WHAT IS THE LOWEST MAGNITUDE THRESHOLD AT WHICH AN EARTHQUAKE CAN BE FELT OR HEARD, OR OBJECTS THROWN INTO THE AIR?

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ABSTRACT

This article is a reflection on effects produced by earthquakes at both ends of intensity scales: II ('Scarcely felt') and XII ('Completely devastating').

Now that most seismic regions—at least in developed countries—are monitored by seismic networks with magnitude thresholds close to magnitude 1, less attention is paid to reports of abnormal phenomena such as vibrations or noises. The alleged reason is that, if the event has not been detected by monitoring networks, there was no event at all. This point of view is discussed in the light of recent examples in South-East France, where tectonic earthquakes with a very shallow focus (sometimes only 300-m deep) can be heard and felt, whereas the nearby (less than 20 km) seismic stations could not record the events. Our study concludes that events with a magnitude smaller than 1, and even negative magnitudes, can be felt, thus making the human being an instrument eventually much more sensitive than monitoring networks.

Another type of remarkable observation which has been reported during earthquakes is the upthrow of objects into the air. Such observations are evidence of ground acceleration exceeding gravity. Although this type of observation is associated with an intensity of XII on the modified Mercalli intensity scale, we show that earthquakes of magnitude as low as 6 can produce such effects.

INTRODUCTION

The question of the lowest magnitude threshold at which an earthquake can be felt or heard is of particular importance when small historical events are used for delineating active zones in moderately seismic areas. The answer provided by most encyclopaedia and earth-science primers is that earthquakes are usually felt for shocks with magnitudes 3 and above. Actually, most authors of seismology textbooks are reluctant to tackle the question. Although Richter (1958) clearly states that 'the smallest shocks reported felt by persons are near magnitude 2', he does not expatiate on key parameters such as focal depth or population density.

Samuel Johnson (Boswell 1791) had a poor opinion on the accuracy and usefulness of popular reactions after an earthquake. Upon Boswell's reporting to him a small earthquake which had just happened in Staffordshire (England), he replied: 'Sir, it will be much exaggerated in popular talk: for, in the first place, the common people do not accurately adapt their thoughts to the objects; nor, secondly, do they accurately adapt their words to their thoughts: they do not mean to lie; but, taking no pains to be exact, they give you very false accounts. A great part of their language is proverbial. If any thing rocks at all, they say *it rocks like a cradle*; and in this way they go on.' This peremptory, extreme, although clever statement is an early (14 Sept. 1777) critical analysis of earthquake descriptions by lay persons. Fortunately, seismologists have long since reconsidered this viewpoint and, using appropriate precautions, now value such accounts.

Browsing Web pages can supply a wealth of information on felt earthquakes as shown for instance by the Community Internet Intensity Map developed by Wald (2006) at USGS, but low-magnitude events are rarely included in such lists because persons experiencing a faint rattle seldom bother to report it. If they ever do, the information is often judged insignificant and not deserving publication. However, out of the many Web sites providing information on felt earthquakes, the Australian Seismology Research Centre (http://www.seis.com.au) is one of the few to list carefully small events felt in Australia. Over the last seven years, the smallest magnitude value they report is an M_L (Richter local magnitude) 1.3 earthquake felt in 2000 in the suburbs of Melbourne.

There are good reasons to believe that this magnitude threshold can be still lower. Feeling small-magnitude shocks is perhaps not that unusual, the main problem being only how to collect this kind of information. Small earthquakes which occur in mines when the upper soil layers are depleted are often reported heard because they emit acoustic energy in the 200–1,000-Hz frequency range. Audible acoustic waves in the 50–70-Hz range have also been reported for many tectonic earthquakes (e.g. Hill et al. 1976; Tosi et al. 2000). Sylvander and Mogos (2005) analysed a macroseismic regional database which contains detailed reports of sounds heard for $M_L < 4$ earthquakes. They demonstrate that, in the Pyrenees, 'events with M_L as low as 1.0 (and perhaps even smaller) may be perceived under very favourable conditions'.

We will not discuss here the now-recognized audibility of small shocks, but rather address the question of repetitive occurrence of earthquakes, another factor which increases the sensitivity of the population. Long aftershock series or swarm earthquakes often further a flow of information, even though the phenomena are faintly felt or heard. We present two cases of low-energy, unusually-shallow seismic activity reported felt in 2002–3 and 2006 in South-East France. Records obtained at temporary stations only tens of metres from epicentres demonstrate that, under particular circumstances, even negative magnitude values can be associated with felt events.

At the other end of the gamut of effects produced by earthquakes, the upthrow of objects was thought for a long time to be an exceptional event encountered in great earthquakes only. The first such documented account was made by Oldham from field observations following the great Assam earthquake of 1897. Oldham reported that in some areas stones had been tossed in the air 'like peas on a drum' (Oldham 1899; Bolt and Hansen 1977).

The magnitude of the great Assam earthquake is estimated to have been close to 8.1 (Ambraseys and Bilham 2003). Reflecting the view that the upthrow of objects in earthquakes is exceptional, 'Objects thrown in the air' are listed as evidence of intensity XII on the modified Mercalli intensity scale. In this article, we will discuss observations of upthrown rocks and boulders produced by earthquakes with magnitudes much smaller than 8.

IN QUEST OF SMALL FELT EVENTS IN SOUTH-EAST FRANCE

Since the Sismalp monitoring network run by the Grenoble Observatory was set up in the 1980s (Thouvenot et al. 1990; Thouvenot and Fréchet 2006), the original procedure proposed by Richter (1935) has been used to compute the local magnitude M_L of earthquakes: the velocity seismogram is first integrated; the magnification value of the Mark Product L4C or L4C-3D 1-Hz sensors and the field recording gain are then taken into account to compare the displacement seismogram to the signal that would have been recorded by a Wood–Anderson torsion pendulum (Fréchet and Thouvenot 2000). In this stage, we use the 2,800 magnification value given for the Wood–Anderson. Uhrhammer and Collins (1990) found out that this value had been calculated on the basis of wrong assumptions on the suspension geometry, and a more correct value would be 2,080. We might therefore underestimate the size of events by 0.13 (Bormann et al. 2002), but we have not introduced this correction in the present study. We use the same attenuation law as that used by Richter although this law has been established for California. However Kradolfer and Mayer-Rosa (1988) analysed a set of earthquakes in and around Switzerland, and concluded that Richter's law was also suitable for the western Alps. Magnitudes computed by Sismalp and the Swiss Seismological Service usually differ by less than 0.2.

A Gutenberg–Richter's (1956) analysis of the 11,777 earthquakes located by Sismalp in the western Alps between 1989 and 2005 shows that events with a magnitude larger than ~ 1.3 can be confidently located (Marsan et al. 2007). Out of those 11,777 events, 725 (43 per year) have a magnitude larger than 2. If we follow Richter in his vague 1958 assumption, these events could be felt. We have checked this since 1996 by directly appealing to testimonies for most $M_L > 2$ earthquakes that occurred in the French Alps, instead of letting information reach us. This was done mainly through telephone calls to gendarmeries, municipal services, and hotels. In recent years, Internet accounts spontaneously sent to us made this quest dispensable. Out of the 128 $M_L > 2$ earthquakes we checked, 123 (96%) were felt. The five events that were not reported felt had magnitudes between 2.0 and 2.3; they either occurred in remote mountainous areas or had a focus deeper than ~ 10 km. Although common farther east in Italy, such 'deep' earthquakes seldom occur in the French Alps, where the seismogenic zone is mostly restricted to the first 10 km of the crust (Thouvenot and Fréchet 2006).

There is also fair evidence that protracted aftershock series favour the perception of still smaller magnitudes. We have in mind two recent destructive earthquakes, viz. the $M_L = 5.3$ 1996 Annecy earthquake, and the $M_L = 3.5$ 1999 Laffrey earthquake (Fig. 1). The Annecy earthquake (maximum MSK intensity VII–VIII) had its epicentre in the NW suburbs of the prefecture town of Haute-Savoie. Its focus was shallow (~ 2 km), within the Mesozoic sedimentary cover. The densely-inhabited epicentral zone was formerly a marsh area whose loose sediments amplified ground acceleration by a factor close to 10 in the 1–10-Hz frequency range (Thouvenot et al. 1998). The strike-slip mainshock generated aftershocks for more than 3 years, a much longer span than what could be anticipated for a 5.3 magnitude. Many aftershocks were locally felt that were recorded only by a temporary station maintained in operation at the epicentre. Since our Gutenberg–Richter's analysis shows that all $M_L > 1.3$ events can be located, we conclude that those aftershocks recorded at a single station probably have a magnitude smaller than 1.3.

A second example is the Laffrey earthquake (maximum EMS intensity V–VI), 15 km south of Grenoble (Isère). Besides the fact that its focus was similarly shallow (~ 3 km) although here located in the pre-Triassic micaschist basement, it should be also pointed out that: (*i*) it also involved strike slip; (*ii*) glacial deposits along the Drac river also produced site effects; (*iii*) it also generated a long series of aftershocks over more than 15 months (Thouvenot et al. 2003), again an unusual span for a 3.5 magnitude. Many of these aftershocks were locally felt, although the information that reached us by e-mail (no on-line questionnaire was then available) is necessarily biased. The smallest aftershock that could be located and was also reported felt occurred 3 days after the mainshock. For this event, we compute a magnitude of 1.1 only, whereas we estimate a maximum intensity of IV from the fragmented received testimonies.

At short epicentral distance, the routine computation of the M_L magnitude can be questioned: Richter (1935) dealt with earthquakes assumed to be sited at a depth of 15 km, and his flat attenuation curve for the first 5 km of epicentral distance expresses this assumption. In the case of the aforementioned event, 4 Sismalp stations at distances of 10, 35, 58, and 100 km were available for M_L computation, which yielded the respective values of 0.86, 1.29, 1.00, and 1.11 (mean value: 1.07 ± 0.18). Although the 0.86 value obtained at a distance of 10 km is the lowest of the series, it does not deviate significantly from the mean value if we take the standard deviation into account. However at still shorter epicentral distance we can expect problems: what would be the meaning of an M_L-magnitude computation for a station sited just above a 300-m-deep focus? The question seems academic, but such instances are encountered when small, ultra-shallow earthquakes are felt or heard.

THE 2002–3 TRICASTIN EARTHQUAKE SWARM AND THE 2006 CONAND AFTERSHOCKS

The 2002–3 Tricastin earthquake swarm

The first instance of such small, ultra-shallow earthquakes is provided by the earthquake swarm that occurred in 2002–3 in Tricastin (France) close to Saint-Paul-Trois-Châteaux (Drôme). This area of the middle ‘Sillon Rhôdanien’ (Fig. 1), between the French Massif Central to the west and the Alps to the east, has been known for centuries as the seat of long-lasting earthquake swarms. In 1772–3 such a swarm visited the village of Clansayes where the church tower was knocked down by the strongest event of the sequence (maximum intensity: VII–VIII); in 1933–6 another swarm visited several villages close to La Garde-Adhémar, which suffered slight damage (maximum intensity: VII) during the 1934 climax (Rothé 1936).

The 2002–3 earthquake swarm initiated at the beginning of December 2002 by shocks perceived as explosions by the inhabitants of a ~ 20-house hamlet close to Clansayes. These abnormal sounds were not at once identified as earthquakes by the inhabitants because local earthquakes are inexistent in the inter-swarm quiescence periods, and—to our knowledge—the latter felt swarm dates back to 1933–6. A temporary velocimetric station was installed in the basement of one of the houses at the end of December 2002; thirteen more stations were installed later in January after we identified the phenomenon as seismic.

Several scores of events could be located over a few weeks monitoring. Although activity was maximum right beneath the hamlet, other shocks were detected along a north–south-trending, ~ 7-km-long zone. Available geological maps identify no corresponding fault. On several seismograms recorded by the station installed in the hamlet, we observed an S – P difference of only 45 ms (Fig. 2). The massive coral-limestone formation that outcrops in the vicinity can be assigned a velocity of 5,000 m s^{-1}. Consequently the corresponding focal depth for those ultra-shallow earthquakes is 300 m at most (Jenatton et al. 2004).

Because of their small magnitude, most of these swarm earthquakes could not be located by the *permanent* monitoring networks, although the Clansayes permanent station could detect some of them. Only two events could be located (14 Dec. 2002, M_L = 1.5 and 1 Jan.

2003, $M_L = 1.7$), whereas in December explosions were reported heard sometimes as frequently as several times a day. The same observation was made in 1934 (Rothé 1936) when earwitnesses described 'véritables canonnades' and 'tirs de barrage'.

The 2006 Conand aftershocks

The $M_L = 3.5$ earthquake that occurred on the south-western flank of the French Jura on 11 Jan. 2006 at 11.32 local time is one of the many events that—just like the Annecy or Laffrey earthquakes—regularly strike the external domain of the Alps (Fig. 1). The epicentral zone is sited amidst NW–SE-trending ranges where Dogger (Middle Jurassic) limestone outcrops. The earthquake was felt up to a distance of ~ 20 km, but reached EMS intensity IV in 5 villages only. A maximum intensity of VI was assigned to Conand (Ain), where more than half of the startled 72 inhabitants left their dwellings. A chimney was knocked down. The church pavement was cracked on both sides of the aisle, and rock flour was expelled from the fissures. Drinking water was turbid for two days, and a falling in of stones blocked a small road (Bureau Central Sismologique Français 2006).

These effects, unusual for a 3.5 magnitude, were followed by vibrations and explosions in the next days. Such phenomena were of course reported by the residents to the prefectoral services, which then addressed the seismological networks. As the magnitude of the corresponding shocks was much below any detection level, the obvious answer was that no seismic activity had been observed, hence leaving the Conand inhabitants in perplexity. It actually took 10 days before we realized that something unusual was happening. A temporary velocimetric station installed in the village soon recorded aftershocks which proved very shallow: with S – P = 0.12 s, and by assuming a 5,000 m s^{-1} velocity for P waves in Dogger limestone, we compute a hypocentral distance of 900 m. From the P-wave amplitude recorded on the vertical and horizontal components, we estimate the station to be sited at ~ 50 m from the epicentre, while the focal depth is ~ 900 m.

The largest recorded aftershock occurred on 10 Feb. 2006, one month after the mainshock. This event was heard as a loud explosion. Vibrations were also reported. It was not recorded by the surrounding monitoring networks although the closest permanent Sismalp station is only 15 km away. This station, installed in a mushroom cave bored in Dogger limestone, has a low noise level; however it is only triggered by an STA/LTA algorithm (no continuous recording).

If we use the seismograms obtained at the Conand local station (Fig. 3) for computing the M_L magnitude of the 10-Feb. earthquake, our routine processing infers a value of 2.3. This is obviously overestimated because Richter's assumption of a 15-km focal depth does not apply here with a station at the epicentre and a shallow focus. To ascertain the seismic moment of this earthquake, we theoretically modelled the S-wave pulse which has a frequency close to 20 Hz and an amplitude of 280 µm s^{-1}. We assumed a 900-m-deep source with a focal mechanism similar to that of the mainshock (pure normal faulting, N135°E-trending horizontal tension axis). We adopted P- and S-wave velocities of 5,000 and 2,900 m s^{-1}, and a density of 2,500 kg m^{-3} for Dogger limestone. We found that a 55°-dipping, 40 m x 50 m source where a 2-mm slip propagated at 2,000 m s^{-1} with a rise time of 12 ms fitted reasonably well the observed S-wave pulse. The seismic moment M_0, obtained by multiplying the rigidy, the fault surface, and the slip, is 8.4 10^{10} N m. To convert it to local magnitude, we use the relation advocated by Bakun (1984) for $M_L < 3$ earthquakes:

$$\log_{10} M_0 = 1.2 M_L + 10 .$$

Hence, under the assumed conditions, M_L is found equal to 0.75.

In February and March 2006, a total of 16 events were recorded by the Conand station. On 28 Mar. 2006 at 07.34 in the morning, two late aftershocks were felt. They were described as two explosions separated by 10 s, the first louder than the second. This doublet was recorded by the local station (Fig. 4). The S – P differences (0.135 and 0.140 s) are slightly larger than for the 10-Feb. earthquake (0.120 s), but we will assume that the difference in focal depth is not significant. By scaling the maximum displacement amplitudes with that of the 10-Feb. shock, we find that the corresponding magnitudes for these two felt events were – 0.2 and – 0.7.

The large discrepancy between the magnitude value computed by routine Richter's technique (2.3) and that computed through the evaluation of the seismic moment M_0 (0.75) demonstrates—if ever it were necessary—that Richter's technique cannot be safely used for shallow ($z < \sim 15$ km) events observed at short ($D < \sim 15$ km) epicentral distance.

However, a very large uncertainty on magnitude values computed here is brought by the conversion from M_0 to M_L. Kanamori's (1977) relation does not apply here because it addresses great earthquakes and involves the so-called moment magnitude. (Were it applied, it would provide a 1.3 value for the magnitude of the 10-Feb. event.) Other empirical relations similar to Bakun's have been proposed, for instance by Hainzl and Fischer (2002) in their study of an earthquake swarm with magnitudes between – 0.5 and 3.2:

$$\log_{10} M_0 = 1.05 M_L + 11.3.$$

This relation would provide an M_L = – 0.35 value for the 10-Feb. event, still smaller than the 0.75 value computed with Bakun's relation. This conversion problem set aside, it seems anyway rather clear that the two 28-Mar. events had very small, most probably negative magnitudes.

THE UPTHROW OF ROCKS

Documented observations of upthrown rocks and boulders are relatively scarce. They include the M = 6.9 1984 Western Nagano, Japan, earthquake (Umeda et al. 1987), the M = 7.8 1990 Philippine earthquake (Umeda 1992), the M = 6.0 1997 Colfiorito, Italy, earthquake (Bouchon et al. 2000), the M = 6.6 2003 Bam, Iran, earthquake (Jackson et al. 2006). One of the interests of these observations is that they provide direct evidence that vertical ground acceleration locally exceeded gravity during these earthquakes. Reports of the upthrow of man-made objects are somewhat more common but, as shown by Newmark (1973) and Bolt and Hansen (1977), they do not necessarily entail vertical ground acceleration greater than gravity.

Recordings of vertical ground accelerations in excess of 1 g during earthquakes are still sparse and uncommon. To date, only half a dozen such records have been documented (Anderson 2006). Remarkably, the best recorded large earthquake to date, the M = 7.6 1999 Chi-Chi earthquake, although it produced surface breaks locally exceeding 7 m in height, generated vertical ground accelerations well below 1 g at all the near-fault accelerometric stations (Lee et al. 2001). Furthermore, although much field work was done following this earthquake, no observation of upthrown rocks was reported.

The smallest-magnitude event for which the upthrow of rocks is well documented is the M = 6.0 1997 Colfiorito, Italy, earthquake. This earthquake has been the largest shock of a series of earthquakes that shook central Italy for several weeks in the autumn of 1997. After this earthquake, it was observed that thousands of stones and rocks, which are numerous in this region of smooth hills and scattered limestone outcrops, had been freshly fractured and broken. Some of the broken stones were lying isolated on soft detritic soil (Fig. 5) while others had been piled up together, probably a long time ago to clear the land for farming (Fig. 6). Broken rocks and stones were found everywhere throughout a zone which covers an area of about 1 km by 1 km, and is located near the heavily damaged village of Annifo, where the maximum shaking intensity (IX) of the earthquake was registered (Camassi et al. 1997). Freshness of cuts and fractures, visible in Figures 5 and 6, and the consistency of the observations for thousands of rocks and stones indicate that these rocks were tossed into the air during the earthquake, with breakage occurring at the time of impact. In several places, the old imprint of the stone in the soil was still visible. A similar phenomenon, although not as extensive, occurred in a second area, located about 4 km away from the first zone, near the village of Colle-Croce, which was also heavily damaged.

This earthquake, like most of the shocks in this sequence, had a normal-fault mechanism typical of the extension regime that characterizes the present-day tectonics of this region. The hypocentre was located at a depth of about 7 km near the bottom of the aftershock zone that delineates the fault plane (Amato et al. 1998). The fault dip was about 40° (Amato et al. 1998). The lack of surface ruptures clearly associated with the earthquake fault plane (Cinti et al. 1999) and the near-disappearance of seismicity at depths shallower than 2 km (Amato et al. 1998) suggest that significant slip during the earthquake was confined to depths larger than 2 km. Satellite radar interferometry data of the area and local GPS measurements (Stramondo et al. 1999) combined with the modelling of the rupture show that the zones of upthrown rocks were located in the area where the largest vertical ground displacement occurred. Vertical displacement inferred in the zones of upthrown rocks is about 30 cm. The relatively moderate size of this event suggests that the upthrow of rocks during earthquakes is a much more common phenomenon than is usually thought.

CONCLUSIONS

Our study concludes that earthquakes much smaller than those commonly assumed, and even with negative magnitudes, can be felt in the case of ultra-shallow earthquakes (those with a focus less than 1 km deep). It means that magnitudes for these events should not be overestimated in historical-seismicity studies whenever such testimonies are used. On the other side, we believe that reports of such phenomena—whether in the past or at present time—should not be neglected. They pinpoint the activity of local faults much more precisely than studies of large earthquakes with complicated isoseismal curves. Felt events with negative magnitudes, usually below the detection threshold of seismometers, finally demonstrate that the human being is an instrument eventually much more sensitive—and perhaps cheaper to maintain—than dense monitoring networks. Awfully, this fact reduces to populated areas the places where the occurrence of such earthquakes can be asserted.

At the other end of remarkable effects, we showed that earthquakes of relatively moderate size ($M = 6.0$) associated with near-fault ground displacement of a few tens of centimetres and no surface break can produce vertical ground accelerations exceeding gravity, and toss objects and rocks into the air. Conversely, some great earthquakes, such as the $M = 7.6$ Chi-Chi event which generated vertical ground displacements more than 10 times higher and a 100-km-long surface break, do not produce vertical ground accelerations exceeding gravity. Both sets of observations are difficult to conciliate. They provide a formidable challenge to seismologists and earthquake engineers for the years to come.

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FIGURE CAPTIONS

Figure 1. Map of South-East France, with the 4 earthquakes discussed in the text: Annecy (15 Jul. 1996, $M_L = 5.3$), Laffrey (11 Jan. 1999, $M_L = 3.5$), Tricastin earthquake swarm (Dec. 2002–Mar. 2003), and Conand (11 Jan. 2006, $M_L = 3.5$).

Figure 2. Example of ultra-shallow swarm earthquake recorded in Tricastin by a temporary station (vertical, N–S, and E–W components of a 2-Hz velocimeter; 200-Hz sampling rate). This 4-s window shows P- and S-wave arrivals only 45 ms apart (S waves better observed on the E–W component). Focal depth is about 300 m. Amplitude window for each component is ± 300 μm s^{-1}.

Figure 3. Felt Conand aftershock (200-Hz sampling rate) used for computing seismic moment and corresponding M_L-0.75 magnitude (4-s window). S – P = 120 ms; focal depth is about 900 m. Amplitude window for each component is ± 300 μm s^{-1} (same amplification as Fig. 2).

Figure 4. Aftershock doublet felt at Conand ($M_L = -0.2$ and -0.7), 25-s time window, 200-Hz sampling rate. Amplitude window for each component is ± 30 μm s^{-1}. Note that the maximum amplitude is here reached on the E–W component, whereas it is observed on the N–S component for Figure 3. It indicates either a slight difference in the position of the epicentre or a difference in source mechanism.

Figure 5. Typical pictures of isolated stones (fragile marly limestone) found throughout a 1-km^2 zone following the $M = 6.0$ Colfiorito earthquake. The two original stones on the left were broken into several pieces while the one on the upper right was completely shattered. The rock on the lower right had its top partly scaled (the white areas), likely at impact. (After Bouchon et al. 2000.)

Figure 6. General typical view of a rock pile (upper left) and three detail views near the heavily-damaged village of Annifo following the Colfiorito earthquake. Most of the stones in the piles (fragile marly limestone) were freshly fractured or broken. (After Bouchon et al. 2000.)

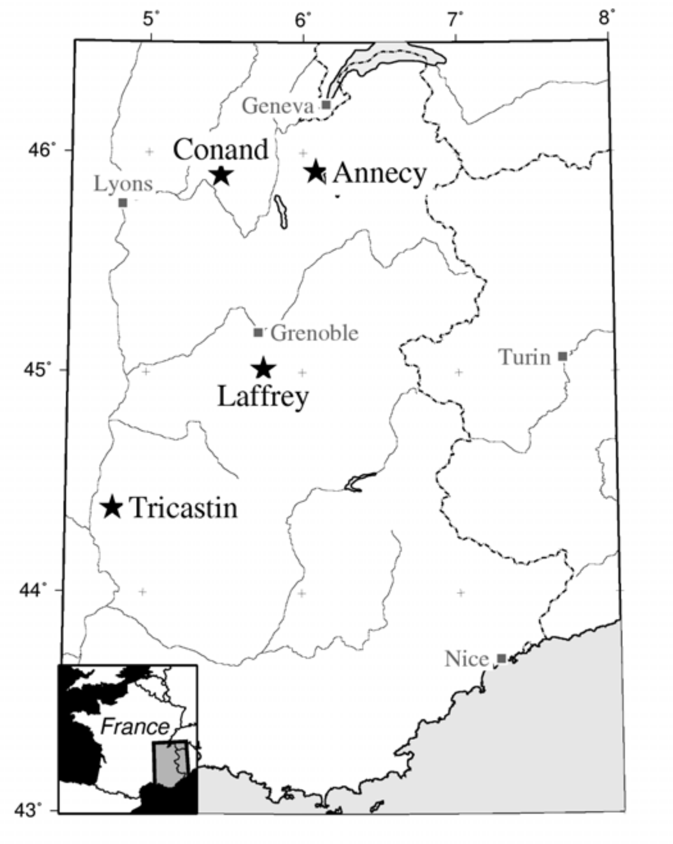

Figure 1.

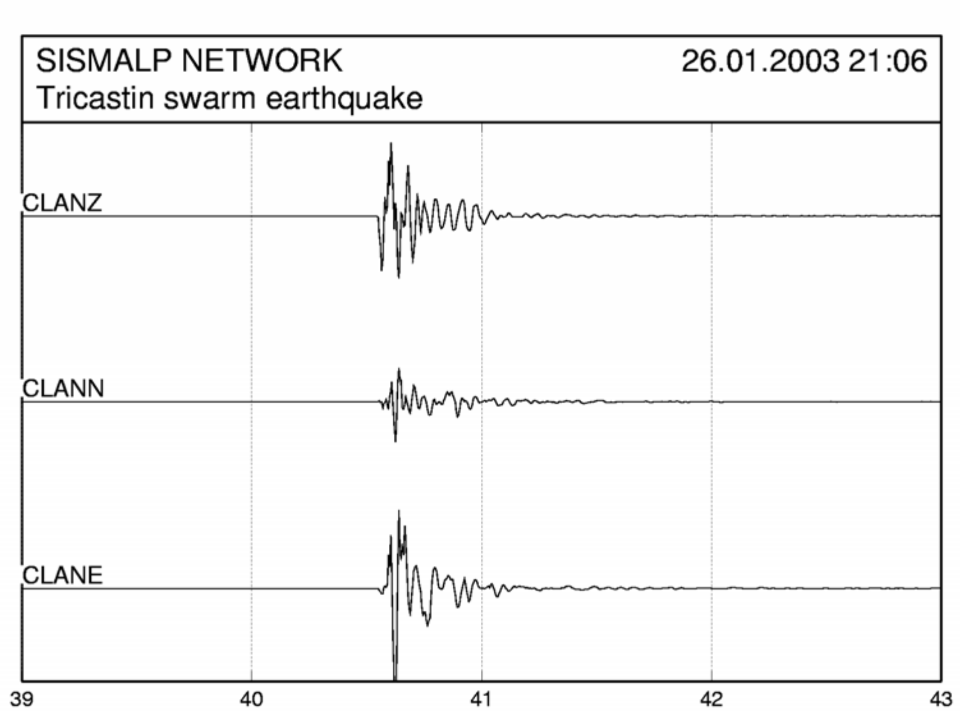

Figure 2.

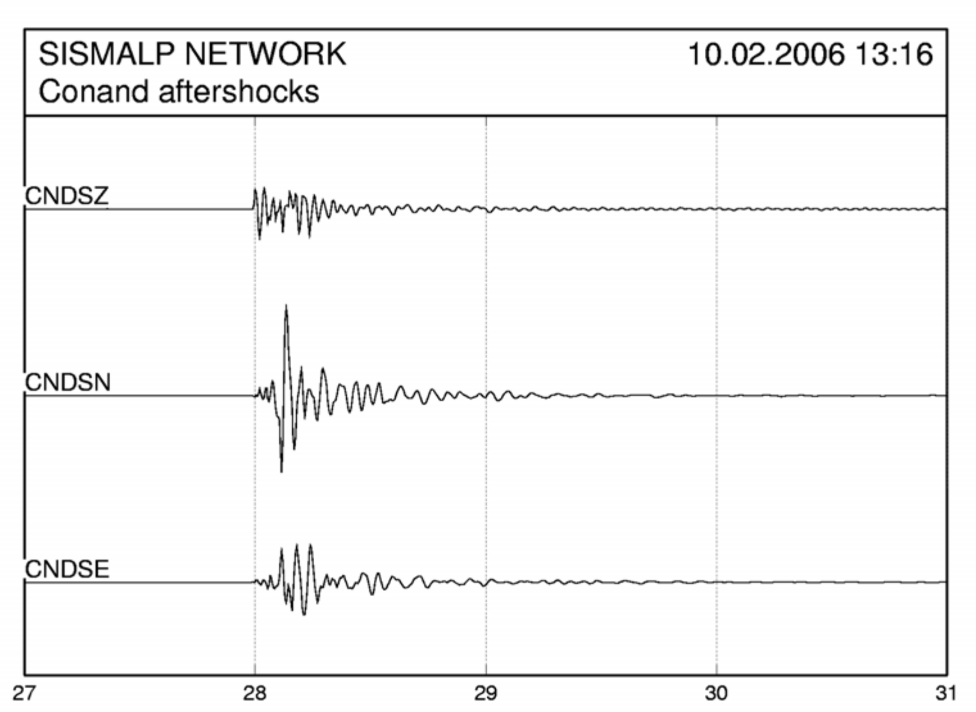

Figure 3.

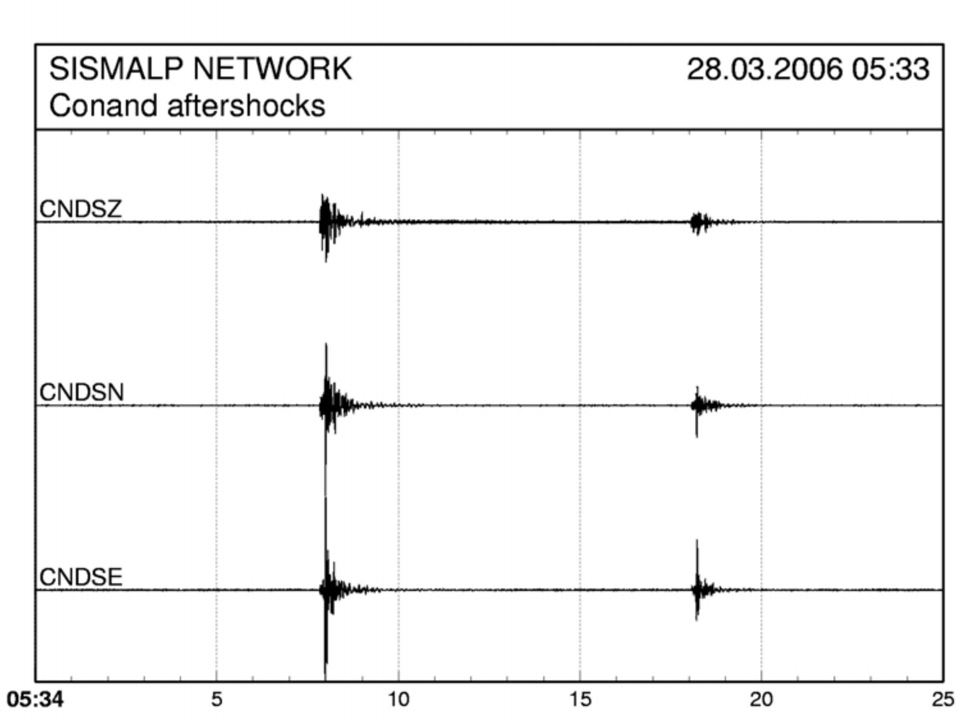

Figure 4.

Figure 5.

Figure 6.