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The 1993 Killari earthquake in central India: A new fault in Mesozoic basalt flows?

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Abstract. The M_w=6.1 1993 Killari earthquake in central India was one of the deadliest earthquakes to occur in a stable continental area. It had a centroid 2.6 km deep and ruptured to the surface. A good fit of teleseismic waveforms is obtained with a simple 4.2-s-long pulse that releases a moment Mo=1.8x10^{18} nm from a reverse fault rupture with strike, dip, and rake of 126°, 46°, and 100°, respectively. A ≈10-km-wide meizoseismal area is tightly defined by relative casualty count; in the center of this area, aftershock hypocenters (Baumbach et al., 1994) outline a southwest dipping rupture that extends 5.5-8 km along strike and from the surface 8-10 km along dip. These dimensions correspond with an average displacement of 1.5-0.8 m and a stress drop of 10-4.5 MPa, respectively. We mapped a 1-km-long zone of compressional scarps which are spatially correlated with the rupture outlined by hypocenter.. In the central portion of this zone, the scarps are multiple and face in opposite directions. Deformation features indicate north-northeast directed shortening of at least 0.5 m. A leveling profile of an irrigation canal about 3 km northwest of the scarps displays a one-wavelength warp, about 1 km long and 1.3 m in peak-to-peak amplitude, that may reflect deformation associated with the rupture. Two trenches across the scarps exposed faults that offset the soil-rock interface as much as 50 cm. We found no convincing evidence suggesting that these faults existed prior to the earthquake. In the basalt, these faults reactivated exfoliation fractures in shear but were not associated with a zone of preexisting breccia and mineralization. The hypothesis that the 1993 rupture is a new fault in the Deccan Traps is consistent with lack of geomorphic evidence of prior faulting and with the lack of accumulated deformation or tilting in basalt layers exposed as close as few tens of meters from the scarp. No convincing clues about a seismogenic fault have yet been detected from either historic seismicity or geology in the 1993 source area, but a burst of seismicity 1 year before the mainshock and 2.5 years after the first impounding of a nearby reservoir was interpreted by some as a possible precursor to a larger earthquake before the 1993 mainshock. Precursory seismicity and artificial effects that cause significant mechanical changes in the crust may provide an indication of future potentially damaging earthquakes in stable continental areas.

Introduction

Intraplate continental areas generally exhibit geologic evidence of neotectonic stability; nevertheless, earthquakes do occur in these areas. The rate of seismicity in stable continental regions (SCR) is low in comparison to that at plate boundaries [e.g., *Johnston*, 1989]. However, SCR earthquakes can be sufficiently large and frequent to pose a substantial hazard. Furthermore, they indicate contemporary tectonic deformation that needs to be considered in light of the geologic characterization of these areas as "stable." The apparent inconsistency between short- and long-term manifestations of what is thought to be a stationary process not only is symptomatic of a poor understanding of intraplate seismotectonics, but also is problematic in terms of the current use of earthquake and geologic data for hazard assessment.

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The juxtaposition of significant seismogenic moment release and lack of faults that have substantial neotectonic displacement suggests an SCR tectonic environment that contains many potentially seismogenic faults, each of which produces earthquakes rarely [*Coppersmith and Youngs*, 1989; *Seeber and Armbruster*, 1993]. This hypothesis is consistent with recent findings from paleoseismic studies of surface-rupturing SCR earthquakes, which show that seismogenic faults have either very low deformation rates [e.g., *Crone et al.*, 1992; *Machette et al.*, 1993] or no neotectonic deformation at all prior to a historic rupture [e.g., *Adams et al.*, 1991].

The overall rate of seismicity is proportional to the number of active faults and to the average rate of seismicity on these faults. Because rates on individual faults are very low, the known faults alone cannot come close to producing a long-term level of seismicity similar to the one during the historic period. Thus such a level of seismicity can be reconciled with geologic evidence in SCR only if many more faults are potentially seismogenic than the ones already identified. A corollary of this hypothesis is that most future damaging earthquakes are going to occur on previously unknown faults [*Seeber and Armbruster*, 1993].

The 1993 Killari earthquake in peninsular India is another example of an earthquake from a fault with little neotectonic displacement in a tectonically stable region [*Gupta*, 1993].

Furthermore, it is a tragic demonstration of how deadly intraplate earthquakes can be and how geologic and earthquake data available before these events can give a false sense of security. The Killari epicenter is in the middle of a large area that was considered tectonically stable because it had little historical seismicity and is covered by nearly flat-lying 65-m.y.-old basalt flows of the Deccan Traps (Figure 1). There is no evidence for the Killari source in the earthquake catalog before 1992. Furthermore, this event ruptured the surface, but we found no evidence of a preexisting fault in the 65-m.y.-old rocks displaced by the 1993 rupture [*Seeber et al.*, 1994]. Thus negative paleoseismic results from this and from other intraplate faults suggest that not only the earthquake catalog, but also geologic features may not reveal many potential sources of future earthquakes. The tragic consequences of the 1993 Indian earthquake combined with the low hazard previously ascribed to the source area is a dramatic reminder that our understanding of seismogenesis in stable continents is poor. Many fundamental questions concerning seismogenesis in SCR are still unanswered; yet strategies for earthquake hazard reduction need to be urgently implemented.

The 1993 Killari Earthquake in Central India

On September 29, 1993, an earthquake of intermediate magnitude and shallow depth occurred near the village of Killari in central India, a typical SCR (Figure 1). This earthquake is assigned magnitudes M_s=6.4 (USGS), M_w=6.1 and a centroid depth of 2.6 km (this study). The teleseismic moment tensor solution (Figure 2) yields nearly pure reverse faulting. The two conjugate planes strike west-northwest and dip north-northeast and south-southwest at similar angles. The latter plane represents the rupture and has a strike, dip, and rake of 126°, 46°, 100°, respectively. Reverse faulting at intermediate dip angles and a shallow centroid depth are typical of shallow intracratonic earthquakes with surface ruptures [*Choy and Bowman*, 1990; *Crone et al.*, 1992; *Machette et al.*, 1993; *Johnston*, 1989].

The Killari earthquake occurred in the Deccan Plateau, in a cultural setting typical of rural peninsular India. Yet it was one of the most lethal earthquakes ever in SCR. It caused about 8000 deaths and the near complete destruction of 20 villages in a ≈15-km-wide area (MM intensity VIII [*Jain et al.*, 1994;

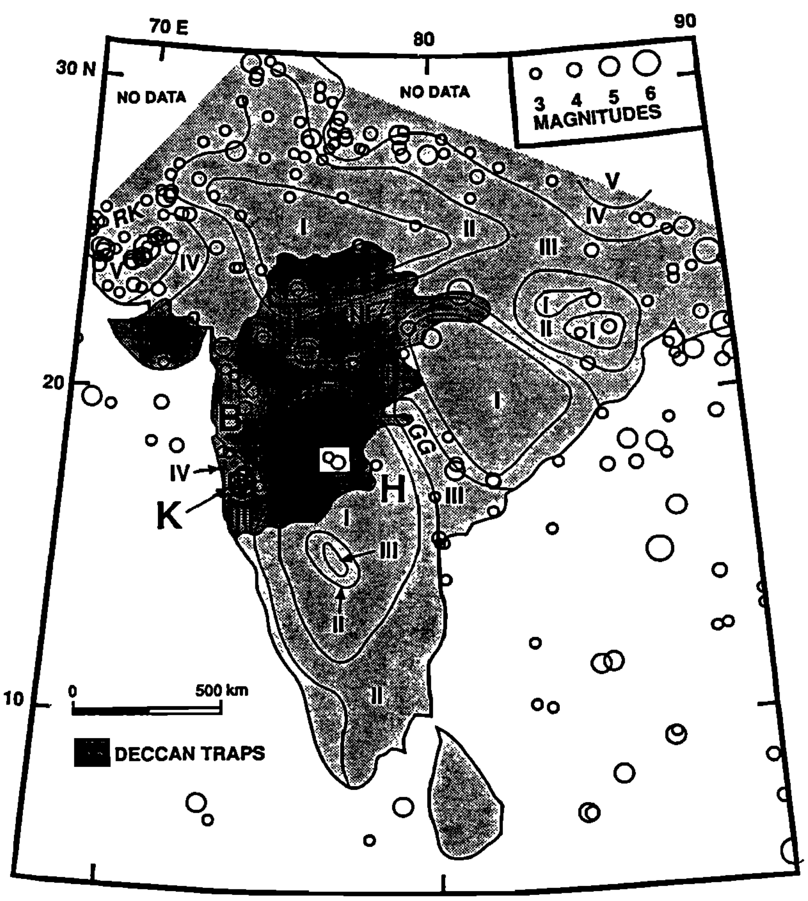

Figure 1. Seismicity map of peninsular India (1900-1992, compiled from Global Hypocenter and Preliminary Epicenter Determination files of the National Earthquake Information Center) and earthquake hazard zones [*Krishna*, 1992]; hazard increases from I to V. Hazard zonation is based on historic seismicity and on repeatedly deformed structural features, such as the Rann of Kutch (RK), the Narabada fault zone (NFZ), and the Godvari graben (GG). B is Bombay, H is Hyderabad, and K is Koyna. The Killari source (unshaded box) is in the southeastern part of the Deccan Traps (dark shading) and is marked by two epicenters (M=4.4 and M=3.9) from the precursory burst of felt earthquakes in 1992.

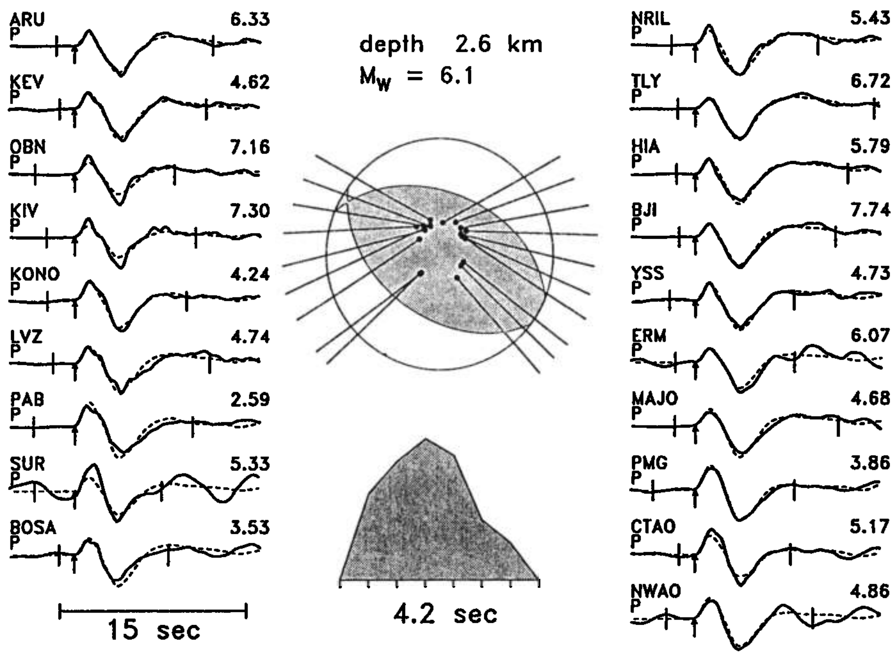

Figure 2. Moment tensor solution for the Killari mainshock in lower hemisphere, equal area projection. Solid circles mark ray paths crossing the hemisphere. The corresponding broadband teleseismic *P* waveforms are shown as solid lines, and the theoretical waveforms corresponding to the focal mechanism, source time function, and centroid depth are dashed. The southwest dipping nodal plain has strike, dip, rake of 126°, 46°, 100°, respectively. The maximum amplitude (in microns) is to the right of each waveform pair. Small arrows show picked arrival times of the *P* phase, and vertical bars show the time window used in the inversion. The rate of moment release (arbitrary scale) and the duration (4.2 s; tick marks are 0.6 s apart) are shown by the normalized source time function below the focal mechanism.

Gupta, 1993] (see also Figure 3a). This high death toll reflects in part the vulnerability of traditional stone houses to shaking damage and the high population density in rural India, but it also points to the need of better understanding of earthquakes in SCR.

On the basis of the low level of historic seismicity and the geologic evidence of neotectonic stability, earthquake hazard was inferred to be low in a large area of central India including the 1993 epicenter (Figure 1). Nevertheless, the population of Killari and neighboring villages apparently considered evacuating the epicentral area in 1992, after a burst of local seismicity which included a damaging earthquake. Many of the $M \geq 6$ SCR earthquakes were part of sequences that initiated prior to the mainshock. The spatial correlation of the 1992 seismicity with the newly impounded Tirna reservoir (Figure 3a) may have contributed to the uneasiness in the Killari area: earthquakes from Koyna (Figure 1) and from other sources of reservoir-induced seismicity have been frequently felt over the Deccan Plateau during the last decade [e.g., *Gupta*, 1992]. Evacuation plans for Killari, however, were branded as alarmist in the press and were aborted.

Historical seismicity and geology were consistent in that they showed no evidence of intraplate tectonic deformation in the Killari epicentral area [e.g., *Krishna*, 1992]. In retrospect, these long-term characteristics were misleading in terms of the area's true hazard. In contrast, the sudden onset of seismicity in 1992 was suggestive of a short-term change in the mechanical conditions of the crust and was interpreted by some as raising the possibility of more damaging future earthquakes. This tragic case suggests that new approaches are needed for earthquake hazard assessments in SCR and that damaging earthquakes can occur in seismically inactive and geologically stable areas of the continents.

We present results from a 10-day investigation of the meizoseismal area of the 1993 Killari earthquake starting 2 weeks after the mainshock, and we address some of the issues of seismogenesis in SCR. Our field work was initially concerned with both engineering and geological effects of the earthquake, but after discovering evidence of surface rupture, we focused on the rupture.

The Meizoseismal Area: Implications About the Mainshock Rupture

Both provincial governments on either side of the Tirna River promptly provided statistics on the disaster. These data were an important guide during our field work. From the count of human casualties in each village, both dead and wounded, we mapped the "relative casualty count" which we defined as (deaths+1/3 wounded)/total population (Figure 3a). Unlike the casualty count, data such as destroyed houses and people hospitalized were inconsistent across provincial boundaries, suggesting discrepancies in measurement criteria. The relative casualty count clearly delineates a 10-km-wide meizoseismal area. Preliminary results from ongoing intensity surveys [e.g., *Gupta*, 1993; *Mohan and Rao*, 1994] show meizoseismal areas generally colocated with the one delineated by the data in Figure 3a.

Relative casualty count is probably a sensitive indicator of building collapse. Within each village, building collapse was

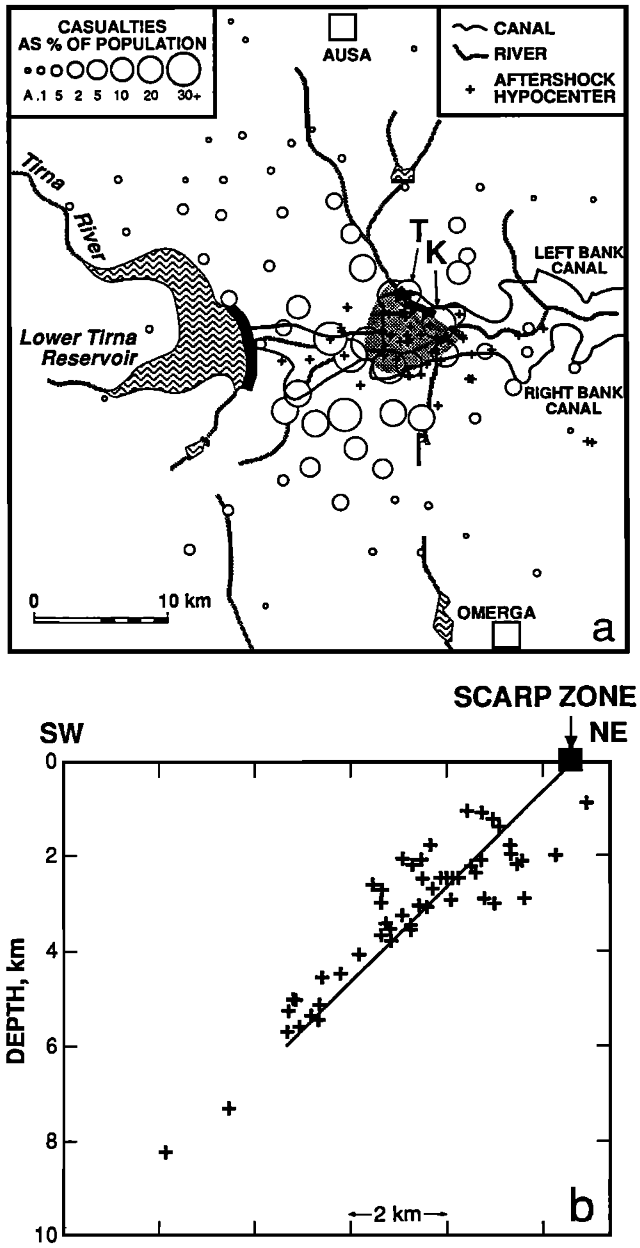

Figure 3. (a) Map and (b) vertical section (no vertical exaggeration) of data constraining the 1993 Killari mainshock rupture. The percentage of casualties in each village (Figure 3a; see text) delineates a 10-km-wide meizoseismal area. the surface rupture (heavy line; barbs on uplifted side of the buried fault) is west of Killari (K); near-surface rupture is inferred to warp the left-bank canal near Talni (T). Aftershock hypocenters [from *Baumbach et al.*, 1994] are clustered below the meizoseismal area on a plane dipping southwest and intersecting the surface rupture. The half ellipse in Figure 3a represents our best estimate of the location and size of the rupture (see text); it is centered about 10 km downstream of the lower Tirna dam (thick solid line).

strongly dependent on construction style. The majority of the houses had thick stone-and-mud walls and thick mud roofs supported by wooden beams that rested on the walls. Most of these buildings collapsed in the meizoseismal area. In contrast, collapses were rare among the few well-constructed frame buildings and the pole-and-grass huts, which housed the wealthiest and the poorest inhabitants, respectively. The variations in construction style within a village contrasted with a pronounced similarity in the overall characteristics of construction between villages. Thus differences in relative casualties among villages were more likely to reflect the level of shaking than cultural factors.

The level of shaking can also be strongly dependent on site response. The Killari area, however, is characterized by low topographic relief and by a near absence of Quaternary sediments. Most of the villages, including the ones on the bank of the Tirna River, are built on bedrock whereas areas with deep soil are generally dedicated to agriculture. Thus basin response and topographic response probably did not systematically bias the distribution of building collapses and of the relative casualty count. By default, then, source distance and radiation pattern remain as the likely major factors controlling the level of shaking and building collapse in each village. The time available to evacuate a building between the onset of shaking and the collapse may be an additional factor controlling the casualty count. This time increases rapidly away from the epicenter of a shallow source. Thus casualty count may be a more sensitive indicator of proximity to the epicenter than the level of destruction, particularly when this level reaches saturation, as it did in the 1993 meizoseismal area.

The relative casualty count in Figure 3a increases gradually toward a well-defined maximum in a 10-km-wide area centered on the axis of the Tirna River valley. About 20-25% of the people in this area were killed by the earthquake. In general, the high death toll was concentrated along river banks, where most villages were located. In the aftermath of the earthquake, this correlation was widely construed to indicate site amplification effects in the low-lying areas of the valleys. Actually, most of the villages along the Tirna River in Figure 3a are built on basaltic bedrock where large amplification effects are not likely. South of the Tirna River, where sampling allows a comparison between river banks and higher ground, the casualty count does not seem to reflect this difference in setting. In summary, the location of the rupture is likely to be the major factor controlling the relative casualty count in each village, and sites on high ground in the Killari area do not appear to be necessarily safer from earthquake damage than sites along the axes of river valleys.

Tectonic Setting

The 1993 epicentral area is within the Deccan Plateau, which is characterized by the Deccan Traps, a sequence of late Cretaceous basalt flows that cover much of central west India (Figure 1). In most of the plateau, these flows are nearly flat lying and show little evidence of deformation [e.g., *Wadia*, 1961, pp. 292-304]. The total thickness of the Deccan Traps in the epicentral area is poorly constrained but is thought to be at least several hundred meters. Individual flows are generally several to tens of meters thick and are separated by relatively thin interflow layers characterized by clay-rich paleosols.

The Deccan Traps lie unconformably on Precambrian basement of the Indian shield. Prominent continental-scale belts of late Precambrian to Paleozoic age deformation, such as the Narabada fault zone and the Godvari graben (Figure 1), have long been recognized in exposed portions of the shield [e.g., *Wadia*, 1961]. Reactivation of these structures during younger deformation events suggests that they tend to remain zones of weakness over long geologic intervals and they have been inferred to have relatively high level of earthquake hazard (Figure 1). Similar structures could exist below the Deccan Traps; several have been hypothesized, although most are poorly identified and none had been proposed to exist in the epicentral area before the mainshock. Nevertheless, brittle faults are common in exposed basement worldwide [e.g., *Isachsen et al.*, 1983], and the 1993 rupture could have occurred on such a preexisting fault buried below the Deccan basalts.

Geomorphic Setting and Shaking-Induced Deformation

After the 1993 mainshock, evidence of permanent ground deformation and liquefaction was rare, even in the meizoseismal area (Figure 3a). Generally, deformation features associated with ground failure were confined to artificial embankments; no prominent land or rock slides were reported. Shaking was apparently the main cause for the destruction of buildings [*Jain et al.*, 1994; *Mohan and Rao*, 1994].

The geomorphology of the Killari area is typical of the Deccan Plateau, where flat-laying layers of Cretaceous basalt produce a staircaselike topography and suggest lack of neotectonic deformation. The 1993 epicentral area is at an elevation of about 0.5 km and is traversed by the Tirna River; yet it is characterized by remarkably low relief. Topographically high areas are generally no more than 50 m above broad river valleys. The basalt has high resistance to mechanical erosion, which may prevent much downcutting by rivers. Near-absence of alluvial sediments is another characteristic of this area. The basalt is highly susceptible to chemical weathering and tends to become saprolite and residual soil. Low relief and lack of Quaternary sediments are probably the main factors contributing to the general scarcity of ground deformation induced by shaking and of liquefaction features in the 1993 meizoseismal area.

The Surface Rupture

A zone of ground deformation features were noticed by local farmers along the north bank of the Tirna River between the villages of Killari and Talni (Figure 3a). These features were prominent against a background relatively free of deformation. The scarps were best developed in a rare untilled field (Figure 4), where a constant stream of curious visitors testified to the uniqueness of these features in the meizoseismal area. We mapped discontinuous scarps across several fields over a distance of about 1 km in a east-west direction (Figure 5). Deformation features had been severely degraded by monsoon rains and by plowing during the 2 weeks between the mainshock and our investigation. Some of the discontinuities between scarp segments, however, could be ascribed to structural complexity rather than to scarp degradation. Surface deformation related to the rupture may have extended west of the area mapped in Figure 5 into a tributary valley of the Tirna River, where deep soil may have favored superficial warping rather than scarp formation (Figure 3a). The eastward continuation of the scarp into an area of high elevation and thin soil was less likely.

The pattern of deformation associated with the 1993 rupture is complex and discontinuous (Figure 5a). Where best developed, this deformation zone is dominated by a double-verging system of thrust faults and monoclines with a strike generally consistent with the overall westerly strike of the deformation zone (Figure 5b). This zone also includes transfer faults and extensional fissures that were probably related to buckling. The most prominent compressional scarp faces south (south verging; Figure 4a) and reaches a height of 70 cm. Generally, the scarps face down the local topographic slope, which is 2°-3° or less (Figure 5). Many of the compressional scarps are marked by a strip of uplifted ground that is as much as twice as high as the topographic step associated with the scarps (Figure 6). These ridges may reflect antithetic thrust faults and/or folds associated with changes in fault dip near the surface, where weathering of the basalt is intense (Figure 7).

The height and morphology of the scarps vary considerably over short distances (Figures 5b and 6), suggesting local variations in fault geometry and amount of slip. Three transfer faults connect compressional scarps in the area of Figure 5b, one left-lateral (T_2) and two right-lateral (T_1 and T_3). These transfer faults are characterized by secondary compressional and extensional features, which clearly indicate the sense of slip. T_2 and T_3 connect scarps with opposite vergence and the sense of horizontal slip on these transfer faults is consistent with shortening on the scarps. Fault T_2 offsets left-laterally by 53±10 cm a berm marking a field boundary and provides a minimum value for horizontal shortening (Figure 4b). The transfer faults are parallel and indicate that shortening is consistently in a north-northeast direction in spite of the structural complexity.

Finally, a prominent extensional fracture that had a maximum aperture of 4 cm and a length of more than 10 m was mapped on the uplifted side of the highest scarp (Figure 5b). Sediment-laden surficial runoff water poured into the open fracture during rain. The fact that after 2 weeks of daily rains, the fracture was not yet filled by sediment suggests that the fracture was deep. Postseismic deformation might also have contributed to keeping the fracture open, although we saw no evidence of rejuvenation along any of the other deformation features. This extensional fracture was subparallel to compressional scarps on either side and was centered on a broad uplift; it could be the result of tension related to buckling of the basalt layer (Figure 7).

On the basis of the intensity, geodetic (see below) and structural data, and the focal mechanism of the mainshock, we have argued that the deformational features in Figure 5 are the direct surface expression of the mainshock rupture and that the rupture dips to the southwest [*Seeber et al.*, 1993; *Jain et al.*, 1994; *Seeber*, 1994]. After the publication of aftershock data [*Baumbach et al.*, 1994], our early conclusions are now widely accepted because aftershock hypocenters delineate a plane with the expected attitude that reaches the surface at the deformation features (Figure 3b). The following evidence was available prior to the aftershock data:

1. The zone of deformation is complex, but it is narrow and linear compared to its 1-km length. The overall trend of this zone and of most individual scarps is west to west-northwest, and the direction of horizontal shortening is north-northeast (Figures 5a and 5b). This geometry is consistent with the focal mechanism of the mainshock (Figure 2).

2. The rupture zone is much smaller than the meizoseismal area and is located on the northeast side of this area (Figure 3a). This spatial relation between the scarp zone and the meizoseismal area is consistent with a rupture corresponding with the nodal plane dipping 46° southwest (Figure 2).

a

b

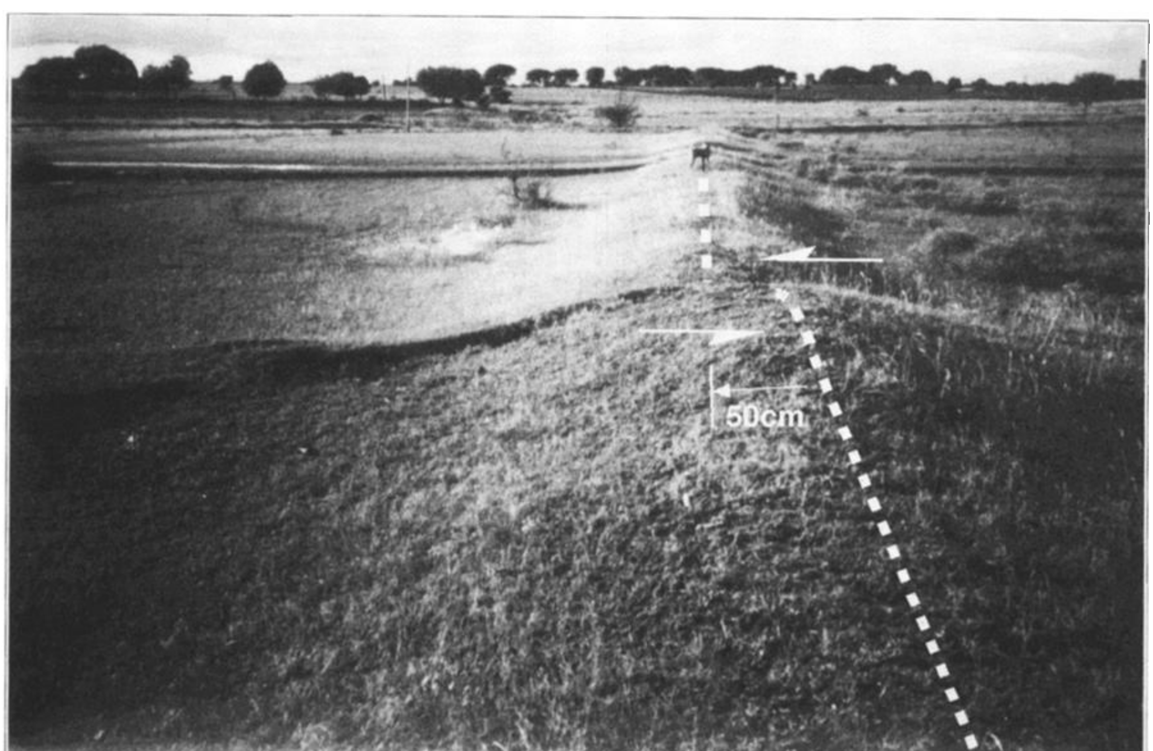

Figure 4. Photographs of scarps in the area of Figure 5b. (a) South facing scarp looking west from the site of trench 1 (marked by 10x6 cm compass on the ground). Tire tracks cross the scarp, which is marked by arrows. The double arrow marks the uplifted ground along profile 2 (Figure 5b). Note low relief in the background. (b) Northwest view along earthen berm (dashed line) offset 50 cm left-laterally by transfer fault T2 (Figure 5b).

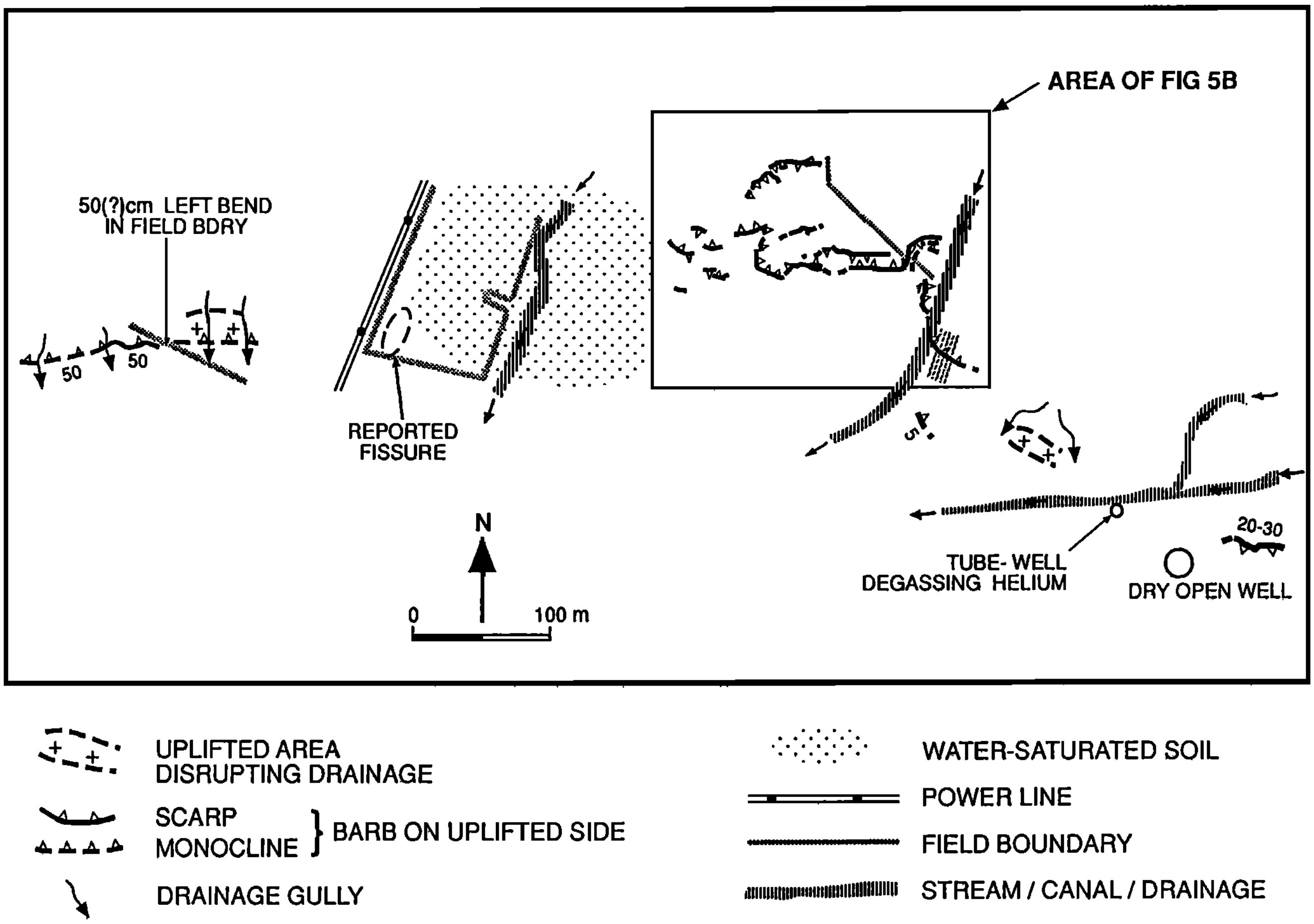

Figure 5a. Map of deformation thought to be the direct surface manifestation of the 1993 mainshock rupture. Surface rupture might have continued westward in a low-lying area that had been extensively tilled after the mainshock.

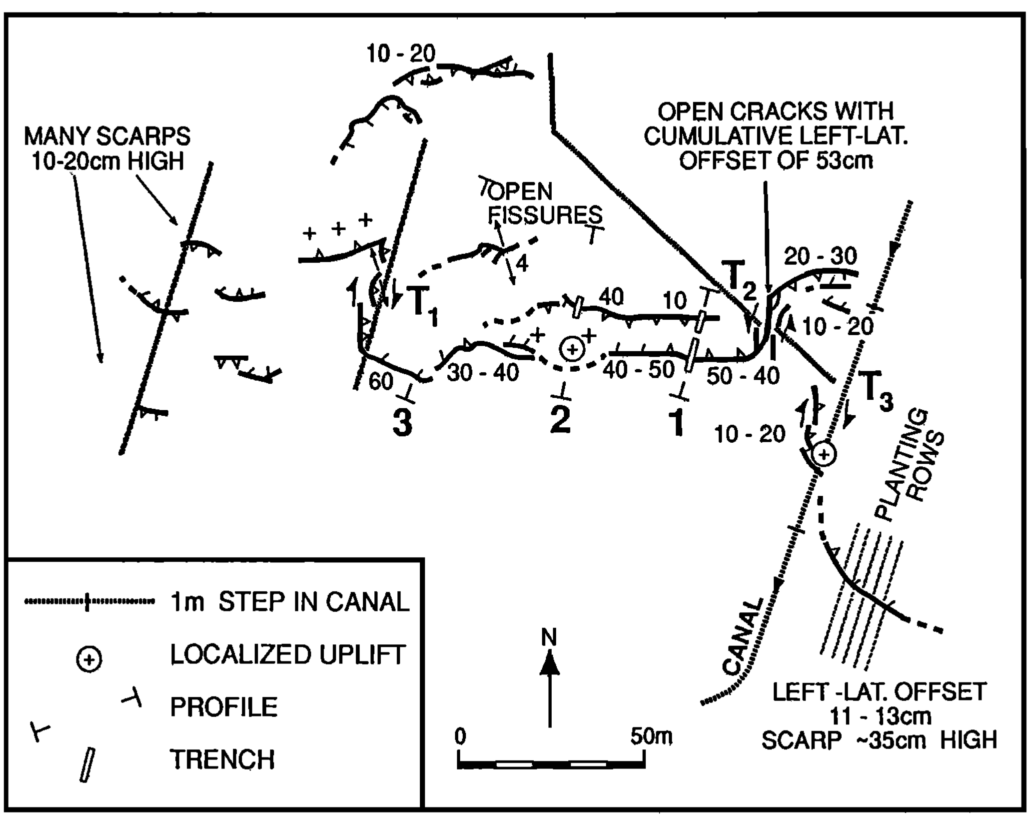

Figure 5b. Detailed map of well-developed and preserved portion of the surface rupture in untilled field. T_1, T_2, and T_3 are transfer faults. Trenches 1 and 2 (Figure 9) are along profiles 1 and 2 (Figure 6).

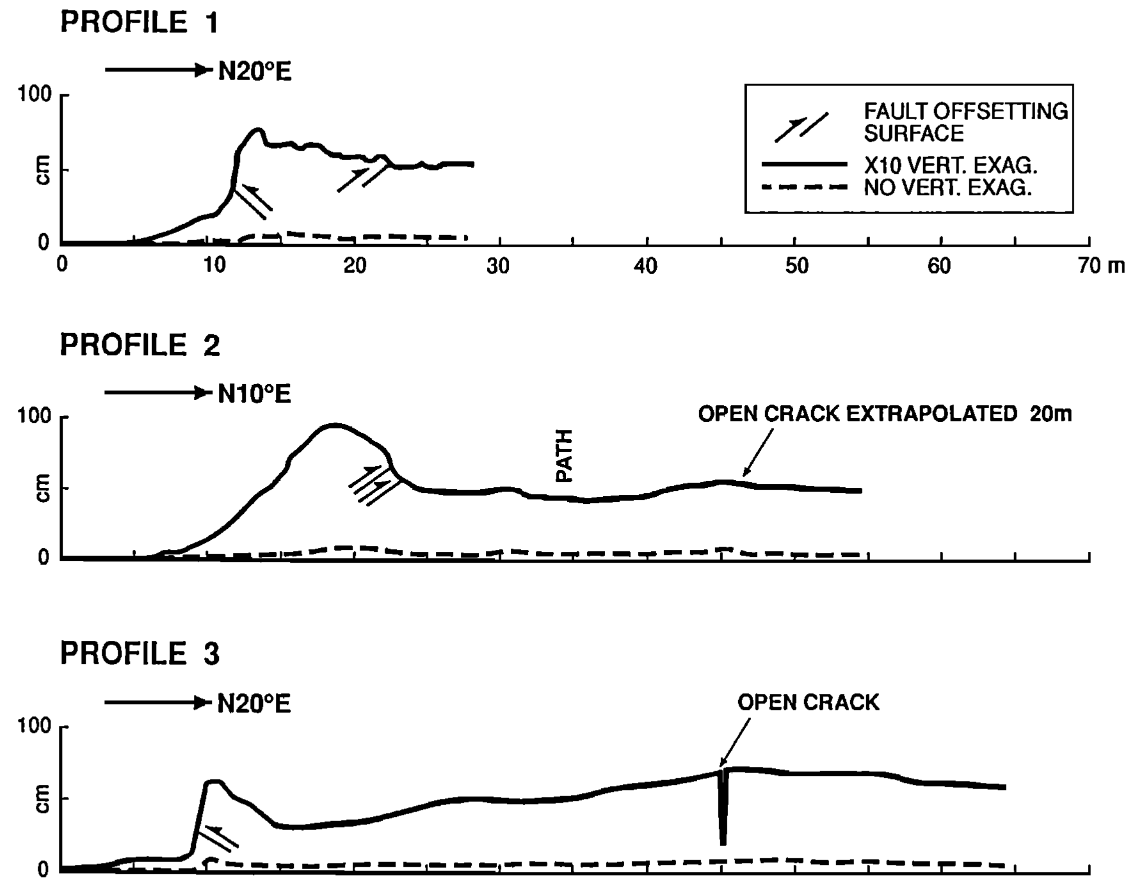

Figure 6. Topographic profiles across scarps. Profiles are located in Figure 5b.

3. The scarps are most prominent where soil cover is thin, suggesting bedrock involvement in the deformation. For three scarps, spanning the full gamut of scarp sizes, bedrock involvement was demonstrated by trenching (see below).
4. Shaking-induced downslope slip is an unlikely cause of the scarps because, (1) while there is clear evidence of shortening across them, there is no evidence of corresponding upslope extension, (2) the surface slope along the scarps is low (1°-3° (Figure 4)), and (3) slope failure was not observed on steeper slopes within a few hundred meters of the scarps.
5. After the mainshock, an anomalous amount of helium was degassing from the soil and from a tube well near the scarp (Reddy et al., 1994 ; Figure 5a). Helium is a by-product of nuclear decay in crustal rocks and is often associated with deep-reaching active faults that may serve as conduits for the gas.

The complex system of scarps which are along the surface projection of the southwest dipping rupture include prominent south facing scarps (Figure 5a) and show a tendency for scarps to face down slope. This discrepancy between the geometry of the rupture and surface faulting may reflect vertical strength-anisotropy in the Deccan Traps. In the "layer cake" structure of the traps, where brittle basalt layers alternate with thin plastic interflow layers rich in montmorillonite [*Sahasrabudhe et al.*, 1969], the deformation associated with the thrust fault may be partitioned into horizontal shortening and differential vertical motion by a combination of horizontal slip along the weak layers and buckling or thrusting of the stiff layers (Figure 7). This pattern of deformation may occur more readily in weathered rocks near the surface. Thus the vergence of faults in superficial layers of the Deccan Traps may reflect local conditions, such as topography-induced stresses, rather than underlying fault geometry.

Evidence of Coseismic Warping?

After the 1993 mainshock, farmers noticed topographic changes and disrupted drainage in some of their fields near the village of Talni (Figures 3a and 8). Ponded water was as much as 1 m deep in the adjacent portion of the left bank canal. This area is about 3 km northwest of the scarp zone. Steady rain and tilling in the deep soil could have easily masked coseismic fractures by the time of our investigation, and the surface rupture could have extended in the lowland from the scarp zone to the area in Figure 8, as shown in Figure 3a.

The irrigation canal in Figure 8 is along the crest of an earthen embankment that rises 5-10 m above the valley floor. The canal was still under construction at the time of the mainshock. It was designed to have a slope of 1:6000 toward the east; it had been graded but was not yet lined. The postearthquake profile of the canal deviated greatly from the expected uniform slope. A very prominent warp spanned the entire 1.3-km-long profile and was superimposed on short-wavelength "anomalies" 0.1-0.2 km long. The warp had an amplitude of 1.3 m, which is only 35 cm less then the water depth in the canal at maximum design flow (Maharashtra Irrigation Department, written communication, 1994). It seems unlikely that such a warp could be a feature in the preearthquake

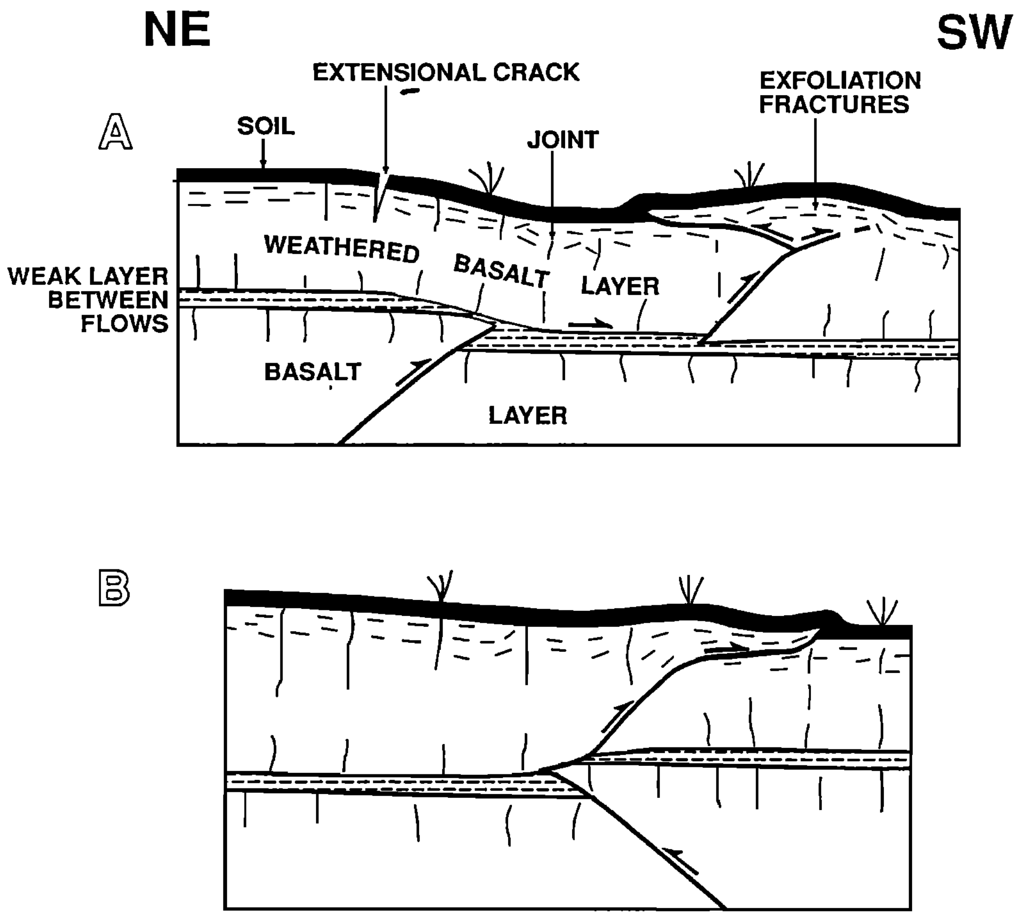

Figure 7. Schematic sections of hypothetical structures at the shallow end of the 1993 fault rupture, which is estimated to dip 46° to the southwest (Figure 3b). This structural complexity is ascribed to interbedded strong and weak layers in the Deccan Traps and probably extends deeper than is shown here. Slip on the horizontal weak layers may be responsible for the double vergence and the complex pattern of surface faulting. Sections A and B are inspired by profiles 2 and 1 (Figure 5b), respectively.

profile of the "graded" canal. Unfortunately, we have no data on the preearthquake shape of the canal.

Ground failure in the embankment was evidenced by longitudinal fissures suggesting the development of incipient slumps along the flanks. These slumps did not seem to affect the floor of the canal along the axis of the embankment, but short-wavelength "anomalies" along the profile could result from localized shaking-induced settlement of the embankment. The prominent long-wavelength warp along the canal, however, is unlikely to be the result of differential settlement because (1) the profile shows a large drop from west to east (Figure 8), opposite to the direction in which the height of the embankment increases and settlement would be expected to increase as well, and (2) fissures related to slumps on the flanks of the embankment were evenly developed along most of the profile and were not concentrated along the low portion of the profile. It is at least plausible therefore that the long-wavelength component of the profile in Figure 8 reflects warping in response to slip on a buried fault. Assuming reverse motion on a fault dipping at 45°, the fault would need to dip to the south to account for the warp. The fault displacement required to produce the warp depends on fault structure and is probably nonunique (e.g., a listric fault or strain partitioning of the type shown in Figure 7); modeling of the warp is therefore unwarranted. The displacement on the inferred fault would need to be substantially larger than the minimum displacement implied by 0.5 m of shortening in the scarp zone; we estimate that reverse displacement of 1.5-2 m on a fault dipping 45° southwest would probably be consistent with the amplitude of the warp.

Rupture Characteristics

We inverted broadband teleseismic waveforms of the Killari mainshock using the method of *Ekstrom* [1989] to determine focal mechanism, source depth, and time history of slip and moment release. The inversion is stabilized by including the long-period data set used in the standard Harvard CMT (centroid-moment tensor) algorithm. Broadband data improve constraints on the source depth and on the history of moment release. The nodal planes in our solution (Figure 2) are very similar to the ones in the CMT solution [*Dziewonski et al.*, 1995], but our centroid depth is much shallower (2.6 km versus 15 km), and the scalar moment is about 20% smaller. Good agreement between the data and the synthetic waveforms for a point source suggests a simple seismic rupture confined to the upper 5-7 km of the crust. We were unable to resolve any directivity of the rupture, but this is not surprising, given the short duration (4.2 s) and small dimension of the source.

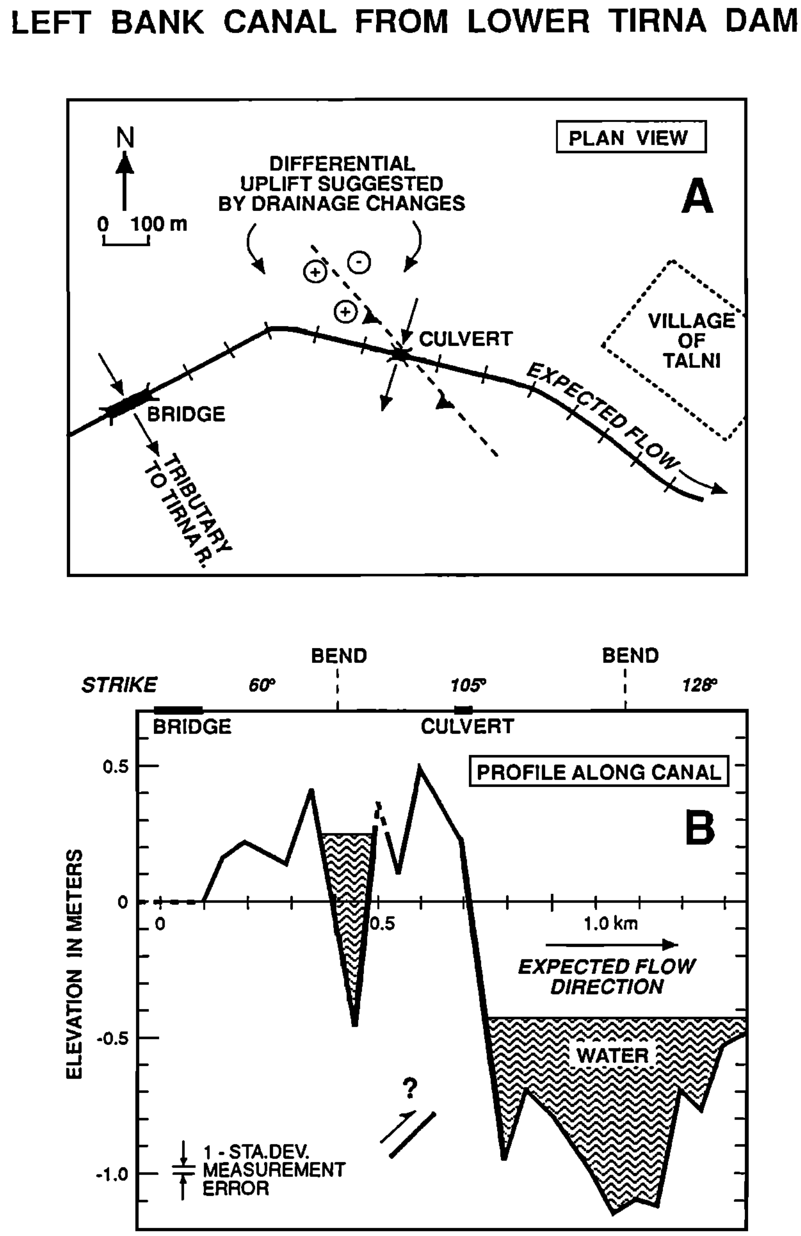

Figure 8. (a) Map and (b) profile of portion of left (north)bank canal from the Tirna Dam. The profile was measured with a hand-held level; the error bar is 1-standard deviation for a set of repeated reversed measurements.

The overlap of the direct *P* and the surface-reflected *pP* and *sP* phases at teleseismic distances weakens depth constraints for very shallow earthquakes. Inverse algorithms, such as the one we use, match the observed waveforms by modeling the interference between these phases under the assumption that the moment release occurs at a point. We therefore interpret our results in terms of a rupture that is centered at 2.6±1 km depth. *Chen and Kao* [1995] have also modeled the Killari earthquake using a similar set of teleseismic broadband seismograms, and obtain a source depth of 7±1 km. They based their depth estimate on the relative delay with respect to the *P* phase of the onsets of the *pP* and *sP* phases, not on the total pulse shapes. Their estimate therefore refers to the onset of rupture, not to the centroid of the radiated energy, and is consistent with our estimate of centroid depth. Finally, the very shallow rupture implied by the 2.6-km-deep centroid is consistent with the depth distribution of aftershocks, which can be directly associated with the presumed rupture and are mostly confined between a depth of 6 km and the surface (Figure 3b; *Baumbach et al.* [1994]).

In summary, the combined instrumental, macroseismic, and geologic data provide a consistent and detailed picture of the seismic rupture. The distributions of casualties, aftershocks, and surface scarps define a rupture centered in the Tirna River valley and dipping southwest (Figure 3a). The strikes and dips of 135°, 42° from aftershock hypocenters [*Baumbach et al.*, 1994] and 126°, 46° from teleseismic data (this paper) are consistent within the resolution of the data. The strike of the 1-km-long zone of surface rupture is 100° (Figure 5a), but the overall trend from the rupture zone to the warped canal is 135°, which is consistent with aftershocks and focal mechanism. The aftershock distribution (Figure 3b), the 2.6-km-deep centroid, and the scarps constrain the rupture to extend from a depth of 6-7 km to the surface. Such a rupture is consistent with the 7-km deep onset of rupture [*Chen and Kao*, 1995], assuming that the rupture started at the downdip end and propagated upward. The aftershock distribution suggests a rupture at least 5.5 km along strike and centered at the zone of scarps (Figure 3a; *Baumbach et al.* [1994]).

We assume a rigidity of 3.5×10^{10} n/m^2, a moment $Mo=1.8\times10^{18}$nm (Figure 2) and a semielliptical rupture with semiaxes r_1 and r_2 along the surface and along the dip of the rupture, respectively. By symmetry, the displacement and stress drop on such a shallow rupture are the same as the ones on the elliptical rupture of the same dimensions in an infinite medium, if r_2 is vertical, but the moment on the semielliptical rupture is half the moment on the elliptical rupture. We assume that the properties of the two ruptures are similar even in the case of the Killari rupture, where r_2 dips 45°. Thus the stress drop $\Delta S=7Mo/(r_1+r_2)^3$ and the average displacement $D=2Mo/\mu\pi r_1 r_2$ are obtained by modifying the formulas for a circular rupture [*Kanamori and Anderson*, 1975]. If the possible sizes of r_1 and r_2 range from a minimum of 2.75 km and 8 km [*Baumbach et al.*, 1994] to a maximum of 4 km and 10 km, respectively (Figure 3), we obtain a range in stress drop from 10 to 4.5 MPa and an average displacement from 1.5 m to 0.8 m. These results are consistent with fault displacements inferred from surface observations, 1.5-2 m from the possible warping of the canal (Figure 8) and more than 0.7 m from the scarps (at least 0.5 m of horizontal shortening; Figures 5b, 9, and 10b, and 10c). Our source parameters for the 1993 Killari mainshock are also consistent with results of *Baumbach et al.* [1994], who obtain a stress drop of 7 MPa and a displacement of 1.5 m for a rectangular rupture 5.5 km along strike and 7 km along dip. Finally, our results are consistent with worldwide averages for SCR earthquakes [e.g., *Johnston*, 1994].

Trenches

We exposed the uppermost 1 m of the subsurface in three hand-dug trenches in the fallow field where the scarps were most prominent. One trench crossed a south-facing scarp, and the other two crossed north-facing scarps (Figures 5b, 9 and 10). These trenches exposed weathered basalt below a 10- to 30-cm-thick residual soil. The soil horizon appeared well mixed and uniform. Soil characteristics and the sharp boundary between soil and basalt probably reflect many centuries of agricultural activity and are unlikely to preserve any record of deformation from prehistoric earthquakes. The basalt was so strongly weathered and fractured that it could easily be excavated by pick and shovel. This weathering varied spatially; it was strong near the surface but showed a tendency

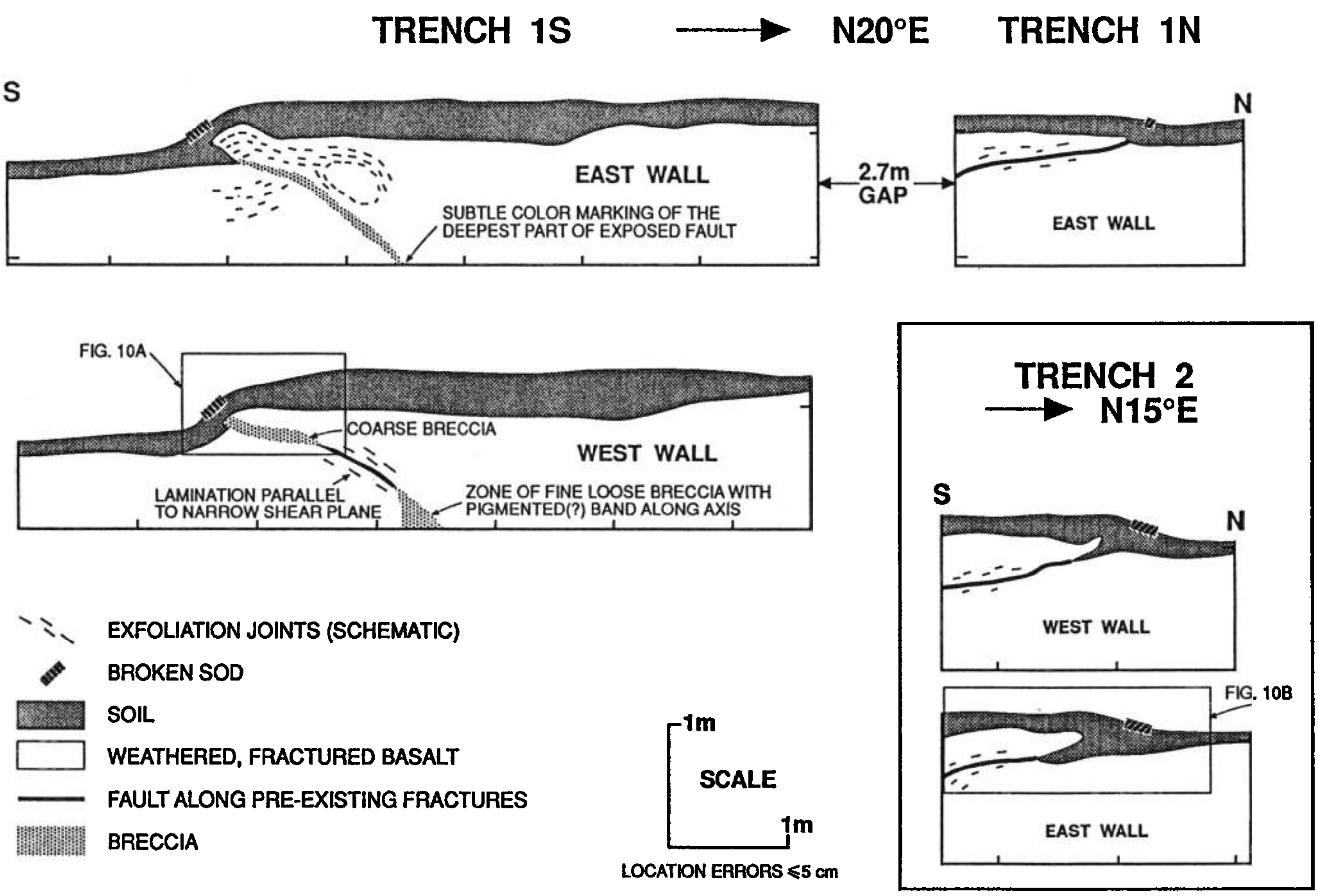

Figure 9. Maps of hand-excavated trenches across scarps associated with the mainshock rupture. All features shown (except exfoliation joints) were mapped to better than ±5 cm accuracy. Trenches 1 and 2 were along profiles 1 and 2 shown in Figure 5b. Soil was generally thicker, and weathering of basalt more pronounced, in the uplifted block between opposing thrusts. We did not have sufficient exposure to understand the significance of the loose breccia zone with subtle pigmentation at the bottom of the trench (see text).

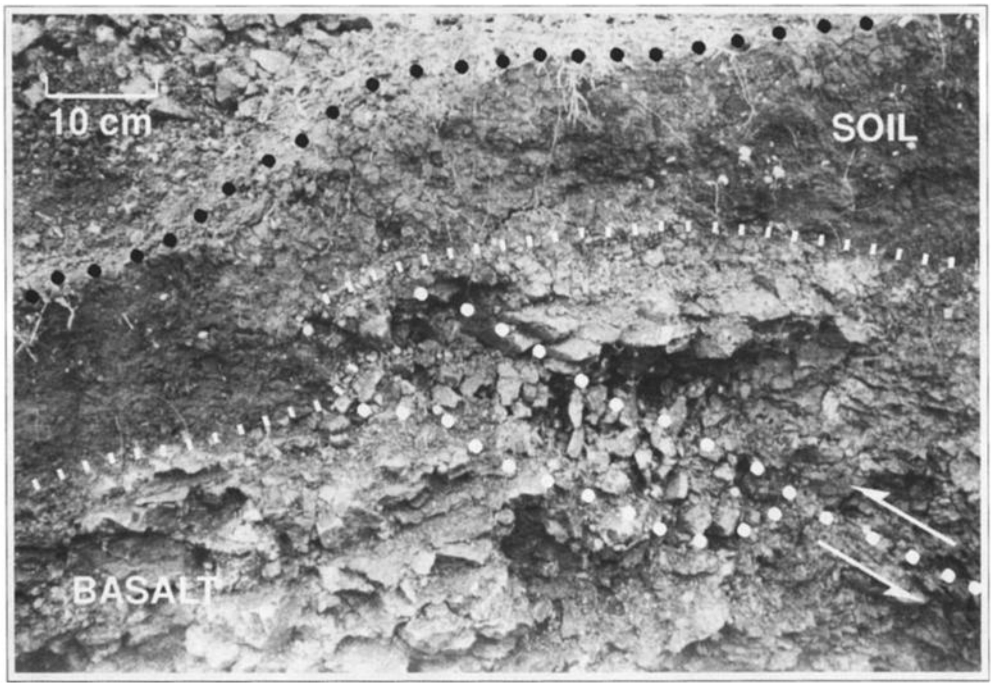

Figure 10a. Photograph of part of west wall in trench 1S showing the south facing scarp and the north dipping fault related to the scarp. (See Figure 9 for location). Here and in Figures 10b and 10c, black dots, dashed line, and white dots mark the surface, the soil-rock boundary, and the 1993 faults or fault zones in the basalt, respectively; arrows indicate the sense of motion. Shear is accommodated by sliding along pre-existing fractures in a 1- to-2 cm-wide zone in the lower part of the frame, but widens into a 10-cm-wide zone of breccia with rotating clasts resulting in a monocline near the surface. The breccia is composed of rock fragments formed by weathering-related exfoliation. Restoring the 1993 deformation would probably return these clasts into their original position in the exfoliation fabric. Thus these clasts were not part of a fault breccia before 1993.

to increase downward in the deeper part of trench 1. Such variations, including a tendency for reverse weathering profiles, seem to be characteristic of individual basalt layers exposed in the Deccan Plateau. Faults associated with the scarps were clearly visible in the basalt; they displaced the soil-rock interface with a reverse sense of throw by amounts ranging from 2-3 cm in trench 1N to 50 cm in the west wall of trench 2 (Figures 10b-c).

Two kinds of fractures were prominent in the trenches. The largest fractures were subvertical and confined to areas where weathering was less intense. They had a preferred north-northeast orientation similar to the direction of shortening; they may be interpreted as extensional fractures parallel to the direction of the maximum compressive stress responsible for the earthquake (Figure 11). A second set of small, pervasive fractures probably resulted from exfoliation related to

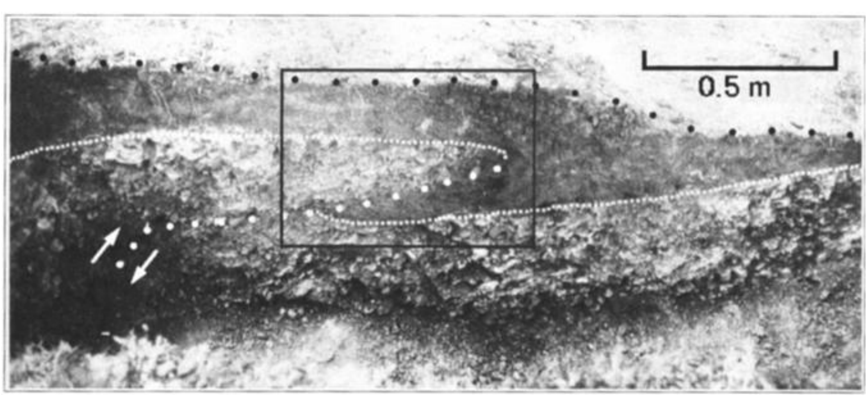

Figure 10b. West wall of trench 2 showing south-dipping fault and related north facing scarp. In contrast to the fault in Figure 10a, this fault reaches the surface along preexisting fractures and is not associated with a breccia. Local variations in the geometry of exfoliation fractures may control some complexities in the pattern of faulting.

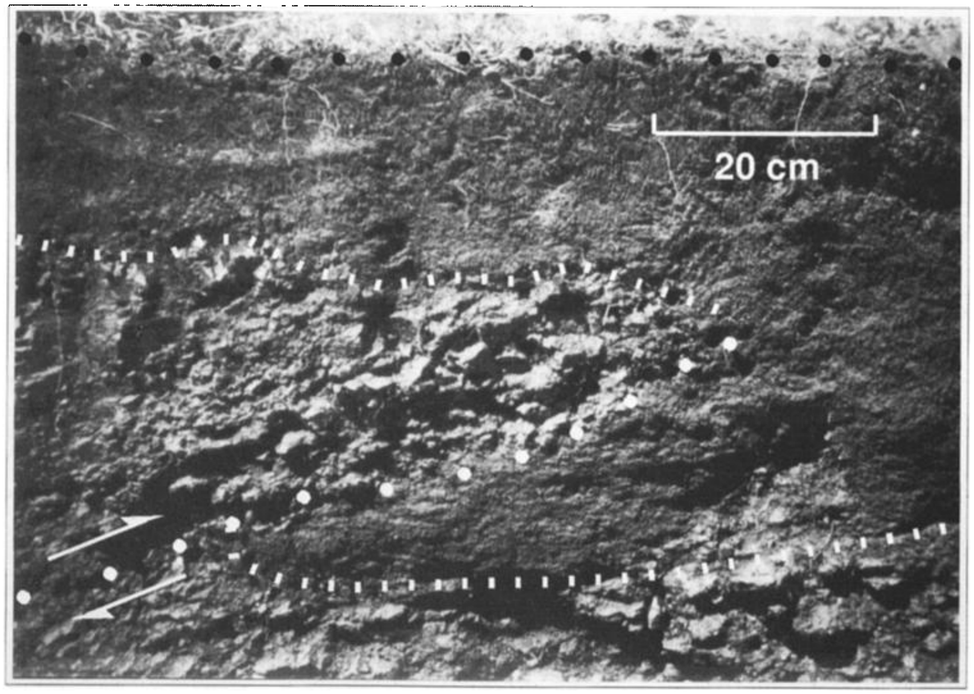

Figure 10c. Detailed view of the fault in Figure 10b, showing weathered basalt overriding soil by about 50 cm.

weathering. They were organized in bands of subparallel fractures smoothly wrapping around less-weathered and less-fractured rock; near the surface, these fractures tended to be horizontal and parallel to the rock-soil interface (Figure 10). These fractures performed two functions in accommodating fault movement. Where fractures were subparallel to the faulting, they provided low-resistance sliding surfaces and the motion was taken up by slip on one or several fractures within a narrow zone that was typically less than 3 cm wide (e.g., Figures 10b and 10c). Where the faulting was at a high angle to the fractures, the deformation was accommodated in a much broader zone of shear which was typically 10-20 cm wide and in which preexisting exfoliation fragments were rotated by slip on the exfoliation fractures (Figure 10a). The sense and amount of rotation of the fragments from their inferred original position in the exfoliation structure seem to conform with the 1993 deformation as defined by the distorted soil-rock interface. Thus the fault breccia in Figure 10a probably formed in the 1993 event.

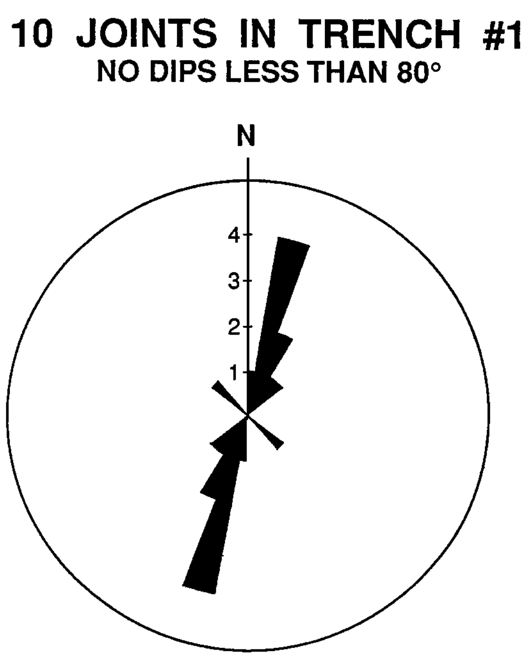

Figure 11. Strike distribution of 10 subvertical joints (all with dip angles ≥80°) in trench 1S.

A New Fault?

Unfortunately, no reference horizon was visible in the basalt exposed in the trenches, and we could not measure directly the displacement accumulated on the faults that slipped in 1993 since the basalts were emplaced. The configuration of breccia along the fault in trench 1S (discussed above) and lack of fault breccia on the other exposed faults (Figures 9 and 10) suggest that total accumulated displacements on these faults are small, possibly no more than the 1993 displacements. We saw no evidence of pre-1993 deformation along any of the 1993 faults, as might have been expressed by slickensides or concentrated weathering and mineralization. One possible exception is a zone of loose breccia and unusual pigmentation along the fault which was very poorly exposed on the waterlogged floor of trench 1S (Figure 9). Another possible suggestion of a preexisting fault is that the soil above the uplifted block between the divergent thrust faults is thicker than the soil above the footwall blocks (Figure 9). From the limited trench exposure, we cannot evaluate whether the soil thickness on the uplifted block is unusual. The thicker soil may reflect a zone where the basalt is more fractured, as a result of deformation in a hanging wall block. Alternatively, a new fault might have propagated into an area of more intensely weathered and fractured basalt, following the path of least resistance.

The geomorphology in the area of the 1993 scarps contains no evidence of preexisting scarps or accumulated vertical deformation. The zone of 1993 scarps is on the north flank of the Tirna valley and correlates with a gentle rise in topography from south to north that develops over a distance of several kilometers. This rise is opposite to the topographic effect that would be caused over that distance by displacement on a south-dipping reverse fault. Absence of pre-1993 Cenozoic warping and related tilting is also indicated by subhorizontal interflow layers exposed in the vicinity of the scarp: one only a few tens of meters south of the scarp (in the dry well, southeast corner of Figure 5a), another less than 1 km north of the scarp. Judging from the complexity of the 1993 deformation, much tilting and warping would be expected in the deformation field generated by the superposition of numerous such events.

The available evidence does not rule out the possibility that the 1993 rupture developed on a preexisting Cenozoic fault. This issue could be conclusively examined by mapping distinct subsurface horizons in the Deccan Traps across the 1993 fault. However, the evidence does rule out a prominent fault with substantial neotectonic displacement. If a preexisting fault is still not evident after a surface-rupturing earthquake, it will be very difficult to recognize such a fault by regional geological investigations before the fault is defined by seismicity. Given the very low rate of long-term seismicity expected from this fault, historic data are also unlikely to reveal the fault.

Was the Killari Earthquake Reservoir-Induced?

The first known seismicity in the vicinity of Killari was in 1992 (Figure 1) [*Baumbach et al.*,1994], about 2.5 years after the lower Tirna reservoir was initially impounded. The maximum water levels attained in 1989, 1990, 1991, and 1992 and the level at the time of the 1993 mainshock were 12.2, 14.7, 14.9, 12.2, and 11.4 m above the minimum ground elevation at the dam, respectively (deduced from written communication from Maharashtra Irrigation Department, 1994). The inferred 1993 rupture is centered below the axis of the valley and about 10 km downstream of the lower Tirna dam (Figure 3a). The rupture of the 1967 mainshock associated with the Koyna reservoir was also centered in the valley, about 8 km downstream of the dam (Figure 12). Maximum depth in the Koyna reservoir is about 100 m, and seismicity there is widely recognized to be reservoir induced [*Gupta*, 1992; *Rastogi*, 1994].

In 1993, the state of Maharashtra (approximately the southern three quarters of the Deccan Traps area in Figure 1) operated about 35 "major" reservoirs, including the lower Tirna reservoir (Maharashtra Irrigation Department, written communication, 1994). Thus the chance that an independent earthquake sequence would occur in Ar, the area within a distance r from any of the dams, is $Ar/(At-Ar) \approx 1/20$, where $r=10$ km and At is the surface area of Maharashtra. Two $M \geq 6$ earthquakes have occurred in this state during the last century, excluding the 1967 Koyna earthquake as being reservoir induced but not the 1993 Killari earthquake. The chance of starting a $M \geq 6$ earthquake sequence within 2.5 years following the first impounding of a reservoir is 1/20. Thus the combined spatial and temporal correlation between the seismicity at Killari and the lower Tirna reservoir has about a 1/400 chance of being fortuitous. The chance of a fortuitous correlation at Killari may be even lower, since the level of historic seismicity in the entire state of Maharashtra is higher than that in the Killari area. Thus the possibility that the seismicity at Killari was triggered by the reservoir needs to be considered [e.g., *Jain et al.*, 1994; *Seeber*, 1994].

The main factor detracting from the hypothesis that the 1992-1993 Killari sequence was triggered by the lower Tirna reservoir is that the maximum water depth of 14.9 m in 1991 is near the lower limit of the range of reservoirs known to have triggered earthquakes [*Rastogi*, 1994; *Johnston et al.*, 1994]. However, earthquakes occurred near the similarly shallow Ghirni reservoir in the Deccan Plateau [*Gupta*, 1992] and water depth changes as small as 1 m have been observed to trigger earthquake swarms at some reservoirs [*Simpson*, 1986]. Both the 1992 and the 1993 sequences followed periods of unusually rapid water level rise in the reservoir (Figure 13); [Rastogi, 1994]. The 1993 aftershock sequence had a magnitude distribution similar to the one for noninduced seismicity in peninsular India and was poorer in small-magnitude earthquakes than typical reservoir-riggered seismicity. The decay rate of the aftershock sequence, however, was slow, which is typical of reservoir-triggered seismicity [*Rastogi*, 1994].

Whether or not the seismicity at Killari was triggered by the lower Tirna reservoir is an open question. The change in stress at seismogenic depths from such a shallow reservoir is probably very small compared to stress level at failure. This change, however, could have been just enough to induce failure, acting as a "trigger," not as a "cause," of the earthquake. The Killari earthquake is fundamentally a tectonic event that might have occurred even without the reservoir, possibly at a different time [*Johnston et al.*, 1994]. Thus it is appropriate to compare the Killari earthquake with other SCR earthquakes that are strictly tectonic in origin.

Implications for Intracratonic Seismogenesis

The 1993 Killari earthquake is one of a family of $M \geq 6$ earthquakes in SCR characterized by very shallow fault ruptures that reach the surface. The seismogenic faults responsible for these earthquakes can be unequivocally identified and investigated geologically. Recent investigations of some of these fault ruptures have shown them to be generally on faults that have not had surface slip for a long time prior to the last surface rupture. The 1989 Ungava, Canada, earthquake is thought to have ruptured Archean rock that had not previously experienced brittle failure [*Adams et al.*, 1991]. The Meers fault, in the SCR of the central United States, ruptured to the surface twice during the Holocene; yet it is not associated with historic earthquakes and it offsets by similar ammounts Holocene and middle Pleistocene deposits, indicating that pre-Holocene faulting occurred more than 100,000 years ago [*Crone and Luza*, 1990]. On the basis of the absence or extensive degradation of possible prehistoric scarps, the rupture prior to the 1986 mainshock at Marryat Creek is inferred to be at least 100,000 years old [*Machette et al.*, 1993]. From their results for the Tennant Creek earthquakes and other available paleoseismic results from SCR worldwide (e.g., 1968 Meckering and 1970 Calingiri, Australia [*Gordon and Lewis*, 1980]; Guinea, west Africa [*Langer et al.*, 1987], *Crone et al.* [1992, p.35], conclude that "...recurrence intervals of 100,000 yr or more are typical of stable interior of continents." Finally, the few other intraplate faults known to have ruptured only in the subsurface also seem to have little

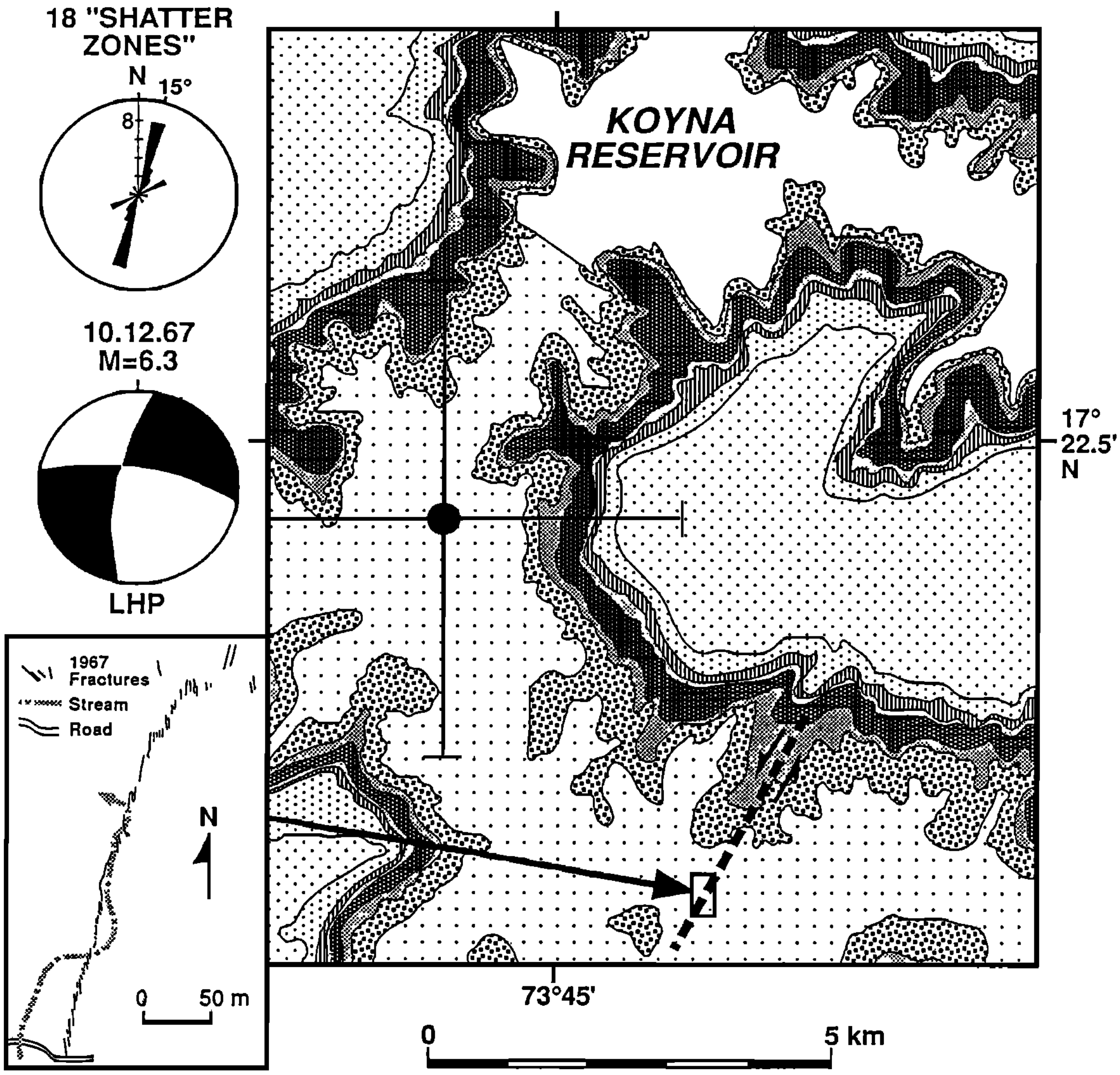

Figure 12. Map of the Koyna dam area at the western border of the Deccan Plateau (modified from *Sahasrabudhe et al.*, [1976]). Eight Late Cretaceous basalt flows (each represented by a distinct pattern) mark contours lines (e.g., the water level in the reservoir) and show no evidence of fault offsets, folding, or tilting. The main structural features are "shatter zones," steeply dipping planar zones of intense fracturing in the basalts. The orientation of these shatter zones is clustered in a north-northeast direction (rose diagram on the left). Right-stepping fissures produced by 1967 earthquake (inset) were mapped along dashed line. They suggest surface rupture with left-lateral component of motion on a steep fault. Such a rupture is consistent with the focal mechanism of the M_s=6.3 mainshock [*Sykes*, 1970] and is subparallel to the majority of the shatter zones. Of several widely scattered epicenters proposed for this event, the one in this figure (solid circle [from *Gupta*, 1992]) is consistent with aftershock epicenters.

neotectonic displacement (e.g., the Dobbs Ferry fault in New York [*Dawers and Seeber*, 1991; *Seeber and Armbruster*, 1993]. Thus the lack of convincing evidence for prior Cenozoic displacement on the fault that ruptured in the Killari earthquake is consistent with known characteristics of seismogenic faults in SCR.

The 1967 Koyna earthquake is the most recent and best documented $M \geq 6$ event in the Deccan Traps area of peninsular India prior to the Killari earthquake. The temporal pattern of seismicity at Koyna is strongly correlated with reservoir loading, yet the P axis of the mainshock focal mechanism is consistent with inferred regional stress directions [e.g., *Gupta*, 1992]. Thus even at Koyna, the effect of the reservoir is probably superimposed on a tectonic environment that might have eventually lead to brittle failure, and the 1967 mainshock may be comparable to any other earthquake in SCR in terms of its relation to structure. *Sahasrabudhe et al.* [1969] mapped the Koyna area in detail for evidence of neotectonic deformation. They did not find direct evidence for pre-1967 faulting. Furthermore, they traced basaltic flows exposed in the area along topographic contours (Figure 12) and concluded that the flows had not been significantly tilted, folded, or faulted after their emplacement in late Cretaceous time. Nevertheless, they mapped a linear zone of "ground fissures," which were coseismic with the 1967 mainshock and extended for at least 3 km across the Koyna valley, about 8 km downstream of Koyna dam (Figure 12). This zone is located 5 km southeast of the 1967 instrumental epicenter, but this distance is similar to

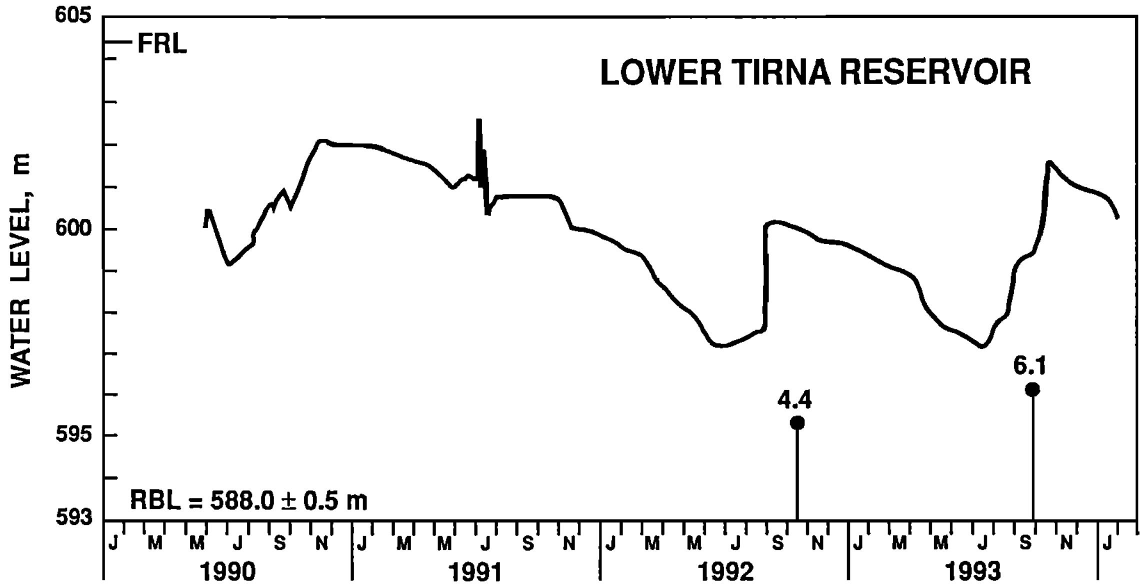

Figure 13. Water level in the lower Tirna reservoir and the timing and magnitudes of the mainshocks in the 1992 and 1993 sequences at Killari (modified from *Rastogi* [1994]). FRL is full reservoir level; RBL is river bed level.

possible location error [*Gupta*, 1992]. This zone is characterized by right-stepping fractures, which are indicative of left-lateral slip (inset, Figure 12), and is parallel to the nodal plane with left-lateral slip in the focal mechanism [*Sykes*, 1970]. Finally, the zone of *ground fissures* appeared to be a fault where exposed by a recent excavation (Superintending Engineer, Koyna Monitoring Cell, personal communication, 1994). The above evidence suggests that the coseismic fracture zone was the surface rupture of the 1967 main shock, although it was not recognized as a fault by *Sahasrabudhe et al.* [1969]. Similarly, the scarp zone at Killari was not widely recognized as a fault by local geologists prior to the evidence from trenches and from aftershock hypocenters [*Baumbach et al.*, 1994]. In both cases, hesitation in recognizing a primary rupture seems to derive from the lack of any geomorphic and structural evidence of a preexisting fault.

The main structural features found by *Sahasrabudhe et al.* [1969] in the Deccan Traps of the Koyna valley are a number of "shatter zones," which they characterized as steeply dipping planar zones of intense fracturing that could be traced horizontally for as much as 0.8 km and vertically for as much as 0.27 km but which showed no evidence of significant accumulated displacement and were not classified as faults. These shatter zones are reminiscent of zero-displacement faults found in cratonic terranes [e.g., *Isachsen et al.*, 1983]. Most of the shatter zones in the Koyna valley have a north-northeast strike (rose diagram in Figure 12), which is similar to the strike of 1967 fracture zone (Figure 12). Thus the geometry and the displacement characteristics of the shatter zones and of the fault that ruptured in 1967 are similar, suggesting that they may either have a common origin or that the 1967 rupture may have reactivated one of the shatter zones as a preexisting zone of weakness. We conclude that the 1967 main shock ruptured to the surface on a small-displacement fault that had no more than a few tens of meters of displacement accumulated during Cenozoic time. Thus the structural context of the 1967 Koyna earthquake resembles the one of the 1993 Killari earthquake and, generally, of $M \geq 6$ SCR earthquakes.

If the available geologic data on seismogenic faults and the historic rate of seismicity are representative of the long-term behavior of SCR, then the rate of earthquakes from individual faults can be reconciled with the rate of intraplate seismicity only by the presence of a large number of potentially seismogenic faults, the vast majority of them still unidentified [*Seeber and Armbruster*, 1993]. Ten earthquakes with $M \geq 6.0$ occurred in the stable continental portions of Australia and India from 1960 to 1993; of these, eight ruptured the surface. Let F be the number of faults that can produce surface ruptures, R be the average rate of these ruptures on an individual fault, and N be the number of surface ruptures in the time T. Then,

$$F = N/RT.$$

If $N=8$, $T=33$ years, and $R=10^{-5}$/year, then $F \approx 2 \times 10^4$ faults. Furthermore, these faults would need to be scattered over most of the stable continental portion of India and Australia, if the spacing between them allows a 20x20 km area for each of these faults. *Crone et al.* [1992] stress that data are scarce but they suggest that a rate of 10^{-5}/year may be at the high end of the range typical of SCR, which would make 2×10^4 faults a minimum value. These numerical results are very uncertain, but they point out that the low rate of paleoseismicity on individual faults combined with the historic rate of $M \geq 6$ earthquakes suggests many low-displacement but potentially seismogenic faults scattered widely over SCR. Such a tectonic regime is very different from the concentration of moment release on few large faults observed at plate boundaries. While along plate boundaries the determination of earthquake hazard on the basis of individual active faults may be a realistic goal, such an approach is likely to be misleading in SCR. In these areas, not only may potentially active faults be numerous and ubiquitous, but they are also likely to remain unknown until they produce a surface-rupturing earthquake because they are very difficult to identify from regional geological investigations.

Not only the widely scattered distribution, but also the depth distribution of SCR seismicity suggests a qualitative difference between interplate and SCR seismogenesis. The depth distribution of moment release along continental plate boundaries, such as the San Andreas fault zone in California or the Himalayan front in northern India, is generally consistent with the notion that the strain rate is uniform with depth and that earthquakes accommodate most of the deformation above the brittle-ductile transition, except in the upper few kilometers of the crust. Aseismic deformation may relax the stress in the highly fractured rock of the shallow subsurface along plate boundaries [*Scholz*, 1990, p. 95]. In contrast, most well-known $M \geq 6$ earthquake ruptures in SCR are concentrated in the upper few kilometers of the crust. A strong upper crust that can store elastic energy for a long time and is little affected by aseismic deformation is evidenced by near-surface faults that have very long repeat times and stress drops similar to ones that characterize faults in the lower crust. Furthermore, a depth distribution of moment release which is biased to the upper crust, seems inconsistent with tectonic models in which seismicity reflects uniform strain through the lithospheric thickness in response to far-field stresses.

In summary, a major component of the seismicity in SCR is concentrated in the upper crust, is widely scattered on many small-displacement faults, and accommodates horizontal shortening. Local processes in addition to forces applied at the boundary of the continental lithosphere may be required to account for widely scattered active faults and tectonic strain concentrated in the upper crust. The strain rate required to relax differential stresses due to the removal of overburden by denudation has been shown by *Anderson* [1986] to be of the same order of magnitude as the rate of deformation implied by the moment release in SCR. This mechanism could also account for the shallow depth range, the wide distribution of many active faults with low slip rates, and the prevalence of horizontal maximum compression axes.

Sources of Future Potentially Damaging SCR Earthquakes

Historic seismicity and geology were not helpful in identifying the Killari earthquake source prior to the event. The Killari area was included in the lowest of four hazard levels for peninsular India in the pre-1993 hazard map (Figure 1) [*Krishna*, 1992]. This map represents a perfectly justifiable interpretation of the historic earthquake data and of the geologic data in terms of a stationary regime, that is, assuming that "regions that have experienced earthquakes in the past, will also do so in the future during the expected life time of man-made constructions" [*Krishna*, 1992, p. 17]. Historic seismicity interpreted to reflect a stationary regime is the basis for most hazard maps worldwide. The main problem with this approach is that times between episodes of seismicity on individual SCR faults seem to be very long, as much as 3 orders of magnitude longer than the historic period [e.g., *Crone et al.*, 1992].

The Killari earthquake is typical of a class of shallow earthquakes in SCR that ruptured the surface and yet originated on faults that had not been recognized from seismological or geologic data. The data from these earthquakes suggest that the sources of future SCR earthquakes are also likely to remain unnoticed if the interpretation of the data is based on the assumption of stationarity. In retrospect, the burst of seismicity that preceded the 1993 Killari mainshock by a year and followed the impounding of the nearby reservoir can be considered a possible precursor of the mainshock [e.g., *Gupta*, 1992]. While the seismological community was not in a position to evaluate the significance of this possible precursor, the population in the Killari area apparently did sense the increased probability for more damaging earthquakes after the first one in 1992. Methodologies to interpret seismicity in terms of a probability gain for a damaging earthquake have been developed for interplate regions [e.g., *Jones*, 1985]. The results from Killari and from other areas suggest that phenomena such as an unprecedented burst of seismicity or the impounding of a reservoir, both of which may represent significant mechanical changes at seismogenic depths, should be included in hazard analysis. They could be considered as time-dependent components of the hazard to be superimposed on a stationary hazard distribution based on historical seismicity and geology.

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