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Ground effects and hydrological changes in the Southern Apennines (Italy) in response to the 23 July 1930 earthquake (M_S=6.7)

E. Esposito[1], **R. Pece**[2], **S. Porfido**[1], **and G. Tranfaglia**[3]

[1]Istituto per l'Ambiente Marino Costiero (CNR), Porto di Napoli, Calata Porta di Massa, 80133 Naples, Italy
[2]Dipartimento di Scienze della Terra, Università di Napoli Federico II, Largo S. Marcellino 10, 80138 Naples, Italy
[3]Istituto Superiore per la Protezione e la Ricerca Ambientale, Via Curtatone 3, 00185 Rome, Italy

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Abstract. The 23 July 1930 earthquake (M_S=6.7) in the Southern Apennines (Italy) was a catastrophic event that produced many effects such as surface faulting, fractures, landslides, settlements, hydrological changes, variations in chemical/physical activity related to the volcanic and/or thermal zones and also acoustic and optical phenomena. It is the first great earthquake of the twentieth century that was studied, thanks to the hydrological monitoring network of the Italian Hydrographic Survey (IHS) set up from 1925 to 1929. For this earthquake we analysed the initial IHS hydrometric and pluviometric data, looking for significant anomalies in springs, water wells and mountain streams. Hydrological data relative to rivers, springs and water wells indicate that some changes can be correlated with the earthquake: a post-seismic excess discharge in some streams, pre- and coseismic decreases in stream flows and water levels in wells, pre- and post-seismic increases in discharges. The pre- and co-seismic stresses and the tectonic deformations were studied in order to find a possible model of interaction between stress state and hydrological variations. The anomalies found in this work can be considered "rebound anomalies", which are the most common precursor reported by many authors and related to increases in porosity and permeability caused by the fracturing that precedes an earthquake. An estimation of the total excess discharge (0.035 km^3) caused by the M_S=6.7 Irpinia earthquake is consistent with the excess discharge of about 0.01 km^3 determined for the M_w=6.9 Loma Prieta earthquake.

Correspondence to: R. Pece
(pece@unina.it)

1 Introduction

In tectonic crustal structures, where moderate to strong ($M \geq 5$) earthquakes take place, also temporary or permanent environmental changes (geomorphic, hydrogeological and structural coseismic features) are generated. The repeated occurrence of these features leaves a geological signature in the recent stratigraphy and topography of an area, unequivocally related to the intensity of the local seismicity. Accurate surveys and interpretations of surface effects represent a valid back-analysis tool for assessing the actual vulnerability of the environment, and to forecast its future behaviour at the occurrence of a significant earthquake. This can be done through the construction of a data base of all the permanent and temporary modifications of the physical environment induced by well-documented earthquakes from selected sample areas.

For this purpose, for each great earthquake that occurred in the last century in the Southern Apennines of Italy (Fig. 1), a complex carbonate formation, we collected well-founded information and records of hydrological phenomena that cannot be ascribed to environmental or anthropic causes (Onorati and Tranfaglia, 1994; Onorati et al., 1994; Esposito et al., 1998, 1999, 2000, 2001; Pece et al., 1999; Porfido et al., 2002, 2007).

The 23 July 1930 Irpinia earthquake took place in the northern part of the Southern Apennines, which are a complex curved structure elongated from the Abruzzo-Molise to the Basilicata-Calabria border (Fig. 1a). The Apennines are a Neogene and Quaternary thrust and fold belt located in the hanging-wall of the west-plunging Adria plate (Cinque et al., 1991; Doglioni et al., 1996).

Several kilometers of vertical displacement occurred on the Tyrrhenian margin mainly along southwest-dipping normal and oblique slip faults. The extensional movements, due

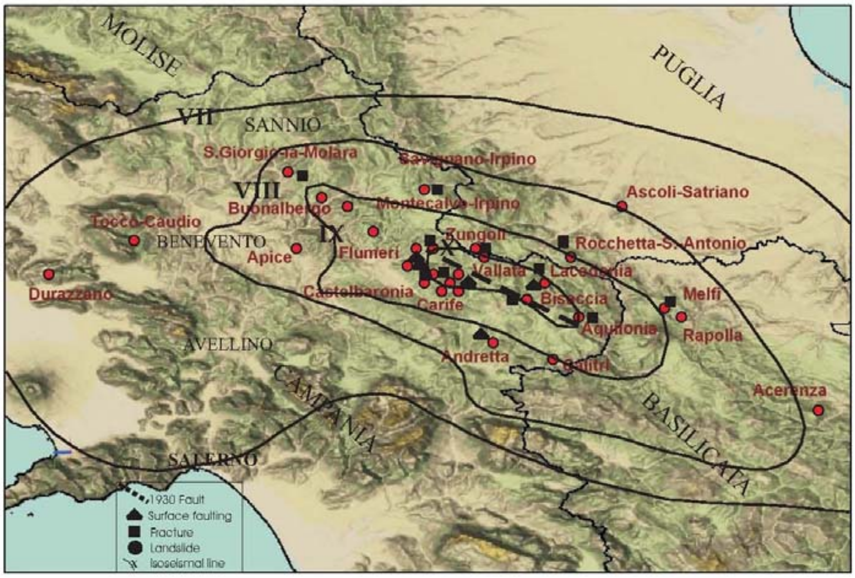

Fig. 1a. Shaded relief of the study area with ground effects, macroseismic field and fault of the 23 July 1930 earthquake.

to their progressive shift toward the eastern sectors of the still uplifting Apennines, created deep tectonic basins elongated north-west.

Studies of active tectonics and paleoseismicity confirm that extensional tectonics is still active in the Southern Apennines, with slip-rates of several tens of millimeters per year, mostly for active faults from late Holocene until now (Vittori et al., 1991; Westaway, 1992, 1993; Pantosti et al., 1993; Ascione et al., 2003). The present-day tectonic setting of the mountain belt is governed by a system of Quaternary faults responsible for frequent moderate to strong crustal earthquakes, with typical hypocentral depths of 7–20 km (Amato et al., 1997).

The 23 July 1930 (M_S=6.7) earthquake affected an area of over 6000 km^2 from the Campanian Plain to the Bradanic Foredeep (Fig. 1a, in the isoseismal area VIII MCS (Mercalli-Cancani-Sieberg scale), SW of Acerenza). The district of Avellino experienced an intensity of X MCS. Many surficial phenomena were reported, and in this paper the major ground effects and hydrological changes are studied.

For this earthquake we analysed hydrometric and pluviometric data of the Italian Hydrographic Survey (IHS), looking for significant anomalies in springs, water wells and mountain streams.

Changes in the water-rock interaction are caused by the seismic stresses in the area where the tectonic deformation leads to the seismic event. For earthquakes in Southern Italy, the pre- and co-seismic stresses and the tectonic deformations have been correlated with the hydrologic changes in order to find a possible model of interaction between stress state and hydrological variations (Onorati et al., 1994; Esposito et al., 2001).

Table 1. I_0 epicentral intensity; M macroseismic magnitude according to CPTI04 (2004), (*surface wave magnitude).

Year	Month	Day	Hour-Min	Epicentral Area	I_0	M
1688	06	05	15:30	Sannio	XI	6.7
1694	09	08	11:40	Irpinia-Basilicata	X-XI	6.9
1702	03	14	05:00	Sannio-Irpinia	IX-X	6.3
1732	11	29	07:40	Irpinia	X-XI	6.6
1805	07	26	21:00	Molise	X	6.6
1930	07	23	00:08	Irpinia	X	(*)6.7
1962	08	21	18:19	Irpinia	IX	(*)6.2
1980	11	23	18:34	Irpinia	X	(*)6.9

2 The 23 July 1930 Irpinia earthquake

The 23 July 1930 earthquake, local hour 01:08:29, occurred in the districts of Benevento and Avellino, in the most seismic part of the Southern Apennines. Seven other earthquakes with I≥IX MCS and macroseismic magnitude (M) or surface wave magnitude (M_S) greater than 6.0 have occurred in this area in the last four centuries (Table 1).

The epicenter of the main shock (Fig. 1a) was located at 41°05′ N and 15°37′ E in Irpinia (Boschi et al., 1995; Freeman, 1930). The earthquake affected a very wide area, 36 000 km^2 (Alfano, 1931), comprising the regions of Campania, Puglia and Basilicata. The main shock was particularly destructive, resulting in 1425 fatalities, about 10 000 injured and more than 100 000 homeless people, 22 villages

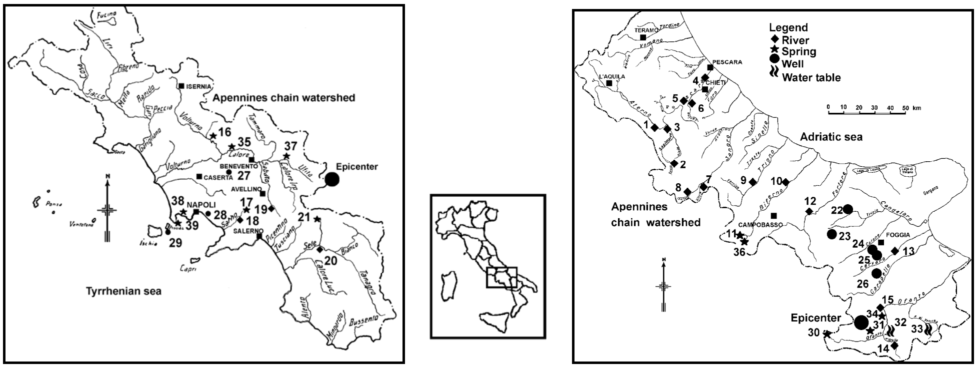

Fig. 1b. Hydrological changes of the 23 July 1930 earthquake. Numbers indicate the sites listed in table 3. The Tyrrhenian side and the Adriatic side can be merged in the Epicenter point.

destroyed and about 40 000 dwellings damaged (Spadea et al., 1985). The epicentral zone of I_{MCS}=X is elliptical and extends over an area of 180 km^2 with the major axis of 34 km parallel to the Apennine trend (WNW-ESE; Fig. 1a). The area of the greatest effects, primary (surface faulting) and secondary effects (slope movements, ground cracks, hydrological anomalies) ($I_{MCS} \geq$VIII) is also elliptical, extending over about 6000 km^2.

Many foreshocks and aftershocks accompanied the main event. At least two foreshocks preceded it at 23:30 on 22 July and at 00:30 (Oddone, 1931; Alfano, 1930). The aftershocks with destructive effects occurred until 1931, also with intensity $I_{MCS} >$VI (Spadea et al., 1985). Surface wave magnitudes (M_S) in the range 6.2–6.7 have been estimated for the 23 July 1930 earthquake. Margottini et al. (1993) utilized 41 observations of amplitude and period, and calculated M_S=6.6±0.3.

Whereas Westaway (1992) determined a seismic moment of Mo=3.2×10^{25} dyne-cm, Jimenez et al. (1989), on the basis of seismograms recorded at Jena (Germany), calculated Mo=2×10^{25} dyne-cm. The fault plane orientation was WNW-ESE (Apennine chain trend), the fault length was 32.6 km and the depth 15 km, estimated on the basis of the equivalent ray of the major isoseismal lines (Martini and Scarpa, 1983; Gasperini et al., 1999).

3 Ground effects

The studies of ground effects benefited from detailed descriptions of the seismic event from numerous historical and scientific sources, and allowed recognition of primary effects (surficial faulting), secondary effects (fractures, landslides, settlements, hydrological changes, variations in the chemical/physical activity related to the volcanic and/or thermal zones) and also acoustic and optical phenomena.

A NW-SE trending fracture several kilometers long was observed (Fig. 1a). This fracture awakened the interest of several scientists because of its exceptional length and the displacement up to 40 cm and was interpreted by Oddone (1931, 1932) to be a fault reactivation. It is drawn in Fig. 1a with a dashed line within the I_{MCS}=X isoseismal line showing a NW-SE trend, for a length of about 38 km (Porfido et al., 2007; Serva et al., 2007).

The earthquake caused many types of sliding phenomena, which mainly affected the rural area and, to a lesser extent, the towns around the epicentral area. Twenty-six landslides were observed (Table 2). Ten of them were rotational slides and slump-earth flows; for the other slides the kind of movement and/or topographic position was not clear (Porfido et al., 2002, 2007; Serva et al., 2007). Slope movement involved considerable volumes of material. In the Benevento district in the San Giorgio la Molara village a notable landslide 3 km long had a front of 1 km (Vari, 1931). It has been reactivated with the strong earthquakes of 1688, 1805, 1962 and 1980 (Esposito et al., 1998). It ha been also possible for the localities listed in Table 2 to assess the Intensity according to ESI2007 scale (Michetti et al., 2007).

Landslide distribution indicates (Fig. 1a) that the area of maximum density is consistent with the I_{MCS}=IX-X damage level. In Fig. 2 the cumulative number of landslides is plotted against the fault rupture segment distances. Most of the landslides occurred near the fault (0–20 km), their number decreasing sharply with distance. The maximum distance from the fault rupture segment was 52 km.

The ground effects of this earthquake were not enhanced by the presence of saturated soils because the seismic event occurred in the summer and the level of the water table located in wedge soil landslides is characterised by a low discharge. Ground failure, in particular sliding and fracturing, was concentrated in the I_{MCS}=IX–X areas of the macroseismic field, and surface faulting trended NW–SE, consistent with the isoseismal pattern of intensity distribution.

Table 2. Ground effects of the 23 July 1930 earthquake.
LEGEND: Tsr – tectonic surface rupture, C – compaction, Sef – slump-earth flow, Fr – fractures, Rs – rotational slide.

No.	Site	Altitude (m)	Latitude (N)	Longitude (E)	Type of landslide	Fault distance (km)	Epicentral distance (km)	Intensity I_{MCS}	Intensity ESI 2007
1	Acerenza	831	40.800	15.950	Slide	49.0	56.0	7.5	8
2	Andretta	840	40.933	15.317	Tsr, slide	7.0	15.5	7.0	7
3	Anzano di Puglia	800	41.117	15.283	Slide	9.0	9.0	9.5	?
4	Apice	225	41.117	14.933	Slide	11.0	37.0	8.0	>4
5	Aquilonia	772	40.983	15.483	Rs-sef, Fr	5.5	10.5	10.0	8
6	Ariano Irpino	778	41.150	15.083	Rs-sef, Fr	1.5	27.5	8.0	10
7	Ascoli Satriano	425	41.200	15.567	Slide	30.0	22.5	8.0	7
8	Bisaccia	860	41.017	15.383	C, Fr	2.5	6.0	8.0	8
9	Buonalbergo	555	41.217	14.983	Sef,	11.0	39.0	8.0	7
10	Calitri	525	40.900	15.433	Rs-sef	9.0	19.5	8.0	7
11	Carife	640	41.033	15.217	Slide	4.0	14.0	8.0	7
12	Castelbaronia	740	41.050	15.183	Slide	3.5	16.0	9.0	7
13	Durazzano	286	41.067	14.450	Slide	52.0	77.5	8.0	7
14	Flumeri	440	41.083	15.150	Sef,Tsr	2.0	18.0	8.5	10
15	Lacedonia	756	41.050	15.417	Slide, Fr	9.0	4.0	10.0	7
16	Melfi	531	41.000	15.650	Slide, Fr	21.0	25.0	9.0	7
17	Montecalvo Irpino	625	41.200	15.033	Sef, Fr	5.5	32.5	9.0	9
18	Rapolla	439	40.983	15.683	Slide	22.0	27.5	8.0	7
19	Rocchetta S. Antonio	630	41.100	15.467	Slide, Fr	14.5	8.0	9.0	8
20	S. Giorgio la Molara	667	41.267	14.917	Rs-sef, Fr	19.0	45.0	7.0	8
21	S.Sossio Baronia	650	41.067	15.200	Slide, Tsr	0.0	14.0	8.0	10
22	Savignano Irpino	698	41.233	15.183	Slide	14.0	25.0	8.0	7
23	Scampitella	775	41.100	15.300	Slide	7.0	7.5	9.5	9
24	Tocco Caudio	497	41.133	14.617	Slide	36.0	63.0	8.0	7
25	Trevico	1090	41.050	15.233	Sef, Fr	1.0	12.5	9.0	8
26	Vallata	870	41.033	15.250	Sef	2.0	10.5	8.0	8
27	Vallesaccarda	650	41.067	15.250	Sef, Fr	1.0	10.0	7.5	7
28	Villanova del Battista	742	41.117	15.167	Tsr	2.0	19.0	10.0	10
29	Zungoli	657	41.117	15.200	Fr	5.0	15.5	8.0	7

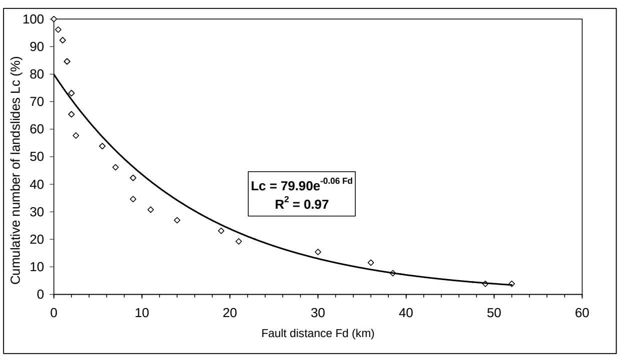

Fig. 2. The cumulative number of landslide Lc (%) versus fault segment distance Fd (km) in the 1930 earthquake.

In many villages various kinds of rumbles were heard (from loud thunder to prolonged whistles). In areas closer to the epicenter (VIII$\leq I_{MCS} \leq$X) optical phenomena were observed ("seismic flashes"; Alfano, 1930).

4 Hydrological changes and hydrothermal phenomena

Hydrological changes were observed in the whole macroseismic field, mostly in the far field, near the main carbonate aquifers in a widespread karstic environment. They include flow increases both in springs and wells, turbid water and drying up of springs, appearance of new springs and variations in the chemical parameters of waters.

From the data collected by the IHS (Annales 1925–1940) 29 anomalies (Table 3, Fig. 1b) were chosen by evaluating the shape and timing of hydrological changes in 151 sites both in the Tyrrhenian and in Adriatic watersheds (48 wells,

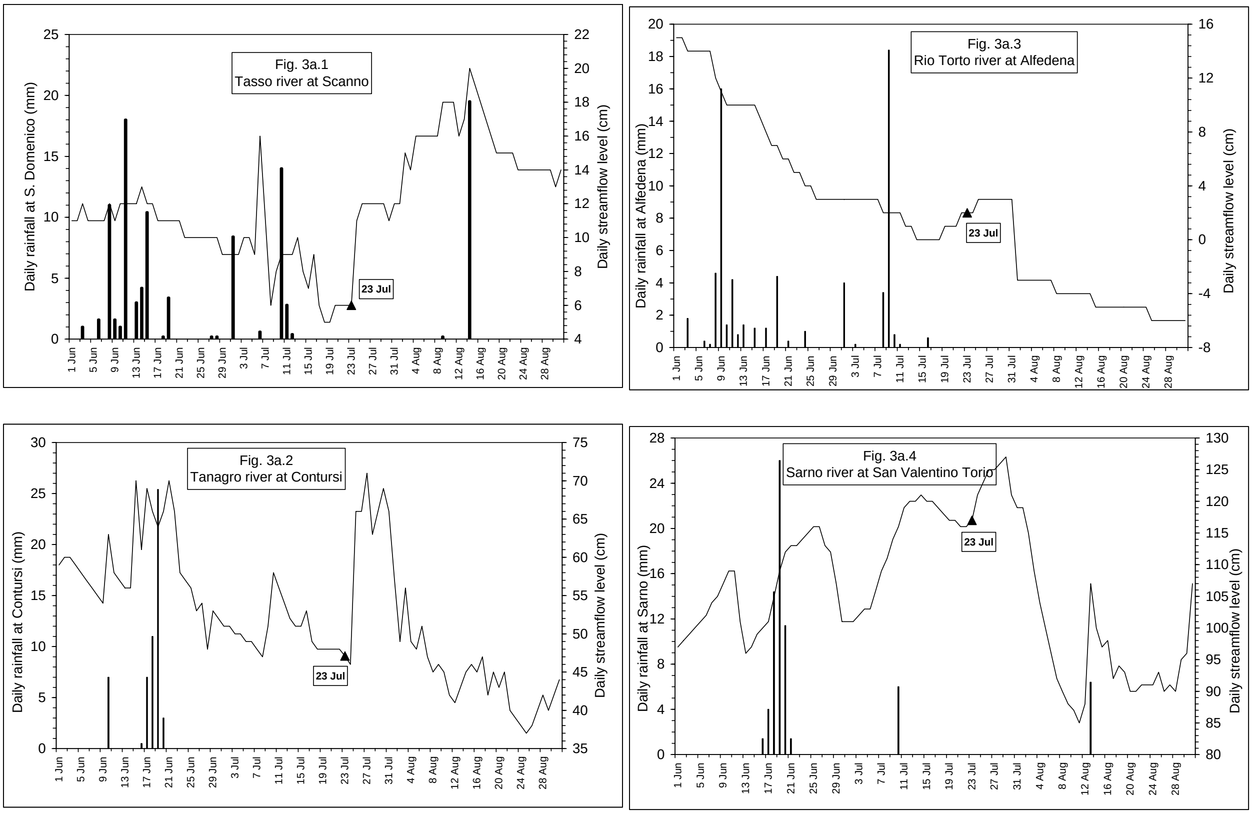

Fig. 3a. Hydrological changes in rivers correlated to the 23 July 1930 earthquake. Black bars indicate rainfalls (left scale).

88 stream gauge stations, 15 springs). Other anomalies are reported elsewhere and are listed from 30 to 39 in Table 3.

Spring flow increases occurred at Madonna del Carmine (Fig. 1b, point 30), at Monte della Guardia (Fig. 1b, point 31) and at Telese (Fig. 1b, point 35) (Majo, 1931a; Boschi et al., 1995). Spring flow at Madonna del Carmine increased from 10 l/min to 40 l/min after the earthquake, and at Monte della Guardia increased from 5 l/min to 16 l/min at the end of August.

At Solfatara, a volcanic crater in Pozzuoli (Fig. 1b, point 38), at a distance of about 100 km from the epicenter, variations in endogenous activity were observed for about 20 days after the earthquake. Alfano (1931) and Majo (1931b) report a temporary decrease in fumarolic gases, diffuse H_2S emanation from the soil, strong gas bubbling in a mud pool, and a notable temperature increase in monitored points. The possible influence of seismicity on gas release from depth can be demonstrated on the occasion of a small earthquake felt locally on 12 August 1930 near Pozzuoli. Table 4 reports the temperatures measured in various sites inside the Solfatara crater (a) before 23 July, (b) during the period 28 July–8 August and (c) during the period 13–26 August (Majo, 1931b). In a nearby site, at Stufe di Nerone (Fig. 1b, point 39), a considerable increase in CO_2 and temperatures was also observed.

Another site with fumaroles and mud pools, Ansanto Valley (Fig. 1b, point 37), situated about 20 km from the epicenter, presented an increase in gas emission and mud boiling, together with light flashes (Alfano, 1930; Ricciardelli, 1930). This is a very sensitive site since such phenomena also occurred here for other earthquakes in the Southern Apennines (Balderer et al., 2000; Italiano et al., 2000).

Near Venosa (the Vulture volcanic complex, not far from the epicentral zone, Fig. 1b, point 33) an increase in soil temperature was measured; Oddone (1932) imputed it to chemical reaction produced by a water table uplift in layers with Fe and S.

Effects on fumarole activity at distances of about 100 km from the epicenter seems hard to explain, but correlations between seismicity and volcanic phenomena are well known (Wakita et al., 1985; Hill et al., 2002; Husen et al., 2004).

Hill et al. (2002) pointed out that earthquakes and volcanoes are linked through plate tectonics and large earthquakes are capable of triggering eruptions within a matter of minutes or days at nearby volcanoes. In USA, a series of earthquakes as large as M=6.3 on 25–28 May 1980, caused turbidity and temporary increases in the discharge of hot springs in the Long Valley caldera of East-Central California. These earthquakes had other obvious effects on the hydrothermal system, including emptying and refilling of boiling pools and

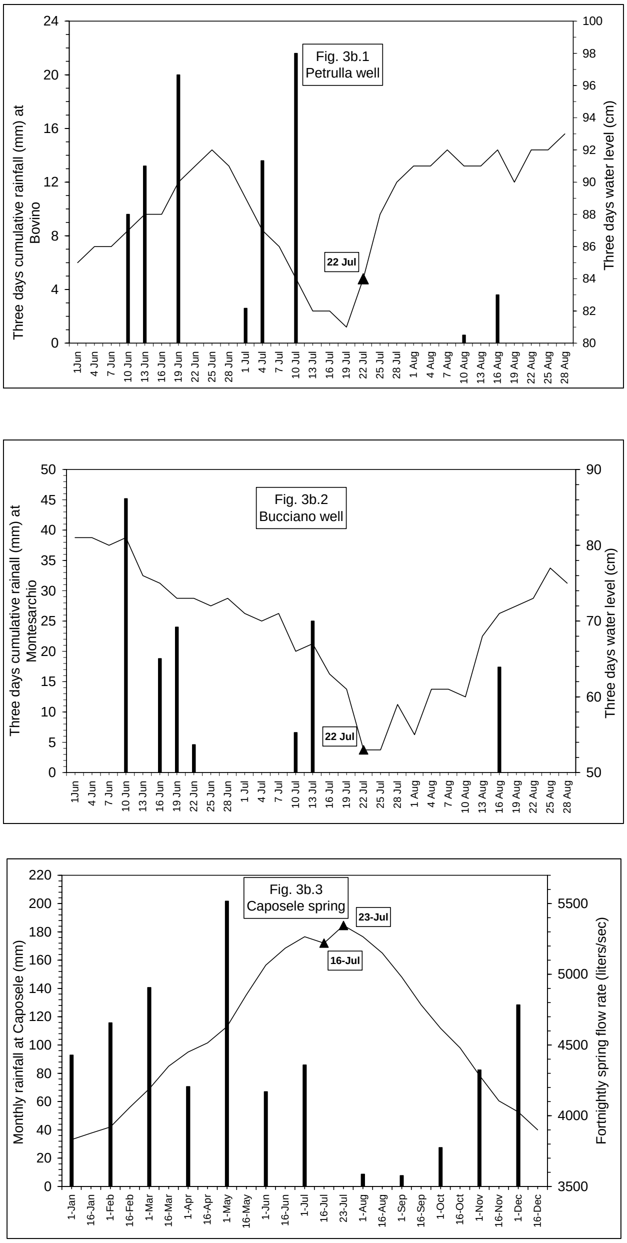

Fig. 3b. Water level in 2 wells, and the fortnightly flow rate of Sanità spring at Caposele correlated to rainfall. Black bars indicate rainfall (left scale).

temporary increases in fumarolic activity (Sorey and Clark, 1981). In Central Japan anomalies in gas compositions were observed at fumaroles (at an epicentral distance of 9 km) and three mineral springs (at epicentral distances of 50, 71 and 95 km) about 1–3 months prior to an inland earthquake of M=6.8 on 14 September 1984 (Sugisaki and Sugiura, 1986).

Husen et al. (2004) report changes in geyser eruption behavior in Yellowstone National Park at very large distances (more than 3000 km from the epicenter) for Denali fault earthquake (Alaska), M=7.9. They interpreted these changes as being induced by dynamic stresses associated with the arrival of large-amplitude surface waves. They reported also an increase of seismic activity in Yellowstone Park and suggest that this seismicity were triggered by the redistribution of hydrothermal fluids and locally increased pore pressure.

It is plausible that such effects would occur in Southern Italy which is affected by young active tectonics with frequent strong earthquakes and many volcanically active areas (Pece et al., 1999).

5 Discussion

Many anomalous behaviours of aquifers have been noted before, during and after a seismic event: sudden increases/decreases in spring flows, changes in piezometric levels in water wells, and increases in the emanation of deep gases (Gordon, 1970; Sorey and Clark, 1981; Whitehead et al., 1984; Wakita et al., 1985; Igarashi et al., 1992; Briggs, 1994; Curry et al., 1994; Rojstaczer and Wolf, 1992, 1994; Quilty et al., 1995; Schuster and Murphy, 1996; Balderer et al., 2000; Italiano et al., 2000; Thorson, 2001; Montgomery and Manga, 2003; Husen et al., 2004).

Characterizing the behaviour of aquifers and detecting anomalies in the late 1930s may be easier than in subsequent years since water resources were less exploited at that time. They are: pre- and co-seismic decreases in stream flows and water levels in wells; post-seismic increases in most of the discharges; only in some cases are they pre-seismic.

In this work we illustrate the features of 7 types of the hydrological changes that we consider anomalous and connected with the 1930 earthquake.

The first category of anomalous behaviour consists of decreases in stream flows before the earthquake, followed by increases after the seismic event. Fig. 3a and b shows the data collected daily at two stream gauges (located very near great springs) on the Tyrrhenian side and at two stream gauges on the Adriatic side. In Fig. 3b the water levels measured with a 3-day frequency in one well in the Adriatic watershed and one on the Tyrrhenian side are reported, as well as the flow rate of Sanità Spring at Caposele. In Fig. 3a.1 (Table 3, point 2) the anomaly consists of a sharp decrease in stream flow a few days before the seismic event, even if high rainfall preceded this decrease. The increase after the seismic event seems imputable to an anomalous discharge of the tributary springs that lasted for more than 10 days. In Fig. 3a.2 (Table 3, point 20), after the decreasing summertime trend, with a minimum reached on 24 July, there is a notable post-seismic increase from 25 July to 12 August due to contributions from numerous large springs. In Fig. 3a.3 (Table 3, point 8) the anomaly is a temporary increase of a few cm after the earthquake.

Table 3. Hydrological changes of the 23 July 1930 earthquake.

No.	Site	Altitude (m)	Latitude (N)	Longitude (E)	Fault distance (km)	Epicentral distance (km)	Intensity I_{MCS}
1	Aterno river at Molina	442.5	42.133	13.733	154.0	179.0	3.0
2	Tasso river at Scanno	925.6	41.900	13.883	127.5	154.0	4.0
3	Sagittario river at Capo Canale	280.0	42.133	13.867	142.5	167.5	3.0
4	Pescara river at Maraone	240.5	42.250	13.817	150.0	200.0	3.0
5	Pescara river at Scafa	87.5	42.267	13.983	153.0	178.0	3.0
6	Lavino river at Scafa	85.0	42.250	14.017	150.0	175.0	3.0
7	Zittola river at Montenero	822.0	41.750	14.083	102.0	129.0	3.0
8	Rio Torto river at Alfedena	923.0	41.733	14.033	106.0	107.5	3.0
9	Trigno river at Trivento	211.0	41.800	14.567	86.5	107.0	5.0
10	Biferno river at Guardialfiera	120.0	41.800	14.817	92.5	96.0	5.0
11	Biferno river at Boiano (Maiella spring)	500.0	41.483	14.467	62.5	90.0	3.5
12	Fortore river at Stretta di Occhito	155.0	41.633	15.000	55.0	70.0	5.5
13	Cervaro river at Beccarini bridge	3.6	41.517	15.900	80.0	72.5	6.0
14	Atella river at Atella	406.0	40.850	15.633	21.0	30.5	8.0
15	Ofanto river at Rocchetta S. Antonio	212.0	41.067	15.550	17.5	14.0	9.0
16	Pila spring at Pontelatone	20.6	41.200	14.283	84.0	95.5	7.0
17	Rio Palazzo spring at Sarno	18.0	40.817	14.617	62.5	71.0	7.0
18	Sarno river at S. Valentino Torio	16.0	40.800	14.567	58.0	69.0	7.0
19	Sabato river at Serino (Urciuoli spring)	349.0	40.883	14.850	36.0	46.0	6.5
20	Tanagro river at Contursi	42.0	40.633	15.233	41.0	50.0	5.0
21	Caposele (Sanità spring)	426.0	40.800	15.217	22.5	31.0	6.0
22	Torretta well	64.8	41.617	15.433	62.5	62.5	6.0
23	Mercaldi Vecchio well	40.6	41.600	15.650	63.0	59.0	6.0
24	Tuoro di Massella well	128.5	41.400	15.500	48.5	40.0	6.0
25	Petrulla well	44.1	41.567	15.683	62.0	57.5	7.0
26	Mortelito well	141.6	41.317	15.567	37.5	31.0	7.0
27	Bucciano well	266.3	41.067	14.567	42.5	68.0	7.5
28	Pomigliano well	1.4	40.950	14.383	100.0	121.0	6.0
29	Procida well	24.6	40.750	14.017	97.5	119.0	6.0
30	Vallata (Madonna del Carmine spring)	870.0	41.033	15.250	2.0	10.5	8.0
31	Aquilonia (Monte della Guardia spring)	732.0	40.983	15.483	5.5	10.5	10.0
32	Atella (water table)	500.0	40.883	15.650	22.5	32.5	8.0
33	Venosa (water table)	415.0	40.967	15.817	34.5	38.5	7.5
34	Rocchetta S. Antonio (spring)	650.0	41.100	15.567	14.5	8.0	9.0
35	Telese (spring)	64.0	41.200	14.517	45.0	73.5	7.0
36	Boiano (spring)	488.0	41.483	14.467	61.0	90.0	3.5
37	Mefite d'Ansanto (boiling spring)	750.0	40.950	15.167	13.0	20.0	8.0
38	Pozzuoli (Solfatara) (boiling spring)	93.0	40.833	14.133	86.0	109.0	6.0
39	Bacoli (Stufe di Nerone) (boiling spring)	30.0	40.800	14.083	91.0	113.0	6.0

Table 4. Influence of 23 July 1930 earthquake on gas release.

Site	Town	(a) T (°C)	(b) T (°C)	(c) T (°C)
Fangaia (mud pool)	Pozzuoli	99.5	104.5	99.4
Bocca Grande (main Fumarole)	Pozzuoli	162.5	163.8	162.0
Pietra Spaccata	Pozzuoli	98.0	101.5	98.2
Stufe di Nerone	Bacoli	92.0	98.0	93.0

Of great interest is the post-earthquake behaviour of 3 springs that contribute to the Sarno river (Fig. 3a.4; point 18 in Table 3 and Fig. 1b). The measurements carried out at San Valentino Torio (Table 3, point 18), where the total contribution of the 3 springs is measured, indicate a stream flow increase with a maximum of 127 cm on 29 July (6 days after the earthquake) followed by a decrease to a minimum of 88 cm on 12 August, a minimum level never reached before.

These types of variations have been observed for many earthquakes all over the world. In the USA, Whitehead et al. (1984) observed many significant hydrologic changes after an earthquake on 28 October 1983 in Idaho (M=7.3). Discharge measured at 10 springs and 48 stream gauging stations of the Big Lost River and surrounding watersheds increased in some instances by more than 100%. The Loma Prieta earthquake (17 October 1989) with M_w=6.9 produced hydrogeological effects reported by several authors who analysed the records of many gauging stations. Briggs (1994) analysed the hydrological effects of this earthquake in Waddell Creek watershed near Santa Cruz (California) at about 38 km from the epicenter. Numerous new springs appeared, and many inactive springs resumed flow; the springs maintained an exponential recession with minimal rain interference until they ceased flowing abruptly. As a consequence, post-seismic discharge near the mouth of Waddell Creek rose to 12.5 times the pre-earthquake discharge, followed by a gradual recession which was obscured by rain runoff beginning after about 50 days. Also Curry et al. (1994) observed very significant and unexplainable increases in the San Francisco peninsula and Santa Cruz Mountains watersheds immediately after the main shock of the Loma Prieta earthquake. For the watersheds of Lorenzo and Pescadero, Rojstaczer and Wolf (1992, 1994) observed that stream flows increased at most gauging stations within 15 min after the earthquake. Groundwater levels in the upland parts of watersheds were locally lowered by as much as 21 m within weeks to months after the earthquake.

Levels in water wells exhibited a general post-seismic increase. At Petrulla (Fig. 3b.1, point 25) and Bucciano (Fig. 3b.2, point 27) June and July rainfall did not influence the summer decreasing trend, and the increase lasted throughout August; note that at the Petrulla well the increase started 3 days before the earthquake. Fig. 3b.3 (point 21) shows the flow rate of Caposele spring at 22.5 km from the fault (Table 3, point 21). A discharge increase of 150 liters/sec (about 3%) was measured a few hours after the seismic event, compared to the measurement on 16 July 1930, a week before the earthquake (Celentani Ungaro, 1931).

Schuster and Murphy (1996) describe an analogous hydrogeological effect for the Draney Peak earthquake, M_w=5.9 in Idaho-Wyoming (USA), on 3 February 1994: a marked increase in groundwater flow (from 4527 to 5695 l/min) occurred at the spring for the Auburn Fish hatchery, 5 km NE of the epicenter.

Also for the Idaho earthquake, Whitehead et al. (1984) analyzed water levels in 69 wells: those near the epicenter generally increased rapidly after the earthquake, by as much as 3 m. Igarashi et al. (1992), for the 2 February 1992 Tokyo Bay earthquake (M=5.9), reported possible precursor water level changes detected by the long-term groundwater observation sites. Three observation wells, about 90–110 km away from the hypocenter, showed anomalous changes: a rise and fall in water levels of 3–10 cm which began simultaneously 1–1.5 days before the earthquake. They excluded that rainfall or pumping could produce this change. The water level fall began to recover about 6 h before the earthquake, followed by a coseismic rise of about 20 cm.

In Fig. 3 the rainfall is shown. Analysis of the yearly rainfall from 1925 to 1940 shows that 1930 had slightly less than average rainfall. Moreover, the epicentral area was less rainy than the mountainous part of the Apennines and watersheds on the Tyrrhenian side (Esposito et al., 2000). The absence of rain on the days preceding and following the event shows that the increase in the level of the aquifer was totally due to variations in spring flow rates that flow down to the riverbeds.

It is difficult to assess the anomalous variations (negative or positive). In some instances the stream flow data are sufficient to permit estimates of the total “excess” stream flow derived from a particular seismic event. Using the extensive USGS hydrological network it was estimated that the Hebgen Lake earthquake (17 August 1959; M=7.5) apparently produced about 0.3 km^3 of water, the Borah Peak earthquake (28 October 1983; M=7.3) about 0.5 km^3 of water, and the Loma Prieta earthquake (17 October 1989; M_w=6.9) only about 0.01 km^3 of water (Muir-Wood and King, 1993; Rojstaczer and Wolf, 1994).

We performed an evaluation of the stage-discharge rating curves for 11 streams for which sufficient data were available (Table 5). By assuming that the daily values collected in 1930 were constant in the 24 h time frame, we calculated the average discharge in the entire anomalous period (Qav) and, obviously, the total discharge (Qtot) in this period. This permits a rough quantification of excess discharge (about 0.035 km^3 for these 11 streams) which does not appear to be correlated with the distance from the epicenter.

Gordon (1970), following the Meckering earthquake (Western Australia) of 14 October 1968 (mainshock M_l=6.9), reported an increase (of about 11 cm) in water level in three boreholes 110 km west of the epicenter, which started 90 min prior to earthquake motion and lasted about six hours. For the 18 November 1755 Cape Anne historic earthquake in New England (USA) with an epicentral intensity MM=VIII, Thorson (2001) reported hydrological responses up to 275 km from the epicenter, consisting in coseismic, abrupt, long-term changes in the flow rate and chemistry of water wells from five towns in Connecticut.

The 39 anomalies reported in Table 3 were evaluated to determine whether there were patterns of hydrologic change related to epicentral or fault distance. Figure 4 shows that: (a) most of the phenomena lie between 30–120 km from the epicenter, whereas the maximum distance was 200 km; (b) most hydrological changes occurred within 30–110 km from the fault rupture segment. The maximum distance of such variations from the fault rupture was 155 km. Note that few hydrological anomalies occurred near the fault or near the epicenter (<30 km).

Hydrological changes depend on both the structure of the aquifer and the strain that an earthquake induces on the area of the fault rupture.

Table 5. Excess discharge of the 23 July 1930 earthquake. Qav and Qtot are, respectively, the average and the total discharge in the entire anomalous period.

No.	Site	Anomalous period (days)	Qav (m^3/**d**)	Q tot (m^3**)**	Epicentral distance (km)
1	Aterno river at Molina	12	162174	1946084	179.0
2	Tasso river at Scanno	28	62610	1565255	154.0
3	Sagittario river at Capo Canale	6	317486	1269946	167.5
4	Pescara river at Maraone	9	1498499	13486494	200.0
6	Lavino river at Scafa	23	74608	1715990	175.0
7	Zittola river at Montenero	25	9480	236995	129.0
8	Rio Torto river at Alfedena	15	50574	758614	107.5
9	Trigno river at Trivento	13	134317	1746116	107.0
10	Biferno river at Guardialfiera	18	536461	9656293	96.0
12	Fortore river at Stretta di Occhito	11	202098	2223075	70.0
15	Ofanto river at Rocchetta S. Antonio	8	104047	832375	14.0

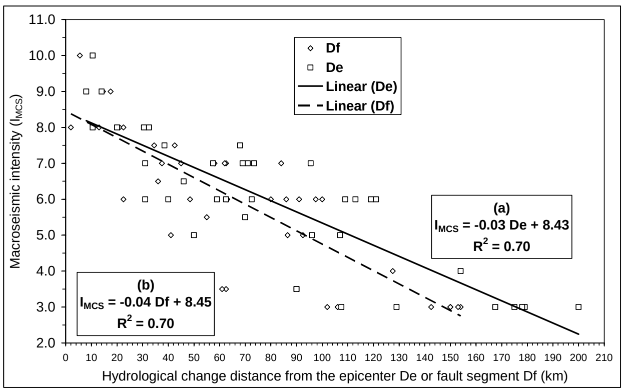

Fig. 4. Distance to epicenter **(a)** and to fault segment **(b)** versus intensity (I_{MCS}). A clear negative linear regression is visible.

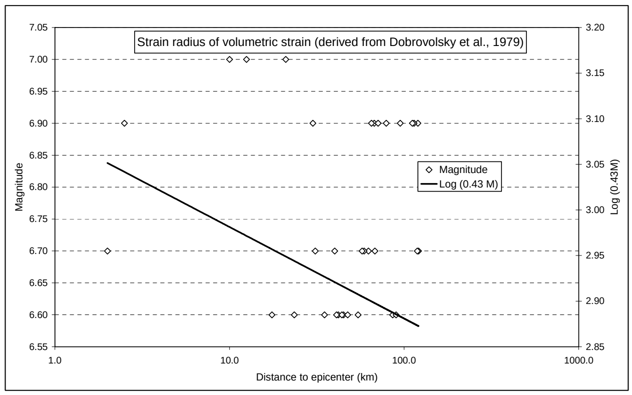

Fig. 5. Distance to epicenter for hydrological changes in wells versus magnitude of eight historical earthquakes (Table 1). The strain radius (Dobrovolsky et al., 1979) is reported (see Montgomery and Manga, 2003).

Dobrovolsky et al. (1979) give a theoretical relation regarding earthquake magnitude, distance from the epicenter and volumetric strain. The "strain radius" R_s of a circle centered on the epicenter, in which precursor deformations and other physical phenomena occur, is given by:

$$R_s = 10^{0.43\,M}, \text{ that is } R_s \approx e^M \tag{1}$$

This exponential curve divides the areas where strain is lower than 10^{-8} and greater than 10^{-8}. For strain=10^{-8} water level changes are only 1 cm. The data of some earthquakes in Irpinia (Table 1 in Onorati and Tranfaglia, 1994; Tables 2, 4 and 6 in Porfido et al., 2007) are plotted in Fig. 5 and the strain radius is indicated.

Montgomery and Manga (2003) suggest that the stream flow changes are attributable to liquefaction of valley bottom deposits. Papadopoulos and Lefkopoulos (1993) give an empirical maximum distance to the epicenter at which liquefaction can occur as a function of earthquake magnitude:

$$M = -0.44 + 3^{*}10^{-8} D_e + 0.98 \log D_e, \tag{2}$$

where D_e is the distance to the epicenter in cm. Figure 6 shows distance to epicenter for hydrological changes in rivers versus magnitude of Irpinia earthquakes in 1930, 1980 and 1984 (Porfido et al., 2007; Onorati and Tranfaglia, 1994). The liquefaction curve determined by the above relation is reported. Because many hydrological changes are at distances greater than the liquefaction curve determined for valley bottom deposits, they can be caused by preseismic fracturing of carbonatic aquifers in the Apennine Chain.

6 Conclusions

The study of the geochemical and hydrodynamic characteristics of aquifers is acknowledged to make a valid contribution to understanding the natural processes connected

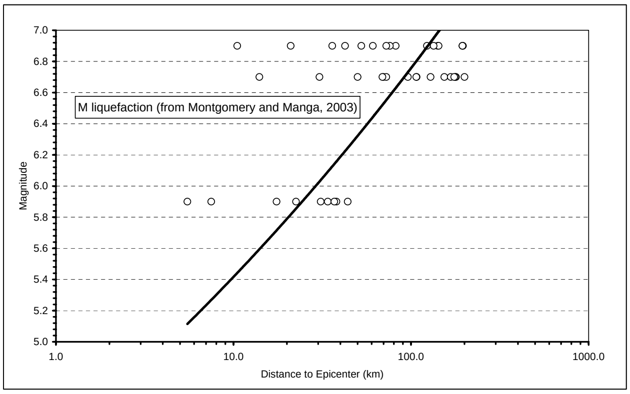

Fig. 6. Distance to epicenter for hydrological changes in rivers versus magnitude of some Irpinia earthquakes (Onorati and Tranfaglia, 1994; Porfido et al., 2007). The liquefaction curve (Papadopoulos and Lefkopoulos, 1993) is reported (see Montgomery and Manga, 2003).

to earthquakes (King et al., 1981, 1994; King, 1985; Bredehoeft et al., 1987; Roeloffs et al., 1989; Kissin and Grinevsky, 1990; Quilty and Roeloffs, 1997; Ingebritsen and Sanford, 1999). Changes in the water-rock interaction are caused by the seismic stresses in the area where the tectonic deformation leads to the seismic event (Rojstaczer and Wolf, 1992, 1994; Muir-Wood and King, 1993; Quilty and Roeloffs, 1997; Roeloffs, 1998; Ingebritsen and Sanford, 1999; Manga, 2001).

Various mechanisms have been invoked to explain earthquake-related changes in water tables and in spring and stream discharges:

1. Large (as much as 20 m), near-field (probably <50 km from the epicenter) water level declines can sometimes be related to near-surface permeability enhancement due to ground motion (Rojstaczer and Wolf, 1992, 1994). These authors limit the validity of the relationship between seismic intensity and areas with water increases only to normal fault earthquakes.

2. Muir-Wood and King (1993) proposed a model of deformation in complex fault systems with different mechanisms and orientations. In the model, the coseismic dislocations during strong earthquakes produce a deformation of the crust which directly influences the surficial aquifers. Furthermore, in the inter-seismic periods in areas undergoing crustal extension, there is an increase in pore volume, which is then filled by percolating fluids with a consequent decrease of levels in underground waters. After a normal-faulting earthquake, the stress release produces a decrease in pore volumes and hence an increase in water levels. This hydrological behaviour can be considered a precursor anomaly. In areas undergoing inter-seismic compression, the anomalous behaviour has the opposite sign.

3. Quilty and Roeloffs (1997) analyzed co-seismic changes in water level in nine wells near Parkfield, California, produced by an earthquake on 20 December 1994 (M=4.7), in order to test the hypothesis that coseismic water level changes (which for nine wells ranged from −16 to +34 cm) are proportional to coseismic volumetric strain.

4. According to Cooper et al. (1965), most of the coseismic water level oscillations observed at larger distances are resonance phenomena caused by particular fracture patterns of the formations where wells are located, that act to amplify a very small crustal strain signal.

The negative anomalies found in this work can be considered "rebound anomalies", which are the most common precursor reported by many authors and are related to increases in porosity and permeability caused by fracturing that precedes an earthquake (Roeloffs, 1988; Igarashi et al., 1992).

The total excess discharge (0.035 km^3) caused by the Irpinia earthquake (M_s=6.7) of 11 streams is comparable with the excess discharge of about 0.01 km^3 for the Loma Prieta earthquake (M_w=6.9).

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