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ORIGINAL PAPER

Possible evidence of remotely triggered and delayed seismicity due to the 2001 Bhuj earthquake (Mw = 7.6) in western India

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Abstract Remote triggering by large earthquakes at regional distances is a globally observed phenomenon. However, there are no reports of observations of dynamic triggering at regional distances of several source lengths associated with the large Mw = 7.6 Bhuj earthquake of January 26, 2001, in western India. In the present study, a swarm of over 140 microearthquakes that occurred about 500 km southeast of Bhuj, in the geothermal province of the Western Ghats in the Deccan volcanic province (DVP) of India, immediately after the occurrence of the Bhuj earthquake in 2001 is investigated. The post-Bhuj seismicity ($M < 2.0$) occurred in three bursts spread over 2 months with each burst of intense activity lasting for 2–3 days. All the three bursts of seismicity occurred in the same volume along a 5-km-long NW–SE trending fault. The temporal coincidence and the sudden rise in seismicity that interrupts the characteristically low background seismicity strongly suggest that the Bhuj earthquake may have remotely triggered this activity. The triggered seismicity began approximately 2.5 h after the onset of the Bhuj mainshock and continued well after the passage of the surface waves, suggesting that the dynamic stresses possibly gave rise to secondary time-dependent mechanisms leading to the triggering. It is proposed that the triggered and delayed seismicity is possibly a consequence of the redistribution in pore fluid pressure due to the Bhuj earthquake. This is the first documented observation of remotely triggered seismicity at regional distances due to the Bhuj earthquake.

Keywords Triggered seismicity · Coulomb stress · Geothermal · Western Ghats

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1 Introduction

In the past half century, several intraplate earthquakes with Mw $\geq$ 6 have occurred predominantly in the Deccan volcanic province (DVP) of India. The strong earthquakes (Fig. 1) that occurred in DVP are the Mw $\sim$ 6.3 at Koyna (1967), the Mw $\sim$ 6.2 at Latur (1993), the Mw $\sim$ 5.8 at Jabalpur (1997) and the Mw $\sim$ 7.6 at Bhuj (2001). Large earthquakes like the January 26, 2001, Bhuj earthquake are expected to alter the stress in the surrounding crust, leading to aftershocks and triggered events. Seismicity around the epicenter of Bhuj earthquake and up to distances of about 200 km continues unabated till date (Rastogi et al. 2012). However, there are no reports of observations of triggered seismicity associated with the Bhuj earthquake at regional distances beyond several source lengths. Due to sparse seismic networks in India, it is unclear whether the strong earthquakes in DVP have triggered earthquakes beyond the source region. Remotely triggered earthquakes are undetected primarily due to paucity of seismic networks with capabilities of capturing the microseismicity associated with large earthquakes. In the present study, a seismic network that was setup near Nasik (Fig. 1) a year prior to the Bhuj earthquake recorded an unusual swarm of microearthquakes that occurred immediately after the Bhuj earthquake. The swarm of microearthquakes continued for about 2 months after the Bhuj mainshock and then subsided. The sudden rise in the microseismicity and the timing strongly suggest that the Bhuj earthquake may have triggered this activity. The fact that the post-Bhuj microseismic activity occurs at a distance of about 500 km SE of Bhuj makes it an interesting data set to investigate for the possibility of remote triggering. This observation is quiet significant as it is the first documented evidence of remote triggering at

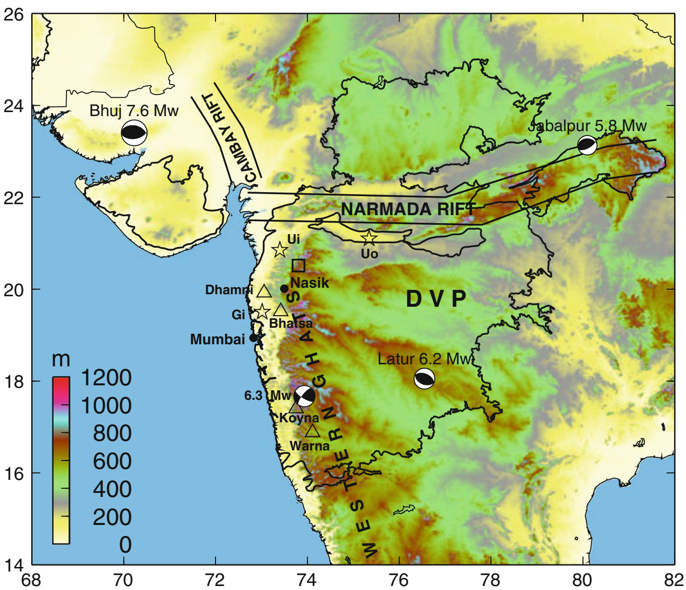

Fig. 1 Seismotectonic map of western India with the focal mechanisms of major earthquakes. The outlined region represents the Deccan volcanic province (DVP). The *open square* represents the study area, and the *triangles* represent the dams (Bhatsa, Dhamni, Koyna and Warna) associated with significant seismicity. The geothermal springs Unai(Ui), Unapdeo(Uo) and Ganeshpuri(Gi) are represented by *stars*

regional distances due to the Bhuj earthquake and secondly because it is one of the few examples of remote triggering in a stable continent by an intraplate earthquake.

2 Dynamic triggering

It is well known that the stress changes associated with an earthquake can induce or retard seismic activity in the surrounding region or trigger other earthquakes at great distances (Freed 2005). Earthquake triggering is primarily classified into two categories: static and dynamic. Earthquakes that occur within a few fault lengths from the causative event are due to static Coulomb failure stress changes (ΔCFS) (Reasenberg and Simpson 1992; Harris et al. 1995; Harris 1998). However, the ΔCFS are observed to be too small to trigger earthquakes at remote distances of a couple of source lengths (Hill et al. 1993). On the other hand, the transient, oscillatory stress changes transmitted as seismic surface waves, that is, dynamic stresses, which do not permanently change the applied load can trigger earthquakes at remote distances by altering the mechanical state or properties of the fault zone (Kilb et al. 2000). Thus, dynamic stress changes that are several orders of magnitude larger than ΔCFS have been associated with remotely triggered earthquakes, a process called dynamic triggering (Hill et al. 1993, 1995; Anderson et al. 1994; Gomberg and Bodin 1994; Gomberg 1996; Hill and Prejean 2007). The most well-studied phenomenon of dynamic triggering at remote distances (1,250 km) is the 1992 Landers earthquake of M $\sim$ 7.3, which initiated swarms at several places in Western United States (Hill et al. 1993; Anderson et al. 1994). Subsequently, dynamic triggering was shown to occur after the passage of surface waves from a large earthquake, for example, the 2002 Denali earthquake of Mw 7.9, which triggered swarms at distances of about 4,000 km (Eberthart-Phillips et al. 2003; Gomberg et al. 2004; Prejean et al. 2004; Hough 2007), and the 2004 Sumatra earthquake (Mw = 9.1)(West et al. 2005; Velasco et al. 2008). Although remotely triggered seismicity has been primarily observed to occur in active plate boundary regions (Gomberg et al. 2004; Brodsky and Prejean 2005; Hill and Prejean 2007), several studies have found that such remotely triggered events occur in intraplate regions with relatively low background seismicity (Hough et al. 2003; Hough 2007; Peng et al. 2009). Triggered seismicity at remote distances beyond the influence of static stresses from a large earthquake is observed especially in geothermal and volcanic provinces (Prejean et al. 2004). However, the mechanism, frequency, controlling factors and the global extent of dynamic triggering are yet to be fully understood (Velasco et al. 2008). There are several observations of the triggered activity persisting hours to 1 week or more after the seismic waves from the Landers event subsided, which emphasizes that the triggered activity is not driven solely by the dynamic stresses (Hill et al. 1993). Mechanisms for dynamic triggering fall into two broad categories (Hill 2008; Hill and Prejean 2007; Gonzalez-Huizer and Velasco 2011): (1) models related to the excitation of crustal fluids or aseismic creep or (2) frictional models described by Coulomb frictional failure. A range of physical processes have been proposed by several researchers to explain remote earthquake triggering which include changes in the mechanical state or properties of existing faults (Gomberg et al. 1998; Voisin 2002), redistribution in pore fluid pressure of a region's hydrothermal system (Hill et al. 2002; Brodsky and Prejean 2003), changes in the state of magma bodies (Linde et al. 1994; Hill et al. 2002) and changes in fluid pressure as bubbles oscillate or rise through fluid or fluid saturated rock (Hill et al. 1993; Linde et al. 1994; Linde and Sacks 1998; Brodsky et al. 1998).

3 Tectonic overview of Bhuj earthquake

The January 26, 2001, Bhuj earthquake in western India is one of the largest intraplate events (Mw = 7.6) in this tectonically active region (Fig. 1), which is a part of a network of failed rifts that developed as a consequence of the breakup of the southern supercontinent of Gondwana (Biswas 1982; Talwani and Gangopadhyay 2001). The Bhuj earthquake, which lies in a tectonically active thrust belt, occurred at a depth of 23.4 km (Harvard hypocenter) and did not exhibit any prominent surface ruptures along any fault (Malik et al. 2000). The absence of primary surface rupture indicates movement along a blind fault that has been accompanied by widespread liquefaction. This mainshock was followed by several hundreds of aftershocks. Fault plane solution using teleseismic studies represents thrust motion with a small strike-slip component on moderately dipping planes striking east–west (strike 82°, dip 51° and rake 77°) (Antolik and Dreger 2003). The rupture zone was calculated to be 375 km^2 with the primary rupture directivity toward the northwest with a larger amount of slip (8.2 m) confined at the center of the rupture as compared to the slip (1.7 m) in the surrounding part (Antolik and Dreger 2003).

4 Study area

The study area (Fig. 1) is located in the Western Ghats at an elevation of 800 m above m.s.l. in the Deccan volcanic province. The Deccan basalts are about 1.5 km thick and predominantly composed of tholeiites. Several geothermal springs exist along the Western Ghats and the Son-Narmada lineament (Fig. 1) with the closest thermal spring at Unai located about 50 km north of study region (Fig. 1). The Western Ghats escarpment runs parallel to the west coast of India and is host to over 33 dams. Microseismicity that occurs all along the Western Ghats is observed to be associated with several dams (Fig. 1) e.g., Koyna, Bhatsa, Dhamni (Gupta 2002; Guha 2000; Rastogi et al. 1986, 1997). While reservoir-induced seismicity is the preferred explanation for the seismicity near dams (Gupta 1992), several other explanations were proposed for the seismicity related to the upliftment of the Western Ghats due to isostasy and lithospheric flexuring (Catherine et al. 2007).

5 Seismic network and data analysis

A seismic network comprising of eight digital stations was established temporarily in phases during the period 2000–2002 in a 30 km × 30 km region about 50 km north of Nasik in DVP (Fig. 2). The network consisted of one Guralp CMG 3T broadband seismometer with a natural period of 100 s and seven 1-Hz short-period seismometers. The data were recorded in a continuous mode with a sampling frequency of 50 Hz. In the two-year period from March 2000 to March 2002, about 217 microearthquakes with earthquake magnitude referred to as M $\sim$ 1.0–3.7 were recorded by the network. Most microearthquakes (181) were of magnitude M $<$ 1.5, 27 events were of M $\sim$ 1.5–2.5 and 9 were of M $>$ 2.5. Hypocenters were computed for 79 events recorded by four or more stations with HYPOCENTRE program using the P velocity model obtained from the deep seismic sounding (DSS) profiles at Koyna (Kaila et al. 1981) in DVP about 150 km south of Nasik. A Poisson's ratio of 0.26 (Tiwari et al. 2006) obtained through receiver function studies at a broadband station (MPAD) in the seismic network (Fig. 2) was used to constrain the velocity structure. Most microearthquakes were near-surface events confined to shallow

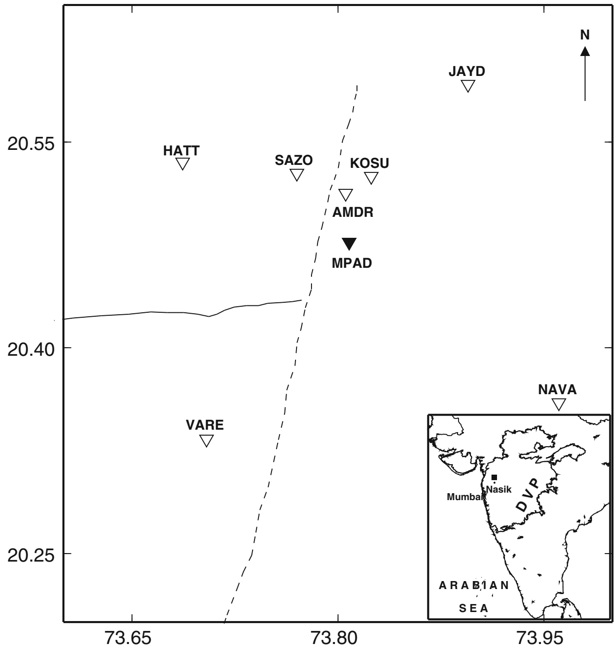

Fig. 2 Location map of the seismic network. *Open* and *filled inverted triangles* denote short-period and broadband stations, respectively. The *dashed line* denotes a lineament, and the *solid line* represents a mafic intrusion (GSI 2000). The inset map of India shows the outline of the Deccan volcanic province with the study area marked by a *filled square*

depths of less than 2 km. The hypocentral parameters were determined with small rms errors of $\leq$0.15 for most events. The events recorded by single or two stations were analyzed using the approach of predictive coherence method (Roberts et al. 1989) to calculate azimuths, and the time differences between S and P phases were used to determine epicentral distances.

Transient stresses caused by the passage of surface waves play an important role in triggering events at remote distances. To identify the local events triggered within the coda, the raw seismograms were filtered and analyzed. A 5–20-Hz bandpass filter was applied to the seismogram recorded at station AMDR to effectively eliminate the mainshock coda waves and enhance the triggered local earthquakes (Fig. 3). The raw data was also filtered in the range 0.033–0.2 Hz to enhance the 5–30-s period surface waves in the record. The filtered records do not indicate any local shocks corresponding to the low-frequency Rayleigh waves. The spectrogram (Fig. 3) of the unfiltered record also does not indicate any sharp, prominent excursion in the frequency bands that might correspond to local events implying absence of local triggering within the coda. Although dozens of aftershocks occurred immediately after the Bhuj mainshock, a 20-h normalized record of day 1, that is, January 26, 2001, plotted in UTC along with the corresponding spectrogram (Fig. 4) reveals the presence of several local shocks that occurred much after the Bhuj mainshock. The spectrogram is dominated by the aftershocks that to a great extent mask the local events. Taking the spectrogram of the local shocks in a one-hour interval on January 28, 2001, during which the aftershocks of Bhuj mainshock were not dominant, reveals the frequency content of the local shocks clearly (Fig. 4).

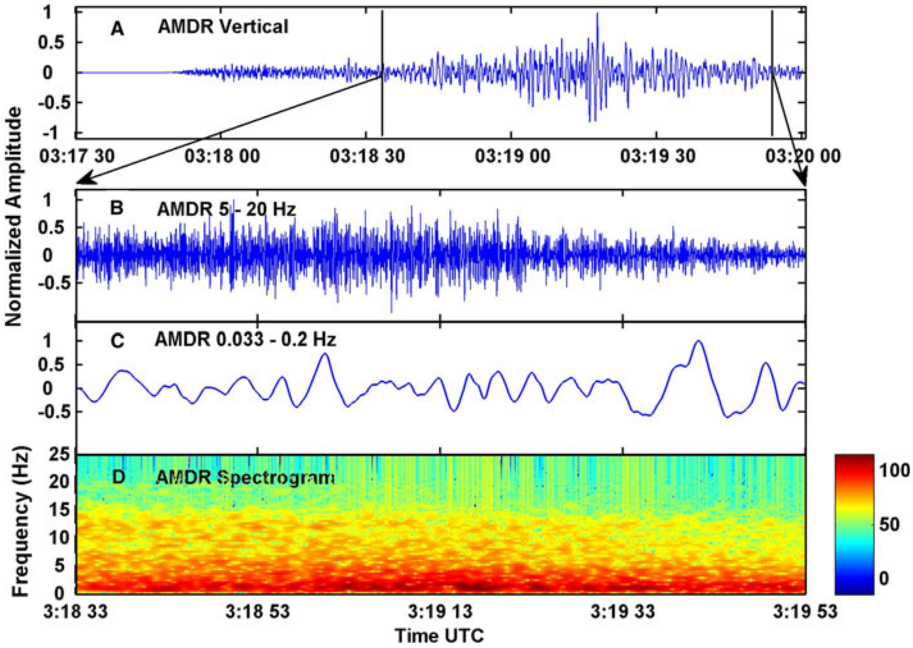

Fig. 3 Seismic record of the Bhuj earthquake at station AMDR. **a** Vertical component of seismic record from the short-period station AMDR, plotted as a function of UTC on January 26, 2001. **b** Expanded view of surface waves filtered in the bandpass 5–20 Hz to identify high-frequency local shocks, **c** Expanded view of surface waves bandpass filtered from 0.033 to 0.2 Hz (5–30 s) to identify the large-amplitude low-frequency surface waves, **d** Spectrogram showing the frequency content of the unfiltered signal, indicating the absence of local shocks in the coda

6 Spatial and temporal distribution of local seismicity

The spatial distribution of the local seismicity that occurred during the period 2000–2002 along with the histogram of seismicity is plotted in Fig. 5. The pre-Bhuj background seismicity ($M \approx 1.0$–3.1) is not confined to any particular fault and occurs predominantly outside the seismic network. Out of the 26 pre-Bhuj events, only 5 microearthquakes were centered within the network. About 12 events were of $M \geq 2.0$, with the largest earthquake recorded being of $M \sim 3.1$, near Nasik, which occurred one day prior to the Bhuj earthquake. In contrast, the local seismicity triggered immediately after the Bhuj mainshock, on January 26, 2001, referred to as triggered seismicity, continued for a period of 2 months, and occurred in a small source volume within the network (Fig. 5). Most of the triggered seismicity (143 events) is seen to be spatially confined to a narrow 5-km-long elongated zone. The zone of seismicity trends NW–SE intersecting a 100-km-long NE–SW lineament. The microseismicity is seen to be confined to shallow depths of about 2 km, that is, within the Deccan Traps. Interestingly, the post-Bhuj (2 months after the mainshock) seismicity (48 events) is clustered along the northern portion of the NE–SW lineament and also coincides partially with the zone of triggered seismicity.

The temporal distribution of seismicity in the period 2000–2002 is shown in Fig. 6. In the entire one year preceding the January 26, 2001, Bhuj event, merely 26 local earthquakes were recorded by the seismic network. These events were temporally spread

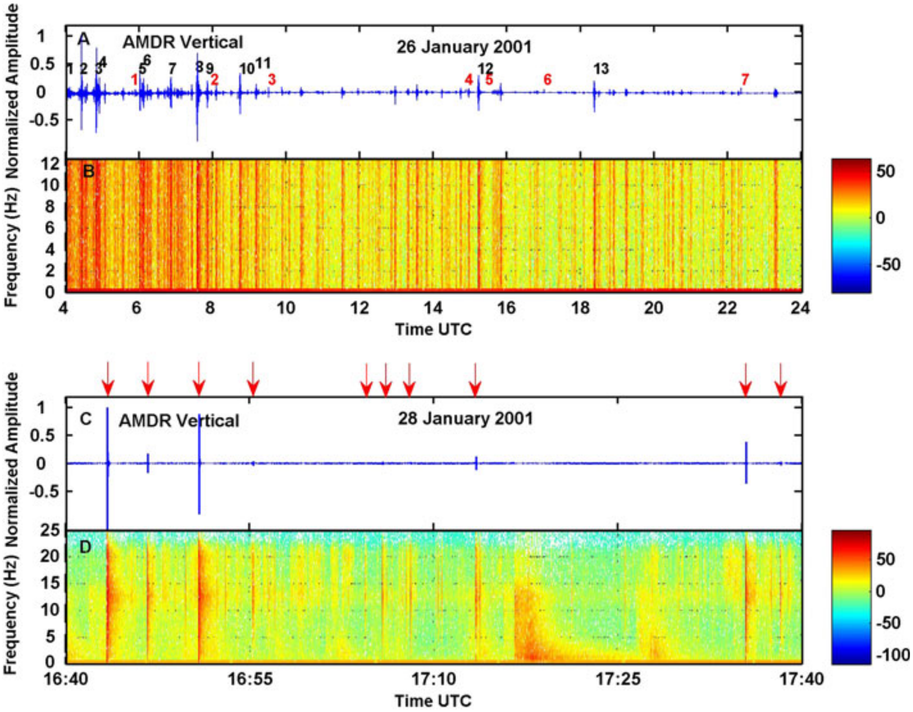

Fig. 4 Normalized seismic record (downsampled to 25 Hz) of 20-h duration at station AMDR on January 26, 2001, post-Bhuj mainshock exhibiting the Bhuj aftershocks (M > 4.0) and the local shocks numbered sequentially in *black* and *red*, respectively, along with the corresponding spectrogram. In the absence of strong aftershocks, the local shocks in a one-hour interval on January 28, 2001, are seen clearly in the record and their frequency content seen in the corresponding spectrogram. The *arrows* indicate timing of the local events

out and were not clustered together. Similarly, the local seismicity from April 2001, 2 months after the Bhuj earthquake, remained feeble with about 48 microearthquakes being recorded in the entire year 2001–2002. Most of the microearthquakes were of magnitude <2.0 and were well separated in time. Thus, the seismicity that occurred pre-Bhuj and two month post-Bhuj does not exhibit any temporal clustering of events.

An anomalous increase in the seismicity is observed immediately after the occurrence of the Bhuj mainshock (Fig. 6). The abrupt increase in seismicity within a span of just 2 months is unique in the catalog. The trend of the cumulative number of earthquakes from March 2000 to March 2002 also shows a dramatic increase immediately after the Bhuj mainshock, which persists for about 2 months and then returns to the normal background trend till March 2002 (Fig. 6). In the two month post-Bhuj period, 143 local events were recorded starting from the day of occurrence of the Bhuj mainshock. The overall background seismicity for the area monitored for the period of 2 years is quiet low (Fig. 6) as compared to the triggered seismicity, which contributes to nearly 65 % of the total catalog. A distinct difference between the triggered and the background seismicity is that while 50 % of the background seismicity comprised of events with M > 2.0, only 10 % of the triggered seismicity were of magnitude M > 2.0.

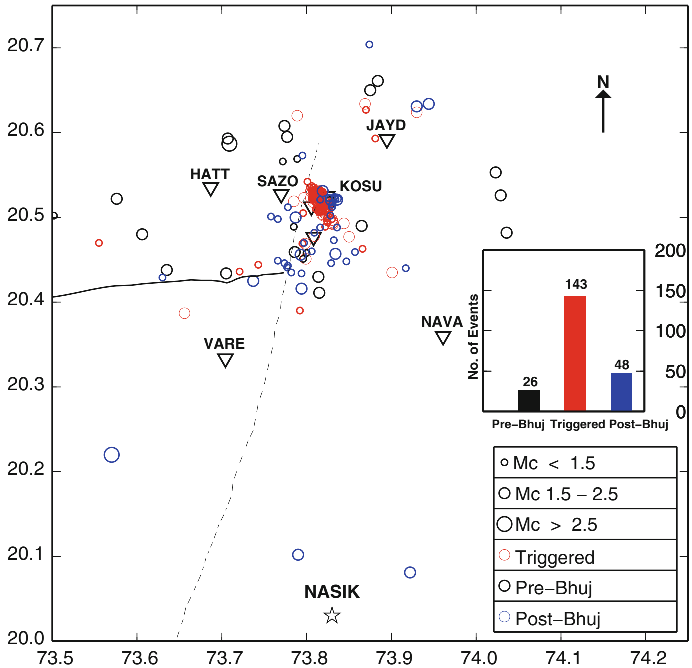

Fig. 5 Spatial distribution of local microseismicity during the period 2000–2002. Pre, triggered and post refer to the local seismicity relative to the Bhuj mainshock. The seismicity pre-Bhuj, triggered (post-Bhuj for 2 months) and the post-Bhuj (2 months after Bhuj mainshock) are denoted by *blue*, *red* and *black open circles* along with the corresponding histograms (*inset*)

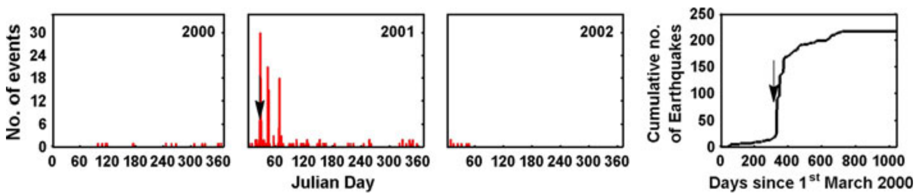

Fig. 6 The count of local events with time that occurred pre- and post-Bhuj earthquake during 2000–2002 (*left*). Cumulative number of locatable earthquakes beginning from March 1, 2000 (*right*). The time of occurrence of the Bhuj earthquake is indicated by an *arrow*

The post-Bhuj triggered seismicity was temporally distributed in three clusters spread over 2 months with each cluster of intense activity lasting for about 2–3 days (Fig. 7). The first local microearthquake occurred approximately 2.5 h after the Bhuj mainshock was recorded, after which 6 more microshocks were recorded on the same day. The seismicity increased the next day (27 January) with around 30 microearthquakes with Mc 1.0–2.2 occurring on a single day (Fig. 7). The first significant microearthquake was triggered at 00:54 UTC on January 27, 2001, with a magnitude 1.5, 22 h after the Bhuj mainshock. The post-Bhuj microseismic activity was intense for the initial 3 days with 54 local events being recorded (Fig. 7). Similarly, there were two more bursts of intense microseismicity in the period February 12–14, 2001, and March 7–8, 2001, during which 38 and

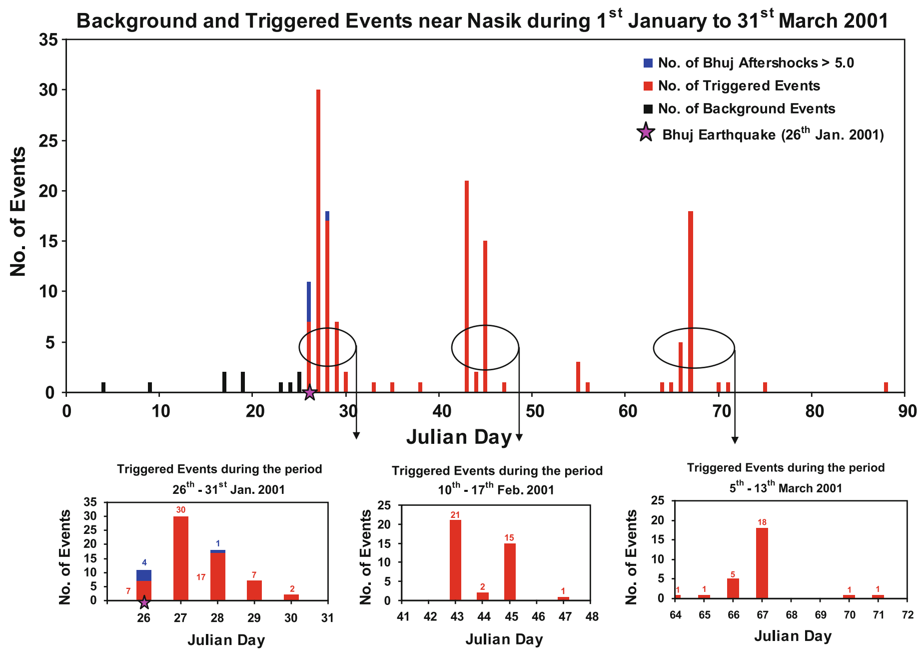

Fig. 7 Histograms of daily occurrence of microearthquakes in the 3 months (January–March 2001) (*top*). The detailed histograms for the three swarm bursts in January, February and March 2001 after the January 26, 2001, Bhuj earthquake (*bottom*)

23 microearthquakes were, respectively, recorded. The seismicity diminished gradually over the subsequent months marked by absence of abnormal clusters of microearthquakes.

7 Triggering mechanism

Microseismicity temporally correlatable with large earthquakes at regional distances has generally been attributed to variations in the dynamic stresses that are much larger than the static Coloumb failure stress changes (ΔCFS). The static stress due to the Bhuj earthquake is confined to a very small area near the source region, and due to the high stress drop (Antolik and Dreger 2003), the predicted ΔCFS at distances >500 km are insufficient to remotely trigger microearthquakes. The intensity in the study area due to the Bhuj earthquake is VI, indicating strong shaking, and corresponds to peak acceleration of 9.2–18 % of g (Hough et al. 2002). Thus, dynamic stress transients associated with low-frequency surface waves could play a major role in the triggering mechanism. However, there is no evidence of dynamic triggering resulting from the passage of surface waves associated with the Bhuj mainshock. In fact, the local microseismicity occurred well after the passage of the surface waves and continued for about 2 months post-Bhuj mainshock, implying that the transient dynamic stresses may not have a direct effect on the triggering process. Thus, there should be some other process by which the dynamic stress transients are transformed into sustained stress changes that are responsible for the prolonged triggered seismicity. For example, large seismic swarms were observed hours to days after the Denali fault earthquake in Western United States, suggesting a delayed response (Prejean et al. 2004). Hill et al. 1990 note that earthquake swarms of this type (spasmodic bursts) are distinctly different in character from mainshock–aftershock earthquake sequences and are commonly observed in geothermally or volcanically active environments. A variety of time-dependent stress transfer mechanisms, such as viscous relaxation, poroelastic rebound, or afterslip or reduction in fault friction are proposed to explain delayed earthquake triggering ranging from a few seconds to decades (Freed 2005).

The most favorable mechanism that can explain delayed seismicity from a few hours after the causative event to a few days is redistribution in pore fluid pressure (Hill et al. 2002; Brodsky and Prejean 2003) since (1) water level fluctuations were locally observed, although not documented in the study area and it is well known that large earthquakes affect the water levels in wells thousands of kilometers away from the epicenter (e.g., Coble 1965; Roeloffs et al. 2003), (2) earthquakes are triggered remotely by changes in hydrothermal systems in hydrothermally/volcanically active regions (Hill et al. 1993, 2002), (3) the study area is within the geothermal province of the Western Ghats and the Narmada Tapti region with the closest thermal spring located just 50 km north of the area of seismicity (Fig. 1), (4) it is possible that the dynamic stresses in turn lead to secondary time-dependent processes leading to the triggered seismicity (Prejean et al. 2004). Finally, delayed seismicity at Yellowstone, Mount Rainer and the Long Valley are examples that suggest that more than one mechanism generate remotely triggered earthquakes (Prejean et al. 2004). In the present study, the three bursts of post-Bhuj seismicity have occurred along a 5-km-long NW–SE trending fault, with the first episode seen along the entire fault while the second and third bursts were limited to approximately half and one-fourth the fault length. It is likely that in response to the seismic waves from distant earthquakes, the shallow rupturing of the basalts provided a pathway for fluids from depth to rise to areas of low pore pressure, possibly leading to unstable rupture and triggering earthquakes. Alternate mechanisms for generation of swarms cannot be unequivocally ruled out as observations over the past one decade reveal that compared with pre-Bhuj, the post-Bhuj seismicity shows a significant increase in the regions surrounding Bhuj in Kutch and Saurashtra (Fig. 1), leading to theories on viscous relaxation being a probable cause for the triggered seismicity (Rastogi et al. 2012). Studies on earthquake swarms (Srivastava and Dube 1996) indicate that swarms that are confined to shallow depths are non-precursory while events continuing deeper up to about 8–10 km could result in large earthquakes and hence are precursory in nature. In the last 10 years, no major earthquake has been reported in the study region, suggesting that the seismicity in the study area appears to be non-precursory.

8 Conclusions

The characteristically low background seismicity in the study region in the Western Ghats of the DVP is interrupted by three bursts of swarms over a period of 2 months post-Bhuj mainshock of January 26, 2001. The intense microseismicity was confined to a 5-km-long NW–SE trending fault within the basalts. The temporal coincidence between the Bhuj earthquake and the swarms about 500 km away from the epicenter of the Bhuj earthquake is suggestive of remote triggering at regional distances. The triggered seismicity began and continued well after the passage of surface waves, suggesting that the dynamic stresses possibly gave rise to secondary time-dependent mechanisms leading to the triggering. It is proposed that the delayed earthquake triggering from a few hours to a few days after the causative event is possibly a consequence of the redistribution in pore fluid pressure due to the Bhuj earthquake.

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