

Understanding Satellite-Observed Mountain-Wave Signatures Using High-Resolution Numerical Model Data

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ABSTRACT

Prior work has shown that pilot reports of severe turbulence over Colorado often occur when complex interference or crossing wave patterns are present in satellite water vapor imagery downstream of the Rocky Mountains. To improve the understanding of these patterns, a high-resolution (1-km) Weather Research and Forecasting (WRF) model simulation was performed for an intense mountain-wave event that occurred on 6 March 2004. Synthetic satellite imagery was subsequently generated by passing the model-simulated data through a forward radiative transfer model. Comparison with concurrent Moderate Resolution Imaging Spectroradiometer (MODIS) water vapor imagery demonstrates that the synthetic satellite data realistically captured many of the observed mesoscale features, including a mountain-wave train extending far downstream of the Colorado Front Range, the deformation of this wave train by an approaching cold front, and the substantially warmer brightness temperatures in the lee of the major mountain ranges composing the Colorado Rockies. Inspection of the model data revealed that the mountain waves redistributed the water vapor within the lower and middle troposphere, with the maximum column-integrated water vapor content occurring one-quarter wavelength downstream of the maximum ascent within each mountain wave. Due to this phase shift, the strongest vertical motions occur halfway between the locally warm and cool brightness temperature couplets in the water vapor imagery. Interference patterns seen in the water vapor imagery appear to be associated with mesoscale variability in the ambient wind field at or near mountaintop due to flow interaction with the complex topography. It is also demonstrated that the synergistic use of multiple water vapor channels provides a more thorough depiction of the vertical extent of the mountain waves since the weighting function for each channel peaks at a different height in the atmosphere.

1. Introduction

Atmospheric turbulence is a major aviation hazard responsible for 609 fatalities, 823 injuries, and an estimated property loss of \$134M from 1983 to 1997 (Eichenbaum 2000). Aircraft turbulence is often associated with rapid thunderstorm development (e.g., Lane et al. 2003), upper-tropospheric folding events (e.g., Endlich 1964; Koch et al. 2005), and topographically induced mountain-wave and rotor phenomena (e.g., Reiter and Foltz 1967; Clark and Gall 1982; Clark et al. 2000; Doyle and Durran 2002). Ground-based and aircraft observations are useful for depicting and understanding the 3D structure and time evolution of mountain waves, but only provide this information for

select locations and time periods (Jiang and Doyle 2004; Doyle et al. 2005; Grubišić and Billings 2007). These sophisticated observations are often collected in association with field experiments and may not be available for use in operational turbulence diagnosis and forecasting. In some cases, the existence of clear-air turbulence associated with mountain waves is “validated” by aircraft that experience the turbulence (Clark et al. 2000), incidents which can cause serious injury to crew/passengers and damage to the aircraft itself.

High-resolution satellite water vapor imagery from the Moderate Resolution Imaging Spectroradiometer (MODIS) allows forecasters to identify the presence of mountain waves and infer the potential for hazardous turbulence. Mountain waves are often clearly evident in MODIS 6.7- μm water vapor channel imagery, but appear only to a limited extent in visible and 11- μm infrared window imagery where moisture and vertical

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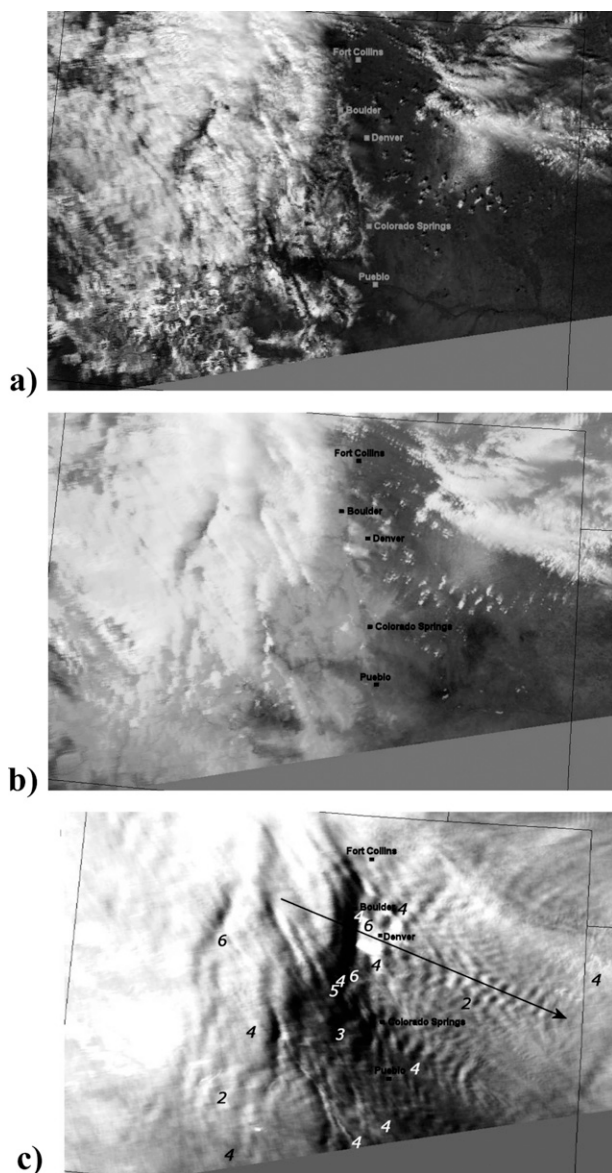


FIG. 1. MODIS (a) visible, (b) infrared window, and (c) water vapor band imagery illustrating the utility of the $6.7\text{-}\mu\text{m}$ WV channel for observing mountain-wave features. Plotted numbers in (c) represent the PIREP turbulence intensity within ± 2 h of the MODIS image. Generally, any number between and including 4 and 7 is considered moderate to severe turbulence.

motions are sufficient to form clouds (Fig. 1). In a climatological study of mountain-wave events over Colorado, Uhlenbrock et al. (2007) found that pilot reports of severe turbulence often occur when complex interference or crossing wave patterns are evident in the MODIS water vapor imagery. Satellite images with simpler linear wave features or no interference patterns were associated with far fewer turbulence reports.

This paper presents a unique methodology for com-

binning numerical weather prediction (NWP) model output and satellite water vapor imagery (both observed and synthetic) to better understand satellite-observed mountain-wave signatures. A high-resolution (1 km) Weather Research and Forecasting (WRF) model simulation was performed for an intense mountain-wave event that occurred on 6 March 2004 in order to gain new insight into the underlying dynamics and mesoscale moisture variability associated with mountain waves. WRF model-simulated temperature and moisture profiles are used to produce synthetic satellite water vapor imagery at high spatial and temporal resolutions using an infrared radiative transfer model. This provides the direct link between the model dynamic fields and the synthetic satellite imagery required to perform this study.

This event was chosen because it exhibited a well-developed interference pattern in the MODIS imagery (Fig. 1c) and was characterized by the highest frequency of pilot reports of severe turbulence in this region during 2004. Figure 1c includes pilot reports (PIREPs) of turbulence reported around the time of the satellite observation. These reports categorize turbulence on a subjective numerical scale with zero equal to no turbulence and nine equal to extreme turbulence. The scale reflects the “bumpiness in flight” and, generally, any number between and including four and seven is considered moderate to severe turbulence.

Section 2 describes the WRF model simulation and infrared radiative transfer model used to produce the synthetic satellite brightness temperatures. Section 3 describes the results of the WRF simulation and compares observed and synthetic MODIS water vapor imagery. Section 4 provides a discussion on additional applications that could be done to better diagnose mountain-wave structure as well as a look forward toward the capability of the future Geostationary Operational Environmental Satellite-R (GOES-R) Advanced Baseline Imager (ABI) for observing mountain waves. The results of this study are summarized in section 5.

2. Model descriptions

The WRF model is a sophisticated numerical weather prediction model that solves the compressible nonhydrostatic Euler equations cast in flux form on a mass-based terrain-following vertical coordinate system. Prognostic variables include the horizontal and vertical wind components, various microphysical quantities, and the perturbation potential temperature, geopotential, and surface pressure of dry air. High-resolution global datasets are used to initialize the model topography and other static surface fields. A

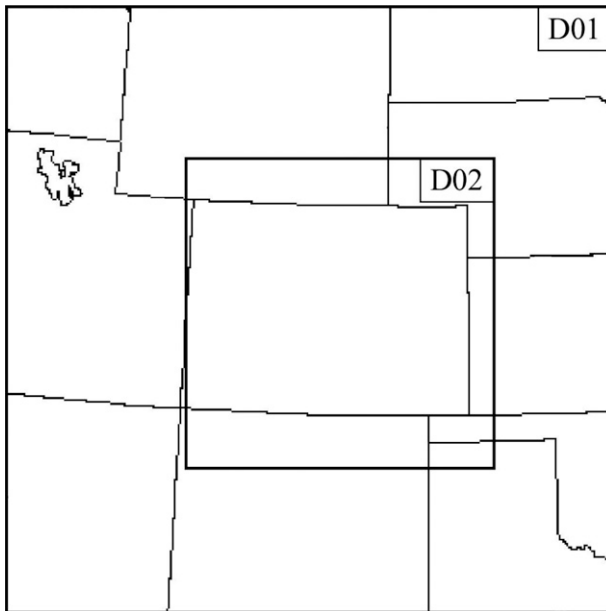


FIG. 2. WRF model domain configuration: D1 contains 260×260 grid points with 5-km horizontal resolution and D02 contains 661×661 grid points with 1-km resolution.

complete description of the WRF modeling system is contained in Skamarock et al. (2005).

A high-resolution model simulation was performed using version 2.2 of the WRF model to study the underlying structure of the mountain-wave pattern seen in Fig. 1. The simulation was initialized at 0000 UTC on 6 March 2004 using 1° Global Data Assimilation Sys-

tem (GDAS) analyses and then integrated for 24 h on two nested domains containing 5- and 1-km horizontal grid spacing, respectively (Fig. 2). Simulated atmospheric fields from the outer 5-km domain provided high-resolution lateral boundary conditions for the higher-resolution inner domain. Figure 3 shows the model topography for the inner domain, which was initialized using the high-resolution (30 s) U.S. Geological Society (USGS) topography dataset. The simulation contained 65 vertical levels with the model top set to 10 hPa. The vertical resolution decreased from <100 m in the lowest 1 km to ~ 440 m from 7 km to the model top. Subgridscale processes were parameterized using the Thompson et al. (2008) mixed-phase cloud microphysics scheme, the Mellor–Yamada–Janjić planetary boundary layer scheme (Mellor and Yamada 1982), and the Dudhia (1989) shortwave and Rapid Radiative Transfer Model (RRTM) longwave radiation (Mlawer et al. 1997) schemes. Surface heat and moisture fluxes were calculated using the Noah land surface model. No cumulus parameterization scheme was used; therefore, all clouds were explicitly predicted by the microphysics scheme.

WRF model output, including the surface skin temperature, atmospheric temperature, water vapor (WV) mixing ratio, and the mixing ratio and effective particle diameters for five hydrometeor species (cloud water, rainwater, ice, snow, and graupel), were ingested into the Successive Order of Interaction (SOI) forward radiative transfer model (Heidinger et al. 2006) to generate synthetic top of atmosphere brightness temperature

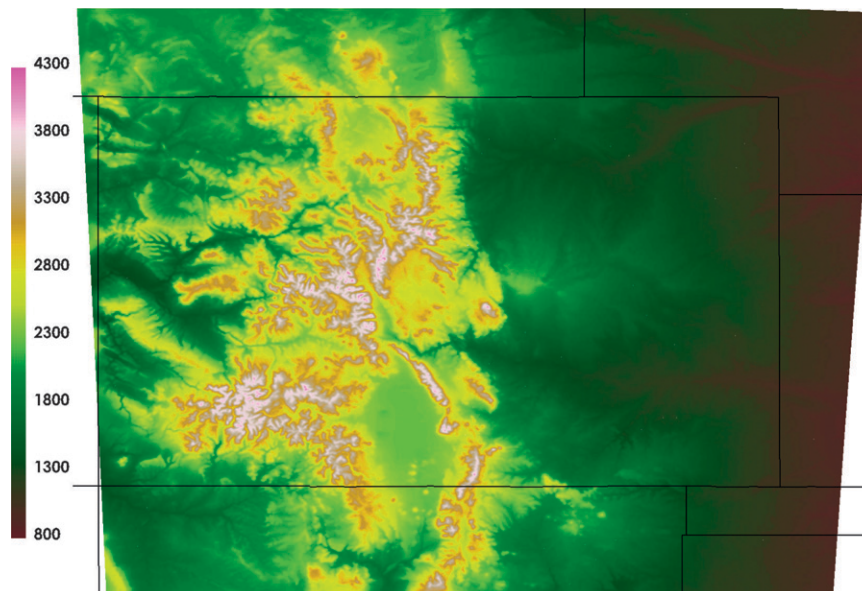


FIG. 3. The 1-km resolution topography of the CO Rockies used as input within the WRF–Advanced Research WRF (ARW) simulation.

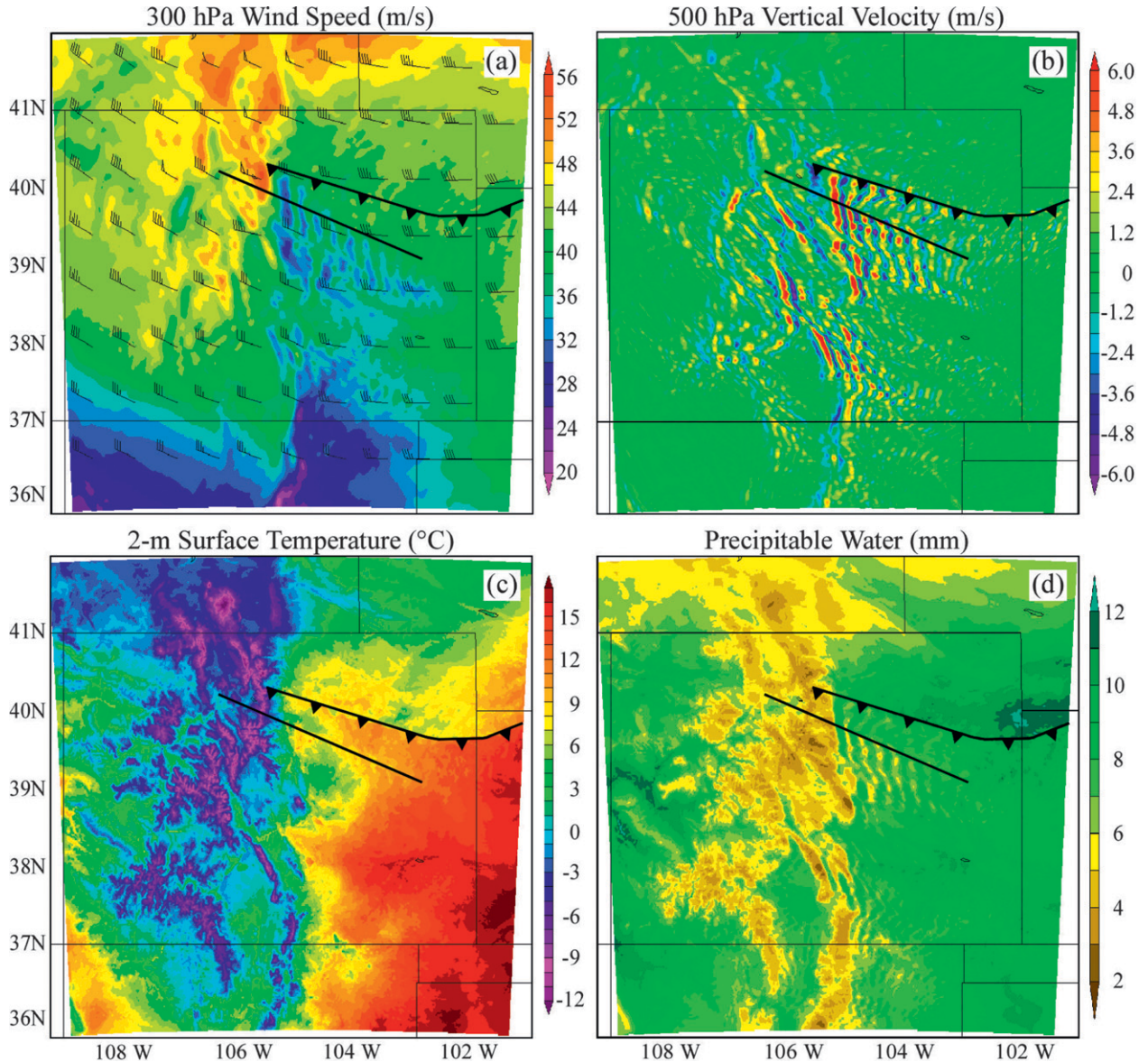


FIG. 4. WRF-simulated (a) 300-hPa wind speed (m s^{-1}) and wind barbs, (b) 500-hPa vertical velocity (m s^{-1}), (c) 2-m surface temperature ($^{\circ}\text{C}$), and (d) precipitable water (mm) valid at 2000 UTC 6 Mar 2004. The location of the surface cold front is indicated by the curved black line while the straight black line indicates the location of the cross section shown in Fig. 5.

(T_B) imagery. Gas optical depths were calculated for each MODIS infrared band using the National Oceanic and Atmospheric Administration (NOAA) Community Radiative Transfer Model. Ice cloud absorption and scattering properties were obtained from Baum et al. (2006), whereas the liquid cloud properties were based on Lorenz–Mie calculations.

3. Results

This section begins with an overview of the model-simulated atmospheric conditions at 2000 UTC 6 March

2004. At 300 hPa (Fig. 4a), the large-scale flow was nearly perpendicular to most of the major mountain ranges across central Colorado (refer to Fig. 3). Strong upward momentum transport during the previous 10 h (not shown) had shifted the jet stream vertically upward so that relatively weak winds were now present over eastern Colorado. The wind pattern over east-central Colorado is a manifestation of an impressive high-amplitude wave train extending ~ 300 km downstream of the Front Range (Fig. 4b). Vertical velocities within the first few wavelengths exceeded 4 m s^{-1} in the midtroposphere, with progressively weaker

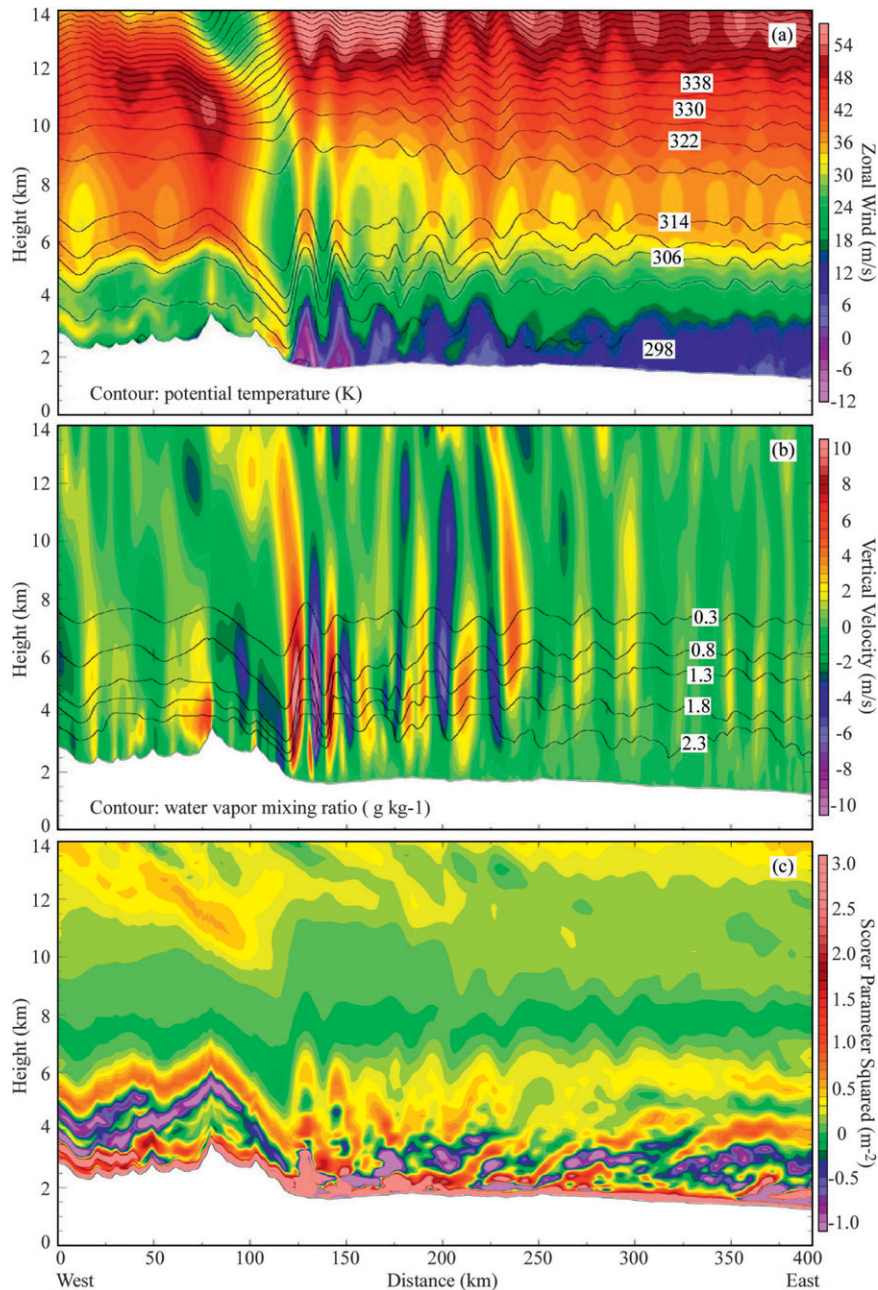


FIG. 5. Vertical cross section of WRF-simulated (a) zonal wind speed (m s^{-1} ; colors) and potential temperature (K; contoured every 4 K and labeled every 8 K), (b) vertical velocity (m s^{-1} ; colors) and water vapor mixing ratio (g kg^{-1} ; contoured every 0.5 g kg^{-1} from 0.3 to 2.3 g kg^{-1}), and (c) Scorer parameter squared (m^{-2} ; 10^{-8}) valid at 2000 UTC 6 Mar 2004. The location of the vertical cross section is shown in Fig. 4.

ascent–descent couplets farther downstream. The northern end of the mountain-wave train was beginning to weaken and shift toward the south as unfavorable conditions for trapped lee waves began to move into the region behind a low-level cold front advancing from the north (Fig. 4c). Due to the long duration of this

event and the quasi-stationary location of each individual circulation pattern, the mountain-wave train was able to fundamentally alter the horizontal distribution of the column-integrated water vapor content (Fig. 4d).

Vertical cross sections through the middle of the wave train at 2000 UTC are shown in Fig. 5. The large-

scale environment was characterized by an elevated stable layer near the mountaintop, a deep well-mixed layer in the upper troposphere, and moderately strong zonal winds. The thermodynamic structure was conducive to the development of trapped lee waves since the Scorer parameter rapidly decreased with height above the stable layer both upstream and downstream of the Front Range. As the large-scale flow crossed over the mountain barrier, strong downslope winds developed in the immediate lee of the mountains, while a long train of trapped lee waves occurred farther downstream. The depth and quasi-steady location of each circulation pattern allowed a given mountain wave to redistribute the WV within the entire troposphere such that the maximum mixing ratio occurred one-quarter wavelength downstream of the maximum ascent. Since satellite WV imagery is sensitive to such changes, the coldest (warmest) T_B will generally occur within the crest (trough) of a given wave where the WV mixing ratio is the highest (lowest). It should also be noted that the phase shift between the vertical velocity and WV mixing ratio means that the maximum ascent and descent responsible for modulating the vertical moisture distribution occur roughly halfway between the warmest and coldest T_B .

To evaluate the realism of the model-simulated mountain-wave pattern seen in Fig. 4, synthetic MODIS T_B values were generated using the SOI forward radiative transfer model. Since the WRF model generated excessive upper-level cirrus cloud downwind of the Rocky Mountains, which would obscure the mountain-wave patterns in the synthetic satellite water vapor imagery, all of the cloud hydrometeor mixing ratios were set to zero before conducting the forward model calculations. Although the resultant “clear sky” T_B will generally be warmer than the corresponding “cloudy” temperatures for cloudy grid points, this bias should be relatively small for clouds occurring within the mountain-wave region since the MODIS WV bands are not highly sensitive to low-level clouds.

Comparison of synthetic and observed MODIS 6.7- μm T_B imagery (Fig. 6) indicates that the WRF simulation provides a realistic representation of the complex mountain-wave pattern over eastern Colorado. The warmest T_B s in both images correspond to regions of strong descent in the lees of major mountain ranges. These areas of descent match up very well in both the synthetic and observed imagery since the topography is well represented by the model. Wave signatures are evident both within and in the lee of the Rockies. Lee waves extend far downstream (~ 300 km) of the mountains in both images and exhibit the complex “interference” patterns described by Uhlenbrock et al. (2007).

The low-level cold front in NE Colorado appears as a subtle discontinuity that appears to interfere with and suppress waves. Though not explicitly shown in this paper, the cold bias in the synthetic data appears to be due to excessive upper-level WV in the GDAS initialization dataset, which causes the synthetic satellite WV signal to peak at a higher (i.e., colder) atmospheric level due to enhanced water vapor absorption.

Figure 7 compares the observed and simulated MODIS 6.7- μm WV T_B transects drawn in Fig. 6. Although the leeside warming downstream of the Front Range (near 125 km) is much larger for the MODIS observations, the location and width of this feature are similar for each transect. Farther downstream, the average horizontal wavelength (~ 17 km) and position of the waves is also similar between the observation and the synthetic T_B data. The simulated wave amplitude (as measured by the peak-to-trough T_B difference) is larger than observed within the approximately 140–220 km range, but closely matches the observations for the remainder of the transect.

WRF-simulated temperature and WV profiles were used to compute weighting function profiles for each MODIS WV band to gain a greater understanding of the approximate height and multispectral characteristics of the mountain-wave signatures observed by MODIS. A weighting function profile indicates the relative contribution of an atmospheric level to the signal detected by a satellite for a given spectral band. Figure 8 shows representative weighting function profiles from the model simulation for two MODIS WV absorption bands (6.7 and 7.3 μm) and the infrared (IR) “window” band (11 μm), as well as the three WV bands that will be included on the future GOES-R ABI sensor (which will be discussed in section 4). The WV emission over a relatively deep layer in the middle and upper troposphere contributes most of the detected signal for the MODIS WV bands, whereas the IR window band primarily observes the boundary layer and surface characteristics. For this example, the peak of the 6.7- μm weighting function profile occurred at ~ 7 km, which is ~ 2.5 km above the peak of the corresponding 7.3- μm weighting function. The vertical distribution and magnitude of the total-column WV strongly influences the shape and peak of these weighting functions. For instance, decreasing (increasing) the WV content in the middle and upper troposphere causes more of the detected signal to originate from lower (higher) in the atmosphere, thereby producing a warmer (cooler) T_B . Clouds can also modulate the shape of the weighting function profile, producing a more discrete peak at the top of the cloud.

Figure 9 shows a vertical cross section of the poten-

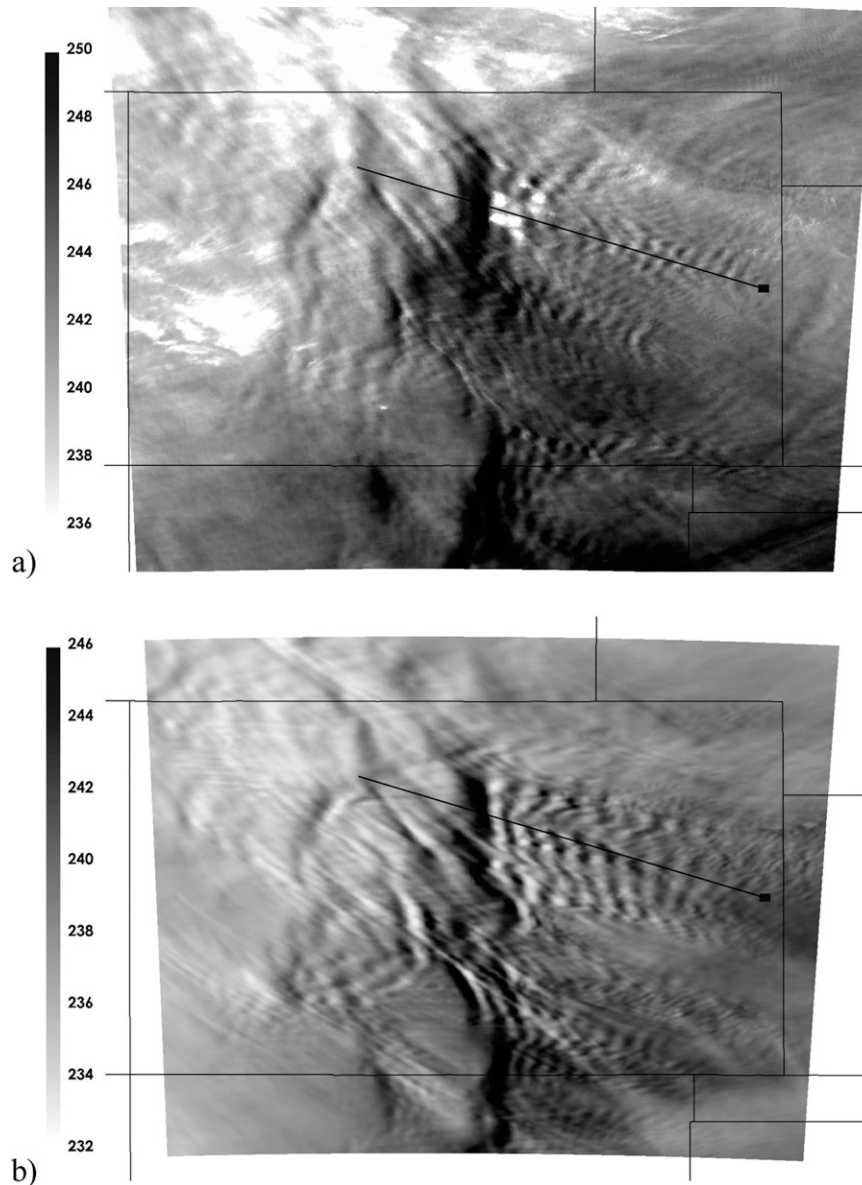


FIG. 6. (a) Observed and (b) synthetic MODIS band 27 ($6.7\ \mu\text{m}$) WV T_B fields. The same enhancement table was used for both sets of imagery, but was shifted 4 K colder to account for a temperature bias in the synthetic WV field.

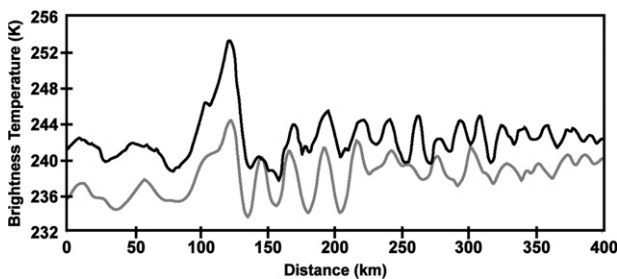


FIG. 7. Observed (black) and synthetic (gray) MODIS $6.7\text{-}\mu\text{m}$ WV T_B transects along the line shown in Fig. 5.

tial temperature through the mountain-wave train, as well as the heights of the weighting function peaks for both of the synthetic MODIS WV bands. Large variations in the height of the peak WV signal are evident, which underscores the difficulty in assigning a static height to mountain-wave signatures observed in clear-sky WV imagery. The $6.7\text{-}\mu\text{m}$ peak transmission height varies from ~ 6 to ~ 7 km with relatively little variability within the mountain-wave train itself, whereas the $7.3\text{-}\mu\text{m}$ signal originates much lower in the atmosphere and contains substantially more spatial variability. Com-

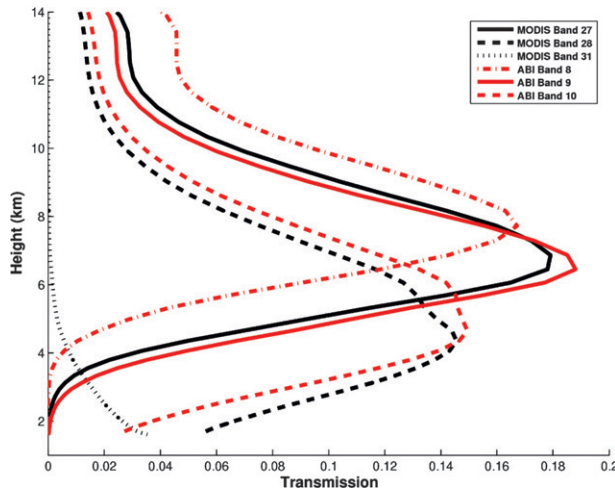


FIG. 8. An example of synthetic clear-sky weighting functions for MODIS WV absorption bands 27 ($6.7\ \mu\text{m}$) and 28 ($7.3\ \mu\text{m}$), and the IR window band 31 ($11\ \mu\text{m}$), in addition to WV bands 8 ($6.2\ \mu\text{m}$), 9 ($7.0\ \mu\text{m}$), and 10 ($7.3\ \mu\text{m}$) for the future GOES-R ABI instrument.

parison of the potential temperature and peak weighting function fields indicates that the $6.7\text{-}\mu\text{m}$ and $7.3\text{-}\mu\text{m}$ bands can be used to observe different vertical levels of the mountain-wave structure. The enhanced spatial variability of the $7.3\text{-}\mu\text{m}$ data, which more closely follows the undulations of the isentropic surfaces, is due to larger horizontal WV gradients between 4 and 7 km (refer to Fig. 5). The relative importance of each channel in diagnosing mountain waves may be a function of the wave structure and vertical moisture distribution.

As mentioned in the introduction, the 6 March 2004 mountain-wave case study was chosen due to its high number of severe aircraft turbulence reports and the presence of a “herringbone” or wave interference pattern in the MODIS WV imagery. Although a thorough

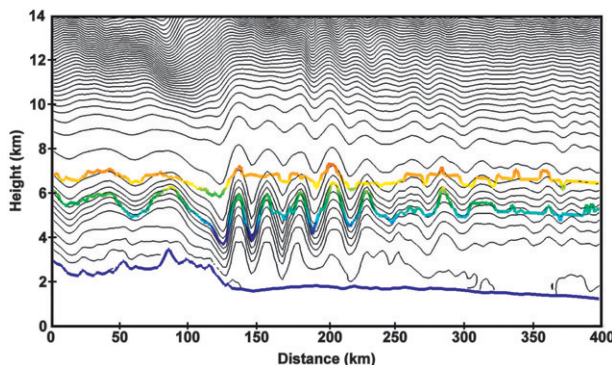


FIG. 9. The weighting function peak height for synthetic MODIS WV bands 27 (warm colors) and 28 (cool colors), overlaid upon a potential temperature (black lines) and terrain (dark blue line) cross section along the line shown in Fig. 6. Potential temperature is contoured every 2 K.

evaluation of the causes of this interference pattern is beyond the scope of this study, some aspects of the pattern will be examined in this section. Figure 10 shows the synthetic $7.3\text{-}\mu\text{m}$ T_B , the simulated 650-hPa wind speed and direction, and the model topography above 2.5 km. The $7.3\text{-}\mu\text{m}$ band was chosen for this analysis as it better observes the high-amplitude mountain-wave structure (refer to Fig. 9), which allows for an easier visual identification of complex interference patterns. The 650-hPa level was chosen since it corresponds to flow at or slightly above the top of the mountain ranges. Inspection of the synthetic imagery reveals several regions of wave suppression and interference across eastern Colorado (Fig. 10a). The nearly parallel orientation of the mountain-wave train to the 650-hPa flow indicates that the wind direction at the top of a mountain range can be inferred from the general orientation of the downstream mountain-wave field. Two corridors of suppressed wave activity over southeastern Colorado correspond to low wind speed regions downstream of Pikes Peak and the Wet Mountains in central Colorado. The isolated nature of Pikes Peak allows the air to flow around it rather than over it, which suppresses the mountain-wave activity and causes a weak wind “wake” to develop downstream of the obstacle (Smith and Gronas 1993; Ólafsson and Bougeault 1996). Although the Wet Mountains have a much longer ridge dimension, the orientation of the flow parallel to the ridgeline reduces the cross-sectional area of the mountains; thus, they behave in a similar manner to an isolated peak.

The wave pattern exhibits substantial deformation along the periphery of the wake regions and near other localized wind speed minima–maxima. Some of the localized fluctuations in the wind field may be due to small-scale topographic effects (such as isolated peaks and valleys within the individual mountain ranges), interactions between waves with different orientations, or changes in the stability and momentum profiles along the leading edge of the cold front. Though no single cause can be assigned, the complex shape and orientation of the Colorado Rockies and the complex interaction between the topography and the prevailing flow could create a very complex wave pattern in certain situations. Additional sensitivity tests will need to be performed to thoroughly evaluate the causes of the wave pattern.

4. Discussion

Additional information about mountain-wave vertical extent and temporal persistence could be inferred through analysis of wave patterns in $6.7\text{-}\mu\text{m}$ and $7.3\text{-}\mu\text{m}$

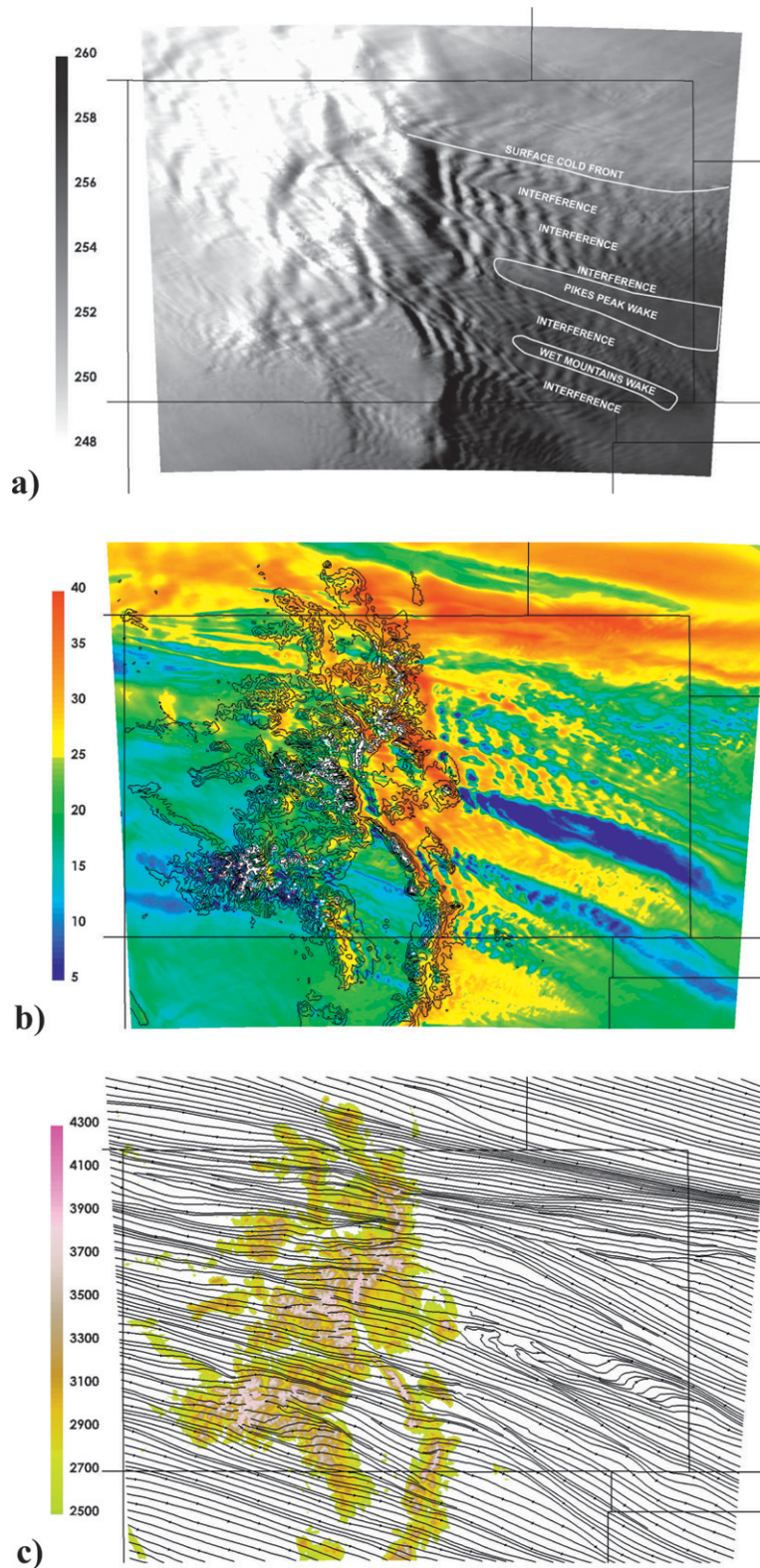


FIG. 10. (a) Synthetic MODIS band 28 (7.3 μm) imagery valid at 2000 UTC 6 Mar 2004. Wake regions and wave interference patterns are labeled with white text. (b) WRF 650-hPa wind speed (colors) with topography above 2500 m (black contours). Locations with topography above the 650-hPa level are shown in white. (c) The 650-hPa wind streamlines with colored topographic height (m).

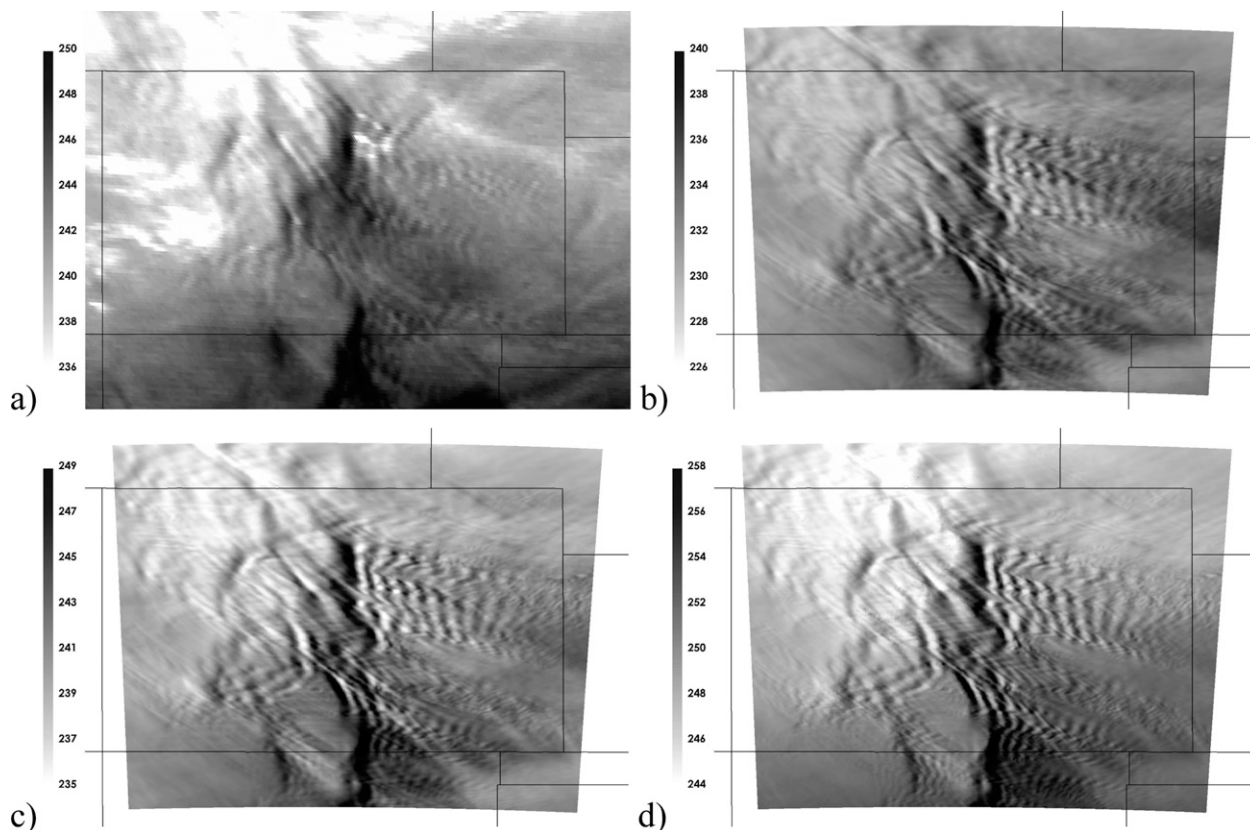


FIG. 11. (a) *GOES-12* 4-km 6.5- μm WV band observations at 2002 UTC. (b)–(d) Synthetic 2-km resolution imagery at 2000 UTC for the three GOES-R ABI WV bands: (b) band 8 (6.2 μm), (c) band 9 (7.0 μm), and (d) band 10 (7.3 μm). The grayscale enhancement is fixed to the warmest temperature in the image and is applied over a 14-K range to make the wave features apparent in all spectral bands.

imagery coupled with retrievals of atmospheric moisture content. As in this event, if a wave signature is clearly evident in both images, this implies a deep, vertically oriented, trapped wave extending through much of the troposphere. Waves are likely confined to the middle to lower troposphere if the wave signature is only observed in 7.3- μm imagery. Also, as the WRF-simulated total precipitable water (TPW) field mirrors the wave signatures, this implies that TPW derived at 1–2-km resolution from a satellite sensor could be another feature indicating the amplitude and temporal persistence of mountain waves.

The three WV bands that will be available on GOES-R ABI will provide additional information on the wave vertical extent, as these bands peak at three disparate levels in the troposphere (Schmit et al. 2005). Figure 11 shows synthetic imagery from the three future ABI WV bands at the 2-km ABI spatial resolution (Figs. 11a–c), in comparison to 6.5- μm WV imagery observed by the current *GOES-12* Imager. Wave presence and spatial variability in the wave pattern are evident in the northern half of the domain in the ABI band

8 (6.2 μm) imagery, which peaks at the highest level of the three WV bands (refer to Fig. 8). Features in the southern half appear much clearer in ABI bands 9 (7.0 μm) and 10 (7.3 μm). This suggests that variability in wave vertical extent exists across the domain and that some combination of imagery from the three ABI channels could help infer this information. Waves are evident in *GOES-12* imagery as well, but the relatively coarse 4-km spatial resolution (at nadir, ~ 5.5 km over Colorado) and single WV channel cannot capture the important spatial details and vertical structure information that will be observed by the ABI.

5. Conclusions

A 1-km horizontal-resolution WRF model simulation was performed for a significant aircraft turbulence event over Colorado to better understand mountain-wave signatures observed in satellite WV imagery. WRF thermodynamic profiles were combined with gaseous absorption profiles within a sophisticated IR radiative transfer model to produce synthetic satellite T_{BS} for the MODIS and future GOES-R ABI spectral bands.

Overall, the synthetic imagery realistically captured many of the features observed within the concurrent MODIS WV imagery, including a mountain-wave train extending far downstream of the Colorado Front Range, the deformation of this wave train by an approaching cold front, and the substantially warmer T_B s in the lee of the major mountain ranges composing the Colorado Rockies. Inspection of the model data revealed that the mountain waves redistributed the WV within the entire atmospheric column, with the maximum column-integrated WV content occurring one-quarter wavelength downstream of the maximum ascent within each mountain wave. Due to this phase shift, the strongest vertical motions occur halfway between the locally warm and cool T_B couplets in the WV imagery. The warmest WV T_B regions correspond to descent on the lees of mountain ranges oriented perpendicular to the flow. Interference patterns seen in the WV imagery are associated with finescale variability in the ambient wind field at or near mountaintop due to flow interaction with the complex topography.

It is also demonstrated that the synergistic use of multiple WV channels provides a more thorough depiction of the vertical extent of the mountain waves since the weighting function for each channel peaks at a different height in the atmosphere. Analysis of synthetic GOES-R ABI imagery, with its 2-km spatial resolution and three WV absorption band channels, suggests that ABI will provide a significant improvement over the current *GOES-12* Imager for observing mountain waves.

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REFERENCES

- Baum, B. A., P. Yang, A. J. Heymsfield, S. Platnick, M. D. King, Y.-X. Hu, and S. T. Bedka, 2006: Bulk scattering properties for the remote sensing of ice clouds. Part II: Narrowband models. *J. Appl. Meteor. Climatol.*, **44**, 1896–1911.
- Clark, T. L., and R. Gall, 1982: Three-dimensional numerical model simulations of airflow over mountainous terrain: A comparison with observations. *Mon. Wea. Rev.*, **110**, 766–791.
- , W. D. Hall, R. M. Kerr, D. Middleton, L. Radke, F. M. Ralph, P. J. Neiman, and D. Levinson, 2000: Origins of aircraft-damaging clear-air turbulence during the 9 December 1992 Colorado downslope windstorm: Numerical simulations and comparison with observations. *J. Atmos. Sci.*, **57**, 1105–1131.
- Doyle, J. D., and D. R. Durran, 2002: The dynamics of mountain-wave-induced rotors. *J. Atmos. Sci.*, **59**, 186–201.
- , M. A. Shapiro, Q. Jiang, and D. L. Bartels, 2005: Large-amplitude mountain wave breaking over Greenland. *J. Atmos. Sci.*, **62**, 3106–3126.
- Dudhia, J., 1989: Numerical study of convection observed during the Winter Monsoon Experiment using a mesoscale two-dimensional model. *J. Atmos. Sci.*, **46**, 3077–3107.
- Eichenbaum, H., 2000: Historical overview of turbulence accidents. MCR Federal Rep. TR-7100/023-1, McLean, VA, 59 pp.
- Endlich, R. M., 1964: The mesoscale structure of some regions of clear-air turbulence. *J. Appl. Meteor.*, **3**, 261–276.
- Grubišić, V., and B. J. Billings, 2007: The intense lee-wave rotor event of Sierra Rotors IOP 8. *J. Atmos. Sci.*, **64**, 4178–4201.
- Heidinger, A. K., C. O'Dell, R. Bennartz, and T. Greenwald, 2006: The successive-order-of-interaction radiative transfer model. *J. Appl. Meteor. Climatol.*, **45**, 1388–1402.
- Jiang, Q., and J. D. Doyle, 2004: Gravity wave breaking over the central Alps: Role of complex terrain. *J. Atmos. Sci.*, **61**, 2249–2266.
- Koch, S. E., and Coauthors, 2005: Turbulence and gravity waves within an upper-level front. *J. Atmos. Sci.*, **62**, 3885–3908.
- Lane, T. P., R. D. Sharman, T. L. Clark, and H.-M. Hsu, 2003: An investigation of turbulence generation mechanisms above deep convection. *J. Atmos. Sci.*, **60**, 1297–1321.
- Mellor, G. L., and T. Yamada, 1982: Development of a turbulence closure model for geophysical fluid problems. *Rev. Geophys. Space Phys.*, **20**, 851–875.
- Mlawer, E. J., S. J. Taubman, P. D. Brown, and M. J. Iacono, 1997: Radiative transfer for inhomogeneous atmospheres: RRTM, a validated correlated-k model for the longwave. *J. Geophys. Res.*, **102**, 16 663–16 682.
- Ólafsson, H., and P. Bougeault, 1996: Nonlinear flow past an elliptic mountain ridge. *J. Atmos. Sci.*, **53**, 2465–2489.
- Reiter, E. R., and H. P. Foltz, 1967: The prediction of clear air turbulence over mountainous terrain. *J. Appl. Meteor.*, **6**, 549–556.
- Schmit, T. J., M. M. Gunshor, W. P. Menzel, J. J. Gurka, J. Li, and S. Bachmeier, 2005: Introducing the next-generation Advanced Baseline Imager on GOES-R. *Bull. Amer. Meteor. Soc.*, **86**, 1079–1096.
- Skamarock, W. C., J. B. Klemp, J. Dudhia, D. O. Gill, D. M. Barker, W. Wang, and J. G. Powers, 2005: A description of the Advanced Research WRF Version 2. NCAR Tech. Note/TN-468+STR, 88 pp.
- Smith, R. B., and S. Gronas, 1993: The 3-D mountain airflow bifurcation and the onset of flow splitting. *Tellus*, **45A**, 28–43.
- Thompson, G., P. R. Field, R. M. Rasmussen, and W. D. Hall, 2008: Explicit forecasts of winter precipitation using an improved bulk microphysics scheme. Part II: Implementation of a new snow parameterization. *Mon. Wea. Rev.*, **136**, 5095–5115.
- Uhlenbrock, N. L., K. M. Bedka, W. F. Feltz, and S. A. Ackerman, 2007: Mountain wave signatures in MODIS 6.7- μ m imagery and their relation to pilot reports of turbulence. *Wea. Forecasting*, **22**, 662–670.