Value of a Dual-Polarized Gap-Filling Radar in Support of Southern California Post-Fire Debris-Flow Warnings

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ABSTRACT

A portable truck-mounted C-band Doppler weather radar was deployed to observe rainfall over the Station Fire burn area near Los Angeles, California, during the winter of 2009/10 to assist with debris-flow warning decisions. The deployments were a component of a joint NOAA–U.S. Geological Survey (USGS) research effort to improve definition of the rainfall conditions that trigger debris flows from steep topography within recent wildfire burn areas. A procedure was implemented to blend various dual-polarized estimators of precipitation (for radar observations taken below the freezing level) using threshold values for differential reflectivity and specific differential phase shift that improves the accuracy of the rainfall estimates over a specific burn area sited with terrestrial tipping-bucket rain gauges. The portable radar outperformed local Weather Surveillance Radar-1988Doppler (WSR-88D) National Weather Service network radars in detecting rainfall capable of initiating post-fire runoff-generated debris flows. The network radars underestimated hourly precipitation totals by about 50%. Consistent with intensity–duration threshold curves determined from past debris-flow events in burned areas in Southern California, the portable radar-derived rainfall rates exceeded the empirical thresholds over a wider range of storm durations with a higher spatial resolution than local National Weather Service operational radars. Moreover, the truck-mounted C-band radar dual-polarimetric-derived estimates of rainfall intensity provided a better guide to the expected severity of debris-flow events, based on criteria derived from previous events using rain gauge data, than traditional radar-derived rainfall approaches using reflectivity–rainfall relationships for either the portable or operational network WSR-88D radars. Part of the reason for the improvement was due to siting the radar closer to the burn zone than the WSR-88Ds, but use of the dual-polarimetric variables improved the rainfall estimation by ~12% over the use of traditional $Z-R$ relationships.

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1. Introduction

In 2005, the National Oceanic and Atmospheric Administration (NOAA) and the U.S. Geological Survey (USGS) began a demonstration project to deliver timely warnings to the public for debris flows that are generated in response to high rainfall intensities over recent wildfire burn areas (NOAA–USGS Task Force 2005; Restrepo et al. 2008). A component of this effort is collection of data necessary to improve the existing rainfall intensity–duration thresholds that National Weather Service (NWS) forecasters use to identify likely debris-flow conditions from near-real-time rain gauge data and radar-derived quantitative precipitation estimates (QPE) (Cannon et al. 2010). Radar-based QPE has been used for decades to estimate rainfall but has a number of uncertainties as discussed by Krajewski et al. (2010). In particular, for Southern California, a recent National Academy of Science study determined that the Southern California S-band radars (i.e., KSOX and KVTX) likely suffer from underestimation of near-coastal QPE due to beam overshooting because they are sited at relatively high altitudes coupled with their lowest elevation angle of 0.5° (National Research Council 2005). Thus, a need exists for “gap filling” radars to monitor high rainfall rates that are approaching and falling over debris-flow-vulnerable areas, particularly those that border populated areas. Moreover, there is increasing evidence that short-term rainfall intensities, rather than long-term precipitation loading, is important for triggering debris flows within burned watersheds (Cannon et al. 2008). To this end, area-wide monitoring of precipitation intensity by radar could be an important tool for the warning decision process and for differentiating rainfall–runoff response from recently disturbed steep terrain.

Although weather radars would seem particularly well suited for monitoring flash-flood conditions over small watersheds because of their high temporal and spatial resolutions, in reality the rainfall estimates, which have historically been based on classical single-polarization measurements, have quantitative limitations (Anagnostou et al. 1999; Smith et al. 1996). These limitations are well known (Battan 1973; Doviak and Zrnic 2006) and involve uncertainties associated with the choice of appropriate atmospheric reflectivity–rainfall relationships for the storms being studied; radar receiver calibration uncertainties; attenuation of reflectivity by intervening precipitation; contamination of weather echoes by ground clutter and other nonmeteorological scatter such as birds, insects, and chaff; beam blockage problems, particularly in complex terrain; and uncertainty interpreting weather echoes from beams above or at the freezing level in terms of surface precipitation rates.

To evaluate mitigation of some of the uncertainties associated with traditional nonpolarimetric radar-derived QPE, during the winter rain season of 2009/10 a truck-mounted dual-polarized (DP) C-band Doppler weather radar system was deployed to monitor rainfall over the area burned by the 2009 Station Fire near fire Los Angeles, California. The radar system, called C-band polarimetric (CPOL), was operated during nine rainfall events from December 2009 to February 2010, three of which produced substantial debris flows from several watersheds. The Station Fire was the largest wildfire in the history of Los Angeles County, burning from 26 August 2009 to 16 October 2009 (NCDC 2010) in the San Gabriel Mountains north of Los Angeles. This human-ignited fire consumed over 160 000 acres and destroyed 209 structures, including 89 homes (NCDC 2010). The proximity of the burn to highly populated areas along the I-210 corridor, including the communities of La Cañada Flintridge and La Crescenta-Montrose, prompted the CPOL deployment (Fig. 1) to the Burbank, California, airport. The radar data provided rainfall information for storms that both did and did not initiate debris flows in our effort to refine the existing intensity–duration thresholds used in debris-flow warning decisions. Data were also transmitted in near–real time, via an Internet cell phone connection, to the local NWS forecast office in Oxnard, California. Figure 1 also shows instrument sites installed within the burn area by the USGS consisting of tipping-bucket rain gauges, soil moisture probes, flow stage, and pore pressure instruments. Flow stage was measured using both ultrasonic and laser distance meters suspended over the channel. Both devices measure the distance from the sensor to the surface below. Pore pressures were measured by pressure transducers vented to the atmosphere to measure fluid pressures in the channel bed material directly beneath the stage gauge. These measurements, in addition to poststorm field observations to evaluate processes of erosion and deposition, constitute the validation of debris-flow occurrence that can be correlated to the radar-derived rainfall and rain gauge observations.

Section 2 describes the datasets used in the study and data processing methodology to estimate precipitation from the radar data. Section 3 describes the results of the study.

2. Radar characteristics and data processing methodology

a. Radar parameters and scanning characteristics

Technical parameters of the truck-mounted radar upgraded to dual-polarization capability by funds provided by the NOAA/National Severe Storms Laboratory (NSSL), the University of Oklahoma (OU), and the National Science Foundation are given in Table 1.
Prior to the polarization upgrade, the radar was a joint development effort of NSSL and several universities and termed the Shared Mobile Atmospheric Research and Teaching Radar (SMART-R) (Biggerstaff et al. 2005). The locations of the three radars used in this study relative to the debris basin downslope of the Dunsmore Canyon watershed monitored for debris flows are given in Table 2.

To provide complete volume scanning of atmospheric reflectivity over the complex terrain of Southern California requires multiple elevation angles to avoid low-level beam blockage by high topography. Volume

![Map of instrument locations](image)

**FIG. 1.** (top) Los Angeles basin area and (bottom) expanded map of instrument locations. The locations of the two NWS WSR-88Ds (KSOX and KVTX) are shown in (top) with their 50-km range rings as dashed lines. CPOL location with its 15-km range ring is shown near the Station Fire burn area (blue outline). The black box is the domain of the expanded map. County boundaries are in light gray and major freeways are the red lines. In (bottom), the Station Fire burn area is the white outline. Major freeways are shown with labels. The CPOL (sited at the Burbank airport) is denoted by the red square. The yellow outlined area is the Dunsmore watershed. The rain gauges used in the study are shown as white triangles. Image in (bottom) is courtesy of Google Earth.
Coverage Pattern 12 (VCP-12) was used for CPOL (Brown et al. 2005) to be consistent with what many NWS Weather Surveillance Radar-1988 Doppler (WSR-88D) radars use in rain conditions because it provides for denser vertical sampling at lower elevation angles. The elevation angle sequence is shown in Table 1 and is repeated approximately every 5 min. The two Southern California WSR-88D radars used in this study (KSOX and KVTX) also used VCP-12 scanning during the rain events documented in this study, but their repeat times were slightly shorter, ~4–5 min.

Calibration of reflectivity is critical for accurate determination of rainfall using the standard $Z-R$ relationship. CPOL is routinely calibrated by injecting known signal strength into the waveguide (both horizontal and vertical channels) before and after each deployment. No reflectivity calibration differences were noted after the debris-flow deployment. However, because the signal generator injects power in the waveguide near to the receiver, there are still unknown losses due to the rest of the waveguide as well as the rotary joint and feed horn losses. To calibrate those losses, reflectivity comparisons with a collocated NSSL WSR-88D radar (KOUN) in Norman, Oklahoma (which is also calibrated frequently using a signal generator), was done for thunderstorms exhibiting a range of maximum reflectivity. Comparisons to reflectivity calculated from a disdrometer located about 12 km away were also made. The “extra” loses were ~4 dB. We feel that in an absolute sense the radar was well calibrated to ~1 dB.

### b. Attenuation correction

C-band radar is known to attenuate in heavy rain (Bringi et al. 2001; Capsoni et al. 2001; Delrieu et al. 1999), leading to a reduction in returned power for echoes at farther range. However, the availability of the dual-polarization parameter differential phase $\phi_{DP}$ allows for some mitigation of reflectivity $Z$ and differential reflectivity $Z_{DR}$ due to path attenuation. Several correction methods for this attenuation have been proposed (Bringi et al. 2001; Gorgucci et al. 1998; Scarchilli et al. 1998). Because we have restricted our observations to those that are below the freezing level, the method used follows that of Silvestro et al. (2009) for homogeneous rain,

$$Z_C = Z_{raw} + \alpha [\phi_{DP}(r) - \phi_{DP}(0)],$$  \hspace{1cm} (1)

where $Z_C$ is the corrected reflectivity estimate at range $r$; $Z_{raw}$ is the original reflectivity estimate from the horizontal channel; $\phi_{DP}(r)$ and $\phi_{DP}(0)$ are the differential phase at range $r$ and at the first minimum range bin (determined by taking the mean of the first 5 bins), respectively; and $\alpha$ is a constant that depends on the character of the drop size distribution and can be estimated by empirically fitting straight lines to pairs of $(Z_{raw}, \phi_{DP})$ data points (the negative of the slope of the line is $\alpha$). The analogous attenuation correction relationship for differential reflectivity $[=10 \log(Z_{h}/Z_{v})]$, where $Z_{h}$ is the horizontal channel reflectivity and $Z_{v}$ is the vertical channel reflectivity $Z_{DR_{c}}$ is

$$Z_{DR_{c}} = Z_{DR_{raw}} + \beta [\phi_{DP}(r) - \phi_{DP}(0)],$$  \hspace{1cm} (2)

where $\beta$ is an analogous constant to $\alpha$ in (1). Following Ryzhkov et al. (2005) and Silvestro et al. (2009), the two coefficients, $\alpha$ and $\beta$, are determined by Bringi and Chandrasekar (2001) by scattering simulations using gamma fits to typical drop size distributions at C band and are set, for this study, at the midpoint of their typical ranges: that is, $\alpha = 0.08$ and $\beta = 0.015$.

### Table 2. Radars used in this study.

<table>
<thead>
<tr>
<th>Radar</th>
<th>Location relative to Dunsmore</th>
<th>Alt (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CPOL</td>
<td>−12 km to southwest</td>
<td>216</td>
</tr>
<tr>
<td>KSOX</td>
<td>−75 km to southeast</td>
<td>923</td>
</tr>
<tr>
<td>KVTX</td>
<td>−88 km to west-northwest</td>
<td>831</td>
</tr>
</tbody>
</table>

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c. Construction of hybrid reflectivity scans

Rainfall is derived from range bins assembled from reflectivity and dual-polarization measurements from each volume scan into a fixed polar grid with resolution of 1° in azimuth by 0.075 km in range out to the maximum unambiguous range of 111 km. At each range bin, the tilt angles used are pieced together from the various tilt levels of the VCP-12 scan following the methodology of O’Bannon (1997), and this is termed a “hybrid scan.” The choice of tilt level depends upon local terrain variations as determined by the USGS digital elevation model (DEM) with a 90-m resolution (3 arc second). The 90-m resolution DEM was used to match the gate spacing of the radar (100 m). The criterion for which tilt angle to use for each polar grid point is the lowest tilt that clears the ground by at least 50 m, unless that beam is blocked more than 50% by terrain. The beam blockage calculation was based on standard atmospheric ray propagation (Battan 1973) within the radar main lobe (3-dB main lobe).

The beam height above the ground is shown in Fig. 2 for the three radars used in this study (CPOL, KVTX, and KSOX). The beam heights are similar for all three radars in the vicinity of the rain gauges (~500 to ~1250 m), except for the gauge labeled CCC, for which the CPOL beam elevation is ~1850 m above local topography from CPOL. That station was sufficiently high to be above the freezing level during most of the rain events as was not used very often. The CPOL beam heights generally increased with increasing range because of the need to use higher tilts (>5° elevation angle) to shoot over the nearby Verdugo Mountains that

**Fig. 2.** Height of the radar beam minus the ground surface topography (m) for (a) CPOL, (b) KVTX, and (c) KSOX. The Station Fire boundary is shown as the white outline. Rain gauge locations are shown as yellow triangles with their three-letter identifier. CPOL location is shown as a red dot.
were situated between the burn area in the San Gabriel Mountains and the CPOL site at the Burbank airport.

There are several discontinuities in beam heights shown in Fig. 2 caused by beam obstructions that cause a switch to a higher tilt. The discontinuities in radar parameters, however, were not nearly as severe as those indicated for beam heights because the vertical gradients in reflectivity and the polarimetric variables typically are not that sharp, at least below the freezing level. Figure 3 is an example of a 5-min hybrid scan from on the more intense rain periods [intensive observation period 8 (IOP-8); at 1155–1200 UTC 6 February 2010]. Discontinuities in reflectivity can still be seen, but their effect on rain calculation is felt to be minimal, at least in the vicinity of the burn area.

d. Rainfall rate estimator from dual-polarization parameters

We evaluated several rainfall rate estimators discussed by Bringi and Chandrasekar (2001) for C-band radar and follow the approach of Silvestro et al. (2009) for combining the various relationships based on thresholds. The estimation of rainfall rate \( R \), using dual-polarization radar parameters following attenuation correction (for convenience, we drop the \( C \) subscript), \( Z_h \) (the horizontal channel reflectivity), \( Z_{DR} \) (differential reflectivity; \( =10 \log(Z_h/Z_v) \)), where \( Z_v \) is the vertical channel reflectivity), and \( K_{DP} \) (specific differential phase) are given by

\[
R(K_{DP}, Z_{DR}) = 37.9K_{DP}^{0.89}10^{-0.072Z_{DR}}, \quad (3)
\]

\[
R(Z_h, Z_{DR}) = 0.0058Z_h^{0.91}10^{-0.209Z_{DR}}, \quad (4)
\]

\[
R(K_{DP}) = 31.4K_{DP}^{0.70}, \quad \text{and} \quad (5)
\]

\[
R(Z_h) = 0.017Z_h^{0.714}. \quad (6)
\]

Estimator (6) is the \( Z-R \) relationship for convective rain used by the NWS (Fulton et al. 1998). The coefficients of relationships (3)–(5) are given by Bringi and Chandrasekar (2001) for C-band and for different gamma drop size distributions over a realistic range of parameters of the distribution such as the intercept parameter, water content, and mean volume diameter. Negative \( K_{DP} \) values were set to zero.

Calibration of \( Z_{DR} \) was accomplished following the methodology of Gourley et al. (2006), which involves pointing the radar antenna vertically (i.e., a “bird bath”
orientation) during rainfall and spinning the dish through several rotations. Because the orientation of falling rain drops should not be biased toward $Z_h$ or $Z_v$ at vertical incidence, an average of $Z_{\text{DR}}$ through the spin cycle represents a $Z_{\text{DR}}$ bias that must be removed before the rainfall rate estimators are calculated. The estimate of $Z_{\text{DR}}$ bias is done for ranges over 0.5 km up to the freezing level and typically represents $>10^4$ values where $Z_h > 25$ dBZ. The precision (i.e., standard deviation about the mean) of the $Z_{\text{DR}}$ bias was typically $\sim(0.25-0.40)$ dB, approximately what Gourley et al. (2006) found for an operational C-band radar. The bias was $\sim0.6$ dB and did not vary (i.e., was within the precision of the measurements) during the 2009/10 winter rain season.

The methodology used to choose which rainfall rate estimator [Eqs. (3)–(6)] to use makes use of thresholds on $K_{\text{DP}}$ and $Z_{\text{DR}}$. The basic idea is to use an estimator in a range where it theoretically performs best. For example, in regions of light precipitation (e.g., $Z_h < 25$ dBZ), $K_{\text{DP}}$ and $Z_{\text{DR}}$ are often very noisy (Gorgucci et al. 1994), so the $Z - R$ relation [estimator (6)] is used when $Z_h < 25$ dBZ. Radar calibration errors on $Z_h$ and attenuation uncertainties at high rainfall rates make using a $Z - R$ relationship problematic because small changes in $Z$ make for large variations in $R$. For high rainfall rate areas, it is therefore more appropriate to use one of the other dual-polarization estimators [(3)–(5)]. The quantity $K_{\text{DP}}$ is calculated within the Vaisala/Significant Meteorological Information (SIGMET) signal processor (SIGMET 2005) as a least squares estimate of the slope of the differential phase within a moving 5-km range window. However, the trend of differential phase along the beam is noisy as well, affected by aliasing and backscatter differential phase issues. The consistency of $K_{\text{DP}}$ was checked by plotting $Z$ versus $K_{\text{DP}}$ (not shown), and the trends are consistent with plots from other research radars (Bringi and Chandrasekar 2001).

### e. Threshold determination

The algorithm to pick the appropriate estimator (3)–(6) follows the methodology of Silvestro et al. (2009). Given thresholds for $K_{\text{DP}}$ and $Z_{\text{DR}}$ ($T_{K\text{DP}}$ and $T_{Z\text{DR}}$, respectively), relationship (6) is used for $R$ if $K_{\text{DP}} < T_{K\text{DP}}$ and $Z_{\text{DR}} < T_{Z\text{DR}}$; relationship (3) is used if $K_{\text{DP}} \geq T_{K\text{DP}}$ and $Z_{\text{DR}} \geq T_{Z\text{DR}}$; relationship (4) is used if $K_{\text{DP}} \leq T_{K\text{DP}}$ and $Z_{\text{DR}} \leq T_{Z\text{DR}}$; and relationship (5) is used if $K_{\text{DP}} > T_{K\text{DP}}$ and $Z_{\text{DR}} \geq T_{Z\text{DR}}$. The determination of the thresholds $T_{K\text{DP}}$ and $T_{Z\text{DR}}$ was achieved by computing normalized biases (NBs; defined as the sum over all gauges of radar-derived hourly accumulated rainfall minus hourly gauge accumulations divided by the sum of the gauge accumulations) for all the events for combinations of trial thresholds for hourly accumulated radar-derived and gauge-measured rainfall, then determining the thresholds that minimized the biases. Trial thresholds for $T_{K\text{DP}}$ were $0.1^\circ \text{ km}^{-1}$ steps from $0.1^\circ$ to $5^\circ \text{ km}^{-1}$ and for $T_{Z\text{DR}}$ were $0.1$-dB steps from 0.1 to 2.0 dB. To determine the radar-derived rainfall over each gauge listed in Table 3, a Cressman interpolation (Cressman 1959) was performed with a radius of influence of 200 m from the radial data bins contained in the hybrid reflectivity scans. Based on the 1000 threshold combinations, the minimum normalized gauge biases were for the thresholds

$$T_{K\text{DP}} = 0.7^\circ \text{ km}^{-1} \quad \text{and} \quad T_{Z\text{DR}} = 1.0 \text{ dB}.$$
In actual practice, using the above thresholds, we found that relationship (6) was used about 80% of the time with the other relationships used about equally the rest of the time. The frequency that each relationship was used in the three IOPs that had debris flows is shown in Fig. 4. For rainfall rates up to moderate (~45 dBZ) the $R(Z)$ relation was used most often, indicating that both $K_{DP}$ and $Z_{DR}$ did not exceed their respective thresholds. For the highest rainfall rates, the $R(K_{DP}, Z_{DR})$ relationship was used most for IOP-3 and IOP-8, indicating that both thresholds were exceeded. Interestingly, for IOP-4, the strongest storm in terms of total rainfall and strength of individual dBZ cells, the $R(Z, Z_{DR})$ relationship was used just as often, indicating that the $K_{DP}$ threshold was not exceeded but $Z_{DR}$ was. We speculate that this indicates that the drop size distribution of IOP-4 was skewed, relative to the other storms, toward bigger drops rather than higher number concentrations of moderate sized drops (reflectivity is proportional to the sixth power of drop diameter, with $K_{DP}$ only proportional to the third power).

Additionally, we evaluated several $Z–R$ relations using the same hourly accumulated gauge bias determinations as for $T_{KDP}$ and $T_{ZDR}$. The $Z–R$ relations evaluated were ($R$ in mm h$^{-1}$ and $Z$ in mm$^6$ m$^{-3}$)

\[
Z = 300R^{1.4} \quad (\text{Fulton et al. 1998}), \tag{7}
\]
\[
Z = 200R^{1.6} \quad (\text{Marshall et al. 1955}), \quad \text{and} \tag{8}
\]
\[
Z = 101R^{1.76} \quad (\text{Matrosov et al. 2007}). \tag{9}
\]

Equation (7) is actually the estimator (6) used by the NWS for convective rain. Equation (8) is the traditional stratiform rainfall relationship, whereas Eq. (9) is a relationship derived from winter season precipitation measured by a disdrometer in Northern California (Gourley et al. 2009). Rain gauge normalized bias estimates for the three relationships [(7)–(9)] were fairly close, indicative of the relatively small difference in radar-derived rainfall for all three relationships within the range of maximum observed reflectivity (typically <50 dBZ) in Southern California rain storms. We therefore chose Eq. (7) to use as it is currently utilized by the NWS as a standard relation for convective rainfall estimation by the WSR-88D radar network.

3. Results

a. Radar-derived rainfall accumulations compared to rain gauges

We produced radar-derived rainfall estimates (for all IOP events shown in Fig. 5) using the dual-polarization estimator described in section 2d from each 5-min volume scan and accumulated the individual rain rates to produce hourly rain totals over each gauge site. Similarly, rain gauge observations were accumulated from their raw reporting frequency (Table 3) to hourly amounts. A scatterplot of CPOL-derived hourly accumulated rain versus gauge amounts is shown in Fig. 6 for both the dual-polarization estimate and $Z–R$-estimated rain amounts.

The correlation coefficient shows a slight improvement in correlation using the dual-polarization estimates compared to the traditional $Z–R$ approach and a moderate improvement in the underestimation of radar-derived dual-polarization amounts compared to $Z–R$. The normalized bias improved from a $-14.8\%$ underestimation to only a $-2.6\%$ underestimation. Although it is not surprising that the threshold algorithm approach produced the lowest normalized bias (we engineered the approach to minimize bias), it is somewhat surprising to
FIG. 5. Rainfall time series from the DUN3 gauge. Vertical blue lines are rain intensity [mm (5 min)$^{-1}$; left vertical axis] and the black line is rain accumulation (mm; right vertical axis). Shaded areas are the time periods of CPOL data collection (IOPs). Shown are the periods (a) 6–15 Dec 2009, (b) 17–27 Jan 2010, and (c) 2–12 Feb 2010. There was no significant rainfall for the periods between the graph limits.
see the traditional Z–R method consistently underestimate rainfall accumulation. This underestimation points to a consistent Z–R bias in the relationship used, perhaps because of a difference in California orographic winter rain drop size distribution from that used to define the Z–R relation.

The improvement that a gap-filling radar in rain determination over the two NWS network radars, KSOX and KVTX, utilizing only the Z–R relationship [Eq. (7)], is shown in Fig. 7. Although the correlation coefficients to the gauge amounts are only slightly lower than the CPOL coefficients, the normalized bias is considerably more negative. The network radars underestimated the precipitation by about 50%. Although beam height above local topography was comparable for all three radars (Fig. 2), the long range of the network radars to the gauge sites (vertical beamwidth of 1° implies a beam that is ~1.4 km in vertical extent at 80 km range) meant that the WSR-88D radar beams, for most storms, extended above the melting level given the rather low freezing levels of most Southern California storms in winter. The relatively rapid reduction in radar reflectivity with height in these orographically generated Southern California rainstorms produced a decided underestimate of the surface rainfall rates. An appropriate vertical profile of reflectivity correction could be applied (Gray et al. 2002; Zhang and Qi 2010) to correct for brightband and ice effects, but that was not done with either the CPOL or WSR-88D radar data in this study, because the focus of this paper is on radar observations of rain.

b. Basin total rainfall

Although comparison of CPOL estimates of rain directly over gauges showed an improved correlation and normalized bias over those estimates obtained from the distant WSR-88D-derived estimates, the character of orographically forced Southern California winter rain tends to be organized as small cells of intense rain within a general background of lighter precipitation. This character is shown in an example of an hourly accumulation of dual-polarization-estimated rain in Fig. 8 for a storm that occurred on 18 January 2010. Rain cells of accumulated rain of 10–15 mm can be seen in the general background field of 2–10 mm. In general, these relatively intense cells are only 1–2 km in size. Although a comprehensive study of the scale of precipitation cells in Southern California winter storms has not been completed, a hint of the small-scale nature of the cells can be seen in the rain gauge and radar maps shown in Neiman et al. (2004).

To evaluate whether the dual-polarization radar provides an improvement in rainfall detection accuracy over a wide domain, a calculation of accumulated rain in the watershed upslope of the Dunsmore basin outlet (watershed boundary shown in red outline in Fig. 8) was performed. The Dunsmore watershed is a 2.1-km² area
(519 acres) that was instrumented with three USGS rain gauges and flow sensors (the stations labeled DUN1, DUN2, and DUN3 in Fig. 1).

For the rain accumulation over the watershed, any Cartesian grid point from CPOL, KSOX, or KVTX within the Dunsmore watershed was included in the total rain calculation. The hourly estimate was divided by the domain area to yield an average rain amount that was then compared to the average hourly rain accumulation from the three gauges. The scatterplot from this comparison is shown in Fig. 9. Consistent with the gauge comparisons, the dual-polarization rain estimates show improved correlation coefficients and lower normalized biases than the WSR-88D \( Z-R \)-derived values. Interestingly, the dual-polarized rain estimates overestimated the three-gauge average by 5.8%, although given the small-scale nature of the rain cells it is possible the gauges themselves missed some of the precipitation detected by the CPOL radar.

c. Use of gap-filling radar-derived precipitation in debris-flow warning decisions

Debris-flow warning decisions are based on rain intensity–duration thresholds for the San Gabriel Mountains of Southern California determined empirically from USGS studies using rain gauges and poststorm field observations of debris flows (Cannon et al. 2008, 2010; Restrepo et al. 2008). These thresholds are used in the NWS Flash-Flood Monitoring and Prediction (FFMP) system (Filiaggi et al. 2002) to alert forecasters whenever radar-derived or rain gauge–measured amounts exceed the threshold for a given time period in a particular watershed. To compare the effectiveness of dual-polarization-derived rainfall estimates to those from the WSR-88D \( Z-R \) relationship, a calculation of rain intensity–duration amounts for the three IOPs that produced debris flows was performed (Table 4). Figure 10 shows the results for CPOL- and WSR-88D-derived rainfall relative to intensity–duration thresholds identified by Cannon et al. (2010). The thresholds are from first-year wildfires in the San Gabriel Mountains of Southern California using rain gauges sited within various burn areas and includes data from the Station Fire area. The rainfall accumulation calculation over the Dunsmore watershed was done for the 12 h prior to each of the recorded debris–flow event, except for IOP-3, which only had CPOL data for about 9 h prior to the event. The average rainfall intensity was calculated over a specified duration by summing the radar-derived rainfall over the burn area.
for the given duration and then dividing by the duration. For durations less than one hour, the maximum rainfall intensity in any 10-, 15-, or 30-min period up to one hour prior to the debris-flow event was used. This was done to allow for some delay in the debris-flow response from the short-period rainfall. The threshold rainfall conditions that relate to debris-flow events of different magnitudes are reproduced from Cannon et al. (2010) and shown in Table 5. Debris-flow magnitude is defined in terms of the volumes of individual debris flows, consequences of debris flows and floods in an urban setting, and spatial extents of the hydrological response. Note that the potential for post-fire debris flows will decrease with time (i.e., the second year following the fire) as revegetation stabilizes hill slope and soil and rock are removed from canyons by debris flows and fluvial transport.

The strongest debris-flow response (IOP-8; Table 4) seen by CPOL was well above the threshold for a magnitude-II event at all times except for the shortest 10-min intensity. Even KSOX- and KVTX-derived rainfall intensity was above that threshold beyond ~15 min, although KVTX fell below the threshold after ~30 min. The USGS estimate of the response was magnitude III (Table 4), and neither the CPOL nor the WSR-88D observed rainfall in excess of that threshold. For the moderately strong IOP-4 response, only CPOL was above the threshold for a magnitude-II event and only after a 30-min accumulation. The CPOL rainfall accumulation came close to the magnitude-II threshold for the short time periods (10- and 15-min accumulations) during IOP-3 but did not exceed any threshold for longer time periods. Neither WSR-88D-derived rainfall exceeded the magnitude-II–III threshold for any time period during either IOP-3 or IOP-4, which is consistent with the earlier results that showed a general underestimation of rain by the network radars in comparison to the rain gauges and CPOL. It is worth noting that the spike in KSOX IOP-8 rain at 30 min may not have been rainfall. Inspection of KSOX reflectivity maps for the hour prior to the measured debris flow showed what appeared to be an extensive brightband contamination (i.e., enhanced reflectivity due to melting snow) between the small, more intense convective cells, even though the center of the beam was below the melting level.

The USGS magnitude thresholds are subjective and are roughly correlated to rainfall accumulations determined by radar as shown in Fig. 10. The characteristics of individual basins affect the response. The

![Fig. 9. As in Fig. 6, but for the Dunsmore watershed average precipitation from the CPOL DP estimate (blue dots), KVTX Z–R relationship estimates (red squares), and KSOX Z–R relationship estimates (green triangles).](image)

**Table 4.** The magnitude of the debris-flow event was determined by post-event surveys conducted by the USGS using the definitions described in Table 5.

<table>
<thead>
<tr>
<th>IOP</th>
<th>Time–date of debris flow</th>
<th>Magnitude</th>
<th>Gauge detected</th>
<th>Damage</th>
</tr>
</thead>
<tbody>
<tr>
<td>3</td>
<td>~0900 UTC 13 Dec 2009</td>
<td>I</td>
<td>DUN1</td>
<td>Small debris flow detected at monitoring site; it did not travel to basin outlet.</td>
</tr>
<tr>
<td>4</td>
<td>~2000 UTC 18 Jan 2010</td>
<td>II</td>
<td>DUN1 and DUN2</td>
<td>Debris flow up to 3 m deep destroyed monitoring site; it traveled into debris basin at basin outlet.</td>
</tr>
<tr>
<td>8</td>
<td>~1100 UTC 6 Feb 2010</td>
<td>III</td>
<td>DUN1 and DUN2</td>
<td>Several debris flows up to 4 m deep that occurred over several hours detected at monitoring site; it traveled into debris basin at basin outlet: 86 residences damaged and millions of dollars in property damage.</td>
</tr>
</tbody>
</table>
NWS, however, does not make distinctions in severity when issuing flash-flood guidance. Typically, if anticipated or measured rainfall exceeds a magnitude-I threshold, a warning is issued. Based on that criteria, both radar systems, CPOL and the two WSR-88Ds, detected the two stronger events, IOP-4 and IOP-8. Only CPOL, however, was able to measure rainfall above the warning threshold for the weak event, IOP-3.

Although the CPOL data were made available to the Los Angeles NWS forecasters for use in their debris-flow warning decisions, it is difficult to quantify the radar’s impact on warning decisions. NWS forecasters use a variety of data sources in their subjective determination to issue warnings, especially looking “upstream” from the burn area.

### 4. Conclusions and future work

A dual-polarized, mobile Doppler weather radar was strategically sited to observe rainfall over a recent wildfire burn area during the 2009/10 winter season in Southern California. A procedure to combine several dual-polarization estimates of precipitation using thresholds was shown to produce improvements in accuracy, in terms of higher correlation coefficients and lower normalized biases (for observations below the freezing level), compared to traditional empirical reflectivity–rainfall relationships when compared to rain gauges sited within the burn area. Moreover, the gap-filling radar improved rainfall estimates over those obtained from nearby WSR-88D radars. Finally, the gap-filling radar was able to
correctly identify rainfall conditions associated with varying magnitude debris-flow events, whereas the WSR-88D consistently underestimated area-wide precipitation in the Dunsmore watershed and would have been of marginal value in warning decisions for all but the highest magnitude events.

The major unanswered question is whether the improvements in CPOL radar-derived rainfall estimates over those provided by the WSR-88D comes as a result of using dual-polarimetric variables or as a consequence of being sited closer to the rain gauges. Comparison of the normalized biases in Figs. 6 and 7 show a dramatic reduction in normalized bias when using CPOL $Z-R$ over WSR-88D $Z-R$. This reduction is likely due to siting, assuming a correct calibration of all radars. The additional improvement in normalized bias (from $-14.8\%$ to $-12.6\%$) between CPOL $Z-R$ and CPOL dual-polarization parameters can probably be attributed to the use of dual-polarimetric parameters. Similar improvements in area-wide precipitation estimates of $10\%$–$20\%$ have been seen using an S-band radar (Ryzhkov et al. 2005).

Although the dual-polarization parameters provided more accurate estimates of precipitation compared to NWS network radars in this study, many avenues of continued improvements can be pursued to improve measurement accuracy further. As outlined by Matrosov et al. (2005), drop size measuring instruments (disdrometers) can be used to adjust the coefficients of the dual-polarimetric relationships to “tune” them for the particular character of the rain of Southern California. When the national WSR-88D network radars are dual-polarized in the near future, knowledge about local drop size distributions in typical rain storms can be used to improve the dual-polarimetric rain estimators. Corrections for brightband effects and extrapolating radar measurements made above the freezing level [vertical profile of reflectivity corrections (VPRs)] to WSR-88D observations can also be made that improve surface rainfall estimates (Matrosov et al. 2007).

Even without the dual-polarization upgrade to the WSR-88D radars, there are approaches that can lead to improved precipitation estimates. The Multisensor Precipitation Estimator software (MPE; Seo 1998) can be used to bias correct hourly rainfall estimates using reporting rain gauges and apply these bias values to the high-resolution precipitation estimates that can feed the automated flash-flood alert software used at most NWS warning offices. This approach shows promise in reducing the underestimation of precipitation by the WSR-88D.

Finally, although work is underway on the development of methods for estimating probabilities and volumes of debris flows relative to watershed and rainfall conditions and to identify areas that can be impacted by these events (Cannon et al. 2009), more work needs to be done to quantify uncertainties associated with such methods for characterizing debris-flow occurrence and how the information is conveyed to emergency managers (Restrepo et al. 2008). Moreover, combining empirically determined rainfall thresholds with physical measurements and models of debris-flow processes offers the hope of enhancing and expanding warnings to other regions, both burned and unburned. Such a lofty goal, however, cannot be implemented in an operational system at present.

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REFERENCES


