Dynamics and Variability of Foehn Winds in the McMurdo Dry Valleys Antarctica

Dissertation

Presented in Partial Fulfillment of the Requirements for the Degree Doctor of Philosophy in the Graduate School of The Ohio State University

By
Daniel Frederick Steinhoff, M.S.
Graduate Program in Atmospheric Sciences

The Ohio State University
2011

Dissertation Committee:
David H. Bromwich, Advisor
Jay S. Hobgood
Jialin Lin
Jeffrey Rogers
Abstract

The McMurdo Dry Valleys (“MDVs”) are the largest ice-free region in Antarctica, featuring perennially ice-covered lakes that are fed by ephemeral melt streams in the summer. The MDVs have been an NSF-funded Long-Term Ecological Research (LTER) site since 1993, and LTER research has shown that the hydrology and biology of the MDVs are extremely sensitive to small climatic fluctuations, especially during summer when temperatures episodically rise above freezing. However, the atmospheric processes that control MDVs summer climate, namely the westerly foehn and easterly sea-breeze regimes, are not well understood. The goals of this study are to (i) produce a coherent physical mechanism for the development and spatial extent of foehn winds in the MDVs, and (ii) determine aspects of large-scale climate variability responsible for intraseasonal and interannual differences in MDVs temperature. Polar WRF simulations are run for a prominent foehn case study at 500 m horizontal grid spacing to study the mesoscale components of foehn events, and 15 summers at 2 km horizontal grid spacing to analyze event and temporal variability. The Polar WRF simulations have been tailored for use in the MDVs through modifications to the input soil conditions, snow cover, land use, and sea ice. An objective foehn identification method is used to identify and categorize events, as well as validate the model against LTER AWS observations.
The MDVs foehn mechanism consists of a gap wind through a topographic constriction south of the MDVs, forced by pressure differences on each side of the gap and typically set up by cyclonic flow over the Ross and Amundsen Seas. Significant mountain wave activity over the gap modulates the flow response over the MDVs themselves, and pressure-driven channeling drives foehn flow down-valley. During strongly forced events, mass accumulation east of the MDVs from flow around Ross Island is responsible for easterly intrusions, and not a thermally forced sea breeze as previously thought. A variety of ambient flow directions and associated synoptic-scale patterns can result in MDVs foehn, but adequate forcing is necessary to activate the foehn mechanism. The warmest foehn events are associated with amplified circulation patterns that are not associated with particular interannual modes of variability, but instead related to intraseasonal variability forced by the extratropical response to a stagnant MJO. Implications of the findings upon current MDVs paleoclimate theories on the existence of huge melt lakes at the LGM are also presented.
Dedication

To my parents, for their endless love and support.
Acknowledgments

This research was supported by the National Science Foundation via NSF Grant ANT-0636523. I thank my advisor Dave Bromwich, first for taking a chance on me when I showed up to his research group in autumn 2004, and then for sticking with me in the years since. I appreciate his research insight, constructive criticism, and tenacity. Andrew Monaghan first conceived this research project and provided both computing and intellectual support. Johanna Speirs has been a tremendous source of knowledge about the McMurdo Dry Valleys, and a pleasure to collaborate with on this project. I also thank Sheng-Hung Wang for assistance with the EOF analysis and numerous other computational issues, Ryan Fogt for discussions on Southern Hemispheric climate variability and other research group members (Julien Nicolas, Aaron Wilson, Keith Hines, Le-Sheng Bai, and Francis Otieno) for their help and camaraderie. UCAR Subcontract S01-22901 (the AMPS project) supported my second trip to McMurdo Station, Antarctica, in January-February 2009. Computing support has been provided by NCAR/CISL under account 36091018, and by the Ohio Supercomputing Center through grant PAS0535-1. On a personal note, I deeply appreciate the love and support of my family, as well as friends in Columbus and elsewhere. Finally I thank God for His continued blessings upon me.
Vita

February 8, 1981 ............................................Born, Milwaukee, Wisconsin, U.S.A.

2003................................................................B.S., Atmospheric and Oceanic Sciences,
University of Wisconsin-Madison

2005 to present...............................................Graduate Research Associate,
The Ohio State University

2008 ...............................................................M.S., Atmospheric Sciences, The Ohio State
University

Publications


Fields of Study

Major Field: Atmospheric Sciences
Table of Contents

Abstract ........................................................................................................................................ ii
Dedication .................................................................................................................................... iv
Acknowledgments ....................................................................................................................... v
Vita ................................................................................................................................................ vi
List of Figures .............................................................................................................................. xii
List of Tables ................................................................................................................................ xx
Chapter 1: Introduction ............................................................................................................. 1
Chapter 2: Literature Review .................................................................................................... 6
  2.1 Physical Setting ................................................................................................................ 6
  2.2 Wind Regimes .................................................................................................................. 11
  2.3 Climate Variability .......................................................................................................... 15
    2.3.1 ENSO ..................................................................................................................... 17
    2.3.2 SAM ..................................................................................................................... 25
    2.3.3 ENSO-SAM Interactions ....................................................................................... 27
    2.3.4 South Pacific Wave and Intraseasonal Variability .............................................. 29
    2.3.5 The Austral Summer of 2001/2002 ..................................................................... 31
    2.3.6 Blocking ................................................................................................................ 32
List of Figures

Figure 2.1. Map of the Ross Sea region of Antarctica. The McMurdo Dry Valleys AWS network: Lake Vida (VV), Lake Vanda (WV), Lake Brownworth (WB), Beacon Valley (BV), Taylor Glacier (TTa), Lake Bonney (TB), Lake Hoare (TH), Howard Glacier (THo), Canada Glacier (TCa), Lake Fryxell (TF), Commonwealth Glacier (TCo) and Explorer’s Cove (TE). Landsat ETM+ image captured 21 Nov 2001. From Speirs et al. (2010), Fig. 1. In top left inset, “EAIS” refers to East Antarctic Ice Sheet, and “WAIS” refers to West Antarctic Ice Sheet. ................................................................. 62

Figure 2.2. Conceptual model of summertime flow in Wright Valley. From McKendry and Lewthwaite (1990), Fig. 13........................................................................................................ 63

Figure 2.3. Schematic of upper-tropospheric height anomalies over the Pacific Ocean during the early stage of an ENSO event in the austral winter. Stippling shows region of enhanced convection over central equatorial Pacific and arrows indicate westerly wind anomalies in jet streams. From Karoly (1989), Fig. 11a. .............................................. 64

Figure 2.4. Schematic of atmospheric circulation response to ENSO warm (top) and cold (bottom) events superimposed on the corresponding SST composites. From Yuan (2004), Fig. 8..................................................................................................................... 65

Figure 2.5. Streamlines over an infinite series of sinusoidal ridges for (a) evanescent waves ($m^2 > 0$, $l^2 > k^2$, or $N/U > k$) and (b) vertically propagating waves ($m^2 < 0$, $l^2 < k^2$, or $N/U < k$). Airflow is from left to right across diagram. Dashed line in (b) represents line of constant phase. From Durran (1990), Fig. 4.2...................................................... 66

Figure 2.6. Streamlines over a bell-shaped ridge. (a) Narrow ridge and evanescent wave ($l^2 << k^2$ or $al << 1$), (b) intermediate ridge and nonhydrostatic vertically propagating wave ($l^2 \approx k^2$ or $al \approx 1$), (c) wide ridge and hydrostatic vertically propagating wave ($l^2 >> k^2$ or $al >> 1$). Airflow is from left to right across diagrams. From Durran (1986b), Fig. 20.2................................................................................................................................. 67

Figure 2.7. Streamlines over an isolated bell-shaped ridge for trapped waves. Airflow is from left to right across diagram. From Durran (1990), Fig. 4.4............................................... 68

Figure 2.8. Isentropes for airflow from left to right in a two-layer atmosphere when the mountain height is fixed at 500 m and the interface between the two layers is at (a) 1000 m, (b) 2500 m, (c) 3500 m, and (d) 4000 m. From Durran (1986a), Fig. 5................................. 69
Figure 2.9. (a) Thermodynamical foehn theory, where air rises on windward slope, becomes saturated and cools moist adiabatically, descends and warms at dry adiabatic lapse rate. (b) Current foehn theory, where air near the surface in the lee of the ridge originates from about 2 km AGL upstream. From Seibert (1990), Fig. 9.

Figure 2.10. Semi-idealized vertical cross sections through Wipp Valley near Innsbruck of potential temperature (solid lines), wind speed along cross section (shaded), and wind vectors along cross section when ambient wind direction is (a) SSE, (b) S, (c) SW, and (d) WSW. From Zängl (2003b), Fig. 10.

Figure 2.11. Summary of dynamics pertaining to a northerly foehn event. From Jiang et al. (2005), Fig. 16.

Figure 3.1. 2.5° latitude x 5.0° longitude bin-averaged (a) sea ice thickness (m), and (b) snow thickness on top of sea ice (m), from the ASPeCt program observations.

Figure 3.2. Terrain height (shaded and contoured, m) from (a) 32-km, (b) 8-km, (c) 2-km, and (d) 0.5-km Polar WRF domains. Black rectangles represent nested domains. Relevant geographical features labeled.

Figure 3.3. Actual station locations (green dots) in relation to Polar WRF (a) Terrain height (shaded and contoured, m), and (b) Land use category (White: Snow/Ice, Brown: Barren or Sparsely Vegetated, Blue: Water) for 2-km domain.

Figure 3.4. Same as Fig. 3.3, except for 0.5-km domain.

Figure 3.5. Vertical cross-section through a basin with zero large-scale wind of potential temperature (contoured, K) after 12 hours simulation with (a) diffusion along model vertical levels and (b) truly horizontal diffusion. Dashed lines represent initial temperature field. From Zängl (2003b), Fig. 1.

Figure 3.6. 2-km domain initial subsurface temperature (color shaded, °C) at 0000 UTC 1 November 2005 (beginning of one-year spinup simulation) for (a) 0-10 cm layer, (b) 10-40 cm layer, (c) 40-100 cm layer, and (d) 100-200 cm layer. Black dots represent LTER AWS station locations.

Figure 3.7. 2-km domain initial subsurface soil moisture (color shaded, m³ m⁻³) at 0000 UTC 1 November 2005 (beginning of one-year spinup simulation) for (a) 0-10 cm layer, (b) 10-40 cm layer, (c) 40-100 cm layer, and (d) 100-200 cm layer. Black dots represent LTER AWS station locations.

Figure 4.1. AMPS SLP and near-surface wind vector 2 year composites (2006 and 2007) for (a) 688 non-foehn cases and (b) 688 foehn cases, (c) SLP difference (foehn – non-foehn) and (d) pressure gradient difference (foehn-non-foehn). In (c) and (d) stipling is for positive differences, hatching is for negative differences. Light stipling/hatching refers to 90% confidence level, heavy stipling/hatching refers to 95% confidence level. Star denotes location of MDVs and arrow in (d) shows the direction of the pressure gradient.

Figure 4.2. Same as Fig. 4.1 except for summer (DJF, 96 cases).
Figure 4.3. Same as Fig. 4.1 except for winter (JJA, 298 cases). .......................... 141

Figure 4.4. AMPS SLP and near-surface wind vector annual composites (a) 2006 and (b) 2007, (c) SLP difference (2007 – 2006) and (d) pressure gradient difference (2007 – 2006). In (c) and (d) stipling is for positive differences, hatching is for negative differences. Light stipling/hatching refers to 90% confidence level, heavy stipling/hatching refers to 95% confidence level. Star denotes location of MDVs and arrow in (d) shows the direction of the pressure gradient............................................... 142

Figure 4.5. Synoptic overview plots of (left) sea-level pressure (hPa, contours), 3 m temperature (°C, color shading), and 3 m wind vectors (arrows) and (right) 500 hPa geopotential height (m, contours), relative vorticity (10^-5 s^-1, color shading) and wind vectors (arrows) for the labeled times. Sea-level pressure not plotted over areas exceeding 500 m elevation.............................................................................................. 143

Figure 4.6. Time series of 3 m temperature (°C, purple), relative humidity (%), wind direction (degrees, orange), and wind speed (m s^-1, green) from LTER observations (dark colors) and representative location in Polar WRF (light colors) over the case study period. Note scaling differences between stations. ................................................................. 145

Figure 4.7. Polar WRF 3 m wind speed (m s^-1, color shading) and streamlines at (a) 0000 UTC 29 December 2006 and (b) 1800 UTC 29 December 2006. ............................................................................. 147

Figure 4.8. Orientation maps for 500-m grid spacing simulations. Terrain height (m) shaded, in (a) white transect represents gap profile, magenta transect represents cross-gap pressure difference calculations, red transect represents cross sections in Figs. 4.14, 4.19, and 4.22, yellow transect represents cross section in Fig. 4.15, light blue transect represents cross section in Fig. 4.23, green transect represents cross section in Fig. 4.31, white averaging box used for near-surface gap wind speeds, blue averaging box used for inverse Froude number calculations, and orange dots represent LTER AWS sites. In (b), red transect is used in along-gap reduced pressure calculations, and green dots represent LTER AWS sites............................................................................................................. 148

Figure 4.9. Profile slice of the gap south of the McMurdo Dry Valleys along the white transect in Fig. 4.8a. “EAIS” is East Antarctic Ice Sheet, “MF” is Mount Feather (2985 m), “Gap” represents the approximate width of the gap, of which the lowest point and vertical profile varies with latitude, and “RSR” is the Royal Society Range................. 149

Figure 4.10. Polar WRF domain 2 sea-level pressure (hPa, contours), 3 m temperature (°C, color shading), and wind vectors (arrows) at 1800 UTC 29 December 2006. Sea-level pressure not plotted over areas exceeding 500 m elevation................................................................. 150

Figure 4.11. Polar WRF reduced pressure difference across gap along the magenta transect in Fig. 4.8a (red, hPa), averaged 3 m wind speed in the gap (m s^-1, blue) using the white box in Fig. 4.8a, 600 hPa upstream wind direction (degrees, green), and 600 hPa upstream wind speed (m s^-1, gold). ....................................................................................... 151

Figure 4.12. Polar WRF level 10 wind speed (m s^-1, color shading) and streamlines at (a) 1800 UTC 29 December 2006. ............................................................................. 152
Figure 4.13. Vertical profiles of wind speed (m s$^{-1}$, top) and wind direction (degrees, bottom) at origin of the red cross section in Fig. 4.8a at 1800 UTC 29 December 2006. ................................................................. 153

Figure 4.14. Vertical cross section of potential temperature (K, contours), wind speed along the red cross section in Fig. 4.8a (m s$^{-1}$, color shaded), and circulation vectors at 1800 UTC 29 December 2006. “Gap” is the elevated gap, “KH” is Knobhead”, “FG” is Ferrar Glacier, “TV” is Taylor Valley, “AR” is Asgard Range, “WV” is Wright Valley, “OR” is Olympus Range, and “VV” is Victoria Valley. ................................................ 154

Figure 4.15. Vertical cross section of potential temperature (K, contours), wind speed along the yellow cross section in Fig. 4.8a (m s$^{-1}$, color shaded), and circulation vectors at 1800 UTC 29 December 2006. “RSR” is the Royal Society Range, “CR” is Cathedral Rocks, “FG” is Ferrar Glacier, “KH” is Kukri Hills, “TV” is Taylor Valley, and “AR” is Asgard Range. .................................................................................................................. 155

Figure 4.16. Polar WRF domain 2 sea-level pressure (hPa, contours), 3 m temperature (°C, color shading), and wind vectors (arrows) at 1800 UTC 30 December 2006. Sea-level pressure not plotted over areas exceeding 500 m elevation. ..................... 156

Figure 4.17. Polar WRF 3 m wind speed (m s$^{-1}$, color shading) and streamlines at (a) 1800 UTC 30 December 2006. ....................................................................................... 157

Figure 4.18. Vertical profiles of wind speed (m s$^{-1}$, top) and wind direction (degrees, bottom) at origin of the red cross section in Fig. 4.8a at 1800 UTC 30 December 2006. .................................................................................................................. 158

Figure 4.19. Vertical cross section of potential temperature (K, contours), wind speed along the red cross section in Fig. 4.8a (m s$^{-1}$, color shaded), and circulation vectors at 1800 UTC 30 December 2006. “Gap” is the elevated gap, “KH” is Knobhead”, “FG” is Ferrar Glacier, “TV” is Taylor Valley, “AR” is Asgard Range, “WV” is Wright Valley, “OR” is Olympus Range, and “VV” is Victoria Valley. ................................................ 159

Figure 4.20. Polar WRF 3 m wind speed (m s$^{-1}$, color shading) and streamlines at (a) 0300 UTC 01 January 2007. .......................................................................................... 160

Figure 4.21. Vertical profiles of wind speed (m s$^{-1}$, top) and wind direction (degrees, bottom) at origin of the red cross section in Fig. 4.8a at 0300 UTC 1 January 2007.... 161

Figure 4.22. Vertical cross section of potential temperature (K, contours), wind speed along the red cross section in Fig. 4.8a (m s$^{-1}$, color shaded), and circulation vectors at 0300 UTC 1 January 2007. “Gap” is the elevated gap, “KH” is Knobhead”, “FG” is Ferrar Glacier, “TV” is Taylor Valley, “AR” is Asgard Range, “WV” is Wright Valley, “OR” is Olympus Range, and “VV” is Victoria Valley. ................................................ 162

Figure 4.23. Vertical cross section of potential temperature (K, contours), wind speed along the light blue cross section in Fig. 4.8a (m s$^{-1}$, color shaded), and circulation vectors at 0300 UTC 1 January 2007. “FG” is Ferrar Glacier, “KH” is Kukri Hills, and “TV” is Taylor Valley................................................................. 163
Figure 4.24. (a) Polar WRF near-surface wind speed (m s$^{-1}$, color shaded) and streamlines, (b) Polar WRF sea-level pressure (hPa, color shaded) at 0900 UTC 30 December. (c) and (d) same as (a) and (b), respectively, but for zoomed in area of MDVs. Large dots in (c) and (d) for Explorer’s Cove. 164

Figure 4.25. Polar WRF reduced pressure (hPa, red), near-surface wind speed (m s$^{-1}$, green), near-surface wind direction (degrees, gold), and terrain height (m, black) along the transect shown in Fig. 4.8b at 0900 UTC 30 December. 165

Figure 4.26. (a) Polar WRF near-surface wind speed (m s$^{-1}$, color shaded) and streamlines, (b) Polar WRF sea-level pressure (hPa, color shaded) at 2100 UTC 30 December. (c) and (d) same as (a) and (b), respectively, but for zoomed in area of MDVs. Large dots in (c) and (d) for Explorer’s Cove. 166

Figure 4.28. (a) Backward trajectories run from 1800 UTC 29 December 2006. Trajectories run at 3 km, 4 km, and 5 km ASL for 84 hours, arrows plotted every 24 hours. (b) Potential temperature (K, red), equivalent potential temperature (K, blue), precipitable water (mm, green), and height (m, black) along the 5 km trajectory. 168

Figure 4.29. (a) Backward trajectories run from 1800 UTC 30 December 2006. Trajectories run at 3 km, 4 km, and 5 km ASL for 84 hours, arrows plotted every 24 hours. (b) Potential temperature (K, red), equivalent potential temperature (K, blue), precipitable water (mm, green), and height (m, black) along the 5 km trajectory. 169

Figure 4.30. (a) Backward trajectories run from 0600 UTC 1 January 2007. Trajectories run at 3 km, 4 km, and 5 km ASL for 84 hours, arrows plotted every 24 hours. (b) Potential temperature (K, red), equivalent potential temperature (K, blue), precipitable water (mm, green), and height (m, black) along the 5 km trajectory. 170

Figure 4.31. (a) Vertical cross section of potential temperature (K, contours), wind speed along the green transect in Fig. 4.8a (m s$^{-1}$, color shaded), and circulation vectors at 1200 UTC 31 December 2006. (b) Same as (a) except zoomed in to rotor. 171

Figure 5.1. Interannual variability of foehn days at LTER AWS sites for all foehn events identified in Polar WRF (“WRF ALL”, green), all foehn events identified in observations (“LTER ALL”, blue), and matched events in both Polar WRF and LTER observations (“BOTH”, red). “Corr” refers to correlation between LTER ALL and BOTH. “% of LTER” refers to the percentage of observed events in the BOTH set. 203

Figure 5.2. Spatial variability along transects in Taylor Valley (TG to EC), Wright Valley (LVA and LBr), and in Victoria Valley (LVi) for all foehn events identified in Polar WRF (“WRF ALL”, green), all foehn events identified in observations (“LTER ALL”, blue), and matched events in both Polar WRF and LTER observations (“BOTH”, red). 204

Figure 5.3. Averaged differences between Polar WRF and LTER observations (Polar WRF – LTER) at each site for event-average temperature (“TEMPAVG”), event warming (“WARMING”), degree days above freezing (“DDAF”), event-average relative humidity (“RHAVG”), and event-average wind speed (“WSAVG”). 206
Figure 5.4. Number of missed criteria for foehn events at each LTER site for (a) events in Polar WRF not in LTER and (b) events in LTER not in Polar WRF. ................................. 207

Figure 5.5. Scatterplot of cross-gap reduced pressure difference (hPa, x-axis) vs. near-surface gap wind speed (m s$^{-1}$, y-axis). Values color-coded according to gap wind speed and wind direction at 600 hPa. Cross-gap transect and wind speed and wind direction averaging area shown in Fig. 4.8a. .................................................................................................. 208

Figure 5.6. Orientation map for 2-km grid spacing simulations. Terrain height (m) shaded, LTER AWS sites shown by green dots, white transect used in cross-gap pressure calculations, white averaging box used for near-surface gap wind speed, and blue averaging box used for inverse Froude number calculations........................................... 209

Figure 5.7. Scatterplot of 600 hPa upstream wind direction (degrees, x-axis) vs. nondimensional mountain height (y-axis) for all observed foehn events at two or more sites. Averaging areas for wind direction and nondimensional mountain height shown in Fig. 4.8a. ................................................................................................................. 210

Figure 5.8. Scatterplots of 600 hPa wind direction (degrees, x-axis) vs. event-average wind speed (m s$^{-1}$, y-axis) for all observed foehn events at two or more sites for Lake Vanda (top left), Lake Brownworth (top right), Lake Bonney (bottom left), and Lake Fryxell (bottom right). ................................................................. 211

Figure 5.9. Scatterplots of 600 hPa wind direction (degrees, x-axis) vs. event-maximum temperature ($^\circ$C, y-axis) for all observed foehn events at two or more sites for Lake Vanda (top left), Lake Brownworth (top right), Lake Bonney (bottom left), and Lake Fryxell (bottom right). .................................................................................................................. 212

Figure 5.10. Scatterplot of time (x-axis) vs. event-maximum temperature ($^\circ$C, y-axis) for all observed foehn events at two or more sites for Lake Vanda (top left), Lake Brownworth (top right), Lake Bonney (bottom left), and Lake Fryxell (bottom right). Color-coding refers to 600 hPa upstream averaged wind speed, according to the key on bottom of the plot (m s$^{-1}$). ................................................................. 213

Figure 5.11. Barplot of the number of foehn events exceeding +3$^\circ$C maximum temperature at Lake Hoare based on the average upstream 600 hPa wind direction during the foehn period. ................................................................................................................. 214

Figure 5.12. Scatterplot of positive 500 hPa blocking indices from ERA-Interim for the 73 warmest foehn events at Lake Hoare (those with maximum observed temperature $>$ +3$^\circ$C) according to the blocking index of Wright (1994) at longitudes indicated on the x-axis. Numbers at the top of the plot correspond to the number of events with positive blocking index values at this longitude, and the percentage of the 73 total events. .... 215

Figure 5.13. Average 600 hPa wind vectors for all 73 foehn events at Lake Hoare exceeding +3$^\circ$C maximum temperature along latitudinal transects from 90°E to 90°W. Dashed lines represent zero phase lines of meridional wind. Reference vector shown to bottom right of plot. ................................................................. 216
Figure 5.14. JRA-25 MSLP differences for negative SAM seasons minus positive SAM seasons over the 1980-2008 period. The zero line marked in bold and negative MSLP shown by dashed line. From Speirs et al. (2011)........................................................... 217

Figure 5.15. Monthly standardized foehn anomaly compared with the standardized air temperature anomaly for Lake Hoare. Data area smoothed with a 5 month moving average. From Speirs et al. (2011)................................................................................. 218

Figure 5.16. ERA-Interim NDJF 2001 anomalies, relative to the 1989-2009 period, for (a) 500 hPa geopotential height (m) and (b) SST (K)................................................................................. 219

Figure 5.17. First four leading Varimax-rotated EOFs of DJF 500-hPa geopotential height anomalies from 1989-2010 ERA-Interim. Percentage of variance explained in top right of each plot. Signs of loading centers are arbitrary............................................... 220

Figure 5.18. Time series plots of the amplitude of each eigenvalue of the four leading EOFs presented in Fig. 5.15. The mean of each component time series is subtracted to give the amplitudes shown.............................................................................................. 221

Figure 5.19. ERA-Interim 500 hPa geopotential height anomalies for (a) November 2001, (b) December 2001, (c) January 2002, (d) February 2002. .......................................................... 222

Figure 5.20. 500 hPa geopotential height (contours, m) at 0000 UTC 15 December 2001 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity (10^{-5} \text{s}^{-1}), color shaded) and wind vectors....................................................................................... 223

Figure 5.21. Daily MJO index phase space plot from 1 December 2001 to 15 January 2002. X-axis represents PC for RMM1 and Y-axis represents PC for RMM2 (see text for explanation). Geographic locations along axes pertains to the approximate locations of enhanced convective signal of MJO for that respective phase. Dots represent daily values, with text labels of day of month every 5 days. Values inside the circle represent weak MJO activity. ......................................................................................................... 224

Figure 5.22. 500 hPa geopotential height (contours, m) at 0000 UTC 01 January 2002 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity (10^{-5} \text{s}^{-1}) color shaded) and wind vectors....................................................................................... 225

Figure 5.23. 500 hPa geopotential height (contours, m) at 0000 UTC 11 January 2002 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity (10^{-5} \text{s}^{-1}), color shaded) and wind vectors....................................................................................... 226

Figure 5.24. ERA-Interim November 1997 anomalies, relative to the 1989-2009 period, for (a) 500 hPa geopotential height (m) and (b) SST (K).......................................................... 227

Figure 5.25. 500 hPa geopotential height (contours, m) at 0000 UTC 10 November 1997 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity (10^{-5} \text{s}^{-1}), color shaded) and wind vectors....................................................................................... 228

Figure 5.26. Daily MJO index phase space plot from 15 October 1997 to 30 November 1997. X-axis represents PC for RMM1 and Y-axis represents PC for RMM2 (see text for explanation). Geographic locations along axes pertains to the approximate locations of
enhanced convective signal of MJO for that respective phase. Dots represent daily values, with text labels of day of month every 5 days. Values inside the circle represent weak MJO activity. ......................................................................................................... 229

Figure 5.27. ERA-Interim January 1996 anomalies, relative to the 1989-2009 period, for (a) 500 hPa geopotential height (m) and (b) SST (K)........................................................................................................ 230

Figure 5.28. 500 hPa geopotential height (contours, m) at 1200 UTC 28 January 1996 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity (10^{-5} \text{ s}^{-1}, color shaded) and wind vectors....................................................................................... 231

Figure 5.29. Daily MJO index phase space plot from 15 December 1995 to 31 January 1996. X-axis represents PC for RMM1 and Y-axis represents PC for RMM2 (see text for explanation). Geographic locations along axes pertains to the approximate locations of enhanced convective signal of MJO for that respective phase. Dots represent daily values, with text labels of day of month every 5 days. Values inside the circle represent weak MJO activity. ......................................................................................................... 232

Figure 5.30. MJO phases 10 days prior to (a) the 73 warmest foehn events at Lake Hoare and (b) all foehn events with southwesterly ambient flow (600 hPa winds from a direction > 203°). ........................................................................................................................................ 233
List of Tables

Table 3.1. Gridpoints in 2-km Polar WRF domain used for comparison with LTER AWS observations. Distance is the distance from the actual station location. 

Table 3.2. Gridpoints in 0.5-km Polar WRF domain used for comparison with LTER AWS observations. Distance is the distance from the actual station location. 

Table 3.3. MDV LTER AWS observation locations. 

Table 5.1. Sources of extra foehn events in WRF compared to LTER observations. Values represent average number of foehn events per year in WRF not found in LTER observations. Column categories described in text. Values in parentheses in second row of WRF ALL represent average number of observed events per year. 

Table 5.2. Sources of extra foehn events in LTER observations compared to WRF. Values represent average number of foehn events per year in LTER observations not found in WRF observations. Column categories described in text. Values in parentheses in second row of LTER ALL represent average number of events per year in WRF.
Chapter 1: Introduction

The McMurdo Dry Valleys (hereafter “MDVs”) are the largest ice-free region in Antarctica, a continent otherwise characterized by a perennial snow cover and extensive ice sheets. The MDVs are most notable for being one of the few areas on the continent with extensive biological activity, including flora and microorganisms existing in the soil, rocks, ephemeral streams, and melt lakes (see Fountain et al. 1999a and references therein). As such, the MDVs have been an NSF-funded Long-Term Ecological Research (LTER) site since 1993. Because the MDVs receive little precipitation (e.g., Bromley 1985; Monaghan et al. 2005; Fountain et al. 2009), melt from the surrounding higher-elevation glaciers provide most of the liquid water for melt lakes and streams in the MDVs (e.g., Chinn 1981; Clow et al. 1988; Fountain et al. 1998, 1999b), so there are important glaciological and hydrological applications in the region. The sharp temperature and moisture gradients across the MDVs are of interest for studies of soil properties (e.g., Campbell et al. 1998; Ikard et al. 2009). Furthermore, the strong winds and loose dry soil of the MDVs result in aeolian processes (e.g., Speirs et al. 2008). The preceding list highlights the interdisciplinary research in the MDVs. Meteorological studies have primarily been in support of the aforementioned applications, focusing on
wind regime variability (e.g., Doran et al. 2002a; Nylen et al. 2004) and surface energy balance (e.g., Thompson et al. 1971a; Dana et al. 1998; Hoffman et al. 2008).

Meteorological observations of the MDVs have consisted of continuous automatic weather station observations since the mid-1990’s, and before that only sporadic surface and lower-tropospheric meteorological field studies within individual valleys (e.g., McKendry and Lewthwaite 1990, 1992). Numerical modeling studies of MDVs meteorology have also been limited because of the complex terrain and the high spatial resolution necessary to properly represent the mesoscale and microscale effects. As a result of the limited observations and modeling efforts, meteorological research in the MDVs has been stagnant, limited to basic climatology and conjecture regarding the responsible physical processes. Still, this lack of sound physical understanding of MDVs meteorology has not prevented researchers from suggesting a link between the MDVs and large-scale climate, including climate change (Doran et al. 2002b) and Southern Hemisphere climate variability (Bertler et al. 2004, 2006). The hydrologic and biological conditions of the MDVs are sensitive to even minor climate fluctuations, so it is suggested that the amplified response of the MDVs can aid in detecting larger-scale climate changes.

With advances in computing power and numerical model development efforts both for complex terrain and polar environments, a realistic representation of the MDVs in numerical models is now possible. The work here provides a better understanding of MDVs meteorology and climate through a series of model simulations using a polar version of the Weather Research and Forecasting model (Polar WRF), a mesoscale
numerical weather prediction model tailored for high-latitude applications. The temporal and spatial coverage and resolution of the Dry Valleys region in the model simulations is set up to effectively study the various scales of processes, ranging from the effects of small-scale terrain features on wind and precipitation fields to effects of large-scale climate variability on MDVs climate. There are also a host of questions about the paleoclimate of the MDVs – primarily the presence of huge melt lakes during the last glacial maximum (LGM) – that are unresolved at this time. Although specific simulations addressing these paleoclimate questions could not be undertaken, the findings of this study allow for a fresh reexamination of theories. Initial progress towards modeling MDVs meteorology and climate has been done through use of the Antarctic Mesoscale Prediction System (AMPS), with results presented in Speirs et al. (2010). While the broad features of the winter foehn wind regime, and the synoptic-scale influences upon foehn winds in the Dry Valleys were found, the fine-scale features of the microclimates, the summer conditions, and the long-term climate variability cannot be studied with AMPS.

This study begins with a broad literature review first of the MDVs physical environment, including the intra-valley climate differences, wind regimes, and land surface (energy balance and soil properties). Climate variability is then discussed, focusing on large-scale modes of variability like ENSO and SAM, as well as intraseasonal variability. The last half of the chapter discusses mountain waves, as Speirs et al. (2010) identified wave activity during a prominent winter strong wind event in the MDVs. After basic theory is presented, downslope windstorms are discussed, including
mountain wave breaking, which is a prominent component of severe downslope windstorms. Foehn winds are then discussed, with the literature review pointing to the Alps, where considerable observational and modeling efforts have been undertaken over the past decade or so. It will be shown later that the Austrian Alps are a good analog to foehn effects in the MDVs.

The analysis in Chapter 4 focuses on high-resolution Polar WRF simulations of a summer MDVs foehn event in December 2006 / January 2007, where the case study is used to identify several dynamical features leading to strong winds and warming. New ideas are also presented for the down-valley extent of foehn winds and some easterly wind intrusions. In Chapter 5, we bring in 15 summers (November to February) of Polar WRF simulations and use the foehn event identification method of Speirs et al. (2010) to produce a critical evaluation of the ability of Polar WRF to properly simulate foehn conditions in the MDVs. While large discrepancies in wind speed and the number of foehn events exist, we show that Polar WRF can still be used to study conditions in the MDVs. The catalog of foehn events is then used to show that several of the features identified in Chapter 4 are robust across many foehn events. The chapter then concludes with effects of large-scale climate variability on MDVs climate, particularly towards foehn events, since the temporal distribution of foehn events primarily determines temperature variability. Unlike previous studies of Antarctic climate variability, the focus here is on intraseasonal timescales, since foehn variability also occurs on this timescale.
The end result of the research proposed here is a better understanding of the meteorology and the important summer climate in the MDVs, and how both vary spatially and temporally. A natural extension is then how climate change and variability in the MDVs relate to larger-scale continental changes. Central to the research are conceptual models of the prominent foehn winds, which are incomplete at this time. Once these wind regimes are understood, the MDV can be properly placed in the context of large-scale climate variability and climate change.
Chapter 2: Literature Review

This section begins with an overview of the McMurdo Dry Valleys (hereafter “MDVs”), specifically the geography and the physical characteristics. Discussion of the two prevailing wind regimes of the MDVs – the westerly foehn and easterly sea breeze – follows, as these wind regimes largely define MDVs climate. The extreme sensitivity of the MDVs to climate fluctuations has been exploited in Antarctic climate variability research, and this research is reviewed. Finally, theories on the existence of huge melt lakes in the MDVs during the last glacial maximum are discussed.

Background material on mountain wave theory and characteristics is then presented, as the MDVs are strongly affected by orographic effects. First, the linear theory of mountain waves is presented, followed by current explanations of downslope windstorms. Wave breaking and boundary layer effects are also discussed, as both can modulate the surface wind speed response. The chapter concludes with a recent review of Alpine foehn research, which is perhaps the best analog to foehn in the MDVs.

2.1 Physical Setting

The MDVs were first discovered by Robert Falcon Scott (Scott 1905), and have been the focus of wide-ranging scientific research, including biological activity (e.g.,
Moorhead and Priscu 1998), glaciers (e.g., Fountain et al. 1998), hydrology (e.g., Chinn 1981), aeolian processes (Ayling and McGowan 2006; Speirs et al. 2008), soil properties, and meteorology. The MDVs consist of three northeast-southwest oriented ice-free valleys situated to the west of Ross Island, within the Transantarctic Mountains (Fig. 2.1, from Speirs et al. 2010). Collectively covering an area of approximately 4800 km², the MDVs are the largest ice-free area of Antarctica. The MDVs are surmised to be snow-free because of a decrease in elevation of the inland ice sheet and rock thresholds along the western boundary of the valleys that block flow into the region (McKelvey and Webb 1962; Chinn 1990). However, the MDVs are located in a precipitation shadow of the Royal Society Range (Monaghan et al. 2005), and meteorological controls strongly affect accumulated snowfall. The widths of the valleys vary between 5-10 km, while the length of each valley is 40-50 km. The valley floor elevations range from 0-400 m ASL, and the valleys are separated by mountain ranges that rise to elevations upwards of 2000 m ASL. The MDVs generally slope upwards to the west from McMurdo Sound to the East Antarctic Ice Sheet. However, it is not a continuous rise up-valley, as the eastern ends of Wright and Victoria Valleys are blocked by Wilson Piedmont Glacier, and the eastern and western portions of Taylor Valley are separated by Nussbaum Reigal, a 700 m elevation hill.

In the valley basins, the MDVs feature several perennially ice-covered lakes, fed by ephemeral streams that flow from the surrounding glaciers during the summer melt season (Fountain et al. 1998, 1999a). Perennially ice-covered lakes require sufficiently low mean summer temperatures, ice sublimation rates high enough to match the rate of
freezing at the base of the ice cover, and peak summer temperatures above 0°C so that meltwater from local glaciers is available to flow into the lakes (Wilson 1982). The depth, temperature and salinity vary greatly between lakes, but the thickness of ice in each lake ranges from 3.5 m to 6 m, except for Lake Vida, which is frozen solid and ice-anchored to the bottom at 11 m depth (McKay et al. 1985). Yearly changes in lake level reflect variations of inflow, rather than ablation (sublimation), with ablation rates varying from 15 cm yr\(^{-1}\) to 50 cm yr\(^{-1}\) (Clow et al. 1988).

Campbell et al. (1998) divide the MDVs into five distinct soil environments. The coastal regions (1) feature sandy to silty soil, with only 1-3% clay. Soil temperatures here can exceed +14°C in summer, with diurnal amplitudes of 12°C. There are few freeze-thaw cycles, and the surface can be unfrozen for up to 40 days at a time. Gravimetric moisture content is ~1% in the upper 3 cm of soil, increases to 10% at the base of the active zone (30-60 cm), then rises abruptly in the underlying permafrost. The valley floors (2), up to 800 m elevation, consist of gravel along the slopes and sand on valley floors. Soil temperature extremes are greater than at the coast, and away from moisture sources gravimetric soil moisture is only about 0.5% down to 40 cm. Wetted margins along melt lakes or streams in the valleys range from about 1 to 11 m in horizontal length with lateral moisture gradients of about 5% per meter (Northcott et al. 2009). Along the valley sides (3), soils are sandy to silty, with about 1% gravimetric moisture content at depth. The upland valleys (4) feature a colder climate, with about 1.5% gravimetric moisture content, and slightly more clay than to the east. The plateau fringe (5) has the coldest climate, with coarse soil and little known about the moisture
content. As implied above, permafrost is ice-cemented along the coast and at lower elevations inland, with dry-frozen permafrost (permanently frozen soil that contains no ice) elsewhere in the region (Bockheim et al. 2007). The depth (or even the existence) of ground ice is dependent upon snow recurrence (McKay 2009). Dry permafrost in portions of Beacon Valley suggests a snow recurrence interval of 10 years. Soil moisture distributions are more complex near the melt lakes, with values near saturation at the shoreline to dry (<5 % gravimetric) only 10 m from shore (Ikard et al. 2009). Contrary to lower tropospheric air temperatures, no long-term trends in soil temperature have been observed (Hunt et al. 2010).

The contrasting climate of the surrounding areas (McMurdo Sound vs. the East Antarctic Ice Sheet), the ice-free valleys, and the complex topography of the MDVs set up microclimates that differ greatly from the surrounding regions, and often lead to varying meteorological conditions within individual valleys themselves. Mean annual temperatures recorded from several automatic weather stations (AWS) throughout the valleys range from -14.8°C (Lake Hoare) to -30°C (Lake Vida), with an overall north-to-south gradient across the valleys (Doran et al. 2002a). The annual range of temperatures in the MDVs (specifically Vanda station), both mean and extreme, are the highest of all stations on the continent (Keys 1980). Temperatures are about 7°C warmer in mid-summer and 5-7°C colder in mid-winter than adjacent snow-covered areas (Thompson et al. 1971a), and even on a daily timescale, air temperatures do not correlate with those at nearby Scott Base (Keys 1980). Mean annual relative humidity ranges from 55% at Lake Vanda to 74% at Explorers Cove, with relative humidity decreasing with distance from
McMurdo Sound at a rate of 0.3% per km (Doran et al. 2002a). Annual average values of downward shortwave radiation across several MDV observing sites are 84-117 W m\(^{-2}\) for 1994-1995 (Dana et al. 1998). These values are lower than most other sites around Antarctica (annual average value for Mawson, along the East Antarctic coast, is 160 W m\(^{-2}\), for example). Higher values elsewhere are due to the lower latitudinal position of some sites, greater transparency and aridity, and lack of orography. Thompson et al. (1971a) measure a positive radiation balance of about 37,600 J cm\(^{-2}\) annually (about 12 W m\(^{-2}\)) in Wright Valley, whereas over snow-covered areas at this latitude, there is normally a net loss. Variability between MDV sites is highly dependent upon topographic shading, and the variability becomes greater in spring and autumn when sun angle is lower. North facing slopes receive more solar radiation than south facing slopes.

The MDVs are a polar desert, with annual precipitation values ranging between 3 mm and 50 mm weq, and the gradient directed towards the coast. Only few light rain showers have ever been observed, so almost all precipitation is in frozen form (Bromley 1985). The snowfall can be episodic and highly variable between years (Bromley 1985), showing little seasonality inland, but generally greatest in winter near the coast (Fountain et al. 2009). Snowfall does not significantly contribute to the hydrology of the MDVs, since it quickly sublimes (Chinn 1981). Ablation easily exceeds accumulation in the valley basins, with about 70% of mass loss resulting from evaporation and sublimation (Fountain et al. 1998; Lewis et al. 1998; Fountain et al. 1999a). Fountain et al. (2009) estimate that the annual ratio of precipitation to accumulation is about 50%, as winds transport snow from valley sidewalls to the valley floors. Accumulated snow disappears
rapidly in summer, but in winter it can remain for a few days to a week in Taylor Valley, and can persist all winter in Victoria Valley (Fountain et al. 2009). Strong westerly winds can remove an extensive snow cover in 2-3 hours (Bromley 1985). However, precipitation is still important, as precipitation at higher elevations and wind-blown snow downvalley are the primary source of meltwater (Fountain et al. 1999a) and snow increases the albedo of the surface, thereby changing the surface energy balance such that less energy is available for melt (Dana et al. 1998; Fountain et al. 2009).

Glacial melt is the primary source of water to streams and melt lakes, and Hoffman et al. (2008) studied the surface energy balance over Taylor Glacier using 11 years of observations. Only 42 days featured melt over the 11 years, as ablation was dominated by sublimation. Melt is driven by solar radiation, but interestingly, above freezing temperatures did not always result in melt, and in some instances melt occurred during sub-freezing temperatures. Low wind speed was found to be critically important for melt initiation, as 76% of melt days occurred when wind speeds were 4 m s^{-1} or less. 14% of melt days occurred during overcast conditions, when downward longwave radiation exceeded the decrease in incident solar radiation. Over the annual cycle, solar radiation (55%) and sensible heat flux (40%) provide the most energy to the glacier, while most energy is lost to upward longwave radiation (79%) and latent heat flux (21%).

2.2 Wind Regimes

The controlling factor on MDVs climate is the winds. The wind direction is typically bimodal in summer, driven either by westerly foehn winds or easterly sea
breezes, with the latter being more common (McKendry and Lewthwaite 1992; Doran et al. 2002a). There is a diurnal cycle of summer winds, being easterly from late morning into evening, and westerly at night into the early morning, when the sun angle is lower (Bull 1966; McKendry and Lewthwaite 1992; Nylen et al. 2004). This diurnal pattern can be interrupted during strong westerly foehn events. In winter, these westerly foehn events are more common and are stronger. When foehn winds are not occurring, wind speeds are light more often than in the summer, and a weak katabatic flow develops down the surrounding slopes (Riordan 1975; Nylen et al. 2004 refer to this as “glacier flow”).

Understanding the foehn and sea breeze wind regimes is critical for determining the role that MDVs climate has on glaciers, hydrology, and biological production. Foehn winds are responsible for impressive warming events and strong wind speeds, particularly in the western portion of the MDVs. Foehn events are most dramatic in winter, and are described in detail by Bromley (1985) and Speirs et al. (2010). Durations of foehn events vary, but are typically 1-24 hours. The events commence rapidly, as wind speeds increase out of the west up to speeds of 45 m s\(^{-1}\), temperatures can rise by 25-35°C, sometimes by 50°C, and relative humidity falls to 6-10%, even to 1-2%. While adiabatic compression is largely responsible for the dramatic warming, the forcing for these events has been disputed. Some researchers (Bull 1966; Clow et al. 1988; Doran et al. 2002a; Nylen et al. 2004) have referred to the winds as “katabatic winds”, associated with katabatic surges from the polar plateau (e.g., Parish and Bromwich 1998), although little discussion has been presented as to the vertical structure and source regions for such events. Others (Thompson et al. 1971a; Riordan 1975; Bromley 1985; McKendry and
Lewthwaite 1990, 1992; Speirs et al. 2010) describe the winds as a foehn effect. Thompson et al. (1971a) and Bromley (1985) note the deflection of ambient southerly to southwesterly flow aloft into the MDVs. McKendry and Lewthwaite (1990, 1992), based on the only observational study of MDVs atmospheric vertical structure to date, support the foehn explanation, as the westerlies are confined to the valleys (particularly the western ends), whereas katabatic surges would be expected to be more widespread. They note that an inversion near ridge level satisfies an important pre-condition of downslope windstorms (e.g., Durran 1986a), and find evidence of a particular synoptic-scale meteorological sequence enabling downward deflection of flow in the western portions of the MDVs. This sequence is explained in Speirs et al. (2010), and will be discussed in Chapter 4.

In winter, cold pools in the valleys often modulate the onset of foehn winds at surface MDV observing sites. As a result of the cold pools, observed temperatures at glacier sites (along valley side walls) are about 10°C warmer than valley sites (Nylen et al. 2004). These cold pools form due to radiative heat flux divergence, which is eventually balanced by upward ground heat flux, and a drying mechanism such as water vapor deposition that prevents the formation of fog (Zängl 2005a). Downslope flows along the side slopes of a basin are of minor importance because the downward vertical motion is stifled by the strong static stability of the cold air pool (Clements et al. 2003). Cold pools can delay onset of foehn winds at certain valley observing sites, leading to large and nonlinear temperature variations across the MDVs. Doran et al. (2002a) and Nylen et al. (2004) hypothesize that cold pooling is responsible for Lake Brownworth
experiencing foehn onset earlier than Lake Vanda, as the strong westerly winds “ride over” Lake Vanda prior to erosion of the cold pool. An even stronger cold pooling effect appears to be responsible for less frequent foehn events at Lake Vida compared to other sites.

While foehn wind speeds and warming are most dramatic in winter, they are of greatest climatic importance in summer. Summer foehn events often push temperatures above freezing, and the dry conditions allow for substantial incident solar radiation, sublimation, and melt. At Taylor Glacier, 50% of the annual ablation occurs between mid-November and late January. Net solar radiation is substantially higher over the bare terrain, due to the lower albedo compared to snow-covered surfaces. Even a thin sediment layer on top of a snow cover can potentially double melt (Lewis 2001). In effect, the summer foehn regime drives the hydrologic cycle of the MDVs (Doran et al. 2002a). During quiescent synoptic-scale weather conditions in summer, a sea breeze forms during the daytime (Bull 1966; Clow et al. 1988), when the snow-free MDVs warm in comparison to the relatively cold water of McMurdo Sound to the east, and thermally generated pressure differences result in an on-shore, up-valley flow. Easterly flow results in cool, damp, and often cloudy conditions, that are reflected in climatological records (Doran et al. 2002a). The vertical structure of summertime flow is examined for the Wright Valley by McKendry and Lewthwaite (1990), from their observations of two case studies. A conceptual model from their study is shown in Fig. 2.2. An easterly “tongue” of cool and damp air penetrates inland, and is overridden by warm westerly winds, forming an inversion at the interface. Another inversion is found
above ridge level, separating the westerly winds associated with the topographically modified flow from ambient flow aloft. The processes responsible for the movement of the “front” between the westerly foehn and easterly sea-breeze are not well-understood, nor is the spatial variability across the valleys, as the observations from the McKendry and Lewthwaite studies are only done at one site in Wright Valley.

Orographic effects within individual valleys can modulate wind regimes in the MDVs. Fountain et al. (1999a) note that Nussbaum Reigal effectively confines weather conditions to each end of Taylor Valley, while Fountain et al. (2009) find that the gradient of accumulation (magnitude, east to west) drops from -2.6 mm weq km$^{-1}$ east of Nussbaum Reigal to -0.3 mm weq km$^{-1}$ to the west. In Wright and Victoria valleys, low clouds can form as easterlies rise over Wilson Piedmont Glacier (Bull 1966), resulting in lower RH values compared to Taylor Valley.

2.3 Climate Variability

Even small changes in the climate of the MDVs can have significant impacts upon meltwater production and subsequent biological activity (Chinn 1981; Dana et al. 1998; Fountain et al. 1999a). Summer temperatures are largely controlled by changing solar radiation levels, although diminished wind speeds result in surface warming as sensible heat fluxes are reduced (Clow et al. 1988). Changes in cloudiness directly impact the amount of incident solar radiation (Dana et al. 1998). The amount of incident solar radiation also affects the thickness of the ice cover over the valley lakes, which in turn impacts biological production (McKay et al. 1985). Snowfall, while not persisting
on the ground for long, does affect the albedo and net solar radiation (Dana et al. 1998; Fountain et al. 1999a). The sensitive nature of meltwater and biological production to climate fluctuations is illustrated by Doran et al. (2002b). Lake levels across the MDVs receded and lake ice cover thickened over the 1986-1999 period, while during this same time period, summer and autumn temperature decreases of 1.2°C/decade and 2.0°C/decade, respectively, occurred at Lake Hoare. While seasonally averaged wind speeds decreased, incident solar radiation increased 8.1 W m⁻²/decade. The result was decreases of biological production in lakes of 6-9% per year and in soils of 10% per year. This apparent paradox of increasing solar radiation and decreasing temperatures is resolved by the strong influence of winds on temperature in the MDVs. Doran et al. (2008) find that the anomalously warm austral summer of 2001/2002 restored lake level losses from the previous 14 years. Summer temperatures were on average 2.4°C warmer than the previous year, with no statistically significant difference in solar radiation. Adlam et al. (2009) found an anomalously deep active layer depth during this summer, with interannual variability in both timing and depth of maximum thaw strongly influenced by mean summer air temperatures. Hoffman et al. (2008) recorded ablation of 21 cm weq. at Taylor Glacier during summer 2001/02 (Dec.-Jan.), in contrast to 4-10 cm weq. during a typical summer. Especially for inland sites, melt events result from an increased frequency of westerly foehn winds, as there is a positive relationship between wind speed and degree-days above freezing.

Climatic changes in the MDVs can reflect changes in large-scale climate variability. The two primary modes of large-scale variability that influence Antarctica
are the Southern Annular Mode (SAM, Thompson and Wallace 2000) and the El Niño-
Southern Oscillation (ENSO, Turner 2004). Discussion of these Southern Hemisphere
teleconnections follows, with emphasis on processes that affect climate of the MDVs.

2.3.1 ENSO

The El Niño-Southern Oscillation (ENSO) refers to the unusually warm ocean
current along the west coast of South America near Peru and Ecuador, extending
westwards towards the International Date Line, every few years during the austral
summer (El Niño), and the associated atmospheric response (Southern Oscillation). The
opposite of El Niño is La Niña, which is an unusually cold ocean current along the west
coast of South America and the eastern and central tropical Pacific, and the associated
atmospheric response. In accordance with the variations in SST between El Niño and La
Niña events, several definitions of ENSO pertain to SST in key regions of the Pacific
Ocean. An often-used definition is the “Nino 3.4” region (Trenberth 1997), which refers
to time-averaged SST anomalies in the region bounded by 5°N-5°S and 120°W-170°W.
Other definitions are similar, with varying geographic regions.

The atmospheric response to ENSO involves a zonal oscillation of pressure
between the western Pacific (near Indonesia) and the central Pacific. Accordingly, the
Southern Oscillation Index (SOI) is defined as the twice-normalized difference in surface
pressure between Tahiti and Darwin. Positive phases of the SOI correspond to La Niña
(cold) events, as cold SSTs in the eastern and central Pacific shift the primary region of
tropical convection westward, and lower pressures are found in the region of Indonesia
and northern Australia than over the central Pacific. Likewise, negative phases of the SOI correspond to El Niño (warm) events, where warm SSTs in the eastern and central Pacific shift the primary region of tropical convection eastward, and lower pressures are found over the central Pacific.

While effects of ENSO can be observed globally, the focus here is on the teleconnection of ENSO to southern high latitudes and Antarctica. Recent reviews of the ENSO teleconnection to the Antarctic can be found in Carleton (2003) and Turner (2004). Karoly (1989) identifies a Rossby wave train extending from the tropics to high southern latitudes, as deep convection near the equator generates waves in the upper-tropospheric geopotential height field. A schematic of this Rossby wave pattern during the early period of an ENSO event is shown in Fig. 2.3. This Rossby wave train is more formally known as the Pacific South American (PSA) teleconnection (e.g., Mo and White 1985; Mo and Higgins 1998; Mo 2000; Mo and Paegle 2001). The PSA consists of two modes: the first (PSA1) is the Southern Hemisphere response to ENSO, operating on time scales of 40-48 months, and PSA2 is associated with the quasi-biennial oscillation (QBO) on timescales of about 26 months (Mo 2000). ENSO events typically develop during the austral winter, with small but increasing anomalies of SST and convection in the central equatorial Pacific. ENSO events reach their mature stage during the following austral summer. The results of Houseago et al. (1998) support the wave-train pattern explanation through Hovmöller analysis, while also noting the high degree of variability between individual ENSO events (both warm and cold). They bring up the possibility that the anomalous response of the 1982 El Niño event is due to wave trains propagating
equatorward from Antarctica. Garreaud and Battisti (1999) agree with the role of the PSA towards interannual climate variability in the Southern Hemisphere, while also finding interdecadal (ENSO-like) climate variability associated with wave-mean flow interactions in high latitudes.

More recent studies summarize the proposed mechanisms for ENSO teleconnection to the southern high latitudes. Liu et al. (2002) propose that an equatorward shift in the subtropical jet during El Niño events shifts the storm track equatorwards. The mid-latitude Ferrel cell is modified from associated changes in the meridional eddy heat flux and latent heat release, which affects conditions in the high latitudes through changes in the mean meridional heat flux. However, it appears that the teleconnection can be explained more succinctly through Rossby wave propagation. Kidson and Renwick (2002) again find that Rossby waves are initiated by convection in the central equatorial Pacific and force an anomalous anticyclonic circulation near 60°S in the eastern Pacific during El Niño events. Lachlan-Cope and Connolley (2006) analyze the forcing for Rossby waves, based on the vorticity equation. Forcing is largest in areas of divergence, divergent wind, absolute vorticity, and gradients of absolute vorticity. The forcing is largest east of Australia. Areas of large-scale ascent are necessary for Rossby wave generation, as the required SST increase to force ascent in regions of weak ambient ascent or descent is not realistic. Overall, the authors state that the processes described by Liu et al. (2002) are equivalent to those described by Rossby wave propagation.
ENSO teleconnections to the southern extratropical latitudes are most prominent in the South Pacific, and one such manifestation is the split jet near New Zealand. The split jet consists of the subtropical jet (STJ) near 30°S and the polar front jet (PFJ) near 60°S. The split jet is most pronounced in winter and early spring (Trenberth 1975), and is thought to exist due to the displacement of Antarctica from being centered at the South Pole (James 1988). Chen et al. (1996) find that interannual variations of the split jet are highly correlated with SST anomalies in the central equatorial Pacific. The STJ strengthens (weakens) and the PFJ weakens (strengthens) during El Niño (La Niña) events. Changes in the STJ result from an increase in eddy momentum convergence, while PFJ changes result from poleward synoptic eddy momentum flux and equatorward 7-30 day eddy momentum flux. Bals-Elsholz et al. (2001) find that the split jet varies in magnitude between years, but not in location. They do not find a strong correlation between split flow and ENSO. Instead, split flow is strongly modulated by synoptic-scale effects on the PFJ that regulate cold air extrusions from Antarctica.

As previously explained, ENSO effects are teleconnected to the Antarctic via Rossby wave trains, and the effects are largest in the Pacific sector. Kwok and Comiso (2002a) construct composites of El Niño and La Niña events and find the strongest high latitude correlations between ENSO and various meteorological parameters in the Bellingshausen, Amundsen, and Ross seas. Lower (higher) sea-level pressure is found in the Amundsen Sea during La Niña (El Niño) events. The associated circulation pattern differences between ENSO regimes result in climate variations over West Antarctica. Cullather et al. (1996) analyze the moisture convergence into a sector of coastal West
Antarctica, where nearly 40% of moisture transport into Antarctica occurs. The position of the Amundsen Sea low is the primary factor influencing West Antarctic precipitation variability associated with ENSO. During normal or La Niña events, the low is weaker and located to the west in the Ross Sea. Direct moisture transport occurs into West Antarctica. In contrast, for El Niño events the low is located farther east, so that moisture transport is directed towards the Antarctic Peninsula region. There is a positive correlation between ENSO and West Antarctic precipitation prior to 1990, which becomes negative thereafter (Cullather et al. 1996; Bromwich et al. 2000). In the 1990s there is an eastward shift in the tropical convection associated with ENSO, and an easterly shift in the SE Pacific blocking pattern that modulates moisture transport to West Antarctica (Bromwich et al. 2004; Guo et al. 2004). Harangozo (2000) also finds the importance of ENSO-induced circulation variations in the SE Pacific to be important for west Antarctic Peninsula climate in winter. Tropical SST values are found not to be reliable for determining the teleconnection response to the Antarctic Peninsula region. Instead, the coldest winters are associated with reduced SST cooling in the central tropical Pacific (see also Vera et al. 2004) and with increased pre-winter sea ice extent. Near-surface air temperatures in the Peninsula region are modified indirectly through changes to the local sea ice extent from atmospheric circulation changes. Pezza et al. (2008) analyze Southern Hemisphere cyclone activity during different ENSO phases, and find a variable response around Antarctica. Cyclone system density and depth both tend to increase in the Bellingshausen Sea and decrease in the Ross Sea for La Niña conditions.
ENSO effects on the broader continent of Antarctica are mainly dependent upon
effects of the circulation over the Pacific sector on the continental interior. Smith and
Stearns (1993) find a change in surface pressure and temperature anomalies across El
Niño events. The Ross Sea and Ross Ice Shelf area shift from positive pressure and
negative temperature anomalies before the SOI minimum to negative pressure and
generally weak negative to slightly positive temperature anomalies after the SOI
minimum. However, these results are based on only manned station observations, so the
analysis is questionable in data-sparse regions of the continent.

Because of the effects on the atmospheric circulation in the South Pacific, ENSO
variability has been linked to that of sea ice. Carleton (1989) finds that much of the
connection between ENSO and sea ice in previous studies results from the temporal
autocorrelation of sea ice and SOI. When this autocorrelation is removed, much of the
sea ice extent variability can be explained by chance, however in regions of strongest
correlation (Ross and Weddell sea sectors), the sea ice extent lags the SOI by several
months. Simmonds and Jacka (1995) also find that the strongest relationships occur
when SOI leads the sea ice. In the SW Pacific, sea ice in fall and early winter is
positively related to SOI in the previous 12 months. In the SE Pacific, sea ice extent is
negatively correlated with SOI in the previous 6 months. Ledly and Huang (1997) again
find a lagged correlation for both SST and sea ice extent with ENSO, with increased SST
and reduced sea ice extent in the Ross Sea sector during three El Niño events from 1989-
extent and atmospheric fields from the NCEP-NCAR reanalysis. Up to 34% of the
variance in sea ice extent is linearly related to ENSO, as sea ice extent has high correlations with tropical Pacific precipitation and tropical Indian Ocean SSTs. Correlation of sea ice extent with global surface temperature reveals four teleconnections: an ENSO-like pattern in the tropics, teleconnection between the eastern Pacific region of the Antarctic and the west-central tropical Pacific, an Antarctic dipole across the Drake Passage, and meridionally-oriented teleconnections in the central Pacific and Atlantic from the polar regions to the tropics. The Antarctic dipole (ADP) has received further attention (Yuan and Martinson 2001; Renwick 2002; Yuan 2004). It is an out-of-phase relationship of sea ice extent and surface air temperature anomalies between the eastern Pacific and Atlantic sectors of the Antarctic. It is the dominant interannual variability pattern in sea ice extent and surface air temperature in the Pacific and Atlantic south polar regions. The ADP responds more to La Niña than to El Niño, likely because of the large variability in atmospheric response to El Niño events. The formation of the sea ice anomalies differs between El Niño and La Niña events. For El Niño, a blocking high is established in the SE Pacific during austral spring which generates the ADP anomalies. For La Niña, the ADP anomalies are not established until summer. The main ENSO impact on sea ice concentration occurs near the sea-ice edge, due to stationary eddy heat fluxes. Transient eddies have little effect upon the sea ice distribution. Eastward propagation of sea ice concentration anomalies comprising the ADP is associated with the PSA (Udagawa et al. 2009). The atmosphere creates the ADP, but the ocean carries it eastward.
Figure 2.4, from Yuan (2004), provides a summary of the ENSO teleconnection to high southern latitudes, from the perspective of atmospheric circulation changes, that leads to formation of the ADP. Recall that an equivalent explanation for the surface pressure anomalies in the SE Pacific can be explained through Rossby wave propagation and the PSA. Variability of the split jet components follows from Chen et al. (1996), and variability in the Ferrel cell and storm track follows from Liu et al. (2002). A positive feedback between the jet stream, the storm track, and stationary eddy activity in the SE Pacific prolongs and maintains the blocking high-pressure region associated with the PSA during El Niño events.

While large-scale climate variability has generally not been analyzed on a scale that differentiates the MDVs from greater Antarctica, a few recent studies have postulated on large-scale climate variability effects in the MDVs. Bertler et al. (2004) claim that the cooling trends in the MDVs found by Doran et al. (2002a) are due to ENSO variability. Over most of the 1986-2000 period of MDVs cooling, a warm ENSO phase (El Niño, SOI index < 0) was in place, resulting in the Amundsen Sea low being shifted eastward compared to cold ENSO phases (Cullather et al. 1996; Bromwich et al. 2000, Marshall 2009). This eastward shift directs flow across West Antarctica and onto the Ross Ice Shelf, bringing colder air to the MDVs, compared to La Niña phases, when more maritime easterly flow impacts the MDVs. Bertler et al. (2006) reconstruct summer temperature and moisture variability over the previous four decades from stable oxygen isotopes and snow pits near the MDVs. They find phase-dependent statistically significant correlations between SOI and summer temperatures in the MDVs. Still, the
assumptions in these studies are broad, and the modes of large-scale climate variability need to be placed in context with the meteorological processes responsible for foehn winds, which are not well understood. Detailed, long-term simulations are necessary to determine the role of ENSO-related circulation and sea ice variability on MDVs climate.

2.3.2 SAM

The dominant mode of atmospheric circulation variability in the high latitudes of the Southern Hemisphere is the Southern Annular Mode (SAM), also known as the high-latitude mode or the Antarctic Oscillation (AAO) (Thompson and Wallace 2000). The SAM is represented by the leading empirical orthogonal function (EOF) of the monthly 500-hPa geopotential height field (e.g., Rogers and van Loon 1982; Kiladis and Mo 1998). Alternatively, the SAM can be represented by the Antarctic Oscillation Index (AOI), defined as the difference between the normalized zonal mean sea-level pressure at 40°S and 65°S (Gong and Wang 1999). The SAM is manifested as zonal pressure/height anomalies having opposing signs in the mid-latitudes and high-latitudes. The positive phase implies higher pressure in mid-latitudes and lower pressure over Antarctica, resulting in stronger circumpolar westerlies. The pressure/height anomalies translate to a greater density of cyclones around coastal Antarctica for positive SAM than for negative SAM (Pezza et al. 2008). The negative phase results in a slackening of the pressure gradient and weaker circumpolar westerlies. Positive SAM is associated with cold anomalies over most of Antarctica, as the strengthened zonal flow isolates the continent from warm air intrusions (Thompson and Wallace 2000; Kwok and Comiso 2002b). The
exception is the northern portion of the Antarctic Peninsula, which receives a greater maritime influence during positive SAM. The SAM is forced internally – forcing SST to the long-term mean in model simulations does not produce extensive differences in circulation (Limpasuvan and Hartmann 2000). Zonal wind variability associated with the SAM is maintained by transient eddies on synoptic timescales, while the Coriolis effect maintains zonal wind anomalies against friction near the surface (Limpasuvan and Hartmann 1999, 2000). The SAM has a more symmetrical pattern in summer, attributed to the seasonal cycle of near-surface temperature over Australia (Kidston et al. 2009).

The SAM has trended towards the positive phase over the last few decades (Thompson et al. 2000), with the greatest trends occurring in austral summer (Thompson and Solomon 2002; Marshall 2003; Marshall et al. 2004). The SAM trend has been attributed to ozone loss in the lower stratosphere, which results in cooling and a strengthening of the polar vortex (e.g., Turner et al. 2009). However, Marshall et al. (2004) note that the summer trend in the SAM began at least a decade before ozone loss, so that anthropogenic forcing is also an important component of the SAM trend. When the SAM is in a positive phase, a northward surface Ekman drift in the ocean leads to downwelling at 45°S and upwelling along the Antarctic coast, leading to greater ice extent (Lefebvre et al. 2004). However, the non-annular component of the SAM has the greatest influence on sea ice, with the largest effects in the Ross, Amundsen, and Bellingshausen seas. During positive SAM phases, ice area increases (decreases) in the Ross and Amundsen (Weddell) seas. The long-term trends in sea ice, while agreeing
with those occurring during positive SAM, are apparently not related to the trend in the SAM.

2.3.3 ENSO-SAM Interactions

The most recent research on large-scale Southern Hemisphere climate variability focuses on interactions between ENSO and SAM. Kwok and Comiso (2002b) find that the combined trend towards positive SAM and negative SOI over recent decades has lead to a cooling trend over much of East Antarctica and warming over the Antarctic Peninsula. While the SAM accounts for much of the surface temperature variability over sea ice regions, ENSO accounts for more of the changes in ice extent and SST variability in the Pacific sector of the Antarctic. Carvalho et al. (2005) find that negative (positive) SAM phases are dominant when tropical SST and convection anomalies reflect El Niño (La Niña) conditions. Changes in deep convection in the central equatorial Pacific, related to either El Niño or an eastward-propagating Madden-Julian Oscillation (MJO, see Matthews and Meredith 2004; Pohl et al. 2010), favor negative SAM phases during the austral summer. Fogt and Bromwich (2006) make the case that the decadal variability in the ENSO teleconnection to the South Pacific observed during the 1980s and 1990s is related to interactions with the SAM. It is found that ENSO and SAM must be in phase relative to each other (i.e., SAM+/SOI- or SAM-/SOI+) in order for the ENSO teleconnection to be effective in the South Pacific. The lack of coupled SAM/ENSO events during austral spring in the 1980s resulted in Rossby waves not modifying the zonal circulation, and disrupting the transport of the ENSO signal to high
latitudes. The southeastward shift in the teleconnection in summer from the 1980s to the 1990s resulted from changes in the structure of the SAM and strong ENSO events in the late 1990s that produced an amplified PSA2 pattern. The authors support previous findings that SAM trends are explained not just by ozone depletion, but also by anthropogenic and natural climate variability.

While Sterl et al. (2007) question the robustness of ENSO teleconnection changes with time, due to the short study period of Fogt and Bromwich (2006), results from the 50-year study period of Fogt et al. (2010) support the concept of ENSO-SAM interactions. The ENSO-SAM phase relationships are significant and occur more often than expected by chance. Gregory and Noone (2008) support the ENSO-SAM connection, finding that the linkage between ice core isotope records and ENSO in the Amundsen and Bellingshausen seas only appears when SAM and SOI are in phase. Correspondence is seen in cyclone density response between La Niña, negative SAM, and negative sea ice extent anomalies around coastal Antarctica (Pezza et al. 2008). The South Pacific ENSO teleconnection results from the interaction of anomalous transient eddy momentum fluxes from both SAM and ENSO (L’Heureux and Thompson 2006; Fogt et al. 2010). The ENSO teleconnection to high southern latitudes from anomalous wave activity is reduced by opposing strong SAM events. Gong et al. (2010) expand upon the ENSO-SAM connection by noting that the zonal-mean background flow associated with La Niña (El Niño) events preconditions the atmosphere for strong (weak) anticyclonic Rossby wave breaking on the equatorward side of the eddy-driven jet, which excites positive (negative) SAM events. Stammerjohn et al. (2008) find observed
changes in ice season duration of -85 ± 20 days in the Antarctic Peninsula/Bellingshausen Sea region and +60 ± 10 days in the western Ross Sea from 1979-2004. Ice season changes (changes in timing of annual retreat and growth) are primarily forced by atmospheric circulation changes. Changes in the zonal circulation associated with SAM and ENSO phases lead to the aforementioned observed changes in sea ice extent.

The effects of large-scale modes of climate variability on the MDVs, especially the SAM and SAM-ENSO interactions, have received virtually no attention in the literature. Such localized effects in the MDVs cannot be resolved from current and previous research efforts that utilize relatively coarse numerical model output and observational datasets. Therefore it is necessary to perform high-resolution numerical modeling on climate time-scales to properly assess large-scale climate variability in the MDVs.

2.3.4 South Pacific Wave and Intraseasonal Variability

The fourth EOF mode in the Southern Hemisphere has been termed the “South Pacific Wave” (SPW) by Kidson (1999), and features a wavetrain extending from northern Australia southeastward to the Antarctic Peninsula region, and then curving northwards and eventually into the Indian Ocean (e.g., Kiladis and Mo 1998; Kidson 1999). It is a wavenumber 3 pattern during winter and a wavenumber 4 pattern during summer, and represents an amplification or suppression of the climatological stationary waves over the Southern Hemisphere. Frederiksen and Zheng (2007) separate 500 hPa geopotential height fields into intraseasonal and slowly varying components, and also
find the fourth EOF in both winter and summer to represent the SPW. They find it to be contemporaneously correlated with SST in the subtropical Indian Ocean.

Intraseasonal variability has been studied using 10-50 day bandpass filtered geopotential height anomalies (e.g., Kiladis and Mo 1998). Kidson (1999) finds wavenumber 4 and 5 patterns confined to the Southern Hemisphere storm track along 50°S. Frederiksen and Zheng (2007) find the largest loadings in the South Pacific region and zonal wavenumber 3 (winter) and 4 (summer) patterns after the dominant SAM pattern. Intraseasonal variability has primarily been attributed to the Madden-Julian Oscillation (MJO) (e.g., Kiladis and Mo 1998; Zhang 2005). The MJO is an eastward-propagating tropical convection disturbance that has a period in the range of 30-60 days, but the cycle is highly episodic, and is better described as a discrete pulse-like event than a wave. It originates over the central Indian Ocean, and moves eastward at 5-10 m s⁻¹ towards the maritime continent and then into the western Pacific. The peak season for the MJO is in austral summer, when the strongest convective signals are just south of the equator. During El Niño events, the eastern edge of the warm pool extends eastward, as does the MJO. The MJO is particularly strong prior to the peak of an ENSO warm event (i.e., the “gold standard” MJO of 1996-1997, prior to the onset of the 1997-1998 El Niño event), but the interannual variability is driven more by internal atmospheric dynamics than SST anomalies.

The extratropical response to the MJO tropical convection anomalies is a Rossby wavetrain that extracts energy from the mean flow (Matthews et al. 2004; Matthews and Meredith 2004). It takes one to two weeks for the extratropical response to be set up
from tropical convective forcing, but this extratropical response is the dominant mode of intraseasonal variability. During austral summer, MJO effects extend into the southern subtropics (Wheeler and Hendon 2004), where the response over Antarctica is generally muted (Pohl et al. 2010). Yet, tropically forced Rossby wave trains can have direct influence on near-surface temperatures over coastal Antarctica on intraseasonal timescales (Yu et al. 2011).

2.3.5 The Austral Summer of 2001/2002

The austral summer of 2001/2002 featured anomalously large glacial melt in the MDVs, as described earlier in this section. This season also featured warm conditions elsewhere in Antarctica, such as interior East Antarctica (see next subsection), as well as the Antarctic Peninsula region. Gamble (2003) provides a climatic summary of the 2001/2002 austral summer. Near-neutral ENSO conditions continued from earlier in 2001, while in December an OLR index focused on the western Pacific is in its most negative phase since October 1997, related to an active MJO phase. SSTs are above normal in the western equatorial Pacific, and an anomalously strong blocking index is found at almost all longitudes, strongest in January and February in the Australian region. Turner et al. (2002) attribute the warm conditions to a deeper than normal circumpolar trough, especially from the Amundsen Sea to the Weddell Sea, and an amplified wavenumber-3 pattern that brought more mid-latitude intrusions of warm maritime air into the continent. Massom et al. (2006) note that the period featured large positive SAM conditions.
2.3.6 Blocking

Blocking is defined as persistence for a week or more of a major split in upper level flow, a significant positive geopotential anomaly, or anomalous easterly 500-hPa geostrophic flow (Sinclair 1996). Persistent blocking is found to only occur in two regions of the Southern Hemisphere: southeast of New Zealand (associated with the split jet) and west of South America near 55°S (to be discussed). Simultaneous blocking in both regions is rare (Sinclair et al. 1997). Blocking frequency in the southeast Pacific increases with El Niño, while it varies less systematically over the southwestern Pacific (Renwick 1998; Renwick and Revell 1999; Wiedenmann et al. 2002; Sáez de Adana and Colucci 2005). Renwick (2005) uses a cluster analysis of NCEP-NCAR and ERA-40 reanalysis output to find a zonal-wave 3 pattern that has some indication of ENSO forcing, but weaker and in an opposite sense to that of SE Pacific blocking. The wave propagation pattern for such events involves a strong negative geopotential height anomaly in the SE Pacific concurrent with a blocking high in the SW Pacific. Hobbs and Raphael (2010) explore the month-to-month variability in zonal wavenumbers 1 and 3 patterns in the southern hemisphere. The climatological high in the SW Pacific represents variability in west Antarctica that is not captured by SAM. They attribute the strength and position of the SW Pacific anticyclone to the PSA1 mode, and the SE Pacific anticyclone to the PSA2 mode.

Blocking highs have been attributed to enhanced meridional flow and increased precipitation over interior East Antarctica. Bromwich et al. (1993) find that upper-level
ridging over Wilkes Land and a blocking anticyclone south of Australia and New Zealand, with a split-jet in the southwest Pacific leads to enhanced cold-air drainage from East Antarctica. Murphy and Simmonds (1993) show that this synoptic-scale pattern results in a strengthened pressure gradient across Wilkes Land and a continental wavenumber-4 pattern. Goodwin et al. (2003) similarly find that enhanced ridging weakens zonal westerly winds and produces an anomalous meridional wind component across Wilkes Land. Massom et al. (2004) look at three case studies, which all feature an anomalously strong blocking high south of New Zealand. Just a few episodes of blocking can contribute a significant portion of the small annual precipitation to inland regions. Two deep incursions in December 2001 and January 2002 brought 27% of the mean annual accumulation over a combined 11-day period. The amplified longwave patterns associated with blocking activity bring warm air into interior Antarctica, as Vostok recorded a temperature of -16.5°C on 11 January 2002, nearly a record high. South Pole reached -18.1°C on the same date. Fujita and Abe (2006) similarly find that significant precipitation events at Dome Fuji are associated with warm intrusions. Scarchilli et al. (2010) note that about 50% of the strong precipitation events at four ice core sites in East Antarctica feature blocking events in the Tasman Sea-eastern Indian Ocean at a higher latitude than normal, with warm and moist air directed into the continent.

2.4 Paleoclimate
Melt lakes within the MDVs were much deeper (up to 450 m greater depth) and larger (up to 20 times larger, filling some entire valleys) during the last glacial maximum (LGM) than at present (e.g., Denton et al. 1989; Hendy 2000; Hall et al. 2001, 2002). At the same time, the Ross Ice Shelf (better described as the Ross Ice Sheet) extended several hundred kilometers farther northward than it currently does. The region just east of the MDVs was ice-covered in the summer, rather than being the open water of McMurdo Sound as it is today. These differences had profound effects upon the climate of the MDVs, resulting from the elimination of the summer sea-breeze circulation, and from modification of the regional storm track (Morse et al. 1998). Ice core records from Taylor Dome (just west of the MDVs) indicate that annual mean temperatures were about 4°-8°C colder than at present (Steig et al. 2000), and that the region was much more arid than today (snow accumulation of about 20% of contemporary values).

More extensive melt lakes during a time of a colder and more arid climate is perplexing. There are two theories as to why the seemingly contradictory conditions existed. The first is that solar radiation was greater, due to the arid conditions and reduced maritime influence (e.g., Denton et al. 1989; Hall and Denton 1996; Hall et al. 2001, 2002). Greater solar insolation could be responsible for greater melt, even with colder temperatures, as a significant portion of meltwater can form just below the surface of a glacier from absorption of solar radiation, even at temperatures below freezing (Fountain et al. 1998; Hendy 2000). Decreased precipitation would also increase surface heating by reducing or removing the damping effect of snow albedo upon the surface energy balance. The second theory attributes melt lake generation to more frequent
episodic warming events from MDV foehn winds (Doran et al. 2002b, 2008). As previously discussed, greater melt and streamflow results from more frequent westerly wind events in the MDVs. While Doran et al. (2002b) did find an increase in solar radiation over their 1986-2000 study period, Doran et al. (2008) note that the differences in solar radiation between “flood” and “non-flood” years are insignificant. A westerly wind of greater horizontal extent would be expected with no competing sea breeze, possibly explaining warmer conditions across the MDVs.

Estimation of the high-latitude Southern Hemisphere circulation during the LGM was done by Drost et al. (2007) using the HadCM3 climate model. The amplitude of the semi-annual oscillation during the LGM was found to be smaller than present (represented by the NCEP-NCAR reanalysis), associated with an enhanced wavenumber 3 pattern and equatorward shift of the subpolar trough during the LGM. The SAM was slightly stronger, with stronger westerly winds at high latitudes than present. However, these results are taken with caution because the simulations contained numerous model-related anomalies.

2.5 Mountain Waves

In this section, we diverge from the MDVs and Antarctica to a more general description of mountain wave theory and applications. Discussion will be limited to topics that allow for a physical interpretation of simulated mountain waves in the MDVs and that pertain to the Antarctic environment. Since observations and simulations of mountain waves in the vicinity of the MDVs are almost non-existent, real-world
applications focus on two regions – the front range of the Rocky Mountains (Boulder, Colorado) and the Alps – where observational and modeling efforts have been extensive. A review of recent mountain wave studies associated with the Mesoscale Alpine Programme (MAP) can be found in Smith et al. (2007).

2.5.1 Linear Theory

Several reviews of linear mountain wave theory have been done (e.g., Smith 1979, 1989; Gill 1982; Durran 1986b, 1990; Lin 2007), and material from those reviews is presented here. We start by linearizing the governing equations, and assuming a steady state, two-dimensional (x and z), adiabatic, nonrotating, inviscid, Boussinesq flow over a small-amplitude mountain, to get

\begin{align}
  & u_0 \frac{\partial u'}{\partial x} + \frac{d u_0}{d z} w' + \frac{1}{\rho_0} \frac{\partial p'}{\partial x} = 0 \tag{2.1} \\
  & u_0 \frac{\partial w'}{\partial x} + \frac{1}{\rho_0} \frac{\partial p'}{\partial z} - g \frac{\theta'}{\theta_0} = 0 \tag{2.2} \\
  & u_0 \frac{\partial \theta'}{\partial x} + \frac{N^2 \theta_0}{g} w' = 0 \tag{2.3} \\
  & \frac{\partial u'}{\partial x} + \frac{\partial w'}{\partial z} = 0 \tag{2.4}
\end{align}

which are the horizontal momentum equation (one-dimensional), the vertical momentum equation, a density equation, and the continuity equation, respectively. Primes indicate a perturbation quantity, and 0-based subscripts indicate basic-state (mean) quantities. \( N \) is the buoyancy frequency (or Brunt-Väisälä frequency), defined as

\[ N^2 = \frac{g}{\theta_0} \frac{d \theta_0}{dz}. \tag{2.5} \]

These equations are combined into a single equation for \( w' \) to get
\[
\frac{\partial^2 w'}{\partial x^2} + \frac{\partial^2 w'}{\partial z^2} + I^2 w' = 0, \tag{2.6}
\]

where

\[
I^2 = \frac{N^2}{u_0^2} - \frac{1}{u_0} \frac{d^2 u_0}{d z^2} \tag{2.7}
\]
is the Scorer parameter. We assume for now that \(u_0\) and \(N\) are both constant with height (the second term on the RHS of (2.7) then goes to zero), and that air is flowing over an infinite series of periodic ridges described as

\[
h(x) = h_m \cos kx, \tag{2.8}
\]

where \(h_m\) is the maximum height of the ridges and \(k\) is the horizontal wavenumber. With sinusoidal forcing by the topography, we look for solutions to (2.6) of the form

\[
w'(x,z) = \hat{w}_1(z) \cos (kx) + \hat{w}_2(z) \sin (kx), \tag{2.9}
\]

where the \(\hat{w}\) terms are only functions of height. Substituting (2.9) into (2.6) and simplifying yields

\[
\frac{d^2 \hat{w}_i}{dz^2} + (I^2 - k^2) \hat{w}_i, i = 1, 2 \tag{2.10}
\]

We now have an ordinary second-order linear partial differential equation, the solution of which depends upon the sign of the vertical wavenumber,

\[
m(z) = (I^2 - k^2)^{1/2}. \tag{2.11}
\]

If \(m^2\) is negative \((I^2 < k^2, \text{ or } N/U < k)\), we find solutions to (2.10) of the form

\[
\hat{w}_i = A_i e^{\mu z} + B_i e^{-\mu z}, \tag{2.12}
\]

where \(\mu^2 = -m^2\). To determine the coefficients, we first note that it is unphysical for the wave amplitude to grow exponentially with height, so \(A_i = 0\). To determine \(B_i\), the lower boundary condition is set as
meaning that the flow follows the terrain at the lower boundary. Substituting (2.12) and (2.13) into (2.9), we obtain the solution

\[ w'(x, z) = -u_0 h_m k e^{-\mu z} \sin kx. \]  

Equation (2.14) states that the wave amplitude decays with increasing height. Such waves are known as *evanescent waves*, and are shown schematically in Fig. 2.5a (from Durran 1990). Evanescent waves have vertical constant phase lines. There is no pressure drag and no net vertical energy transport. Such wave regimes occur when the basic flow has strong wind and/or weak stability, or when the topographic wavelength is sufficiently short.

If \( m^2 \) is positive, \((l^2 > k^2, \text{ or } N/U > k)\), we instead find solutions of the form

\[ \hat{w}_i = A_i \cos mz + B_i \sin mz. \]  

The lower boundary condition of (2.13) is used, and the *radiation condition* is used for the upper boundary condition (Eliassen and Palm 1960). The radiation condition requires that all waves at some arbitrarily great height above the mountain must be transporting energy upwards (away from the ground, which is the energy source). This condition requires that solutions with lines of constant phase of \((kx + mz)\) be used rather than those with \((kx - mz)\). The former feature phase lines that tilt *upstream* with height, and the waves transport energy upwards and momentum downwards. Further explanation of the choice of solutions can be found in the references given at the beginning of the section.

The solution is

\[ w(x, 0) = u_0 \frac{\partial h}{\partial x} = -u_0 k h_m \sin kx, \]  

\[ w'(x, z) = -u_0 h_m k e^{-\mu z} \sin kx. \]
and is shown schematically in Fig. 2.5b (from Durran 1990). These waves propagate vertically without loss of amplitude. In this setup, there is high (perturbation) pressure on the up-slope, and low pressure on the down-slope, which leads to a pressure drag on the mountain of

\[ D = \frac{1}{2} \rho_0 u_0^2 h_m^2 k \sqrt{l^2 - k^2} \]  

(2.17)

The mountain exerts an equal and opposite drag on the atmosphere.

Rather than dealing with an infinite series of ridges, an isolated ridge can be represented by a Fourier transform of the perturbation vertical velocity and the topography. The Fourier transform of a real function is

\[ \tilde{\phi}(k) = F[\phi(x)] = \frac{1}{\pi} \int_{-\infty}^{\infty} \phi(x) e^{-ikx} dx \]  

(2.18)

where \( \phi \) is an arbitrary function, and the tilde represents the Fourier transformed variable. An inverse transform is then used to express the solution in physical space:

\[ \phi(x) = F^{-1}[\tilde{\phi}(k)] = \text{Re} \int_{-\infty}^{\infty} \tilde{\phi}(k) e^{ikx} dk \]  

(2.19)

Taking the Fourier transform of (2.6), we obtain an equation identical to (2.10), where the tilde \( w \) terms represent the Fourier transformed variable. Applying the same upper and lower boundary conditions as in the infinite series case, taking the inverse Fourier transform, and expressing the solution in terms of the vertical streamline displacement \( \eta \),

\[ w'(x, z) = -u_0 \frac{\partial h(x, z)}{\partial x} \]  

(2.20)

we obtain the following solution in physical space:

\[ \eta(x, z) = \text{Re} \left[ \int_{-\infty}^{\infty} \hat{h}(k) e^{ikx} e^{i\eta} dk + \int_{-\infty}^{\infty} \hat{\eta}(k) e^{-ikx} e^{i\eta} dk \right] \]  

(2.21)
where the first integral represents an upward propagating wave, and the second integral represents an evanescent wave. The terrain is typically represented by a *Witch of Agnesi* profile, given by

\[
h(x) = \frac{h_m a^2}{x^2 + a^2},
\]

(2.22)

where \(a\) is the mountain half-width. The Fourier transform of the terrain is

\[
\hat{h}(k) = h_m ae^{-ka}.
\]

(2.23)

For the extreme case \(l^2 << k^2\) (or \(al << 1\), narrow mountains, weak stability, strong winds), the solution to (2.21), using (2.23), is

\[
\eta(x,z) = \frac{h_m a(z + a)}{x^2 + (z + a)^2},
\]

(2.24)

and shown schematically in Fig. 2.6a (from Durran 1986b). The wave pattern is centered over the crest and decays with height, similar to the evanescent waves described earlier.

For the other extreme case \(l^2 >> k^2\) (or \(al >> 1\), wide mountains, strong stability, weak winds), the solution is

\[
\eta(x,z) = \frac{h_m a(a \cos lz - x \sin lz)}{x^2 + a^2},
\]

(2.25)

and shown schematically in Fig. 2.6c. Note that (2.24) does not contain any dependence upon the horizontal wavenumber, and waves such as those in Fig. 2.6c are known as *hydrostatic mountain waves*. The wave is confined to a column directly above the mountain, with lines of constant phase tilting upstream with height. For the case \(l \approx k\) (or \(al \approx 1\), the solutions are more complicated, and the reader is referred to Smith (1979) or Lin (2007) for explanation, but the solution is shown in Fig. 2.6b. Such waves are *nonhydrostatic propagating waves*, as there is a wavenumber dependence in the solution. The nonhydrostatic modes can be seen in the dispersive tail downstream. For a narrow
mountain ridge where the hydrostatic approximation is not completely valid, the gravity wave response is best described as a superposition of vertically propagating waves and evanescent waves (Zängl 2003a). Only a portion of the wave spectrum excited by the mountain is able to propagate vertically, so the horizontal wavelength of the vertically propagating waves appear longer than the horizontal dimension of the mountain.

Variations in atmospheric properties with height, namely the mean wind speed $u_0$ and buoyancy frequency $N$, result in different wave patterns than those presented above for constant conditions with height. Such variations are prevalent and enhanced in Antarctica, due to the ubiquitous near-surface temperature inversion and katabatic winds. Considering a two-layer atmosphere, where the Scorer parameter is larger in the lower layer ($l_L > l_U$), solutions for the vertical streamline displacement are found in a similar manner to those previously discussed, with the same upper and lower boundary conditions, and appropriate conditions upon the layer interface. Details can be found in Smith (1979) and Lin (2007). Scorer (1949) showed that the necessary condition for trapped waves is

$$l_L^2 - l_U^2 > \frac{\pi}{4H^2},$$

where $H$ is the depth of the lower layer. The result is a train of waves of constant amplitude downstream of the ridge that are confined to the lower layer (Fig. 2.7, from Durran 1990). These waves are totally reflected from the Scorer parameter interface and from the lower boundary. Trapped waves are only possible within the nonhydrostatic limit.
Variability of the mean flow with time affects trapped waves, depending upon the degree to which transition in the background flow changes wave amplitude, and the differences in group velocities before and after flow transition (Nance and Durran 1997). This nonstationarity can result in a gradual downstream drift of the wave pattern, a gradual change in horizontal wavelength, or irregular changes in wavelength and amplitude. Chen et al. (2005) use a nonlinear numerical model to study how a slowly evolving synoptic-scale flow affects mountain wave momentum fluxes. The maximum momentum flux generated by waves in the slowly evolving accelerating flow is stronger than in cases with a stronger constant background flow, due to nonlinear processes. Flow variations even on the order of two days can affect mountain-wave-induced momentum flux profiles. Furthermore, nonlinearity cannot be adequately quantified using an instantaneous measurement because drag increases for accelerating cross-mountain flow. Similarly, nonlinear effects such as wave breaking cause transient effects upstream, on a slow enough timescale to be confused with permanent changes (Garner 1995).

2.5.2 Downslope Windstorms

Downslope windstorms occur in mountainous regions all over the world, including Antarctica (e.g., Steinhoff et al. 2008; Plougonven et al. 2008), and can result in property damage while also posing a serious aviation hazard. Three theories exist regarding the development of strong leeside winds. The first applies linear theory to an atmosphere of varying $u_0$ and/or $N$, but the hydrostatic approximation is made so that total wave reflection is not possible, and variations in $u_0$ and/or $N$ must occur over
shallow vertical depths. Based on observations of the 11 January 1972 Boulder Colorado windstorm, Klemp and Lilly (1975) applied linear mountain wave theory to a three-layer atmosphere. The lowest layer was about 2 km deep and had strong stability, the middle layer was about 6 km deep and had weak stability, and the upper layer (representing the stratosphere) had strong stability and strong winds. The wave amplification depends solely on $N$, but the mean wind speed $u_0$ largely determines the vertical wavelength. The maximum wave amplification leading to downslope winds occurs when the thickness of the lower two layers is $\frac{1}{4}$ of the vertical wavelength of each respective layer. When the system is properly tuned in this way, wave reflection at the layer boundaries, and a second reflection at the ground, result in maximum wave amplification through enhanced energy transport and maximum winds located about halfway down the lee slope.

When terrain obstacles are large enough so that linear theory is invalid, nonlinear effects become important. Hydraulic theory has been applied to large-amplitude mountain waves in order to account for these nonlinear effects. This model was first proposed by Long (1953), suggesting a connection between downslope windstorms and hydraulic jumps. This connection is developed further by Durran (1986a), who used a numerical model to study the role of hydraulics and vertically propagating internal waves on the development of large-amplitude waves. Durran finds that a two-layer structure, with the lower layer having larger $N$ and depth $\frac{1}{2}$ the vertical wavelength (which in linear theory is detuned, for a weak amplification), actually results in an almost six-fold increase in nondimensional pressure drag compared to the linear response. If the lower layer depth is fixed, and the mountain height increased, an amplified response is found,
where the largest pressure drag (strongest downslope winds) moves farther downslope. Nonlinear effects are thus responsible for strong winds beyond the lee slope. As the depth of the lower layer is increased, the resemblance of the response progresses from supercritical flow, to a propagating hydraulic jump, to a stationary hydraulic jump, then to subcritical flow (Fig. 2.8, from Durran 1986a). Similar results are found in the regime diagrams of Vosper (2004, his Figs. 9 and 10). Durran relates the nonlinear response to the horizontal pressure gradient associated with displacement of the layer interface. As the nondimensional mountain height increases, the contribution from the displacement of the interface towards the total pressure gradient increases. Hence, for strongly nonlinear flow, a large portion of the lee-side flow is governed by the hydraulic analog. In this amplified state, the lower stability in the upper layer prevents the pressure perturbation associated with the interface displacement from decelerating the flow, leading to supercritical flow in the lee. This amplification process was important for the 11 January 1972 Boulder windstorm. Interestingly, when the tropopause (an important reflecting mechanism) is removed from these simulations, a strong downslope windstorm still forms.

Similar to trapped lee waves, variations in atmospheric properties with height can lead to wave breaking. But unlike trapped waves, which generally do not lead to downslope windstorms, wave breaking has been attributed to the formation of downslope windstorms. In a series of papers, Clark and Peltier (1977, 1984), Peltier and Clark (1979, 1983), and Clark and Farley (1984) suggest that a “wave-induced critical layer” is formed by waves overturning and breaking aloft, leading to highly turbulent conditions.
A critical level is where the mean flow speed equals the wave phase speed. The critical level causes wave reflection and resonance if the critical layer is properly tuned to the mountain. While wave breaking is a nonlinear process, the authors solve for the linear wave equations with a fixed horizontal critical level that reflects all waves. Smith (1985) also assumes that the wave-breaking region traps energy, but instead solves Long’s equation for nonlinear steady flow with an upper boundary condition based on a “dividing streamline” between the flow below and within the wave breaking region. The different mathematical formulations lead to different conditions for wave amplification: Clark and Peltier’s solution results in amplification only when the critical level is positioned at \( \frac{1}{4} + \frac{n}{2}, n=0,1,\ldots \) vertical wavelengths above the ground, whereas Smith’s solution predicts amplification in the entire range between \( (1/4 + n) \) and \( (3/4 + n) \) vertical wavelengths above the ground.

Both Durran and Klemp (1987) and Bacmeister and Pierrehumbert (1988) test these theories by varying the height of the critical level and of the mountain. These results show that with a critical layer located at a vertical wavelength that does not support resonance from linear theory, wave amplification and breaking is possible as the mountain height is increased. Wave amplification is indeed possible over a range of critical level heights. This terrain dependence supports the results of Smith, based on a relationship between the height of the dividing streamline and a transition from subcritical to supercritical flow in hydraulic theory. These results are supported by Colle and Mass (1998b), who analyze past events and perform idealized simulations for strong wind events in the Cascade Mountains (Washington State, USA). While many events
featured a mid-tropospheric critical level, the presence of a critical level was not important for windstorm development with slow (< 10 m s⁻¹) cross-barrier flow, when low-level wave breaking forces a transition to supercritical flow in the lee. Strong winds can develop with no critical level when the shallow near-surface stable layer extends above crest level. Bond et al. (2006) observe and simulate a bora-type “Taku wind” event near Juneau, Alaska that features a critical level near 500 hPa, as a low pressure system is centered over the region. Fudeyasu et al. (2008) also find the presence of a mean-state critical level aloft during “Hirodo-Kaze” downslope winds off of Mt. Nagi in Japan. Wang and Lin (1999) attempt to clarify the discrepancies between the earlier theoretical studies on downslope windstorm mechanisms. They find that wave breaking creates a wave duct between the ground and the wave-breaking region. This turbulent mixing region expands due to nonlinear effects, and waves accelerate through this duct without loss of energy, since the turbulent mixing region above is a perfect reflector. The high-drag state is then maintained through the hydraulic mechanism.

Diabatic heating and frontal processes can affect the vertical profile of stability and the resulting mountain wave structure. Durran and Klemp (1982) find that moisture can modify the height of wave trapping and affect wave resonance. Doyle and Shapiro (2000) model a downslope windstorm in Norway using COAMPS, and find that strong downslope winds occur when an approaching warm front is modulated by the complex terrain. Strengthening of mesoscale thermal gradients, associated with frontogenesis processes like differential vertical motion and tilting, and vertical wind shear associated with development of a low-level jet in the complex terrain, increase wave trapping and
lead to a resonant amplification of mountain waves. Latent heat release in the warm
frontal ascent destabilizes the mid- to upper-troposphere and further increases Scorer
parameter layering, and sensitivity studies without diabatic heating show that wave
amplitudes are markedly reduced. Similar results are found by Doyle and Smith (2003),
where upstream differential diabatic heating leads to intensified Scorer parameter
layering.

2.5.3 Wave Breaking

As described in the previous section, wave breaking can be important in the
generation of downslope windstorms. Therefore, some additional explanation of wave
breaking is presented here. Gravity wave breaking occurs when wave amplitudes
increase, so that the wave eventually becomes unstable and overturns. The result is a
turbulent mixed layer above the wave breaking region. In the troposphere, wave
breaking often occurs when upward propagating mountain waves encounter a critical
level. For a stationary mountain wave, a critical level occurs when the mean flow speed
is zero, since the phase speed of the wave is also zero. In more complex situations, with
several wave modes or three-dimensional wave dispersion, individual critical levels exist
that comprise a critical layer. The wave response to the critical level depends upon the
Richardson Number,

$$Ri = \frac{N^2}{(\partial U/\partial z)^2},$$

(2.27)

which is the ratio of buoyant suppression of turbulence to the generation of turbulence by
vertical wind shear. When the Richardson Number is large (> 2), waves are effectively
absorbed by the critical layer. As the Richardson Number approaches 0.25, a significant portion of the wave energy is reflected. Lower than 0.25, the established limit for shear instability, overreflection occurs as the wave extracts energy and momentum from the mean flow. If the reflected waves are in phase with the upward incident waves, amplification will occur through resonance.

Nonlinear effects such as terrain height and shape also influence wave breaking. Miles and Huppert (1969) solve Long’s equation under hydrostatic conditions for a bell-shaped terrain profile and constant upstream conditions, and find that the critical value of the nondimensional mountain height for wave breaking (inverse Froude Number) is 0.85. For isolated three-dimensional orography, this critical height increases to 1.1 (Smith and Grønås 1993) due to horizontal wave dispersion. Miller and Durran (1991) find that wave breaking is favored for steep terrain (large aspect ratio). It can now be seen how wave breaking relates to the hydraulic theory and wave-breaking mechanisms for downslope windstorms presented in the last section. Wave breaking leads to generation of a “self-induced critical layer” that can reflect wave energy downward. Tuned responses with upward propagating waves can lead to resonance and wave amplification. In the same way that a transition to supercritical flow on the lee slope can lead to downslope windstorms, wave breaking can occur from similar nonlinear effects, leading to flow separation below the dividing streamline that resembles shallow water flow.

Most studies on vertically propagating waves, including those specifically dealing with wave breaking, typically assume a wide mountains (e.g., \( Na/U >> 1 \), where \( a \) is the mountain half-width), where hydrostatic conditions are assumed, and vertically
propagating waves are assured. Zängl (2003a) performs numerical experiments for mountain widths near the nonhydrostatic limit of vertical propagation ($Na/U \sim 1$) to find out why strong wind maxima can form in the lee of narrow mountains. He finds that there is a range of mountain heights in the $Nh_m/U = 1$ to 1.5 range where wave breaking extends the hydrostatic flow regime that features a pronounced leeside wind maximum. Higher nondimensional mountain heights cause surface flow separation and shift the wind pattern back to nonhydrostatic (wind maximum at crest level only). A narrower mountain ($Na/U < 1$) eliminates wave breaking and nonlinear amplification, again shifting the flow to nonhydrostatic.

Recent research on gravity wave breaking has focused on observations as well as fine-scale aspects of the wave breaking process. Doyle et al. (2000) compare simulated wave breaking from a suite of mesoscale numerical models for the 11 January 1972 Boulder windstorm, to determine the feasibility of using high-resolution numerical models to predict the location and strength of breaking mountain waves. While the vertical and horizontal locations of wave breaking are similar between models, the fine-scale structure of wave breaking is sensitive to variations in initial conditions in the stratosphere. The fine-scale structure is also affected by the means of horizontal diffusion in the models, with a decrease in wave breaking for greater smoothing.

A theoretical study by Jiang and Smith (2003) using a two-layer hydrostatic model over three-dimensional bell-shaped terrain focuses on the role of vertical shear towards wave breaking. Vertical wind shear controls the vertical distribution of wave breaking. When the low-level flow is fast, waves can propagate aloft, where layers of
slow winds will promote breaking. Slow moving layers near the ground lead to wave breaking there, so that wave breaking does not occur aloft. An observational study by Jiang and Doyle (2004) from the central Alps focuses on the role of complex terrain towards wave breaking. Wave breaking in this event is associated with a critical level, a layer of vertical wind shear (in both speed and direction), and a local Richardson Number less than unity near the critical level. The multiscale, complex terrain is shown to promote faster wave breaking, with stronger turbulence and downslope winds, from nonlinear wave-wave interactions. Such effects are certainly possible in the complex terrain of the MDVs.

Wave breaking over Greenland is both observed and simulated by Doyle et al. (2005). The horizontal flux minimum occurs in a layer where the vertical shear of the cross-mountain wind component is reversed, which is conducive to wave breaking (Smith 1989). Similar to previous results, wave breaking is stronger for steeper lee slopes, as the effective Rossby Number increases. In a finding applicable to Antarctica, surface-based diabatic cooling results in a 15-20% increase in wave amplitude. However, for narrow ridges (around 10 km width), the effects of diabatic cooling are negligible. Wave breaking can even occur for large-scale easterly flow over Greenland, as found by Ólafsson and Ágústsson (2009).

2.5.4 Boundary Layer Effects

Early studies on mountain waves were simulated with no boundary layer and free-slip conditions. Recent studies have exposed the impact of boundary layer effects on
different wave regimes and wave breaking. Ólafsson and Bougeault (1997) and Vosper (2004) find that friction suppresses wave breaking for all nondimensional mountain heights. The value of the pressure drag is reduced for low nondimensional mountain heights (unblocked flow), but increased for high nondimensional mountain heights (blocked flow), in essence bringing pressure drag closer to the predicted linear value (simulations are done for uniform upstream velocity and stability).

The effects of boundary layer stability on mountain waves were first identified by Smith et al. (2002), who found that a stagnant boundary layer absorbed downward reflected waves, preventing the formation of lee waves. Jiang et al. (2006) and Smith et al. (2006) expand upon earlier findings, noting that a stable boundary layer is more efficient in absorbing trapped waves than a turbulent convective boundary layer. Friction decreases the wavelength of trapped waves, and wave amplitude decreases downstream. A negative heat flux (surface cooling) inhibits turbulence and decreases the boundary layer depth, but lee waves decay faster than with a positive heat flux (surface heating) with a deeper and more turbulent boundary layer. A stagnant boundary layer thus acts as a “sponge”, partially absorbing downgoing wave energy. Stagnant boundary layers also occur in the MDVs when cold pools form during periods of low incident solar radiation.

For hydrostatic three-dimensional waves, the reduced boundary layer wind speed increases downslope winds while shifting the pattern upstream, and wave breaking aloft is reduced (Smith 2007). Smith and Skyllingstad (2009) use an LES simulation at 10 m horizontal grid spacing, and find that negative surface fluxes instead increase downslope winds and rotor formation, from interaction with a jet in the stratified region capping the
boundary layer. The discrepancies between the Jiang et al. and Smith and Skyllingstad studies pertain to the different lee wave characteristics. In the former study, the waves are forced to greater heights, resulting in larger vertical energy propagation, so that the waves are more affected by internal wave dynamics than the latter study, where turbulent processes are more important.

Jiang et al. (2008) study boundary layer effects on vertically propagating mountain waves using COAMPS and an analytical boundary layer model. In general, the flow induced by a representative force balance in the boundary layer shifts wave patterns upstream, weakens waves aloft, and reduces drag and momentum fluxes through both inertial and frictional adjustment processes. Frictional adjustment is only important for wide terrain, as flow is essentially inviscid for narrow terrain. Boundary layer effects decrease with increasing nonlinearity.

Diurnal effects on the boundary layer, primarily heating of the boundary layer during daytime, impact the structure of mountain waves. Jiang and Doyle (2008) find that a convective boundary layer during daytime, characterized by a shallow wind shear layer near the surface and a deep well-mixed layer aloft, can significantly weaken mountain waves and reduced momentum fluxes by up to 90%. In contrast, during nighttime a shallow stable boundary layer develops, and flow in the stable nocturnal boundary layer is governed by hydraulics.

2.6 Foehn
The concept of the “foehn” wind originated with the famous warm downslope wind in the Alps, and is known by other monikers in different parts of the world. The most recognized and best observed foehn regimes are in the Alps and the “Chinook” in the Front Range of Colorado. The WMO (1992) defines foehn as a “wind warmed and dried by descent, in general on the lee side of a mountain.” In a general sense (in many meteorological textbooks), foehn is used to describe thermodynamical warming of an airmass as it ascends a mountain, becomes saturated and cools moist-adiabatically, condenses, and warms dry-adiabatically in the lee (Fig. 2.9a, from Seibert 1990). This is a limited definition of the foehn, distorted from the original description, that is too restrictive for the situation in the MDVs and elsewhere. A detailed argument for the broader definition of foehn is presented by Seibert (1990). She states that the original foehn definition of Hann (1866) has been modified over the years, resulting in the current limited definition. Seibert notes that there are foehn events without cloud cover upwind, and precipitation, which is required for a net gain of energy in the lee, does not always occur (at least 50% of foehn events at Innsbruck are not associated with precipitation on the windward side). Even with precipitation present, the quantitative effects of precipitation on the observed warming are small. Equivalent potential temperature differences are large (up to 14 K) between windward and leeward sites, whereas differences are clustered around zero for a mountain top site and a leeward site. This implies that foehn air is forced downward from about 2 km altitude. A schematic of this forced descent mechanism leading to foehn winds is shown in Fig. 2.9b. This forced descent can be caused by air descending over lee slopes from mountain waves (e.g.,
Klemp and Lilly 1975; Jiang et al. 2005) or by blocking of low-level winds windward of mountain barriers, forcing descent over the lee slope from ridge level (e.g., Parish 1983; Gohm et al. 2004). Note that foehn differ from “bora”, which are cold downslope winds, often found southwest of the Dinaric Alps along the Adriatic Sea (e.g., Smith 1987) and in the Washington Cascades (e.g., Colle and Mass 1998a). However, the mountain wave dynamics involved in both foehn and bora are similar (Klemp and Durran 1987).

The best analog of foehn winds in the MDVs is the Alps, which will be the focus of this section. The Alps and MDVs both feature narrow valleys, complex topography, and conceptually similar winter atmospheric structures, among other similarities. Alpine foehn encompasses multiple scales – ranging from local processes (radiative cooling, local pressure gradients, cold pools), to mesoscale (valley tributaries, flow splitting), and synoptic (large-scale pressure gradients, mountain waves) (Drobinski et al. 2003; Jaubert and Stein 2003). These processes are also prevalent in the MDVs as well. Foehn research in the Alps has a long history, and recent reviews can be found in Zängl (2003b) and Drobinski et al. (2007). Recent studies, both observational and model-based, provide a wealth of detail about orographic wind regimes in the Alps. The following subsections each contain discussion of various aspects of Alpine foehn.

2.6.1 Gap Flow

Because Alpine foehn typically occur in north-south-oriented valleys originating from a mountain pass, the wind regimes are best described as a gap flow (Zängl et al. 2004a,b). Gap flows are associated with a pressure gradient in the along-flow direction
(Overland and Walter 1981), caused by a synoptic-scale disturbance, a cross-ridge temperature difference, or upstream blocking and subsidence in the lee. Shallow water flow is a good approximation to gap flow in the Alps because in all gap flow regimes, a thickening nearly neutral and stagnant layer above the gap flow decouples the gap flow from the overlying flow (Gohm and Mayr 2004; Armi and Mayr 2007; Mayr et al. 2007).

Gap flows and foehn in the Alps are characterized by the depth of cold air in the upstream reservoir. “Deep foehn” occurs when the upstream reservoir of cold air exceeds the mountain height, and southerly flow is present throughout much of the depth of the troposphere. Deep foehn are associated with large amplitude gravity waves and stronger winds compared to “shallow foehn” (or “shallow gap flow”). Shallow foehn are forced by the cross-Alpine pressure gradient caused by the pooling of cold air windward of the Alps (Zängl 2003b).

Gap winds are primarily forced by the low-level pressure difference due to upstream blocking (Zängl 2002a). Thus, a vertical constriction has a greater effect than a horizontal constriction (Pan and Smith 1999). Mesoscale circulations, such as pressure drag from mountain waves and upstream flow blocking, are at least as important as the ambient synoptic-scale pressure gradient (Gaberšek and Durran 2006). Colle and Mass (1998b) show that along-gap pressure gradients can be enhanced by 25-50% from leeside wave amplification. The downstream extent of gap winds past the gap exit suggests additional forcing besides the cross-gap pressure difference. Pan and Smith (1999) find that the flow through the gap does not transition to critical flow like the flow over adjacent ridges does, and thus the gap flow avoids Bernoulli loss in hydraulic jumps (see
also Gaberšek and Durran 2004). Even in an elevated gap, the lower nondimensional mountain height extends the shooting flow farther downstream than over the adjacent ridges.

2.6.2 Pre-foehn

Foehn initiation at Innsbruck is preceded by a steady westerly wind in 70% of all foehn cases investigated by Seibert (1985). This “pre-foehn” wind regime was originally thought to be associated with cold air outflow from the valleys, supporting foehn initiation. However, Zängl (2003b) finds that the pre-foehn is a localized feature in the vicinity of Innsbruck. This westerly flow is forced by gradients in perturbation pressure caused by an east-west asymmetry in gravity wave activity. Consistent with this explanation, the direction of the large-scale wind has a larger impact on the strength of the pre-foehn wind than the intensity of cold pooling in the valleys. Interestingly, Speirs et al. (2010) find a similar wind speed and temperature increase leading up to foehn events in the MDVs. No physical explanation has been given yet for the existence of this pre-foehn wind regime.

2.6.3 Foehn Breakthrough

The initiation of foehn winds at observing sites in valleys is often associated with the destruction of cold pools, hence the term “breakthrough” being used for foehn initiation. The strong static stability associated with the cold pools prevents warming from both the downward transport of adiabatically warmed air from above and from
turbulent mixing – hence the potential temperature difference between the layers of cold and warm air determines the timescale of cold pool erosion (Zhong et al. 2003). Cold pools are formed by radiative heat flux divergence at the surface, rather than downslope flows along the side slopes of the basin, which is found to have only minor importance (Clements et al. 2003). A steady-state is attained when the slanted upper boundary compensates for the ambient horizontal pressure gradient (Zängl 2003c) and radiative heat flux divergence is balanced by upward ground heat flux. In order for the radiative heat flux divergence to be maintained, a small heat conductivity of the ground surface, and a mechanism for extracting moisture from the surface, are both required (Zängl 2005a). These requirements are met through a fresh snow pack (lowering heat conductivity) and water vapor deposition, respectively.

Downvalley outflow is the most effective mechanism for cold pool destruction. If the ambient wind favors the flow of cold air out of the valley, the cold-air pool will not persist (Zängl 2005b). However, if the ambient wind instead favors flow into a valley constriction, then the cold pool is more likely to be maintained. Similarly, valley geometry and mesoscale pressure perturbations from mountain waves can also force cold air out of valleys (Zängl 2003c; Flamant et al. 2006). For deep valleys considered in the above studies, turbulent vertical mixing plays a minor role in cold-pool erosion.

Another factor influencing foehn breakthrough is upstream flow blocking. Zängl and Gohm (2006) simulate flow in the Inn Valley (Austria) using a modified version of MM5 with 467 m grid spacing in the region of interest. Foehn in the Inn Valley is associated with flow through the Wipp Valley, which “T-bones” and exits into the Inn
Valley near Innsbruck. The orographic feature just north of the Inn Valley associated with flow blocking is called the Nordkette. Presence of the Nordkette is found to accelerate foehn breakthrough compared to a sensitivity simulation with lowered topography, as a barrier flow response is found. However, completely removing the Nordkette accelerates foehn breakthrough even more, as the “real” Nordkette results in decelerated flow and reduced gravity wave amplitude.

2.6.4 Synoptic-scale Setup and Vertical Structure

The overlying synoptic-scale weather situation for south foehn in the Alps is largely consistent between cases, and is summarized by Seibert (1990). Leading up to the foehn event, the Alpine region is typically under large-scale south-southwesterly flow associated with an upper-level trough over the eastern Atlantic Ocean and a surface cyclone over the British Isles. Cool air with stable to moist adiabatic stratification in low-levels exists on both sides of the Alps. Advection of warm air with the low-level south to southwesterly flow leads to a temperature gradient across the Alps that is the forcing for shallow foehn. As the low-level cyclone moves north of the Alps, and the upper-level trough approaches from the west, the flow over the Alps becomes south to southwesterly throughout the troposphere and the foehn intensity increases. Foehn rapidly ceases with passage of a cold front associated with the upper-level trough, as flow turns westerly. This sequence of events is rarely homogeneous across the Alps, however. For example, Zängl and Hornsteiner (2007a) describe an east-west spatial gradient in
foehn across the Alps resulting from different origin airstreams and the formation of a mesoscale cyclone north of the Alps.

Zängl (2003b) models south foehn near Innsbruck (Wipp/Inn Valleys) using a modified version of MM5 using idealized, but realistic, atmospheric conditions. Vertically propagating gravity waves are forced by flow over the surrounding mountains, with the vertical structure strongly dependent upon the direction of the large-scale wind (Fig. 2.10). Large wave amplitudes are generated only when the wind direction is approximately normal to the ridgeline. Oblique flow angles to the terrain ridge leads to a smaller incident wind speed, which increases flow blocking and decreases the effective mountain height. The sensitivity simulations of Zängl (2003b) also show that the layer of strong stability near ridge height is greatly reduced for the off-angle cases. Diagnosing mountain wave effects in such complex topography is difficult, however, due to three-dimensional wave dispersion (Smith 1980; Zängl 2002a), as the group velocity of the waves depends on the angle between the phase surfaces (determined by the terrain) and the wind direction.

Zängl et al. (2004a) perform a similar study, but instead for the Rhine Valley and a real case. Similar to the Wipp Valley, there is three-dimensional wave propagation into the valley from surrounding ridges. Also found is down-valley warming, which is related to turbulent mixing of stably stratified air. This mechanism is most effective where the valley widens, and mass continuity requires sinking motion, which strengthens the stable stratification necessary for downward heat flux. Other features include wave trapping aloft, associated with positive vertical wind shear, and a hydraulic jump occurring at the
leading edge of the foehn where it meets colder air. This hydraulic jump structure is
different than those that occur in a single air mass (e.g., Durran 1986a), and is similar to
the hydraulic jump found near McMurdo in the model results of Steinhoff et al. (2008).
Zängl and Hornsteiner (2007b) find an interesting south foehn event where trapped
mountain waves appear to be responsible for a strong foehn windstorm. A deep neutral
layer in the upper troposphere is responsible for the trapping, which occurs throughout
the region during the time period studied. Strong winds are found beneath wave troughs
for trapped waves, which transport potentially warm air towards the surface.

A connection has been made between the hydraulic theory for mountain waves
and downslope windstorms (section 2.5.2, Durran 1986a) and foehn in the Alps by Gohm
and Mayr (2004). Foehn vertical structure can be approximated by the shallow water
model when foehn in the lower troposphere occurs beneath a stable layer (this stable
layer is the “free surface” in shallow water theory). A statistical analysis of 33 foehn
events shows that flow typically transitions from subcritical to supercritical flow at a
vertical topographic contraction. The model correctly captures spatial inhomogeneities in
surface wind speed across the Wipp Valley, which is related to the complex valley
geometry.

The Alpine north foehn is less frequent and less studied than the south foehn
counterpart, but some case studies have been presented recently in the literature. North
foehn features a different synoptic-scale setup compared to the southerly foehn. The case
from Jiang et al. (2005) features a large anticyclone west of France, and cut-off upper-
level lows northeast (Austria) and southeast (southern Italy) of the Alps, resulting in
northerly flow across central Europe. Flow blocking is apparent in low-levels north of the Alps. A slightly different setup is found in the event from Zängl (2006), where flow is northwesterly throughout the troposphere over the Alps, between an anticyclone over northwest Africa and a cyclone over eastern Russia.

Zängl (2006) models a northerly foehn event over the Inn Valley and finds mainly a trapped wave response, primarily caused by a decrease in static stability with height, as vertical wind shear is weak. The strongest winds and warming are located in large-amplitude troughs. Sensitivity tests with atmospheric moisture show that dry conditions feature stronger wind speeds and warming than moist conditions. This results from the evaporation of precipitation, which stabilizes the atmosphere in lower levels and prevents gravity waves from penetrating to the valley surface. A similar sensitivity to moisture is found by Zängl and Hornsteiner (2007a). Jiang et al. (2005) use a suite of observations and COAMPS simulations to study a different northerly foehn event over a larger portion of the Alpine massif. A schematic of the dynamical processes is shown in Fig. 2.11 (their Fig. 16). Similar to the Zängl (2006) case, trapped waves are forced above the small-scale topography in the northern portion of the Alps, beneath a weak stability layer. The broader main Alpine ridge forces stationary hydrostatic waves that amplify and break above the lee slope.
2.7 Figures

Figure 2.1. Map of the Ross Sea region of Antarctica. The McMurdo Dry Valleys AWS network: Lake Vida (VV), Lake Vanda (WV), Lake Brownworth (WB), Beacon Valley (BV), Taylor Glacier (TTa), Lake Bonney (TB), Lake Hoare (TH), Howard Glacier (THo), Canada Glacier (TCa), Lake Fryxell (TF), Commonwealth Glacier (TCo) and Explorer’s Cove (TE). Landsat ETM+ image captured 21 Nov 2001. From Speirs et al. (2010), Fig. 1. In top left inset, “EAIS” refers to East Antarctic Ice Sheet, and “WAIS” refers to West Antarctic Ice Sheet.
Figure 2.2. Conceptual model of summertime flow in Wright Valley. From McKendry and Lewthwaite (1990), Fig. 13.
Figure 2.3. Schematic of upper-tropospheric height anomalies over the Pacific Ocean during the early stage of an ENSO event in the austral winter. Stippling shows region of enhanced convection over central equatorial Pacific and arrows indicate westerly wind anomalies in jet streams. From Karoly (1989), Fig. 11a.
Figure 2.4. Schematic of atmospheric circulation response to ENSO warm (top) and cold (bottom) events superimposed on the corresponding SST composites. From Yuan (2004), Fig. 8.
Figure 2.5. Streamlines over an infinite series of sinusoidal ridges for (a) evanescent waves \((m^2 > 0, \hat{f}^2 > \kappa^2, \text{or } N/U > k)\) and (b) vertically propagating waves \((m^2 < 0, \hat{f}^2 < \kappa^2, \text{or } N/U < k)\). Airflow is from left to right across diagram. Dashed line in (b) represents line of constant phase. From Durran (1990), Fig. 4.2.
Figure 2.6. Streamlines over a bell-shaped ridge. (a) Narrow ridge and evanescent wave ($l^2 << k^2$ or $al << 1$), (b) intermediate ridge and nonhydrostatic vertically propagating wave ($l^2 \approx k^2$ or $al \approx 1$), (c) wide ridge and hydrostatic vertically propagating wave ($l^2 >> k^2$ or $al >> 1$). Airflow is from left to right across diagrams. From Durran (1986b), Fig. 20.2.
Figure 2.7. Streamlines over an isolated bell-shaped ridge for trapped waves. Airflow is from left to right across diagram. From Durran (1990), Fig. 4.4.
Figure 2.8. Isentropes for airflow from left to right in a two-layer atmosphere when the mountain height is fixed at 500 m and the interface between the two layers is at (a) 1000 m, (b) 2500 m, (c) 3500 m, and (d) 4000 m. From Durran (1986a), Fig. 5.
Figure 2.9. (a) Thermodynamical foehn theory, where air rises on windward slope, becomes saturated and cools moist adiabatically, descends and warms at dry adiabatic lapse rate. (b) Current foehn theory, where air near the surface in the lee of the ridge originates from about 2 km AGL upstream. From Seibert (1990), Fig. 9.
Figure 2.10. Semi-idealized vertical cross sections through Wipp Valley near Innsbruck of potential temperature (solid lines), wind speed along cross section (shaded), and wind vectors along cross section when ambient wind direction is (a) SSE, (b) S, (c) SW, and (d) WSW. From Zängl (2003b), Fig. 10.
Figure 2.11. Summary of dynamics pertaining to a northerly foehn event. From Jiang et al. (2005), Fig. 16.
Chapter 3: Data and Methods

This chapter begins with a description of Polar WRF, the mesoscale numerical weather prediction model used in this study, including the input datasets, domains, model physics parameterizations, and modifications specifically done for the McMurdo Dry Valleys (MDVs). A brief description of the Antarctic Mesoscale Prediction System (AMPS) is then presented, which is used in Chapter 4. The primary observational dataset used throughout this work is automatic weather station (AWS) observations from the McMurdo Dry Valleys Long-Term Ecological Research (MDV LTER) program, and information pertaining to station locations, observations used, and processing methods is provided. The chapter concludes with a description of the MODIS visible satellite imagery data used for some case study applications.

3.1 Polar WRF

The Advanced Research Weather Research and Forecasting Model (WRF-ARW) is a fully compressible (volume of an air parcel changes with time), nonhydrostatic model that solves for the Euler equations (horizontal and vertical momentum, mass continuity, thermodynamic energy, equation of state, conservation of water vapor, and an equation for geopotential) in perturbation and flux form. Details of the model formulation can be
found in the WRF-ARW technical note (Skamarock et al. 2008). Spatial discretization of the governing equations is done on an Arakawa-C grid, where the $u$ and $v$ velocity variables are staggered one-half grid length from the scalar (mass) variables. The C-grid staggering has the advantage that finite differences can be calculated over a distance of one grid length, in effect doubling the resolution of an unstaggered grid (Kalnay 2003). Fifth-order horizontal advection is used for both momentum and scalar variables, which is implicitly diffusive. The vertical discretization is done on a terrain-following hydrostatic-pressure vertical coordinate, defined as

$$
\eta = \frac{(p_h - p_{ht})}{\mu},
$$

where $\mu = p_{hs} - p_{ht}$. $p_h$ is the hydrostatic pressure, $p_{hs}$ is the surface hydrostatic pressure, and $p_{ht}$ is the model-top hydrostatic pressure. The model levels range from 1.0 (surface) to 0.0 (model top, constant pressure). Third-order vertical advection is used. Temporal discretization is done using a time-split integration scheme, where the high-frequency acoustic modes are integrated over a shorter time step than the low-frequency meteorologically relevant modes (here, the acoustic timestep is four times smaller than the low-frequency timestep). The low-frequency modes are integrated with a third-order Runge-Kutta (RK3) scheme, while the acoustic modes are integrated within the RK3 loop.

Modifications have been made to the WRF code for applications in polar regions by the Polar Meteorology Group (Polar WRF). Details of these modifications are presented in Hines and Bromwich (2008), Bromwich et al. (2009), and Hines et al. (2011). Most of the modifications deal with treatment of the land surface processes.
When a surface is largely snow covered (greater than 97%), thermal diffusivity of snow is used in ground heat flux calculations rather than that of soil. Values of snow density, thermal conductivity, and heat capacity for specific layers in the land surface model are provided in Hines and Bromwich (2008). Snow emissivity is set to 0.98, and surface roughness is set at 0.1 mm over glacial ice. Over sea-ice and glacial land points, the maximum depth of the top snow cover layer is set to the depth of the upper subsurface layer, so that a deep snowpack is represented by multiple subsurface layers, rather than as one layer on top of underlying soil layers. Latent heat flux calculations using the Penman-Monteith approach are modified to include sublimation from frozen surfaces, and the proper latent heat constant is determined using the skin temperature. The surface temperature over snow is calculated by iteratively solving the full surface energy balance equation, taking upward longwave radiative flux as a function of surface temperature rather than the lowest-level atmospheric temperature. This adjustment can be important under strong stable stratification over ice surfaces.

For sea-ice points, the albedo is set to 0.81 and the emissivity is set to 0.98. The base subsurface temperature for sea ice is adjusted to 271.36 K. A fractional sea ice scheme is used, where sea ice concentration values are input to the model, and these values alone are used to determine sea ice locations (surface temperature is not used). The surface layer parameterization scheme is called separately for each fraction of ice and open water. The ice-fraction surface exchange coefficients and surface fluxes from the surface scheme are used for the land-surface model call. Over open water, surface fluxes and other surface variables are calculated directly from the surface scheme. The
ice and open-water variables are then aerially averaged to get final values for the fractional sea ice grid box. Variable thickness of sea ice and snow on top of sea ice is allowed, and the climatological values used are described later in this section.

3.1.1 Input Datasets

Several input datasets are used in the Polar WRF simulations. Input conditions for the atmosphere, some surface fields, and lateral boundary conditions are supplied by ERA-Interim, which is ECMWF’s most recent global reanalysis product (Simmons et al. 2007). ERA-Interim features several improvements upon ERA-40, including 12 hour 4D-Var data assimilation, T255 (~0.70°) horizontal resolution, improved humidity analysis, variational bias correction of satellite radiance data, more extensive use of satellite radiances, and additional observations. Important data sources for high-latitude regions that are ingested into the ECMWF analyses used in ERA-Interim from 2001 onwards include MODIS winds, AIRS radiances, and COSMIC GPS radio occultation soundings (Andersson 2007). ERA-Interim output is reprocessed and distributed in the United States by the NCAR Data Support Section (DSS) on a regular 512 x 256 N128 Gaussian grid. Output is available every six hours from 1989 through 2010.

Sea surface temperature (SST) data is taken from daily optimally interpolated 0.25° x 0.25° grid spacing analyses based on AVHRR and AMSR-E (2002 onwards) SST data by NOAA/NESDIS/NCDC (Reynolds et al. 2007). *In situ* data from buoys and ships are also used, and bias correction of the satellite data utilizes this *in situ* data. The AMSR-E SST retrievals are most useful as a near-all-weather measurement, as IR
methods fail in regions of extensive cloud cover. Sea ice data is used as a proxy for SST when sea ice concentrations are greater than 50%. This combined AVHRR-AMSR-E product from 2002 onwards shows the greatest improvement over the AVHRR-only product in the mid- to high-latitude Southern Ocean, where *in situ* data is generally lacking.

Two different sea ice concentration products are used, depending on the length of the simulation. The first is the Bootstrap sea ice concentration algorithm, which is used with measurements from the Special Sensor Microwave/Imager (SSM/I) instrument onboard the DMSP -F8, -F11, and -F13 satellites (Comiso 1999). This dataset spans from 1978 to 2007, so it is used with the summer climatological 2-km simulations. Data are obtained from the National Snow and Ice Data Center (NSIDC), and version 2 of the product is used, on a polar stereographic 25 x 25 km grid. Retrieval accuracy is assumed to be within 5-10% of actual conditions. The algorithm uses different channels (19V and 37V GHz) for the extensive seasonal sea ice of the Antarctic than for the multi-year Arctic ice field. The other algorithm is the ASI algorithm, which uses the AMSR-E instrument onboard the Aqua spacecraft (Spreen et al. 2008). The horizontal resolution of the ASI algorithm using the 89 GHz channels is up to four times greater than that of the 19 and 37 GHz channels used in the Bootstrap algorithm. In the ASI algorithm, lower frequency channels are used as weather filters. Ice concentration is calculated by the polarization difference of brightness temperatures between horizontal and vertical polarizations. Output is obtained directly from the University of Bremen (www.iup.uni-bremen.de:8084/amsr/amsre.html) on a polar stereographic grid at 6.25 km grid spacing.
This dataset only goes back to 2003, so it is used for the short-duration 0.5-km simulations. It is also used for 2008 and 2009 in the summer climatological simulations, as the Bootstrap sea ice dataset ends in 2007.

Andersen et al. (2007) perform a comparison amongst several passive microwave sea ice concentration retrieval algorithms over high-ice-concentration regions of the Arctic. The algorithms using 85+ GHz channels (like the ASI algorithm) consistently give a better agreement with synthetic aperture radar (SAR) ice concentrations and ship observations than those that do not use such channels. While the 85+ GHz channels are more sensitive to atmospheric effects than lower frequency channels, this is found to be secondary compared to the variability of surface emissivity effects.

For sea ice thickness and snow thickness on top of sea ice, climatological values are taken from the Antarctic Sea Ice Processes and Climate program (ASPeCt, Worby et al. 2008). ASPeCt comprises over 23,000 ship and aircraft sea ice observations from 81 voyages in all seasons from 1981-2005. Ship observations are representative of an area about 1 km around the ship. To ensure independent observations, the dataset is edited to remove observations taken within 6 nautical miles of the previous observation, reducing the number of observations to 14,557. Snow thickness data is only taken on level ice, so it represents a lower bound. For both ice and snow, errors increase with decreasing thickness, from ±20% for thick (> 30 cm) ice to ±50% for thin (< 10 cm) ice. Since some Antarctic coastal regions feature more observations during certain seasons, an annual climatology was constructed by averaging all values in 2.5° latitude by 5.0° longitude bins. Averaged fields of sea ice thickness and snow thickness on sea ice are shown in
Figs. 3.1a,b. Because Antarctic sea ice is a complex mixture of different ice types and thicknesses, the calculation of surface fluxes over sea ice should be more realistic with this climatological dataset, compared to using constant values across the entire model grid.

Terrain height is interpolated from the 200 m grid spacing RAMP digital elevation model (DEM) dataset (Liu et al. 2001). This field is smoothed in the WRF preprocessing using a smoother-desmoother that prevents gridpoints that were previously not negative from becoming negative. The smoothing is done in order to avoid computational instability with steeply sloping topography. A special land use dataset is used for the McMurdo Dry Valleys region, which is interpolated from USGS GeoTiff files by Kevin Manning of NCAR MMM. This dataset identifies bare ground at 9 second grid spacing, and is interpolated to the model grid.

3.1.2 Model Domains

A series of nested domains is used in this study. This strategy is used so that the initial and boundary conditions from ERA-Interim can be properly scaled down to high resolution around the MDVs region. A 32-8-2 km grid spacing nesting structure is used, and the grids are shown in Fig. 3.2a-c. The 32-km grid (200 x 200 gridpoints) covers all of Antarctica and much of the surrounding Southern Ocean, while the 8-km grid (161x121 gridpoints) covers portions of East Antarctica, the Ross Sea, and the Ross Ice Shelf. The 2-km grid (193x165 gridpoints) shown in Fig. 3.2c is the domain primarily used for the climatological aspects of this study, as it covers the greater MDVs region.

79
There are 40 full-eta levels in the model, the model top is at 50 hPa, the lowest model level is approximately 13.4 m above the surface, and timesteps of 150s, 37.5s, and 9.375s are used for the three domains. Forcing between domains is one-way. The terrain height field for all three domains is smoothed twice using the smoother-desmoother described in the previous section.

A 0.5-km (693 x 493 gridpoints) domain is nested just inside the 2-km grid (covering all but a width of 5 gridpoints along the edge of the 2-km grid), and the interior portions of this grid are shown in Fig. 3.2d. This domain is used for the December 2006 / January 2007 case study in Chapter 4, and is run separately from the three outer domains, with one-way forcing from the 2-km domain at hourly intervals using WRF’s “ndown” program. Disadvantages to this method of nesting include having lateral boundary conditions updated hourly (rather than every timestep), and that the microphysics variables are not used for boundary conditions (only for initial conditions). However, the crucial advantage of this method is that it allows for different namelist specifications from the coarser domains. This allows for different horizontal diffusion options, which will be explained in the next section. For the 0.5-km grid, there are 55 vertical levels, and a ½ second timestep is used. The lowest model level is approximately 26.5 m above the surface, as having the lowest model level closer to the surface caused computational stability issues with the diffusion option being used. The terrain field is smoothed three times.

For comparing model output with AWS observations, it is important that the model gridpoints used to represent observing sites feature representative terrain height,
terrain slope, and land use characteristics. Figure 3.3a shows AWS station locations within a subset of the 2-km domain, and Figure 3.3b shows land use categories for this subset area. From Fig. 3.3a,b it is apparent that terrain height and land use can change rapidly across adjacent gridpoints. Because of the complex terrain of the MDVs, and potentially wide-ranging conditions between adjacent gridpoints, it was decided to not use any kind of spatial averaging of gridpoints surrounding an observation. Furthermore, the nearest gridpoint is not always representative of actual station conditions. The model gridpoint used to represent an observing station is instead chosen more subjectively. Terrain height and land use values for the four surrounding grid points are used along with spatial plots of the terrain field, to determine the location of each gridpoint relative to the valley orientation. The chosen gridpoint has to have the correct land use and be properly located in the valley geometry. From these points, the one closest in elevation and distance from the actual station location is chosen. Table 3.1 contains information about each model gridpoint used for the 2-km grid, and Table 3.2 has the same information for the 0.5-km grid. For the 2-km grid, all model points are within 2 km of the observing site, and most are within 100 m in elevation. Differences in both elevation and distance from observing site are generally smaller for the 0.5-km grid, as expected from Figure 3.4a,b, which shows terrain height and land use categories for a subset of the 0.5-km domain, with AWS locations overlain. Fine-scale orographic features within the MDVs are more apparent in the 0.5-km domain, and glaciers and melt lakes represented by the snow/ice land use category are better resolved than in the 2-km domain.
3.1.3 Model Physics and Nudging

Important components of any numerical weather prediction system, including Polar WRF, are the parameterization schemes. The parameterization schemes aim to include effects of physical processes that cannot be explicitly resolved by the numerical model. In this section, we step through each choice of parameterization scheme used in the Polar WRF simulations. For each class of parameterization scheme, a brief discussion of the chosen scheme is provided, including any motivating factors for its choice, and corresponding references.

The microphysics scheme handles the explicitly resolved water vapor, cloud, and precipitation processes. The scheme chosen here is the WRF Single-Moment 6-class scheme (WSM6, Hong and Lim 2006). The WSM6 scheme is a bulk single-moment scheme, which means that it uses a specified function for particle size distribution, and does not predict particle mixing ratio. The WSM6 differs from the WSM5 scheme in that a predictive equation for graupel is added. The WSM5 and WSM6 schemes feature more realistic melting and freezing processes, which occur over a deeper layer than in the WSM3 scheme, where such processes are instantaneous at the freezing level. The additional prognostic variables in the WSM6 become more important at higher spatial resolution, where distinct differences can be found compared to the WSM3 and WSM5 schemes.

Radiation schemes provide both the total radiative flux at the surface (needed for surface energy budget calculations) and the vertical radiative flux divergence in the free atmosphere (Stensrud 2007). Because the shortwave and longwave portions of the
electromagnetic spectrum are rather distinct, they can be handled separately. The longwave radiation scheme used here is the Rapid Radiative Transfer Model (RRTM, Mlawer et al. 1997). The RRTM handles integration of the absorption coefficient $k$ over the optical path by linearly interpolating stored sets of $k$ from a line-by-line radiative transfer model (LBLRTM). Integration of $k$ over frequency is handled by the correlated-$k$ method, which involves mapping $k$ from spectral space to a cumulative probability function $g$. The distribution of $g$ is then partitioned into subintervals that each correspond to a limited range of $k$, and a characteristic absorption coefficient value is obtained. Heating/cooling rates obtained using RRTM generally agree well with those from a full LBLRTM (Mlawer et al. 1997). An empirically-based shortwave radiation scheme is used, the Dudhia scheme (Dudhia 1989). Downward shortwave radiation received at the surface takes into account atmospheric attenuation from scattering and absorption from both clouds and the clear air. Cloud fraction in each gridbox is binary (0 or 1). Cloud back-scattering and absorption is bilinearly interpolated from theoretical functions, clear air water vapor absorption is calculated as a function of water vapor path (a function of solar zenith angle), and clear air scattering is uniform and proportional to atmospheric mass path length (also a function of solar zenith angle).

The planetary boundary layer (PBL) parameterization scheme is responsible for representing the effects of vertical subgrid-scale turbulent fluxes of momentum, heat, and moisture. The Mellor-Yamada-Nakanishi-Niino (MYNN) Level-2.5 scheme is used in this study (Nakanishi 2001; Nakanishi and Niino 2004, 2006). As a level-2.5 (1.5-order) scheme, the PBL scheme not only solves for the mean quantities of momentum, heat, and
moisture, as is done in first-order turbulence closure, but also for the turbulent kinetic energy (TKE), so that the eddy diffusivities are functions of TKE rather than just wind shear and stability (Stensrud 2007). Additional terms that are solved prognostically in the 2nd-order closure scheme, like the potential temperature variance, are solved diagnostically in the 1.5-order scheme. We use the 1.5-order closure scheme here because it is 40% less computationally expensive than the 2nd-order scheme (Nakanishi and Niino 2004). Nakanishi (2001) formulated a new master length scale $L$, used in dissipation and diffusion terms, that is a function of stability rather than just height. This new diagnostic equation for $L$ sums length scales for the surface layer, the turbulent structure of the PBL, and the buoyancy effect. Additionally, new closure constants were derived, and buoyancy effects upon the pressure covariance terms are accounted for. These changes help the MYNN scheme alleviate underestimation of mixed-layer depth and TKE magnitude and better resolve Kelvin-Helmholtz instability compared to the original Mellor-Yamada scheme.

Zängl et al. (2008) analyze the influence of PBL scheme choice and the vertical distribution of model levels on Alpine foehn simulations with MM5. An assessment of PBL scheme performance in forced mechanical turbulence under statically stable conditions has not received significant attention previously. TKE-based 1.5-order schemes are found to perform better than first-order schemes, as diffusion coefficients are more realistic near the surface. Wind speed is found to be strongly dependent upon PBL scheme choice. However, the placement of the lowest model level had a greater influence on results than the PBL scheme choice. Low-level temperatures steadily
decrease with decreasing lowest model level for all PBL schemes. The placement also affects flow separation in the boundary layer, and the height at which this occurs affects foehn and gravity wave structure. A lowest-level height of 7 m is found to compare better with observations than 36 m, however in some aspects a lowest-level height of 22 m is better than 7 m. Olson and Brown (2009) compare the MYNN PBL scheme with the Mellor-Yamada-Janjić (MYJ, Janjić 1994) scheme for a hybrid barrier jet over southeastern Alaska. The MYNN better simulates temperature, wind direction, and TKE, but MYJ features more realistic wind speed and jet structure. The MYNN scheme does resolve the long-standing issue of underestimated mixed layer depth in the MYJ scheme. Overall, the authors state that MYNN may better parameterize mixing of heat and momentum throughout a wide variety of atmospheric conditions. This relates to the fact that MYJ is tuned to a limited set of observations, while MYNN is tuned to large eddy simulations (LES).

The primary function of the land-surface model (LSM) is to provide surface heat and moisture fluxes over land and sea-ice points. The LSM uses atmospheric information such as surface layer variables (exchange coefficients from the surface layer scheme), radiation forcing, and precipitation as input to represent heat and moisture fluxes within the soil layers, evapotranspiration, runoff, and canopy layer processes. The Noah LSM is used in this study, which results from efforts between NCEP, Oregon State University, the U.S. Air Force, and the Hydrologic Research Laboratory (Chen and Dudhia 2001). The Noah LSM is based on a coupling of a diurnally dependent Penman potential evaporation approach, a multi-layer soil model, a primitive canopy model, canopy
resistance, and snow and sea-ice treatment. The Noah LSM has 4 soil layers with thicknesses (top to bottom) of 10, 30, 60, and 100 cm. Modifications to the Noah LSM for Polar WRF were described in the previous section, and additional details on initialization of the land-surface state for the MDVs are provided in the next section.

Cumulus parameterizations handle subgrid-scale convection and the associated vertical fluxes from updrafts and downdrafts. The Grell-Devenyi cumulus scheme is used here (Grell and Devenyi 2002). The cumulus scheme is only used with the 32-km domain, as it is assumed that convection will be explicitly resolved for the finer-scale domains. Even so, it is doubtful that convection is important for the high-latitude inner domains.

The surface layer scheme provides exchange coefficients and friction velocities for surface fluxes (LSM) and surface stress (PBL scheme). In WRF, surface schemes are paired with a corresponding PBL scheme, and the MYNN surface scheme is thus used here. Details on surface scheme formulation in WRF can be found in Skamarock et al. (2008).

In the MDVs, the slope and shadowing effects of the complex terrain are important for calculation of incident solar radiation (Dana et al. 1998). These effects are incorporated into the shortwave radiation parameterization in WRF (Garnier and Ohmura 1968). As of WRF version 3.2, this option is available with all shortwave radiation schemes (before it was only available with the Dudhia scheme).

While vertical diffusion is handled by the PBL scheme, horizontal diffusion is done in a separate formulation in the model. Typically, diffusion in WRF is handled
along model coordinate surfaces to second order. However, diffusion on model surfaces can lead to problems in regions of steep terrain. Any temperature gradients along valley slopes, which can occur in the MDVs during periods of strong solar forcing (i.e., an unstable layer near the ground), or during winter (strong radiational cooling at the surface, leading to stable stratification), will be diffused because sloping model surfaces are taken to be horizontal. Diffusion of water vapor along model surfaces tends to dry valleys and moisten air above mountains. The problem is most apparent for the MDV cold pools in winter, where positive model temperature biases can exceed $20^\circ$C with diffusion along model surfaces. The solution is to perform diffusion truly horizontally, as was done by Zängl (2002b) with MM5. For horizontal levels fully above terrain, Zängl treats diffusion truly horizontally, whereas for levels that intersect terrain, a transition from horizontal to model-level diffusion is done. This horizontal diffusion prevents spurious horizontal temperature gradients that lead to breaking gravity waves in the lower troposphere. Zängl (2004a,b) perform sensitivity experiments with horizontal diffusion options in MM5, and find that performing diffusion along model surfaces leads to incorrect timing of foehn breakthrough in the Wipp and Rhine Valleys of the Alps. These errors were more prevalent than any discrepancies caused by decreasing the horizontal resolution of the model.

For the maintenance of cold pools, Zängl (2003b) finds truly horizontal diffusion to be crucially important, as shown in Fig. 3.5a,b. In Fig. 3.5a, diffusion is treated along model surfaces, and the initial cold pool has been largely destroyed by diffusion in 12 hours. In Fig. 3.5b, using truly horizontal diffusion, the cold pool remains. Proper
representation of cold pools in the MDVs is important for the proper timing of foehn effects, and as such we are implementing truly horizontal diffusion in select Polar WRF simulations. WRF has an option to perform diffusion in three dimensions, rather than along model surfaces. The 3D diffusion is designed for LES applications, where horizontal grid spacing is small enough to resolve boundary layer eddies and use of a PBL scheme is not appropriate. For the 0.5-km simulations, we have kept the PBL scheme turned on, while using 3D diffusion. The PBL scheme will handle vertical turbulent diffusion, while horizontal diffusion will be handled truly horizontally. Some brief test simulations have shown that this combination of model settings is computationally stable and better resolves MDVs cold pools in early winter than simulations using diffusion along model surfaces. In order to implement this diffusion option for the 1 km simulations, these simulations have to be run separately from the outer domains using WRF’s ndown program, as explained in the previous section. For the 2 km simulations, we retain the fully nested structure and do not use the horizontal diffusion option, as it becomes numerically unstable along steep slopes in the 32 km and 8 km domains.

WRF includes an option for an explicit sixth-order horizontal diffusion operator on model coordinate surfaces, described in Knievel et al. (2007). This scheme is designed to remove poorly-resolved features of wavelength 2-4 times the model grid spacing when the implicit diffusion in the fifth-order upwind-biased advection is ineffective due to weak winds. Such undesirable features can occur in weakly forced environments with low wind speeds, such as boundary layer circulations in regions of
heterogeneous land surface characteristics. Knievel et al. (2007) test this diffusion option for lake breeze and salt breeze environments in Utah, and find that it reduces static stability in the lowest layers of the lake breeze. Overall, the sixth-order diffusion option eliminates a significant amount of noise without otherwise negatively affecting relevant portions of the model simulation. While this option shows promise for summer sea-breeze circulations in the MDVs, test simulations show that it prevents cold pool formation during periods of strong surface radiational cooling, and can diffuse heat out of valleys during summer heating. Therefore, the 6th order diffusion is not used in the simulations performed here.

In weather prediction models, upward propagating gravity waves need to be absorbed in upper levels, in order to prevent unphysical downward reflection from the model top. Several options are available in WRF, and here we use implicit Rayleigh damping of vertical velocity, described by Klemp et al. (2008). The damping is applied only to vertical velocity as a final adjustment at the end of each acoustic timestep. This is different than adding an explicit damping term directly into the equation for vertical velocity. Tests by Klemp et al. (2008) for an idealized squall line and of a real mountain wave simulation in the Rocky Mountains show that this damping option works better over larger horizontal scales than a traditional Rayleigh damping layer.

Instead of reinitializing the model simulations at short intervals to update the initial and boundary conditions (for example, AMPS is reinitialized every twelve hours), here we only reinitialize the 2 km simulations twice a month. This saves considerable computational expense because the number of 12-hour model spinup intervals necessary
for each reinitialization are reduced. However, after only a few days, model “drift” from reality becomes a problem in WRF, especially in the Antarctic. To combat model drift, we use four-dimensional data assimilation (FDDA), also known as nudging or Newtonian Relaxation, throughout our simulations. Nudging is a method of data assimilation originally designed for dynamic initialization of numerical weather prediction models (Hoke and Anthes 1976) that forces a numerical solution towards observations. This technique is less complex and computationally demanding than full four-dimensional data assimilation. It is also used to improve the accuracy of forecasts during the nudging period (Stauffer and Seaman 1990). Grid nudging forces the solution to a gridded analysis point-by-point, and can be used to keep a long-term model simulation in line with the forcing dataset (Skamarock et al. 2008).

Nudging involves an extra time tendency term in the solution for a given variable:

\[
\frac{\partial \theta}{\partial t} = F(\theta) + G_\theta W_\theta (\hat{\theta}_0 - \theta)
\]

where \( \theta \) is a given variable, \( F(\theta) \) is contributions from the normal forcing terms, \( G_\theta \) is a time-scale parameter, \( W_\theta \) is a weighting term, and the hat-term is the analysis field that the forecast is being nudged to. The time-scale parameter is typically set to the time scale of the slowest physical adjustment process in the model (Stauffer and Seaman 1990). A value of \( 3 \times 10^{-4} \text{s}^{-1} \) is used here, corresponding to one hour, for all nudged variables. For the simulations here, only the 32 km outermost grid is nudged to the six-hourly ERA-Interim reanalyzer output. The inner domains are not nudged because it can have a detrimental effect if relatively coarse-resolution forcing such as ERA-Interim is applied to much finer scale grids, especially with strong fine-scale forcing like that found in
regions of complex terrain (Stauffer and Seaman 1994). The purpose of the 32 km domain is to provide accurate boundary conditions for the 8-km, and subsequently 2-km, domains. Nudging keeps the 32-km domain in step with ERA-Interim, while the 32 km domain itself provides better simulation of mesoscale effects and higher temporal resolution than ERA-Interim. The $u$ and $v$ components of the wind vector and temperature are nudged, while water vapor is not, as it is generally a passive scalar and it is desired that atmospheric moisture be driven by both strongly- and weakly-forced processes in WRF. Nudging is done for the model vertical atmospheric column down to the 13th lowest model level. Experimentation shows that this allows for orographic forcing to be properly represented in the 32 km grid (which is important upstream of the MDVs), while still keeping the large-scale dynamical forcing in line with ERA-Interim. Such strong constraint throughout the model vertical depth is necessary, as simulations performed with nudging only in the top few model levels often lead to the generation of spurious cyclones over the Ross Ice Shelf and Ross Sea. Nigro et al. (2011) use self-organizing maps (SOMs) to evaluate AMPS/Polar WRF for specific weather patterns, and find large bias and RMSE values for situations of strong cyclonic forcing over the Ross Ice Shelf and Ross Sea with forecasts past a few days.

3.1.4 Modifications Specific to the MDVs

In this section modifications made to the Polar WRF simulations specifically for applications in the MDVs are discussed. The first change is the removal of the snow surface from the ERA-Interim initial conditions over model gridpoints featuring bare-
ground landuse. The primary purpose is to correct the surface albedo, to allow for thermally-generated circulations and to avoid an overestimation of the diurnal air temperature cycle like that from Valkonen et al. (2010). In the metgrid files, which represent initial conditions horizontally interpolated to the WRF grid, the snow field was simply set to zero over bare-ground points. Within the Noah LSM code, water-equivalent snow depth over bare-ground points is limited to 0.2 cm. This is done because WRF cannot simulate the removal of snow by strong foehn winds, as often occurs in the MDVs.

The second change deals with the soil categorization. There is no pre-set soil category for Antarctica, so during initialization, Antarctic soil is set to the default category, “Silty Clay Loam”, which is 10% sand, 56% silt, and 34% clay. While soil category is largely irrelevant for the extensive Antarctic ice sheet environment, it is of crucial importance for the MDVs, as the soil category affects soil heat capacity and soil moisture potential, which in turn affect ground heat and moisture fluxes. A more realistic soil category of “Loamy Sand”, comprising of 82% sand, 12% silt, and 6% clay, is used instead. This soil distribution agrees well with soil properties observed by Campbell et al. (1998) and Northcott et al. (2009).

The wind speed at the surface, like that measured by most LTER AWS sites at 3 m height AGL, is strongly dependent upon the characteristics of the ground surface. These characteristics are manifested in the roughness length, which is a measure of the aerodynamic roughness of the surface, and proportional to the height of surface roughness elements (Oke 1987). During periods of at least moderate winds and weak
stability, wind speed will be slower over a rough surface than over a smooth surface, because of the dependence on roughness length in the logarithmic wind profile equation. However, during periods of strong near-surface stability (which often occurs in the winter), a rougher surface leads to an enhanced mechanical turbulence, which can lower the Richardson Number to the point where mechanical turbulence dominates the stabilizing buoyancy effect. Correct simulation of turbulence regimes is important during foehn events, where foehn breakthrough at the surface depends upon the surface layer wind speed and stability. Surface wind speeds can be enhanced over rougher surfaces in low Richardson Number regimes, due to the downward momentum flux associated with mountain waves. Inadequate accommodations of roughness length in complex terrain can lead to poor model estimates of wind speed (Valkonen et al. 2010).

In WRF, the most convenient way to specify roughness length is to assign a constant value to each land use category. This is problematic in the MDVs, however. Lancaster (2004) measures roughness lengths at 17 MDVs sites, taking 8 30-m transects radiating from an anemometer at each site. Mean roughness lengths range from 0.0005 m at a site in Victoria Valley to 0.036 m at Lake Hoare. Even at individual sites, roughness lengths can vary by up to two orders of magnitude. As it is not feasible to specify roughness length on a gridpoint-by-gridpoint basis, the roughness length is set to 0.005 m for the bare ground land use category. This value is larger than that often used for sand, and represents a “middle” value for barren soil (Oke 1987). Polar WRF test simulations where roughness length is varied show that this value broadly represents the MDVs. Nonetheless, it is acknowledged that this value may not be truly representative of some
observing sites. Similar issues exist for surface albedo. Different soil compositions and soil moisture across the MDVs lead to varying albedo between sites. Thompson et al. (1971a) find an average albedo of about 0.2 at Vanda, while Campbell et al. (1998) find a range of albedo values in the MDVs, from about 0.06 in dark soils to 0.26 for sandstone. Hunt et al. (2010) give values ranging from 0.06 to 0.14. The base albedo for barren ground in WRF is set to 0.18.

To properly simulate surface heat and moisture fluxes in the MDVs, accurate profiles of soil heat and moisture are needed, as ground heat fluxes are important for the surface energy balance. Initial Polar WRF test simulations in the MDVs indicated that the input ERA-Interim soil variable fields were not adequate, as ERA-Interim does not have the spatial resolution to resolve the MDVs. It was decided that a soil state needed to be spun-up for the MDVs. Rather than run an offline land surface model (i.e., HRLDAS, Chen et al. 2007), it was deemed easier to run a one-year simulation to spin-up a land-surface state. The 32-8-2 km Polar WRF simulations were run from 1 November 2005 to 31 October 2006, and the land surface variables from 1200 UTC 31 October 2006 are used as the initial soil conditions for all 2-km summer simulations. Note that we are assuming that the soil conditions do not change dramatically between years. To make this one-year spinup effective, we replaced the ERA-Interim initial conditions with a modified soil state for the beginning of the spinup runs at 0000 UTC 1 November 2005. The soil temperature observations of Thompson et al. (1971b) are used as a basis for the soil temperatures in the MDVs. The monthly average values at Vanda Station from October and November 1969 and 1970 are averaged to obtain a base vertical temperature
profile for November 1 at Vanda. A third-order polynomial function is fitted to this profile,

\[ T_{\text{wv}}(z) = -1 \times 10^{-6}z^3 + 0.0008z^2 - 0.1662z - 10.688, \]  

(3.3)

where \( T_{\text{wv}} \) is soil temperature (°C) at Vanda, and \( z \) is depth below ground. From this function, soil temperatures for the four Noah LSM soil layers are calculated for Vanda. For all four soil layers in WRF, soil temperatures across the MDVs are assumed to be a function of terrain height and distance from the coast:

\[ T(z,D) = T_{\text{wv}}(z) - (0.01 \times (H - H_{\text{wv}}) + (0.09 \times (D - D_{\text{wv}})), \]  

(3.4)

where \( T(z, D) \) is the ground temperature at a gridpoint, \( z \) is the depth (m), \( H \) is terrain height (m), \( D \) is the great-circle distance from the coast (defined as the 164.1°E longitudinal), and the \( \text{wv} \) subscript denotes values at Vanda Station. The second RHS term in (3.4) represents a dry adiabatic lapse rate adjustment, and the third RHS term models the increase in temperature from the coast, based on Doran et al. (2002a). For the deep soil temperature (at 8 m) in WRF, the annual average 3 m ground temperature at Vanda from Thompson et al. (1971b) is used for all MDVs points. The resulting ground temperature field is used for bare ground grid points only. Figure 3.6a-d shows the initial soil temperature field.

For the initialized soil moisture field, the intent is to model the data from Campbell et al. (1998). Here, different formulations are used for each of the four soil layers, starting at the top:
where $M_1$ is volumetric soil moisture content for the top soil layer, $\lambda$ is longitude (degrees), and $H$ is terrain height (m). To account for enhanced soil moisture values along the shores of the melt lakes (Ikard et al. 2009), volumetric soil moisture content values for grid points representing LTER AWS melt lake sites are set to 0.15. Similar functions to (3.5) are set for layers 2, 3, and 4:

\[
M_2(\lambda) = \begin{cases} 
0.15, & \lambda > 163.5^\circ E \\
0.02 + (0.13 \times (\lambda - 162.5^\circ E)), & 162.5^\circ E < \lambda < 163.5^\circ E \\
0.04, & \lambda < 161^\circ E and H > 1500 \\
0.02, & otherwise
\end{cases}
\]

\[
M_3,4(\lambda) = \begin{cases} 
S, & \lambda > 163.5^\circ E \\
0.1 + ((S - 0.1) \times (\lambda - 162.5^\circ E)), & 162.5^\circ E < \lambda < 163.5^\circ E and H < 500 \\
(\lambda - 162.5^\circ E), & 500 < H < 1000 \\
0.02 + ((S - 0.02) \times (\lambda - 162.5^\circ E)), & 162.5^\circ E < \lambda < 163.5^\circ E and H > 1000 \\
0.1, & 161^\circ E < \lambda < 162.5^\circ E and H < 500 \\
0.02, & 161^\circ E < \lambda < 162.5^\circ E and H > 1000 \\
0.04, & \lambda < 161^\circ E and H > 1500 \\
0.02, & otherwise
\end{cases}
\]

where $S$ is saturation soil moisture content (based on a porosity of 0.41, McKay et al. 1998), and

\[
\epsilon = \left(0.02 + 0.08 \times \left(\frac{1000 - H}{500}\right)\right)
\]

(3.8)

Lakeshore grid point values are set to 0.25 for layer 2, and saturation for layers 3 and 4.

Figure 3.7a-d shows the initial soil moisture field.
The final modification to Polar WRF for the MDVs involves the use of fractional
sea ice with the MYNN surface scheme. MYNN PBL/surface scheme options were
added to WRF in version 3.1, but coupling to the fractional sea ice option was never
done. A wrapper program was added to module_surface_driver.F to handle the separate
surface layer calls for sea ice and open water. An additional problem related to fractional
sea ice was rectified with the help of model improvements to WRF version 3.2.1. When
there is a land/sea mask mismatch between the input sea ice dataset and WRF, where a
gridpoint is land in the input sea ice dataset but water in WRF, fractional sea ice values
get set to extremely low values (around 0.02 to 0.03) in the metgrid spatial interpolation.
With these small ice fractions, the resulting skin temperature is unphysical (about 20 K),
and the surface heat fluxes become unrealistically large causing the model to repeatedly
-crash. A new option in WRF 3.2.1, “tice2tsk_if2cold”, sets the skin temperature of ice to
the land skin temperature at that point, avoiding the interpolation based on the incorrect
fractional sea ice concentration values.

3.2 Antarctic Mesoscale Prediction System (AMPS)

The Antarctic Mesoscale Prediction System (AMPS, Powers et al. 2003) is an
experimental real-time forecasting system jointly developed and run by the Polar
Meteorology Group of the Byrd Polar Research Center at The Ohio State University and
the Mesoscale and Microscale Meteorology division of the National Center for
Atmospheric Research (NCAR) in support of United States Antarctic Program (USAP)
operations. For the time periods considered here, AMPS employs Polar MM5, a version
of the fifth-generation Pennsylvania State University-NCAR Mesoscale Model (Grell et al. 1994) optimized for use in polar regions by the Polar Meteorology Group at the Byrd Polar Research Center, The Ohio State University (Bromwich et al. 2001, Cassano et al. 2001). Polar MM5 includes a modified parameterization for the prediction of ice cloud fraction, improved cloud-radiation interactions, an optimal stable boundary layer treatment, improved calculation of heat transfer through snow and ice surfaces, and the addition of a fractional sea-ice surface type.

AMPS output used in this case study is at 20-km resolution, on a grid domain covering Antarctica and much of the surrounding Southern Ocean, and at 2.2-km resolution on a grid encompassing the Ross Island area, extending into the McMurdo Dry Valleys. There are 31 vertical half-sigma levels, with 11 levels in the lowest 1000 m to capture the complex interactions in the planetary boundary layer. The lowest half-sigma level is about 13 m above the surface. For comparisons with AWS observations at 3 m above the surface, AMPS wind components are interpolated logarithmically from the lowest model level to 3 m. Air temperature at 3 m is interpolated linearly between the surface and the lowest model level, while relative humidity at 3 m is estimated with the interpolated temperatures and calculated pressure.

AMPS Polar MM5 is initialized twice daily at 0000 and 1200 UTC. The initial and boundary conditions are derived from the National Centers for Environmental Prediction Global Forecasting System (GFS) model. AMPS uses three-dimensional variational data assimilation (3DVAR). The observations available for assimilation into AMPS include reports from radiosondes, surface SYNOP reports, AWS observations,
ship and buoy reports, atmospheric motion vector winds from satellites, and GPS radio occultation soundings. AMPS ingests sea ice data daily from the National Snow and Ice Data Center for its fractional sea ice depiction.

Guo et al. (2003) evaluate Polar MM5 performance over Antarctica for a one-year period (1993) on a 60-km resolution domain and show that the intra- and interseasonal variability in pressure, temperature, wind, and moisture are well-resolved. Bromwich et al. (2005) evaluate 2-years of AMPS Polar MM5 forecasts on the 30-km domain and show that the same variables are well resolved at synoptic time-scales. Tastula and Vihma (2011) validate standard WRF simulations using different parameterization scheme combinations and Polar WRF against station observations around Antarctica for a single winter month. Polar WRF does better than standard WRF for a 9-day experiment and in a 20-km grid spacing simulation over Dronning Maud Land, but considerably worse with temperatures at Vostok. However, results are highly dependent upon the choice of physics schemes. With either model configuration, wind speed correlations are especially poor at observing sites in complex terrain with an outer 100-km grid spacing domain, highlighting the importance of higher spatial resolution in regions of complex terrain. Monaghan et al. (2005) reviewed the climate of the McMurdo region (including the MDVs) in the 3.3 km grid domain and show that AMPS captures important temporal and spatial aspects of the region’s climate. Additionally, Steinhoff et al. (2008) demonstrated that the 3.3 km domain is valuable on an event basis in the analysis of a down-slope windstorm at Ross Island.
3.3  McMurdo Dry Valleys LTER AWS Observations

The primary source of observations for the MDVs is the McMurdo Dry Valleys Long-Term Ecological Research (LTER) program (Doran et al. 1995). A map of LTER AWS sites is shown in Fig. 2.1, and information pertaining to each station is presented in Table 3.3. All data is obtained from the MDV LTER website at http://www.mcmlter.org, where additional details on the MDV LTER program can also be found. The measured quantities vary between each observing site, but the following variables are used in this study and exist at all stations: air temperature, relative humidity, wind speed, wind direction, ground temperature, and incoming shortwave radiative flux. Relative humidity is calculated with respect to ice when air temperatures are below freezing. Observations are taken at 3 m height, except for Canada Glacier, where measurements are taken at 2 m height. To facilitate comparison with most AWS observations, selected variables are output at 3 m height in Polar WRF by adjusting the height that exchange coefficients are calculated at in the surface layer scheme. The sampling interval for observations varies with time and between sites. Before November 1995, most sites sampled all variables at 30-second intervals, and were averaged for output over intervals ranging from 10 minutes to 3 hours. Most stations switched to sampling wind every 1 second and all other variables every 30 seconds in November 1995. Between November 1997 and January 1998, all stations transitioned to and remain sampling wind at 4 second intervals and all other variables at 30 second intervals except for Lake Bonney, which has remained at 1 s / 30 s sampling frequency. Currently, all variables are averaged and output every 15 minutes. All data is quality-controlled by LTER investigators prior to distribution.
3.4 Figures and Tables

Figure 3.1. 2.5° latitude x 5.0° longitude bin-averaged (a) sea ice thickness (m), and (b) snow thickness on top of sea ice (m), from the ASPeCt program observations.
Figure 3.2. Terrain height (shaded and contoured, m) from (a) 32-km, (b) 8-km, (c) 2-km, and (d) 0.5-km Polar WRF domains. Black rectangles represent nested domains. Relevant geographical features labeled.

Continued
Figure 3.2: Continued
Figure 3.3. Actual station locations (green dots) in relation to Polar WRF (a) Terrain height (shaded and contoured, m), and (b) Land use category (White: Snow/Ice, Brown: Barren or Sparsely Vegetated, Blue: Water) for 2-km domain.
Figure 3.4. Same as Fig. 3.3, except for 0.5-km domain.
Figure 3.5. Vertical cross-section through a basin with zero large-scale wind of potential temperature (contoured, K) after 12 hours simulation with (a) diffusion along model vertical levels and (b) truly horizontal diffusion. Dashed lines represent initial temperature field. From Zängl (2003b), Fig. 1.
Figure 3.6. 2-km domain initial subsurface temperature (color shaded, °C) at 0000 UTC 1 November 2005 (beginning of one-year spinup simulation) for (a) 0-10 cm layer, (b) 10-40 cm layer, (c) 40-100 cm layer, and (d) 100-200 cm layer. Black dots represent LTER AWS station locations.
Figure 3.7. 2-km domain initial subsurface soil moisture (color shaded, m$^3$ m$^{-3}$) at 0000 UTC 1 November 2005 (beginning of one-year spinup simulation) for (a) 0-10 cm layer, (b) 10-40 cm layer, (c) 40-100 cm layer, and (d) 100-200 cm layer. Black dots represent LTER AWS station locations.
Table 3.1. Gridpoints in 2-km Polar WRF domain used for comparison with LTER AWS observations. Distance is the distance from the actual station location.

<table>
<thead>
<tr>
<th>Station</th>
<th>WRF X</th>
<th>WRF Y</th>
<th>WRF Lat. (°N)</th>
<th>WRF Lon. (°E)</th>
<th>Elev. Diff. WRF – LTER (m)</th>
<th>Distance (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EC</td>
<td>117</td>
<td>86</td>
<td>-77.5841</td>
<td>163.4940</td>
<td>0.82</td>
<td>1.90</td>
</tr>
<tr>
<td>TF</td>
<td>114</td>
<td>83</td>
<td>-77.6209</td>
<td>163.1760</td>
<td>7.82</td>
<td>1.12</td>
</tr>
<tr>
<td>TH</td>
<td>111</td>
<td>81</td>
<td>-77.6399</td>
<td>162.8810</td>
<td>94.51</td>
<td>1.67</td>
</tr>
<tr>
<td>TCo</td>
<td>114</td>
<td>86</td>
<td>-77.5684</td>
<td>163.2500</td>
<td>20.84</td>
<td>0.88</td>
</tr>
<tr>
<td>TCa</td>
<td>111</td>
<td>83</td>
<td>-77.6049</td>
<td>162.9320</td>
<td>278.65</td>
<td>1.21</td>
</tr>
<tr>
<td>TB</td>
<td>108</td>
<td>76</td>
<td>-77.7108</td>
<td>162.5090</td>
<td>190.24</td>
<td>1.13</td>
</tr>
<tr>
<td>TTa</td>
<td>104</td>
<td>73</td>
<td>-77.7409</td>
<td>162.1030</td>
<td>6.63</td>
<td>0.60</td>
</tr>
<tr>
<td>WB</td>
<td>105</td>
<td>91</td>
<td>-77.4328</td>
<td>162.6470</td>
<td>10.08</td>
<td>1.13</td>
</tr>
<tr>
<td>WV</td>
<td>96</td>
<td>83</td>
<td>-77.5202</td>
<td>161.7210</td>
<td>9.74</td>
<td>1.34</td>
</tr>
<tr>
<td>VV</td>
<td>95</td>
<td>90</td>
<td>-77.3945</td>
<td>161.8230</td>
<td>10.94</td>
<td>1.93</td>
</tr>
<tr>
<td>BV</td>
<td>90</td>
<td>63</td>
<td>-77.8325</td>
<td>160.6800</td>
<td>251.82</td>
<td>0.74</td>
</tr>
</tbody>
</table>

Table 3.2. Gridpoints in 0.5-km Polar WRF domain used for comparison with LTER AWS observations. Distance is the distance from the actual station location.

<table>
<thead>
<tr>
<th>Station</th>
<th>WRF X</th>
<th>WRF Y</th>
<th>WRF Lat. (°N)</th>
<th>WRF Lon. (°E)</th>
<th>Elev. Diff. WRF – LTER (m)</th>
<th>Distance (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EC</td>
<td>427</td>
<td>305</td>
<td>-77.5862</td>
<td>163.4136</td>
<td>78.49</td>
<td>0.29</td>
</tr>
<tr>
<td>TF</td>
<td>417</td>
<td>297</td>
<td>-77.6080</td>
<td>163.1609</td>
<td>-19.00</td>
<td>0.38</td>
</tr>
<tr>
<td>TH</td>
<td>406</td>
<td>289</td>
<td>-77.6283</td>
<td>162.8869</td>
<td>-23.55</td>
<td>0.45</td>
</tr>
<tr>
<td>TCo</td>
<td>420</td>
<td>308</td>
<td>-77.5639</td>
<td>163.2894</td>
<td>-22.89</td>
<td>0.19</td>
</tr>
<tr>
<td>TCa</td>
<td>409</td>
<td>293</td>
<td>-77.6149</td>
<td>162.9731</td>
<td>-49.83</td>
<td>0.27</td>
</tr>
<tr>
<td>TB</td>
<td>393</td>
<td>265</td>
<td>-77.7153</td>
<td>162.4684</td>
<td>71.29</td>
<td>0.14</td>
</tr>
<tr>
<td>TTa</td>
<td>380</td>
<td>255</td>
<td>-77.7408</td>
<td>162.1367</td>
<td>-36.74</td>
<td>0.21</td>
</tr>
<tr>
<td>WB</td>
<td>385</td>
<td>327</td>
<td>-77.4341</td>
<td>162.7003</td>
<td>-32.94</td>
<td>0.10</td>
</tr>
<tr>
<td>WV</td>
<td>344</td>
<td>295</td>
<td>-77.5160</td>
<td>161.6718</td>
<td>-72.24</td>
<td>0.13</td>
</tr>
<tr>
<td>VV</td>
<td>340</td>
<td>326</td>
<td>-77.3758</td>
<td>161.7962</td>
<td>6.54</td>
<td>0.25</td>
</tr>
<tr>
<td>BV</td>
<td>321</td>
<td>215</td>
<td>-77.8280</td>
<td>160.6530</td>
<td>27.67</td>
<td>0.15</td>
</tr>
<tr>
<td>Station</td>
<td>Latitude (°N)</td>
<td>Longitude (°E)</td>
<td>Elevation (m)</td>
<td>Install Year</td>
<td>Location in relation to lake</td>
<td></td>
</tr>
<tr>
<td>---------</td>
<td>---------------</td>
<td>----------------</td>
<td>--------------</td>
<td>--------------</td>
<td>-----------------------------</td>
<td></td>
</tr>
<tr>
<td>EC</td>
<td>-77.5887</td>
<td>163.4175</td>
<td>26</td>
<td>1997</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>TF</td>
<td>-77.6109</td>
<td>163.1696</td>
<td>19</td>
<td>1993</td>
<td>Island</td>
<td></td>
</tr>
<tr>
<td>TH</td>
<td>-77.6254</td>
<td>162.9004</td>
<td>78</td>
<td>1993</td>
<td>Island</td>
<td></td>
</tr>
<tr>
<td>TCo</td>
<td>-77.5637</td>
<td>163.2801</td>
<td>290</td>
<td>1993</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>TCa</td>
<td>-77.6127</td>
<td>162.9634</td>
<td>264</td>
<td>1994</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>TB</td>
<td>-77.7144</td>
<td>162.4641</td>
<td>64</td>
<td>1993</td>
<td>East Lobe</td>
<td></td>
</tr>
<tr>
<td>TTa</td>
<td>-77.7402</td>
<td>162.1284</td>
<td>334</td>
<td>1994</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>WB</td>
<td>-77.4335</td>
<td>162.7036</td>
<td>279</td>
<td>1994</td>
<td>West Shore</td>
<td></td>
</tr>
<tr>
<td>WV</td>
<td>-77.5168</td>
<td>161.6678</td>
<td>296</td>
<td>1994</td>
<td>East Shore</td>
<td></td>
</tr>
<tr>
<td>VV</td>
<td>-77.3778</td>
<td>161.8006</td>
<td>351</td>
<td>1995</td>
<td>West Shore</td>
<td></td>
</tr>
<tr>
<td>BV</td>
<td>-77.8280</td>
<td>160.6568</td>
<td>1176</td>
<td>2000</td>
<td>-</td>
<td></td>
</tr>
</tbody>
</table>

Table 3.3. MDV LTER AWS observation locations.
Chapter 4: Foehn Mechanism

A major weakness of previous research on the McMurdo Dry Valleys (MDVs) environment concerns the physical explanations of the strong westerly wind events and the opposing cool easterly breezes in summer. Use of the MDVs in a climate change or large-scale climate variability context will remain ad hoc until robust physical concepts are presented. The purpose of this chapter is to present a case study that illustrates the physical processes involved in MDVs foehn and sea breeze events. New insights will be provided that discard the katabatic concept from previous MDVs literature.

The chapter begins with an overview of the large-scale synoptic situations present for foehn events, based on the study by Speirs et al. (2010) using two years (2006 and 2007) of AMPS output. Then, a case study of a prominent summer foehn event in December 2006 / January 2007 is presented. Concepts such as gap flow, mountain wave structure, pressure-driven channeling in valleys, and terrain blocking are shown to be important towards describing MDVs winds. Output from 500 m grid-spacing Polar WRF simulations are utilized for this case study to better represent small-scale features. The features identified in the case study analysis of this chapter will carry over to Chapter 5, where these features are shown to be robust across many events.
4.1 Large-scale Setup

While several earlier studies of MDVs meteorology have related strong westerly wind events to the greater Antarctic katabatic wind regime, Speirs et al. (2010) showed that the synoptic-scale circulation over the Ross Ice Shelf and the Ross Sea is largely responsible for foehn events and foehn variability in the MDVs. Figure 4.1a,b shows the AMPS composite sea-level pressure and near-surface wind vectors for 172 non-foehn days and 172 foehn days, respectively, in 2006 and 2007. Immediately apparent is the well-defined low-pressure center off of Marie Byrd Land in the foehn composite that is absent in the non-foehn composite. Sea-level pressure is significantly lower over most of the Ross Ice Shelf and eastern Ross Sea for foehn events (Fig. 4.1c), resulting in a stronger pressure gradient over the Ross Ice Shelf (Fig. 4.1d). Speirs et al. (2010) implicated increased cyclone activity in the Ross Sea, and associated strengthened southerly geostrophic winds, towards increased foehn activity in the MDVs via mountain wave forcing. This is supported by Doran et al. (2008), who find a stronger 500 hPa geopotential height gradient over the MDVs region during foehn events. Similarly, Fudeyasu et al. (2008) in their study of the downslope wind “Hirodo-Kaze” over Mt. Nagi in Japan find that the windstorms always occur as an intense cyclone moves through the vicinity.

The differences in the synoptic-scale situation are largely independent of season, as shown in Fig. 4.2a-d (summer) and Fig. 4.3a-d (winter). Similar sea-level pressure patterns can be found in both summer and winter (Fig. 4.2a,b and Fig. 4.3a,b) as in the annual composites. Although foehn-non-foehn sea-level pressure differences are not
statistically significant in winter (Fig. 4.3c) as they are in summer (Fig. 4.2c), likely due to increased cyclone activity near the Antarctic coast in winter (e.g., Simmonds et al. 2003; Hoskins and Hodges 2005), in both seasons the pressure gradient and associated southerly geostrophic winds are stronger over the Ross Ice Shelf (Fig. 4.2d and Fig. 4.3d).

Figure 4.4a,b shows the AMPS composite sea-level pressure and near-surface wind vectors for 2006 and 2007, respectively. 2007 featured 10% more foehn days than 2006, and this difference is reflected in the tightening of the isobars in 2007 compared to 2006. The difference is shown in Fig. 4.4c, and is statistically significant over the western Ross Ice Shelf and western Ross Sea. A statistically significant pressure gradient difference from 2006 to 2007 (Fig. 4.4d) suggests stronger southerly geostrophic winds over the western Ross Ice Shelf.

4.2 29 December 2006 – 2 January 2007 Case Study

To illustrate many of the physical features that comprise MDV foehn events, and to show how characteristics of foehn events vary with the synoptic-scale circulation, we present a case study from late December 2006 / early January 2007. This was a prominent foehn event that lasted for several days, and affected all LTER valley AWS sites. Figure 4.5 shows a series of surface and 500 hPa charts, every 18 hours, spanning the event from 0000 UTC 29 December 2006 to 1800 UTC 1 January 2007. At 0000 UTC 29 December 2006, an upper-level low is located over the northern Ross Ice Shelf, and a weak low-level cyclone is centered just off of Siple Coast, downstream of an upper-
level trough. Over the next 18 hours, the low-level cyclone moves northwest across the Ross Ice Shelf. By 1200 UTC 30 December, the low-level cyclone moving equatorward across the Ross Ice Shelf has dissipated, and a stronger system with a barotropic structure moves poleward to the tip of Marie Byrd Land. The cyclone moves westward to the north-central Ross Ice Shelf by 1200 UTC 31 December, and weakens as it moves into the southwestern Ross Sea at 0000 UTC 1 January. By 1800 UTC 1 January, the aforementioned system has merged with another disturbance dipping southeastward from Adélie Land, and moves eastward. Through the sequence of the two cyclones during this time period, the greater MDVs region features a variety of flow orientations (wind speed and direction), both near the surface and aloft, which will be shown to determine conditions in the MDVs themselves.

While it is known that MDVs foehn events often coincide with the presence of cyclones over the Ross Ice Shelf, it will now be shown that foehn conditions in the MDVs are highly dependent upon the ambient upstream wind direction. Figure 4.6 shows time series comparisons of all valley LTER AWS (light colors) and Polar WRF (dark colors) at 15 minute intervals for 3-m air temperature, relative humidity, wind direction, and wind speed from the 500-m grid spacing simulations. This figure, along with the surface and 500 hPa plots in Fig. 4.5, will be referred to throughout the case study discussion. Note that Taylor Glacier AWS was not operational during this period.

4.2.1 Foehn Initiation – Gap Flow
While wind direction is already westerly at Lake Bonney and Lake Vanda at 0000 UTC 29 December, foehn conditions begin around 1200 UTC 29 December at all observing sites (Fig. 4.6). While the timing of foehn onset in Polar WRF is good (much better than the 2-km grid spacing simulations, where foehn onset is about 6 hours too early at all sites, not shown), model wind speeds are too high, especially in Taylor Valley. While a gradual warming and drying had been occurring in the MDVs since 25 December, foehn onset corresponds with a temperature and wind speed increase. Figure 4.7a shows near-surface streamlines and wind speed at 0000 UTC 29 December. While previous studies have noted that foehn events often begin in the western MDVs first (e.g., Speirs et al. 2010), the reason for this has not been elucidated. From Fig. 4.7a, where wind direction has already shifted to southwesterly at Lake Bonney and Lake Vanda in the model and in the observations, the flow appears to be approaching Taylor Valley from the south, and is deflected eastward down-valley. This becomes clearer by 1800 UTC 29 December (Fig. 4.7b), as southerly flow stagnates in western Taylor Valley, and down-valley flow increases near Taylor Glacier.

The southerly flow results from a gap wind between the Royal Society Range and the East Antarctic ice sheet. Gap winds can occur in level gaps like straits, or associated with mountain passes. Inspection of the Antarctic terrain field in Fig. 4.8a shows a gap just west of the Royal Society Range and south of western Taylor Valley. Figure 4.9 shows a cross section of terrain height across the gap corresponding to the white transect shown in Fig. 4.8a. The width of the gap is approximately 30 km, with the terrain height lowest on the eastern end. However, as apparent from Fig. 4.8a, the terrain is extremely
complex along the gap, and there is no straight pass of lowest terrain height through the
gap. While gap winds are primarily forced by the component of the ambient pressure
gradient along the gap axis (Overland and Walter 1981), mesoscale effects like mountain
waves can have dominating effects, especially with vertical constrictions (Pan and Smith
1999; Gaberšek and Durran 2006).

The cross-gap pressure difference is influenced by both the ambient pressure
gradient (that associated with synoptic-scale weather systems in the region) and by
mesoscale effects like pressure drag (the pressure difference exerted by the mountain
wave on the obstacle) and flow blocking. As mentioned earlier, MDVs foehn events
often coincide with a cyclone over the Ross Sea/Ross Ice Shelf, and one potential source
of a cross-gap pressure difference is upstream flow blocking. This is apparent from
several of the surface plots in Fig. 4.5, where geostrophic flow impinges on the Royal
Society Range, which juts out onto the Ross Ice Shelf. Figure 4.10 shows a surface plot
from 1800 UTC 29 December from the 8-km grid spacing domain 2. A mesoscale high-
pressure region is located along the Transantarctic Mountains near Skelton Glacier. The
pressure gradient in the western sector of the cyclone is stronger than to the east or north,
clearly indicating the influence of the Antarctic continent. The western Ross Ice Shelf is
well-known for barrier winds, associated with low-level cyclonic flow being blocked
against the Transantarctic Mountains and directed northward by an evolving mesoscale
pressure gradient (O’Connor et al. 1994; Steinhoff et al. 2008). In Fig. 4.10, a westerly
ageostrophic component to the surface flow is evident in the western sector of the
cyclone, partly associated with barrier effects.
To estimate the pressure difference across the gap, the *reduced pressure* along a transect across the gap is calculated. Because of the varying surface elevation along the transect, surface pressure cannot be used. Reduction of pressure to sea-level is problematic in regions where there are strong variations in the vertical profile of virtual temperature, because a constant lapse rate (usually U.S. standard lapse rate, or dry adiabatic lapse rate) is used. This problem is exacerbated in Antarctica, where sea-level pressure is meaningless over higher elevations because of strong inversion conditions in the boundary layer. The method used here is similar to that of Mayr et al. (2002), who measured pressure differences across the Brenner Gap in the Alps in an instrumented automobile. Instead of reducing pressure to sea level, pressure is reduced to a constant height, defined here to be the highest point along the transect. In Mayr et al. (2002), a *slantwise reduction* was performed, where a pseudo-vertical profile was constructed along the automobile route upstream or downstream of the point used for reduction. Because we are using model output, and have vertical profiles of atmospheric conditions along the entire transect, such a slantwise reduction is not necessary. Instead, pressure is adjusted from the surface up to a constant height level using the virtual temperature profile in the model output at each point along the transect. Reduced pressure differences between the starting and ending points of the transect are computed to give the cross-gap pressure difference. Figure 4.11 shows the reduced pressure difference across the gap (in red), along the *magenta* transect in Fig. 4.8a. The pressure difference increases during 29 December to about 2.8 hPa by early 30 December, before dropping under 2 hPa later in the day. After that, the pressure difference increases substantially to 3.5-4.5 hPa after
1200 UTC 31 December. To see if there is a connection between this pressure difference and the gap wind speed, the area-averaged (white averaging box in Fig. 4.8a) near-surface gap wind speed is shown in Fig. 4.11 (in blue). The increases during 29 December and the maximum around 0000 UTC 1 January both match between the variables. However, strong gap winds (around 20 m s\(^{-1}\)) exist late 30 December when the pressure difference is moderate at about 2.8 hPa. On the other hand, gap wind speeds decrease early 31 December while the cross-gap pressure difference is increasing. While the broad distribution of the gap wind speed scales with the cross-gap pressure difference, there are fluctuations in the gap wind speed that are independent of the cross-gap pressure difference.

Figure 4.11 shows 600 hPa wind direction (in green) averaged for all model points in the domain south of the gap. This level is chosen to represent the ambient flow, as it is above the boundary layer across the domain. The period late on 30 December features an ambient south-southwesterly wind direction, meaning flow is generally off-continent. Thus, a strong cross-gap pressure difference would not be expected because of less blocking south of the Royal Society Range. When ambient flow has a westerly component, the flow upstream of the gap originates at a higher level, so that it is less apt to be blocked upstream of the gap. The dashed line in Fig. 4.11 approximates the gap axis at 203°, and Fig. 4.11 shows the 600 hPa domain-averaged wind speed (in orange). Late on 30 December, strong gap flow results from flow nearly parallel to the gap axis, with ambient wind speeds around 15 m s\(^{-1}\). This contrasts with the period around 0000 UTC 1 January, where gap flow is just as strong, but the ambient flow is more easterly.
(less of a cross-gap wind component). While the ambient wind speed largely approximates the gap wind speed, the gap flow setup differs depending on the ambient wind direction. When the ambient flow is south-southwesterly, it can directly flow into the gap, with a cross-gap pressure gradient enhanced by blocking west of the Royal Society Range. When the ambient flow becomes more easterly, the cross-gap wind component is smaller, but a strong cross-gap pressure difference forces the gap flow.

4.2.2 Mountain Waves

While gap flow provides significant forcing for flow into the MDVs from the south, Fig. 4.11 shows that the cross-gap pressure difference does not completely describe the wind speed in the gap. Figure 4.8a shows the complex topography in the gap, and that the gap is best described as a pass, or a saddle point in the terrain field. Vertical constrictions in the gap are surmised to be more important to the flow response than horizontal constrictions (Pan and Smith 1999), and mountain wave activity in both elevated and level gaps significantly affect the resulting flow through momentum fluxes and pressure drag (Colle and Mass 1998b; Gaberšek and Durran 2004; Gohm and Mayr 2004; Gohm et al. 2008). Bromley (1985) and Fountain et al. (2009) observe wind-blown snow into the MDVs from valley walls to the south, supporting the concept of mountain waves over the MDVs. While mountain waves were identified in AMPS output by Speirs et al. (2010), here we provide more detail to structure and features.

Starting with 1800 UTC 29 December, we find relatively strong ambient flow aloft from a southerly direction (Fig. 4.5), associated with a low-pressure disturbance
over the north-central Ross Ice Shelf (Fig. 4.5, Fig. 4.10). However, Fig. 4.7b, which shows near-surface wind speed and streamlines at 1800 UTC 29 December, shows weaker flow (< 10 m s\(^{-1}\)) upstream of the gap from a westerly direction. The reason for this westerly flow with ambient southerly geostrophic forcing lies with blocking effects of the Transantarctic Mountains well upstream. Figure 4.12 shows wind speed and streamlines from model level 10 (approx. 610 m AGL) at 1800 UTC 29 December. Flow diversion occurs along the Transantarctic range well south of 80°S, and extends northward to the Royal Society Range. Because peaks in the Transantarctic Mountains extend upwards of 4500 m, blocking effects would be expected to be vertically extensive. Thus, ambient southerly forcing results in weaker upstream flow into the gap. The westerly near-surface flow into the region just upstream of the gap is associated with blocking phenomena. In the absence of synoptic-scale forcing (as occurs south of the gap, due to blocking effects), the near-surface wind field will be forced by localized pressure gradients resulting from flow incident upon terrain features over the ice slopes (Parish and Cassano 2003). Note that significant katabatic effects would not be expected in summer (Parish and Cassano 2003; van den Broeke and van Lipzig 2003). Figure 4.7b shows blocking to the west, and downslope flow in the lee approaching the upstream gap region. This flow is then blocked by the Royal Society Range, leading to a decrease in near-surface wind speed. The point to be made here is that ambient southerly flow results in weak flow in the lower troposphere upstream of the gap, due to blocking effects. Figure 4.13 shows a vertical profile of wind speed and wind direction at the origin of the red transect in Fig. 4.8a at 1800 UTC 29 December. Up to almost 2 km ASL, wind speed
decreases from about 10 m s\(^{-1}\) to 2 m s\(^{-1}\). The westerly wind at the surface turns to southerly through this layer, associated with decreasing surface forcing and better representing the blocking effects of the upstream Transantarctic Mountains with ambient southerly flow. Wind speed recovers to about 20 m s\(^{-1}\) at about 3.5 km ASL, as blocking effects decrease with height. Wind direction remains generally southerly up to about 6 km ASL before turning southwesterly aloft.

Figure 4.14 shows a cross section of potential temperature and along-transect wind speed at 1800 UTC 29 December along the red transect in Fig. 4.8a. Interpretation of wave effects in complex multiscale terrain is difficult, but some features can still be identified and explained. At first glance, the flow in Fig. 4.14 looks to be nonlinear, evidenced by the large wave amplitudes, strong leeside acceleration, and wave breaking regions. To quantify this, we calculate the *nondimensional mountain height* (also known as the *inverse Froude number*) for a volume upstream of the gap. The nondimensional mountain height is expressed as

\[
M = \frac{Nh_m}{U},
\]

where \(U\) is the mean-state wind speed, \(N\) is the mean-state Brunt-Väisälä frequency, and \(h_m\) is the terrain height. Here we calculate \(M\) over the blue-outlined volume depicted in Fig. 4.8a, up to 2000 m. As recommended by Reinecke and Durran (2008), the average \(N\) in this layer is used to estimate the mean-state \(N\) in (4.1), rather than the bulk difference between the top and bottom of the layer. \(M\) is a measure of the nonlinearity of the flow, with an established threshold of 1.1 for continuously stratified, hydrostatic flow over a three-dimensional axisymmetric Gaussian hill (Smith and Grønås 1993). Values
above 1.1 induce wave breaking above the lee slope of the obstacle. A value of 1.62 is found at 1800 UTC 29 December, which is reasonable considering the wave response over the gap (the first ridge in Fig. 4.14), and qualitatively agrees with idealized simulations with a similar value of $M$ (e.g., Lin 2007, p. 127, 129). Blocking is apparent upstream of the ridge, with transitional flow at the crest and acceleration in the lowest 1 km in the lee. Wave steepening occurs over the lee slope associated with flow stagnation at 4-5 km elevation. The flow pattern indicates a hydrostatic vertically propagating wave regime, due to the relatively long length scale of the gap, and the single wave pattern over the ridge.

Flow accelerates in the lee of the first ridge in Fig. 4.14, underneath the region of steepened isentropes. This behavior is best explained by hydraulic theory, consisting of a layer of homogeneous fluid bounded by a free surface flowing over a ridgelike obstacle (e.g., Long 1954; Durran 1990). This is a rather special situation in the atmosphere, best matched by a layer of constant density underneath a strong inversion. A better hydraulic analog for most real atmospheric conditions consists of an infinitely deep, continuously stratified fluid with a single interface of discontinuous stratification (Durran 1986a). The layer of continuous stratification beneath the density interface acts as the “fluid” in hydraulic theory, and can transition from subcritical to supercritical flow if the depth of the layer decreases or speed increases sufficiently (Durran 1990). The correspondence of the continuously stratified system with the shallow water system (defining a similar Froude Number) holds when the depth of the lower layer is less than one vertical wavelength (Durran and Klemp 1987). Thus, a hydraulic analog has been used in the
explanation of downslope windstorms. Layering is seen in the lee of the gap in Fig. 4.14, where a layer of strong stability up to about 4 km ASL underlies a region of neutral stability in the wave steepening region. A wave steepening region once again forms above Knobhead (KH), but lower, centered at about 3 km ASL. The layer of “fluid” (the layer of strong stability above the surface) thins and accelerates in the lee of KH, corresponding to supercritical flow. This “shooting flow” rapidly adjusts to subcritical conditions downstream in a turbulent hydraulic jump just upstream of the Asgard Range. This concept was suggested by Doran et al. (2002a), but incorrectly interpreted as a katabatic jump. Downstream, the flow separates from the surface, and low-level wave breaking occurs in the lee of the Asgard Range (Wright Valley) and in the lee of the Olympus Range (Victoria Valley), where strong downslope winds are also found. Low-level wave breaking occurs with slow moving layers near the ground (when the flow is blocked and becomes nonlinear), as vertical shear controls the vertical distribution of wave breaking (Jiang and Smith 2003). The concept of low-level wave breaking over the MDVs is supported by McKendry and Lewthwaite (1990), who observed stagnant flow at about ridge level height above Wright Valley. Wave amplitudes are dampened above the wave breaking regions, as significant energy is dissipated in the wave breaking process.

Figure 4.7b shows that surface flow is weak in the lee (north) of the Royal Society Range, creating what appears to be a wake region that extends north to eastern Taylor Valley. Figure 4.15 shows a cross section of potential temperature and wind speed along the yellow transect in Fig. 4.8a, east of the gap at 1800 UTC 29 December. Flow is clearly blocked by the Royal Society Range, which extends above 2.5 km ASL. Blocked
flow results in a lower effective mountain height, which decreases wave amplitude (Jiang et al. 2005). This leads to low-level wave breaking in the lee, and the layer of strong flow thins to below 2 km ASL. A hydraulic jump occurs in the lee of Cathedral Rocks (CR), and no further wave effects exist downstream. Thus, the eastern MDVs do not directly experience strong gap flow from the south, except in rare instances when the ambient flow turns more easterly (which will be discussed later in the chapter). Similar to the explanation of Pan and Smith (1999), the gap flow south of the western MDVs is differentiated from flow on either side because it avoids Bernoulli loss associated with hydraulic jumps occurring over the lee slopes. The Bernoulli function can be written as

\[
B = c_p T + \frac{v^2}{2} + \Phi
\]

where \( B \) is the Bernoulli function, \( c_p \) is the specific heat at constant pressure, \( T \) is temperature, \( v \) is velocity, and \( \Phi \) is geopotential (e.g., Gill 1982; Pan and Smith 1999; Doyle and Shapiro 2005). This equation expresses the total energy in the form of the internal energy, kinetic energy, and potential energy (the three RHS terms respectively). Air parcels that pass through wave breaking regions or hydraulic jumps experience a decrease in \( B \) due to dissipative processes – a loss of total energy manifested as reduced wind speeds in a downstream wake.

As apparent from Fig. 4.5, the ambient upper-level flow becomes more westerly after 1800 UTC 29 December. Figure 4.16 shows near-surface temperature, sea-level pressure, and wind vectors at 1800 UTC 30 December. A strong cyclone located just offshore of Marie Byrd Land sets up southwesterly flow over the Ross Ice Shelf. Figure 4.17 shows near-surface streamlines and wind speed at 1800 UTC 30 December. Here,
the large-scale flow comes directly from the southwest into the gap, unlike the case at 1800 UTC 29 December, when the ambient flow was more southerly and was blocked. As such, the nondimensional mountain height is lower at 1800 UTC 30 December, at only 0.78. This means that nonlinear effects like flow stagnation and wave breaking will not be as significant as 24 hours earlier. Figure 4.18 shows vertical profiles of wind speed and wind direction at the red cross section origin at 1800 UTC 30 December. Westerly flow of almost 25 m s\(^{-1}\) exists above the surface, which decreases to 5 m s\(^{-1}\) at 2 km ASL. Speeds then recover to about 15 m s\(^{-1}\) at 5 km ASL, while flow direction remains a general southwesterly with height. Differences from the previous 24 hours are increased low-level wind speed and a more westerly wind direction. Figure 4.19 shows a cross section of potential temperature and wind speed along the red cross section at 1800 UTC 30 December. The near-surface flow at the upstream end of the cross section is faster than 24 hours earlier, and the vertical wavelength is greater (the vertical wavelength scales negatively with the nondimensional mountain height, Zängl 2003a). In the southern half of the cross section (up to about 45 km on the x-axis), a shallow layer of increased stability (representing an inversion) exists just above the surface. This appears to decouple the surface flow from aloft, and would behave in a manner similar to shallow-water flow. Accordingly, the wave activity aloft upstream of the Asgard Range is dampened. It is unclear what is occurring over Ferrar Glacier (FG) and Taylor Valley (TV), upstream of the Asgard Range. What appear to be trapped lee waves extend up to about 7 km ASL, where a layer of strong wind speed could be trapping wave energy. Flow stagnates somewhat just upstream of the Asgard Range, as the shallow inversion
layer is now gone, and strong leeside flow occurs beneath a wave breaking region above about 4 km ASL. Low-level wave breaking is present over both Wright Valley and Victoria Valley. Wave activity is greatly amplified in the vertical, downstream of the Asgard Range. The key differences from 24 hours earlier are the stronger ambient flow, less upstream blocking, which means less low-level wave breaking. A hydraulic jump does not form just upstream of the Asgard Range, and with a more westerly incoming component to the upstream flow, the strong winds reach farther down-valley. Observed wind speeds at this time are the strongest of the event at most sites (Fig. 4.6).

Through 31 December, the ambient wind direction turns southerly and then southeasterly, as the cyclone that was located near Roosevelt Island moves westward and then north into the Ross Sea (Fig. 4.5). Figure 4.20 shows near-surface streamlines and wind speed at 0300 1 January 2007. Strong south-southeasterly winds exist across the domain, with enhanced wind speeds in the gap and in the lee of several terrain features. The upstream nondimensional mountain height at 0300 UTC 1 January is 0.46, indicating minimal nonlinear effects. Figure 4.20 shows that the gap flow appears to originate over the Ross Ice Shelf, through Skelton Glacier, and then into the gap. Vertical profiles of wind speed and wind direction at the red cross section origin at 0300 UTC 1 January are shown in Fig. 4.21. Wind speeds increase from about 11 m s\(^{-1}\) at the surface up to about 32 m s\(^{-1}\) at 7 km ASL, while wind direction changes from southerly to SSE. Above 7 km ASL, the wind speed decreases to 6 m s\(^{-1}\) while wind direction shifts to easterly by 8 km ASL. Figure 4.22 shows a cross section of potential temperature and along-transect wind speed along the red transect at 0300 UTC 1 January. Immediately noticeable is the deep
(almost 3 km) layer of strong (> 25 m s⁻¹) wind speeds in the gap. The strongest gap wind speeds occurring when the ambient flow is directed the most oblique to the gap axis points to the importance of the cross-gap pressure difference. The southerly flow near the surface results from the gap wind, as ambient flow becomes more easterly with height. Figure 4.11 shows that the cross-gap pressure difference is largest during the period around 0000 UTC 1 January, as a result of mass accumulation just upstream of the MDVs and an ambient pressure gradient directed across the valleys (Fig. 4.5).

Orographic blocking south of the MDVs induces a southerly flow near the surface, which then turns to easterly with height. The result is a critical layer aloft, where the cross-gap wind speed goes to zero. A critical level is defined as the level where the mean-state flow is equal to the horizontal phase speed of a wave. With the complex terrain dealt with here, there are several distinct wave modes, propagating at different phase speeds, forming a critical layer. A critical layer effectively traps wave energy below, as the negative vertical shear promotes wave breaking. The critical level acts as a “dividing streamline” between the strong winds below and stagnant air in the wave breaking region. (Smith 1985). The critical layer acts in a manner analogous to a free surface in shallow-water hydraulic theory (Durran and Klemp 1987; Bacmeister and Pierrehumbert 1988; see also Chapter 2). The large cross-gap pressure difference drives flow across the gap, with strong leeside winds forced as the stagnant layer aloft descends to about 5 km ASL (Fig. 4.22). Wave breaking occurs above the critical layer, and this “wave-induced critical layer” descends in the lee of the gap. The layer of strong near-surface winds thins along the transect, but remains coherent (no hydraulic jumps), as flow
remains supercritical. A layer of weak stability between 2 and 4 km ASL traps wave energy near the surface over the MDVs themselves, with only weak vertically propagating waves above.

Returning to Fig. 4.22, we see that no hydraulic jumps are forced in the western MDVs by WRF, and that the gap flow remains coherent well downstream. Also, this is one of the rare times when the eastern MDVs directly experience foehn, as flow traverses the orography east of the Royal Society Range and descends into the valleys. This flow pattern is also borne out in the station observations (Fig. 4.6). Explorers Cove and Lake Fryxell shift to a southerly to even south-southeasterly flow from about 1800 UTC 31 December to about 1600 UTC 1 January. Figure 4.23 shows a cross section of potential temperature and along-transect wind speed along the light blue transect in Fig. 4.8a at 0300 UTC 1 January. Flow accelerates through the gap between obstacles into Ferrar Valley, where what appears to be a wave-breaking region forms above 3 km ASL. The flow decelerates into Taylor Valley, with trapped waves above the narrow ridges between valleys. With this flow regime, wind speed variations in eastern Taylor Valley result from wave characteristics over the Kukri Hills, particularly the placement and strength of hydraulic jumps. In Fig. 4.23, lower ridge heights allow for relatively strong flow into Taylor Valley. However, just to the west, over higher portions of the ridge, the flow separates from the surface along the lee slope. Lake Hoare and Lake Bonney experience variable wind directions with low wind speeds, due to hydraulic jumps well upstream. Lake Brownworth winds are also variable during the period, as it appears to be near the dividing streamline between flow through the gap and flow around the Royal Society
Range from the east. Lake Vanda and Lake Vida experience similar wind speeds as earlier in the period, but wind directions become more southerly.

To summarize, the mountain wave patterns across the gap and into the MDVs vary with upstream ambient wind direction (similar to the results of Zängl 2003b for Wipp Valley). When the ambient wind direction is southerly, blocking along the Transantarctic Mountains well upstream of the MDVs leads to lower near-surface wind speeds, greater nonlinearity, and low-level wave breaking. Hydraulic jumps occur upstream of Taylor Valley, leading to lower observed wind speeds at most MDV sites. When the ambient wind direction is more westerly, nonlinearity decreases, meaning less low-level wave breaking and greater vertical propagation of wave activity. The response is stronger wind speeds at most MDVs sites. For an easterly ambient wind, southerly gap flow forcing is strong, and a critical layer forms aloft, which is primarily responsible for wave amplification and moderate foehn winds across the MDVs.

4.2.3 Event Variability: Foehn Propagation to Eastern Valleys and Easterly Intrusions

Often in the MDVs, westerly foehn events are interrupted for brief periods (on the order of a few hours), especially at eastern sites, from easterly intrusions. These easterly intrusions lead to a decrease in temperature, an increase in relative humidity, and usually a decrease in wind speed. On the whole, easterly intrusions shorten foehn events at eastern MDV sites, leading to the cooler average summer temperatures. These intrusions have previously been attributed to sea-breeze effects. A sea-breeze can form over the
MDVs when ground temperatures over land are substantially higher than over adjacent water or snow surfaces, leading to thermally-generated mesoscale pressure perturbations. Since easterly intrusions also occur during winter (see Speirs et al. 2010 for an example), the lack of solar radiation requires a different explanation. Here we present evidence of easterly intrusions formed by the synoptic-scale flow and blocking effects of Ross Island and the Antarctic continent in general.

During periods of strong southerly flow along the western Ross Ice Shelf, Ross Island exerts considerable control on the low-level flow in the surrounding region, including McMurdo and Scott Bases (O’Connor and Bromwich 1988; Seefeldt et al. 2003). This concept has not been considered before in relation to flow in the MDVs. LTER AWS observations at Explorer’s Cove, Lake Fryxell, and Lake Brownworth show easterly intrusions around 0900 UTC 30 December 2006, when foehn conditions are occurring in other portions of the MDVs. Polar WRF has an easterly intrusion to Explorer’s Cove about this time as well. Figure 4.24a shows easterly winds exceeding 8 m s$^{-1}$ in McMurdo Sound, east of the MDVs and west of McMurdo. This easterly flow results from southerly flow upstream of Ross Island that is blocked and is forced around. A mesoscale high-pressure region forms in Windless Bight, the northerly indentation between Mt. Erebus and Mt. Terror, as shown in Fig. 4.24b. The flow around Ross Island is largely driven by the mesoscale pressure gradients. As this easterly flow approaches the Antarctic coast, it is also blocked, and surface pressure increases along the coast (Fig. 4.24b). This appears to be responsible for the cessation of the easterly flow as it approaches the Antarctic coast in Fig. 4.24a. While this cessation prevents the
easterly flow from penetrating into the MDVs, the localized pressure perturbations caused by the easterly flow do affect the down-valley extent of foehn winds.

Figure 4.24c,d provide a closer look at the near-surface streamlines and sea-level pressure at 0900 UTC 30 December. Down-valley winds cease near the coast, associated with a convergence zone along the MDVs shoreline between the westerly foehn and easterly wind around Ross Island (Fig. 4.24a). A localized high-pressure region forms just east of Explorer’s Cove (Fig. 4.24d), associated with the convergence zone. This convergence zone was also identified by Fountain et al. (2009), associated with a precipitation maximum. Also seen is a high pressure region in western Taylor Valley, associated with the mountain waves from the south, and flow being blocked upstream of the Asgard Range. The flow within Taylor Valley itself will be dictated by the pressure gradient force at a particular point, which has been described as “pressure-driven channeling” (Whiteman and Doran 1993). Winds are driven by the component of the pressure gradient along the valley axis, and the force balance is best described as antitriptic, where the pressure gradient force is primarily balanced by friction, as winds are constrained along the valley walls. Additional warming may result from turbulent mixing beneath stably stratified air (Zängl et al. 2004a). Because of the wake zone formed by the Royal Society Range, downward momentum transport from mountain waves would not be of substantial influence in Taylor Valley, as it was for the Tennessee Valley in Whiteman and Doran (1993). Figure 4.25 shows reduced pressure (red), near-surface wind speed (green), near-surface wind direction (gold), and terrain height (black) along the transect shown in Fig. 4.8b. Wind speed increases from a southwesterly
direction just downstream of the high pressure region formed by the mountain waves in the western MDVs. However, pressure then increases again just offshore, associated with the convergence zone previously described. At this point, wind speed decreases sharply as wind direction shifts to easterly, and wind speeds then recover to the east. The key point from Fig. 4.25 is that the wind speeds in the valley are strongly tied to the mesoscale pressure perturbations associated with terrain blocking.

A different situation occurs just twelve hours later at 2100 UTC 30 December. Ambient flow has shifted more westerly, and westerly foehn winds are recorded at all LTER AWS sites. This is reflected in the Polar WRF near-surface streamlines and wind speed (Fig. 4.26a). With an offshore wind, the flow around Ross Island is directed around to the east, with a negligible high pressure region in Windless Bight (Fig. 4.26b). The close-up plots in Fig. 4.26c,d confirm this, with westerly winds through Taylor Valley, and no localized high-pressure offshore as occurred 12 hours earlier. Figure 4.27 shows the reduced pressure, near-surface wind speed, near-surface wind direction, and terrain height along the same transect as in Fig. 4.25, but 12 hours later at 2100 UTC 30 December. Now there is a single high-pressure perturbation along the transect, at the western end associated with blocking upstream of the Asgard Range, with a steady but small decrease to the east. Wind speed again increases just downstream of the high pressure region, then gradually decreases to the east. Wind direction retains a southwesterly component along most of the transect, with deviations mainly caused by the varying orientation of the valley along the transect.
The preceding analysis shows the influence that dynamic pressure perturbations have on flow in the MDVs. An easterly intrusion into Taylor Valley results from flow around Ross Island and blocking of this flow by the Antarctic continent. The easterly intrusion occurred at about 10 PM local time only a few days after the summer solstice, so that thermally-generated effects cannot be ruled out. However, this easterly intrusion extends well into the night, when easterly intrusions forced by a sea breeze would be expected to die off. As Fig. 4.6 shows at Explorer’s Cove, during periods of strong dynamical forcing, there is no clear diurnal cycle of easterly intrusions. Thus for strongly forced events, sea breeze effects are negligible. This agrees with McKendry and Lewthwaite (1992), who find that the synoptic-scale circulation determines the evolution of the sea-breeze front, rather than the local thermal forcing. Easterly intrusions are primarily forced by flow deflection around Ross Island, which requires a southerly wind component upstream. The large-scale flow direction once again plays a large role in determining the localized flow response in the MDVs. While the example here focuses on Taylor Valley, similar arguments can be made for Wright Valley. Polar WRF shows a high pressure region in western Wright Valley, formed by mountain waves from the south and flow being blocked by the Olympus Range. High pressure along the Antarctic coast provides an opposing PGF during periods of blocking. However, the pressure-driven channeling is complicated by the Wilson Piedmont Glacier along the coast and the mountain wave activity from the south within Wright Valley. Doran et al. (2002a) note events where strong winds occur at Lake Brownworth with calm conditions at Lake Vanda. They attributed this behavior to a rather vague explanation of foehn flow “riding
over” Lake Vanda, possibly associated with cold-air pooling preventing foehn breakthrough. Such an argument is not valid during summer, and instead the weak flow at Lake Vanda results from an upstream hydraulic jump, with pressure-driven channeling bringing strong winds downvalley to Lake Brownworth.

While sea-breeze effects are deemed negligible for strongly forced events, the situation changes for weakly forced events with a more westerly ambient flow. Like the situation shown at 2100 UTC 30 December, easterly flow around Ross Island is nonexistent without a southerly ambient wind. With a weak high pressure perturbation in the western MDVs, thermally-generated pressure perturbations east of the MDVs would provide a legitimate opposing PGF.

4.2.4 Event Variability: Temperature Variation

From Fig. 4.6, it is clear that the temperature at all of the LTER AWS sites increases during the case study period, independent of the diurnal cycle. Maximum observed temperatures occur early UTC 2 January, which is during the peak of daytime heating. However, the warmer temperatures at this time compared to a day earlier rule out increased amplitude of mountain waves (stronger foehn effect) as being responsible for warming throughout the period, as ambient forcing and mountain waves both decrease in strength. Another explanation for the warming is simply advection – that warmer air is being transported to the MDVs. With the substantial terrain effects upstream of the MDVs, and the foehn effect itself, it is difficult from synoptic charts alone to decipher source regions for the air arriving at the surface of the MDVs.
Therefore, trajectories have been calculated for air arriving at different levels just upstream of the MDVs. Since we have to use the outer 32 km domain for these trajectories, the focus is only on source regions for the foehn air, and not on the magnitude of foehn descent.

Trajectories are run backwards from a point just upstream of the MDVs at 3 km, 4 km, and 5 km ASL for 84 hours, and shown in Fig. 4.28a ending at 1800 UTC 29 December. All trajectories originate over the central Ross Sea and traverse the Ross Ice Shelf before curling into the MDVs, associated with the weak cyclone in the area on the 29th. Figure 4.28b shows time series of variables along the 5 km trajectory. Height is shown in the bottom in black, and this trajectory rises from just over 3 km at the beginning. Potential temperature (top, in red) starts at about 287 K and drops to about 284 K at the terminus, likely from mixing in of colder air. Equivalent potential temperature (in blue) is slightly warmer at the beginning, around 290 K, associated with the maritime environment over the Ross Sea. Equivalent potential temperature approaches the actual potential temperature over the western Ross Ice Shelf, as precipitable water (in green) decreases associated with precipitation. The trajectory at 1800 UTC 30 December (Fig. 4.29a) originates from West Antarctica at lower levels, but from East Antarctica at 5 km. The 5 km trajectory descends from 7 km to 5 km, with very little change in potential temperature, equivalent potential temperature, or precipitable water. The potential temperature of 288 K at the terminus is slightly warmer than 24 hours earlier, likely associated with the large-scale descent over East Antarctica. A different situation occurs at 0600 UTC 1 January, for which trajectories are shown in
Fig. 4.30a. The 3 km trajectory originates in the Amundsen Sea, but the 4 km and 5 km trajectories originate north of the Ross Sea. All trajectories arrive upstream of the MDVs from almost due east. Figure 4.30b shows that the 5 km trajectory rises from about 1.1 km at the beginning of the cross section, and is very warm and moist, based on equivalent potential temperature being over 300 K (almost 15 K warmer than the potential temperature), and precipitable water values of almost 18 mm. Unlike the previously discussed trajectories, here the potential temperature increases along the transect. This occurs when the trajectory traverses Marie Byrd Land, and rises moist adiabatically, as evidenced by the near-constant equivalent potential temperature and decreasing precipitable water along the second half of the trajectory. At the terminus, the potential temperature is almost 294 K, and substantially warmer than previous times. A similar pattern continues into 2 January (not shown). Therefore, the warmest foehn conditions are associated with the advection of warm, moist maritime air into the MDVs region aloft, which gains enthalpy as it ascends moist adiabatically and then descends dry adiabatically as foehn into the MDVs.

4.2.5 Rotors

While this last section does not directly deal with the foehn mechanism, Polar WRF output identifies an important mountain meteorology feature in the MDVs. Atmospheric rotors are associated with large amplitude mountain waves and most often occur in regions of boundary layer separation beneath trapped mountain waves (Doyle and Durran 2002, 2004). The adverse pressure gradient beneath a ridge of a trapped
mountain wave and surface friction both contribute to boundary layer separation and rotor formation. A thin layer of horizontal vorticity along the lee slope generated by surface friction is lifted aloft in the boundary layer separation process and into the rotor (Doyle and Durran 2007). It is this shear vorticity that is responsible for the dangerous turbulence, as rotors have been implicated in several aviation disasters (Darby and Poulos 2006).

While there are no direct observations of rotors over or near the MDVs, perusal of Polar WRF output identifies the presence of rotors during the December 2006 / January 2007 case study. Figure 4.31a shows a cross section of wind speed along the cross section, circulation vectors, and potential temperature along the green transect segment in Fig. 4.8a (a portion of the longer transect used in cross sections earlier in the chapter) in western Taylor Valley near the entrance to Ferrar Glacier at 1200 UTC 31 December 2006. Trapped lee waves form downstream of the gap at this time, while an area of reversed (approximately northerly) flow exceeding 6 m s⁻¹ is found just above the surface over the span of a few kilometers under the wave ridge, centered at about 8 km on the X-axis. Figure 4.31b provides a closer look at the rotor, showing the reversed flow on the lowest two model levels. A clockwise rotation of the circulation vectors in the plane of the cross section can be seen as identification of the rotor. This is a Type 1 rotor as categorized by Hertenstein and Kuettner (2005), which form beneath trapped lee waves (as opposed to rare Type 2 rotors, which form aloft associated with hydraulic jumps). These authors identify an upstream near-mountain-top inversion and vertical shear within that inversion as being important for rotor formation, and both of these components are
met here. As rotors typically form beneath trapped lee waves, Scorer parameter layering is necessary for such waves to form (see Chapter 2). Trapped waves seem favored for the southwest Taylor Valley region, as waves launched from the narrow-width terrain features just south of Taylor Valley (i.e., Knobhead) are allowed to propagate aloft (i.e., no low-level wave breaking), and waves are reflected from a Scorer parameter interface, associated with high wind speeds with the propagating waves aloft. A second rotor was identified in the 2-km grid spacing output at 0400 UTC 30 December, but not in the 500-m grid spacing output. Note that the 2-km output has a lower lowest model level, so that it may actually better resolve rotors than the 500-m output. Because of the danger that rotors pose towards aviation (i.e., helicopters that frequently fly into the MDVs), rotors in the MDVs region certainly warrant further study.
4.3 Figures

Figure 4.1. AMPS SLP and near-surface wind vector 2 year composites (2006 and 2007) for (a) 688 non-foehn cases and (b) 688 foehn cases, (c) SLP difference (foehn – non-foehn) and (d) pressure gradient difference (foehn-non-foehn). In (c) and (d) stipling is for positive differences, hatching is for negative differences. Light stipling/hatching refers to 90% confidence level, heavy stipling/hatching refers to 95% confidence level. Star denotes location of MDVs and arrow in (d) shows the direction of the pressure gradient.
Figure 4.2. Same as Fig. 4.1 except for summer (DJF, 96 cases).
Figure 4.3. Same as Fig. 4.1 except for winter (JJA, 298 cases).
Figure 4.4. AMPS SLP and near-surface wind vector annual composites (a) 2006 and (b) 2007, (c) SLP difference (2007 – 2006) and (d) pressure gradient difference (2007 – 2006). In (c) and (d) stipling is for positive differences, hatching is for negative differences. Light stipling/hatching refers to 90% confidence level, heavy stipling/hatching refers to 95% confidence level. Star denotes location of MDVs and arrow in (d) shows the direction of the pressure gradient.
Figure 4.5. Synoptic overview plots of (left) sea-level pressure (hPa, contours), 3 m temperature (°C, color shading), and 3 m wind vectors (arrows) and (right) 500 hPa geopotential height (m, contours), relative vorticity (10^-5 s^-1, color shading) and wind vectors (arrows) for the labeled times. Sea-level pressure not plotted over areas exceeding 500 m elevation.
Figure 4.5: Continued
Figure 4.6. Time series of 3 m temperature (°C, purple), relative humidity (%), red, wind direction (degrees, orange), and wind speed (m s⁻¹, green) from LTER observations (dark colors) and representative location in Polar WRF (light colors) over the case study period. Note scaling differences between stations.

Continued
Figure 4.6: Continued
Figure 4.7. Polar WRF 3 m wind speed (m s\(^{-1}\), color shading) and streamlines at (a) 0000 UTC 29 December 2006 and (b) 1800 UTC 29 December 2006.
Figure 4.8. Orientation maps for 500-m grid spacing simulations. Terrain height (m) shaded, in (a) white transect represents gap profile, magenta transect represents cross-gap pressure difference calculations, red transect represents cross sections in Figs. 4.14, 4.19, and 4.22, yellow transect represents cross section in Fig. 4.15, light blue transect represents cross section in Fig. 4.23, green transect represents cross section in Fig. 4.31, white averaging box used for near-surface gap wind speeds, blue averaging box used for inverse Froude number calculations, and orange dots represent LTER AWS sites. In (b), red transect is used in along-gap reduced pressure calculations, and green dots represent LTER AWS sites.
Figure 4.9. Profile slice of the gap south of the McMurdo Dry Valleys along the white transect in Fig. 4.8a. “EAIS” is East Antarctic Ice Sheet, “MF” is Mount Feather (2985 m), “Gap” represents the approximate width of the gap, of which the lowest point and vertical profile varies with latitude, and “RSR” is the Royal Society Range.
Figure 4.10. Polar WRF domain 2 sea-level pressure (hPa, contours), 3 m temperature (°C, color shading), and wind vectors (arrows) at 1800 UTC 29 December 2006. Sea-level pressure not plotted over areas exceeding 500 m elevation.
Figure 4.11. Polar WRF reduced pressure difference across gap along the *magenta* transect in Fig. 4.8a (red, hPa), averaged 3 m wind speed in the gap (m s⁻¹, blue) using the *white* box in Fig. 4.8a, 600 hPa upstream wind direction (degrees, green), and 600 hPa upstream wind speed (m s⁻¹, gold).
Figure 4.12. Polar WRF level 10 wind speed (m s$^{-1}$, color shading) and streamlines at (a) 1800 UTC 29 December 2006.
Figure 4.13. Vertical profiles of wind speed (m s$^{-1}$, top) and wind direction (degrees, bottom) at origin of the red cross section in Fig. 4.8a at 1800 UTC 29 December 2006.
Figure 4.14. Vertical cross section of potential temperature (K, contours), wind speed along the *red* cross section in Fig. 4.8a (m s⁻¹, color shaded), and circulation vectors at 1800 UTC 29 December 2006. “Gap” is the elevated gap, “KH” is Knobhead”, “FG” is Ferrar Glacier, “TV” is Taylor Valley, “AR” is Asgard Range, “WV” is Wright Valley, “OR” is Olympus Range, and “VV” is Victoria Valley.
Figure 4.15. Vertical cross section of potential temperature (K, contours), wind speed along the yellow cross section in Fig. 4.8a (m s$^{-1}$, color shaded), and circulation vectors at 1800 UTC 29 December 2006. “RSR” is the Royal Society Range, “CR” is Cathedral Rocks, “FG” is Ferrar Glacier, “KH” is Kukri Hills, “TV” is Taylor Valley, and “AR” is Asgard Range.
Figure 4.16. Polar WRF domain 2 sea-level pressure (hPa, contours), 3 m temperature (°C, color shading), and wind vectors (arrows) at 1800 UTC 30 December 2006. Sea-level pressure not plotted over areas exceeding 500 m elevation.
Figure 4.17. Polar WRF 3 m wind speed (m s\(^{-1}\), color shading) and streamlines at (a) 1800 UTC 30 December 2006.
Figure 4.18. Vertical profiles of wind speed (m s$^{-1}$, top) and wind direction (degrees, bottom) at origin of the red cross section in Fig. 4.8a at 1800 UTC 30 December 2006.
Figure 4.19. Vertical cross section of potential temperature (K, contours), wind speed along the red cross section in Fig. 4.8a (m s\(^{-1}\), color shaded), and circulation vectors at 1800 UTC 30 December 2006. “Gap” is the elevated gap, “KH” is Knobhead”, “FG” is Ferrar Glacier, “TV” is Taylor Valley, “AR” is Asgard Range, “WV” is Wright Valley, “OR” is Olympus Range, and “VV” is Victoria Valley.
Figure 4.20. Polar WRF 3 m wind speed (m s$^{-1}$, color shading) and streamlines at (a) 0300 UTC 01 January 2007.
Figure 4.21. Vertical profiles of wind speed (m s\(^{-1}\), top) and wind direction (degrees, bottom) at origin of the red cross section in Fig. 4.8a at 0300 UTC 1 January 2007.
Figure 4.22. Vertical cross section of potential temperature (K, contours), wind speed along the red cross section in Fig. 4.8a (m s\(^{-1}\), color shaded), and circulation vectors at 0300 UTC 1 January 2007. “Gap” is the elevated gap, “KH” is Knobhead”, “FG” is Ferrar Glacier, “TV” is Taylor Valley, “AR” is Asgard Range, “WV” is Wright Valley, “OR” is Olympus Range, and “VV” is Victoria Valley.
Figure 4.23. Vertical cross section of potential temperature (K, contours), wind speed along the light blue cross section in Fig. 4.8a (m s$^{-1}$, color shaded), and circulation vectors at 0300 UTC 1 January 2007. “FG” is Ferrar Glacier, “KH” is Kukri Hills, and “TV” is Taylor Valley.
Figure 4.24. (a) Polar WRF near-surface wind speed (m s$^{-1}$, color shaded) and streamlines, (b) Polar WRF sea-level pressure (hPa, color shaded) at 0900 UTC 30 December. (c) and (d) same as (a) and (b), respectively, but for zoomed in area of MDVs. Large dots in (c) and (d) for Explorer’s Cove.
Figure 4.25. Polar WRF reduced pressure (hPa, red), near-surface wind speed (m s$^{-1}$, green), near-surface wind direction (degrees, gold), and terrain height (m, black) along the transect shown in Fig. 4.8b at 0900 UTC 30 December.
Figure 4.26. (a) Polar WRF near-surface wind speed (m s\(^{-1}\), color shaded) and streamlines, (b) Polar WRF sea-level pressure (hPa, color shaded) at 2100 UTC 30 December. (c) and (d) same as (a) and (b), respectively, but for zoomed in area of MDVs. Large dots in (c) and (d) for Explorer’s Cove.
Figure 4.27. Polar WRF reduced pressure (hPa, red), near-surface wind speed (m s\(^{-1}\), green), near-surface wind direction (degrees, gold), and terrain height (m, black) along the transect shown in Fig. 4.8b at 2100 UTC 30 December.
Figure 4.28. (a) Backward trajectories run from 1800 UTC 29 December 2006. Trajectories run at 3 km, 4 km, and 5 km ASL for 84 hours, arrows plotted every 24 hours. (b) Potential temperature (K, red), equivalent potential temperature (K, blue), precipitable water (mm, green), and height (m, black) along the 5 km trajectory.
Figure 4.29. (a) Backward trajectories run from 1800 UTC 30 December 2006. Trajectories run at 3 km, 4 km, and 5 km ASL for 84 hours, arrows plotted every 24 hours. (b) Potential temperature (K, red), equivalent potential temperature (K, blue), precipitable water (mm, green), and height (m, black) along the 5 km trajectory.
Figure 4.30. (a) Backward trajectories run from 0600 UTC 1 January 2007. Trajectories run at 3 km, 4 km, and 5 km ASL for 84 hours, arrows plotted every 24 hours. (b) Potential temperature (K, red), equivalent potential temperature (K, blue), precipitable water (mm, green), and height (m, black) along the 5 km trajectory.
Figure 4.31. (a) Vertical cross section of potential temperature (K, contours), wind speed along the green transect in Fig. 4.8a (m s$^{-1}$, color shaded), and circulation vectors at 1200 UTC 31 December 2006. (b) Same as (a) except zoomed in to rotor.
Chapter 5: Robustness of Foehn Features and Large-Scale Variability

Climate variability is apparent in the McMurdo Dry Valleys (MDVs) both temporally, as evidenced by anomalously warm summers leading to increased melt that can replenish decades of decreased lake levels (Doran et al. 2008), and spatially from along-valley wind regime variations. This variability has enormous impact upon biological activity in the MDVs. Inferences have been made that large-scale Southern Hemisphere climate variability (ENSO) is largely responsible for interannual differences in melt (Bertler et al. 2004, 2006).

In this chapter, we explore MDVs climate variability using output from 15 summers of Polar WRF simulations, LTER AWS observations, and ERA-Interim reanalysis output. A large focus is placed upon foehn winds, since the warming associated with foehn events is often responsible for above-freezing temperatures in the MDVs (Speirs et al. 2011, submitted to *Int. J. Climatol.*). A foehn identification scheme, based on that used in Speirs et al. (2010), is used with both LTER and WRF output, and this scheme is explained in the following section. Differences between LTER and WRF foehn event distributions are then presented. The foehn identification scheme is used as a validation tool of the WRF output, while the discrepancies between the LTER and WRF distributions are characterized to establish the causes of the differences.
Once some measures of model validation have been presented, a representative set of foehn events is used to explore climate variability in the MDVs. First we show that several of the foehn features described for the December 2006 / January 2007 case study event in Chapter 4 are robust across this set of cases. Then the interannual and intraseasonal aspects of foehn variability are explored. While SAM and ENSO effects are in play on interannual timescales, we show that MJO forcing on intraseasonal timescales is important for extended periods of foehn activity that ultimately lead to individual warm seasons in the MDVs.

5.1 Foehn Identification

The foehn identification scheme used in this study was developed by Johanna Speirs, and was used in Speirs et al. (2010). Foehn events in LTER AWS observations are identified using the same criteria as Speirs et al. (2010). Using observations at 15-minute frequency, all of the following must occur over any given 6 hour period:

- Temperature: Increase of 1°C over a 1 hour period OR any reading above 0°C
- Relative Humidity: Decrease of 5% over a 1 hour period OR any reading below 30%
- Wind Speed: Greater than 5 m s\(^{-1}\) for at least 80% of the observations
- Wind Direction: Greater than 180° and less than 315° (360° for Lake Vida) for at least 80% of the observations.

Note that the wind speed observation used is the 15 minute maximum, and all times are defined using local time (NZDT, which is UTC+13 hours).
For Polar WRF output, which is also at 15 minute frequency the criteria were adjusted slightly to account for the different characteristics of model output compared to observations:

- Temperature: Increase of 1°C over a 2.5 hour period OR any reading above 0°C
- Relative Humidity: Decrease of 5% over a 2.5 hour period OR any reading below 30%
- Wind Speed: Greater than 3.5 m s⁻¹ for at least 80% of the observations
- Wind Direction: Greater than 180° and less than 315° (360° for Lake Vida) for at least 80% of the observations.

The longer time period for the temperature increases and relative humidity decreases allows for slower developing foehn onset, which might result from the smoothed terrain in the model. The decreased wind speed threshold is an attempt to compensate for the use of maximum wind speed (gust) in the observations, for which there is no corresponding variable in WRF.

For both LTER and WRF, each foehn event is placed into a six-hour bin, centered on 0, 6, 12, and 18 NZDT. Only one event in the six-hour bin is required for declaration of a “foehn event” for that six-hour block, and the number of foehn instances in each six-hour bin is not recorded. Note that this differs slightly from the “foehn day” entity used in Speirs et al. (2010), which bins over daily intervals. For each six-hour foehn event, the following variables are recorded: maximum temperature (TEMPMAX), average temperature (TEMPAVG), maximum amount of warming over the threshold (1 or 2.5 hours) period (WARMING), degree days above freezing (DDAF), minimum relative
humidity (RHMIN), average relative humidity (RHAVG), maximum wind speed (WSMAX), and average wind speed (WSAVG).

5.2 Foehn Validation

In this section we present several metrics pertaining to the performance of Polar WRF in identifying foehn events in the surface observations. As mentioned in the previous section, several accommodations are made for Polar WRF in the foehn identification scheme. Similarly, accommodations are made in the validation methods to allow for a fair comparison. Validation statistics are recorded over fifteen summer seasons (1994/95 to 2008/09).

First, we compare distributions of foehn events at two or more sites, using all valley sites and Taylor Glacier. This is similar to the “foehn day” criteria used in Speirs et al. (2010), where all statistics are based on foehn days being recorded at three or more stations. This threshold is meant to obtain a set of “significant” foehn events. During the validation period, 740 matches are found (both Polar WRF and LTER observations record events), 239 events are found only in LTER observations, and 1034 events are found only in Polar WRF. So 24% of observed foehn events are not in Polar WRF, and 58% of Polar WRF cases are not in the observations. While Polar WRF identifies a clear majority of the real foehn events, it also records far too many events that are not real.

Next, we change the validation parameters to only utilize a set of four “core” sites – Taylor Glacier, Lake Bonney, Lake Hoare, and Lake Vanda, and only require a foehn event to be recorded at one of these sites. These four sites have a similar foehn
climatology (see Fig. 8 in Speirs et al. 2010), so this pooling method is an aim to eliminate poorly resolved mesoscale and subgrid-scale effects that presumably have a large impact on foehn conditions in the MDVs. Lowering the event count threshold to one attempts to account for any systematic discrepancies (underestimates) of foehn event identification by Polar WRF at these sites with a high frequency of foehn occurrence. There are 1383 matches, 418 LTER-only events, and 1441 WRF-only events. The result is 23% of observed cases not found in WRF, and 51% of WRF cases not in the observations. This metric provides a more favorable comparison for Polar WRF, but still illustrates that Polar WRF has too many foehn events across the MDVs.

A validation analysis is now done across individual observing sites. While this is a stricter comparison than the pooling methods above, it does allow for some explanation of why WRF misses some observed events and also why there are too many events in Polar WRF. Figure 5.1 shows the interannual variability of foehn events for each individual observing site. Once again it is clear that Polar WRF has too many foehn events across all sites. Overall, Polar WRF captures a good amount of interannual variability in the observed foehn event counts, especially at Taylor Glacier, Lake Fryxell, Lake Vanda, Lake Brownworth, and Lake Vida. Noticeable discrepancies occur primarily in 2005, 2007, and at Lake Hoare. Figure 5.2 shows the spatial variability of foehn events along west-east transects (TG to EC, LVa to LBr, and LVi) for each year. The decrease of foehn events from west to east is well-captured by Polar WRF, although Polar WRF has proportionally too many events at Lake Hoare (where 2003, 2004, and 2007 actually have more events at Lake Hoare than up-valley at Lake Bonney).
For the foehn events that match between LTER and Polar WRF, we explore the differences in conditions between the two. Figure 5.3 shows the annually averaged differences (Polar WRF minus LTER) for several variables across sites. Polar WRF is too cold during foehn events across most sites except Lake Fryxell, but especially so at western sites. The positive temperature change is generally smaller in Polar WRF as well, except for Taylor Glacier, which likely results from the frozen surface. The degree days above freezing (DDAF) is also lower in Polar WRF, corresponding to the cold bias at most sites. Relative humidity is too low in Polar WRF across the MDVs, but worst at Lake Hoare and Lake Fryxell. A possible reason for the negative bias overall, and the especially poor performance at Lake Hoare and Lake Fryxell, is the method of horizontal diffusion used in these model simulations. As mentioned in Chapter 3, horizontal diffusion on model levels was used, which has a tendency to diffuse moisture up the valley slopes. The relatively narrow valley positions of Lake Hoare and Lake Fryxell, along with too frequent dry foehn conditions, likely lead to worse performance at these two sites. In contrast, Lake Vida receives less of a sea-breeze contribution, and is in a more exposed location in Victoria Valley. Wind speed is too high across the MDVs during foehn events, especially so at western sites and Lake Vida.

Finally, we look to why Polar WRF misses some observed events, and why it generally has too many events. Figure 5.4a shows the total number of times each criterion is not found in LTER observations for events that occur only in Polar WRF. This number of additional instances of a particular criterion would be necessary for the number of foehn events in the observations to match the number of events in Polar WRF.
Across all sites, wind speed and wind direction are the primary culprits. Therefore, Polar WRF is generating “phantom” foehn events, with strong winds out of a westerly direction, that are not found in observations. Temperature and relative humidity criteria are less of a problem, implying that there may be some record of a foehn event in many of the missed events (warming and/or drying), but the winds remain incorrect. For a more detailed look at why there are too many events in WRF, Table 5.1 shows the average number of extra Polar WRF events per year at each site in the left-most column. Several types of missed events, shown in the next three columns, are then subtracted from this total. However, events overlap between the categories, which is taken into account for the final result in column “WRF NEW.” “WRF WEAK” represents observations that suggest a foehn event, but are too weak to meet the identification criteria. To identify these weaker events, all of the criteria thresholds for observations are halved (except for wind direction), and the number of LTER periods that switch from non-foehn to foehn with this weaker set of observed criteria is recorded. Visual inspection of LTER and Polar WRF time series shows that Polar WRF is either premature on event initiation, or extends events too long. The values in column “WRF EXTEND” are obtained when an observed foehn event occurs within 24 hours before or after an event that only occurs in Polar WRF. Finally, if there is an observed event at another station at the same time that only Polar WRF identifies an event, then “WRF OTHER” is iterated. When we subtract these three types of missed events, the new quantity of missed events is shown in column “WRF NEW”. At all stations, there is a substantial reduction in the number of missed events. The point is that Polar WRF is not “making up” foehn events, but rather the
discrepancies from LTER observations are functions of the criteria thresholds. As shown in Table 5.1, the foehn events only in Polar WRF are also present in the observations, but foehn events in Polar WRF are either stronger (WRF WEAK), start early or last too long (WRF EXTEND), or are too widespread (WRF OTHER).

Figure 5.4b shows the number of missed criteria for observed events that Polar WRF does not identify. At more western sites (Taylor Glacier, Lake Bonney, Lake Hoare, Lake Vanda, and Lake Vida), temperature and wind speed have the largest counts. There are smaller totals at the eastern sites, where wind speed and wind direction have the largest totals. Table 5.2 shows the average number of missed events per year in the “LTER ALL” column. Due to the cold bias in Polar WRF temperature and DDAF during foehn events (Fig. 5.3), the 0°C temperature threshold for foehn events may be the culprit. The “LTER ABOVE 0” column shows the number of LTER-only events where the temperature criterion is met through the 0°C threshold. To analyze the wind speed criterion, the “LTER WEAK WIND” column shows the LTER events where the only missing criteria in Polar WRF is wind speed. This appears to happen due to misplaced features like hydraulic jumps or flow blocking in the model. When these events are removed, the new average number of events that Polar WRF misses per year is shown in the “LTER NEW” column. Thus, many of the observed events that Polar WRF misses are due to general model biases in this region that are not related to the model’s ability to identify foehn forcing.

In summary, Polar WRF has too many foehn events compared to LTER observations, and the number of extra Polar WRF events exceeds the number of events
that the model misses. Polar WRF generally represents the interannual and spatial
variability of foehn events in the MDVs, but the model is too cold, too dry, and too windy
during foehn events. Most of the extra foehn events in Polar WRF result from the
positive wind speed bias and foehn conditions lasting for too long and being too
widespread. Observed events that WRF misses result from systematic model biases.

5.3 Foehn Features Across Events

In Chapter 4, prominent meteorological features of a foehn case study were
illustrated. Here, we show the robustness of some of these features across many foehn
events. The foehn criteria described in section 5.1 are used on LTER AWS data to
produce a set of foehn events. This set corresponds to “LTER ALL” of the previous
section. Hence, we are using Polar WRF output corresponding to observed events. Even
though there are some foehn events in observations not identified in the model, the
analysis in the last section showed that Polar WRF is primarily missing the local response
at specific sites, and not the large-scale setup leading to foehn. So we feel confident in
using the model to represent conditions during observed foehn events.

5.3.1 Gap Wind

In section 4.2.1, a gap flow was implicated as the mechanism to bring southerly
flow and to force mountain waves over the western MDVs. The case study showed that
gap wind forcing is set up by flow blocking south of the Royal Society Range, and that it
plays a larger role in events featuring an ambient easterly wind than events with offshore
flow more from the west. To see if this concept is robust across other foehn cases, a scatter plot of the reduced pressure difference across the gap against wind speed within the gap is presented in Fig. 5.5. The cross-gap transect for the pressure difference and the averaging box for the wind speed (both white) are shown in Fig. 5.6. In Fig. 5.5, the data are color coded by 600 hPa wind direction and wind speed upstream of the MDVs. Purple points represent wind speeds below 5 m s⁻¹, red points represent wind speeds larger than 5 m s⁻¹ and wind direction less than 203° (defined to be the gap axis in Chapter 4), and green points represent wind speeds above 5 m s⁻¹ and wind direction greater than 203°. The purple points, representing weak ambient forcing, are clustered in the lower left corner of the plot, with weak gap wind speeds (generally less than 10 m s⁻¹) and weak gap forcing (cross-gap reduced pressure differences generally less than 2 hPa). Understandably, as the gap wind speed increases, more red (southeasterly forcing) and green (southwesterly forcing) points are found. The red points feature a stronger relationship between gap pressure difference and gap wind speed. As cross-gap pressure difference increases, the gap wind speed increases. This relationship is weaker for the green points, where cross-gap pressure differences remain relatively weak (less than 3 hPa), even for strong gap wind speeds approaching 20 m s⁻¹.

These results confirm the notion in section 4.2.1 that gap forcing is more important for ambient easterly forcing. Flow blocking south of the Royal Society Range and the ambient synoptic-scale pressure gradient are primarily responsible for the cross-gap pressure difference, which drives the gap flow into the western MDVs, even when the ambient wind direction is oblique to the gap axis. As ambient winds become more
westerly, flow blocking is greatly reduced because flow is offshore and originates at a higher elevation over the East Antarctic ice sheet. In these cases a cross-gap pressure difference is not necessary for the gap flow. Still, the weak relationship seen between cross-gap pressure difference and gap wind speed for the green points in Fig. 5.5 reflects blocking of westerly flow by the Royal Society Range, which still sets up a weak cross-gap pressure difference.

5.3.2 Ambient Wind Direction Dependence

For the December 2006 / January 2007 case study presented in Chapter 4, the foehn characteristics changed with the direction and speed of the ambient wind upstream of the MDVs. The response of mountain waves over the gap south of the MDVs depended on the nondimensional mountain height, also known as the inverse Froude Number. Upstream flow from a generally southerly direction was blocked well upstream of the MDVs region, resulting in lower wind speeds and a larger nondimensional mountain height. This lead to low-level wave breaking and hydraulic jumps upstream of Taylor Valley. When the ambient wind direction shifted more westerly, the nondimensional mountain height decreased, and wave breaking was not as pronounced. Observed wind speeds at LTER AWS sites were larger without hydraulic jumps occurring upstream.

To see if this relationship between upstream wind direction and the nondimensional mountain height holds for more cases, a scatter plot of upstream wind direction versus nondimensional mountain height is plotted in Fig. 5.7. Recall that a
theoretical nondimensional mountain height threshold of 1.1 separates linear flow at lower values from nonlinear flow with wave breaking at higher values. Note that flow direction is not taken into account for the calculation of wind speed for nondimensional mountain height. A large variance in nondimensional mountain height values is seen for more easterly wind directions. There is some difficulty in defining the upstream nondimensional mountain height for easterly ambient flow because flow is blocked yet a strong cross-gap pressure difference leads to acceleration into the gap itself. Thus, for these wind directions, the cross-gap pressure difference might be a better indicator of flow response over the gap, as discussed in the previous subsection.

For southerly flow (approximately 180° to 205°), nondimensional mountain heights generally remain above 1.1, indicating nonlinear flow and wave breaking, as was found for the case study in the previous chapter. However, with wind directions greater than about 205°, nondimensional mountain heights decrease, with now most values under 1.1. This again agrees with the case study from last chapter, where a westerly offshore wind is not blocked and results in vertically propagating amplified mountain waves downstream of the gap. Nondimensional mountain heights increase past 270° for the few weak events that have a northerly component.

In the case study, observed wind speeds in the MDVs were larger when the ambient flow was southwesterly. Figure 5.8 shows scatterplots of upstream ambient wind direction versus observed wind speed for a sample of stations. For stations in Wright Valley (Lake Vanda and Lake Brownworth), there is a clear relationship between upstream wind direction and wind speed. Station wind speeds are generally larger as the
ambient wind direction turns more westerly. Ambient southwesterly flow does not flow through the gap, but instead comes from the East Antarctic plateau and is channeled down-valley. There are no hydraulic jumps like there are for southerly gap flow, and wind speeds are higher. A similar relationship is seen at Lake Vida in Victoria Valley (not shown). This relationship does not hold for Taylor Valley sites, as illustrated for Lake Bonney and Lake Fryxell in Fig. 5.8. Here, there are more events with easterly ambient flow than for the other valleys. However, the average wind speed increases as flow becomes more westerly. For the strongest events (wind speeds greater than 10 m s\(^{-1}\)), there are ambient wind directions from both southeasterly and southwesterly, but more from the southwest. Also notice a dearth of strong events in the range of 155° to 205°. This likely results from upstream flow blocking along the Transantarctic Mountains, resulting in lower incident wind speeds and a weaker response at sites in the MDVs.

5.3.3 Foehn Maximum Temperature

In Chapter 4, it was found that temperatures in the MDVs were related to advection patterns aloft, and not to differences in mountain wave structure. We expand upon this one case by looking at event maximum temperature based on ambient wind direction aloft, as shown for selected stations in Fig. 5.9. While there is a tendency for the warmest few events to feature southwesterly ambient flow aloft, there are a good portion of events with temperature above 0°C and southeasterly winds (as occurred for the case study in Chapter 4, where warmest temperatures occurred with easterly ambient
flow). This suggests once again that advection, and not mountain wave structure, is most important for temperature during foehn events in the MDVs.

The case study presented in Chapter 4 featured strong forcing throughout the period, associated with two cyclonic disturbances that passed over the Ross Ice Shelf, and strong upper-level flow over the MDVs. Is strong forcing (strong upper-level flow) required for particularly warm foehn events? Figure 5.10 shows scatterplots of event-maximum temperature with time for selected sites. The points are color-coded by 600 hPa wind speed. Several points can be made from these plots. First, it is clear that the warmest events are associated with strong winds aloft, as there are few events of ambient wind speed over 12 m s\(^{-1}\) that feature below-freezing temperatures. Especially for the eastern MDVs sites (Lake Fryxell and Lake Brownworth in Fig. 5.10), there are few events with above-freezing temperatures that have ambient winds below 12 m s\(^{-1}\). Such strong winds also appear necessary for early season (November) events.

Interestingly, there are more above-freezing events in November than in February at all sites, despite average temperatures being similar if not slightly cooler during November at McMurdo and several coastal Antarctic stations (King and Turner 1997). Figure 5.10 shows more strongly-forced events in November than February at all sites shown, as a majority of the February foehn events are weak and below freezing. This suggests that the summer cyclone activity minimum around coastal Antarctica (e.g., Simmonds et al. 2003) and weaker upper-level flow during summer come into play in February.
5.4 Synoptic and Large-Scale Foehn Variability

Climate variability in the McMurdo Dry Valleys is important not just for the biology and hydrology of the immediate region, but also has supposed links to climate variability of greater Antarctica. The premise is that climate variability in the MDVs is amplified compared to other parts of the continent, and that it can provide a window into possible Southern Hemisphere climate change. However, previous studies (e.g., Doran et al. 2002b; Bertler et al. 2004, 2006) are only speculative because the physical mechanisms responsible for the MDVs warming events were not known, and thus neither were the implications of the large-scale modes of Southern Hemisphere climate variability. Here we focus on abnormally warm summer seasons and events to study the synoptic-scale patterns responsible for these warm events, and then relate those synoptic-patterns to both intraseasonal and interannual climate variability.

5.4.1 Synoptic-scale Flow During Extreme Warming

In the case study presented in Chapter 4, temperature changes during the event, independent of diurnal effects, were attributed to differences in advection aloft and the source regions of the air. The warmest period in the case study corresponds to an intrusion of warm, moist maritime air advected aloft over Antarctica and then brought to the surface in the MDVs by a foehn mechanism. Here we discuss the synoptic-scale patterns associated with the warmest foehn events, and the large-scale modes of climate variability that cause these events.
In the last section it was shown that foehn events in the MDVs can occur when the ambient wind direction ranges anywhere from $90^\circ$ (easterly) to $270^\circ$ (westerly). A valid question is if a particular ambient wind direction favors extremely warm events. First, we compile a set of warm events as any observed event at Lake Hoare that exceeds $+3^\circ$C maximum temperature during the event. Lake Hoare is chosen because it is centrally located in Taylor Valley and also has a reliable and long observation record. With these criteria, 73 events for the 1994/1995 to 2008/2009 NDJF period are found. Figure 5.11 shows the average upstream 600 hPa wind direction for these events. Warm events are found throughout the range of wind directions mentioned above. This shows that there is not a specific wind direction and/or physical process associated with ambient winds from that direction that lead to warm events. As shown in Chapter 4, the foehn mechanism is similar whether the wind direction is primarily easterly, southerly, or westerly. Most warm events occur when the ambient wind direction is south-southeasterly or south-southwesterly. This is somewhat surprising based on previous literature (Bertler et al. 2004), where warm seasons were attributed to maritime influence, and cold seasons to continental flow. Still, this argument is not completely invalid, because there are several warm events with ambient flow from the east.

While the finding that the warmest foehn events can occur through a wide range of wind directions appears robust, it should be noted that the events that comprise Fig. 5.11 are not independent, as most groupings of events occur during consecutive days or even consecutive 6-hour periods. Of the 979 total observed foehn events considered, 498 of them (51%) are part of an “extended” foehn event, where the considered event is
embedded in five straight days of foehn conditions in the MDVs. This point is interesting in that the warm events are not scattered somewhat randomly across a season, but instead occur in what are assumed to be organized synoptic-scale patterns. In Chapter 2 it was mentioned that intrusions of warm, moist air into East Antarctica are often associated with blocking highs south of Australia and New Zealand. To explore this further, we compute a blocking index at various longitudes for the 73 warmest events at Lake Hoare. The blocking index used is that of Wright (1994), defined as

$$BI = 0.5(U_{25} + U_{50} + U_{55} + U_{60} - U_{40} - U_{50} - 2U_{45}),$$

where $U$ is the zonal wind component at 500 hPa, and the subscripts indicate Southern Hemisphere latitudes. Positive values indicate well-developed zonal flow at low and high latitudes, with a weakly zonal component in mid-latitudes, indicative of blocking. Positive blocking values for all of the warm events at longitudes ranging from 90°E to 90°W every 15° are shown in Fig. 5.12. Blocking events are most common and strongest between 150°E and 120°W, which agrees with the blocking climatologies referenced in Chapter 2. About half of the warmest events feature a positive blocking index in this sector, suggesting that an amplified circulation pattern is responsible for advecting warm midlatitude air into Antarctica.

The range of ambient wind directions for warm events in Fig. 5.11 precludes any specific weather patterns from being responsible, so a straight compositing of geopotential height is not done. Instead, we average the 600 hPa winds across a transect from 90°E to 90°W through a range of longitudes between 70°S and 80°S, which is shown in Fig. 5.13. Immediately apparent is a wave-2 pattern of meridional winds across
the transect, with alternating northerly and southerly winds, and phase lines tilting eastward to the north. This pattern suggests high pressure over East Antarctica near 150°E, low pressure near 180°, and high pressure near 135°W. This is an amplified circulation pattern over Antarctica, which differs markedly from “typical” Antarctic circulation patterns of storm systems circling the continent and Antarctica itself being isolated from the rest of the Southern Hemisphere. The southerly wind component found in most of the warm events is obvious from Fig. 5.13, and the easterly component seen for other warm events is apparent from the easterly tilt in the vectors at all latitudes from approximately 135°E to 135°W. This easterly component is clearly independent from the wave-2 pattern, as it is opposite to the general westerly ambient upper-level flow around the continent. So the easterly component also suggests an amplified circulation pattern. Amplified circulation patterns (enhanced meridional winds) over Antarctica are responsible for warm conditions at high-elevation stations over continental Antarctica (see section 2.3.5). Therefore, maritime intrusions appear to be responsible for warm events in the MDVs, as they are over the rest of the continent.

5.4.2 Interannual Variability

In the previous subsection we showed that the warmest foehn events in the MDVs typically feature an amplified circulation pattern over Antarctica, with ambient flow direction over the MDVs either from the south-southwest associated with high-pressure intrusions over Adélie Land, or an easterly component associated with high-pressure over Antarctica and lower pressure to the north. Is there any evidence that large-scale modes
of interannual climate variability like SAM or ENSO might be causing these synoptic-scale patterns to occur?

The dominant mode of interannual variability in the Southern Hemisphere is the Southern Annular Mode (SAM), so it is expected that the SAM has some influence on MDV climate variability. A positive relationship between MDV foehn frequency and the SAM was inferred by Speirs et al. (2010), as foehn frequency differences between 2006 and 2007 were attributed to a larger magnitude low-pressure anomaly in the Amundsen Sea, and a corresponding increase in the SAM. Speirs et al. (2011) find a statistically significant correlation of +0.75 between DJF foehn frequency averaged from several LTER MDV sites and the SAM index. An insignificant correlation of +0.40 is found between Lake Hoare temperature and the SAM index. Note that DJF is the only season with statistically significant positive correlation between foehn frequency and SAM. Similar to the findings of Speirs et al. (2010), Fig. 5.14 shows MSLP differences for negative SAM seasons minus positive SAM seasons for the 1980-2008 period from JRA-25. In DJF, MSLP differences are maximized in the Ross Sea, corresponding to lower pressure during positive SAM summers. While the relationship featuring increased foehn frequency with positive SAM summers is robust, it does not explain all of the foehn frequency variability in the MDVs.

Speirs et al. (2011) quantify a positive relationship between foehn frequency and air temperature anomalies, as had been inferred from earlier studies. Figure 5.15 shows a time series of standardized foehn frequency anomalies and standardized air temperature anomalies for Lake Hoare. There is a statistically significant correlation of +0.58
between the two quantities, which is clear from Fig. 5.15. Particularly warm summer seasons and high foehn frequency are seen for 1995/1996 and 2001/2002, and opposite cooler and less frequent foehn events during 2000/2001 and 2003/2004. What can be said about the large-scale modes of climate variability these seasons and other isolated periods in the observed record?

5.4.3 Intraseasonal Variability

The first season considered is summer 2001/2002. As seen in Fig. 5.15, this is one of the warmest summers on record. Monthly temperature anomalies at Lake Hoare from November 2001 to February 2002 are -0.50°C, +1.50°C, +1.36°C, and +1.13°C, respectively. The significance of this season is discussed in Chapter 2, as it resulted in enormous melt and streamflow in the MDVs. As discussed by Massom et al. (2006), this season features a positive SAM index throughout, with monthly values of 2.14, 1.17, 2.11, and 2.79 for November 2001 through February 2002, respectively (data from the Marshall (2003) station-based SAM index). However, ENSO forcing is unorganized and weak during this summer (Turner et al. 2002). Southern Oscillation Index (SOI) values, the normalized Tahiti – Darwin surface pressure, are 0.7, -0.8, 0.4, and 1.1 for November 2001 through February 2002, respectively (data from NOAA/NCEP CPC). Both Turner et al. (2002) and Massom et al. (2006) note the strong circumpolar trough and enhanced wavenumber-3 pattern of high-pressure in the midlatitudes. The latter feature brought warm-air intrusions into the continent (Turner et al. 2002; Massom et al. 2004). This pattern is apparent from the 500 hPa geopotential height anomalies for November 2001-
February 2002 compared to the corresponding period from 1989-2009 from ERA-Interim in Fig. 5.16a. Geopotential heights are lower around Antarctica, especially in the region of the Amundsen Sea low, and an enhanced wavenumber-3 pattern is seen in mid-latitude positive anomalies. Figure 5.16b shows sea-surface temperature anomalies for the 2001/2002 summer season from ERA-Interim. As expected, there are no large organized tropical Pacific SST anomalies associated with ENSO. But there is a broad region of above-average SSTs, up to +2.5 K, extending from 150°E to 150°W in the midlatitudes, east of New Zealand. The role, if any, of this midlatitude positive SST anomaly is not clear. Schneider et al. (2011) find statistically significant correlations between SSTs in a region just north and west of here and temperatures in West Antarctica in austral spring, and attribute trends in these quantities to the extratropical response to tropical convection.

So far, there is nothing to suggest more than positive SAM conditions for greater foehn frequency and warm conditions in the MDVs during summer 2001/2002. One possible explanation for the warm conditions is the South Pacific Wave (SPW). This is the fourth EOF of geopotential height or streamfunction interannual variability, as discussed by Kiladis and Mo (1998), Kidson (1999), and Frederiksen and Zheng (2007). The forcing has been attributed to contemporaneous SST-forcing in the subtropical Indian Ocean, with wavetrain source over northern Australia (Frederiksen and Zheng 2007, see their Figs. 3d and 5d). The wavetrain pattern in Fig. 5.16a is similar to the fourth slowly-varying EOF in Frederiksen and Zheng (2007). The pattern is also consistent with the strong blocking high northeast of the Antarctic Peninsula that brings warm air into the region from the north (Massom et al. 2006). Furthermore, the SST
anomalies in Fig. 5.16b correspond with the one-point DJF SST-EOF4 correlations in Frederiksen and Zheng (2007, their Fig. 5d), in particular the meridional dipole in the subtropical Indian Ocean, negative anomalies south of Australia, and positive anomalies south and east of New Zealand. While it must be remembered that this is the fourth EOF, and explains on the order of 5-8% of variability, there is also weak ENSO forcing during this season. The lack of ENSO forcing may allow for the SPW pattern to show up. For a more detailed look at the modes of variability, an EOF analysis was performed on the DJF 1989-2010 period of ERA-Interim 500 hPa geopotential height. The first four Varimax rotated EOFs are retained, and shown in Fig. 5.17. The loading patterns are broadly similar to those found for austral summer in other studies (e.g., Kiladis and Mo 1998; Kidson 1999; Fogt and Bromwich 2006; Frederiksen and Zheng 2007), with varying percentage variance explained and geographical position of loading centers. EOF1, explaining 27.8% of variance, resembles the SAM, although the positive loadings around Antarctica appear shifted towards East Antarctica compared to that in other studies. EOF2 and EOF3 resemble the PSA1 and PSA2, respectively, and explain 15.5% and 14.8% of variability. EOF4 is a wavenumber-4 mode similar to that shown in Fogt and Bromwich (2006, their EOF4 in Fig. 10) and Frederiksen and Zheng (2007, SS-EOF4 in their Fig. 3), who refer to this mode as the SPW. A time series of these EOFs for DJF from 1989-2010 is shown in Fig. 5.18. Immediately apparent is the large negative score of EOF4 during summer 2001/2002. This suggests that SPW effects have a substantial influence for the warm summer 2001/2002 in the MDVs. Note that this year has the largest score of any for the unrotated EOFs as well. Interestingly, despite weak ENSO
conditions, both EOF2 and EOF3 have large negative scores for summer 2001/2002. An explanation for the substantial EOF2 and 3 effects is not clear. Tropical forcing not directly related to ENSO, like the SPW or higher-frequency variability, may be projecting onto these modes.

MDV foehn events are sporadic and clustered together, and warm conditions during warm seasons are typically not uniform with time. There are 23 six-hourly periods of foehn conditions with temperatures greater than +3°C at Lake Hoare during this season, occurring on only 11 days, and during only three distinct periods. This suggests that there is considerable intraseasonal variability associated with foehn events. Figure 5.19a-d shows monthly 500 hPa geopotential height anomalies for the 2001/2002 summer season. In all plots, several features are consistent amongst themselves and Fig. 5.16a: negative anomalies over southeast Australia and the Amundsen Sea, and a wavenumber-3 pattern of high-pressure in midlatitudes. However, the blocking high-pressure region in the southwest Pacific shifts between months, from east of New Zealand in November and December, to south of New Zealand in January and February. January is particularly interesting, with a weakened negative anomaly in the Amundsen Sea that is shifted west, and a positive intrusion extending poleward across Adélie Land. Overall, Fig. 5.19a-d suggests combined SAM+ and SPW forcing, with the position of individual features shifted between months.

Because the dominant modes of intraseasonal variability are related to MJO forcing (e.g., Kiladis and Mo 1998; Matthews et al. 2004), it is of interest to gauge any possible MJO influence on MDVs warm events. Here we show examples of 500 hPa
plots from the three particularly warm periods during the anomalously warm 2001/2002 austral summer, and information about the MJO phase corresponding to each time.

Figure 5.20 shows 500 hPa geopotential height plots from ERA-Interim and the 32-km grid Polar WRF output at 0000 UTC 15 December 2001, which is one of the warm foehn events exceeding +3°C at Lake Hoare. A broad cyclonic circulation is centered over the Ross Sea and extending from the Tasman Sea to West Antarctica. There is a wavenumber-4 pattern around Antarctica, associated with an amplified flow pattern. A confluence zone between the cyclonic circulation and a ridge over East Antarctica exists over the MDVs, with either source region being warm, maritime air. The deep low-pressure over the Ross Sea is consistent with the SAM+ conditions during December 2001, also reflected by the 500 hPa geopotential height anomalies for December 2001 (Fig. 5.19b). But as mentioned, weak ENSO forcing exists for this month, so it is possible that a higher-frequency mode of variability might be in play.

Figure 5.21 shows values of the real-time MJO index of Wheeler and Hendon (2004), which are based on the projection of daily observed data onto multivariate EOFs with the annual cycle and components of interannual variability removed. The index in Fig. 5.21 is shown in the phase space of the normalized PC’s of the two EOFs, RMM1 and RMM2, with the 8 phases corresponding to the PC time series values and the associated position of the tropical negative OLR anomalies. Values in the circle represent weak MJO forcing. There is a one to two week lag between the phase of the MJO index and the extratropical response (Matthews et al. 2004). Taking into account this lag, for early December, the index increases in magnitude within phase 5 and moves into phase 6 after
8 December, corresponding to anomalously strong convective activity in the Maritime Continent and extreme western Pacific regions. While the response in the Southern Hemisphere mid- and high-latitude polar regions is complex and not directly related to the MJO, the MJO phase composites of Pohl et al. (2010, their Fig. 4) show anomalously low 700 hPa geopotential heights over Australia, the Ross Sea, and West Antarctica, with higher geopotential heights just east of New Zealand for both phases 5 and 6. This pattern corresponds with both the 500 hPa geopotential height plots in Fig. 5.20 and the December 2001 500 hPa geopotential height anomalies in Fig. 5.19b.

Another particularly warm foehn event occurs from 30 December 2001 to 3 January 2002, when temperatures clear +7°C at Lake Hoare. Figure 5.22 shows 500 hPa geopotential height from ERA-Interim and Polar WRF at 0000 UTC 1 January 2002, in the middle of the extended foehn event. There is a wavenumber-5 pattern around Antarctica, with intrusions of high-pressure into Antarctica over East Antarctica (near 120°E), West Antarctica (near 130°W), and the Antarctic Peninsula. South-southwesterly winds are present above the MDVs, once again on the edge of a low-pressure system and a ridge. The amplified ridge into East Antarctica advects maritime air from the Indian Ocean over the continent. The MJO index phase plot for December 2001 in Fig. 5.21 shows a stagnant MJO cycle from mid-December through the end of the month in phases 7 and 8. In fact, the index remains in phase 7 or 8 for 29 days, when the nominal period of two phases of a typical MJO cycle is only 12 days. The MJO 700 hPa geopotential height composites of Pohl et al. (2010) hint at decreased heights
southeast of Australia and west of the Antarctic Peninsula and increased heights north of
the Ross Sea.

The circulation pattern becomes even more amplified for the third warming event
10-12 January 2002, for which 500 hPa geopotential heights are shown in Fig. 5.23 at
1200 UTC 10 January. Temperatures at Lake Hoare exceed +9°C during this event. A
pronounced blocking high is positioned southwest of New Zealand, with poleward flow
extending from southeast Australia to interior East Antarctica. Flow over the MDVs is
confluent between southerly flow from the South Atlantic or the aforementioned blocking
high. A massive cyclonic circulation exists over the western hemisphere. Returning to
the MJO phase plot in Fig. 5.21, we see strong phase 7/8 activity through 4 January
before weakening activity. It is possible that the persistent phase 7/8 activity is
responsible for the amplification seen during this event. Note that the phase 8 composite
of Pohl et al. (2010) is similar to the SPW pattern previously mentioned, where the
blocking high south of New Zealand and cyclone in the Amundsen Sea lead to amplified
meridional flow over the MDVs.

November 1997 was an active month for foehn events (Fig. 5.15), including six 6-
hourly periods with temperatures above +3°C at Lake Hoare. The monthly average
temperature was over 4°C higher than the November climatological average. A much
different circulation pattern exists for November 1997, as shown in the 500 hPa
geopotential height anomalies in Fig. 5.24a. Positive height anomalies surround
Antarctica and the adjacent Bellingshausen Sea, with negative anomalies north of the
continent and south of New Zealand. This is almost an opposite pattern to 2001 (Fig.
5.16), as November 1997 features strong negative SAM (-3.17 SAM index) and El Niño (-2.0 SOI index) conditions, the latter obvious from the SST anomalies for this month in Fig. 5.24b. An example of 500 hPa geopotential height from ERA-Interim and Polar WRF from one of these warm events at 0000 UTC 10 November 1997 is shown in Fig. 5.25. A trough extends from the Ross Sea northwards to the east of New Zealand, with an anticyclone over East Antarctica, bringing east-southeasterly flow into the MDVs. This is clearly a different circulation pattern than the warm events in 2001/2002 shown earlier, illustrating that MDVs foehn events can occur through a variety of different synoptic patterns. MJO forcing is consistent for most of November 1997, which may lead to the amplified circulation pattern. From Fig. 5.26, relatively strong phase 1 conditions lasted until 17 October, before returning to phase 8 only 11 days later on 27 October. Conditions then remained in phase 8 or 1 and increased in strength through 11 November. The persistent MJO forcing in phase 8 and 1 from mid-October through early November may be responsible for the active foehn conditions during the first three weeks of November. The regression maps of 200-hPa streamfunction response to OLR during the first portion of the MJO cycle in Matthews et al. (2004) support lower geopotential heights in the Ross Sea.

Our final example comes from January 1996, for which 500 hPa geopotential height anomalies are shown in Fig. 5.27a. The pattern is similar to January 2001, with a broad region of positive anomalies southwest of New Zealand and a localized negative anomaly in the Amundsen Sea. This month features a weak positive SAM (+0.55), and a relatively strong La Niña (SOI of +1.6). An example during a particularly warm period
(greater than +3°C at Lake Hoare) is shown for 1200 UTC 28 January 1996 in Fig. 5.28. A deep cyclone (under 4900 gpm) is located in the Amundsen Sea, with a ridge extending into Adélie Land from south of New Zealand. The MDVs feature southerly flow on the western edge of the cyclone. This warm event is in the midst of an extended foehn event from 25 January to 1 February, and once again is associated with a stagnant MJO cycle. From 20 December 1995 to 16 January 1996, only phase 8 or 1 conditions exist (with two weak instances of phase 2 towards the end of this period (Fig. 5.29). Additionally, phase 7 conditions existed for 11 straight days before this period.

The preceding examples have not only highlighted the amplified circulation patterns associated with particularly warm foehn events in the MDVs, but the varied large-scale forcing associated with periods of persistent foehn events. Months featuring persistent foehn events with above-average MDVs temperatures and a few extremely warm events can occur under either phase of the SAM or the SOI. However, a tendency was also found for stagnant MJO modes to occur corresponding to the active foehn periods at the appropriate 1-2 week lag. This finding appears robust, as for the 73 warmest events at Lake Hoare, 63 of them feature a stagnant (remaining in one phase or retrograding before subsequently returning) MJO phase for 9 or more days in the period 6-30 days prior to the foehn event. For these warmest events, the phase does not seem to particularly matter, as shown by the MJO index 10 days prior to each event in Fig. 5.30a. Warm events occur in any phase, but most are outside or towards the outside of the inner circle, and are thus “significant” MJO events. For the 979 total foehn events, there is no discernible pattern (not shown). However, the foehn events with southwesterly ambient
forcing (600 hPa wind direction greater than 203°) are clustered more towards phases 4-7, indicative of enhanced convection over the Maritime Continent and the Western Pacific (Fig. 5.30b). Enhanced convection in this region may be necessary for the specific synoptic-scale pattern of a blocking high south of New Zealand and an enhanced low-pressure region in the Ross-Amundsen Seas for these events. Unlike in the Northern Hemisphere, in the Southern Hemisphere the extratropical response is extremely sensitive to the exact location of the tropical convection (Matthews et al. 2004). But overall, no one specific synoptic-scale pattern is responsible for foehn events in the MDVs, which may explain the lack of preference for specific MJO modes for the total foehn event set. Stagnant MJO forcing is perceived to enhance the extratropical response to tropical forcing (which may be necessary in the austral summer, when this response is typically weak), and also keep an amplified circulation pattern in place for an extended period of time (i.e., one week or more), which allows time for warm air to reach Antarctica from midlatitudes.

5.5 Summary

The foehn criteria of Speirs et al. (2010) was used on both LTER observations and Polar WRF output at gridpoints representing observation sites to gauge the model’s ability to represent foehn conditions in the MDVs. Polar WRF has far too many foehn events across all observing sites and all years considered. The extra foehn events in Polar WRF primarily result from a positive wind speed bias and foehn conditions lasting too long and being too widespread across the valleys. Thus, Polar WRF is generally not
“making up” foehn events, and does capture much of the interannual and spatial variability of MDVs foehn.

The primary foehn features identified in the case study of Chapter 4 are confirmed to be robust across a set of foehn events. A gap wind is the primary mechanism for southerly flow into the MDVs during periods of easterly large-scale forcing, while it is less important for westerly ambient flow. The ambient flow direction also affects mountain wave characteristics upstream of the MDVs, with nonlinear effects like wave breaking more prominent for easterly and southerly ambient flow. The warmest events are not dependent upon the ambient flow direction, but instead on the ambient wind speed aloft, which needs to be high enough to force large-amplitude mountain waves, and the source region of the flow, determined by the degree of amplification of the planetary wave pattern around the Southern Hemisphere extratropics.

The foehn activity during a particular season is significantly related to the MDVs temperature anomalies during that season, and statistically significant positive correlations are found between foehn activity and the SAM. This makes sense because positive SAM implies a stronger Amundsen Sea low and lower pressures around the Antarctic continent, both of which can lead to an amplified planetary wave pattern around the continent. Correlations with ENSO are not significant during summer, which is somewhat surprising considering the effect ENSO forcing has on the extratropical wavetrain pattern around the continent. Examples are shown of particularly warm conditions and increased foehn activity during both ENSO phases, which likely explains the poor linear correlations between ENSO and foehn activity. Note that stagnant MJO
phasing is also present in the weeks leading up to the December 2006 / January 2007 case study presented in Chapter 4. An EOF analysis on DJF 500 hPa geopotential heights show large PC values for PSA and SPW modes during seasons when ENSO forcing is negligible. Several cases are found of stagnant MJO phasing preceding extended foehn events, suggesting that an organized extratropical response to persistent tropical convective activity may be responsible for the “foehn outbreaks” mentioned in Speirs et al. (2010).
Figure 5.1. Interannual variability of foehn days at LTER AWS sites for all foehn events identified in Polar WRF (“WRF ALL”, green), all foehn events identified in observations (“LTER ALL”, blue), and matched events in both Polar WRF and LTER observations (“BOTH”, red). “Corr” refers to correlation between LTER ALL and BOTH. “% of LTER” refers to the percentage of observed events in the BOTH set.
Figure 5.2. Spatial variability along transects in Taylor Valley (TG to EC), Wright Valley (LVa and LBr), and in Victoria Valley (LVi) for all foehn events identified in Polar WRF ("WRF ALL", green), all foehn events identified in observations ("LTER ALL", blue), and matched events in both Polar WRF and LTER observations ("BOTH", red).
Figure 5.2: Continued
Figure 5.3. Averaged differences between Polar WRF and LTER observations (Polar WRF – LTER) at each site for event-average temperature (“TEMPAVG”), event warming (“WARMING”), degree days above freezing (“DDAF”), event-average relative humidity (“RHAVG”), and event-average wind speed (“WSAVG”).
Figure 5.4. Number of missed criteria for foehn events at each LTER site for (a) events in Polar WRF not in LTER and (b) events in LTER not in Polar WRF.
Figure 5.5. Scatterplot of cross-gap reduced pressure difference (hPa, x-axis) vs. near-surface gap wind speed (m s$^{-1}$, y-axis). Values color-coded according to gap wind speed and wind direction at 600 hPa. Cross-gap transect and wind speed and wind direction averaging area shown in Fig. 4.8a.
Figure 5.6. Orientation map for 2-km grid spacing simulations. Terrain height (m) shaded, LTER AWS sites shown by green dots, white transect used in cross-gap pressure calculations, white averaging box used for near-surface gap wind speed, and blue averaging box used for inverse Froude number calculations.
Figure 5.7. Scatterplot of 600 hPa upstream wind direction (degrees, x-axis) vs. nondimensional mountain height (y-axis) for all observed foehn events at two or more sites. Averaging areas for wind direction and nondimensional mountain height shown in Fig. 4.8a.
Figure 5.8. Scatterplots of 600 hPa wind direction (degrees, x-axis) vs. event-average wind speed (m s$^{-1}$, y-axis) for all observed foehn events at two or more sites for Lake Vanda (top left), Lake Brownworth (top right), Lake Bonney (bottom left), and Lake Fryxell (bottom right).
Figure 5.9. Scatterplots of 600 hPa wind direction (degrees, x-axis) vs. event-maximum temperature (°C, y-axis) for all observed foehn events at two or more sites for Lake Vanda (top left), Lake Brownworth (top right), Lake Bonney (bottom left), and Lake Fryxell (bottom right).
Figure 5.10. Scatterplot of time (x-axis) vs. event-maximum temperature (°C, y-axis) for all observed foehn events at two or more sites for Lake Vanda (top left), Lake Brownworth (top right), Lake Bonney (bottom left), and Lake Fryxell (bottom right). Color-coding refers to 600 hPa upstream averaged wind speed, according to the key on bottom of the plot (m s⁻¹).
Figure 5.11. Barplot of the number of foehn events exceeding +3°C maximum temperature at Lake Hoare based on the average upstream 600 hPa wind direction during the foehn period.
Figure 5.12. Scatterplot of positive 500 hPa blocking indices from ERA-Interim for the 73 warmest foehn events at Lake Hoare (those with maximum observed temperature > +3°C) according to the blocking index of Wright (1994) at longitudes indicated on the x-axis. Numbers at the top of the plot correspond to the number of events with positive blocking index values at this longitude, and the percentage of the 73 total events.
Figure 5.13. Average 600 hPa wind vectors for all 73 foehn events at Lake Hoare exceeding +3°C maximum temperature along latitudinal transects from 90°E to 90°W. Dashed lines represent zero phase lines of meridional wind. Reference vector shown to bottom right of plot.
Figure 5.14. JRA-25 MSLP differences for negative SAM seasons minus positive SAM seasons over the 1980-2008 period. The zero line marked in bold and negative MSLP shown by dashed line. From Speirs et al. (2011).
Figure 5.15. Monthly standardized foehn anomaly compared with the standardized air temperature anomaly for Lake Hoare. Data area smoothed with a 5 month moving average. From Speirs et al. (2011).
Figure 5.16. ERA-Interim NDJF 2001 anomalies, relative to the 1989-2009 period, for (a) 500 hPa geopotential height (m) and (b) SST (K).
Figure 5.17. First four leading Varimax-rotated EOFs of DJF 500-hPa geopotential height anomalies from 1989-2010 ERA-Interim. Percentage of variance explained in top right of each plot. Signs of loading centers are arbitrary.
Figure 5.18. Time series plots of the amplitude of each eigenvalue of the four leading EOFs presented in Fig. 5.15. The mean of each component time series is subtracted to give the amplitudes shown.
Figure 5.19. ERA-Interim 500 hPa geopotential height anomalies for (a) November 2001, (b) December 2001, (c) January 2002, (d) February 2002.
Figure 5.20. 500 hPa geopotential height (contours, m) at 0000 UTC 15 December 2001 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity (10^{-5} s^{-1}, color shaded) and wind vectors.
Figure 5.21. Daily MJO index phase space plot from 1 December 2001 to 15 January 2002. X-axis represents PC for RMM1 and Y-axis represents PC for RMM2 (see text for explanation). Geographic locations along axes pertains to the approximate locations of enhanced convective signal of MJO for that respective phase. Dots represent daily values, with text labels of day of month every 5 days. Values inside the circle represent weak MJO activity.
Figure 5.22. 500 hPa geopotential height (contours, m) at 0000 UTC 01 January 2002 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity ($10^{-5}$ s$^{-1}$, color shaded) and wind vectors.
Figure 5.23. 500 hPa geopotential height (contours, m) at 0000 UTC 11 January 2002 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity (10^-5 s^-1, color shaded) and wind vectors.
Figure 5.24. ERA-Interim November 1997 anomalies, relative to the 1989-2009 period, for (a) 500 hPa geopotential height (m) and (b) SST (K).
Figure 5.25. 500 hPa geopotential height (contours, m) at 0000 UTC 10 November 1997 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity ($10^{-5} \text{ s}^{-1}$, color shaded) and wind vectors.
Figure 5.26. Daily MJO index phase space plot from 15 October 1997 to 30 November 1997. X-axis represents PC for RMM1 and Y-axis represents PC for RMM2 (see text for explanation). Geographic locations along axes pertain to the approximate locations of enhanced convective signal of MJO for that respective phase. Dots represent daily values, with text labels of day of month every 5 days. Values inside the circle represent weak MJO activity.
Figure 5.27. ERA-Interim January 1996 anomalies, relative to the 1989-2009 period, for (a) 500 hPa geopotential height (m) and (b) SST (K).
Figure 5.28. 500 hPa geopotential height (contours, m) at 1200 UTC 28 January 1996 from (a) ERA-Interim and (b) Polar WRF. (b) also contains absolute vorticity (10^{-5} s^{-1}, color shaded) and wind vectors.
Figure 5.29. Daily MJO index phase space plot from 15 December 1995 to 31 January 1996. X-axis represents PC for RMM1 and Y-axis represents PC for RMM2 (see text for explanation). Geographic locations along axes pertains to the approximate locations of enhanced convective signal of MJO for that respective phase. Dots represent daily values, with text labels of day of month every 5 days. Values inside the circle represent weak MJO activity.
Figure 5.30. MJO phases 10 days prior to (a) the 73 warmest foehn events at Lake Hoare and (b) all foehn events with southwesterly ambient flow (600 hPa winds from a direction > 203°).
Table 5.1. Sources of extra foehn events in WRF compared to LTER observations. Values represent average number of foehn events per year in WRF not found in LTER observations. Column categories described in text. Values in parentheses in second row of WRF ALL represent average number of observed events per year.

<table>
<thead>
<tr>
<th>Stations</th>
<th>WRF ALL</th>
<th>WRF WEAK</th>
<th>WRF EXTEND</th>
<th>WRF OTHER</th>
<th>WRF NEW</th>
</tr>
</thead>
<tbody>
<tr>
<td>Explorers Cove</td>
<td>12.3 (7.9)</td>
<td>4.8</td>
<td>5.3</td>
<td>4.5</td>
<td>0.4</td>
</tr>
<tr>
<td>Lake Bonney</td>
<td>46.6 (53.5)</td>
<td>18.0</td>
<td>21.8</td>
<td>9.6</td>
<td>9.6</td>
</tr>
<tr>
<td>Lake Brownworth</td>
<td>36.2 (24.1)</td>
<td>11.2</td>
<td>15.4</td>
<td>13.6</td>
<td>2.8</td>
</tr>
<tr>
<td>Lake Fryxell</td>
<td>26.2 (19.7)</td>
<td>11.0</td>
<td>7.8</td>
<td>10.0</td>
<td>2.2</td>
</tr>
<tr>
<td>Lake Hoare</td>
<td>65.0 (34.1)</td>
<td>21.0</td>
<td>34.4</td>
<td>18.0</td>
<td>8.8</td>
</tr>
<tr>
<td>Lake Vanda</td>
<td>48.2 (39.1)</td>
<td>24.4</td>
<td>31.2</td>
<td>3.6</td>
<td>8.4</td>
</tr>
<tr>
<td>Lake Vida</td>
<td>21.2 (22.9)</td>
<td>9.4</td>
<td>9.8</td>
<td>5.8</td>
<td>2.4</td>
</tr>
<tr>
<td>Taylor Glacier</td>
<td>76.0 (81.8)</td>
<td>37.2</td>
<td>45.4</td>
<td>2.2</td>
<td>20.4</td>
</tr>
<tr>
<td>Stations</td>
<td>LTER ALL (13.0)</td>
<td>LTER ABOVE 0</td>
<td>LTER WEAK WIND</td>
<td>LTER NEW</td>
<td></td>
</tr>
<tr>
<td>-------------------</td>
<td>----------------</td>
<td>--------------</td>
<td>----------------</td>
<td>----------</td>
<td></td>
</tr>
<tr>
<td>Explorers Cove</td>
<td>3.0</td>
<td>1.8</td>
<td>0.3</td>
<td>0.8</td>
<td></td>
</tr>
<tr>
<td>Lake Bonney</td>
<td>29.2 (74.1)</td>
<td>15.8</td>
<td>7.8</td>
<td>11.2</td>
<td></td>
</tr>
<tr>
<td>Lake Brownworth</td>
<td>6.4 (52.4)</td>
<td>1.4</td>
<td>1.0</td>
<td>4.2</td>
<td></td>
</tr>
<tr>
<td>Lake Fryxell</td>
<td>7.8 (38.8)</td>
<td>3.4</td>
<td>1.2</td>
<td>3.6</td>
<td></td>
</tr>
<tr>
<td>Lake Hoare</td>
<td>9.2 (87.3)</td>
<td>3.6</td>
<td>3.8</td>
<td>3.0</td>
<td></td>
</tr>
<tr>
<td>Lake Vanda</td>
<td>21.6 (80.6)</td>
<td>14.6</td>
<td>13.2</td>
<td>3.2</td>
<td></td>
</tr>
<tr>
<td>Lake Vida</td>
<td>6.8 (33.3)</td>
<td>4.4</td>
<td>3.4</td>
<td>1.2</td>
<td></td>
</tr>
<tr>
<td>Taylor Glacier</td>
<td>40.4 (121.5)</td>
<td>14.2</td>
<td>14.0</td>
<td>18.4</td>
<td></td>
</tr>
</tbody>
</table>

Table 5.2. Sources of extra foehn events in LTER observations compared to WRF. Values represent average number of foehn events per year in LTER observations not found in WRF observations. Column categories described in text. Values in parentheses in second row of LTER ALL represent average number of events per year in WRF.
Chapter 6: Conclusions

This study is the first to comprehensively examine the physical processes responsible for foehn events in the McMurdo Dry Valleys, Antarctica. Foehn warming events are largely responsible for climatological warm anomalies in the MDVs, so the dynamics responsible for such events, and the large-scale circulation patterns that support the development of these events, are studied here. A major hindrance to the credibility of climate studies featuring the MDVs is that the mesoscale processes unique to this region are not well understood. This led to conjecture on the effects of large-scale circulation variability on the MDVs. The goal of this study was to present a coherent explanation of foehn processes in the MDVs, and to gain some fresh insight as to how the synoptic-scale patterns and climatological modes of variability in the Southern Hemisphere affect the frequency and strength of MDVs foehn winds.

A case study of a prominent foehn event from December 2006 / January 2007 is used in Chapter 4 to illustrate the physical mechanisms associated with foehn in the MDVs. Foehn events are initiated in the western portions of the valleys, typically associated with southerly flow. The southerly flow is a gap wind, forced by reduced pressure differences on each side of an elevated gap just south of the MDVs, between the East Antarctic plateau and the Royal Society Range. Examples are shown where the
southerly gap flow is present even with ambient (600 hPa) flow having significant
easterly or westerly components. The cross-gap pressure differences are set up primarily
from flow blocking into the Antarctic continent (for easterly flow) or into the Royal
Society Range (for southerly and westerly flow).

Because of the complex terrain and the gap upstream of the MDVs being
elevated, there is mountain wave forcing over the MDVs. Specifically over the gap,
vertically propagating waves are forced that are typically nonlinear, and foehn occurs as
waves are amplified in the lee. The characteristics of these mountain waves change
especially with incoming wind direction. Southerly flow is of course the most direct path
through the gap, but upstream wind speeds are reduced because of blocking along the
Transantarctic Mountains well upstream of the MDVs. Nonlinear effects (measured
through the nondimensional mountain height) are thus greater, leading to wave breaking
in the lee of the gap. Flow with a greater westerly component is off continent, and
originates at a higher level than the gap. Therefore, flow blocking is negligible, and
nonlinear effects are reduced. These waves propagate to greater heights vertically, and
become more amplified than those from southerly flow. Shooting flow typically reaches
the western MDVs, and near-surface wind speeds in the valleys are larger. In contrast,
wave structure for easterly ambient forcing is similar to southerly flow (because of the
strong gap wind forcing), but is stronger and also features a mean-state critical layer,
which is formed by the turning of the wind with height. The critical layer changes the
wave dynamics, and leads to wave breaking aloft and a hydraulic analog to downslope
windstorms in the lee.

237
Because of the blocking effects of the Royal Society Range, only in rare instances do the eastern MDVs experience direct foehn forcing from downslope windstorms, like those found in the western MDVs. Pressure-driven channeling is the mechanism responsible for bringing the warm, dry air downvalley (and for the westerly winds observed during such warming events). As the gap flow/downslope winds reach western Taylor Valley, it is to some extent blocked by the Asgard Range, so pressure increases. No such blocking effects occur downvalley, so a pressure gradient force is established, directed downvalley. On small spatial scales within individual valleys, the force balance is antitriptic, as the Coriolis effect will not be in play. In Wright Valley and Victoria Valley, similar arguments apply when southerly flow reaches this far northward. If not, or if the ambient flow is more westerly, upstream flow from the west can be deflected downvalley through blocking effects upstream.

During foehn events, easterly intrusions often occur in the eastern MDVs, where foehn conditions are interrupted for a time by cool, moist maritime flow. These easterly intrusions have previously been attributed to thermodynamic effects between the heated bare ground land surfaces and water/ice surfaces (a sea-breeze effect). However, easterly intrusions can occur during winter, when no sea-breeze effects occur. An alternative explanation presented here is that during episodes of strong synoptic-scale forcing over the Ross Ice Shelf (as occurs during many foehn events), strong southerly flow impinges upon Ross Island. Flow blocking by Ross Island leads to a localized high-pressure region in Windless Bight, and flow is deflected around the island. The flow around the west of Ross Island is then blocked by the continent, and surface pressure increases along the
coast, just east of the MDVs. When this westward-directed pressure gradient force exceeds the easterly-directed component in the western MDVs, easterly intrusions can occur that mimic sea-breezes in thermodynamic properties. Only when flow around Ross Island is weak, which typically occurs when the ambient flow is off continent, are true sea-breeze effects important. In summer, quiescent conditions result in a sea-breeze occurring frequently, as documented by previous studies. However, because most foehn events are associated with significant synoptic-scale forcing (typically cyclonic circulations over the Ross Sea or Amundsen Sea), easterly intrusions during foehn events are better described by blocking effects rather than thermally-generated circulations.

A valid question is the role that foehn winds have on meltwater generation in the MDVs. Hoffman et al. (2008) note that solar radiation primarily drives melt (as long as surface temperatures are near or above freezing), and that weak winds are necessary to reduce sensible heat fluxes from the surface. Since foehn are often associated with storm systems and clouds, and strong winds, how can they be implicated in generating melt? First, foehn winds bring temperatures above freezing, which is not trivial even over a bare-ground surface like the MDVs. Without opposing forcing, the sea-breeze regime will prevent valley regions from warming substantially above freezing. The sea-breeze can also increases wind speeds during periods of weak forcing, promoting sensible heat loss from the surface. Even though foehn is often associated with synoptic-scale cyclones, the drying effect of foehn descent will clear away cloud cover. Even with clouds, there are cases where downward longwave fluxes can exceed what downward solar forcing would be with no clouds (Hoffman et al. 2008), presumably during twilight
periods. The analysis in Chapter 4 showed that hydraulic jumps are frequent throughout the valleys, as shooting flow rapidly adjusts from supercritical to subcritical flow. This can occur along either lee slopes or downstream along the slopes of the next mountain encountered. In these turbulent hydraulic jumps, foehn air still gets mixed in, and solar forcing warms the underlying surface with little sensible heat loss. Foehn allows for above freezing temperatures over frozen glacial surfaces, promoting melt.

Several of the components of the foehn conceptual model presented above are novel to the discussion of MDVs meteorology. As explained in the introduction, several earlier studies attributed the warm, dry westerly flow to adiabatically-warmed katabatic flow, while others suggested a foehn mechanism. Clear evidence for the latter theory was presented by Speirs et al. (2010), and the foehn concept has been further expounded in this study. The explanation of Speirs et al. (2010) left off at mountain waves and flow-deflection downvalley as the foehn mechanism, so the gap flow, pressure-driven channeling, and coastal blocking from flow around Ross Island are all new concepts in MDVs meteorology, at least to the author’s knowledge. Additional details on mountain waves (i.e., wave breaking and critical levels) are also new to MDVs meteorology. It is hoped that the presentation of this coherent and physically consistent conceptual model of MDVs foehn will cease further description in the literature of MDVs warming events as “katabatic.”

The mesoscale meteorological features of the MDVs have broader appeal beyond just the MDVs or Antarctic meteorology. Foehn in the MDVs may be the best analog to foehn in the Austrian Alps, which also features gap flow (Brenner Gap) into another
valley (Inn Valley) where it is blocked (the Nordkette) and deflected (down Inn Valley to Innsbruck). As such, the MDVs can provide additional verification for the foehn concepts in the much-studied Alps. Additionally, there are key differences between foehn in the MDVs and the Alps. The biggest difference is most prominent in the austral winter, when the extremely stable boundary layer over the Antarctic ice sheet can provide a hydraulic analog for mountain waves unrivaled elsewhere in the world. Even though stable boundary layer effects have been documented for Alpine mountain waves, such effects in Antarctica are stronger and more persistent. Boundary layer effects (particularly stagnant boundary layers) on mountain waves have received considerable research interest recently. It is also hypothesized that critical layers are more common in the MDVs than in the Alps, resulting from the incessant blocking effects of the Transantarctic Mountains.

While it is obvious that the MDVs will never be a data-rich area like many parts of the Alps, an expanded observational program in and around the MDVs would tremendously aid meteorological studies. Currently, the limited AWS surface observations and one previous field study in Wright Valley (McKendry and Lewthwaite 1990, 1992) are the only available observations. Verification of gap flow upstream of the MDVs would be as simple as an AWS unit with pressure measurements on each side of the gap. Similarly, a temporally limited field program with vertical wind profiling at just a few locations could, along with numerical model output, completely describe the mountain wave structure over the MDVs. Such combined observational-modeling
studies are commonplace for the Alps and have lead to great advances in mesoscale meteorology research.

Questions still remain as far as the influence of interannual and intraseasonal large-scale climate variability on the MDVs. It is shown in Chapter 5 that the situation is more complex than that suggested by Bertler et al. (2004, 2006). Speirs et al. (2010) showed that a strong Amundsen Sea Low is commonplace during MDVs foehn events, as it leads to strong winds throughout the troposphere along the western Ross Ice Shelf. In this study we additionally found that many foehn events are associated with a significant offshore ambient flow component, which shows the importance of the upper-tropospheric circulation over East Antarctica, particularly an anticyclonic circulation. This pattern of anticyclonic circulation over East Antarctica and a strong Amundsen Sea low implies an amplified circulation pattern, which results in both strong winds and intrusions of warm midlatitude air. The warmest foehn events are associated with an amplified circulation pattern where warm air is advected aloft over interior Antarctica, with foehn activated by the strong winds and being the mechanism that transports this warm air to the surface. Such a dependence of foehn maximum temperature on amplified circulation is also shown by Zängl and Hornsteiner (2007a), where a very warm and persistent foehn event is associated with a meridionally elongated trough that advects Saharan air into central Europe. The MDVs are in the most direct path of warm intrusions of mid-latitude air of any high-latitude region at such low elevations. Otherwise, warm signatures will be affected by terrain blocking and boundary layer effects. As shown in the case study examples in Chapter 4, the warm, moist midlatitude air ascends moist adiabatically along
the Antarctic coast, and descends dry adiabatically as foehn over the MDVs. This results in a net warming along the parcel trajectory. This may explain why the warmest events tend to have a greater westerly (off-continent) ambient wind component, as parcels traversing higher elevation East Antarctica are drier than those over West Antarctica from the east. Indeed, precipitation events at nearby Taylor Dome originate from the east (over West Antarctica), as off-continent flow is too dry (Scarchilli et al. 2010).

The large-scale variability patterns favorable for foehn conditions, and hence anomalously warm conditions on seasonal timescales, are therefore those that amplify the circulation patterns in the Ross Sea region. On interannual timescales, a positive SAM phase would be expected to be favorable for foehn, as lower pressure/geopotential height around the Antarctic continent promotes stronger pressure gradients and stronger winds, which activate foehn conditions. ENSO would also be expected to influence foehn, based on the associated PSA wavetrain patterns. While foehn correlates positively with SAM in the austral summer, a weak or even opposite relationship is found in other seasons. Correlations of foehn and ENSO are never significant during the period studied. Given the pervasiveness of SAM/ENSO effects on Antarctic climate illustrated in the literature, it is surprising that stronger relationships between foehn and these modes of variability are not found. However, Speirs et al. (2010) show that foehn events are highly episodic, and that even a few strong foehn events can significantly alter the seasonal temperature anomalies in the MDVs. This suggests that forcing on intraseasonal timescales may be more important for the MDVs. The MJO, being the dominant mode of intraseasonal variability, might then influence MDVs foehn. Several examples are found
of stagnant MJO phasing 1-2 weeks prior to an extended MDVs foehn event. These extended foehn events can occur in either phase of SAM or ENSO. Persistent stationary convective activity associated with the MJO is hypothesized to force extratropical Rossby wavetrains that amplify circulation patterns around Antarctica.

Although paleoclimate applications were not explored in this study, it is interesting to revisit theories for huge melt lakes in the MDVs at the LGM with the new ideas presented in this study. As stated in section 2.4, the two prevalent theories to explain the huge melt lakes during the LGM are (1) greater solar insolation due to the arid conditions brought on by a more expansive Ross Ice Sheet and the associated storm track changes, and (2) more frequent and extensive foehn events with no competing sea breeze. Based on the findings of this study, neither theory is completely correct, but instead aspects of each appear likely to be true. MDV foehn frequency appears tied to the greater SH storm track, based on the importance of the Amundsen Sea Low and blocking south of New Zealand. Either high-latitude effects like a more extensive ice sheet or tropical effects like different convection and SST patterns or oscillations could affect SH storm tracks. Morse et al. (1998) use radar stratigraphy, ice flow modeling, and the Taylor Dome ice core record to find arid conditions and a reversed accumulation gradient (greater accumulation to the north) at the LGM. They attribute these conditions to an expanded Ross Ice Sheet, and the associated changes in storm tracks. The reversed accumulation gradient suggests different synoptic-scale patterns being responsible for precipitation, and that cyclones in the Ross Sea advect less moisture into the MDVs region, but exactly how the storm tracks respond to the expanded ice sheet is unclear.
A more amplified circulation pattern would result in more foehn, and appears likely according to Drost et al. (2007), who find a stronger wavenumber-3 circulation pattern in their LGM simulations. More arid conditions could also favor greater warming by reducing cloud cover during foehn events, as foehn events are often associated with increased cloudiness to some extent. The reduced sea-breeze effect is probably negligible, as a thermally driven circulation still forms even without open water. Additional model simulations perturbed to LGM conditions and other sensitivity simulations would help to answer these paleoclimate questions now that the foehn effect is better understood. Related to MDVs paleoclimate questions is the role of the Taylor Dome ice core record in reconstructing past MDVs climate. As noted by Scarchilli et al. (2010), Taylor Dome precipitation originates in the Ross Sea and the West Antarctic sector. This certainly corresponds with the trajectories for the warmest period of the case study in Chapter 4, where warm conditions in the MDVs originate in the Ross Sea. Also, most foehn events feature an easterly ambient wind component. However, we have shown here that there are also many foehn events associated with off-continent ambient flow, which are presumed dry and not associated with precipitation at Taylor Dome. The cooler, more arid conditions implied by the Taylor Dome ice core (Steig et al. 2000) with massive MDVs melt lakes at the LGM suggest that more foehn events with a westerly off-continent ambient forcing occurred. This would correspond with the extremely warm events in January 2002 discussed in Chapter 5, with an amplified large-scale circulation and increased blocking south of New Zealand. The limited precipitation received at Taylor Dome probably does not coincide with foehn conditions, and is not representative
of paleoclimate within the MDVs themselves. A related paleoclimate question is why the MDVs are dry. Clow et al. (1988), Chinn (1990), and Fountain et al. (2009), among others, postulate that glacial flow being blocked upstream of the MDVs causes the bare-ground conditions. The unique meteorological characteristics of the MDVs – being located in a strong gap flow and foehn regimes and in the lee of a precipitation shadow (Monaghan et al. 2005) – suggests that meteorological conditions may be more important. The blue ice regions in western Taylor Valley and Ferrar Glacier are clear indicators of aeolian processes, and glaciological explanations can’t resolve the cessation of Taylor Glacier in the middle of Taylor Valley.

The MDVs have been suggested to be a bellwether of climate change, owing to the sensitivity of meltwater generation to small climatic changes. The reasoning is that because the MDVs temperatures are near 0°C most of the summer, even a small increase in average summer temperature of high southern latitudes will lead to an amplified response of increased melt. Such a concept is not supported by the results of this study. The warmest MDVs foehn events are associated with an amplified circulation pattern on intraseasonal timescales, and a simple small increase in average temperatures will not result in more foehn events and will not substantially increase maximum temperatures observed during foehn events. An argument could be made that more intrusions of warm midlatitude air would result in a general warming of Antarctica. However, this warming would be on regional scales, and the MDVs would not be a unique bellwether, as warm conditions for the austral summer 2001/2002 were also observed over interior East Antarctica and in the Antarctic Peninsula region.
As MDVs foehn involves spatial scales ranging from microscale to planetary, so do the possibilities for future research on this topic. As mesoscale modeling appears to be the best vehicle to study MDVs foehn processes, a solution to the problem identified in Chapter 4 of winds being too strong during foehn events is needed. An obvious candidate is increased horizontal and vertical model resolution, but the 500 m grid spacing simulations suffered from similar issues as the 2 km simulations. Another possibility is the method of turbulence parameterization and diffusion calculation in the model simulations (Doyle et al. 2000; Zängl et al. 2004b; Gohm et al. 2008). Short test simulations using combinations of anisotropic/isotropic mixing lengths and different turbulence parameterization methods in physical space at 2 km horizontal grid spacing are inconclusive. However, tests need to be done at finer grid spacing (i.e., 500 m) with these different options. As noted by Zängl et al. (2008), the placement of the lowest model level strongly influences mountain wave response at the surface, particularly in the placement of hydraulic jumps. Attaining lowest model levels close to the ground in high-resolution simulations is difficult without violating CFL criteria, but further testing is underway.

The other glaring future research application to come out of this work is the influence of the MJO on the high southern latitudes. This has been looked at by Matthews and Meredith (2004) and Pohl et al. (2010), but no studies have specifically considered summer or systematically analyzed the resulting extratropical circulation during different MJO phases of different length and during overarching interannual
modes like SAM and ENSO. This intraseasonal variability is not only important for the MDVs, but also operational forecasting at coastal stations around Antarctica.


Carvalho, L. M. V., C. Jones, and T. Ambrizzi, 2005: Opposite phases of the Antarctic Oscillation and relationships with intraseasonal to interannual activity in the tropics during the austral summer. *J. Climate*, 18, 702-718.


251


Kidson, J. W., 1999: Principal modes of Southern Hemisphere low-frequency variability obtained from NCEP-NCAR reanalyses. J. Climate, 12, 2808-2830.


——, and S. Grønås, 1993: Stagnation points and bifurcation in 3-D mountain airflow. 
_Tellus_, **45A**, 28-43.

——, S. Skubis, J. D. Doyle, A. S. Broad, C. Kiemle, and H. Volkert, 2002: Mountain 
waves over Mont Blanc: Influence of a stagnant boundary layer. _J. Atmos. Sci._, 
**59**, 2073-2092.

——, Q. Jiang, and J. D. Doyle, 2006: A theory of gravity wave absorption by a 

——, J. D. Doyle, Q. Jiang, and S. A. Smith, 2007: Alpine gravity waves: Lessons from 
MAP regarding mountain wave generation and breaking. _Quart. J. Roy. Meteor. 
Soc._, **133**, 917-936.

surrounding the minimum in the Southern Oscillation Index. _J. Geophys. Res._, 
**98**, 13 071-13 083.

Speirs, J. C., H. A. McGowan, and D. T. Neil, 2008: Meteorological controls on sand 
transport and dune morphology in a polar-desert: Victoria Valley, Antarctica. 

——, D. F. Steinhoff, H. A. McGowan, D. H. Bromwich, and A. J. Monaghan, 2010: 
Foehn winds in the McMurdo Dry Valleys, Antarctica: The origin of extreme 
warming events. _J. Climate_, **23**, 3577-3598.

variability driven by foehn winds in the McMurdo Dry Valleys, Antarctica. _Int. J. 
Climatol._, submitted.

Spreen, G., L. Kaleschke, and G. Heygster, 2008: Sea ice remote sensing using AMSR-E 

in Antarctic annual sea ice retreat and advance and their relation to El Niño-
Southern Oscillation and Southern Annular Mode variability. _J. Geophys. Res._, 

Stauffer, D. R., and N. L. Seaman, 1990: Use of four-dimensional data assimilation in a 
limited-area mesoscale model. Part I: Experiments with synoptic-scale data. 


270


