Topographically Generated Cloud Plumes

QINGFANG JIANG

University Corporation for Atmospheric Research, Monterey, California

JAMES D. DOYLE

Naval Research Laboratory, Monterey, California

(Manuscript received 22 April 2005, in final form 19 October 2005)

ABSTRACT

Two topographically generated cirrus plume events have been examined through satellite observations and real-data simulations. On 30 October 2002, an approximately 70-km-wide cirrus plume, revealed by a high-resolution Moderate Resolution Imaging Spectroradiometer (MODIS) image and a series of Geostationary Operational Environmental Satellite (GOES) images, originated from the Sierra Nevada ridge and extended northeastward for more than 400 km. On 5 December 2000, an approximately 400-km-wide cloud plume originated from the Southern Rocky Mountain massif and extended eastward for more than 500 km, the development of which was captured by a series of GOES images. The real-data simulations of the two cirrus plume events successfully capture the presence of these plumes and show reasonable agreement with the MODIS and GOES images in terms of the timing, location, orientation, length, and altitude of these cloud plumes. The synoptic and mesoscale aspects of the plume events, and the dynamics and microphysics relevant to the plume formation, have been discussed. Two common ingredients relevant to the cirrus plume formation have been identified, namely, a relatively deep moist layer aloft with high relative humidity and low temperature ($\approx -40^\circ C$ near the cloud top), and strong updrafts over high terrain and slow descent downstream in the upper troposphere associated with terrain-induced inertia–gravity waves. The rapid increase of the relative humidity associated with strong updrafts creates a high number concentration of small ice crystals through homogeneous nucleation. The overpopulated ice crystals decrease the relative humidity, which, in return, inhibits small crystals from growing into large crystals. The small crystals with slow terminal velocities ($<0.2 \text{ m s}^{-1}$) can be advected hundreds of kilometers before falling out of the moist layer.

1. Introduction

The Moderate Resolution Imaging Spectroradiometer (MODIS) is a key instrument aboard the polar-orbiting satellite Terra and provides high radiometric sensitivity (12 bit) in 36 spectral bands ranging in wavelength from 0.4 to 14.4 $\mu$m. Two bands are imaged at a nominal resolution of 250 m at nadir. Such unprecedented high-resolution images have broad applications in geosciences. Shown in Fig. 1 is a Terra MODIS 250-m resolution true color image of the West Coast of the United States valid at 1830 UTC 30 October 2002. The northwest–southeast-oriented high ridge is the Sierra Nevada mountain range, the average height of which is approximately 3 km ASL. A spectacular cloud plume originated from the Sierra massif and extended northeastward for more than 350 km with a mean width of approximately 70 km (hereafter, referred to as the “Sierra plume”). On first inspection it could easily be mistaken for a smoke plume, possibly emanating from a forest fire. Further analysis of the MODIS image and a series of Geostationary Operational Environmental Satellites (GOES) images indicates that the plume is clearly composed of cirrus clouds induced by the high terrain underneath. Although there are some thin cloud filaments upstream (to the west) of the Sierra Nevada massif, the western edge of the much denser clouds is located immediately over the high Sierra Nevada ridge.
suggestive of the importance of orographic forcing in the plume formation.

Cirrus clouds, which regularly cover about 35% of the earth’s surface, have important influence on the global climate through modulating the radiation budget (Ramanathan et al. 1983; Liou 1986). In addition to synoptic-scale motion and radiative cooling, high cirrus could also be generated by updrafts associated with mountain waves. In fact, long cloud plumes, trailing from major barriers such as the Cascade, Sierra, Rockies, Andes (Kahn et al. 2003), and Alps (Jiang et al. 2005), are often observed by satellites. So far, these terrain-generated plumes have received relatively little attention, and the mechanisms and favorable atmospheric conditions for long distance cirrus plume generation are still poorly understood. Based on Television Infrared Observation Satellite (TIROS) images, Conover (1964) classified orographically induced clouds into six categories. The cloud plumes examined in this study resemble his “fibrous plumes” class. However, regarding the mechanisms, his interpretations were limited and descriptive in character.

A number of observational and numerical studies of cirrus clouds have been conducted focusing on the microphysical characteristics and radiative properties of ice particles. For example, the in situ measurements by Heymsfield and Miloshevich (1995) show the existence of large numbers of small ice particles (<3 μm) in mountain wave–induced cirrus clouds. The crystal number concentrations and sizes are sensitive to the air temperature, relative humidity (RH), and vertical velocity (Heymsfield and Miloshevich 1993, 1995). When the air temperature is low (−36°C or lower), ice particles are primarily nucleated through the homogeneous nucleation process. The terminal velocities of the ice particles in cirrus clouds and their sublimation in ice sub-saturated environment have been the subject of several studies (e.g., Hall and Pruppacher 1976; Lin et al. 1998; Heymsfield and Iaquinta 2000). In terms of dynamics, while major mountain barriers such as the Rockies and Andes may extend their influence thousands of kilometers downstream through planetary waves associated with the earth’s rotation, the atmospheric responses to mesoscale terrain are often confined to the immediate vicinity of the terrain. There are some exceptional conditions under which mesoscale terrain may extend its influence far downstream even in the absence of the rotation of the earth as discussed by Smith et al. (1997). These exceptions include long island/mountain wakes induced by wave breaking–related dissipation (Pan and Smith 1999), and trapped waves. In a recent study by Jiang et al. (2003), the long and persistent banner clouds trailing from the Massif Centrale observed during two mistral events were connected to low-level wave breaking or a hydraulic jump over a single Alpine peak (i.e., Monte Lozère). In addition, low-level cumulus cloud plumes have been observed trailing from small islands such as Nauru (Nordeen et al. 2001). In contrast to the aforementioned studies, the cloud plume revealed by the MODIS image in Fig. 1 was located in the upper troposphere, likely induced by inertia–gravity waves excited by underlying terrain. Wave-induced vertical motions have been found to contribute to the polar stratospheric cloud formation as well (e.g., Dörnbrack and Leutbecher 2001).

Satellite images suggest that high terrain may play a crucial role in the initiation of these plumes, and motivate a number of interesting scientific questions regarding the dynamics, thermodynamics, and cloud physics.
governing the cirrus plumes. Specifically, the issues we seek to address include the following: 1) what are the key factors that contribute to cloud plume formation, 2) what determines the length of these plumes, 3) how sensitive are the plumes to the terrain height and slope, and 4) can these plumes be accurately simulated by mesoscale models?

The remainder of this paper is organized as follows. The development of two cloud plume events over the Sierra Nevada and Rocky Mountain ranges are described in section 2. The numerical setup is introduced in section 3. Large-scale aspects of the two events are examined in section 4 based on diagnosis of two mesoscale model simulations. In section 5, the control simulations are further diagnosed from a mesoscale perspective. Possible mechanisms for cloud plume generation and maintenance are discussed in section 6 using the linear wave theory and parcel theory. The results are summarized in section 7.

2. Cloud plumes seen from space
   a. The Sierra plume

A number of interesting cloud characteristics are evident in the Sierra plume case, as illustrated by the close-up MODIS image shown in Fig. 1b. Superimposed on the main plume, there are a series of five enhanced (bright) banners, which are oriented along the plume with a cross-plume width of the order of 10 km, likely corresponding to the high mountain peaks underneath (from north to south): Mt. Morgan (3190 m), Mt. Humphreys (4263 m), Mt. Goddard (4144 m), Mt. Pelisade (4341 m), and Mt. Whitney (4418 m). The bright banners, presumably corresponding to dense clouds, decay rapidly over approximately the first 50 km. The height of the cloud plume can be estimated from the well-defined cloud shadow on the surface using \[ H = \frac{d}{\tan(\alpha)} \]
where \( H \) is the plume height above the surface, \( \alpha \) is the sun angle, and \( d \) is the shadow length in the cross-plume direction. For 5 December, \( \alpha \) is 36° and \( d \) is approximately 10 km, which yields a cloud level of approximately 7 km above the surface, or 10 km ASL. The southern portion of the plume, located approximately over the Mt. Whitney, only extends 20–50 km downstream, likely due to the time lag as the plume advects southward. A thin cloud plume is located slightly to the north, trailing from the peaks of the White Mountains and eventually merges with the main plume.

The narrow but long-extended Sierra plume is captured by a series of GOES imagery as well. Figure 2a shows a GOES water vapor image valid at 1900 UTC 30 October 2002. Warm colors indicate the presence of cloud ice or high water vapor content in the upper troposphere. A long, moist plume originates from the California coast and extends northeastward over four states: California, Nevada, Utah, and Colorado. The moist plume is more than 1500 km long and approximately 150 km wide, and the angle between the moist plume and the latitude circle is approximately 20°. Embedded in the moist plume, a much smaller plume indicated by the orange color in Fig. 2a originates from the Sierra massif and extends across the border between the states Nevada and Utah. The inner plume coincides with the cirrus plume shown in the MODIS image (Fig. 1), indicating that the stronger reflection (i.e., orange color) is likely associated with the ice particles in the cirrus plume. The animation of the GOES water vapor imagery indicates that the moisture plume originates from a moist patch located off the coast of North America (Fig. 2a), associated with strong con-
vection over the Pacific Ocean. The moisture plume forms around 1200 UTC approximately oriented along the 40°N latitude and is much wider but shorter than that observed at 1900 UTC. It moves slowly southward while it is being stretched. The inner plume first becomes distinguishable around 1630 UTC downstream of the White Mountains (see Fig. 1 for location). As the moisture plume moves slowly southward, a more spectacular cloud plume appears downstream of the main Sierra Nevada ridge and extends farther downstream with time. At approximately 2200 UTC, the cloud plume reaches its maximum length, approximately 420 km, and extends eastward over Utah. Subsequently, the plume detaches from the Sierra Nevada ridge and then advects eastward over the Intermountain West and eventually dissipates. During the course of the plume development, as it drifts southward and extends farther downstream, the upwind edge of the cirrus plume always coincides with the main Sierra Nevada ridge prior to 2200 UTC, indicative of the importance of topographic forcing on the plume formation.

Two GOES visible snapshots are shown in Fig. 3. At 1800 UTC, the cloud plumes are evident over both the White Mountains and the main Sierra Nevada peaks. The plume located downstream of the main Sierra Nevada ridge is much brighter, and the plume trailing from the White Mountains is much longer. At 1900 UTC, the cloud plume from the main Sierra Nevada ridge becomes significantly longer than observed 1 h earlier. The southward-drifting speed and growth rate of the plume length can be estimated by comparing the two images, which are approximately 5 and 35 m s⁻¹, respectively.

b. The Southern Rockies plume

Shown in Fig. 2b is a GOES water vapor image valid at 1902 UTC 5 December 2000. Two distinct cloud patches are located immediately to the east of the Southern Rockies, separated by an approximately 100-km-wide cloud-free zone. We are particularly interested in the spectacular southern cloud plume, the left edge of which is parallel to the main massif of the Sangre de Cristo Range. The plume is approximately 500 km long and 400 km wide, and spreads over five states: Colorado, Kansas, New Mexico, Texas, and Oklahoma. Along the plume, there are two reflectivity maxima; one is located immediately downstream of the Sangre de Cristo ridge, and the other is located approximately 250 km farther downstream. The animation of the half-hour interval GOES water vapor imagery indicates that the development of the cloud plume is associated with the passing of a moist air mass advected northeastward from the state of Arizona. The cloud plume first appears in the lee of the Southern Rocky Mountain ridge around 1330 UTC, and grows rapidly in both length and width. The plume reaches its maximum in terms of the cloud cover area around 1900 UTC, and dissipates gradually afterward. The left edge of the plume coincides with the Sangre de Cristo Massif, and is fairly stationary over the whole period. The growth rate or advection speed of the plume is approximately 40 m s⁻¹ as estimated from the GOES water vapor images.
According to the GOES visible images (Fig. 4), the plume is approximately 300 km long at 1645 UTC. A well-defined foehn window is located upstream of the cloud plume, indicating strong flow descent and implying the importance of the Southern Rocky Mountain ridge in the cloud plume formation. At 1902 UTC (Fig. 4b), the plume is longer, and the small-scale variation of the brightness along the Rocky Mountain ridge is evident, likely corresponding to the variation of the ridge height. Although the two cloud plume events occur over different barriers, they share some remarkable similarities as summarized below.

- The plumes originate from high ridges, with their upwind edges coincident with high peaks throughout the two events, strongly indicating that these plumes are topographically generated.
- Both plumes are located in the upper troposphere and oriented approximately along the wind direction.
- Both plumes show multiple scale features, namely, hundreds of kilometers along the plume, and fine-scale variations along the ridges.

3. Numerical setup

The atmospheric component of the Navy’s Coupled Ocean–Atmospheric Mesoscale Prediction System (COAMPS\(^1\)) is used to simulate the two topographically generated cirrus plume events. COAMPS is a mesoscale model that makes use of finite-difference approximations to represent the fully compressible, non-hydrostatic equations that govern atmospheric motions (Hodur 1997). The physical parameterizations include 1.5-order turbulence kinetic energy (Mellor and Yamada 1974), surface flux parameterization (Louis 1979), and ice cloud physics (Rutledge and Hobbs 1983, 1984). The cloud physics scheme includes six water species: water vapor, cloud water, rainwater, cloud ice, snow, and graupel. The processes that are most relevant to the cirrus formation are ice initiation, depositional growth of cloud ice at the cost of water vapor, ice crystal fallout, and sublimation in an ice subsaturated environment. For temperatures less than 269.16 K, the initiation of cloud ice follows Fletcher (1962),

\[
n_i = n_o \exp(\beta(T_o - T)),
\]

where \(n_i\) is the number concentration of the ice nuclei, \(T\) is the temperature, \(T_o = 273.16 \, \text{K}\), \(\beta = 0.6 \, \text{K}^{-1}\), and \(n_o = 10^{-7} \, \text{m}^{-3}\). Following Stephens (1979), the rate of ice crystal formation is given by \(M_i n_i/\Delta t\), where \(\Delta t\) is the time step, and \(M_o = 10^{-12} \, \text{kg}\) is the initial mass of cloud ice crystals.

The growth of an ice crystal through vapor deposition is given by

\[
\frac{dM_i}{dt} = \frac{4D_i(S_i - 1)n_i}{A^* + B^*},
\]

where \(D_i = 16.3M_i^{0.2}\) is the crystal diameter, \(M_i\) is the average ice crystal mass, \(S_i\) is the saturation ratio relative to ice, and \(A^*\) and \(B^*\) are given by Pruppacher and Klett (1978).

The fall speed of a crystal, \(V_f\), is given by

\[
V_f = k_r R_i (P_o/P)^{0.5},
\]

where \(k_r = 304\) is a constant, \(P_o = 1000 \, \text{hPa}\), and \(P\) is the pressure.

The computational domains for the two simulations are configured with three nested grids with correspond-
ing horizontal spatial resolutions of 27, 9, and 3 km, respectively. The topographic data are taken from a 1-km resolution terrain dataset. The numbers of the horizontal grid points are (from coarse to fine grids) $181 \times 181$, $151 \times 121$, and $181 \times 151$ for the Sierra plume simulation, and $151 \times 121$, $151 \times 121$, and $181 \times 151$ for the Southern Rocky Mountain plume simulation. Because of the large physical dimensions of these plumes, the innermost domain only includes the ridges at the upwind edges of the plumes and some portion of the plumes. There are 55 vertical levels and the terrain-following coordinate is stretched with $\Delta z_{\min} = 20$ m near the surface. The model top is at 31 km with Rayleigh damping applied to the upper 11 km. The initial and boundary conditions are specified using the Naval Operational Global Atmospheric Prediction System (NOGAPS) analysis. An incremental update data assimilation procedure that enables mesoscale phenomena to be retained in the analysis increment fields is used for model initialization. The model was initialized at 0000 UTC 30 October 2002, and 0000 UTC 5 December 2000, respectively.

4. Synoptic-scale perspective

a. Sierra plume

Figure 5 shows the horizontal wind speed, potential temperature, and RH profiles derived from the 1200 UTC 30 October 2002 Oakland, California (37.75°N, 237.78°E), sounding. Based on the launch time and the GOES images (i.e., Fig. 2a), the sonde likely ascended through the moist plume shown in the water vapor images. The wind direction is northwesterly below 7 km and approximately westerly above. The tropopause is located approximately at 12.5 km, below which the troposphere shows a two-layer structure in terms of relative humidity and stability. The lower layer, from the surface to 8 km ASL, is drier and more stable with an average buoyancy frequency of 0.0126 s$^{-1}$. In the lower layer, the wind speed almost linearly increases from 10 m s$^{-1}$ at the 1-km level to approximately 20 m s$^{-1}$ at the 8-km level. A sharp increase of RH and decrease of stability occurs at 8 km ASL. The humidity relative to ice is close to 100% and the average (dry) buoyancy frequency is 0.009 s$^{-1}$ between 8 km and the tropopause (~12.5 km). The altitude of the moist layer is consistent with the cloud-top level estimated from the cloud shadow. The air temperature in the moist layer decreases from $-31^\circ$C at the bottom (i.e., 8 km) to $-64^\circ$C near the top, implying that the cloud plume is primarily composed of ice particles.

Shown in Fig. 6 are the geopotential height contours, wind vectors, and RH with respect to ice (in grayscale) at 300 hPa derived from the COAMPS coarse mesh (i.e., $\Delta X = 27$ km) valid at 1800 UTC. A pressure trough is located off the British Columbia coast, and a cutoff low is centered around the state of Montana. Between the two intense low pressure systems, there are a pair of pressure ridges: one is located over British Columbia and the other is located off the west coast of the United States. A slowly moving cold front is located between 35° and 40°N with a maximum wind speed of 30 m s$^{-1}$ at the 300-hPa level. The COAMPS-simulated RH field matches the GOES water vapor image reasonably well (Fig. 2a). For example, both show the moist region located off the North America coastline, and dry region located over the Southern United States. The COAMPS also captures the presence of the moisture plume, which is in reasonable agreement with the GOES water vapor images in terms of the timing, orientation, and length of the plume. Animation of the COAMPS plots (same as in Fig. 6 but for different time) indicates that the moist air is entrained into the deformation zone from the moist patch off the coast. It is stretched and extends farther inland from 1500 to 2000 UTC while it moves slowly southward, which is consistent with the series of GOES images. The wind
direction in the deformation zone is westerly with wind speed between 20 and 30 m s\(^{-1}\). It is remarkable that the COAMPS model forecasts the long, narrow moist plume correctly, considering that the plume originates from the data-sparsed Pacific Ocean. It should be noted that the inner ice cloud plume trailing from the Sierra, as shown in the GOES water vapor images, is absent from the COAMPS coarse mesh.

### b. Southern Rockies plume

Vertical profiles derived from a sounding launched at 1200 UTC 5 December 2000 from Albuquerque, New Mexico (35.05°N, 253.39°E), located upstream of the Southern Rockies plume, are shown in Fig. 7. Similar to the Sierra plume event, the troposphere shows a two-layer structure in terms of the RH and stability. The air below 7 km ASL is much drier and more stable with an average buoyancy frequency of 0.0127 s\(^{-1}\). The tropopause is located at 9.5 km, much lower than observed during the Sierra plume event. A sharp increase of the RH occurs around 7 km ASL, above which the air is more moist and less stable. In fact, the high relative humidity layer extends well into the lower stratosphere with the RH relative to ice close to 100%. Above 2 km ASL, the wind is almost strictly westerly with little directional shear and the wind speed increases from 11 m s\(^{-1}\) at 3 km to 20 m s\(^{-1}\) at 7 km. In the moist layer, the average wind speed is approximately 20 m s\(^{-1}\). The air temperatures at 7 and 9.5 km are \(-25^\circ\) and \(-43^\circ\)C, respectively.

Shown in Fig. 8 are the geopotential height contours, relative humidity with respect to ice (grayscale), and wind vectors at 300 hPa derived from the COAMPS 27-km grid mesh valid at 1800 UTC 5 December 2000. The jet stream is located over the central United States, between a deep low pressure system over southern Canada and a pressure ridge over Mexico. Associated with the pressure ridge and a closed low located off the coast of California, a subtropical jet transports warm and moist air from the Pacific Ocean toward the southern Rocky Mountains. The wind direction over the states of New Mexico and Arizona is southwesterly and turns into westerly in the vicinity of the Southern Rocky Mountain ridge. A second moist air mass over the North America is located farther north associated with a shortwave embedded in the Canadian trough. A dry air zone is located in the deformation zone, which separates the two moist air masses. The cloud patches...
in Figs. 2b and 4 and the moist air masses in Fig. 8 match each other reasonably well.

5. Mesoscale perspective

a. Cross-sectional analysis

Shown in Fig. 9 are the vertically integrated cloud ice \[ \text{VIC}(x, y) = \int \rho_a q_i \, dz \] and horizontal wind vectors at 9 km ASL derived from the innermost mesh valid at 1900 UTC 30 October 2002. Areas with model terrain above 3 km are hatched. A well-defined plume trailing from the main Sierra ridge is evident. The plume is more than 250 km long and 70 km wide, and there is a small angle (\( \sim 20^\circ \)) between the plume orientation and the wind direction at 300 hPa (i.e., westerly). The estimated plume growth and southward-drifting speeds are 25 and 6 m s\(^{-1}\), respectively, which are approximately in agreement with the estimation from the series of GOES images. The cloud ice mixing ratio, ice liquid water equivalent potential temperature, wind speed, and relative humidity (relative to ice) in a vertical cross section oriented along the plume (see Fig. 9 for location) are shown in Fig. 10a. The ice liquid water equivalent potential temperature is defined as (Tripoli and Cotton 1981)

\[
\theta_l = \theta - \left[ 1 - \frac{L_{lw} q_l}{C_p \max(T, 253)} - \frac{L_{li} q_i}{C_p \max(T, 253)} \right],
\]

where \( C_p \) is the air specific heat at a constant pressure, \( L_{lw} = L_{li} + L_f \), \( L_{li} \) is the latent heat of fusion, and \( q_l \) and \( q_i \) are mixing ratios of liquid water and cloud ice, respectively. Along this section, the main ridge width is of the order of 100–150 km and the lee slope is fairly steep. A cloud ice plume is located between 8 and 11 km, which is consistent with the cloud level estimated from the cloud shadow. While the cloud ice is present both upstream and downstream of the ridge, the terrain enhancement effect is still evident. The equivalent potential temperature field indicates the presence of windward blocking as indicated by the interception of isentropes with terrain. The RH field indicates that a moist layer is located between 8.5 and 11 km, where the air is ice saturated or nearly saturated even on the upstream side of the Sierra Nevada ridge. An inertia–gravity wave (IGW) is launched from the highest peak and penetrates the moist layer in the upper troposphere. Near the cloud plume level, the IGW is characterized by a sudden updraft over the peak and a slow descent in the tail. The ice saturation is a sufficient condition for crystal formation in a cold environment in the model (Fletcher 1962), which is a necessary but not a sufficient condition in nature according to observational studies (e.g., Heymsfield and Miloshevich 1995). This issue will be further discussed in the coming sections. Nevertheless, the relative humidity seems to be increased in the lee associated with IGW-induced updrafts.

In Fig. 11, a widespread cloud ice plume originating from the Sangre de Cristo ridge is evident. The plume extends more than 300 km westward and is approximately oriented along the wind direction. The COAMPS-forecasted cloud ice field agrees with the GOES imagery very well in terms of the location, orientation, and size of the cloud plume. Figure 12 shows the vertical cross section of the equivalent potential temperature, cloud ice, and wind speed, oriented along the plume. Areas with the relative humidity relative to ice greater than 90% are hatched. The highest peak of the Sangre de Cristo ridge is about 3 km ASL. Again, there is a near-saturated layer located approximately between 8 and 12 km. As airflow passes the ridge, IGWs are launched, the amplitude of which decays with the increasing altitude over the ridge. To the right (i.e., downstream side) of the ridge, the cloud ice mixing ratio is significantly larger, indicative of the crucial role of the Southern Rocky Mountains in the plume formation.
b. Trajectory analysis

Figures 13 and 14 show some results from trajectory analysis using the 3-km grid data. The trajectory of a Lagrangian parcel is computed using $x_{i+1} = x_i + v(x_i, t_i)\Delta t$, where $x = (x, y, z)$ is the position vector, $v = (u, v, w, i)$ is the four-dimensional wind field linearly interpolated from half-hour interval COAMPS 3-km grid data, and $\Delta t$ is the time interval.

For the Sierra plume event, parcels are launched at 1900 UTC from 36.55°N, 243.5°E, which is located within the cloud plume and approximately 150 km downstream of the main Sierra ridge. The initial parcel altitudes are 6, 7, 8, 9, and 10 km, respectively, with $\Delta t = -0.25$ min; a negative time interval corresponds to a reverse trajectory. The parcel altitudes, relative humidity (with respect to ice), and cloud ice mixing ratio along the trajectories are plotted versus horizontal distance. The terrain higher than 2 km ASL and underneath the 6-km parcel is plotted for reference. Clearly, for all the parcels, strong ascent ($\Delta t$ = 600–1200 m) occurs over the lee slope of the main Sierra Nevada ridge and the parcel relative humidity increases accordingly. The parcels initially from the 7-, 8-, and 10-km levels are drier (RH < 65%) upstream of the Sierra Nevada ridge and become ice saturated over the highest peak. The parcel originating upstream from the 9-km level has greater RH and becomes saturated over the windward slope associated with weak upslope ascent. The parcel experiences a sharp descent over the highest peak where the RH decreases to 80%, and becomes saturated subsequently over the lee slope associated with a sudden ascent.

Theoretically, in stratified flow, a parcel that tempo-
rarily departs vertically from its equilibrium level tends to return to its level of origin because of buoyancy force. However, the parcels launched from the 7–10-km levels stay aloft over more than 150 km downstream, and, correspondingly, remain saturated or near saturated. The parcel launched from the 6-km level (i.e., below the cloud plume) is not saturated and experiences a net descent downstream. The topographic control of the cloud ice formation is evident in Fig. 13c. For parcels initially launched from 7 to 10 km (i.e., within the cloud plume), the highest peak of the Sierra ridge separates the no-ice upstream state to ice-contained state downstream. The ice mixing ratio of the parcel from 8 km indicates an oscillation with a horizontal wavelength of approximately 50 km.

A similar set of trajectories computed from the Southern Rockies plume simulation are shown in Fig. 14. The parcels are launched at 1900 UTC from 35.5°N, 258°E, which is located within the cloud plume and approximately 300 km downstream of the Rockies, and traced back to upstream using the backward trajectory calculation method with $\Delta t = -0.25$ min. Clearly, the parcels launched from 7, 8, and 10 km experience ascent over the lee slope of the Southern Rocky Mountain ridge and stay aloft for more than 300 km downstream. The parcels from the 8–10-km layer are ice saturated downstream of the Rocky Mountain ridge where cloud ice appears accordingly. In addition to the net ascent, some parcels oscillate in the vertical as they drift downstream, with a horizontal wavelength of approximately 100 km. Accordingly, the cloud ice mixing ratios exhibit similar oscillation for parcels from the 7–10-km layer.
In summary, the trajectory analysis clearly indicates the topographic control of the cloud ice formation during the two plume events examined. Cloud ice is generated in the lee of the major ridges associated with a sudden ascent during which parcels become ice saturated. Farther downstream, parcels may stay aloft with a net positive vertical displacement or oscillate in the vertical, likely associated with inertia–gravity waves as will be discussed in the next section.

c. Microphysical aspects

The domain-averaged crystal number concentrations, crystal diameters, and terminal velocities for the two control simulations are listed in Table 1 and the corresponding frequency distributions of the grid-mean crystal diameters and terminal velocities are shown in Fig. 15. The domain-averaged quantities are derived from the COAMPS simulations and averaged over grid points with $q_i > 5 \times 10^{-6} \text{ kg kg}^{-1}$, approximately corresponding to crystals with a grid-mean diameter of less than 10 $\mu$m. It is interesting that the domain-averaged crystal properties and the frequency distributions of grid-mean diameters and vertical velocities derived from the two control simulations are almost identical. For approximately 80% of the grid points, the grid-mean crystal diameters are between 50 and 150 $\mu$m and the terminal velocities are between 2 and 10 cm s$^{-1}$. As expected for high cirrus clouds, the crystal sizes are relatively small for both plumes and correspondingly the vertical velocities are small as well, which is consistent with the marked length of the two plumes. Despite the simplicity of the ice-phase physics parameterization used in COAMPS, the simulated crystal number concentrations, sizes, and vertical velocities are comparable to those aircraft measurements of cirrus clouds at a similar altitude (e.g., Heymsfield 1975; Heymsfield and Iaquinta 2000).

To examine the sensitivity of the cirrus plumes to microphysical parameterization, additional simulations of the Southern Rockies plume event have been carried out with the numerical setup identical to the control run except for the coefficient $n_o$ in Eq. (1) and $k_t$ in Eq. (3). Results from two pairs of such simulations with $k_t = 3040$ and 30.4, and $n_o = 1$ and $10^{-4}$ m$^{-3}$, respectively, are included in Table 1 and Fig. 16. Because the crystal number concentration is only a function of the air temperature, the domain-averaged number concentration is almost independent of $k_t$ (Table 1). Corresponding to the increase of $k_t$ by a factor of 10, the domain-averaged terminal velocity is increased approximately by a factor of 8, and the mixing ratio of cloud ice and diameter are significantly reduced, likely due to the fallout of large crystals. Compared to the control run, the cloud cover area as indicated by the vertically integrated ice shrinks considerably (Fig. 16a). As $k_t$ is decreased to one-tenth of its original value, the domain-averaged mixing ratio and crystal diameter increase slightly and the terminal velocity decreases to approximately one-tenth of that of
the control run. As expected, associated with much slower falling speed, the cloud cover area is larger, and the decay of denser clouds is much slower (Fig. 16b). For a smaller $n_0$, the crystal number concentration is proportionally smaller, and the domain-averaged crystal diameter and terminal velocity increase approximately as $n_0^{-1/2}$. Correspondingly, the plume is smaller than that in the control run (Fig. 16c). For a larger $n_0$, the crystal diameter and terminal velocity decrease approximately as $n_0^{-1/2}$ due to the increase of crystal number concentration. Associated with slower fallout of crystals, the simulated plume is larger and more intense (Fig. 16d) similar to the simulation with $k_t = 30.4$.

6. Discussion

Clearly, both dynamical and microphysical processes are involved in the formation of the cloud plumes aloft. Ascent motion at cloud levels, likely associated with perturbations introduced by mesoscale terrain and propagated into the upper troposphere through IGWs, is necessary for creating an ice-saturation environment in the upper troposphere. In addition to dynamics, the ice-phase microphysics plays an important role in the plume formation and determination of the plume length. In this section, the IGW effect is examined using a linear wave model, and the relevant microphysical processes are discussed based on published results and using a Lagrangian parcel argument.

a. Inertia–gravity wave dynamics

For stratified unidirectional flow with a constant wind speed $U$, the linear steady IGW equation can be written as (Queney 1948)

$$
(\partial_x^2 + \partial_y^2 + \partial_z^2)w_{xx} + f^2 U^2 w_{zz} + \frac{N^2}{U^2} (\partial_x^2 + \partial_y^2)w = 0,
$$

(5)

where $w$ is the vertical velocity and $f$ is the Coriolis parameter. Equation (5) has been solved for stratified airflow past an idealized ridge using the Fast Fourier Transformation (FFT) method (Smith 1980). The computational domain is two-dimensional including 1024 points in the horizontal and the resolution is 1 km with periodic boundary conditions applied along the lateral boundaries. At the surface ($i.e., z = 0$), we have $w = Uh$, where $h(x) = h_m \exp(-x^2/a^2)$ is a Gaussian ridge defined by a ridge height $h_m$ and a half-ridge width $a$. The atmosphere is separated into three discrete layers.
with a constant buoyancy frequency $N$ and ambient wind speed $U$ in each layer. The lowest two layers are used to represent the troposphere and the third stable layer is assumed to be infinitely deep with a radiation condition applied at the top. The layer depths and other parameters, derived from the two soundings shown in Figs. 5 and 7 are listed in Table 2 for reference. The ambient flow is in geostrophic balance (see Smith et al. 2002 for more details about the three-layer linear wave model). To illustrate the dynamical role of IGWs in the cloud plume formation, four linear IGW solutions associated with stratified flow past an idealized ridge are shown in Fig. 17. Figure 17a shows a solution with $f = 10^{-4}$ s$^{-1}$ and an asymmetric ridge with $h_m = 1000$ m, and $a = 60$ km in the windward side and 15 km in the lee side. The asymmetric ridge is used to mimic the Sierra ridge cross section shown in Fig. 10. Between 7 and 10 km, the vertical displacement in the lee of the ridge peak indicates a sharp ascent and then a slow descent downstream. The length of the positive vertical displacement area (PVDA; shaded) is approximately 120 km. Clearly, the level to have relatively large PVDA is located at $1.5 \lambda_c$, where $\lambda_c = 2\pi U/N$ is the vertical wavelength of hydrostatic waves.

The presence of the PVDA in the lee of the ridge is consistent with the real-data COAMPS simulations. Especially, the sharp-ascent and slow-descent pattern might be critical for the formation of the long cloud plumes. The second wave crest is located approximately 400 km downstream of the ridge with its amplitude reduced to one-fifth of the first wave. In the absence of rotation, the dominant waves are hydrostatic and characterized by a slow descent and sharp ascent pattern between 7 and 10 km ASL in the vicinity of the ridge and the positive vertical displacement area in the rotational solution is absent (Fig. 17b). The positive vertical displacement area is still present in the solution with a less steep lee slope (i.e., $a = 60$ km on both upstream and lee sides, Fig. 17c) and the leading edge of the PVDA is gentler accordingly, indicative of weaker ascending motion. Figure 17d shows a solution with the Albuquerque sounding and a narrower ridge (i.e., $a = 30$ km). Again, the solution indicates the presence of a PVDA with a sharp ascent and subsequent slow descent. The along-wind dimension of the PVDA is approximately 200 km.

There are two velocities relevant to the development of the plumes: the advection speed $U$ and the horizontal component of the IGW group velocity $C_{gx}$. In two-dimensional (i.e., $x$-$z$) hydrostatic limit, the dispersion relation of IGWs can be written as

$$\omega = \frac{Uk \pm \sqrt{k^2 N^2 + m^2 f^2}}{m},$$

where $m, k$ are vertical and horizontal wavenumbers, and the $\pm$ signs correspond to downstream- and upstream-propagating waves, respectively. We are only interested in waves that are stationary relative to the terrain (i.e., the horizontal wave phase speed $C_{phx} = \omega/k$ is zero), which gives

![Fig. 15. Frequency distributions of (a) crystal diameter and (b) crystal terminal velocity derived from the two control runs. The frequency $f_i (f_j)$ is defined as the ratio of the number of grid points in each diameter (terminal velocity) bin to the total number of grid points with $q_i > 5 \times 10^{-3}$ g kg$^{-1}$. The bin sizes are 10 $\mu$m for the diameter and 1 cm s$^{-1}$ for the terminal velocity separately.](image)
Using (6), (7), and $C_{gx} = \frac{\partial u}{\partial k}$, we obtain

$$m^2 = \frac{k^2 N^2}{U^2 k^2 - f^2}. \quad (7)$$

where $R_k = \frac{Uk}{f}$ is the Rossby number based on the horizontal IGW wavenumber. If we define the Rossby number as $R_L = \frac{U}{(fL)}$, where $L$ is the wavelength, the nondimensional horizontal component of the group velocity is $(4\pi^2 R_L^2)^{-1} = 0.025 R_L^2$. As an example, if we choose $U = 20 \text{ m s}^{-1}$, $L = 350 \text{ km}$ (estimated from Fig. 14a), and $f = 10^{-4} \text{ s}^{-1}$, it follows that $R_L = 0.6$ and $C_{gx} = 1.4 \text{ m s}^{-1}$. Clearly, the observed rates at which the cirrus plumes extend downstream are comparable to the cloud-level advective speeds, and much faster than the estimated IGW horizontal group velocity.

<table>
<thead>
<tr>
<th>Location</th>
<th>Depth (km)</th>
<th>Layer 1</th>
<th>Layer 2</th>
<th>Layer 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oakland</td>
<td>8</td>
<td>3.5</td>
<td>$\infty$</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$u (\text{m s}^{-1})$</td>
<td>15</td>
<td>25</td>
<td>28</td>
</tr>
<tr>
<td></td>
<td>$N (\text{s}^{-1})$</td>
<td>0.012</td>
<td>0.008</td>
<td>0.018</td>
</tr>
<tr>
<td>Albuquerque</td>
<td>7</td>
<td>2.5</td>
<td>$\infty$</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$u (\text{m s}^{-1})$</td>
<td>15</td>
<td>20</td>
<td>22</td>
</tr>
<tr>
<td></td>
<td>$N (\text{s}^{-1})$</td>
<td>0.012</td>
<td>0.008</td>
<td>0.018</td>
</tr>
</tbody>
</table>
Instead of examining the microphysical process in detail, we try to put this study in the proper perspective by comparing with relevant published results. Previous studies indicated that the nucleation processes in the cold environment are sensitive to both air temperature and relative humidity (Heymsfield and Miloshevich 1995). In the upper troposphere, heterogeneous nucleation was found relatively inefficient (Hobbs and Rango 1985; Heymsfield et al. 1991). For example, when the air temperature is lower than \( -36^\circ C \), liquid droplets were absent in aircraft in situ measurements, suggestive of the domination of homogeneous ice nucleation in high cirrus clouds (Sassen and Dodd 1988; Heymsfield and Sabin 1989). According to the aircraft
measurements made in cirrus during the First International Satellite Cloud Climatology Project (ISCCP) Research Experiment (FIRE) II, the lower bound on the relative humidity required for cirrus formation \(RH_c \approx -RH_{hn} - 10\) (Heymsfield and Miloshevich 1995). Here \(RH_{hn}\) is the peak relative humidity with respect to water in wave clouds, which decreases from 100% for temperature above \(-39^\circ C\) to 73% at \(-56^\circ C\) approximately following

\[
RH_{hn} = 188.92 + 2.81T + 0.013336T^2,
\]

where \(T\) is the temperature in degrees Celsius. Here we consider an air parcel that ascends adiabatically. The minimum vertical displacements \(\delta_c\) and \(\delta_{hn}\) required for the parcel to reach \(RH_c(T)\) and \(RH_{hn}(T)\) can be obtained using (9). The derived \(\delta_c\) and \(\delta_{hn}\) and the corresponding RH with respect to water \(RH_w\) and ice \(RH_i\) using the Oakland and Albuquerque soundings are shown as a function of the initial parcel altitude (Fig. 18). Clearly, for the Oakland sounding, the minimum vertical displacement required to form cirrus (i.e., the \(\delta_c\)) is more than 1 km for air parcels initially located below 8 km ASL, and is much smaller (i.e., \sim 300 m) in the moist layer. Especially near the tropopause, \(\delta_c\) is close to zero, which is consistent with the presence of thin cirrus clouds upstream of the Sierra ridge (Fig. 1). The corresponding saturation ratio with respect to ice is between 110% and 130%. The minimum vertical displacement \(\delta_c\) computed using the Albuquerque sounding is of the order of 700 at the bottom of the moist layer (i.e., \sim 7 km) and linearly decreases to zero near the tropopause (i.e., \sim 9.5 km). The corresponding relative humidity is in the range of 120%–140% in the moist layer.

Karcher and Strom (2003) found that vertical air motion is a key factor in determining ice crystal concentrations and size distribution in cirrus clouds. Associated with strong updrafts (i.e., \(w > 0.1\) m s\(^{-1}\)), most young cirrus clouds typically contain a high number density of small (diameter <20 \(\mu m\)) ice crystals, and likewise, with slower updrafts, cirrus clouds more likely contain low concentrations of larger crystals. The presence of large numbers of small ice crystals in mountain wave–induced cirrus clouds associated with rapid cooling have been reported by other observational and numerical studies as well (e.g., Lin et al. 1998). It was demonstrated by Heymsfield and Miloshevich (1993) that corresponding to the increase of vertical motion from 0.25 to 2 m s\(^{-1}\), the ice crystal number concentrations approximately increase by 100 times. Using the vertical displacement in Fig. 18 and the 2D linear steady relation \(w_c = \frac{U \partial \delta_c}{\partial x} = 2\pi U \delta_c/L\), where \(L\) is the horizontal wavelength corresponding to the vertical motion, we can estimate the minimum updraft (i.e., \(w_c\)) for homogeneous nucleation and cirrus formation. When \(U = 25\) m s\(^{-1}\), \(L = 60\) km, and \(\delta_c = 300\) m (i.e., approximately corresponding to the Sierra plume event), \(w_c\) is approximately 0.8 m s\(^{-1}\). Similarly, using \(U = 30\)
m s\(^{-1}\), \(L = 30 \text{ km}\), and \(\delta_c = 300 \text{ m}\) (relevant to the Southern Rockies plume), we obtain \(w_c \sim 2 \text{ m s}^{-1}\). The simulated updraft maxima over the high peaks and at cloud levels are larger than the estimated \(w_c\) values for the two plume events. According to linear mountain wave theory, for given wind and stratification profiles, the mountain wave amplitude (i.e., vertical velocity \(w\) or vertical displacement \(\delta\)) is proportional to the terrain height and the threshold values of \(\delta_c\) and \(w_c\) estimated here correspond to certain critical mountain heights to generate these perturbations, implying that these cloud plumes only form over relatively high ridges.

Apparently, the terminal velocities of cirrus crystals are of primary importance in determining the decay of the cirrus plumes. Once falling out of clouds, ice particles are subjected to sublimation in an ice-subsaturated environment. While laboratory and observational studies have highlighted the dependence of crystal terminal velocity on the crystal shapes, in general, the terminal velocity is approximately proportional to the crystal diameter (e.g., Fig. 3 of Heymsfield and Iaquinta 2000). As revealed by the MODIS visible image, the length of denser clouds, likely associated with larger crystals, is \(\sim 50 \text{ km}\) and the length of the whole cloud plume is \(\sim 400 \text{ km}\). Using \(U = 25 \text{ m s}^{-1}\), we obtain two time scales: 2000 and 16 000 s. Assuming that the cloud depth is 2 km, the average fallout velocity for large crystals is approximately 1 m s\(^{-1}\), and for small crystals, the average fallout velocity is approximately 0.12 m s\(^{-1}\), corresponding to crystal sizes of 200–400 \(\mu\text{m}\) and \(<50 \mu\text{m}\), respectively. The survival of crystals falling into subsaturated environment has been studied by Hall and Pruppacher (1976). They found that the survival time of crystals in an ice-subsaturated environment is very sensitive to the relative humidity; the higher the RH, the longer the crystal can survive. In addition to fallout, crystals in cirrus plumes could also sublimate in an ice-subsaturated environment created by wave-induced descent. The relatively deep moist layer (\(\sim 2 \text{ km}\)) observed during the two plume events and the slow descent associated with IGWs apparently help maintain the high RH environment around small crystals and likely contribute to the maintenance of the long plume tails.

c. Thermodynamics and parcel argument

While the presence of the long PVDA in the IGW solutions is likely relevant to the cloud plume formation, the application of linear wave theory to the cirrus problem could be complicated by diabatic cooling or warming associated with ice physics. It has been shown in previous studies that dry wave theories are still applicable to a moist atmosphere involving diabatic processes if the dry buoyancy frequency \(N\) is replaced with a moist buoyancy frequency \(N_m\) (Lalas and Einaudi 1974). In the presence of cloud ice, we can define a moist \(N_m\) as \(N_m = gd(ln \theta)/dz\). This approach may become problematic due to the fallout of ice crystals or microphysical hysteresis. The term hysteresis here is used loosely to refer to the time scale difference in ice crystal growth and sublimation (i.e., sublimation is slower), or the relative humidity difference required for crystal formation (\(\text{RH}_c > 110\%\) in the upper troposphere according to Fig. 18) and for sublimation (\(\text{RH}_s < 100\%\)). The use of moist \(N\) is also inappropriate as the microphysical-related time scale is comparable or larger than \(N^{-1}\).

Following a Lagrangian parcel, the increase of parcel potential temperature due to the growth of ice crystals at the cost of water vapor is

\[
\frac{d\theta}{\theta} = -\frac{L_{\text{vap}}dq_{\text{i}}}{C_p T},
\]

where \(\theta\) is the potential temperature, \(T\) is the parcel temperature. For \(dq_{\text{i}} = 0.15 \text{g kg}^{-1}\) (estimated from Figs. 12 and 13), \(\theta = 320 \text{ K}\), the corresponding potential temperature increase is \(\sim 1 \text{ K}\), which is fairly significant considering the relatively weak stability in the moist layer. For example, if the potential temperature gradient in the ambient flow is \(3 \text{ K km}^{-1}\) (i.e., \(N \sim 0.01 \text{ s}^{-1}\)), a 1-K increase in potential temperature allows the parcel to reach a new equilibrium level, which is approximately 300 m above its original level.

To further illustrate the effect of ice cloud hysteresis...
on vertical motion, we consider a moist Lagrangian parcel, the vertical motion of which is governed by
\[
\frac{ Dw }{ Dt } = - \frac{ \theta_p - \theta_a }{ \theta } g, \tag{11}
\]
where \( \theta_a \) and \( \theta_p \) are the ambient and parcel potential temperatures, respectively, and \( w \) is the vertical velocity. Initially, the air parcel is in equilibrium (i.e., \( \theta_p = \theta_a = \theta_0 \)), and the vertical velocity is \( w_v > 0 \), corresponding to mountain-induced ascent. A few possible solutions to Eq. (11) are schematically shown in Fig. 19 and elaborated below. It is assumed that instant conversion from water to ice occurs whenever the relative humidity relative to water reaches a certain critical value of \( \text{RH}_c \).

1) When the initial updraft is weak, the parcel maximum vertical displacement is less than \( \delta_c \), the minimum vertical displacement required for the parcel to reach \( \text{RH}_c(T) \), and no cirrus cloud is generated.

2) If \( w_v \) is sufficiently large and the parcel is lifted higher than the threshold \( \delta_c \), ice forms and the air parcel potential temperature increases (i.e., path A–B–C). The parcel potential temperature keeps increasing associated with further ascent and more ice conversion from water vapor, until \( w \) is reduced to zero. If ice particles are large enough and fall out of the parcel, the parcel will maintain its potential temperature \( \theta_p(C) \) during its descent phase until a new equilibrium is reached with \( \theta_p(D) = \theta_p(C) > \theta_{pc} \). Associated with a net increase in potential temperature, the new equilibrium level is higher than the initial level, implying a permanent increase in altitude. The cloud plume length is equal to \( U(\tau_{AC} + \tau_f) \), where \( \tau_{AC} \) is the time for the parcel to travel from A to C and \( \tau_f \) is the ice particle fallout time. As \( \tau_{AC} \) is usually small, and so is \( \tau_f \) for large particles, a short cloud plume is generated in this case.

3) Same as 2 with the exception that the ice particles are too small to fall out. The parcel follows path A–B–C–D and reaches a temporary equilibrium at point D. Then associated with sublimation due to IGW-induced descent, \( \theta_p \) decreases and the parcel gradually descends back to its original level (i.e., A). The cloud plume length is approximately \( U[\tau_{AD} + \max(\tau_{sub}, \tau_f)] \), where \( \tau_{AD} \) is the time for the parcel to travel from level A to D, and \( \max(\tau_{sub}, \tau_f) \) is the larger one of the sublimation time scale \( \tau_{sub} \) and the IGW-induced descent time scale \( \tau_f \). In this case, a long plume composed of small-sized particles is generated.

4) Initially, the parcel is saturated relative to ice and its RH relative to water is still too low for homogeneous nucleation (i.e., \( \text{RH} < \text{RH}_c \)). As the parcel is lifted above the threshold level, similar to the scenario 3, the parcel travels through states A’–B’–C’–D’. Subsequently, slow sublimation and descent occur until a new equilibrium state E’ is reached. The net increase of altitude from A’ to E’ is due to the supersaturation of the parcel relative to ice at the initial point A’, and the cloud length in this case is infinity.

According to the satellite images and model trajectory analysis, processes 2, 3, and 4 are likely relevant to the generation and maintenance of the studied cirrus plumes.

7. Summary

Two cirrus plumes trailing from the Sierra Nevada ridge and Southern Rocky Mountains have been examined based on the analysis of satellite images and diagnosis of the real-data simulations. The synoptic conditions for the two plume events are quite similar. Both events are associated with the presence of a moist layer in the upper troposphere with moisture originating from over the Pacific Ocean, likely related to deep convection and cumulus anvils. Typically, the number concentration of natural ice nuclei is relatively low in the upper troposphere (i.e., above \( \sim 7 \) km). However, as demonstrated, with high relative humidity and low air temperature, it only takes moderate ascent to trigger the homogeneous nucleation mechanism and generate cirrus crystals.

The second common ingredients for the topographically induced cirrus plumes is a relatively strong cross-barrier flow with little directional shear in the troposphere, which is known to favor the excitation, and vertical propagation of the gravity waves. At the 1.5\( \lambda_c \) level, where \( \lambda_c = 2\pi U/N \) is the vertical wavelength of hydrostatic waves, the inertia–gravity wave response is characterized of a strong updraft over the terrain peak and slow descent farther downstream, as shown in both the real-data simulations and the linear solutions with vertical wind and stability profiles derived from the real soundings. The dependence of the crystal number concentrations and size distributions on the amplitude of updraft has been examined in a few previous studies. Strong updrafts at the upwind edge of the waves favor the generation of populated small ice crystals, which reduce the relative humidity and limit the growth of crystals. Small-sized ice crystals with relatively large number concentrations have been found in observational studies of mountain wave–induced cirrus. Clearly, the long cirrus plume tail is mainly composed of small crystals with small terminal velocities, which can be advected over a long distance before falling out.
of the moist layer. The slow descent downstream helps maintain a high relative humidity environment and slow down the ice crystal sublimation. The effect of ice particle fallout, sublimation, and ice physics hysteresis have been discussed in section 6c using the Lagrangian parcel argument. Time and length scales relevant to the cloud plumes have been identified.

Sampling of GOES visible images over the western U.S. states in 2004 indicates that long cirrus plumes can be frequently seen trailing from the Cascade Mountain, Sierra Nevada, and Rocky Mountains. Considering that some cirrus clouds, mainly composed of small crystals (Kahn et al. 2003), might be too thin to be distinguishable from visible images, the occurrence of these cirrus plumes could be significantly more frequent. As noted in previous studies, small ice crystals in cirrus clouds have significant impact on global radiation balance. The impact of these topographically generated long cirrus plumes on global radiation budget needs to be evaluated in future studies.

The COAMPS model exhibits skill in simulating the two plumes. The coarse (i.e., 27 km) grid successfully captures the formation of the long, narrow moisture tongue below the tropopause observed during the Sierra plume event. The cirrus cloud plume generated by the Sierra Nevada ridge is captured by the 3-km grid and the topographic effect in plume formation is evident in the trajectory analysis. However, the cloud ice plume is absent from the 27- and 9-km grids, likely because the simulated vertical motion is weaker in the coarse grids where the terrain is not as well resolved as in the 3-km grid. Similarly, the moisture field derived from the Southern Rockies plume simulation shows reasonable agreement with the GOES water vapor images, and the 3-km grid successfully captures the formation of the ice cloud plume over the mountain peaks, which is not captured by the coarser grids.

The absence of the plumes in the coarse grids suggests the sensitivity of the plume formation to localized vertical motion. The analysis in section 6 suggests a simple approach to parameterize terrain-induced cirrus plumes in large-scale models. According to linear theory, the vertical motion in the upper troposphere simulated in a coarse resolution model could be parameterized using $w = w_\ell h_b/h_t$, where $w_\ell$ is the vertical velocity in the global model, and $h_b$ and $h_t$ are the major barrier heights derived from a higher-resolution terrain dataset and a lower-resolution global model, respectively. The relative humidity as a function of temperature associated with leeside adiabatic ascent can be estimated using the corrected vertical motion and vertical displacement and compared with cirrus formation condition $\text{RH} \geq \text{RH}_c = \text{RH}_{h_b} - 10$, where $\text{RH}_{h_b}$ is given in Eq. (9). If the condition is satisfied, cirrus clouds could form and their impact on radiation should be included in the global model.

Acknowledgments. This research was supported by the Office of Naval Research (ONR) Program Element 0601153N. The authors are indebted to Dr. Steven Miller who brought our attention to the MODIS image. We want to thank Dr. Jerome Schmidt for helpful discussions regarding microphysics. The authors also benefited from discussions with Dr. Ronald Smith. The simulations were made using the Coupled Ocean–Atmospheric Model Prediction System (COAMPS) developed by U.S. Naval Research laboratory.

REFERENCES


