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Hydrological triggering of the seismicity around a salt diapir in Castellane, France

Laurent Bollinger *, Marc Nicolas, Sylvie Marin

Département Analyse Surveillance Environnement, CEA, DAM, DIF, 91297 Arpajon, France

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ABSTRACT

The subalpine belts in Provence (France) present a low level of background seismicity. However, Castellane and some surrounding villages in the middle of one of the most renown belts, the Castellane arc, have been damaged twice in a century by significant earthquakes, in 1855 and 1951. The macroseismic database acquired after these events suggests that the sources were moderate ($M_L \sim 4.5$) but very shallow (~1 km). A recent instrumental microseismic catalogue further attests to the shallow seismicity in the area. A 58 year hydrological record of the Verdon river discharge allows us to test whether this seismicity is modulated by some transient local state-of-stress changes induced by large aquifer and artificial-lake loadings. We show that 41% of the extreme discharges (those with a probability of exceedance at 0.1%) are followed by at least one seismic event within a 7–28 day optimal interval and that this correlation, which represents 8% of the earthquakes, is not due to chance. We consider several natural and artificial hydrological mechanisms, including artificial reservoir and aquifer elastic loading and induced pore-pressure variations. We favour a model in which the local salt–gypsum domes respond to the aquifer forcing.

1. Introduction

Subalpine belts in Southeastern France present a low level of background seismicity (e.g. Nicolas et al. 1998). However, Castellane, in the Provence Alps, as well as its surrounding villages, Taulanne, Chasteuil, and Taloire, has been damaged twice by locally destructive seismic events, occuring within one hundred years, in 1855 and 1951, with similar inferred locations (Rothé and Dechevoy, 1967) (Fig. 1).

Apart from these two spectacular and renown events, generating field surveys by the most renown French seismologists of these periods (e.g. Perrey, 1856; Rothé and Dechevoy, 1967), the instrumental as well as historical seismicity in this subalpine region remains poorly investigated due to its paucity. Castellane was settled very early at least since ligurian tribes' presence in the region then grew during the roman period, partly from the presence of salted springs, surrounding the site, giving it its former roman name of salina or salinae. This rich historical inheritage should have helped keep a record of past catastrophic events which is unfortunately not the case.

In this paper, we investigate Castellane region's historical and instrumental seismicity. We first describe its tectonic and seismotectonic setting. We discuss the seismic event characteristics, both macroseismic and instrumental, leading to a determination of their magnitude and depth of occurrence. We then confront the time structure of the nearby Verdon river discharge to the occurrence of the seismic events. We finally describe local stress loading–unloading mechanisms encompassing local hydrological forcing potentially triggering these seismic events.

2. Regional tectonic and seismotectonic setting

Castellane has given its name to one of the most renown perialpine structures in Southeastern France, the Castellane arc. This arc is a fold-thrust belt of arcuate geometry extending for more than 50 km between the Digne Nappe complex to the North and the external folds South of Audibergue Plateau (Figs. 1 and 2). This fold-thrust belt has grown mainly in the Miocene. Its geometry appears largely controlled by a major decollement level which occurs in the middle to late Triassic (e.g. Goguel, 1936). Its kinematic growth encompasses SW-directed tectonic transports (e.g. Aubourg et al., 1999).

About 40 individual tectonic units have been surveyed and are incorporated in balanced cross sections through the fold-thrust belt (e.g. Ford et al., 1999). The local tectonic frame in Castellane appears therefore complex, with almost 6 slivers in the villages' vicinity (Fig. 2), each accommodating a shortening around 2 km (Laurent et al., 2000).

All of these tectonic units are partly delineated by Triassic evaporitic outcrops and underlined by cumulated Triassic given to reach frequently 400 m (Laurent et al., 2000), locally much more (Kerchkove and Roux, 1978). The presence of Triassic outcrops nearby dome structures of 1–2 km radius at Castellane, Brandis and Rougon (Fig. 2) suggest that some of these domes are underlined by salt–gypse diapirs (e.g. Guébhard, 1914; Kerchkove and Roux, 1978), relatively familiar in the southwestern subalpine belts (e.g. Mascle et al., 1988; Dardeau and de Graciansky, 1990). Unfortunately, to our

* Corresponding author.
E-mail address: laurent.bollinger@cea.fr (L. Bollinger).

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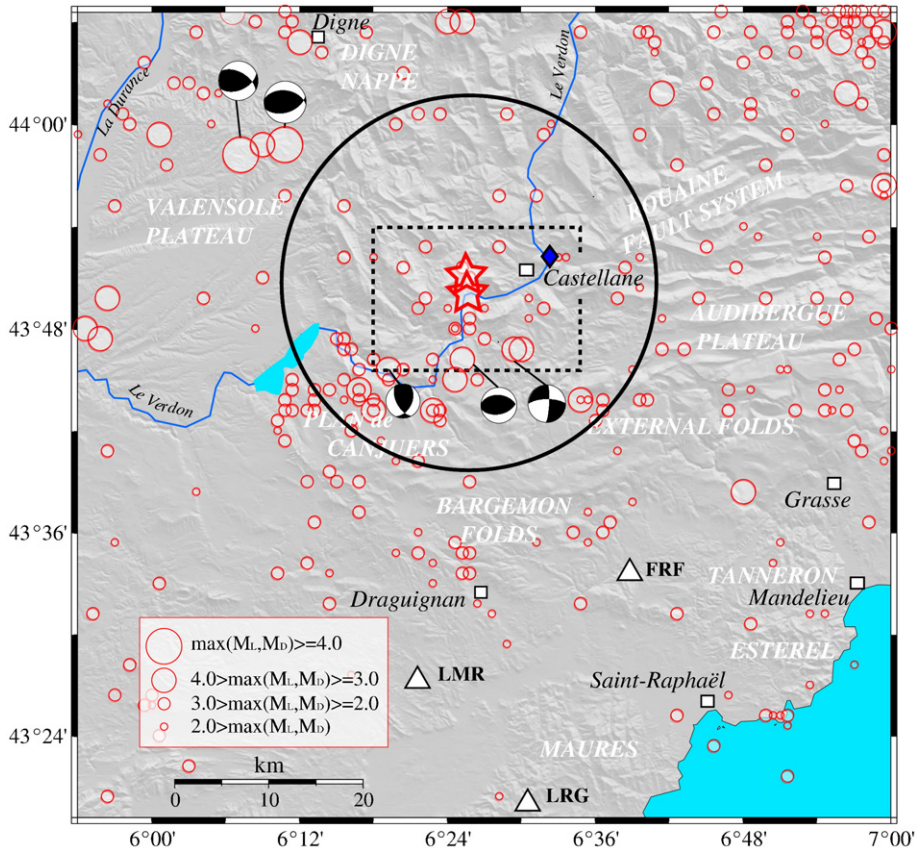

Fig. 1. Castellane in its regional frame. 1962–1993 instrumental seismicity from Nicolas et al. (1998) completed from 1993 to 2004 using the same methodology. Red stars correspond to the 12/12/1855 and 30/11/1951 macroseismic epicentres. Focal mechanisms from Nicolas et al. (1990), Madeddu et al. (1996) and Baroux et al. (2001)). White triangles refer to LDG seismic stations. Blue diamond materialises Demandolx hydrological station. Circle (radius 20 km) delineates the extent of the seismic catalogue tested.

knowledge, no gravimetrical data exist on these structures to demonstrate their salt–gypse diapirical nature.

Although the tectonic frame is described in detail in the literature, the late kinematics in the area remains largely unconstrained due to a lack of geologic records allowing age constrains on the tectonic activity of each sliver. Very few indices of neotectonic activity have been documented in the area (Neopal, 2007), but some suggest a Pleistocene tectonic activity at the front of the Dignes Nappes.

3. Historical seismic events around Castellane

SISFRANCE database, the most exhaustive intensity catalogue for historical earthquakes felt in France (Lambert et al., 1997; Sisfrance, 2007) documents more than 20 earthquakes felt in Castellane vicinity over the last 2 centuries. Overall, the macroseismic observations, all interpreted in MSK scale (Sponheuer and Karnik, 1964 in (Karnik, 1969)), come from both primary and secondary sources of variable quality. They include observations collected in newspaper articles or reports as well as, for the most recent earthquakes, from macroseismic forms sent to the local administrations and archived by Bureau Central Sismologique Français (BCSF). Among the seismic events evaluated in the database, two historical seismic sequences, culminating with destructive earthquakes and preceded by precursor events are documented.

The first seismic sequence begins with a shock that occurred on November the 23rd 1855 at 15h15. It caused cracks in walls in three (unidentified) villages in Castellane vicinity, triggered a large landslide and was noticed within 15 km. This event is given to be a foreshock, with an epicentral intensity VII, of the December 12th earthquake. On this day, at 20h40, a very damaging event, felt within 90 km, destructed nearly all the houses in Taulanne and Chasteuil. An epicentral intensity VIII and a location at Chasteuil were associated to this event. It was followed with almost 5 shocks also felt in the area until December 14th, increasing the alarm of the population. At 10 km from Castellane, it was said that a large crack formed, emanating sulfite rich vapors; the surrounding soil seemed hotter than elsewhere. The crack as well as the vapour emanation is attested by Perrey (1856) himself who went to the outcrop after the event. The distance between Castellane and this site, as well as the description of its location on a snowy mountain around Castellane suggests that this crack was probably somewhere near the crest, culminating at 1740 m, immediately North to Chasteuil–Villars–Brandis on Fig. 3.

Strikingly, given the very low seismicity rate in the whole region, on November 30th 1951 at 6h08, a violent shaking induced strong damages to the same villages. In Chasteuil, most of the houses were severely damaged, walls cracked and roofs collapsed. Villages within 5 km were strongly damaged. The departmental road D952 was cut by the collapse of a large mass of rocks between kilometres 6 and 9, partially damaging the retaining wall between kilometres 6.4 and 6.9 (Fig. 3). Turbid springs were mentioned in Talloire, Rougon as well as in Taulanne (Fig. 3). This seismic event came after two foreshocks on November 29th at 14 h as well as on November 30th at 4h00, and was followed by aftershocks on November 30th and December 10th. All these events were felt in Chasteuil, suggesting a small epicentral distance and/or presence of local site effects.

Among the damage report forms sent by the BCSF to 561 Southeastern France towns following the main seismic event, 479 were returned, including 123 felt observations (Rothé and Dechevoy, 1967). 111 were assigned an intensity greater than II (Sisfrance, 2007). No information being given in this database on the areas where

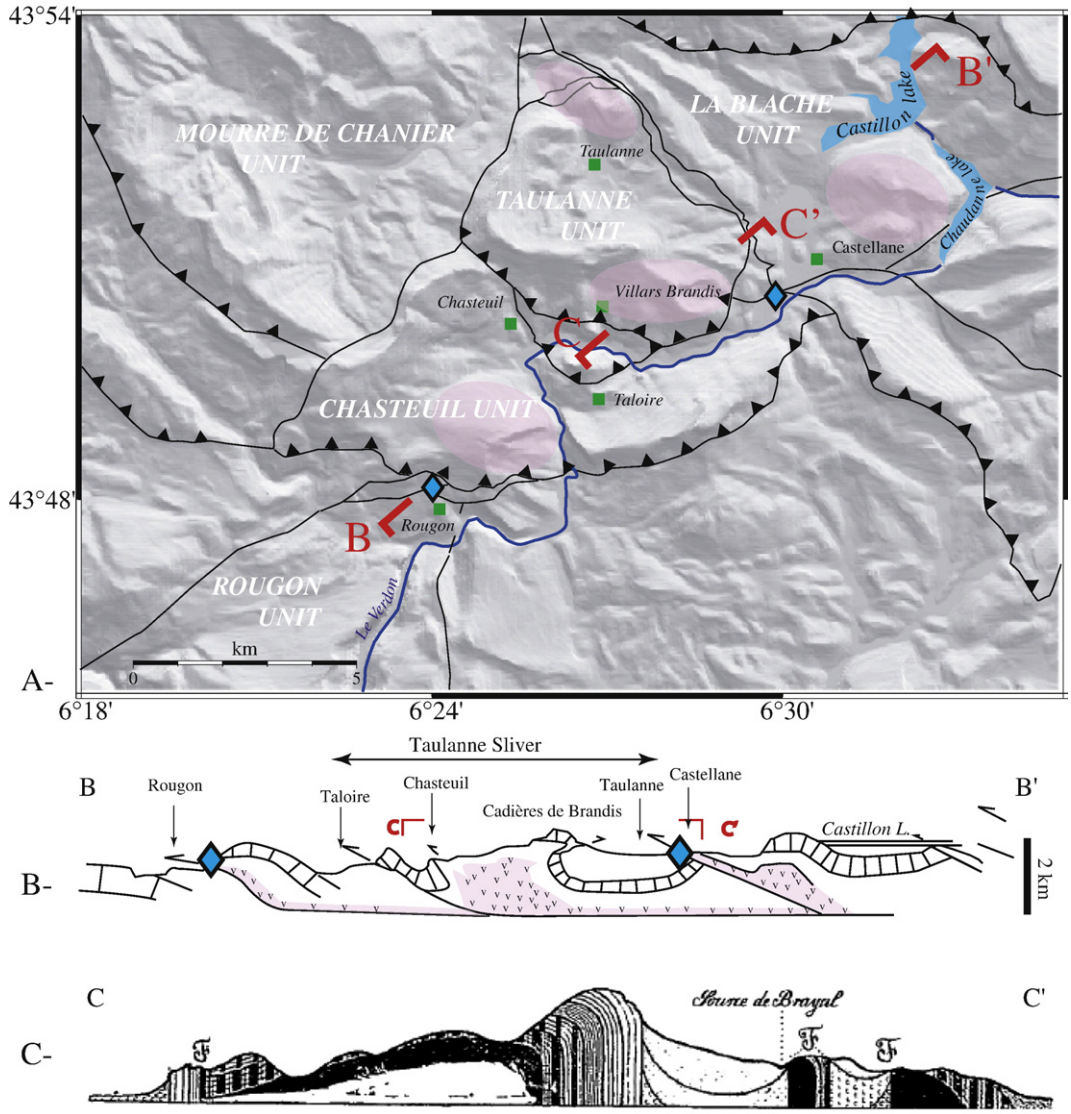

Fig. 2. Tectonic setting. A— Tectonic map generalised from Roux (1972), Kerchkove and Roux (1976) and Laurent et al. (2000). In pink, inferred subsurface location of salt–gypse diapirs and thick Triassic evaporitic deposits. C–C′ materialises the tectonic cross section presented in B—. Blue diamonds locate the salted springs. C— Geological cross section through the main salt dome from Guébhard (1914).

this event was unfelt, we completed the macroseismic database with 84 additional locations presented in Rothé and Dechevoy (1967).

Despite the high intensities (up to VII–VIII) reached within 5 km from the macroseismic epicentre, the radius of the intermediate intensities (i.e. IV–V) appears small. Furthermore, 20 km East and Westward and 30 km Northward, the main shock remained mostly unfelt. This observation is corroborated with the very sharp decrease in intensity within the first kilometres (Fig. 4). The number of the data available in the near field, as well as the geology underlying the most damaged villages (Fig. 2) prevents us to interpret the peak intensity as purely induced by site effects. On the contrary, macroseismic intensities available in the intermediate and far field appear widely distributed, ranging from I to IV–V at 30 km some of them being most likely correlated with site effects (e.g. at Draguignan and Toulon, Fig. 4).

The shape of the attenuation, very strong within the first 20 km, suggests a very shallow depth. We compare the 30/11/1951 observed intensities with predicted intensities for a 4 km deep event (corresponding to the mean focal depth in the Provence region) and a shallow event at 1 km using Marin et al. (2004) attenuation law (Fig. 4). According to our modelling, the reported intensities are best fitted for an $M_L = 4.5$ at 1 km than an $M_L = 5.1$ at 4 km in moderate to far field. A depth of 1 km or less is therefore most likely and the event may present a magnitude of around $M_L = 4.5$.

The damages and effects of 1855 main shock, although available sparsely (3 evaluations within 5 km, 3 additional ones within 80 km), appear similar with the 1951 main shock estimates (Fig. 4). We therefore suggest that the intensity attenuation as a function of the epicentral distance is similar to the 1951 event and that this event was also very superficial with a local magnitude around 4.5.

None of the other 18 seismic events catalogued in the SISFRANCE database within 20 km from Chasteuil over the last 2 centuries appear as destructive as the two 1855 and 1951 mainshocks. Furthermore, the sparse documentation available on these complementary events, some of them only felt at a single observation point, prevents for the moment any further interpretation.

4. Instrumental seismicity around Castellane

The instrumental catalogue considered here is derived from a combination of data sources from the ISC catalogue (originally from e.g. ETH, ING, ReNaSS and Sismalp seismological networks) and from

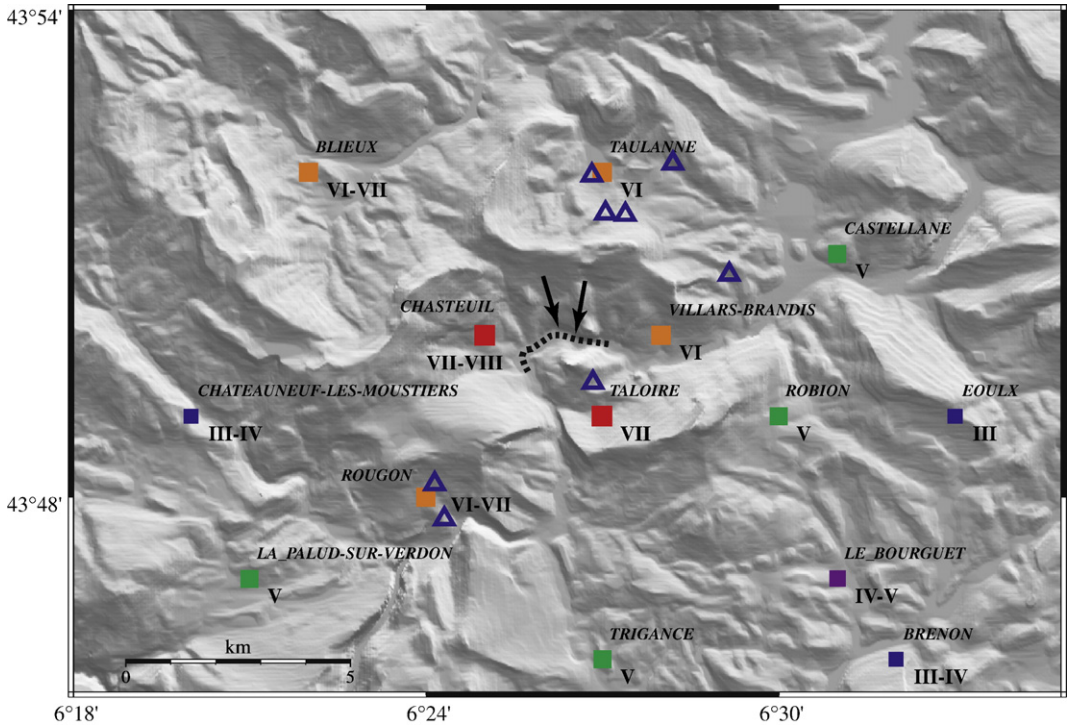

Fig. 3. Natural response, damage and MSK macroseismic near field observations induced by the 30 November 1951 seismic event. Figure and square colours refer to the MSK64 intensities (Sisfrance, 2007) deduced from BCSF archive (Rothé and Dechevoy, 1967). Dotted thick black line for D952 road section affected by rock collapse, black arrows corresponding to inferred landslide trajectories. Blue triangle materialises Taulanne, Taloire and Rougon springs, at least one spring in each village was turbid after the event.

CEA/LDG. This revised catalogue of the Western Alps initially covers 31 yrs of seismicity, from 1962 to 1993 (Nicolas et al., 1998). It is completed up to 2005 using the methodology, including location algorithms and crustal models, described in Nicolas et al. (1998). Due to several significant evolutions of the contributing seismological network geometries over the period considered (the number of operating stations in Western Europe over the last 50 yrs was multiplied by more than 7), the detection threshold for the subalpine belts in Provence considerably evolved with time. The first regional contributor to these changes is the CEA/LDG, after the setup of 5 stations along the French Riviera since 1962. These stations are part of a French network developed since 1957, but were only centralized and completely telemetered in 1977. Since then, SISMALP network improved the azimuthal coverage in the region, adding 44 stations mainly to the Northeast of the subalpine Provence belts after its actual geometry was reached, in 1994. These evolutions lead to a gradual decrease of the completeness magnitude of the instrumental catalogue from $M_L \sim 3.5$ between 1962 and 1977 down to $M_L \sim 2.5$

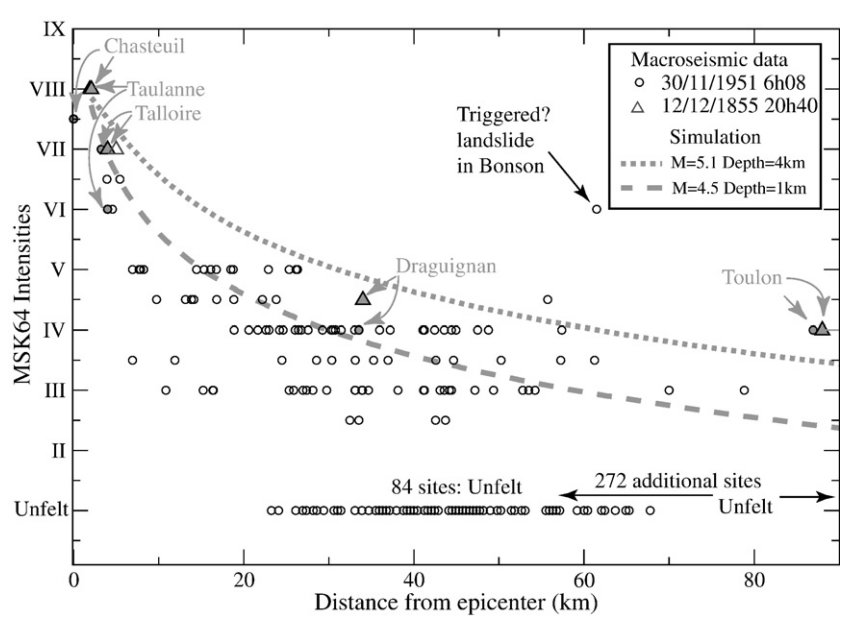

Fig. 4. Comparison of observed and predicted intensities for the 30/11/1951 and 12/12/1855 Chasteuil earthquakes. Observed MSK1964 intensities from Sisfrance (2007) and interpreted from BCSF archive (Rothé and Dechevoy, 1967). Predicted MSK1964 intensities for $M_L = 5.1$ (depth = 4 km) and $M_L = 4.5$ (depth = 1 km) events using Marin et al. (2004)) attenuation law.

since then (Marin et al., 2004). In the meantime, the average location uncertainties decreased significantly but they still remain greater than 1 km, according to the mislocation of nearby quarry blast events. The typical uncertainty of hypocentral depth is large, greater than 5 km for the pre-1987 events. However, although poorly constrained, the depths of the events in the region appear systematically superficial, within the first 8 km. Some of these events are even clearly located at very shallow depths, probably within the first 2 km as attested from the depletion of the high frequency part of their source (cf. Fig. 5C for example). This depletion is typical of seismic sources filtered by the complex surface geology, and is a usual quarry blast/natural event discriminant.

The scarcity of events and large location uncertainties prevent finding of spatial relationships between them (Fig. 1) or associating them with a particular tectonic structure (Fig. 2). The focal mechanisms of 3 of the largest events have been previously determined (Nicolas et al., 1990; Madeddu et al., 1996; Baroux et al., 2001) complementing the disparate and poorly constrained CMT dataset in a region where the state-of-stress is still a matter of debate (e.g. Baroux et al., 2001; Delacou et al., 2004; LePichon et al., 2007).

In the same way, the time structure appears insufficiently dense to describe any sequence and even to quantify the biases induced by the large time variations of the detection threshold.

Nevertheless, the three major shocks occurred in November while the seismicity appears to be mostly generated in late fall.

This observation, as well as the presence of nearby large artificial lakes potentially modulating the seismicity leads us to investigate the potential relationship between local hydrology and the seismicity.

5. A new case of hydrologically triggered seismicity?

5.1. Verdon river and Castillon–Chaudanne artificial lakes

The Verdon river flows from Allos region in the southern French Alps to Vinon-sur-Verdon on a 175 km course through several artificial lakes, including Castillon, Chaudanne, Sainte-Croix and Quinson. Near Castellane, the river drains a catchment basin around 700 km^2 (655 km^2 at Demandolx and 1820 km^2 at Vinon). Its natural discharge is low, on average 27.1 m^{3}s^{-1} at outlet in Vinon-sur-Verdon and 13 m^{3}s^{-1} at Castillon above Castellane (Hydrofrance, 2007). The 150,000,000 m^3 Castillon artificial reservoir water filling covering a surface of approximately 500 ha took place in 1948, after 19 yrs of intermittent works. The water outlet at the dam is 500 m^{3}s^{-1} with a flood discharge potential of 1200 m^{3}s^{-1}. The water goes through the downstream 27,000 m^3 Chaudanne lake filled in 1953 (cf. Fig. 2).

5.2. Hydrological records of the Verdon river at Demandolx

The hydrological record at Demandolx, on Castillon lake is available online under the HydroFrance database (referred hereafter as Hydrofrance (2007)), the most exhaustive hydrological database publicly available in France. The proximity of these available hydrological dataset is therefore a good opportunity to test whether or not the seismicity is modulated by the hydrology in this region. The time series of the daily discharge at Demandolx considered here covers 21,185 days between 1948 and 2005. We were unable to find or reconstruct any discharge prior to 1948 to cover the historical part of our seismicity catalogue from 1855. The mean as well as the median daily discharge is 13.0 m^{3}s^{-1} while the upper 1% of the daily discharge is 66.5 m^{3}s^{-1}. The mean monthly discharge depicts significant variations, with two maximums in May and November, respectively at 30.3 and 15.0 m^{3}s^{-1}, and two minimums in August and January, respectively at 4.9 and 8.3 m^{3}s^{-1}. The maximum daily discharge was measured on November 5th 1994 at 406.5 m^{3}s^{-1} while the maximum mean monthly discharge, 77.7 m^{3}s^{-1}, was reached in November 1951. Both these exceptional events, generating devastating floods downstream along the Verdon, show an occurrence probability around 1 over 100 yrs.

5.3. Testing the correlation between discharge series and time structure of the seismicity

Strikingly, the peaks of these 2 extreme discharge events, at 285 m^{3}s^{-1} on November 11th 1951 and 406.5 m^{3}s^{-1} on November

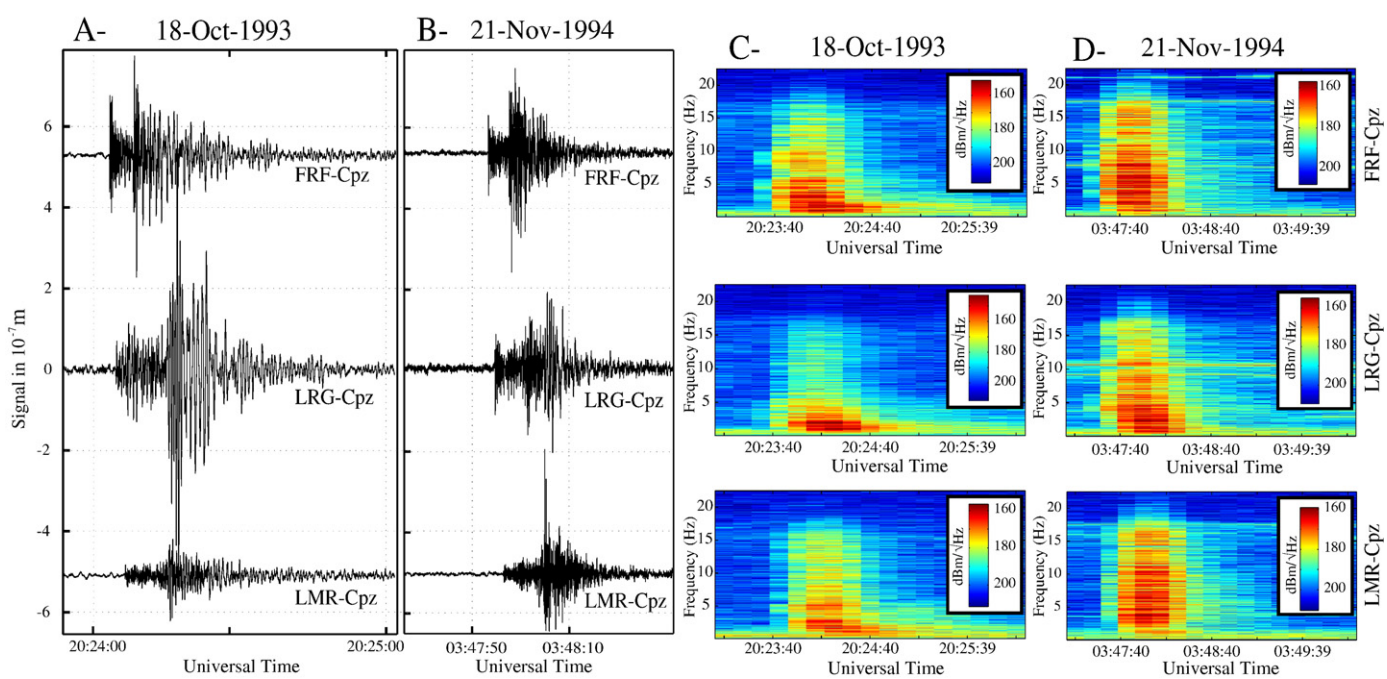

Fig. 5. An example of instrumental records: for the 18/10/1993, $M_L = 2.7$ (A−) and 21/11/1994, $M_L - 2.5$ (B−) events at FRF, LRG and LMR, the 3 nearest stations respectively at 35 ± 2, 42 ± 2 and 56 ± 2 km (cf. Fig. 1 for location). C− and D− are the respective Time–Frequency analysis of the two events. The high frequency depletion in C− suggests a very shallow origin for the 18/10/1993 event.

5th 1994, were reached respectively 18 and 16 days before the two largest instrumental events located in the region (Fig. 6). Furthermore, although not available in 1855, the Verdon river discharge is suspected to have been exceptionally high in October–November 1855. Major floods were affecting the lower Rhône valley from Arles to Beaucaire (with a Rhône river highest level of 4.8 m in Arles (Pichard, 1995) while the discharge reached 7550 $m^3 s^{-1}$ in Beaucaire – on 21/10/1855, one month before Castellane 1855th swarm – the 25th biggest daily discharge over the last 150 yrs). Lower Rhône tributaries, including the Durance–Verdon, are given to be important contributors to this flood (Champion, 1862). Furthermore, several other extreme discharge events are followed by earthquakes within the same delays (Fig. 6 and Table S1, Fig. S2 in supplementary data material).

To test whether or not any correlation could exist between the local hydrology and the seismicity, we first compare the hydrological database and the seismic catalogue. This catalogue results from the combination between the macroseismic and instrumental catalogue. Typical uncertainty on the macroseismic epicentre or instrumental event location in the area being large (between 1 and 5 km), hydrological forcing potentially affecting a broad region (5–10 km), we select every event within 20 km from Chasteuil in the seismic catalogue. As a consequence, a total of 97 earthquakes, including 20 macroseismic events are selected. To circumvent contamination from aftershocks we suppressed their effects in our catalogue using the declustering method proposed by Reasenberg (1985). We adjusted the spatial (5 km horizontal, 10 km vertical, and inter-event separation $\leq$20 km) and temporal ($1 \leq T \leq 10$ days) input parameters to exclude events whose sequential occurrence occurred with $P = 0.95$ confidence in the catalogue. This depleted catalogue contains 89 main shocks.

We then restricted our test between 1948 and 2005, corresponding to the period covered by the continuous hydrological time series at Demandolx, a few kilometres upstream along the Verdon river. As detailed in the instrumental seismicity section, the completeness of the earthquake catalogue dramatically changed during that period. However, a high cut off in magnitude should prevent any statistically significant test between discharges and seismic events. Furthermore, testing the correlation between decadal floods and earthquakes requires a catalogue covering several decades of events, preventing us to test the correlation on the very last years of

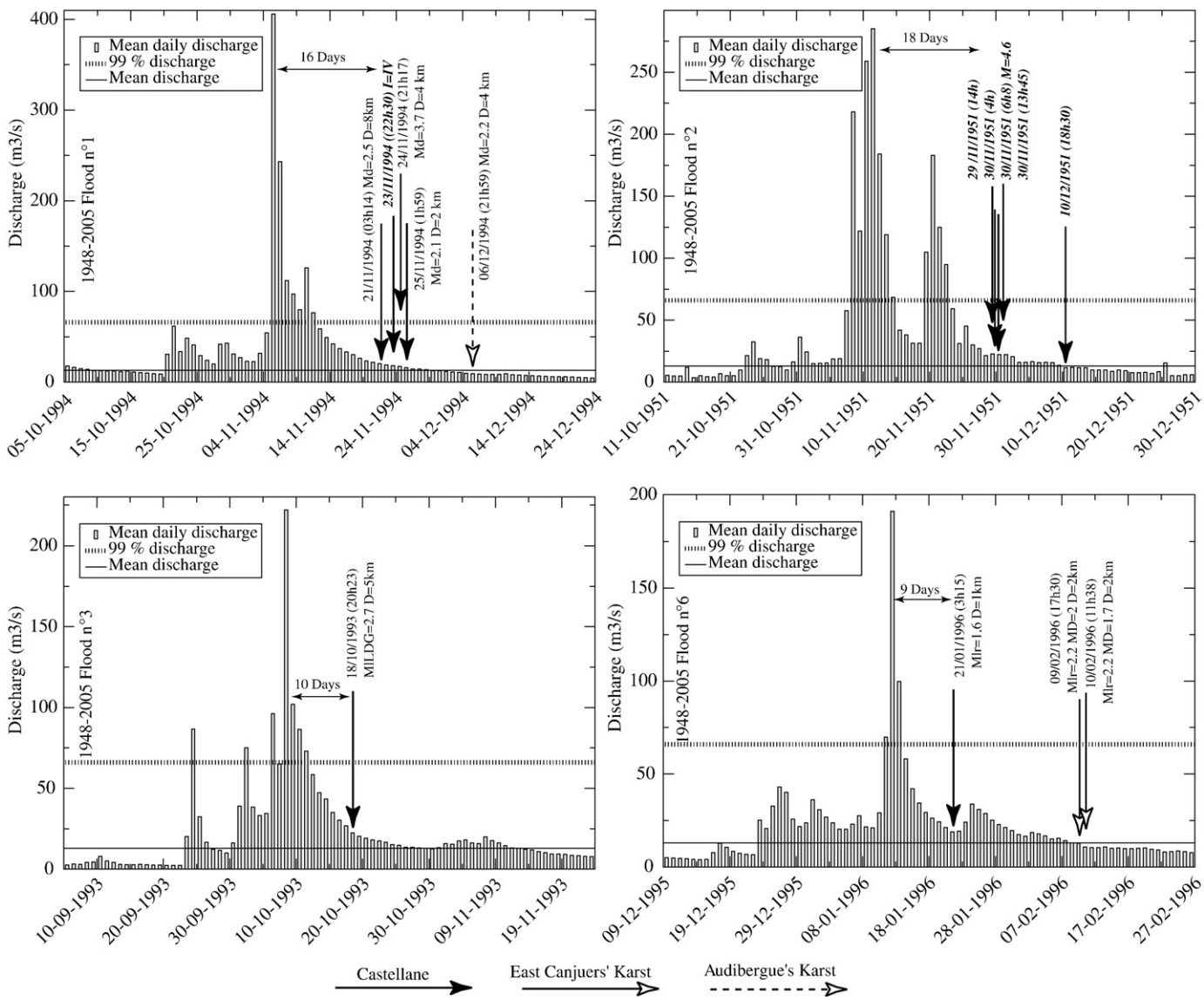

Fig. 6. Time structure of the mean daily Verdon river discharge at Demandolx: (Hydrofrance, 2007) during major floods (cumulated discharge events corresponding to centennial to decadal floods rank 1–2–3 and 6 within the 58 yrs discharge series considered (Table S1 in supplementary data material)) followed within 30 days by seismic events. Note that none of these hydrological events were preceded by seismic events within months. Some additional extreme discharge events followed by seismicity within the same delays are presented in Fig. S2 and Table S1.

monitoring. We therefore decided to test the correlations between this incomplete catalogue, biased by very long term variations of completeness magnitude, and the complete daily discharge time series at high frequencies (periods from days to months). The incompleteness of the earthquake catalogue may distort the statistical analysis and the observed correlation coefficients, most probably minimizing the correlation.

Despite the incompleteness of the earthquake catalogue, a first confrontation between the complete discharge time series and the seismicity turns out to be enigmatic. The 10-day moving average discharge rate histogram for 1948–2005 at Demandolx appears highly asymmetrical, enhancing the very low probability of events over 50 $m^3 s^{-1}$ (Fig. 7A). Assuming that the seismic event is equi-probable in time, the probability of occurrence for a seismic event during a discharge greater than this value is less than 2% while 10 $m^3 s^{-1}$ events should represent nearly 50% (Fig. 7A). Assuming that the seismic events postdate by 10 days the 10-day moving average discharge highs, we shift-forward the smoothed discharge series by 10 days and report the discharge corresponding to each earthquake occurrence. More than 50% of the seismic events correspond to discharges greater than 50 $m^3 s^{-1}$ (Fig. 7). This distribution of the number of earthquakes for a given discharge appears strikingly different from the one generated using a 10,000 event purely random seismic catalogue (Fig. 7A), suggesting a correlation between high daily discharge and seismicity.

With the seismic catalogue being small, this a priori correlation might be due to chance, and the strong a priori hypothesis used (i.e. seismic events 10 days after the 10-day averaged discharge highs) appears arbitrary. We therefore systematically assessed correlations between extreme discharge series and the seismic catalogue. We take into account two parameters in the measure of the correlation between the discharge time series and the earthquake occurrence: the time interval during which an induced seismic event is given to occur (corresponding to a potential "realisation" time) Δt, and a time gap, τ, corresponding to the delay necessary for the "system loading", delay between the date of occurrence of an extreme discharge event and the beginning of the time interval Δt. The partial correlation associated to a given discharge is 1 if at least an event occurs during $[t_{\text{discharge}}+\tau; t_{\text{discharge}}+\tau+\Delta t]$ and to 0 otherwise. The total correlation is the sum of the partial correlations divided by the number of extreme discharges in the time series considered. Maximum correlation for $\tau=[0;15]$ days and $\Delta t=[0;30]$ days for various extreme discharge thresholds is illustrated in Table 1. We restrain first our analysis to the upper 0.1% of the discharge time series, corresponding to the largest 19 discharge events, all greater than 148.5 $m^3 s^{-1}$, larger thresholds in terms of probability of exceedance corresponding to a number of events too limited. This first correlation analysis appears to be partly biased, due to multiple exceedances of the extreme discharge threshold considered during a single flood (Table S1). An independent flood time series is difficult to establish properly due to the sparse hydrological parameters available to us. To overcome this problem we created independent extreme discharge catalogues after depletion from the original time series of all but the highest discharge exceedance. This depletion is performed over one month sliding windows, this time period being larger than any of the individual flood events. The upper 0.1% of the independent discharges corresponds to the events greater than 147 $m^3 s^{-1}$. These independent extreme discharges no more belong to the same flood events (Table S1). This catalogue is therefore selected to further statistical tests. Maximum correlation for $\tau=[0;15]$ days and $\Delta t=[0;30]$ days is 0.41. The correlation sensitivity to τ and Δt is illustrated in Fig. 8. Most logically, the correlation increases with the time interval Δt considered. The "optimal solution" minimizing both τ and Δt while maximising the correlation is $\tau^*=6$ days and $\Delta t^*=22$ days with 41% of the independent extreme discharges greater than 147 $m^3 s^{-1}$ followed by an earthquake in this optimal 7–28 day time period.

To test whether this correlation is significant or due to chance we generate randomized earthquake catalogues encompassing the time structure observed. To randomize the occurrence time of earthquake in a catalogue with respect to extreme discharge series, the catalogue is first translated by 10 day steps, a total of more than 2100 catalogue correlations being tested for each (Δt; τ). We also generate purely random catalogues. Since results of the tests performed using both approaches are almost identical, we only present here the results of the tests using purely random catalogues. The decision rule is based on the significance level α, defined as the probability to reject the hypothesis of a correlation due to chance, when it is actually true. Usually, the significance level is chosen to be $\alpha=0.05$. For $\tau^*=6$ days and $\Delta t^*=22$ days, the probability of wrongly rejecting the null hypothesis is <0.05 (Fig. 8B, and Fig. S3 in supplementary data

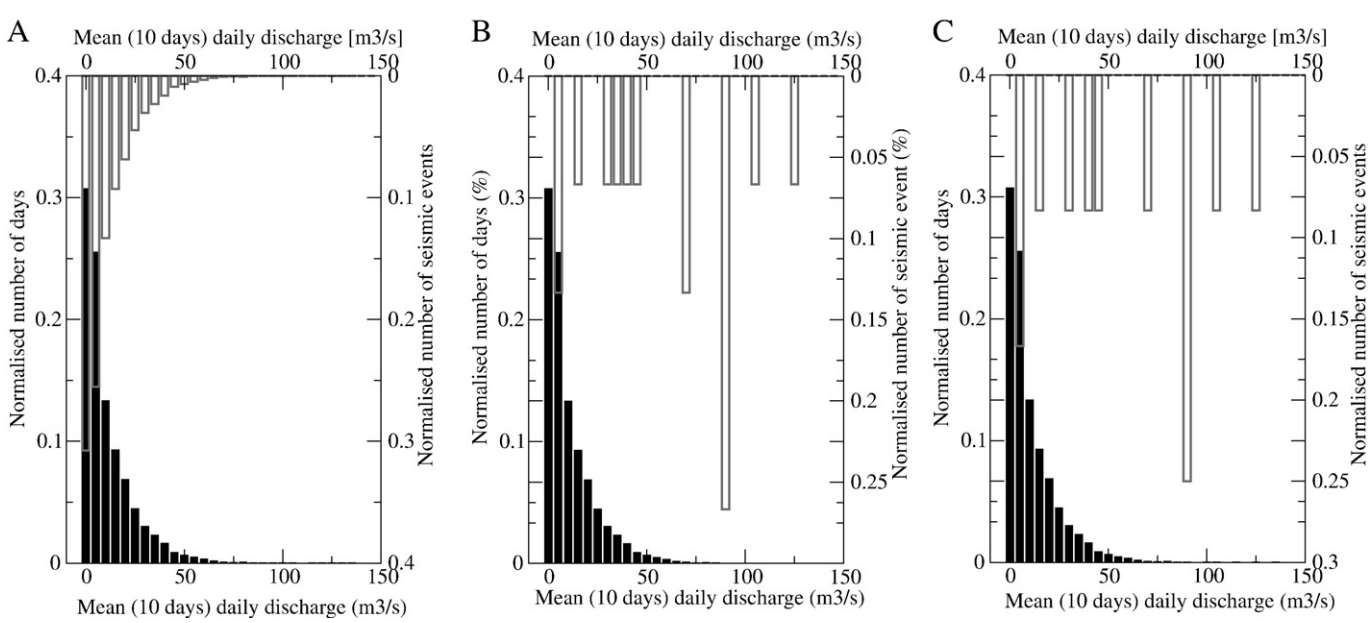

Fig. 7. Seismic event occurrences within 10 km from Chasteuil for a given river discharge. 10 day mean discharge rates distribution at Demandolx (black histogram) versus A— a random seismic event catalogue B— the seismic catalogue backward shifted by 10 days and C— the aftershock declustered and 10 day backward shifted seismic catalogue.

Table 1
Maximum correlation between 1948–2005 'extreme' discharges series (>66.5, 148.5, 183.5 and 259.4 $m^3 s^{-1}$) and the seismicity within the window $\tau = [0;15]$ days and $\Delta t = [0;30]$ days. The correlation analysis performed on independent extreme discharge catalogues, depleted from multiple exceedances of the discharge threshold considered during a single flood leads to slightly different results, with 0.1% of the largest independent discharges corresponding to 147 $m^3 s^{-1}$ and a maximum correlation at 0.41 (cf. text for further details).

Discharge threshold ($m^3 s^{-1}$)	Probability of exceedance	Number of discharges	Maximum correlation observed
66.5	1%	185	0.23
148.5	0.1%	19	0.53
183.5	0.05%	10	0.80
259.4	0.01%	2	1.0

material for the distribution of the correlation with purely random catalogues for $\tau^* = 6$ days and $\Delta t^* = 22$ days). We therefore accept the hypothesis that the 0.41 correlation is not due to chance for $\tau^* = 6$ days and $\Delta t^* = 22$ days.

Moreover, for $\tau^* = 6$ days the probability of wrongly rejecting the null hypothesis being lower than 0.05 when Δt^* is [0;30] days, the correlation is not due to chance whatever time interval between 0 and 30 days is considered.

To check whether the time gap $\tau^* = 6$ days might be significant, we test $\tau^* = 0$ day for $\Delta t = [0;30]$ days (cf. Fig. S4 in supplementary data material). The probability of wrongly rejecting the null hypothesis reaches values lower than 0.05 only for $\Delta t^* = [7;30]$ days. We cannot reject the hypothesis that the correlation is due to chance for $\Delta t =$ [0;6] days, the explanation for this is the low correlation values for this period close to the values of random series.

We therefore show that 41% of the independent extreme discharges (>147 $m^3 s^{-1}$) are followed by at least one seismic event within a 7–28 day optimal interval and that this correlation which represents 8% of the earthquakes on the period 1948–2005 is not due to chance.

6. Discussion–conclusion

We report evidence for the occurrence of seismic events after the major floods of the Verdon river nearby Castellane. We propose that these seismic events are triggered by a hydrological forcing.

The first idea is that the filling and overflowing of the Castillon and Chaudanne artificial lakes trigger or induce superficial seismicity in the region. Artificial reservoir induced seismicity is indeed a well known contributor of local seismicity nearby high dams, 90 of them being identified worldwide as potentially inducing or triggering earthquakes (e.g. Gupta, 2002, Talwani et al., 2007). Processes of stress modification in these settings have been suggested as the mechanisms responsible for earthquake triggering by large reservoir including elastic stress changes induced by the direct effect of loading and pore-pressure variations (with an identification of fluid flows along individualised faults for some of them). Unfortunately, only a couple of these reservoirs are adequately instrumented for seismic surveillance, it is therefore difficult to determine in which proportion and through which physical mechanisms the seismicity detected nearby is induced.

In France, only 2 artificial water reservoirs induced seismicity cases were suggested. The first one was proposed for Vouglans (Jura), following a magnitude 4.4 event located nearby the artificial reservoir 3 yrs after the filling of the lake (Rothé, 1970). The lack of further description of any seismic sequence following the filling as well as the absence of any constrained physical mechanism makes this correlation unconstrained. The second one, at Monteynard (Vercors massif), appears statistically more robust after an unusual seismic sequence following the 1961 filling of the lake, encompassing 20 shallow events of $M_L >= 3.5$, felt in a region where no seismic activity was reported for five centuries (Grasso et al., 1992).

Despite the presence of high strain zones under Castillon lake, including highly fractured and faulted sequences (e.g. Chenevas-Paule, 1987), as well as significant seasonal lakes oscillations, this scenario appears inadequate to account for all the seismic events

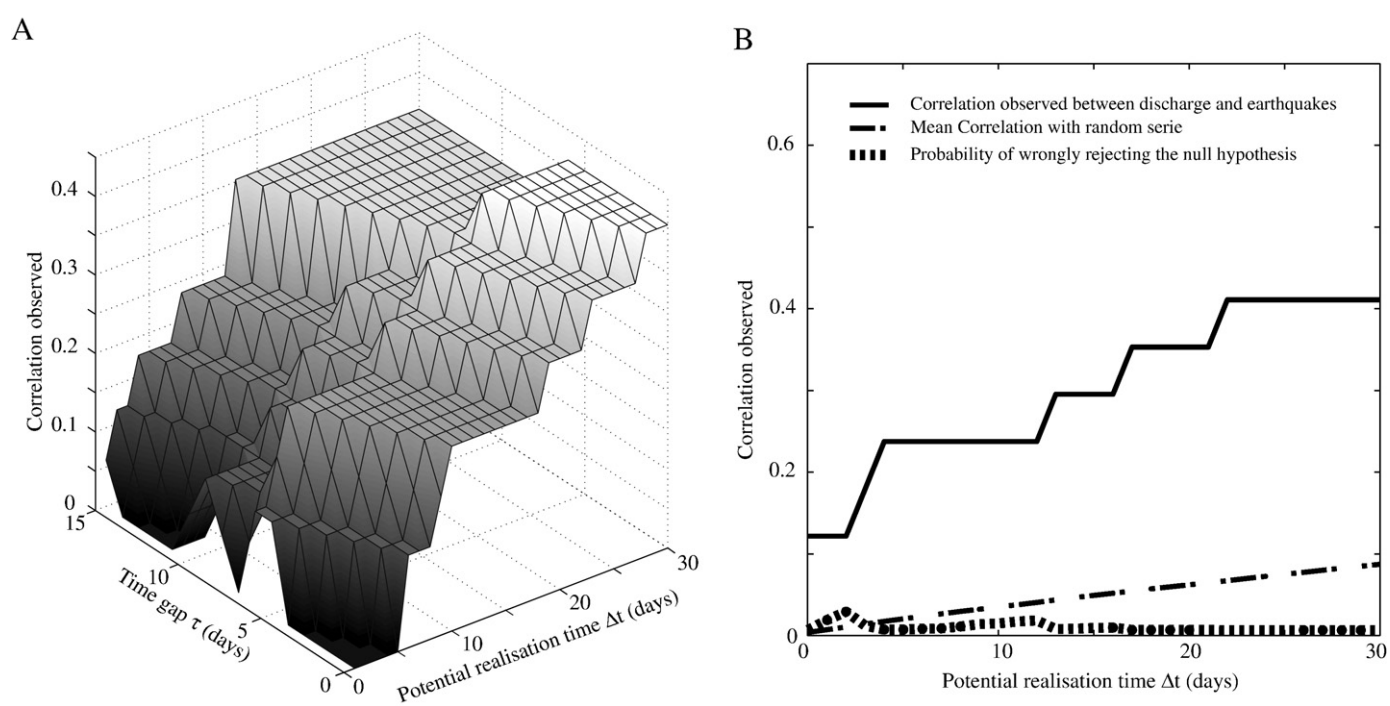

Fig. 8. Correlation between extreme discharge series and the seismicity: A— Correlation between the independent extreme discharges (>147 $m^3 s^{-1}$) and the seismicity as a function of a time gap, τ, corresponding to the delay necessary to the system loading and the time interval Δt during which an induced seismic event is given to occur or potential realisation time. B— Same correlation for time gap $\tau^* = 6$ days (solid line) compared to the mean correlation between extreme river discharges and purely random earthquake catalogue (dashed–dotted line, see text for detail) associated with the probability of wrongly rejecting the null hypothesis (dashed line).

triggered nearby. It cannot explain the November 1855 cluster, following within a month a centennial flood event, at a time when no dam nor lake was there. Furthermore, with no background micro-seismicity being available pre-1948, corresponding to the filling of the lake, this scenario involving artificial forcing appears difficult to test.

Significant natural hydrological modulations of the stress field have been suggested in several contexts to explain modulations of the number of seismic events detected along rivers (Costain and Bollinger, 1991; Roth et al., 1992) or karst (Rigo et al., 2008), within intraplate contexts (Costain, 2008), tectonically active regions (e.g. Wolf et al., 1997; Noir et al., 1997; Ohtake and Nakara, 1999; Heki, 2003; Husen et al., 2007) or locally along major faults such as the Main Himalayan Thrust and the San Andreas fault (Christiansen et al., 2007; Bollinger et al., 2007; Bettinelli et al., 2008). At smaller scale, the use of local seismological networks monitoring the micro-seismicity as well as dense streamflow time series further allowed assessment of the pore-pressure variations at volcanoes (Jimenez and Garcia-Fernandez, 2000; Saar and Manga, 2003; Christiansen et al., 2006). Mt. Hochstaufen (Bavaria) and its meteorologically induced, 10 days delayed, shallow seismicity (Hainzl et al., 2006; Kraft et al., 2006) is nevertheless probably a better analogue to Castellane. With its peri-alpine situation, decollement in the remarkably thick salt sequences, fractured and karstified limestone and evaporitic permo-triassic outcrops as well as evidences for mass movements, the geological environment at Mt. Hochstaufen (cf. detail in Kraft et al., 2006) is indeed very similar to Castellane. In both environments, the seismogenesis could therefore be linked to pore-pressure, Karst formation and salt dissolution. At Mt. Hochstaufen, Kraft et al. (2006) carried out a seismo-meteorological study examining the migration of hypocentres with pore-pressure diffusion models. They demonstrated that pore-pressure diffusion was also a possible forcing mechanism in this context using more than 1000 microearthquakes recorded by a local seismological network. Castellane micro-seismicity time structure and location is not known with sufficient details to properly constrain a diffusion model. However, assuming a mean event depth around 1 km, and considering a time(t)–depth(z) dependence of the triggering pore-pressure front with $z = \sqrt{4\pi Dt}$ (following Kraft et al. (2006)), with hydraulic diffusivities D around 1–10 $m^2 s^{-1}$ appear compatible with the generation of seismic events 10–20 days after the hydrological event. These hydraulic diffusivities appear within the 0.1–10.0 $m^2 s^{-1}$ range obtained from analysis of seismicity patterns by pore-pressure diffusion modelling (e.g. Roeloffs, 1988; Saar and Manga, 2003; Kraft et al., 2006, Husen et al., 2007, Talwani et al., 2007, Rojstaczer et al., 2008).

With fluid arrivals down to depths around 0.5–2 km, where massive salt formations are inferred, salt dissolution processes generating cavities are suspected. These cavity creations may generate salt dome response and/or collapses (although cavity collapses appear inconsistent with the double couple focal mechanisms determined for three of the events) and mass movements.

Whereas some of these plausible hydrological forcing mechanisms on the geological setting generate unique seismic time structures, or particular seismic spatial patterns, their respective contribution to the triggering of the seismic events emphasized here cannot be resolved further due to lack of available seismic data. Any further investigation will require dense seismic network data acquisition as well as complementary hydrological studies of the salt and turbid springs described nearby.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl.2009.11.051.

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