Original

Bulletin of the Seismological Society of America, Vol. 86, No. 6, pp. 1804–1819, December 1996

Rare, Large Earthquakes at the Laramide Deformation Front—Colorado (1882) and Wyoming (1984)

by William Spence, C. J. Langer*, and George L. Choy

Abstract The largest historical earthquake known in Colorado occurred on 7 November 1882. Knowledge of its size, location, and specific tectonic environment is important for the design of critical structures in the rapidly growing region of the Southern Rocky Mountains. More than one century later, on 18 October 1984, an m_b 5.3 earthquake occurred in the Laramie Mountains, Wyoming. By studying the 1984 earthquake, we are able to provide constraints on the location and size of the 1882 earthquake. Analysis of broadband seismic data shows the 1984 mainshock to have nucleated at a depth of 27.5 $\pm$ 1.0 km and to have ruptured $\sim$2.7 km updip, with a corresponding average displacement of about 48 cm and average stress drop of about 180 bars. This high stress drop may explain why the earthquake was felt over an area about 3.5 times that expected for a shallow earthquake of the same magnitude in this region. A microearthquake survey shows aftershocks to be just above the mainshock's rupture, mostly in a volume measuring 3 to 4 km across. Focal mechanisms for the mainshock and aftershocks have NE–SW-trending T axes, a feature shared by most earthquakes in western Colorado and by the induced Denver earthquakes of 1967.

The only data for the 1882 earthquake were intensity reports from a heterogeneously distributed population. Interpretation of these reports also might be affected by ground-motion amplification from fluvial deposits and possible significant focal depth for the mainshock. The primary aftershock of the 1882 earthquake was felt most strongly in the northern Front Range, leading Kirkham and Rogers (1985) to locate the epicenters of the aftershock and mainshock there. The Front Range is a geomorphic extension of the Laramie Mountains. Both features are part of the eastern deformation front of the Laramide orogeny. Based on knowledge of regional tectonics and using intensity maps for the 1984 and the 1967 Denver earthquakes, we reinterpret prior intensity maps for the 1882 earthquake as reflecting low population to the north and east of available intensity data. We estimate that, had there been a more evenly distributed population, the 1882 earthquake would have been felt over an area of about 850,000 km^2, with a corresponding moment magnitude (M_w) of 6.6 $\pm$ 0.6. Our study, in the context of regional tectonics, implies that rare earthquakes of magnitude up to M_w 6.6 $\pm$ 0.6, at depths from shallow to mid-crustal, are possible throughout the Laramie Mountains and the Front Range, approximately from Casper, Wyoming, to Pueblo, Colorado.

KEY WORDS: Colorado, Earthquake, Earthquake Hazard, Seismic Hazard, Earthquake Potential, Seismic Potential, Colorado Earthquakes, Denver Earthquakes.

Introduction

Improved understanding of the origins of the earthquake felt throughout Colorado on 7 November 1882 is important for determining the appropriate earthquake-resistant design for critical structures in the rapidly developing Rocky Mountain region. There have been considerable difficulties with using intensity data to infer the location and size of this earthquake. For example, separate studies have suggested several widely separated locations of this earthquake: at the northwestern corner of Colorado (McGuire *et al.*, 1982), in north-central Colorado at the site of the "Denver" earthquakes of 1967 (Hadsell, 1968), or in the northern Front Range, west of Fort Collins (Kirkham and Rogers, 1985, 1986).

*Present address: Box 2354, Hailey, Idaho 83333.

Analysis of the origins of the 1882 earthquake is aided in this article by study of the Laramie Mountains earthquake of 18 October 1984 (m_b 5.3) and its aftershocks. The 1984 earthquake occurred in the northern Laramie Mountains, Wyoming, which are the northern extension of Colorado's Front Range (Fig. 1). These features together represent part of the eastern deformation front of the Laramide orogeny and mark the dramatic physiographic boundary between the Rocky Mountains and the High Plains. The dividing line between the Laramie Mountains and the Front Range is the northern boundary of the Colorado lineament (Warner, 1978), shown in Figure 1. We will use the inference that the 1882 and 1984 earthquakes occurred within analogous tectonic structures to provide a combined interpretation of the 1882 and 1984 earthquakes.

Figure 1 shows the tectonic setting and known earthquakes of Colorado and southern Wyoming, occurring from 1867 to 1993. Generally, these earthquakes are less than magnitude 5 and occur diffusely throughout the Southern Rocky Mountains, with slightly increased activity at the eastern boundary of the Colorado Plateau (Spence *et al.*, 1988). Locations of historical earthquakes prior to the 1920s are based on intensity data. After the 1920s, earthquakes larger than about magnitude 5 could be instrumentally located, but it was not until 1964 that this capability extended down to the magnitude 3 range. Earthquakes within the Colorado Plateau and in the Colorado Rocky Mountains are characterized by NE–SW-trending extension (Wong and Humphrey, 1989). In Figure 1, the earthquakes clustered east of the Front Range mostly were induced in the mid-1960s by high-pressure injection of toxic fluids in a 3.7 km (2.3 mi)-deep disposal well at the Rocky Mountain Arsenal, just northeast of Denver (Evans, 1966). The three largest were in 1967, with body-wave magnitudes (m_b) of 4.7, 5.0, and 5.1 (International Seismological Commission Catalog, ISC), depths of 3 to 5 km, and focal mechanisms showing normal faulting having NE–SW-trending extension (Healy *et al.*, 1968; Herrmann *et al.*, 1981). In Figure 1, the 1984 earthquakes are shown in the northern Laramie Mountains, and the Kirkham and Rogers' (1985, 1986) location of the 1882 mainshock is shown by the large circle in the Front Range.

1984 Mainshock

The seismic moment, M_0, of the 1984 earthquake is 1.1 $\times$ 10^{24} dyne-cm (ISC), and the corresponding moment magnitude, M_w, is 5.3. Based on 60 observations, the m_b also is 5.3 (ISC). Because of its relatively small size, it has been difficult to analyze the 1984 earthquake using available analog seismic records. The preliminary focal mechanism determined from P-wave first motions (Gordon, 1987), for instance, is inconsistent with many of these first-motion data (Fig. 2). These inconsistencies have arisen for a number of reasons: (1) the earthquake was not well recorded at teleseismic distances and thus first-motion interpretations, especially for stations near nodal planes, often are equivocal; (2) many close-in stations are located near travel-time triplications, which lead to complicated signals; and (3) lack of resolution arising from recordings from narrow-band, analog instruments. For these reasons, it is preferable to model the rupture process using digitally recorded broadband data. From the major digitally recording seismograph networks, we found only one well-recorded seismogram, at station RSNY, at a distance of 22.6°. We will use the entire suite of amplitude and phase information in the broadband records of this P waveform to model the rupture process, with the intent of using these reliable results to help place limits on the size of the 1882 earthquake.

We processed the RSNY record, using the method of Harvey and Choy (1982), to obtain a broad bandwidth record that was flat to displacement over the frequency range of 0.01 to 5 Hz. We then modeled the broadband data to refine the focal mechanism and to estimate focal depth and source dimensions by using the method of Choy and Kind (1987). For a single shallow source, the far-field P-wave displacement, U, is

$$U(\mathrm{x}, t) = G_P * \Omega_P + G_{pP} * \Omega_{pP} + G_{aP} * \Omega_{sP}, \quad (1)$$

where G_i and Ω_i are the propagation and source operators for body waves $i = P$, pP, and sP. For simple coherent sources, synthetic displacements are computed using triangular source functions. To obtain a propagation operator, we use the full-wave method described by Cormier and Choy (1981) that can account for the interference head waves and diffracted rays that arise from the interaction of the P waveform with the upper mantle. Our velocity model at the source is that of Jackson and Pakiser (1965), which also is the crustal model used for locations of aftershocks in the next section. The T7 model of Burdick and Helmberger (1978) is used for the P-wave velocity in the upper mantle. The T7 model was derived for the western United States, and its crustal and upper mantle velocities coincide with those derived by Jackson and Pakiser (1965) for central Wyoming.

The P-wave displacement and corresponding velocity records for the mainshock are shown in Figure 3. The solid lines are data, and the dashed lines are synthetics. The doublets in each direct P and pP pulse arise from the interference of two $T - \Delta$ triplications. From the differential time between the first arrivals of pP and P of 7.8 sec, we obtain a depth of 27.5 km. The differential T_{pP-P} is resolvable to better than ± 0.15 sec, which constrains the depth to better than ± 1.0 km. The modeling process consists of finding the theoretical seismogram that best fits the data. Theoretical seismograms are computed in a grid search around the strike, dip, and rake of the preliminary P-wave first-motion solution of Gordon (1987), shown by the dashed lines in Figure 2. The focal mechanism that produced the best fit to the waveforms is the oblique, strike-slip solution shown by the solid lines in Figure 2. This focal mechanism also is inconsistent with many of the preliminary P-wave, first-motion data.

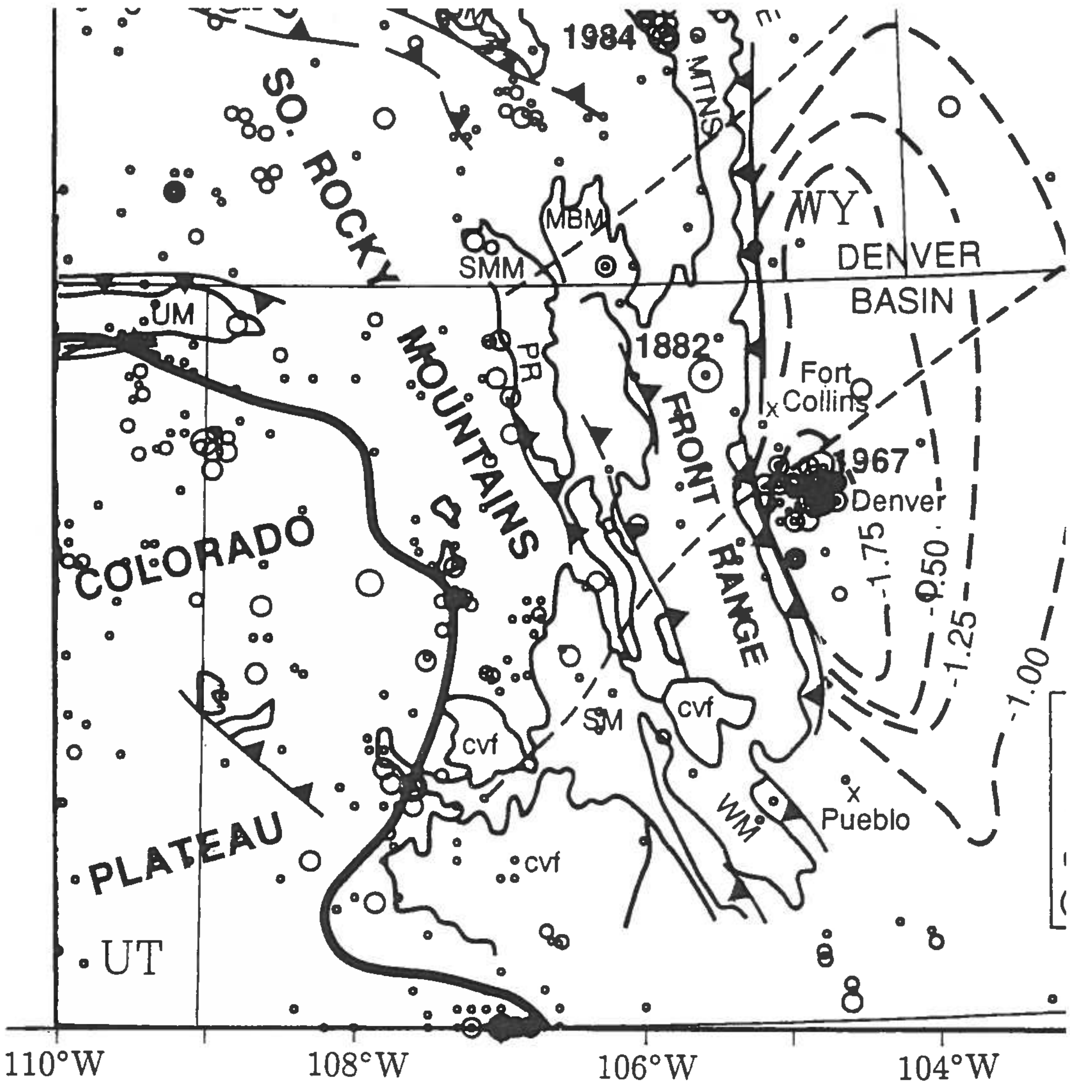

Figure 1. Seismicity (USGS database) and major geological features (from Brewer *et al.*, 1982) of the central Rocky Mountains region. Northeast-trending dashed lines represent northern and southern boundaries of the Colorado lineament in the study region (Warner, 1978). The famous Colorado mineral belt is parallel to but broadly includes the Colorado lineament. Toothed lines represent thrust or reverse faults, primarily of Laramide age and younger. Mountain ranges are OCM, Owl Creek Mountains; WRM, Wind River Mountains; GM, Granite Mountains; BH, Black Hills; MBM, Medicine Bow Mountains; UM, Uinta Mountains; SMM, Sierra Madre Mountains; PR, Park Range; SM, Sawatch Mountains; and WM, Wet Mountains. Intervening sediment-filled basins are described by Dickinson *et al.* (1988). Cenozoic volcanic fields are labeled cvf. Seismicity is diffuse throughout these mountain ranges and the intervening basins. Depth contours (km) of the western Denver basin from MacLachlan and Kleinkopf (1969). The induced Denver earthquakes are east of the Front Range. The 1882 earthquake is shown by large circle in the northern Front Range, and the 1984 earthquakes are in the northern Laramie Mountains.

These inconsistencies are attributable to the reasons given earlier in this section and also because the waveform modeling solution represents the average source properties, whereas the conventional *P*-wave, first-motion solution reflects the very initial source properties. Uncertainty in the waveform-based solution from using a single station is dominated by the level of signal to noise. Synthetics with acceptable amplitude ratios between direct *P* and the depth phases constrain the strike, dip, and rake to better than $\pm 5°$ each. The faulting is predominantly strike slip, with an updip component (right-lateral slip on the ENE-trending plane or left-lateral slip on the NNW-trending plane).

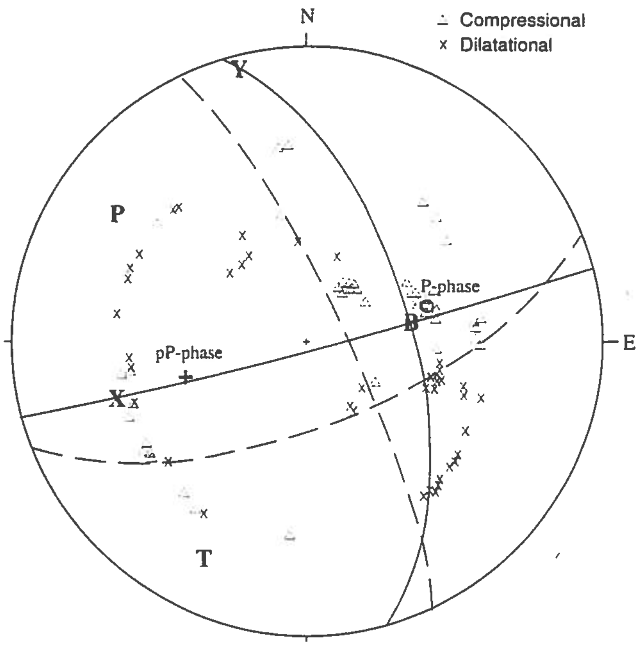

Figure 2. Focal mechanism solutions for 1984 mainshock, shown on lower hemisphere, equal area projection. Preliminary solution of Gordon (1987) is shown by dashed lines, with his accompanying first-motion, *P*-wave data. Final solution (shown by solid lines) is constrained by relative amplitudes of the *P* and *pP* phases at digitally recording station RSNY. Nodal plane *X* strikes N75°E and dips 87 °S; nodal plane *Y* strikes N343°W and dips 60 °E; rake is 210°. Azimuth and plunge of *T* and *P* axes are 205 °SW, 18.4° and 303 °NW, 23°, respectively.

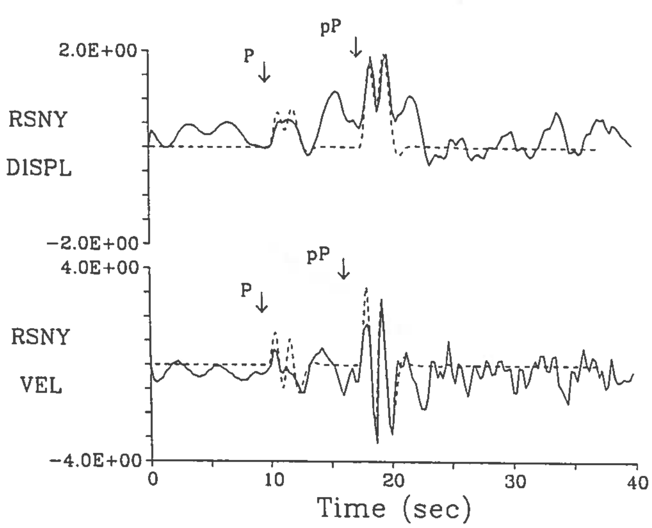

Figure 3. Records of displacement (microns) and velocity (microns/sec) at RSNY (located at the northern edge of the Adirondack State Park, New York), showing arrivals of phases *P* and *pP*. Solid lines indicate data; dashed lines indicate synthetics.

The source function for the *P*-wave has a rise time of 0.5 sec and a total duration of 1.0 sec. The relationship of Das and Kostrov (1986) for rise time of the source function, $\tau \approx \alpha/\upsilon$ (where α is the asperity's radius and υ, the rupture velocity, is assumed to be 75% of the shear wave velocity), implies a small rupture dimension, $D = 2\alpha$, of about 2.7 km. With only one station, it is difficult to quantify directivity and thus choose the causal nodal plane. However, the *pP* arrival, which is an upgoing ray, has a higher-frequency content than the downgoing *P* wave, suggesting again that the rupture had a significant updip component, similar to the rupture for some other intraplate earthquakes (Choy and Bowman, 1990). This rupture is shown schematically on a depth section of aftershocks in Figure 6b.

We obtain an average stress drop using the relationship

$$\langle \Delta\sigma \rangle = 8\mu\langle u \rangle/\pi D, \tag{2a}$$

where the average slip is

$$\langle u \rangle = M_0/\mu A, \tag{2b}$$

with shear modulus μ taken to be 4.0×10^{11} dyne/cm^2, and rupture area, A, assumed to be circular. For $A = 5.7$ km^2, we obtain approximate average slip and average stress drop of 48 cm and 180 bars, respectively. Hence, teleseismic data suggest that the 1984, Laramie Mountains mainshock was a high stress drop earthquake. Two other North American earthquakes in the mid-to-lower crust also have stress drops on the order of 200 bars, similar to that for the 1984 Wyoming earthquake: the earthquakes of southern Illinois (9 November 1968, $M_w = 5.4$, $h = 22$ to 25 km; Somerville *et al.*, 1987) and Saguenay, Quebec (25 November 1988, $M_w = 5.9$, $h = 27$–29 km; Boore and Atkinson, 1992; Hanks and Johnston, 1992; Hartzell *et al.*, 1994; North *et al.*, 1989).

The mainshock's SSW-trending tension axis is similar to that for the stress field within the Colorado Plateau (Wong and Humphrey, 1989; Zoback and Zoback, 1989), in the Colorado Rocky Mountains (Wong and Humphrey, 1989), and for the induced Denver earthquakes (Herrmann *et al.*, 1981). With an average crustal thickness for the Laramie Mountains of 37 km (Prodehl and Pakiser, 1980; Allmendinger *et al.*, 1982), this earthquake's depth of nucleation of 27.5 ± 1.0 km places it in the lower part of the upper crust. Its depth also may qualify the Laramie Mountains earthquake as the deepest known U.S. earthquake with a magnitude exceeding 5.0 and occurring within the North American plate, west of the Mississippi River.

Aftershocks of the 1984 Earthquake

The first station of our temporary seismograph network was operational about 24 hr following the mainshock. Due to various difficulties, it took 3 more days to have a fully operational network. The final network consisted of 16 vertical-component analog and seven three-component digital systems, whose locations are shown in Figure 4. Field operations continued for 6 more days, during which 46 aftershocks were well recorded.

Both *P*- and *S*-wave data were read for all aftershocks. For hypocenter determinations, we used a *P*-wave velocity of 6.2 km/sec for the 37-km-thick crust (Jackson and Pakiser, 1965; Prodehl and Pakiser, 1980) with the computer program HYPOELLIPSE (Lahr, 1984). A Wadati diagram (Fig. 5) indicates that $V_P/V_S = 1.70$. The corresponding *S*-wave velocity of 3.6 km/sec was used to redetermine hypocenter locations with both *P*- and *S*-wave data. Except for one event, these locations are near the center of our network, and the quality factors and error statistics mostly are excellent. Event 10 is located about six mainshock source dimensions southwest of the primary aftershock cluster (Fig. 4) and probably is not an aftershock. The average axis of the 68% confidence ellipsoids is 0.3 km, horizontally, and 0.2 km, vertically; average RMS travel-time residual is 0.03 sec. Aftershock data computed using the single-layer model are shown in Table 1 (Langer *et al.*, 1991).

We recomputed hypocenters using the layered model of Prodehl and Lipman (1989): 4.0 km/sec, 0 to 2 km; 6.3 km/sec, 2 to 30 km; 6.9 km/sec, 30 to 38 km; and 7.9 km/sec half-space. This yielded slightly deeper foci (about 0.3 km overall), while the epicenters and location error statistics remained virtually unchanged from those determined with HYPOELLIPSE. This test assured us that our model of a single layer over a half-space introduced no significant error in the aftershock hypocenters.

Magnitudes are duration-based values, M_d, computed using average network signal durations (Lee and Stewart, 1981) and calibrated to be equivalent to local M_L values. Three of these aftershocks had $M_d \geqq 3.0$. In the 4 days prior to our monitoring, there were five earthquakes with $M_L > 3.0$, the largest being an M_L 4.2 earthquake occurring one-half hour after the mainshock. During the 1 1/2 years after our monitoring, there were four additional aftershocks of M_L 2.9 to 3.3. These pre- and post-network aftershocks are not used in this study because of their relatively inaccurate locations.

Epicentral locations that are based on our network data are plotted by magnitude in Figure 4 and, in more detail, with corresponding error ellipses, in Figure 6a. The epicenters are concentrated in an area measuring about 2 by 4 km, similar to the source dimension computed for the mainshock. In Figure 6b, hypocenters are projected onto vertical sections with the same strike as the nearly vertical fault planes determined for the mainshock. All aftershocks are in the depth range 20 to 25 km. Of the 46 aftershocks, 41 have focal depths between 21.0 and 23.8 km; the average focal depth for this aftershock cluster is 22.5 km. Because the depths of the aftershocks and the mainshock all are well constrained, we infer that the mainshock nucleated at a depth of 27.5 ±

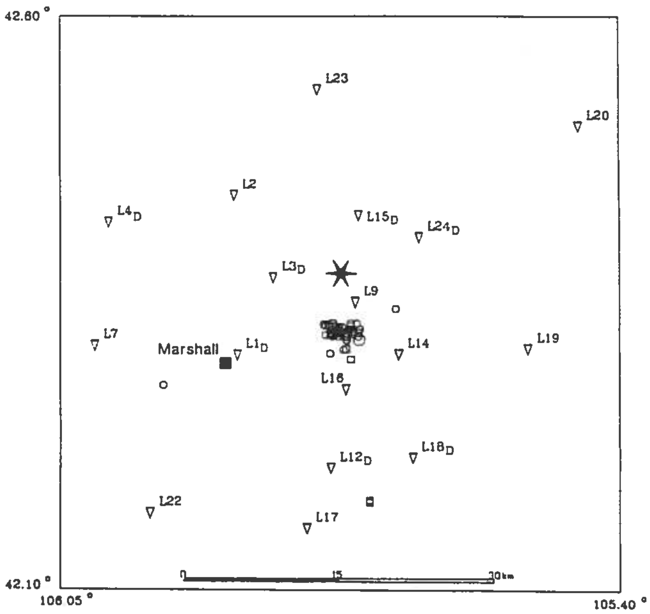

Figure 4. Map of epicenters of 1984 mainshock and aftershocks. USGS mainshock epicenter (star) is biased ~6 km north of the aftershock cluster. Temporary seismographic stations shown by triangles; station numbers without subscripts indicate short-period vertical instruments, and those subscripted by D indicate three-component digital instruments. Aftershock epicenters are keyed to duration magnitude: small squares represent $M_d < 2.0$, small circles represent $2.0 \leqq M_d \leqq 3.0$, and larger circles represent $M_d \geqq 3.0$. The two events southeast of station L12 are quarry blasts (not included in Table 1) and the event southeast of station L7 probably is not an aftershock.

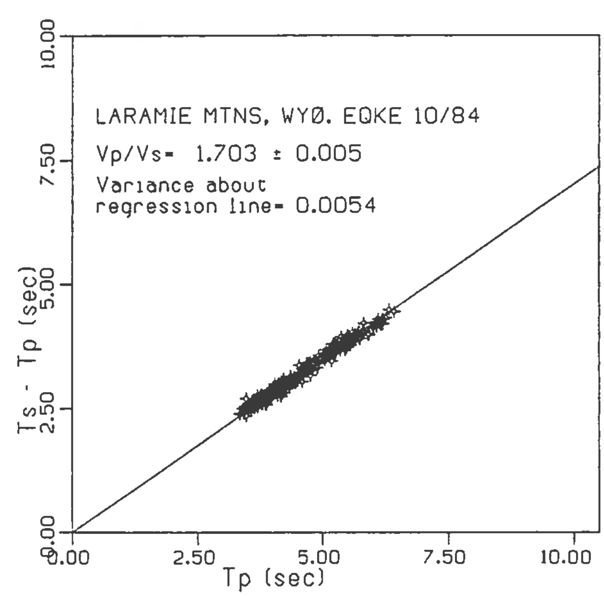

Figure 5. V_P/V_S ratio of 1.703 determined from Wadati diagram of *P*-wave travel time (T_P) vs. $T_S - T_P$ for aftershocks. Data used are restricted to arrivals with *S*-wave travel-time residuals ≦ 0.20 sec and *P*-wave travel-time residuals ≦ 0.10 sec.

Table 1
Laramie Mountains Aftershock Hypocentral Parameters and Duration Magnitudes, Computed using HYPOELLIPSE (Lahr, 1984)

Earthquake Number	DATE Oct-84	Origin Time (UTC) (Hr Min Sec)	Lat. (°N)	Long. (°W)	Depth (km)	Mag. (M_d)
1	22	1846 46.78	42.318	105.716	22.7	2.8
2	22	1852 19.64	42.345	105.658	24.6	2.4
3	22	2007 50.98	42.325	105.709	23.5	2.5
4	22	2316 30.65	42.323	105.709	21.3	2.0
5	22	2323 37.06	42.325	105.726	21.9	2.4
6	23	0015 34.45	42.332	105.711	22.0	2.2
7	23	0534 53.89	42.322	105.740	22.7	2.1
8	23	0549 56.66	42.332	105.704	23.8	2.3
9	23	0848 07.03	42.330	105.735	22.5	1.8
10	23	0934 53.74	42.277	105.933	20.3	2.1
11	23	1049 27.04	42.327	105.712	21.8	2.5
12	23	1143 05.13	42.321	105.725	23.2	2.2
13	23	1301 46.76	42.327	105.700	21.9	2.2
14	23	1333 18.73	42.324	105.702	22.1	2.0
15	23	1458 15.84	42.309	105.719	24.8	2.5
16	23	1543 34.70	42.321	105.732	21.9	2.7
17	23	1944 21.69	42.325	105.729	20.3	2.8
18	23	2223 24.41	42.318	105.700	23.3	3.0
19	24	0203 32.67	42.321	105.702	23.2	2.1
20	24	0408 54.10	42.324	105.733	22.2	2.4
21	24	0719 32.69	42.326	105.722	22.3	3.1
22	24	0832 13.66	42.330	105.736	22.8	2.0
23	24	0903 54.75	42.322	105.717	22.1	3.3
24	24	1439 26.89	42.325	105.727	22.1	2.0
25	24	1538 58.63	42.324	105.735	22.1	2.2
26	24	1813 03.96	42.331	105.742	22.4	2.7
27	25	0034 11.69	42.330	105.740	22.1	2.2
28	25	0817 41.17	42.329	105.729	22.2	2.2
29	25	0927 26.6	42.324	105.722	22.6	1.8
30	25	1233 16.32	42.306	105.734	25.6	2.4
31	26	1356 44.41	42.323	105.732	21.3	1.8
32	26	1358 39.49	42.331	105.737	22.3	2.2
33	26	1429 16.98	42.324	105.727	22.5	2.0
34	27	1404 24.52	42.325	105.713	23.3	2.3
35	27	1535 22.28	42.330	105.710	21.4	2.0
36	27	2055 08.34	42.320	105.717	22.9	1.6
37	27	2134 14.19	42.325	105.721	21.4	2.3
38	27	2156 06.56	42.326	105.707	21.7	2.3
39	28	0149 24.72	42.310	105.715	23.0	2.0
40	28	0428 33.01	42.332	105.731	22.6	1.2
41	28	0443 13.95	42.328	105.725	21.8	2.0
42	28	0516 51.29	42.321	105.735	22.2	1.6
43	28	0951 12.43	42.301	105.710	24.0	1.5
44	28	1352 17.51	42.323	105.727	21.2	2.5
45	28	1600 58.31	42.323	105.716	20.8	2.7
46	28	1749 46.56	42.318	105.715	22.6	2.4
47	28	1928 01.47	42.317	105.715	22.7	2.2

1.0 km and ruptured updip 2.5 to 3.0 km to beneath what became the zone of aftershocks. It is observed generally that aftershocks occur on the periphery of asperities that fail in the mainshock (Mendoza and Hartzell, 1988; Choy and Bowman, 1990; Boyd *et al.*, 1995; Langer and Spence, 1995), consistent with the hypothesis that aftershocks are due to a redistribution of stress following the rupture of controlling asperities. However, our results and those for the Saguenay, Quebec, earthquake of 25 November 1988 (m_b = 5.9, h = 29 km; North *et al.*, 1989) suggest that aftershocks of deep, intraplate earthquakes can occur primarily above the mainshock's rupture.

The broad aperture of our seismic network relative to the aftershocks allowed computation of single-event focal mechanisms for each aftershock (Langer *et al.*, 1991). These focal mechanisms indicate a nearly even distribution between strike-slip and normal faulting. The focal mechanisms exhibit stable T axes, trending NE–NNE with subhorizontal plunges, similar to the T axis for the mainshock. The P axes are stable azimuthally, trending WNW–NW, but have

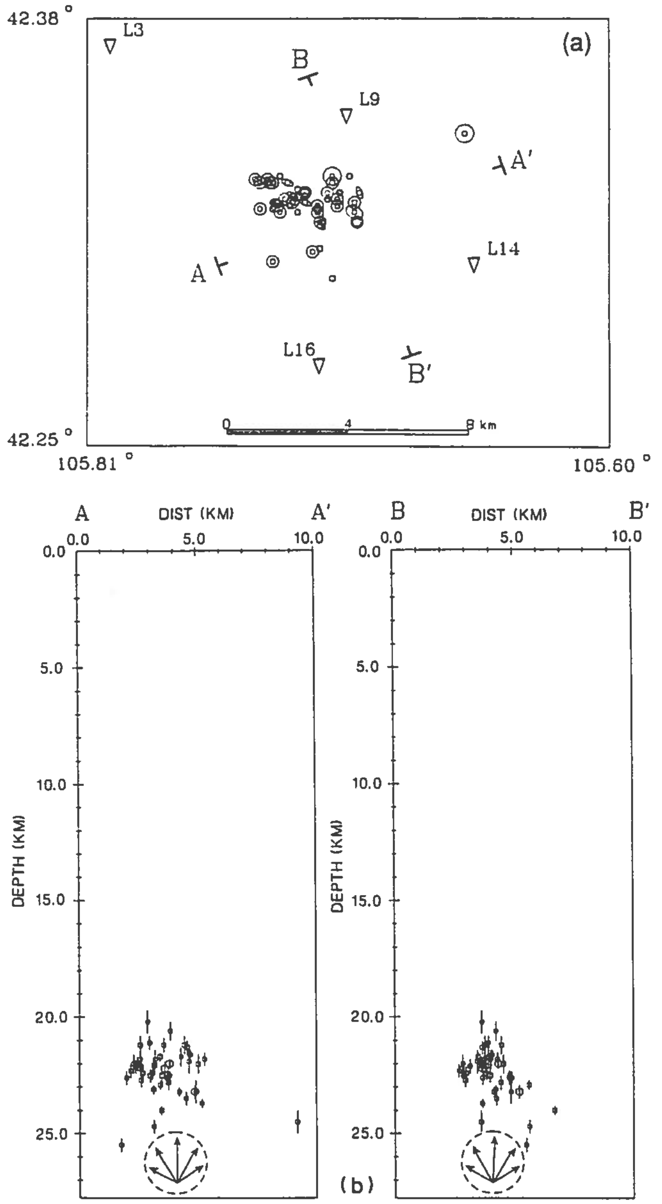

Figure 6. (a) Aftershock epicenters plotted with corresponding error ellipses. A-A′ and B-B′ coincide with orientations of the nodal planes for the mainshock, on which aftershock hypocenters are projected. (b) Vertical section plots of hypocenters projected onto planes having strikes the same as nodal planes for the mainshock. Hypocenter symbols keyed to magnitudes, as in Figure 4; vertical error bars based on 68% confidence ellipsoids (Lahr, 1984). Dashed circle beneath aftershocks shows zone containing mainshock rupture, inferred from analysis of broad bandwidth, digital data of station RSNY.

plunges ranging from subhorizontal to subvertical, corresponding to strike-slip and normal faulting, respectively. The mutual occurrence of both strike-slip and normal faulting suggests that the magnitudes of the intermediate and maximum principal stresses are very close to the same value.

Comparison of the 1984 and 1882 Earthquakes

Because our knowledge of the 1882 event is derived from sparse intensity data that primarily were observed in the geologically complex Rocky Mountains, it is important to interpret these data in the context of intensity data from more recent earthquakes. The deep crustal focal depth and high stress drop of the 1984 earthquake requires us to consider similar values for the 1882 earthquake. Most prior estimates of the felt area for the 1882 earthquake were about 470,000 km^2, with an associated magnitude of M_L 6.2 $\pm$ 0.2 (Dames & Moore, 1981; McGuire *et al.*, 1982; Kirkham and Rogers, 1985, 1986). Under the constraint of this total felt area and using four seismologists to independently recontour the intensity data of the 1882 earthquake, Bollinger (1994) estimated the most likely magnitude of this earthquake to be M_W 6.4 $\pm$ 0.3. The remainder of our article addresses intensity data for the 1882 and modern earthquakes, leading to new constraints on the source properties of the 1882 event.

Intensity Map for the 1984 Mainshock

Figure 7a is a map of Modified Mercalli (MM) intensities for the 1984 mainshock, derived from several hundred standard intensity questionnaires from within a 400-km radius of the epicenter, sent out immediately after the earthquake (Stover, 1985). Intensities have been contoured into a map of isoseismal lines, showing a felt area of 287,000 km^2 and a moderate-sized zone of maximum intensity VI. This intensity map has good radial symmetry, indicating that attenuation and site response vary only moderately in the region surrounding the Laramie Mountains.

However, this felt area is anomalously large for a magnitude 5.3 earthquake in this region. Kirkham and Rogers (1985) performed a regression analysis between felt area and magnitude, M_L for 58 earthquakes felt in Colorado, of magnitudes $2.5 < M_L \leqq 7.7$ and depths mostly $\leqq$ 10 km, giving the relationship

$$M_L = 1.22 \log_{10} A_{\text{felt}} - 0.68. \qquad (3)$$

Because M_L here is normalized to magnitudes determined from regional and distant seismic stations, this relationship roughly accounts for average attenuation of intensity for the region. Equation (3) implies that a typical M_L = 5.3 earthquake in the Southern Rocky Mountains region would have a felt area of 79,700 km^2; thus the observed felt area of 287,000 km^2 is 3.6 times what would be expected. Conversely, using equation (3) with the observed felt area implies M_L = 6.0, or 0.7 of a magnitude unit greater than the actual value. In contrast, for the two largest 1967 Denver earthquakes (Fig. 7b), equation (3) yields M_L values equal to the observed m_b values of 5.0 and 5.1. Can the discrepancy for the 1984 mainshock be explained by this earthquake's unusually deep focal depth or its high stress drop? From purely geometric considerations, the effects of mid to deep crustal source depths include diminishment of a sharply defined inner zone of highest intensity (for distances less than about 10 times the focal depth), and a broadening of the remaining high-intensity zones, even though the total felt area is essentially the same as for a very shallow shock of the same magnitude and stress drop (Evernden and Thomson, 1988).

Theoretical work by Boore (1983) relates maximum acceleration, α_{max}, to M_W and $\Delta\sigma$:

$$\log a_{max} \sim 0.31\ M_W + 0.80\ \log\Delta\sigma. \qquad (4)$$

This implies that a factor of 2 increase in $\Delta\sigma$ would be offset by a magnitude reduction of 0.77 to maintain the same α_{max}. Assuming that the perception of an earthquake's intensity is related to the accelerations that it produces, equation (4) implies that the felt area is a strong function of $\Delta\sigma$. Thus the magnitude discrepancy of 0.7 of a magnitude unit implied by equation (3) for the 1984 earthquake appears explainable by its stress drop being about twice that of the ~100 bars for an average shallow earthquake.

Intensity Map for the 1882 Mainshock and Arguments for Its Epicenter Being in the Front Range

Local intensities reported for the 1882 earthquake have been documented in a series of studies by Hadsell (1968), Docekal (1970), Dames & Moore (1981), McGuire *et al.* (1982), Kirkham and Rogers (1985, 1986) and Oaks and Kirkham (1986). Even after these multiple efforts to procure additional intensity data for this earthquake, the final data set does not provide a clear picture of this important event. The sparseness of regional population, poor construction practices, and other factors have led to particular difficulty in interpretation of this earthquake's intensity data. A typical intensity map developed in these studies for the 1882 earthquake is shown by the long, dashed curves in Figure 8.

In historical perspective, this earthquake occurred during the second opening of the American West (Stegner, 1953). Deposits of gold and silver drew fortune seekers into Colorado's mountains, along the Colorado mineral belt (Fig. 1). Population was concentrated in the mountains' mining camps and the corridor that bounds the Front Range to the east; only Denver, Colorado Springs, Leadville, and Silver Cliff exceeded 4000. Population in likely epicentral zones was very low (McGuire *et al.*, 1982; Kirkham and Rogers, 1986). The earthquake was reported along the Union Pacific line (Fig. 8), but there was little settlement (and virtually no surviving intensity data) farther to the north (McGuire *et al.*, 1982). Likewise, there is a dramatic lack of intensity data in eastern Colorado and western Kansas and Nebraska. Settlements were still operating on local, sun time, and special care was needed to reduce reports from this earthquake to standard times. Interestingly, the mainshock occurred at about 6:30 p.m. Denver time on a mid-term, national election day, a situation that both competed with and led to some dramatization of reporting of the earthquake.

Literal interpretations of the earlier intensity map led to the conclusion that the epicenter of the 1882 mainshock was near the center of this map, in western Colorado. There was an account of high intensity reported by mountain travelers in northwestern Colorado, with reported rockfalls and an immense, smoking crater. This account was not corroborated at the time. Moreover, extensive field investigations (e.g., McGuire *et al.*, 1982) failed to identify a geologic candidate for the reported zone, and an instrumental seismic study for the suspect area failed to show modern microearthquake activity (Clift, 1986).

Rockfalls or small landslides were reported in two Front Range locations near Pueblo, at DeBeque Canyon east of Grand Junction, and at the notoriously unstable slopes of the Douglas Pass area, in northwestern Colorado. Three of the four rockfall sites have been assigned intensity as felt (*F*) (McGuire *et al.*, 1982; Kirkham and Rogers, 1986), because of a lack of independent data to assign a numerical intensity and a recognition that rockfalls can be triggered by low levels of ground shaking. The rockfalls at Debeque Canyon and Douglas Pass thus do not provide support for the 1882 earthquake being in western Colorado.

The 1882 mainshock does not show a clearly defined area for the highest reported intensity (Fig. 8). Intensity MM VII was observed at several scattered locations in Denver, near Pueblo, and, possibly, near Fort Collins (McGuire *et al.*, 1982; Kirkham and Rogers, 1985, 1986). At Denver's electricity-generating plant, a 1-in. driveshaft bolt was snapped and another was bent. In parts of northern Denver, the earthquake was felt so strongly that many families ran from their houses, whereas in other parts of northern Denver, the earthquake was barely felt (Kirkham and Rogers, 1986). The MM VI zone has remarkable E–W extent, due largely to reports from along the Union Pacific rail line. In the Salt Lake Valley, Utah, there were several observations of MM IV to V. The most consistent observations of high intensity exist for the northern Front Range: in Laramie, near Fort Collins, Boulder, and Denver (McGuire *et al.*, 1982; Kirkham and Rogers, 1986).

Of the prior investigations of the 1882 earthquakes, the studies of Kirkham and Rogers (1985, 1986) are based on the largest set of intensity data and contain the most internally consistent interpretations. They provided key evidence to constrain the 1882 mainshock's location by utilizing intensity data for a large aftershock (magnitude about 5) occurring on the morning of 8 November 1882, about 10½ hr after the mainshock. These intensity observations (encircled data in Fig. 8) were used to estimate the aftershock's primary felt area, as shown by the screened oval in Figure 8. This aftershock also was felt near Meeker (40 °N, 108 °W), although no intensity was assigned. It is reasonable to place the epicenter of this moderate aftershock near the center of its primary felt area, as observed for the 1984 earthquake. A shallow earthquake of magnitude 6.5 to 7.0 typically would have a rupture length of 10 to 20 km (e.g., Tocher, 1958); a deeper earthquake would have a smaller rupture dimension. Because aftershocks occur near the edges of mainshock rup-

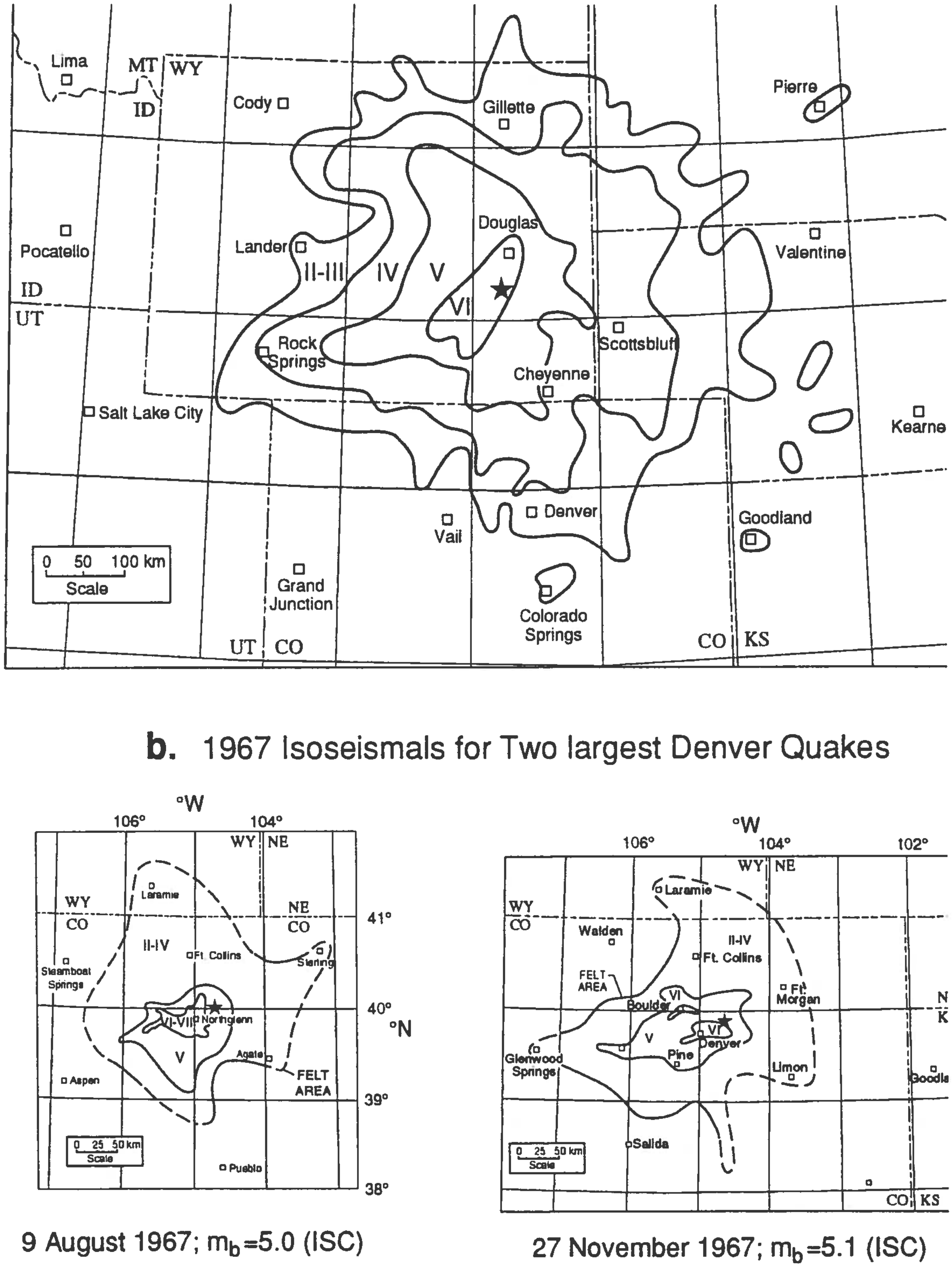

Figure 7. (a) Isoseismal map for the 1984 Laramie Mountains mainshock (Stover, 1985); epicenter shown by star. Felt area is 287,000 km^2. General circular shape of intensity map indicates that attenuation and site response do not have large regional variation. (b) Isoseismal maps for the two largest of the Denver earthquakes (Kirkham and Rogers, 1985); epicenters (ISC) shown by stars. These shallow earthquakes (depths < 5 km) have respective felt areas of 50,000 and 56,000 km^2 [maps plotted at same scale as (a)]. Note irregular felt areas, E–W extended zones of strong shaking, and felt areas extending to the north and east, into the Denver basin. See text for discussion.

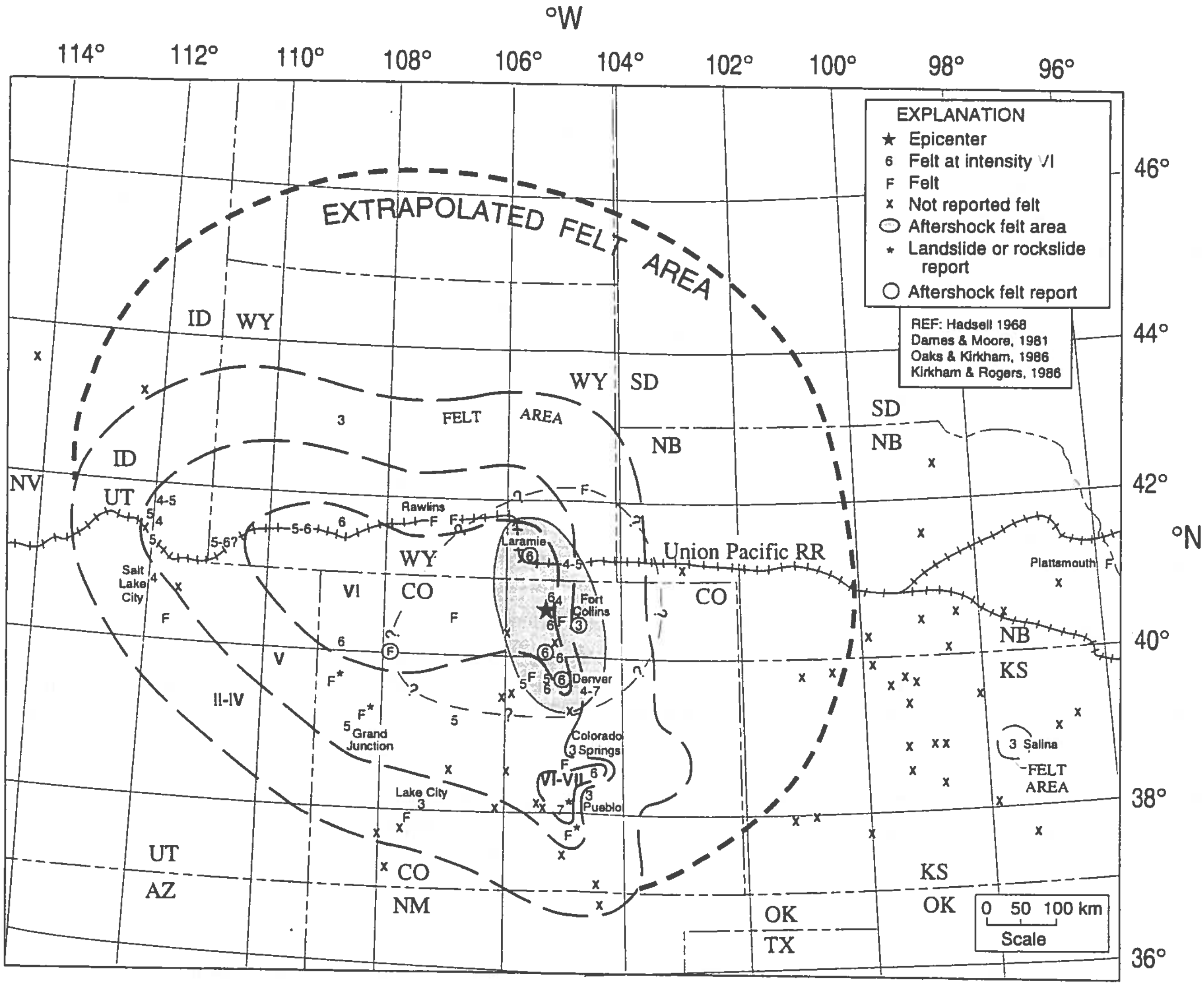

Figure 8. Isoseismal map for the 7 November 1882 Front Range mainshock (plotted at same scale as Figs. 7a and 7b). Interpretation of Kirkham and Rogers (1985, 1986) shown by long, dashed isoseismals, which are flattened at the north and east [area 470,000 km^2; M_L 6.2 ± 0.2)]. Route of Union Pacific Railroad taken from Powell (1879). Modified Mercalli (MMI) intensities shown by Arabic numbers and contoured intensity zones by Roman numerals. Specific locations *not reporting* feeling the earthquake shown by x's. As discussed at length in Kirkham and Rogers (1985, 1986), x's do not necessarily imply that in fact the earthquake was not felt there. Similarly, an intensity of F does not mean that the earthquake was felt minimally, but rather that it is not known to what degree the earthquake was felt. Kirkham and Rogers' estimate of the felt area of primary aftershock on 8 November shown by stippling, with the - -?- -?- - contour representing a possible limit to which this aftershock could have been felt, given sufficient population to the east and north. Kirkham and Rogers place the epicenters of this aftershock and mainshock at the center of the stippled zone, at 40.5 °N and 105.5 °W, with uncertainties of at least 0.5°. Our estimate of the potential felt area of the 1882 mainshock, given adequate reporting population to the north and east, is shown by the heavy dashed curve that extends the former maps in those directions. Revised area is 850,000 km^2, and our estimated magnitude is M_W 6.6 ± 0.6.

tures (e.g., Mendoza and Hartzell, 1988; Choy and Bowman, 1990; Boyd *et al.*, 1995; Langer and Spence, 1995), an aftershock of the 1882 mainshock would be expected to be within 20 km of the mainshock's epicenter. This inference that the mainshock was fairly near the center of the aftershock's felt area, in the northern Front Range, agrees with the conclusion of Kirkham and Rogers (1985, 1986), who assigned locations for both the aftershock and mainshock at 40.5 °N, 105.5 °W, with an accuracy no greater than $\pm 0.5°$.

Although negative evidence, two other lines of reasoning support the interpretation that the 1882 earthquakes were not in western Colorado. First, earthquake lights (Derr, 1973) were observed near the time of the mainshock at Greeley and at Long's Peak (both near Fort Collins), but not in western Colorado (McGuire *et al.*, 1982; Kirkham and Rogers, 1986). Also, Grand Junction, a city in western Colorado that is known for particularly high site response (Kirkham and Rogers, 1985, p. 97), reported the mainshock only at intensity MM V to VI (Kirkham and Rogers, 1986; Dames & Moore, 1981) and did not report feeling the large aftershock.

If the mainshock was in the Front Range region, then the prior intensity map (Fig. 8) is anomalous with regard to recent models of attenuation and with intensity maps for more recent earthquakes of the region. Seismic attenuation ($1/Q$) is known to decrease from west to east in the conterminous United States. Based on a Q-dependent scattering model to explain the coda waves of local and near-regional earthquakes, Singh and Herrmann (1983) approximated Q by Q_0 and developed a crude regionalization where Q_0 increased from about 400 in the Colorado Plateau to 600 to 800 in the Rocky Mountains and the High Plains. The prior intensity map for the 1882 earthquake, given that this earthquake originated in the northern Front Range, contradicts such models of attenuation for this region, as its isoseismals are elongated into the zone presumed to have high attenuation and are collapsed in front of the zone thought to have low attenuation.

The intensity maps of McGuire *et al.* (1982) and Kirkham and Rogers (1985, 1986) for the 1882 earthquake (Fig. 8) are flattened to the north and east, particularly when compared with the isoseismals of the smaller earthquakes of 1967 and 1984 (Fig. 7). This suggests that the 1882 felt area relative to an epicenter in the northern Front Range would have been much greater than shown by these earlier maps, had sufficient population existed to the north and east. A felt report far to the east (at the second floor of a brick building) at Salina, Kansas (McGuire *et al.*, 1982), was confirmed by Oaks and Kirkham (1986); also, there is a felt report north of Salina, at Plattsmouth, Nebraska. These extreme eastern felt reports lie east of sparsely populated sites for which this earthquake was "not reported" to have been felt (see caption for Fig. 8).

We project that had significant population existed to the north and east, the heavy, dashed line in Figure 8 would reasonably approximate the isoseismal for the felt area of the 1882 earthquake. This projection is more consistent with the shapes of the isoseismal maps for the 1967 and 1984 earthquakes. Factors that effect estimation of the potential felt area of the 1882 earthquake from intensity data, other than a very heterogeneous regional population, include poor construction practices and site responses complicated by various geological effects. The eastern radius of this isoseismal would be about 2/3 the western radius, constrained by the cluster of "not reported felt" data west of Salina, Kansas, and similar to the ratio of corresponding radii for the 1967 earthquakes. No attempt is made to sketch corresponding isoseismals for zones of higher intensity. In the drawing of the 1882 earthquake's potential felt area, care is taken not to violate the group of "not reported felt" locations to the west and north of Salina any more than the earlier maps violated such data to the south. The extrapolated potential felt area for the 1882 mainshock is 850,000 km^2.

Unresolved Aspects of the Intensity Distribution of the 1882 Earthquake

That the extrapolated eastern radius of the 1882 intensity map is less than its western radius still contradicts our notion of seismic attenuation in this region. It is possible that attenuation is modified by features such as the enormous Denver basin (130,000 km^2, see Fig. 1), which extends hundreds of kilometers to the east (MacLachlan and Kleinkopf, 1969; Tweto, 1975). Kirkham and Rogers (1985) suggest the higher apparent eastward attenuation of energy from the 1967 Denver earthquakes may be due to the acoustically soft, 1000-m-thick Pierre shale of the Denver basin, the underlying additional 500 to 1000 m of sediments, and the deeper deformed basement. Conversely, it is possible that relatively lowered attenuation to the west could be linked with the Colorado lineament and mineral belt (Fig. 1). The intensity map for the 1984 earthquake, which is north of both the Denver basin and the Colorado lineament and mineral belt, shows a much greater radial symmetry than the maps of the 1967 earthquakes or that of the 1882 earthquake.

Although the E–W elongation of the inner isoseismals of the 1882 earthquake was influenced by observations of high intensities along the Union Pacific railway, this elongation may be real. The intensity maps of the 1967 Denver earthquakes, which are based on comprehensive sets of observations, also show pronounced westward extension of their high-intensity zones (cf. Figs. 7b and 8). The westward extending zones of higher intensities cut across the tectonic grain of the Front Range and along the Colorado mineral belt. In addition, to test a possible link between intensity distribution and radiation pattern, we examined both the *P*-wave and *SH*-wave radiation patterns (the latter because the *SH* wave may be directly linked to degree of ground shaking) for the 1967 and 1984 earthquakes. The radiation patterns from the normal-faulting 1967 earthquakes (Herrmann *et al.*, 1981) do not provide explanation of the E–W-trending contours of the higher intensities, for although their observed Love waves and the theoretical *SH* waves show E–W-trending radiation lobes, there are equally strong radiation lobes that are N–S trending. The general circular symmetry of intensity observed for the 1984 event (Fig. 7a) implies little dependency on its focal mechanism (Fig. 2).

Kirkham and Rogers (1985) note that numerous Colorado earthquakes show intensity patterns that indicate narrow felt zones clearly separated by "not felt" reports. They suggest that the E–W-trending, narrow felt zones result from seismic energy that has undergone soil amplification in the populated valleys of streams that drain the Front Range eastward from the Continental Divide. The suggestions that the E–W elongation of intensities may be explained by their association either with soil amplification of fluvial deposits or with the Colorado mineral belt could benefit from further study. Also, there is suggestion that soil amplification may occur where the Denver basin pinches out. This could explain the high intensity observed for the 1984 earthquake at the boundary between the Denver basin and the foothills in Golden, Colorado (Stover, 1985), nearly 300 km south of that epicenter and the high intensities observed near Pueblo from the 1882 earthquake.

Could the 1882 Earthquake Have Been Mid-crustal?

Before we attempt to use the projected larger felt area of the 1882 earthquake to estimate its magnitude, we ask if this earthquake could reasonably have originated in the mid-to-deep crust and have had a high stress drop, analogous to the 1984 earthquake. Most earthquakes of the Intermountain West occur in the upper crust, mostly shallower than 10 km (Smith and Bruhn, 1984; Wong and Chapman, 1990). However, there are examples of earthquakes in the lower crust and upper mantle of the western United States, as summarized by Wong and Chapman (1990). These include the earthquakes of Crownpoint, northwestern New Mexico ($h = 44$ km, $M_L = 4.2$; $h = 41$ km, $M_L = 4.4$), Randolph, northern Utah ($h = 90$ km, $M_L = 3.8$), many small earthquakes throughout the 50-km-thick crust and some in the upper mantle of the central Colorado Plateau, and the 1984 Laramie Mountains event.

Hanks and Johnston (1992) tentatively associate larger-than-usual stress drops with greater-than-usual focal depths for continental earthquakes. Not only will these high stress drop earthquakes exhibit broadened felt areas but, because of their depth, the zones of higher intensities also will be broadened, as we found for the 1984 Laramie Mountains earthquake. Anomalously large felt areas also were observed for the deep crustal earthquakes of southern Illinois and Saguenay. The felt areas of the latter earthquakes are compared to M_W in Figure 9. Both events occurred in the stable continental region, which is characterized by lower attenuation than the region of the 1984 earthquake. The lack of known surface faulting associated with the earthquake of 1882 (Kirkham and Rogers, 1986), the broadened character of its isoseismal maps, and the lack of sharply defined zones of highest intensity all imply that the 1882 earthquake also could have originated deep in the Earth's crust, as suggested by Oaks *et al.* (1985) and Kirkham and Rogers (1985, 1986), and have had a high stress drop.

Theoretically, the maximum depth at which lithospheric earthquakes are possible is governed by the temperature vs. depth profile. Beyond a critical temperature at some depth, deformation is accommodated by ductile flow rather than brittle failure. The source zone of the Laramie Mountains mainshock appears to be in a region of low heat flow, where the temperature at the 1984 hypocenter probably is less than 300 °C (Wong and Chapman, 1990; Decker *et al.*, 1980). The heat flow in the Front Range, as corrected for near-surface radioactive sources, is about the same as for the Laramie Mountains (50 to 60 mW/m^2) (Decker *et al.*, 1984). Knowing that the 1984, Laramie Mountains earthquake originated at about 27.5 km depth in a 37-km-thick crust (Prodehl and Pakiser, 1980; Allmendinger *et al.*, 1982), it is plausible that the 1882 earthquake was at least as deep in the 52-km-thick crust (Prodehl and Pakiser, 1980) of the Front Range.

Size of the 1882 Mainshock

We use Figure 9 (based on Hanks and Johnston, 1992) for the analysis of magnitude estimation rather than equation (3) because it is based on a comprehensive, global data base for stable continental regions and a large data base for California earthquakes. The resulting slopes of log(A_f) vs. M_W thus are better constrained than that of equation (3), which has a steeper slope. This constraint is important because equation (3) is based on relatively sparse data for the Rocky Mountain region, and these data are subject to effects of soil amplification, possible topographic effects, and other causes of variations in attenuation.

In Figure 9, we plot the 1984 Laramie Mountains earthquake based on its magnitude, M_W, of 5.3. We also plot a corresponding point at magnitude 6, for an earthquake of the same felt area but with shallow depth and typical stress drop, as implied by equation (3). Through these two points, we have drawn lines roughly parallel to the graphs for area vs. magnitude for earthquakes in the stable continental region (upper, dotted line) and, for comparison, earthquakes in California (lower, dashed line). Our two solid lines bound the limits for felt area vs. magnitude for earthquakes in the Southern Rocky Mountains with stress drops that vary from about 180 bars (upper line) to about 100 bars (lower line).

We cannot reasonably exclude the possibility that the 1882 earthquake was a mid-crustal, high stress drop event like the 1984 earthquake. This possibility would lead to a minimum magnitude estimate ($M_W \sim 6.3$). Also, even in the absence of known surface faulting that can be associated with the 1882 earthquake, we cannot exclude the possibility that this earthquake was fairly shallow (with $M_W \sim 6.9$). To estimate the magnitude of the 1882 earthquake, we assume that this earthquake's focal depth and stress drop were intermediate between those of the 1984 earthquake and typical shallow earthquakes of the region. Using the potential felt

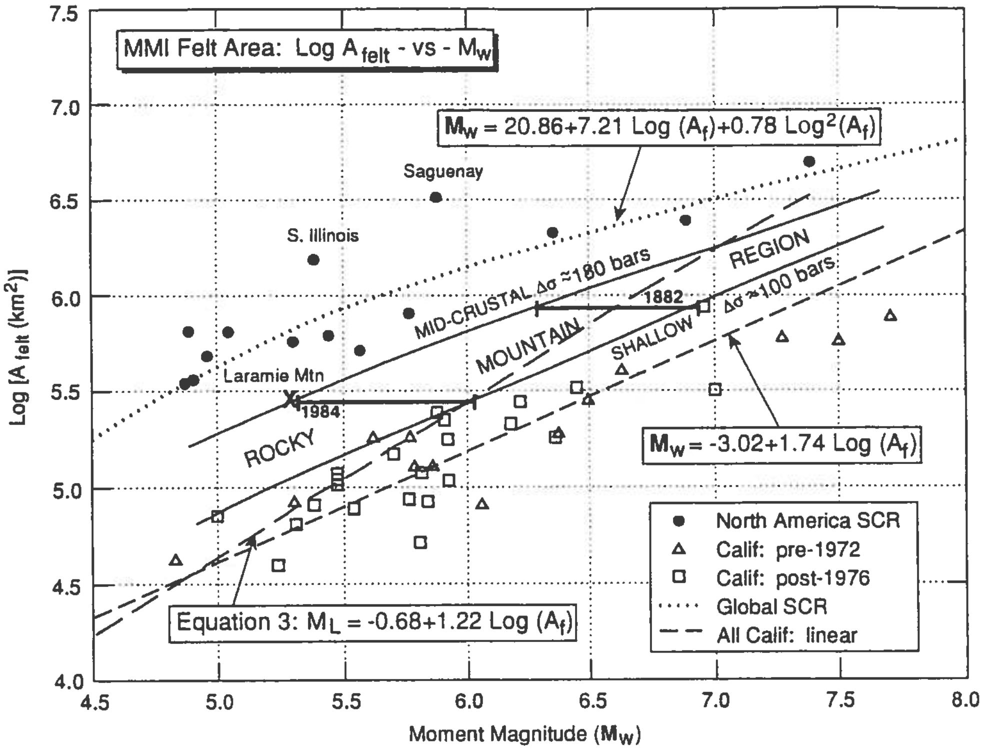

Figure 9. Log (felt area, A_f) vs. M_W for the earthquakes of this study relative to that for earthquakes of the stable continental region, SCR (upper, dotted line), and for California (lower, dashed line). Two inner, subparallel solid lines bound log(A_f) vs. M_W inferred for earthquakes in the Southern Rocky Mountains. The slopes of these lines are constrained by the slopes of the SCR and California lines, because the latter are similar to each other and are based on a far more robust data set than is available for the southern Rocky Mountains. The Southern Rocky Mountains data yield equation (3), which is plotted by the short dashed line that cuts across SCR and California data (see text); here M_L is taken as roughly equivalent to M_W. The lower solid line corresponds to earthquakes of typical focal depth (~10 km) and typical stress drops of about 100 bars. The upper solid line corresponds to earthquakes of deep crustal focal depths (~25 to 30 km) and stress drops of about 180 bars. These lines are positioned by analysis of the 1984 Laramie Mountains earthquake; the horizontal bar from the point of the actual 1984 earthquake extends to M_W 6.0, which is the magnitude that its felt area would correspond to for a shallow crustal earthquake of typical stress drop. An interpolation midway between the solid lines is used to infer M_W 6.6 for the 1882 earthquake. Base figure adapted from Hanks and Johnston (1992).

area for the 1882 earthquake of 850,000 km^2 and taking the point midway between the solid lines in Figure 9 implies $M_W = 6.6$. We note that using equation (3) for local magnitude, M_L, fortuitously also yields magnitude 6.6.

Uncertainties in estimation of magnitude based on felt area arise from uncertainties in our estimates of focal depth, stress drop, and regional attenuation. The range of uncertainty due to our estimates of depth and stress drop, shown by the spread to the regional lines in Figure 9, is ± 0.3. In Figure 9, we also assumed that the same attenuation applies to the total felt area for the 1882 event. The variability of individual intensity maps for many regional earthquakes indicates that we lack a detailed knowledge of regional attenuation. For example, we have not considered possible high attenuation in northwestern Wyoming, suggested by closely spaced isoseismals there for the M_S 7.5 (M_L 7.7), 1959 Hebgen Lake earthquake (Steinbrugge and Cloud, 1962). Also, Figure 9 employs a near-linear relationship between log(A_f) vs. M_W; other relationships exist (Johnston, 1996; Frankel, 1994), but for our purposes, Figure 9 permits a reasonable and straightforward analysis. We estimate an uncertainty of ± 0.2 magnitude units to account for unknown variations in regional attenuation and assumptions in using Figure 9. Finally, even if we could accurately account for variations in regional attenuation, the true felt area might vary from our estimate by $\pm 20\%$ or more ($\pm 170{,}000$ km^2 or more), giving an additional magnitude uncertainty of ± 0.1. Taking the root of the sum of the squares of these largely independent uncertainties, our estimate of the magnitude for the 1882 mainshock is M_W 6.6 ± 0.6.

Discussion

The NE–SW trending T axes for earthquakes within the Colorado Plateau and western Colorado, the Denver earthquakes, and the 1984 earthquake (Wong and Humphrey, 1989; Zoback and Zoback, 1989) imply a small amount of tectonic extension. At first this may seem puzzling, because the North American lithospheric plate is translating west-southwestward, implying ENE–WSW coherence of P axes for regional earthquakes. One explanation for this extension is that the uplifted zone of the Colorado Plateau and Southern Rocky Mountains is undergoing gravitational spreading that is independent of, but riding upon, the larger-scale translation of the North American plate. Much of the modern terrain of the Colorado Plateau and the Rocky Mountains is the result of 1.5 to 3.0 km of uplift of unknown origin, primarily during Laramide and post-Laramide times (Dickinson *et al.*, 1988; Morgan and Swanberg, 1985; Epis and Chapin, 1975; Gregory and Chase, 1994). The initial Laramide uplift of the Front Range was Late Cretaceous (Tweto, 1975), while that of the Laramie Mountains was about 10 m.y. later, in Late Paleocene (Blackstone, 1975). The model of gravitational spreading has been applied to the more mature extension that has resulted from uplift in the Basin and Range province (Coney, 1987; Sonder *et al.*, 1987). Such extension in the Colorado Plateau and Southern Rocky Mountains (Wong and Humphrey, 1989; Jones *et al.*, 1996) may bridge into the Laramie Mountains and the Front Range, which are the deeply rooted eastern limit of the Laramide deformation front. Although the 1984 and 1882 earthquakes may reflect deep-seated extension in this region, we see no obvious reason to exclude the possibility of shallow earthquakes due to this extension, and future earthquakes in the Laramie Mountains or the Front Range may possibly occur from shallow to mid-crustal depths, approximately from Casper, Wyoming, to Pueblo, Colorado (see Fig. 1).

Given that most earthquakes of the Intermountain West are in the upper crust, it is anomalous that the two largest earthquakes of the Front Range and Laramie Mountains may have occurred in the deep crust. Modern seismic monitoring indicates a low-to-moderate rate of seismicity in the Front Range and Laramie Mountains. An assessment of seismicity in the Southern Rocky Mountains of Colorado indicates a recurrence interval of 1400 $\pm$ several hundred years for a magnitude 6.5 earthquake (Wong *et al.*, 1994; I. Wong, personal comm., 1995). The 1882 and 1984 earthquakes, which almost certainly involved the rupture of distinct zones, themselves are statistically insufficient to make meaningful statements about the probability of future damaging mid-crustal earthquakes at the Laramide deformation front. However, the rate of deformation in the Front Range and Laramie Mountains may be influenced by the boundary condition of the eastern limit of the Rocky Mountains. For example, the great extension at the southern Rio Grande rift has been argued to be related to its boundary condition as the eastern edge of the Basin and Range (see Spence and Gross, 1990).

Conclusions

The 18 October 1984, Laramie Mountains, Wyoming, earthquake occurred at a depth of 27.5 $\pm$ 1.0 km. This magnitude 5.3 earthquake had an average stress drop of about 180 bars and an average fault slip of 48 cm, distributed over a source area of about 5.7 km^2; the aftershocks were located just above the area that ruptured during the mainshock. The mainshock's high stress drop accounts for a felt area about 3.5 times greater than would be typical for a shallow, magnitude 5.3 earthquake of the Southern Rocky Mountains. The 7 November 1882 earthquake is Colorado's largest known historic earthquake. Impediments to the correct interpretation of the intensity data of this historic earthquake have been addressed. Our methodology involves considering regional variations in site response and attenuation, knowledge of regional tectonics, and inferring possible source properties by analogy with recent earthquakes. Our inferences are that the 1882 mainshock was located in the northern Front Range and had a magnitude of $M_W = 6.6 \pm 0.6$. Future similar earthquakes throughout the region of the Laramie Mountains and the Front Range pose significant seismic hazard to structures and lifelines of this rapidly developing region. Although such earthquakes may be as large as magnitude, M_W, 6.6 $\pm$ 0.6, they must be very rare events.

Acknowledgments

For their assistance in the field, the authors are grateful to U.S.G.S. colleagues Tom Bice, Kathy Haller, Dave Hampson, Regina Henrisey, and George Snyder, and also to U.S. Bureau of Reclamation colleagues Rick Martin and C. K. Wood. For reviews and/or discussions, we thank Selena Billington, Jim Dewey, Jack Evernden, Margaret Hopper, Arch Johnston, Robin McGuire, Russ Needham, Lou Pakiser, Dave Perkins, Pat Rogers, Ivan Wong, and an anonymous reviewer.

References

Allmendinger, R. W., J. A. Brewer, L. D. Brown, S. Kaufman, J. E. Oliver, and R. S. Houston (1982). COCORP profiling across the Rocky Mountain Front in southern Wyoming, Part 2: Precambrian basement structure and its influence on Laramide deformation, *Geol. Soc. Am. Bull.* **93**, 1253–1263.

Blackstone, D. L., Jr. (1975). Late Cretaceous and Cenozoic history of Laramie Basin region, southeast Wyoming, in *Cenozoic History of the Southern Rocky Mountains*, B. F. Curtis (Editor), *Geol. Soc. Am. Memoir 144*, 249–279.

Bollinger, G. A. (1994). Reevaluation of the epicenter and magnitude for the 1882 Colorado earthquake, in *Seismic Hazard Analysis for Rocky Flats Plant, Appendix C*, final report prepared for EG&G Rocky Flats, Inc., September 1994.

Boore, D. M. (1983). Stochastic simulation of high-frequency ground motions based on seismological models of the radiated spectra, *Bull. Seism. Soc. Am.* **73**, 1865–1894.

Boore, D. M. and G. M. Atkinson (1992). Source spectra for the 1988 Saguenay, Quebec, earthquakes, *Bull. Seism. Soc. Am.* **82**, 683–719.

Boyd, T. M., E. R. Engdahl, and W. Spence (1995). Seismic cycles along the Aleutian arc: analysis of seismicity from 1957 through 1991, *J. Geophys. Res.* **100**, 621–644.

Brewer, J. A., R. W. Allmendinger, L. D. Brown, J. E. Oliver, and S.

Kaufman (1982). COCORP profiling across the Rocky Mountain Front in southern Wyoming, Part I: Laramide structure, *Geol. Soc. Am. Bull.* **93,** 1242–1252.

Burdick, L. J. and D. V. Helmberger (1978). Upper mantle P-wave velocity structure of the western United States, *J. Geophys. Res.* **83,** 1699–1712.

Choy, G. L. and J. R. Bowman (1990). Rupture process of a multiple main shock sequence: analysis of teleseismic, local, and field observations of the Tennant Creek, Australia, earthquakes of January 22, 1988, *J. Geophys. Res.* **95,** 6867–6882.

Choy, G. L. and R. Kind (1987). Rupture complexity of a moderate-sized (m_b 6.0) earthquake: broadband body-wave analysis of the North Yemen earthquake of 13 December 1982, *Bull. Seism. Soc. Am.* **77,** 28–46.

Clift, A. C. (1986). Seismotectonic evaluation of the Dudley Gulch graben in the Piceance Creek Basin, in *Contributions to Colorado Seismicity and Tectonics—A 1986 Update,* W. P. Rogers and R. M. Kirkham (Editors), *Colo. Geol. Surv. Spec. Pub. 28,* 65–74.

Coney, P. J. (1987). The regional tectonic setting and possible causes of Cenozoic extension in the North American continent, in *Continental Extensional Tectonics,* M. P. Coward, J. F. Dewey, and P. L. Hancock (Editors), *Geol. Soc. Am. Spec. Pub. 28,* 177–186.

Cormier, V. F. and G. L. Choy (1981). Theoretical body wave interactions with upper mantle structure, *J. Geophys. Res.* **86,** 1673–1678.

Dames & Moore (1981). Geologic and seismologic investigations for Rocky Flats Plant: unpublished report prepared for U.S. Department of Energy under DOE Contract DE-AC04-80AL10890, Los Angeles.

Das, S. and B. V. Kostrov (1986). Fracture of a single asperity on a finite fault: a model for slow earthquakes?, in *Earthquake Source Mechanics, Geophysical Monograph Series 37,* S. Das, J. Boatwright, and C. H. Scholz (Editors), American Geophysical Union, Washington, D. C., 91–96.

Decker, E. R., K. R. Baker, G. J. Bucher, and H. P. Heasler (1980). Preliminary heat flow and radioactivity studies in Wyoming, *J. Geophys. Res.* **85,** 311–321.

Decker, E. R., G. J. Bucher, K. L. Buelow, and H. P. Heasler (1984). Preliminary interpretation of heat flow and radioactivity in the Rio Grande rift zone and in central and northern Colorado, *New Mexico Geological Society Guidebook,* Vol. **35,** 45–50.

Derr, J. S. (1973). Earthquake lights: a review of observations and present theories, *Bull. Seism. Soc. Am.* **63,** 2177–2187.

Dickinson, W. R., M. A. Klute, M. J. Hayes, S. U. Janecke, E. R. Lundin, M. A. McKittrick, and M. D. Olivares (1988). Paleogeographic and paleotectonic setting of Laramide sedimentary basins in the central Rocky Mountain region, *Geol. Soc. Am. Bull.* **100,** 1023–1039.

Docekal, J. (1970). Earthquakes of the stable interior, with emphasis on the midcontinent, *Ph.D. Dissertation,* University of Nebraska, Lincoln, 332 pp.

Epis, R. C. and C. E. Chapin (1975). Geomorphic and tectonic implications of the post-Laramide, Late Eocence erosion surface in the Southern Rocky Mountains, in *Cenozoic History of the Southern Rocky Mountains,* B. F. Curtis (Editor), *Geol. Soc. Am. Mem. 144, 45–74.*

Evans, D. M. (1966). The Denver earthquakes and the Rocky Mountain Arsenal disposal well, Mountain Geologist **3,** 23–26.

Evernden, J. F. and J. M. Thomson (1988). Predictive model for important ground motion parameters associated with large and great earthquakes, *U.S. Geol. Surv. Bull.* **1838,** 27 pp.

Frankel, A. (1994). Implications of felt area—magnitude relations for earthquake scaling and the average frequency of perceptible ground motion, *Bull. Seism. Soc. Am.* **84,** 462–465.

Gordon, D. W. (1987). Seismicity and tectonics of the Wyoming foreland (abstract), *Seism. Res. Lett.* **58,** 97–98.

Gregory, K. M. and C. C. Chase (1994). Tectonic and climatic significance of a late Eocene low-relief, high-level geomorphic surface, Colorado, *J. Geophys. Res.* **99,** 20141–20160.

Hadsell, F. A. (1968). History of earthquake activity in Colorado, in *Geophysical and Geological Studies of the Relationship between the Denver Earthquakes and the Rocky Mountain Arsenal Well,* J. C. Hollister and R. J. Weimer (Editors), *Colo. Sch. Mines Quarterly* **63** (1), 57–72.

Hanks, T. C. and A. C. Johnston (1992). Common features of the excitation and propagation of strong ground motion for North American earthquakes, *Bull. Seism. Soc. Am.* **82,** 1–23.

Hartzell, S., C. Langer, and C. Mendoza (1994). Rupture histories of eastern North American earthquakes, *Bull. Seism. Soc. Am.* **84,** 1703–1724.

Harvey, D. and G. L. Choy (1982). Broadband deconvolution of GDSN data, *Geophys. J. R. Astr. Soc.* **69,** 659–668.

Healy, J. H., W. W. Rubey, D. T. Griggs, and C. B. Raleigh (1968). The Denver earthquakes, *Science* **161,** 1301–1310.

Herrmann, R. B., S.-K. Park, and C.-Y. Wang (1981). The Denver earthquakes of 1967–1968, *Bull. Seism. Soc. Am.* **71,** 731–745.

Jackson, W. H. and L. C. Pakiser (1965). Seismic study of crustal structure in the Southern Rocky Mountains, *U.S. Geol. Surv. Profess. Pap. 525-D,* 85–92.

Johnston, A. C. (1996). Seismic moment assessment of earthquakes in stable continental regions—III: New Madrid 1811–1812, Charleston 1886, and Lisbon 1755, *Geophys. J. Int.* **126,** 314–344.

Jones, C. H., J. R. Unruh, and L. J. Sonder (1996). The role of gravitational potential energy in active deformation in the southwestern United States, *Nature* **381,** 37–41.

Kirkham, R. M. and W. P. Rogers (1985). Colorado Earthquake Data and Interpretations—1867 to 1985, *Colo. Geol. Surv. Bull.* **46,** 105 pp.

Kirkham, R. M. and W. P. Rogers (1986). An interpretation of the November 7, 1882 Colorado earthquake, in *Contributions to Colorado Seismicity and Tectonics—A 1986 Update,* W. P. Rogers and R. M. Kirkham (Editors), *Colo. Geol. Surv. Spec. Pub. 28,* 122–144.

Lahr, J. C. (1984). HYPOELLIPSE/VAX—A computer program for determining local earthquake hypocenter parameters, magnitude, and first motion pattern, *U.S. Geol. Surv. Open-File Rept. 84-519,* 60 pp.

Langer, C. J. and W. Spence (1995). The 1974 Peru earthquake series, *Bull. Seism. Soc. Am.* **85,** 665–687.

Langer, C. J., R. A. Martin, C. K. Wood, G. L. Snyder, and G. A. Bollinger (1991). The Laramie Mountains, Wyoming, Earthquake of 18 October 1984: A Report on its Aftershocks and Seismotectonic Setting, *U.S. Geol. Surv. Open-File Rept. 91-258,* Denver, Colorado, 41 pp.

Lee, W. H. K. and S. W. Stewart (1981). *Principles and Applications of Microearthquake Networks,* Academic Press, New York, 293 pp.

MacLachlan, J. C. and M. D. Kleinkopf (Editors) (1969). Configuration of the Precambrian surface of Colorado, *Mountain Geologist* **6,** 193–197 (with insert map).

McGuire, R. C., A. Krusi, and S. D. Oaks (1982). The Colorado earthquake of November 7, 1882: size, epicentral location, intensities, and possible causative fault, *Mountain Geologist* **19,** 1–23.

Mendoza, C. and S. H. Hartzell (1988). Aftershock patterns and main shock faulting, *Bull. Seism. Soc. Am.* **78,** 1438–1449.

Morgan, P. and C. A. Swanberg (1985). On the Cenozoic uplift and tectonic stability of the Colorado Plateau, *J. Geodyn.* **3,** 39–63.

North, R. G., R. J. Wetmiller, J. Adams, F. M. Anglin, H. S. Hasegawa, M. Lamontagne, R. DuBerger, L. Seeber, and J. Armbruster (1989). Preliminary results from the November 25, 1988 Saguenay (Quebec) earthquake, *Seism. Res. Lett.* **60,** 89–93.

Oaks, S. D. and R. M. Kirkham (1986). Results of a search for felt reports for selected Colorado earthquakes, *Colo. Geol. Surv. Information Series* **23,** 89 pp.

Oaks, S. D., M. G. Hopper, L. M. Barnhard, and S. T. Algermissen (1985). November 7, 1882, Colorado earthquake reinterpreted in light of the October 18, 1984, Wyoming earthquake (abstract), *Earthquake Notes* **56,** 24.

Powell, J. W. (1879). *Lands of the Arid Region of the United States with a More Detailed Account of the Lands of Utah, with Maps,* report of U.S. Geographical and Geological Survey of the Rocky Mountain Region, Washington, D.C., 195 pp.

Prodehl, C. and P. W. Lipman (1989). Crustal structure of the Rocky Mountain region, in *Geophysical Framework of the Continental United*

States, L. C. Pakiser and W. D. Mooney (Editors), *Geol. Soc. Am. Mem.* **172**, 249–284.

Prodehl, C. and L. C. Pakiser (1980). Crustal structure of the Southern Rocky Mountains from seismic measurements. Part 1, *Geol. Soc. Am. Bull.*, **91**, 147–155.

Singh, S. and R. B. Herrmann (1983). Regionalization of crustal coda Q in the continental United States, *J. Geophys. Res.* **88**, 527–538.

Smith, R. B. and R. L. Bruhn (1984). Intraplate extensional tectonics of the eastern Basin-Range: inferences on structural style from seismic reflection data, regional tectonics, and thermal-mechanical models of brittle-ductile transition, *J. Geophys. Res.* **89**, 5733–5762.

Somerville, P. G., J. P. McLaren, L. V. LeFevre, R. W. Burger, and D. V. Helmberger (1987). Comparison of source scaling relations of eastern and western North American earthquakes, *Bull. Seism. Soc. Am.* **77**, 322–346.

Sonder, L. J., P. C. England, B. P. Wernicke, and R. L. Christensen (1987). A physical model for Cenozoic extension of western North America, in *Continental Extensional Tectonics*, M. P. Coward, J. F. Dewey, and P. L. Hancock (Editors), *Geol. Soc. Am. Spec. Pub. 28*, 187–201.

Spence, W., P. Chang, and R. A. Martin (1988). Seismicity across the southern boundary between the Colorado Plateau and the Southern Rocky Mountains (abstract), *Geol. Soc. Am. Prog. (with abstracts)*, **20**, 415.

Spence, W. and R. S. Gross (1990). A tomographic glimpse of the upper mantle source of magmas of the Jemez lineament, New Mexico, *J. Geophys. Res.* **95**, 10829–10849.

Stegner, W. (1953). *Beyond the Hundredth Meridian. John Wesley Powell and the Second Opening of the American West*, Houghton Mifflin Co., Boston, 438 pp.

Steinbrugge, K. V. and W. K. Cloud (1962). Epicentral intensities and damage in the Hegben Lake, Montana earthquake of August 17, 1959, *Bull. Seism. Soc. Am.* **52**, 181–234.

Stover, C. W. (1985). Preliminary isoseismal and intensity distribution for the Laramie Mountains, Wyoming earthquake of October 18, 1984, *U.S. Geol. Surv. Open-File Rept. 85-137*, 9 pp.

Tocher, D. (1958). Earthquake energy and ground breakage, *Bull. Seism. Soc. Am.* **48**, 147–153.

Tweto, O. (1975). Laramide (Late Cretaceous–early Tertiary) orogeny in the Southern Rocky Mountains, in *Cenozoic History of the Southern Rocky Mountains*, B. F. Curtis (Editor), *Geol. Soc. Am. Mem. 144*, 1–44.

Warner, L. A. (1978). The Colorado lineament: a middle Precambrian wrench fault system, *Geol. Soc. Am. Bull.* **89**, 161–171.

Wong, I. G., J. D. J. Bott, and J. Ake (1994). Earthquakes and tectonic stresses in western Colorado (abstract), *Geol. Soc. Am. (abstracts with programs)*, **26**, 69.

Wong, I. G. and D. S. Chapman (1990). Deep intraplate earthquakes in the western United States and their relationship to lithospheric temperatures, *Bull. Seism. Soc. Am.* **80**, 589–599.

Wong, I. G. and J. R. Humphrey (1989). Contemporary seismicity, faulting, and the state of stress in the Colorado Plateau, *Geol. Soc. Am. Bull.* **101**, 1127–1146.

Zoback, M. L. and M. D. Zoback (1989). Tectonic stress field of the continental United States, in *Geophysical Framework of the Continental United States*, L. C. Pakiser and W. D. Mooney (Editors), *Geol. Soc. Am. Mem.* **172**, 523–537.

National Earthquake Information Center
Central Geologic Hazards Team
U.S. Geological Survey, M.S. 967
Denver, Colorado 80225
(W.S. [e-mail: spence@gldfs.cr.usgs.gov], G.L.C.)

U.S. Geological Survey
Branch of Earthquake and Landslide Hazards
Box 25046, M.S. 966
Denver, Colorado 80225
(C.J.L.)

Manuscript received 18 August 1995.