The Diurnal Variation of Precipitation in California and Nevada

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ABSTRACT

The diurnal variation of precipitation across California and Nevada has been studied by means of a harmonic analysis of 35 years of hourly precipitation data for 347 stations, and a regional probability of precipitation analysis for grouped stations. Results are shown for the cool (November-April) and warm (May-October) seasons.

For all measurable (>0.25 mm) precipitation events, the phase of the cool season diurnal cycle tends to peak between 0300-0900 LST along the coast, between 1000-1300 LST in the coastal mountains and the Sierra Nevada and in the Sacramento and San Joaquin valleys and between 0800-1200 LST in most of Nevada. Similar times are found during the warm season along the coast and in the coastal mountains while there is a shift to 1400-2000 LST inland across the Sacramento and San Joaquin valleys, the Sierra Nevada and Nevada.

For the heavier (>2.5 mm) precipitation events, the phase of the cool season diurnal cycle tends to peak from 1300-1800 LST at many locations. Exceptions include the eastern Sierra Nevada and Nevada, where a maximum near 1000 LST is found and along the coast and coastal mountains where a predawn maximum occurs. The transition from a late morning precipitation maximum along the coast and in the Sacramento Valley to an early evening maximum in the San Joaquin Valley is well defined in the gap in the coastal mountains through the San Francisco Bay area.

During the warm season, precipitation tends to maximize between 1400-2200 LST at most interior locations and in portions of the coastal mountains in phase with the diurnal heating cycle. Coastal areas from Santa Barbara to San Diego retain a 0300-0700 LST precipitation maximum for the lighter precipitation amounts, characteristic of the well-known coastal summer stratus regime.

1. Introduction

The purpose of this paper is to document the diurnal variability of precipitation in California and Nevada from an analysis of the available hourly precipitation data. Topographic forcing (see Fig. 1) plays a major role in the well-known annual distribution of precipitation across the region. Briefly, annual totals vary from under 200 mm in the drier, southern portion of the region to well in excess of 1 m at more northern mountainous California locations well exposed to Pacific airstreams. A well-defined winter precipitation maximum is noted almost everywhere in California. In the much drier Nevada region and California desert area a second precipitation maximum is found in summer when convective activity may be present.

Recent interest in precipitation variability analysis was sparked by Wallace's (1975) comprehensive analysis of the diurnal variability of precipitation and thunderstorm frequency across the conterminous United States, based on a 10-year database for a representative sample of stations. Englehart and Douglas (1985) and Easterling and Robinson (1985) have updated this work on a national basis. Wallace's (1975) work stimulated us to undertake a number of regional studies of diurnal precipitation variability in order to unmask possible local influences on the rainfall distribution. Schwartz and Bosart (1979), Landin and Bosart (1985) and Riley et al. (1987) examined the diurnal rainfall characteristics across Florida, the northeastern United States and the Central Rockies and adjacent Great Plains, respectively. The influence of large water bodies and varied terrain on the diurnal distribution of convective and stratiform precipitation as a function of the time of the year was clearly revealed in these analyses.

The complex physiography of the western United States provides researchers rich opportunities to study the seasonal and diurnal variability of regional weather regimes. To cite only a few examples, Mass (1982), Mass et al. (1986) and Mass and Albright (1987) have looked at the effect of topography on the diurnal variability of precipitation over western Washington and the topographic and environmental control of the onshore push of marine air during the warm season. Tang and Reiter (1984) compared the influence of the North American and Tibetan plateaus on the monsoon circulations that so strongly influence the climate in these regions. Our interest in cool season precipitation vari-

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Fig. 1. Smoothed terrain map for California and Nevada. Height contours are drawn for 300 (dashed) and 1500 (solid) meters with elevations greater than 1500 meters stippled. Locations of key places mentioned in the text are shown.

atation as well leads us to concentrate on the California-Nevada area, given the marked winter rainfall contrasts across the region.

The analysis method is similar to that used by Wallace (1975) and Horn and Bryson (1960) with the details given in Landin and Bosart (1985). We will present cool and warm season results for the first harmonic of the diurnal rainfall distribution for selected precipitation amount categories, followed by a more detailed discussion of the monthly diurnal rainfall characteristics for homogeneously grouped stations. We conclude with a physical perspective on the cool and warm season results.

2. Data processing

The data used in this study consists of hourly precipitation measurements recorded throughout California and Nevada, and covering the approximate period 1948–83. Over the 35-year period of record a number of stations switched from the standard tipping bucket
raingage (sensitive to the nearest 0.25 mm) to Fisher–Porter gages, which record rainfall to the nearest 2.5 mm. Overall a total of 347 stations were available for analysis. Of these 347 stations, 165 reported all measurable precipitation events. A total of 134 stations converted to Fisher–Porter recorders (primarily before 1970) during the period of record. Each of these stations had 10 or more years of reporting all measurable events and they were incorporated into the >0.25 mm precipitation amount analyses. A total of 48 stations reported Fisher–Porter data exclusively. Consequently, a total of 299 stations were analyzed for all measurable precipitation events and 347 stations were available for analysis of the heavier (>2.5 mm) precipitation amounts. The use of the Fisher–Porter gages introduces a timing uncertainty because the gage records only after accumulating 2.5 mm. Landin and Bosart (1985) chose not to use Fisher–Porter data in their analysis of northeastern United States precipitation because of the relatively high frequency of lighter rainfalls in that region. On the other hand Riley et al. (1987) were able to use the Fisher–Porter data because of the strong convective signal in the warm precipitation maximum across the Central Rockies and adjacent Great Plains. Here the use of Fisher–Porter data is restricted to the >2.5 mm rainfall categories.

The number of occurrences of precipitation for each month during each hour were calculated from the categories of >0.25 mm (measurable), >1.0 mm, >2.5 mm, >5.0 mm and >10.0 mm for stations whose locations are shown in Fig. 2. A harmonic analysis was performed on the data to obtain amplitudes and phases of the diurnal (1st harmonic) and semidiurnal (2nd harmonic) cycles for each category. The corresponding amplitudes were then normalized by dividing them by the 24-hour mean of the parameter in question. Only results for the diurnal variation of the >0.25 mm and 2.5 mm categories will be presented here.

The data were grouped seasonally as follows: (i) winter (November–April), and (ii) summer (May–October). This procedure combined with the long record length of the data gave a representative view of the rainfall variations across the region. A station record of at least 10 years was required in order for it to be included in the analysis. The same significance level as used by Wallace (1975), i.e., at least 1% of the hourly observations had to report measurable precipitation, was used in this study. Precipitation amount categories of >0.25 mm and >2.5 mm are described for the monthly analysis of homogeneously grouped stations. These thresholds allow for an assessment of the modulation of stratiform and convective precipitation by the diurnal cycle.

3. Results

a. Harmonic analysis of the diurnal cycle

Figures 3–6 depict the cool and warm season normalized amplitude and phase of the diurnal cycle (first harmonic) in the frequency of precipitation events across the 0.25 mm (measurable) and 2.5 mm thresholds for California and Nevada. The prevailing winter pattern for all measurable (>0.25 mm; recall that no Fisher–Porter gage data is included in this grouping) events shown in Fig. 3 is a weak predawn (0400–0600 LST) maximum along the immediate coastal strip (exceptions from Monterey south to San Luis Obispo) that becomes an equally weak late morning maximum (1000–1100 LST) in the coastal mountains north of Los Angeles. A transition from a predawn to midmorning maximum occurs from the San Francisco Bay area eastward through the gap in the coastal mountains. A continuation of the pattern of a weak morning maximum is seen across much of the remainder of California, including the Sacramento and San Joaquin valleys, and Nevada. Notable exceptions are found along the eastern slopes of the southern Sierra Nevada and in some of the mountains east of Los Angeles where a return to a predawn maximum is found. While there is no distinct pattern in the strength of the normalized amplitude of the first harmonic, there is some suggestion of a slightly stronger signal across the coastal mountains, eastern slopes of the Sierra Nevada and in much of northern and central Nevada where a midto late-morning maximum is observed.

Subtle changes take place across the 2.5 mm threshold shown in Fig. 4. The Fisher–Porter stations included in these analyses did not lead to any significant increase in noise based upon a subjective determination from the raw plots. An occasional data point, however, may either be noise or a reflection of a local topographical feature. The signal along the immediate coastal strip is more chaotic as the normalized amplitudes of the first harmonic are reduced and the phase becomes more variable. There is still a slight tendency for a predawn maximum along the coast north of San Francisco and south of Santa Barbara. Farther inland the precipitation maximum in the north end of the Sacramento Valley shifts to early afternoon (1200–1300 LST). A clear transition from an early morning to midafternoon precipitation maximum occurs from west to east across the San Francisco area into the western end of the interior valley. This precipitation maximum shifts later in time southward through the San Joaquin Valley. To the east of the San Francisco Bay area the precipitation maximum occurs from early to midafternoon (1400–1600 LST), becoming early evening (1800–1900 LST) just to the north of Modesto. This early evening maximum continues into the mountains at the south end of the San Joaquin Valley.

Many stations in the vicinity of the Sierra Nevada lack a distinct signal but a few stations along the western (eastern) slopes of the mountains observe an early evening (late morning) precipitation maximum. A number of mountain stations east of the Los Angeles–San Diego corridor report a shift to a weak early to late afternoon precipitation maximum for the >2.5 mm events but this pattern is not common elsewhere.
Across much of Nevada and the southeastern California desert region where hourly rainfalls > 2.5 mm are comparatively rare, the phase signal is more chaotic (and may be contaminated by noise) with some stations observing a near midnight precipitation maximum in contrast to late afternoon at other stations.

For the cool season normalized amplitude and phase results for hourly precipitation events > 10.0 mm (not shown), there is the suggestion of a late afternoon to late evening (1600-2200 LST) precipitation maximum in an otherwise noisy pattern across much of California. This is particularly true in much of the Sacramento and Joaquin valleys, the western foothills of the Sierra Nevada and the coastal strip from San Francisco to Santa Maria. A near midnight to near dawn precipitation maximum is observed at a number of locations in the Sierra Nevada and the coastal mountains south of San Francisco, particularly from north of Santa Barbara to east of San Diego and along the immediate coastal strip south of Santa Barbara. The only area...
with a persistent late morning precipitation maximum for the heaviest events is found in the coastal mountains north of San Francisco to the Oregon border.

Figures 5 and 6 show the pattern of warm season diurnal precipitation variation for all measurable and >2.5 mm hourly categories respectively. The warm season patterns differ considerably from the winter results shown in Figs. 3–4. A distinct early to late afternoon rainfall maximum is found for all measurable events across much of Nevada, the Sierra Nevada and the Sacramento Valley. At several stations along the eastern slopes of the Sierra Nevada an early evening maximum (2000–2200 LST) is found, suggestive of a transition from late afternoon to early evening from west to east across the Sierra Nevada.

A predawn to sunrise precipitation maximum
(0300–0600 LST) is found along most of the immediate coastal strip, in the San Joaquin Valley, in portions of the western foothills in the southern Sierra Nevada and in the mountainous areas east of Los Angeles and San Diego. The late morning to near noon (0900–1200 noon LST) precipitation maximum encompasses much of the coastal mountains north of Santa Barbara and reaches the immediate coast from Monterey to San Luis Obispo. Comparison of Figs. 5 and 6 reveals that during the warm season there is a progressive shift toward a late afternoon and early evening precipitation maximum for much of Nevada, the Sierra Nevada, the Sacramento and San Joaquin valleys and the coastal mountains north of Santa Barbara for the >2.5 mm hourly precipitation category.

Along the immediate coast south of Santa Barbara and near San Francisco a predawn to midmorning precipitation maximum is retained for this category. At a number of mountain locations east of Los Angeles, San Diego and Monterey, and to the north of San
Francisco, the transition to a noon to early afternoon maximum is apparent. For the 10.0 mm threshold (not shown) there are very few events, particularly in coastal California. At interior mountain locations the noisy signal is dominated by a prominent 1600–2000 LST precipitation maximum.

b. Regional probability of precipitation analysis

On the basis of the results shown in Figs. 3–6, regional groupings of stations were subjectively prepared to show the probability of precipitation (POP) across the 0.25 mm and 2.5 mm thresholds as a function of the time of day. The groupings are outlined in Fig. 2. Other groupings are obviously possible. Our strategy was to group the stations with an eye toward topographical homogeneity and a rough balance of the number of stations among the different groups. Figures 7–10 show results for the different groupings. These plots complement the harmonic analyses of the diurnal cycle presented earlier.
The Coastal Zone POP patterns (Fig. 7a) illustrate the prominent seasonal variation between a wet cool season and a dry warm season. During the cool season there is a broad precipitation frequency maximum for all measurable events from 0200–0900 LST. This maximum, while less prominent for the 2.5 mm precipitation threshold, is still evident during the cool season. A similar signal is found during the warm season.

The Coastal Mountain area (Fig. 7b) is similar to the Coastal Zone (Fig. 7a) on a seasonal basis, but with the well-known higher POPs characteristic of this wet region. The diurnal maximum for the measurable precipitation category broadens to cover the period from near sunrise to early afternoon, suggestive of several physical processes at work producing the rainfall maximum. Indeed, for the 2.5 mm threshold there is evidence for weak, twin maximum centered near 0600 and 1700 LST. This diurnal pattern continues into the warm season with a slightly greater emphasis on the afternoon maximum for the 2.5 mm category.
East of the Coastal Mountain zone there is a decrease in the POP frequency through the Sacramento and San Joaquin valleys (Fig. 8a), especially for the heavier 2.5 mm hourly rainfalls. Unlike the Coastal Zone (Fig. 7a), however, a double maximum in the POP frequency during the cool season for all measurable events becomes a broad afternoon maximum for the 2.5 mm POP amounts. The summer pattern reflects both the very dry conditions and near absence of any diurnal signal for all measurable events. A very weak late afternoon and early evening maximum is indicated for the 2.5 mm category, but the number of events going into this analysis is very small.

Precipitation frequency increases once again eastward into the Western Sierra Mountains (Fig. 8b) with the highest measurable POP frequency in excess of 10% found in this region. Otherwise the pattern looks like the Coastal Zone (Fig. 7a) with the exception that the afternoon peak is more prominent in both seasons. A dramatic decrease in the POP frequency, especially in winter, is seen for the Eastern Sierra Mountain Region (Fig. 9a) with the hourly measurable POP frequency near 5% in winter. A cool season midmorning maximum (~1000 LST) is found for both thresholds and there is no evidence that this maximum extends into the afternoon hours. For the homogeneously grouped stations there is little difference across the Sierra Nevada during the warm season except that it is somewhat drier to the east, especially for the 2.5 mm category.

The High Desert grouping (Fig. 9b) resembles the Eastern Sierras (Fig. 9a) region except for a somewhat drier cool season and a slightly wetter warm season. The cool season diurnal precipitation variation features twin POP maxima centered near 0300–0400 and 0900–1000 LST for the measurable precipitation category with a slight late morning maximum for the 2.5 mm category. During the warm season there is a prominent midafternoon POP frequency maximum near 1500 LST for both rainfall categories.

In the Desert Valley (Fig. 10a) we see the expected reduction in POP frequency for both seasons, but particularly during the winter. There is no diurnal varia-
tion in POP frequency during the cool season with a weak warm season maximum for both categories.

Finally, Fig. 10b presents the hourly POP frequency results for the southeastern mountains which encompasses the area east of Los Angeles and San Diego. This area is drier overall than the Coastal Mountains (Fig. 7b), reflecting its more southerly location relative to the mean westerlies. An important difference, however, is the cool season early afternoon POP frequency maximum for the 2.5 mm precipitation category. Otherwise, the warm season POP pattern is well correlated with the other regions removed from the coast.

4. Discussion

Our analysis of the diurnal variability of precipitation across California and Nevada has established the substantial topographical influence on the rainfall distribution across this region. Superimposed on this orographic signal is a diurnal variation whereby the relative precipitation frequency tends to maximize at night in most areas during the cool season and during the mid- and late afternoon during the warm season. This latter pattern is broadly consistent with the large scale distribution found by Wallace (1975) and Easterling and Robinson (1985).

a. Cool season

California has the typical Mediterranean type of climate which features a prominent wet winter and a dry summer. Superimposed weakly on this regime is the continental pattern of a summer precipitation maximum. This latter feature is most prominent from the Sierra Nevada eastward across Nevada. Figures 1–10 confirm the existence of this dual pattern regime. During the cool season the highest relative precipitation frequency is found along the western margins of the Sierra Nevada and the coastal mountains. These areas are well exposed to moist airstreams from the Pacific. The relative precipitation frequency is somewhat reduced along the immediate coastal strip and further diminishes eastward across the Sacramento and San Joaquin valleys to the eastern slopes of the Sierra Nevada and into Nevada.

Despite the considerable variability in cool season relative precipitation across the region, however, a
broad early to midmorning (0300–0900 LST) relative precipitation maximum for all measurable events can be found. A 0300–0400 LST maximum is most prominent along the coast, across the mountains east of Los Angeles and San Diego and in the high desert area of Nevada. Across the coastal mountains, the Sacramento and San Joaquin valleys and Sierra Nevada, this 0300–0400 LST relative precipitation frequency maximum is masked by a broad maximum that extends from the early morning to early afternoon hours. Its continued presence can be inferred, however, from the increase in relative precipitation frequency after 2100 LST nearly everywhere. A similar pattern was found by Landin and Bosart (1985) and Riley et al. (1987) across the northeastern United States and Central Rockies and adjacent High Plains regions respectively. The existence of a 0300–0400 LST relative precipitation frequency maximum in widely spaced locations suggests a larger scale origin to the phenomenon. While no obvious mechanisms come to mind, it is possible that radiative cooling from cloud tops is helping to augment precipitation through air mass destabilization as discussed by Gray and Jacobson (1977) for deep convective systems in the tropics. Here the impact is mostly on the lighter precipitation amounts with the exception of the southeastern coastal mountains (Fig. 10b) and the coastal mountains (Fig. 7b). In the latter zone there is a modest 0600 LST maximum for precipitation events > 2.5 mm.

From the coastal mountains eastward across the Sacramento and San Joaquin valleys to the western Sierra Nevada (Figs. 7b, 8a and 8b) a more prominent and broad 0600–1500 LST relative precipitation frequency maximum is observed for all measurable precipitation events. This broad maximum is interrupted by a relative minimum near 1100 LST across the Sacramento and San Joaquin valleys (Fig. 8a). Across the eastern Sierra Mountains and High Desert regions (Figs. 9a and 9b) the maximum is sharper and concentrated near 1000 LST. Along the Coastal Zone (Fig. 8a) this same maximum may be apparent as an extension of the 0300–0400 LST maximum to 0700 LST. A similar, but weaker, diurnal signal is also present for the 2.5 mm precipitation category at many of these locations.

Several physical processes appear to be important in explaining this signal. A contributor to the cool season diurnal precipitation distribution across the Sac-
ramento and San Joaquin valleys might be a mesoscale mountain–valley circulation forced by differential diabatic heating between the flanking coastal and Sierra Nevada ranges. The relative precipitation minimum in the valley just before noon is consistent with the existence of a subsiding branch of a mountain–valley circulation. This idea has been used by Mass (1982) and Landin and Bosart (1985) to account for the diurnal precipitation variation in the Puget Sound lowlands and portions of the northeastern United States.

The corresponding nocturnal drainage of cold air from the adjacent mountains into the great interior valley of California may possibly explain the modest 0800 LST maximum seen there in Fig. 8a. This effect is apparent only for all measurable precipitation events which suggests a diurnal modulation of lighter precipitation from stratiform clouds. (Note that these details are not captured in the corresponding normalized amplitude and phase map for the first harmonic shown in Fig. 3. The maps and POP graphs must be used together to help interpret the results.) It is also possible that the broad early morning maximum along the immediate coastal strip (Fig. 7a) is modulated by cold air drainage from the nearby coastal mountains. In this interpretation the nocturnal cold air drainage would help to reinforce horizontal convergence along the immediate coast and, together with the differential roughness across the coastline, help to augment light precipitation from shallow stratiform clouds. Another possibility is that nocturnal cold air drainage from the coastal mountains might be enhanced under an easterly large scale flow regime associated with cyclonic flow offshore and aloft. In such situation the cloud layers would be deeper and there might be a very modest diurnal signal on the heavier (>2.5 mm) precipitation amounts. A weak 0600 LST maximum in this heavier rainfall category along the coast (Fig. 7a) might possibly represent some evidence for this interpretation.

From the coastal mountains eastward across the Sacramento and San Joaquin valleys to the western Sierra Nevada a very modest measurable precipitation maximum is centered near 1500 LST. It is not a prominent peak but its existence can be inferred from the weak relative measurable precipitation frequency minimum centered near 1100 LST in the San Joaquin Valley combined with the slightly more prominent early
to midafternoon precipitation frequency peak for the 2.5 mm hourly precipitation threshold for the adjacent coastal and Sierra Nevada areas. A well-known characteristic of the California cool season climate is the showery nature of post–cold front precipitation associated with the passage of the principal synoptic scale trough aloft and embedded migratory short-wave troughs. Convective elements are frequently observed in such weather regimes with relatively shallow cumulonimbus tops reaching to the 500–400 mb level. The embedded convective elements might be expected to undergo a modest diurnal variation tied to the diurnal heating cycle and orographic uplift. For example, Wallace (1975) showed that winter thunderstorms in southern California strongly cluster around 1600 LST. Some support for this interpretation is provided by a 1300–1400 LST relative precipitation frequency maximum in the >2.5 mm category for the high mountain area east of Los Angeles and San Diego shown in Fig. 10b. Additionally, coastal sea breeze regimes might be expected to augment boundary layer convergence and orographic uplift during the time of maximum afternoon heating inland. Careful inspection of Fig. 4 shows the transition from a predaawn precipitation maximum in the Los Angeles basin to a late afternoon maximum over and to the east of the mountains that ring the city for the heavier rainfall category.

Thunderstorms, heavier convective showers, or a combination would be favored anytime a moist Pacific airstream containing a weakly conditionally unstable layer or layers was forced to ascend the mountain barriers. The effect should be relatively insensitive to the time of day unless it can be demonstrated that the airflow at the top of the boundary layer is stronger at night when the free atmosphere above is relatively decoupled from the airflow in the planetary boundary layer. This line of reasoning has been used by many investigators (e.g., see Wallace 1975; Landin and Bosart 1985) to possibly account for the observed nocturnal precipitation maximum in many parts of the world. The predaawn precipitation maximum along the California coast and adjacent coastal mountains occurs in a region of relatively weak surface diurnal heating. Typically the large scale static stability in this region is relatively small ahead of and accompanying mobile synoptic disturbances aloft (marine air near the surface, cold air aloft) so that the impact of local diurnal circulations might be augmented through enhanced vertical motions. Why this should be manifest as a predaawn precipitation maximum we cannot say beyond what we have discussed above and what has been reported elsewhere in the literature.

Across the Sacramento and San Joaquin valleys and eastward to the western slopes of the Sierra Nevada is a modest rainfall frequency maximum centered near 1500 LST for both precipitation categories. Given that the normal daily diurnal temperature range in this part of California is large compared to coastal locations, it is probable that the afternoon precipitation maximum can be attributed to the diurnal maximum of surface-based convective activity. Convective rainfall is often present along and to the rear of Pacific cold frontal cloud bands as cold air aloft sweeps over air of marine origin below that is subject to the diurnal heating cycle. No evidence in our regionally averaged data supports the assertion that convective elements generated over the interior valley of California later sweep eastward across the mountains, as there is no apparent phase lag in the precipitation frequency maximum between these regions (compare Figs. 8a and 8b).

Over the eastern Sierra Nevada eastward through the Great Basin (Figs. 9a and 9b) there is no evidence for a precipitation maximum tied to the afternoon surface heating maximum in either rainfall category. Instead there is a modest precipitation frequency maximum centered near 1000 LST in both rainfall categories. (Given the horizontal scale of the event, attributing this behavior to a mountain–valley circulation is difficult.) In certain locations, however, air moving upslope after sunrise can possibly trigger the growth of precipitation elements in a favorable synoptic scale environment. Another possibility is that the late morning rainfall maximum can be attributed to the eastward advection of precipitation elements associated with the predaawn maximum along the coast. A third, and possibly more probable, explanation is that the surface-based diurnal heating cycle is sufficient to produce a convective temperature by late morning over these semiarid regions. Typically, there is a surge of low level warm air northward into the Great Basin from the Gulf of California ahead of strong troughs aloft and their associated cold fronts approaching from the west. With very cold air aloft overspreading Nevada, the tropospheric lapse rate can become very steep (static stability ~0) and convective activity can be triggered in the presence of sufficient moisture and synoptic scale ascent ahead of the advancing trough.

The only region without any obvious cool season diurnal precipitation variation is the interior desert valley (Fig. 10a). In this dry region the average measurable hourly precipitation frequency is >1.5% (<0.25% for hourly amounts > 2.5 mm). The individual data (not shown) do reveal a tendency for a predaawn maximum but the signal is so weak as to be barely evident in Fig. 10a.

A word of caution is in order about the interpretation of the higher precipitation events where most of the precipitation falls as snow at the higher elevation stations in the Sierra Nevada during the cool season. Snowfall is notoriously difficult to measure. There is some evidence from the raw observations that the liquid water equivalent of the measured snowfall was determined by dividing the snowfall amount by 10 and then dividing the measurable precipitation equally across all the hours that snow was observed for a number of stations. The problem was noted at both tipping bucket and Fisher–Porter precipitation sites. Such a procedure, if widespread, would act to mask any diurnal signal by
smearing out the diurnal variation across many hours and it may help to explain the relatively noisy signal for the winter > 10.0 mm precipitation category (not shown) in mountainous locations. Unfortunately it was impossible to correct this problem simply (such as throwing away suspicious data) because of the effort involved to search individual station records and make subjective decisions on data suitability.

The horizontal maps for the cool season diurnal precipitation variation shown in Figs. 3–4 also reveal a subtle difference in the timing of the diurnal precipitation maximum for the 0.25 mm, 2.5 mm and 10.0 mm (not shown) rainfall categories between the Sacramento and San Joaquin valleys. The averaging procedure used in the construction of the probability of precipitation composites described previously will mask the difference that leads to a late morning or early afternoon (late afternoon or early evening) precipitation maximum in the interior valley north (south) of the San Francisco Bay area. Our interpretation of the difference in timing of the phase of the maximum precipitation rests on the characteristic airflow in winter through the gap in the coastal mountains across San Francisco Bay.

Ahead of an approaching cold front there will typically be a strong south or southeasterly airflow along both the coast and in the interior valleys. With the passage of the cold front a westerly air stream flows through the San Francisco Bay area into the interior valley. Subsequently the air stream is channeled northward through the Sacramento Valley and southward through the San Joaquin Valley by the configuration of the topography. Pronounced wind shifts from south to north accompany the cold front passage south of Modesto while to the north there is very little interruption in the prevailing flow as low level winds remain southerly both before and after the frontal passage. Such a situation would suggest that the diurnal variation of precipitation in the northern end of the Sacramento Valley and adjacent mountains would exhibit little difference and respond more to larger scale forcing as appears to be the case in Figs. 3–5. Consequently the low level frontal convergence might be expected to be stronger to the south where the frontal wind shift is pronounced.

We will hypothesize that this effect could lead to enhanced convective rainfall with the passage of a cold front when the frontal wind shift is abrupt in the presence of synoptic scale ascent of air of weak static stability, particularly when the cold frontal passage is timed to the diurnal heating cycle. The convective elements, once formed, would propagate northeastward along the front while the front moves southward. As the low level cool air behind the front is channeled southward through the San Joaquin Valley, new convective elements would be able to erupt along the frontal boundary so long as the layer of low level convection remained sufficiently deep in a favorable synoptic environment for ascent. Individual case studies will be needed to test this hypothesis.

A number of investigators (e.g., see Carbone 1982; Parish 1982; Marwitz 1983) have studied the role of the Sierra Nevada in forcing a barrier wind below mountaintop level in winter. The typical barrier wind blows roughly parallel to the mountains from the south-southeast at speed of 15–30 m s⁻¹ with the axis of the low level jet usually below 2000 m. These barrier winds develop with the approach of a Pacific cold front and migratory short-wave trough aloft from the west. A layer of cold air (quasi-isothermal near 0°C) is produced by diabatic cooling associated with the melting of snow along the western slopes of the Sierra Nevada. The net result is an increase in mean sea level pressure by several mb and resulting cold air damming upwind of the mountain barrier. Marwitz (1983) has shown that there is a dynamical feedback process operating between the microphysics of snow melting and the dynamics of orographic airflow whereby the strengthening dome of trapped, low-level cold air adjacent to the barrier causes further ascent and precipitation with additional melting reinforcing the cold air dome. Nocturnal cold air drainage adjacent to the mountain slopes may act to strengthen the shallow cold air dome, which in turn may alter the low-level wind field and modulate the diurnal precipitation signal. Individual case studies are needed to test this hypothesis.

b. Warm season

Precipitation across California and Nevada is infrequent during the warm season. The diurnal precipitation cycle is keyed to the well-known coastal fog and stratus regime and the occasional inland convective episodes associated with the summer monsoon circulation in the southwestern United States. This is typified by a close comparison of Figs. 5, 6, 7a and 9b. At most coastal locations there is a broad measurable precipitation maximum centered between 0300 and 0900 LST. Inspection of Fig. 7a shows that this diurnal characteristic almost disappears for the 2.5 mm probability of precipitation threshold, which supports the above assertion that the morning diurnal maximum reflects light precipitation amounts from coastal fog and stratus decks. Other evidence to support this interpretation is given by Wallace’s (1975) analysis of trace only rainfall events in which he demonstrates a strong sunrise maximum for several California coastal stations.

Comparison of Figs. 5, 6 and 7b also reveals a weak mid- to late morning measurable precipitation maximum in the California coastal mountains which shifts to early and midafternoon across the 2.5 mm rainfall threshold. Figure 7b depicts a broad flat peak from 0800–1600 LST for all measurable events in the coastal mountains with a slight 1500–1600 LST maximum for the 2.5 mm hourly rainfall category. This suggests to us that the following physical processes are combin-
ing to mask the diurnal precipitation cycle during the warm season in the California coastal mountains: (i) light precipitation amounts from stratus clouds shortly after sunrise on days when the marine layer is sufficiently deep to encompass the coastal mountains, (ii) late morning light precipitation amounts associated with an upslope mountain—San Joaquin Valley circulation on days with a deep marine layer, and (iii) infrequent afternoon convective activity and heavier hourly rainfall associated with a moist southerly airflow during active Arizona summer monsoon periods.

Figures 5 and 6 also reveal a time shift from near noon to 1800 LST for the peak rainfall from the western Sierra Nevada foothills eastward across the mountains into extreme western Nevada. The signal is best defined for all measurable precipitation events but it is still present in northern California from the Sacramento Valley eastward into Nevada for the 2.5 mm rainfall threshold. We speculate that the near-noon rainfall events along the western slopes of the Sierra Nevada are probably relatively light precipitation episodes associated with the eastern ascending branch of the vast mountain–interior valley circulation system described previously on days with a deep marine layer in the Sacramento and San Joaquin valleys. The time shift to mid- and late afternoon eastward across the Sierra Nevada is probably a manifestation of the formation of convective elements over the mountains associated with a moist south or southwesterly air stream aloft when the Arizona summer monsoon is active. These convective elements in turn would drift off the mountains into western Nevada leading to a progressively later peak rainfall time into the early evening. Farther east across Nevada there is very little spatial variability in the time (1400–1700 LST) of the afternoon warm season precipitation maximum. Locally generated convective activity in response to the diurnal heating under Arizona summer monsoon conditions is indicated.

5. Conclusions

The diurnal variability of precipitation across California and Nevada has been studied by means of a harmonic analysis of 35 years of hourly precipitation data for 347 stations, and a regional probability of precipitation analysis for grouped stations. The results are shown for the cool (November–April) and warm seasons (May–October) for all measurable and >2.5 mm hourly rainfall categories.

During the cool season there is broad, weak 0300 LST rainfall maximum nearly everywhere that is superposed on a distinct predawn maximum along the immediate coast and coastal mountains, and a broader maximum that extends until midafternoon over the eastern Sierra Nevada. This broad maxima has two components. One appears to be associated with a widespread tendency for a nocturnal precipitation maximum. It also may be associated with nocturnal cold air drainage and land breeze convergence along the coast, enhanced boundary layer flow at night, and air mass destabilization associated with nocturnal radiative cooling over cloud tops. The other component appears to be tied to convective forcing by the diurnal heating cycle inland. Over the Sacramento and San Joaquin valleys the broad maximum is interrupted by a prenoon minimum, probably in association with air descending over the valley in an average sense as part of a broad mountain–valley circulation. Along the eastern slopes of the Sierra Nevada and over the Great Basin an afternoon precipitation maximum is absent. Instead a late morning maximum may be a combination of the eastward drift of predawn rainfall elements to the west and new precipitation cells triggered before the time of maximum diurnal surface heating of mobile short-wave troughs aloft.

The summer regime is characterized by considerably less rainfall overall with a predawn to sunrise maximum along the coast and over the coastal mountains in association with the well-known coastal stratus regime. Elsewhere a modest afternoon maximum represents convective activity tied to the diurnal heating cycle. Over the Sacramento and San Joaquin valleys a summer diurnal rainfall signal is absent.

Two important cool season topographical effects are noted. The first is a shift of the rainfall maximum from late morning to late afternoon eastward through the gap in the coastal mountains east of San Francisco. Likewise, there is a transition from a late morning to a midevening rainfall maximum from north to south through the great interior valley of California. The second is a transition from a predawn to sunrise rainfall maximum in the Los Angeles basin to an afternoon maximum on the northern and eastern slopes of the mountains that ring the basin. Both of these features are likely influenced by land and sea breeze circulations, mountain–valley circulations, topographically channeled flow and orographically induced convective activity.

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REFERENCES

