow in run CNTL, as opposed to a broader and more N-S oriented high PV ridge in run UPV. As the intrusion moves eastward, the low OLR distribution have a different pattern in each run, and the PV ridge is more pronounced in run CNTL. Also, the PV gradient downstream (where the cloudiness is located) is much weaker in run UPV compared with CNTL. We want to investigate how these differences are related to latent heat and its interaction with PV evolution itself.

Figure 4.10 shows the contributions of LH only and of interaction between PV and LH to total PV (shaded values are negative). By 14 January 12UTC there is not significant PV due to LH or INTR, because convection has not occurred yet. As the system evolves though, the pure contribution of LH is restricted to the northern edge of the domain and is one order of magnitude lesser than total PV. This suggests that LH alone is not enough to cause significant changes in PV. The maps of interaction of PV with latent heat show that the feedback between these two features is responsible for the narrower trough in the control run, with negative PV found in the west side of the trough. On 15 January 1987 12UTC, a negative PV area centered around 20°N 212.5°E is nearly coincident with the area of low OLR (not shown). The negative PV area become larger as the convection...
evolves (16 January 12UTC), and is basically placed downstream of the intrusion, suggesting that the interaction between the upper level PV anomaly and latent heat associated with condensation controls the shape of the PV tongue, by “eating away” PV downstream.

These results are confirmed by inspecting Fig. 4.11, which shows cross sections at 20°N of PV and RH of 40 and 80%, for CNTL and UPV. Comparison of maps of CNTL and UPV shows that for the time before the convection (14 January 12UTC) both PV fields are nearly identical, except for some small scale features. However, as the convection develops, there is erosion of the PV tongue and sharpening of PV gradient in the region of high RH in CNTL (e.g., 16 January 12 UTC, middle column), and PV contours closely follow the border of relative humidity of 40%. The (negative) contribution of LH extends to a large area as the convection fully develops on 16 January (Fig. 4.12), however, the INTR is more important, since it is one order of magnitude larger that LH.

The same analysis was performed for the same four cases as in section 4.7. Figure 4.13 shows PV at 200hPa and OLR for runs CNTL and UPV at stages in the intrusion when the convection was fully developed. This figure also shows the contribution of the interaction of PV and latent heat (pure contribution of latent heat was one order of magnitude smaller at this level and is not shown).

The cases of January 1999 and January 2003 show a similar picture as seen in the case of January 1987: PV is being destroyed upstream of the intrusion causing a tightening of the PV gradient (Fig. 4.13c,d). However, the cases of February 1991 and January 1997 do not show such pattern: the interaction of PV and latent heat is shown here to produce PV downstream, though more to the north. There is no signal of positive PV being generated east of the PV “trough” (Fig. 4.13a,b). However, the PV evolution for CNTL show that there is still a tightening of the PV gradient downstream and a lot of fine scale structures, which are not present in run UPV. Using exclusion arguments, turbulent friction should be then the most important factor.
Looking from another perspective however, from Fig. 4.14 we can see that the interaction of PV and latent heat still have a significant contribution in eroding the vertical penetration of the intrusion - in all cases. Notice from the INTR column in Fig. 4.14 that the PV destruction occurs in lower levels than the lat×lon cut shown in Fig. 4.13. Also, notice that there is production of PV at lower levels in case (a), corresponding to the effect of convection described earlier.

This analysis does not provide a complete picture of the evolution of PV or its material change due to convection, because here we are accessing only the contribution of latent heat within the diabatic effect. Although the effect of radiative cooling is generally recognized to be more important for shallow clouds (e.g., Hartmann 1994) there is no way to guarantee its effect would be negligible in the present case by the type of analysis and simulations we performed. Also, the results above indicate that when convection is vigorous the turbulent effects may have a significant impact on the PV evolution, perhaps with the same order of magnitude as the interaction term.
Figure 4.8: Time sequence of area-averaged (a) OLR (W.m\(^{-2}\)), (b) dry static stability \( \mathcal{S} = -g \frac{d\theta}{dp} \) (in K.m\(^2\).kg\(^{-1}\)), (c) CAPE (J.kg\(^{-1}\)), (d) vertical velocity \( w = \frac{dz}{dt} \) (cm.s\(^{-1}\)), (e) latent heat flux (W.m\(^{-2}\)), and (f) \( \theta_e \) (K), for cases (i) February 1991, (ii) January 1999, (iii) January 1997, and (iv) January 2003. Results from MM5 simulation, including \( q_U' \) and latent heat release (solid line), including \( q_U' \) but no latent heat (dashed line), including latent heat but no \( q_U' \) (dotted line), and removing both \( q_U' \) and latent heat (dot-dashed line).
Figure 4.8: Continued.
Figure 4.8: Continued.
Figure 4.9: Potential Vorticity (1, 2, 4 and 8 PVU) at 200hPa and OLR (<210 W.m$^{-2}$) given by control run (first row), and run with $q_U$ only (no latent heat, second row), on (a) 14 January 1987 12UTC, (b) 16 January 1987 12UTC, and (c) 18 January 1987 00UTC.

Figure 4.10: Potential Vorticity at 200hPa: contribution of LH only to total PV (first row), and contribution of interaction of LH with PV (second row), for (a) 14 January 1987 12UTC, (b) 16 January 1987 12UTC, (c) 18 January 1987 00UTC. PV contours are ±0.1, 0.2, 0.4, 0.8 PVU for LH, and 1, 2, 4 and 8 PVU for INTR.
Figure 4.11: Cross-section at 20°N of potential vorticity (1, 2, 4 and 8 PVU) and relative humidity (40% light gray, and 80% dark gray) given by control run (first row), run with \( q'_U \) only (no latent heat, second row). (a), (b) and (c) the same dates as in Fig. 4.9

Figure 4.12: Cross-section at 20°N of contribution of LH to total PV (top row) and contribution of INTR (bottom row). PV contours ±0.1, 0.2, 0.4, 0.8 PVU for LH and ±1, 2, 4 and 8 PVU for INTR. (a), (b) and (c) the same dates as in Fig. 4.9
Figure 4.13: Potential Vorticity (1, 2, 4 and 8 PVU) at 200hPa and OLR (<210 W.m$^{-2}$) given by control run (first column), run $\text{UPV}$ only (no latent heat, second column), and the contribution of the interaction term to PV (third column, negative values shaded), on (a) 14 February 1991 00UTC, (b) 17 January 1997 12UTC, (c) 25 January 1999 00 UTC, and (d) 30 January 2003 00UTC.
Figure 4.14: Cross-section of potential vorticity (1, 2, 4 and 8 PVU) and relative humidity (40% light gray, and 80% dark gray) given by runs CNTL (first column) UPV (no latent heat, second column), and contribution of INTR to PV (third column, negative values shaded). Dates in (a)-(d) are the same as in Fig. 4.13, and the cross-sections are at: (a) 20°N, (b) 20°N, (c) 15°N, (d) 12.5°N.
4.9 Summary

We used MM5 to study the impact of upper level PV intrusions into the tropical troposphere on the development of convection. Since PV is not conserved as convection develops, we also examined how latent heat would affect the PV evolution. MM5 is able to simulate intrusions and convection ahead of high PV tongues. Simulations are more sensitive to boundary conditions than to physical parameterization choices, gridsize or initial conditions.

Results from MM5 simulation and numerical modeling analysis corroborates the results obtained from PV inversion analysis. For deep intrusions $q_U'$ has the dominant contribution to the quantities that characterize convection (CAPE, $S$, $w$). When the intrusion is shallower, the larger contributions are still from $q_U'$ or the interaction term, however there is not any evident pattern of when either is more or less important. Table 4.6 summarizes the contribution of $q_U'$ obtained here and by using PV inversion diagnostics.

The effect of latent heat on the evolution of PV is also examined. Agusti-Panareda et

<table>
<thead>
<tr>
<th>Event date</th>
<th>$\frac{dS_f'}{dt}$</th>
<th>CAPE</th>
<th>$w$</th>
</tr>
</thead>
<tbody>
<tr>
<td>(deep)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>15 Jan 1987 [12,12]</td>
<td>100</td>
<td>67</td>
<td>100</td>
</tr>
<tr>
<td>12 Feb 1991 [12,12]</td>
<td>100*</td>
<td>65</td>
<td>61</td>
</tr>
<tr>
<td>(shallow)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>23 Jan 1999 [12,06]</td>
<td>98</td>
<td>89</td>
<td>100</td>
</tr>
<tr>
<td>16 Jan 1997 [12,00]</td>
<td>93</td>
<td>0</td>
<td>38</td>
</tr>
<tr>
<td>28 Jan 2003 [12,06]</td>
<td>100</td>
<td>76</td>
<td>99</td>
</tr>
</tbody>
</table>

Table 4.6: Relative contribution (in %) of $q_U'$ to the static stability tendency for 48h prior to convection, CAPE, and vertical velocity ($=dz/dt$) at 500hPa, for the dates indicated in the left column. The values between brackets are the time (UTC) for PV inversion and numerical simulations, respectively. Values marked with * correspond to the percentage relative to the total negative contributions.
al. (2004) showed that latent heat release was very important to lower surface pressure at early stages of the system evolution. Posselt and Martin (2003) showed that diabatic heating due to latent heat release caused dilution of upper tropospheric PV as a surface cyclone develops. In the present case there was no significant surface feature, therefore we investigated the impact of latent heat and its feedback with PV itself on the intrusion evolution.

For the intrusion of January 1987 it is clear that the interaction of $q'_U$ and LH defines the shape and tightens the PV gradient on the downstream side of the PV tongue. In this particular event, PV destruction was very well defined both in a horizontal plane at 200 hPa and in the vertical cross-sections, showing that there is indeed erosion of PV as it penetrates downwards as well. These results are somewhat comparable with the findings of Ahmadi-Givi et al (2004). They used MM5 to simulate a mid-latitude event of rapid cyclogenesis and found that latent heat caused a horizontal and vertical constraint in PV evolution. They could not however distinguish between the latent heat effect as its direct effect to the effect of non-linear interaction with PV, like it was done in this work using the FSM. In the intrusion case of January 1987, latent heat plays an important role indirectly, through its interaction with PV, both for the outbreak of convection and in the evolution of the system as a whole.

For the other intrusion events studied here, only the vertical profile matched the previous results. The diabatic heating due to latent heat is likely localized lower in the troposphere, and therefore, the (absolute) maximum of negative PV was found lower in the troposphere in some of the cases. The results presented also suggest that since neither latent heat itself or its interaction with PV could cause the PV erosion at higher levels, it is likely that turbulent mixing caused lateral redistribution of PV at those levels. This issue was not completely solved in this analysis. If it would be indeed reasonable that the friction term in Eq. (4.6) is small, then the diabatic term would be responsible for all the redistribution
of PV in a given volume. However, the study done here suggest that this may not be the case, and friction by turbulent mixing may in fact have a large role in PV evolution. One way to solve this question would be to estimate each of the terms during the model run, and calculate directly the contribution of each term to changes in PV (e.g., Lamarque and Hess 1994).
Chapter 5

Summary and Conclusions

In this work we addressed aspects of the connection between convection and extratropical (ET) intrusions into the tropical troposphere over the eastern tropical Pacific in the northern hemisphere winter. These intrusions occur in the so-called “westerly ducts” within the Equatorial easterly jet. Several theoretical and observational studies have shown that these regions of westerlies allow disturbances of ET origin to propagate into the tropical region (Webster and Holton 1982, Tomas and Webster 1984, Hoskins and Ambrizzi 1993), and these disturbances in turn act as forcing for deep convection (Kiladis and Weickmann 1992a,b, Kiladis 1998). Here we address issues that were not explored in these previous studies. In particular, we examined the three-dimensional structure of these intrusions, the quantitative impact of these intrusions as a forcing to deep convection in this region, and finally, the impact of latent heat on the evolution of these upper-level disturbances.

In Chapter 2, we formed composites of PV at various levels for these events, based on the climatology of deep tropical intrusions formed by Waugh and Polvani (2000) and found that they have a remarkably similar three-dimensional structure. The composite high-PV “trough” at 350K on day 0 has a slight tilt to the southwest-northeast relative to the N-S direction. Also, its structure is nearly barotropic from 350K to 430K. However, they do not
penetrate deep vertically, with very few events (around 10%) with 2PVU contour reaching 20°N at 330K.

The simultaneous analysis of OLR showed a relatively low signal at the leading edge of the intrusion. There is OLR less than 200 W.m\(^{-2}\) in 90% of the events. Furthermore, the high-PV “trough” amplifies prior to the signal of low OLR suggesting that convection is being forced by the intrusion. This is consistent with results obtained by Kiladis and Weickmann (1992a,b) using another diagnostic analysis. This point was further corroborated in our study by identifying “low OLR events”, i.e., identifying days or periods in which there was low OLR in a fixed position and forming composites of PV for such dates. This analysis showed that there is high-PV protruding from the ET prior to the day of minimum OLR, confirming the former results.

The above analysis showed that there is a strong link between the PV intrusions and tropical convection. This link has been suggested to act in a similar way to the baroclinic development in the mid-latitudes. From a potential vorticity perspective, one can think of an upper-level positive PV disturbance inducing lower static stability and promoting enhanced vertical motion and convection downstream. Our next step then was to use concepts of PV thinking and PV invertibility principle (Hoskins et al. 1985) to access the amount of destabilization and vertical motion directly related to the positive PV anomaly associated with the intrusion.

In Chapter 3 we used the PV inversion technique developed by Davis and Emanuel (1991) to address the issue of attribution. Using this methodology we obtained the relative contribution from the upper-level PV anomaly \(q'_{U}\) to the anomalous vertical velocity \(\omega\), static stability \(S\) and CAPE, for selected intrusion events. We found that when the anomaly is strong and deep, the hypothesis proposed to explain the link between the intrusion and convection is clearly observed in the (NCEP) data. When the anomaly is weaker and shallower the trend is still detected but the overall picture is more blurred. Quantitative results
are strongly dependent on the intensity and vertical penetration of the intrusion. When
the PV anomaly was strong and deep results showed that it contributed to about 50% of
CAPE accumulation prior to convection, \( \sim 100\% \) of static stability decrease in the of layer
850-500hPa, and \( \sim 100\% \) of vertical velocity at 500hPa. When the anomaly is weaker and
shallower these numbers are much more modest.

These PV inversion results confirm the hypothesis that the positive PV advection ahead
of the intrusion promotes atmospheric destabilization and enhanced vertical motion, coin-
cident with the region of low OLR. This methodology however does not take into account
the non-linear effects of diabatic heating due to convection on the PV itself. In Chapter 4
we used a limited area mesoscale model (the Penn State University/ NCAR MM5 model)
to confirm the results obtained by PV inversion (Chapter3) and to investigate the impact of
latent heating on PV evolution.

MM5 simulations were able to reproduce fairly well the observed PV structure of the
intrusions and also able to produce an area of low OLR ahead of the intrusion. However, the
convective area was consistently smaller compared to the OLR data. A series of simulations
with PV intrusion and/or latent heat removed were performed to isolate the contributions of
the upper level PV anomaly \( (q_U') \) and latent heat for areas of low OLR. These simulations
showed that \( q_U' \) is the dominant factor to decrease OLR and static stability, and to increase
CAPE and vertical velocity. The latent heat effect by itself had minor contribution to the
total field, but it is indirectly important through the interaction term, which becomes more
important as the convection develops. These model results compared reasonably well with
the results of PV inversion diagnostics (Chapter 3), at least for the cases of deep vertical
penetration. For such cases, the results show that the upper level PV anomaly induces OLR
and static stability decrease, CAPE increase and positive vertical velocity. The simulation
results show a strong dependency on the depth of the penetration: When the intrusion
is shallower, the relative contributions of different factors have no apparent pattern. The
general result however is that the upper level PV anomaly associated with the intrusion is a necessary feature for convection to occur.

The last issue addressed was the impact of latent heat on the PV evolution. PV is materially conserved in an adiabatic and frictionless flow, but in the evolution of intrusion and outbreak of convection, there is diabatic heating through latent heat release and turbulent mixing and PV is no longer conserved. Analysis of simulations showed that the non-linear interaction of latent heat release and the upper-level PV intrusion acted to destroy PV downstream of the intrusion, causing the high-PV “trough” to be narrower and more N-S oriented. Also, destruction of PV due to the interaction term occurs over a layer (depth), coincident with the area of higher relative humidity.

**Future Work**

There are several issues that we did not examine that deserve further attention. One of these issues is the calculation of PV budget (Lamarque and Hess 1994) and calculation of cross-tropopause mass and ozone transport. It would be interesting to investigate how strong the coupling between convection and irreversible transport of trace constituents is in these subtropical events.

Another aspect that was not explored here was the momentum budget, energy propagation and wave activity associated with these intrusion events. Transient eddy wave activity was found to likely be an important component of the momentum balance in the tropical eastern Pacific (Kiladis 1998), and may have an impact on the intraseasonal variability of NAO (Abatzoglou and Magnusdottir 2005). It would be interesting to investigate the momentum flux and wave activity during intrusion events, and whether there are significant differences between different events, e.g. shallow or deep intrusions.

Although the connection of convection and high-PV is very robust over the eastern tropical Pacific, for both South Pacific and South Atlantic regions this connection is not
detected. It would be interesting to investigate the differences and similarities in the dynamical fields between the events in the Southern and Northern Hemisphere, and whether there is any correlation between intrusions events that happen in the Northern and Southern hemispheres.

The absence of convection at the leading edge of intrusions in the Southern Hemisphere raises questions as for the water vapor distribution and transport for these events and controlling mechanism of convection. Also, we do not know whether there is cross-tropopause transport of mass, ozone, and other constituents. The study of these issues would help form a more complete global picture of the impact of such intrusions on the tropical dynamics.
References


Appendix A

PV inversion equations

The inversion of PV field is tied to the choice of balance for the flow (Hoskins et al., 1985). In the diagnostic system and method to solve it proposed by DE91, the authors used the so-called “Charney balance equation” (after Charney, 1955).

First, recall that the horizontal velocity ($v$) can be decomposed into an irrotational plus a non-divergent parts (Helmoltz decomposition):

$$v = v_\chi + v_\psi = \nabla \chi + k \times \nabla \psi,$$

where $\chi$ is the velocity potential and $\psi$ the streamfunction. Defining $R_\psi = V_\psi / f_o L$ and $R_\chi = V_\chi / f_o L$, the basic assumption for the flow is that $V_\psi \ll V_\chi$ and terms of $O(R_\psi)$ are retained while those of $O(R_\chi)$ are neglected. In spherical coordinates, the Charney balance is then expressed as

$$\nabla^2 \Phi = \nabla. (f \nabla \psi) + \frac{2}{a^4 \cos^2 \phi} \frac{\partial (\partial \psi / \partial \lambda, \partial \psi / \partial \phi)}{\partial (\lambda, \phi)}, \quad (A-1)$$

where $\Phi$ is the geopotential, $\lambda$ the longitude, $\phi$ the latitude and $a$ the radius of the earth. This balance is very accurate in flows with large curvature.

The second equation of the system is given by the approximate Ertel’s PV equation:
\[ q = -\frac{g \kappa \pi}{p} \left( \eta \frac{\partial \theta}{\partial \pi} - \frac{1}{a \cos \phi} \frac{\partial v}{\partial \pi} \frac{\partial \theta}{\partial \lambda} + \frac{1}{a} \frac{\partial u}{\partial \phi} \frac{\partial \theta}{\partial \phi} \right) \tag{A-2} \]

where \( \kappa = \frac{R_d}{C_p} \), \( p \) is the pressure, \( \pi \) is the Exner function \( [C_p(p/p_o)^\kappa] \) serving as the vertical coordinate, \( \eta \) is the vertical component of absolute vorticity and the hydrostatic approximation was incorporated. Performing a similar scaling of Eq.(A-2) as was used to obtain Eq.(A-1), a relation between \( q, \psi \) and \( \Phi \) is obtained:

\[ q = \frac{g \kappa \pi}{p} \left[ (f + \nabla^2 \psi) \frac{\partial^2 \Phi}{\partial \pi^2} - \frac{1}{a^2 \cos^2 \phi} \frac{\partial^2 \psi}{\partial \lambda \partial \pi} \frac{\partial^2 \Phi}{\partial \lambda \partial \pi} - \frac{1}{a^2} \frac{\partial^2 \psi}{\partial \phi \partial \pi} \frac{\partial^2 \Phi}{\partial \phi \partial \pi} \right] \tag{A-3} \]

Equations (A-1) and (A-3) from a coupled non-linear system to be solved for \( \Phi \) and \( \psi \). Boundary conditions are required for the lateral, upper and bottom edges. Either Dirichlet or Neumann conditions may be used, but most authors have chosen Dirichlet conditions for \( \psi \) and \( \Phi \) for the lateral boundaries and the second one for horizontal boundaries, \( \partial \Phi/\partial \pi = f_o (\partial \psi/\partial \pi) = -\theta \) (hydrostatic approximation). Griffith et al. (2000) chose Dirichlet boundary conditions for top and bottom, and in a sensitivity study, showed that the results are not significantly different from one choice of boundary condition or the other.

The method of solution consists of solving equations obtained by summing and subtracting non-dimensionalized forms of Eqs. (A-1) and (A-3):

\[ \left[ \sin \phi + \frac{\partial^2 \Phi^{(v)}}{\partial \pi^2} \right] \nabla^2 \psi^{(v+1)} = q + \nabla^2 \Phi^{(v)} - \sin \phi \frac{\partial^2 \Phi^{(v)}}{\partial \pi^2} - N_2[\psi^{(v)}, \Phi^{(v)}] - N_1[\psi^{(v)}, \Phi^{(v)}] \tag{A-4} \]

\[ \nabla^2 \Phi^{(v+1)} + \left[ \sin \phi + \nabla^2 \psi^{(v+1)} \right] \frac{\partial^2 \Phi^{(v+1)}}{\partial \pi^2} = q + \sin \phi \nabla^2 \psi^{(v+1)} - N_2[\psi^{(v+1)}, \Phi^{(v)}] + N_1[\psi^{(v+1)}, \Phi^{(v)}] \tag{A-5} \]

In the above equations the variables \( \psi, \Phi \) and \( q \) are non-dimensional, and the operators \( N_1 \) and \( N_2 \) are defined as
\[ N_1(a,b) = \frac{2}{(\cos^2 \phi)} \frac{\partial(a,b)}{\partial(\lambda, \phi)}, \quad N_2(a,b) = -\frac{1}{\cos^2 \phi} \frac{\partial^2 a}{\partial \lambda \partial \pi} \frac{\partial^2 b}{\partial \lambda \partial \pi} - \frac{\partial^2 a}{\partial \phi \partial \pi} \frac{\partial^2 b}{\partial \phi \partial \pi} \]

The equations (A-4) and (A-5) are elliptic in \( \psi^{n+1} \) and \( \Phi^{n+1} \), provided that the static stability and absolute vorticity are positive, and their finite difference form are solved individually by using standard successive overrelaxation.
Appendix B

Overview of the PSU/NCAR Mesoscale Modeling System MM5

The Fifth-Generation Penn State/NCAR Mesoscale model is a fully compressible, non-hydrostatic three-dimensional mesoscale model, with multiple nesting capabilities, making it extremely versatile and suitable for a broad range of applications. The short description given here is mostly taken from the User’s Guide, available at: http://www.mmm.ucar.edu/mm5/On-Line-Tutorial/interpf/interpf.html (as of 31 May 2005), and the reader is referred to this link for a complete description of the model.

The MM5 modeling system is composed of several modules, to be completed in a defined sequence: (1) TERRAIN: Set up the mesoscale domain (user defined), (2) REGRID: “Creates” the first guess files from the gridded global or regional analysis, putting it into the MM5 grid, (3) LITTLE_R (or, optionally, RAWINS): Incorporates surface observations and radiosonde data to the first guess, (4) INTERPF: Create boundary condition files, interpolating from pressure to $\sigma$ levels, and (5) MM5: Predictive part of the system, solves the equations of momentum, continuity and thermodynamics, given respectively by Eqs. (B-1) to (B-5). Other available modules are INTERPB (interpolates the output given in $\sigma$ levels...
into pressure levels), NESTDOWN (horizontally interpolates the $\sigma$-coordinated data from a coarse grid to a fine grid), and GRAPH/RIP (graphical display). Figure B.1 shows a flow chart of the MM5 modeling system.

The model uses finite-differences to solve the equations of momentum (Eq. B-1 - B-3), continuity (compressible, Eq. B-4), and thermodynamics (Eq. B-5):

Momentum x-component:

$$\frac{\partial u}{\partial t} + \frac{m}{\rho} \left( \frac{\partial p'}{\partial x} - \frac{\sigma}{p^*} \frac{\partial p'}{\partial \sigma} \right) = -V \cdot \nabla u + v \left( f + u \frac{\partial m}{\partial y} - v \frac{\partial m}{\partial x} \right) - \frac{\alpha}{\sin \alpha} - \frac{uv}{r_{earth}} + D_u$$  \hspace{1cm} (B-1)

Momentum y-component:
\[
\frac{\partial v}{\partial t} + \frac{m}{\rho} \left( \frac{\partial p'}{\partial y} - \sigma \frac{\partial p'}{\partial \sigma} \right) = -\mathbf{V} \cdot \nabla v + u \left( f + u \frac{\partial m}{\partial y} - v \frac{\partial m}{\partial x} \right) + \rho g p' \frac{\partial}{\partial y} \left( \frac{\partial p'}{\partial y} - \sigma \frac{\partial p'}{\partial \sigma} \frac{\partial p'}{\partial \sigma} \right) + e w \sin \alpha - \frac{vw}{r_{earth}} + D_v
\]

Momentum z-component:

\[
\frac{\partial v}{\partial t} - \frac{\rho_0 g p'}{\rho p*} \frac{\partial p'}{\partial \sigma} + \frac{g p'}{\gamma p} = -\mathbf{V} \cdot \nabla w + g \left( \frac{p_0 T'}{T} \frac{R_d p'}{c_p p} \right) + e (u \cos \alpha - v \sin \alpha) + \frac{u^2 + v^2}{r_{earth}} + D_w
\]

Continuity (anelastic):

\[
\frac{\partial p'}{\partial t} - \rho_0 g w + \gamma p \nabla \cdot \mathbf{V} = -\mathbf{V} \cdot \nabla p' + \frac{\gamma p}{T} \left( \frac{\dot{Q}}{c_p} + \frac{T_0}{\theta_0} D_0 \right)
\]

Thermodynamics:

\[
\frac{\partial T}{\partial t} = -\mathbf{V} \cdot \nabla T + \frac{1}{\rho c_p} \left( \frac{\partial p'}{\partial t} + \mathbf{V} \cdot \nabla p' - \rho_0 g w \right) + \frac{\dot{Q}}{c_p} + \frac{T_0}{\theta_0} D_0
\]

In the above equations, \( e = 2\Omega \cos \lambda \), \( \alpha = \phi - \phi_c \), \( \lambda \) is latitude, \( \phi \) is longitude, \( \phi_c \) is central longitude, \( m \) is the map-scale factor, defined as the ratio between the distance on the grid and the actual distance on earth. The terms \( u \partial m/\partial y \) and \( v \partial m/\partial x \) represent curvature effects. In the model, the last term in the right hand side of Eq. (B-4) is not included.

The vertical grid is a terrain-following sigma-coordinate, defined as:

\[
\sigma = \frac{(p - p_t)}{(p_s - p_t)}
\]

where \( p \) is the pressure, \( p_t \) is a constant top pressure (user specified) and \( p_s \) is the surface pressure.

As input data, MM5 requires gridded atmospheric data of sea-level pressure, wind, temperature, relative humidity and geopotential height, at these pressure levels: surface,
Figure B.2: Schematics of the direct interaction of parameterizations in MM5.

1000, 850, 700, 500, 400, 300, 250, 200, 150, 100hPa; it also requires topography data and landuse (in categories), and surface report and sounding data. Like other mesoscale models, MM5 requires lateral boundary conditions in addition to initial conditions. The lateral boundary conditions must have specified horizontal winds, temperature, pressure and moisture fields. MM5 has 3 options to handle the lateral boundary conditions: fixed, time-dependent (values specified for all predicted fields in the outer two rows and columns), and relaxation/inflow-outflow (outer row and column are specified, and adjacent 4 points are relaxed towards the boundary values with a relaxation constant decreasing linearly away from the boundary).

MM5 is an open source system and therefore there has been several contributions and adaptations to its original form. The basic MM5 “package” accommodates a number of physical parameterizations: Cumulus parameterization (8 types, plus option to allow shallow cumulus), planetary boundary layer (8 types, with additional option for moist vertical diffusion and/or changing the thermal roughness length for heat/moisture for some of the schemes), cloud microphysics (8 types), radiation schemes (5 types) and surface schemes.
(5 types, with additional options for snow and soil moisture for some of the schemes). Figure B.2 shows a schematics of the direct interaction of these parameterizations.

MM5 output consists of 14 3D and 36 2D predicted fields, as well as 11 constant 2D fields. Those relevant for the present study are u-wind, v-wind, vertical velocity (in $\sigma$-levels), temperature, water vapor mixing ratio, surface fluxes (latent heat and sensible), outgoing longwave radiation at the top of the atmosphere, sea surface temperature. Other fields used in this study, such as sea surface pressure, potential vorticity and (geopotential) height, were derived from the above.

The latest version of MM5 is the 7th, at which point it has been frozen. The extent of the use of the MM5 modeling system and its predecessors can be seen, e.g., at the link: http://www.mmm.ucar.edu/mm5/Publications (as of 31 May 2005).
Vita

1. EDUCATION

Ph.D. in Atmospheric Sciences, 2005: Stratospheric intrusions and transient convection in the tropical eastern Pacific. The Johns Hopkins University, Baltimore, MD.
Advisor: Prof. Darryn W. Waugh.

M.A&Sc., 2002. The Johns Hopkins University, Baltimore, MD.

M.Sc. in Meteorology, 1999: Synoptic-dynamic study of cyclogenesis using potential vorticity. Brazilian National Institute for Space Research (INPE), São José dos Campos, SP, Brazil. Advisors: Dr. Manuel A. Gan and Dr. Ernesto S. Caetano Neto.

B.Sc. in Meteorology, 1992. Astronomical and Geophysical Institute (IAG) of University of São Paulo, São Paulo, SP, Brazil.

2. PROFESSIONAL EXPERIENCE


3. CO-CURRICULAR ACTIVITIES


Measurement Campaign: field work in the International Pantanal Experiment (IPE-0) team, a project developed and managed by the Meteorological Sciences Division of INPE, at Passo do Lontra, MS, Brazil, 28 Sep - 6 Oct 1996.


4. FELLOWSHIPS AND AWARDS

- Hans P. Eugster Research Fund, for laboratory work support (2002).
- Brazilian National Council for Research (CNPq) RHAE-DTI grant, for research at INPE (1996-1997).

5. PUBLICATIONS

Papers:


**Conference Abstracts:**


**Funatsu, B.M.,** and D. W. Waugh. The connection between stratospheric intrusions and tropical convection. 12th Conference on the Middle Atmosphere, San Antonio, TX, 4-7 Nov 2002.


133