Seismogenic properties of the crust inferred from recent studies of reservoir-induced seismicity – Application to Koyna

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Recent studies of reservoir-induced seismicity have led to a better understanding of the seismogenic crust. These studies reveal that many parts of the seismogenic crust are critically stressed and small strength changes induced by reservoir impoundment can lead to seismicity. In the vicinity of the reservoir impoundment causes an immediate undrained response to the loading, and a delayed response due to the diffusion of pore pressure. Analytical calculations of strength changes at Monticello Reservoir, South Carolina show that diffusion is the predominant mechanism in triggering seismicity. Seismicity associated with pore pressure diffusion occurs in a crust with seismogenic permeability, k_s, ~ 2 to 200 mD. Analyses of seismicity data at Koyna reveal that the rocks may have stress-memory (they remember the largest load that they were subjected to), and that seismicity occurs within a block enclosed by NNE-SSW trending Koyna River fault zone on the west, NE-SW trending Patan fault and a subsidiary fault on the east and a series of NW-SE trending faults to the north and south. Continued monitoring of water levels in observation wells and their correlations with lake levels will likely produce the next breakthrough in understanding the protracted seismicity in the Koyna–Warna area.

Introduction

In recent years there has been a marked improvement of our understanding of the nature of the seismogenic crust and the physics of the earthquake process. Induced seismicity, especially reservoir-induced seismicity (RIS), has contributed greatly to this understanding, because the temporal and spatial association of induced earthquakes with a causative source can usually be inferred. Some of the parameters determined for induced earthquakes are also common to tectonic earthquakes and thus are applicable to the processes responsible for the generation of earthquakes. In this paper I will briefly review some recent findings about the physics of the earthquake process as inferred from a study of induced seismicity. Observations from one of the most enigmatic cases of RIS in the Koyna–Warna area are presented, with suggestions for further studies.

Induced or triggered?

As an aside, there has been some discussion on whether the earthquakes resulting from reservoir impoundment are induced or triggered. McGarr and Simpson[1] arbitrarily chose the magnitude of the stress (or pore pressure) change as a discriminant. They used the adjective 'induced' to apply to seismicity that results 'from a *substantial* change in crustal stress or pore pressure from its ambient state'. They noted that 'induced seismicity' is that for which the causative activity can account for . . . *most of the energy* required to produce the earthquakes. They contrasted it from 'triggered' seismicity which is applied to 'a situation for which the crust is sufficiently close to failure state due to natural tectonic processes that *only a small change* in either stress or pore pressure stimulates earthquakes'. This change accounted for only '*a small fraction* of the stress change or energy associated with the earthquake' (emphasis added). Another idea was that if the earthquakes would have occurred anyway, and the impoundment of the reservoir caused them to happen earlier than they normally would have, they could be called triggered. If, on the other hand, it was only due to the change in the stress condition brought on by human activity (e.g. impoundment of a reservoir, fluid injection in wells, quarrying of mines) that the earthquake occurred, we will call them induced.

I find this discrimination artificial, since *a priori* we cannot prove that the earthquake would have occurred without human activity. Would the ~ 100,000 earthquakes at Koyna and ~ 10,000 earthquakes at Monticello Reservoir have occurred without the impoundment of reservoirs? I use a more literal definition, induce – to bring about, or cause; and do not distinguish it from trigger, to initiate something. Consequently, I will use induced and triggered interchangeably for reservoir-stimulated seismicity.

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Critically stressed seismogenic crust

Some of the earliest associations of reservoir impoundment and pursuant seismicity suggested that very small changes in the stress field were adequate to induce earthquakes. Impoundment of a 100 m deep reservoir causes a stress change of 1 MPa at the bottom of the reservoir. Stress changes at hypocentral locations are smaller. There are several cases of reservoirs shallower than 100 m which have been associated with earthquakes[2]. Stress changes as small as 0.01 MPa have been associated with RIS[3]. Calculations of strength changes at the hypocentral location of earthquakes at Monticello Reservoir showed that changes as low as 0.1 MPa were associated with the initial seismicity[4]. In recent years, small stress changes caused by an earthquake have been shown to enhance or retard seismicity on neighbouring faults (e.g. Harris, this issue). Experimental induction of earthquakes at a depth of ~ 9 km in the KTB well showed that pore pressure changes of ~ 1 MPa were enough to trigger earthquakes[5]. Grasso and Sornette[6] have reviewed several cases wherein small stress perturbations (≤ 1 MPa) are associated with induced seismicity at reservoirs, injection wells, and locations of hydrocarbon withdrawal and those of geothermal activity. All these observations underscore the conclusion that at many locations in the seismogenic crust the rocks are critically stressed and small stress or pore pressure changes can trigger earthquakes.

Mechanism of RIS

Impoundment of a reservoir, and changes in lake levels can induce seismicity in two ways. One, the rapid response due to undrained effect and two, the delayed response to diffusion of pore pressure (for a detailed discussion, see e.g. Simpson *et al.*[7], Rajendran and Talwani[8], Talwani[9] and Chen and Talwani[4]). Here we briefly summarize the two effects. RIS is caused by shear failure on a pre-existing fault plane. According to Coulomb's law, ΔS, the strength change along a pre-existing fault plane due to reservoir impoundment is given by[10]

$$\Delta S = \mu\,(\Delta s_{\mathrm{n}} - \Delta P) - \Delta\tau, \qquad (1)$$

and

$$\Delta P = \Delta P_{\mathrm{u}} + \Delta P_{\mathrm{diff}}, \qquad (2)$$

where Δs_{n} and Δt are the changes in normal and shear stresses, and ΔP_{u}, ΔP_{diff} and ΔP are the undrained, diffusion induced and total pore pressure changes, and μ is the coefficient of friction. Negative values of ΔS signify weakening while positive values imply strengthening.

Neglecting nonlinear effects the subsurface responds elastically to the reservoir loading by a change in normal and shear stresses on the fault plane. An increase in normal stress strengthens the subsurface fault, while a change in shear stress may weaken or strengthen the fault depending on the orientation of the fault relative to the stress field. The instantaneous, or undrained change in strength, ΔS_{u}, results from Δs_{n}, Δs and ΔP_{u},

$$\Delta S_{\mathrm{u}} = \mu\,(\Delta s_{\mathrm{n}} - \Delta P_{\mathrm{u}}) - \Delta t\,. \qquad (3)$$

The total change in strength ΔS, includes both the instantaneous undrained ΔP_{u}, and the delayed increase in pore pressure due to diffusion, ΔP_{diff} (eqs (1) and (2)).

Strength changes associated with impoundment – The predominant role of diffusion

In most examples of RIS, we note that there is a time lag between impoundment and the onset of seismicity. This is usually attributed to the delayed effect of diffusion[7–9]. Simpson *et al.*[7] suggested that seismicity occurs at some reservoirs primarily because of undrained strength

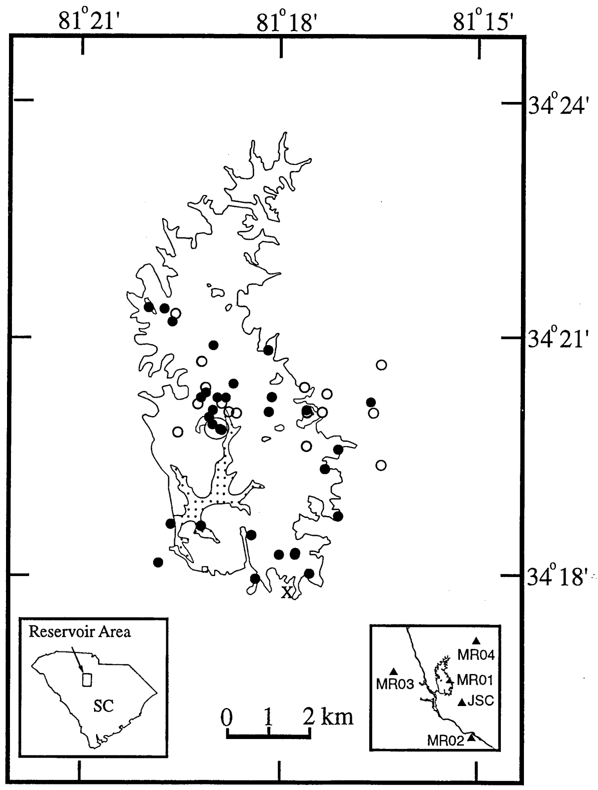

Figure 1. Locations of initial seismicity from December 1977 to January 1978 at Monticello Reservoir. Insets show the location of the reservoir in South Carolina and the seismic stations used to locate the earthquakes. Open and solid circles show locations of earthquakes where the undrained effect due to impoundment of the reservoir (ΔS_{u}) resulted in weakening and strengthening, respectively. X shows the location of the earthquake with the largest strengthening. The dotted area shows the deepest part of the reservoir where no initial seismicity was observed (from Chen and Talwani[4]).

changes ΔS_u, whereas diffusion of pore pressure is responsible at others.

Recently, Chen and Talwani[4] analysed the strength changes, ΔS and ΔS_u associated with the impoundment of Monticello Reservoir. ΔS and ΔS_u were calculated at hypocentral locations of 53 events with B location quality (HYPO71 output) during the first month following impoundment. Of these, 16 events were associated with weakening due to loading (ΔS_u) (Figures 1 and 2 *a*). Most of these were shallow ($Z < 1$ km) and about half of them occurred outside the periphery of the rising lake levels. At the hypocentral locations of all other events impoundment of the lake resulted in strengthening (positive ΔS_u, Figures 1 and 2 *a*) and would therefore not result in the observed seismicity. Total strength changes, ΔS, for different values of hydraulic diffusivity, C, show that in turn weakening at hypocentral locations occurs as a result of diffusion (Figure 2 *b–d*). Figure 2 thus illustrates that except for a few shallow earthquakes that occurred on the periphery of the reservoir in response to impoundment, RIS was predominantly due to pore pressure changes associated with diffusion and that at Monticello Reservoir the hydraulic diffusivity of the fractured rocks was ~ 5 to 10 m^2/s.

Bad Creek experiment

In a field experiment near Bad Creek Reservoir in northwestern South Carolina, Talwani *et al.*[11] compared water level changes in the reservoir with correlative changes in an observation well connected to the reservoir by a 250 m long, ~ 1 m thick shear zone. Analyses of the data for a five-year period showed that pore pressure fluctuations at the bottom of the reservoir in response to lake level fluctuations were transmitted to the bottom of the well through the shear zone by diffusion. The depth and amount of diffused pore pressure depend on the frequency of lake level fluctuations. The experiment also showed the feasibility of determining *in situ* hydraulic diffusivity and permeability of the shear zone. The permeability of the shear zone was found to lie at the lower end of the range for seismogenic permeability, discussed later.

The nature of RIS

Talwani[9] divided RIS into two categories, initial and protracted, according to its time history. This section is taken

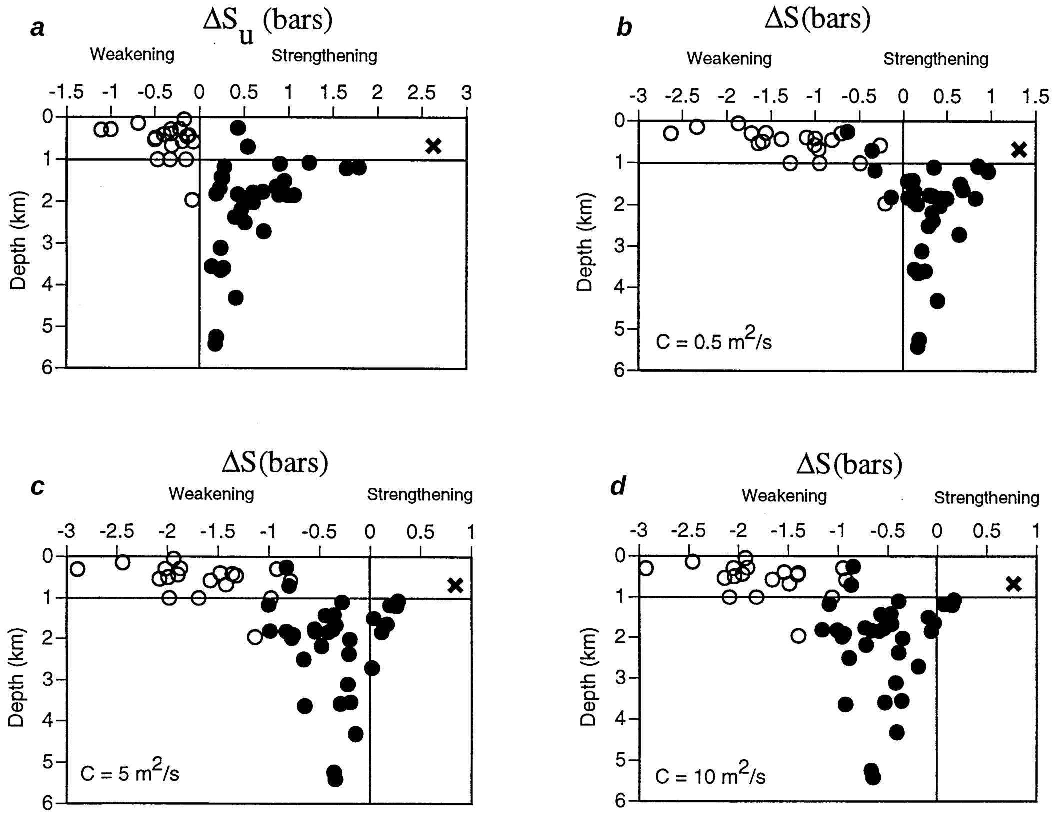

Figure 2. Comparison of undrained strength changes, ΔS_u (***a***) with coupled poroelastic strength changes, ΔS for $C = 0.5$, 5 and 10 m^2/s (***b–d***). Open circles are events that show weakening due to undrained effect and whose locations are shown in Figure 1. Weakening increases with increasing C. X shows the earthquake with largest strengthening. Horizontal scale indicates the change in strength in bars. Zero, positive and negative values indicate no change in strength, strengthening and weakening, respectively (from Chen and Talwani[4]).

from that review. Initial seismicity was ascribed to the coupled poroelastic response of the reservoir to initial filling or water level changes. It is characterized by an increase in seismicity above pre-impoundment levels, large event(s), general stabilization and (usually) a lack of seismicity beneath the deepest part of the reservoir, widespread seismicity on the periphery, migrating outward in one or more directions. With time, there is a decrease in both the numbers and magnitudes of earthquakes, with the seismicity returning to pre-impoundment levels.

Protracted seismicity was used to describe the situation at some reservoirs which continue to be active several years after impoundment. Protracted seismicity was found to depend on the frequency and amplitude of lake level changes, reservoir dimensions and the hydraulic diffusivity of the substratum. Longer period lake level changes were found to cause deeper and larger earthquakes than shorter period changes. Earthquakes occur at reservoirs where the lake level changes are comparable or a large fraction of the least depth of water. The seismicity is likely to be more widespread and deeper for a larger reservoir than for a smaller one. The continuing seismicity at Koyna is the most distinctive case of protracted seismicity, and I will address it again in later sections.

Seismogenic permeability

As we have seen earlier, diffusion of pore pressure plays an integral role in RIS. The physical parameter associated with the time of transmission of pore pressure and its magnitude is hydraulic diffusivity, C. This quantity C is linearly related to the intrinsic permeability of the fractured rock matrix, k, and inversely to its storativity. The latter depends on the effective compressibility of the rock matrix (b_{eff}) which in turn depends on its porosity, f. C depends strongly on k and weakly on f.

In well-documented cases of RIS, where the timing and location of lake level change and resultant seismicity are known (or can be inferred), C can be calculated from the hypocentral distance from the reservoir, the time lag between the lake level changes and onset of seismicity[12]. Seismogenic fractures were found to be associated with a hydraulic diffusivity lying between ~ 0.1 and 10 m^2/s. For reasonable estimates of j and b_{eff}, k can be calculated. For over 50 cases Talwani and Chen[13] found k to lie in the range 2 to 200 mD (2×10^{-15} to 2×10^{-13} m^2). (In nature the permeability of rocks varies between 10^{-24} m^2 and 10^{-9} m^2 (refs 14 and 15).) Talwani and Chen[13] labelled the permeability associated with RIS, seismogenic permeability, k_s, and suggested that it is a characteristic value for fractured rocks where the seismicity is associated with an increase in pore pressure.

If $k < k_s$, application of additional loads results in elastic deformation of the rock matrix with negligible increase in pore pressures (undrained response). If $k > k_s$ the application of additional stresses results in fluid flow without an increase in fluid pressure (drained response). Only when $k = k_s$, is there diffusion of excess pore pressure from the source of the additional stress through the rock matrix, to hypocentral locations.

Stress memory in rocks – Evidence from Koyna

Following the impoundment of Shivajisagar (Koyna Reservoir) seismicity has been monitored and observed continuously near Koyna in Maharashtra, since 1963. Through May 1995 more than 90,000 earthquakes ($M \geq 1.5$) have been recorded at an average of over 4000 events annually[16]. These include more than 1400 events with $M \geq 3.0$, 80 events with $M \geq 4.0$ and 8 events with $M \geq 5.0$ including the destructive M 6.3 event on 10 December 1967 (UTC) (4.21 a.m. on 11 December 1967 local time). The observed seismicity is the longest known sequence of RIS[16]. Unlike most other cases there are two unique aspects of RIS at Koyna. The amplitudes of lake level changes are of the same order as the initial height of impoundment $H1$, and the continuing occurrence of M 4 and 5 events long after the initial impoundment (Figure 3).

After the 1967 events and before the start of the impoundment of the Warna Reservoir ~ 30 km to the south, two other episodes of $M \geq 5.0$ occurred in 1973 and 1980. Including the 1967 events, we shall refer to these as episodes I–III. Earlier attempts to explain the three episodes include suggestions of anomalously faster filling rates[17], and the occurrence of pairs of earthquakes of $M \geq 4.0$ before $M \geq 5.0$ events[18]. Examination of over 30 years of data shows that there are several instances of increased filling rates and M 4.0 pairs without pursuant M 5.0 events. Further examination revealed a possible temporal association of these episodes with maximum lake levels.

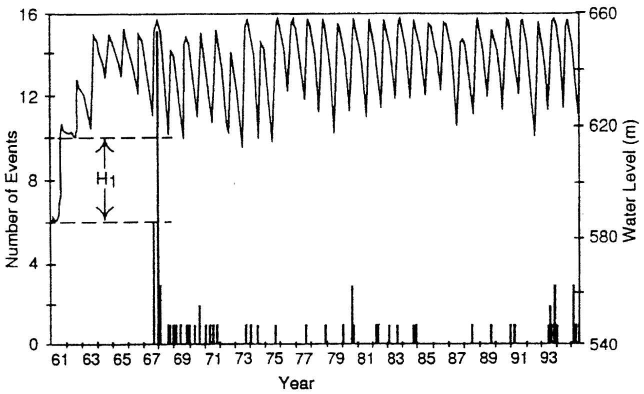

Figure 3. Daily water levels (elevations above mean sea level) at Koyna from 1961 to 1995 and $M \geq 4.0$ events. Three episodes of $M \geq 5.0$ activity occurred in 1967, 1973 and 1980. H_1 is the initial height of the water level and it represents the lowest levels during any cycle of filling and emptying (from Talwani[16]).

Figure 4 shows the maximum water levels (H_{max}) and times of episodes I–III. The dates when the new H_{max} is reached (solid dot) and exceeded (cross) are shown. The horizontal bar indicates the time during which the previous H_{max} was not exceeded (see also Table 1). Filling started in 1961 and the date H_{max} was reached and the H_{max} value are given in columns 2 and 3 in Table 1. The date in the next year when H_{max} was exceeded (11 July 1962) is given in the first column and shown by a cross in Figure 4. The date that H_{max} was reached in 1962 (08-15-1962) is given in column 2 and its value in column 3. This is shown by a solid dot in Figure 4. This initial filling stage lasted until 1965.

H_{max} in 1965 (655.16 m) was first equalled and exceeded on 30 August 1967. It was followed by a large event (event 1, Table 1) causing some destruction on 13 September 1967. (The magnitude of this event has been debated and lies between 4.5 and 5.2 (refs 16, 19).) The highest water level attained was on 4 October 1967 and the largest earthquake associated with a reservoir (event 2) occurred on 11 December 1967 (LT). It was followed by two aftershocks with $M \geq 5.0$ (events 3 and 4). The last aftershock (with $M > 5.0$) occurred on 28 October 1968. The 1967 mainshock, foreshock and aftershocks constitute episode I. The pattern of activity is similar to RIS observed at other reservoirs following initial impoundment.

H_{max} reached in 1967 (656.99 m) was next exceeded on 2 September 1973 and new H_{max} (657.98 m) was reached on 27 September 1973. It was followed by a M 5.1 event (no. 5) on 17 October 1973. No other $M \geq 5.0$ occurred between event 4 and this event. This constitutes episode II and was located near the reservoir[19,20].

The 1973 H_{max} was next exceeded on 1 September 1980 (Figure 4). It was followed by a $M \sim 5.0$ event on 2 September 1980 (event 6), located ~ 25 km away. H_{max} (658.13) was reached on 3 September 1980 and two other

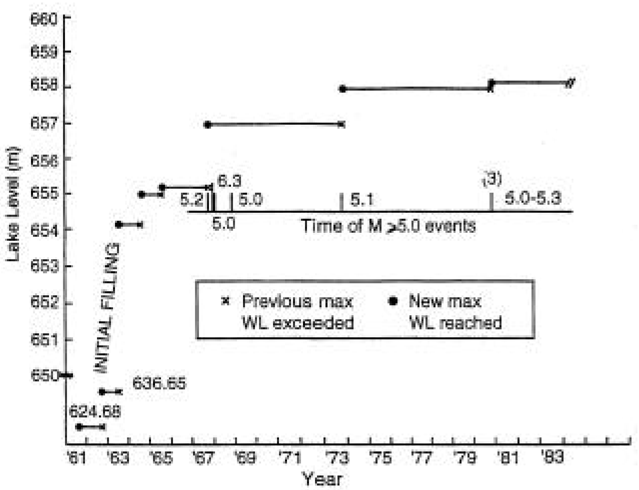

Figure 4. Maximum water levels and times of $M \geq 5.0$ events (see the text for details).

events with $M \sim 5.0$ occurred on 20 September 1980. Events 6 to 8 constituting episode III occurred near Warna River ~ 30 km from Koyna.

These three episodes are the *only* incidences of earthquakes with $M \geq 5.0$ until 1985 and they all occurred within about 2 months after the water level in Koyna Reservoir had exceeded the previous maximum. (The next $M \geq 5.0$ in the area did not occur until after the impoundment of Warna Reservoir began in 1985.)

The increases in water levels over the previous maximum ~ 2 m, 1 m, 0.1 m for the three episodes correspond to extra loads of 0.02, 0.01 and 0.001 MPa, respectively, and yet the rocks responded with $M \geq 5.0$ events – apparently displaying stress memory. This behaviour – the rocks displaying stress memory was also observed at Nurek Reservoir[21], and is very similar to Kaiser Effect observed in the laboratory[22,23].

These observations could underscore an important phenomenon at Koyna and other reservoirs – the rocks are highly stressed and close to failure, they may have stress memory and very small changes in strength may trigger large earthquakes.

Seismotectonics of the Koyna–Warna area

Since the inception of seismicity, several studies described the seismicity in the Koyna region. A comprehensive review of the literature through about 1990 (ref. 2), presented an epicentral map based on routine analysis up to that time (figure 4.32 *b* in that study). Unfortunately the epicentres were based on erroneous station locations and poorly constrained seismic parameters. Those locations and reported hypocentral depths ranging from near surface to over 60 km were inadequate to infer seismological features.

Talwani[16] relocated over 300 earthquakes (1963–1995) with $M \geq 3.0$ using revised location parameters (station locations, velocity model, station delays and V_p/V_s ratio). Integration of these revised locations with available geological, geophysical and geomorphological data has led to the identification of several seismogenic features in the Koyna–Warna area (Figure 5). These include the 40–50 km long N 10–15°E trending and dipping steeply to the west, Koyna River Fault Zone (KRFZ). This zone was interpreted from a combination of aeromagnetic, first motion and seismicity data. Open fissures that extended for ~ 15 km were associated with the 1967 M 6.3 Koyna earthquake that occurred on this fault. Well located hypocentres showed that the larger earthquakes ($M \geq 3.0$) that occurred on the KRFZ occurred at depths between ~ 6 and 13 km in a region interpreted to contain fluid-filled fractures[16]. KRFZ was also identified as the western boundary of the seismicity.

Seismicity was bounded to the east by the NW dipping and NE-SW trending Patan fault. The location of the

Table 1. Lake levels and episodes of $M \geq 5.0$ events at Koyna

Previous H_{max} exceeded on M D Y	New H_{max} reached on M D Y	New H_{max} (m)	Episode	Number	Event Date M D Y	Magnitude	Remarks
	07 11 61	624.68					Initial filling
07 11 62	08 15 62	636.65					
07 09 63	08 14 63	654.10					
08 01 64	08 13 64	654.94					
07 22 65	07 22 65	655.16					
08 30 67	10 04 67	656.99		1	09 13 67	5.2; 4.5	Foreshock
				2	12 11 67	6.3	Mainshock
			I	3	12 24 67	5.0; 5.2	Afterschock
				4	10 28 68	5.0	Afterschock
09 02 73	09 27 73	657.98	II	5	10 17 73	5.1	
09 01 80	09 03 80	658.13		6	09 02 80	4.9; 5.5	m_o, M_s
			III	7	09 20 80	4.9	m_o
				8	09 20 80	5.3	m_o

Patan fault which extended from an anomalous trend in the Koyna River to Bhogiv on the Warna River was based on the geomorphic features and mapped fractures. At least three NW-SE trending features joining these two features were also identified (Figure 5).

Based on offsets of retaining walls, about one km of the KRFZ has been identified as the Donichiwada fault[24]. The Donichiwada fault is located where KRFZ crosses the Koyna River. Monitoring of microseismicity with improved instrumentation led to the detection of ~ > 9000 events ($M \geq 0.0$) between 1993 and 1995 (refs 25, 26) and the identification of two NNE-SSW fault zones about 5 to 10 km apart[26], named the Koyna and Bhogiv faults. Comparison of their location shows that the Koyna and Bhogiv faults are the same as the previously described KRFZ and Patan fault (Figure 5). Results of recent shallow drilling in KRFZ near the Koyna River and helium measurements[27] confirmed the existence of the fault and interpreted steep westerly dip. The authors, however, chose to refer to it as the Donichiwada fault zone. In order to avoid further confusion, I suggest that KRFZ and Patan fault be used to describe the two NNE to NE-SSW to SW trending fault zones. A recent study using digital recording confirmed earlier depth estimates[28].

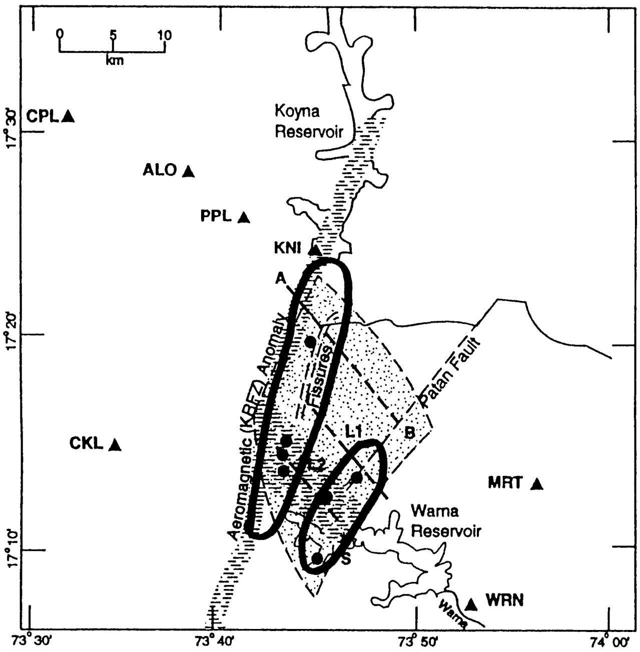

Figure 5. Interpreted seismotectonic features in the Koyna–Warna area. The Koyna River Fault Zone (KRFZ) is based on the pattern of aeromagnetic anomalies, seismicity and fissures associated with the 1967 earthquake. The NW-SE pattern of aeromagnetic anomalies includes a portion of the Warna River and lies along lineament L_2. Other NW-SE block boundaries are indicated by lineaments L_1 and L_2 and line AB. The better-located seismicity (1993–1995) is roughly located within a seismogenic zone (stippled pattern) enclosed by the KRFZ to the west and Patan fault to the east (and a parallel fault to its south-east). The larger events in 1993–1995 are shown by solid dots. The zone of fissures associated with the 1967 earthquake lies to the east of KRFZ (from Talwani[16]). Recent well-located seismicity from Rastogi *et al.*[26] and identified by them as Koyna and Bhogiv faults, is enclosed in solid ellipses. Three-letter codes next to triangles show location of seismic stations.

Future studies

Based on the preceding sections, we can summarize our current view of the seismicity in the Koyna–Warna area. The protracted seismicity is primarily associated with the annual filling cycle of the reservoirs (June–August). Pore pressure changes occur in the vicinity of the reservoir in response to lake level changes. Available fractures facilitate pore pressure diffusion away from the reservoir. The most efficient conduit for pore pressure diffusion appears to be the fluid filled ~ 5 km wide, ~ 6 to 7 km deep, (primarily between depths of ~ 6 to 13 km) KRFZ. The absence of significant seismicity ($M \geq 3.0$) shallower than ~ 6 km depth along the KRFZ[16] can possibly be explained as follows. The earlier episodes of larger events (1967,

1973) successfully released the locked-up stresses at shallow depths and consequently significant seismicity occurred below a ~ 6 km depth. A similar lack of larger events at shallow depths after the initial burst of seismicity was observed at Lake Jocassee[29]. Alternatively, the lack of shallow significant seismicity could be because the earlier episodes of large events have enhanced the permeability to a value greater than k_s, the seismogenic permeability. Permeability enhancement following seismicity has been observed, for example, at Loma Prieta[30]. The seismogenic part of KRFZ is associated with seismogenic permeability and in the vicinity is critically stressed such that small perturbations in fluid pressures trigger earthquake activity.

The seismicity should be monitored further. In order to better understand the physics of the protracted seismicity, the water levels in the observation wells and reservoirs should be further analysed. At Bad Creek, South Carolina only one fracture zone was monitored. At KRFZ, pore pressures should be monitored at one or more depths depending on the presence of intersecting fractures. Cross-correlation of the water level time series at observation wells, especially those drilled on KRFZ[25], with the reservoir lake levels will help describe the hydraulic properties of KRFZ and allow for quantitative estimates of the fluid pressures involved. This can lead to a clearer understanding of the physics of protracted seismicity.

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