The Impact of the Sierra Nevada on Low-Level Winds and Water Vapor Transport

JINWON KIM
Department of Atmospheric and Oceanic Sciences, University of California, Los Angeles, Los Angeles, California

HYUN-SUK KANG*
Department of Geography, University of California, Los Angeles, Los Angeles, California

(Manuscript received 6 April 2006, in final form 29 November 2006)

ABSTRACT
To understand the influence of the Sierra Nevada on the water cycle in California the authors have analyzed low-level winds and water vapor fluxes upstream of the mountain range in regional climate model simulations. In a low Froude number (Fr) regime, the upstream low-level wind disturbances are characterized by the rapid weakening of the crosswinds and the appearance of a stagnation point over the southwestern foothills. The weakening of the low-level inflow is accompanied by the development of along-ridge winds that take the form of a barrier jet over the western slope of the mountain range. Such upstream wind disturbances are either weak or nonexistent in a high-Fr case. A critical Fr (Fr_c) of 0.35 inferred in this study is within the range of those suggested in previous observational and numerical studies. The depth of the blocked layer estimated from the along-ridge wind profile upstream of the northern Sierra Nevada corresponds to Fr_c between 0.3 and 0.45 as well. Associated with these low-level wind disturbances are significant low-level southerly moisture fluxes over the western slope and foothills of the Sierra Nevada in the low-Fr case, which result in significant exports of moisture from the southern Sierra Nevada to the northern region. This along-ridge low-level water vapor transport by blocking-induced barrier jets in a low-Fr condition may result in a strong north–south precipitation gradient over the Sierra Nevada.

1. Introduction
Orography in California, which is characterized by two nearly parallel two-dimensional (2D) mountain ranges, the Coastal Range and the Sierra Nevada, plays an important role in shaping the regional water cycle through terrain-induced wind disturbances. It is well known that winter rainfall in California is primarily generated by orographic lifting of moisture-rich low-level inflow from the eastern Pacific over the western slopes of these mountain ranges (Chen et al. 1994; Soong and Kim 1996; Kim 1997; Chung et al. 1998; Neiman et al. 2002; Kim and Lee 2003; Grubišić et al. 2005). Extreme elevation changes in the Sierra Nevada further complicate the regional water cycle through their influence on the local snow budget and cloud formation (Kim 2001; Kim et al. 2006). Among these orographic effects, the low-level wind disturbances induced by these mountain ranges play the most fundamental role in determining the precipitation distribution in California, especially in association with winter storms.
In the presence of the effect of the earth’s rotation, there are two elements in mesoscale wind disturbances induced by a 2D ridge: the vertical motion induced by orographic lifting of the low-level inflow perpendicular to the ridge (cross-ridge component, U_c hereafter) and the along-ridge wind below the ridge crest upstream of the mountain range (along-ridge component, U_p hereafter) induced by orographic blocking (Pierrehumbert and Wyman 1985; Smith 1980, 1988, 1989; Smolarkiewicz and Rotunno 1990; Kim and Mahrt 1992a; Rotunno and Ferretti 2001). The vertical motion induced by orographic lifting of U_c over the windward slope of a 2D ridge primarily determines the amount of local precipitation during storms. The well-known precipitation pattern in California, which is characterized

* Current affiliation: Department of Atmospheric Sciences, Yonsei University, Seoul, South Korea.

Corresponding author address: Jinwon Kim, Department of Atmospheric and Oceanic Sciences, University of California, Los Angeles, 405 Hilgard Avenue, Los Angeles, CA 90095-1565. E-mail: jkim@atmos.ucla.edu

DOI: 10.1175/JHM599.1

© 2007 American Meteorological Society
by heavy precipitation over the western slopes of the Coastal Range and the Sierra Nevada accompanied by rain shadows east of them, results from orographic lifting of westerly winds over the western slopes of these mountain ranges (Kim 1997; Grubišić et al. 2005). The impact of terrain-induced vertical motion on precipitation has received the most attention so far since the robust relationship between the vertical motion and local precipitation can be used to predict precipitation in these mountainous regions (Rhea 1978; Kyriakidis et al. 2004; Neiman et al. 2002; Smith 2003; Smith and Barstad 2004).

The along-ridge component of the low-level wind disturbances induced by orographic blocking, often called a barrier jet, can also be important in precipitation over mountainous regions. Low-level moisture transport by barrier jets induced by significant mountain ranges has been shown to strongly influence the distribution of precipitation. In a numerical modeling study, Rotunno and Ferretti (2001) showed that the low-level moisture transport by $U_p$ induced by the blocking of southerly winds by the Alps can cause intense precipitation in the southwestern Alps. Barrier jets can affect precipitation over California mountain ranges in a similar way. Parish (1982) showed that barrier jets are frequent along the western slope of the Sierra Nevada during winter. Low-level moisture transported by barrier jets upstream of these mountain ranges is eventually subject to orographic lifting and results in precipitation in the regions favored by the orography. In a numerical modeling study, Kim and Soong (1996) suggested that along-ridge moisture transport over the foothills and western slope of the Sierra Nevada could have contributed to heavy rainfall during an extreme precipitation event in February 1986 over the Feather River basin in the northern Sierra Nevada. Details of the impact of $U_p$ induced by the Sierra Nevada on precipitation according to large-scale inflow conditions, however, remain to be understood.

The two components of terrain-induced wind disturbances, $U_p$ and $U_c$, are related, and their occurrences and intensities are affected by both the lower tropospheric thermal structure and the inflow speed, in conjunction with the characteristics of the barrier. In the unblocked case (Fig. 1a), inflows near the surface can rise over the barrier and $U_p$ is nonexistent. In the blocked case, in conjunction with the effect of the earth’s rotation (Fig. 1b), low-level inflows decelerate ($U_c$ decreases) and turn left into the ridge-parallel direction (positive $U_p$ develops) as they approach a 2D barrier in the Northern Hemisphere. In this case, a stagnation point ($U_c = 0$) occurs upstream of a ridge (Smith 1980; Pierrehumbert and Wyman 1985; Smolarkiewicz and Rotunno 1990). Thus, only the flow originating above the blocked layer is subject to orographic lifting.

The occurrence of orographic blocking and the depth of a blocked layer can be approximated in terms of the upstream wind speed and the strength of stratification below the crest of a 2D barrier. In an observational study of upslope winds at several locations along the coast of California, Neiman et al. (2002) found that most of the occurrences of orographic blocking are associated with passages of cold fronts during which low-level stratification is generally strong. In a study of mountain waves over the Coastal Range of former Yugoslavia, Kim and Mahrt (1992a) inferred from the profiles of gravity wave stress that the displacement of the streamline at the surface can be approximated in terms of the critical Froude number ($Fr_c$) as

$$D_b = \max \left( h - Fr_c \frac{U}{N}, 0 \right),$$

where $h$ is the height of the ridge, $Fr = U/Nh$, $N$ is the buoyancy frequency of upstream flow calculated between the surface and the crest level, and $U$ is the upstream wind speed at the crest level.

The critical Froude number $Fr_c$ is the value of $Fr$ at which flow stagnation, that is, orographic blocking, oc-
curs upstream of an obstacle. Since linear solutions are only qualitatively valid for $Fr \leq Fr_c$ (Smith 1988), the value of $Fr_c$ is usually estimated in a numerical or laboratory study. On the basis of inviscid 2D linear mountain wave theory, Smith (1988) estimated $Fr_c = 0.77$ with $f = 0$, where $f$ is the Coriolis parameter. In a numerical study with $f = 0$, Smolarkiewicz and Rotunno (1990) estimated $Fr_c \approx 0.5$ for a 3D bell-shaped mountain. Pierrehumbert and Wyman (1985) found $Fr_c \approx 0.67$ for $f = 0$ and that $Fr_c$ varies with the Rossby number $Ro$ when $f \neq 0$ in a numerical study of stratified flows over a 2D barrier. Their results show that $Fr_c$ varies from 0.33 for $Ro = 1$ to 0.45 for $Ro = 2$. For a mesoscale mountain range, such as the Sierra Nevada whose cross-ridge length scale is 50–100 km, $Ro$ is between 1 and 2 in the midlatitudes. On the basis of measured gravity wave momentum flux profiles, Kim and Mahrt (1992a) suggested $Fr_c = 0.32$. This value is smaller than the estimates from numerical studies with $f = 0$ (e.g., Smolarkiewicz and Rotunno 1990) but is close to the value corresponding to $Ro = 1$ in Pierrehumbert and Wyman (1985). This is perhaps the only observational estimate of the value of $Fr_c$. Thus, it may be possible to characterize low-level wind disturbances and the associated moisture fluxes induced by the Sierra Nevada, as well as their impact on precipitation over the mountain range, in terms of $Fr$ calculated from suitably defined upstream flow parameters.

This study aims to understand low-level wind disturbances and the associated moisture transport upstream of the Sierra Nevada and their influence on precipitation during winter storms according to large-scale atmospheric conditions characterized by the values of $Fr$. An outline of the numerical experiment and the criteria used to identify the winter storm cases representative of high- and low-$Fr$ regimes are presented in section 2. The low-level winds and the associated moisture fluxes upstream of the Sierra Nevada and precipitation over the mountain range simulated in the selected cases of contrasting $Fr$ are discussed in sections 3, 4, and 5, respectively. Summary and conclusions follow in section 6.

2. Experimental design

a. Model description

The Mesoscale Atmospheric Simulation (MAS) model (Soong and Kim 1996; Kim 2005; Kim et al. 2006) coupled with the Noah land surface scheme (Kim and Ek 1995; Chang et al. 1999) was employed in this study. The model domain covers California and Nevada with an $84 \times 84$ point horizontal mesh at an 18-km grid spacing (Fig. 2), with 20 atmospheric and 4 soil layers in the vertical. The MAS model is a primitive equation, limited-area atmospheric model based on $\sigma$ coordinates. The advection equation is solved using a third-order accurate finite difference scheme (Takacs 1985), which is characterized by minimal phase errors and numerical dispersion. A modified version of the bulk microphysics scheme of Cho et al. (1989) and the simplified Arakawa–Schubert scheme (Pan and Wu 1995; Hong and Pan 1998) are used to compute grid-scale precipitation and convection, respectively. The effects of vertical turbulent mixing are computed using the bulk aerodynamic scheme of Deardorff (1978) at the surface and $K$ theory within the model atmosphere. The eddy diffusivities for the $K$-theory method are computed using a scheme that combines the nonlocal PBL scheme of Troen and Mahrt (1986) and the local scheme by Louis et al. (1982) above the boundary layer in conjunction with the asymptotic mixing length obtained in the observational study of Kim (1990) and Kim and Mahrt (1992b). Atmospheric radiative transfer including the impacts of clouds and atmospheric aerosols is computed using the 5-2-4-stream Fu–Liou scheme (Fu and Liou 1993; Gu et al. 2003; Kim et al. 2006). A four-layer version of the Noah land surface scheme is coupled with the MAS model to compute the land surface processes and surface fluxes. The Noah model predicts the volumetric soil moisture content, both frozen and unfrozen, and soil temperature.
within model soil layers. It also predicts the canopy water content and snow water equivalence at the surface. The temperature and specific humidity for calculating the surface sensible and latent heat fluxes, outgoing longwave radiation, and ground heat fluxes are calculated by iteratively solving a nonlinear form of the surface energy balance equation. For more details of the MAS and Noah models, readers are referred to Mahrt and Pan (1984), Pan and Mahrt (1987), Kim and Ek (1995), and Soong and Kim (1996). This model has been used previously in a number of regional modeling studies for the western and continental United States as well as East Asia (e.g., Soong and Kim 1996; Kim 1997, 2001, 2005; Kim et al. 2002; Kim and Lee 2003; Kim et al. 2006).

b. Simulation and case selection

A winter season simulation has been performed over the 4-month period December 1997–March 1998 using large-scale atmospheric and sea surface temperature (SST) forcing data from National Centers for Environmental Prediction–Department of Energy (NCEP–DOE) reanalysis version 2 (Kanamitsu et al. 2002). To investigate orographic blocking by the Sierra Nevada and its impact on low-level winds, moisture transport, and precipitation over the mountain range, we have selected two sets of cases based on the presence or absence of significant $U_p$ upstream of the Sierra Nevada when precipitation occurred. We further required that the 700-hPa winds over the eastern Pacific be free of strong curvature and that the direction of the large-scale flow be within $30^\circ$ of the ridge-cross direction. These additional constraints were necessary to avoid complications in low-level winds according to the incident angle of upstream flow (Zängl 2004) and the influence of large-scale pressure gradients on the low-level winds that are difficult to analyze for realistic terrain. Note that these criteria do not explicitly consider Fr. On the basis of these constraints, six cases, three each for the presence or absence of a barrier jet, have been selected for investigation (Table 1). The selected cases compose about 10% of the total number of days on which precipitation occurred in some part of the Sierra Nevada during the 4-month period.

<table>
<thead>
<tr>
<th>TABLE 1. The high- and low-Fr storm cases selected for investigation.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>High-Fr cases</strong></td>
</tr>
<tr>
<td><strong>Low-Fr cases</strong></td>
</tr>
</tbody>
</table>

To investigate the relationship between the upstream Fr and the characteristics of mesoscale wind disturbances induced by the Sierra Nevada, we have calculated Fr using a reference mountain height of 2500 m. This value corresponds approximately to the average height of the Sierra Nevada crest in the simulation, which varies from 2000 m in the northern region to over 3000 m in the southern region. The buoyancy frequency $N$ is calculated using the gradients and the mass-weighted averages of the potential temperature between the lowest model layer and the 2500-m level sampled directly from model results. The 2500-m level wind $U$ is calculated using a variation of the method by Koffi (1994) and Georgelin and Richard (1996) as follows. First, model winds are decomposed into $U_p$ and $U_c$. Subsequently, the values of $U_c$ at 2500 m are calculated from wind profiles obtained from the linear regression of $U_c$ between the lowest model layer and 6000 m using the least squares method. Calculations of Fr in this way have yielded similar values for three different reference crest heights, 2000, 2500, and 3000 m.

3. Low-level winds

The Fr values over the eastern Pacific are smaller than 0.35 in all three cases in which a barrier jet is evident and are larger than 0.35 when it is not (Fig. 3), suggesting Fr$_c$ = 0.35 for the occurrence of orographic blocking. This value of Fr$_c$ is similar to those suggested in previous studies for a 2D orographic barrier by Pierrehumbert and Wyman (1985) and Kim and Mahrt (1992a) that range from 0.32 to 0.45. Note that a value of Fr$_c$ in the range of 0.33–0.45 corresponding to the values of Ro 1–2 in Pierrehumbert and Wyman (1985) is applicable to the cases investigated in this study. We further selected the storms of 1200 UTC 14 January (Fig. 3a) and 0000 UTC 2 February (Fig. 3d) as typical examples of high- and low-Fr inflow cases, respectively, for further analysis below.

In the far-upstream region, that is, over the eastern Pacific near the left edge of Fig. 3, low-level wind directions are similar in both cases; however, they are significantly different near the coast and the Sierra Nevada according to Fr. Low-level winds in the low-Fr case (Fig. 3a) are characterized by the significant turning of the southwesterly inflow over the eastern Pacific toward the north near the coastline and the Sierra Nevada with a stagnation point, which appears over the foothills of the southern Sierra Nevada (marked with a cross in Fig. 3a). The appearance of a stagnation point upstream of the southern Sierra Nevada is a common feature in all three low-Fr cases (Figs. 3a–c). To the south of the stagnation point, low-level flows turn southward, that is, negative $U_p$ develops. This deflection of low-level inflows at the stagnation point is con-
sistent with the results in previous numerical studies of low-Fr flows over an isolated 2D barrier (Georgelin and Richard 1996; Zängl 2004). The turning of low-level inflows in the low-Fr case starts to occur over the eastern Pacific at a distance of about 5–6 times the half-width of the Sierra Nevada (i.e., 250–300 km) from the mountain range. Over the foothills and the western slope of the Sierra Nevada, low-level winds are nearly parallel to the mountain range, indicating the development of a barrier jet. Positive $U_p$ is weak near the stagnation point but intensifies away from it. These characteristics of low-level winds upstream of the Sierra Nevada in the low-Fr case are similar to those found in previous numerical studies of low-Fr flows over an iso-

Fig. 3. The low-level winds averaged over the lowest two sigma layers, about 150 m thick, and the values of Fr in the (a)–(c) low- and (d)–(f) high-Fr cases selected for investigation. The contour intervals are 0.05. Shading indicates the region where Fr is larger than 0.35. The cross in (a)–(c) indicates the locations at which a stagnation point appears.
lated 2D barrier elongated in the cross-stream direction in the presence of the effects of the earth’s rotation (Georgelin and Richard 1996; Rotunno and Ferretti 2001; Zängl 2004). These previous studies also showed that 1) a stagnation point occurs upstream of a 2D barrier; 2) low-level inflows are deflected toward the left and right, with respect to the inflow direction, on either side of the stagnation point; 3) the deflection of low-level inflow starts to occur at a distance 5–6 times the half-width of the orographic barrier; 4) $U_c$ weakens rapidly from its upstream value as it approaches a 2D barrier; and 5) $U_p$ induced by blocking intensifies farther away from the stagnation point with a maximum near the edges of the barrier. Since over 60% of column-integrated water vapor resides within the lowest 2000 m over the eastern Pacific during the storm events analyzed in this study (not shown), $U_p$ induced by orographic blocking can transport a significant amount of moisture into the northern Sierra Nevada region in the low-Fr case. In the high-Fr case (Fig. 3d), the inflow direction remains largely unchanged as it passes over the Coastal Range and the Sierra Nevada. In addition, low-level inflow speed decreases over the foothills and the western slope of the Sierra Nevada and increases to the east of the crest. These variations in the low-level winds in the vicinity of the Sierra Nevada in the high-Fr case are qualitatively similar to those associated with internal gravity waves forced by a 2D barrier in the absence of significant blocking (Smith 1977; Kim 1986; Smolarkiewicz and Rotunno 1990).

To examine low-level wind disturbances upstream of the Sierra Nevada, the $U_p$ field at 500, 1500, and 2500 m above sea level (ASL) are plotted in Fig. 4. The shaded areas in Fig. 4 indicate the regions where the model terrain is above the corresponding level. Here $U_p$ varies in the cross-ridge direction much more rapidly in the low-Fr case (Figs. 4b,d,f) than in the high-Fr case (Figs. 4a,c,e), especially on the lowest two levels, that is, below 1500 m. In both cases, especially in the low-Fr case, the gradient of $U_p$ in the cross-ridge direction increases near the Sierra Nevada. This variation in $U_p$ in the cross-ridge direction decreases rapidly in the vertical, especially in the low-Fr case. The cross-ridge component $U_p$ on the 2500-m level (and above) varies primarily in the north–south direction, that is, in the along-ridge direction, with only weak gradients in the cross-ridge direction in both cases.

According to Fr, $U_p$ below the crest varies significantly (Fig. 5). Important features of $U_p$ in the high-Fr case (Figs. 5a,c,e) include the following: 1) $U_p$ is weaker than the corresponding $U_c$ and is generally negative on all levels, 2) $U_p$ varies primarily in the cross-ridge direction on the lowest two levels and in the along-ridge direction on the 2500-m level, and 3) spatial variation in $U_p$ is weaker compared to that in $U_c$ on all levels. In the low-Fr case, $U_p$ varies primarily in the along-ridge direction with negative (positive) values in the southern (northern) part of the ocean on all levels (Figs. 5b,d,f); $U_p$ develops significant spatial variations near the two mountain ranges, especially in the cross-ridge direction upstream of the Sierra Nevada. The largest value of $U_p$ occurs over the western slope of the southern Sierra Nevada in which $U_p$ exceeds 12 m s$^{-1}$. In addition, positive $U_p$ occurs only on the lowest two levels (Figs. 5b,d).

Low-level positive $U_p$ weakens rapidly in the vertical above 1500 m to become negative, that is, in the same direction as the large-scale flow, on the 2500-m level (Fig. 5f). Thus, significant along-ridge winds, especially the southerly barrier jets, are confined to the close upstream region of the Sierra Nevada well below its crest, despite the fact that positive $U_p$ starts to develop over the eastern Pacific far from the mountain range. It is interesting to note that, over most of the foothill and western slope of the Sierra Nevada north of a stagnation point, $U_p$ on the lowest level (500 m ASL) also varies strongly in the along-ridge direction with a maximum over the western slope of the northern Sierra Nevada and a minimum over the foothill of the southern Sierra Nevada near the stagnation point in the low-Fr case. Similar variations in $U_p$ in the along-ridge direction over the western slope of the Sierra Nevada also occur on the 1500-m level; however, along-ridge variations in $U_p$ are largely confined near the stagnation point. Such strengthening of $U_p$ away from a stagnation point is one of the important characteristics of low-level wind perturbations under low-Fr conditions in the presence of the Coriolis forcing (Georgelin and Richard 1996; Zängl 2004).

Low-level winds, both $U_c$ and $U_p$, vary in the cross-ridge direction much more rapidly in the low-Fr case than in the high-Fr case (Figs. 4 and 5). The rapid decrease of $U_c$ as inflows approach the coastline, accompanied by strengthening of $U_p$, shows that the Coastal Range blocks low-level inflows below the 1000-m level in the low-Fr case. Occurrence of low-level blocking by the Coastal Range on this day was also found in the observational study of Neiman et al. (2002), although they focused on a much smaller spatial scale. This blocking effect by the Coastal Range decreases rapidly upward above the 500-m level because most of the Coastal Range, as represented in the simulation, is below 1000 m. Thus, in the low-Fr case, westerly winds from the eastern Pacific may be influenced primarily by the Sierra Nevada in the lowest 2500 m, with additional blocking effects by the Coastal Range below the 500-m level. It is not possible to separate the impact of the
Coastal Range from that of the Sierra Nevada within
the context of this experiment, however. The differ-
ences in the cross-ridge direction variation in low-level
winds between the high- and low-Fr cases are clearly
seen in the characteristic length scales inferred from the
spatial lagged autocorrelation of $U_c$ and $U_p$ on the
1500-m level (Fig. 6) along the line A1–A2 in Fig. 2.
Characteristic length scales estimated from the $e$-

![Figure 4](image)

Fig. 4. The cross-ridge component $U_c$ (m s$^{-1}$) on the 500-, 1500-, and 2500-m levels in the
(a), (c), (e) high- and (b), (d), (f) low-Fr inflow cases. Dark shading indicates the area where
the model terrain is above the corresponding layer.

$-$

folding distance of lagged autocorrelation values are
about 50 and 200 km for $U_c$ in the low- and high-Fr
cases, respectively. The scale separation between the
high- and low-Fr cases in terms of lagged autocorrela-
tion for $U_p$, however, is not as pronounced as that of $U_c$.
Estimated in the same way, length scales of 170 and 200
km were obtained for $U_p$ in the low- and high-Fr cases,
respectively.
Vertical variations in low-level winds near upstream of the northern Sierra Nevada also differ significantly according to Fr (Fig. 7). Note that the height of the northern Sierra Nevada is about 2000 m and the mean terrain height along the line B1–B2 in Fig. 2 is 200 m. In both cases, $U_c$ is characterized by significant vertical variations near the surface. The thickness of the low-level shear layer is noticeably larger in the low-Fr case (700 m) than in the high-Fr case (400 m). The difference in the depth of shear layers between the two cases, in conjunction with the presence of significant positive $U_p$ in the low-Fr case, suggests that the low-level shear is induced primarily by surface friction in the high-Fr case and by a combination of surface friction and orographic blocking in the low-Fr case. In the high-Fr case (Fig. 7a), $U_p$ is much smaller than $U_c$ on all levels and is negative except on the lowest model layer. The origin of the positive $U_p$ near the surface that occurs only in

Fig. 5. Same as in Fig. 4, but for $U_p$ (m s$^{-1}$).
one of the three high-Fr events selected for analysis, is not clear. It may suggest that localized blocking can occur even in the high-Fr regime depending on local orography and the large-scale flow. In the low-Fr case (Fig. 7b), a well-defined layer of positive $U_p$ appears within the lowest 1500 m from the surface. The along-ridge component $U_p$ is negative over the eastern Pacific and above the 3000-m level in both cases; thus, the positive $U_p$ is most likely to be forced by the Sierra Nevada. Positive $U_p$ in the low-Fr case attains its maximum intensity at 300 m above the ground and subsequently decreases upward above this level. Unlike in the high-Fr case, the magnitude of positive $U_p$ exceeds that of $U_c$ within the lowest 500 m. This suggests that the low-level moisture transport associated with $U_p$ can be significant compared to that by $U_c$ in the low-Fr case. The low-level moisture transport associated with the $U_c$ upstream of the Sierra Nevada is discussed in the following section. Composites of $U_c$ and $U_p$ over the low- and high-Fr cases (Fig. 7) show that these vertical variations in low-level $U_c$ and $U_p$ according to Fr occur commonly in all corresponding cases.

Taken together, low-level wind variations upstream of the Sierra Nevada in these two cases show the characteristics of low-level wind disturbances induced by a 2D orographic barrier under high- and low-Fr conditions found in previous studies. Interestingly, the depth
of the positive $U_p$ layer in the low-Fr case (Fig. 7b) could be approximated using Eq. (1) and the values of $Fr_c$ suggested by Pierrehumbert and Wyman (1985) and Kim and Mahrt (1992a). The depth of a blocked layer based on Eq. (1) in conjunction with the values of upstream wind speed at the 2500-m level ($15 \text{ m s}^{-1}$), the height of the northern Sierra Nevada (2000 m), and buoyancy frequency ($0.0125 \text{ s}^{-1}$) ranges from 1660 to 1460 m for a value of $Fr_c$ between 0.3 and 0.45, respectively. The good agreement between the depth of the positive $U_p$ layer estimated from the $U_p$ profile (Fig. 7b) and Eq. (1) also supports that the positive $U_p$ in the low-Fr case is induced by orographic blocking by the Sierra Nevada.

4. Low-level water vapor fluxes

Low-level water vapor fluxes upstream of the Sierra Nevada are closely related to precipitation over the mountainous region (Kim 1997; Grubišić et al. 2005). The moisture fluxes in the cross- and along-ridge directions within the lowest 3000-m layer ($Q_c$ and $Q_p$, respectively, hereafter), where most of the water vapor is contained, are shown in Figs. 8 and 9. Here $Q_c$ and $Q_p$ were calculated by vertically integrating water vapor fluxes over the lowest nine model layers, which correspond approximately to the lowest 3000-m layer over the Pacific, as

$$Q_c = \sum_{i=1}^{9} q_i U_{ci} \delta p_i / g$$

(2)

$$Q_p = \sum_{i=1}^{9} q_i U_{pi} \delta p_i / g,$$

(3)

where $\delta p$, $q$, and $g$ are the pressure thickness, specific humidity, and the gravitational acceleration, respectively, and the subscript $i$ indicates $i$th model layer.

In the high-Fr case (Fig. 8a), the $Q_c$ upstream of the Sierra Nevada varies primarily in the along-ridge direction largely due to the spatial variation of low-level water vapor, which decreases toward the north with only a minor variation in the east–west direction over the eastern Pacific (not shown). The spatial variations in $U_c$ in the high-Fr case are small (Fig. 4); $Q_c$ varies in the cross-ridge direction off the northern California coast in the high-Fr case as well, however, this variation is much smaller than that in the low-Fr case (Fig. 8b) and is largely due to precipitation over the Coastal Range, which reduces atmospheric water vapor content and moisture fluxes. In the low-Fr case (Fig. 8b), $Q_c$ varies strongly in the cross-ridge direction with relatively minor variations in the along-ridge direction.

This variation of $Q_c$ in the low-Fr case is primarily related with the rapid changes in $U_c$ in the cross-ridge direction due to blocking effects of the Sierra Nevada, especially below the 1500-m level (Fig. 4). Smaller $Q_c$ over the northern Coastal Range is due to local precipitation as in the high-Fr case. The cross-ridge variation in water vapor over the eastern Pacific is small in both cases; thus, Fig. 8 shows that the variation of low-level moisture flux in the cross-ridge direction is strongly affected by the cross-ridge variations in $U_c$ ac-
cording to the presence or absence of orographic block-

In the high-Fr case, \( Q_p \) (Fig. 9a) is smaller than the corresponding \( Q_c \) (Fig. 8) and is negative in most of the upstream region. Also, the along-ridge variation of \( Q_p \) in the high-Fr case (Fig. 9a) is much weaker than in the low-Fr case (Fig. 9b). As a result, the low-level moisture transport into the Sierra Nevada is mostly associated with \( Q_c \) (i.e., the water vapor fluxes perpendicular to the mountain range) in the high-Fr case. In the low-Fr case (Fig. 9b), \( Q_p \) is also negative in the far-upstream region; however, positive \( Q_p \) occurs upstream of the northern Sierra Nevada with a maximum over the western slope of the northern Sierra Nevada where \( U_p \) is strong. Unlike in the high-Fr case, \( Q_p \) in the low-Fr case becomes as large as 70% of \( Q_c \) over the foothills and the western slope of the northern Sierra Nevada (Fig. 9b). The low-level moisture flux divergence associated with \( Q_p \) in the low-Fr case is characterized by a well-defined dipole structure upstream of the Sierra Nevada, with \( \partial Q_p / \partial x_p < 0 \) over the northwestern slope of the Sierra Nevada (Fig. 10) and \( \partial Q_p / \partial x_p > 0 \) over the southwestern slope of the ridge. The differentiation in the along-ridge direction \( \partial / \partial x_p \) was calculated as follows. The \( Q_p \) field was interpolated first onto a rotated coordinate system defined in such a way that the direction of one of its axis is the same as the along-ridge direction. Subsequently, the differentiation \( \partial / \partial x_p \) was calculated by applying a centered finite difference method in the rotated coordinate system. Thus Fig. 10 illustrates the low-level moisture budget associated with the \( Q_p \) component in the along-ridge direction. The most significant loss of moisture due to the along-ridge transport upstream of the southern Sierra Nevada occurs to the north of the stagnation point. Thus, blocking-induced along-ridge winds (\( U_p \)) and the associated moisture fluxes (\( Q_p \)) in the low-Fr case result in the gain of low-level moisture over the western slope of the
northern Sierra Nevada, accompanied by the loss of moisture over the western slope of the southern Sierra Nevada. Such redistribution of low-level moisture between the northern and southern Sierra Nevada by $U_p$ is insignificant in the high-Fr case.

The vertical profiles of the cross- and along-ridge moisture fluxes ($qU_c$ and $qU_p$, respectively, hereafter) differ significantly according to Fr as well (Fig. 11). In the high-Fr case, $qU_c$ (solid line in Fig. 11a) is positive with a maximum near the 500-m level (or 300 m above the ground), while $qU_p$ (dashed line in Fig. 11a) is negative except in the lowest model layer. A composite over the three high-Fr cases (lines with circles in Fig. 11a) shows that similar vertical variations in low-level moisture fluxes occur in all of the selected high-Fr cases. In the low-Fr case, $qU_c$ (solid line in Fig. 11b) is positive with a maximum at 900 m above the ground. Unlike in the high-Fr cases, $qU_p$ is characterized by positive values within the lowest 1500 m, approximately corresponding to the depth of a blocked layer in this region as estimated in section 3, with a maximum at 200 m above the ground (dashed line in Fig. 11b). A composite over the three low-Fr cases (lines with circles in Fig. 11b) also shows that low-level moisture fluxes vary in a similar way in all three low-Fr cases.

5. Precipitation over the Sierra Nevada

The simulated daily precipitation agrees reasonably with the NCEP Unified Raingauge Data (URD; Higgins et al. 2000); however, the model tends to overestimate precipitation in both cases (Fig. 12). The difference in precipitation over the southern end of the Sierra Nevada between the URD analysis and the simulation may be due to model errors or the lack of rain gauges in this high-elevation region. The lack of high-elevation rain gauges is one of the major sources of systematic errors in precipitation datasets based on analysis of rain gauge data and often results in underestimation of precipitation in high-elevation regions in the western United States (Kim and Lee 2003). In the high-Fr case (Fig. 12a), precipitation occurs over a large part of the Sierra Nevada as well as over the Coastal Range. In the low-Fr case (Fig. 12b), precipitation occurs mostly over the northern Sierra Nevada compared to the high-Fr case, perhaps due to the low-level water vapor transport by $U_p$ that is expected to bring extra water vapor into the northern Sierra Nevada at the expense of the southern region. Precipitation over the Sierra Nevada in all of the low- and high-Fr cases examined in this study varies similarly over the Sierra Nevada according to Fr (not shown).

6. Summary and conclusions

To understand the impact of the Sierra Nevada on low-level wind disturbances, moisture transport, and precipitation during winter storms, we have analyzed data from a seasonal climate simulation. A value of the critical Froude number of 0.35 inferred in this study on the basis of the presence or absence of a barrier jet is
close to those obtained in previous observational and modeling studies. This suggests that the simulated low-level wind disturbances upstream of the Sierra Nevada take on the characteristics of blocked or unblocked flow by the Sierra Nevada according to the Fr values of the inflows over the eastern Pacific.

The simulated low-level winds upstream of the Sierra Nevada vary distinctively according to Fr. Low-level wind directions remain similar upstream of the Sierra Nevada in the high-Fr case but vary sharply near the Sierra Nevada in the low-Fr case. To the north (south) of a stagnation point, significant southerly (northerly) along-ridge winds develop over the foothills and the western slope of the Sierra Nevada. The deflection of the low-level inflow in the low-Fr regime is most notable over the western slope of the northern Sierra Nevada where a southerly barrier jet occurs. These variations in the low-level winds upstream of the Sierra Nevada closely resemble one of the characteristics of mesoscale wind disturbances upstream of a 2D barrier in the absence or presence of orographic blocking, in conjunction with the effects of the earth’s rotation, according to Fr found in previous studies (Pierrehumbert and Wyman 1985; Georgelin and Richard 1996; Zängl 2004). The cross-ridge direction variations in low-level winds are more rapid in the low-Fr case than in the high-Fr case.

The vertical variations in low-level winds are characterized by a presence and absence of significant positive $U_p$ upstream of the Sierra Nevada below the crest level in the low- and high-Fr regimes, respectively. Positive $U_p$ in the low-Fr case varies significantly in the along-ridge direction with a maximum over the western slope of the northern Sierra Nevada, similar to the findings in the numerical studies by Georgelin and Richard (1996) and Zängl (2004). The thickness of the layer $U_p > 0$ in the low-Fr regime could be estimated as a function of the critical Froude number (Fr$_c$), buoyancy frequency, and $U_c$, near the crest level using the relationship and the value of Fr$_c$ suggested by Stern and Pierrehumbert (1988) and Kim and Mahrt (1992a). Thus the characteristics of the low-level wind disturbances investigated in this study are related to a presence or absence of blocking by a 2D orographic barrier according to Fr.

**Fig. 12.** The simulated (MAS) and observed (URD) daily precipitation in the (a),(b) high- and (c),(d) low-Fr cases.
Low-level wind disturbances generated by the Sierra Nevada result in contrasting low-level moisture transport upstream of the ridge according to Fr. In the high-Fr case, $Q_p$, is significantly smaller than the corresponding $Q_c$; thus, low-level moisture transport into the Sierra Nevada is primarily associated with the cross-ridge winds. In the low-Fr case, $Q_p$ is positive within the lowest 1500 m, with a maximum at 300 m above the ground, before it turns negative (i.e., in the direction of the large-scale flow) above the level upstream of the Sierra Nevada. In this case, $Q_p$ increases toward the northern Sierra Nevada to the north of a stagnation point. Divergence of $Q_p$ shows that blocking-induced barrier jets export a significant amount of moisture into the northern Sierra Nevada out of the southern region in the low-Fr case. These low-level wind disturbances and the associated moisture fluxes upstream of the Sierra Nevada may affect the north–south distribution of precipitation over the Sierra Nevada according to Fr. In the high-Fr case, the simulated precipitation occurs over a wide region with relatively small differences between the northern and southern Sierra Nevada. In the low-Fr regime, precipitation over the northern Sierra Nevada is much larger than over the southern region, perhaps due to the southerly low-level moisture transport by $U_p$.

The Coastal Range appears to affect the low-level wind and moisture flux below the 500-m level according to Fr (e.g., Neiman et al. 2002) in a similar way; however, the impact of the Coastal Range could not be separated from that of the Sierra Nevada within the context of this study. The influence of the Coastal Range on low-level wind disturbances and the associated moisture transport on precipitation may be small compared to that of the Sierra Nevada.

Results from this study show that low-level wind fields and the associated moisture fluxes upstream of the Sierra Nevada, as well as precipitation over the mountain range, vary systematically according to Fr. As Fr is determined by the characteristics of large-scale circulation, in particular wind speed and low-level stratification, alterations in static stability and/or wind speed in the lower troposphere can affect precipitation over the Sierra Nevada. The global warming induced by the increase in greenhouse gases, for example, may influence systematically the precipitation distribution and water cycle over the Sierra Nevada through alterations in the lower-tropospheric thermal structure over the eastern Pacific, in addition to changes in storm tracks, atmospheric moisture content, and temperatures.

Acknowledgments. The authors thank Steven Markus and Bjorn Stevens for careful reading and comments that were valuable for improving this paper. This study was supported by the grants from NOAA/GAPP (NA03OAR4310012), NASA/SENN (NAG5-13248), and NASA-ESE/IDS (NAG5-11363).

REFERENCES


——, 2005: A projection of the effects of the climate change in-
duced by increased CO₂ on extreme hydrologic events in the western U.S. *Climate Change,* 68, 153–168.


———, and W. Wu, 1995: Implementing a mass flux convection parameterization package for the NCEP medium-range forecast model. NMC Office Note, 40 pp. [Available from NCEP/EMC, 5200 Auth Road, Camp Springs, MD 20744.]


