The Role of Upstream Midtropospheric Circulations in the Sierra Nevada Enabling Leeside (Spillover) Precipitation. Part II: A Secondary Atmospheric River Accompanying a Midlevel Jet

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ABSTRACT

The synoptic structure of two case studies of heavy "spillover" or leeside precipitation—1–2 January 1997 and 30–31 December 2005—that resulted in Truckee River flooding are analyzed over the North Pacific beginning approximately 7 days prior to the events. Several sequential cyclone-scale systems are tracked across the North Pacific, culminating in the strengthening and elongation of a polar jet stream's deep exit region over northern California and Nevada. These extratropical cyclones separate extremely cold air from Siberia from an active intertropical convergence zone with broad mesoscale convective systems and tropical cyclones. The development of moisture surges resulting in leeside flooding precipitation over the Sierra Nevada is coupled to adjustments within the last wave in the sequence of cyclone waves. Stage I of the process occurs as the final wave moves across the Pacific and its polar jet streak becomes very long, thus traversing much of the eastern Pacific. Stage II involves the development of a low-level return branch circulation [low-level jet (LLJ)] within the exit region of the final cyclone scale wave. Stage III is associated with the low-level jet's convergence under the upper-level divergence within the left exit region, which results in upward vertical motions, dynamic destabilization, and the development of mesoscale convective systems (MCSs). Stage IV is forced by the latent heating and subsynoptic-scale ridging caused by each MCS, which results in a region of diabatic isallobaric accelerations downstream from the MCS-induced mesoridge. During stage IV the convectively induced accelerating flow, well to the southeast of the upper-level jet core, organizes a midlevel jet and plume of moisture or midlevel atmospheric river, which is above and frequently out of phase with (e.g., southeast of) the low-level atmospheric river described in Ralph et al. ahead of the surface cold front. Stage V occurs as the final sequential midlevel river arrives over the Sierra Nevada. It phases with the low-level river, allowing upslope and midlevel moisture advection, thus creating a highly concentrated moist plume extending from near 700 to nearly 500 hPa, which subsequently advects moisture over the terrain.

When simulations are performed without upstream convective heating, the horizontal moisture fluxes over the Sierra Nevada are reduced by −30%, indicating the importance of convection in organizing the midlevel atmospheric rivers. The convective heating acts to accelerate the midlevel jet flow and create the secondary atmospheric river between −500 and 700 hPa near the 305-K isentropic surface. This midlevel moisture surge slopes forward with height and transports warm moist air over the Sierra Nevada to typically rain shadowed regions on the lee side of the range. Both observationally generated and model-generated back trajectories confirm the importance of this convectively forced rapid lifting process over the North Pacific west of the California coast −12 h and −1200 km upstream prior to heavy leeside spillover precipitation over the Sierra Nevada.
1. Introduction

Primary sources of water vapor over the North American Pacific coast during heavy precipitation events are “atmospheric rivers,” first theorized by Newell et al. (1992) and analyzed by Ralph et al. (2004). The flux of water vapor in these “rivers of air” is comparable to the flux of water within the Amazon River, thus justifying the term “atmospheric river.” The magnitude of the flux in these atmospheric rivers has been calculated by Newell et al. (1992). Their calculations, based on European Centre for Medium-Range Weather Forecasts (ECMWF) analyses, were made in conjunction with the development of midlatitude cyclones over the Pacific Ocean during the Measurement of Air Pollution from Satellites (MAPS) project. They found that anomalously large water vapor fluxes were concentrated in filaments or tropospheric rivers. Often, there were four or five of these filaments leading from the tropics to midlatitudes in both hemispheres as precursors to the heavy rainfall events. Cloudiness estimates from the Total Ozone Mapping Spectrometer (TOMS) instrument were consistent with the number and placement of the filaments. These initial results have been examined in greater detail by Zhu and Newell (1994, 1998).

The basic idea of atmospheric rivers was further explored by Ralph et al. (2004) in conjunction with the California Landfalling Jets (CALJET) experiment and the Pacific Landfalling Jets (PACJET) experiment projects that included more detailed observational data. An array of special observations and instruments were used in these projects; among them were the following: dropsondes, C-band radar, wind profilers, Geostationary Operational Environmental Satellites (GOESs), Special Sensor Microwave Imager (SSM/I), ground-based observational stations (rain gauges, surface observations), and rawinsondes. A principal result from these studies was the discovery of the existence of two frontal regions—one to the north and one to the south of the low-level atmospheric river. In association with the south-positioned front was a moisture plume extending from the surface to 600 hPa. The plume neighbored strong convection, and the flux maximum was associated with a low-level jet (LLJ). The moisture flux on the warm side of the north-positioned front was significantly weaker than that associated with the south-positioned front. Ralph et al. (2004) derived key features of the atmospheric river using products from the polar-orbiting and geostationary satellites. They used the SSM/I measurements to estimate the integrated water vapor (IWV). Several hundred composite structures of IWV were found that indicated the presence of long, narrow plumes where IWV > 2 cm. Composites of rain rate (RR) and cloud water (CW) indicated that the widths of these features were about half that associated with the IWV. They conjectured that these results indicated vertical circulations driving the cloud and rain features were roughly half the size of the circulations driving the IWV plume.

Further analysis of the CALJET and PACJET data by Ralph et al. (2004) concentrated on diagnosing conditions that led to heavy orographic precipitation along California’s Coast Range. They found that the strong winds associated with the LLJ led to large water vapor flux linked to the extreme rainfall over the coastal mountains. They also found that orographic flooding was associated with a moist-neutral stability of the air that extended from the surface to elevations well above the LLJ and coastal mountain tops.

The objective of this study is to test the hypothesis that midlevel moisture transport resulting in large part from convective and jet streak forcing over the Pacific Ocean is the critical mechanism providing moisture for spillover or leeside precipitation during the extreme flood events of 1–2 January 1997 and 30–31 December 2005 in northwestern Nevada and northeastern California. These events were associated with the flood-producing increase in the Truckee River streamflow in the Reno, Nevada (REV), metropolitan region that caused millions of dollars in property damage in a semiarid region (Fig. 1). This study will focus on the evidence for a secondary atmospheric river in the midtroposphere, only indirectly coupled to the primary river of Newell et al. (1992) and Ralph et al. (2004). This midlevel secondary river is hypothesized to be an important source of water vapor for heavy orographic rainfall that spills over the Sierra Nevada far removed from the Pacific coast and ultimately swells the Truckee River, resulting in major flooding. Underwood et al. [2009 (Part I)] described the composite climatological forcing leading to spillover precipitation for many historical case studies. This study will provide additional finer-scale synoptic/dynamical evidence for the midlevel moisture transport processes discussed in Part I that sustain this elevated moisture plume or secondary atmospheric river from the Pacific Ocean ~1200 km west of the California coast to Reno, Nevada, and the Truckee River drainage basin.

In the next section of the paper, we will first provide a brief discussion of how the moisture, wind, and stability profiles are different over Reno, Nevada, from the low-level atmospheric river model of Ralph et al. (2004) for two extreme Truckee River flooding events. Section 3 will then provide a detailed overview with observations of the multiday precursor features of these two flood case studies on the leeside of the Sierra Nevada. This will include detailed observationally generated back trajectories and synoptic analyses, indicating when, where, and
FIG. 1. National Weather Service (NWS) multisensor precipitation analyses (in.) valid for (a) 1–3 Jan 1997 and (b) 30–31 Dec 2005. White lines on (b) are terrain elevation contours.
why parcels become elevated into the midtroposphere prior to arriving in Reno, Nevada, during the heavy lee-side rainfall events. This will be followed by section 4, which describes numerical simulation sensitivity studies to determine in more detail the effect of upstream convection on midlevel jet formation and parcel ascent that resulted in strong 3D moisture fluxes and secondary midlevel atmospheric river formation centered on and above the sloping 305-K isentropic surface. Lastly, in section 5 we will summarize our findings.

2. Atmospheric structure during two Truckee River flooding events

Both case studies contained extreme precipitation on the windward and lee sides of the Sierra Crest and Carson Ranges in northern Nevada and California (see Fig. 1), resulting in extensive and damaging flooding of the Truckee River in Reno and Sparks, Nevada. The heavy rainfall in these two events occurred 1–3 January 1997 and 30–31 December 2005 for the 1997 and 2005 flood events, respectively, and exceeded 50 mm downstream from the Sierra Nevada and nearly 400 mm over the upslope of the Sierra Nevada.

Figure 2 is adapted from the Fig. 13 conceptual model found in Ralph et al. (2004). The modified profiles in Fig. 2 are consistent with the soundings in Fig. 3 and have some substantial differences from the composite profiles of wind, stability, and moisture fluxes from Fig. 13 in Ralph et al. (2004), which were evaluated for a larger sample of cases during CALJET and PACJET upstream along the Pacific coast of northern California. The modified profiles in Figs. 2 and 3a–c show a midlevel (~500–700-hPa or ~6–3-km MSL) wind maximum ~3.5–0.5 km above the low-level jet signal in the composites from Ralph et al. (2004; when evaluated relative to the ~1.5-km MSL surface elevation of the rawinsonde location at REV). This elevated wind maximum is accompanied by a much deeper and higher moist neutral layer relative to the composite of Ralph et al. (2004) in Fig. 2 and a secondary maximum of moisture and heat flux nearly coincident with the midlevel jet, which averages ~30–35 m s$^{-1}$ in the two flood case studies. It is the process resulting in this difference between the ~1-km deep feature relative to MSL or primary low-level atmospheric river of Ralph et al. (2004) as well as Newell et al. (1992) and Zhu and Newell (1994, 1998) and the terrain-relative ~3.5–0.5 km deep feature or secondary midlevel atmospheric river that is the focal point of this paper. Importantly, secondary midtropospheric jets and moisture flux maxima—noted on the soundings in Figs. 3d,e,h—are typical both upstream at Oakland, California (OAK), and over REV. This demonstrates that the secondary or midlevel river is in existence ~200 km upstream from REV at a location near sea level—that is, OAK—indicating that the midlevel rivers form prior to reaching both the Pacific coast as well as the Sierra Nevada and thus are likely distinct features from the low-level river before reaching North America.

3. Observational analyses/synoptic structure of the upstream atmosphere for two Truckee River flood events

a. Data sources and analyses procedures

In this observational analysis, General Meteorological Package (GEMPAK) software is employed to use 1) National Centers for Environmental Prediction (NCEP) global reanalysis data at 2.5$^\circ$ x 2.5$^\circ$ horizontal resolution; 2) Aviation (AVN) analysis data at 1.0$^\circ$ x 1.0$^\circ$ horizontal resolution; and 3) North American Regional Reanalysis (NARR) analysis data at ~3 times the resolution of the AVN data (e.g., Kalnay et al. 1996; Mesinger et al. 2006). Supplementing the GEMPAK analysis are satellite data from SSM/I observing integrated water vapor as well as Tropical Rainfall Measuring Mission (TRMM) rainfall data. Moisture flux from rawinsonde observations is calculated employing the formulation used in Ralph et al. (2004): \( \rho Vq \times \Delta Z \times \Delta L \), where \( \rho \) is the density of air, \( V \) is the total wind speed, \( q \) is the mixing ratio, \( \Delta Z \) is the difference between the 50 hPa heights of the layer, and \( \Delta L \) is the width of the region that the flux is evaluated over—assumed here to be 1 km.

This observational analysis will examine how moisture is transported from the tropics up into the midtroposphere to interact with midlatitude waves. A back
trajectory analysis is constructed and diagnosed in an effort to determine the subsynoptic details of air transport. This is accomplished employing the Hybrid Single Particle Lagrangian Integrated Trajectory Model (HYPLIT; Draxler and Hess 1997, 1998) code for backward trajectories using the NCEP global reanalysis datasets (Kalnay et al. 1996). Back trajectories are assumed to originate above REV in the region of heavy leeside precipitation.

b. 1997 case study

1) OBSERVED MOISTURE FLUX STATISTICS SUPPORTING DUAL ATMOSPHERIC RIVERS AT REV

Horizontal moisture flux values were calculated for the period from 0000 UTC 1 January to 1200 UTC 1 January (denoted 00/1 and 12/1, and similarly throughout) employing OAK and REV sounding data in Fig. 3 as well as the technique using the moisture flux formulation specified in Ralph et al. (2004). At both times, the OAK sounding in the 850–900-hPa layer (Figs. 3d,e) indicates a low-level river signal with huge moisture fluxes greater than $12 \times 10^4$ kg s$^{-1}$ accompanying low-level flow from nearly due south and greater than 20 m s$^{-1}$. However at 00/1 a distinct elevated separate layer of moisture flux greater than $5 \times 10^4$ kg s$^{-1}$ can be seen between 650 and 700 hPa, with larger wind velocities from the west-southwest. The midlevel maximum of moisture flux reflects larger wind velocities and smaller mixing ratio values than the low-level maximum. This second layer, with a maximum of $\sim 60\%$ of the low-level moisture flux maximum, is a candidate for the midlevel river because the 305-K surface passes right through the base of this layer and is coincident with the arrival of the first along-gradient surge of moisture (depicted in Fig. 9k) to be described later. The separate midlevel wind and moisture flux maxima, with substantially differing airflow trajectories upstream from REV at OAK, support the contention that two rivers exist and that the midlevel river has two surges at different times embedded within wave 5 (to be described later, evident in Fig. 7b’s satellite data). Additionally, the stronger midlevel airflow is nearly parallel to trajectories 1–3 (Fig. 7a) starting from REV at 12/1 that pass over OAK before arriving at REV, as described later. The 00/1 REV sounding indicates that the largest magnitude horizontal moisture fluxes greater than $8 \times 10^4$ kg s$^{-1}$ exist just above 305 K within the 625–675-hPa layer, where nearly 30 m s$^{-1}$ southwesterly flow exists similar to the upstream data at OAK near 305 K.

By 12/1 the OAK sounding in Fig. 3e indicates an increase of both the low-level south-southwesterly and midlevel southwesterly air flows, resulting in a deep and relatively uniform horizontal moisture flux zone extending over several 50-hPa deep layers from 775 to 925 hPa greater than $12 \times 10^4$ kg s$^{-1}$ with little if any distinct maximum aloft in the midtroposphere. Peak values exceeding $15 \times 10^4$ kg s$^{-1}$ are observed between 875 and 925 hPa, which is typical of a low-level river (e.g., Ralph et al. 2004). However, at REV in Fig. 3b, the peak midlevel river signal is evident at this time as values greater than $7 \times 10^4$ kg s$^{-1}$ can be seen near or just above 305 K between 600 and 700 hPa where wind speeds approach 40 m s$^{-1}$ from the southwest. These horizontal moisture flux values at REV are coincident with the 00/1–00/2 soundings during the period of 700–400-hPa precipitable water increases approaching or exceeding 20 mm, with unambiguous maxima associated with a midlevel river (Fig. 4).

2) OBSERVED PACIFIC SYNOPTIC EVOLUTION OVERVIEW

Figure 5 depicts the multiday polar jet and precipitable water analyses structure over the North Pacific Ocean starting at the time of the heavy rainfall sounding location at REV and working backward for both case studies. Figure 6 depicts the midtropospheric sounding location at REV and working backward for both case studies. Figure 6 depicts the midtropospheric cold pools from a northern Pacific perspective that support each cyclone-scale wave for both case studies over the same time period as Fig. 5. The cyclone-scale waves (wavelengths $\sim 2500$ km) are defined by a polar jet maximum accompanying a wind shift with the 300-hPa trough/ridge system from southwesterly flow downstream and northwesterly flow upstream, a 500-hPa thermal ridge downstream and thermal trough upstream, synoptic-scale warm air advection downstream in the thermal ridge at 500 hPa, and cold air advection upstream in the thermal trough at 500 hPa as well as a distinct synoptic-scale plume of column-integrated precipitable water greater than 30 mm just downstream of the 300-hPa trough. A summary of tropical wave activity, which supplied moisture to these extratropical waves in Figs. 5 and 6, can be found in Tables A1–A2.

The 1997 flood event is associated with five of these cyclone-scale waves (only three are shown). As each wave is tracked backward from the Sierra Nevada in time, these waves possess a 300-hPa wind speed maximum and southwest–northeast tongue of precipitable water exceeding 30 mm extending from the northernmost location of the wave back to the tropics in Fig. 5. The waves all are associated with a 300-hPa jet streak of wavelength greater than 2500 km, which acts to extend and strengthen the cross-Pacific baroclinic zone and polar jet exit region toward the Pacific coast of North America.
during the later period. This results in a gradual upscale extension of the polar jet streak consistent with the elongation and strengthening of the south–north baroclinic zone over the North Pacific over time (notice the extension of the cross-jet temperature gradient over the Pacific in Figs. 6a,d). Viewing the sequence backward in time from REV, the weakening of the south–north baroclinic zone across the Pacific can be seen. Each cold pool in Figs. 6a–c, d–f accompanying each wave in Fig. 5 is depicted incrementally in time upstream across the Pacific from the North American coastline to east of Asia. The extremely cold pools strengthen as they are analyzed upstream toward Asia across the Pacific. However, the net effect of all five waves and their attendant individual cold pools backward in time is to reduce the horizontal length of this baroclinic zone upstream. Also, this includes the weakening of this baroclinic zone, as can be seen in the upstream precession of cold pools depicted in reverse order in time in Fig. 6. Wave 5 in the sequence, which is depicted in Figs. 5a and 6a, is the last of the
sequence to reach the Pacific coast because all five waves develop in various locations within the Pacific since very cold Siberian air supports cyclone development. Additionally, this backward weakening reflects the modification of the early pools of cold polar air in the lower troposphere as a result of the upward sensible and latent heating within the planetary boundary layer as they initially migrate across the Kuroshio Current near Asia.

Waves 5, 4, and 3 are unique in the wave train, because it is during their landfall that precipitable water values at REV and OAK both show substantial increases, as can be seen in Fig. 5. During the period encompassing 0000 UTC 26 December–0000 UTC 27 December 1996 (00/26–00/27), wave 3 landfalls force a relative increase in the precipitable water at REV (>10 mm with average values ~13 mm) and OAK (>20 mm with average values ~25 mm) when compared to the antecedent and subsequent rawinsonde reports. These values are much higher at both locations, ~60% at OAK and ~100% at REV, when compared to the previous 60-h period, which included the landfall of waves 2 and 1. By 00/29 wave 4 results in a second episode of precipitable water less than 10 mm with average values ~14 mm at REV and less than 25 mm with average values ~29 mm at OAK through 12/31 (1200 UTC 31 December). An even higher surge of water vapor arrives during 00/1–2 with average values at REV of ~20 mm and at OAK of ~39 mm accompanying wave 5 during the heavy precipitation. The greatest increase in local water vapor
accompanies wave 5 at REV, with the greatest midlevel warming and strongest southwesterly jet greater than 40 m s\(^{-1}\) within the 500–700-hPa layer. Pulses of midlevel warming and wind acceleration also accompany waves 3 and 4 at REV; however, they are not as strong as wave 5.

3) OBSERVED PACIFIC BACKWARD TRAJECTORY ANALYSES

As can be seen in Tables A1, A2, tropical convection is widespread during the entire period of all five active cyclone-scale wave lifetimes encompassing 12/20–00/01. However, the ascent of parcels that provided water, warmth, and momentum to the midtroposphere in the REV sounding in Fig. 3a viewed backward in time occurs primarily between 12/01 and 00/31, which is consistent with precipitable water increases mentioned earlier in Fig. 4 accompanying wave 5 at REV. Figure 7a depicts HYSPLIT backward trajectories calculated from coarse-resolution NCEP global reanalysis data valid from 12/26–12/01 1996–97 and originating above REV between ~550 and 725 hPa or within the midlevel jet noted in Fig. 3a at REV. Juxtaposed with the trajectories in Fig. 7b is SSM/I water vapor imagery valid on 12/31. Evident in Fig. 7a are three key trajectories that begin over REV in the elevated moist neutral layer between 550 and 723 hPa in Fig. 3a traced backward in space and time.
Trajectory 3 is traced back to a large tropical anticyclone on 12/24 at ~725 hPa near 0°, 165°E, trajectory 2 is traced back to outflow from Tropical Storm (TS) Fern at ~650 hPa on 12/24 near 10°N, 160°E, and trajectory 1 is traced back to an easterly wave and subsequent low-level jet at 850 hPa near 10°N, 115°W on 12/24. These three trajectories arrive at REV at 550 hPa, 624 hPa, and 723 hPa, respectively, on 12/01, where mixing ratio values are ~2.5, 5.3, and 6.4 kg kg$^{-1} \times 10^3$, respectively, and column precipitable water maximizes in Figs. 3a and 4. They converge in horizontal space just south-southeast of 30°N, 140°W or just northeast of the Hawaiian Islands during ~06/31–12/31. This convergence occurs as they begin to undergo their maximum 18–24-h lifting, rising ~75, 125, and 175 hPa, respectively, during the period including ~24–0 h back in time or ~2400–0 km upstream of REV. Figure 7b shows multiple SSM/I-derived moisture plumes or rivers that coalesced relatively close to the Pacific coast of California that arrived from the southwestern and south central tropical Pacific on 12/31 close to the trajectory convergence near 30°N, 140°W. Their region of origin most closely corresponds to trajectories 1 and 3 in Fig. 7a, including air from west of 160°E and east of 160°W, respectively. The trajectories
and the satellite imagery indicate that moisture from the tropical western, central, and eastern Pacific on either side of the Hawaiian Islands arrived at REV during the period of heaviest precipitation, resulting in the Truckee River flood. The eastern Pacific moisture plume arriving at 723 hPa is much larger in magnitude and is lifted over a greater depth. The trajectories indicate that the parcel east of the Hawaiian Islands, which originated within the lower troposphere, is lifted over the longest vertical distance. This parcel is the wettest after being converged with air parcels rich in moisture from farther west in the Pacific. This is consistent with the satellite imagery in Fig. 7b indicating multiple moist airstreams converge near 30°N, 140°W, likely comprising the moisture arriving at REV on 12/1, resulting from strong confluence, lifting, and zonal accelerations within wave 5.

4) OBSERVED PACIFIC SUBSYNOPTIC DYNAMICS ACCOMPANYING THE MIDLEVEL ATMOSPHERIC RIVER

The period of confluence of the three trajectories described earlier began near or just south-southeast of 30°N, 140°W at ~12/31, which would be in a region and time controlled by wave 5. The subsynoptic-scale dynamics within this confluent zone are crucial to the understanding of how the moisture arrives at REV at 12/1. As mentioned earlier, although waves 3 and 4 produced substantial moisture increases and some precipitation at
REV and OAK, the flooding precipitation occurred over and downstream from the Sierra Nevada with wave 5 from 12/31 to 00/1. This is particularly true at REV, where moistening was, on a percentage basis, much more impressive than at OAK during this period—that is, more than 70% at REV relative to 12/31. The focused analysis region and time will be within the rectangle formed by 25°–35°N, 160°–140°W on about 00/31–12/31.

Key features in Figs. 8 and 9 include the following: 1) three maxima of convective precipitation occupying most of this region (Fig. 8); 2) the arrival of a strong polar jet exit region at 300 with 500 hPa ascent in the left exit region as well as to the right—that is, southeast of the exit region (Figs. 9a,b); 3) a low-level ageostrophic return branch circulation with a 925-hPa LLJ directed north-northeastward orthogonal to and under the 300-hPa exit region, which supports low-level mass flux convergence of air and low-level moisture advection where the 500-hPa omegas are negative and therefore upward (Figs. 9b–d); 4) dual significant subsynoptic wind maxima at 305 K south of the polar jet (Figs. 9e,f); 5) northward-directed ageostrophic midlevel flow with each of the southeastward shifted subsynoptic wind maxima in Figs. 9g,h resulting from subsynoptic ridging; and 6) 305-K moisture advection boundaries or discontinuities in Figs. 9i,l above and displaced to the southeast of the 925-hPa ageostrophic wind maximum and moisture tongue in Fig. 9d.

The centroid of this focused analysis region is near 30°N, 150°W or ~750 km upstream from the region of trajectory 1–3 confluence at ~06/31–12/31, which is also when the massive decrease in all three trajectory parcel pressures begins to occur close to the locus of confluence south-southeast of 30°N, 140°W. This average decrease in parcel pressures of ~125 hPa in ~24 h is ~500 km downstream from the time and location of convective precipitation in Fig. 8. The North American Regional Reanalysis (NARR) 3-h analyzed convective precipitation that occurs between 21/30 and 00/31 supports the juxtaposition of the propagating polar left jet exit region at 300 hPa, upward vertical motion at 500 hPa, and the low-level convergence at 925 hPa, resulting in the organization of widespread convection in the region encompassing 25°–35°N, 160°–140°W during the early period of 00/31. This is just upstream from the 30°N, 140°W location of trajectory 1–3 confluence. Thus, the convection is organized where the exit region transverse ageostrophic circulation is observed; however, the subsynoptic wind maxima on 305 K are generally south and east of the convection at this time.

There are at least two 305-K midlevel wind maxima extending southeastward from the location of wave 5 and the jet exit region during 00/31–18/1. The first midlevel jet at 305 K forms to the southeast of the dual mesoscale convective systems (MCSs) near 35°N, 150°–140°W just before 00/31. This midlevel jet propagates northeastward and reaches the OAK region by 18/31 and is associated with a surge in midlevel moistening at REV, particularly above 650 hPa, where values increase from ~4 kg kg⁻¹ × 10³ to greater than 5 kg kg⁻¹ × 10³ over less than six hours. Figures 9k,l depict the vertical moisture cross sections from 00/31 to 12/31. These forward sloping (with height) moisture fronts, accompanying wind perturbations 1 and 2, develop ahead of the convective precipitation southeast of the polar jet’s left exit region. Heating (not shown) from this convection likely intensified the slope of the midlevel (305 K) isentropic surface, as diagnosed from the gradient of pressure on that surface in Figs. 9g,h. The strongest along-river gradient of moisture, as can be seen in the vertical cross sections in Figs. 9k,l, traverses the region from 30°N, 140°W to REV between 00/31 and 12/31 within the 400–800-hPa layer slightly in advance of the low-level moisture flux. This low-level
moisture flux is located below 925 hPa farther west of the midlevel gradient. By 12/31–18/31, a second midlevel jet and attendant moisture flux maximum develops southeast of the more southwestern MCS near 27.5°N, 137.5°W and results in the thrusting upward of moisture, creating a second along-river gradient above 700 hPa that arrives at REV at ~06/1, increasing the 700-hPa moisture values there to ~7 kg kg⁻¹ × 10³. These two midlevel rivers are not in phase with the low-level river, because they lay to the southeast of the 925-hPa jet and moisture maxima. Although just before the heaviest rainfalls over the Sierra Nevada and REV, upslope moisture maxima at the base of the Sierra Nevada become more closely aligned with the midlevel flux zone—that is, after 06/1.

These observations support the likelihood of finer-scale convective lifting of air parcels that enhanced larger-scale quasigeostrophic ascent within and just to the southeast of the polar jet’s left exit region as it traversed a course from ~30°N, 150°W to the Sierra Nevada.
during a 30–36-h period within 31 December 1996–1 January 1997. The period of these adjustments is coincident with the arrival of substantial moistening at OAK and REV during the heavy spillover precipitation, as indicated in Figs. 3a and 4 accompanying wave 5. It also indicates that this lifting commences long before the arrival of the low-level atmospheric river along the California coast, as indicated in the coincident moistening at OAK and REV in Figs. 3 and 4 extending as high as 750 hPa in OAK and 550 hPa in REV during 00/31–12/1 as wave 5 arrives onshore. This region in between OAK and REV is encompassed by a sloping isentropic surface centered on ~650 hPa. Two distinct midlevel moisture surges accompany the two distinct midlevel wind surges with out-of-phase along-river gradients created above the low-level river described by Ralph et al. (2004). These midlevel features occur as ageostrophic winds develop southeast of the convectively induced thermal ridges embedded within the exit region of the jet streak accompanying wave 5.

c. 2005 case study

1) OBSERVED MOISTURE FLUX STATISTICS SUPPORTING DUAL ATMOSPHERIC RIVERS AT REV

Fortuitously, the OAK balloon data were available at asynoptic times for 31 December 2005 with soundings at 0300, 0600, and 0900 UTC (shown in Fig. 3). The availability of asynoptic soundings reveal that between 0000 and 0600 UTC above the coastal mountains near OAK, a secondary midlevel wind maximum exists centered on ~305 K between 750 and 650 hPa. The stronger winds exemplified by this maximum transport a layer of
3–5 kg kg$^{-1}$ $\times 10^3$ of moisture up to and above 650 hPa. This layer observed at OAK is consistent with the strong along-flow gradient of water vapor arriving in the vertical cross section (to be described later in Fig. 12p) over 35°N, 135°W just ahead of the second midlevel wind surge (close to that location in Fig. 12m). The moisture fluxes at OAK at 03/31 are $-13 \times 10^4$ kg s$^{-1}$ centered on 900 hPa and $-8 \times 10^4$ kg s$^{-1}$ centered on 700 hPa just below 305 K. By 06/31 there is little difference between the 900–850- and 700–650-hPa layer, shown in Fig. 3h, with both fluxes $\sim 10 \times 10^4$ kg s$^{-1}$. This indicates that ahead of the second midlevel momentum surge (which is over the Pacific in Fig. 12m) the sounding at OAK indicates a separate midlevel river of magnitude similar to the low-level river. The low-level and midlevel rivers become less distinct by 0900 and 1200 UTC 31 December as the moisture plumes become more vertically coupled in Figs. 3i,j. By 12/31 there is little difference in the fluxes at REV when compared to OAK at 06/31 in the 700–650-hPa layer with greater than $8 \times 10^4$ kg s$^{-1}$ within both layers, again as in the 1997 case study, reflecting the importance of midlevel moisture fluxes passing above OAK and arriving at REV. These observations further support the contention that an elevated moist layer is generated upstream of the Sierra Nevada over the Pacific.

2) OBSERVED PACIFIC SYNOPTIC EVOLUTION OVERVIEW

The 2005 flood event shares many similarities and a few differences when compared to the 1997 event. Like the 1997 event, several cyclone-scale waves—that is, at least five—propagate across the North Pacific, culminating in a long and even more intense polar jet separating tropical from extremely cold Siberian air. Each wave is associated with very strong jet stream momentum, plumes of integrated precipitable water, and extremely cold air. Each wave contains a strong low-level jet transporting warm moist air to the Pacific coast, culminating with moistening of the troposphere over the Sierra Nevada. The polar jet in the 2005 case is much stronger than in the 1997 case study, achieving a wind maximum in excess of 100 m s$^{-1}$ and a length scale spanning most of the Pacific. The moisture plumes are not, however, as strong as in the 1997 case study. Additionally, the tilt of the long wave trough in the western
Pacific is somewhat different from that in the 1997 case study; the jet is oriented more southwest–northeast because the subtropical ridge of high pressure is flatter, accompanying less pronounced tropical storm activity relative to the 1997 case. In the 2005 case, waves 5–3 can be diagnosed in Figs. 5–6 in a sequential manner similar to the 1997 case and the flooding rainfalls do occur only with the final wave in the sequence—that is, wave 5; however, the maximum moisture in the REV sounding is not associated with that wave but with one of the precursor waves. Like the 1997 case study, the heaviest rainfalls are coupled to the jet exit region accompanying the final wave in the sequence, as can be inferred from Fig. 5, with the period of maximum rainfall and arrival of the strongest jet exit region on 31 December 2005. Figure 5 depicts the extraordinarily strong jet core (~100 m s$^{-1}$) and long jet exit region (>3000 km). As was the case with the 1997 flood, the effect of the five cold troughs that transit the North Pacific is to support the strong polar jet with a very strong south–north temperature gradient that ultimately encompasses the entire North Pacific from Asia to North America by 00/31.

Figure 4 shows that the largest precipitable water in the REV sounding does not occur during the heavy rainfall on 31 December 2005. The sequence of precipitable water observations indicates that wave 5’s average values ~15 mm at REV during the period of 12/30–12/31 are less than that accompanying wave 4, which is ~17 mm during the period of 12/27–12/28 as well as wave 1, which is ~15 mm during the period of 00/20–12/22. Although the transport was likely very strong with all five waves, the size and areal coverage over the region of influence of wave 5 exceeded the other four waves during its tenure over the Sierra Nevada. This likely facilitated the lifting that resulted in the heavy rainfall with wave 5.

3) OBSERVED PACIFIC BACKWARD TRAJECTORY ANALYSES

When compared to the 1997 flood event, back trajectories calculated from REV during the period of heavy rainfall encompassing 00–12/31 have a longer cross-Pacific fetch at lower pressure levels relative to midtropospheric levels, as shown in Fig. 10a. This indicates that the larger and more powerful polar jet in this case study likely allows the confluent forcing, analogous to that occurring near 35°N, 140–130°W for the 1997 case, to occur somewhat farther upstream in the 2005 case. This reflects a stronger and longer pattern of linkage between low-level jet forcing and pan-Pacific transport in the 2005 case relative to the 1997 case. This can be seen in Fig. 10b, in which the SSM/I water vapor shows a more coherent single plume of water vapor than the 1997 case study exhibited. However, this SSM/I imagery, when examined at higher resolution in Fig. 10b, still reveals additional feeder bands or conduits of moisture to the southeast of the main plume on 31 December. The parcel vertical pressure changes indicate that the strongest lifting and most dramatic confluence of parcel trajectories occurs near 35°N, 150°–140°W ~18–24 h before reaching the 600–800-hPa layer over REV around 06/31. This is consistent with stronger zonal kinetic energy in this case. The midlevel lifting is even more pronounced in Fig. 10a than in the 1997 case study (Fig. 7a), with parcels arriving above 700 hPa at REV on 00/1 emanating from below 900 hPa on 00/31. This stronger lifting and longer range transport is consistent with the stronger jet forcing, weaker tropical storm activity, and different tilt of the Rossby wave in the 2005 flood event. Like the 1997 event, however, parcel transport analyses indicate extreme turning and lifting farther west near 30°–35°N, 150°W north-northeast of the Hawaiian Islands.

4) OBSERVED PACIFIC SUBSYNOPTIC DYNAMICS ACCOMPANYING THE MIDLEVEL ATMOSPHERIC RIVER

In a manner quite similar to the 1997 flood case study, we will focus on the final wave’s (wave 5) convectively forced features at the subsynoptic scale ~30 h and ~2500 km upstream of REV leading up to the heavy rainfall event. Widespread convection develops within the left side of wave 5, as can be seen in Fig. 11’s TRMM imagery; 500-hPa ascent at this time extends to 150°W, where the TRMM data in Fig. 11 shows an MCS along the leading edge of the low-level moisture plume at 925 hPa and just south of the surface cold front near 35–40°N, 150°W. The powerful (~100 m s$^{-1}$) jet streak is associated with this ascent by 12/28, centered in between 180° and 160°W just south of 37.5°N. By 06/30–12/30, as in the 1997 case study, the 925-hPa low-level return branch circulation, which covers the huge region between 180° and 150°W and south of 35°N, produces low-level convergence under diverging flow in the left exit region at 300 hPa (Figs. 12a–d).

By 12/30 in Fig. 12b, the 500-hPa ascent extends back well to the southwest of the Pacific coast to 35–40°N, 150°W. The strongest low-level return branch circulation is near 35°N, 155°W, as evident from the 925 hPa 25 m s$^{-1}$ wind barb crossing the 750-m isohelght. The 925-hPa moisture plume is advancing into this region ahead of the surface cold front, as noted by the arrow in Fig. 12d, whereas TRMM indicates convection near 37.5°N, 145°W in Fig. 11. This region of low-level moisture transport is located under the eastward-extending/redeveloping jet streak’s left exit region accompanying wave 5 between 155–135°W and 35°–40°N in Fig. 12a.
FIG. 9. NARR (a) 300-hPa height (m), wind barbs (pennant = 25 m s$^{-1}$; long barb = 5 m s$^{-1}$; short barb = 2.5 m s$^{-1}$), and isotachs (shaded; knots); (b) 500-hPa omegas (shaded in Pa s$^{-1}$ with a line segment centered on key ascent); (c) 925-hPa height (m), wind barbs [as in (a)], and isotachs (shaded; knots); (d) 925-hPa mixing ratio (kg kg$^{-1}$ with an arrow oriented along the key moist tongue) all valid at 0000 UTC 31 Dec 1996; (e),(f) 305-K wind barbs (long barb = 10 m s$^{-1}$; short barb = 5 m s$^{-1}$) and isotachs (shaded; m s$^{-1}$); (g),(h) 305-K ageostrophic wind barbs [as in (e)] and 305-K pressure (solid in hPa); (i),(j) NARR 305-K mixing ratio (solid; kg kg$^{-1}$ × 10$^3$); (k),(l) vertical mixing ratio (solid; kg kg$^{-1}$ × 10$^3$) cross section from 30$^\circ$N, 140$^\circ$W—REV valid at 0000 and 1200 UTC 31 Dec 1996. Thick dotted–dashed lines in (e)–(l) represent the locations of midlevel wind surges and supporting features.
Figures 12e–p depict similar fields on the 305 K surface as in the 1997 flood case study in Fig. 9. In a somewhat similar manner to the 1997 flood case study, there is a nexus among MCS formation at the tip of the 925-hPa ageostrophic south-southwesterly low-level jet, the 300 hPa diffuent flow in the exit region of the final jet streak, and the 500 hPa ascending omega field all near 37.5°N, 145°W at 12/30 (notice Figs. 11 and 12a–d). Embedded within this region are analogous features to the 1997 case study, including subsynoptic jets on 305 K extending to the southeast of the MCS ahead of the surface cold front shown in Figs. 12e,f,m. These jets accompany southerly ageostrophic wind maxima and subsynoptic pressure perturbations in Figs. 12g,h,n as well as moisture perturbations, discontinuities or fronts oriented in an along-stream manner in Figs. 12i–l,o,p. Additionally, these figures show the accompanying along-flow moisture gradients extending up to 400 hPa between ~35°N,
145°W and REV—that is, developing near the confluence zone of backward trajectories (Fig. 10a) and then building downstream toward REV.

As can be seen/inferred in Fig. 12, a secondary wind maximum on 305 K develops before 00/30 near 35°N, 145°W and propagates to 1) 37.5°N, 140°W at 00/30 to near 2) 37.5°N, 137.5°W at 12/30; 3) 40°N, 135°W at 18/30; 4) 40°N, 125°W at 00/31; 5) 40°N, 120°W at 06/31; and 6) over Reno by 12/31, consistent with the midlevel jet in Fig. 3c. By 18/30, a second wind maximum develops near 35°N, 132.5°W and moves toward Reno. Each wind maximum in Figs. 12e,f,m is supported by a subsynoptic pressure ridge in Figs. 12g,h,n in proximity to ongoing convection ahead of the polar jet and embedded within the southeastward-propagating jet exit region aloft. The wind maximum develops coincident with northward-directed cross-stream ageostrophic flow from the south that forms on 305 K in proximity to this ridge in Figs. 12g,h,n, as was the case with the 1997 flood case study. Consistent with that case study, each wind maximum is just upstream of an along-flow moisture gradient that steepens with height above 700 hPa to as high as nearly 500 hPa.

As the wind maximum proceeds downstream through the trajectory confluence zone, the lifting with the midlevel jet between 12/30 and 12/31 transports the increasing along-flow moisture gradient up to and over the Sierra Nevada (Figs. 3c and 10a). Two distinct pulses of moisture accompany two wind maxima on 305K during this period in the Figs. 12k,l,p moisture vertical cross sections at 0000, 1200, and 1800 UTC 30 December. The midlevel along-flow plume is alternately in and out of phase and to the southeast of the low-level moisture plume. As was true with the 1997 case study, the upslope surge of moisture at 12/31 over the windward side of the Sierra Nevada is more in phase with the midlevel plume accompanying the second along-flow moisture gradient on and above 305K in Fig. 12p at 1800 UTC over ~35°N, 130°W. This allows significant moisture to extend downstream to REV, as is observed in the rawinsonde depicted in Fig. 3c.

d. Summary of observations

Midlevel moisture plumes arrive within the jet exit region in the final of a sequence of cyclone-scale waves that traverse the North Pacific over several days. Within the jet exit region, recurrent MCSs develop that produce subsynoptic-scale ridge perturbations, creating highly accelerative and finer-scale midlevel wind maxima. The lifting of moist plumes by the convection allows moisture to be injected into the horizontal midlevel wind maxima near 305 K. The along-flow moisture plumes arrive over the Sierra Nevada often out of phase—that is, well southeast of stronger low-level plumes that are embedded within the LLJ. The elevated moisture plumes are
associated with rawinsonde-derived maxima of horizontal moisture fluxes that—although less than the low-level moisture flux maxima in magnitude—still account for substantial increases in column precipitable water, particularly at REV and high above the Sierra Nevada, thus increasing the spillover moisture by ~40%–60% relative to earlier observed values.

4. Model experiment design and simulation results
   a. Numerical experiment overview

   Numerical simulations were employed in an effort to increase the temporal resolution of datasets, in particular for trajectory calculations, as well as to test the importance of convective forcing in midlevel river formation. The numerical modeling system employed was the Operational Multiscale Environment Model with Grid Adaptivity (OMEGA) developed by Science Applications International Corporation (e.g., Bacon et al. 2000, 2003; Boybeyi et al. 2001; Gopalakrishnan et al. 2002; Boybeyi et al. 2007; Marzette et al. 2007). OMEGA is a multiscale nonhydrostatic atmospheric simulation model with an adaptive grid that permits a spatial resolution ranging from roughly 100 km to less than 1 km without the need for nested grids (see Table 1 for details).

   Both the 1997 and 2005 simulations were initialized with the NCEP–National Center for Atmospheric Research (NCAR) reanalysis dataset (Kalnay et al. 1996). The 1997 case study was integrated for 36 h, starting prior to the heavy rain events—that is, on 0000 UTC 31 December 1996 through 1200 UTC 1 January 1997. The 2005 simulation was integrated for 42 h, starting at 1800 UTC 29 December 2005 through 1200 UTC 31 December 2005. Both simulations were run in a statically adaptive grid mode to increase horizontal resolution over complex terrain. The region of integration spanned the east Pacific Ocean, California, and Nevada from 180° to 115°W and from 0° to 60°N with a horizontal resolution ~30 km over the Pacific, statically adaptive higher resolution over complex terrain, and a variable vertical resolution of 72 sigma levels, putting the top level well into the stratosphere. Vertical resolution was gradually reduced from the lower to upper troposphere.

   Simulation sensitivity studies are performed in this section, where in a no-latent-heating simulation (NLH) the convective parameterization and microphysical schemes were turned off for comparison with a full-physics simulation (LH). In the NLH, the model simulations are not allowed to produce clouds and precipitation. Although turning these processes off affects multiple variables—for example, radiation exchange and transfer (both longwave and shortwave) and latent heat release—the interest of this section is on the first order effects—that is, primarily on latent heat release (e.g., Uccellini et al. 1987; Kaplan et al. 1998; Hamilton et al. 1998). The simulation sensitivity studies focused on the specific influence on the back trajectories by latent heat release in convective regions that developed along the cold front, in particular how latent heat release may lift moist parcels into the midtroposphere as well as accelerate the parcels downstream over the Sierra Nevada. Therefore, a comparison was carried out between the control, full physics (LH) simulation for both case studies, and the experiment simulation (NLH), with condensation and evaporation turned off to diagnose the role of latent heat release for both case studies.

   b. 1997 and 2005 case study comparison of NLH and LH experiments

   The results of the 1997 and 2005 sensitivity experiments comparing the role of latent heating in parcel dynamics and midlevel river formation are rather similar. Figure 13 depicts the LH and NLH backward trajectory parcel elevation comparison for both case studies. The timing for plotting each parcel height comparison is every 1 h or 6 times the frequency of the observationally generated trajectories. Additionally (not shown), vertical moisture flux maxima were calculated, replacing horizontal wind speed with vertical wind speed in the horizontal flux formulation described in section 3a. The results show that the final nine hours of the backward trajectories indicate elevation increases between 2000 and 2500 m greater for the LH versus the NLH simulations or ~150–200-hPa increases in approximately half the time that the observed parcels increased ~125 hPa. This indicates how convection can boost the parcel location by increasing the outflow and dramatically lifting accompanying midlevel jet flows. Consistent with this are ~20%–40% increases in the calculated horizontal as well as 100% increases in the vertical moisture flux maxima in the LH versus the NLH model simulations.

   The parcels rise because of the acceleration of the midlevel jet as a result of latent heat release. This has been alluded to earlier in this paper and also analyzed in Kaplan et al. (1998) and Hamilton et al. (1998). Figures 14a–d depict comparisons of LH/NLH simulated back trajectories versus 700-hPa winds approximately nine hours before 12/1 and 12/31 for the 1997 and 2005 case studies, respectively. This is near 35°N, 130°W, approximately halfway between 1) the 30°N, 145°W backward parcel trajectory confluence location in Figs. 7a and 10a back ~24 h and 2) REV at backward parcel trajectory initiation time. This 9-h period before parcel initiation in the midtroposphere over REV is the period when the simulated parcels start to develop their strongest upward velocity component. This period represents
stronger parcel lifting than in the observations, albeit in rather similar locations and times as a result of diabatic heating rates that are somewhat larger at finer scales than the observations. The parcel ascent occurs ahead of a midlevel secondary (700 hPa) zonal wind maximum. This secondary maximum is a subsynoptic maximum in the LH simulation southeast of the larger-scale or primary 700-hPa wind maximum, which is not in the NLH simulation southeast of the larger-scale 700-hPa wind maximum.

The LH simulation depicts similar features to the observations, with stronger vertical motions and horizontal accelerations. For example, the simulation shows the zonal wind maximum increase of 5–10 m s$^{-1}$ between
140° and 132°W relative to the NLH. The increasing midlevel jet that results from the convective heating forces a confluence of parcels relatively upstream and diffuence of parcels relatively downstream from the convection by 00/31 and 00/1 in the two case studies, respectively. These mass adjustments accompanying the convergence and divergence of the wind result in a midlevel ascent zone near 35°N, 130°W and ~4 km MSL (not shown). This is unlike the NLH simulation where the midlevel lifting and rapid increase in midlevel zonal wind are missing. The midlevel jet, which is described in the observations, is replicated in the simulations downstream from the model-simulated convection between 140° and 135°W and just north of 30°N in both case studies (Fig. 14). This region is within the southeastward flank of the exit region of each final wave—that is, wave 5 for 1997 and for 2005; however, the convective heating maxima is more directly implicated in producing the strongest diffluent wind and velocity divergence than is the larger quasigeostrophic flow within the jet exit region. The polar jet flow maximum is at a higher level than 700 hPa, as is evident in both the LH and NLH simulations. Eulerian Rossby number calculations (not shown) indicate midtropospheric values approaching or exceeding 1.0 in the LH simulations versus much weaker values in the NLH simulations.
indicating strong oscillations from confluent to diffluent flow not as apparent at higher levels within the polar jet exit region (e.g., Uccellini et al. 1987). Such highly accelerative flow is consistent with the diabatic isallobaric term on isentropic surfaces, in which subsynoptic wind fluctuations are forced by latent heat release (e.g., Wolf and Johnson 1995).

This midlevel accelerating flow due to diabatically induced ridge building is also critical for creating the midlevel moisture surges seen earlier in the observed cross sections and much more evident in the vertical cross sections in both case studies from the LH as opposed to the NLH simulations. As one approaches 12/1 and 12/31 in the two case studies, the midlevel along-flow moisture

<table>
<thead>
<tr>
<th>TABLE 1. OMEGA model overview. Table of model physics used in OMEGA (Bacon et al. 2000).</th>
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<tr>
<td><strong>An overview of OMEGA</strong></td>
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<tr>
<td>Governing equations</td>
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<td>Grid structure</td>
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<td>Grid adaptivity</td>
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<td>Cumulus parameterization</td>
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<td>Upper boundary</td>
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<td>Lateral boundaries</td>
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<tr>
<td>Initialization</td>
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<td>Dispersion</td>
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gradient fields (Figs. 15a,b) indicate stronger along-flow gradients that are lifted up to the Sierra Nevada in the LH simulations but not lifted in the NLH simulations (not shown). This reflects the upstream transition, described in the previous paragraph, from confluent to diffluent flow accompanying the convectively forced midlevel wind maximum. As air converges, accompanying the increase in zonal kinetic energy, downstream diffluence creates the upward vertical motions that allow moisture within the LLJ to be lifted into the midlevels, supporting the midlevel river fluxes evident in the observations traversing OAK to REV at this time; that is, 3–6 h before the heavier spillover precipitation in both case studies. The moisture surges in Figs. 15a,b are reflected as warm and moist bulges in the 550–700-hPa layers in the LH soundings at REV in Figs. 15c,d that are similar to the observations in Figs. 3b,j.

c. Summary of simulation experiments

The simulation experiments support the hypothesis that latent heat release significantly modifies the structure of the midlevel airflow, therefore modifying the along-flow moisture fluxes. The upstream diabatic heating perturbation accelerates the midlevel wind field, creating a southeastward-shifted stronger zonal flow at the expense of the meridional flow. By doing so, it imposes a subsynoptic-scale region of upstream confluence and downstream diffluence within the southeastern polar jet exit region, creating a midlevel wind surge, midlevel ascent zone, and secondary moisture flux maxima with distinct along-flow moisture gradients analogous to the observations. The strong upstream zonal response is more evident in simulations than observational analyses. Its absence when latent heating is turned off strongly supports upstream convective forcing as being essential to creating a midlevel moisture flux maximum downstream over the Sierra Nevada and thus contributing to spillover precipitation.

5. Overall summary and conclusions

Figure 16 includes three simple schematic diagrams based on observations and simulation experiments that describe the key sequence of events, creating a midlevel moisture flux maximum during periods of extreme Sierra Nevada precipitation for two case studies that are members.
of the population of 14 storms that comprised the climatology in Part I (i.e., Underwood et al. 2009). They depict a sequence of processes triggered by MCSs embedded within the exit region of a polar jet streak transiting the eastern North Pacific. Quasigeostrophic and diabatically forced subsynoptic (highly ageostrophic) processes focus midlevel moisture fluxes during a 12–24-h period supporting heavy upslope and leeside precipitation.

Several cyclone-scale waves propagate across the Pacific, with the cold surge accompanying each wave, building a broad polar jet exit region. The final wave results in a very strong and expansive lifting region within the left exit region and a low-level return branch circulation (Fig. 16a). As the low-level moisture is transferred from the tropics toward the midlatitudes accompanying this low-level jet, destabilization and column moistening occur and result in large MCSs ~24–36 h before the midlevel river arrives over the Sierra Nevada. The MCSs heat the mesoscale environment, producing a cross- and downstream-directed pressure gradient force ahead of each diabatically generated ridge. Strong increases in zonal kinetic energy results in a mesoscale midlevel jet, strong upstream confluence, and downstream diffuence (Fig. 16b). Parcel trajectory analyses indicate that air parcels become confluent as they converge within and diffuently just ahead of this jet downstream from these MCSs. The diffuence results in significant ascent, which transports moisture from the LLJ up into the midtroposphere and to the southeast of the LLJ. The result is southwesterly–westerly, along-flow, and midtropospheric moisture flux gradients and maxima that arrive just before heavy upslope and leeside Sierra Nevada precipitation (Fig. 16c). The midlevel moisture surges start out misphased relative to the low-level surge; however, they eventually arrive in phase with the LLJ moisture surge and create an extremely favorable environment for heavy upslope and spillover precipitation during these two major Sierra Nevada flood events.

To summarize the findings of this paper, the analyses strongly support the hypothesis that midlevel moisture transport resulting in large part from combined convective and jet streak forcing over the Pacific Ocean

**Fig. 14.** NLH-simulated vs LH-simulated parcel back trajectories for the parcels originating in Fig. 13 at REV (light solid lines with arrows) valid at (a) 0600 UTC 1 Jan 1997 and (b) 0600 UTC 31 Dec 2005 near 35°N, 130°W, with the 700-hPa wind barbs (as in Fig. 9) and isotachs (thick solid lines in m s⁻¹) superimposed on top.
is the critical mechanism providing moisture for spill-over or leeside precipitation during the extreme flood events of 1–2 January 1997 and 31 December 2005 in northwestern Nevada and northeastern California:

1) The nature of these circulation systems is to favor a massive jet exit region forced by a very long and powerful along-stream gradient of kinetic energy. Such forcing can substantially shift the lifting by the low-level return branch within the exit region’s transverse ageostrophic indirect circulation well southeastward from the jet core. This forces parcels rich in tropical moisture to ascend a substantial vertical distance over periods of only 12–24 h.

2) This lifting and its accompanying dynamic destabilization produces mesoscale convective complex systems that further add to the downstream acceleration and lifting of the moist airstreams on isentropic surfaces, so that surges of unusually moist air can rise above the major mountain regions of the West, such as the Sierra Nevada, and be available for both heavy upslope and leeside (spillover) precipitation.

3) Furthermore, this southeastward displacement process tends to separate the moisture transport and lifting process from the strongest anticyclonic shear on the immediate right side of the jet where the isentropic surfaces slope the most because of the gradient of thermal wind, thus separating the lifting process by substantial distances from the regions of inertial and symmetric instability more typical of compact and smaller-scale jet/front systems. Baroclinic forcing by diabatic ageostrophic processes ultimately compensate...
for the spatial shift away from quasigeostrophic baroclinic processes accompanying the thermal wind and symmetric instability.

**Acknowledgments.** This material is based upon work supported by the National Science Foundation under Grant 0447416 as well as UCAR/COMET Grant S06-58387, DOD/Army Grant N61339-04-C-0072, and a DRI/IPA grant. We thank David Bacon and Ananthakrishna Sarma at Science Applications International Corporation (SAIC) for letting us use OMEGA for our research purposes. We would also like to thank Rhett Milne and James Wallman of the Reno National Weather Service Forecast Office for collaborating with us on various aspects of this research. John Abatzoglou of San Jose State University provided the NWS multisensor precipitation analyses. Cassie Hansen and Ben Reagan provided assistance with figure preparation.

**Fig. 16.** Schematics depicting (a) transverse ageostrophic circulation in the jet exit region, (b) midlevel jet formation due to convective heating, and (c) subsequent moisture surges downstream resulting in a secondary midlevel atmospheric river.
APPENDIX

Tabular Description of Tropical Features for the 1997 and 2005 Case Studies

**TABLE A1.** Tropical evolution in 1997. Terms are abbreviated as follows: typhoon (Ty), tropical cyclone (TC), and tropical depression (TD).

<table>
<thead>
<tr>
<th>Date (UTC/day/month/year)</th>
<th>Tropical activity</th>
<th>Type of moisture transport or outflow</th>
</tr>
</thead>
<tbody>
<tr>
<td>1200 UTC 20 Dec 1996</td>
<td>Central Pacific, convection</td>
<td>South-southeasterly low-level flow</td>
</tr>
<tr>
<td></td>
<td>Western Pacific, two tropical disturbances: one developing, one weakening</td>
<td>Weak convective outflow</td>
</tr>
<tr>
<td>0000 UTC 22 Dec 1996</td>
<td>Eastern Pacific, convection</td>
<td>Weak southerly low-level flow</td>
</tr>
<tr>
<td>1200 UTC 24 Dec 1996</td>
<td>Eastern Pacific, convection</td>
<td>Outflow from TS Fern</td>
</tr>
<tr>
<td></td>
<td>Western Pacific, two TS: Greg (upstream) and Fern</td>
<td>Weak southerly low-level flow</td>
</tr>
<tr>
<td>0000 UTC 27 Dec 1996</td>
<td>Central/eastern Pacific, convection</td>
<td>South-southwesterly low-level flow</td>
</tr>
<tr>
<td></td>
<td>Western Pacific, Ty Fern, TS Greg dissipated</td>
<td>Outflow from TC Fern</td>
</tr>
<tr>
<td>0000 UTC 29 Dec 1996</td>
<td>Central/eastern Pacific, convection</td>
<td>South-southwesterly low-level flow</td>
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<tr>
<td></td>
<td>Western Pacific, Ty Fern weakening</td>
<td>Southerly low-level flow east of Ty Fern, outflow from Ty Fern</td>
</tr>
<tr>
<td>0000 UTC 31 Dec 1996</td>
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<td>South-southwesterly low-level flow</td>
</tr>
<tr>
<td></td>
<td>Western Pacific, TD Fern.</td>
<td>Southerly low-level flow east of TD Fern, weak outflow</td>
</tr>
<tr>
<td>0000 UTC 1 Jan 1997</td>
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**TABLE A2.** Tropical evolution in 2005.

<table>
<thead>
<tr>
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<th>Outflow development</th>
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</thead>
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<td>0000 UTC 23 Dec 2005</td>
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<td></td>
<td>Bay of Bengal, convection</td>
<td>South-southwesterly low-level flow, outflow</td>
<td>Divergence</td>
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<tr>
<td>1200 UTC 24 Dec 2005</td>
<td>Western/central Pacific, convection</td>
<td>South-southwesterly low-level flow, outflow</td>
<td>Inertial instability divergence</td>
</tr>
<tr>
<td>0000 UTC 26 Dec 2005</td>
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<td>Inertial instability divergence</td>
</tr>
<tr>
<td></td>
<td>Western/central Pacific, convection</td>
<td>South-southwesterly low-level flow, outflow</td>
<td>Inertial instability divergence</td>
</tr>
<tr>
<td></td>
<td>Southeast Asia, convection</td>
<td>South-southwesterly low-level flow, outflow</td>
<td>Inertial instability divergence</td>
</tr>
<tr>
<td>1200 UTC 27 Dec 2005</td>
<td>Eastern Pacific, convection</td>
<td>South-southwesterly low-level flow, outflow</td>
<td>Inertial instability divergence</td>
</tr>
<tr>
<td></td>
<td>Central Pacific, convection</td>
<td>South-southwesterly low-level flow, outflow</td>
<td>Inertial instability divergence</td>
</tr>
<tr>
<td></td>
<td>Western Pacific, Southeast Asia, convection</td>
<td>South-southwesterly low-level flow, outflow</td>
<td>Inertial instability divergence</td>
</tr>
<tr>
<td>1200 UTC 28 Dec 2005</td>
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<td>South-southwesterly low-level flow, outflow</td>
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<tr>
<td></td>
<td>Central Pacific, convection</td>
<td>South-southwesterly low-level flow, outflow</td>
<td>Inertial instability divergence</td>
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<tr>
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<td>Western Pacific, Southeast Asia, convection</td>
<td>South-southwesterly low-level flow, outflow</td>
<td>Inertial instability divergence</td>
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<td>Central/eastern Pacific, convection</td>
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REFERENCES


