Tracking the spatio-temporal evolution of foreshocks preceding the Mw 6.3 2009 L'Aquila Earthquake

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Abstract

How faulting processes lead to a large earthquake is a fundamental question in seismology. To better constrain this pre-seismic stage, we create a dense seismic catalog via template matching to analyze the precursory phase of the Mw 6.3 L'Aquila earthquake that occurred in central Italy in 2009. We estimate several physical parameters in time, such as the coefficient of variation, the seismic moment release, the effective stress drop, and analyze spatio-temporal patterns to study the evolution of the sequence and the earthquake interactions. We observe that the precursory phase experiences multiple accelerations of the seismicity rate that we divide into two main sequences with different signatures and features: the first part exhibits weak earthquake interactions, quasi-continuous moment release, slow spatial migration patterns, and a lower effective stress drop, pointing to aseismic processes. The second sequence exhibits strong temporal clustering, rapid spatial expansion of the seismicity and larger effective stress drop typical of a stress transfer process. We interpret the differences in the seismicity behavior between the two sequences as distinct physical mechanisms that are controlled by different physical properties of the fault system. We conclude that the L'Aquila earthquake is preceded by a complex preparation, made up of different physical processes taking place over different time scales on faults with different physical properties.

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Key Points:

- We build a new, denser catalog of foreshocks before the M6.3 L'Aquila earthquake
- We reveal a complex two stages evolution of the precursory seismicity
- The precursory seismicity is driven by different processes as external forcing and stress interaction

Abstract

How faulting processes lead to a large earthquake is a fundamental question in seismology. To better constrain this pre-seismic stage, we create a dense seismic catalog via template matching to analyze the precursory phase of the Mw 6.3 L'Aquila earthquake that occurred in central Italy in 2009. We estimate several physical parameters in time, such as the coefficient of variation, the seismic moment release, the effective stress drop, and analyze spatio-temporal patterns to study the evolution of the sequence and the earthquake interactions. We observe that the precursory phase experiences multiple accelerations of the seismicity rate that we divide into two main sequences with different signatures and features: the first part exhibits weak earthquake interactions, quasi-continuous moment release, slow spatial migration patterns, and a lower effective stress drop, pointing to aseismic processes. The second sequence exhibits strong temporal clustering, rapid spatial expansion of the seismicity and larger effective stress drop typical of a stress transfer process. We interpret the differences in the seismicity behavior between the two sequences as distinct physical mechanisms that are controlled by different physical properties of the fault system. We conclude that the L'Aquila earthquake is preceded by a complex preparation, made up of different physical processes taking place over different time scales on faults with different physical properties.

Plain Language Summary

In this work we study the seismicity before the Mw 6.3 2009 L'Aquila earthquake. We first catalog almost 5000 events from the continuous seismic record. We then analyze the spatio-temporal evolution of this sequence through several physical parameters. We observe that the sequence is divided in two main sequences. Our results indicate that several different physical mechanisms (e.g., aseismic deformation, stress transfer due to earthquake interactions) and potential heterogeneities in the fault system (e.g., distance between seismic regions) controlled how the earthquake sequence played out. Our observations show a complex spatiotemporal evolution during the precursory phase and challenge classic fault models that explain earthquake initiation as a process along a homogenous planar fault.

1 Introduction

The characterization of the physical processes occurring before major earthquakes is a fundamental challenge in seismology. So far, physical models have been proposed to explain the processes that lead to large seismic events, including cascade, pre-slip, and progressive or migratory localization (Ellsworth & Beroza, 1995; McLaskey, 2019; Kato & Ben-Zion, 2020). Which one of these mechanisms best represents the physics of the precursory phase of earthquakes is still under debate.

One of the most powerful tools to study the physical processes taking place before the occurrence of significant earthquakes are foreshocks: small earthquakes that precede some large mainshocks (Dodge et al., 1996; Bouchon et al., 2013). Foreshocks were first observed more than a century ago (Omori, 1908). Since then, many laboratory studies have focused on the precursory moment release (Acosta et al., 2019), aseismic slip and stress changes (McLaskey & Kilgore, 2013; McLaskey & Lockner, 2014) and other characteristics of the foreshocks during the initiation of laboratory earthquakes (McLaskey, 2019 and references therein). In addition, direct seismological observations in different seismotectonic settings such as strike-slip faults (e.g., Dodge et al., 1995, 1996; Bouchon et al., 2011; Chen & Shearer, 2013; Ellsworth & Bulut, 2018; Tape et al., 2018; Yoon et al., 2019; Shelly, 2020; Durand et al., 2020), subduction zones (Kato et al., 2012; Bouchon et al., 2013; Ruiz et al., 2014, 2017), and extensional regimes (Sugan et al., 2014; Sánchez-Reyes et al., 2021) have been carried out to assess which model best explains the occurrence of foreshocks and the physical processes occurring during the precursory phase of large earthquakes. More recently, some studies have taken advantage of high-resolution detection methods such as template matching and/or machine learning (e.g., Gardonio et al., 2019; Ross et al., 2019; Yoon et al., 2019; Durand et al., 2020; Shelly, 2020; Sánchez-Reyes et al., 2021) and the availability of better field data (e.g., more stations near faults. See Savage et al., 2017; Tape et al., 2018; Meng and Fan, 2021; Simon et al., 2021) to study foreshocks. These studies reveal an increased spatiotemporal complexity (i.e., fault interactions, volumetric processes, heterogeneous fault properties) of the processes taking place before large earthquakes. This complexity, mainly revealed by foreshocks patterns, is hard to reconcile with a single physical explanation of the precursory phase (cascade, pre-slip or progressive localization). In addition, the observed foreshocks patterns challenge the actual laboratory scale and theoretical models, which treat earthquake initiation as a process along a homogenous planar fault (Dieterich, 1992; Marone, 1998; Liu & Rice, 2005; Rubin & Ampuero, 2005) or a combination of several planar fault segments (Shimizu et al., 2021), although some cases with non-planar fault geometry exist (e.g., Zhang et al., 2014; Dutta et al., 2021).

To gain insight about the ongoing physical processes occurring near to the nucleation region, before a large earthquake, we study the Mw 6.3 2009 L'Aquila earthquake and its foreshock sequence. This event, which struck central Italy on 6 April 2009 (01:32 UTC) causing damage and fatalities, was preceded by more than 500 small (M>0.5) earthquakes (Chiaraluce et al., 2011). Based on the locations of the events, Chiaraluce et al. (2011) reported that the sequence of foreshocks took place in two different faults: (1) a main fault, where the mainshock (Fig. 1) occurs on 6 April 2009, that hosts most of the seismicity occurring from the beginning of January until 30 March; and (2) an antithetic fault that is activated on 30 March 2009 by a Mw 3.9 foreshock (hereafter F1, Fig. 1). On 5 April 2009 (five hours before the mainshock), the seismicity migrates back to the main fault after the occurrence of another Mw 3.9 foreshock (hereafter F2, Fig. 1, Chiaraluce et al., 2011). The co-seismic rupture took place in the Paganica fault (Falcucci et al., 2009; Cheloni et al., 2010), generating exposed ground deformation (Falcucci et al., 2009; Boncio et al., 2010) and maximum surface displacements of 8.1 cm and 16.5 cm in the vertical and horizontal directions, respectively (Cheloni et al., 2010). Joint inversion using GPS, strong motion, and Synthetic Aperture Radar (SAR) data indicate that the maximum slip on the fault is of the order of 1.4 m (Cirella et al., 2012). According to different rupture models (e.g., Cirella et al., 2009, 2012; Cheloni et al., 2010; Scognamiglio et al., 2010), the slip was concentrated in two main asperities: a small patch updip from the hypocenter, and a second, larger asperity located to the southeast along strike. In this context, the foreshocks were located at the base of the activated fault plane in a region where almost no slip occurred during the mainshock rupture (Valoroso et al., 2013).

Here we complement previous studies of foreshocks of the L'Aquila earthquake (Chiaraluce et al., 2011; Valoroso et al., 2013; Sugan et al., 2014; Vuan et al., 2018), by estimating quantitative parameters of the spatiotemporal evolution of the foreshocks sequence. We focus on an area of 10 km x 10 km surrounding the epicenter (Fig. 1). We then densify the catalog of seismicity before the L'Aquila earthquake by using template matching (Gibbons & Ringdal, 2006) to scan 6 months of data before the main shock. We use a frequency band between 5-30 Hz. The inclusion of high frequencies (>20Hz) compared to previous studies (Sugan et al., 2014; Vuan et al., 2018) permitted us to detect more small events (~M < 1.0), which are best captured at high frequency. Our final catalog that covers from 6 October 2008 to 6 April 2009, contains 4978 events, with the first event occurring on 3 January 2009. No seismicity was detected from 6 October 2008 to 2 January 2009.

Using this new catalog, we analyze the seismic sequence of foreshocks by tracking the time evolution of temporal clustering (earthquake interactions), seismic moment release, and effective stress drop. We also study the spatio-temporal evolution of the events to better characterize the precursory phase of the L'Aquila earthquake. Based on these results, we discuss the physical mechanisms that controls the foreshock sequence, ultimately leading to the mainshock.

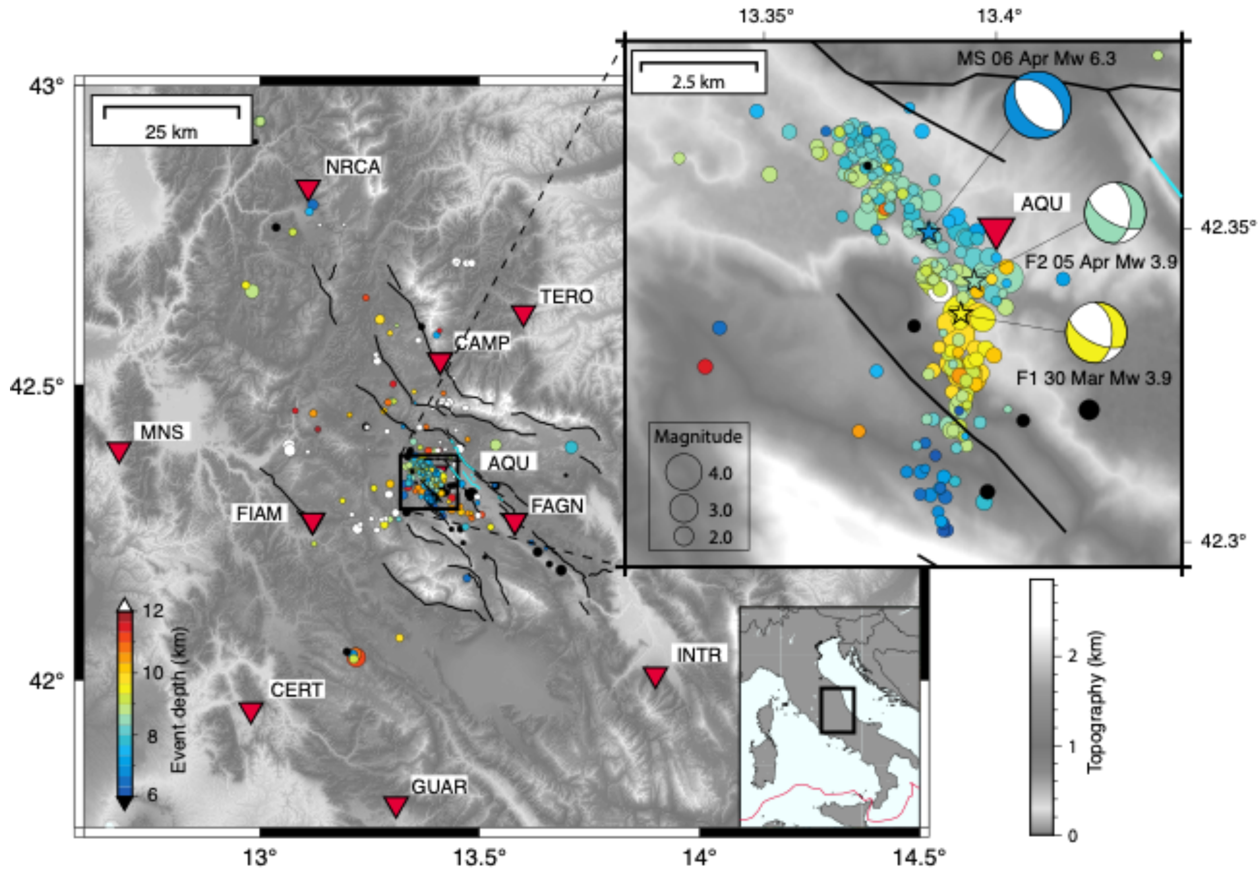

Figure 1: Location map for the L'Aquila earthquake showing the precursory seismicity detected by Chiaraluce et al. (2011); our 267 template events are drawn from this earthquake catalog and are shown in the zoom. The broadband stations we analyzed are shown by the red triangles. Black and cyan thin lines represent traces of the active mapped faults and co-seismic surface ruptures, respectively (Boncio et al., 2010). Upper-right zoom: 267 events used as templates to scan continuous data color-size coded according to depth and magnitude, respectively. Beachballs (compressional quadrants in colors) represent source mechanisms (reported by INGV) for the mainshock (MS 6 April) and the two foreshocks Mw 3.9 one week (F1 30 March) and five hours (F2 5 April) before it, discussed in this study. All of them correspond to normal (extensional) mechanisms.

2 Extending the Seismic Catalog

We apply template matching (Gibbons & Ringdal, 2006) to continuous seismic data collected by the Istituto Nazionale di Geofisica e Vulcanologia (INGV) from 6 October 2008 to 6 April 2009 (6 months). We use 10 broadband three-component stations (red triangles in Fig. 1) from the Italian Seismic Network (INGV Seismological Data Centre, 2006) and the Mediterranean Very Broadband Seismographic Network (MedNet Project Partner Institutions, 1990). Data was continuously recorded at a sampling rate of 100 Hz. Before using the data to study earthquakes, we performed a visual inspection of the spectrograms (Fig. S1), to find the frequency range which is less affected by the strong anthropogenic noise existing in the Apennines (Poli et al., 2020). From this analysis we choose to filter the continuous data from 5 to 30 Hz. The dataset was then organized into 24-hour continuous files with all gaps filled with zeros.

We consider 512 foreshocks reported by Chiaraluce et al. (2011) as potential templates, which have a relative horizontal and vertical location errors about 40 m and 80 m, respectively (Chiaraluce et al., 2011). We identify the highest-quality events by estimating the signal-to-noise ratio (SNR) of each event as the ratio between the RMS velocity during the first 3 s of the P and S waves (for vertical and horizontal components, respectively), and the RMS velocity during a 3 s of noise before the P and S wave arrival times (e.g., Frank et al., 2017; Cabrera et al., 2021). A signal is retained as a final template if it has a SNR ≥ 2 for at least 12 components. We finally retained 267 template event waveforms (inset in Fig. 1), defined as the 3.5 s time windows that starts 0.5 s before the P- and S-wave arrivals at each station for the vertical and horizontal components, respectively, and filtered in an identical manner to the continuous data (bandpassed between 5-30 Hz). The template waveforms are then correlated against a sliding window of continuous data using a GPU-architecture and the Fast Matched Filter algorithm (Beaucé et al., 2018) to obtain daily correlation functions. We search sample-by-sample considering a detection threshold that is 12 times the median absolute deviation (MAD) of the correlation function averaged over all stations and channels to detect events significantly similar to the template. We defined this detection threshold to minimize the occurrence of false detections by first scanning the continuous data using the templates flipped in time (see an example in Fig. S2). With this approach the data are scanned using non-physical and acausal templates unlikely to detect anything, but with the same frequency content as the original templates. We test the number of detections using NxMAD with N in the range 9-12 (see Fig. S3), and we decided to use N=12 as this threshold provides only one false detection during the whole period of time (6 October 2008 to 6 April 2009). To remove double detections over the same time window, we merge consecutive detections with differential times less than 4 s; we keep the detection with the largest average correlation coefficient as the final detection.

We estimate the magnitude of each new event by computing the mean P- and S-wave amplitude ratio between the template event and the detection over the components with a SNR $\geq$ 2. Using the template event's catalog magnitude as a reference, the magnitude of a detected event is determined, assuming that a ratio of 10 of the amplitude ratio corresponds to a variation of one-unit of magnitude (e.g., Peng & Zhao, 2009; Frank et al., 2017; Cabrera et al., 2021).

We further attempt to relocate the newly detected seismicity, respect to the templates. For this scope, we use pair-wise cross-correlation (CC) between each template and its detections, to measure differential delay times. For each event pair, we use waveforms windows of 2 s starting 1 s before the P- and S-waves, respectively. We then relocate each family of detections (a template and its detections) with GrowClust (Trugman & Shearer, 2017). An event pair is only used if its cross-correlation coefficient ($rmin$) is $\geq$0.6 with a maximum source-receiver distance ($delmax$) of 80 km. We also considered a maximum root-mean-square differential time residual for a proposed cluster merger to be allowed during the relocation algorithm ($rmsmax$) $\leq$0.2 (see Trugman and Shearer, 2017. For more details). This procedure resulted in 722 events relocated, or ~17% of the original catalog (Fig. S4). Although low, this percentage is not surprising given the configuration of the network. For example, Ross et al. (2019) relocated 38.7% of events using a denser array of stations in California and Simon et al. (2021) relocated 11.6% of their catalog in Switzerland, in both cases after using template matching. This data reduction is due to the fact that double difference relocations rely on high quality correlations at a single station, while template matching leverages an average correlation across the entire network to identify events that would otherwise go unnoticed. This means that some events could be detected by template matching can have relatively low correlation coefficients that are not necessarily suitable for relocation. Although it is possible to increase the number of relocated events by relaxing for example the $rmin$ and $rmsmax$ parameters, we decided to rather use values similar to previous works (e.g., Trugman and Shearer, 2017; Ross et al., 2019) to prevent a degradation of the quality of the relocation.

As small number of events can be relocated with the approach described above, the new events are considered to occur at the same hypocenter (determined by Chiaraluce et al., 2011) as the template. However, we got an estimation of the distance between the initial location of the detections and the relocated position of new detections. On average, horizontal and vertical distances between templates and new detections are in the order of 83 m and 66 m, respectively (see Figs. S5-6). These values are similar to other studies (~100-200 m, Ross et al., 2019; Simon et al., 2021).

Our final catalog contains 4978 events with magnitude ranging from -0.4 to 3.9 (Fig. 2). We estimate the magnitude of completeness (Mc) of our catalog, using the Lillefors test implemented by Herrmann and Marzochi (2020), which in general provides conservative values of the Mc (see examples in Herrmann and Marzochi, 2020) and allows us to ensure the stability of our further analysis. We use a binning of $\Delta M=0.01$ and we also test Mc for two significance level of $\alpha=0.05$ and $\alpha=0.01$, obtaining Mc=0.8 and Mc=0.9, respectively. As indicated by Herrmann and Marzochi (2020), choosing $\alpha=0.01$ is conservative in a statistical sense (Clauset et al., 2009). We therefore prefer Mc=0.9, a more conservative value for the magnitude of completeness to show the stability of our further analysis (see text S1 for more details). Our catalog presents a decrease in the magnitude of completeness in comparison with Vuan et al. (2018), which catalog exhibits a Mc=1.8 considering the same estimation described above.

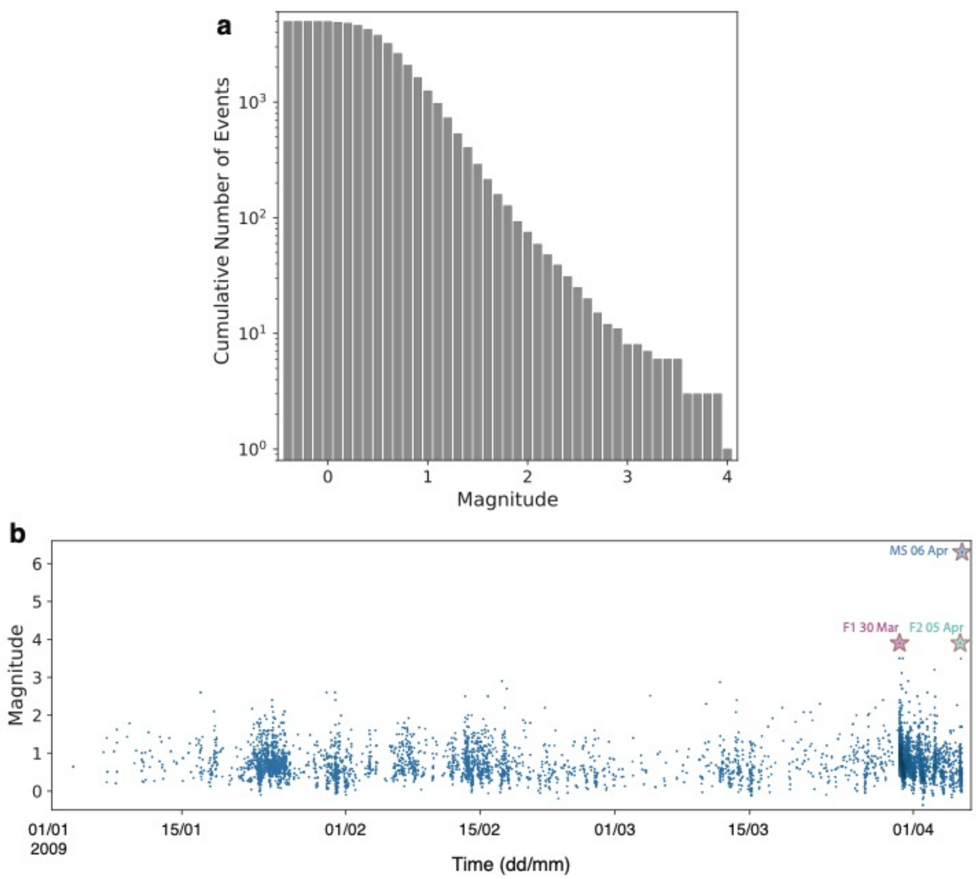

Figure 2: (a) Magnitude-frequency distribution (0.1 bin) for events detected. (b) Estimated magnitudes (see "Extending Seismic Catalog" section for more details).

This new catalog is the largest catalog for this precursory sequence up to date (Sugan et al., 2014 has reported 3571 events and Vuan et al., 2018 extended using one station up to 3786 events), and is created using many constrains to ensure high quality of the detections, such as the selection of the templates based on the SNR criteria for P and S waves, higher frequency band, N-value threshold selection using non-physical acausal templates, the relocation to measure the distance between templates and detections, and a longer period of time scanned.

Figure 3a shows that the seismicity starts on the 3 January and lasts until the 6 April when the mainshock occurs on the main fault. No seismicity is detected in the period between 6 October 2008 to 2 January 2009, so we consider the seismicity starting on the 3 of January as foreshocks of the 9 of April Mw 6.3 earthquake (Chiaraluce et al., 2011; Valoroso et al., 2013; Sugan et al., 2014; Vuan et al., 2018). We observe that the rate of events strongly increases after the occurrence of a Mw 3.9 foreshock on 30 March (F1), which is activating an antithetic fault (Chiaraluce et al., 2011; Valoroso et al., 2013). This activation of the seismicity on the antithetic fault is evidenced in Fig. 3b, which shows a summary of the vertical normalized waveforms for the AQU station (the closest one to the mainshock epicenter, see Fig. 1) aligned on the P-wave arrival. A significant difference of the S-wave arrivals is observed after F1, at the same time as the spatial evolution reported by Chiaraluce et al. (2011) and Valoroso et al. (2013) (see also movie S1). As a first analysis, we split the seismicity before and after F1 on 30 March into two sequences (hereafter S1 and S2, respectively). We observe that the respective cumulative event counts (Fig. 3a and c) of sequences S1 and S2 reveal different time evolutions. The seismicity during S1 is characterized by a slow time evolution, with several accelerations occurring over few days (Fig. 2, 3a and c) and without any clear mainshock driving them (Fig. 2b). On the other hand, the cumulative number of events in S2 evolves with a log-like behavior similar to an Omori law (Utsu & Ogata, 1995).

In the following parts of this work, we track the spatio-temporal evolution of several parameters that describe the style of the seismicity and provide hints about the physical processes active during the foreshock sequence. The mainshock is excluded from this analysis.

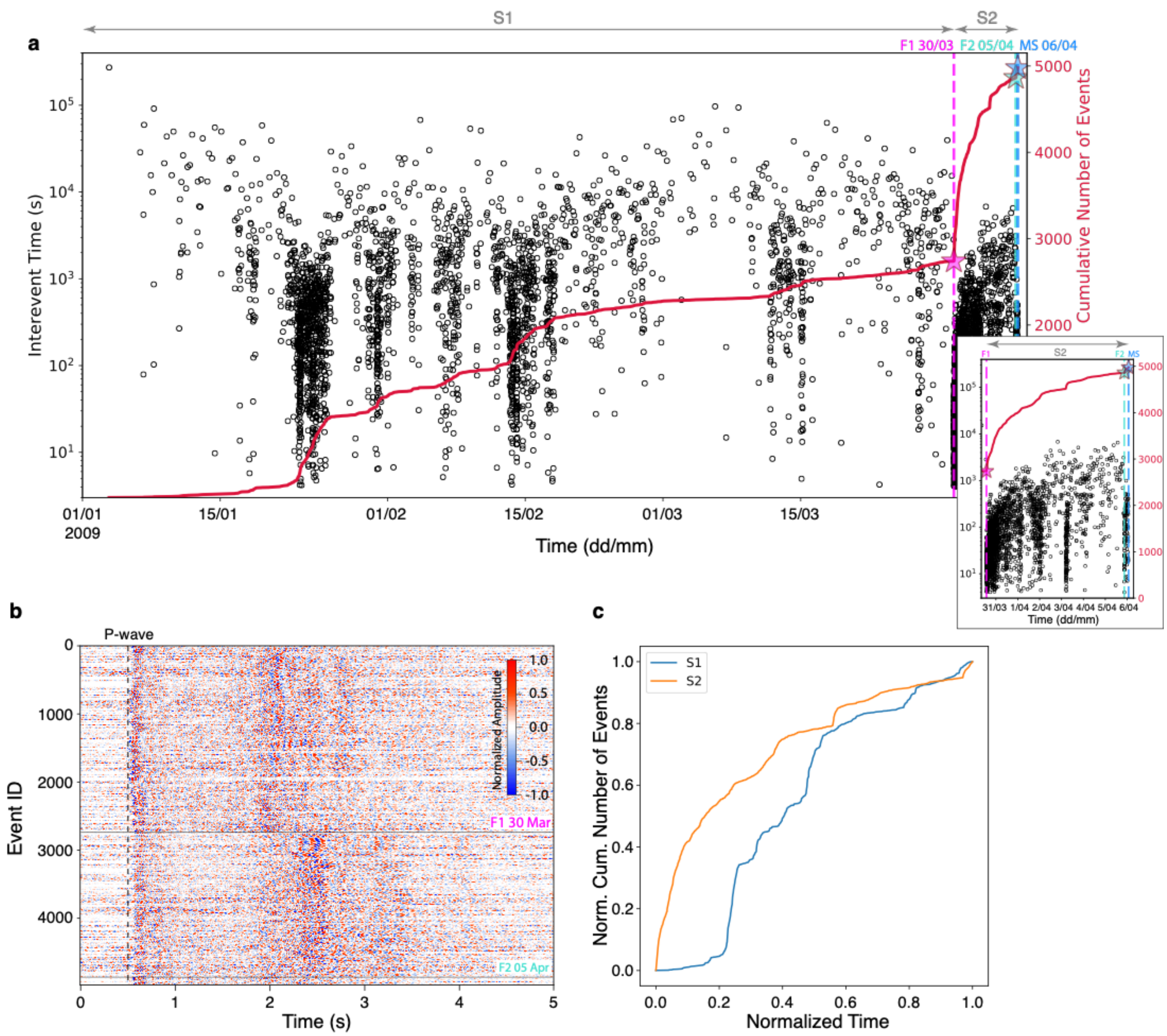

Figure 3: Catalog generated using template matching. (a) Interevent times are plotted using black circles, defined as the elapsed time between consecutive events. Red line represents the cumulative number of events, and fuchsia, turquoise and blue vertical lines show the time of F1, F2 and mainshock (MS) events, respectively. Inset: closer look showing the time interval between F1 and the mainshock. (b) Normalized waveforms of the catalog for the vertical component of the AQU station, aligned 0.5 s before estimated P-wave arrival (black vertical dashed line). Event ID is chronologically ordered (i.e., the vertical axis is time-ordered). Time of occurrence of F1 and F2 are also indicated with black horizontal lines. (c) Comparison between the normalized cumulative of events for S1 and S2.

3 Analysis

We study and discuss the spatio-temporal evolution of the seismicity by tracking the time development of several parameters that characterize the style of seismicity. The parameters are estimated using moving windows of 100-events with a 99-events overlap i.e., the first estimate considers the first 100-events and each subsequent estimate is just shifted by one event in time. This approach allows us to characterize the general evolution of the sequence rather than just focusing on specific time periods. At this point, it is necessary to consider potential effects of the magnitude of completeness and the number of events used in each time window. To that scope, we performed tests considering only events with magnitudes larger than the magnitude of completeness and assess the effect of varying numbers of events for windows-lengths and overlaps, to evaluate the stability of the results (see Fig. S7-10). A jack-knife process was also carried out, removing 20% of the catalog in 100 realizations, to assess the uncertainties for each parameter (Fig. S7-10). Considering the robustness of the tests mentioned above, we present here the results for the entire catalog (Figs. 4, 5, 6).

3.1 Temporal Clustering

Temporal clustering of seismicity, i.e., how past events affect the occurrence of the future ones, is a key feature of seismicity, and is thought to be principally related to static or dynamic stress transfer (Freed, 2005). Therefore, the study of temporal clustering probes the degree to which earthquakes interactions drive the propagation of seismic sequences over external forcing or other physical processes (Schoenball & Ellsworth, 2017).

To quantify the level of time clustering of the seismicity, we estimate the coefficient of variation (COV) of the interevent times (τ) plotted in Fig. 2a, as $COV(\tau) = \sigma_\tau/\bar{\tau}$, where σ_τ is the standard deviation and $\bar{\tau}$ is the average of the interevent times within the window (Kagan & Jackson, 1990). The COV is 0 for a periodic occurrence of seismicity, 1 for completely random Poisson occurrence, and larger than 1 for temporally clustered earthquakes; put plainly, the larger the COV is, the stronger the time clustering is (Kagan & Jackson, 1990; Schoenball & Ellsworth, 2017; Sánchez-Reyes et al., 2021).

Fig. 4a shows the temporal evolution of the COV. During S1 we see slow oscillations of the COV, with generally low values (ranging from 1 to 2.5). We observe that decreases of the COV are often associated with accelerations of seismicity (Fig. 3a). The lowest values (COV~1) for S1, are observed during an increase of the seismicity rate starting on 21 January (light cyan dots in Fig. 4a) and another increase occurring on ~15 February. This observation suggests that the increment of seismicity rate is not due to interevent stress triggering (e.g., seismicity is not driven by a mainshock), and an external mechanism should act to increase the number of events (Beaucé et al., 2019). On the other hand, periods with increased seismicity rates within S2 exhibit strong, punctual temporal clustering followed by random seismicity akin to mainshock-aftershock sequences (Schoenball & Ellsworth, 2017). The general evolution of the COV reflects an evolution of the seismicity style as a function of time, and in particular a clear change of time clustering when moving from S1 to S2.

3.2 Evolution of Seismic Moment Release

The time evolution of the seismic moment (Mo) release reflects the behavior of different types of seismic sequences and offers insights about the processes on activated faults (Vidale & Shearer, 2006). While a stable and gradual moment release by many earthquakes without a dominant large magnitude event is observed for swarm-type sequences (Vidale & Shearer, 2006), most of the moment is released at once during mainshock-aftershocks sequences (Mogi, 1963).

To analyze the time evolution of the seismic moment release, we estimate for each 100-event sliding window the ratio of the maximum seismic moment to the cumulative seismic moment (MaxMo/$\sum Mo$, see Fig. 4b). Values close to 1 are related to windows where the largest event represents most of the Mo released, whereas values close to 0 is observed for windows without a dominant event in terms of moment release. We estimate the moment for each event using the definition of Mw of Kanamori (1977).

Our analysis shows a smooth evolution of the moment ratio during S1 (Fig. 4b), with values ranging from 0.1 to 0.6. This implies that the seismic moment is released nearly uniformly within the window, rather than impulsively by some dominant event. From the beginning of the sequence until the 15 February, the moment ratio is generally low, despite during some periods with increased seismicity rates when larger events occur (Fig. 4b). From 15 February until F1 we observe an increase of the moment ratio associated to the occurrence of larger events M2.7 and M 2.9 (17 February and 11 March, respectively. See Fig. 2a and 4b). During S2, the evolution of moment release is more discontinuous, with large rapid releases of moment, mainly associated with the occurrence of the largest foreshocks (e.g., F1 and F2). A comparison of Figs. 4a and 4b, reveals that the coefficient of variation and seismic moment release have similar patterns, especially during S2, with peak values associated with the occurrence of the largest foreshocks followed by a rapid decrease of COV and moment release. It is important to note that for the moment ratio, the size of the selected window has an effect on the observed level of smoothing. i.e., smaller windows enhance the detection of smaller mainshock-aftershocks sequences (e.g., Fig. S9), which makes its behavior more episodic and less smooth; this effect is diminished for larger windows (e.g., Fig. S10).

3.3 Spatial Evolution

To assess the spatial evolution of the seismicity as function of time we track the along-strike and along-dip position of the events. We first project all the seismicity onto the main fault plane for S1 (strike N133°E and Dip 50° according to Valoroso et al., 2013) and an orthogonal fault plane for S2 (where the antithetic fault is active), obtaining along-strike and along-dip distances measured from the position of the mainshock. We thus track the position of the seismicity centroid (estimated as the average of all event locations) for each time window, and extract the along-strike and along-dip coordinates, which are plotted in Fig. 4c. In addition, the 3D evolution is presented in the supplementary material (see Movie S1).

As previously observed by Sugan et al. (2014), the seismicity starts on the north-west segment of the fault (Fig. 4c). In that region, some slow but significant movement of the centroid along strike and dip are observed, mainly during increases of the seismicity rate until the 13 February. From such observations along with the 3D spatial evolution presented in Movie S1, seismicity appears to be re-rupturing the same fault segment. After the initial activity in the NW segment of the fault, a prominent along-strike migration occurs towards the south-east, observed in Movie S1 and also tracked by the large along-strike variation of the centroid (Fig. 4c). This migration begins around 13 February, accelerates on 17 February covering ~1.2 km in less than 24 hours, and fades on 18 February. At this point, the along-dip and along-strike position of the seismicity stabilize until the end of S1. As an example of the migratory behavior of the seismicity during S1, the event migration starting on 21 January (group of turquoise dots in Fig. 4) is shown in Fig. 5a, where velocities on the order of kilometers per day are required to reproduce the seismicity front (we discuss more details in section 4).

During S2 the antithetic fault is activated after the occurrence of F1, and Fig. 4c shows that the centroid of the seismicity is confined between 1.5 and 2.5 km from the hypocenter both along dip and strike. In addition, the 3D evolution of the seismicity during S2 (Movie S1) does not exhibit slow migrations as observed in S1, but rather a rapid spread of the seismicity on the antithetic fault.

This latter behavior is exemplified by Fig. 5b-c, with longer distances (kilometers) rapidly covered in seconds by seismicity after the occurrence of major events (e.g., F1, F2, and another event magnitude 3.2 in the middle of the sequence indicated by a yellow star). These distances are longer than the expected rupture length of M3.9 earthquakes which are about ~400-800 m (e.g., Udias et al., 2014; Dascher-Cousineau et al., 2020). The strong clustering that we observe from the COV (Fig. 4a) at the occurrence of large foreshocks in S2 (as F1) together with the rapid large scale spreading of the seismicity (Vel.=3 km/s), suggest that stress triggering is the main mechanism driving the seismicity during these bursts (Freed, 2005).

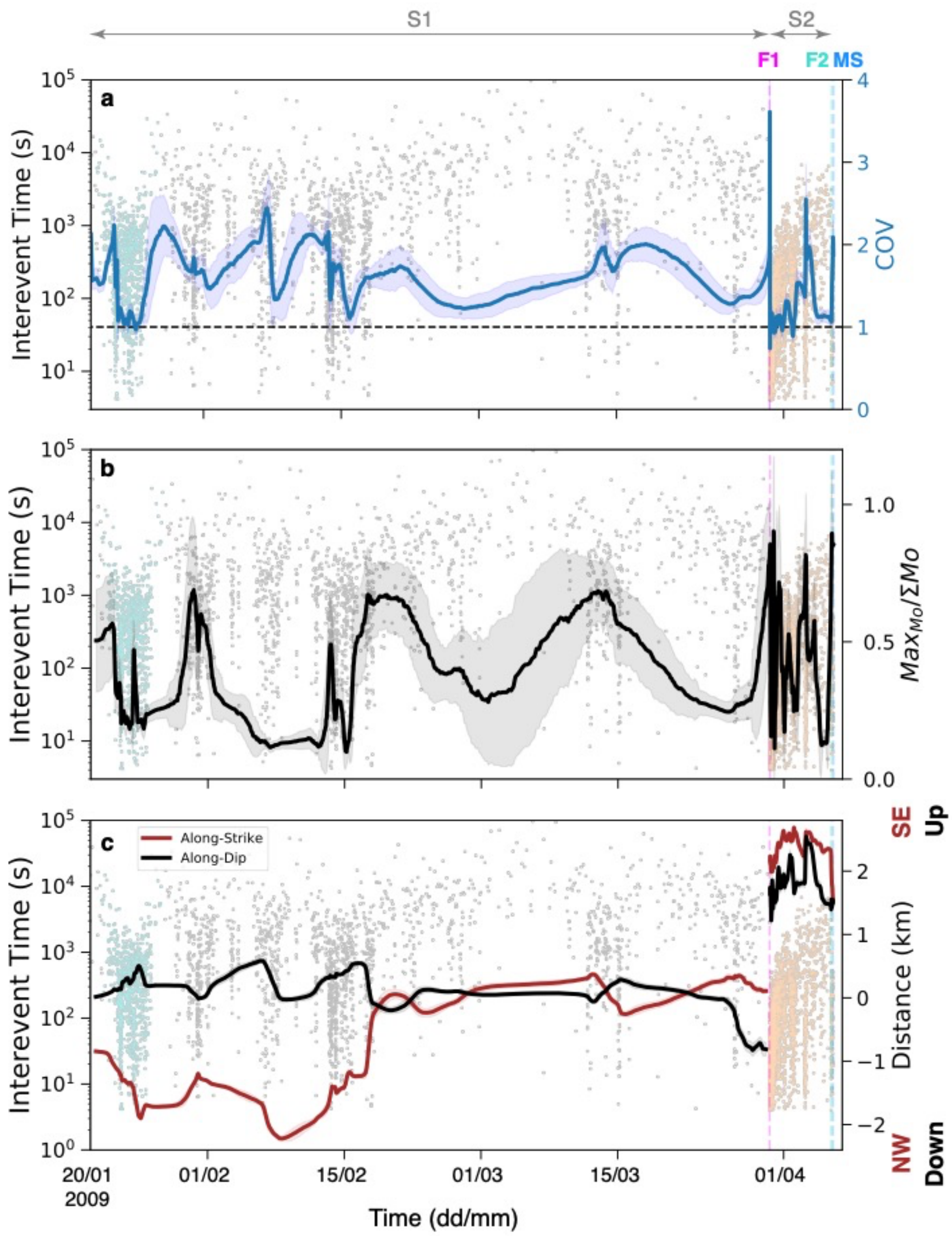

Figure 4: Parameters temporal evolution with a sliding 100-event window length with a 99-event overlap for: (a) Coefficient of variation of the interevent times (b) Ratio between the maximum value of Mo and its total amount within the window (c) Average along-strike and along-dip location of the seismicity measured relative to the mainshock (MS) and projected on the main fault.

Time corresponds to the time of the last event within the 100-event window (see text). Interevent times plotted in Fig. 3a are also indicated in Figs. a, b and c (grey dots). Fuchsia, turquoise, and blue vertical dashed lines show the time of F1, F2, and MS events, respectively. Turquoise and light orange dots represent zoom in presented in Fig. 5.

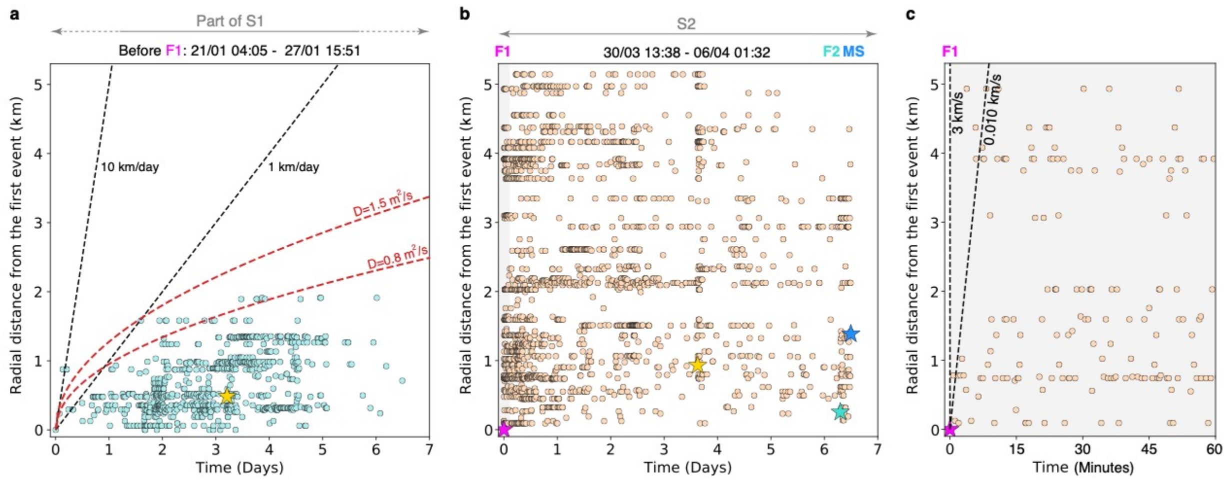

Figure 5: (a) Example of radial distance distribution (measured from the first event of the sequence) for a burst during S1 (see Fig. 4). Yellow star represents the largest event within the sequence (M2.4 according to Chiaraluce et al., 2011). Red dashed lines represent fitted fluid diffusion law (Shapiro et al., 1997) for hydraulic diffusivity of 1.5 m2/s (all the seismicity in a) and 0.8 m2/s (95-percentile of the seismicity in a). (b) Example of radial distance distribution for S2 (see Fig. 4). Fuchsia, turquoise, and blue vertical stars show the time of F1, F2, and MS events, respectively, and yellow star represent another event magnitude 3.2 in the middle of the sequence. (c) Zoom for the first 60 minutes plotted in (b).

3.4 Event Index evolution of the Sequence

The migration style of earthquake locations is generally considered to be an important characteristic to distinguish earthquake swarms from aftershock sequences (Fischer and Hainzl, 2021). While swarms typically show hypocenter migration that depends on the mechanism driving the swarm (e.g., pore-pressure diffusion (Shapiro et al., 1997), hydraulic fracture growth (Dahm et al., 2010), or slow slip (Schwartz and Rokosky, 2007)), aftershocks usually occur immediately across the entire fault plane and along the edges of the mainshock rupture due to stress transfer (e.g., Freed, 2005) although some longer migrations can be linked to afterslip (e.g., Perfettini et al., 2018). Usually, the way to analyze such migration patterns is in the distance-time domain, where the independent variable is typically the time. However, as shown by Fischer & Hainzl (2021) a complementary analysis tool is to use the event order (e.g., event index) as the independent variable, which is also termed natural time (Rundle et al., 2018). While using time as the independent variable permits to resolve if time controls the seismogenic process, using the event-index indicates if the seismogenic process itself controls the seismicity i.e., every rupture opens the way for nucleation of the further rupture (Fischer & Hainzl, 2021). Fischer & Hainzl (2021) showed that an index-plot migration is linear or square-root for either external processes such as pore-pressure diffusion, hydraulic fracture, and slow slip, or in case of an internal process, such as the creation of pore-space during ruptures. In contrast to the random (in space) occurrence of aftershock hypocenters along the mainshock fault plane.

Fig. 6 shows the comparison between the time (Fig. 6a) and the event-index (Fig. 6b) plots for the along-strike position of the seismicity, centered in the mainshock. We can observe that the patterns of spreading seismicity are observed during S1 (e.g., Fig. 6 a and b, red dots), indicating that the active area is increasing due to the occurrence of an external seismicity mechanism. On the other hand, during S2 (Fig. 6 a and b, light orange dots) the event-index plot does not show any migration even removing the time dependence. Instead, we observe a continuous occurrence of events likely resulting from stress transfer, for which no migration patterns is expected (Fig. 6b, Helmstetter & Sornette (2003)).

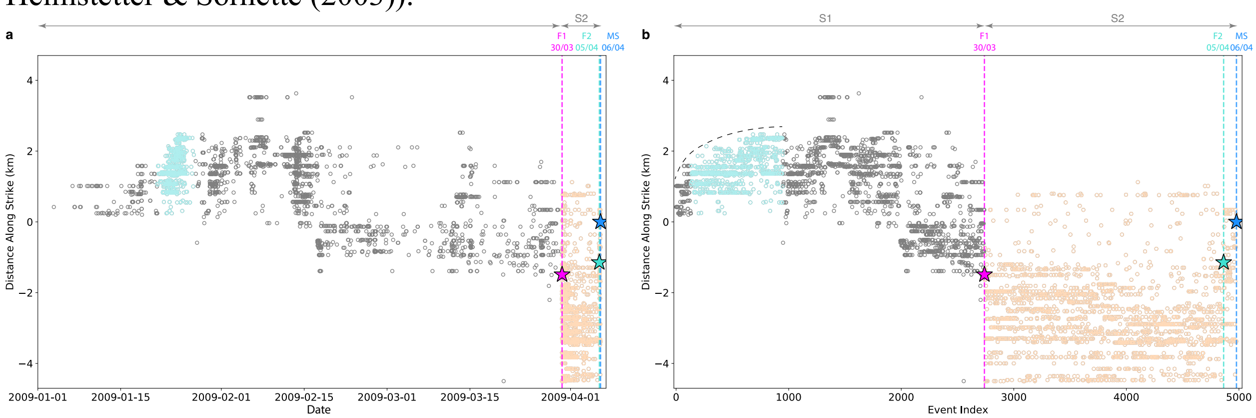

Figure 6: Along-strike distribution of the seismicity centered in the mainshock for (a) time plot and (b) event-index plot. The groups of seismicity shown in Fig. 5a are highlighted in turquoise (part of S1) and light orange (S2). Fuchsia, turquoise and blue vertical lines with stars show the time of F1, F2 and mainshock (MS) events, respectively. Note that the strike considered is the one of the main fault N133°E (Valoroso et al., 2013).

3.5 Effective Stress Drop

We further track the temporal evolution of the effective stress drop ($\Delta\sigma_{eff}$) measured by comparing the cumulative seismic moment and the areal extent of the sequence (Roland & McGuire, 2009; Fischer & Hainzl, 2017). The region enclosing the seismic events was measured using a Delaunay triangulation, after projecting all the seismicity onto the main fault plane for S1 (strike N133° and dip 50° according to Valoroso et al., 2013) and an orthogonal fault plane for S2 (where the antithetic fault is active). An example of this process is shown in Fig. S11. Following Fisher & Hainzl (2017), we impose a distance threshold between neighboring events to avoid outliers, with a maximum triangle leg length of 2.5 km according to the size of the hypocenter cloud. As for the previous analysis, the initial window contains 100 events to estimate the rupture area and the cumulative seismic moment ($\sum M_o$). We then accumulate event by event, and for each window we derive the effective stress drop as $\Delta\sigma_{eff} = \frac{7}{16}\frac{\sum M_o}{r^3}$ (Fisher & Hainzl, 2017). r is here the radius of an assumed circle with the same area as estimated from the triangulation. This procedure was carried out individually for each fault, and their respective results are plotted in Fig. 7.

During S1, we observe a rapid increase in both the radius (the region enclosing seismicity) and the cumulative seismic moment (Fig. 7a) until ~25 January. Then, both parameters become more stable until reaching F1. The first part (S1) of the sequence releases a total seismic moment of 2.9×10^{14} Nm (~Mw 3.6, without considering F1). Different is the behavior of S2, where both $\sum M_o$ and r rapidly grow, reaching a radius and cumulative seismic moment greater than the values for S1 in a shorter time. Fig. 7b shows the time evolution of the effective stress drop for both S1 and S2. The comparison between the effective stress drops for S1 and S2 highlights that during S1, the seismicity take place in an area that is much larger in comparison to the seismic moment released (Fischer & Hainzl, 2017). This leads to a lower effective stress drops of ~0.01 MPa for S1. Whereas in S2, the higher effective stress drop (~0.1 MPa) indicates that most of the area enclosing the seismicity is seismically active. These values are of the order of effective stress drops estimated by Roland & McGuire (2009) for seismic swarms along Southern California and East Pacific Rise transform faults. In addition, the difference of almost one order of magnitude between S1 and S2 is also in concordance with differences in the effective stress drop observed by Fischer & Hainzl (2017) for different seismic sequences such as injection-induced seismicity, natural earthquake swarms, and mainshock-aftershock sequences. In addition, we analyze the cumulative radius against cumulative M_o (Fig. 7c). We observe that our measurements are characterized by a cubic scaling of seismic moment with earthquake cluster radius ($M_o \propto r^3$) but following different constant stress drop values. Such scaling is predicted in the case of faults models with brittle or mixed (brittle and ductile) rheology and homogeneous prestress (Fischer & Hainzl, 2017). Furthermore, the evolution of the cumulative seismic moment release as a function of the cluster radius and differences in the effective stress drops can be used to discriminate different physical processes driving a seismic sequence (e.g., Fischer & Hainzl, 2017). We discuss more in details in the discussion section.

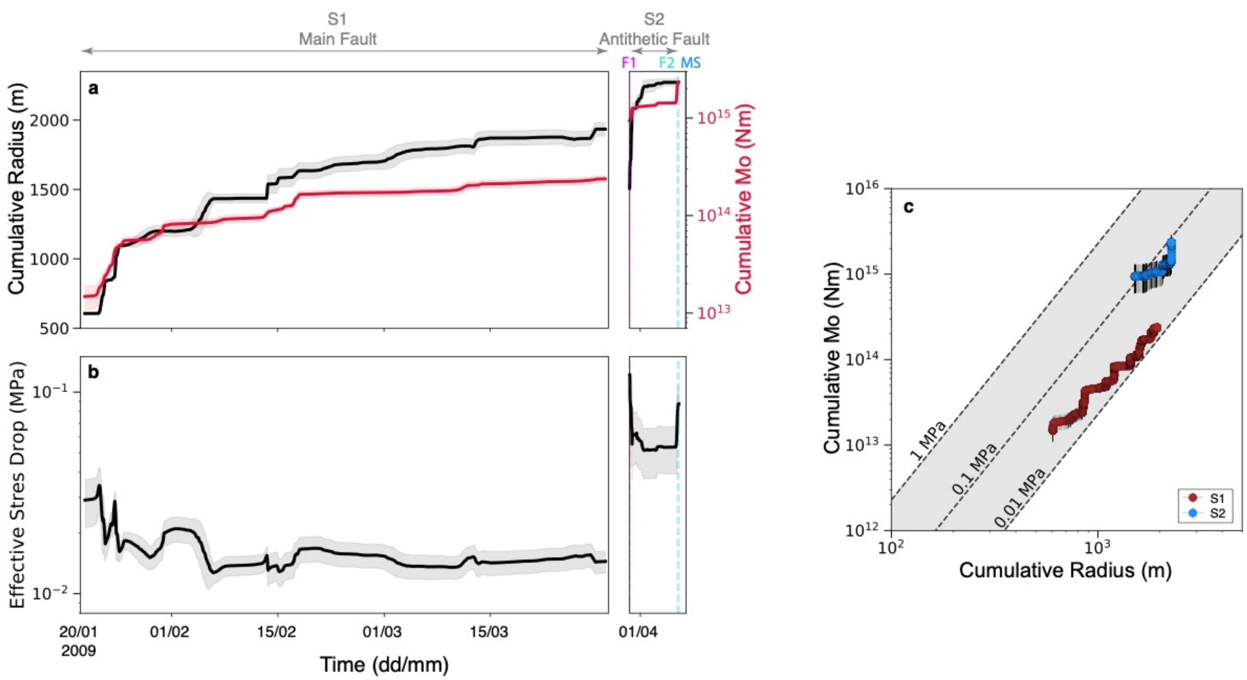

Figure 7: Cumulated moment, radius, and effective stress drop evolution. We use 100-events windows-length and 99 events overlapping for: (a) Cumulated radius (black line) and cumulated moment (red line). (b) Effective stress drop. Time corresponds to the time of the last event within the 100-events window (see text). (c) Scaling between cumulated radius and cumulated moment for the first part of the sequence (S1, red dots) and the second part (S2, blue dots).

4 Discussions

The analysis of the seismicity preceding the Mw 6.3 L'Aquila earthquake reveals a sudden increase of earthquake activity ~3 months prior to the mainshock, in January 2009 (Fig. 2a). From the beginning of the seismicity to the mainshock, almost 5000 foreshocks released a seismic moment of ~3x10^{15} Nm (~Mw 4.3, Fig. 7a). Based on our estimated parameters (Section 3) we observe that the foreshocks sequence develops in two distinct phases and features a complex spatio-temporal evolution. The two stages behavior that we report (mostly aseismic, S1, then mostly seismic, S2) was observed in several other studies, in different tectonic settings (e.g., Kato et al., 2012; Ruiz et al., 2014, 2017; Socquet et al., 2017; Durand et al., 2020).

The first part of the sequence (S1) is characterized by a relatively low temporal interaction of the seismicity (Fig. 4a), smooth moment release (Fig. 4b) and a slow but significant movement of the centroid of the seismicity (Fig. 4c). We also observe migrations lasting up to 7 days (Fig. 5a and Fig. 6). The linear velocity of these migrations ranges from 1-10 km/day (Fig. 5a); these velocities are similar to those associated with seismic swarms driven by aseismic slip (e.g., De Barros et al., 2020). Finally, we observe migrations in time-space and event-index-space (Fig. 6), which is indicative that an external seismogenic process controls the seismicity (Fischer and Hainzl, 2021).

If seismicity is a byproduct of aseismic slip, its intermittent time evolution (Fig. 3A) reflects a variable rate of aseismic slip during the first part of the sequence. A similar behavior is observed during slow slips in subductions zones, with bursts of aseismic slips mainly occurring in rapid episodes associated with bursts of tremors and/or low frequency earthquakes (e.g., Rousset et al., 2019; Jolivet & Frank, 2020). However, confirming the occurrence of aseismic slip using independent data as GNSS is difficult, as the expected surface displacement expected during the bursts of seismicity is smaller than the environmental signals often observed in GNSS data along the Apennines (Amoruso et al., 2017).

The observed migrations (Fig. 5a) may also be explained by fluid diffusion (Shapiro et al., 1997; e.g., Ruhl et al., 2016), considering hydraulic diffusivities of 0.8 and 1.5 m2/s, which are within expected values for the crust (Scholz, 2019; Talwani & Acree, 1985). If this was the case, it would be in agreement with the significant role of fluids reported in the region by several authors (e.g., Antonioli et al., 2005; Lucente et al., 2010; Savage, 2010; Terakawa et al., 2010; Poli et al., 2020).

The second part of the sequence (S2) starts with a magnitude 3.9 event (F1) on the 30 of April 2009, activating an antithetic fault (Chiaraluce et al., 2011; Valoroso et al., 2013) similarly to other recent normal fault earthquakes in the region (e.g., Sánchez-Reyes et al., 2021). The activation of several faults highlights that the precursory process for this event is a complex volumetric process (Savage et al., 2017; Ben-Zion & Zaliapin, 2020), and is not limited to the fault plane.

S2 is characterized by a high temporal clustering (Fig. 4a) and large moment release (Fig. 4b). These parameters suggest a strong interaction between seismic events, likely governed by stress triggering (Freed, 2005). No migration is inferred from the event-index analysis (Fig. 6), and the speed at which seismicity spreads in time is completely different from that observed during S1. Figs. 5b, c show that after the occurrence of F1, the seismicity covers distances of kilometers in seconds to minutes, and similar patterns are observed after the occurrence of another magnitude 3.2 event in the middle of S2 (yellow star in Fig. 5b) and after F2. These velocities are not compatible with mechanisms such as fluid diffusion or aseismic slip, but rather are likely governed by static or dynamic stress transfer (Freed, 2005).

The respective effective stress drops estimated for S1 and S2 are on the order of 0.01 and 0.1 MPa (Fig. 7b). These values are in agreement with estimations in other seismotectonic contexts (e.g., Roland & McGuire, 2009; Fischer & Hainzl, 2017; Schoenball & Ellsworth, 2017), and the difference of $\Delta\sigma_{eff}$ between S1 and S2 (Fig. 7b) provide new insights about the physical mechanisms that might take place during the precursory phase of the studied earthquake. Fischer & Hainzl (2017) estimated the effective stress drops for several seismic sequences to be in a range from 8x10-5 to 3 MPa. They showed that some sequences such as hydraulic stimulations of geothermal reservoirs, seismic swarms and mainshock–aftershock-type are associated with effective stress drops from 0.1 to 3.0 MPa, while smaller values (from 8x10-5 to 0.018 MPa) correspond to sequences that points to a dominating aseismic deformation (e.g., hydraulic fracturing). Considering the above classification, the low effective stress drop (~0.01 MPa) of S1

(Fig. 7b) suggests a dominant role of aseismic deformation during the first part of the sequence, with seismicity occurring over a large area with only a small fraction of the area occupied by asperities releasing seismic energy. In this model, aseismic slip is the main mechanism triggering the activation of distant asperities (Fischer & Hainzl, 2017). Following the models proposed by Fischer & Hainzl (2017) we define S1 as 'mixed' model, as it implies a significant ductile region of the fault with low asperity density. On the other hand, the larger effective stress drop up to ~0.1 MPa after F1 (Fig. 7b) indicates that most of the area enclosing the seismicity is seismically active. In this case the proximity of asperities favors the stress triggering as mechanism for time clustering of events (Fig. 4a) over short time scales (Fig. 5b-c). Given these properties, we call this second model 'brittle'.

Both S1 and S2 show a similar cumulative moment versus radius scaling ($M_o \propto r^3$). This scaling is observed either in the case of brittle fault rheology or in the mixed fault rheology models with homogeneous prestress, but with the different stress drop values discussed above (Fischer & Hainzl, 2017). However, in the case of a partly ductile fault with heterogeneous prestress, the seismic moment only scales with the square of the radius $M_o \propto r^2$, which is not consistent with our observations (Fischer & Hainzl, 2017, Fig. 7c). Considering that the mixed model is representative of S1, and the brittle model of S2 due to the variations of the effective stress drop (Fig. 7c), we discuss possible differences between the fault rheologies in S1 and S2.

In the case of brittle asperities embedded in a ductile environment (mixed model during S1), numerical simulations indicate that two scenarios might occur. Either the asperities rupture simultaneously as a single earthquake or separately as individual events, depending on the distance between the asperities and the frictional strength of the ductile region (Kaneko et al., 2010; Dublanchet et al., 2013; Yabe and Ide, 2017). Thus, high density of the asperities and/or a small $a - b$ frictional parameter in the ductile region lead to simultaneous rupturing of the asperities, while a lower asperity density leads to isolated ruptures, producing a sequence of ruptures with diminished time interaction between each other (Kaneko et al., 2010; Dublanchet et al., 2013; Yabe and Ide, 2017). During S1, the low effective stress drop (Fig. 7b) is resulting from void fault areas deformed aseismically among adjacent ruptures, which did not contribute to the seismic moment release (Fischer & Hainzl, 2017, 2021). In this scenario, the existence of large inter-asperities distances is also consistent with the low seismicity interactions inferred from the COV values (Fig. 4a)

For the case of a brittle fault rheology (S2), the fault segment consists of densely distributed asperities that can rupture individually (Fischer & Hainzl, 2017). For this, some mechanism that prevents the simultaneous rupture of the entire segment and leads to a piecewise rupturing of the fault segment by numerous small earthquakes is needed. Following Yamashita (1999) and Aki and Richards (2002), possible mechanisms might be the presence of barriers, inhomogeneous loading, or dilatancy due to pore creation, a process suggested by Lucente et al., (2010) after the occurrence of F1. In this model, due to the proximity between asperities, the elastic stress plays an important role during the rupture process. This corresponds closely to what our observations indicate during S2: larger COV values (Fig. 4a), episodic and rapid releases of the seismic moment (Fig. 4b) and seismicity covering larger distances of kilometers in short time from seconds to minutes (Fig. 5b, c).

The observed cubic scaling between the accumulated seismic moment and radius is also indicative of re-rupturing for the two models mentioned above (Fischer & Hainzl, 2017). The re-rupturing implies significant overlap between regions hosting subsequent seismic events. This behavior is observed during S1, as reactivation of earthquake families during multiple accelerations of seismicity (Fig. 4c, Movie S1, Fig. S12).

The models of Fischer & Hainzl (2017) suggest that the rerupturing process is expected to continue until the stress is fully released within the whole fault segment. Interestingly, although the seismicity of S1 occurs on the fault plane that slipped during the mainshock (Chiaraluce et al., 2011), there is no overlap between the coseismic slip and the foreshocks (Valoroso et al., 2013, Fig. S13). This suggest that this part of the fault released the full stress in an intermittent fashion through foreshocks (Fig. 4b), as the localized fault properties prohibit the nucleation of a large slip episode. Similar behavior has been observed in modeling, where small events appear at the transition from the locked to creeping behavior toward the bottom of the seismogenic zone with decreasing values of the characteristic slip distance of the friction law (Lapusta & Rice, 2003).

5 Conclusion

The analysis of our high-resolution seismic catalog highlights a number of different physical mechanisms that played a role during the precursory phase of the L'Aquila earthquake. Our results also demonstrate how the faults involved in the sequence present quantitative differences in the earthquake activity they host. While the seismicity occurring on the main fault up to one week before the mainshock (S1) exhibits small time clustering, smooth moment release, slow migrations and a lower effective stress drop, the seismicity occurring on the antithetic fault after F1 (S2) shows strong punctual clustering and moment release, a rapid spreading of the seismicity and larger effective stress drop. Such differences in the seismicity behavior indicate that while an external process (aseismic or fluid diffusion, or likely a combination of both) is driving the seismicity in S1, stress transfer is the dominant mechanism during S2. A comparison of our observations with recent seismic swarm models (Fischer and Hainzl, 2017) indicates that during S1 a mixed rheology model of brittle asperities embedded in a ductile environment and large inter-asperity distances is likely. On the other hand, a brittle fault rheology with a dense population of asperities and small inter-asperity distances is more plausible for the antithetic fault during S2.

Our study shows a complex coalescence of different physical processes occurring during the precursory phase of a large earthquake. Moreover, we highlight how the quantitative analysis of spatio-temporal evolution of microseismicity, can unveil complex precursory behaviors, which differ from nucleation models based on simple planar faults models (Dieterich, 1992; Marone, 1998; Liu & Rice, 2005; Rubin & Ampuero, 2005) aiming for more complex scenarios (e.g., Zhang et al., 2014; Dutta et al., 2021, Shimizu et al., 2021).

Data Availability

Data was downloaded from the Istituto Nazionale di Geofisica e Vulcanologia (INGV, 2006) using obspyDMT (https://github.com/kasra-hosseini/obspyDMT, Hosseini and Sigloch, 2017). The fast matched filter (Beaucé et al., 2018) used in this study can be found at https://github.com/beridel/fast_matched_filter. Computations were performed using the University of Grenoble Alpes (UGA) High-Performance Computing infrastructures CIMENT (https://ciment.univ-grenoble-alpes.fr/wiki-pub/index.php/Welcome_to_the_CIMENT_site!).

The catalog generated here is available at https://doi.org/10.5281/zenodo.4776701 (last accessed 20 May 2021).

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