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A review of recent studies of triggered earthquakes by artificial water reservoirs with special emphasis on earthquakes in Koyna, India

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Abstract

Triggering of earthquakes by filling of artificial water reservoirs is known for over six decades. As of today, over 90 sites have been globally identified where earthquakes have been triggered by filling of water reservoirs. The question of earthquakes triggered by artificial reservoirs has been addressed and reviewed in a number of papers and books. In the present review, the book "Reservoir-Induced Earthquakes" by Gupta [Gupta, H.K., 1992. Reservoir-Induced Earthquakes. Elsevier, Amsterdam, 364 pp.], which contains all the necessary information on this topic till 1990, has been taken as the base. An effort has been made to add information on this important topic gathered over the last 10 years. Koyna, India continues to be the most significant site of artificial-water-reservoir-triggered earthquakes. During 1990s, two events exceeding M 5 and several smaller events occurred in the vicinity of Koyna, and recently impounded Warna Reservoir. Detailed studies have addressed the relocation of earthquakes, stress drop, nucleation, migration and other important aspects of these earthquakes. In a unique experiment, twenty-one 90- to 250-m deep borewells have been drilled in the seismically active Koyna–Warna region and the water levels are continuously monitored. Step-like coseismic changes of several centimeters have been observed in some wells associated with a few $M \geq 4$ events. Detailed tomographic studies conducted on one of the best recorded triggered earthquake sequence at Lake Oroville in California revealed that this sequence was associated with a southwest dipping structure characterized by low velocity, while in adjacent areas, seismic activity occurs in regions of higher velocity. Similar investigations in Aswan showed that shallow activity is associated with low P-wave velocity. Several new reservoir sites that have triggered earthquakes have been reported during 1990s. The most important being Srinagarind Dam in Thailand, which had an M 5.9 earthquake and the sequence had all the characteristics of triggered earthquake sequences. New theoretical work, particularly the effect of pore fluid pressure in anisotropic rocks and its implication in triggered seismicity is an important development. However, much more needs to be done to fully comprehend the role of artificial water reservoirs in triggering earthquakes.

Keywords: triggered earthquakes; artificial water reservoirs; pore pressure; earthquake nucleation

1. Introduction

Huge artificial water reservoirs are created all over the world for generation of hydroelectric power, flood

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control and irrigation purposes. Triggering of earthquakes by artificial water reservoirs was for the first time pointed out by Carder (1945) at Lake Mead in the United States of America. Damaging earthquakes exceeding M 6 occurred at Hsinfengkiang, China in 1962; Kariba, Zambia–Zimbabwe Border in 1963; Kremasta, Greece in 1966; and Koyna, India in 1967. The December 10, 1967, Koyna earthquake of M 6.3 is so far the largest and most damaging reservoir-triggered earthquake. It claimed about 200 human lives, injured about 1500 and rendered thousands homeless. Civil works at Koyna town suffered major damage. The Hsinfengkiang and Koyna earthquakes caused damage to dams themselves. Other reservoir-triggered earthquakes such as those at Kariba, Zambia; Kremasta, Greece; Oroville, CA; Aswan, Egypt; and Srinagarind, Thailand had caused damage in nearby towns and villages. The occurrence and potential of triggered earthquakes has caused major modifications of civil works and engineering projects. The Hsinfengkiang Dam (Shen et al., 1974) in the People's Republic of China was strengthened twice before the occurrence of the March 20, 1962 induced earthquake of magnitude 6.1. The disposal of waste fluid through injection into the ground at Rocky Mountain Arsenal had to be discontinued because of triggered earthquakes (Evans, 1966). The possibility of triggered seismicity was one of the major factors in terminating the construction of Auburn Dam in California (Allen, 1978). The design parameters of the proposed 260-m-tall Tehri Dam in the foothills of Himalaya, India continue to be questioned and reexamined to ensure the dam's capability to resist earthquakes. Earlier, as pointed out by Allen (1982) and Simpson (1986), there is a general reluctance on the part of the Engineering Committee, globally, to accept the significance or even existence of the phenomenon of reservoir-triggered seismicity. However, there has been some change in this attitude. In a recent report of the US Committee on Large Dams (USCOLD, 1997), it is concluded that there can be an increase in occurrence of reservoir-triggered seismicity (RTS) during the period of reservoir water level changes and the possibility of RTS should be considered for every reservoir deeper than 80–100 m. Allen (1982) stated 18 years ago that "From a purely economic point of view, not to speak of public safety, the problem of reservoir-induced earthquakes deserves far more attention than it currently is receiving in most parts of the world". This observation still holds good.

Most of the work carried out on various aspects of artificial-water-reservoir-induced seismicity till 1990 has been reviewed by Gupta (1992). Since 1990, many workshops and symposia have been organized on this topic, and a wealth of papers have appeared in several volumes. (e.g. Gupta and Chadha 1995; Knoll and Kowalle, 1996; Talebi, 1997, 1998). Additionally, there are some good articles by Meade (1991), USCOLD (1997) and McGarr and Simpson (1997).

In the following, an effort has been made to review important work reported during the 1990s on the topic of artificial-water-reservoir-induced earthquakes. Koyna, on west coast of India, continues to be the most significant site of reservoir-triggered seismicity (RTS). Several new investigations and reports appeared during the 1990s. A triggered earthquake sequence has been reported from Thailand at Srinagarind reservoir with the main shock magnitude of M 5.9. Important work reported from Oroville, CA and Aswan, Egypt is commented upon. Seeber et al. (1996) have made a case for the Latur earthquake of 1993 in south India to be a triggered event. However, others do not consider it triggered (Rastogi, 1994).

2. Triggered vis-a-vis induced earthquakes

In a very interesting paper, McGarr and Simpson (1997) have deliberated on induced and triggered seismicity. They point out that the adjectives "triggered" and "induced" are very often used interchangeably whenever we talk of artificially stimulated earthquakes, and it would be advantageous to draw a distinction between the two. They propose that the term "triggered seismicity" shall be used only when a small fraction of stress changes or energy associated with earthquakes is accounted for by the causative activity, whereas "induced seismicity" shall be used where the causative activity is responsible for most of the stress changes or most of the energy required to produce the earthquakes. Thus, one would see that in case of triggered seismicity, tectonic loading plays a primary role. Under this classification, the earthquakes occurring in the vicinity of artificial water reservoirs, as a consequence of impoundment, would fall in the category of "triggered seismicity" since the stress level changes associated with the filling of some of the deepest reservoirs are of the order of 1 MPa or so, whereas the stress drops associated with earthquakes are much larger. Therefore, it is appropriate that all cases so far referred to as "reservoir-induced seismicity" (RIS) should now be classified, following the above criteria, as "reservoir-triggered seismicity" (RTS). We accept and follow the above-mentioned classification in this review.

3. Worldwide distribution

As of now, there are 95 sites globally where RTS has been reported (Table 1). These can be grouped in the following categories:

(i) Sites of $M \geq 6$ (4 cases);
(ii) Sites of M 5–5.9 (10 cases);
(iii) Sites of M 4–4.9 (28 cases);
(iv) Sites of $M < 4$ (53 cases).

In addition to the above list, decrease in microseismic activity has been reported as a consequence of impounding at 8 reservoir sites, and there are another 16 sites of suspected cases of triggered earthquakes (Table 1, Fig. 1). Reservoir sites where triggered earthquakes of magnitude less than 4 occurred are too numerous, and hence all of them are not included in Fig. 1.

Efforts to understand the correlation of reservoir-triggered earthquakes with reservoir characteristics and the prevalent geological conditions have continued. In some instances, the temporal correlations with reservoir level changes are very persuasive and dramatic (Gupta, 1983); whereas in many others, the correlation may be less distinct. It has also been noted that several large reservoirs have not triggered any earthquakes whereas many small ones have led to notable seismic activity. While geological and tectonic factors may be important in deciding the potential of a site, reservoir characteristics are equally significant. The depth of the water column and the reservoir volume are two important factors that control the triggered earthquakes (Baecher and Keeney, 1982). A review of the global examples gives further credence to the observation that the depth of the water column plays the most important role in triggering of earthquakes.

4. Koyna, India

Globally, Koyna continues to be the most significant site of artificial-water-reservoir-triggered seismicity. Earthquakes began to occur soon after the impoundment of the Shivaji Sagar Lake created by Koyna Dam in Western India in 1962. Over the past 38 years, over 10 earthquakes of $M \geq 5$; over 150 earthquakes of $M \geq 4$ and over 100,000 earthquakes of $M \geq 0$ have occurred. The site also has the distinction of having so far the largest and most damaging reservoir-triggered earthquake of M 6.3 on December 10, 1967 (Gupta and Rastogi, 1976; Gupta et al., 1997).

Major bursts of seismic activity associated with earthquakes exceeding M 5 in Koyna region occurred during 1967, 1973, 1980 and 1993–1994. Correspondence between the earthquake occurrence and factors like rate of increase of water level in the reservoir, maximum water level reached and the duration for which high water levels are retained has been underlined by Gupta et al. (1972a,b). It has been also pointed out that a rate of loading exceeding 12 m/week is necessary, but not a sufficient condition for $M \geq 5$ events to occur in Koyna Region (Gupta, 1983).

Till 1992, the earthquakes in the vicinity of the Koyna dam were mostly confined to a 20-km-long seismic zone extending south of the Koyna Dam. A southward shift in seismicity was observed during 1993–1994, which is attributed to the filling of the Warna Reservoir (Rastogi et al., 1995) located SSE of Koyna at a distance of about 35 km (Fig. 2). The Warna reservoir was filled to a depth of 60 m following the monsoon rains of 1993. The 1993–1994 spurt of seismic activity consisted of two events of $M \geq 5$: several events of $M \geq 4$ and several thousand smaller events (Rastogi et al., 1997b). During 1993, seismic monitoring of Koyna–Warna earthquakes was improved through the operation of a number of new seismic stations (Fig. 2, Chadha et al., 1997) and two trends of clusters were identified (Fig. 3). An experiment was mounted deploying a state-of-the-art digital seismic network of 10 stations for 20 months period during April 1966 through December 1997 (Rai et al., 1999). A total of 20 sites over an area of 60×80 km^2 were occupied in a leap frog manner with at least 50% overlap of seismic stations. They have located 406 events during this period confined to a zone of

Table 1
Reported cases of reservoir-associated changes in seismicity

Name of the dam/reservoir	Country	Height of dam (m)	Reservoir volume (10^6 m^3)	Year of impounding	Year of largest earthquake	Magnitude/intensity	References[a]
Sites where earthquakes having magnitude ≥ 6.0 were triggered							
Hsinfengkiang	China (PRC)	105	13,896	1959	1962	6.1	1, 2, 3
Kariba	Zambia–Zimbabwe	128	175,000	1958	1963	6.2	1, 2, 6
Koyna	India	103	2780	1962	1967	6.3	1, 2, 4, 5
Kremasta	Greece	160	4750	1965	1966	6.2	1, 2, 4, 5
Sites where earthquakes having magnitude between 5.0 and 5.9 were triggered							
Aswan	Egypt	111	164,000	1964	1981	5.6	2, 7
Benmore	New Zealand	110	2040	1964	1966	5.0	1, 2, 8
Charvak	Uzbekistan	148	2000	1971	1977	5.3	55
Eucumbene	Australia	116	4761	1957	1959	5.0	2
Geheyan	China	151	3400	1993	1997	VI	48a
Hoover	USA	221	36,703	1935	1939	5.0	1, 2, 10
Marathon	Greece	67	41	1929	1938	5.7	1, 2, 4, 5
Oroville	USA	236	4400	1967	1975	5.7	2, 11
Srinagarind	Thailand	140	11,750	1977	1983	5.9	65
Warna	India	80	1260	1987	1993	5.0	51
Sites where earthquakes having magnitude between 4.0 and 4.9 were triggered							
Aksombo Main	Ghana	134	148,000	1964	1964	V	2, 9
Bajina Basta	Yugoslavia	90	340	1966	1967	4.5–5.0	2, 5
Bhatsa	India	88	947	1981	1983	4.9	39
Bratsk	Russia	100	169		1996	4.2	48b
Camarillas	Spain	49	37	1960	1964	4.1	2, 4, 5
Canelles	Spain	150	678	1960	1962	4.7	2, 4, 5
Capivari–Cachoeira	Brazil	58	180	1970	1971	VI	31
Clark Hill	USA	60	3517	1952	1974	4.3	2, 12
Dahua	China (PRC)	74.5	420	1982	1993	4.5	56
Danjiangkou	China (PRC)	97	16,000	1967	1973	4.7	29
Foziling	China (PRC)	74	470	1954	1973	4.5	29
Grandval	France	88	292	1959	1963	V	1, 2, 4, 5
Hoa Binh	Vietnam	125		1988	1989	4.9	48c
Kastraki	Greece	96	1000	1968	1969	4.6	2
Kerr	USA	60	1505	1958	1971	4.9	1, 2, 9
Komani	Albania	130	1600	1985	1986	4.2	57
Kurobe	Japan	186	149	1960	1961	4.9	2, 13
Lake Baikal	Russia					4–4.8	49a
Lake Pukaki	New Zealand	106	9000	1976	1978	4.6	41
Manicouagan 3	Canada	108	10,423	1975	1975	4.1	2
Marimbondo	Brazil	94	6150	1975	1975	IV	32
Monteynard	France	155	275	1962	1963	4.9	1, 2, 4, 5
Nurek	Tadjikistan	317	1000	1972	1972	4.6	1, 2, 14
P. Colombia/V. Grande	Brazil	40/56	1500/2300	1973–1974	1974	4.2	33
Piastra	Italy	93	13	1965	1966	4.4	2, 4, 5
Pieve de Cadore	Italy	116	69	1949	1950	V	2, 15
Shenwo	China (PRC)	50	540	1972	1974	4.8	29
Vouglans	France	130	605	1968	1971	4.4	2, 4, 5
Sites where earthquakes having magnitude < 4.0 were triggered							
Acu	Brazil	31	2400	1983	1994	2.8	59
Blowering	Australia	112	1628	1968	1973	3.5	2

Table 1 (*continued*)

Name of the dam/reservoir	Country	Height of dam (m)	Reservoir volume (10^6 m^3)	Year of impounding	Year of largest earthquake	Magnitude/ intensity	References[a]
Sites where earthquakes having magnitude < 4.0 were triggered							
Capivara	Brazil	59	10,500	1976	1976	3.7	31
Carmo do Cajuru	Brazil	22	192	1954	1972	3.7	32
Contra	Switzerland	220	86	1963	1965	3.0	2, 4, 5
Dhamni	India	59	285	1983	1994	3.8	50
Donjiang	China (PRC)	157	81	1986	1990	3.2	48d
Emborcacao	Brazil	158	17,600	1981	1984	~ 2.0	36
Emmosson	Switzerland	180	225	1973	1973	3.0	2, 42
Fierza	Albania	167	2800	1978	1981	2.6	60
Gandipet	India	36	117	1920	1982	3.5	16
Grancarevo	Yugoslavia	123	1280	1967	1967	3.0	2, 4, 5
Hendrik Verwoerd	South Africa	66	5000	1970	1971	2.0	1, 2, 17
Huangshi	China (PRC)	40	610	1970	1974	2.3	29
Hunanzhen	China (PRC)	129	2060	1979	1979	2.8	61
Idukki	India	169	1996	1975	1977	3.5	2, 18
Itezhitezhi	Zambia	65	5000	1976	1978	3.8	2
Jocasse	USA	107	1431	1971	1975	3.2	2, 19
Kamafusa	Japan	47	45	1970	1970	3.0	1, 2, 20
Katse	Lesotho	185	1950	1995	1996	3.1	52
Keban	Turkey	212	31,000	1973	1973	3.5	2, 9
Kouris	Cyprus				1994–1995	3.0	49b
Kurupsai	USSR	~ 100	500	1981	1983	Microearthquakes	30
Lake Gordon–Lake Peddar	Australia	140	13,500	1974	1978	Microearthquakes	44
LG 3	Canada	80		1981	1983	3.7	40
Lubuge	China (PRC)	103	110	1988	1988	3.4	62
Makio	Japan	105	75	1961	1978	Earthquake swarm	38[b]
Monticello	USA	129	500	1977	1979	2.8	12, 19, 21
Mula	India	56	1017	1972	1972	1.0	1, 2, 18
Nagawado	Japan	155	123	1969	1969	Earthquake swarm	38[c]
Nanchong	China (PRC)	45	15	1969	1974	2.8	29
Nanshui	China (PRC)	81	1220	1969	1970	2.3	63
Novo Ponte	Brazil		128		1995	3.7	49c
Oued Fodda	Algeria	101	225	1932	1933	3.0	2, 4
Paraibuna–Paraitinga	Brazil	94/105	4700	1975–1976	1977	3.0	34
Qianjin	China (PRC)	50	20	1970	1971	3.0	29
Ridracoli	Italy	103	33	1981	1988	3.5	53
Salanfe	Switzerland				1995	2.5	49d
Schlegeis	Austria	117	128	1970	1971	2.0	2
Shasta	USA	183	5615	1944	1944	3.0	2, 9
Shengjiaxia	China (PRC)	35	4	1980	1984	3.6	58
Shuikou	China (PRC)	101	2350	1993	1994	3.2	58
Sobradinho	Brazil	43	34,100	1977	1979	~ 2.0	35
Sriramsagar	India	43	32,000	1983	1984	3.2	37
Talbingo	Australia	162	935	1971	1973	3.5	2, 14, 22
Thomson	Australia			1983	1990	3.0	54
Toktogul	Kirghizia	215	19,500	1977		2.5	2, 23
Tongjiezi	China (PRC)	74	30	1992	1992	2.9	64
Tucurui	Brazil	100	45,800	1984	1985	3.4	36
Vajont	Italy	262	150	1960	1960		1, 2, 4, 15
Zhelin	China (PRC)	62	7170	1972	1972	3.2	29

(continued on next page)

Table 1 (*continued*)

Name of the dam/reservoir	Country	Height of dam (m)	Reservoir volume (10^6 m^3)	Year of impounding	Year of largest earthquake	Magnitude/ intensity	References[a]
Sites where earthquakes having magnitude < 4.0 were triggered							
Wujiangdu	China (PRC)	165	2140	1979	1985	2.8	58
Yantan	China (PRC)	110	2430	1992	1994	3.5	56
Decrease in microearthquake activity							
Anderson	USA	72	110	1950			24
Bhakranangal	India	226	9868	1958			2, 25
Flaming Gorge	USA	153	4674	1962			2, 26
Glen Canyon	USA	216	33,305	1963			2, 26
Ikawa	Japan	104	151	1957			38
Mangla	Pakistan	116	7250	1967			2, 27
Tarbela	Pakistan	143	13,960	1974			2, 28, 66
Tsengwen Reservoir	Taiwan	128	708	1973			45
Other possible cases							
Cabin Creek	USA	49					2
Clark Canyon	USA	40					9
Coyote Valley	USA	50					2
Disposal Wells, northeastern Ohio	USA				1986	5.0	43
El Grado	Spain	130					2, 9
Ghirni	India	16					2, 9
Kinnersani	India	61					2, 9
Palisades	USA	82					2, 9
Parambikkulam	India	73					2, 9
Rockey Reach	USA						9
San Luis	USA	116					9
Sefid Rud	Iran	106					2
Sleepy Hollow Oil field	Canada						46, 47
Ukai	India	69					2, 9
Warragamba	Australia	137					2, 9

[a] References: 1 = Gupta and Rastogi (1976); 2 = Packer et al. (1979); 3 = Shen et al. (1974); 4 = Rothe, (1970, 1973); 5 = Bozovic (1974); 6 = Gough and Gough (1970b); 7 = Toppozada (1982); 8 = Adams (1974); 9 = Simpson (1976); 10 = Carder (1945); 11 = Bufe et al. (1976); 12 = Talwani (1976); 13 = Hagiwara and Ohtake (1972); 14 = Soboleva and Mamadaliev (1976); 15 = Caloi (1970); 16 = Gupta et al. (1982); 17 = Green (1973); 18 = Guha (1982); 19 = Talwani et al. (1980); 20 = Suzuki (1975); 21 = Zoback and Hickman (1982); 22 = Muirhead et al. (1973); 23 = Simpson et al. (1981); 24 = Bufe (1975); 25 = Chaudhury and Srivastava (1978); 26 = Mickey (1973); 27 = Gupta (1984); 28 = Jacob et al. (1979); 29 = Oike and Ishikawa (1983); 30 = Simpson and Leith (1988); 31 = Berrocal (personal communication, 1990); 32 = Veloso et al. (1987); 33 = Berrocal et al. (1984); 34 = Ribotta (1989); 35 = Berrocal et al. (submitted for publication); 36 = Veloso and Assumpcao (1986); 37 = Rastogi et al. (1986b); 38 = Ohtake (1986); 39 = Rastogi et al. (1986a); 40 = Anglin and Buchbinder (1985); 41 = Reyners (1988); 42 = Bock and Rosa-Mayer (1980/1981); 43 = Nicholson et al. (1988); 44 = Shirley (1980); 45 = Wu et al. (1979); 46 = Rothe and Lui (1983); 47 = Evans and Steeples (1987); 48a = Chen et al. (1996), 48b = Pavlenov and Sherman (1996), 48c = Tung (1996), 48d = Yang et al. (1996); 49a = Djadkov (1997), 49b = Constantinou et al. (1997), 49c = Sellami et al. (1997), 49d = Veloso et al. (1997); 50 = Rastogi et al. (1997a); 51 = Rastogi et al. (1997b); 52 = Bell and Haskins (1997); 53 = Piccinelli et al. (1995); 54 = IASPEI General Assembly (1994); 55 = Plotnikova et al. (1992); 56 = Guang (1995); 57 = Muco (1991b); 58 = Hu et al. (1996); 59 = Joaquim et al. (1995); 60 = Muco (1991a); 61 = Hu et al. (1986); 62 = Jiang and Wei (1995); 63 = Xiao and Pan (1984); 64 = Guo (1994); 65 = Chung and Liu (1992); 66 = Kalpna and Chander (1997).

[b] Ohtake has reported a swarm, including a magnitude 5.3 earthquake in October 1978. In absence of a detailed report, this case is placed in the category of RIS < 4.0.

[c] Ohtake has reported a swarm, including a magnitude 5 in August 1969. In absence of a detailed report, this case is placed in the category of RIS < 4.0.

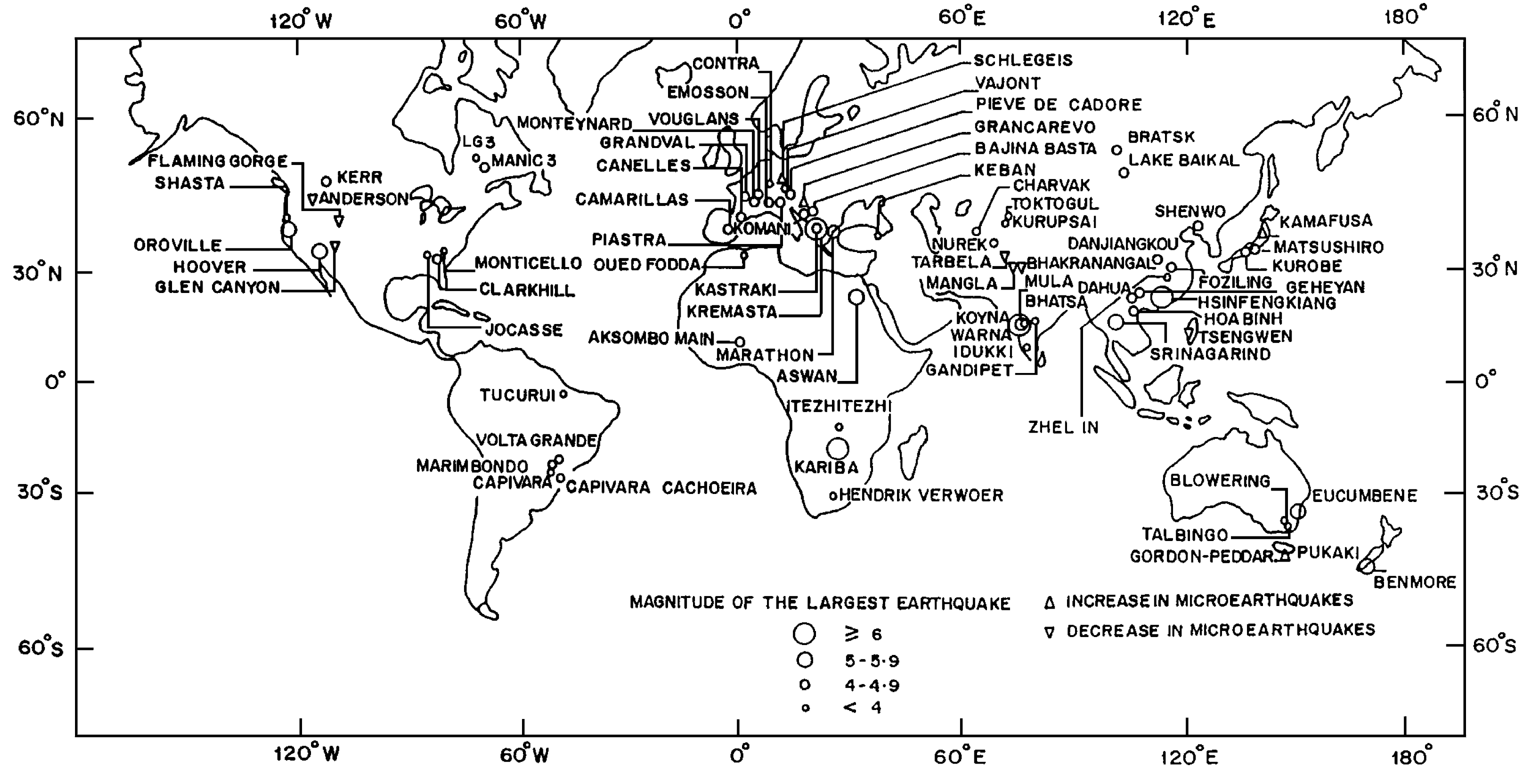

Fig. 1. Worldwide distribution of reservoir-triggered changes in the seismicity. The figure is updated from Gupta (1992) to include all sites where triggered earthquakes of $M>4$ occurred. Sites where RTS events of $M<4$ occurred are too numerous and not updated. For details, please see Table 1.

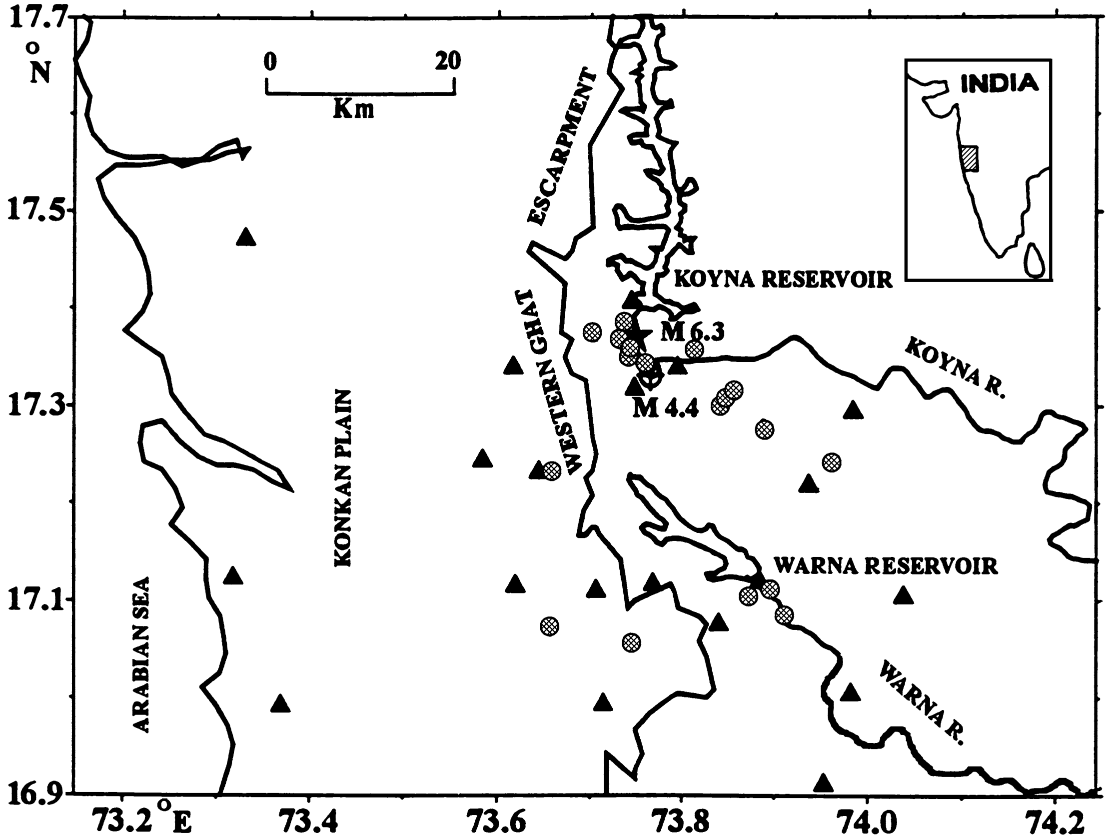

Fig. 2. The location of Koyna–Warna reservoirs in west India. Triangles indicate the sites where seismic stations are operated. Circles show the locations of borewells. The largest triggered earthquake of *M* 6.3 on December 10, 1967 is shown by a star. A circle and a plus sign show the *M* 4.4 earthquake of April 25, 1997, for which the residual pore pressure fluctuations are included in Fig. 10.

11×25 km^2. The hypocentral parameters were computed using a suite of velocity models and finally, a model with an eight-layer crust has been accepted. Ninety percent of events have an epicentral resolution of better than 0.2 km and depth resolution of better than 1.0 km. The depth distribution delimits a seismogenic zone with the base at 10 km. Rai et al. (1999) have identified three seismic zones as the north escarpment zone (NEZ), southern escarpment zone (SEZ) and the Warna seismic zone (WSZ).

4.1. Relocation of epicentres

Earthquakes in Koyna region have been relocated by Rajendran et al. (1996) and Talwani (1997a). Rajendran et al. (1996) relocated 207 earthquakes of magnitude >3 during the period 1983–1993. Out of these 207 relocated events, 125 are of B and C quality. There was a substantial improvement in epicentral location but hypocentral depths continue to be poor. An interesting observation has been made regarding spatial distribution of epicenters. They observed that soon after refilling, earthquakes occur in a widespread area (June through December) whereas during the later periods, (January through May) the area of earthquake activity shrinks. Fig. 4 gives the spatial pattern of relocated seismicity for the period 1983 to 1993 for events of magnitude larger than 3 during these two periods. It can be seen that following the refilling, earthquakes are much more widespread and concen-

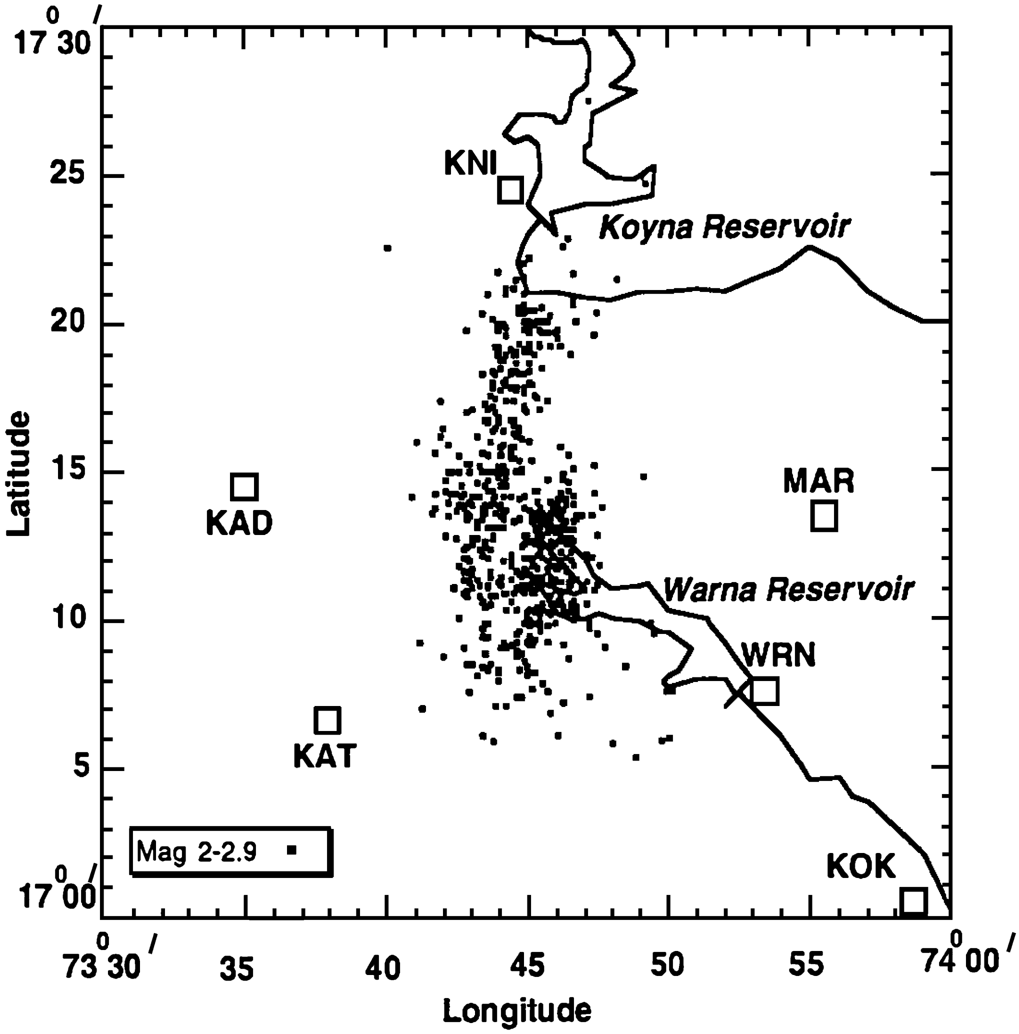

Fig. 3. Map showing epicenters of earthquakes in the magnitude range 2–2.9 during October 1993 to December 1994 located with NGRI seismic station network. Two NNE–SSW trends are clearly observed (figure from Chadha et al., 1997).

trated closer to the Warna reservoir (June to December). However, during the months of January to May, influenced by the draw down, the spatial extent of the earthquakes is much smaller and they concentrate near the Koyna Dam.

Talwani (1997a) has relocated over 300 earthquakes of $M \geq 3$ that occurred during the period 1963 through 1995 using revised location parameters such as velocity model, seismic station locations and delays and V_p/V_s ratio. The hypocentral data thus generated has been integrated with the available geological and geophysical data to delineate and identify the geometry of seismogenic structures. Talwani (1997a) has inferred several seismogenic crustal blocks underlain by a fluid filled fracture zone. The seismicity is bound in the west by the Koyna River Fault Zone (KRFZ), and in the east by NE–SW trending Patan Fault (Fig. 5). According to Talwani (1997a), the area between the KRFZ and the Patan Fault is intersected by a number of NW–SE trending fractures, which may be extending from near surface to the hypocentral depths and may provide conduits to fluid pressure flow to hypocentral depths.

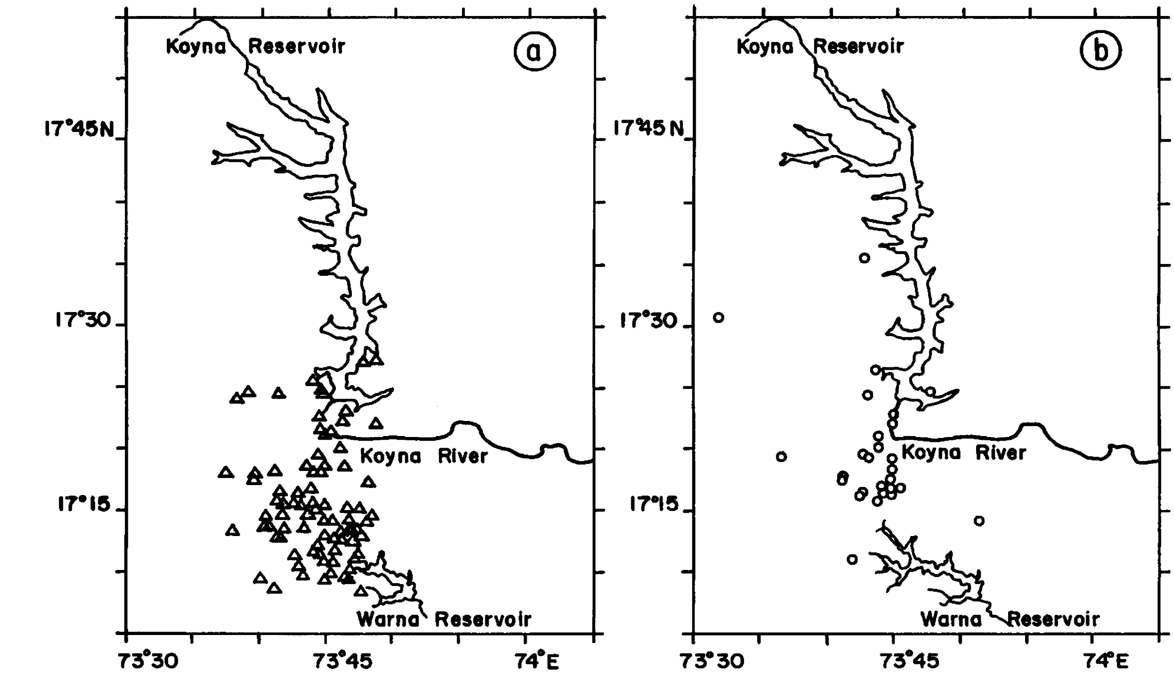

Fig. 4. The relocated earthquakes of $M > 13$ (B and C quality only) for the period 1983–1993 at Koyna. (a) For the period June through December, and (b) January through May (figure after Rajendran et al., 1996; for details, see the text).

4.2. Reservoir level fluctuations and earthquakes

There were significant changes in the water levels in the Koyna and Warna reservoirs during 1993 preceding the enhanced seismic activity in the vicinity of Koyna–Warna Reservoirs (Rastogi et al., 1997b). During July 1993, the water column in Warna reservoir had exceeded 60 m, and in August 1993, a spurt of seismicity was evinced. The maximum rate of loading at Koyna was 10 m/week while at Warna it was 16 m/week. As pointed out by Rastogi et al. (1997b), the total increase of water level at Koyna was 36.20 m during the period June 15, 1993 through September 5, 1993, while it was 44.15 m at Warna during the period June 11, 1993 through August 4, 1993 (Fig. 6). The peak water level at Warna, observed on August 4, 1993, was followed with a burst of seismic activity starting on August 18, 1993 and occurrence of two earthquakes of $M \geq 5$ and several smaller events. During 1994 and 1997, the minimum to maximum water level changes at the Warna Reservoirs were 41.58 and 40.82 m, respectively, compared to 44.15 m during 1993, and the seismic activity has been much less during the following 2 years.

As the maximum water level was attained at the Warna Reservoir 15 days before the enhanced seismicity of 1993, the current seismicity appears to be more influenced by the Warna Reservoir compared to Koyna Reservoir.

4.3. Stress drop of Koyna earthquakes

Mandal et al. (1998) have estimated the seismic moment and stress drop of 193 selected earthquakes of M varying from 1.5 to 4.7, which were recorded during the period October 1994 through June 1995. These estimates were made using the digital data acquisition systems deployed in the region. The seismic moments are estimated to vary from 10^{11} to 10^{16} Newton meters (N m) and the source radii was estimated from 94 to 538 m. The stress drop is estimated from 0.03 to 19 MPa for earthquakes in the magnitude range of 1.5 to 4.7. However, it may be noted that with the exception of two events, the stress drop of all other events is less than 3 MPa.

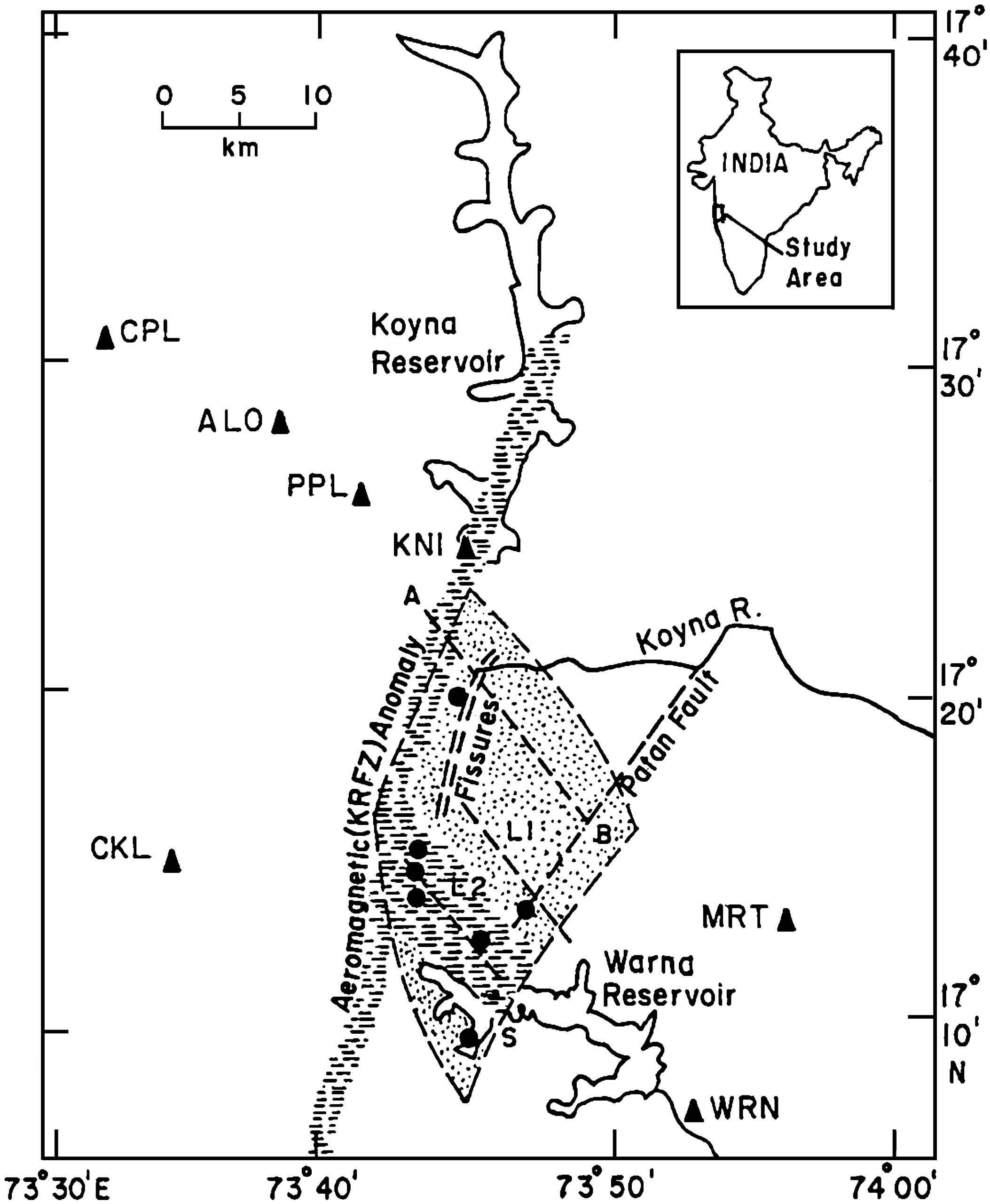

Fig. 5. The structural boundaries inferred from the aeromagnetic data in the vicinity of Koyna and Warna reservoirs are shown. The Koyna River Fault Zone is interpreted on the basis of aeromagnetic anomalies. The NW and SE pattern of aeromagnetic anomalies include portion of Warna river and lie along lineament L2. Other NW–SE block boundaries are indicated by L1, L2 and the line AB. The better located earthquakes for the period 1993–1995 are located within the seismic zone enclosed by the Koyna River Fault Zone in the west and Patan Fault in the east (stippled pattern). Solid dots indicate larger events in 1995. Fissures associated with December 10, 1967 earthquake are also shown (figure after Talwani, 1997a,b).

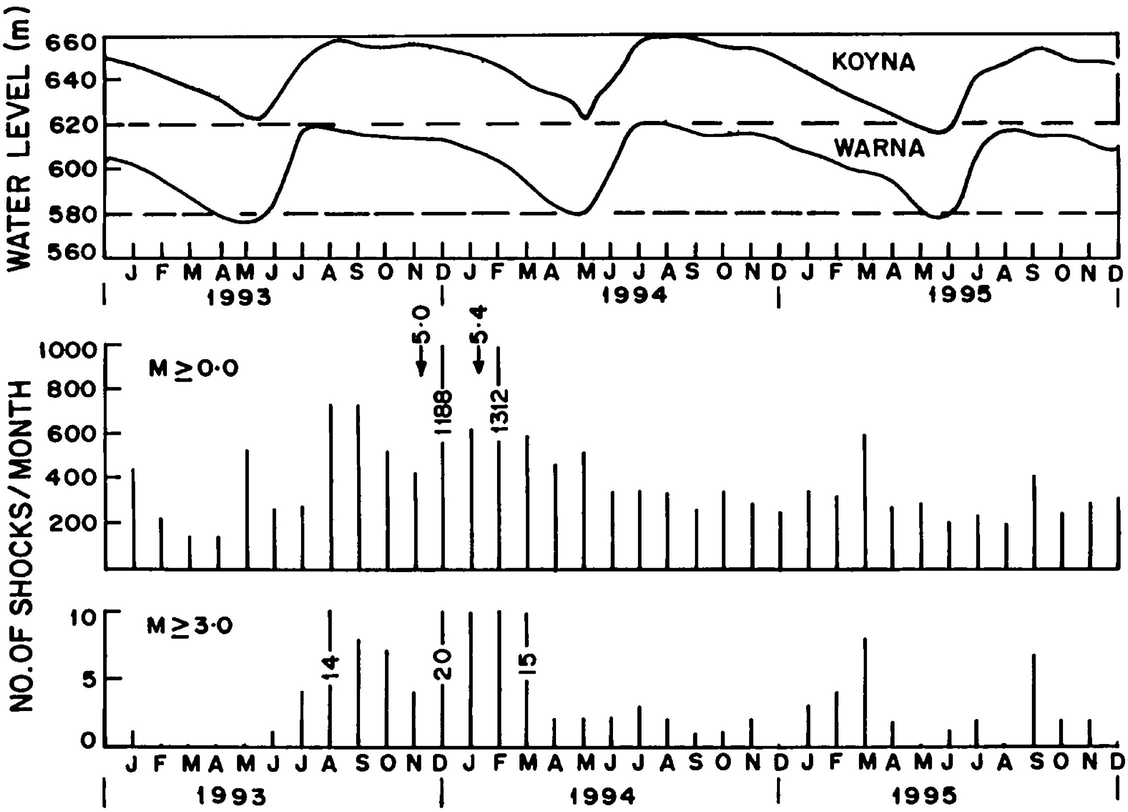

Fig. 6. Water levels in Koyna and Warna reservoirs and the seismic activity in the Koyna and Warna reservoirs for the period 1993 through 1995. The middle figure shows the monthly frequency of $M>0.0$ and the bottom one for $M>3.0$ (figure updated from Rastogi et al., 1997a,b; for details, see the text).

4.4. Low velocity layers and seismicity

An association of earthquakes in the Koyna region with low velocity zone has been noted (Talwani, 1997a). Mandal et al. (1998) reported that there is a concentration of earthquakes at shallow depths of less than 1 km. The seismicity is low in the depth range of 1 to 5 km and then again there is an increase in seismic activity between 5 and 10 km. In Fig. 7, the depth distribution of earthquakes (from Mandal et al., 1998) and the velocity depth section, reported by Krishna et al. (1989) from deep seismic sounding conducted along two profiles in the Koyna region, are shown. It is noteworthy that the concentration of earthquakes in 0–1 km and again in 5–10 km depth ranges occur in the low-velocity layers inferred from seismic profiling.

4.5. Anatomy of surface rupture zone

The epicentre of December 10, 1967 Koyna earthquake and the distribution of its aftershocks suggest that the earthquake was associated with a NNE–SSW striking Donechiwada fault designated by Harpster et al. (1979). Fault plane solutions have shown the fault to be of strike-slip nature and the surface observations indicate a left-lateral movement. Whether the fault dips towards WNW or ESE was not known. This problem has been addressed by Gupta et al. (1999). Encouraged by the results reported by Wakita et al. (1978) and the work in the Latur region (Reddy et al., 1994), soil-helium surveys were carried out along 12 traverses at a depth of 1.6 and 5 m interval (Fig. 8) to ascertain whether the surface fissures were indeed an expression

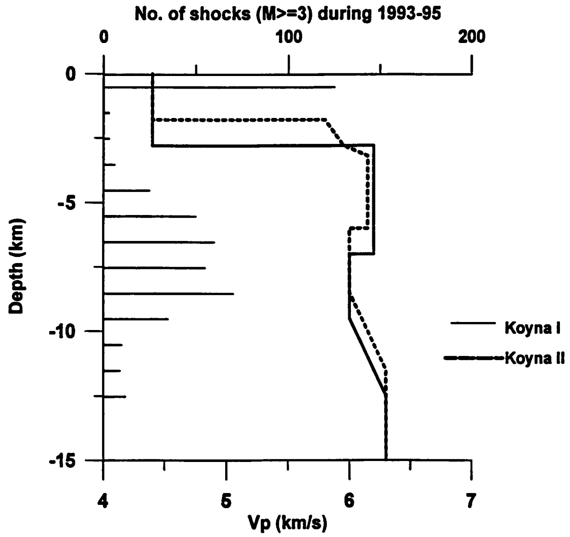

Fig. 7. Depth distribution of the shocks in the vicinity of Koyna Warna region of $M > 3$ for the period 1993–1995 and the P-wave velocity profile reported by Krishna et al. (1989). The number of events for every kilometer is plotted. For details, see the text.

of seismic fault. The peak values of the concentration of these traverses vary from 1 to 7 ppm above the atmospheric abundance of 5.24 ppm (Fig. 8). The surface fissures are indeed an expression of the seismogenic fault borne by the fact that the helium concentrations fall to regional background value of 0.2 ppm within 40 to 60 m on either side of the fissure zone (Gupta et al., 1999). On the basis of $^3He/^4He$ ratio, the source of the soil gas helium is inferred to be crustal. Core drilling at inclined holes was undertaken in the fissure zone to intersect the fault and ascertain the dip of the fault plane (Fig. 9). The fault was found to be dipping towards WNW at 60°. This work ascertained that the surface ruptures observed at the Donechiwada fault zone are indeed the surface manifestation of the causative fault of the 1967 Koyna earthquake.

4.6. Foreshocks and nucleation of Koyna events

In the recent years, there has been an effort of identifying and detecting nucleation processes that precede earthquakes. Several papers providing the experimental and theoretical background have appeared, e.g. Ohnaka (1992), Dodge and Beroza (1995) and Dodge et al. (1996).

With an improved location of hypocentres in the Koyna region, it has now become possible to investigate nucleation preceding main shock events. Rastogi and Mandal (1999) have reported nucleation preceding five Koyna main shocks of magnitude varying from 4.3 to 5.4 during 1993 to 1996. They investigated the space–time pattern of foreshocks within 8 km and 500 h prior to the main shock. Nucleation process is identified to occur in two phases, that is quasi-static and quasi-dynamic, before the dynamic rupturing of the main shock. The foreshock nucleation zone is estimated to grow at a rate of 0.5–10 cm/s till it reaches a diameter of 10 km before the occurrence of the main shock. The nucleation is initiated at shallow depths of less than 1 km and then gradually deepens to depths of 8–11 km in the seismogenic layer. The beginning of nucleation at shallow depths is attributed to the effect of increasing pore pressure caused by reservoirs of Koyna and Warna dams in subhydrostatic conditions. The propagation of fractures towards the seismogenic layer is inferred to be controlled by local stress concentrations along the fault zone and pore pressure diffusion to greater depths. These are some of the first studies anywhere in the world where nucleation process is so well demonstrated.

4.7. Pore pressure monitoring

Anomalous well level fluctuations associated with earthquakes have been reported for a number of sites. These could be pre-, co-, and post-seismic (Roeloffs, 1996, 1998; and others). These fluctuations may reflect pore pressure changes associated with redistribution of stress in the near and far field of dislocation sources. However, due to lack of appropriate data, these fluctuations are not well understood. Beginning in 1995, under a collaborative programme between the National Geophysical Research Institute, India and the University of Bonn, Germany, twenty-one 90 to 250 m deep borewells (Fig. 2) have been drilled in the vicinity of the seismically active Koyna–Warna region and well level sensors for a continuous monitoring have been installed (Chadha et al., 1997; Grecksch et al., 1999; Gupta et al., 1999). Most of the well data exhibit tidal signals implying that the wells are sensitive to small strain changes in the associated rock formations. This implies that well levels will also be sensitive to water level changes in the Koyna and Warna reservoirs

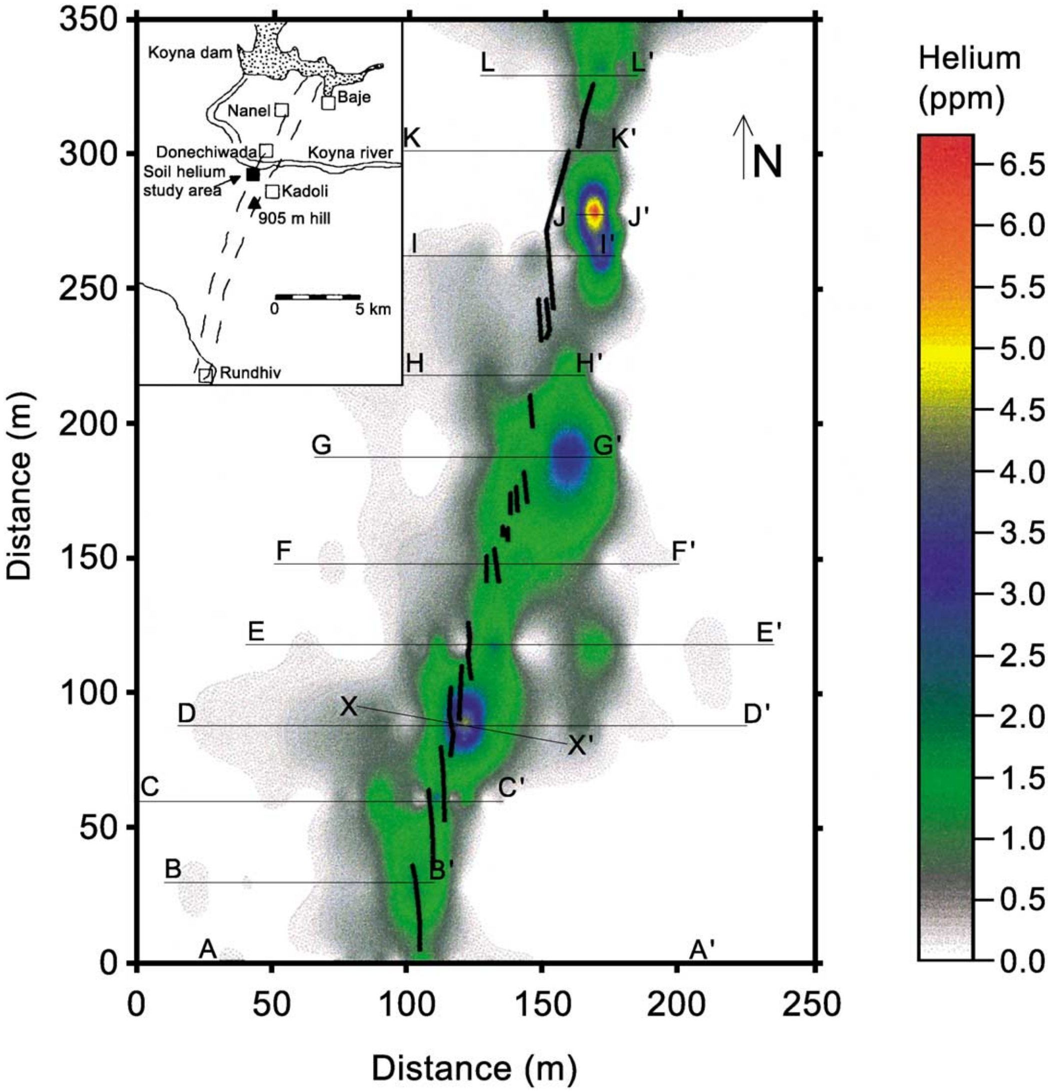

Fig. 8. Map of Donechiwada fault zone—line in the inset SE of the Koyna Dam shows the location of area where soil helium were conducted. Solid rectangles indicate coincidence of helium anomalies with coseismic and en-echelon fissures created by 1967 earthquake (thick lines) near Kodoli in Donechiwada fault zone. Anomalies were established by soil helium measurements along 12 traverses (AA′ to LL′) across the fissure system. The peak values of DD′, II′ and JJ′ are 4.5, 4.5 and 6.7 ppm respectively. XX′ is the traverse where two borewells were drilled (Fig. 9; figure updated from Gupta et al., 1999).

in the area. This is the first experiment of its kind anywhere in the world where wells have been drilled in a region of intense localized seismicity to investigate the role of pore pressure in triggering earthquakes.

Preliminary results show step-like changes of 2.5 and 6.0 cm amplitude at the Govare and Taloshi wells coincident with the *M* 4.4 earthquake of April 25, 1997 at epicentral distances of 1.8 and 2.4 km, respectively (Fig. 10). Similar features were also seen for another *M* 4.4 event of February 11, 1998 (Janssen, 1998). More data are being analyzed that will improve our understanding of the part played by pore-fluid pressures in

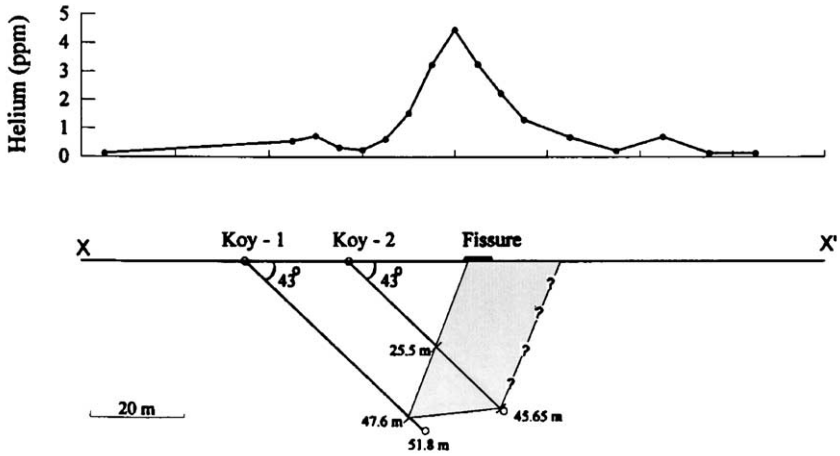

Fig. 9. Koyna seismic fault reconstructed by drilling. The two boreholes drilled are on a WNW–ESE traverse, perpendicular to the general strike of the fissure zone. The helium profile on DD′ (Fig. 8) is also shown. Solid circles show the bottom of the boreholes. The brecciated zone was met with in the subsurface between 47.6 and 51.8 m along KOY-1. In KOY-2, it was met with between 25.5 and 45.65 m along the hole. A 60° dip towards WNW for the fault is obtained by connecting the points of first intersection of the breccia zone in the two boreholes and its extrapolation to the surface fissure. The apparent large thickness of the fault zone is guided by the KOY-2 data (figure from Gupta et al., 1999).

triggering earthquakes. It may, however, be pointed out that step-like groundwater level changes in confined aquifers within a few source dimensions of an earthquake epicentre could be due to poroelastic response to the static strain field of the earthquake (Quilty and Roeloffs, 1997; Wakita 1975).

4.8. Koyna seismicity models

Occurrence of continued triggered earthquakes for the past 38 years at Koyna have led to some conceptual and physical models. We shall briefly mention here the models of Talwani (1995, 1997a,b) and that of Rajen-

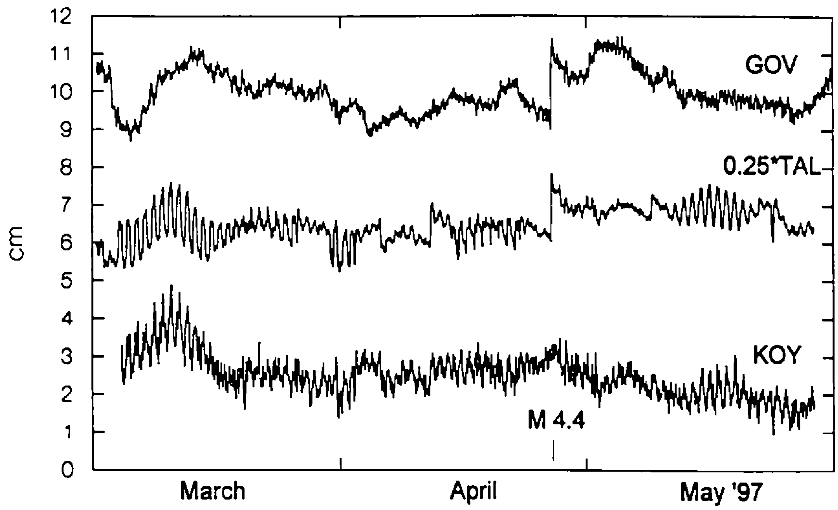

Fig. 10. Residual pore pressure fluctuations as observed in Govare (GOV), Taloshi (TAL) and Koyna (KOY) wells. Co-seismic anomaly due to the earthquake of M 4.4 on April 25, 1997 (epicentre shown in Fig. 2) is clearly observed at GOV and TAL, which are located 1.8 and 2.4 km from the epicenter of the earthquake. Tick marks on vertical axis are cm for GOV and KOY and 4 cm for TAL (after Gupta et al., 2000).

dran et al. (1996) and Rajendran and Harish (2000). Talwani (1995) has proposed a heuristic model based on various physical parameters observed in Koyna region. His model consists of two essential elements: intersecting faults near Koyna, which provide a means for stress build up in response to plate tectonics, and the annual reservoir loading cycle and corresponding changes in ground water table, which perturbs the stress build up by an influx of pore pressure in a fluid filled media. The spatial and temporal distribution of pore pressure and earthquakes are governed by hydromechanical properties of faults and fractures.

Rajendran et al. (1996) suggest that the mechanism of Koyna earthquakes is controlled by stability changes within a continuously degenerating fault zone. The annual loading of the reservoir continues to weaken the fault through enhanced fluid pressure and thereby failure may occur in response to even smaller changes in stress. Reservoir water level changes of the order of 1 to 1.5 m are known to have affected spatial pattern of seismicity at Lake Jocassee in the United States as well as at Nurek dam in Tadjikistan. Rajendran and Harish (2000) have presented a model that incorporates continued failure at Koyna in response to increased fluid pressure within an isolated fault zone (Fig. 11). The fault is near vertical and extends to a depth of about 10 km. This assumption is supported by

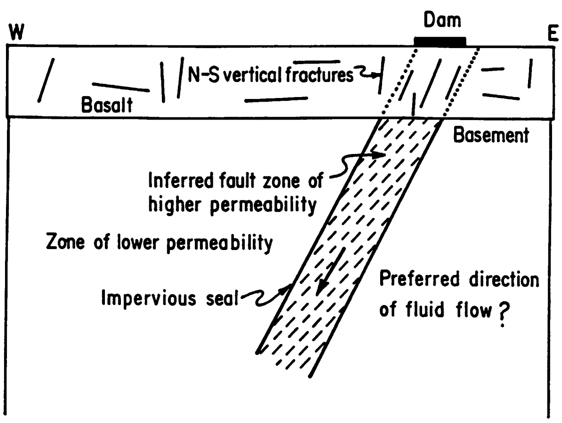

Fig. 11. A schematic diagram illustrating the geological and hydrogeological conditions in the Koyna region. The top layer of basalt is assumed to permit flow only through the N–S vertical fractures. Enhanced fluid flow occurs within the fault zone; lateral flow is restricted due to the lower permeability of the country rock and occurrence of an impervious seal between the fault and the country rock (figure after Rajendran and Harish, 2000).

NNE striking faults inferred from a variety of data as well as focal depths of earthquakes. Rajendran and Harish (2000) assume that the north–south fractures are more conductive and help recharge the fault zone. The permeability along the fault zone is presumed to be much higher than the lateral permeability into the country rock as suggested by Snow (1982). An important feature of this model is the existence of a low permeability seal that separates the fault zone and the country rock. With these assumptions, the seismogenic fault at Koyna could be considered to be constantly under hydrostatic pressure. The disposition of western ghat escarpment, which is parallel to Koyna river, could further help generation of large regional hydrostatic potential, which will generally complement the weakening process. If the fault zone is assumed to be overpressurized and consequently weakened, it could fail in response to small changes in shear stress. With the passage of time, smaller changes in shear stresses could be adequate to trigger failure. This model is supported by spatial and temporal pattern of seismicity in Koyna region, enhanced conductivity of N–S vertical fractures as opposed to the poor conductivity of the E–W fractures (Snow, 1982), the steep dip of 60° observed during the drill and existence of a thick gouge zone (Gupta et al., 1999) and lack of co-seismic response in water levels at some wells to earthquakes.

There are inherent problems with both these models. The Talwani's intersection model requires occurring of earthquakes at the same intersection over a period of 30 years, which is unusual for this region. Moreover, the earthquakes do not cluster close to an intersection. The model proposed by Rajendran and Harish would require earthquakes to trigger with smaller changes in shear stresses with passage of time. However, that is not the case with Koyna earthquakes. It must be mentioned that both these models and the arguments and data presented to support the models provide necessary inputs and a backdrop to develop a more robust model that could explain the continued seismicity.

5. Aswan, Egypt

In the last 10 years or so, a series of papers have appeared which have relevance to triggered earthquakes in the vicinity of Aswan Dam, where an earthquake of m 5.6 occurred on November 14, 1981. Particularly, there are about a dozen papers in a special volume of the Journal of Geodynamics, edited by Vyskocil and Jacoby (1991), which deal with different aspects of geodynamics of the region and triggered earthquake at Aswan Dam.

Kebeasy et al. (1991) have modelled the crustal structure for the northern part of the Aswan Lake area using seismic waves generated by local earthquakes and seismic profiles. The area is found to be underlain by a normal continental crust of 28 ± 2 km thickness and a normal mantle with a P-wave velocity of 8.1 km/s. Throughout the region, a seismic boundary exist at a depth of about 16 km, separating the upper crust and lower crust with velocities of 5.1 and 6.2 km/s, respectively.

In a more recent paper, Awad and Mizoue (1995a) have analyzed the data of earthquakes in the vicinity of Aswan region and proposed that the seismic activity in the region could be grouped into two earthquake clusters. The Aswan region is characterized by heterogeneous crust with the existence of low and high velocity zones in shallow and deep portions of the crust. Based on the analysis of the earthquake data for the period June 1982 through end of 1991, they demonstrated several characteristic features. The shallow earthquake group at a depth of less than 10 km is characterized by a swarm-like activity whereas the deeper seismicity is characterized by typical foreshock, mainshock–aftershock sequences. The shallow seismicity correlates well with water level fluctuations in the reservoir, which is not found for the deeper events. The questions of relation between the occurrence of shallow and deeper earthquakes cannot be resolved due to lack of adequate data. The Aswan seismicity deviates from the classical concept of triggered seismicity because of the presence of deeper seismic zone as well as a long time delay between the start of filling and the occurrence of November 14, 1981 main shock.

In another interesting study, Awad and Mizoue (1995b) have conducted a tomographic inversion of three-dimensional seismic velocity structure in Aswan region using the Thurber iterative simultaneous inversion programme. There exists a prominent velocity contrast in the shallow portions, which shows a remarkable correlation with the local surface geology. The tomographic inversion involves using three-phase observation at 13 stations from 89 local earthquakes, and an initial velocity model deduced from local velocity experiments (Kebeasy et al., 1991). A heterogeneous crust consisting of a shallow low velocity zone and a deep high velocity anomaly has been identified. The well-resolved low velocity zones appears within the upper 6 km of the crust. This low-velocity zone covers an area of about 35×40 km located along the western bank of the lake. Another major finding of this work is the location of a deep high velocity anomaly, which coincides with the occurrence of seismic activity in the lower crustal layers. The observation that shallow seismic activity occurs within the low velocity zone and the deeper seismic activity is concentrated with the high-velocity anomaly is consistent with their earlier finding where they observed different characteristics of earthquakes occurring in the two zones.

6. Srinagarind, Thailand

During 1990s, one major triggered earthquake sequence was reported from the Srinagarind Reservoir located about 90 km northwest of Bangkok (Chung and Liu, 1992). On April 22, 1983, a large area of Thailand and the adjacent Myanmar was shaken by a rare earthquake of m_b 5.8 (Ms 5.9). The region is relatively stable and had experienced only a few earthquakes in the past. Chung and Liu (1992) report that the mainshock was preceded by foreshocks and was followed by numerous aftershocks. The largest foreshock of m_b 5.2 occurred about 1 week before the mainshock and the largest aftershock of m_b 5.3 took place about 3 h after the mainshock. The seismic moment and stress drop of the mainshock is estimated to be 3.86×10^{24} dyn cm and 180 bars, respectively. Chung and Liu (1992) point out that these events correlate well with the unloading of the Srinagarind Reservoir and the fault motion during the mainshock and the foreshock are mainly of thrust type, whereas the largest aftershock had a large strike slip component. A detailed study of characteristics of the sequence revealed that this sequence belongs to Type II of Mogi's model. Also, the difference between the magnitudes of the mainshock and the largest aftershock is small and the magnitude ratio of largest aftershock to the main shock is large. These characteristics, in combination with the spatial distribution of the events in the earthquake sequence with the reservoir and the record of previous seismicity, led to the conclusion that the sequence is reservoir triggered. The April 22, 1983 earthquake is among the six largest reservoir-triggered earthquakes in the world. The occurrence of earthquake after the delay time of 68 months may correlate well with the relative low permeability of the crystalline metamorphic rock type in the epicentral region.

7. Oroville, California

Among the cases of triggered earthquake sequences associated with artificial water reservoirs, the earthquake sequence in the vicinity of Oroville Dam in California with the mainshock magnitude of 5.7 on August 1, 1975 is perhaps one of the best studied reservoir-triggered earthquake sequence anywhere in the world. It is to be noted that Oroville has triggered only a one-time burst that was significant. Since 1975, this region has not been particularly active, although the pattern of cyclic loading in the reservoir has been recurring. Triggered seismicity at Oroville has been noted as a case of 'delayed' response (Simpson et al., 1988) and the seismicity attributed to diffusion of pore pressure to seismogenic depths. Another factor that discriminates Oroville from other cases of triggered sequences is the distance to the epicentre (about 11 km) from the reservoir and hypocentral depth (8 km) for the mainshock.

Trace of the normal fault that moved in the 1975 sequence is located to the west of the reservoir. The 3D image of P-wave velocities by Rajendran et al. (1993) identified a southwest dipping velocity low, which was identified as a probable expression of this fault. Spatial association of earthquakes with zones of velocity gradient was observed by them.

The role of pore pressure in weakening fault zones has been debated for long (e.g. Scholz, 1990). Low V_p anomalies have been reported from many segments of the San Andreas Fault and an increased pore fluid pressure is considered to be the likely reason for such anomalies (Michael and Eberhart-Phillips, 1991). V_p anomalies observed in the close proximity of reservoirs suggest that the reduction in strength and the consequent decrease in P-wave velocity (Todd and Simmons, 1972) are both consequences of the pore pressure diffusion from the reservoir. Low V_p anomalies observed at Oroville and low P-wave velocities noted at Koyna and Aswan testify to the general reduction in P-wave velocities caused by reservoir impoundment.

8. Common characteristics of reservoir-triggered earthquakes

By early 1970s, over a dozen cases of reservoir-triggered earthquake sequences were known. Gupta et al. (1972a,b) discriminated several characteristics of reservoir-triggered earthquake sequences which were common to these sequences and which discriminate them from natural earthquakes occurring in the same region. These characteristics are:

(1) The foreshock *b* value is higher than the aftershock *b* value, both being, in general, higher than the *b* values for natural earthquake sequences in the regions concerned and the regional *b* values.
(2) In addition to a high *b* value, the magnitude ratio of the largest aftershock to the main shock is also high.
(3) Aftershocks have a comparatively slow rate of decay.
(4) The foreshock–aftershock patterns are identical and correspond to Type II of Mogi's Model, whereas the natural earthquake sequences in the regions in question belong to Type I of Mogi's Model.

The above-mentioned factors are governed by the mechanical properties of the media and their deviation from normal earthquake sequences indicate changes in these properties. Gupta (1992) has elaborated on laboratory experiments and implications of the changes in these mechanical properties.

In the recent years, several more cases of triggered sequences showing the above-mentioned characteristics are reported. Here we mention two of them.

Chung and Liu (1992) reported that the earthquake sequences associated with the April 22, 1983 earthquake of Ms 5.9 in the vicinity of the Srinagarind Reservoir in Thailand belonged to Type II of Mogi's classification, the difference between the magnitude of the mainshock and aftershock is small being only 0.5, the magnitude ratio of aftershock to mainshock is large being about 0.9 and the decay was slow. Similarly, Rastogi et al. (1997a) have reported a case of triggered earthquakes around Dhamni Dam, Maharashtra, India. This earthquake sequence had a maximum magnitude of 3.8 and the *b* value, foreshock–aftershock pattern, decay of aftershocks indicate characteristics similar to those identified for reservoir-triggered earthquake sequences.

Velocity structure is not very precisely known at many sites of RTS. However, we now have reliable information from at least three sites: Oroville, Aswan and Koyna. As reported earlier, at Oroville, the zone of triggered earthquakes is characterized by a low velocity structure (Rajendran et al., 1993). At Aswan, two low-velocity zones are inferred to exist within the upper 6 km of the crust (Awad and Mizoue, 1995a,b) and more than 50% of the triggered earthquakes occur within the upper 10 km of the crust. There are deeper earthquakes centered around a depth of 20 km; however, unlike the shallow earthquakes, they do not show correlation with water level variations. At Koyna, India, concentration of triggered earthquakes is discovered at depths of 0–1 and 5–10 km (Mandal et al., 1998), which coincides with the low velocity zones. The *b* value for shallow events in Koyna is reported to be 0.81 and for the deeper events 0.74. It is well known that increased pore pressure reduces P-wave velocity (Todd and Simmons, 1972; Michael and Eberhart-Phillips, 1991). It would be interesting to see whether other sites of triggered seismicity are characterized by similar velocity anomalies. However, as of now, this information is not available for most of the RTS sites.

9. Mechanism of triggered earthquakes

9.1. Background

The foundation for the understanding of the phenomenon of triggered seismicity was laid through the observation of fluid injection induced earthquakes at the Rocky Mountain Arsenal near Denver, CO, during the early 1960s, and application of Hubbert and Rubey's (1959) work on the mechanism of triggering earthquakes by fluid pressure increase to explain fluid injection induced earthquakes near Denver by Evans (1966). The role of reservoir load in triggering earthquakes may have been considered earlier, but the first definite quantitative analysis was undertaken by Gough and Gough (1970a,b) at Lake Kariba at the Zambia–Zimbabwe border. Bell and Nur (1978) and Roeloffs (1988) also considered the role of reservoir load in triggering earthquakes and pointed out that reservoir load induced stresses at seismogenic depths are very small and can only perturb the ambient stress field. Gupta et al. (1972a,b) identified the rate of increase of the reservoir water level, duration of loading, maximum levels achieved and duration of retention of high levels among the important factors affecting the frequency and magnitude of earthquakes near artificial lakes. Snow (1972) investigated the influence of pore fluid pressure in inducing earthquakes in greatly simplified models of reservoir with finite depth, but infinite width (1D case). More sophisticated models were considered by Withers and Nyland (1976) and Bell and Nur (1978) based on Biot's (1941) consolidation theory (Rice and Cleary, 1976).

The much needed field verification of the theoretical and model developments of concepts of triggered seismicity were provided by in situ measurements of physical properties and examination of physical mechanisms controlling triggered seismicity at the Monticello Reservoir, South Carolina (Zoback and Hickman, 1982). The effect of lake level changes and related derived parameters on triggered seismicity are reported for Nurek Dam, Tadjik Republic (Simpson and Negmatullaev, 1981) and Koyna (Gupta, 1983). The part played by pore pressure diffusion in triggering earthquakes has been addressed by Talwani and Acree (1984/1985). The fault stability changes induced by cyclic variation in water level below a reservoir has been reported by Roeloffs (1988). The rapid response and delayed response type of triggered seismicity was identified by Simpson et al. (1988) and the effect of inhomogeneities in rock properties on triggering earthquakes was addressed by Simpson and Narasimhan (1990).

Bell and Nur (1978) as well as several other investigators have underlined the three main effects of reservoir loading relevant to triggering of earthquakes.

(i) The elastic stress increase that follows the filling of the reservoir;

(ii) The increase in pore fluid pressure in saturated rocks (due to decrease in pore volume caused by compaction) in response to elastic stress increase; and
(iii) Pore pressure changes related to fluid migration. In areas where the water table is low prior to impoundment of the reservoir, the flow of water from the reservoir into unsaturated strata and thereby raising the groundwater table becomes another important factor.

The work mentioned in the above paragraphs are landmarks in our understanding of mechanism of reservoir triggered earthquakes till 1990. It has been included here for the sake of completeness and continuity. In the following paragraphs, we comment upon some important work done during the past 10 years.

9.2. Important questions

In a very interesting article, McGarr and Simpson (1997) drew attention to the adjectives "induced" and "triggered" seismicity and concluded that reservoir-induced seismicity (RIS) should be renamed as reservoir-triggered seismicity (RTS). This has been already discussed earlier in this paper. They also posed the following four questions that need to be answered in an attempt to classify stimulated seismicity:

(i) What is the essential mechanism causing the seismicity?
(ii) Is the corresponding stress change large or small compared to typical seismic stress drops (e.g. Brune, 1970)?
(iii) Does the stimulated seismicity fall in the category of "triggered" or "induced"?
(iv) What factors determine the maximum size of a stimulated earthquake?

After reviewing the cases of reservoir triggered seismicity, McGarr and Simpson (1997) came up with the following answers.

(i) The causal mechanism responsible for reservoir-triggered seismicity can involve all the three factors, namely, purely elastic increases in shear and normal stress, pore-pressure increase due to diffusion and coupled poroelastic increase in pore pressure due to compaction.

(ii) The load imposed by the deepest reservoirs is no more than a few MPa and when extrapolated to shallow hypocentral depths, the stress change may be of the order of 0.1 MPa, which is significantly smaller than the stress drop of the triggered earthquakes which is of the order of 1–10 MPa.

(iii) As already stated, the low induced stresses place reservoir seismicity under the triggered category.

(iv) The maximum size of reservoir-triggered earthquake would depend primarily on the state of stress in the reservoir region. For significant earthquakes of $M \sim 5$ to be triggered, preexisting faults of adequate size to generate such earthquakes must exist in the vicinity of the reservoir.

9.3. Rapid, delayed and continued seismicity

Simpson et al. (1988) had grouped artificial-water-reservoir-triggered earthquakes in two categories. Under the 'rapid response' category, an immediate increase in seismicity on the first filling of the reservoir, or an abrupt change in seismicity following a rapid change in water level is observed. Classic examples are Nurek and Kariba for large reservoirs, and Monticello and Manic 3 for relatively small reservoirs. Under 'delayed response', major seismic events occur relatively late in the life of the reservoir. Aswan, Koyna and Oroville being classical examples where the reservoirs have undergone a number of similar cycles of water level changes before the dominant earthquakes occur. Simpson et al. (1988) further observed that at some sites such as Koyna and Lake Mead, both kinds of responses could be seen. Another important observation made was that in all identified cases of delayed response, the highest level of seismicity appears to be triggered by short-term changes associated with the seasonal maximum of water level. Low magnitude, shallow earthquakes occurring below or in the immediate vicinity of the reservoir characterize the rapid response seismicity, whereas the delayed response events are usually located at a distance of ≥ 10 km, are larger, and occur at greater depths. There are evidences that the fault zone associated with delayed seismicity is well connected with the reservoir.

Delayed response seismicity is largely dependent on diffusion of pore pressure from the reservoir into the hypocentral zone whereas rapid response seismicity is related to changes in elastic stress and related pore pressure changes and does not depend on diffusion of water from the reservoir.

Rajendran and Talwani (1992), while examining the role of elastic, undrained and drained responses in triggering earthquakes at the Monticello Reservoir, South Carolina noted that during the impoundment period, the instability is caused dominantly by elastic and undrained responses. Later, the seismicity showed a consistent pattern associated with diffusion of pore pressure.

In continuation with Rajendran and Talwani's (1992) work, Talwani (1997b) has classified seismic response of a reservoir into two temporal categories. The first, 'initial seismicity', which is most commonly observed, is associated with initial impoundment, or large lake level changes as well as with the lake level increase above the highest level attained previously. The second, which is rare, persists for many years without a decrease in frequency or magnitude is classified as 'protracted seismicity'. Talwani (1997b) has carried out two dimensional calculations, similar to Roeloffs (1988), and inferred that 'protracted seismicity' depends upon frequency and amplitude of lake level changes, reservoir dimensions and hydromechanical properties of the substratum. Longer period water level changes, say, of the order of 1 year, are expected to trigger larger and deeper earthquakes compared to shorter period water level changes. Also, the seismicity is likely to be more widespread and deeper for a large reservoir than for a smaller one.

According to Talwani (1997b), the classification of Simpson et al. (1988) of triggered seismicity into 'rapid' and 'delayed' response are both an integral component of 'initial seismicity'. He has cited Koyna seismicity to be a classical example of 'protracted seismicity'.

We prefer to classify seismic responses into three types: 'rapid response seismicity', 'delayed response seismicity' and 'continued seismicity'. There are reservoir sites that have one or combination of two or three categories of the above-mentioned responses. We are replacing Talwani's 'protracted seismicity' by 'continued seismicity' because the triggered seismicity continues at a given reservoir site, year after year, or at some places after a gap of few years depending upon the state of stress at the seismogenic fault and the level of stimulus provided. The coinage 'protracted seismicity' connotes that once the earthquakes are triggered, they continue, which is not the case. Way back in 1970s, Gupta and Combs (1976) had observed the 'continued seismicity' at Koyna, which we believe is the best site to demonstrate the occurrence of 'rapid', 'delayed' and 'continued seismicity' (discussed later).

Another example is Nurek Dam, which we consider to be a site of continued seismicity. The following observation was made by Simpson and Negmatullaev (1981). Seismicity at Nurek is not a direct function of absolute water levels. However, seismicity does depend on the water level limited to the extent that major bursts of activity occur when the water level is near or above any previous maximum. The major increases in activity in 1972, 1973, 1976 and 1977 occurred when the water level reached a new maximum. Lake Mead is another classical example of 'continued seismicity'. Fig. 12 (after Carder, 1945) shows the variations in Lake Mead water levels and the frequency of tremors. The maximum seismic activity, including the strongest earthquake of M 5 occurred in May, 1939 when the lake level rose again after attaining the normal level. The two other significant upraises in water level occurred during 1941 and 1942 and both were followed by significant seismic activity within a few weeks of the peak levels. Other bursts of seismic activity also follow upraises in water level.

9.3.1. Continued seismicity at Koyna

Prior to the filling of the Shivaji Sagar Lake in 1962, formed by the Koyna Dam, there were no seismic stations operating in the region. However, many senior citizens, questioned within 50 km radius of the lake, categorically denied having experienced any earthquake (Gupta and Rastogi, 1976). The nearest seismic station was at Pune located at a distance of 115 km from the Koyna Dam. This station, equipped with Benioff seismometers, had been operating since 1950. A scrutiny of the seismograms showed no seismic activity, which could be assigned to Koyna region. After filling started in 1962, mild tremors accompanied with sound became prevalent, especially near the dam site. This could be termed as 'rapid response'. The frequency of tremors increased considerably from middle of 1963. Four seismic stations were set up in and around the Koyna Dam during 1963–1964. Guha et al. (1968) reported that there is a general tendency for the tremors to begin 20 km upstream, north of the

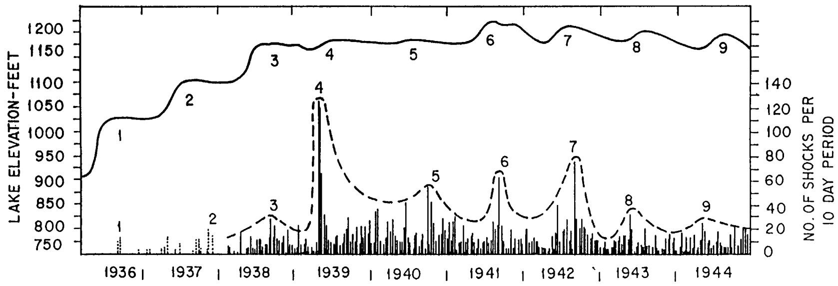

Fig. 12. Lake Mead water levels and the local seismicity. For 1936 and 1937, only the felt shocks are plotted. The rises in water levels and the corresponding bursts of seismic activity are numbered. General trend of tremor-frequency variation is shown by dotted lines (after Carder, 1945).

dam and then epicentres shift to south, before going back again to north. The most damaging December 10, 1967 earthquake of M 6.3 occurred 4 months after the peak water levels, exceeding the previous highs reached in July 1967 (Fig. 13). Gupta and Combs (1976) reported 'continued seismic activity' in Koyna region and observed that although the reservoir was filled to maximum capacity during September 1973, and this was followed by a conspicuous increase in seismic activity including an M 5.1 earthquake, the activity during 1973 was much less severe than during 1967. One possible reason could be an increase in 'threshold level' for relatively large magnitude earthquakes to occur at Koyna. Later, Gupta (1983) pointed out that a necessary, but not a sufficient condition for an $M \geq 5$ earthquake to occur in Koyna region is that the weekly rate of loading should be ≥ 12 m.

The seismic activity at Koyna has continued and earthquakes of $M \geq 4$ occur frequently. Fig. 14 depicts typical variations of water level in Koyna Reservoir (represented by unloading/loading cycles of 1973, 1980 and 1993) and monthly frequency of earthquakes of $M \geq 4$ for the period 1970 through 1999. As can be noted from Table 2, over 150 earthquakes of $M \geq 4$ have so far occurred in Koyna region. It is therefore appropriate to assess the level of seismicity in Koyna region by temporal distribution of $M \geq 4$ earthquakes. Following the heavy Monsoon rains in June/July, the lake level rises rapidly in July and the seismic activity is at a peak during September. It may be noted that although lake level is at a peak during September and October, the frequency of earthquake decreases through January. As also pointed out by Rajendran et al. (1996), there is a mild increase in seismicity during February—related to unloading of the reservoir. However, the seismicity again decreases after February with a minimum in July. From the above analysis, it would appear that while the peak water levels in the lake are achieved during August, the peak seismicity occurs in the month of September. From Table 2, it can be noted that the first $M \geq 5$ earthquake in Koyna region occurred on September 13, 1967. Then, the largest triggered earthquake of M 6.3 occurred on December 10, 1967: after 4 months of the peak water level having being reached. Following the December 10, 1967 earthquake, there were several aftershocks including two of $M \geq 5$ on December 24, 1967 and December 25, 1967. The December 10, 1967 earthquake clearly qualifies to be a delayed response. The next $M \geq 5$ earthquake occurred on October 29, 1968. For the next 5 years, although $M \sim 4$ and smaller earthquakes continued to occur but no $M \sim 5$ earthquakes occurred. Later, $M \geq 5$ events occurred on October 17, 1973; two on September 2, and another on September 20, 1980 and the latest on December 8, 1993 and February 1, 1994. In a nutshell, after the peak water levels are reached at Koyna in the month of August, the seismicity increases. It is most intense in the month of September. During December 1967 and December 1980, largest events of the corresponding water loading of the reservoir occurred after 4 months of reaching the peak reservoir levels exhibiting a delayed response.

Another interesting feature of Koyna–Warna seismicity is the spatio-temporal distribution of earthquakes (Fig. 15). During the month of September, soon after peak levels are reached, earthquakes occur in the vicinity of both the Koyna and Warna reservoirs

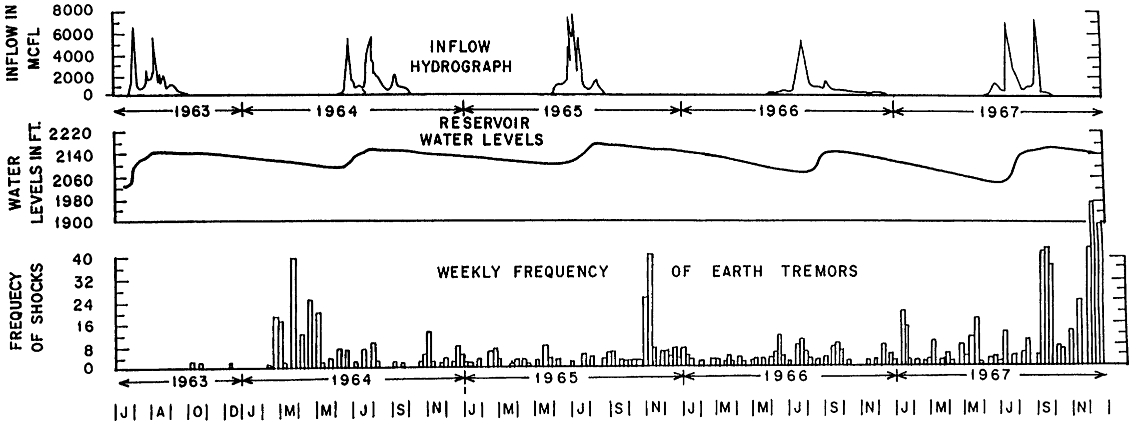

Fig. 13. Inflow hydrograph and water level of the Koyna reservoir compared to the seismic activity in the area (after Guha et al., 1968).

(Fig. 15). During the month of December, the earthquakes are concentrated in the vicinity of Warna Reservoir. However, the unloading related earthquakes in the month of February again spread out and occur in the vicinity of both Koyna and Warna Reservoirs. Another noteworthy observation is that during the months of September and December, shallow (depth ≤ 5 km) events occur, whereas during the month of February, mostly deeper events (depth >5 km) occur.

9.4. In situ measurement of hydraulic properties

Pore pressure diffusion plays a very important role in triggering earthquakes. However, the value of hydraulic diffusivity has been, so far, mostly assumed

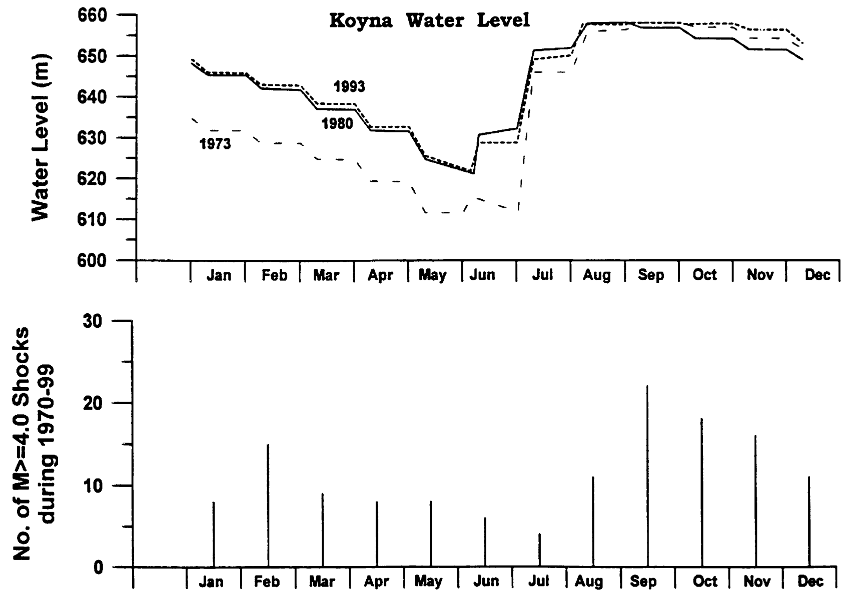

Fig. 14. Water levels in Koyna reservoir for the years 1973, 1980 and 1993 and the monthly number of earthquakes of M 4 and larger for the period 1970 through 1999.

Table 2
Catalog of earthquakes at Koyna of $M \geq 4.0$ since September 1967

D.M.Y. H:M	NGRI Ms	MERI ML	USGS mb	
13.09.67 06:23		5.8		
13.09.67 06:47		4.5		
13.09.67 06:51		4.0		
13.09.67 07:01		4.0		
13.09.67 08:43		4.0		
13.09.67 11:32		4.0		
10.12.67 22:51	6.3	7.0	6.0	
10.12.67 23:52		3.8	5.0	
11.12.67 20:49		5.4	5.2	
12.12.67 06:18		3.6	5.4	
12.12.67 15:48		3.6	5.0	
12.12.67 18:20		4.7		
13.12.67 05:09		4.6		
13.12.67 19:19		4.6		
14.12.67 09:16		4.1		
14.12.67 15:06		4.1		
14.12.67 23:40		4.0		
22.12.67 14:48		4.8		
24.12.67 03:41		4.0		
24.12.67 04:23		4.0		
24.12.67 23:49		5.0	5.5	
25.12.67 00:15		4.2		
25.12.67 00:47		4.2		
25.12.67 17:37		4.6	5.1	
25.12.67 17:59		4.1		
03.01.68 04:35		4.0		
11.01.68 04:37		4.1		
16.01.68 03:33		4.0		
07.02.68 08:09		4.3		
09.02.68 22:52		4.2		
12.02.68 09:13		4.5		
04.03.68 21:36		4.2		
31.08.68 02:53		4.1		
20.09.68 10:11		4.2		
29.10.68 10:00		5.2	5.4	
05.12.68 22:52		4.3		
21.01.69 22:32		4.1	3.6	
13.02.69 18:26		4.2	4.3	
07.03.69 14:28	4.4	4.7		
03.06.69 23:26	4.2	3.5		
27.06.69 20:05	4.5	4.7		
22.07.69 21:49	4.0	3.7		
03.11.69 23:22	4.1	4.5		
04.11.69 05:11	4.2	3.7		
01.01.70 22:30		4.3		
16.04.70 14:46	4.0	3.6		
27.05.70 12:45	4.8	4.4		
08.06.70 05:30	4.1	3.8		
17.06.70 06:48	4.1	3.6		
21.09.70 03:02	4.0	3.7		
25.09.70 04:12	4.6	4.2		
26.09.70 16:36	4.6	4.4		
23.01.71 04:39	4.2	3.7		
14.02.71 01:30	4.0	4.2		
29.05.71 00:27		4.1		
10.08.71 04:30	4.0	3.4		
10.08.71 06:15	4.3	4.2		
22.11.71 10:39		4.2		
01.05.72 21:11	4.2	3.6		
11.05.72 11:49	4.5	3.9		
05.07.72 23:15	4.0	3.2		
11.11.72 04:01	4.1	3.1		
19.04.73 08:45	4.1	3.8		
17.10.73 08:03	4.0	3.2		
17.10.73 14:40	4.1	3.2		
17.10.73 15:24	5.1	5.2	5.0	
24.10.73 18:05	4.6	3.6		
11.11.73 06:48	4.6	2.5		
17.02.74 14:06	4.5	4.7		
28.04.74 09:30	4.0	3.8		
29.05.74 18:26	4.2	3.5		
29.07.74 23:17	4.8	4.3	4.9	
07.08.74 04:23	4.1	3.5		
28.08.74 20:20	4.5	3.8		
11.11.74 15:11	4.3	3.8		
20.12.74 14:16	4.2	3.8		
10.02.75 18:35	4.1	3.6		
02.09.75 23:17	4.2	4.0		
02.12.75 07:40	4.2	3.8		
24.12.75 13:25	4.3	3.6		
14.03.76 05:16	4.8	3.9		
22.04.76 10:46	4.3	3.7		
02.06.76 11:00	4.5	2.4		
16.09.76 14:04	4.2	3.4		
26.09.76 06:48	4.5	3.6		
12.12.76 00:52	4.2	3.9		
19.09.77 00:03	4.7	4.0		
04.11.77 18:57	4.0	3.5		
04.11.77 20:35	4.2	3.6		
04.11.77 20:54	4.4	3.5		
29.11.78 05:30	4.0	3.2		
12.12.78 15:02	4.7	4.0		
26.01.79 19:12	4.0	3.4		
26.09.79 20:02	4.0	3.6		
06.02.80 22:13	4.6	4.4		
19.08.80 22:32	4.3	3.6		
02.09.80 16:39	5.3	4.3	4.9	5.5MS
02.09.80 16:47	4.5	3.8		
20.09.80 07:28	5.5	4.7	4.9	
20.09.80 10:45	5.8	4.9	5.3	
20.09.80 11:22	4.5	3.5		
20.09.80 14:27	4.0	3.6		
20.09.80 23:44	4.2	3.6		
21.09.80 00:00	4.0	3.2		
21.09.80 03:52	4.0	3.4		
21.09.80 08:18	4.0	3.5		
21.09.80 18:19	4.2	3.7		

Table 2 (*continued*)

D.M.Y. H:M	NGRI Ms	MERI ML	USGS mb
22.09.80 11:59	4.3	3.6	
25.09.80 13:38	4.3	3.6	
27.09.80 08:54	4.5	3.5	
30.09.80 13:37	4.2	3.6	
03.10.80 15:20	4.4	3.9	
04.10.80 16:37	5.1	4.1	4.5
04.10.80 19:10	4.3	3.6	
05.10.80 16:09	4.0	3.2	
16.10.80 21:26	4.1	2.2	
17.10.80 21:47	4.1	3.8	
21.10.80 05:02	4.2	3.3	
26.10.80 01:32	4.5	3.7	
26.10.80 01:33	4.4	3.6	
25.01.81 20:30	4.0	3.7	
25.04.82 23:04	4.4	4.3	
05.05.82 07.32	4.2	4.0	
10.09.82 02:42	4.4	3.9	
05.02.83 22:53	4.3	4.4	4.2
21.03.83 15:02	4.1	3.8	
13.05.83 05:53	4.1	3.5	
28.05.83 18:08	4.2	3.8	
25.09.83 18:55	4.8	4.6	4.6ISC
10.11.83 08:55	4.5	3.9	
25.09.84 07:47	4.6	3.8	4.2
14.11.84 11:58	4.7	4.4	4.6
21.12.84 17:26	4.1	4.0	
27.05.85 06:58	4.1	3.7	
29.10.85 08:33	4.0	3.5	
29.10.85 13:59	4.2	3.6	
15.11.85 07:03	4.3	3.8	
21.11.85 09:28	4.0	3.8	
21.11.85 11:39	4.1	3.6	
15.12.85 13:11	4.3	3.9	
28.12.85 14:52	4.0	3.7	
24.07.88 05:33	4.8	4.1	
15.08.88 22:16	4.0	3.4	
15.08.88 23:27	4.1	3.4	
11.09.88 20:40	4.4	3.8	4.3
29.10.89 07:30	4.2	3.6	
03.07.90 13:28	4.1		
06.01.91 22:14	4.4		4.4
20.02.92 05:43	4.1		
01.04.92 01:35	4.0		
27.08.93 22:13	4.1	3.5	4.1
28.08.93 04:27	5.3	4.9	4.9
28.08.93 08:30	4.0	3.6	
03.09.93 23:03	5.0	4.7	4.6ISC
04.09.93 00:53	4.3	4.0	
22.10.93 01:15	4.1	4.3	4.4
08.12.93 01:42	5.2	5.1	5.0
21.12.93 10:10	4.1	4.0	
22.01.94 11:12	4.0	3.7	
01.02.94 09:31	5.5	5.4	5.0
29.03.94 07:51	4.3	3.8	
31.10.94 06:10	4.2	3.9	
12.03.95 08:22	5.1	4.7	4.7
13.03.95 03:09	4.9	4.4	
20.03.95 06:58	4.1	3.7	
15.04.95 22:59	4.1	3.7	
05.09.95 00:53	4.0	3.8	
08.11.95 20:06	4.1	3.9	
15.11.95 04:09	4.1	3.8	
25.02.96 01:03	3.9	3.6	
26.04.96 12:19	4.6	4.4	
25.04.97 16:22	4.3	4.4	
11.02.98 01:09	4.7	4.3	
14.02.98 00:59	4.4	4.3	
07.06.99 15:45	4.3	4.2	

It may be noted that from 1969 onwards, NGRI Ms magnitudes are taken to quantify an earthquake on Ms scale.

or inferred from the migration of seismicity. In an interesting paper, Talwani et al. (1999) have provided the results of in situ measurements of hydraulic diffusivity and Skempton's coefficient from the Bad Creek Reservoir in South Carolina, USA. At an observation well, located 250 m away from the reservoir, the water levels were monitored. The bottom of the well is connected with the bottom of the reservoir by a shallowly dipping 1-m-wide shear zone. The ratio of the change in water level in the observation well to the water level change in the reservoir is a measure of the pore pressure transmitted. Initially, the time lag between the change in water level in well with respect to that in the reservoir was 98 h, which subsequently decreased and stabilized at 72 h. This is interpreted to be due to flushing of fines in the shear zone. A frequency independent diffusivity of $\sim 0.076\ m^2\ s^{-1}$ is estimated for the shear zone, and the Skempton coefficient was calculated to be 0.66 for the undrained response of the reservoir. These observations have a great significance in understanding the pore pressure diffusion.

9.5. Fluid flow and natural earthquakes

Several years ago, Nur and Booker (1972) had proposed that large shallow earthquakes can produce changes in pore fluid pressure comparable to stress drops on faults. The redistribution of pore pressure as a result of fluid flow slowly decreases the strength of the

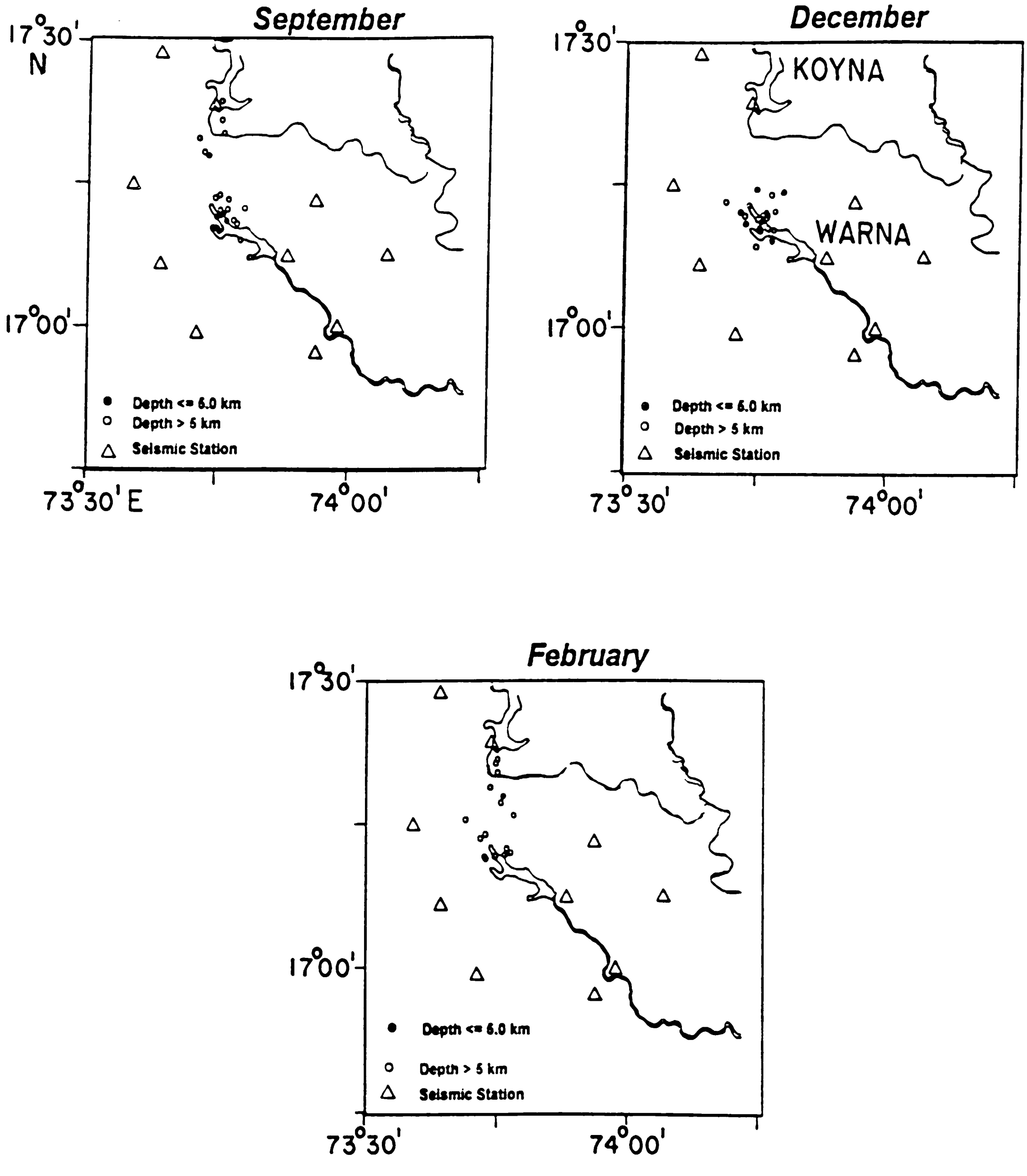

Fig. 15. $M > 2.9$ earthquakes during months of September, December and February for the years 1992 through 1999. Note that during September, earthquakes spread from Koyna to Warna reservoir. In December, they are mostly in the vicinity of Warna reservoir and in February, they again spread out. During February, most earthquakes have depths of more than 5 km. For details, see the text.

rock and may result in delayed fractures. This was an attractive mechanism to explain aftershocks following shallow earthquakes. Nur and Booker (1972) underlined that pore pressure measurements in borewells in seismically active areas would provide a critical test of their suggestion. Recently, Noir et al. (1997) have reported fluid flow triggered migration of events in an earthquake sequence in Central Afar where principal events, including 10 shocks of $M \geq 5.5$, propagated 50 km northwestward in $\sim$ 50 h. Roeloffs (1996) had elaborated on poroelastic techniques in the study of earthquake-related hydrological phenomena. Later, Roeloffs (1998) also reported persistent water level changes in a well near Parkfield, CA, due to local and distant earthquakes. We believe that the kind of results reported by Talwani et al. (1999) and the experiments being conducted at Koyna (Gupta et al., 2000) will improve our understanding of the part played by pore-pressure diffusion in triggering earthquakes.

9.6. Anisotropic models

Most of the theoretical models discussed so far for examining the effect of pore fluid pressure in triggering earthquakes are for isotropic rocks (for example, Bell and Nur 1978; Roeloffs, 1988; and others). Chen and Nur (1992) have investigated the effect of pore fluid pressure in anisotropic rocks and its implication in triggered seismicity. They derived an anisotropic effective stress law and coefficients for transversely isotropic and orthotropic rocks. Through numerical modelling and Mohr diagram analysis, they show that pore fluid pressure in anisotropic rocks causes much larger shear stress variations, either through pore pressure build up or decline, compared to isotropic rocks. Chen and Nur (1992) point out that the deviatoric effects of pore fluid pressure in anisotropic rocks have wide application in comprehending triggered seismicity, earthquake precursors and aftershocks. They show that transient high pore fluid pressure causes transient high shear stresses, which more readily reduce the shear strength, thereby causing the failure, compared to conventional mechanism for the isotropic rocks. Chen and Nur's (1992) approach needs to be applied to specific cases of RTS.

9.7. Other studies

Here, we briefly mention a few important studies related with the triggering of earthquakes by artificial water reservoirs reported during the last few years.

Chander and Kalpna (1997) have explored the notion of fault stability as a measure of the interplay of frictional stresses mobilized and the resolved shear stresses acting on a fault. They suggested that in the simulation of earthquake near a reservoir, the fault stability induced by reservoir (S_r), and at least notionally, the stability induced by the ambient stresses (S_a) should be considered, and the numerical sum of S_a and S_r should be zero on the causative fault at the hypocentre at the time of occurrence of an earthquake. The 2D analysis of Bell and Nur (1978), Roeloffs (1988), and the 3D consideration of Chander and Kalpna (1997) emphasized that filling of a reservoir may stabilize some parts and destabilize other parts of the same fault.

In the past 10 years, limited work has been carried out in the direction of theoretical developments in porous elastic medium exclusively dealing with the simulation of reservoir-induced earthquakes (Shen and Chang, 1995; Kalpna and Chander, 1997, 2000). Kalpna and Chander (1997) solved for stresses and pore pressure induced in a medium comprising of impervious elastic layer resting on a water-saturated porous elastic half space, when the upper surface of the layer is acted upon by a harmonically varying normal stress field. This problem is relevant for those cases of RTS, where a direct hydraulic connection between the reservoir and the hypocentral region of triggered earthquakes does not exist because of intervening impervious rocks and soil, although the hypocentral region may be located in a water saturated zone. Kalpna and Chander (2000) developed an algorithm for simulation of stresses and pore pressure using Green's function method for more realistic laterally finite 3D models of reservoirs. An important aspect of this work is providing the Green's function solutions for the inhomogeneous diffusion equation.

Lee and Wolf (1998) developed mathematical models to study the propagation of excess pore pressures in heterogeneous and fractured rocks. The time required for the pore-pressure front to migrate downwards is directly proportional to the square of the depth and inversely proportional to the permeability of the rock. The models are used to estimate the possible hypocentral depths of periodic seismicity observed near Mt Ogden on the Alaska–British Columbia border. Lee and Wolf (1998) observe that the time lag between hydrological loading and earthquakes is of the order of a few days to weeks, indicating a quick response. The earthquakes are inferred to be triggered hydrologically. If a high degree of vertical interconnectivity of fractures exists, the earthquakes could be triggered at depths of several kilometres.

10. Concluding remarks

In this article, we have reviewed the recent developments in the field of artificial-water-reservoir-triggered earthquakes. Considering the small changes in the stress regime caused by filling of the deepest reservoirs, compared with the stress drop of the associated earthquakes globally, it is appropriate to replace the term 'reservoir-induced seismicity' (RIS) by 'reservoir-triggered seismicity' (RTS). The RTS, based on space, time distribution, and magnitude of events could probably be classified into three categories, namely rapid response seismicity, delayed response seismicity and continued response seismicity. Many reservoir sites have shown more than one type of the response mentioned above. Koyna is a classical case, which has witnessed rapid, delayed and continued response.

For the first time, important hydraulic properties have been measured at the Bad Creek Reservoir in South Carolina, USA. A frequency-independent diffusivity of 0.076 $m^2 s^{-1}$ is estimated for the shear zone and the Skempton coefficient is calculated to be 0.66 for undrained response of the reservoir. These results and the measurement of water levels at a number of bore wells in Koyna, India and their correspondence with RTS have a great significance in understanding the role of pore pressure diffusion in triggering earthquakes. The notion of fault stability as a measure of the interplay of frictional stresses mobilized and the resolved shear stresses acting on a fault has been explored. Filling of a reservoir may stabilize some and destabilize other parts of the same fault. Investigations of the effect of pore fluid pressure in anisotropic rocks and its implication in triggered seismicity is a very important development. Pore fluid pressure changes in anisotropic rocks cause much larger shear stress variations compared to isotropic rocks. It has been shown that transient high pore fluid pressure causes transient high shear stresses, which more rapidly reduce the shear strength, thereby causing failure, compared to conventional mechanism for the isotropic rocks. The approach needs to be applied to comprehend RTS.

Koyna continues to be the most significant site of RTS. So far, over 150 earthquakes of $M \geq 4$ have occurred. During 1990s, two earthquakes of $M \sim 5$ occurred in addition to several smaller earthquakes. In a number of studies, the parameters of earlier earthquakes at Koyna have been redetermined. Deployment of modern seismic data acquisition system has improved the determination of hypocentral parameters. During 1990s, over two dozen important papers addressing various aspects of Koyna seismicity were published.

Earthquake prediction has been one of the most cherished goals of seismologists all over the world for more than 120 years. However, as stated by Max Wyss (personal communication, 1999) "Earthquake prediction research is stagnant by excessive claims of success and excessive criticism". It cannot be overlooked that a few earthquakes have been predicted correctly and there is no question that precursors, such as foreshocks, exists before main shocks. Some of the theoretical models such as self-organised criticality predict a precursory stage. As a matter of fact, what we need to do is to conduct suitable experiments at appropriate sites where earthquakes occur frequently and the site is relatively isolated. We believe Koyna–Warna in western India is one such site where earthquakes have been occurring in a small area of 30×15 km^2 and there are no seismically active regions in the near vicinity. The epicentral region is easily accessible for all kinds of experiments and observations. Earthquakes of $M \sim 4$ occur every year and there is an increase in seismic activity some 15 days before an $M \sim 4$ earthquake. With an improved seismic network, which is now being installed in Koyna region, it is hoped that enhanced foreshock activity would be inferred in real time and should nucleation be detected preceding earthquakes of M 4–5, it may be possible to forecast them with a lead time of 2 days. This would be a very important step in achieving the goal of making scientifically acceptable earthquake forecasts.

We would like to close this review by stating that the study of reservoir-triggered earthquakes provides an exceptionally good input into understanding the physics of earthquakes and should be pursued with a missionary zeal, not only to find safer sites for building artificial water reservoirs, but also for providing a step forward in the long cherished goal of earthquake forecasting.

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