

1 **The role of fault network geometry on the complexity of**
2 **seismic cycles in the Apennines**

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13

14 **Abstract.** Estimating the recurrence intervals and magnitudes of earthquakes for a given fault is essential for
15 seismic hazard assessment but often challenging due to the long recurrence times of large earthquakes. Fault
16 network geometry (i.e. spatial arrangement between faults) plays a key role in modulating stress interactions and,
17 consequently, earthquake recurrence and magnitude. Here, we investigate these effects of fault network geometry
18 using earthquake cycle models to generate numerous earthquakes on two different networks of normal faults in
19 Italy: the Central Apennines, characterised by a wide network of faults offset across strike, and the Southern
20 Apennines, a narrow fault network where faults are predominantly arranged along strike. For each region, we ran
21 an earthquake cycle simulation on systems of seven normal faults generating approximately 150 earthquakes. In
22 the Central Apennines, co-seismic stress transfer between faults promotes more heterogeneous stress, more partial
23 ruptures, greater M_w variability and less periodic behaviour of large earthquakes (coefficient of variation of
24 recurrence time, CV 0.1-0.9). In contrast, faults in the Southern Apennines experience more homogeneous stress
25 loading, leading to a higher proportion of full-fault ruptures with more regular recurrence intervals (CV 0-0.4). In
26 both fault networks, high long-term slip rate amplifies the effects of fault interactions: faults with higher long-
27 term slip rate are more sensitive to positive stress perturbations from nearby faults compared to slower-moving
28 faults. These results highlight that incorporating stress interactions from fault network geometry into seismic
29 hazard models is particularly important for networks of faults offset across strike, where rupture behaviour is more
30 variable.

31 **1 Introduction**

32 Earthquake recurrence intervals and magnitude distributions are key inputs for probabilistic and time-dependent
33 seismic hazard assessment (Gerstenberger et al., 2020), therefore determining the mechanisms behind their
34 variability is necessary to effectively mitigate associated risks. Fault stress interactions, which occur as stress is
35 redistributed following earthquakes (Harris and Simpson, 1998; King et al., 1994; Stein, 1999), influence the
36 recurrence and magnitude of earthquakes. These interactions can be either permanent (Coulomb static stress
37 transfer, CST) or transient (associated with seismic wave propagation; Freed, 2005), and may either promote or
38 inhibit rupture depending on the relative location and orientation of faults (e.g., Harris and Simpson, 1998; King
39 et al., 1994; Nicol et al., 2005; Stein, 1999).

40 Several studies have investigated how fault network geometry shapes stress interactions (Cowie et al., 2012;
41 Rodriguez Piceda et al., 2025a; Sgambato et al., 2020a, 2023). In particular, here we distinguish two basic spatial
42 arrangements of fault networks: “along-strike fault networks” in which faults are co-planar, aligned in the along-
43 strike direction, and “across-strike fault networks” in which faults are non-co-planar, offset in the across-strike
44 direction. Sgambato et al. (2020, 2023) used CST analysis of historical earthquakes in Italy to show that faults
45 with more across-strike interactions develop more irregular stress loading histories, less dominated by interseismic
46 stress loading, whereas faults with fewer across-strike interactions tend to have more regular stress histories
47 controlled by interseismic loading. While these patterns suggest a link between stress loading history and the
48 regularity of earthquake recurrence, it remains difficult to isolate the effects of network geometry using historical
49 data alone, since other processes such as dynamic stress transfer from seismic waves, postseismic stress
50 redistribution and fluid pressure changes (Freed, 2005) can obscure stress patterns driven by network geometry.

51 To explore the effects of network geometry under controlled conditions, numerical modelling approaches have
52 been employed. For instance, studies with elastic-brittle lattice models (e.g., Cowie et al., 2012) found that more
53 across-strike interactions lead to shorter earthquake recurrence intervals. However, these models do not include
54 the effects of time-dependent stress healing and frictional memory as observed in natural fault rocks (Marone,
55 1998b), resulting in less realistic stress-loading histories, nor they address the effects of fault geometry in
56 earthquake magnitudes and the partition between seismic and aseismic slip modes.

57 These effects are accounted for in models of sequences of earthquakes and aseismic slip (SEAS), incorporating
58 rate-and-state friction, enabling a more realistic representation of nucleation, healing and slip mode variability
59 (Erickson et al., 2020; Jiang et al., 2022; Lapusta et al., 2000; Rice, 1993). For example, Romanet et al., (2018)
60 modelled two partially overlapping strike-slip faults with equal long-term tectonic slip rate and found that smaller
61 across-strike separations can produce complex earthquake sequences and slow slip events. Yin et al. (2023)
62 extended this to three overlapping strike-slip faults, showing that fault interactions promote aperiodic cycles and
63 partial ruptures (i.e. ruptures that do not break the whole fault). Rodriguez Picada et al. (2025a) focussed on the
64 role of along-strike and across-strike separation on the earthquake cycle using a simplified system of two parallel
65 normal faults with identical loading conditions. They showed that across-strike separation has a greater impact on
66 recurrence intervals than along-strike separation, with closer across-strike faults characterised by more complex
67 and non-periodic seismic cycles.

68 Natural fault networks, however, are geologically more complex than in generic models: they are often formed
69 by more than three faults, are more geometrically diverse, and long-term slip rates vary between faults (e.g. Nicol
70 et al., 2016; Papanikolaou et al., 2005; Papanikolaou and Roberts, 2007; Roberts, 2007; Roberts and Michetti,
71 2004). The combined effect of such complexities on earthquake timing, magnitude and slip modes remains poorly
72 understood, as does the role of slip rate in modulating interaction effects. These research gaps have direct
73 implications for seismic hazard assessment.

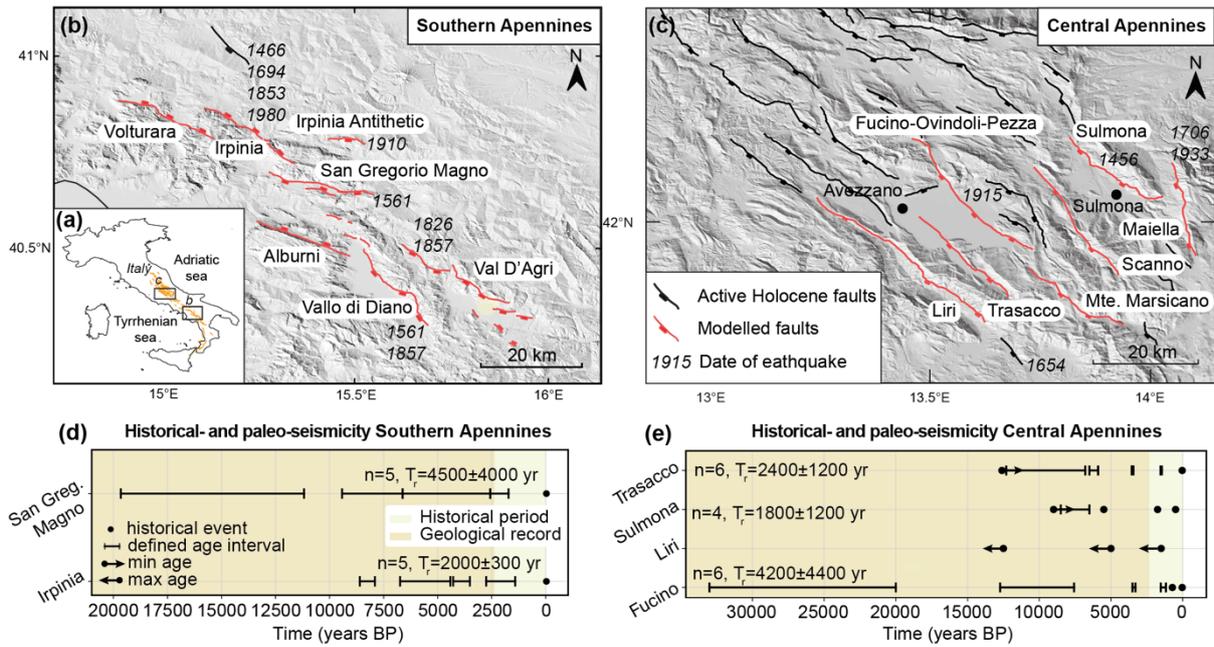
74 Here we address these questions by introducing a novel approach developing 3D SEAS simulations of over 100
75 earthquakes based on the Southern and Central Apennines normal-fault networks in Italy, an ideal natural
76 laboratory due to its diverse network geometries and fault slip rates constrained by field data (Faure Walker et al.,
77 2021; Mildon et al., 2019; Sgambato et al., 2020b) and exceptional historical and paleoseismic records (Galadini
78 and Galli, 1996, 1999a; Galli et al., 2015; Galli, 2020; Guidoboni et al., 2019; Pantosti et al., 1993; Rovida et al.,
79 2020). This approach overcomes several of the simplifications of previous studies by combining realistic fault
80 network geometries, field-derived loading rates and spontaneous rupture propagation within a continuum
81 framework. We investigate the effects of network geometry and long-term slip rate on earthquake recurrence,
82 magnitude distributions, and slip behaviour, and discuss the implications for incorporating such variability into
83 time-dependent seismic hazard models.

84 **2 Geological background**

85 The Apennines fold-and-thrust belt developed during the Neogene and Quaternary due to the convergence
86 between the Eurasian and African plates (Anderson and Jackson, 1987; Doglioni, 1993; Jolivet et al., 1998). The
87 thrusting phase was followed by extensional tectonics during the Pliocene (~2-3 Ma; Cavinato & Celles 1999;

88 Roberts & Michetti 2004). This extensional phase, continuing to the present day, led to the formation of NW-SE
89 oriented normal faults that run nearly parallel to the fold-and-thrust belt. The tectonically active faults have formed
90 surface offsets preserved since the Last Glacial Maximum (12-18 Ka), allowing the estimation of throw rates from
91 exposed Holocene fault scarps (Papanikolaou and Roberts, 2007; Roberts and Michetti, 2004). In the Southern
92 Apennines, average throw rates on individual faults range between 0.2 and 0.67 mm/yr (Faure Walker, 2010;
93 Papanikolaou and Roberts, 2007; Roberts and Michetti, 2004; Sgambato et al., 2020a, b), while in the Central
94 Apennines, they range between 0.2 and 1.56 mm/yr (Faure Walker et al., 2009; Faure Walker, 2010; Faure Walker
95 et al., 2019a, 2021; Mildon et al., 2019; Papanikolaou and Roberts, 2007; Roberts and Michetti, 2004). The
96 different slip rates among faults allowed us to assess how slip rates modulate or amplify stress transfer and
97 interaction between faults.

98 A subset of faults in the Southern and Central Apennines was chosen for computational efficiency. The chosen
99 fault networks differ in the number, length and orientation of faults. The chosen sector in the Central Apennines
100 is characterised by a wide fault network, with up to 8 faults arranged NW-SE (Fig. 1b), whereas the Southern
101 Apennines has a narrower fault network with fewer faults accommodating the extension, with a maximum of 3
102 across-strike faults (Fig. 1c). These faults are associated with moderate to large magnitude earthquakes (Mw 5.5-
103 7; e.g. Bagh et al., 2007; Chiaraluce et al., 2005, 2022; Guidoboni et al., 2019). In the Southern Apennines, the
104 historical record documents 9 earthquakes of Mw>5.6 since 1466 associated with the studied faults (Fig. 1c, Table
105 B1; Cello et al., 2003; Galli et al., 2006, 2014; Galli and Peronace, 2014; Giardini et al., 1996; Rovida et al., 2020;
106 Westaway, 1993; Westaway and Jackson, 1987). Two multi-fault seismic sequences have occurred in the region:
107 one in 1857 of estimated Mw 7.1 involving either the Vallo di Diano and the Val D'Agri faults (Benedetti et al.,
108 1998; Cello et al., 2003; Galli et al., 2006a) or the Caggiano-Montemurro fault (Bello et al., 2022); and a second
109 in 1980 of Mw 6.81 that ruptured the Irpinia (also known as Monte Marzano fault, e.g., Galli, 2020), and Irpinia
110 antithetic faults, as well as possibly the San Gregorio fault (Sgambato et al., 2025). This rupture included a
111 northeast dipping segment of the Irpinia fault known as the Pantano di San Gregorio segment (D'Adezzio et al.,
112 1991, Fig. 1b). The 1980 event was the largest earthquake instrumentally recorded in the Apennines (Bernard and
113 Zollo, 1989). In the studied segment of the Central Apennines, the historical record includes 7 earthquakes of
114 Mw>5.6 since 1456 associated with the studied faults, including the 1915 A.D Fucino earthquake. Paleoseismic
115 data is available for 1 and 4 of the modelled faults in Southern and Central Apennines, respectively (Fig. 1d-e,
116 Table B2).



117

118 **Figure 1: (a) Location of selected fault regions in Italy; (b-c) Map of the fault regions in the Apennines and**
 119 **historical earthquakes (post 1400AD). Active Holocene fault traces in the (b) Southern Apennines (based**
 120 **on Sgambato et al., 2020b) and (c) Central Apennines (based on Faure Walker et al., 2021), showing the**
 121 **different fault network geometries between the regions, with few across-strike faults in the Southern**
 122 **Apennines and multiple across-strike faults for Central Apennines. Dashed black lines show the debated**
 123 **link between Vallo di Diano and Auletta faults and Auletta and Caggiano-Montemurro faults (see section**
 124 **3.1). (d-e) Chronology of paleoseismic and historical events with $M > 6$ in modelled faults of the (d) Southern**
 125 **Apennines and (e) Central Apennines (see Table B2 for details; (Galadini and Galli, 1999a, a; Galli et al.,**
 126 **2008, 2015, 2016; Pace et al., 2020; Pantosti et al., 1993). Whisker plots represent the estimated time range**
 127 **of seismic events. The number of paleoseismic events per fault with defined aged brackets (n), and their**
 128 **estimated mean and standard deviation of recurrence time (T_r) is also shown.**

129 3 Methods

130 3.1 Model set-up

131 We use the boundary-element software QDYN (Luo et al., 2017) to model SEAS on the Southern and Central
 132 Apennines fault networks, each composed of 2D normal faults governed by rate-and-state friction, embedded in
 133 a 3D elastic medium (see Appendix A for a description of the governing equations).

134 We model 7 faults in the Southern Apennines (Alburni, Irpinia, Irpinia Antithetic, San Gregorio Magno (here
 135 referred as “San Gregorio”), Val D’Agri, Vallo Di Diano and Volturara faults (Fig. 2a)) and 7 faults in the Central
 136 Apennines (Fucino-Ovindoli-Pezza (here referred as “Fucino”, also known as San Benedetto dei Marsi-Goia dei
 137 Marsi segment by Galadini and Galli (1999)), Liri, Maiella, Monte Marsicano, Scanno, Sulmona (also known as
 138 Monte Morrone fault by Galli et al. (2015)) and Trasacco faults (Fig. 2b)). All selected faults are active, with
 139 documented Holocene throw (Faure Walker et al., 2021; Sgambato et al., 2020a; Valentini et al., 2017). In
 140 preliminary trials, we identified that the inclusion of three other active faults in the Central Apennines (Parasano-

141 Pescina, Tremonti and San Sebastiano faults) led to unrealistically high stress concentrations in the Fucino fault,
142 which caused numerical instability forcing the simulation to stop prematurely. For these computational reasons,
143 we exclude the three active faults from the Central Apennines model. The potential implications of this exclusion
144 are addressed in the Discussion section.

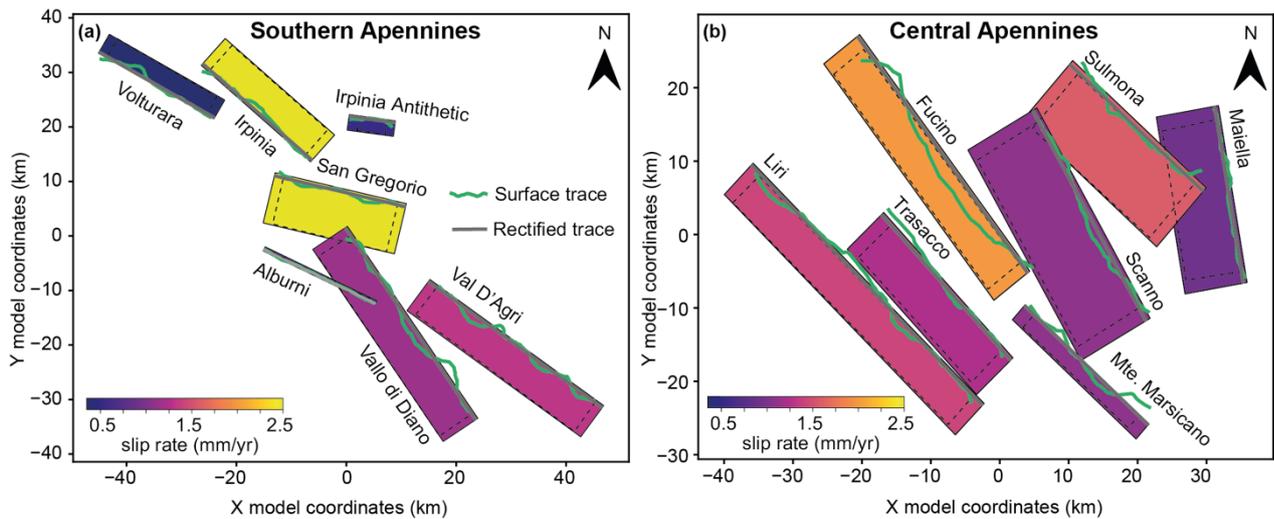
145 The trend and average dip of each fault is taken from published sources (Fault2SHA database, Faure Walker et
146 al. (2021); Mildon et al., (2019); and Sgambato et al., 2020b) (Table 1). For the Central Apennines, we use the
147 definition of “main fault traces” of Fault2SHA, which represent how faults segments have been interpreted to be
148 linked at depth, at the scale recommended for input into hazard models (Faure Walker et al., 2021). For the
149 Southern Apennines, which are not covered by the Fault2SHA database, we utilise fault traces from Sgambato et
150 al., (2020b). The geometry and extent of some of these fault traces is debated. For instance, in some interpretations
151 the NW segment of Vallo di Diano fault (also known as Auletta fault or Caggiano fault) is a separate segment
152 from this fault and continues SE (Galli et al., 2006a) or even joins with Val D’Agri fault as part of the Caggiano-
153 Montemurro fault system (Bello et al., 2022) (Fig. 1b). Here, we follow the interpretation of Sgambato et al.
154 (2020b), where the NW segment is part of Vallo di Diano fault based on the field-based evidence of the slip vector
155 orientations (Papanikolaou and Roberts, 2007).

156 To accommodate QDYN current restriction to a uniform strike for each dip-slip fault, we rectify the fault traces
157 (Fig. 2). The fault depths are set in most of the cases to 15 km, which is taken as approximately the depth of the
158 brittle-ductile transition and is consistent with hypocentral depths observed in the region (Chiarabba et al., 2005;
159 Chiaraluce et al., 2005; Frepoli et al., 2011). We make two exceptions for fault geometry parameters. The Irpinia
160 Antithetic fault, being relatively short (length < 15 km), is assigned a depth of 8 km to maintain an aspect ratio of
161 1 (Nicol et al., 1996). This modification has implications for the generation of full ruptures, as addressed in the
162 Results and Discussion sections. We limited the Alburni fault to a depth of 3 km to prevent numerical instabilities
163 caused by the intersection with the Vallo di Diano fault. This approximation may lead to an underestimation of
164 its potential involvement in multi-fault rupture sequences and stress transfer, especially near the Vallo di Diano
165 fault. However, the Alburni fault is not thought to have ruptured in historical times, but it is considered to have
166 been active between late Pliocene and late Pleistocene (Gioia et al., 2011; Soliva et al., 2008).

167 We assign a constant loading rate value to each fault, corresponding to the maximum Holocene slip-rate (Figure
168 2; Cinque et al., 2000; Faure Walker, 2010; Faure Walker et al., 2019, 2021; Galli et al., 2014; Morewood and
169 Roberts, 2000; Papanikolaou et al., 2005; Roberts and Michetti, 2004; Valentini et al., 2017). Preliminary trials
170 conducted in this study and from previous work (e.g., Yin, 2022), show that in the implicit LSODA solver used
171 here (Yin et al., 2023), slip rates below 1 mm/yr can cause velocity values to drop below numerical precision,
172 especially when neighbouring faults also exhibit low slip rates. We suspect that this is due to the difference
173 between the minimum and maximum slip rates present in the system, which span many orders of magnitude,
174 causing precision underflow when jointly solving the system of equations. Therefore, we scale up the loading rate
175 by a factor of 100 which represents the minimum value required to ensure numerical stability across the full fault
176 network, while remaining as close as possible to the geologically inferred loading rates. Based on these trials and
177 on previous studies (Yin, 2022) we find that this scaling has minimal impact on seismicity statistics, once we

178 correct simulations outputs (such as recurrence times) for this upscaling factor (see Appendix B). All time scales
 179 reported hereafter account for this correction factor.

180 The set up does not contain further complexities such as variable fault geometry, heterogeneous slip-rate
 181 distribution along the fault strike, and or complex heterogeneous distribution of frictional properties. Although
 182 these elements may generate heterogeneous stress concentrations that might act as barriers to rupture propagation
 183 or as regions of earthquake nucleation (Delogkos et al., 2023; Hillers et al., 2007; Luo and Ampuero, 2018; Mildon
 184 et al., 2019; Rodriguez Piceda et al., 2025a), we chose to omit these complexities to primarily focus on the impact
 185 of fault network geometry and fault stress interactions on rupture dynamics.



186
 187 **Fig. 2:** Map view of the model geometry for the (a) Southern Apennines and (b) Central Apennines fault
 188 network. Fault planes are colour-coded according to their Holocene slip rate. Black dashed lines show the
 189 velocity-weakening asperities. Surface fault traces (green) derived from (Faure Walker et al., 2021; Mildon
 190 et al., 2019; Sgambato et al., 2020b) and rectified traces (grey) used in the model are also shown.

191
 192 **Table 1:** Geometry and long-term slip rate of modelled fault sources in the Southern and Central Apennines.
 193 References for slip rate: [1] (Valentini et al., 2017) [2] (Faure Walker, 2010) [3] (Galli et al., 2014) [4] (Cinque
 194 et al., 2000) [5] Fault2SHA database (Faure Walker et al., 2021) [6] (Morewood and Roberts, 2000) [7] (Roberts
 195 and Michetti, 2004) [8] (Faure Walker et al., 2019a) [9] (Papanikolaou et al., 2005)

Modelled region	Fault	Map trace length (km)	Average fault dip (deg)	Maximum long-term slip rate (mm/yr)
	Alburni fault	23	74	0.7 [1] [2]

Southern Apennines	Irpinia fault	27	66	2.5 [1] [3]
	Irpinia Antithetic fault	8.5	70	0.53 [1] [2]
	San Gregorio fault	24	58	2.5 [1] [3]
	Val D'Agri fault	39	64	1.3 [1] [2]
	Vallo Di Diano fault	42	63	1.15 [1] [4]
	Volturara fault	24	75	0.35 [1] [2]
Central Apennines	Fucino fault	40	66	1.99 [5] [6]
	Liri fault	47	68	1.41 [5] [7]
	Maiella fault	24	58	0.92 [5] [7]
	Monte Marsicano fault	24	7	1.08 [5] [7]
	Scanno fault	33	52	1.08 [5] [8]
	Sulmona fault	26	54	1.57 [5] [9]
	Trasacco fault	27	64	1.22 [5] [7]

196

197 The simulations run for 11 kyrs to ensure a sufficient number of seismic events for statistical analysis, with the
198 first 500 years discarded as the spin-up phase. To compute the duration of a seismic event, we consider that it

199 starts when one fault element has a slip rate larger than 0.1 m/s and stops when the slip rate of all the elements
200 slip drops below 0.01 m/s.

201 We additionally simulate SEAS on each individual isolated fault included in the Southern and Central Apennines
202 networks to determine their reference behaviour in the absence of stress interactions with other faults. These
203 simulations use the same parametrization as the full fault network simulations described above.

204 3.2 Fault network and seismic cycle characteristics

205 To quantify the effect of across-strike faults, we compute an across-strike interaction index (AI) for each fault i
206 as:

$$207 \quad AI_i = \sum_{j \neq i} \frac{1}{s_{ij}} \quad (1)$$

208 where j are the indices of other across-strike faults, s_{ij} is the across-strike separation between fault i and fault j
209 (see Appendix C for a detailed definition of the separation between faults). The inverse weighting $\frac{1}{s_{ij}}$ ensures that
210 faults that are closer contribute more to the index than those farther away. Faults with a larger number of across-
211 strike interactions have a higher across-strike interaction index. We focus on across-strike density since previous
212 work (Rodriguez Picada et al., 2025a) showed that across-strike interactions dominate over along-strike
213 interactions at comparable distances.

214 To characterise the complexity of seismic cycles, we compute three metrics: the coefficient of variation of
215 recurrence times of individual faults (CV_{Tr}), the normalised number of partial ruptures (N_p') and the coefficient
216 of variation of rupture lengths (CV_{RL}').

217 CV_{Tr} is calculated as:

$$218 \quad CV_{Tr} = \frac{std(T_r)}{mean(T_r)} \quad (2)$$

219 where T_r is the distribution of time intervals between consecutive events on the same fault. $CV_{Tr} = 0$ indicates
220 strictly periodic seismic cycles; $0 < CV_{Tr} < 0.5$, strongly periodic; $0.5 \leq CV_{Tr} \leq 1$, weakly periodic; $CV_{Tr} = 1$ indicates
221 that event timing is random and independent of other events; and $CV_{Tr} > 1$ implies event clustering (Boschi et al.,
222 1995).

223 N_p' is calculated as:

$$224 \quad N_p' = \frac{N_p}{N (W_s/L_\infty)} \quad (3)$$

225 where N_p is the number of partial ruptures, N the total number of events for each fault, W_s the seismogenic width
226 and L_∞ the nucleation length (Eq. A5) introduced by Rubin and Ampuero (2005).

227 CV_{RL}' is calculated as:

228

$$CV_{RL}' = \frac{std(RL)/mean(RL)}{(W_s/L_\infty)} \quad (4)$$

229

where RL is the distribution of rupture lengths.

230

Both N_p' and CV_{RL}' are normalised by W_s/L_∞ to account for fault-to-fault differences in seismogenic width W_s and the nucleation length L_∞ , enabling comparison of partial ruptures and rupture length variability (Cattania, 2019; Barbot, 2019). Overall, faults with larger CV_{Tr} , N_p' , CV_{RL}' have seismic cycles characterised by less periodic recurrence, more frequent partial ruptures and a wider range of rupture sizes.

234

235 4 Results

236 4.1 Seismic cycles on isolated faults

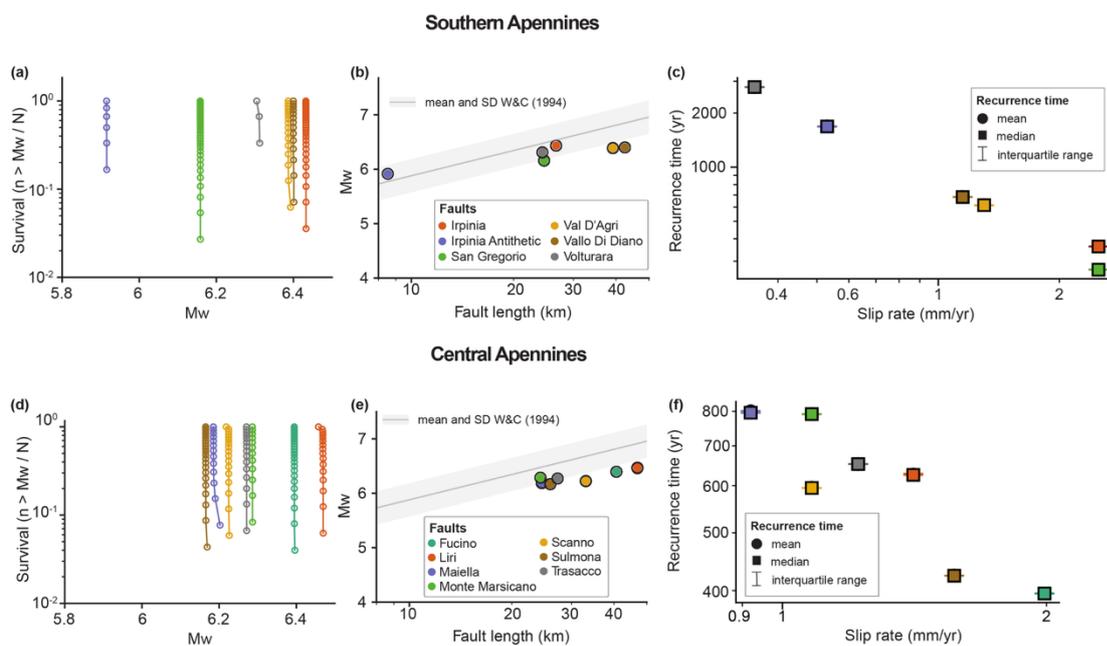
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In all single-fault simulations, faults rupture with a characteristic magnitude M_w (Fig. 3a,d), which is linearly related to the logarithm of fault length (Fig. 3b,e). In the Southern Apennines, M_w on individual faults ranges from 5.9 for the Irpinia Antithetic fault to 6.4 for the Irpinia and Vallo di Diano faults (Fig. 3a). In the Central Apennines, M_w values range from 6.2 on the Sulmona fault to 6.5 on the Liri fault (Fig. 3d). Notably, most earthquakes in both fault networks have magnitudes smaller than those predicted by empirical relationships of subsurface rupture length vs. magnitude (Fig. 3b,e; Wells and Coppersmith, 1994).

243

The resulting seismic cycles are periodic, with recurrence intervals inversely correlated to the prescribed long-term slip rate (Fig. 3c,f). Mean recurrence times range from 300 years in San Gregorio fault to ~1700 years in Volturara fault in the Southern Apennines (Fig. 3c); and from 400 years in Fucino fault to 800 years in Maiella fault in the Central Apennines (Fig. 3f).

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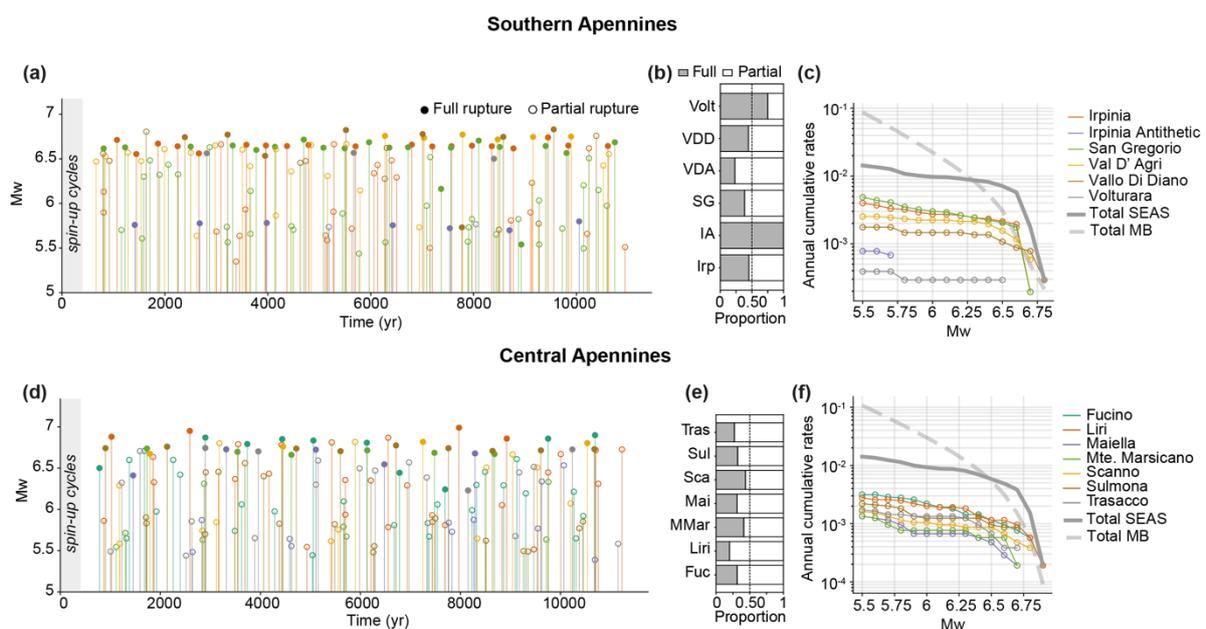


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248 **Figure 3: Earthquake cycle simulations of isolated faults in the (a-c) Southern and (d-f) Central Apennines.**
 249 **(a, d) Magnitude-frequency distributions of earthquakes on each fault (shown as survival functions:**
 250 **number of events with a Mw larger than a given value, normalised by the total number of events).** Their
 251 **near vertical appearance indicates all isolated faults in this model have characteristic-earthquake**
 252 **behaviour, with very narrow range of Mw. (b, e) Comparison between modelled seismic events and mean**
 253 **and standard deviation of empirical relationships of subsurface rupture length vs. Mw (Wells and**
 254 **Coppersmith, 1994); (c, f) Mean, median and interquartile range of earthquake recurrence times vs. long-**
 255 **term tectonic slip rate of individual faults. Note how the mean and median are equal, and the interquartile**
 256 **range is below the marker size for all faults across both regions.**

257 4.2. Seismic cycles on fault networks

258 We generated two synthetic seismic catalogues for the interacting fault networks, containing 150 events in the
 259 Southern Apennines and 154 events in Central Apennines (Fig. 4a,d). The magnitude range produced by faults
 260 within the networks is broader than that of isolated faults: from Mw 5.3 to 6.8 in the Southern Apennines (Fig.
 261 4c) and from Mw 5.3 to 7 in the Central Apennines (Fig. 4f), broadly matching observed earthquake magnitudes.
 262 In our Southern Apennines model, the Alburni fault does not produce any earthquakes. This is likely due to its
 263 shallow seismogenic zone (down to a depth of 2.65 km). Unlike the isolated fault models, not all ruptures extended
 264 for the whole length of the faults, with some ruptures terminating as partial ruptures. Most faults generate both
 265 full and partial ruptures, with a larger proportion of partial ruptures (Fig. 4b,e). Due to the generation of full and
 266 partial ruptures, the magnitude-frequency distributions show a truncated Gutenberg-Richter distribution (Stirling
 267 et al., 1996). The largest magnitude events are limited by the length of the longest fault, in our case, the Vallo di
 268 Diano fault in Southern Apennines (Fig. 4c) and the Liri fault in Central Apennines (Fig. 4f). The Irpinia Antithetic
 269 fault is the only fault to consistently generate full ruptures with a characteristic Mw of 5.8. This is likely due to
 270 its small fault dimensions (8 km in length and depth), which limits its potential for partial ruptures in the model
 271 (Barbot, 2019; Cattania, 2019; Cattania and Segall, 2019) (Fig 4b,c).

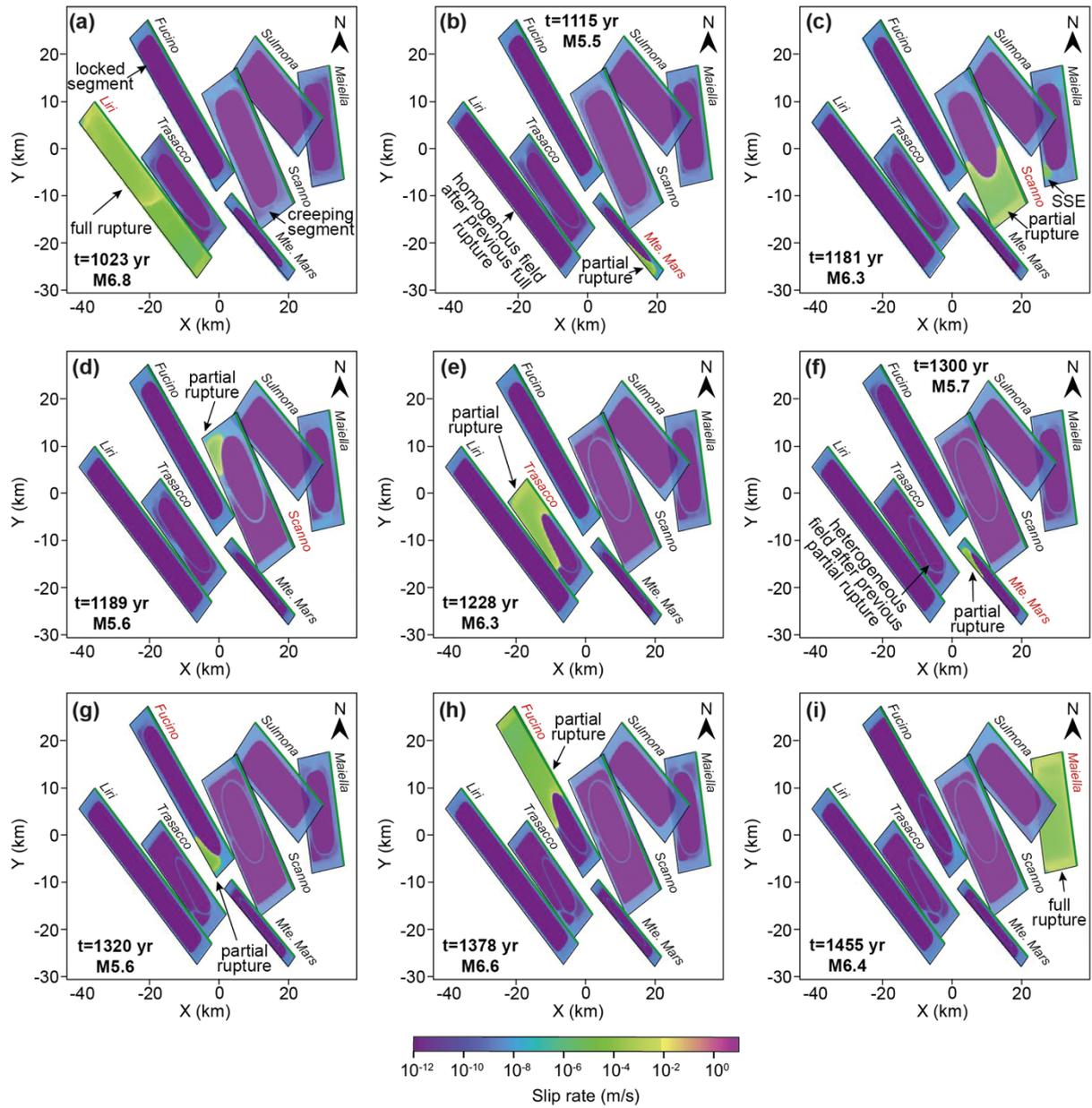


272

273 **Figure 4: Earthquake statistics from the model results for the Southern (a-c) and Central (d-f) Apennines**
274 **fault networks. Time distribution of simulated full- and partial-rupture events with stems and markers**
275 **colour-coded by fault in the (a) Southern Apennines and (d) Central Apennines. Earthquakes in the first 5**
276 **yr correspond to spin-up phases, thus are not included in the analysis. (b, e) proportion of full and partial**
277 **ruptures relative to the total number of events per fault in the (b) Southern Apennines and (e) Central**
278 **Apennines. (c, f) Magnitude–frequency distributions of seismic events expressed as annual cumulative rates**
279 **(number of annual occurrences with M_w greater than a given value) for individual faults in the (c) Southern**
280 **Apennines and (f) Central Apennines. Annual cumulative rates for the entire fault network derived from**
281 **the SEAS models, together with estimates obtained using the moment-budget method (MB; Appendix H),**
282 **are also shown. Colour legend for each fault is shown in panels (c) and (f).**

283

284 To highlight the rupture styles observed in our simulation, Figure 5 shows a subset of events that occur in the
285 Central Apennines between two full ruptures, one on the Liri fault and one on the Maiella fault. While a full
286 rupture occurs on one fault, the remaining faults stay locked (Fig. 5a). The full rupture affects both the locked and
287 surrounding creeping segment of the fault (Fig. 5a), and it is followed by partial ruptures in the remaining faults
288 (Fig. 5b-h). Some of these partial ruptures occur consecutively on the same fault (e.g. Fucino fault, Fig. 5g-h),
289 with some separated by hours (e.g. two events on the Vallo di Diano fault at 5450 yr and two events on the Monte
290 Marsicano fault at 8360 yr, not shown in Fig. 5, Videos S2, S3; (Rodriguez Piceda et al., 2026)). Subsequent
291 events in this type of sequence commonly rupture the fault segments that were not involved in the prior earthquake.
292 Simulated full and partial ruptures nucleate typically at the base of the locked seismogenic zone (Videos S2-S3,
293 (Rodriguez Piceda et al., 2026)). Overall, while full-rupture events homogenise the slip rate field in locked fault
294 patches (Fig. 5b), partial ruptures introduce a heterogeneous slip rate field in these patches (Fig. 5f), a direct
295 consequence of the stress concentration left behind by the arrested partial ruptures. The heterogeneous slip rate
296 then influences the nucleation and propagation of subsequent events (Video S2; (Rodriguez Piceda et al., 2026)),
297 acting either as nucleation sites or barriers where ruptures terminate. In addition to the observed earthquakes,
298 aseismic slip in the form of slow slip events sometimes occurs simultaneously with seismic events on other faults
299 (Fig. 5c, Video S2, (Rodriguez Piceda et al., 2026)). The occurrence of full and partial ruptures, as well as slow-
300 slip events, shows the more diverse slip behaviour of faults within a network compared to the isolated fault models.



301

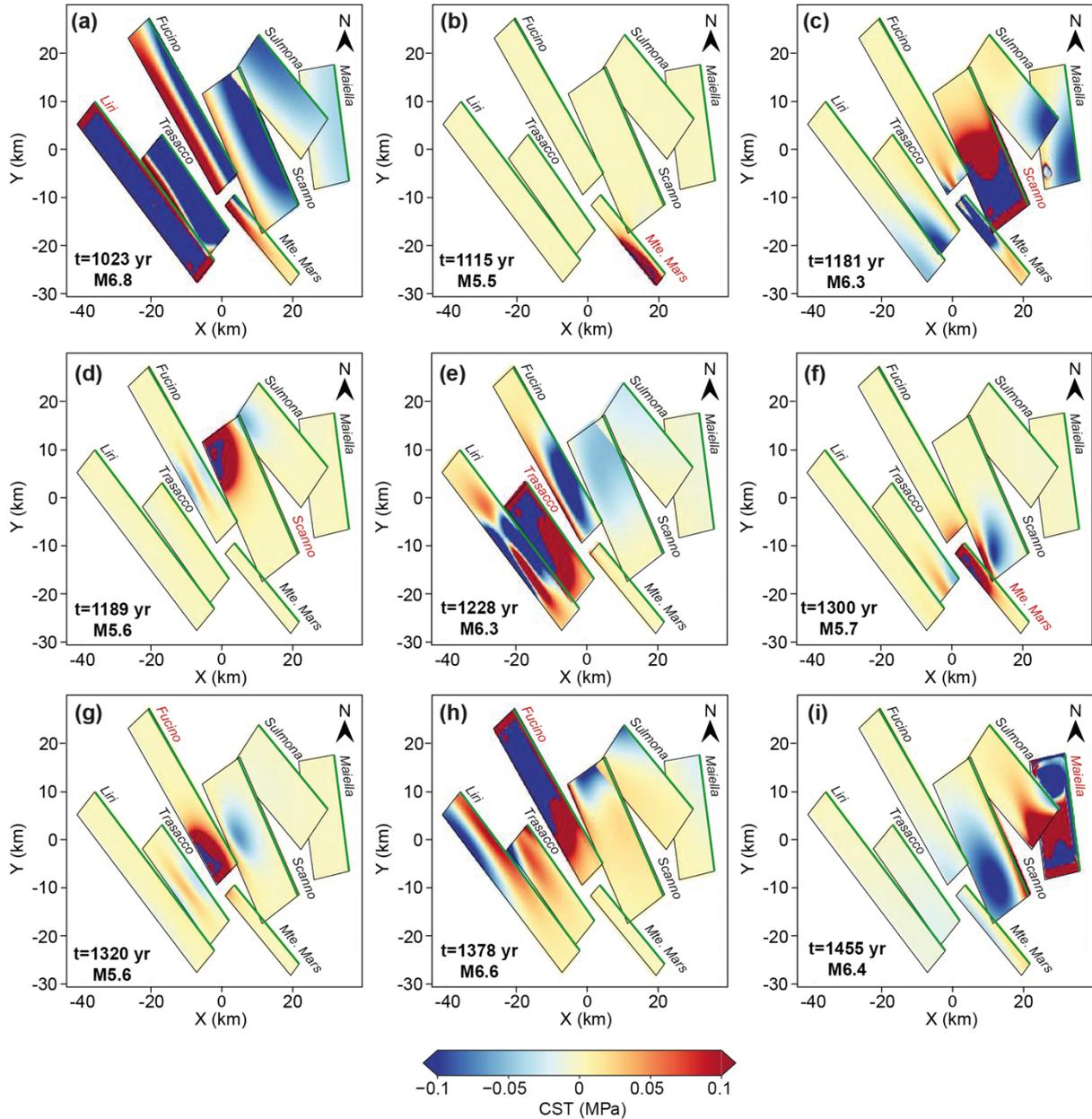
302 **Figure 5: Map view of slip-rate snapshots showing the final timestep of the coseismic phase for each event**
 303 **in a subset of modelled ruptures in the Central Apennines including full ruptures, partial ruptures and**
 304 **slow slip events (SSE). Monte Marsicano fault is abbreviated as Mte. Mars.**

305 To illustrate the stress changes introduced by each seismic event we computed the coseismic Coulomb stress
 306 transfer (CST) for each seismic event as (King et al., 1994):

$$307 \quad \Delta C = \Delta \tau - \mu(\theta, V) \Delta \sigma \quad (5)$$

308 where $\Delta \tau$ is the shear stress-change, $\Delta \sigma$ is the normal stress change before and after the earthquake and $\mu(\theta, V)$
 309 the rate-and-state friction coefficient at the receiver fault location. Figure 6 shows the CST for the same subset of
 310 modelled ruptures as in Fig. 5, and the CST evolution from all events in both fault network models are shown in
 311 Figs. S5 and S6. Most full and partial ruptures introduce a heterogeneous stress change on nearby faults, due to

312 the range of fault strike and partial overlaps between faults in the network (Fig. 6). In the rare cases where faults
 313 are almost parallel and with a near 100% overlapping area, a full-rupture event on one fault introduces a
 314 homogenous stress change. This can be observed for the Liri and Trasacco faults, where a full-rupture event on
 315 Liri fault negatively loads Trasacco fault (Fig. 6a).



316

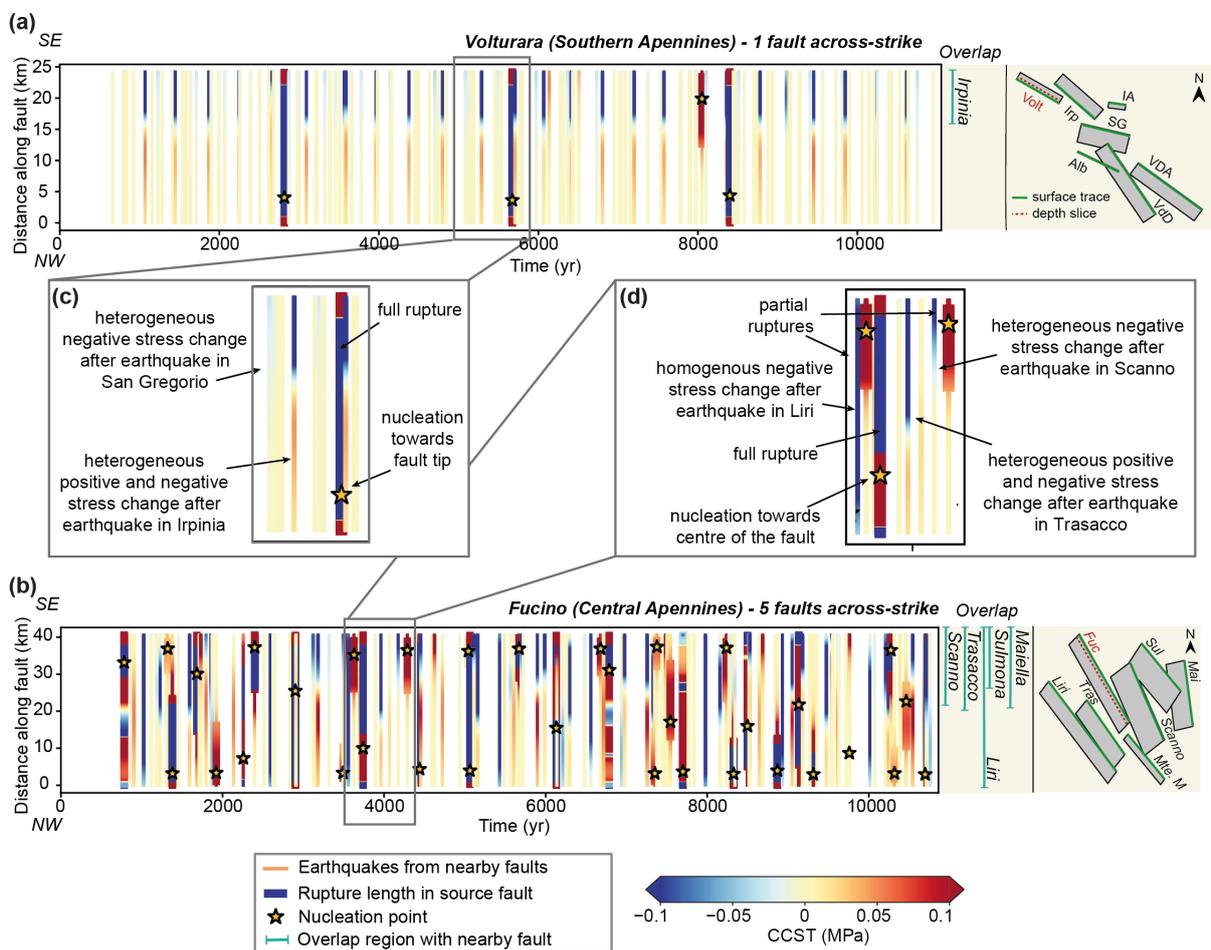
317 **Figure 6: Coseismic Coulomb Stress Transfer (CST) for a subset of modelled earthquakes in the Central**
 318 **Apennines shown in Fig. 5, with fault planes projected to a horizontal surface.**

319

320 Figure 7a-b shows the evolution of CST and nucleation point of each event along a depth slice near the top of the
 321 velocity-weakening asperities (~1700 m) of the Volturara and Fucino faults, compared to the extent of overlap
 322 with adjacent faults in the fault network (Figures S7 and S8 show the same but for the remaining faults). Faults

323 such as Volturara, which have few nearby, partially overlapping, across-strike faults, tend to show coseismic stress
 324 changes that remains broadly similar through successive earthquake cycles. Seismicity comprises mainly full
 325 ruptures that nucleate near fault tips (Fig. 7a, c). These nucleation locations correspond to regions of elevated
 326 stressing rates produced by the backslip loading method, which remain highest near fault tips despite the inclusion
 327 of velocity-strengthening buffer regions (Appendix A).

328 Conversely, faults with multiple neighbouring across-strike faults, such as Fucino fault (Fig. 7b, d), show a more
 329 spatially heterogeneous stress evolution and complex earthquake cycles consisting of full and partial ruptures.
 330 Additionally, the nucleation of full and partial ruptures is not necessarily confined to fault tips. Instead, partial
 331 ruptures that are either isolated or the first in a sequence often occur in areas where faults overlap. Therefore, the
 332 number and arrangement of across-strike faults, and the heterogeneous coseismic stress changes they induce, have
 333 a strong control on the earthquake cycle of individual faults.

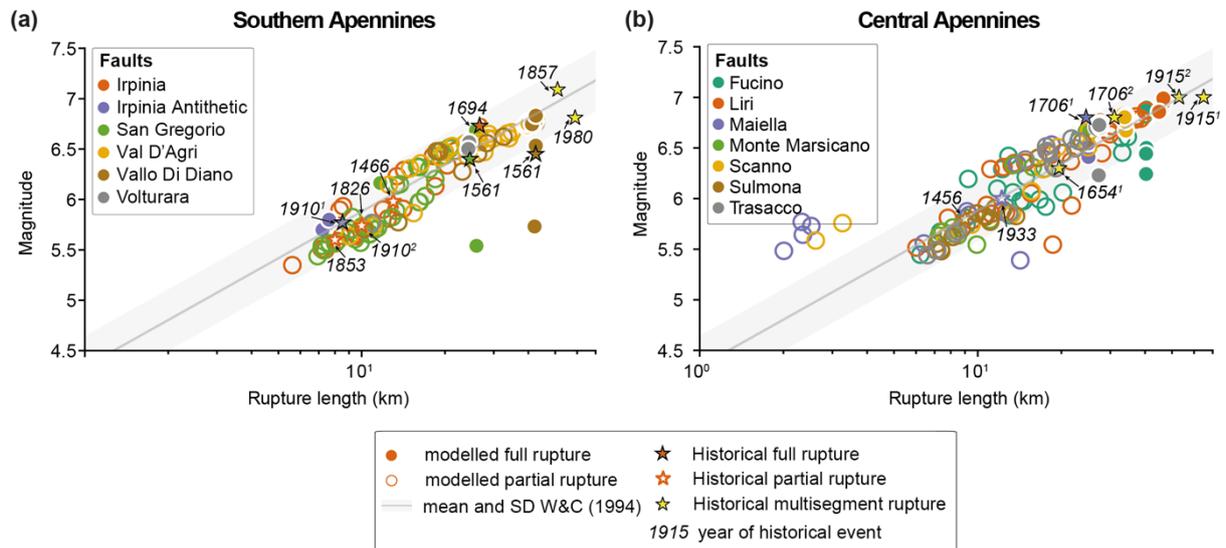


334

335 **Figure 7: (a) Coseismic Coulomb stress transfer (CST) on the Volturara fault, which has 1 across-strike**
 336 **fault that overlaps; (b) CST on the Fucino fault, which has 5 neighbouring across-strike faults that overlap.**
 337 **Coloured backgrounds show the stress changes following each event along a depth slice close to the top of**
 338 **the velocity weakening asperity (location example shown in right panels of Fig. 7a,b). Coseismic CST due**
 339 **to ruptures on the fault are indicated by the thick vertical slices that coincide with a star (nucleation point**
 340 **on the fault); thin vertical slices without corresponding star represent the coseismic CST due to ruptures**

341 on neighbouring faults. On the right side of panels (a) and (b) we indicate the fault segments that are
 342 overlapping with neighbouring faults (taken as the maximum length measured along-strike from the NW
 343 fault tip, see Fig. D2, Appendix D), including an inset with map view of fault surfaces. (c) and (d) are zoom-
 344 ins highlighting contrasting behaviours between the faults. The Volturara fault shows spatially
 345 homogeneous stress changes and simple seismic cycles, with full ruptures and nucleation near fault tips.
 346 The Fucino fault exhibits spatially complex coseismic CST patterns, partial ruptures, and nucleation that
 347 is more distributed across the fault. Faults with limited across-strike interaction tend to show simpler
 348 rupture behaviour, while those with multiple across-strike interactions show a more heterogeneous stress
 349 evolution and complex earthquake cycle.

350 We compared the magnitudes and rupture lengths of the modelled events with historical seismicity (Table B1,
 351 Appendix B) and empirical relationships between magnitude and subsurface rupture length (Wells and
 352 Coppersmith, 1994); Fig. 8a-b). Our models are able to reproduce the magnitude and rupture length of the
 353 historical single-fault events, including the two proposed scenarios for the 1910 earthquakes in Southern
 354 Apennines (Galli and Peronace, 2014). However, we are unable to reproduce multi-fault events as recorded in the
 355 historical seismicity catalogue (e.g. the 1857 and the 1980 seismic sequences in the Southern Apennines; one of
 356 the proposed scenarios for the 1706 sequence and the 1915 sequence in the Central Apennines; Table B1,
 357 Appendix B). Compared to the empirical relationships of magnitude vs. subsurface length, most of our modelled
 358 events (99% in Southern Apennines and 92% in Central Apennines) fall within the range marked by the mean and
 359 standard deviation of Wells & Coppersmith (1994).



360

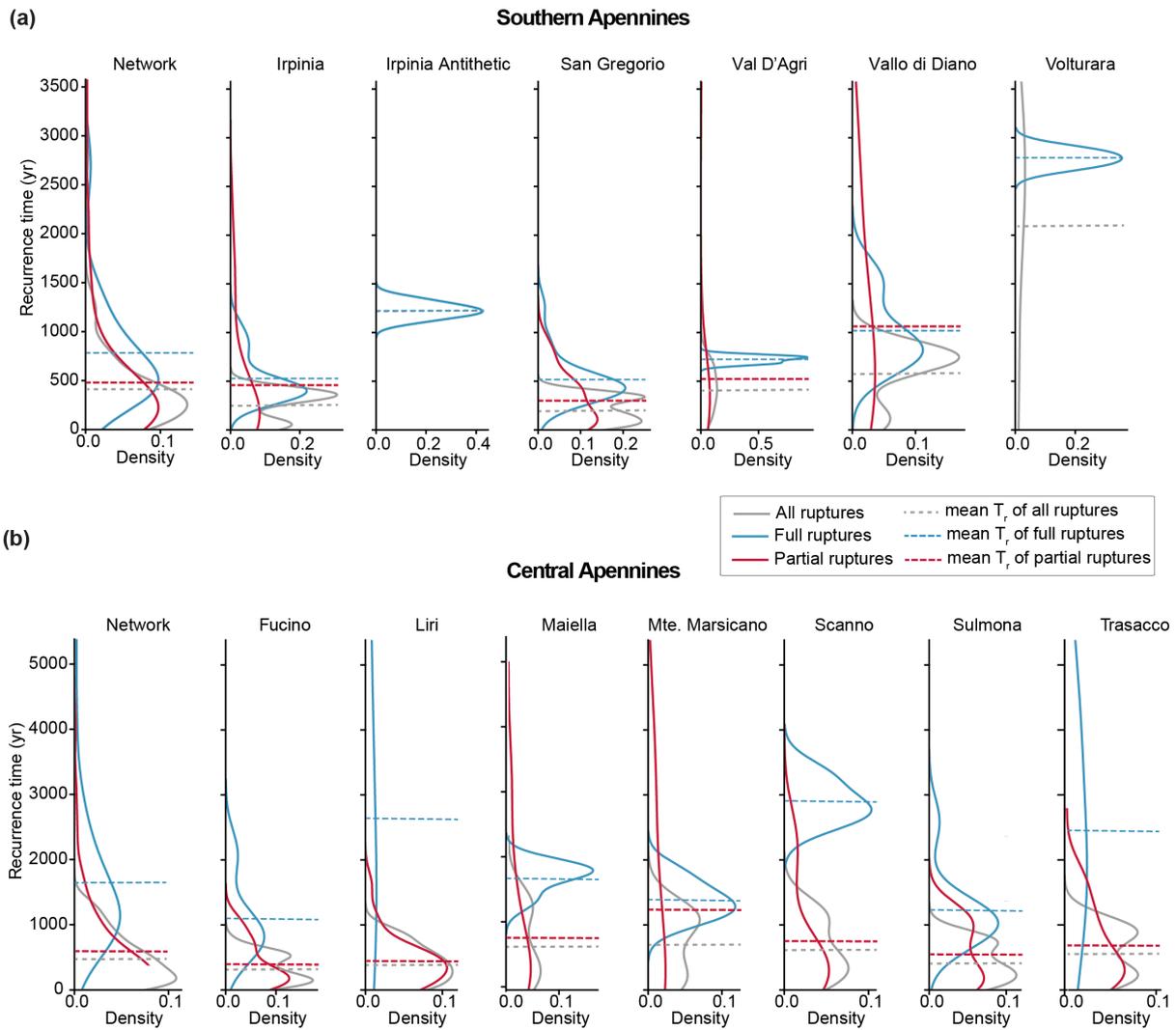
361 **Figure 8: Comparison between modelled seismic events, historical ruptures and empirical relationships of**
 362 **subsurface rupture length vs. Magnitude (Wells and Coppersmith; 1994) for (a) Southern Apennines and**
 363 **(b) Central Apennines. Magnitude refers to M_w for SEAS simulations and empirical scaling relationships;**
 364 **magnitudes of historical earthquakes are defined in Table B1 (Appendix B). Modelled and historical single-**
 365 **fault events are colour-coded according to the source fault. Superindices on dates of historical events refer**
 366 **to alternative scenarios. 1910¹: scenario with rupture of entire Irpinia antithetic (Galli and Peronace, 2014);**
 367 **1910²: scenario with 10-km rupture of Irpinia fault (Galli and Peronace, 2014); 1654¹: scenario with 13-km**

368 **rupture of Southern section of Liri and entire Fibreno faults (outside of study area; (Guidoboni *et al.* 2019);**
369 **1706¹: scenario with rupture of entire Maiella fault (Guidoboni *et al.*, 2019); 1706²: scenario with rupture**
370 **of entire Maiella and Palena faults (outside of study area; (Guidoboni *et al.*, 2019); 1915¹ scenario with**
371 **rupture of entire Fucino, Parasano and San Sebastiano faults (Michetti *et al.*, 1996); 1915² scenario with**
372 **rupture of entire Fucino, Luco and Trasacco faults (Michetti *et al.*, 1996) (see Table B1, Appendix B).**

373

374 Recurrence times vary between the two modelled regions, and depend on whether the full catalogue is considered,
375 or whether it is split between full and partial ruptures. Full catalogue recurrence times for the system and for
376 individual faults are positively skewed, with some faults showing bi-modal distributions (Fig. 9). When the
377 Southern and Central Apennines are compared, greater variability is observed in the mean recurrence times of the
378 Southern Apennines, which range from 250 (San Gregorio fault) to 2100 (Vulturara fault) years in the Southern
379 Apennines compared to 300 (Fucino fault) and 700 (Monte Marsicano fault) years in the Central Apennines (Fig.
380 9).

381 Where full and partial rupture catalogues are compared, partial ruptures tend to have shorter recurrence times than
382 full ruptures (Fig 9a-b). Across both regions, recurrence time distributions of partial ruptures in individual faults
383 and the system are positively skewed, spanning a wide range, from hours to years. In contrast, the recurrence time
384 distributions of full ruptures are narrower, typically following a bimodal or normal distribution (Fig 9).



385

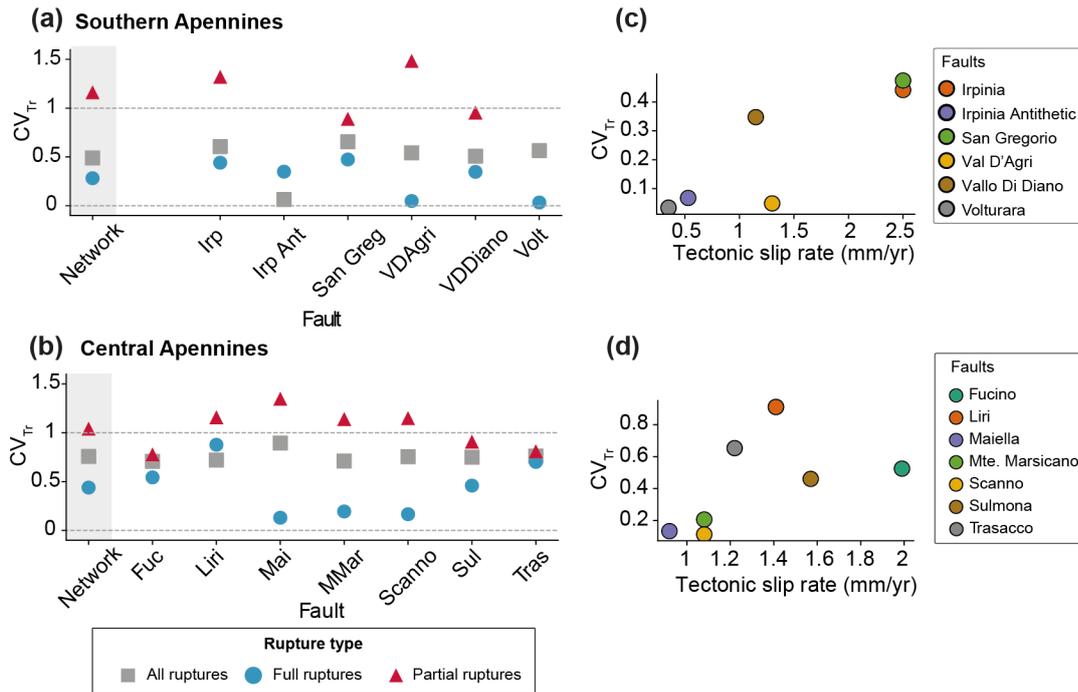
386 **Figure 9: Variation of recurrence time (T_r) for the (a) Southern and (b) Central Apennines, shown as kernel**
 387 **density plots (KDE) and mean recurrence time, for the entire fault network and for individual faults**
 388 **considering the ‘All-ruptures’, ‘Full-ruptures’ and ‘Partial-ruptures’ catalogues. Volturara fault produced**
 389 **only 1 partial rupture, thus no KDE for its partial-rupture catalogue could be computed. As this single**
 390 **partial rupture is included in the ‘All-ruptures’ catalog but absent from the ‘Full-rupture’ catalog, the two**
 391 **distributions are different. Note how the density scale varies by fault, and that partial ruptures typically**
 392 **display a greater range in recurrence time than full ruptures.**

393 The catalogue periodicity is assessed through the coefficient of variation of recurrence time CV_{T_r} . For the all-
 394 rupture catalogues, both regions display weakly periodic behaviour ($CV_{T_r}=0.5-1$, Fig. 10a-b). The exception is the
 395 Irpinia Antithetic fault, which maintains a more periodic recurrence (Fig. 10a). Overall, the Central Apennines
 396 exhibits less periodic seismic cycles ($CV_{T_r}=0.8$, Fig. 10b) compared to the Southern Apennines ($CV_{T_r}=0.5$, Fig.
 397 10a).

398 Periodicity differs between full and partial ruptures as well as between regions. Faults generate full rupture events
 399 with varying degrees of periodicity. In the Southern Apennines (Fig. 10a), all faults show strongly periodic full-
 400 rupture cycles ($CV_{T_r}<0.5$). In the Central Apennines (Fig. 10b), the behaviour is more variable: the Maiella,

401 Monte Marsicano and Scanno faults show strongly periodic full-rupture cycles whereas the Trasacco, Liri,
 402 Sulmona and Fucino faults are weakly periodic ($0.5 \leq CV_{Tr} < 1$). Partial ruptures in both networks show less periodic
 403 behaviour than full ruptures (Fig. 10a-b). Their recurrence ranges from weakly periodic ($0.5 \leq CV_{Tr} < 1$; e.g., San
 404 Gregorio, Fucino, Sulmona and Trasacco faults) to random ($CV_{Tr} = 1$, e.g., Vallo di Diano fault) or clustered
 405 ($1 < CV_{Tr} < 1.5$, e.g., Irpinia, Val D'Agri, Liri, Maiella, Monte Marsicano and Scanno faults). CV_{Tr} of full-rupture
 406 events tends to increase with increasing long-term slip rate (Fig. 10c-d).

407



408

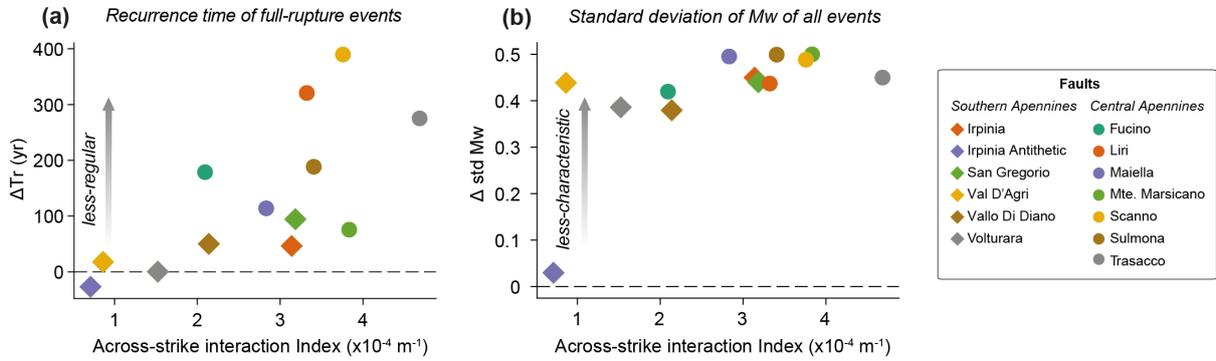
409 **Figure 10: (a-b) Coefficient of variation of recurrence times CV_{Tr} of seismic events (all events, full ruptures**
 410 **and partial ruptures) for individual faults and entire fault network in the (a) Southern Apennines and (b)**
 411 **Central Apennines. Horizontal dashed lines mark the CV_{Tr} values of perfectly periodic ($CV_{Tr}=0$) and lower**
 412 **limit of random ($CV_{Tr}=1$) seismic cycles. (c-d) Long-term slip rate vs. CV_{Tr} of all events for the (c) Southern**
 413 **Apennines and the (d) Central Apennines. Seismic cycles of full-rupture events are either strongly or**
 414 **weakly periodic, while cycles of partial ruptures are weakly periodic, random or clustered. The fault**
 415 **network in the Central Apennines has less periodic seismic cycles than in the Southern Apennines.**

416 4.3 Relationships between seismic cycle characteristics and fault network geometry

417 In the Southern Apennines, where the fault network has fewer across-strike faults and larger distances between
 418 them, the across-strike interaction index (AI) is less than in the Central Apennines, which has multiple closely-
 419 spaced across-strike faults, resulting in a higher AI due to both fault number and proximity. Fault interactions via
 420 stress transfer affect recurrence intervals and magnitudes of full-rupture events, compared to the isolated fault
 421 case (Fig. 11). Most faults produce full ruptures with longer recurrence times and larger magnitudes compared to

422 the reference cycles on isolated fault (Fig. 11). These differences become more pronounced with an increase in
 423 the number of nearby across-strike faults in the Central Apennines, as opposed to the Southern Apennines.

424

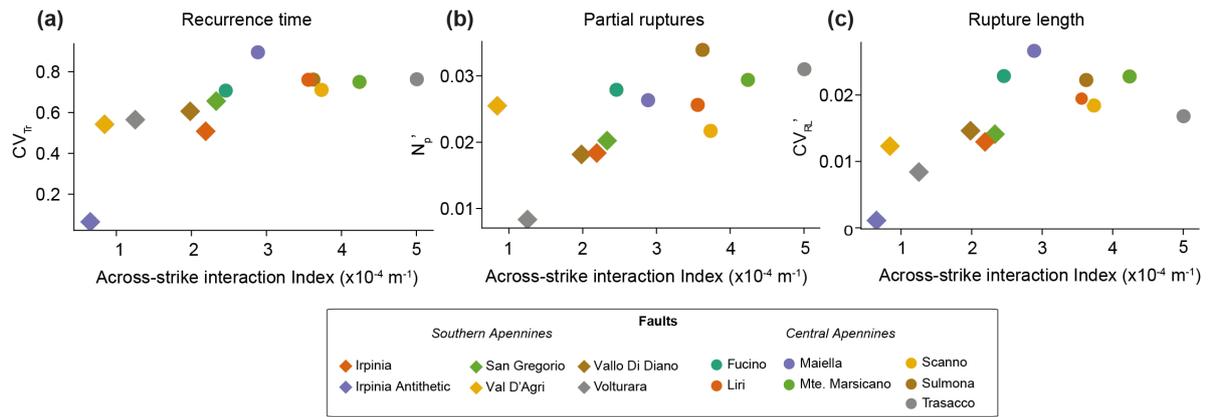


425

426 **Figure 11: Comparison of earthquake cycles between single-fault simulations and fault-network**
 427 **simulations in terms of event (a) recurrence time (Tr) of full-rupture events ($\Delta Tr = \text{mean } Tr_{\text{network}} -$**
 428 **$\text{mean } Tr_{\text{single}})/\text{mean } Tr_{\text{single}}$) and (b) standard deviation of magnitude of full-rupture events**
 429 **$\Delta \text{std } Mw = \text{std } Mw_{\text{network}} - \text{std } Mw_{\text{single}}$) as a function of the across-strike interaction index (eq. 1).**
 430 **$\Delta Tr=0$ indicates regular recurrence, while positive (negative) values correspond to longer (shorter)**
 431 **recurrence times in the fault-network simulations relative to the single-fault simulations. $\Delta \text{std } Mw=0$**
 432 **indicates characteristic magnitude distribution, while increasingly positive values indicate broader**
 433 **magnitude-frequency distributions in the fault-network simulations. Faults with multiple faults across-**
 434 **strike (larger AI) show larger differences in recurrence time and standard deviations of magnitude**
 435 **compared to single-fault simulations, indicating a greater deviation from their characteristic periodic**
 436 **behaviour.**

437 Differences in earthquake cycle properties between the Southern and Central Apennines are influenced by fault
 438 network geometry (Fig. 12). When considering both networks together, we observe consistent positive
 439 correlations between the across-strike interaction index AI and the three metrics of seismic complexity (CV_{Tr} , N_p' ,
 440 and CV_{RL}') with mean Spearman values ranging from 0.46 to 0.84 (Table F1, Appendix F). When analysed
 441 separately, the Southern Apennines show strong correlations for CV_{Tr} and CV_{RL}' at some depths (Table F1), but no
 442 consistent trend for N_p' . In contrast the Central Apennines show weak or no correlations for all metrics (Table
 443 F1). This suggest that the geometric effects become clearer when sampling a broader range of network
 444 configurations across both regions.

445



446

447 **Figure 12: Relationships between fault network geometry, described by the across-strike interaction index**
 448 **(AI), and (a) coefficient of variation of recurrence times (CV_{Tr}), (b) number of partial ruptures (N_p^a) and**
 449 **coefficient of variation of rupture lengths (CV_{RL}^a) for faults in the Southern and Central Apennines. AI is**
 450 **calculated at a depth of 7.5 km, approximately the middle fault depth. Plots corresponding to depths of 0**
 451 **km and 15 km are shown in Fig. F1 (Appendix F).**

452

453 5 Discussion

454 5.1 Comparison between simulated and natural seismic cycles in Italy

455 Italy's exceptional historical and paleoseismic record (D'Addezio et al., 1991; Galadini and Galli, 1996, 1999;
 456 Galli et al., 2006, 2008, 2016, 2014; Galli and Peronace, 2014; Pace et al., 2020; Pantosti et al., 1993) provides a
 457 valuable basis for evaluating our models. Our simulations are, to our knowledge, the first 3D continuum
 458 earthquake cycle models to combine rate-and-state friction with fault interactions across more than three faults,
 459 resolving nucleation and rupture self-consistently.

460 To compare with natural data, we focused on $M \geq 6.5$ events and recurrence intervals from paleoseismic trenching
 461 (Table B2; Pace et al., 2016). After scaling due the high slip rates used in our models, natural recurrence intervals
 462 are 1.3-9.4x longer than simulated ones, consistent with our use of maximum along-strike long-term slip rates,
 463 which likely overestimate loading and shorten cycles. We expect that incorporating realistic tapered profiles
 464 would bring simulated intervals closer to observations (Delogkos et al., 2023; Faure Walker et al., 2019).

465 Paleoseismic recurrence times show greater variability than our nearly periodic modelled seismic cycles for full
 466 ruptures. This may reflect incomplete long-term records (e.g., Lombardi et al., 2025; Mouslopoulou et al., 2025),
 467 temporal changes in slip rates documented by cosmogenic dating (Benedetti et al., 2013; Cowie et al., 2017;
 468 Mildon et al., 2022; Roberts et al., 2024, 2025; Sgambato et al., 2025), the greater geometrical complexity of
 469 natural faults, and possibly also a larger ratio W_s/L_∞ (Barbot, 2019; Cattania, 2019), all of which could enhance
 470 stress heterogeneity and variability.

471 While our models reproduce the magnitude range of historical single-fault events, they do not produce multi-fault
 472 ruptures as documented in the historical catalogue in the region (Benedetti et al., 1998; Cello et al., 2003; Galli et

473 al., 2006). This likely stems from the quasidynamic approximation, which does not simulate dynamic triggering
474 by seismic waves and limits stress drops, slip velocities and rupture jumping compared to fully-dynamic
475 simulations (Thomas et al., 2014). However, studies using other quasidynamic simulators (e.g. RSQSim, Dieterich
476 and Richards-Dinger, 2010) have reported multi-fault rupture scenarios (Herrero-Barbero et al., 2021; Milner et
477 al., 2021; Shaw et al., 2025). In these models, faults are more closely spaced and subject to higher effective normal
478 stresses than our QDYN simulations (~100 MPa), which increases rupture energy favouring multi-fault ruptures.
479 In addition, model approximations of RSQSim, such as discrete state transitions and prescribed slip rates during
480 the sliding stage (Shaw et al., 2025) might also promote larger ruptures. In comparison, our simulations resolve
481 continuous rate-and-state friction, and include velocity strengthening barriers at fault tips and lower effective
482 normal stress, all of which would favour rupture arrest at the fault boundaries. This limitation likely leads to an
483 underestimation of the largest seismic events and biases the upper tail of the magnitude-frequency distribution
484 toward single-fault ruptures. While non-periodic multi-fault ruptures would likely increase the recurrence
485 variability and cycle complexity of individual faults, it is not expected to change the first-order role of fault
486 network geometry identified here. Multi-fault ruptures could be promoted by stronger velocity weakening
487 asperities or higher normal stresses, but such parameter choices would increase the computational cost, limiting
488 their feasibility in these regional-scale simulations. A second alternative to address this would be using simulated
489 cycles as initial conditions for fully dynamic rupture models (Galvez et al., 2020).

490 Despite these limitations, our results show that SEAS models can capture key behaviours of complex fault
491 networks and generate synthetic earthquake catalogues that are directly comparable with paleoseismic data. This
492 provides a strong basis for future simulations integrating more geologically realistic features, such as along-strike
493 variations of slip rate and strike.

494 **5.2 Effect of stress redistribution within a fault network on simulated seismic sequences**

495 Our simulations show that earthquakes on networks with multiple across-strike faults, such as those in the Central
496 Apennines, generate spatially heterogeneous stress perturbations on nearby faults. These interactions promote
497 more partial ruptures, greater variability in rupture lengths, magnitude and nucleation locations, and less periodic
498 behaviour of large earthquakes. In contrast, faults within more along-strike networks, like in the Southern
499 Apennines, experience more uniform stress loading and tend to produce more periodic and characteristic seismic
500 cycles, behaviours more similar to simulations of isolated faults (Figs. 4a, 12,13). These results build upon prior
501 numerical modelling work of (Rodriguez Picada et al., 2025a), which showed that complex and non-periodic
502 seismic cycles emerge in a system of two across-strike normal faults. They also extend prior CST modelling work
503 (Sgambato et al., 2020b, 2023), which showed that relatively isolated faults experience more regular stress loading
504 histories dependent on interseismic loading than networks with multiple faults arranged across-strike.

505 The effects of fault network geometry can be isolated when other factors such as long-term slip rate are considered
506 constant. A clear example of the influence of fault network geometry comes from the Central Apennines, where
507 the Scanno and Monte Marsicano faults have similar slip rates (Table 1), yet they have contrasting seismic cycle
508 characteristics. The Monte Marsicano fault exhibits more irregular rupture timing, partial ruptures and rupture
509 extent variability than the Scanno fault (Fig. 12), likely due to the influence of multiple closely-spaced CST

510 sources (Maiella, Sulmona, Scanno and Trasacco, Fig. E2e,f; E4, Appendix E), and therefore higher across-strike
511 interaction index.

512 Fault networks also show longer recurrence times and larger magnitudes of full-rupture events, than the same
513 faults modelled in isolation (Fig. 11). This results from stress interactions delaying full ruptures, allowing more
514 time for fault healing and strength recovery, which leads to larger stress drops and higher seismic moments.
515 Consequently, the magnitude versus rupture length scaling for fault networks (Fig. 8) better matches natural
516 variability (Wells and Coppersmith, 1994), indicating that stress interaction among faults may contribute to the
517 scatter seen in empirical scaling relationships.

518 Although not included in the simulations, the Parasano-Pescina, Tremonti and San Sebastiano may
519 influence the seismic sequences of the Central Apennines fault network. Due to their limited area, the
520 Parasano-Pescina and Tremonti faults would likely produce full-rupture events, similar to the Irpinia
521 Antithetic fault in the Southern Apennines (Figure 4a-c). The inclusion of these tree faults would
522 increase the number of across-strike interacting fault segments, thus promoting more stress
523 heterogeneities in neighbouring faults such as Fucino, Scanno and Trasacco, and potentially leading to
524 more complex seismic cycles or multi-fault ruptures. Given the limitations of the modelling framework,
525 it is currently not feasible to investigate these interactions within a network-wide approach

526 Our focus is on normal fault networks, thus a remaining question is whether the findings will be applicable to
527 strike-slip and thrust faults. We speculate that similar effects could occur in these settings, with the outcome
528 depending on the degree of development of the fault network. Fault networks in late stage of development tend to
529 evolve into more localized structures, reducing the fault overlap and the extent of interactions. In contrast,
530 interaction effects may be more pronounced in early-stage networks, where a higher number of closely spaced
531 overlapping faults are more common.

532 **5.3 Role of long-term slip rate in stress interaction effects**

533 When faults have few stress interactions with others, as in the Southern Apennines, their seismic cycle is primarily
534 controlled by their long-term slip rate, with faster moving-faults having shorter recurrence times than slower-
535 moving faults (Fig. G1, Appendix G).

536 In both Central and Southern Apennines, faults with higher long-term slip rates are more responsive to positive
537 stress perturbations from nearby faults. For a given across-strike interaction index, faster-loading faults show less
538 periodic recurrence of full ruptures and more variable rupture lengths (Fig. 12). Indeed, as these faults approach
539 failure more often, they are more likely to be triggered by even small CST perturbations from nearby faults (e.g.,
540 Sulmona, Liri and Trasacco faults in Central Apennines, Fig. 12). In contrast, slower-loading faults, such as the
541 Volturara fault in the Southern Apennines (Fig. 7a), accumulate stress more slowly, experience stronger healing
542 and show more regular recurrence.

543 This finding partially contrasts with results of elastic-brittle lattice models (e.g. Cowie et al., 2012), which
544 associate higher CV with slower long-term slip rates in faults with multiple across-strike interactions. In those

545 models the absence of time-dependent stress healing means that slower loading allows heterogeneity to
546 accumulate over longer timescales, increasing CV. In our rate-and-state friction models, by contrast, we isolate
547 loading-rate effects at a fixed network geometry and find that CV increases with higher slip rate, consistent with
548 reduced interseismic healing for the same time interval and higher sensitivity to small CST perturbations.
549 Although the long-term slip-rate trends differ, both approaches agree that more complex fault networks lead to
550 greater recurrence variability.

551 **5.4 Implications of fault network geometry on seismic hazard assessment**

552 Physics-based earthquake simulators based on SEAS models provide an alternative framework to constrain
553 earthquake rates and magnitude-frequency distributions for seismic hazard assessment (SHA, e.g., Milner et al.,
554 2021; Shaw et al., 2018, 2022). To explore this potential, we compare magnitude-frequency distributions derived
555 from our SEAS simulations with those obtained from an approach (Pace et al., 2016) based on seismic moment
556 conservation and a truncated Gutenberg-Richter distribution (“fault-based method”, Appendix H).

557 For both fault networks, the two approaches yield systematically different magnitude-frequency distributions (Fig.
558 4c,f). The fault-based method predicts lower recurrence rates than the SEAS simulations for $M_w < \sim 6.3$, but higher
559 rates at larger magnitudes. One likely explanation is that, unlike the fault-based method which assumes that the
560 entire long-term slip rate is released seismically, the SEAS models allow part of the moment budget to be
561 accommodated aseismically through creep or slow slip transients (e.g. Fig. 5c), thereby reducing the seismic
562 budget for small earthquakes (Rodríguez Piceda et al., 2025c). Additionally, the tested frictional properties in our
563 SEAS models yield a W_s/L_∞ ratio (Eq. A5) that may favour more characteristic rupture behaviour, limiting the
564 occurrence of smaller events relative to Gutenberg-Richter assumptions (Barbot, 2019; Cattania, 2019).

565 At the high-magnitude end, SEAS simulations produce larger magnitudes and higher rate of events than the fault-
566 based method. This difference is likely due to the fault-based method assuming stationary recurrence and a priori
567 prescribed maximum magnitude (Pace et al., 2016), while in the SEAS models rupture sizes emerge dynamically
568 from stress evolution over a finite catalogue duration. Therefore, this comparison, although limited, shows how
569 physics-based models can provide independent constraints both on the shape of the magnitude-frequency
570 distribution and the maximum magnitude for SHA.

571 Beyond these general applications, our results also show how fault-network geometry modulates earthquake
572 recurrence and magnitude variability challenging the common practice of applying uniform recurrence parameters
573 (e.g. a single mean recurrence time or coefficient of variation) across an entire fault system (e.g., Nishenko and
574 Buland, 1987, Ellsworth et al., 1999; Matthews et al., 2002). In networks with numerous across-strike faults and
575 high long-term slip rates, recurrence can vary greatly between faults (e.g., Central Apennines, Fig. 9b),
576 suggesting that hazard models should allow for broader epistemic uncertainty in recurrence and magnitude
577 distributions. Therefore, probabilistic SHA could further benefit from integrating network-derived metrics, such
578 as the across-strike interaction index, as quantitative guides for weighting logic-tree branches. While such metrics
579 do not directly prescribe recurrence or magnitude parameters, they may provide physically grounded constraints
580 to better reflect the complexity of fault interactions. Physics-based simulators are well suited to quantify the
581 recurrence and magnitude variability (Milner et al., 2021; Shaw et al., 2018, 2022), and to inform the weighting

582 of alternative fault-based SHA models. In addition, time-dependent SHA approaches, such as those based on
 583 Coulomb stress transfer histories (Chan et al., 2010; Iacopetti et al., 2021; Mignan et al., 2018; Stein et al., 1997;
 584 Toda et al., 1998), should also account for spatial variability in stress changes. Finally, our simulations indicate
 585 that overlap zones between faults can act as preferred nucleation sites and constrain rupture extent (Fig. 7),
 586 highlighting their importance for assessing rupture scenarios and directivity effects (Spagnuolo et al., 2012;
 587 Thompson and Worden, 2017).

588 **6 Conclusions**

589 We numerically simulated seismic cycles on two fault networks in the Southern and Central Apennines to examine
 590 how changes in stress interactions caused by fault network geometry influence earthquake recurrence rates and
 591 magnitudes. Increased number of fault interactions leads to a greater departure from the characteristic and periodic
 592 behaviour of an isolated fault, with higher variability of earthquake recurrence, nucleation location and magnitude
 593 with increasing fault interaction. Fault networks with multiple across-strike faults are characterised by more
 594 complex seismic cycles than networks with fewer across-strike faults. When the number of across-strike
 595 interactions is similar, faults with higher slip rates tend to produce less-periodic earthquakes with more variable
 596 magnitude, meaning that slip rate influences how faults respond to stress changes from nearby ruptures. Our
 597 models demonstrate that, by carefully considering the numerical limitations, simulated earthquake catalogues can
 598 be meaningfully compared to natural earthquake records, highlighting the potential of using earthquake cycle
 599 modelling to assess the seismic hazard of complex normal fault networks.

600

601

602 **Appendix 1: SEAS governing equations and material properties**

603 Fault friction follows the rate-and-state friction law (Dieterich, 1979; Marone, 1998a; Ruina, 1983), where the
 604 shear stress (τ) along the fault is equal to its frictional strength:

$$605 \quad \tau = \mu\sigma(A1)$$

606 μ is the coefficient of friction and σ is the effective normal stress (total normal stress minus pore-fluid pressure).
 607 We adopt the regularised formulation of rate-and-state (Lapusta et al., 2000; Rice and Ben-Zion, 1996) where
 608 friction evolves with slip rate (V) and a state variable (θ) as:

$$609 \quad \mu(\theta, V) = a \sinh^{-1} \left[\frac{V}{2V^*} \exp \left(\frac{\mu^* + b \ln \left(\frac{V^* \theta}{D_c} \right)}{a} \right) \right] \quad (A2)$$

610 where μ^* is the reference coefficient of friction at a reference slip rate V^* ; a and b are constants for the magnitude
 611 of the contributions of the slip rate and fault state to the friction, respectively. D_c is the characteristic slip distance
 612 and it controls how the state variable evolves following the aging law (Dieterich, 1979; Ruina, 1983):

613
$$\frac{d\theta}{dt} = 1 - \frac{V\theta}{D_c} \quad (A3)$$

614 At steady state, $\frac{d\theta}{dt} = 0$, so steady state-friction is:

615
$$\mu_{ss} = a \sinh^{-1} \left[\frac{v}{2V^*} \exp \left(\frac{\mu^* + b \ln \left(\frac{v}{V^*} \right)}{a} \right) \right] \quad (A4)$$

616 In this state, the parameter $(a - b)$ describes the dependence of μ_{ss} with velocity, with positive $(a - b)$
 617 characteristic of velocity-strengthening materials (i.e. steady-state friction increases with increasing velocity) and
 618 negative $(a - b)$ characteristic of velocity-weakening materials (i.e. steady-state friction decreases with
 619 increasing velocity). Velocity-weakening materials can develop stick-slip behaviour, thus they are assumed to be
 620 characteristic of the seismogenic portion of a fault or seismic asperity. To produce unstable sliding, the smallest
 621 dimension (length L or width W) of a segment with velocity strengthening material must exceed a so-called
 622 nucleation length (L_∞) (Rubin and Ampuero, 2005):

623
$$L_\infty = \frac{1}{\pi} \left(\frac{b}{b - a} \right)^2 \frac{GD_c}{b\sigma} \quad (A5)$$

624 where G is the shear modulus of the host rock. If the size of the velocity-weakening fault does not exceed the
 625 nucleation length, aseismic slip will occur (Rubin and Ampuero, 2005).

626 To compute earthquake cycles, QDYN solves the equation of elasto-static equilibrium, where stress and slip rate
 627 V are related by (Rice, 1993):

628
$$\tau_0 + \tau_e - \frac{G}{2c} V = \sigma \mu \quad (A6)$$

629 where τ_0 is the background shear stress, τ_e is the elastic shear stress due to fault slip, $\frac{G}{2c}$ is the radiation damping
 630 term which approximates the inertial effects of seismic waves, c is the shear-wave speed, σ is the effective normal
 631 stress, calculated by summing the initial normal stress σ_0 and the elastic normal stress σ_e from stress interactions:

632
$$\sigma = \sigma_0 + \sigma_e \quad (A7)$$

633 QDYN utilises the back-slip approach, such that the stresses transmitted from a fault element to neighbouring
 634 elements are proportional to their slip deficit relative to the long-term tectonic slip (Heimisson, 2020; Savage,
 635 1983). In this interpretation of backslip, faults are approximately modelled as faults of finite size loaded remotely
 636 by tectonic stresses (Allam et al., 2019; Dieterich and Smith, 2010). When faults are remotely loaded in an elastic
 637 medium, they tend to accumulate stresses indefinitely with increasing slip; however, due to the crust finite
 638 strength, there should be some process of off-fault inelastic deformation to relax these stresses. The backslip
 639 method approximately accounts for these inelastic processes, such that it maintains kinematic consistence with
 640 the long-term slip rate of faults (Allam et al., 2019; Dieterich and Smith, 2010). The backslip approach
 641 implemented in QDYN is the same as in Heimisson (2020).

642 The elastic shear stress at the i -th fault element τ_i^e due to the slip on the remaining fault elements is:

$$643 \quad \tau_i^e = - \sum_j k_{ij}^{\tau} (u_j(t) - V_{PL} t) \quad (A8)$$

644 where V_{PL} is the long-term tectonic slip rate on the fault, u_j is the slip on the j -th cell and k_{ij}^{τ} is the stiffness matrix
 645 for shear stress, which describes the shear stress change on the i -th fault element exerted by a unit slip on the j -th
 646 fault element. The elastic normal stress σ_e from Eq. A7 is calculated similarly than in Eq. A8, but with the stiffness
 647 matrix for normal stress k_{ij}^{σ} :

$$648 \quad \sigma_i^e = - \sum_j k_{ij}^{\sigma} (u_j(t) - V_{PL} t) \quad (A9)$$

650 Both stiffness matrices in eqs. A8 and A9 are calculated using the analytical equations for static stresses induced
 651 by rectangular dislocations in a homogeneous elastic half-space (Okada, 1992). Free surface conditions are
 652 included in the formulations. Because the faults have varying orientations relative to one another, we are unable
 653 to use optimizations that take advantage of the invariant strikes to construct the stiffness matrices, such as Fast
 654 Fourier transforms (Rice, 1993). Instead, we use the implementation by Galvez et al., (2020) of the hierarchical
 655 matrix (H-matrix) compression to the stress transfer component (Bradley, 2014) and the LSODA solver
 656 implemented by Yin et al. (2023). Despite the improved time-stepping efficiency demonstrated in Yin et al.
 657 (2023), the simulations remain computationally demanding, requiring approximately two months to complete.

658 The material and frictional properties are listed in Table A1. Each fault consists of a rectangular patch in the centre
 659 with velocity-weakening properties, bounded by a velocity-strengthening region with a width of 1.5 km in the
 660 Southern Apennines and 1.1 km in the Central Apennines. Additionally, we set a 1 km transition zone with
 661 velocity-strengthening friction properties along the edges of the velocity-weakening regions to prevent infinite
 662 stress rates at the fault edges that could arise from the backslip method (Rodriguez Piceda et al., 2025a).

663 The variation of normal stress with depth follows the approach by Lapusta et al. (2000), where effective normal
 664 stress $\bar{\sigma}_l$ equals the lithostatic pressure minus the hydrostatic pore-fluid pressure at shallow depths, with a
 665 transition to lithostatic pore pressure gradient with a 50 MPa offset at depth (z):

$$666 \quad \bar{\sigma}_l = \min \left\{ \begin{array}{l} 2.8 + 18 * z/km \\ 50 \text{ MPa} \end{array} \right. \quad (A10)$$

667 We account for the dip angle of the normal faults (α) in our simulations:

$$668 \quad \bar{\sigma} = \bar{\sigma}_l \sin(\alpha) \quad (A11)$$

669 In this set up, where multiple faults are interacting, the normal stress can reach negative values near the surface.
 670 To accommodate the possible stress change that could occur during the spin-up phase, the initial minimum normal
 671 stress is increased to 15 MPa (Yin, 2022).

672

673 **Table A1:** Material and frictional properties of the model set up in the Central Apennines (CA) and Southern
 674 Apennines (SA).

Symbol	Description (units)	Value
G	Shear modulus (GPa)	32
λ	Elastic modulus (GPa)	32
c	Shear wave velocity (m/s)	3000
μ^*	Reference friction coefficient	0.6
a	Direct-effect parameter	0.007
b	Evolution effect parameter	0.014 (VW) / 0.0042 (VS)
D_c	Characteristic slip distance (mm)	8 (SA), 10 (CA)
V_{PL}	loading rate	see table 1
V^*	Reference slip rate (m/s)	$=V_{PL}$
V_0	Initial slip rate (m/s)	$0.8 V_{PL}$
D_w, D_x	Cell size along-dip (m), Cell size along-strike (m)	110 (SA), 128 (CA)

675 **Appendix B: Historical and paleoseismicity of the Southern and Central Apennines**

676 **Table B1: Historical seismic events (>0A.D.) based on (Mildon, 2017a; Sgambato, 2022).** *Magnitudes of
 677 seismic events in Central Apennines prior to 1979 A.D. are taken from the Catalogue di Forti Terremoti
 678 (Guidoboni et al., 2019) and are derived from the macroseismic shaking records (Gasperini and Ferrari,
 679 2000) and as such are described as equivalent magnitudes (M_e , based on Mildon, 2017). For more recent

680 earthquakes in the Central Apennines post 1979 A.D. the magnitudes described are from seismological
681 sources. ¹The name NW segment of Vallo di Diano often differs in the literature, being also known as
682 Auletta fault or Caggiano fault (Bello et al., 2022; Galli et al., 2006a)

Earthquake Date	Magnitude*	Source Fault	Proportion of the fault that slips	Reference
Southern Apennines				
15/01/1466	5.98	Irpinia	SE section 13 km	Marturano (2007)
31/07/1561	6.1	San Gregorio	Entire fault	Castelli et al. (2008)
19/08/1561	6.4-6.5	Vallo Di Diano ¹	all	Galli et al. (2006)
08/09/1694	6.73	Irpinia	All	Galli et al., (2006, 2014); Galli and Peronace (2014)
01/02/1826	5.74	Val d'Agri	Northern section 10 km	Rovida et al., (2020)
09/04/1853	5.6	Irpinia	Northern section 8 km	Galli and Peronace, (2014)
16/12/1857	7.12	Vallo di Diano and Val D'Agri	Northern section Vallo Di Diano (12km) + entire Val D'Agri	Benedetti et al., (1998); Cello et al. (2003); Galli et al. (2006)
07/06/1910	5.76	Irpinia Antithetic	Entire fault	Galli and Peronace (2014)
		Irpinia	Southern section of Irpinia (10 km)	
23/11/1980	6.81	Irpinia, San Gregorio? and Irpinia antithetic	Entire faults	Galli and Peronace (2014); Giardini et al. (1996); Rovida et al. (2020); Sgambato et al. (2025); Westaway, (1993); Westaway and Jackson (1987)
Central Apennines				

05/12/1456	5.8	Sulmona	Northern section (9km)	Guidoboni et al. (2019)
24/07/1654	6.3	Liri and Fibreno	Southern section of Liri (13 km) + entire Fibreno fault (6.6 km)	
3/11/1706	6.8	Maiella	Entire fault	Guidoboni et al. (2019)
		Maiella and Palena	Entire faults	
13/01/1915	7	Fucino, Parasano and San Sebastiano	Entire faults	Michetti et al. (1996)
		Fucino, Luco and Trasacco	Entire Fucino and Luco faults + northern section of Trasacco (8km)	
26/09/1933	6	Maiella	Central section (12.2 km)	Pizzi et al. (2010)

683

684 **Table B2: Paleearthquakes and historical events in modelled faults of the Southern and Central**
685 **Apennines. The nomenclature of some of the faults adopted by Faure Walker et al. (2021) and Sgambato**
686 **et al. (2020) and in this study differs from that of the paleoseismic data: ¹part of the Monte Marzano fault**
687 **system by Galli (2020); ²part of this fault is the San Benedetto dei Marsi–Gioia dei Marsi segment in the**
688 **Fucino fault system in Galadini and Galli (1999); ³also known as Monte Morrone fault by Galli et al. (2015).**
689 **Paleoevents in bold are events with defined aged brackets used for the calculations of mean and standard**
690 **deviation of recurrence times in the Discussion section.**

Fault	N Paleoevents	Date of paleoevents	Reference
Southern Apennines			
Irpinia ¹	5	6736-8600 B.P.; 4411-6736 B.P; 3507-4283 B.P.; 1415-2754 B.P.; 1980 A.D. Pantano di San Gregorio Magno segment:	(D'Addezio et al., 1991; Pantosti et al., 1993)

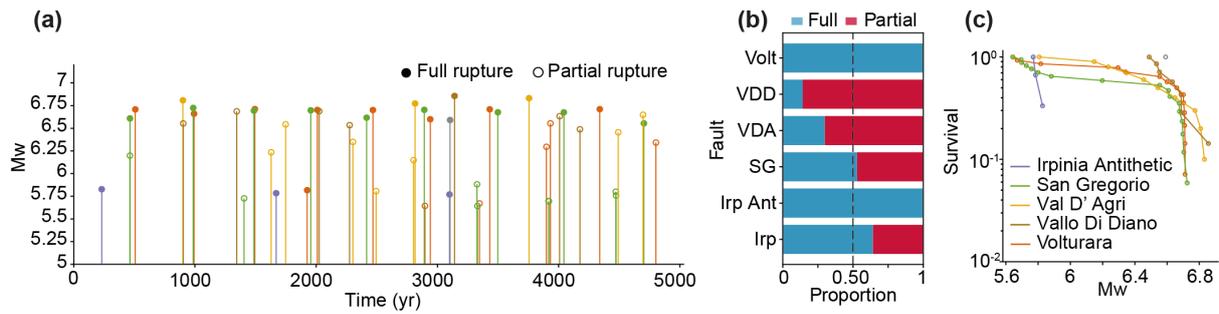
		11,180–19,660 B.P., 6620–9420 B.P., 2570–6620 B.P., 1720–2570 B.P.; 1980 A.D	
Central Apennines			
Fucino-Ovindolli-Pezza ²	6	20,000–32,520 ± 500 B.P.; 12,729–7,576 B.P.; 3,500–3,300 B.P.; 426–782 AD; 1231 AD; 1915 AD	Galadini and Galli (1996, 1999); Galli et al. (2016)
Sulmona ³	6	Before 9,000 B.P.; 6,500–8,500 B.P.; 5,500 B.P.; 200 AD, 1456 AD	Galli et al. (2015)
Trasacco	7	After 12,600 B.P.; 12,600–12,000 B.P.; 7,000–6,600B.P.; 6,500–5,900B.P.; 3,500–3,600 B.P. to 3,400–3,500 B.P.; 1,500– 1,600B.P. or 1,500–1,400 B.P.; 1915 AD	Galadini and Galli, (1999)
Liri	3	Before 12,500 B.P.; before 5,000 B.P.; before 1,500 B.P.	Pace et al., (2020)

691 **Table B3: Comparison between recurrence times (mean and standard deviation in years) derived from**
692 **historical and paleoseismological data (Tr_{paleo}) and modelled seismic events (Tr_{model}). Only paleoseismic**
693 **events with well-defined age brackets were included in the calculation (see Table B2). Tr_{paleo} = recurrence**
694 **times based on historical and paleoseismological data from the past 20 kyr. Scaled Tr_{model} = recurrence**
695 **times from modeled seismic events adjusted by the correction factor.**

Fault	$Tr_{paleo} \pm \sigma$	Scaled Tr_{model}
Irpinia (SA)	1900 ± 186	454±161
Fucino (CA)	4253 ± 4465	900±720
Sulmona (CA)	1866±1231	1440±620
Trasacco (CA)	2469±1259	1217±290

696

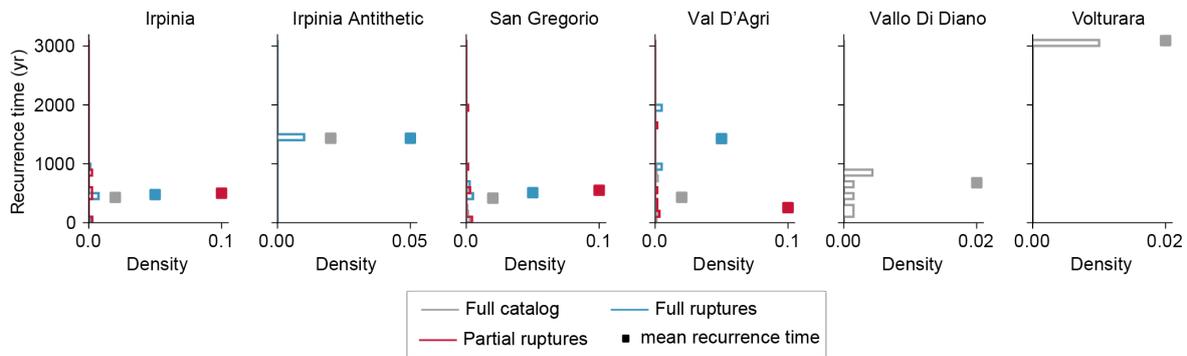
697 **Appendix C: SEAS simulations of Southern Apennines with nominal slip rate**



698

699 **Figure C1: synthetic catalog for a simulation of the Southern Apennines with nominal values of prescribed**
 700 **long-term slip rate. (a) Time distribution of simulated full- and partial-rupture events with stems and**
 701 **markers color-coded by fault (b) (b,e) proportion of full and partial ruptures per fault. (c) Magnitude-**
 702 **frequency distributions of seismic events shown as survival function (number of events with a Mw larger**
 703 **than a given value normalized by total number of events) for each fault. Color legend for each fault is shown**
 704 **in panels (c).**

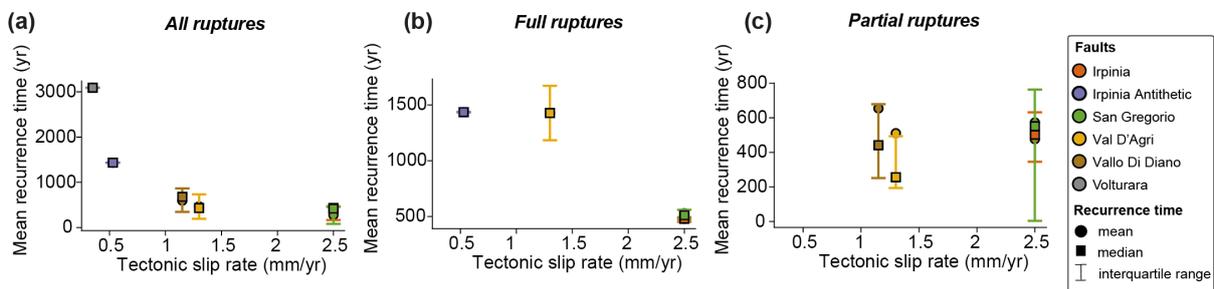
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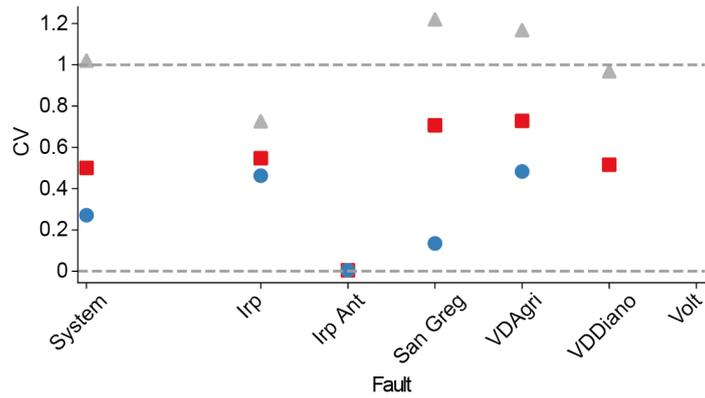
707 **Figure C2: Variation of recurrence time of individual faults (step density histogram and mean recurrence**
 708 **time) and of the entire fault system for seismic events considering the full catalog, full-rupture and partial-**
 709 **rupture events of the fault networks for a simulation of the Southern Apennines with nominal values of**
 710 **prescribed long-term slip rate.**

711



712

713 **Figure C3: Mean, median and interquartile range of recurrence time of seismic events vs. (scaled) long-**
 714 **term tectonic slip rate of individual faults for a simulation of the Southern Apennines with nominal values**
 715 **of prescribed long-term slip rate.**



716

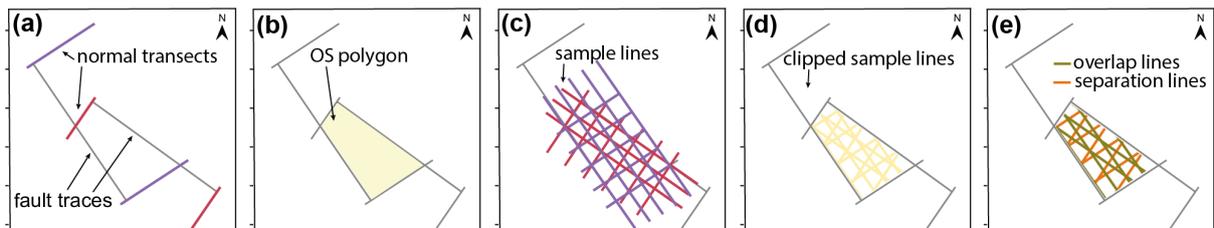
717 **Figure C4: Coefficient of variation of recurrence times CV_{Tr} of seismic events (all events, only full ruptures**
 718 **and only partial ruptures) for individual faults and entire fault system for a simulation of the Southern**
 719 **Apennines with nominal values of prescribed long-term slip rate.**

720

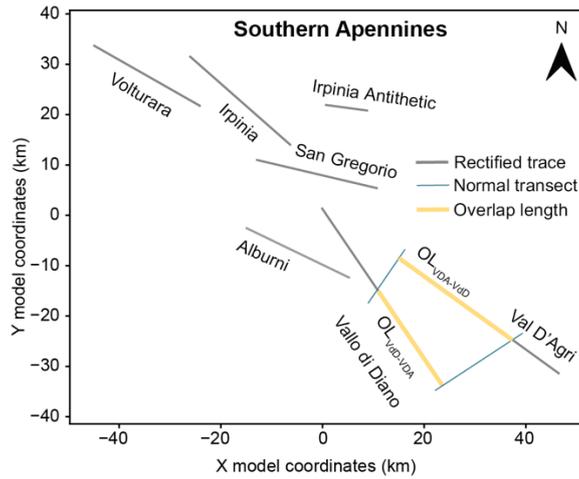
721 **Appendix D: Determination of separation between faults**

722 To determine the separation between two faults at depths of 0, 7.5 and 15 km we followed the following steps.
 723 First, for all fault traces, we created bounding transects normal to the fault tips (Fig. D1a). Second, for each fault
 724 pair, we drew a polygon bounded by two of the transects (one for each fault) and the fault traces (Fig. D1b). Third,
 725 we created sample lines (separation lines) equally spaced by 100 m parallel to the transects (Fig. D1c). Lines that
 726 did not intersect the fault traces or the transects were removed. Fourth, we clipped the sample lines to the polygon
 727 area (Fig. D1d-e). Finally, for each set of overlap and separation lines, we computed the average length (preferred
 728 value) and standard deviation. The workflow was carried out with QGIS (QGIS Development Team, 2009).

729



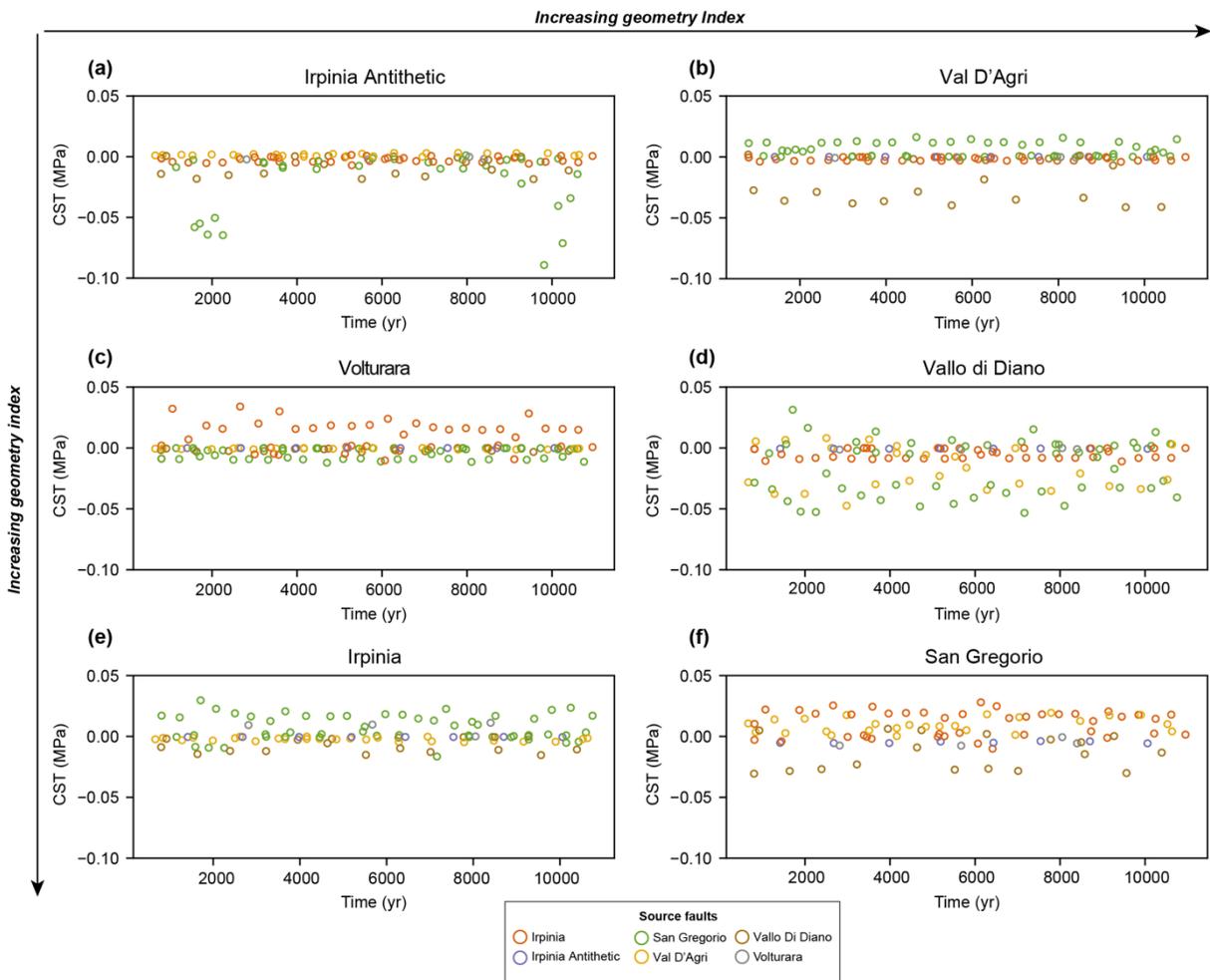
730 **Figure D1: illustrative workflow used to determine the separation between two fault traces at a given depth.**



731

732 **Figure D2: Determination of overlap length between two faults (in this example, Vallo Di Diano and Val**
 733 **D'Agri fault in the Southern Apennines) used for Figure 7 and S7-S8. To determine the overlap length**
 734 **between two given faults (fault 1 and 2), we measure the distance along the strike of fault 1 between the**
 735 **normal transect to one of the tips of fault 1 to the normal transect of to one of the tips of fault 2.**

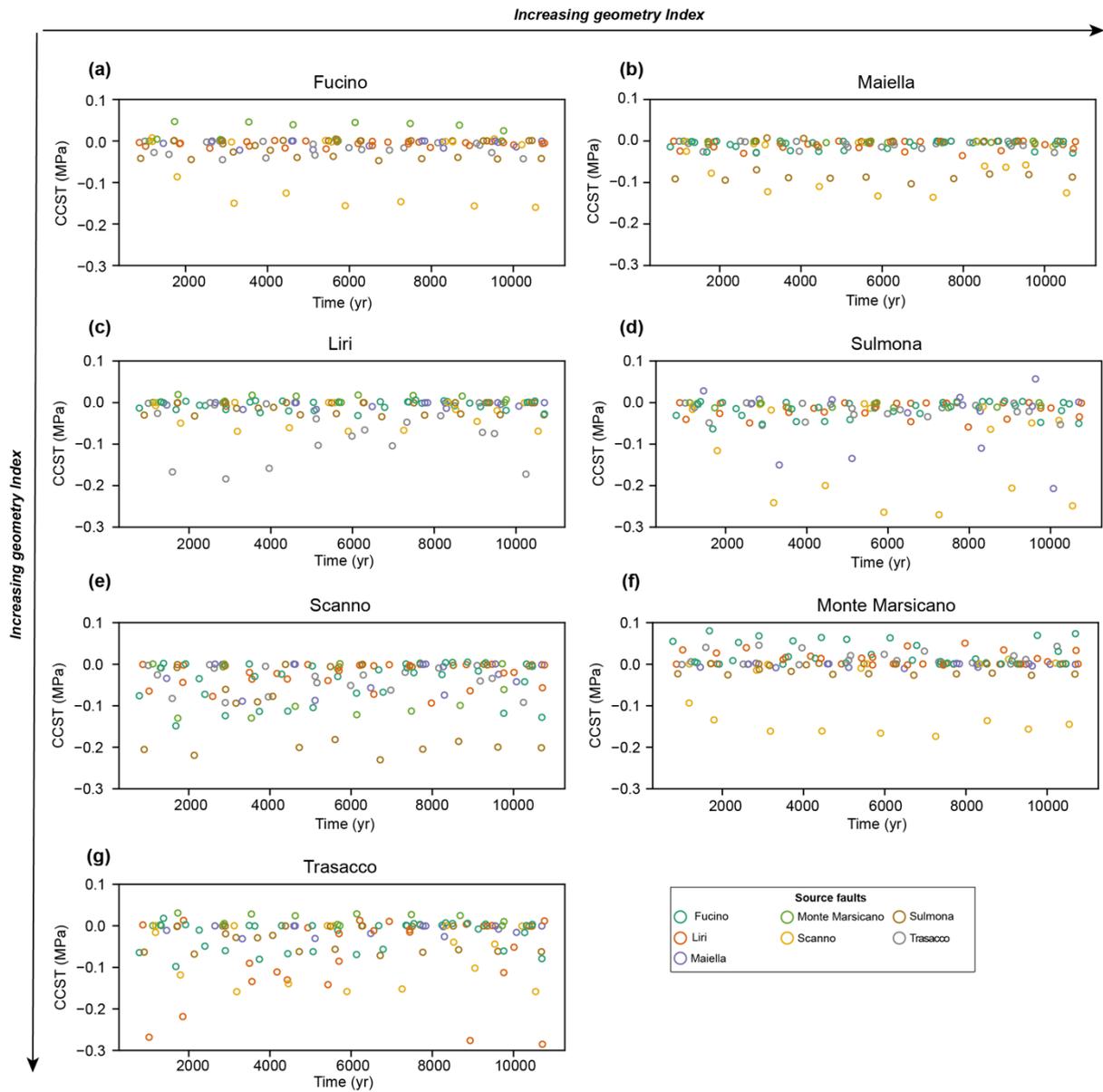
736 **Appendix E: Coseismic coulomb stress transfer**



737

738

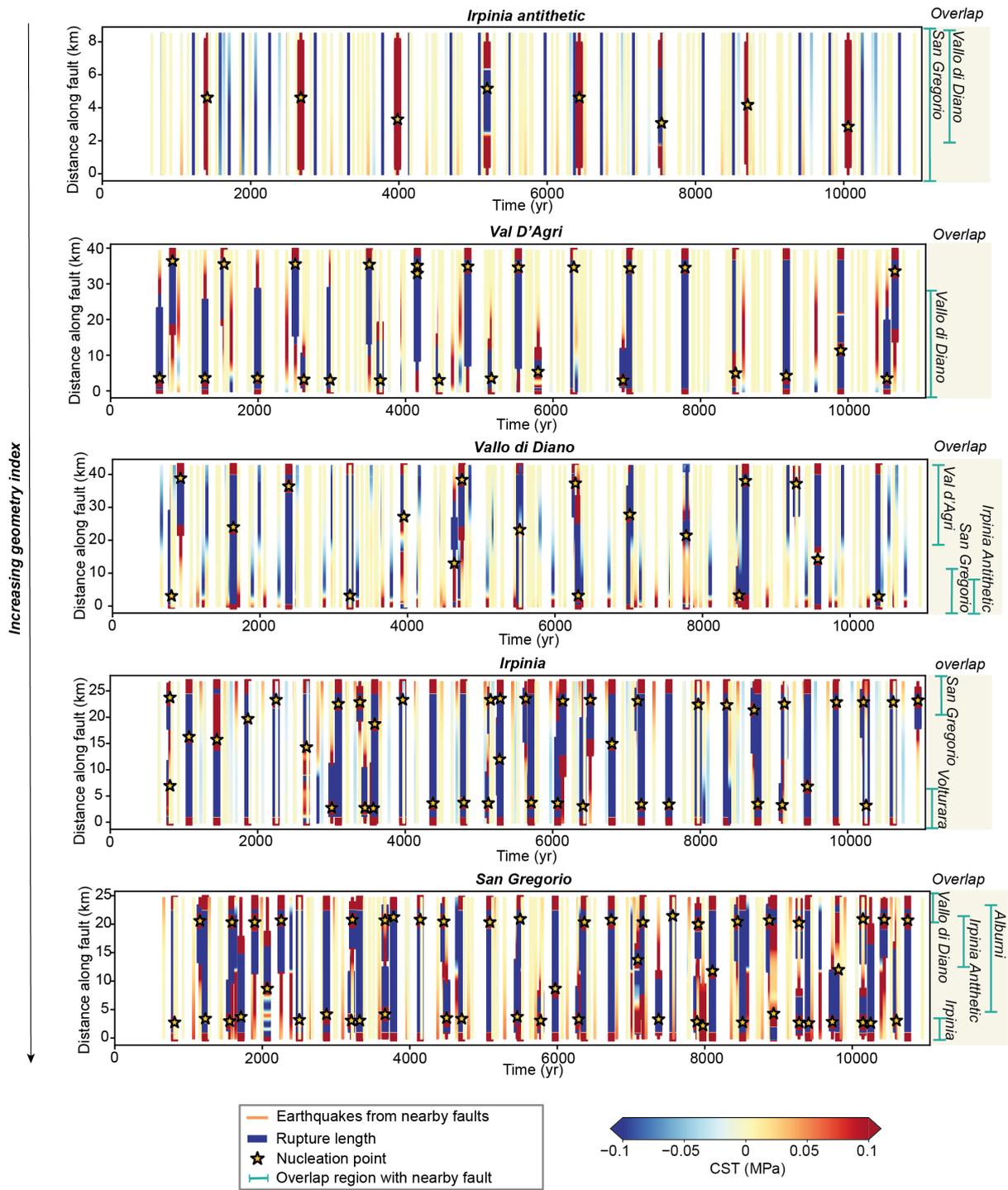
739 **Figure E1: Time series with coseismic Coulomb stress transfer (CST) averaged across the fault surface**
 740 **induced by nearby faults in the Southern Apennines.**



741

742

743 **Figure E2: Time series with coseismic stress transfer averaged across the fault surface induced by nearby**
 744 **faults in the Central Apennines.**



745

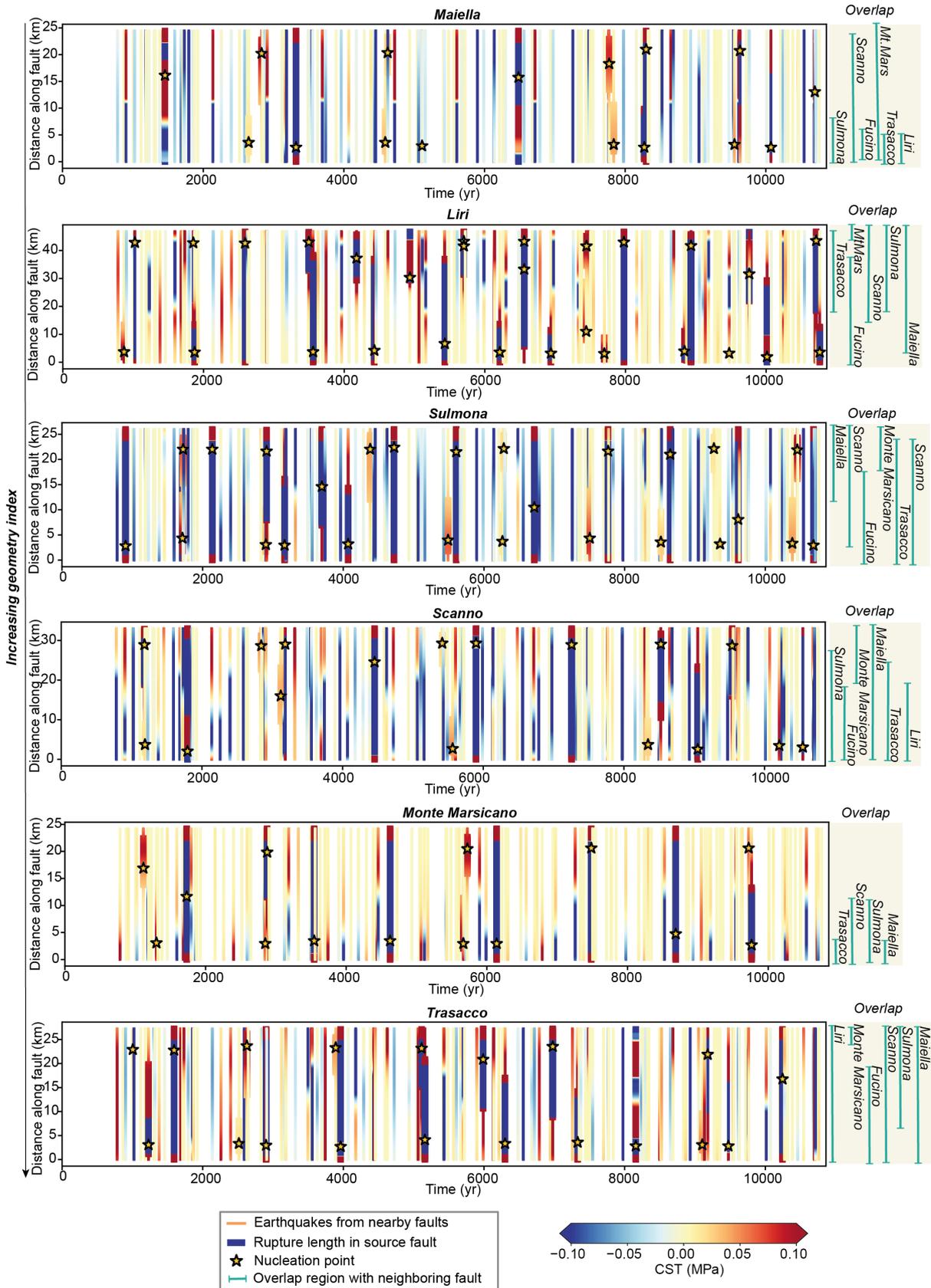
746 **Figure E3: Coseismic stress transfer (CST) in the Irpinia Antithetic, Val D'Agri, Vallo di Diano, Irpinia**
 747 **and San Gregorio faults in the southern Apennines. For the figure explanation, see caption of Figure 8 in**
 748 **the main text.**

749

750

751

752



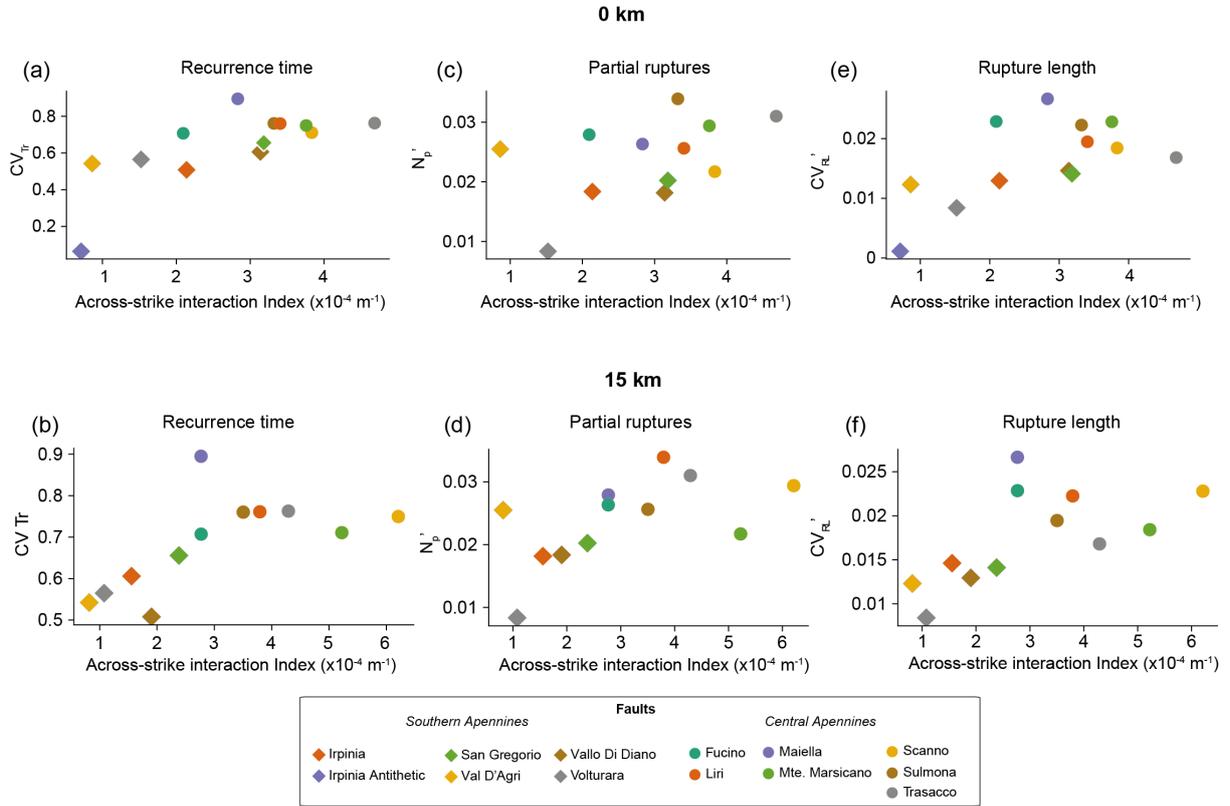
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754

755 **Figure E4: Coseismic stress transfer (CST) in the Maiella, Liri, Sumona, Scanno, Monte Marsicano and**
 756 **Trasacco faults in the Central Apennines. For the figure explanation, see caption of Figure 8 in the main**
 757 **text.**

758

759 **Appendix F: Across-interaction index**



760

761 **Figure F1: relationships between fault network geometry, described by the across-strike interaction index**
 762 **(AI), and (a-b) coefficient of variation of recurrence times (CV_{Tr}), (b-c) number of partial ruptures (N_p')**
 763 **and (d-e) coefficient of variation of rupture lengths (CV_{rl}') for faults in the Southern and Central**
 764 **Apennines. AI index corresponds to a depth of (a,d,e) 0 km and (b,d,f) 15 km.**

765

766 **Table F1: Leave-one-out Spearman's correlation coefficients (mean, standard deviation, minimum,**
 767 **maximum) for relationship between across-strike interaction index (taken at 0 km, 7.5km and 15km) and**
 768 **coefficient of variation of recurrence times (CV_{Tr}), number of partial ruptures (N_p') and coefficient of**
 769 **variation of rupture lengths (CV_{rl}') for faults in the Southern and Central Apennines**

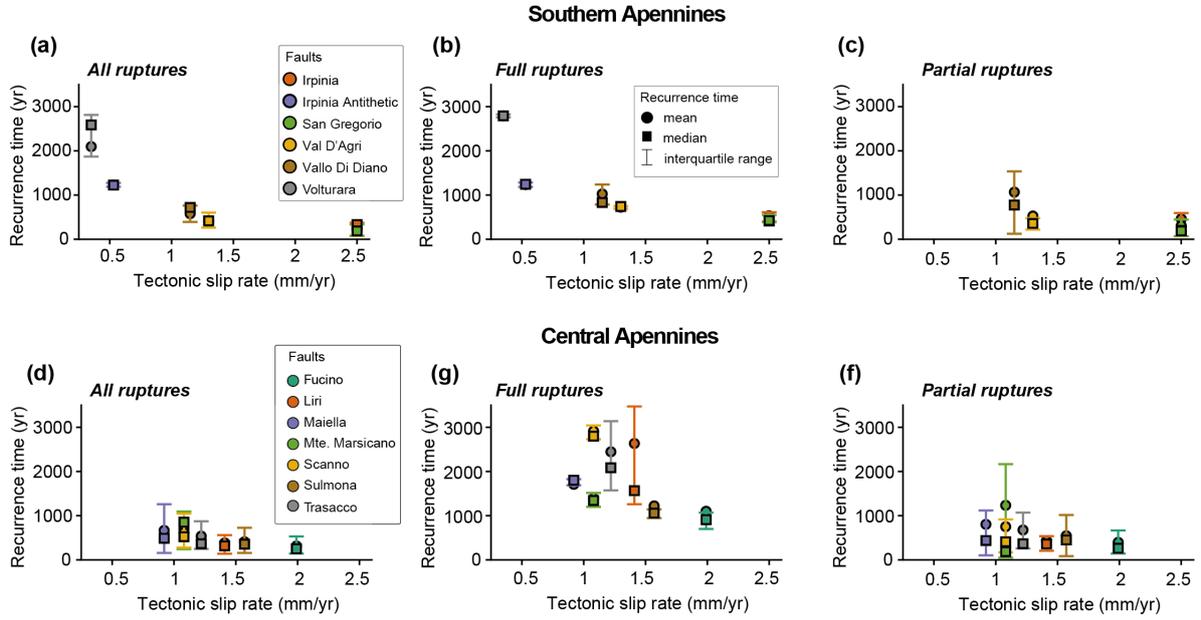
Depth (km)	metric	Mean ρ	Std ρ	Min ρ	Max ρ
Southern Apennines + Central Apennines					
0	CV_{Tr}	0.72	0.05	0.64	0.85
	N_p'	0.46	0.08	0.34	0.58

	CV_{rl}'	0.50	0.09	0.37	0.67
7.5	CV_{Tr}	0.84	0.03	0.80	0.90
	N_p'	0.73	0.05	0.68	0.85
	CV_{rl}'	0.71	0.06	0.64	0.78
15	CV_{Tr}	0.70	0.06	0.62	0.81
	N_p'	0.69	0.06	0.65	0.85
	CV_{rl}'	0.66	0.07	0.57	0.75
Southern Apennines only					
0	CV_{Tr}	0.82	0.13	0.70	1.00
	N_p'	-0.08	0.50	-0.40	0.80
	CV_{rl}'	0.87	0.05	0.80	0.90
7.5	CV_{Tr}	0.40	0.42	-0.20	1.00
	N_p'	0.04	0.54	-0.20	1.00
	CV_{rl}'	0.56	0.17	0.40	0.80
15	CV_{Tr}	0.65	0.23	0.40	1.00
	N_p'	0.04	0.54	-0.20	1.00
	CV_{rl}'	0.75	0.10	0.60	0.90
Central Apennines only					
0	CV_{Tr}	0.09	0.28	-0.43	0.43
	N_p'	0.00	0.24	-0.37	0.31
	CV_{rl}'	-0.84	0.06	-0.94	-0.77
7.5	CV_{Tr}	-0.27	0.25	-0.71	0.14
	N_p'	0.13	0.22	-0.09	0.60
	CV_{rl}'	-0.56	0.20	-0.89	-0.31
15	CV_{Tr}	0.18	0.27	-0.31	0.54
	N_p'	0.31	0.18	0.09	0.60

	CV_{rl}'	-0.70	0.12	-0.89	-0.54
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770

771 **Appendix G: Recurrence time and tectonic slip rate**



772

773 **Figure G1: Mean, median and interquartile range of recurrence time of seismic events vs. long-term**
 774 **tectonic slip rate of individual faults for (a) Southern Apennines and (b) Central Apennines**

775 **Appendix H: Calculation of magnitude-frequency distributions with the moment budget method**

776 We estimate annual cumulative earthquake rates for individual faults and fault networks using the moment-budget
 777 (MB) and activity-rate (AR) tools implement in the seismic-hazard code FiSH (Pace et al., 2016). This approach
 778 is based on the conservation of seismic moment over tectonic time scales, requiring information of the fault
 779 geometry and long-term slip-rate (Table 1). For each fault, the moment rate \dot{M}_0 that must be released by
 780 earthquakes on the fault over time is calculated as:

781
$$\dot{M}_0 = GA\dot{D} \quad (H1)$$

782 Where G is the shear modulus, A is the fault area derived from its geometry and \dot{D} is the long-term slip rate.
 783 Earthquake magnitudes M_w are related to the seismic moment M_0 through the standard relationship of Hanks and
 784 Kanamori (1979):

785
$$\log_{10}(M_0) = 1.5M_w + 9.1 \quad (H2)$$

786 We used the truncated Gutenberg Richter formulation, where the incremental magnitude frequency distribution is
 787 defined as:

788
$$v(M) = C10^{-b(M-M_{min})}, M_{min} \leq M \leq M_{max} \quad (H3)$$

789 Where b is the b -value ($b=1$), M_{min} the minimum magnitude ($M_{min} = 5.5$) and M_{max} the fault-specific maximum
790 magnitude inferred from geometry. The constant C is determined by enforcing the moment conservation, such as
791 the integral of the seismic moment released by the distribution is equal to the long-term moment rate:

$$792 \int_{M_{min}}^{M_{max}} M_0(M)v(M)dM = \dot{M}_0 \text{ (H4)}$$

793 From the incremental rates, we calculate the annual cumulative exceedance rate for each fault defined as the
794 expected annual number of events with magnitude equal or greater than m :

$$795 \lambda(M \geq m) = \int_m^{M_{max}} v(M)dM \text{ (H5)}$$

796 To obtain the exceedance rates per fault network, we summed the cumulative rates assuming independent
797 Poissonian sources. Further details on this approach are provided in (Pace et al., 2016)

798

799

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808 **Code and data availability**

809 Data to build the fault sources was taken from open-access publications (Faure Walker et al., 2019b; Mildon,
810 2017b; Sgambato, 2022; Valentini et al., 2017). QDYN is open source (Luo et al. 2017). The modified code
811 version used in this work can be found in (Rodriguez Piceda et al., 2025b). Input files to reproduce the results of
812 this work are accessible in (Rodriguez Piceda et al., 2026).

813 **Competing interests**

814 The authors declare no conflicts of interest.

815

816 **References**

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