Faulting in the 1986 Chalfant, California, Sequence: 
Local Tectonics and Earthquake Source Parameters

by Kenneth D. Smith and Keith F. Priestley

Abstract  The July 1986, moment magnitude ($M_w$) 6.3 Chalfant, California, earthquake is the largest of a recent series (1978–present) of moderate-sized earthquakes near the Long Valley volcanic region of east-central California. The sequence consists primarily of three moderate-sized strike-slip events. High-quality aftershock relocations and short-period focal mechanisms define the temporal and spatial development of the foreshock-mainshock-aftershock periods of these three events. Faulting during the $M_w$ 5.7 (event I; July 20) and the $M_w$ 6.3 (event II; 21 July) events constitute a set of conjugate strike-slip faults. Event I involved predominantly left-lateral motion on a NE-striking fault plane initiating at shallow depth (7 km). Event II initiated at 10.5 km depth exhibiting right-lateral strike-slip motion on a NW-striking fault dipping moderately to the southwest. An $M_L$ 5.5 strike-slip event on 31 July (event III) extended the aftershock sequence to the south into the White Mountains fault zone. $P$-wave pulse-width stress drops are determined for 185 $M_L$ 2.7–4.0 earthquakes that sample the entire sequence. Higher stress drops are observed near the intersection of event I and II fault planes and at the northern and southern ends of the aftershock zone. The moving-window $b$-value of the temporal magnitude distribution shows a general inverse relationship to stress drop with observed changes in both the average stress drop level and the $b$-value preceding event III. The average aftershock stress drop tends to increase as the sequence progresses suggesting that the faulted volume is equilibrating to the regional stress. Source parameters have been determined for the principal earthquakes from teleseismic body waves, local strong-motion records, and the extent of aftershock activity. The Chalfant sequence appears to be transferring strike-slip motion away from the White Mountains front, contributing to the observed increase in the relative normal offset, from south to north, along the White Mountains and the opening of the White Mountains relative to the Sierra Nevada Range front north of Owens Valley.

Introduction

The 1986 Chalfant, California, earthquake sequence occurred east of the Long Valley Caldera and the Sierra Nevada Range front beneath the Volcanic Tablelands region north of Bishop, California (Figs. 1 and 2). Long Valley is an active Quaternary volcanic center with a geologic history of explosive volcanism (Bailey et al., 1976; Hill, 1996; Hill et al., 1985). The Chalfant sequence is the most recent, and best recorded, in a series of more than 10 moderate-sized earthquakes and aftershock sequences that have occurred in the region since 1978 and includes the largest earthquake of the series (Cramer and Toppazada, 1980; Ryall and Ryall, 1981, 1983; Aki, 1984; Cockerham anditt, 1984; Savage and Cockerham, 1984; Lide and Ryall, 1985; Cockerham and Corbett, 1987; Priestley et al., 1988; Smith and Priestley, 1988; Peppin et al., 1989). Earthquake activity in the caldera has been attributed to magma movement (Julian, 1983; Julian and Sipken, 1985; Wallace, 1984; Hill et al., 1990; Pitt and Hill, 1994; Hill, 1996), and measured deformation has coincided with some of the recent activity (Savage and Clark, 1982; Denlinger and Bailey, 1984; Rundle and Whitcomb, 1984). Seismic activity has remained at a high level since 1978 (Hill, 1996). In addition to potential volcanism, the area includes several Holocene faults with significant Quaternary offsets (Bryant, 1984) indicating that they are capable of strong earthquakes. Several of these Holocene faults are mapped directly within the areas of recent seismicity in the Long Valley-Volcanic Tablelands region. The small town of Chalfant is adjacent to the White Mountains fault zone (WMFZ; Fig. 1). This fault zone is characterized by an approximately equal amount of right-lateral and normal-offset with increased normal-offset north-
ward along its trace ((dePolo, 1989). The increase in normal-offset is expressed as an eastward step, development of normal fault morphology and steepening of the range front north of Chalfant. The overall Quaternary slip rate for the WMFZ is estimated to be 0.5–1.2 mm/yr (dePolo, 1989). The Chalfant mainshock and most of the earthquake sequence took place in the hanging wall block of the WMFZ, although one of the moderate events of the sequence appears to have involved strike-slip faulting along the WMFZ.

Although range-bounding faults are present throughout the region (Figs. 1 and 2), the recent series of moderate earthquakes has shown no significant component of normal faulting. Structurally, the region is characterized by a conspicuous westward step and opening in the Sierra Nevada Range relative to the White Mountains-Inyo Mountains block to the east, coinciding with the Long Valley Caldera volcanic center. Conjugate strike-slip faulting during the Chalfant and the 1984 Round Valley sequences (Priestley et al., 1988) occurring within an overall local E–W extension direction, and in conjunction with N–S striking normal faulting appears to be fundamental to the kinematics of the deformations of the Volcanic Tablelands region.

The Volcanic Tablelands region is mostly comprised of the Bishop Tuff sequence, deposited 0.7 m.y.a during a catastrophic caldera eruption (Gilbert, 1938; Dalrymple et al., 1965; Bailey et al., 1976). Since its deposition, there has been approximately 175 m of vertical displacement across the WMFZ (Sheridan, 1975), and the area has also undergone internal deformation along north-to-north-northeast-striking faults that accommodate both dip-slip and strike-slip motion (Tocher et al., 1963). Bateman (1965) suggested that these faults and their en-echelon geometry are mostly likely the product of east–west extension within the Volcanic Tablelands region within a larger right-lateral shear environment.

Within the context of the regional deformation (Fig. 1), the Chalfant area is located at the north end of the throughgoing faults of the Eastern California Shear Zone (Dokka and Travis, 1990) and southwest of the strike-slip faults of the east-central Walker Lane Belt (Stewart, 1988). Through the Tertiary, between 30 to 75 km of right-lateral strike-slip motion has been accommodated along the Furnace Creek and Fish Lake Valley fault zones (Reheis and McKee, 1991; Reheis and Sawyer, 1997). The Owens Valley fault accounts for 2 to 20 km of cumulative slip (Beanland and Clark, 1994). The Chalfant sequence took place directly north of the surface displacements of the 1872 Mw 7.6 Owens Valley earthquake, a right-oblique predominantly strike-slip event (Beanland and Clark, 1994). The more recent late Tertiary activity and smaller cumulative slip through Owens Valley suggests that primary deformation, previously concentrated on the Furnace Creek system, may be evolving westward to Owens Valley and possibly to the WMFZ (Dixon et al., 1995).

Slip transfer mechanisms that provide some continuity in the strain and connect the primary fault systems have been proposed throughout the region. Based on Very Long Baseline Interferometry (VLBI) and GPS data, Reheis and Dixon (1996) suggest that slip may be transferred by a series of normal fault systems through Eureka Valley and Deep Springs Valley from Owens Valley eastward to Fish Lake Valley (Fig. 1). North of Owens Valley, along the WMFZ and eastern Sierra Nevada Range front, strain is distributed in a complex pattern involving normal faulting along developed range fronts and distributed high-angle strike-slip deformation expressed in the recent earthquake sequences in the Long Valley region. North of the White Mountains, Oldow et al. (1994) suggest that right-lateral strike-slip motion is linked from the Fish-Lake Valley Fault to the east-central Walker Lane Belt through a large scale detachment mechanism. Stewart (1985) proposes that large offsets along EW-trending strike-slip faults account for the observed regional strain fabric, although much of this is pre-Tertiary deformation.

Because the Chalfant sequence occurred within a dense seismic network, we are able to document in detail the spatial and temporal development of the earthquake sequence.
From $P$-wave pulse-width measurements we can map the spatial distribution and temporal progression of stress drops for events greater than $M_L > 2.7$ and relate this to the complex fault geometry of the sequence. The Chalfant sequence provides a unique opportunity to observe the development of complex faulting in a Basin and Range Province earthquake sequence.

**Fault Geometry and Aftershock Distribution**

**Earthquake Location Procedure**

The Chalfant sequence was recorded on the University of Nevada Reno Seismological Laboratory (UNRSL) and U.S. Geological Survey's (USGS) short-period seismic network near Long Valley (Fig. 2). At the time, the network primarily consisted of vertical-component 1 Hz seismometers with some horizontal sensors. The analog signals were telemetered to UNRSL, digitized at 50 Hz, and arrival times and first motions were determined by an analyst.

Earthquakes were first located using FASTHYPO (Herrmann, 1979), a 1D $P$-wave velocity model (Table 1) and only stations within 75 km of the sequence. This model is used at UNRSL in the western Great Basin region. The average travel time residual at each station, relative to the velocity model, was applied as a station correction in a subsequent relocation. The seismicity plots show all $M_L > 3$ events and $M_L < 3$ earthquakes that fit the following criteria: (1) RMS error of the travel time residual less than 0.08 sec; (2) at least 15 stations within 75 km used in the location; (3) a location gap of less than 90 degrees, and (4) at least one station within 10 km of the epicenter. All $M_L > 3$ earthquakes, regardless of location quality are shown. A total of 2847 earthquakes of $M_L > 1.5$ (approximate magnitude threshold of the local network) have been located during the
period from 3 July through 31 October 1986, and of these, 2450 satisfy these conditions.

P-wave first-motion focal mechanisms were determined using the program FPFIT (Reasenberg and Oppenheimer, 1985) for events with 20 or more first-motion polarities.

The earthquake sequence consists of three distinct moderate-sized, dominantly strike-slip faulting events. Events I and II define a conjugate fault geometry in the hanging wall block of the WMFZ, and event III took place within the WMFZ at the southern end of the aftershock zone (de-Polo, 1989). Although an immediate aftershock to event II was the third largest earthquake of the sequence, event III is significant in its direct relationship to the WMFZ. Table 2 summarizes hypocentral coordinates and fault plane solutions determined in this study.

Seismicity

Figures 3–5 show seismicity and cross sections and perspective views summarizing relocations and focal mechanisms. Figure 3 shows the seismicity map for all events occurring in the July–October time period (Fig. 3a) and a cross-sectional view looking along the event II fault plane (Fig. 3b). The sequence extends for approximately 20 km in a general N–S direction with most of the activity lying between 3 and 14 km depth. Aftershocks north of event II are broadly distributed but generally trend N–S. South of event III, aftershock activity also aligns N–S subparallel to the WMFZ. Between events II and III aftershocks trend NW–SE. This epicentral distribution is an expression of the geometry of intersecting and dipping fault planes of the three principal events and an extension of the sequence NE of event II not related to principal faulting.

Also, shown in Figure 3a are the P-wave first-motion fault plane solutions for the three principal events. The preferred fault planes, labeled with arrows, are based on the alignment of aftershock activity associated with each of these events. All three principal earthquakes show dominantly strike-slip motion on steeply to moderately dipping fault planes.

Figure 4a is a perspective view in a best attempt to look down-dip at the principal fault planes, and Figures 4b–c are enlargements of the activity up until the time of event II. Figure 4b shows the conjugate geometry of the event I and event II fault surfaces in the same perspective view as Figure 4a, although only showing activity up to and including event II. The geometry of event II aftershocks and the event I rupture surface with respect to the hypocenter of event II is shown in Figure 4c. Figures 4d–j show seismicity maps as a function of time and illustrate the temporal and spatial evolution of the sequence. In the 20 months prior to 1 July 1986, four earthquakes, all less than Ml 2.5, occurred in the volume of the Chalfant sequence. Event I foreshocks began on 3 July with an Ml 3.5 event and continued until event I (Fig. 4b). The rate of seismicity increased on 18 July and then noticeably decreased prior to event I. During the 17 days proceeding event I, 42 earthquakes were large enough to trigger the monitoring network, including eight events larger than of equal to Ml 3. These events formed a tight cluster around the event I hypocenter at about a depth of 7 km (Fig. 4d). This tight foreshock cluster and initiation of event I appear to have taken place at the intersection of the event I and event II fault planes (Fig. 4c). Foreshocks of event I are updip, and approximately 5 km shallower, and NNE of the eventual event II hypocenter (Fig. 4b). Event I ruptured with left-lateral strike-slip motion on a NE-striking moderately NW-dipping fault plane (Fig. 4c).

Earthquakes occurring between 14:29 GMT on 20 July and 14:42 on 21 July consisted of aftershocks of event I and foreshocks of event II (Fig. 4b–e). Smith and Priestley (1988) showed that aftershocks of event I (Fig. 4b) occurred around the edge of a generally circular, NW-dipping rupture surface (Fig. 4b) corresponding to the NW-dipping, NE-striking fault plane shown in the event I mechanism (Fig. 3a). The bulk of event I aftershocks occurred SW of the event I epicenter and at a greater depth (Fig. 4b), suggesting that rupture propagated down and to the SW and did not extend below a depth of 10 km. Following event I, earthquakes that are clearly off the event I fault surface locate near the event II hypocenter approximately perpendicular to the event I fault plane and at depth of approximately 11

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*UC Berkeley Richter Magnitude.
km (Fig. 4b). In contrast to the foreshocks of event I, only one $M_L 3+$ event took place in the event II foreshock activity near the event II hypocenter (Fig. 4b).

In the nine hours following event II (Fig. 4f), aftershock activity extended to the SE along the event II fault plane intersecting the WMFZ. One of the largest aftershocks of the sequence occurred 10 minutes after event II; $M 5.7$ at 14:51 on 21 July. Although the location of this event is constrained to be SW of event II (Fig. 4f), first arrivals are difficult to identify in the nearly continuous activity following event II. Its location probably marks the southern extent of event II faulting intersecting the WMFZ. Event II aftershock activity also spread to the NW, defining the northern extend of the event II fault plane.

Event III nucleated in the cluster of activity at the southern extent of the event II rupture (Fig. 4i). While there were a number of $M_L 3+$ events in Chalfant aftershock zone in the days following event II (Fig. 4g–i), few occurred in the event III epicentral region in the hours before event III (Fig. 4i). While two $M_L 4+$ earthquake and several $M_L 3+$ events occurred at the northern end of the sequence, aftershocks south of event II in the 31 hours prior to event III were less than $M_L 2$. Figure 4i shows the seismicity in the two months following event III, and this pattern is persistent in the seismicity of the region through the mid-1990s.

The broad zone of activity north of event II (Figs. 3a, 4a) does not conform to either the event I or event II fault planes. This activity does not extend below about 10 km depth (Fig. 3b). Normal faulting mechanisms in this area (Fig. 4a) suggest an extensional or “pull-apart” region. This volume is in fact in an extensional regime resulting from the conjugate fault geometry of events I and II, and is adjacent to the predominantly normal faulting northern portion of the WMFZ. The composite P-T diagram (Fig. 5b) suggests a general east–west extension direction for the sequence, although this does not account for the partitioning of the moment
release for the various orientations of fault planes. Figure 6 summarizes the geometry of faulting during the sequence and shows the intersecting nature of the three principal faulting events.

Savage and Gross (1995) have revised their slip model for Chalfant (Gross and Savage, 1987) to account for conjugate strike-slip faulting geometry. Gross and Savage’s (1987) measurement of $5.2 \times 10^{25}$ dyne cm for a total moment release (which did not include event III) is in good agreement with a cumulative moment of $5.7 \times 10^{25}$ dyne cm that we determine for the July through September 1986 activity. The difference in these total moment estimates is approximately the seismic moment of event III. Table 3 lists the 20 largest events of the Chalfant sequence.

Source Parameters

Teleseismic Modeling

The fault planes of the three principal events are well constrained by the first motion polarities, but we have also inverted the teleseismic $P$ and $SH$ waveforms recorded at stations in the $30^\circ$–$90^\circ$ distance range. We have used the waveform inversion routine from the IASPEI software li-
brary, and details of the processing and inversion are discussed by Nabelek (1984) and McCaffrey and Aber (1988). The results of the body-waveform inversion are shown in Figure 6. Although the $P$ waves for event I (Fig. 7) are small in amplitude, the nodal planes are well constrained by the $SH$ waveforms. The fault strike from the waveform inversion is consistent with that of the $P$-wave first motion and the seismic moment was radiated in a single 4-sec pulse indicating a relatively simple source.

Pacheco and Nabelek (1988) modeled broadband teleseismic waveforms for event II and found a fault plane similar to that shown by the $P$-wave first-motion solution. Their source time function consists of three sharp pulses indicating the breaking of three asperities, and a characteristic time for the whole rupture process of 2.4 sec implying a fault radius of 6.3 km. They also observed a secondary phase about 15 sec after the first $P$-wave arrival, which they attributed to the conversion of energy at a near source, low-velocity zone.

With respect to the wavelengths used in the teleseismic body-wave inversions, events I and II were essentially collocated and event I shows no secondary phase of this type. Lienkaemper et al. (1987) and dePolo and Ramelli (1987) mapped surface fractures found in a NW–SE trending zone across the Volcanic Tablelands, but also found fracturing along the WMFZ following event II and prior to event III, suggesting possible displacement on the WMFZ during event II.

In our inversion of the event II body waves, we have considered the possibility of a secondary source. The solid lines in the inversion result (Fig. 7b) denote faulting asso-
associated with the main moment release and the dashed lines represent faulting from a possible secondary source. Event II has a source duration of about 10 sec with the main moment release occurring over 7 sec. The fault plane of the main moment release is nearly identical to the first-motion solution, but the auxiliary plane has a larger normal component. The moment of the postulated second source is about 15% of the moment of the $M_w 6.3$ mainshock and the near N–S plane is consistent with the orientation of the WMFZ. However, the other plane of this mechanism is similar to the event I fault plane. Savage and Gross (1995) suggest that motion on both event I and II fault planes during event II may account for discrepancies in the seismic and geodetic moment estimates.

The mechanism of event III (Fig. 7c) is constrained by few data. One plane is fixed by the change in polarity of the $SH$ wave between stations KONO and GRFO and is consistent with the first-motion fault-plane solution. The seismic energy was radiated in a single 2-sec pulse indicating a relatively simple source for this event. Table 4 summarizes source parameters of the principal events.

**Principal Events**

For event I we estimate a source radius of about 3 km based on the first 24 hours of aftershock activity. Since this is the first moderate-sized event, an estimate of its source area from the aftershock data is clear, which is not the case for events II and III. Using this fault area the slip is (Brune, 1970, 1971),

$$\langle u \rangle = \frac{M_0}{\mu A} = 63 \text{ cm}$$

where $\mu$ is the rigidity (assumed $3 \times 10^{11}$ dyne/cm$^2$) and the static stress drop is,

$$\Delta\sigma = \frac{7}{16} \frac{M_0}{r^2} = 87 \text{ bars}.$$ 

Early aftershocks of event II roughly define a 15-km-long region that ranges in depth from about 5 to 13 km. Event II nucleated near the base of the seismogenic zone, and appears to have ruptured bilaterally northwest and southeast (Fig.
4a). Complicating the estimation of the extent of rupture of event II is an $M_{L} 5.5$ + event, 10 minutes later at 14:51 GMT, and the overprinting of event I aftershock activity. Assuming an 8-km fault width and a surface-wave moment of $3.9 \times 10^{18}$ N m, the average slip is 0.71 m and the static stress drop for the rectangular geometry is (Kanamori and Anderson, 1975),

$$\Delta \sigma = \frac{2}{\pi} \mu \frac{\langle \omega \rangle}{W} = 26 \text{ bars}$$

where $W$ is the fault width. If slip during the 14:51 $M_{L} 5.7$ aftershock occurred over a surface area equivalent to that of event I, then the estimated rupture area of event II would be decreased, to 90 km$^2$, resulting in an average displacement of 1.5 m and a static stress drop of 35 bars. Pacheco and Nabelek (1988) suggest that the energy release during event II was dominated by three asperities with individual stress drops on the order of 500 to 600 bars. This interpretation could account for additional surface area on the event II fault surface that may have slipped (based on the aftershock distribution) during the 14:51 earthquake. Event III aftershocks, although more difficult to isolate, suggest a fault area similar to that of event I. Applying a fault radius of 3 km and a surface-wave-derived moment of $1.6 \times 10^{17}$ N m, the average slip and stress drop for this event are 25 cm and 40 bars, respectively.

### Estimates of Source Parameters from Strong Motion Records

Table 5 summarizes the source parameters determined from fitting the spectra of acceleration records recorded at the strong motion station at Zack Ranch. Figure 8 shows the horizontal components and the $S$-wave acceleration spectra of the component of PGA. The Zack Ranch station is located on alluvium in Chalfant Valley adjacent to the WMFZ, where significant site effects are likely. We fit the observed spectra to a Brune (1970, 1971) source model by constraining the long-period level to the seismic moments determined from the telesismic records (see Table 4) and the high frequencies to a best estimate of the local value of kappa (Anderson and Hough, 1984). A kappa value of 40 msec provides the best constraint to all spectra in the 5 to 25 Hz band, a band
Table 3
The 20 Largest Earthquakes of the Sequence

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*Year-month day-hour minute-second.
†Degrees and decimal-minutes.

that is assumed to be above the corner frequencies. The corner frequency is then fit by minimizing the error between the smoothed spectra (Fig. 8) and the model, assuming a factor of two for the free-surface amplification and correcting for distance. The source parameters estimated from the strong-motion records are generally consistent with those estimated from the aftershock distribution, although there is an excess of energy in the 2–8 Hz band for event II (Fig. 8b). This additional energy may be due to applying a Brune source model to a nonradial rupture geometry or may result from failures of localized high stress-drop asperities as proposed by Pacheco and Nabelek (1988). Since we have only have one strong-motion station it is not possible to resolve directivity effects in the data. Table 5 summarizes source parameters determined from fitting strong-motion records.

**P-Wave Pulse-Width Stress Drops**

O’Neill and Healy (1973) proposed a simple method for estimating the rupture duration of small earthquakes ($M_L$ 3–4) from clipped short-period seismograms. They corrected the initial $P$-wave pulse width, $\tau_{1/2}$, for the effects of the instrument and path attenuation and estimated rupture durations and stress drops by comparing observed and theoretical pulse widths. Frankel and Kanamori (1983) showed that frequency domain deconvolution could be approximated in the time domain by simply subtracting the pulse width of an empirical Green’s function event $\tau_{1/2}$ from the pulse width of a larger, colocated event $\tau_{1/2}$, giving $\tau_{1/2}$, source. The source radius, $r$, is then (Boatwright, 1980),

$$r = \frac{\tau_{1/2}(source)^2}{1 - (\nu/c) \sin \theta}$$

where $\nu$ is the rupture velocity, $c$ is the $P$-wave velocity, and $\theta$ is the azimuth between the normal to the fault plane and the recording site. (In this study we assume $\nu$ is equal to 80% of a 3.2 km/sec shear-wave velocity and $\theta$ is 45\degree). The stress drop, $\Delta \sigma$, for a radially symmetric source is (Brune, 1970, 1971),

$$\Delta \sigma = \frac{7}{16} \frac{M_0}{r^3}$$

where $M_0$ is the seismic moment.

Smith and Priestley (1988) applied this technique to aftershocks of the 1984 Round Valley, California, sequence, 15 km southwest of Chalfant, and found that for the limited source area of the aftershock zone, pulse broadening for the very smallest earthquakes at an individual station was a site effect. Therefore, establishing a magnitude-based minimum pulse width, $\tau_{1/2}$, source (O’Neill and Healy, 1973) or a correction
for each source receiver pair (Frankel and Kanamori, 1983), is not necessary. Removing the path and instrument effects could be reduced to determining minimum pulse widths at each seismograph site. Frankel and Kanamori (1983) concluded that pulse widths could be measured to one quarter of a sample interval (0.005 sec for 50 Hz data) by interpolating between samples of the first arrival and first zero crossing.

We hand-measured over 3000 $P$-wave pulse widths from 350 events of the Chalfant sequence. The subsample level first-arrival time was interpolated from the rise of the initial pulse and the pulse width was then established as the zero-crossing with the pre-$P$-wave noise level. The nine stations where pulse widths measurements were made are shown in Figure 1 (filled triangles) and four sites are shown in Figure 9. At least to approximately 40-km range, the minimum pulse width does not vary systematically with hypocentral distance. Also, Figure 9 shows pulse widths for a theoretical causal pulse shape for an input delta function for a $Q_p$ of 200 and 50 at distances of 20 to 40 km. For $Q_p$ of 50, whole-path attenuation dominates pulse broadening but has little effect at $<40$ km for a $Q_p$ of 200. Abercrombie (1995) determined an upper to midcrustal $Q_p$ of approximately 1000 in the 2–10 Hz band from deep borehole recordings and showed that surface recordings were dominated by near-surface effects. Considering that $P$-wave

Figure 7. Focal mechanisms (lower-hemisphere projections) for three principal events of the Chalfant earthquake sequence determined from body-wave inversion: (a) 20 July 1986, (b) 21 July 1986, and (c) 31 July 1986. The $P$ and $T$ axes are marked by solid and open circles, respectively. The observed $P$ and $SH$ waveforms (solid lines) are compared with synthetic waveforms (dashed lines) computed for the minimum misfit solution. Label at the left of each waveform identifies the station code, an uppercase letter corresponding to the location on the focal sphere, and a lowercase letter denoting the instrument type ($w = $ WWSSN LP, $d = $ GDSN LP). The source time function is shown below the $P$-wave focal sphere, and below this is the waveform time scale. The waveform amplitude scales are to the left of the focal spheres. For details of the inversion procedure see McCaffrey and Aber (1988).
Table 4

Source Parameters from Teleseismic Waveform Modeling and Extent of Aftershock Distribution

<table>
<thead>
<tr>
<th></th>
<th>Event I</th>
<th>Event II</th>
<th>Event III</th>
</tr>
</thead>
<tbody>
<tr>
<td>P &amp; N (Mₚ N m)</td>
<td>None</td>
<td>2.1 × 10ⁱ⁸</td>
<td>None</td>
</tr>
<tr>
<td>S &amp; P</td>
<td>4.2 × 10¹⁷</td>
<td>2.5 × 10¹⁸</td>
<td>None</td>
</tr>
<tr>
<td>This study</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>M₀ (N m)</td>
<td>5.4 × 10¹⁷</td>
<td>3.9 × 10¹⁸</td>
<td>1.6 × 10¹⁷</td>
</tr>
<tr>
<td>Source Area (km²)</td>
<td>28</td>
<td>120</td>
<td>5.4</td>
</tr>
<tr>
<td>Static Δτ (bars)</td>
<td>87</td>
<td>26</td>
<td>40</td>
</tr>
<tr>
<td>Displacement (cm)</td>
<td>63</td>
<td>108</td>
<td>25</td>
</tr>
</tbody>
</table>

*Source Area (km²) 27 123 17 11
*fo (Hz) 0.40 0.19 0.5 0.63
*Distance (km) 12 18 20 26
*PGA (%g) 28 46 17 7

Source Parameters from Fitting Strong-Motion Spectra

<table>
<thead>
<tr>
<th>Zack Ranch</th>
<th>Event I</th>
<th>Event II</th>
<th>7/21 14:51*</th>
<th>Event III</th>
</tr>
</thead>
<tbody>
<tr>
<td>PGA (%g)</td>
<td>28</td>
<td>46</td>
<td>17</td>
<td>7</td>
</tr>
<tr>
<td>Distance (km)</td>
<td>12</td>
<td>18</td>
<td>20</td>
<td>26</td>
</tr>
<tr>
<td>fo (Hz)</td>
<td>0.40</td>
<td>0.19</td>
<td>0.5</td>
<td>0.63</td>
</tr>
<tr>
<td>Source Area (km²)</td>
<td>27</td>
<td>123</td>
<td>17</td>
<td>11</td>
</tr>
<tr>
<td>Stress Drop (bars)</td>
<td>88</td>
<td>69</td>
<td>72</td>
<td>106</td>
</tr>
<tr>
<td>Displacement (cm)</td>
<td>64</td>
<td>106</td>
<td>41</td>
<td>48</td>
</tr>
</tbody>
</table>

*Source Area (km²) 28 120 22
*S & P 4.2
*fo (Hz) 0.40
*Distance (km) 12 18 20 26
*PGA (%g) 28 46 17 7

Source Parameters from Teleseismic Waveform Modeling and Extent of Aftershock Distribution

Table 5

Source Parameters from Fitting Strong-Motion Spectra

<table>
<thead>
<tr>
<th>Zack Ranch</th>
<th>Event I</th>
<th>Event II</th>
<th>7/21 14:51*</th>
<th>Event III</th>
</tr>
</thead>
<tbody>
<tr>
<td>PGA (%g)</td>
<td>28</td>
<td>46</td>
<td>17</td>
<td>7</td>
</tr>
<tr>
<td>Distance (km)</td>
<td>12</td>
<td>18</td>
<td>20</td>
<td>26</td>
</tr>
<tr>
<td>fo (Hz)</td>
<td>0.40</td>
<td>0.19</td>
<td>0.5</td>
<td>0.63</td>
</tr>
<tr>
<td>Source Area (km²)</td>
<td>27</td>
<td>123</td>
<td>17</td>
<td>11</td>
</tr>
<tr>
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<td>69</td>
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</tr>
<tr>
<td>Displacement (cm)</td>
<td>64</td>
<td>106</td>
<td>41</td>
<td>48</td>
</tr>
</tbody>
</table>

*Source Area (km²) 28 120 22
*fo (Hz) 0.40
*Distance (km) 12 18 20 26
*PGA (%g) 28 46 17 7

We measured pulse widths from 50 of the smallest events that spatially sampled the entire aftershock zone to determine the pulse-width minimum (site correction) for the stations indicated on Figure 1. The minimum pulse width was determined graphically by plotting all pulse-width measurements on a log-log graph to emphasize minimum values and selecting the pulse-width minimum as the intercept of the asymptote of the sorted set with the time axis. In this procedure we were able to justifiably eliminate unphysically narrow pulse-width measurements in the large database that were most likely misinterpreted due to trace noise at low amplitudes or resulting from processing errors.

For each earthquake in the range M₀ 2.7–4.0 with seven or more observations we then computed the corrected pulse width (τ₁/2corrected) for each seismogram. This guarantees good azimuthal coverage. The corrected pulse widths for each event were averaged to establish the source-process time (τ₁/2source). To avoid events with possible unilateral rupture geometry, which violate the assumption of a radially symmetric source, we eliminate events with an average deviation from the mean of greater than 0.02 sec (1 sample). Approximately 300 earthquake stress drops were computed, and of these 185 met this criteria.

Seismic moments were determined from the coda duration magnitudes (Mₚ) in the UNRSL catalog using the moment-magnitude relationship of Chavez and Priestley (1985) for the Great Basin:

\[
\log{M₀} = 1.2M + 17.49
\]

The UNRSL catalog magnitudes were found to be systematically 0.3 M units greater than magnitudes (Mₚ) reported by the USGS for the Bishop area at this time, and we reduced Mₚ by this amount. The actual stress drop may be in question because of the uncertainty in the estimate of the seismic moment, but the relative scaling of the stress drop between events is preserved. Magnitude calculations were consistent through the development of the earthquake catalog.

The spatial distribution of stress drops and earthquake magnitudes (not including principal events) is shown in Figure 10. The concentration of high stress-drop events in the center of Figure 10(b) corresponds to intersection of the event I and II fault planes. There is a second concentration of high stress-drop events along the WMFZ and to the south. The stress drops of earthquakes that we interpret to have taken place on the event II fault plane, immediately north and south of event II, have low to intermediate stress drops. High stress drops are present at the northern end of the aftershock zone in an area dominated by normal faulting but not likely to have been involved in primary faulting.

The short-period data provides a nearly continuous record of the foreshock and the postseismic periods. Figure 11 show the temporal progression of stress drops, magnitudes, and maximum likelihood b-values for 120 days of the sequence. Table 6 summarizes the stress drops for all event I foreshocks, some which precede the beginning of the log-time plot of Figure 11. Earthquakes prior to event I show higher stress drops compared to those that follows events I and II. Prior to event III there are a number of 100+ bar events following a period of lower stress-drop earthquakes. These high stress drops coincide with a decrease in the moving b-value over the same time period. There is a decrease in stress drops and a corresponding upturn in the b-value immediately prior to event III.

This inverse relationship between the mean stress-drop level at any particular time and the b-value is expected if...
Faulting in the 1986 Chalfant, California, Sequence: Local Tectonics and Earthquake Source Parameters

Note Scaling

Event I
M 5.7

14:51 GMT
7/21

Event II
M 6.3

Event III
M 5.5

14:51 GMT
7/21

(a)

(b)

(c)

(d)
Figure 9. Measured P-wave pulse widths at four stations as a function of hypocentral distance. Also shown is the expected pulse broadening for a causal pulse shape with distance for Q values of 50 and 200 at 20 and 40 km (Y. Zeng, UNRSL, written communication, 1995). The locations of these stations used in the pulse width study are shown in Figure 1. Stations shown are (a) ORC, (b) CWC, (c) BHP, and (d) BON. At station BON travel paths exceed 40 km for several events from the southern end of the aftershock zone. At this distant whole path Q effects may be contributing to pulse broadening. Station BHP is located on a thick section of sediments near Bishop, and this is most likely reflected in its larger pulse-width minimum.

There is an increase in stress drop with seismic moment; the moving b-value would be lower with increased numbers of relatively higher magnitude events. Also, it is clear from Figure 11 that times of low b-values are not only characterized by increased numbers of high-magnitude events but a general decrease in the number of Ml 1–3 earthquakes. This is also clear in the post-event II, pre-event III time period. This is not a function of network processing procedures, since the entire sequence was processed at same level of completeness (D. dePolo, UNRSL, personal communication, 1995). The cumulative b-value for the entire sequence is about 1.0. The general upward trend of stress drop tends to reach the levels for pre-event I values shown in Table 6, suggesting that the aftershock zone is equilibrating as the sequence progresses to the preprincipal faulting stress state.

Figure 12 is the moment–stress-drop relationship and includes lines of constant source radii of 500 and 250 m. There is an increase in the minimum stress-drop level with increasing seismic moment consistent with a number of studies of stress-drop scaling relationships (Tucker and Brune, 1974; Boatwright et al., 1990; Hough and Dreger, 1995; Feng and Ebel, 1996). There is a greater variation of stress drop at lower magnitudes for the Chalfant data than is generally seen in these other studies. One notable outlier is the low stress drop for an Ml 3.8 event with a source radius of nearly 1 km. (This event occurred 1 minute before an Ml 5 aftershock on 1 August in the same general hypocentral region.) We confirmed the relatively broad pulse width for this event. The maximum source dimension of approximately 500 m is estimated over the Ml 2.7–4 range and is not an instrument effect; very broad pulses are observed for the very larger earthquakes (Ml > 5, not included in the pulse-width analysis). Also from Figure 11, the average stress drop over this range of magnitudes is about 100 bars. This general shape of the moment–stress-drop relationship is sometimes seen in other aftershock studies and may result from the existence of a range of source sizes with a variety of stress states distributed around some mean stress level. Because the larger events have generally higher stress drops, they dominate the radiated energy from the sequence to a greater
Figure 10. Epicentral locations, magnitudes, and stress drops for events for which aftershock stress drops were determined shown on a Thematic Mapper image: (a) earthquake magnitudes; (b) pulse-width stress-drop values.

degree than would be implied from their larger seismic moments alone. This means that the smaller earthquakes are less of a contributor, although minor in any event, to the total slip in the aftershock period.

Tectonic Interpretations

We observe that the WMFZ transitions northward from a dominantly strike-slip to normal faulting system at Chalfant. Because of its nearly pure strike-slip sense of motion, the Chalfant mainshock appears to be transferring strike-slip motion away from White Mountains front and into the Volcanic Tablelands. This is a mechanism for slip transfer along the lines proposed by Dixon et al., (1995), in which normal faults through Deep Spring Valley and Eureka Valley transfer slip from Owens Valley to Fish Lake Valley. Here at Chalfant we hypothesize that strike-slip motion transferred from the WMFZ is expressed in conjugate strike-slip faulting in a process that contributes to the left-step and the conspicuous opening in the Sierra Nevada Range front in the Long Valley–Volcanic Tablelands region. The E–W extension direction expressed in focal mechanisms the Volcanic Tablelands is indicative of this process. This is in contrast to a prevailing NW extension direction for the central Great Basin (Rogers et al., 1991). Conjugate strike-slip faulting was exhibited in the 1984 Round Valley sequence (Priestley et al., 1988) and intersecting strike-slip faulting often characterizes the relationships between the moderate earthquakes in the immediate Long Valley area. In the Chalfant area in particular, strike-slip faulting is occurring in conjunction with normal faulting as shown by the presence of the north-striking Fish Slough normal fault, offset to the west from the White Mountains Range front (see Fig. 1).

Smith and Biasi (1999) proposed a model in which the volcanic source at Long Valley is being overridden by the Sierra Nevada Range driven by the NW progression of the Sierran block. Regional gravity and P-wave tomography show that the heat source for the Long Valley area may extend in the upper mantle to east. We suggest that the White Mountains transition from a normal faulting to strike-slip system may correspond to the position of this deeper upper mantle hot-weak zone. Therefore, the tectonic processes observed at Chalfant may reflect a change in upper mantle–lower crustal strength in this transition of faulting style.
Discussion and Conclusions

The 1986 Chalfant, California, earthquake sequence represents an eastward expansion of the recent period of earthquake activity in the Long Valley area (1978–present). In terms of the local tectonic framework, the Chalfant sequence would appear to be more closely associated with deformation along the White Mountains Fault Zone rather than caldera deformation and related volcanic processes.

The sequence initiated on 3 July 1986, with a well-defined foreshock period that included 17 $M_L \geq 3$ earthquakes. Event I faulting, based on its aftershock distribution, took place well above the base of the seismogenic zone. The second and largest moderate earthquake ($M_w 6.3$) nucleated at 10.5 km depth and involved nearly pure right-lateral strike-slip motion. From the conjugate faulting geometry (event II and event II fault planes), it appears that event I initiated on the event II fault surface but up-dip and NNW of the event II hypocenter. Event III took place 10 days following event II, on 31 July, and most likely ruptured a section of the WMFZ.

One interpretation of the spatial distribution of stress drops is that the entire volume experienced a stress reduction as a result of the series of moderate-sized earthquakes, and the generally higher levels observed at the northern and southern edges are reflecting the state of regional stress outside of the faulted volume. The high stress-drop values near the central portion of the aftershock zone could result from stress complexities as a consequence of the conjugate fault-
ing geometry and local stress adjustments at primary fault
intersections. In a more speculative interpretation, if deep,
mid to lower crustal afterslip may be driving some after-
shock activity, consistently low stress drops could be ob-
served along fault surfaces that have experienced relatively
complete stress release.

The stress drop–magnitude scaling relationship is simi-
lar to that observed in other studies, providing some relative
confidence in the pulse-width method. A maximum source
size implied from the distribution is around 500 m for events
in this magnitude range. This may represent a generaliza-
tion of the maximum asperity size. Also, there is a general in-
verse relationship between the mean stress-drop level and
the moving window \( b \)-value. This relationship is particu-
larly clear in the days leading up to event III. If stress drops in-
crease with moment, it should be expected that the mean
stress drop, in any particular time period, will increase with
decreasing \( b \)-values.

There is more variation in the stress drops at low mag-
nitudes, and this may result from inaccuracies in the \( M_c \)
estimate and resolving the pulse width for the smaller events.
Overall, the mean stress-drop level for the Chalfant after-
shocks is around 100 bars. The source dimensions of the
principal events were on the order of several to 10 km (as
compared to a maximum of 500 m observed for the after-
shocks), but these events may themselves involve many in-
dividual patches or asperities as in composite source repre-
sentations (Frankel, 1991; Zeng and Anderson, 1995).

The focal mechanisms determined from the teleseismic
data are consistent with short-period results and the teleseis-
mic-moment estimates are consistent with the longer-period
levels of the \( S \)-wave spectra from the strong-motion records.

A second source modeled from the teleseismic data for event
II may indicate coseismic slip on the event I fault surface,
consistent with a model proposed by Savage and Gross
(1995) to account for discrepancies between geodetic and
seismic evidence at Chalfant.

We hypothesize that the inverse relationship between
stress drop and the \( b \)-value may be the consequence of af-
terslip processes resulting from a viscoelastic response to
loading of the ductile midcrust during event II faulting. This
may be contributing to driving the aftershock period. If af-
terslip rates vary, then this may be expressed in variations
in aftershock rates, the temporal development of earthquake
source parameters and the evolving aftershock period in gen-
eral. This hypothesis could be tested in future earthquakes
with the strategic deployment of GPS instruments by com-
paring real-time deformation with the progression of seis-
micity rates and source parameters. In addition, the general
increase in earthquake stress drops as the sequence pro-
gresses may represent a process of equilibrating to the back-
ground stress, or regional stress level, that exists outside the
volume that experienced stress relaxation from principal
faulting.

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References


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