From fault creep to slow and fast earthquakes in carbonates

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ABSTRACT
A major part of the seismicity striking the Mediterranean area and other regions worldwide is hosted in carbonate rocks. Recent examples are the destructive earthquakes of L’Aquila (Mw 6.1) in 2009 and Norcia (Mw 6.5) in 2016 in central Italy. Surprisingly, within this region, fast (~3 km/s) and destructive seismic ruptures coexist with slow (~10 m/s) and nondestructive rupture phenomena. Despite its relevance for seismic hazard studies, the transition from fault creep to slow and fast seismic rupture propagation is still poorly constrained by seismological and laboratory observations. Here, we reproduced in the laboratory the complete spectrum of natural faulting on samples of dolostones representative of the seismogenic layer in the region. The transitions from fault creep to slow ruptures and from slow to fast ruptures were obtained by increasing both confining pressure (P) and temperature (T) up to conditions encountered at 3–5 km depth (i.e., P = 100 MPa and T = 100 °C), which corresponds to the hypocentral location of slow earthquake swarms and the onset of seismicity in central Italy. The transition from slow to fast rupture is explained by an increase in the ambient temperature, which enhances the elastic loading stiffness of the fault, i.e., the slip velocities during nucleation, allowing flash weakening and, in turn, the propagation of fast ruptures radiating intense high-frequency seismic waves.

INTRODUCTION
In Earth’s upper crust, faults release elastic strain energy stored in the wall rocks via different modes of slip. Depending on the velocity of the rupture front (Vr), faults may creep or generate slow (Vr ≤ 10 m/s; Ide et al., 2007), sub-Rayleigh (Vr = 3000 m/s; also called fast ruptures), or supershear (Vr ≥ 4200 m/s) earthquakes (Kanamori and Brodsky, 2004; Bouchon and Vallée, 2003; Passelegue et al., 2013; Marty et al., 2019). The Mediterranean area and several other regions worldwide are affected by moderate- to large-magnitude earthquakes (Chiaraluce, 2012; Valoroso et al., 2013) nucleating and propagating within (4–8 km) carbonate sequences (i.e., limestones and dolostones). This is the case of the Northern and Central Apennines of Italy, which were recently struck by destructive seismic sequences largely hosted within dolomitic rocks (Figs. 1A and 1B; Chiaraluce, 2012; Valoroso et al., 2013). These sequences were characterized by complex spatio-temporal distributions of mainshocks-aftershocks, with (1) most of the seismicity compartmentalized between 10 km and ~3 km depth, and (2) a sharp upper seismicity cutoff at ~3 km depth (Fig. 1B; Chiaraluce, 2012; Valoroso et al., 2013). Remarkably, in this region, destructive fast seismic ruptures coexist with slow (~10 m/s) and nondestructive rupture phenomena (Figs. 1A and 1B; Crescentini et al., 1999; Amoruso et al., 2002). From a rock mechanics point of view, the coexistence of these different modes of slip remains enigmatic because carbonates can accommodate deformation by crystal-plastic processes, such as aseismic mechanical twinning and dislocation glide, even at room temperature (De Bresser and Spiers, 1997). The ability of calcite crystals to deform plastically at low pressure (P) and temperature (T) can explain the lack of acoustic emission activity (microseismicity) during failure of carbonates at shallow depth conditions (Schubnel et al., 2006; Nicolas et al., 2017). To investigate the frictional stability of carbonates, recent experimental studies focused on the frictional behavior of calcite- and dolomite-rich fault rocks sheared at subseismic to seismic slip rates to determine the frictional behavior of carbonates at ambient and crustal temperatures (Verberne et al., 2015; Fondriest et al., 2013; De Paola et al., 2015). Here, we report the results of triaxial experiments (see the method in the GSA Data Repository1) performed on saw-cut samples cored in dolostone blocks of the Mendola Formation (northeast Italy, Upper Triassic in age) (Fondriest et al., 2015).

METHODS
The experiments were conducted in ambient conditions (temperature and pressure) typical of Earth’s crust where all these different slip modes occur (Fig. 1). We studied the influence of both confining pressure and bulk temperature on the stability of the experimental fault system. Experiments were coupled to strain gauges and an acoustic sensor array to discriminate the nature of the seismicity (see detailed methods in the Data Repository).

EXPERIMENTAL RESULTS
At 25 °C, slip initiated when the shear stress reached the peak strength of the fault, corresponding to a static friction, fsc = τc/σn ≈ 0.4 (where τc and σn are the shear stress and the normal stress at the onset of instability, respectively). At this temperature, the fault exhibited typical frictional behavior (dependence of peak shear stress on confining pressure), and strain energy accumulated during loading was released by stable slip (Fig. 1C). Increasing the ambient temperature to 65 °C preserved the frictional behavior of the fault but led to a transition from stable slip to stick-slip motion (Fig. 1D). In this case, while stress-strain curves suggest stick-slip motion, slow ruptures of ~0.1 m/s were observed, and no high-frequency radiations were recorded. At 100 °C, the fault exhibited a different mechanical behavior. Fast ruptures

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1GSA Data Repository item 2019267, detailed methods and supplementary microstructural and mechanical results, is available online at http://www.geosociety.org/datarepository/2019/, or on request from editing@geosociety.org.

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were observed at both 60 and 90 MPa confining pressure, while only slow ruptures were observed at 30 MPa confining pressure. Fast ruptures induced strong high-frequency motions, recorded on both high-frequency acoustic and dynamic strain monitoring systems. Note that coexisting slow ruptures, which do not produce high-frequency motions, were also observed in the first stage of the experiments conducted at 90 MPa confining pressure. For each event, the peak shear stress at the onset of slip increased with normal stress, which shows that fault reactivation respects frictional criteria independent of the temperature and pressure conditions (Fig. 2A).

At low confining pressure (30 MPa), the increase of the peak friction with ambient temperature led to a transition from stable slip to slow rupture. At 60 and 90 MPa confining pressure, a similar trend was observed: An increase in temperature led to an increase of the peak friction coefficient and to a transition from stable slip to slow rupture, but to fast rupture at $T = 100\, ^\circ C$ (Fig. 2A). Moreover, the amount of stress release via the different modes of slip observed depended on the peak shear stress reached during the loading. Larger peak stress led to larger stress drop (Fig. 2B). However, frictional drop remained small during stable slip and slow rupture ($\Delta f = 0.05$). For similar values of initial shear stress, fast rupture released a larger amount of shear stress (Fig. 2B), i.e., larger frictional drop, which ranged from 0.07 to 0.22 (Fig. 2B). This behavior is highlighted by comparing the static stress drop of each event to the related amount of slip. Each mode of slip (i.e., stable, slow and fast rupture) presents a different linear relation between the static stress drop and the fault slip (Fig. 2C). For the same value of slip, the resulting stress drop is larger during fast ruptures than during slow ruptures. Note that slow ruptures observed at both 65 °C and 100 °C followed the same trend, suggesting similar mechanisms. These results suggest that fast ruptures are more dispersive than slow ruptures.

NUCLEATION OF SLOW AND FAST RUPTURES
Using the travel times of the rupture front recorded by the array of strain gauges located along fault strike, estimates for $V_r$ during slow rupture propagation range from 0.1 to 20 m/s (Fig. 3A). These values are in agreement with rupture velocities of natural slow earthquakes (Ide et al., 2007), suggesting that our experimental slow ruptures are similar to those observed in nature. An increase of $V_r$ during rupture propagation was observed along fault strike (Fig. 3A). In addition, increasing the initial shear stress (i.e., the confining pressure) led to larger rupture velocities at the onset of the frictional instability. To further analyze the influence of
the background shear stress, we computed the evolution of stress with slip (weakening stiffness $K = \Delta \tau/\Delta u$, where $u$ corresponds to the slip along the fault) during each event induced during the experiments conducted at 65 °C under 30, 60, and 90 MPa confining pressure, respectively. The weakening process during slow rupture was slip weakening (Fig. 3B), confirming recent expectations (Ikari et al., 2013). In addition, the increase in the peak shear stress along fault strike led to a larger fraction of shear stress released during the slip events (Fig. 3B). Assuming a typical earthquake energy budget (Kanamori and Brodsky, 2004), pure slip weakening behavior is expected to drastically limit the radiated energy during rupture propagation, explaining the smaller stress drop for a given amount of slip compared to fast rupture phenomena.

Fast earthquakes present a complex nucleation behavior. At the onset of slip, dynamic strain recording showed that fault slip accelerates and radiates low-amplitude and relatively low-frequency (20 kHz) acoustic waves (Fig. 3C). At a critical point (red dashed line in Fig. 3C), the shear stress drops abruptly within 20 μs, initiating high-frequency wave radiation. Using piezoelectric transducers as seismic rupture chronometers (Passelègue et al., 2016), we estimated rupture velocities ranging from 1500 to 5200 m/s. Our estimations for ruptures speed are compatible with previous studies (Passelègue et al., 2016) and with the rupture speed of classic earthquakes (Ide et al., 2007). The amplitude and the frequency of the acoustic motions increased with the stress release rate (Figs. 3C and 3D). Note that the radiation of the high-frequency wave front appeared to occur after the release of half of the shear stress (half of the dynamic stress drop). These results suggest that fault weakening initiates before the radiation of high-frequency waves.

POSTMORTEM MICROSTRUCTURES

Fault surfaces recovered from experiments where stable slip and slow ruptures occurred have highly light-reflective patches visible to the naked eye. These patches are similar to the mirror-like slip surfaces previously observed in nature or after friction experiments conducted with subseismic to seismic slip rates on carbonate rock gouges (Fondriest et al., 2013; Verberne et al., 2015). Scanning electron microscope images reveal the extremely smooth topography of mirror surfaces, composed of tightly packed to welded, subrounded nanograins with negligible porosity (Fig. 4A). In contrast, fault surfaces that experienced fast ruptures are on average much rougher than those after stable slip and slow rupture and are pervasively covered by a foam-like material embedding small well-rounded nanograins with an average size of 150 nm (Fig. 4B). The foam-like material locally includes ultrathin (>5 nm in thickness) filaments and patches connecting and wrapping the nanograins (Fig. 4B), recalling frictional melting textures found in silicate-bearing rocks sheared under similar deformation conditions (Passelègue et al., 2016).

INTERPRETATION AND DISCUSSION

Our experiments reproduced the complete spectrum of natural faulting: (1) stable slip at room temperature, (2) slow ruptures at 65 °C, and (3) coexisting slow and fast ruptures at 100 °C. Our experimental approach succeeded in explaining the onset of fault creep and the transition from slow ruptures (Crescentini et al., 1999; Amoruso et al., 2002) to typical seismicity at $P$-$T$ conditions encountered at 3 km depth, as observed in the Central Apennines based on seismological and geodetic investigations (Chiaraluce, 2012). It seems that this first-order similarity between experimental and natural fault-slip modes was mainly controlled by the variation in temperature. However, since all the experiments were performed in dry conditions, we cannot rule out the potential role played by pore fluids.

In our experiments, the transition from stable to unstable slip was promoted by a combination of the confining pressure, which increases the stiffness of the fault (Fig. DR2 in the Data Repository; Leeman et al., 2016), and the fault temperature, which promotes unstable behavior of rocks and carbonates (Blanpied et al., 1995; Brantut et al., 2011; Verberne et al., 2015; Pluymakers et al., 2016). This finding could be counterintuitive, since an increase in ambient pressure and temperature enhances microplasticity in carbonates, which should release part of the stored elastic strain energy and reduce the rupture speed (Rutter, 1972; Nicolas et al., 2017). In our experiments, the stiffness of the fault increased with both bulk temperature and confining pressure (see the Data Repository, and Fig. DR2a therein) and became greater than the experimental fault length (Fig. 4C). This process is at the origin of the transition from a stable to slow rupture front.

The transition from slow to fast rupture is more complicated. First, the transition seems to depend on the peak friction along the fault at the onset of slip (Fig. 2A), as previously observed (Ben-David et al., 2010; Passelègue et al., 2013). However, large values of peak friction alone are
assuming the average slip rate observed during fast on the size of initial asperities, flash decarbon et al., 2014; Aubry et al., 2018). First, based events (Goldsby and Tullis, 2011; Passelègue to activate flash heating phenomena during fast weakening phenomena (900 J kg

\[ V_r = \frac{\rho \cdot c_p \cdot (T_a - T_r) \cdot v \cdot k(D(r)))} {\rho} \]

where \( \rho \) is rock density (2650 kg/m\(^3\)), \( c_p \) is heat capacity (900 J kg\(^{-1}\) K\(^{-1}\)), \( k \) is thermal diffusivity (1.25 \times 10^{-4} m\(^2\) s\(^{-1}\)), \( T_r \) is decarbonation temperature (600 °C), \( T_a \) is ambient temperature, \( D \) is asperity size (50 and 1 μm), and \( \tau_c \) is contact hard ness of dolomite (3 GPa; Goldsby and Tullis, 2011). Color bar corresponds to nucleation length of each event normalized by fault length (L0). Nucleation length was calculated following \( L_0 = (2L \times \mu \times G \times (1 - f_d/f_s))/2 \). (Campillo and Ionescu, 1997), where \( \beta \) is nondimensional shape factor coefficient (>1.158), \( \mu \) is shear modulus of dolomite estimated using strain measurements, and \( G \) is effective fracture energy, \( G = u \times \sigma_f \times (f_d - f_s)/2 \). \( \tau_c \) and \( \sigma_f \) are the peak shear stress and the normal stress at the instability, respectively, \( f_d \) and \( f_s \) are the static and the dynamic friction coefficient, respectively.

not sufficient to induce fast rupture propagation at low confining pressure. As expected theoretically, ruptures speed increases with the fault weakening rate. In our experiments, the weakening rate increased with the peak slip rate reached during rupture propagation (Fig. 4C). For weakening stiffness (\( K \)) above that of the apparatus (\( K < K_c \)), the slip rate became faster than 1 m/s, and could reach up to 10 m/s, within the limit of our resolution (see the Data Repository). This enhancement in the weakening stiffness and weakening rate can be explained by the activation of weakening mechanisms due the increase in the slip rates.

While the activation of plastic mechanisms explains the weakening and transition from stable to unstable behavior in calcite (De Paola et al., 2015; Green et al., 2015; Verberne et al., 2015, 2017; Pozzi et al., 2018), these mechanisms are not dominant in dolomite (see the Data Repository). However, sliding velocities above a critical weakening velocity (\( V_r \)) are expected to activate flash heating phenomena during fast events (Goldsby and Tullis, 2011; Passelègue et al., 2014; Aubry et al., 2018). First, based on the size of initial asperities, flash decarbonation is expected to occur when the slip rate becomes larger than 1 cm/s (Fig. 4C), explaining the low weakening observed during slow ruptures (10^{-4} < V_r < 10^{-2} m/s). Second, assuming the average slip rate observed during fast ruptures (\approx 2.5 m/s), we can state that asperities larger than 0.24 μm are expected to decarbonate during their lifetimes during fast ruptures (see the Data Repository). These results agree with postmortem microstructures, which highlighted nanograins wrapped by a foam-like material that resembles a solidified melt. Note that in carbonate minerals, the melting point decreases dramatically in the presence of CO₂ at room humidity conditions and can be close to the decarbonation temperature, explaining the melting observed at the scale of the asperities (Wyllie, 1965). The activation of flash heating on asperities during instabilities seems to explain both (1) the gap existing between slow and fast ruptures in terms of weakening and slip velocity (Fig. 4C), and (2) the strong enhancement of the weakening rate during the nucleation of fast ruptures, which initiates the radiation of high-frequency seismic waves (Figs. 3C and 3D).

To conclude, our results demonstrate that, in contrast to silicate rocks such as granite, which behave dynamically (fast rupture propagation) without activation of strong weakening processes (Blanpied et al., 1995; Passelègue et al., 2014; Leeman et al., 2016), dynamic rupture and high-frequency radiation require the activation of intense fault weakening in carbonates, such as frictional flash weakening in dolomite or plastic processes in calcite (Verberne et al., 2015, 2017; Pluymakers et al., 2016; Pozzi et al., 2018).

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Supplementary materials for: From Fault Creep to slow and fast Earthquakes in Carbonates

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1. Sample used in this study

Cylindrical samples (diameter: 40 mm, length: 90 mm) were cored from dolostone blocks of the Mendola Formation (Upper Triassic in age), a dolomitized platform carbonate with crystal size in the range 20 to 300 µm and larger crystals filling diagenetic pores (Fondriest et al., 2017). These dolostones are equivalent to the Upper Triassic-Jurassic dolomitized carbonates composing the seismogenic layer of the Central Apennines and were characterized in terms of acoustic and elastic properties in previous studies (Fondriest et al., 2017). The rock cylinders were saw-cut to create an experimental fault at an angle of 30° with respect to \(\sigma_1\) (principal stresses are denoted \(\sigma_1 > \sigma_2 = \sigma_3\)). Fault surfaces were roughened first with grinder and then with coarse sandpaper (grit number P240, which corresponds to ±50 µm in surface roughness) using ethanol to avoid frictional heating during sample preparation. All experiments were conducted on fault surface presenting the same initial geometry and roughness (Figures S1a and S1b).

2. Materiel and methods

Experimental apparatus

The apparatus used in this study is a tri-axial oil medium loading cell (\(\sigma_1 > \sigma_2 = \sigma_3\)) built by Sanchez Technologies. The confining pressure is directly applied by a volumetric servo-pump up to a maximum of 100 MPa. The axial stress is controlled independently by an axial piston controlled by a similar servo-pump. The axial stress can reach 680 MPa on 40 mm diameter samples. Both confining and axial pressure are controlled and measured with a resolution of 0.01 MPa. Axial shortening is measured by averaging the values recorded on three capacitive gap sensors located externally. These sensors record both the sample deformation and that of the apparatus. The resolution of these measurements is 0.1 micron. Both pressure and displacement data are
recorded at sampling rates ranging between 10-to-1000 Hz during experiments. More details can be found in Brantut et al., (2011). Note that because of our sample geometry, increasing the differential stress leads to an increase of both shear and normal stresses.

**Acoustic monitoring system**

During experiments, acoustic activity was monitored through 15 piezo-ceramic sensors which consist of a PZT crystal (PI ceramic PI255 0.5 mm thick and of diameter 5 mm) contained in a brass casing (Figure S1b). The sensors were glued directly on the samples with cyanoacrylate adhesive. Acoustic waveforms were recorded with two different techniques (Passelegue et al., 2017). First, each unamplified signal was relayed to a digital oscilloscope allowing for the recording of macroscopic stick-slip events within a time window of 6.5 ms at 10 MHz (Passelegue et al., 2017). Second, to record low amplitude acoustic emissions activity, signals were amplified at 45 dB though pre-amplifiers. Amplified signals were then relayed to a trigger logic box. Using this second system, AEs were recorded if at least 4 sensors recorded an amplitude larger than a given threshold, that is set at 0.001 Volts. The complete waveform catalogue was then manually analyzed to remove possible triggers from background noise.

**Strain gages measurements**

In addition to the record of regular mechanical data, four strain gauges 1.5 cm equally spaced and recording preferentially shear strain $\varepsilon_T$ were glued at 4 mm from the fault plane along the fault strike (Figure S1b). Mechanical data and strain measurements were recorded at 2.4 kHz sampling rate. In addition to classical strain gages, a Wheatstone bridge strain gages was glued directly on the rock sample close to the fault plane (Figure S1b). The Wheatstone bridge is composed of four resistors ($\Omega=350 \text{ ohms}$) measuring together the differential strain ($\varepsilon_1 - \varepsilon_3$). The signals are relayed to a high frequency strain gage amplifier allowing up to 10 MHz sampling rate. The strain gages are calibrated using low frequency stress and strain measurements during the elastic part prior to each STE, assuming a constant Young's modulus (E= 64 GPa) for Dolomite. Assuming linear elasticity, the measurement at high sampling rate of the differential strain is converted into the dynamic evolution of the differential stress (i.e., $\sigma_1 - \sigma_3$) during STE. The assumption of linear elasticity during short time intervals remains robust because damage is localized during stick-slip instability in comparison with intact specimens and do not affect the bulk of the granite at the measurement location.

3. **Rupture properties**

**Slip measurement**

Axial displacements ($D_{ax}$) were recorded through three gap sensors located outside the cell with a resolution of 0.1 $\mu$m at 2.4k samples per second. External data (collected using the gap sensors) are corrected using the axial deformation of the sample measured using strain gages glued
directly on the rock sample. Up to four pairs of strain gages can be used during each experiment. Each pair of strain gages is composed of two resistors ($\Omega = 120$ ohms) measuring respectively the axial and the radial strain, corresponding to $\varepsilon_1$ and $\varepsilon_3$ in the selected frame of reference. Strains are recorded continuously at a sampling rate of 2.4kHz. Using these measurements, we can estimate the elastic constants of the rock during the elastic part of the experiments and correct the shortening measured externally from the rigidity of the apparatus using the relation:

$$\varepsilon_{\text{ax}}^{FS} = \varepsilon_{\text{ax}}^{\text{sample}} + \frac{\Delta \sigma}{E_{\text{ap}}}$$

Equation 1

where $\varepsilon_{\text{ax}}^{FS}$ is the average axial strain measured on gap sensors, $\varepsilon_{\text{ax}}^{\text{sample}}$ is the axial strain of the sample, $\Delta \sigma$ is the differential stress applied during the loading stage and $E_{\text{ap}}$ is the rigidity of the apparatus. The rigidity of the apparatus was estimated for each conditions tested, and depends of the applied confining pressure and of the ambient temperature.

Using the measurement of the axial shortening by capacitive gap sensors located externally combined with axial strain gage measurements, we are able to estimate the axial displacement from

$$D_{\text{ax}} = \varepsilon_{\text{ax}}^{\text{sample}} L = \left( \varepsilon_{\text{ax}}^{FS} - \frac{\Delta \sigma}{E_{\text{ap}}} \right) L$$

Equation 2

where $L$ is the length of the rock sample. The finite displacement along the fault during stick-slip instabilities, called also final displacement, is then estimate using a simple projection assuming

$$D_f = D_{\text{ax}} / \cos \theta$$

Equation 3

All displacements discussed in the main text correspond to displacement along the fault ($D_f$).

**Estimation of the fault's elastic stiffness**

For each events, the fault elastic stiffness was computed from axial strain measurement using strain gages. Fault elastic stiffness was measured following

$$K_f = \frac{\Delta \sigma}{L_f \varepsilon_1}$$

Equation 4

where $L_f$ is the length of the fault. Assuming these estimations, our experimental results show that the fault system elastic stiffness increases with the confining pressure, but also only with the ambient temperature (Figure S2a). Note that while this fault elastic stiffness is expected to control the rupture mode when its becomes greater than the stiffness of the apparatus (Leeman et al., 2016) (Figure S2b), it is not the case in our experiments and large $K_f$ recorded at 65 °C temperature and 30 and 60 MPa confining pressure induces only slow rupture while $K_f > K_{\text{ap}}$ (Figure S2).
Estimation of the slip velocities

Slip velocities were computed following two different methods due to technical limitations. For stable slip and slow ruptures, slip velocities were estimated from the displacement measurement corrected from the elastic stiffness of the apparatus (Equation 3), and consist simply in the derivative of $D_f$ with time.

During dynamic rupture, our sampling rate of the slip measurement does not allow a direct estimation of the slip velocity because most of the slip occurs within 20 microseconds. In these conditions, an estimate of the slip velocity is computed from the static slip measured at 2.4kHz, divided by the weakening time recorded by dynamic strain gages (Passelegue et al., 2016; Brantut et al., 2016). This estimate of the slip velocity has been confirmed recently using a laser vibrometer which now allow slip velocity measurement at high sampling rate. However, this system was not available at the stage of the experiments presented here.

Inversion of the dynamic rupture speed

While the rupture speed during slow rupture front was simply compute as a function of the travel time of the rupture through the array of strain gages located along the fault (Figures S1b and 2a), the inversion of the rupture speed during fast rupture required an inversion from the acoustic data. Due to the 3D experimental geometry, a complex mixed-mode rupture can develop. Under the applied loading conditions, the principal slip direction is parallel to the fault length (main axis of the ellipsoidal fault). As a consequence, a mode II (in-plane) rupture will propagate along the fault length, while a mode III (antiplane) rupture will propagate along its width. This implies that the rupture velocity, $V_r$, along the fault length (mode II) is such that $V_r < C_R$ during a sub-Rayleigh rupture and $C_S < V_r < C_P$ during a supershear rupture ($C_R$, $C_S$ and $C_P$ being the Rayleigh, shear and compressional elastic wave velocities respectively). In the Mode-III direction, along the fault width, rupture velocity is always $V_r < C_R$. For simplicity we approximate the shape of the rupture front as circular at sub-Rayleigh velocity and as elliptical at supershear velocity, where the ratio of the two axes corresponds to the ratio of the velocities in the in-plane direction.

For each STE, the first wave arrival recorded on each sensor was manually picked for better accuracy. The location and the rupture velocity are then inverted using a least-squares method comparing the experimental arrival times to theoretical arrival times calculated using the approximate rupture front geometry described above. To be more precise, the theoretical arrival times are calculated as a function of (i) the possible rupture velocity $V_r$ along strike (ii) the rupture front geometry (circular rupture front up to $V_r = C_R$, elliptical above $C_S$), (iii) the time of initiation of the event, and finally (iv) the positions of the sensors relative to the nucleation zone of rupture. We assume that the lowest residual time outputs the best solution for (i) the location of the nucleation zone, (ii) the time of initiation and (iii) the average rupture velocity along strike. Thereafter the value of the rupture velocity at a given point of the fault can be estimated during supershear event using
where \( V_R \) is the rupture velocity parallel to the slip direction and \( \alpha \) is the angle between the coordinates of the given point \((X, Y)\) and the mode-III direction perpendicular to the slip direction (Passelègue et al., 2016).

4. **Complementary experiments and microstructures**

Recent experimental studies have shown that in calcite, the transition from stable to unstable slip is explained by the activation of plastic processes. To understand further the link between the microstructures and the mode of slip, the formation of nanograins, and the mechanisms of deformation involved during the interseismic stage of the experiments, we conducted several specific experiments. The first experiments consisted in a simple elastic loading of the experimental fault, submitted at the three confining pressure and each temperature tested (30, 65 and 100 °C). The fault was loaded up to 90% of the expected differential stress leading to instability. Then, this stress was kept constant during a period ranging from 10 minutes to 1 hour. The fault was then unloaded and the sample removed from the apparatus.

The experimental results of these experiments highlighted that fault slip during constant load was observed only at low confining pressure and low temperature conditions, where the elastic fault stiffness present the lowest values. At 100 °C, conditions were dynamic rupture were observed in the experiments presented in the main text, the fault remains completely stick during the entire constant load experiments (no axial shortening), which suggest a pure elastic behavior.

Secondly, the post mortem microstructures conducted after these experiments highlighted that fault creep induced at low confining pressure induces the formation of mirror-like surface and gouge production along clivage. At 100 °C, where no slip was recorded, fault surface looks identical to the starting material and only very rare nanograins aggregates were observed (Figure S3). These results suggest that the nanograins observed after both slow rupture and dynamic ruptures are not present before the onset of the first instability at 100 °C and that plastic processes are not dominant along fault during the loading part of the experiments conducted at 100 °C, at the opposite of the behavior observed in calcite.

5. **Post rupture experiments microstructures**

Post mortem fault surfaces

Remarkably, the fault surfaces recovered from experiments conducted at different ambient temperature and which have experienced stable slip and slow rupture or dynamic rupture are very different even at the naked eyes (Figures S4a, S4b and S4c). Fault surfaces recovered from experiments where stable slip and
slow ruptures occurred have highly light reflective patches visible (Figure S4b). Instead, these extended mirror-like structures are not found on fault surfaces which experienced fast ruptures (Figure S4c). In fact, these fault surfaces are on average rougher but still characterized by the presence of limited in extension and discontinuous smooth patches with visible slicken-lines (See Figure S4d-I for SEM images at the same scale). This remarkable microstructural difference implies that weakening during faulting was induced by distinct processes in slow versus fast rupture propagation. Punctual chemical analysis by energy-dispersive spectroscopy (EDS) shows that residual dolomite grains after dynamic rupture events present similar fraction of O and Ca or O and Mg, with residual fraction of bot Mg and Ca similarly to analysis conducted after high velocity friction experiments (Green et al., 2014). These punctual analyses could suggest that the residual grains are formed of oxydes (MgO and CaO) due to fast rupture events, which may suggest decarbonation-assisted melting (Brooker et al., 1998) of the slip surface due to frictional heating. However, it remains not clear if melting occurs before decarbonation or after.

5. Determination of weakening velocity for flash heating

As stated in the caption of the figure 4d, the weakening velocity allowing flash heating depends of the hardness of dolomite, of the weakening temperature (decarbonation or melting temperature for exemple) and of the size of the asperities. While we have an idea of both the hardness and the weakening temperature of dolomite, an estimate of the size of the asperities acting along the fault prior instabilities remains hard to estimate. From literature, this size depends of minerals, and of the normal stress acting along the fault. In the manuscript, we consider a size of asperities of 50 microns based on microstructural analysis and roughness measurement of the post mortem fault surfaces which experienced stick slip events and strong weakening (Figure S5). However, this calculation is made in the text to show that the slip rate estimated during dynamic rupture is largely faster than critical values allowing flash temperature (Goldsby and Tullis, 2011; Tisato et al., 2012; Violay et al., 2014; Passelegue et al., 2014). According to our estimates of slip rates reached during dynamic ruptures, we can estimate the minimum size of asperities which allow decarbonation of dolomite following

\[ d_{min} = \frac{(\rho C)^2(T_w - T_0)^2\pi\alpha}{V_s} \]

Equation 6

with \( V_s \) the average rupture speed of all dynamic rupture events, \( T_w \) the weakening temperature corresponding to decarbonation. Doing this calculation, we can state that asperities larger than 0.16 microns are expected to decarbonate during fast ruptures, which is largely below the range of the imposed and measured roughness, and below the size of the asperities imaged by SEM analysis along post-mortem fault surfaces.
6. Supplementary figures

Figure 1. Starting material and experimental setup used in this study. (a) Picture of a saw cut specimens of natural dolomite. (b) Scheme presenting the fault geometry and the sensors used to investigate rupture processes. Quarter bridge were recorded at 2.4 kHz sampling rate, allowing to track slow rupture front. Both Wheatstone bridge and acoustic signals were recorded at 10 MHz sampling rate, allowing to track fast motions during dynamic instabilities.
Figure 2. Relation between the state of stress, the bulk temperature and the elastic loading stiffness of the fault (a) and the stiffness of the apparatus (b). Note that the ratio of the two stiffness measured is expected to be a proxy for the mode of slip.
Figure 3. Post mortem microstructures observed after elastic loading of the fault at 90 percent of the peak strength at 60 MPa confining pressure and room temperature (a. and b.), and at 60 MPa and 100 °C (c and d). Note that d corresponds to the only evidence of “nanograins” observed on the fault surface after the elastic loading, suggesting that nanograins are absent, and not required, prior the first dynamic instability.
Figure 4. Post mortem fault microstructures. (a), (b) and (c) corresponds to picture of the fault before, after stable slip or slow rupture, and after dynamic rupture, respectively. Note that macroscopic mirror-
like surface are observed only after stable slip and slow rupture events. (d), (e) and (f) correspond to micrographs of the fault surfaces after stable slip events induced at 100 MPa confining pressure and at room temperature. (g), (eh) and (i) correspond to micrographs of the fault surfaces after dynamic events.
Figure 5. Roughness of the fault that has undergone dynamic rupture events. Note that contact sliding area can be estimate around 25 to 100 microns in size.