Abstract

SUFFERN, PAUL SAMUEL. Interactions of gravity waves and moist convection in the troposphere and stratosphere (Under the direction of Drs. Yuh-Lang Lin and Michael L. Kaplan).

On December 12th and 13th, 2002 a deep large-amplitude tropospheric mesoscale gravity wave formed over Texas and propagated northeastward across several states. This study examines the role of the coupling of geostrophic adjustment, moist convection, and an area of shear instability from the meso-α to the meso-γ scale as the mesoscale gravity wave is formed/maintained. Three main chapters comprise this thesis. Chapter 4 employs observations to describe the mesoscale gravity wave event. Numerical modeling of the mesoscale gravity event is explained in Chapter 5 to further understand the coupling of geostrophic adjustment, moist convection, and the area of shear instability. Chapter 6 analyzes the vertically propagating gravity waves in the lower stratosphere employing observations and numerical modeling.

Observations of the mesoscale gravity wave system allow a detailed representation of the event. As an upper-level jet streak moved into central Texas, moist convection and a corresponding surface low-pressure system began to develop. The development of widespread convection lead to the downstream growth of a secondary jet streak, which in turn continued the imbalance of mass and momentum at the upper-levels across central Texas. Geostrophic adjustment and moist convection are closely correlated with the mesoscale gravity wave as the mesoscale gravity wave forms and moves northeastward all the way to Mississippi. Other observations also show dry air and large amounts of shear
located directly behind the moist convection. The development of an area of possible shear instability behind the moist convection represents mixing and descending momentum in the troposphere and formation/maintenance of the mesoscale gravity wave.

Chapter 5 confirms and enhances the observations of the coupling between geostrophic adjustment, moist convection, and the area of shear instability through numerical simulations. Mass and momentum perturbations from the moist convection are directed upward, upstream, and cross-stream into the jet streak momentum. Momentum in the moist convective process descends toward the stable layer, resulting in mass accumulation near the surface, as simultaneously, the updraft within the strengthening moist convection penetrates into the lower stratosphere. The mesoscale gravity wave(s) is(are) initiated/maintained as the negative buoyancy perturbation and descent of momentum interacts with the stable layer near the surface behind the moist convection. The impact of latent heating and its importance to the development of the coupling of mesoscale gravity wave system are discussed in this chapter employing model sensitivity studies.

Chapter 6 analyzes the development of vertically propagating gravity waves in the lower stratosphere. Observations from the Advanced Microwave Sounding Unit-A (AMSU) polar orbiting satellite show a signal of possible vertically propagating gravity waves above the moist convection across Texas and the Gulf of Mexico. Numerical simulations were performed to diagnose the
environment in which these gravity waves formed in the lower stratosphere as well as their realism to the observations.
INTERACTIONS OF GRAVITY WAVES AND MOIST CONVECTION IN THE TROPOSPHERE AND STRATOSPHERE

By

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A thesis submitted to the Graduate Faculty of North Carolina State University in partial fulfillment of the requirements for the degree of Master of Science

MARINE, EARTH, AND ATMOSPHERIC SCIENCES

Raleigh, NC

2006

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BIOGRAPHY

Paul S. Suffern was born November 1, 1981 in Durham, North Carolina. Upon completing high school through home schooling in 2000, he enrolled at the North Carolina State University in Raleigh, North Carolina. He earned a Bachelor of Science degree in meteorology in 2004, and later that year began the Master’s program at North Carolina State University under the direction of Drs. Yuh-Lang Lin and Michael L. Kaplan. He will officially graduate in May 2007.
ACKNOWLEDGEMENTS

I would first like to thank my Lord and Savior Jesus Christ for giving me the perseverance and courage to complete this thesis. I would like to sincerely thank my committee members, Drs. Yuh-Lang Lin, Michael Kaplan, and Matthew Parker for their encouragement and guidance throughout my research. I am very grateful for the time and energy that both Michael Kiefer and Dr. Kenneth Waight gave me in helping me be able to run the MASS model. Without their help I would still be trying to get my first model run going. I would like to thank Zach Brown, Chad Ringley, and Major Dave Vollmer for their insight into my case study as they are working on similar parts of the project. I would like to thank all the members of the Mesolab that I have worked with for the past two years: Chad Ringley, Crosby Savage, Barrett Smith, Heather Reeves, Katie Robertson, Allison Witcraft, Chenjie ‘Christine’ Huang, Christ Hill, Carrie Larsen, Leigh Jones, David Levin, Dave Vollmer, Morgan Silverman, Mike Kiefer, Christian Cassell, and Zach Brown. I am very thankful for their encouragement, smiles, and discussions, without which I would have been too serious and not enjoyed my time in the Mesolab. I would like to thank all the other meteorology graduate students for their encouragement over the past two years as well, especially Patrick Pyle and his wife Amy, and Kevin Hill. Also, I would like to thank Dong Wu from the Jet Propulsion Laboratory (JPL) for his helpfulness in using and developing the AMSU results. Finally, I would like to thank my parents and two
brothers. Their encouragement and love over the past two years has made this thesis possible.

A few of the simulations for this research were performed on the Beowulf cluster maintained by the NCSU College of Physical and Mathematical Sciences (PAMS) and the High Performance Computing (HPC) Center maintained by North Carolina State University. Both Kevin Smith and Heather Reeves helped to maintain a local system where the rest of the simulations were performed and I am especially grateful to them for their hard work and many overtime hours. This research was performed with funding from the U.S. Air Force under AFRL Contract No. FA8718-04-C-0011.
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1. Introduction

Gravity waves occur frequently throughout the atmosphere, but due to their nature they are hard to find, track, and predict unless we are fortunate enough that a gravity wave forms and moves over our existing data collection sites. Because of this, our understanding of gravity waves still leaves much to be desired. Mesoscale gravity waves have wavelengths of 50-500 km, periods of 0.5-4 h, phase velocities of 15-35 ms$^{-1}$, and amplitudes of 0.5-15 hPa. Gravity waves have been shown to redistribute energy and momentum (e.g., Rauber et al. 2001), initiate and move along with convection (e.g., Zhang et al. 2003), and be a significant factor leading to clear air turbulence (e.g., Lane et al. 2003). Mesoscale gravity wave generation mechanisms have been shown to be associated with jet streaks (Zhang 2004 and references therein), orography (e.g. Clark et al. 2000), and convection (Lane et al. 2003 and references therein). While it is clear that much work has been done in an effort to understand gravity waves, both with real data-initialized cases and idealized studies, they are still not correctly forecast or accounted for, and their direct effect on the atmosphere is not clear. Computer models can resolve the larger wavelength mesoscale gravity waves, however, it is the changes in the initiation of moist convection (Uccellini 1974) and precipitation patterns (Zhang 2004) that still pose short term forecasting problems (Koch et al. 2005). One of the least understood areas remains the interactions between atmospheric turbulence and mesoscale gravity
waves. This research is motivated by that interaction and the problems, i.e. aircraft flight and atmospheric weapons research, it provides to the United States Air Force and their operations.

Understanding the interactions between moist convection, geostrophic adjustment and shearing instability during wave genesis and its implications for lower stratospheric turbulence is the problem that will be addressed. Our hypothesis is that the proximity of moist convection between two jet streaks allowed for juxtaposed unbalanced flow and shearing instability and the formation of a large-amplitude, deep, tropospheric, mesoscale gravity wave. It is possible that a synthesis or phasing of these three components was necessary for the development of the mesoscale gravity wave and a simple coupled synoptic/dynamical paradigm will be shown. Secondly, a group of gravity waves analogous to those that formed in the lower stratosphere in the mesoscale model used in this study were inferred from satellite imagery, and the environment within and around the moist convection allowed for these vertically propagating gravity waves to impact the lower stratosphere. In order to test these hypotheses a case study of an observed gravity wave event that occurred on December 12, 2002 was analyzed with observations and numerical simulations. During this day a large-amplitude tropospheric gravity wave formed and moved from central Texas through northern Mississippi. Fig. 1.1a shows the surface pressure trace from Ledbetter, Texas (LDBT) where the gravity wave caused a drop in pressure of 3 hPa over a one-hour period at 2000 UTC on December 12, 2002. This wave of depression had similar characteristics to a study done by Lin and Goff (1988) and
these similarities will be discussed further in Chapter 2. The wave of depression in this case study was located a few hundred kilometers behind a mesoscale convective system (MCS). By the time this gravity wave moved to Okolona, Mississippi (OKOM) its amplitude remained at a 12 hPa drop over a five-hour period [Fig. 1.1b], only slightly smaller in magnitude than that observed in Texas. On December 12, 2002 not only were there horizontally propagating mesoscale gravity waves observed in the troposphere, but there is evidence that there were vertically propagating gravity waves in the stratosphere as well. Due to the presence of all these mesoscale gravity waves on December 12 and 13, 2002 this date was selected to study the mechanisms accompanying wave genesis, maintenance, and dissipation of the mesoscale gravity waves.

The origin of the mesoscale gravity wave began under a strong upper-level trough that baroclinically amplified and moved southward on December 12, 2002 into central Texas and intersected with a strong jet streak moving eastward over northern Mexico. The interaction between the jet streak and amplifying trough allowed for moist convection to occur along the eastern coast of Texas and the western Gulf of Mexico in association with a strong surface stationary front. As the day progressed, the upper-level trough continued to amplify, the moist convection increased in intensity and coverage, and a surface low pressure center formed in the left exit region of the jet streak moving across northern Mexico. A sub-synoptic outflow jet began to form later in the day associated with and downstream of the convection, which had now organized itself into a squall line, similar to Hamilton et al. (1998). Part of the deep convection had formed north of
stationary frontal boundary, above a good gravity wave duct, and near a jet streak, which is a favorable location for a mesoscale gravity wave to form (Uccellini and Koch 1987) in the troposphere.

Mesoscale gravity wave maintenance and wave genesis in the literature has long been interpreted as two different processes. Wave genesis occurs due to moist convection, shearing instability and geostrophic adjustment, while gravity wave maintenance occurs when there is a possible wave duct for the gravity wave to travel through, through solitary wave maintenance, or through a wave-CISK mechanism (Lin 2007). This research will attempt to show that the wave genesis and maintenance processes are more closely related as momentum continues to be redistributed downward through the moist convection process. For mesoscale gravity waves that travel long distances, the waves are generated and then maintained by the same process or a combination of processes that led to their formation rather than just a gravity wave duct.

The rest of the thesis will be organized as follows. Chapter 2 includes a review of the relevant literature that pertains to this subject. Methods and experiment design are addressed in Chapter 3. Chapter 4 includes observations of the mesoscale gravity waves in this case, both in the troposphere and in the stratosphere using satellite analysis. Results of the tropospheric large-amplitude gravity wave simulations are presented in Chapter 5, addressing the mechanisms for gravity wave generation, maintenance, and dissipation. Also, Chapter 5 assesses the realism and genesis mechanisms of the simulated gravity waves from the various computer model runs. Chapter 6 focuses on the stratospheric gravity
waves associated with the turbulence and moist convection prominent in this December 2002 case study. Finally, some conclusions and future work are presented in Chapter 7.
Fig. 1.1 Surface pressure traces from NOAA/GSD Ground-Based GPS Meteorology (GPS-MET) Real Time Water Vapor Interface for a) Ledbetter, Texas from 0000 UTC 12 December 2002 to 0000 UTC 14 December 2002 and b) Okolona, Mississippi from 0000 UTC 12 December 2002 to 0000 UTC 14 December 2002.
2. Literature Review

2.1 Gravity wave literature

As mesoscale gravity waves have been observed in more detail with improvements in meteorological observational networks, the literature about them has steadily increased. By the 1970s as mesoscale gravity waves could be better observed with the additional surface observational network, and they were beginning to be studied in the ‘real’ world as opposed to just laboratory experiments.

Again a mesoscale gravity wave is defined by amplitudes of 1 to 15 hPa, periods of 1 to 4 hours, and wavelengths of 50 to 500 km (Koch and O’Handley 1997). A simple schematic of the circulations surrounding a mesoscale gravity wave is shown in Fig. 2.1. Convergence at the surface and rising motion occur after the trough in the surface pressure data as the mesoscale gravity wave propagates through the domain. Descending vertical motions and surface divergence are located after a ridge in the surface pressure data correlated with the mesoscale gravity wave. Clouds and possible precipitation located with the mesoscale gravity wave would, in this simple model, occur shortly after the trough in the surface pressure data. The simple dynamics found in Fig. 2.1 will be further examined in Chapter 4 and Chapter 5.

Many gravity wave source and maintenance mechanisms are summarized in Lin (2007) and this review will look at three of the generation and maintenance mechanisms. The generation mechanisms briefly examined are: (1) moist
convection, (2) geostrophic adjustment, and (3) non-linear interactions; and the maintenance mechanisms are (1) linear wave ducting, (2) solitary wave, and (3) wave-CISK. Moist convection can act as a source for gravity waves when it encounters a stable overlying layer like the top of a temperature inversion near the planetary boundary layer (Lin 2007). As downdrafts from a single convective cell or larger convective system hit the stable inversion, the downdrafts vertically displace the stable air and buoyancy acts as a restoring force with wave excitation within the stable layer. In most instances these gravity waves are highly dispersive, however, they can be maintained by several mechanisms that will be discussed shortly. Areas in which geostrophic adjustment is occurring have been areas in which gravity waves have also been observed to form (Kaplan and Paine 1977; Uccellini and Koch 1987; and O’Sullivan and Dunkerton 1995). Inertial-gravity waves can be excited when there is a non-zero divergence tendency of the ageostrophic wind field (Lin 2007). As it takes several inertia periods for a fluid to reach a balanced equilibrium, in this case geostrophic equilibrium, energy will be dispersed away from the source region of the fluid in longwave components. Several nonlinear interactions have also been proposed as gravity wave generation mechanisms. Excitation of Kelvin-Helmholtz waves can induce gravity wave formation when the Kelvin-Helmholtz waves reach large amplitudes and cause significant changes in the mean velocity profiles (Fritts 1982). Also, subharmonic excitation (Davis and Peltier 1979) and eddies that develop with shear instability in the mean flow (McIntyre and Weissman 1978) have been proposed to generate gravity waves through nonlinear interactions. While nonlinear interactions are
important in the generation of gravity waves, both moist convection and geostrophic adjustment play the largest roles in allowing for the generation of the mesoscale gravity wave in this December 12th case study.

Not only is it important to have a process that allows for the formation of a mesoscale gravity wave, but maintenance mechanisms for gravity waves allow for the continued propagation of these gravity waves large distances from their source regions (e.g. Lin and Goff 1988). Linear wave ducting was first proposed by Lindzen and Tung (1976) as a maintenance mechanism for mesoscale gravity waves. A wave duct is characterized as a layer in which a gravity wave can propagate horizontally or vertically without large amounts of wave energy loss as the wave energy is reflected or maintained. This allows for some gravity waves to propagate large distances away from their source regions. Four criteria were proposed by Lindzen and Tung (1976) as necessary for wave ducting. First of all, a stable layer is needed for the mesoscale gravity wave to propagate through. This stable layer effectively traps the mesoscale gravity wave energy allowing for long term propagation. This stable layer is usually located near the ground (Koch and O’Handley 1997) and identified by an inversion in a vertical profile. Next, the stable layer is capped by a dynamically unstable layer with Richardson numbers lower than ¼. The dynamically unstable layer aids in the reflection of the mesoscale gravity wave energy back into the lower stable layer allowing for the mesoscale gravity wave to propagate without the loss of energy. Third, a critical level, where the background wind speed is the same as the propagation speed of the mesoscale gravity wave, is located within the unstable layer. The critical level
again allows for the reflection of mesoscale gravity wave energy back into the lower stable layer. In some instances wave overreflection can occur at the critical level and this would allow a mesoscale gravity wave to amplify as it propagates horizontally. The final criteria for wave ducting from Lindzen and Tung (1976) is that the duct layer should be \((0.25+n/2)\lambda\), where \(\lambda\) is the vertical wavelength and \(n=0,1,2\ldots\). The duct layer needs to be sufficiently deep enough for mesoscale gravity wave propagation so that energy loss does not occur. Fig. 2.2 provides a simple schematic a wave duct with region 1 being the wave duct, region 2 the reflector, and region 3 another stratified layer. Region 1 would be relative stable layer deep enough for mesoscale gravity wave propagation. Region 2 is characterized by an unstable layer with a critical layer causing either the reflection or overreflection of gravity waves from region 1. A more detailed wave ducting environment is shown in Fig. 2.3 with a 1.6 km deep stable layer underlying a shallower convectively unstable layer. The stable layer would be the layer in which the gravity waves would be ducted given that the layer is deep enough, while the convectively unstable layer would reflect or overreflect the wave energy downward into the stable layer keeping the gravity waves ‘ducted’ or trapped.

According to the Lindzen and Tung ducting criteria one would also expect a possible critical level, a level where the wave propagation speed and the environmental flow are equal, within or just above the convectively unstable layer in Fig. 2.3. The Lindzen and Tung (1976) ducting criteria has come under considerable amount of scrutiny recently as vertical wind shear profiles have been examined in several cases and found critical levels within the lower stable layer.
and the wave energy instead of being absorbed was reflected allowing for the continued propagation of the wave (Monserrat and Thorpe 1996). So under certain conditions the wave duct plays a different role than wave reflection and the selection of the dominant trapped mode (Lin 2007). The solitary wave mechanism has been proposed as another maintenance source for mesoscale gravity waves. A solitary wave is defined as a wave of a single elevation or depression that propagates without change in its form or its finite amplitude (Lin 2007). For solitary waves to be maintained, equilibrium between wave dispersion and wave nonlinearity needs to exist. In a fluid system longer waves travel at faster speeds than short waves and therefore the shorter waves are left behind leading to the ‘flattening’ of a wave through wave dispersion over time. Nonlinearity within the wave tends to steepen the overall wave front as it propagates, leading the shorter waves to pass the longer waves. If an exact balance exists between the nonlinearity and wave dispersion then solitary waves can be maintained. A final maintenance mechanism for mesoscale gravity waves discussed is the wave-CISK mechanism. Wave-CISK occurs as latent heating is released by moist convection which causes a large scale disturbance. This large disturbance in turn generates low-level convergence which is supported through internal gravity waves. As the low level convergence continues conditionally unstable air rises and produces latent heat which maintains the moist convection process. The wave-CISK mechanism is similar to the original CISK mechanism in which boundary layer friction aids in the low level convergence rather than internal gravity waves. While many of these source and maintenance mechanisms...
associated with mesoscale gravity waves are seen in this December 12th case, several case studies in the literature also provide helpful insight into the features and effects of mesoscale gravity waves.

Uccellini (1975) analyzed gravity waves that had helped to initiate severe convection in the Midwest. This study found that the mesoscale gravity waves had durations of 6 to 10 hours, amplitudes of 0.5 to 2.5 hPa, and wavelengths on the order of 400 to 500 km. He found that the analyzed waves were a precursor to observed convection, which further acted to redevelop the convection that had decreased from the night before. The mesoscale gravity waves were found to have propagated over a region characterized by a low-level inversion and strong vertical wind shears, which were found in soundings above this inversion as the waves passed through the sounding area. Uccellini (1975) was one of the first to use the network of surface observations and radar to correlate the mesoscale gravity waves with the moist convection. Using this technique he found that the heaviest precipitation occurred in region with locally higher surface pressure traces, and a decrease in the convective intensity was found as the next mesoscale gravity wave approached.

A larger ‘synoptic’ view of favorable areas for mesoscale gravity wave formation was developed by Uccellini and Koch (1987). They observed and analyzed 13 cases of mesoscale gravity waves and tried to find the similar synoptic-scale characteristics in each case. Similar synoptic conditions in each of these cases included a strong 300 hPa jet, at the inflection point of the 500 hPa trough and on the stable side of a front, usually north of a warm front. While each
of these three features seem to be prominent in each of the mesoscale gravity wave cases, the generation mechanisms of each of these sets of gravity waves was less clear. Uccellini and Koch (1987) found that the amount of convection does not directly correlate with the amplitude of the mesoscale gravity wave, for in some of the strongest amplitude in their 13 cases, most were not associated with convection. They suggest that geostrophic adjustment of the jet streaks plays the most important role. Large divergence tendencies at the jet streak level were located in areas of geostrophic adjustment as the mesoscale gravity waves began to form. The simple synoptic model discussed in Uccellini and Koch (1987) will be further examined in Chapter 4.

Observations of a singular wave of depression propagating to the northeast downstream of moist convection were found in Lin and Goff (1988). The synoptic environment in which this mesoscale gravity wave forms is similar to that of Uccellini and Koch (1987). Pressure data from several barographs was examined as the wave of depression propagated to the northeast and a 3 to 6 hPa pressure drop was co-located with the mesoscale gravity wave. Strong moist convection played an important role in this wave of depression as the convection was located near the origin of the wave. Also, a midlevel inversion was located below a weakly stratified layer indicating a favorable environment for wave ducting. An analysis of the solitary wave maintenance mechanism was found to play a more dominant role than wave ducting as nonlinearity and wave dispersion were of the same order of magnitude. The wave of depression observed in Lin and Goff (1988) is similar to the mesoscale gravity wave observed during the December
12th case study both qualitatively and quantitatively and this will be further addressed in Chapter 4.

Source mechanisms for mesoscale gravity waves were further examined in Koch and Dorian (1988). This paper examined a gravity wave event that occurred in the upper-Midwest near the Cooperative Convective Precipitation Experiment (CCOPE) area and soundings and surface meso-network data from this experiment were used to analyze the gravity waves. As previously established in Uccellini and Koch (1987), their 13 gravity waves cases all occurred in similar synoptic conditions near the 300 hPa jet exit region, at the inflection point of the 500 hPa trough, and on the cold side of the surface frontal boundary. Koch and Dorian (1988) found that gravity waves formed under an upper-level jet streak and north of a stationary outflow boundary. Five soundings were observed in CCOPE to be near where the gravity waves had initiated. These soundings and upper-level analyses indicate that the gravity waves were excited in the right exit region of the jet streak as it moved over the High Plains. In a geostrophic balanced flow where there is no curvature in the jet streak we would assume that the right exit region of the jet streak would correspond to descending and decelerating air. The indirect ageostrophic circulation in this region should be turning upper-level parcels to the right and decelerating them, however, Koch and Dorian (1988) found that the ageostrophic winds are directed toward lower heights near the gravity wave source region [Fig. 2.4]. The ageostrophic winds directed toward lower geopotential heights are seen across Idaho and western Montana. Also, notice that the total wind maximum and the geostrophic wind
maximum are out of phase, also indicative of unbalanced flow. The gravity waves excitation region occurred in an area with both leftward directed ageostrophic flow and Rossby numbers greater than 0.5 [Fig. 2.4]. Consistent with Koch and Dorian’s findings Rossby number values of 1.6 were found to accompany strong jet streaks that had gravity waves forming downstream. Zack and Kaplan (1987) and Koch and Dorian (1988) diagnosed Ro values, which ranged from 0.7 to 1.5 (In Section 2.4 we will address the unbalanced flow in further detail). While Uccellini and Koch’s (1987) 13 cases in which similar synoptic conditions were present, Koch and Dorian (1988) have shown that geostrophic adjustment plays a major role in gravity wave formation regardless of other possible synoptic features.

Rauber et al. (2001), part I of the STORM-FEST St. Valentine’s Day mesoscale gravity wave event, studied how the leading edge of a dry air mass possibly played an important role in the mesoscale gravity wave formation. The synoptic setting in this case was similar to Uccellini and Koch (1987) with a developing 500 hPa disturbance, the exit region of strong jet streak at 250 hPa, and north of a warm frontal boundary. A tongue of dry air was collocated within a westerly jet directed from the Rocky Mountains in Colorado toward Kansas. As the westerly jet and dry air interacted with the warm frontal boundary, it helped to initiate the mesoscale gravity wave (Rauber et al. 2001). A convective rain band in the region of the warm front intensified, surface pressure fell and rises began to occur, like one would observe for gravity waves [i.e. Fig. 2.1], and strong vertical motions were observed ahead of the gravity wave trough. Rauber et al. (2001)
found a strong descent of momentum within a profiler dataset correlated with surface pressure falls as the gravity waves propagated over the profiler sites. The momentum from the westerly jet in the larger scale dry-slot seemed to be a driving force behind the initiation of the mesoscale gravity wave in this case, however, it is still unclear how this momentum was mixed and/or advected downward to allow for the gravity wave to form. Rauber et al. (2001) speculate that as the dry air moved above the warm front a favorable wave ducting environment could have been available with the stable lower layer below the warm front and the convectively unstable layer above this.

Yang et al. (2001), part II of the STORM-FEST St. Valentine’s Day mesoscale gravity wave event, further examined the mesoscale gravity wave formation using a Dual Doppler radar analysis looking at the three-dimensional structure of the wave. There is evidence of the strong momentum as the westerly jet from the Rockies overrode the warm frontal feature. As the momentum passes through the Dual Doppler analysis it was seen to be co-located with the pressure falls at the surface. Fig. 2.5 shows a schematic derived from the Dual Doppler analysis with the mesoscale gravity wave and the convective circulation associated with them. Yang et al. (2001) found that the momentum and associated dry convective unstable layer played a large role in allowing for the mesoscale gravity wave to form, and they speculate that the stratiform precipitation was able to mix the momentum downward on the backside of the moist convection. As the momentum mixed downward toward the stable layer, behind the moist convection and below the warm frontal feature, the momentum perturbed the stable layer
releasing gravity waves. Fig. 2.5 is very similar to the mature convective system paradigm by Houze et al. (1989), but the stable layer underlying the moist convection from the dual-Doppler analysis, and the position of the warm front are different than would be expect for the Houze et al. (1989) paradigm. Yang et al. (2001) conclude that there is no strong evidence to suggest that the stratiform precipitation was the dominant source for the mesoscale gravity wave; but that the momentum with the low-level westerly jet could have also been an important factor. The observations of strong momentum at lower levels are also observed in this December 12th case study and will be further examined in Chapter 4 and 5. While the past few studies before Rauber et al. (2001) and Yang et al. (2001) found geostrophic adjustment, shear instability, and convection to be the dominant factors in their mesoscale gravity wave formation, the momentum associated with the midlevel dry westerly wind likely contributes to these mechanisms for gravity wave formation in Yang et al. (2001).

Zhang et al. (2003) used numerical simulations to further understand the mesoscale gravity event on 4 January 1994 and the gravity wave structure and environment. In the gravity wave source region the duct layer was 1.5-2.5 km thick in the cold air north of the warm frontal boundary in his control simulations with a 3-6 km layer containing small static stability. A critical level and a convectively unstable layer, where the Richardson number (Ri) < 0.25 was found, were located above the stable ‘duct’ and this is consistent with the Lindzen and Tung (1976) ducting criteria so that the simulations were assumed to be a good representation of the overall synoptic environment. Latent heating within the
simulations played a large role in the development of the unbalanced jet streak in these simulations, because when sensitivity studies were done without latent heating within the model, the gravity waves did not occur and the unbalanced jet within the inflection point of the trough did not develop. So while the convection itself may or may not be directly associated with the mesoscale gravity wave itself, in this case, the convection was important in the development of the favorable environment for gravity wave formation, and therefore is indirectly associated with the mesoscale gravity waves.

Lane et al. (2004) investigated the relationship between gravity waves in the troposphere and atmospheric turbulence where dropsonde observations were used to verify the numerical simulations of a gravity wave event on 18 February 2001. A 100 ms\(^{-1}\) jet streak was observed in the Pacific Ocean moving eastward along with a baroclinic surface wave. As this feature advected eastward a Gulfstream-IV (G-IV) aircraft flew through the jet streak to obtain high-resolution dropsonde measurements. While flying through the jet streak the G-IV experienced large amounts of turbulence with vertical accelerations that exceeded 4.9 ms\(^{-2}\). Simulations showed that an upper-level front between 800-600 hPa played a role in the gravity wave formation and turbulence. Richardson numbers below critical levels were found behind and above this front, in areas which also had a large amount of momentum with the strong jet streak, and it is suggested that these areas are where the gravity waves began to form. Lane et al. (2004) speculate that gravity waves extracted the energy from the mean flow, through possible Kelvin-Helmholtz instability or mixing. This mixing in areas of possible
gravity waves accounted for large amounts of turbulent kinetic energy, TKE, to be observed both by the G-IV flight and within the computer model simulations of this event. While there were no surface observations to see if the mesoscale gravity waves would have been traceable by surface pressure fluctuations, the turbulence observed by the G-IV is speculated by Lane et al. (2004) to be associated with gravity waves as gravity waves were observed within the computer model simulations of the event.

2.2 Moist convection

As mentioned above moist convection has been observed in areas of mesoscale gravity wave formation and the moist convection can also affect the background environmental flow (e.g. Lin and Goff 1988). Mass and momentum adjustments occur on several length scales as the moist convection perturbs the environmental flow. In Hamilton et al. (1998) moist convection was shown to have a significant impact on the mass and momentum of the environment in an idealized atmosphere. In this idealized atmosphere a midlevel wind maximum occurs downstream of the moist convection. The maximum of latent heating within the convection allows for consecutive pressure falls and rises and the wind field adjusts accordingly (Hamilton et al. 1998). Geostrophic adjustment occurs below the level of maximum heating as air parcels accelerate toward this heating with the increase in pressure gradient force. This in turn with rotational forces leads to the development of vertical vorticity, column stretching, and the
formation or enhancement of a midlevel cyclone. As time continues the
downstream midlevel winds increase and gravity waves occur similar to Lin
(1986). A wind maximum downstream of the moist convection is also observed in
this December 12th case study and will be further discussed in Chapter 4.

In Lin (1986) gravity waves are generated from the level of maximum
heating. As the environmental flow encounters the stationary heat source a trail of
damped oscillations is visible. Upward and downward displacement areas are
located near the heating source with strong divergence and upward displacement
above and downstream of the heated layer similar to the idealized results of
Hamilton et al. (1998). Moist convection is seen in these two idealized studies to
play an important role in the mass and momentum distributions as local wind
maxima and minima occur and gravity waves are released. Similar patterns from
these idealized studies will be observed in the observations from this December
12th case study.

2.3 Geostrophic adjustment and unbalanced jet literature

As mentioned by many of the previous authors, geostrophic adjustment
can play a large role in the development of mesoscale gravity waves within the
jet-front system. Uccellini and Johnson (1979) examine the coupling of the lower-
level jet and the upper-level jet-front system. They proved that the coupling
between the indirect-ageostrophic circulation in the exit region of the upper-level
jet allows for pressure falls at the surface and corresponding cyclogenesis or convective development under the left exit region.

Zack and Kaplan (1987) found in their simulations of moderately curved jet streaks, with small wavelengths compared to the Rossby radius of deformation, that in the exit region of the jet streak instead of the ageostrophic wind blowing toward higher heights, the ageostrophic wind blows toward lower heights [Fig. 2.6]. In this cross section cut perpendicular to the jet exit region the heavy black arrows are the ageostrophic wind component and they are directed down the isentropes toward lower heights instead of being directed up the isentropes and thus causing an acceleration of the upper level jet streak in the exit region. Instead of a parcel decelerating in the exit region of the jet to restore geostrophic balance, a parcel would accelerate there, because the Coriolis force does not have enough time to turn the parcels to the right, because of the shorter wavelength trough. This unbalance flow allows for divergence aloft, which in turn modifies the mass tendencies. This adjustment process changes the mass distribution in the exit region of the jet and causes large scale upward directed vertical motion. If a surface low-pressure system has formed in the exit region, it can continue to strengthen through vertical stretching and positive vorticity advection. The surface low pressure center can act as a lifting mechanism to release convective or potential instability and aid in the growth of moist convection. The latent heating release from the moist convection can then release gravity waves in order to restore geostrophic flow (Zack and Kaplan 1987).

Moore and Abeling (1988) suggest that the movement of a strong jet streak into a
height ridge can also be a region where both geostrophic and gradient wind balance are not maintained, because as the swiftly moving air parcels reach the height ridge the Coriolis force does not act quickly enough to deflect the parcels to the right and maintain geostrophic balance. This again leads to increase divergence aloft, surface pressure falls and possible moist convection in the same way as Zack and Kaplan (1987).

Kaplan et al. (1997) examined mesoscale gravity wave formation across the inter-mountain West where geostrophic adjustment is hypothesized to be a main source the gravity waves. From coarse resolution rawinsonde observation they find that the ageostrophic winds are pointed toward the cyclonic side of the jet’s lower heights, i. e., in the exit region of their jet across western Idaho. The ageostrophic flow unbalance created by the cyclonically directed ageostrophic wind in the exit region of the jet streak helped to induce a mesoscale jet streak across southwestern Idaho and southern Montana through the low-level return branch of the thermally indirect circulation accompanying the upper level jet streak. This mesoscale jetlet strengthened in the same region as the mesoscale gravity wave and before the mesoscale gravity wave occurred. This secondary mesoscale jet streak, which is the thermally indirect circulation, helped to increase divergence aloft, then a lifting mechanism released moist convection, and subsequent mesoscale gravity waves occurred downstream of the convection and mountains (Kaplan et al. 1997).
2.4 Lower stratospheric gravity waves and satellite data

Gravity waves have long been observed in the troposphere; however, as aircraft flight has increased over the past decade it is important to also note vertically propagating gravity waves that reach into the stratosphere can lead to turbulence (Koch et al. 2005 and references therein). Forty-four turbulence cases were studied, including two aircraft accident cases, where severe turbulence was found in jet exit regions with upward synoptic scale vertical motions, and synoptic scale leftward-directed ageostrophic flow (Kaplan et al. 2003). In several of the 44 cases deep moist convection was located near juxtaposed jet streaks and in these cases observed turbulence was more prevalent.

In an idealized numerical modeling study by Pandya and Alexander (1999) thermal forcing perturbations were introduced and studied to see the interactions at the tropopause of vertically propagating gravity waves. As the convective clouds grew into the lower-stratosphere the stratiform anvil caused a low-stability layer in the lower stratosphere. The low-stability layer prevented the vertical propagation of the short-wavelength gravity waves into the lower stratosphere as the shorter wavelength modes were most attenuated by the low-stability layer. When the ratio of stratospheric wave variance was compared with the horizontal wavenumber the longer wavelength gravity waves were able to penetrate into the lower stratosphere without attenuation through the low stability layer, while the shorter wavelengths were attenuated. The dominant horizontal wavelength was between 10 and 100 km in these idealized simulations agreeing
with linear theory for this case. Gravity waves in lower stratosphere with these wavelengths and moist convection will be further examined in Chapter 6.

Another idealized study examined a vertically propagating gravity wave that was caused by the updrafts of deep convection in the tropics (Lane et al. 2001). Sources of these gravity waves became highly variable and depended mainly on the strength of the updraft of the convective cell. However, the gravity waves once in the stratosphere had small amplitudes causing vertical velocity fluctuations of 0.3 m s\(^{-1}\). Generation of gravity waves in the lower stratosphere was greatest once the anvil or upper part of the updraft reached the tropopause as the updraft overshoot its equilibrium level. The stable air around the overshooting top was being displaced laterally and downward around the vertically moving column of air allowing for mixing and the buoyancy and momentum flux terms to be significant in turbulence generation. The vertical acceleration perturbation values also increased as updraft protruded above the tropopause and gravity waves are observed within the potential temperature perturbation fields in the lower stratosphere. As individual convective updrafts exceeded their equilibrium level within the moist convection, idealized gravity waves were observed in the lower stratosphere by Lane et al. (2001) and the lower stratospheric gravity waves associated with the convective updrafts will be further examined in Chapter 6.

Lane et al. (2003) noticed that out-of-cloud turbulence, especially in the stratosphere, was related to the vertically propagating gravity waves breaking and turbulence kinetic energy, which subsequently cascaded into smaller scales.

Before the gravity waves, which propagated upstream away from the convection
broke, they had 6 km horizontal wavelengths, and the gravity waves broke only in a 3 km layer above the convection in the lower stratosphere. Our December 12th case study will be examined to see how far above the moist convection gravity waves are observed between the computer model simulations and satellite observations.

Zhang (2004) simulated an idealized jet-front system where the gravity waves had horizontal wavelengths of 100-200 km and found that increased resolution did not change the wavelengths of the gravity waves in the upper-troposphere. The author speculated that geostrophic adjustment plays a key role in the initiation of mesoscale gravity waves and he diagnosed this by examining the terms of the nonlinear balance equation. The nonlinear balance equation tells how geostrophic balance is disrupted in the environment. Mesoscale gravity waves are often observed in areas in which terms in the nonlinear balance equation are non-negligible. The area of formation for the idealized mesoscale gravity waves was consistent with the Uccellini and Koch (1987) paradigm, however, the mesoscale gravity waves in Zhang (2004) were propagating vertically and horizontally into the stratosphere. A more observationally based analysis of mesoscale gravity waves was done by Koch et al. (2005) where they examined the turbulence event on 17-18 February 2001 as part of Severe Clear-Air Turbulence Colliding with Aircraft Traffic (SCATCAT) using observations from the NOAA Gulfstream-IV (G-IV). They found that gravity waves in the upper-troposphere were initiated near a secondary tropopause fold that formed above a stable layer. This area was conducive for the mixing of tropospheric and stratospheric air masses. Again the
exit region of the jet streak was unbalanced allowing for mesoscale gravity waves to form by geostrophic adjustment in much the same way as the idealized case by Zhang (2004). Koch et al. (2005) also did a wavelet analysis, finding that not only were there gravity waves with wavelength of a few hundred kilometers, but also within those packets they found gravity waves with wavelengths of 1-20 km. This suggests a downward cascade of energy between longer wavelength gravity waves and shorter wavelength gravity waves, as is suggested by L. F. Richardson. This would explain why the G-IV experienced turbulence even though the wavelengths of most gravity waves were observed larger than the aircraft.

Wu and Zhang (2004) used Advanced Microwave Sounding Unit-A (AMSU) data to verify their simulated mesoscale gravity waves in the stratosphere. This study used the raw radiance measurements from polar orbit satellites, such as the AMSU, to infer the perturbations of the air temperature in the stratosphere. Detailed wave-like structures were found in this study from 80 to 2 hPa in the stratosphere. The waves had horizontal wavelengths of 300 to 600 km and vertical wavelengths of 20 to 30 km, and these gravity waves occurred above a strengthening baroclinic jet-front system across the north Atlantic. When this case study was simulated the models qualitatively produced similar gravity waves propagating vertically into the stratosphere. Similar studies by Eckermann and Wu (2005) and Eckermann et al. (2005) have found that the AMSU can detect gravity waves with vertical wavelengths greater than 10 km and horizontal wavelengths of 150 to 200 km with lower stratospheric temperature perturbation amplitudes of 1 to 3 K. Since field experiments can be quite costly these polar
orbiting satellites provide cost-effective and detailed observations of possible
gravity waves vertically propagating into the stratosphere, which can verify case
study model runs.
Fig. 2.1 Schematic depiction of a ducted mesoscale gravity wave with one-half of a vertical wavelength contained between the ground and a critical level. (a) Vertical cross section in the direction of wave propagation, showing wave-induced horizontal and vertical wind motions (arrows), streamlines or isentropes (solid lines), and the critical level (dashed), for a wave that is propagating with intrinsic phase speed $C^*$ faster than the winds in the duct layer. (c) Wave-induced surface pressure perturbations ($p'$) and wind perturbations in the direction of wave propagation ($u'$) drawn for the same wave segment shown in (a). Source: Koch and O’Handley (1997).
Fig. 2.2 Schematic diagram of the model atmosphere. Region 1 is the wave duct, region 2 the “black box” reflector, and region 3 is a semi-infinite propagating region. Source: Lindzen and Tung (1976).
Fig. 2.3 Schematic representation of typical environmental conditions for mesoscale disturbances from Marks (1975). The equivalent potential temperature (θ_e) profile is on the left and the mean velocity profiles are on the right. Source: Lindzen and Tung (1976).
Fig. 2.4 (a) analyzed wind vectors at grid points (whole barb = 10 ms$^{-1}$), 50 ms$^{-1}$ isotach (heavy line), contoured geopotential height field (CI = 60 dam), and depictions of height ridge axis (dotted line) and short-wave disturbances (dashed lines); (b) analysis of geostrophic wind vectors (ms$^{-1}$) and 50 ms$^{-1}$ isotach; (c) analysis of ageostrophic wind vectors (ms$^{-1}$) and isotach of ageostrophic winds in excess of 10 ms$^{-1}$ directed toward lower heights; (d) computed Rossby number Ro > 0.5. Heavy stippling shows subregions that also meet criterion for direction of ageostrophic wind. Source: Koch and Dorian (1988).
Fig. 2.5 (top) Schematics of the evolution of the dry air mass (dark shading) and its relationship to convection (light shading), fronts, and the surface cyclone based on analyses at (a1) 1400 and (a2) 2100 UTC. (bottom) Conceptual model of circulations within the leading edge of the dry air mass, the gravity wave, and the convection. The background shading is the radar reflectivity. The foremost light shading is the dry air mass. The darkest shading is the stable layer below the warm frontal surface. Black arrows denote schematically the circulations both within the dry air mass and ahead of it. The white arrows denote the sense of vertical motion. The dashed lines denote zero vertical velocity. Letters C and D denote convergence and divergence, respectively; H and L denote regions of high and low perturbation pressure, respectively. Source: Yang et al. (2001).
Fig. 2.6 Cross section of potential temperature (bold solid lines), vertical velocity (dashed lines), and tangential ageostrophic wind component (dash-dot lines) at 1730 UTC for simulation 2. Source: Zack and Kaplan (1987).
3. Methods

3.1 Analyses methods and observations

In an attempt to better understand the mesoscale gravity waves that formed on December 12th and 13th, 2002 a mesoscale numerical model will be used to simulate the case. Also, a large-array of observations were fortuitously located along the path of the mesoscale gravity waves and these observations will be examined and compared with the mesoscale numerical model. The observations that were employed include: Automated Surface Observing System (ASOS) data, upper-air soundings, Doppler radars, GPS water vapor interface, and satellite. While NCEP reanalysis was used for the initialization of the NHMASS version 6.3 for our case study, data from both the North American Regional Reanalysis (NARR) and Rapid Update Cycle (RUC) was compared with the output from our NHMASS version 6.3 model runs. Both the NARR and RUC data provide key insights into the atmospheric characteristics where some observational data is missing. Level III radar data from the National Climatic Data Center (NCDC) will be used to show the precipitation patterns associated with the mesoscale gravity waves in space and time, and when compared with the Real Time Water Vapor Interface data from NOAA/GSD Ground-Based GPS Meteorology (GPS-MET) provide a very powerful observational tool. Upper-air sounding data was compared with the model vertical sounding data for a comparison of the NHMASS runs with the observations and to diagnose the
vertical shear in the atmosphere with the passage of the mesoscale gravity wave. ASOS was investigated to find the correct position for the frontal boundary extending along the Gulf Coast east of the surface low-pressure center. AMSU satellite data was examined in the manner described by Eckermann et al. (2005) to ensure that there were observations of vertically propagating gravity waves in the lower stratosphere during the event. Finally, various satellite data sources including: AMSU, MODIS, and GOES-12 were very helpful in detecting the different patterns that emerge with a mesoscale gravity wave both in the troposphere, but also the gravity waves in the lower stratosphere. Chapter 4 will go into further details on the observed mesoscale gravity waves on December 12th and 13th, 2002 using these observational tools.

### 3.2 Numerical experiment overview

The numerical model choice in this case study was the Non-Hydrostatic Mesoscale Atmospheric Simulation System (NHMASS) version 6.3 (Kaplan et al. 2000; MESO Inc. 1994). This version of the model has mixed-phase moisture physics (Lin et al. 1983; Rutledge and Hobbs 1983), 1.5-order Turbulent Kinetic Energy (TKE) PBL physics (Therry and Lacarrerre 1983), Sigma-p terrain-following vertical coordinate system, and the outer grid domains also used the Kain-Fritsch cumulus parameterization scheme (Kain and Fritsch 1993). Table 3.1 has a more detailed explanation of the details of NHMASS version 6.3.
The initial coarse grid used in this study had horizontal resolution of 18 km. One-way nesting of the model output was used for higher horizontal resolution runs including 6 km, 2 km, 667 m, 222 m, and 71 m horizontal grid spacing. All model runs performed for this research are summarized in table 3.2. Fig. 3.1a shows the model domain for the 6 km, 2 km, 667m, and 222m resolution runs and Fig. 3.1b covers the 667m, 222m, and 71m resolution runs. Most of the Midwest and southward into Mexico is covered by the 18 km run so that there should be as little model boundary problems with the Rocky Mountains or Gulf of Mexico with the target area in eastern Texas and Louisiana. For the Air Force project, which has funded this research, an auto-nesting code was developed to find areas of larger turbulence potential, which leads to automatically starting the next finer resolution model run over that area and this code is further described in Kaplan et al. (2005). This auto-nesting capability will allow for a “hands-off” approach with the model, which could save valuable time and resources, and provide a valuable asset to the Air Force. Several different parameters were tested for this auto-nesting code and those parameters will be discussed in detail in Chapters 5 and 6. NCEP reanalysis data was used for initialization of the 18 km run, and then one-way nesting was used for finer horizontal resolution model runs. Upper-air and surface data was combined with the NCEP reanalysis data to provide for the best data for the model. Further descriptions of the NCEP reanalysis data can be found in Kalnay et al. (1996). NWS ETA model analysis data was also utilized for model initialization, however, the model runs using this source performed quantitatively and qualitatively much poorer than the model
runs using the NCEP reanalysis data. To study gravity waves in the lower stratosphere the NHMASS version 6.3 has 90-sigma levels from the surface to 10 hPa so that the upper boundary is far removed from the lower stratosphere where gravity waves can form. Two different versions of the 90-sigma level vertical structure scheme were used to find which one would better represent the real atmosphere and not reflect evanescent gravity waves back into the simulated atmosphere. The first vertical structure scheme had 60 smooth sigma levels with an addition 30 added on top. Therefore there was a discontinuity between the 60th and 61st sigma level. The second vertical structure scheme has a smooth 90-sigma level vertical structure that took care of the previous discontinuity.

3.3 Numerical sensitivity tests

Several sensitivity tests were performed using the NHMASS version 6.3 over the entire domain on the 18 km, 6 km, and 2 km horizontal gridscale runs [Table 3.2]. These sensitivity tests were done to examine the role that latent heating plays in this case study. Since convection seems to be a large contributor to the development of the mesoscale gravity wave(s) in this case, then eliminating the convection by turning the latent heating on and off would allow us to see what kind of role convection could have. To remove the latent heating from the NHMASS version 6.3 the terms involving evapotranspiration in the thermodynamic energy equation were omitted similar to Kiefer (2005). Also, the
parameterization schemes for grid-scale and convective scale precipitation were
turned off at all horizontal resolutions for the sensitivity tests.
Table 3.1 NHMASS model (version 6.3) characteristics

**MODEL NUMERICS**
- Non-hydrostatic primitive equation model with hydrostatic option
- Cartesian grid on polar stereographic map
- Vertical coverage from ~10m to 29600m
- Fourth-order horizontal space differencing on an unstaggered grid
- Time-dependent lateral boundary conditions
- Massless tracer equations for ozone and aerosol transport
- 3-D equations for u, v, T, q, and p
- Sigma-p terrain-following vertical coordinate system
- Energy-absorbing sponge layer near top of domain
- Split explicit time integration schemes (a) forward-backward for the gravity mode and (b) Adams-Bashforth for the advective mode
- Positive-definite advection scheme for scalar variables

**INITIALIZATION**
- First guess/lateral boundary conditions from NCEP Reanalysis data for 18-km run
- First guess from next larger-scale simulation for 6-km and downscale nests
- High resolution terrain database derived from observations
- High resolution satellite or climatological sea surface temperature database
- High resolution land use classification scheme
- High resolution climatological subsoil moisture database derived from antecedent precipitation
- High resolution normalized difference vegetation index

**PBL SPECIFICATION**
- Surface energy budget
- Atmospheric radiation attenuation scheme
- 1.5-order Turbulence Kinetic Energy PBL parameterization
- Soil hydrology scheme

**MOISTURE PHYSICS**
- Grid-scale prognostic equations for cloud water and ice, rainwater, and snow (Lin et al. 1983; Rutledge and Hobbs 1983)
- Kain-Fritsch (1993) convective parameterization scheme for 18 km, 6 km, and 2 km runs
Table 3.2 Summary of NHMASS Simulations performed

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<th>Grid Spacing (km)</th>
<th>Grid Dimensions (x,y,z)</th>
<th>Duration (hours)</th>
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</tr>
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</tr>
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<td>0.233</td>
<td>None</td>
<td>OLD71M</td>
</tr>
</tbody>
</table>

* DRY runs differ from FULL simulations in that diabatic heat absorption or release due to evaporation and condensation has been omitted. Sensible heating remains intact in all simulations.
Fig. 3.1 Comparison of NHMASS grid meshes: a) 6 km Domain (outer grid), 2 km grid mesh (medium shading), 667 m and 222 m grid meshes (light and dark shading, respectively) b) 667 m Domain (Dark green), 222 m, and 71 m grid meshes (lightest shading).
4. Observational Analysis of Mesoscale Gravity Waves

4.1 Introduction

Gravity wave formation and propagation is a common occurrence throughout the atmosphere. A gravity wave is a wave disturbance in which vertical equilibrium is restored through buoyancy (or reduced gravity) and most gravity waves make very small changes to the sensible weather. However, mesoscale gravity waves, with wavelengths between 50 to 500 km and amplitudes of 0.5-15 hPa, can have a direct impact on the sensible weather (e.g. Wu et al. 2004). Unfortunately, most of these mesoscale gravity waves are very hard to identify and when they are identified are not sampled in enough detail by the various observational sites throughout the country to do a correct analysis.

Fortunately, the mesoscale gravity wave that formed in central and southeastern Texas on December 12, 2002 did pass over several observational sites throughout Texas, Louisiana, Arkansas, and Mississippi and will allow for analyses of its characteristics and its environment. To diagnose the deep tropospheric large-amplitude mesoscale gravity wave both sea-level pressure maps and time series of pressure traces will be used to find areas where greater than 2 hPa per hour changes have occurred with a singular wave of depression in the time series and wave-like features are observed in the sea-level pressure maps north of the frontal boundary and surface cyclone behind the linear convective system. A similar analysis was done with a wave of depression in Lin and Goff (1988) and the mesoscale gravity waves in Zhang et al. (2003).
The importance of mesoscale gravity waves has been well noted in the literature. Moist convection that developed on May 18, 1971 occurred as mesoscale gravity waves were moving northeastward along a stationary front (Uccellini 1975). These thunderstorms produced several tornadoes, strong winds, and hail damage. On February 14th and 15th 1992 severe convection developed ahead of a mesoscale gravity wave and these storms produced hail and wind damage as they progressed through Missouri (Trexler and Koch 2000). A final example can be seen on January 4, 1994 where mesoscale gravity waves formed and propagated up the East Coast as a winter storm was occurring (Bosart et al. 1998). These gravity waves helped to modulate the snowfall rates and caused blizzard conditions in some locations. Therefore a correct analysis of the mesoscale gravity wave and its environment will provide helpful insight that could be used by weather forecasters in forecasting other mesoscale gravity waves in real-time and warning their clients of the possible impending effects on the sensible weather.

Our hypothesis, that the proximity of moist convection in between two jet streaks allowed for juxtaposed unbalanced flow and an area of shearing instability and the formation of a large-amplitude, deep, tropospheric, mesoscale gravity wave, asserts that mesoscale gravity waves form from the interaction of several weather phenomena, which occur at various length scales, and therefore multi-scale analysis tools will be used to diagnose the interaction of these different circulations. For instance, an unbalanced jet streak provides a favorable meso-\(\alpha\) scale background environment for gravity waves to form (e.g. Koch and Dorian...
1988) with the large scale upper divergence and background energetics. Moist convection and its interaction with or within the unbalanced jet streak can provide support for further development of a mesoscale gravity wave through latent heat release, buoyancy perturbations, and the development of critical levels within the background flow allowing for overreflection of wave energy within a duct (Uccellini and Koch, 1987; Lin et al. 1998; Rauber et al. 2001). Finally, shearing instability and low Richardson numbers have been observed in areas in which the mesoscale gravity waves propagate and turbulence generation can occur (Koch et al. 2005). Therefore observations from each of these different scales must be assessed. With a jet streak on the order of 100 to 1000 km, moist convection on the order of 10 to 100 km, and shear instability on the order of 1 to 10 km provides a range of length scale requiring use of all available observations to provide an opportunity to understand the genesis and maintenance mechanisms of mesoscale gravity waves as discussed in Chapter 2.

The remainder of the chapter is organized into 4 sections. Section 4.2 will provide an analysis of the mesoscale gravity wave in the troposphere and the surrounding environment in Texas, Louisiana, Arkansas, and Mississippi. Radar, profilers, surface data, and the zero hour forecast of the RUC (Rapid Update Cycle) and NARR (North American Regional Reanalysis) will be tools used for this observational analysis. Section 4.3 contains observational data of gravity waves in the upper troposphere and lower stratosphere where upper-air soundings and satellite data corroborate the existence of gravity wave patterns. Section 4.4 compares the given observational data with the computer model simulations
employed for this case study in an effort to enhance the spatial and temporal detail of the analyses. Finally, Section 4.5 summarizes this chapter and discusses some conclusions.

4.2 Observations of a tropospheric mesoscale gravity wave

On December 12th 2002 gravity waves formed in the lower troposphere and are first recorded by the GPS pressure trace at both AUS and LDBT shortly after 1200 UTC [Fig. 4.1a and b]. While there are many perturbations in the surface pressure over the next seven hours at LDBT, the mesoscale gravity wave of interest for this case study is a singular wave of depression, an internal solitary wave of depression similar to Lin and Goff (1988) that passes over LDBT at 2000 UTC. The gravity wave component of the pressure falls was initially isolated and remained so from the linear convective system as the mesoscale gravity wave occurred 50 to 100 km behind the linear convective system and it moves northeastward, while the linear convective system propagated eastward. The movement of these two features will be further discussed in Section 4.2.2. After passing over LDBT, the mesoscale gravity wave progressed northeastward into eastern Texas, where it was recorded at the GPS pressure trace from PSN shortly after 2200 UTC [Fig. 4.2a]. The mesoscale gravity wave continued to travel to the northeast and into Louisiana and was next recorded at the pressure trace station at WNFL [Fig. 4.2b]. There were a series of gravity waves between 0200 UTC and 0800 UTC on December 13th, 2002 at WNFL, however, the singular wave of
depression of interest to this case study was the first wave of depression, which occurred shortly after 0300 UTC [Fig. 4.2b]. VIC1 next observed the mesoscale gravity wave at 0500 UTC. The GPS pressure trace at VIC1 contained a 5 hPa drop in 2 hours as the mesoscale gravity wave moved through the city [Fig. 4.3a]. As the mesoscale gravity wave continued to travel northeastward it passed over OKOM at 0800 UTC on December 13, 2002 [Fig. 4.3b]. OKOM is the furthest the mesoscale gravity wave of depression will be examined in this case study. Fig. 4.4 shows the propagation of the mesoscale gravity from central Texas through northeastern Mississippi. It is rare to have such a well-observed mesoscale gravity wave without the effort and resources of a field experiment due to the transient and non-linear nature of most gravity waves.

While it is clear from the pressure traces that there were mesoscale gravity waves on December 12\textsuperscript{th} and 13\textsuperscript{th}, 2002, similar to Lin and Goff’s (1988) wave of depression; the background environment also favored mesoscale gravity wave development since it was similar to the Uccellini and Koch (1987) paradigm. Examining the mesoscale gravity wave synoptic environment for December 12-13, 2002, we find these three features to be present. At 1500 UTC on December 12, 2002 at 300 hPa a jet streak was moving eastward into the western Gulf of Mexico and eastern Texas was in the left exit region of this jet streak [Fig. 4.5a]. Also, in northeastern Texas a small region of higher winds was advecting toward the downstream ridge axis. By 1800 UTC at 300 hPa, eastern Texas and the northwestern Gulf of Mexico were between several different jet streaks. Eastern Texas was in the inflection point of the 300 hPa height field
upstream of the ridge axis [Fig. 4.5b]. Finally, at 1500 UTC there was a surface low-pressure centered near Matagorda Bay, Texas with a warm frontal boundary extending northeastward along the Texas and Louisiana Gulf Coast [Fig. 4.6]. Eastern Texas and Louisiana were therefore conducive for mesoscale gravity wave formation according to Uccellini and Koch (1987). The juxtaposition of the jet streaks near warm frontal boundaries on the synoptic scale provided a favorable environment for mesoscale gravity waves over 12-13 December 2002.

The remaining sections will discuss the development of the unbalanced jet streak across Texas around the time of the first mesoscale gravity wave signals in AUS and LDBT, as well as the role of geostrophic adjustment on the synoptic scale environment. While geostrophic adjustment has been proposed as one of the source mechanism for mesoscale gravity waves (e.g. Uccellini and Koch, 1987), moist convection also played a significant role in this event and will be discussed on the meso-β scale. Finally, shear instability and the location of a possible duct will be examined on the meso-γ scale to find the correlation of the pressure perturbation within the moist convection similar to Yang et al. (2001).

4.2.1 Development of unbalanced jet streaks

In previous work done by Uccellini and Koch (1987) mesoscale gravity waves formed below areas of unbalanced jet streaks. Zhang (2004) and Koch et al. (2005) both examined the exit region of jet streaks and found that mesoscale
gravity waves were observed in these regions as the wavelength between jet
streaks at upper levels decreased and the non-linear balance equation (or
divergence tendency) was not equal to zero. This section will examine the
geostrophic adjustment process occurring above the mesoscale gravity wave
across Texas and show that the upper level pattern was similar to that observed by
Uccellini and Koch (1987), Lin and Goff (1988), Hamilton et al. (1998), and
Zhang (2004). At 0000 UTC on December 12th, 2002, a 140 knot subtropical jet
streak began to move into Mexico from the Pacific Ocean at 200 hPa as an upper-
level trough moved eastward into New Mexico [Fig. 4.7a]. As the subtropical jet
moved into Mexico, at 300 hPa a polar jet streak was continuing to advect further
southward toward the base of the trough in New Mexico [Fig. 4.7b]. By 1200
UTC the subtropical jet streak had moved into northern Mexico and the left-exit
region was located over south central Texas and the Western Gulf of Mexico [Fig.
4.8a]. Based on Keyser and Shapiro (1986) we hypothesize there was an area
upper-level divergence in this region. At the same time the axis of the trough at
300 hPa had progressed into western Texas [Fig. 4.8b]. The jet streaks (and their
exit-regions) as well as the upper-level trough were in the correct location for
mesoscale gravity wave formation to occur according to Uccellini and Koch
(1987). As the mesoscale gravity waves were observed at LDBT at 1800 UTC not
only was eastern Texas in the exit region of a jet streak at 300 hPa, but it was near
the right entrance region of another jet streak that had formed across northeastern
Texas and eastern Oklahoma [Fig. 4.9a]. This second jet streak had formed
downstream of the moist convection similarly to Hamilton et al. (1998) where the
latent heat released by the convection aided in a momentum increase downstream of the moist convection. By 2100 UTC as the mesoscale gravity wave was nearing PSN the area of large upper divergence associated with the exit and entrance regions of the jet streaks at 300 hPa had moved further toward eastern Texas [Fig. 4.9b]. From the RUC analysis at 2100 UTC the winds between the two jet streaks were beginning to flow across the geopotential height contours at 250 hPa in eastern Texas, indicating that the flow at 250 hPa was no longer geostrophic [Fig. 4.10a], as the flow was more geostrophic between 0000 UTC and 1200 UTC [Figs. 4.7b and 4.8b] with the wind barbs pointed parallel with the geopotential height contours within the jet streak. As the trough at 300 hPa moved into eastern Texas by 0000 UTC on December 13th, the upper-level winds over eastern Oklahoma increased to 130 knots [Fig. 4.10b]. The winds at 250 hPa continued to flow across the geopotential height lines in eastern Texas in the region of strong upper divergence in between the two jet streaks [Fig. 4.11a]. As the mesoscale gravity wave was close to WNFL the entire state of Louisiana was in a region of strong upper divergence in between the exit region of a jet streak over the northwestern Gulf Mexico and the entrance region of a jet streak in the Ohio Valley [Fig. 4.11b]. At 300 hPa the trough was beginning to move into northeastern Texas and the wind speed over eastern Oklahoma and western Arkansas had decreased substantially from just 3 hours prior [Fig. 4.12a]. With the mesoscale gravity wave moving into central Mississippi by 0600 UTC, the area of upper-level divergence at 300 hPa continued to be above the mesoscale gravity wave in between the two jet streaks [Fig. 4.12b]. So as the mesoscale
gravity wave formed at 1200 UTC December 12th, it was within the exit region of one jet streak and in the entrance region of a subsequent downstream jet streak similar to Zhang (2004). This favorable upper-level pattern for mesoscale gravity waves (e.g. Uccellini and Koch, 1987) persisted for the next 18 hours as the mesoscale gravity wave moved all the way into Mississippi evacuating large amounts of mass out of the column aloft as can be inferred from the upper-level divergence. This meso-α or synoptic scale pattern agreed with previous studies (e.g. Uccellini and Koch, 1987) as a favorable region for mesoscale gravity wave development.

Now that the location of the jet streaks and the upper level trough were in the correct position for the mesoscale gravity wave to form in Texas, and move northeastward, examining the ageostrophic circulation within the jet streaks and several geostrophic balance indicators will show that geostrophic adjustment could be a mechanism that contributed to the mesoscale gravity wave formation. A jet streak in near geostrophic balance has been shown to have a thermally direct ageostrophic circulation in the entrance region, while the exit region of the jet has a thermally indirect ageostrophic circulation (e.g. Keyser and Shapiro, 1986). The thermally indirect circulation in the exit region causes parcels to rise in the left exit region and turn to the right of the mean flow aloft. This turning to the right causes parcels to decelerate as they travel down isentropic surfaces on the anticyclonic side of the jet streak and in turn returns geostrophic balance as the air parcels are balanced between the pressure gradient force and the Coriolis force. At 1200 UTC 12 December, 2002 a cross section was taken through the exit
region of the jet streak in northern Mexico and the circulation vectors in the left
exit region were moving toward the north between 500 hPa and 150 hPa [Fig.
4.13a]. This would indicate that the exit region of this jet streak over eastern
Texas was having a difficult time being in geostrophic balance as the air parcels
were not following the indirect circulation observed in an idealized jet streak exit
region (Keyser and Shapiro, 1986). The ageostrophic circulation in the exit region
of jet streak was advecting parcels to the left or the cyclonic side of the jet.
Geostrophic adjustment was occurring during this time, however, the inertia
response by the Coriolis force was much slower than the direct response by the
pressure gradient force, so the air parcels continued to turn to the cyclonic side of
the jet. Between 1200 and 1500 UTC a second jet streak developed over
northeastern Texas and eastern Oklahoma from the existence of these leftward
directed parcels [Figs. 4.14a and 4.14b] and the development of moist convection
(to be discussed in 4.2.2). The existence of the cyclonically-directed flow in the
exit region continued to occur at 1500 UTC [Fig. 4.13b] with the vectors directed
up the isentropes toward the second jet streak in eastern Texas. At 1800 UTC the
leftward directed vectors were the strongest [Fig. 4.13c] and during the past 3
hours the second jet streak over eastern Oklahoma had increased in speed at 300
hPa by 20 knots [Fig. 4.14c]. Also, note the large increase of negative omega
(upward motion) at 500 hPa in the left exit region across eastern Texas between
1500 and 1800 UTC correlating with a large evacuation of mass and parcel
acceleration in this region [Fig. 4.15a and 4.15b]. The movement of the
cyclonically-directed air parcels aloft in between the two jet streaks continued
over eastern Texas for the next 12 hours and shifted into central Louisiana. By 0600 UTC 13 December, the second jet streak, [Fig. 4.16], once across eastern Oklahoma moved into Missouri and Illinois. The ageostrophic circulation within the exit region of the original jet streak across the northwestern Gulf of Mexico became more thermally indirect with parcels turning toward the right of mean flow and decelerating, facilitating a return to geostrophic balance [Fig. 4.17].

While the ageostrophic circulations within the exit region of the jet streak acted to restore geostrophic balance, several parameters including the Rossby radius of deformation, Lagrangian Rossby number, and non-linear balance equation (NBE) indicated an imbalance in the flow across eastern Texas and the western Gulf of Mexico. These three parameters will be calculated and discussed over the next two chapters.

When the horizontal scale of the disturbance is smaller than the Rossby radius of deformation, then the pressure field will adjust to the momentum field (Matsumoto, 1961). For a stratified fluid the Rossby radius of deformation is given by (Blumen, 1972),

\[
L_N \sim \frac{NH}{f}
\]  

(4.2.1)

and for this case study \(N = 0.02 \text{ s}^{-1}\), \(H = 10 \text{ km}\), and \(f = 10^{-4} \text{ s}^{-1}\), giving \(L_N = 2000 \text{ km}\). The horizontal wavelength of the mesoscale gravity wave in this case study is 200 km, which is well below 2000 km, and therefore the pressure field should adjust to the rapidly changing momentum field on the way to the final balanced state.
The next parameter used to diagnose an imbalance in geostrophic flow in the exit region of a jet streak was the presence of large Lagrangian Rossby numbers (Koch and Dorian 1988) with

\[
\text{Ro}_L = \frac{|DV / Dt|}{|fV|}. \tag{4.2.2}
\]

\(\text{Ro}_L\) greater than 0.5 occurs when the atmosphere is out of geostrophic balance in the exit region of a geostrophic wind maximum (Koch and Dorian 1988). Both Egentowich (2000) and Rozumalski (1997) determined 0.5 to also be the critical \(\text{Ro}_L\) value for large geostrophic imbalances above mesoscale gravity waves, finding \(\text{Ro}_L\) greater than 1 or 2 in some locations of the exit regions of their jet streaks they studied. In the exit region of the jet streak over eastern Texas at 300 hPa the \(\text{Ro}_L\) was below the critical level and near a value of 0.3 at 1200 UTC 12 December [Fig. 4.18a]. By 1400 UTC the \(\text{Ro}_L\) had grown to near 0.9 at 300 hPa in the same region indicating unbalanced flow above the mesoscale gravity wave [Fig. 4.18b]. The unbalanced flow continued in the exit region of the jet streak over the northwestern Gulf of Mexico at 0000 UTC 13 December with \(\text{Ro}_L\) above 2 and close to the value Egentowich (2000) and Rozumalski (1997) observed [Fig. 4.18c].

The non-linear balance equation (NBE) is the final diagnostic that will be used to assess the imbalance of flow in the area of the mesoscale gravity wave formation over Texas. The NBE can be written as

\[
0 = f\zeta + 2J(u,v) - \nabla^2 \phi - \beta u, \tag{4.2.3}
\]
where $f$ is the Coriolis parameter, $\zeta$ is the relative vorticity, $J$ is the Jacobian operator, $\phi$ is the geopotential height, and $\beta$ is the meridional variation of $f$. The four terms of NBE should be close to zero or negligible when the upper-level jet streak is in geostrophic balance (Paine et al. 1975). NBE values of $4 \times 10^{-8}$ s$^{-2}$ were found by Rozumalski (1997) in the development of a cyclone in the exit region of a mesoscale jetlet and Zack and Kaplan (1987) found large increases in upper-level divergence when the NBE was not equal to zero. Similar values of NBE were found across Texas, Louisiana, and Mississippi throughout this event and will be discussed further in Chapter 5. The $R_{0L}$, Rossby radius of deformation, and NBE confirm the magnitude of unbalanced flow at upper-levels during this mesoscale gravity wave event similar to Zhang (2003). This allows a large amount of upper-level divergence to develop along with compensating low-level ageostrophic flow.

### 4.2.2 Mesoscale gravity wave movement and moist convection

As discussed in the previous section, there was a large magnitude of flow imbalance on the meso-\(\alpha\) scale and while this imbalance established a favorable upper-level environment for wave genesis, the observed meso-\(\alpha/\beta\) processes, including the position and movement of the surface cyclone, surface front, and moist convection, were also important in the redistribution of pressure and momentum during the formation of the mesoscale gravity wave over Texas. As
the region of upper-level divergence in the exit region of the jet streak moved into
the western Gulf of Mexico a surface cyclone began to form southeast of CRP at
1000 UTC on December 12th [Fig. 4.19a]. To the north and east of the surface
cyclone a stationary frontal boundary extended into Louisiana with a stable
boundary layer for the mesoscale gravity wave to move along just north of the
stationary front. East of the surface cyclone there was a large amount of
conditional instability indicated by the CAPE values exceeding 2000 Jkg\(^{-1}\) [Fig.
4.19b]. By 1400 UTC the surface cyclone had deepened 3 hPa and had moved
northeastward along the stationary boundary and now was in the northwestern
Gulf of Mexico directly east of CRP [Fig. 4.20a]. With the conditional instability
in eastern Texas having increased in the past four hours [Fig. 4.20b] to over 1500
Jkg\(^{-1}\), sufficient low level moisture over the Gulf of Mexico and a lifting
mechanism along the front, moist convection had developed in and around AUS
and was developing northward [Fig. 4.21a]. There was also a corresponding
pressure rise after the rain moved over the AUS GPS pressure trace, stabilizing
the boundary layer environment [Fig. 4.1a]. As the moist convection increased in
coverage after 1400 UTC the latent heating from the convection was allowing for
the redistribution of the momentum field similar to Hamilton et al. (1998) as the
second jet streak downstream of the convection beings to strengthen [Fig. 4.9a].
As discussed in Section 4.2.1 the upper-levels of the atmosphere were no longer
in geostrophic balance in eastern Texas allowing for upper-level divergence and
evacuation of mass and momentum above the moist convection. This region of
convective outflow enhanced the background state of flow imbalance. The
feedback from a favorable upper level environment aided also in increasing the coverage and intensity of the convection to the northwest of the surface cyclone was seen from the AUS radar at 1504 UTC [Fig. 4.21b] by increasing the scale and magnitude of upward vertical motion. Many small amplitude pressure perturbations that were possibly gravity waves are seen in the GPS pressure trace from LDBT between 1400 and 1800 UTC [Fig. 4.1b], and the radar from AUS indicated short intense banded features within the convection propagating and moving eastward as the convection began to organize into a squall line [Figs. 4.22a and 4.22b]. The largest pressure fall associated with the mesoscale gravity wave began to affect LDBT shortly after 1800 UTC [Fig. 4.1b], the corresponding radar signature [Fig. 4.22c] continued to show banded features within the convection and a small more intense band at the back edge of the precipitation. This feature provided the corresponding pressure rise shortly after 2000 UTC in LDBT. The surface cyclone had progressed to the northeast along the stationary boundary by 1800 UTC [Fig. 4.23a] to just east of Matagorda Bay, and the conditional instability in the warm sector and along the stationary boundary continued to be relatively high with CAPE values over 1500 Jkg⁻¹ [Fig. 4.23b]. At 1800 UTC the squall line along the stationary boundary was nearing HOU [Fig. 4.24a] with the mesoscale gravity wave still back near LDBT. As the squall line matured by 2000 UTC the banded features could again be seen behind the convection to the west of HOU [Fig. 4.24b], and at this time the pressure rose behind the wave of depression in LDBT was beginning to occur, associated with a banded feature and enhanced precipitation [Fig. 4.1b]. The surface cyclone had
moved just east of GLS and slightly intensified [Fig. 4.25]. After this period the mesoscale gravity wave continued to travel northeastward into Louisiana north of the stationary boundary, while the surface cyclone moved along the Gulf Coast in Louisiana with the squall line. The magnitude of the singular wave of depression associated with the mesoscale gravity wave had increased between the 2000 and 2200 UTC before passing through PSN around 2200 UTC [Fig. 4.2a]. A reflection of this intensification was seen in the precipitation from the HOU radar as the banding feature located near LDBT at 2000 UTC intensifies as it moved eastward toward PSN [Figs. 4.26a, b, c, and d]. This enhanced banded precipitation region continued to travel northeastward into Louisiana by 2300 UTC [Fig. 4.26e]. At 0000 UTC on December 13th, the surface cyclone was located south of LFT along the Gulf Coast [Fig. 4.27] with the initial squall line continuing to propagate eastward along the stationary boundary near MSY [Fig. 4.28]. The enhanced precipitation band with the mesoscale gravity wave was located along the Texas/Louisiana border at 0000 UTC on the 13th [Fig. 4.29a] to the northwest of the surface cyclone. Again the banded feature with the mesoscale gravity wave intensified and another squall line began to form across central Louisiana between 0000 and 0230 UTC [Figs. 4.29b, c, d, and e]. The strongest pressure fall associate with the mesoscale gravity wave occurred after 0400 UTC at WNFL [Fig. 4.2b], behind the main squall line which was moving eastward but ahead of a second banded feature [Fig. 4.29f] which was moving northeastward. The mesoscale gravity continued to move northeastward with the same structure into Mississippi with the squall line ahead of a secondary enhanced precipitation
band and the mesoscale gravity wave [Figs. 4.29g, h, and I]. The squall line created a sharp pressure rise at VIC1 after 0500 UTC, but then another wave of depression occurred when the secondary band reached VIC1 shortly after 0600 UTC [Fig. 4.3a]. These features continued to travel northeastward toward OKOM (not shown).

The meso-β scale structure remained consistent throughout this mesoscale gravity wave event with the moist convection, surface cyclone, and front moving eastward, while the mesoscale gravity wave formed and moved with a northeastward component around 100 km behind the squall line. Moist convection began to occur shortly after 1000 UTC on December 12th under the growing divergence in the left exit region of the upper-level jet streak. A surface cyclone strengthened and moved along a stationary boundary located in the northwestern Gulf of Mexico. As the surface cyclone intensified the moist convection began to organize into a squall line. To the northwest of the cyclone banded features with enhanced precipitation began to occur behind the main squall line. These banded features were co-located with the mesoscale gravity wave as it moved from central Texas into Louisiana and Mississippi with coherent structure.

4.2.3 Shear instability

Observations of the meso-α and meso-β structure have revealed the importance of an unbalanced jet streak and moist convection in the case of this mesoscale gravity wave, however, the observations of the meso-β/γ structure will
aid in a more complete understanding of the combination of geostrophic adjustment, moist convection, and possible maintenance mechanisms. While the observations in this case are very coarse and do not contain Richardson number, which is necessary to make an analysis of shear instability directly and other maintenance mechanisms, upper-air soundings, similar to Lin and Goff (1988), will be used to examine the vertical structure of the atmosphere as the mesoscale gravity wave passes in an effort to make a connection between the vertical structure of the momentum field (e.g. Trexler and Koch 2000).

In Lin and Goff (1988) the temperature profiles at nine stations were examined around the time of the passage of the singular wave of depression. Fig 4.30 shows the vertical structure and all the nine stations have shallow inversion layers below 500 hPa. Above 500 hPa each of the temperature soundings has a layer in which the lapse rate is close to dry-adiabatic. In Lindzen and Tung (1976) they discuss the importance of a wave ducting mechanism or critical level for the maintenance of mesoscale gravity waves. Fig. 4.31, from Koch and O’Handley (1997), provides an example of the temperature profile necessary for wave ducting and wave maintenance as discussed in Chapter 2. Again there is a shallow inversion layer below 700 hPa and then a layer with a dry-adiabatic lapse rate above that. Notice also the vertical profile of winds and how just below 700 hPa the winds change direction and speed rapidly, which suggest the possibility of a critical level at 700 hPa in this idealized sounding. Employing a representative sounding from Dallas-Fort Worth, Texas (DFW) since it is north of the surface frontal boundary at 1200 UTC, the height of the inversion was inferred to be
between 2 and 3 km [Fig. 4.32]. Above this inversion was dry-adiabatic layer similar to the Lindzen and Tung (1976) for wave ducting [Fig. 4.32]. The vertical wind structure also changed rapidly below and above the inversion at 2.5 km, going from south-southwesterly and 20 knots to west-southwesterly at 30 to 50 knots indicating a large amount of wind shear was occurring between the inversion layer and the dry-adiabatic layer. A similar wave ducting pattern was visible at 1200 UTC at Shreveport, Louisiana (SHV) with an inversion between the surface and 2 km [Fig. 4.33] and between the inversion layer and the more unstable layer above it the wind again changed direction and speed from an easterly component to a southwesterly component. Another sharp inversion was seen at 3 km at SHV at 1200 UTC [Fig. 4.33] and it is possible that this higher inversion also aided in the wave ducting mechanism during this mesoscale gravity wave event, however, at the time of this sounding the mesoscale gravity wave was still several hundred kilometers to the southwest of SHV, therefore it is difficult to surmise that the midlevel inversion would remain (at 3 km), however, the low-level inversion should remain as long as the surface frontal boundary remains south of SHV and this lower level inversion would provide a more sustainable wave duct mechanism similar to Koch and O’Handley (1997). With the temperature profiles and vertical wind structures in this mesoscale gravity wave even being similar to Lin and Goff (1988) and their singular wave of depression it suggest that the mesoscale gravity wave from December 12th could have been maintained by a wave ducting mechanism.
In the previous three sections of this chapter several observational sources have been used to examine the formation of an environment favorable for a mesoscale gravity wave. Propagation of the mesoscale gravity wave from its initial stages in central Texas, through its mature stages in Louisiana was shown in detail. A final examination of observations included observing the vertical structure located in and around the mesoscale gravity wave. It is evident in the case of the mesoscale gravity wave on December 12th and 13th, 2002 that several key gravity wave source and maintenance mechanisms were present across Texas, Louisiana, and Mississippi and these include: moist convection, geostrophic adjustment, unbalanced flow and its non-linear interactions, and a wave duct, and these mechanisms play a key synergistic role in the formation, propagation, and maturation of the mesoscale gravity wave across Texas, Louisiana, and Mississippi.

4.3 Stratospheric mesoscale gravity waves

Development of a better forecasting method for stratospheric turbulence is essential for aviation safety and while there was a large-amplitude mesoscale gravity wave occurring the troposphere there were also gravity waves in the lower stratosphere. Gravity waves in the stratosphere have been found to occur near and above moist convection, near jet streaks, and above steep orography (e.g. Wu and Zhang, 2004). On 12 and 13 December, 2002 both a strong jet streak and moist convection were observed across the southern United States. The following two
sections will discuss the observations of the growth of moist convection and stratospheric gravity waves by first using radar, then by using the method developed by Eckermann and Wu (2005) and Eckermann et al. (2005) for observing stratospheric gravity with the Advanced Microwave Sounding Unit (AMSU).

4.3.1 Development of moist convection

At 0700 UTC 12 December moist convection began to develop in the western Gulf of Mexico, and this area of convection was associated the exit region of the jet streak across northern Mexico and the development of surface cyclone in the western Gulf of Mexico as previously discussed in Section 4.2. The area of thunderstorms developed in the warm sector east of the surface cyclone by 1200 UTC with the strongest cells located east of CRP [Fig. 4.34]. A sounding from BRO at 1200 UTC had over 1400 Jkg\(^{-1}\) of CAPE with a LI value of -6.2 [Fig. 4.35]. An analysis of the CAPE from the RUC at 1100 UTC was even more informative with higher CAPE values over 2000 Jkg\(^{-1}\) in the western Gulf of Mexico [Fig. 4.36] and the positive CAPE values continued north and westward to the warm front along the Texas Gulf Coast. The developing surface cyclone and accompanying cold front were providing a lifting mechanism to release this instability resulting in the growth of the thunderstorms. Overshooting tops penetrating into the lower stratosphere can be seen in Fig. 4.37 by 1640 UTC as the area of moist convection began to organize into a squall line. The squall line
had matured into an MCS by 2000 UTC and began to advect eastward into southern Louisiana [Fig. 4.24b]. While all these observations are not direct measurements of turbulence or gravity waves, the observations of overshooting tops with and around the moist convection are co-located with the AMSU observations of stratospheric gravity waves in Section 4.3.2.

4.3.2 AMSU observations

A more direct way of measuring turbulence and gravity waves in the lower stratosphere has just been developed using the Advanced Microwave Sounding Unit (AMSU) data. Eckermann and Wu’s (2005) methodology consists of taking the observed microwave radiance values from the cross-track beams of the AMSU (available on the NOAA-15 through NOAA-18 weather satellites) and calculating the brightness temperatures in the stratosphere. Radiance perturbation values can then be obtained and gravity waves with horizontal wavelengths of 150 to 200 km can be observed (e.g. Wu and Zhang, 2004).

On 12 and 13 December, 2002 mesoscale gravity wave like features in the lower stratosphere could be observed using the AMSU observations. Fig. 4.38a shows large amounts of perturbations in the radiance fields and the radiance perturbation have been correlated to vertically propagating gravity waves by Wu et al. (2005). On 12 December above the moist convection wave signals were seen at 1300, 1600, and 2000 UTC, at 200 hPa or 12 km, the same level as the overshooting tops in Fig. 4.37 and above the equilibrium level from the
Brownsville, Texas upper air sounding [Fig. 4.35]. Gravity wave signals in the lower stratosphere were again seen at 150 hPa and they were strongest at 1300 and 1600 UTC [Fig. 4.38b]. The same vertically propagating gravity waves were seen at 90 hPa above the area of moist convection in the northwestern Gulf of Mexico at all three times of 1300, 1600, and 2000 UTC [Fig. 4.38c]. The signal of vertically propagating gravity waves in the lower stratosphere decreased by the time the waves reach the 50 hPa level as the gravity waves were only seen at 1300 UTC and at a smaller magnitude at 2000 UTC [Fig. 4.38d]. Above 50 hPa, there are no signals of vertically propagating gravity waves in the lower stratosphere on December 12th using the AMSU observations. On December 13th, as the moist convection had moved into the northern Gulf and southern Louisiana, gravity waves in the lower stratosphere are again observed at both 90 and 150 hPa at 0030 and 0330 UTC [Figs. 4.39a and b] with wavelike features in the radiance perturbation fields. The vertically propagating gravity waves above the moist convection on the 13th were not as strong as the 12th as there was no signal of gravity waves above 90 hPa.

Using radar and satellite observations on December 12th and 13th, 2002 features assumed to be vertically propagating gravity waves in the lower stratosphere have been observed near and above the moist convection. As the moist convection expanded in aerial coverage and traveled toward the east and northeast so did the signal of vertically propagating gravity waves and turbulence. Further analysis of the environment in which these vertically propagating gravity waves exist will be examined through computer modeling in Chapter 6.
4.4 Comparison of MASS simulation results with observations

The 18 km NHMASS run (MASS18) simulating sea-level pressure, precipitation, and 300 hPa features will be compared with synoptic observations in this section to provide credibility to it and accordingly to finer-scale MASS model nested grid simulations. Sea-level pressure, precipitation, and the 300 hPa upper-level pattern were chosen, because they are three of the most representative features of the environment of the mesoscale gravity wave in the lower troposphere as discussed in the previous sections. At 1000 UTC 12 December, 2002 at 300 hPa the axis of the upper-level trough was in western Texas in both the MASS18 and observations [Figs. 4.40a and b]. A closer examination of the 300 hPa flow does show that the observed upper-level trough was 50 km further to the east than MASS18, however, it is common for mesoscale models to have a time lag relative to nature. By 2100 UTC, after the mesoscale gravity wave had been initiated and was moving toward the northeast, both MASS18 and the synoptic observations had the mesoscale jet streak in the same location over Oklahoma with the same magnitude [Figs. 4.41a and b]. MASS18’s location of the left exit region of the jet streak propagating through northern Mexico, as well as, the negative tilt of the main trough, were also very similar to the synoptic observations. Both sea-level pressure fields, from MASS18 and synoptic observations at 1400 UTC on December 12th, were similar indicating a developing surface cyclone to the east of Corpus Christi, Texas [Figs. 4.42a and
b]. The stationary frontal boundary was located in the same spot along the northwestern Gulf of Mexico and this boundary provided an important focusing mechanism for the moist convection and mesoscale gravity wave. By 2100 UTC the MASS18 surface cyclone had progressed inland while the synoptic observations clearly indicate that the surface cyclone did not progress inland [Figs. 4.43a and b]. Also, the magnitudes of the two surface cyclones differed with the MASS18 surface cyclone being 3 hPa deeper. While the evolution of the mesoscale gravity wave event was similar in both the MASS18 and the observations, the time lag between the two events causes problems for direct comparisons at the surface after 2100 UTC.

The time lag error can again be seen in the precipitation fields. At 1400 UTC the MASS18 simulated the convection north and along the stationary boundary correctly, it was missing the squall line development further to the southeast of the surface cyclone [Figs. 4.44a and b]. By 2100 UTC the squall line feature had developed to the east of the mesoscale gravity wave, however, the MASS18 simulated squall line was in southeastern Texas, while the synoptically observed squall line had already moved into Louisiana [Figs. 4.45a and b]. While there is a time lag error, which is common in mesoscale models, the evolution and development of the observed mesoscale event is correctly simulated within MASS18 and therefore will be examined to further study the initiation and propagation of the mesoscale gravity wave in the lower troposphere. While the results from the mesoscale model are credible, they are still no substitute for direct measurements and observations.
4.5 Summary and conclusions

Chapter 4 summarized the development and movement of the mesoscale gravity waves in the lower troposphere and in the lower stratosphere from an observational standpoint. The lower tropospheric mesoscale gravity wave system was found to have a similar synoptic location as proposed by Uccellini and Koch (1987) as the mesoscale gravity formed in the inflection point in the height field at 500 hPa, the exit region of a jet streak at 300 hPa, and north of a surface frontal boundary. Observations also clearly show gravity waves propagating vertically into the lower stratosphere above the convection. This is similar to the observations described in Wu and Zhang (2004) where they observed vertically propagating gravity waves in the lower stratosphere in the exit region of a jet streak in the north Atlantic. Three key source mechanisms (moist convection, geostrophic adjustment, and non-linear interactions) for mesoscale gravity waves were observed across Texas and Louisiana and a possible wave duct was observed as a maintenance mechanism which allowed for initiation and propagation of the mesoscale gravity wave in the lower troposphere northeastward from Texas into Mississippi.

The first mechanism, geostrophic adjustment, was inferred from observations in the region of a jet streak becoming unbalanced on December 12th, 2002. As this jet streak moved into northern Mexico and the western Gulf of Mexico it interacted with the developing moist convection. The heating and deep
mass perturbation caused by the moist convection aided the ageostrophic
circulation in the exit region of the jet streak at upper-levels to be directed
cyclonically, instead of anti-cyclonically toward higher heights. The cyclonically
directed parcels continued to accelerate northeastward above the convection and
provided an increase of divergence aloft for further subsequent moist convective
development. The moist convection helped to induce the formation of a second jet
streak across northeastern Texas and eastern Oklahoma similar to Hamilton et al.
(1998) and Kaplan et al. (1998). The secondary jet streak, its circulation, and non-
linear interactions in turn allowed for an increase in divergence aloft above the
growing convection in central Texas. This system was caused by a non-linearly
expansive process, which helped to create a favorable upper-level environment
for the evolution of second mechanism that enhanced gravity wave genesis, i.e.,
moist convection.

The evolution of the moist convection was a key process in the
development of the mesoscale gravity wave as it allowed for: 1) the enhancement
of the secondary jet streak as well as 2) latent heat release and the favorable non-
linear interactions within the environment. As we observed, the moist convection
developed near the surface cyclone shortly after 1000 UTC on December 12th. As
the surface cyclone deepened in intensity the area and intensity of the moist
convection expanded to the east and north of the surface frontal boundaries. By
1800 UTC the moist convection had organized into a squall line along the
stationary boundary across the northern Gulf of Mexico. This squall line moved
eastward into Louisiana, but behind it several small wave-like features were
observed in the radar data. These wavelike feature(s) were assumed to be the mesoscale gravity wave(s) north of the surface frontal boundary as they were correlated in time and space with the surface pressure tendencies at several observational stations and are similar to Lin and Goff’s (1988) observations of their singular wave of depression. The mesoscale gravity wave then helped to initiate another smaller band of showers across north central Louisiana that progress into Mississippi on December 13\textsuperscript{th}. A possible wave duct was also examined through upper-air analysis as a maintenance mechanism to allow the mesoscale gravity wave to travel such a long horizontal distance without being dispersed. The combination of each of these source and maintenance mechanisms and the favorable synoptic environment enabled the development and longevity of the mesoscale gravity wave observed on December 12\textsuperscript{th} and 13\textsuperscript{th}, 2002 across Texas, Louisiana, and Mississippi.

Subsequent analyses of observations of the gravity waves in the lower stratosphere were discussed. A method developed by Eckermann and Wu (2005) was used to calculate radiance perturbation in the lower stratosphere using AMSU data. Features assumed to be gravity waves in the lower stratosphere were then observed on both December 12\textsuperscript{th} and 13\textsuperscript{th} above the moist convection across the northwestern Gulf of Mexico and southern Texas.

Finally, a heuristic validation was done between MASS18 and observations to demonstrate the viability of using a mesoscale model in further chapters to provide the correct evolution and environment to the mesoscale system. The upper-level pattern between observations and the mesoscale model
was very similar, however, the same could not be said of the surface cyclone
development and associated moist convection. The surface cyclone was much
stronger in the mesoscale model than what was observed. The precipitation
development was also quite different with the mesoscale model being two hours
behind in time and space relative to the observed squall line feature. However, the
evolution and environment of the mesoscale system was similar enough in the
mesoscale model to the observation so that further details can be extracted from
the mesoscale model.

Chapter 5’s discussion will employ the NHMASS model to describe
further the meso-α to meso-γ environment surrounding the modeled mesoscale
gravity wave. Further analysis of the descent of the momentum behind the moist
convection will also be examined to determine if it plays a key role in the
mesoscale gravity wave environment and evolution.
Fig. 4.1 NOAA Global Positioning System (GPS) surface pressure trace from a) Austin, TX (AUS) GPS site. Valid 0000 UTC 12 Dec 2002 to 0000 UTC 14 Dec 2002. b) Ledbetter, TX (LDBT) GPS site. Valid 0000 UTC 12 Dec 2002 to 0000 UTC 14 Dec 2002.
Fig. 4.2 NOAA Global Positioning System (GPS) surface pressure trace from a) Palestine, TX (PSN) GPS site. Valid 0000 UTC 12 Dec 2002 to 0000 UTC 14 Dec 2002. b) Winnfield, LA (WNFL) GPS site. Valid 0000 UTC 12 Dec 2002 to 0000 UTC 14 Dec 2002.
Fig. 4.3 NOAA Global Positioning System (GPS) surface pressure trace from a) Vicksburg, MS (VIC1) GPS site. Valid 0000 UTC 12 Dec 2002 to 0000 UTC 14 Dec 2002. b) Okolona, MS (OKOM) GPS site. Valid 0000 UTC 12 Dec 2002 to 0000 UTC 14 Dec 2002.
Fig. 4.4 Map of the propagation of the observed mesoscale gravity wave of depression through Texas, Louisiana, and Mississippi on December 12-13, 2002 taken from NOAA Global Positioning System (GPS) surface pressure traces.
Fig. 4.5 Rapid Update Cycle (RUC) 300 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid: a) 1500 UTC 12 Dec 2002, and b) 1800 UTC 12 Dec 2002.
Fig. 4.6 Surface station plots valid for 1500 UTC 12 Dec 2002 from Plymouth State Weather Center.
Fig. 4.7 NARR a) 200 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for 0000 UTC 12 Dec 2002 and b) 300 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for 0000 UTC 12 Dec 2002.
Fig. 4.8 NARR a) 200 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for 1200 UTC 12 Dec 2002 and b) 300 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for 1200 UTC 12 Dec 2002.
Fig. 4.9 NARR a) 300 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for 1800 UTC 12 Dec 2002 and b) 300 hPa wind barbs (ms$^{-1}$) and divergence (convergence), which is the positive (negative) values, (1x$10^{-5}$ s$^{-1}$) (colored shading) valid for 2100 UTC 12 Dec 2002.
Fig. 4.10 RUC a) 250 hPa isoheights (m) (black lines), wind barbs (m s⁻¹), and isotachs (knots) (colored shading) valid for 2100 UTC 12 Dec 2002 and b) North American Regional Reanalysis (NARR) 300 hPa isoheights (m) (black lines), wind barbs (m s⁻¹), and isotachs (knots) (colored shading) valid for 0000 UTC 13 Dec 2002.
Fig. 4.11 RUC a) 250 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for 0000 UTC 13 Dec 2002 and b) NARR 250 hPa wind barbs (ms$^{-1}$) and divergence (convergence), which is the positive (negative) values, (1x10$^{-5}$ s$^{-1}$) (colored shading) valid for 2100 UTC 12 Dec 2002.
Fig. 4.12 NARR a) 300 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for 0300 UTC 13 Dec 2002 and b) NARR 300 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for 0600 UTC 13 Dec 2002.
Fig. 4.13 NARR circulation vectors in a vertical cross section from 35.5 N; –99.2 W to 22.1 N; –92.9 W, valid for: a) 1200, b) 1500, and c) 1800 UTC 12 Dec 2002.
Fig. 4.13 (Continued)
Fig. 4.14 RUC 300 hPa isoheights (m) (black lines), wind barbs (m s$^{-1}$), and isotachs (knots) (colored shading) valid for: a) 1200, b) 1500, and c) 1800 UTC 12 Dec 2002.
Fig. 4.14 (Continued)
Fig. 4.15 RUC 500 hPa Omega (colored shading) multiplied by –1 so that + is correlated with upward motion and – is correlated with downward motion valid for a) 1500 and b) 1800 UTC 12 Dec 2002.
Fig. 4.16 RUC 250 hPa isoheights (m) (black lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for 0600 UTC 13 Dec 2002.
Fig. 4.17 RUC circulation vectors in a vertical cross section from 34.9 N; –98.7 W to 27.1 N; –88.6 W, valid for 0600 UTC 13 Dec 2002.
Fig. 4.18 RUC 300 hPa Lagrangian Rossby Number valid for: a) 1200, b) 1400, and c) 0000 UTC 13 Dec 2002.
Fig. 4.19 RUC a) mean sea-level pressure (black lines) and surface winds (knots) valid 1000 UTC 12 Dec 2002 and b) surface CAPE values valid 1000 UTC 12 Dec 2002.
Fig. 4.20 RUC a) mean sea-level pressure (black lines) and surface winds (knots) valid 1400 UTC 12 Dec 2002 and b) surface CAPE values valid 1400 UTC 12 Dec 2002.
Fig. 4.21 Base reflectivity at National Weather Service (NWS) Austin, TX Radar (red star) valid a) 1400 and b) 1504 UTC 12 Dec 2002.
Fig. 4.22 Base reflectivity at NWS Austin, TX Radar (red star) valid a) 1557, b) 1655, and c) 1800 UTC 12 Dec 2002.
Fig. 4.23 RUC a) mean sea-level pressure (black lines) and surface winds (knots) valid 1800 UTC 12 Dec 2002 and b) surface CAPE values valid 1800 UTC 12 Dec 2002.
Fig. 4.24 Base reflectivity at NWS Houston, TX Radar (red star) valid a) 1804 and b) 2004 UTC 12 Dec 2002.
Fig. 4.25 RUC mean sea-level pressure (black lines) and surface winds (knots) valid 2000 UTC 12 Dec 2002.
Fig. 4.26 Base reflectivity at NWS Houston, TX Radar (red star) valid a) 2024, b) 2049, c) 2124, d) 2159, and e) 2259 UTC 12 Dec 2002.
Fig. 4.27 RUC mean sea-level pressure (black lines) and surface winds (knots) valid 0000 UTC 13 Dec 2002.
Fig. 4.28 Base reflectivity at NWS New Orleans, LA (red star) Radar valid 0000 UTC 13 Dec 2002.
Fig. 4.29 Base reflectivity at NWS Fort Polk, LA Radar (red star) valid a) 0004, b) 0045, c) 0131, d) 0200, e) 0229, f) 0356, g) 0431, h) 0454, and i) 0506 UTC 13 Dec 2002.
Fig. 4.30 Temperature profiles at 1200 UTC, 6 March 1969 for nine stations projected to a surface intersecting the 543 dm height. Source: Lin and Goff (1988).
Fig. 4.31 Idealized thermodynamic sounding for wave ducting, according to the Lindzen and Tung (1976) linear wave theory. Source: Koch and O’Handley (1997).
Fig. 4.32 Upper-air sounding from Dallas-Fort Worth, TX (FWD) for 1200 UTC 12 Dec 2002. Source: University of Wyoming.
Fig. 4.33 Upper-air sounding from Shreveport, Louisiana (SHV) for 1200 UTC 12 Dec 2002. Source: University of Wyoming.
Fig. 4.34 Base reflectivity composite radar for the Southern Plains from Plymouth State valid 1215 UTC 12 Dec 2002.
Fig. 4.35 Upper-air sounding from Brownsville, TX (BRO) for 1200 UTC 12 Dec 2002 with a CAPE of 1473 J kg\(^{-1}\). Source: University of Wyoming.
Fig. 4.36 RUC surface CAPE values valid 1100 UTC 12 Dec 2002.
Fig. 4.37 Terra MODIS visible satellite image valid 1641 UTC 12 Dec 2002.
Fig. 4.38 Advanced Microwave Sounding Unit (AMSU) satellite observations of blackbody temperature perturbation from positive (warm colors) to negative (cool colors) 0.2 K in channels 7-10 and +/- 1.5 K in channels 11-14 valid 12 Dec 2002 for: a) 200 hPa, b) 150 hPa, c) 90 hPa, and d) 50 hPa.
Fig. 4.38 (Continued)
Fig. 4.39 Advanced Microwave Sounding Unit (AMSU) satellite observations of blackbody temperature perturbation from +/- 0.2 K in channels 7-10 and +/- 1.5 K in channels 11-14 valid 13 Dec 2002 for: a) 150 hPa and b) 90 hPa.
Fig. 4.40 300 hPa isoheights (m) (black lines), wind barbs (ms\(^{-1}\)), and isotachs (knots) (colored shading) valid for 1000 UTC 12 Dec 2002 from a) MASS18 and b) RUC.
Fig. 4.41 300 hPa isoheights (m) (black lines), wind barbs (ms⁻¹), and isotachs (knots) (colored shading) valid for 2100 UTC 12 Dec 2002 from a) MASS18 and b) RUC.
Fig. 4.42 Mean sea-level pressure (black lines) and surface winds (knots) valid for 1400 UTC 12 Dec 2002 from a) MASS18 and b) RUC.
Fig. 4.43 Mean sea-level pressure (black lines) and surface winds (knots) valid for 2100 UTC 12 Dec 2002 from a) MASS18 and b) RUC.
Fig. 4.44 Base radar reflectivity (Dbz) and surface winds (ms\(^{-1}\)) 1400 UTC 12 Dec 2002 from a) MASS18 and b) observed radar summary from Plymouth State.
Fig. 4.45 Base radar reflectivity (Dbz) and surface winds (ms$^{-1}$) 2100 UTC 12 Dec 2002 from a) MASS18 and b) observed radar summary from Plymouth State.
5. NHMASS Simulations of the Lower Tropospheric Mesoscale Gravity Wave Event

5.1 Introduction

The fine scale adjustments within the atmosphere occur continually, and due to this, it is difficult for forecasters to know which adjustment processes will produce an impact on sensible weather. Some of the finer adjustments, source and maintenance mechanisms, can lead to the gravity waves that were observed both in the troposphere and lower stratosphere on December 12th and 13th, 2002 across Texas, Louisiana, and Mississippi. The mesoscale gravity wave in the troposphere directly impacted the sensible weather by enhancing precipitation rates in Texas behind the squall line and aiding in the development of another precipitation feature across Louisiana. Observations discussed in the last chapter showed the mesoscale gravity wave developing in a region of upper-level imbalance, associated with moist convection and co-located with shear instability. While mesoscale gravity waves have been shown to form in areas as described by Uccellini and Koch (1987) and having at least geostrophic adjustment and shear instability present, the initial impulse and smaller driving mechanisms behind mesoscale gravity wave formation have been addressed by very few people on the finest scales (e.g. Yang et al. 2001). This chapter will attempt to address the initiation and evolution of the mesoscale gravity wave event in the troposphere on 12 and 13 December, based in large part on inferences from simulations. Again, diagnosing the deep tropospheric large-amplitude mesoscale gravity wave in the
simulations will be done using both sea-level pressure maps and time series of pressure traces. This will include diagnosing areas where greater than 2 hPa per hour changes have occurred with a singular wave of depression (e.g. Lin and Goff, 1998) in the time series and wave-like features are observed in the sea-level pressure maps (Zhang et al. 2003) north of the frontal boundary and surface cyclone.

Within this chapter the mesoscale gravity wave system will be addressed from the larger scales of a favorable background environment to the smaller scales of the vertical motions accompanying the mesoscale gravity wave and its redistribution of horizontal momentum. Section 2 will provide a brief description the MASS model and the assumptions that were used for the simulations shown in the chapter. The third section will discuss the simulated large-scale environment the mesoscale gravity wave forms in, including geostrophic imbalance and the ageostrophic adjustment processes, similar to Uccellini and Koch (1987). At the finer scales moist convection, the impact of latent heating within the modeled atmosphere, and a mid-level dry air impulse similar to Yang et al. (2001) will be presented in the evolution of the mesoscale gravity wave system. The third and final part of Section 3 will investigate shear instability as it is often located in the same area as gravity waves (e.g. Koch et al. 2005). Section 4 will provide results testing the simulated gravity waves and their realism relative to observations. Finally, Section 5 will summarize the chapter and add some concluding remarks.
5.2 Model description

While the observational data in this case study was quite comprehensive, to further resolve the fine scale mechanisms accompanying the genesis and evolution of the mesoscale gravity wave system, a mesoscale model will be used and its output analyzed. As described in Chapter 3, the non-hydrostatic Mesoscale Atmospheric Simulation System (NHMASS) model version 6.3 (Kaplan et al. 2000) was employed to study the mesoscale gravity wave event. In both the 18 km and 6 km grid spacing runs the cumulus parameterization scheme is allowed to run. Data from the grid spacing runs between 18 km and 667 m will be presented in this section and further details on the model specifications can be viewed in Chapter 3.

5.3 MASS simulation results

5.3.1 Meso-α scale

5.3.1a Upper-level imbalance as an indicator of gravity wave genesis

The 18 km control simulation (MASS18) will be examined first in an effort to diagnose the level of imbalance in the jet exit region. As in the observations the upper-levels during MASS18 also exhibited significant imbalance. At 1200 UTC December 12th a 300 hPa jet streak is advecting
eastward in northern Mexico under the base of large-scale trough [Fig. 5.1a]. While the jet streak moves across northern Mexico, over central Texas the formation of a >40 ms$^{-1}$ jet streak is occurring downstream of the incipient moist convection much like the idealized simulations from Hamilton et al. (1998) and Kaplan et al. (1998). By 1600 UTC the downstream jet streak, over northern Texas, has strengthened to >45 ms$^{-1}$ [Fig. 5.1b]. The winds at 300 hPa in the left exit region of the first jet streak, across northern Mexico, are being deflected to the left of mean flow across the isoheights and are accelerating toward the secondary jet streak. Between 1200 and 1600 UTC the tilt of the upper-level trough across western Texas turns progressively more negative setting up a favorable upper-level environment for mesoscale gravity waves consistent with Uccellini and Koch (1987) model of geostrophic adjustment [Figs. 5.1a and 5.1b]. This negative tilt favors the development of upper-level mass flux divergence in the left exit region as the sense of the ageostrophic circulation becomes imbalanced. Winds in the entrance region of the secondary jet streak continue to indicate imbalance as the total winds in this region are directed across the isoheights. By 2200 UTC the upper-level momentum has continued to intensify with the secondary jet streak increasing in magnitude and scale across the Midwest and the upper-level trough continuing to tilt increasingly more negatively [Fig. 5.1c]. The area of upper-level imbalance and increased divergence aloft moves into central Louisiana by 0200 UTC on December 13th, as the secondary jet streak has reached its maximum in magnitude, while the first jet streak is now moving into the northern Gulf of Mexico [Fig. 5.1d].
A way to view unbalanced flow is to diagnose areas that have nonlinear balance equation (NBE) values greater than $1 \times 10^{-8}$ s$^{-2}$ as in Kaplan and Paine (1977) and Rozumalski (1997). The NBE on the 326-K isentropic surface at 1200 UTC December 12th, over southern Texas has values exceeding $3 \times 10^{-8}$ s$^{-2}$ in the exit region of the jet streak across northern Mexico [Fig. 5.2a]. The area of NBE exceeding $1 \times 10^{-8}$ s$^{-2}$ has expanded northeastward into eastern Texas by 1600 UTC as the area of imbalance increases in between the two jet streaks [Fig. 5.2b]. By 1800 UTC the winds in southern Texas are turning to the left and are ascending up the 326-K isentrope from 320 hPa in southern Texas to 230 hPa in northern Texas [Fig. 5.2c]. NBE values greater than $\pm 1 \times 10^{-8}$ s$^{-2}$ are continuing to expand aloft and are found across all of northern and northeastern Texas. At 2200 UTC the largest NBE values are co-located above the strongest moist convective towers in between the two jet streaks with values exceeding $1.5 \times 10^{-7}$ s$^{-2}$, however, all of eastern Texas has magnitudes greater than $2 \times 10^{-8}$ s$^{-2}$ within the region of unbalanced flow [Fig. 5.2d]. As the area of strong upper-level divergence advects into the Louisiana by 0200 and 0400 UTC on December 13th, the degree of unbalanced flow in between the two jet streaks has decreased in magnitude [Figs. 5.3a and 5.3b]. However, NBE values greater than $1 \times 10^{-8}$ s$^{-2}$ over most of Louisiana still implies some unbalanced flow at upper-levels. As was the case in the observations, MASS18 also indicates unbalanced flow in between the two jet streaks allowing for large amounts of velocity divergence aloft. The upper-levels winds are acting as a favorable region of mass flux divergence enhancing the
growth of as well as being further amplified by the moist convection similar to Hamilton et al. (1998).

5.3.2 Meso-β scale

In an effort to define the meso-β features, this section will be utilizing results from the 6 km (MASS6) and 2 km (MASS2) simulations, which were one-way nested from MASS18. Moist convection forms in between the two jet streaks in the meso-α environment, but in the meso-β scale the organization of the moist convection into a squall line north of the frontal boundary is an important process for gravity wave genesis. Simultaneously, a second key feature is a mid-level dry air impulse on the meso-β scale similar to Yang et al. (2001), which is seen behind the moist convection as the mesoscale gravity wave system becomes better organized.

5.3.2a Evolution of moist convection and mesoscale gravity wave movement

With an unbalanced flow environment at upper-levels in eastern Texas, the meso-α scale features setup the development of the surface cyclone and mesoscale gravity wave similar to Uccellini and Koch (1987). The contribution of the moist convection played a synergistic nonlinear role on the meso-β scale in the development of the mesoscale gravity wave as will be described. At 1200
UTC December 12th, MASS6 simulates several small bands of precipitation, one located along the stationary boundary and the other located from west of Matagorda Bay into south-central Texas [Fig. 5.4a]. The 1012 hPa surface cyclone is just beginning to strengthen near CRP [Fig. 5.4b]. As the mesoscale gravity wave begins to develop northeast of Matagorda Bay north of the stationary boundary by 1600 UTC [Fig. 5.5a], some moist convection begins to organize into north-south oriented line segments [Fig. 5.5b]. In the schematic for a mesoscale gravity wave [Fig. 2.1] the moist convection develops after a trough in the surface pressure perturbation with the associated upward vertical motion and this is similar Lin et al. (1998) and the wave-CISK mechanism aiding in the development of new moist convective cells. Wave-CISK assumes the gravity waves released within the latent heating of established moist convective cells aid in the development of newer moist convective cells and the process continues in this cycle (Lin et al. 1998). Fig. 5.6a shows the maximum of vertical motion in between a trough and ridge of the surface pressure at 1900 UTC near Houston, TX. However, as the mesoscale gravity wave system became more established within the model by 0000 UTC in central Louisiana, the vertical motion and surface pressure signal [Fig. 5.6b] are very similar to the schematic representation of a mesoscale gravity wave described in Koch and O’Handley (1997) [Fig. 2.1]. Within the model atmosphere the lighter precipitation is still organized in east-west oriented bands parallel to the stationary boundary, however, two of the heavier convective elements, one just west of HOU, and the other north of Matagorda Bay, are oriented north-south and are beginning to travel eastward just
ahead of the mesoscale gravity wave, as defined by the location of highest frequency mean sea level pressure perturbations similar to Zhang et al. (2003). The surface cyclone center is located to the southwest of HOU at 1800 UTC along the stationary front and there are two mesoscale gravity waves, one directly to the north of HOU with the main mesoscale gravity to the northwest HOU [Fig. 5.7a]. The north-south oriented convective features are now located in the meso-high region just behind both of the mesoscale gravity waves [Fig. 5.7b] with the mesoscale gravity waves remaining north and along the frontal boundary similar to Zhang et al. (2003). As the main mesoscale gravity wave moves to the north of HOU by 2000 UTC, the moist convection has organized into a squall line located in the meso-high region just behind the mesoscale gravity wave [Figs. 5.8a and 5.8b]. At 2200 UTC both the squall line and mesoscale gravity wave has strengthened and traveled northeastward to the Texas/Louisiana border [Figs. 5.9a and 5.9b]. The moist convection continues to remain located in the meso-high region just behind the mesoscale gravity wave as would be expected from Koch and O’Handley (1997) [Fig. 2.1]. By 0000 UTC December 13th the mesoscale gravity wave has moved into central Louisiana northeast of the now 1003 hPa surface cyclone, which is located east of BPT [Fig. 5.10a]. The moist convection continues to propagate just behind the mesoscale gravity wave and the center of the moist convection begins to form a bowing feature [Fig. 5.10b]. The evolution of the moist convection from west-east oriented bands, to cellular convection, to north-south oriented bands, to squall line is very important on the meso-β scale, because the mesoscale gravity wave strengthens further in each stage of the moist
convection organization and moves further north and east ahead of the moist convection. The nonlinear interaction between the moist convection/squall line and the incipient gravity wave aid in the growth of the gravity wave system and while there is an imbalance at the upper-levels similar to Uccellini and Koch (1987), the moist convection in this December 12th case study is much stronger than the Uccellini and Koch (1987) cases. The moist convection is the key to the development of the mesoscale gravity wave system as will be described in the following section.

5.3.2b Impact of latent heating

To further understand the mechanisms and the cause and effect relationships within the mesoscale gravity wave event on December 12th, 2002 a specific sensitivity experiment was performed. The sensitivity experiment was performed to see the importance of latent heating in the development of the mesoscale gravity wave. This experiment removed latent heating by omitting terms in the thermodynamic energy equation involving evapotranspiration as well deactivating both the convective parameterization and grid-scale schemes for precipitation.

Latent heating, as expected for the moist convective case, plays a dominant role in aiding the mass and momentum redistributions, and the formation of the mesoscale gravity wave during this event. At 1200 UTC December 12th in the DRY18 simulation the first jet streak is propagating across
northern Mexico and it is very similar to MASS18, however, the second jet streak has begun to strengthen in MASS18 in central Texas, while DRY18 does not exhibit this strengthening [Figs. 5.11a and 5.11b]. This pattern continues at 1800 UTC with the trough beginning to tilt negative and the strengthening of the secondary jet over Oklahoma to over 50 ms\(^{-1}\) in MASS18, while DRY18 does have a small second jet streak over Oklahoma, but it is much weaker and smaller in horizontal extent than MASS18’s jet streak [Figs. 5.12a and 5.12b]. By 0000 UTC December 13\(^{th}\), there is no secondary jet streak at 300 hPa in DRY18, while the secondary jet streak in MASS18 stretches across much of the eastern Midwest [Figs. 5.13a and 5.13b]. The importance of latent heating and the moist convection is manifest at the upper-levels as the secondary jet streak is much weaker in the no-latent heating runs. As in Hamilton et al. (1998) and Kaplan et al. (1998) the latent heating within the moist convection aids in strengthening a downstream ‘jetlet’ feature. At the surface the differences with and without latent heating appear prominently. In MASS18 at 1200 UTC December 12\(^{th}\) the surface cyclone is beginning to strengthen in southern Texas at 1012 hPa, while in DRY18 the surface cyclone has not quite formed yet, however, the surface pressure is very similar at 1013 hPa in southern Texas [Figs. 5.14a and 5.14b]. By 1800 UTC the differences become much larger between the two simulations with the surface cyclone in MASS18 at 1009 hPa and in the northwest Gulf of Mexico, with the surface cyclone in DRY18 just east of CRP and at 1010 hPa [Figs. 5.15a and 5.15b]. Also, the mesoscale gravity wave has formed in MASS18 and is located west of HOU, while there is no mesoscale gravity wave formation in
DRY18. There is a 5 hPa difference in the strength of the two surface cyclones between MASS18 and DRY18 by 0000 UTC on December 13\textsuperscript{th}, and the surface cyclone in DRY18 is 300 km further to the southwest, still near Matagorda Bay, Texas [Figs. 5.16a and 5.16b]. The mesoscale gravity wave still has not formed in DRY18, while the mesoscale gravity wave in MASS18 has matured and continues to travel across central Louisiana ahead of the moist convection.

While looking at the upper-level and surface pattern did show a difference with and without latent heating, the differences are less in the equations for diagnosing imbalance within and in proximity to a jet streak. This means that the gravity wave generation mechanism of geostrophic adjustment and upper-level imbalance does not have as prominent a role in the development of the mesoscale gravity wave as previously hypothesized. Moist convection remains the dominant generation mechanism for the mesoscale gravity wave in this case study. The Lagrangian Rossby number at 300 hPa at 1200 UTC on December 12\textsuperscript{th} is well above 0.5 across much of southern Texas in MASS18 with some $Ro_L$ as high as 2.1 [Fig. 5.17a]. DRY18 has much lower $Ro_L$ at 0.3 across southern Texas at 12 UTC [Fig. 5.17b]. As the exit region of the 300 hPa jet streak moves into eastern Texas at 18 UTC in DRY18 the Lagrangian Rossby number gets much larger between 0.6 and 1.2, which is well above the critical magnitude associated with gravity wave genesis (e.g. Uccellini and Koch, 1987), however, it is lower than the $Ro_L$ between 1.5 and 2.7 found in MASS18 at the same time [Figs. 5.18a and 5.18b]. The $Ro_L$ remains above critical in DRY18 at 300 hPa even at 0000 UTC on December 13\textsuperscript{th}, however it is still 2 to 4 times less $Ro_L$ than MASS18 at the
same time [Figs. 5.19a and 5.19b]. Magnitudes of NBE have a similar pattern on the 326-K isentrope between DRY18 and MASS18 during the mesoscale gravity wave event. NBE values between $\pm 1 \times 10^{-8}$ s$^{-2}$ and $\pm 4 \times 10^{-8}$ s$^{-2}$ occurs in both DRY18 and MASS18, however, the extent of the higher values is slightly larger in MASS18 [Figs. 5.20a and 5.20b]. By 0000 UTC December 13th, the NBE values have exceeded $+1 \times 10^{-7}$ s$^{-2}$ in MASS18 in between the two jet streaks, while the NBE values in DRY18 remain at or below $\pm 2 \times 10^{-8}$ s$^{-2}$ in the same region [Figs. 5.21a and 5.21b]. Again the upper-level imbalance does not appear to be as big a generation mechanism for the mesoscale gravity wave in this case study as the moist convection.

Finally, the differences with and without latent heating can be seen in cross-sections through the mesoscale system as it travels northeastward from Texas to Mississippi both with and without latent heating. At 1600 UTC December 12th, as the mesoscale system becomes organized as the mass and momentum is being redistributed in the vertical much more effectively in MASS6 than DRY6, also, notice the perturbations of the isentropes on MASS6 were much more frequent [Figs. 5.22a and 5.22b]. By 1800 UTC 25 ms$^{-1}$ of momentum has been redistributed downward to 850 hPa in MASS6, while DRY6 continues to stay more stratified with relation to its mass and momentum field [Figs. 5.23a and 5.23b]. At 0000 UTC December 13th, as the surface cyclones are propagating northeastward, the mass and momentum fields in MASS6 continues to change vertically with 30 to 35 ms$^{-1}$ of momentum transported downward between 400
and 800 hPa near the moist convection [Fig. 5.24a]. DRY6 continues to stay stratified both with the isentropes and with the momentum as the isentropes stay horizontal and none of the 40 to 45 ms\(^{-1}\) momentum at jet level has mixed downward [Fig. 5.24b]. The moist convection is able to redistribute the momentum much more effectively throughout the troposphere as the gravity wave system begins to organize. As the momentum is redistributed into a wave duct (Lindzen and Tung, 1976 and Yang et al. 2001) it is possible for the mesoscale gravity wave to propagate farther downstream in the latent heating runs. In the no-latent heating runs there is no redistribution of momentum and perturbation of isentropic fields, therefore there are no generation or maintenance mechanisms for gravity waves of sufficient magnitude.

The significant impact of simulated latent heating within the atmosphere was demonstrated in this section, as the mesoscale gravity wave did not even form when latent heating was removed from the modeled atmosphere as per the stated criterion for wave genesis earlier in the thesis. Moist convection and the latent heating that it produces played the dominant role in allowing for mesoscale gravity wave formation in this case study, because the exit region of the jet streak in the dry run was out of geostrophic balance as both the Lagrangian Rossby Number and the NBE were large or above their critical values. However, the mesoscale gravity wave only formed when latent heating and moist convection were added to the unbalanced upper-level jet streak.
5.3.2c Mid-level dry air impulse

Much like Rauber et al. (2001) a dry air feature (dry slot) is advecting into the region of mesoscale gravity wave formation and continues to move behind the mesoscale gravity wave as the whole system continued to the northeast. At 1200 UTC on December 12th on the 312-K isentropic surface MASS18 shows a relatively uniform moisture field across much of Texas with relative humidities between 30 and 60 % [Fig. 5.25a]. As the mesoscale gravity wave is generated in MASS18 by 1600 UTC, the moist convective towers can be seen on the 312-K isentrope in central and eastern Texas with relative humidities between 80 and 100 % [Fig. 5.25b]. Across northern Mexico there is the development of dry air tongue with air associated with descent from the 400 hPa to the 500 hPa pressure surface with horizontal velocities of 25 to 50 ms$^{-1}$. This dry air has a relative humidity between 10 and 30 % [Fig. 5.25c], and is descending directly toward the upstream side of the moist convection at mid-levels. The dry-slot has moved northeastward closer to the upstream side of moist convection associated with the mesoscale gravity wave at 1800 UTC [Fig. 5.25d]. The air continues to be co-located with descending motion at 410 hPa in southern Texas through the column down to 550 hPa and even 600 hPa in central Texas with velocities 22 to 32 ms$^{-1}$. A notch of dry air between the moist convective towers develops at 2000 UTC in central Texas where the air continues to descend and be advected eastward at 20 to 30 ms$^{-1}$ [Fig. 5.26]. This notch of dry air was also seen at mid-levels in the water vapor imagery of the observed mesoscale event (not shown). As the first
dry air notch advects into northern Texas on the 312-K isentropic surface by 2200 UTC, a second dry air notch is beginning to form in far southeastern Texas just upstream of the moist convection and the northeastward propagating mesoscale gravity wave [Fig. 5.25e]. By 0200 UTC December 13th the dry slot had moved into central Louisiana with air continuing to descend from 440 to 550 hPa and move forward at 25 ms$^{-1}$ into the upstream side of the moist convection [Fig. 5.25f]. More pronounced development and propagation of a dry-slot at mid-levels upstream of the moist convection occurs in MASS6. At 1600 UTC December 12th the dry air can be seen in northeastern Mexico with relative humidities between 10 and 30 % on the 312-K isentrope [Fig. 5.27a]. At lower-levels on the 296-K isentropic surface at 1600 UTC the winds are ascending the pressure levels from 945 hPa to 795 hPa with a forward velocity of 15-20 ms$^{-1}$ into the moist convection [Fig. 5.27b], so there is a large amount of mass convergence across central Texas. The dry slot becomes more pronounced by 2000 UTC on the 312-K isentropic surface as the mesoscale gravity wave and moist convection have strengthened over the past 4 hours [Fig. 5.27c]. At 2200 UTC the moist convection has organized into a squall line in eastern Texas and the dry slot continues to strengthen with more dry air descending the pressure levels and moving forward at 20 to 30 ms$^{-1}$ [Fig. 5.27d]. Also, there is 20 % relative humidity air directed into the back of the squall line in eastern Texas at 30 ms$^{-1}$. By 0000 UTC December 13th, the dry slot on the 312-K isentropic surface continues to move northeastward behind the mesoscale gravity wave and a notch of dry air develops in northeastern Texas to the northwest of the squall line [Fig.
The dry air from upper-levels that descends the isentropes behind the moist convection is very similar to Rauber et al. (2001) and Yang et al. (2001) and their mesoscale gravity wave event. During this December 12th case study the dry air advects directly behind the moist convection in the mesoscale gravity wave system much like a dry-slot visible in a mature cyclonic system.

At this moment it is not clear whether the dry slot is a causal mechanism in the mesoscale gravity wave system or just visualization of descending momentum from some other source, however, it would a helpful tool for forecasters, because it is co-located with the moist convection and mesoscale gravity wave, and as the dry slot becomes more pronounced the mesoscale gravity wave does intensify. This dry slot has several factors that could be helpful for mesoscale gravity wave genesis or maintenance. First, the dry slot could be an indication of the descending momentum from the jet streak and the momentum could be hitting the stable layer behind the moist convection causing buoyancy perturbations leading to the genesis of more gravity waves or maintaining the large amplitude mesoscale gravity wave as it moves eastward. This speculation will be address in the section on shear instability. Secondly, the dry-air descending directly behind the moist convection would allow for a favorable wave-duct situation as the dry unstable layer would be above the moist stable layer north of the stationary front allowing for mesoscale gravity wave maintenance for westward moving gravity waves in this case. The dry air would be located directed above a relatively cooler stable layer left by the passage of the
moist convection and the cooler stable layer with sufficient depth would be available for wave ducting (Lindzen and Tung 1976).

5.3.3 Meso-γ scale

With the interest of investigating the role of meso-γ scale features in wave genesis and maintenance this section will be using results from the MASS2SW and MASS667SWM simulations to view an isentropic analysis of the mesoscale gravity wave system. An area of possible shear instability and its interactions between the mass and momentum fields will also be examined as its role in the genesis and maintenance of the mesoscale gravity wave system from the meso-α scale down to the meso-γ scale. Formation or continued movement of the mesoscale gravity wave within the moist convection will be examined.

5.3.3a Isentropic analysis of the initiation of the mesoscale gravity wave system

Initially the moist convection is cellular in nature at 1400 UTC on December 12th in MASS2SW [Fig. 5.28a] with most of the precipitation north of the stationary boundary along the Gulf Coast. There is a strong moist easterly component to the wind rising up the pressure levels on the 296-K isentropic surface supplying the cellular moist convection [Fig. 5.28b]. By 1600 UTC the
1008 hPa surface low is west of Matagorda Bay, Texas moving along the stationary boundary to the northeast [Fig. 5.29a]. The moist convection continues to remain cellular in nature with the majority of the precipitation northeast of the surface cyclone [Fig. 5.29b]. Cellular moist convection intensifies by 1700 UTC with two distinct lines of convection, one north of Matagorda Bay, and the other 30 km to the west [Fig. 5.30a]. The moist convection north of Matagorda Bay is located above the surface cyclone, while the moist convection to the west is associated with a convergence line in the surface wind field which is the cold front located to the southwest of the low center [Fig. 5.30b]. Strengthening and development of the mesoscale gravity wave system has begun by 1724 UTC as the moist convection begins to organize into a squall line [Fig. 5.31a]. This strengthening occurs with the increased organization of the moist convection, as the surface low pressure deepens within the modeled atmosphere and as the surface gradient or inflow into the moist convection tightens. Gravity waves develop to the north of the stationary boundary and to the northeast of the surface cyclone similar to Zhang et al. (2004), however, the modeled gravity waves in this December case study have shorter wavelengths [Fig. 5.31b]. However, the main mesoscale gravity wave begins to intensify in an area just north of Matagorda Bay, which is co-located with a sharp decrease in the Richardson number on the 296-K isentropic surface with values below 0.25 [Fig. 5.31c]. The low Richardson number values indicate a possibility of shear instability near the top of the stable layer north of the frontal boundary as speculated in Uccellini and Koch (1987) and discussed in Koch et al. (2005). We speculate that the low Richardson
numbers in this area are an indication of gravity waves breaking which entails neutral stability and Richardson numbers close to 0. The area in which the low Richardson numbers occur is the same region as discussed in Yang et al. (2001) where they had an area of evaporational cooling behind the moist convection. Yang et al. (2001) believed this evaporational cooling was responsible for increased downward momentum behind the moist convection into the stable layer located north of the warm front. This observation will be examined in the following section. This nonlinear process continues by 1736 UTC with the moist convection continuing to intensify and organize [Fig. 5.32a] and the mesoscale gravity wave beginning to travel to the northeast along the stationary boundary [Fig. 5.32b]. The area of Richardson number below critical value on the 296-K isentropic surface continues to be co-located directly above the mesoscale gravity wave [Fig. 5.32c]. By 1754 UTC December 12th, the main mesoscale gravity wave has moved just to the northeast of Matagorda Bay and the Richardson numbers on the 296-K isentrope keep dropping further below critical values now close to 0.10 [Figs. 5.33a and 5.33b]. The moist convection still has two distinct lines, however, it is continuing to organize along a north-south oriented line associated with the mesoscale gravity wave while Zhang et al.’s (2004) mesoscale gravity waves organized along an east-west oriented line [Fig. 5.33c]. At 1830 UTC the main mesoscale gravity wave continues to move toward the northeast distancing itself further to the northeast of the surface cyclone with the Richardson numbers remaining below critical on the 296-K isentrope [Figs. 5.34a and 5.34b]. Moist convection at this time has matured into a squall line
propagating toward the northeast along the stationary boundary [Fig. 5.34c].

Finally, by 1930 UTC the squall line has propagated close to the model boundary [Fig. 5.34d], with the main mesoscale gravity wave and low Richardson numbers on the 296-K isentrope approaching Houston, TX [Figs. 5.35a and 5.35b]. While the Richardson numbers are still below critical values they are higher values than were seen closer to the mesoscale gravity wave formation time near Matagorda Bay and therefore the mesoscale gravity wave is possibly weakening with a less overturning isentropes and higher Richardson numbers.

The evolution of the mesoscale gravity wave system is evident in MASS2SW as the main mesoscale gravity wave forms around 1720 UTC north of Matagorda Bay. As evidenced by the 296-K isentropic surface values of Richardson number and the low Richardson numbers are coincident with the mesoscale gravity wave as in Koch et al. (2005). The mesoscale gravity wave continues to strengthen and move further away from the main surface cyclone and while doing this, the moist convection changes character from cellular in nature to an intensify squall line. Richardson numbers below 0.25 were observed in the organization of the mesoscale gravity wave system and those low values will be studied in vertical cross sections to see if shear instability, as Koch et al. (2005), in the vertical could aid in the formation of this mesoscale gravity wave.
5.3.3b Shear instability

This section will focus on the meso-$\gamma$ scale and shear instability to look for the momentum or mass perturbations that result in the surface pressure perturbations as the mesoscale gravity wave has already formed in the favorable meso-$\alpha$ and meso-$\beta$ environment setup up by geostrophic adjustment and moist convection similar to Uccellini and Koch (1987). Looking at a vertical cross section at 1804 UTC December 12th, taken west to east through the squall line after the mesoscale gravity wave had formed, shows the moist convective tower associated with the squall line and the anvil spread to the east ahead of the squall line [Fig. 5.36a]. Directly below the main updraft the density current is moving eastward with the cold pool from the squall line [Fig. 5.36b]. Behind the main updraft Richardson number values approaching 0, well below critical, are observed with a corresponding downdraft greater than 5 $\text{ms}^{-1}$ [Figs. 5.36c and 5.36d]. The horizontal wind speeds near the downdraft are 25 $\text{ms}^{-1}$ while closer to the stable layer just behind the cold pool horizontal wind speeds are near 15 $\text{ms}^{-1}$ [Fig. 5.36b]. By 1810 UTC the center of the main downdraft is near 700 hPa, 150 hPa closer to the surface than 6 minutes prior, and the horizontal wind speeds just above the stable layer have increased to 30 $\text{ms}^{-1}$ behind the cold pool [Figs. 5.37a and 5.37b]. The downdraft is very dry with relative humidity of 30 to 40 %, and is located just behind the moist main updraft [Fig. 5.37c]. The pattern is again repeated between 1828 to 1834 UTC with Richardson numbers values below critical at 500 hPa co-located with a 10 $\text{ms}^{-1}$ downdraft at 1828 UTC [Figs. 5.38a
and 5.38b]. This pattern of dry air behind the updraft is observed in the Yang et al. (2001) and their mesoscale gravity wave system. Dry air is beginning to descend behind the main updraft again starting at 600 hPa and the momentum from the 35 ms$^{-1}$ horizontal winds are impinging on the back of the updraft at 550 hPa. A downward transfer of horizontal momentum [Figs. 5.38c and 5.38d] is seen shortly after the 35 ms$^{-1}$ of horizontal momentum impinges on the back of the updraft. Four minutes later the low Richardson numbers still remain at 550 hPa, but the center of the downdraft is positioned lower at 650 hPa [Figs. 5.39a and 5.39b]. Two minutes later the 318-K isentrope with the low Richardson number values has folded downward completely from 550 hPa to 700 hPa [Fig. 5.40a]. Areas of dry air behind the main updraft have expanded with the descent between 1832 and 1834 UTC, just as the downdraft has strengthened back to over 10 ms$^{-1}$ [Figs. 5.40b and 5.40c] similar to Yang et al. (2001). As horizontal winds build again behind the main updraft in the squall line 10 minutes later, the pattern of the low Richardson numbers, dry air, and descending momentum in the downdraft continues (not shown). This is consistent with the process of maintaining the mesoscale gravity wave as it travels to the northeast similar to Yang et al.’s (2001) mesoscale gravity wave.

Taking a north-south cross section through the same convective line the descent of momentum can again be seen behind the main updraft in the squall line. At 1814 UTC there is a large amount of dry air between 300 and 700 hPa to the south of the moist convection [Fig. 5.41a]. On the south side of the moist convection the Richardson number is below critical between 500 and 700 hPa and
is co-located with a 5 ms\(^{-1}\) downdraft [Figs. 5.41b and 5.41c]. 30 ms\(^{-1}\) of horizontal momentum is available for the shear instability to redistribute the momentum downward on the south side of the moist convection as well [Fig. 5.41d]. The dry air has been transported toward the updraft by 1818 UTC, and the low Richardson number values with the downdraft are an indication of the isentropes beginning to overturn and redistribute the horizontal momentum downward [Figs. 5.42a, 5.42b, and 5.42c]. At 1822 the dry air was completely transported into the squall line system behind the updraft at 500 hPa [Fig. 5.43a]. The low Richardson number values slope downward from 500 hPa to 750 hPa as the isentropes continue to fold [Fig. 5.43b]. Finally at 1824 UTC the dry air has been transported downward to 650 hPa co-located again with a 10 ms\(^{-1}\) downdraft and low Richardson number values just above indicating the folding of isentropes [Figs. 5.44a, 5.44b, and 5.44c]. The folding isentropes indicate the redistribution of momentum and kinetic energy in the vertical from the environment flow into the mesoscale gravity wave.

As the mesoscale gravity wave system is generated, it remains dominated by cellular convection. The moist convection then organizes into a squall line feature and downdrafts form behind the main updraft. These downdrafts, in areas of Richardson number below critical, transport the momentum downward in the stable layer as described in Yang et al. (2001). As gravity waves are generated through latent heating within the moist convection, isentropes fold over, low Richardson numbers are observed, the momentum is transported downward, and the air becomes drier. On the meso-\(\gamma\) scale shear instability, areas of Richardson
number below 0.25, play an important role in the organization of the mesoscale gravity wave, as Richardson numbers below 0.25 are observed in the same areas in which horizontal momentum is transferred downward. This is the same pattern observed by Yang et al. (2001) and their observations of a mesoscale gravity wave. The kinetic energy from this momentum is transferred into the genesis/maintenance of the mesoscale gravity wave, consistent with high surface winds and pressure rises, i.e., surface wind covariance under the descending wave mode.

5.4 Validation of simulated mesoscale gravity waves

Several comparisons were performed on the NHMASS model results to discover the theoretical realism of the mesoscale gravity waves simulated by the model and these tests were conducted along the guidelines from Kiefer (2005). The first test is between the wave-normal wind (u*) and the perturbation pressure (p’) to see if they are in phase. The wave normal wind is given by,

\[ u^* = -u \sin \alpha - v \cos \alpha, \]  

(5.4.1)

where u and v are the Cartesian wind coordinates and \( \alpha \) is the direction from which the wavefronts are propagating (clockwise from north) (Koch and Golus 1988). Being consistent with Kiefer (2005) the MASS2SW model run was utilized with the strong pressure falls and rises associated with the mesoscale gravity wave. Analysis of the mesoscale gravity wave was done in between
Matagorda Bay and HOU with $\alpha = 260$ degrees in order to test the coherency of the mesoscale gravity waves. Before the mesoscale gravity wave moves through the area of interest $u^*$ and $p'$ have a solid positive correlation of 0.94, while the $u^*/p'$ correlation is –0.50 after the mesoscale gravity wave passage. As the positive $u^*/p'$ correlation turns more negative, a loss of wave energy and coherency occurs across Matagorda Bay, however, this test is consistent with mesoscale gravity wave activity and propagation.

Another test for the modeled mesoscale gravity waves is to look for quadrature between the perturbation vertical velocity ($w'$) and the perturbation potential temperature ($\theta'$) fields. While it would be more helpful to directly have the $w'$ and $\theta'$ fields, no filtering was done in MASS2SW so the $w$ and $\theta$ will be used in the above analysis. Previous work on inertial gravity waves by Powers and Reed (1993) found $w'$ and $\theta'$ fields at 400 hPa are offset by $\frac{1}{4}$ wavelength, while the $w'$ and $\theta'$ are in-phase for convectively induced perturbations (e.g. squall lines). Due to the nature of the mesoscale gravity wave in this case study being between an inertial gravity wave and convectively induced perturbations one would expect the $w$ and $\theta$ to be between in-phase and $\frac{1}{4}$ wavelength off. Figs. 5.45a and b give the $w$ and $\theta$ at a point east of Matagorda Bay as the mesoscale gravity wave travels through the area of interest and as the first wave passes the $w$ and $\theta$ are out of phase by $\frac{1}{4}$ wavelength between 1730 and 1800 UTC at 400 hPa. However, as a second wave moves over this point the given $w$ and $\theta$ field is in-phase at 1900 UTC and this second perturbation would be convectively induced.
Therefore the quadrature between \(w\) and \(\theta\) fields at 400 hPa allows for the
determination of two different signals within the mesoscale gravity wave system,
with a small \(\frac{1}{4}\) wavelength out of phase signal followed by the more convectively
induced signal.

A final test subjectively compares the intrinsic phase speed of the
mesoscale gravity wave and the dispersion relationship of a propagating
hydrostatic gravity wave with Coriolis rotation,

\[
C_i = \frac{\omega_i}{k} = \sqrt{\frac{f^2}{k^2} + \frac{N^2}{m^2}},
\]

where \(\omega_i\) is the intrinsic frequency of the wave, and \(m\) and \(k\) are the vertical and
horizontal wavenumbers, respectively (Zhang 2004). With a vertical wavelength
of 4.17 km, a horizontal wavelength of 175.42 km (determined for the mesoscale
gravity wave at 800 hPa from MASS2SW potential temperature fields), and a
value of Brunt-Vaisalla frequency of 0.02 s\(^{-1}\), a phase speed of 13.57 ms\(^{-1}\) was
estimated using the dispersion relationship. Through a subjective analysis (Kiefer
2005) of wave propagation using the surface pressure tendencies analysis, the
intrinsic phase speed was estimated at 17.88 ms\(^{-1}\). While both these methods
provide reasonable agreement, with the mesoscale gravity wave phase speed
being faster in the surface pressure tendency analysis, we believe that the
mesoscale gravity wave in this case study to be between an inertial gravity wave
and pure gravity waves in character, and therefore a mesoscale gravity wave as
defined by Uccellini and Koch (1987). Each of the three tests discussed in this
section provide reasonable agreement that the mesoscale gravity waves in the
NHMASS simulations are indeed physically reasonable and representative of what would be observed in actual atmosphere.

5.5 Summary and conclusions

This section examined the mesoscale gravity wave event through the simulations from NHMASS and further confirmed the hypothesis that the proximity of moist convection in between two jet streaks allowed for juxtaposed unbalanced flow and shearing instability and the formation of a large-amplitude, deep, tropospheric, mesoscale gravity wave. Several processes, which interacted from the meso- to the meso- scale, became in-phase with one another allowing for the development of the mesoscale gravity wave during December 12th and 13th, 2002. First, the jet streak, which moved into southern Texas from northern Mexico, is out of geostrophic balance. Moist convection begins to develop in the exit region of this jet streak and a second jet streak similar to Hamilton et al. (1998) forms downstream of the moist convection. On the meso- scale the Sutcliffe self-development process is beginning to occur. As the moist convection increases in coverage and intensity there is increased divergence aloft and increased inflow near the surface allowing for continued organization of the convection. As the momentum from the jet streak encounters the moist convection the momentum descends downward into the inversion or stable layer just behind the moist convection north of the frontal boundary similar to Yang et al. (2001). As long as the momentum from the upper-levels continues to be transported
downwards in the highly sheared ducting environment by elevated convection, the
non-linear adjustments will continue and the mesoscale gravity wave continues to
be sustained by reinforcing pulses of momentum. Phasing between all three
mesoscales is the key to this event. On the meso-\(\alpha\) scale the unbalanced jet exit
region (with its upper-level divergence) and the moist convection phase. The
organization and propagation of the moist convection north of the frontal
boundary along with the non-linear feedback aloft allow for phasing on the meso-
\(\beta\) scale. Finally, the phasing between the moist convection and the regions of
shear instability or overturning isentropes on the meso-\(\gamma\) scale are an indication of
the momentum mixing downward. Several processes still need further
investigation. Does the dry air impulse act as a causal mechanism in the formation
of the mesoscale gravity wave or is it just a reflection of the descending
momentum from some other source or subsidence from a larger scale process?
Also, it is clear that moist convection plays a significant role in this case study,
with no development of the mesoscale gravity wave in the no-latent heating
NHMASS runs, but how unbalanced does the exit region of the jet streak need to
be and how efficient does the upper-level divergence need to be above the moist
convection to act as a sufficient “exhaust pipe” for the mesoscale system as a
whole? Further examination of several other mesoscale gravity wave events need
to be made in order to more completely understand these processes.

The next chapter will consider the gravity waves propagating into the
lower stratosphere, their environment, and possible upper-level turbulence
associated with this mesoscale gravity wave event as was shown in the
observations from the lower stratosphere in Chapter 4.
Fig. 5.1 MASS18 300 hPa isoheights (m) (white lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading) valid for: a) 1200, b) 1600, c) 2200, and d) 0200 UTC 13 Dec 2002.
Fig. 5.2 MASS18 326-K isentrope pressure (hPa) (black lines), wind barbs (ms$^{-1}$), and NBE (s$^{-2}$) (colored shading) valid for: a) 1200, b) 1600, c) 1800, and d) 2200 UTC 12 Dec 2002.
Fig. 5.3 MASS18 326-K isentrope pressure (hPa) (black lines), wind barbs (ms$^{-1}$), and NBE (s$^2$) (colored shading) valid for: a) 0200 and b) 0400 UTC 13 Dec 2002.
Fig. 5.4 MASS6 valid for 1200 UTC 12 Dec 2002 for: a) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms$^{-1}$) (wind barbs) and b) mean sea-level pressure (hPa) (white lines) and surface winds (knots).
Fig. 5.5 MASS6 valid for 1600 UTC 12 Dec 2002 for: a) mean sea-level pressure (hPa) (white lines) and surface winds (knots) and b) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms\(^{-1}\)) (wind barbs).
Fig. 5.6 MASS6 time series from 0800 UTC December 12 to 0400 UTC December 13 of surface pressure perturbation (solid line) (hPa) and 900 hPa vertical motion (dashed line) (ms\(^{-1}\)) for a) 31.0 N; -94.0 W and b) 32.0 N; -93.0 W.
Fig. 5.7 MASS6 valid for 1800 UTC 12 Dec 2002 for: a) mean sea-level pressure (hPa) (white lines) and surface winds (knots) and b) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms⁻¹) (wind barbs).
Fig. 5.8 MASS6 valid for 2000 UTC 12 Dec 2002 for: a) mean sea-level pressure (hPa) (white lines) and surface winds (knots) and b) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms$^{-1}$) (wind barbs).
Fig. 5.9 MASS6 valid for 2200 UTC 12 Dec 2002 for: a) mean sea-level pressure (hPa) (white lines) and surface winds (knots) and b) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms⁻¹) (wind barbs).
Fig. 5.10 MASS6 valid for 0000 UTC 13 Dec 2002 for: a) mean sea-level pressure (hPa) (white lines) and surface winds (knots) and b) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms⁻¹) (wind barbs).
Fig. 5.11 300 hPa isoheights (m) (white lines), wind barbs (ms\(^{-1}\)), and isotachs (knots) (colored shading) valid for 1200 UTC 12 Dec 2002 for: a) DRY18 and b) MASS18.
Fig. 5.12 300 hPa isoheights (m) (white lines), wind barbs (ms⁻¹), and isotachs (knots) (colored shading) valid for 1800 UTC 12 Dec 2002 for: a) DRY18 and b) MASS18.
Fig. 5.13 300 hPa isoheights (m) (white lines), wind barbs (ms\(^{-1}\)), and isotachs (knots) (colored shading) valid for 0000 UTC 13 Dec 2002 for: a) DRY18 and b) MASS18.
Fig. 5.14 Mean sea-level pressure (hPa) (white lines) and wind barbs (m s⁻¹) valid for 1200 UTC 12 Dec 2002 for: a) DRY18 and b) MASS18.
Fig. 5.15 Mean sea-level pressure (hPa) (white lines) and wind barbs (ms\(^{-1}\)) valid for 1800 UTC 12 Dec 2002 for: a) DRY18 and b) MASS18.
Fig. 5.16 Mean sea-level pressure (hPa) (white lines) and wind barbs (ms$^{-1}$) valid for 0000 UTC 13 Dec 2002 for: a) DRY18 and b) MASS18.
Fig. 5.17 Lagrangian Rossby number (black lines) valid for 1200 UTC 12 Dec 2002 for: a) MASS18 and b) DRY18.
Fig. 5.18 Lagrangian Rossby number (black lines) valid for 1800 UTC 12 Dec 2002 for:
  a) DRY18 and b) MASS18.
Fig. 5.19 Lagrangian Rossby number (black lines) valid for 0000 UTC 13 Dec 2002 for:
a) DRY18 and b) MASS18.
Fig. 5.20 326-K isentrope pressure (hPa) (black lines), wind barbs (ms⁻¹), and NBE (s⁻²) (colored shading) valid for 1800 UTC 12 Dec 2002 for: a) DRY18 and b) MASS18.
Fig. 5.21 326-K isentrope pressure (hPa) (black lines), wind barbs (ms\(^{-1}\)), and NBE (s\(^{2}\)) (colored shading) valid for 0000 UTC 13 Dec 2002 for: a) DRY18 and b) MASS18.
Fig. 5.22 A vertical cross section west to east through the moist convection valid 1600 UTC 12 Dec 2002 with potential temperature (white lines) and isotachs (knots) (colored shading) for: a) MASS6 and b) DRY6.
Fig. 5.23 vertical cross section through the moist convection valid 1800 UTC 12 Dec 2002 with potential temperature (white lines) and isotachs (knots) (colored shading) for: a) MASS6 and b) DRY6.
Fig. 5.24 vertical cross section through the moist convection valid 0000 UTC 13 Dec 2002 with potential temperature (white lines) and isotachs (knots) (colored shading) for: a) MASS6 and b) DRY6.
Fig. 5.25 MASS18 312-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and relative humidity (%) (colored shading) valid for: a) 1200 UTC 12 Dec 2002, b) 1600, c) 1800, d) 2000, e) 2200, and f) 0200 UTC 13 Dec 2002.
Fig. 5.25 (Continued)
Fig. 5.25 (Continued)
Fig. 5.26 MASS2 valid for 2000 UTC 12 Dec 2002 for a vertical cross section from 29 N; –96 W to 31 N; –94 W, for potential temperature (K) (black lines), relative humidity (90 % red line) and W (ms⁻¹) (dashed lines).
Fig. 5.27 MASS6 312-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and relative humidity (%) (colored shading) valid for: a) 1600 UTC 12 Dec 2002, and MASS6 296-K isentrope for b) 1600 UTC 12 Dec 2002, and MASS6 312-K isentrope for: c) 2000, d) 2200, and e) 0000 UTC 12 Dec 2002.
Fig. 5.27 (Continued)
Fig. 5.27 (Continued)
Fig. 5.28 MASS2SW valid for 1400 UTC 12 Dec 2002 for a) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms⁻¹) (wind barbs) and b) 296-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and relative humidity (%) (colored shading).
Fig. 5.29 MASS2SW valid for 1600 UTC 12 Dec 2002 for: a) mean sea-level pressure (hPa) (black lines) and surface winds (knots) and b) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms⁻¹) (wind barbs).
Fig. 5.30 MASS2SW valid for 1700 UTC 12 Dec 2002 for: a) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms$^{-1}$) (wind barbs) and b) mean sea-level pressure (hPa) (black lines) and surface winds (knots).
Fig. 5.31 MASS2SW valid for 1724 UTC 12 Dec 2002 for: a) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms\(^{-1}\)) (wind barbs), b) mean sea-level pressure (hPa) (black lines) and surface winds (knots), and c) 296-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and Richardson number (colored shading).
Fig. 5.31 (Continued)
Fig. 5.32 MASS2SW valid for 1736 UTC 12 Dec 2002 for: a) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms⁻¹) (wind barbs), b) mean sea-level pressure (hPa) (black lines) and surface winds (knots), and c) 296-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and Richardson number (colored shading).
Fig. 5.32 (Continued)
Fig. 5.33 MASS2SW valid for 1754 UTC 12 Dec 2002 for: a) 296-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and Richardson number (colored shading), b) mean sea-level pressure (hPa) (black lines) and surface winds (knots), and c) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms\(^{-1}\)) (wind barbs).
Fig. 5.33 (Continued)
Fig. 5.34 MASS2SW valid for 1830 UTC 12 Dec 2002 for: a) mean sea-level pressure (hPa) (black lines) and surface winds (knots), b) 296-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and Richardson number (colored shading), c) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms⁻¹) (wind barbs), and d) base radar reflectivity (Dbz) (colored shading) and 950 hPa wind (ms⁻¹) (wind barbs) valid for 1930 UTC 12 Dec 2002.
Fig. 5.35 MASS2SW valid for 1930 UTC 12 Dec 2002 for: a) mean sea-level pressure (hPa) (black lines) and surface winds (knots) and b) 296-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and Richardson number (colored shading).
Fig. 5.36 MASS667MSW valid for 1804 UTC 12 Dec 2002 for a vertical cross section from 28.77 N; –96.25 W to 28.77 N; –95.85 W, for: a) potential temperature (K) (black lines) and relative humidity (%) (colored shading), b) potential temperature (K) (black lines) and horizontal wind speed (knots) (colored shading), c) potential temperature (K) (black lines) and Richardson number (colored shading), and d) potential temperature (K) (black lines) and W (ms\(^{-1}\)) (colored shading).
Fig. 5.36 (Continued)
Fig. 5.37 MASS667MSW valid for 1810 UTC 12 Dec 2002 for a vertical cross section from 28.77 N; –96.25 W to 28.77 N; –95.85 W, for: a) potential temperature (K) (black lines) and W (ms\(^{-1}\)) (colored shading), b) potential temperature (K) (black lines) and horizontal wind speed (knots) (colored shading), and c) potential temperature (K) (black lines) and relative humidity (%) (colored shading).
Fig. 5.37 (Continued)
Fig. 5.38 MASS667MSW valid for 1828 UTC 12 Dec 2002 for a vertical cross section from 28.77 N; –96.25 W to 28.77 N; –95.85 W, for: a) potential temperature (K) (black lines) and Richardson number (colored shading), b) potential temperature (K) (black lines) and W (ms$^{-1}$) (colored shading), c) potential temperature (K) (black lines) and relative humidity (%) (colored shading), and d) potential temperature (K) (black lines) and horizontal wind speed (knots) (colored shading).
Fig. 5.38 (Continued)
Fig. 5.39 MASS667MSW valid for 1832 UTC 12 Dec 2002 for a vertical cross section from 28.77 N; –96.25 W to 28.77 N; –95.85 W, for: a) potential temperature (K) (black lines) and Richardson number (colored shading) and b) potential temperature (K) (black lines) and W (ms$^{-1}$) (colored shading).
Fig. 5.40 MASS667MSW valid for 1834 UTC 12 Dec 2002 for a vertical cross section from 28.77 N; −96.25 W to 28.77 N; −95.85 W, for: a) potential temperature (K) (black lines) and Richardson number (colored shading), b) potential temperature (K) (black lines) and relative humidity (%) (colored shading), c) potential temperature (K) (black lines) and W (ms⁻¹) (colored shading), and d) potential temperature (K) (black lines) and horizontal wind speed (knots) (colored shading).
Fig. 5.40 (Continued)
Fig. 5.41 MASS667MSW valid for 1814 UTC 12 Dec 2002 for a vertical cross section from 29.15 N; –95.98 W to 28.65 N; –95.98 W, for: a) potential temperature (K) (black lines) and relative humidity (%) (colored shading), b) potential temperature (K) (black lines) and Richardson number (colored shading), c) potential temperature (K) (black lines) and W (ms$^{-1}$) (colored shading), and d) potential temperature (K) (black lines) and horizontal wind speed (knots) (colored shading).
Fig. 5.41 (Continued)
Fig. 5.42 MASS667MSW valid for 1818 UTC 12 Dec 2002 for a vertical cross section from 29.15 N; –95.98 W to 28.65 N; –95.98 W, for: a) potential temperature (K) (black lines) and relative humidity (%) (colored shading), b) potential temperature (K) (black lines) and Richardson number (colored shading), and c) potential temperature (K) (black lines) and W (ms⁻¹) (colored shading).
Fig. 5.43 MASS667MSW valid for 1822 UTC 12 Dec 2002 for a vertical cross section from 29.15 N; –95.98 W to 28.65 N; –95.98 W, for: a) potential temperature (K) (black lines) and relative humidity (%) (colored shading) and b) potential temperature (K) (black lines) and Richardson number (colored shading).
Fig. 5.44 MASS667MSW valid for 1824 UTC 12 Dec 2002 for a vertical cross section from 29.15 N; –95.98 W to 28.65 N; –95.98 W, for: a) equivalent potential temperature (K) (black lines) and relative humidity (%) (colored shading), b) potential temperature (K) (black lines) and W (ms⁻¹) (colored shading), and c) equivalent potential temperature (K) (black lines) and Richardson number (colored shading).
Fig. 5.44 (Continued)
Fig. 5.45 MASS2SW time series plots from 1630 UTC 12 Dec 2002 to 2000 UTC 12 Dec 2002 at point 29.40 N; -95.75 W at 400 hPa of: a) W (ms\(^{-1}\)) and b) potential temperature (K).
6. Gravity Waves and Turbulence in the lower Stratosphere

6.1 Introduction

Much like in the lower troposphere where the atmosphere is redistributing mass and momentum allowing for the formation of the mesoscale gravity wave(s), above the moist convection vertically propagating finer-scale gravity waves are being generated to release energy. Turbulence generated from these vertically propagating gravity wave poses a significant risk to the commercial and military aircraft community as numerous turbulence related injuries occur every year on aircraft flights. The Federal Aviation Administration (FAA) has founded guidelines for aircraft to avoid moist convection (thunderstorms) by 37 km horizontally, and flying above thunderstorms by at least 1000 feet for every 5 ms$^{-1}$ cloud-top wind speed. However, as discussed by Lane et al. (2003) flying above the moist convection may still not be the best decision due to the vertical propagating gravity waves that can be induced causing turbulence well above the convective towers into the lower stratosphere. Koch et al. (2005) showed that the intensity of the atmospheric turbulence was highly correlated with the appearance of the packets of vertically propagating gravity waves in the exit region of an unbalanced upper-level jet streak. Therefore it is very important to understand the environmental characteristics above the moist convection in order to determine the correct location of the clear-air turbulence and vertically propagating gravity waves. Vertically propagating gravity waves in this chapter are defined by
perturbations in the isentropes above the moist convection and into the stratosphere. Applying the following equation,

$$\frac{Na}{U}, \quad (6.1.1)$$

where $a$ is the area or length of the disturbance and $U$ is the horizontal wind speed, then for realistic vertically propagating gravity waves in the model and atmosphere our hydrostaticity should be greater than 1. Hydrostaticity can be applied to gravity waves within the modeled atmosphere to see if the gravity waves are a realistic representation of the real world or just evanescent waves that would not be observed in the real atmosphere. If the hydrostaticity is less than 1, then we would expect evanescent waves and no vertical perturbation of the isentropes into the lower stratosphere and no turbulence (Lin 2007). The hydrostaticity does not determine whether modeled gravity waves are hydrostatic or non-hydrostatic, it determines whether the modeled gravity waves are realistic or evanescent.

The rest of the chapter is organized as follows with results from the NHMASS simulations to provide the basis for the study of the vertically propagating gravity waves, their realism, and the background environment surrounding/interacting with them. Section 2 will describe the growth of moist convection across Texas and the northwestern Gulf Mexico on the meso-α and meso-β scale. Vertically propagating gravity waves within and above the moist convection will be examined in Section 3 along with the downscale generation of turbulence in the lower stratosphere. The fourth section will again, as in Chapter
4, briefly compare the NHMASS turbulence in the lower stratosphere to the observations from the AMSU satellite observations. Finally, Section 5 will provide a summary and conclusions of the downscape generation of turbulence in the lower stratosphere within this mesoscale event.

6.2 NHMASS simulation results

This chapter will employ NHMASS runs with horizontal resolutions from 18 km down to 667 m. The large-scale runs will provide the basis for the favorable environment in the lower stratosphere for vertically propagating gravity waves, while the small-scale simulations were an attempt to capture the gravity waves and turbulence employing grid resolutions approaching large-eddy simulation (LES) scales. Again the 667 m NHMASS runs were one-way nested from the larger 2 km, 6 km, and 18 km NHMASS simulations, therefore the finer scale resolution runs are very dependant on their coarser scale counterparts for the correct initial and boundary condition files to correctly simulate the lower stratospheric environment.

6.2.1 Development of moist convection

As discussed in Chapter 5, the moist convection increases in coverage shortly after 1000 UTC 12 December as the exit region of the first jet streak begins to move into southern Texas [Figs. 6.1a and b]. The intensity of the moist
convection starts to increase dramatically in MASS18 at 1800 UTC as the second jet streak begins to strengthen over Oklahoma and the moist convection organizes into a squall line near the surface low-pressure center [Figs. 6.2a and b]. With the growth of the moist convection between 1000 and 1800 UTC, there is a corresponding increase of moisture in the upper troposphere associated with the growth of observed anvil in eastern Texas [Figs. 6.3a and b]. The relative humidity increases on the 330-K isentrope from <30% to a large area >80% in eastern Texas. Several perturbations in the isentropes can be seen just above jet-level when a cross-section is cut through the moist convection at 1800 UTC [Fig. 6.4]. The wavelength of these simulated gravity waves was 100 to 120 km, which was similar to the wavelength of the gravity waves seen in the Advanced Microwave Sounding Unit (AMSU) observations.

Employing MASS6 the growth of the moist convection can be seen even more clearly as the finer horizontal grid scale allows for the better definition of the environment in the lower stratosphere and the vertically propagating gravity waves. By 1900 UTC on December 12th in MASS6 the moist convection has organized into a squall line just west of HOU [Fig. 6.5a] and the upper-level environment continues to favor increased strengthening of the squall line with the favorable location of the downstream upper-level streak and the increasing divergence at 300 hPa [Fig. 6.5b]. A sharp defining moisture boundary has developed in the relative humidity field with the anvil cirrus on the 330-K isentrope at 1900 UTC as air is forced above the equilibrium level and pulled downstream by the upper-level wind flow [Fig. 6.6a]. Cutting a cross-section
through the developing squall line, a strong vertically propagating gravity wave signal is seen between 210 hPa and 150 hPa just above the moist convective towers, where the isentropes are beginning to fold or slope, and smaller gravity waves are seen well into the lower stratosphere to 60 hPa with more finer perturbation along the isentropes [Fig. 6.6b]. Examining both MASS18 and MASS6 a favorable meso-α and meso-β environment has been established for continuing strong development of moist convection as updrafts within the moist convection are overshooting the equilibrium level well into the lower stratosphere seen in the relative humidity fields [Fig. 6.7a]. With the favorable background environment in place, the following section will diagnose the fine-scale development with the squall line, the vertically propagating gravity waves associated with the moist convection, and the downscale mixing, which leads to the turbulence seen in the lower stratosphere.

6.3 Moist convective updrafts and turbulence generation

6.3.1 Convective updrafts and gravity waves

As the squall line grows there are several strong convective updrafts that extend above the tropopause, i.e, above 225 hPa, and well into the lower stratosphere. At 2300 UTC December 12th in a cross-section taken west to east through the squall line there is a convective tower that has grown to 150 hPa in the MASS2 simulation [Fig. 6.7a]. There are vertically propagating gravity waves
associated with this convective tower from 200 hPa to 50 hPa and the waves are sloping backward behind the convective tower. Thirty minutes later the overshooting convective tower has grown larger in horizontal extent and there are a series of two main vertically propagating gravity wave packets located behind the two higher overshooting convective tops [Fig. 6.7b and 6.7c]. To find the hydrostaticity of these modeled gravity waves we assume that \( U = 30 \, \text{ms}^{-1} \), \( N = 0.02 \, \text{s}^{-1} \), and \( a = 10 \, \text{km} \) (the background wind was subtracted from the moving speed of the squall line using the base reflectivity to find \( U \)). A hydrostaticity of 6.67 was calculated and is greater than 1, therefore our modeled gravity waves are realistic in their vertical propagation in this simulation. Looking at the horizontal winds at 2300 UTC in Fig. 6.8a, there are two areas of increased wind speed, one located just ahead of the vertically propagating gravity wave at 150 hPa, and then a second stronger increase in wind speed just behind the gravity wave. The same signal is again seen in the horizontal wind speed field at 2330 UTC with two wind maximums both ahead and behind the vertically propagating gravity wave and a wind minimum located with the gravity wave [Fig. 6.8b]. This same signal will be seen in later runs. A pattern can also be seen in the divergence/convergence fields with the vertically propagating gravity waves. At 2300 UTC between 100 hPa and 200 hPa, there is convergence co-located with the second wind maximum just behind the vertically propagating gravity wave [Fig. 6.9a] and this is in area of downward momentum flux in the \( W \) field [Fig. 6.9b]. Near the ridge axis of the vertically propagating gravity wave between 100 and 200 hPa there is divergence located with positive \( W \) and the local horizontal wind speed minimum. These \( W \)
fields are allowing the isentropes to steepen and break into gravity waves and finer scale turbulence [Figs. 6.9a, b, and 6.8a]. The pattern of convergence, divergence, convergence located near the vertically propagating gravity wave is again seen at 2330 UTC as the gravity wave has traveled more distance in the vertical [Fig. 6.10]. In our simulation there are gravity waves above the moist convective towers as observed in the AMSU observations.

The convergence and divergence patterns are very similar to what one would see with moist flow near a mountain with the convergence on the windward side of the mountain and the environmental flow, with the correct Froude number, rising to the top of the mountain and then allowing for an increase in wind speed on lee side of the mountain and possible down-slope winds. The moist convective process is allowing the release of energy and gravity waves and turbulence is generated above these moist convective towers as the mesoscale system grows scale. Again no vertically propagating gravity waves are seen in the NHMASS runs without latent heating as there is no moist convective process to perturb the isentropes and adjust the mass and momentum fields.

6.3.2 Turbulence generation

While it has been duly noted that there are vertically propagating gravity waves above the moist convection in the lower stratosphere, the areas for turbulence generation near the moist convection and gravity waves can be seen more clearly in the MASS667M simulation. In a cross-section cutting through the
moist convection in MASS667M vertically propagating gravity waves can be seen propagating up to 30 hPa with the same convergence/divergence pattern as in the last section [Fig. 6.11]. Focusing on an individual moist convective tower within the squall line, the turbulence kinetic energy (TKE) production is co-located with the expanding cloud anvil edge and the gravity waves associated with it [Figs. 6.12a, b, and c]. In fact, the area of TKE generation continues to expand over the 6 minute period as the anvil is spreading out and the convective tower begins to overshoot. Large magnitudes of TKE are seen 6 minutes later at 12 km as the moist convective tower and the expanding anvil have penetrated into the lower stratosphere [Figs. 6.13a and b]. TKE is associated with the convective updraft at first as the largest TKE remains near the core of the updraft, however, as the storm grows, the largest values of TKE are then associated with the cloud edge very similar to Lane et al. (2003). Using the TKE tendency equation:

\[ \frac{\partial \varepsilon}{\partial t} = -\overline{u}\frac{\partial \varepsilon}{\partial x} - \overline{v}\frac{\partial \varepsilon}{\partial y} - \overline{w}\frac{\partial \varepsilon}{\partial z} + \frac{g}{\Theta} \left[ \overline{(w' \theta')} - \overline{(u'w)} \frac{\partial \overline{u}}{\partial z} - \overline{(v'w)} \frac{\partial \overline{v}}{\partial z} + \frac{\partial}{\partial z} \left( \frac{\overline{w'}}{\rho} \right) \right] - \varepsilon \]  

(6.3.1)

at 16 km the vertically propagating gravity waves can be seen to be coincident with the mechanical shear term (fifth and sixth terms on the right side of (6.3.1)) of the TKE equation, though 4 km above the cloud top, the gravity waves are a reflection of the turbulence generation directly above the updraft area [Fig. 6.14a]. This area of turbulence occurs 6 minutes after the large amount of TKE at 12 km. Six minutes later at 16 km the turbulence generation area has moved away from the main updraft as the gravity waves travel in all directions well above the convective tower [Fig. 6.14b]. The same signal in the TKE production can be seen
with the vertically propagating gravity waves all the way to 20 km, 8 km above
the convective updraft, however, the TKE production, associated with the
vertically propagating gravity waves, at 20 km is visible 30 minutes after the
initial TKE production at 8 km.

This section briefly discussed the generation of turbulence located in the
cloud edge of the convective updrafts and with the vertically propagating gravity
waves seen in MASS667M. The results were very similar to Lane et al. (2003)
where the down scale mixing from the mechanical shear and buoyancy production
terms of the TKE tendency equation near the cloud edge of the convective
updrafts was the main source of turbulence generation. However, vertically
propagating gravity waves in the lower stratosphere did account for some
turbulence up to 20 km and a 6 minute time lag for every 2 km of vertical
distribution was discovered. A more in depth discussion of the turbulence
generation and the moist convective updrafts in this case study can be found in
Ringley (2006).

6.4 Comparison of NHMASS with AMSU

Vertically propagating gravity waves appeared in all NHMASS runs
above the moist convection as the hydrostaticity was above 1 and fortunately the
AMSU satellite observations confirm the existence of gravity waves in the lower
stratosphere above the moist convection during this case study. The vertically
propagating gravity waves in MASS18 and MASS6 have horizontal wavelengths
between 50 and 100 km [Fig. 6.15] and are similar in scale to the AMSU observation of 100 to 200 km wavelength gravity waves [Fig. 4.39b]. Also, both the NHMASS simulations and AMSU observations have vertically propagating gravity wave signals all the way to 50 hPa (not shown). These findings are similar to both Eckermann et al. (2005) and Wu and Zhang (2004). Unfortunately, the AMSU observations do not have the capability yet of resolving finer scale vertically propagating gravity, as was seen in MASS667M and MASS222M, however, the modeled gravity waves at finer scale are analogous to the gravity waves found in Lane et al. (2003) in wavelength, distribution, and amplitude.

6.5 Summary and conclusions

This chapter examined the simulated vertically propagating gravity waves in the lower stratosphere in proximity to the moist convection. The meso-α and meso-β environment in the troposphere with strong divergence aloft was conducive for the growth and organization of thunderstorms into a linear squall line feature propagating along the Gulf Coast. As the moist convection grew in strength a few of the thunderstorms began penetrating into the lower stratosphere as overshooting tops. Coarser scale NHMASS runs showed the development of vertically propagating gravity waves above the stronger moist convective towers in eastern Texas and western Louisiana. The horizontal and vertical scales of these vertically propagating gravity waves were similar to the scales of the gravity waves observed in the AMSU satellite observations. Both Wu and Zhang (2004)
and Eckermann et al. (2005) had similar results in their AMSU observation of gravity waves in the stratosphere.

At the finer resolution runs of the NHMASS, the model developed more vertically propagating gravity waves that dispersed well into the lower stratosphere. The hydrostaticity of the simulated gravity waves was calculated and found to be greater than 1, therefore the upstream titling gravity waves in the lower stratosphere were internal consistent in the model. Two main sources for the vertically propagating gravity waves were briefly discussed, as the first source region was the convective updrafts themselves penetrating the critical layer at the tropopause and the energy being dispersed through vertically propagating gravity waves. The second much smaller source for the gravity waves was the deflection of upper-levels winds around the moist convection as the overshooting tops acted as a ‘moving mountain,’ to block and deflect the background winds. Deflection of the winds was seen to aid in the dispersion of the gravity waves away from the convective tower and then allowing for wind shear near cloud edge and downscale generation of turbulence. These two sources are similar to Lane et al. (2003) and for this mesoscale gravity wave case study are discussed in much more depth in Ringley (2006). Our second hypothesis was that gravity waves that formed in lower stratosphere in the mesoscale model analogous to those which were observed. The results between the NHMASS simulations and the AMSU observations are consistent in that gravity waves did form in the lower stratosphere during this case study over the northern Gulf of Mexico and Gulf Coast states. The vertically propagating gravity waves were observed from
satellite data and the moist convection as well as its background environment were important in development of the modeled gravity waves in the lower stratosphere similar to that described by Lane et al. (2003). However, our study was unique in that the model simulations were initialized with fully 3-dimensional observed datasets and the propagating squall line exhibited much more complex structure than the simpler convective environment simulated by Lane et al. (2003). This increased realism in the initial conditions and boundary conditions makes our study unique.
Fig. 6.1 MASS18 valid for 1000 UTC 12 Dec 2002 for: a) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms$^{-1}$) (wind barbs) and b) 300 hPa isoheights (m) (white lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading).
Fig. 6.2 MASS18 valid for 1800 UTC 12 Dec 2002 for: a) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms⁻¹) (wind barbs) and b) 300 hPa isoheights (m) (white lines), wind barbs (ms⁻¹), and isotachs (knots) (colored shading).
Fig. 6.3 MASS18 330-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and relative humidity (%) (colored shading) valid for: a) 1000 and b) 1800 UTC 12 Dec 2002.
Fig. 6.4 MASS18 vertical cross section from 30.0 N; -100.0 W to 34.0 N; -89.0 W of potential temperature (K) (white lines) and horizontal wind speed (knots) (colored shading) valid for 1800 UTC 12 Dec 2002.
Fig. 6.5 MASS6 valid for 1900 UTC 12 Dec 2002 for: a) base radar reflectivity (Dbz) (colored shading) and 950 hPa winds (ms$^{-1}$) (wind barbs) and b) 300 hPa isoheights (m) (white lines), wind barbs (ms$^{-1}$), and isotachs (knots) (colored shading).
Fig. 6.6 MASS6 valid 1900 UTC 12 Dec 2002 for: a) 330-K isentrope pressure (hPa) (black lines), total wind (knots) (black wind barbs), and relative humidity (%) (colored shading) and b) a vertical cross section from 28.0 N; -100.0 W to 32.0 N; -92.0 W of potential temperature (K) (white lines) and horizontal wind speed (knots) (colored shading).
Fig. 6.7 MASS2 vertical cross section from 29.5 N; -94.6 W to 31.5 N; -92.6 W of potential temperature (K) (black lines) and relative humidity (%) (colored shading) valid for: a) 2300 and b) 2330 UTC 12 Dec 2002 and c) potential temperature (K) black lines) and W (ms$^{-1}$) (colored shading) for 2330 UTC 12 Dec 2002.
Fig. 6.7 (Continued)
Fig. 6.8 MASS2 vertical cross section from 29.5 N; -94.6 W to 31.5 N; -92.6 W of potential temperature (K) (black lines) and horizontal wind speed (knots) (colored shading) valid for: a) 2300 and b) 2330 UTC 12 Dec 2002.
Fig. 6.9 MASS2 vertical cross section from 29.5 N; -94.6 W to 31.5 N; -92.6 W valid 2300 UTC 12 Dec 2002 for: a) potential temperature (K) (black lines) and divergence (1x10^4 s^-1) (colored shading) and b) potential temperature (K) (black lines) and W (ms^-1) (colored shading).
Fig. 6.10 MASS2 vertical cross section from 29.5 N; -94.6 W to 31.5 N; -92.6 W valid 2330 UTC 12 Dec 2002 for potential temperature (K) (black lines) and divergence (1x10^4 s^-1) (colored shading).
Fig. 6.11 MASS667M vertical cross section from 30.5 N; -94.1 W to 30.6 N; -93.0 W valid 2254 UTC 12 Dec 2002 for potential temperature (K) (black lines) and divergence (1x10^4 s^-1) (colored shading).
Fig. 6.12 MASS667M 10 km TKE (m$^2$s$^{-2}$) (left colored shading) and 10km winds (ms$^{-1}$) (left wind barbs) and 10 km radar reflectivity (Dbz) (right colored shading) and 10 km winds (ms$^{-1}$) (right wind barbs) valid for: a) 2330, b) 2333, and c) 2336 UTC 12 Dec 2002.
Fig. 6.12 (Continued)
Fig. 6.13 MASS667M 12 km TKE (m$^2$s$^{-2}$) (left colored shading) and 12km winds (ms$^{-1}$) (left wind barbs) and 12 km radar reflectivity (Dbz) (right colored shading) and 12 km winds (ms$^{-1}$) (right wind barbs) valid for: a) 2336 and b) 2342 UTC 12 Dec 2002.
Fig. 6.14 MASS667M a) 12 km U-mechanical shear (m²s⁻²) (scaled by 2) (colored shading) and dU/dZ (scaled by 3) (black solid and dashed lines) valid 2342 UTC 12 Dec 2002 and b) 16 km U-mechanical shear (m²s⁻²) (scaled by 2) (colored shading) and dU/dZ (scaled by 3) (black solid and dashed lines) valid 2348 UTC 12 Dec 2002.
Fig. 6.15 MASS6 vertical cross section from 28.0 N; -100.0 W to 32.0 N; -92.0 W valid 1930 UTC 12 Dec 2002 for potential temperature (K) (white lines) and horizontal wind speed (knots) (colored shading).
7. Final Summary and Conclusions

The atmospheric event that occurred on December 12th and 13th, 2002 provided an opportunity for mesoscale gravity waves to be studied from the meso-$\alpha$ to the meso-$\gamma$ scale with the wealth of observations that were fortuitously located along the path of the mesoscale gravity wave. The large-amplitude tropospheric mesoscale gravity wave had a wavelength of $\sim$200 km, a period of 2 hours, a phase velocity of 27.75 ms$^{-1}$, and an average amplitude of $\sim$3-5 hPa (however higher values were observed and simulated), all within the criteria of Uccellini and Koch (1987) for observed mesoscale gravity waves. This mesoscale gravity wave was of particular interest, because of its length of propagation from central Texas northeastward into the Tennessee Valley. A combination of unique synoptic features came together to allow for a favorable environment for the formation and propagation of this mesoscale gravity wave including two different jet streaks, a frontal boundary, and moist convection. On finer scales, the moist convective process, momentum from the jet streak, and an area of possible shear instability played a vital role allowing for the mesoscale gravity wave to strengthen and stay organized.

With the initial work on the favorable synoptic pattern for mesoscale gravity wave development by Uccellini and Koch (1987), followed by further experiments at meso-$\beta$ and meso-$\gamma$ scales by Koch and Dorian (1988), Zack and Kaplan (1997), Trexler and Koch (2000), Rauber et al. (2001), Zhang (2004), and
Koch et al. (2005), the author believes that the hypothesis that the proximity of moist convection in between the two jet streaks allowed for juxtaposed unbalanced flow and shearing instability and the formation of a large-amplitude, deep, tropospheric, mesoscale gravity wave offers a contribution to mesoscale gravity wave research. The modeled correlation between the descending momentum from the jet streak similar to Yang et al. (2001) and the area shear instability similar to Koch et al. (2005) behind the moist convection toward a stable duct provides another source for the mass and momentum discontinuities found in a large-amplitude tropospheric mesoscale gravity wave. Figs. 7.1 and 7.2 compare the original synoptic environment found by Uccellini and Koch and then the synoptic environment found in this December 12th and 13th, 2002 mesoscale gravity wave event. In the December case study, the synoptic environment is very similar to the Uccellini and Koch paradigm with the mesoscale gravity wave occurring near the inflection point of the 500 hPa isoheights, north of a surface frontal boundary and in the exit region of a 300 hPa jet streak. However, the mesoscale gravity wave event in December 2002 does differ from the Uccellini and Koch paradigm in that moist convection and the non-linear processes therein were necessary for the formation of the mesoscale gravity wave. The location of the moist convection, development of the secondary downstream jet streak, and the magnitude of the jet momentum on the synoptic scale were very important in the mesoscale gravity wave formation in the December 2002 case study [Fig. 7.2]. The mesoscale gravity wave on the meso-γ is seen in Yang et al. (2001), [Fig. 7.3] however, the interaction between the moist convection and momentum, and the
reflection of gravity waves overturning through an area of shear instability near
the stable duct is missing. Fig. 7.4 updates Fig. 7.3 to include the interaction
between moist convection, an area of shear instability and the descending
momentum as those three processes were important in the formation/maintenance
of the mesoscale gravity wave on December 12 and 13, 2002. While more work is
necessary to find the differences in small and large amplitude gravity wave
formation environments, this work furthers the understanding in mesoscale
gravity wave formation/maintenance to note the importance of the interaction
between moist convection, shear instability, and jet momentum on the meso-γ
scale.

The mesoscale gravity wave event on December 12 and 13, 2002 was first
discussed in Chapter 4 with the observations of the event from the meso-α to the
meso-γ scale. Chapter 4 summarized the development and movement of the
mesoscale gravity waves in the lower troposphere and in the lower stratosphere
from an observational standpoint. The lower tropospheric mesoscale gravity wave
system was found to have a similar synoptic organization as proposed by
Uccellini and Koch (1987) as the mesoscale gravity formed in the inflection point
in the height field at 500 hPa, the exit region of a jet streak at 300 hPa, and north
of a surface frontal boundary. Observations also clearly show gravity waves
propagating vertically into the lower stratosphere above the convection. This is
similar to the observations described in Wu and Zhang (2004) where they
observed vertically propagating gravity waves in the lower stratosphere in the exit
region of a jet streak in the north Atlantic. Three key sources mechanisms of geostrophic adjustment, moist convection, and shear instability were observed across Texas and Louisiana and the mechanisms allowed for initiation and propagation of the mesoscale gravity in the lower troposphere northeastward into Mississippi.

Chapter 5 examined the mesoscale gravity wave event through the simulations from NHMASS and further confirmed the hypothesis that the proximity of moist convection to the jet streak exit region allowed for juxtaposed unbalanced flow and shearing instability and the formation of a large-amplitude, deep, tropospheric, mesoscale gravity wave. While several studies have found geostrophic adjustment and shear instability to be important in the mesoscale gravity wave system, this study is the first to link the three processes of geostrophic adjustment, moist convection, and shear instability in the formation/maintenance of a mesoscale gravity wave event. These three processes, which interacted from the meso-$\alpha$ to the meso-$\gamma$ scale, became in-phase with one another allowing for the development of the mesoscale gravity wave during December 12th and 13th, 2002. Fig. 7.4 shows a three-part development/maintenance of the mesoscale gravity wave system. In part 1 the moist convection has already become organized north of the frontal boundary under a favorable synoptic environment and momentum from the jet streak is beginning to interact with the moist convection [Fig 7.4a]. Mass perturbations from the moist convection are directed upward and cross-stream into the jet streak momentum allowing for the perturbation of the isentropic field. The vertical
pressure gradient force tries to restore the mass and momentum fields quickly with a shear instability beginning to develop near the back of the moist convection as the downdraft begins to mix some momentum downward toward the stable layer [Fig. 7.4b]. The mesoscale gravity wave(s) is initiated/maintained as the downdraft interacts with the stable layer near the surface behind the moist convection [Fig. 7.4c]. Also, vertically propagating gravity waves begin to develop above the moist convection as momentum from both the jet and the moist convective updraft penetrating into the lower stratosphere [Fig. 7.4c].

An analysis of the modeled vertically propagating gravity waves was done in Chapter 6. Gravity waves were seen to form above the moist convection in a similar fashion as modeled by Lane et al. (2003). Momentum from the updraft and the deflections of the environmental winds was seen to aid in shear production near the gravity waves and further the generation of turbulence [Fig. 7.4c]. However, this part of the mesoscale gravity wave case study is discussed in much more depth in Ringley (2006). While similarities existed between our Chapter 6 and Lane et al. (2003), our study was unique in that the model simulations were initialized with fully 3-dimensional observed datasets and the propagating squall line exhibited much more complex structure than the simpler convective environment simulated by Lane et al. (2003).

The completely mesoscale gravity wave system from the meso-α to meso-γ scale was examined in this December 12 and 13, 2002 event as the non-linear interactions between geostrophic adjustment, moist convection, and shear instability allowed for the formation of a deep large-amplitude mesoscale gravity
wave. However, more work needs to be done to better understand mesoscale gravity waves. For instance, is the dry air a causal mechanism in the mesoscale gravity wave system, or is it a signal of momentum mixing downward or subsidence? Do other mesoscale gravity wave events, beside Yang et al. (2001) and this study, have a signal of dry air behind the moist convection in the exit region of the jet streak? What is the difference in a mesoscale gravity wave event that leads to severe weather and an event that does not? Noting the favorable background environment and causal mechanisms on all scales will allow for a better understanding of mesoscale weather and will provide a better forecast of mesoscale gravity wave events and associated damages to property and loss of life.
Fig. 7.1 Diagram showing the favorable meso-\(\alpha\) scale environment for mesoscale gravity waves with the 300 hPa jet streak, 500 hPa inflection point, and surface frontal boundaries. Source: Uccellini and Koch (1987).
Fig. 7.2 Updated favorable meso-α scale environment for mesoscale gravity waves with the 300 hPa jet streaks, 500 hPa isoheights, surface frontal boundaries, and moist convection location for the December 12 and 13, 2002 case study.
Fig. 7.3 (top) Schematics of the evolution of the dry air mass (dark shading) and its relationship to convection (light shading), fronts, and the surface cyclone based on analyses at (a1) 1400 and (a2) 2100 UTC. (bottom) conceptual model of circulations within the leading edge of the dry air mass, the gravity wave, and the convection. The background shading is the radar reflectivity. The foremost light shading is the dry air mass. The darkest shading is the stable layer below the warm frontal surface. Black arrows denote schematically the circulations both within the dry air mass and ahead of it. The white arrows denote the sense of vertical motion. The dashed lines denote zero vertical velocity. Letters C and D denote convergence and divergence, respectively; H and L denote regions of high and low perturbation pressure, respectively. Source: Yang et al. (2001).
Fig. 7.4 Favorable meso-γ scale environment for mesoscale gravity waves with: a) momentum from the jet streak propagating into upward and cross-stream propagating mass perturbations from the moist convective towers, b) momentum from the jet streak being directed around the moist convective towers and shear instability helping to transfer momentum downward toward the stable layer with mass accumulations, and c) the downward cascade of momentum continuing with the reinforcement of the deep large-amplitude tropospheric mesoscale gravity wave and the development of vertically propagating gravity waves in the lower stratosphere above the moist convective towers.
8. References


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