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Title:

Localized afterslip at geometrical complexities revealed by InSAR after the 2016 Central Italy seismic sequence

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25 Keywords:

- 26 - Postseismic
- 27 - InSAR time-series
- 28 - 2016-2017 Amatrice-Norcia Seismic Sequence
- 29 - Geometrical complexity

30

31 Key points:

- 32 - We monitor pre and post-seismic deformation of the 2016 seismic sequence using two-
- 33 year InSAR time-series
- 34 - Centimetre scale post-seismic surface displacements are detected after October 30,
- 35 2016 Mw 6.5 mainshock (Norcia earthquake)
- 36 - Localized shallow afterslip occurred at structural complexity that may have hindered
- 37 the propagation of seismic ruptures

38

39

40 **Abstract :**

41 The Mw 6.5 Norcia earthquake occurred on October 30, 2016, along the Mt Vettore
42 fault (Central Apennines, Italy), it was the largest earthquake of the 2016-2017 seismic
43 sequence that started two months earlier with the Mw 6.0 Amatrice earthquake (August, 24).
44 To detect potential slow slip during the sequence, we produced Interferometric Synthetic
45 Aperture Radar (InSAR) time-series using 12 to 6-day repeat cycles of Sentinel-1A/1B images.
46 Time-series indicates that centimetre-scale surface displacements took place during the 10
47 weeks following the Norcia earthquake. Two areas of subsidence are detected: one in the
48 Castelluccio basin (hanging wall of the Mt Vettore fault), and one in the southern extent of the
49 Norcia earthquake surface rupture, near an inherited thrust. Poroelastic and viscoelastic
50 models are unable to explain these displacements. In the Castelluccio basin, the displacement
51 reaches 13.2 ± 1.4 mm in the ascending line of sight (LOS) on January 06, 2017. South of the
52 Norcia earthquake surface rupture (a zone between the Norcia and Amatrice earthquakes),
53 the post-seismic surface displacements affect a smaller area, but reach 35.5 ± 1.7 mm in
54 ascending LOS by January 2017 and follow a logarithmic temporal decay consistent with post-
55 seismic afterslip. Our analysis suggests that the structurally complex area located south of the
56 Norcia rupture (30 October) is characterized by a conditionally stable frictional regime. This
57 geometrical and frictional barrier likely halted rupture propagation during the Amatrice
58 (August 24) and Norcia (October 30) earthquakes at shallow depth (<3-4 km).

59

60 1 Introduction

61 Monitoring the spatial and temporal variations of the slip on a fault enables researchers
62 to better assess stress build-up on seismic asperities and slip released during seismic cycle
63 (Avouac, 2015; Bürgmann, 2018; Chen & Bürgmann, 2017; Harris, 2017). In the rate and state
64 formulation, rupture propagation can be hindered by rate-strengthening sections of a fault,
65 which tend to slip via creep rather than in seismic rupture (e.g., H. Perfettini et al., 2010; Hirose
66 et al., 2010). Such barriers are also often associated with structural complexities – such as a
67 change of strike, secondary faulting or interaction with inherited faults (e.g., King & Nabelek,
68 1985; King, 1986; Wesnousky, 1988). These structural complexities can act as a geometrical
69 barrier, and are characterized by an increased equivalent strength (Nielsen & Knopoff, 1998).
70 Locating aseismic slip on the fault and comparing these locations with those of seismic slip and
71 fault segmentation is of pivotal importance to better characterize the frictional behavior of a
72 fault system and its relation with structural complexities.

73 A notable seismic sequence occurred in the Central Apennines (Italy) in 2016-2017
74 (Chiarabba et al., 2018; Perouse et al., 2018; Cheloni et al., 2017; Huang et al., 2017; Cirella et
75 al., 2018; Civico et al., 2018; Scognamiglio et al., 2018; Ragon et al., 2019), with four main
76 events: the 24th August 2016 M_w 6.0 Amatrice event, the 26th October 2016 M_w 5.9 Visso
77 event, the 30th October 2016 M_w 6.5 Norcia event, and the 18th January 2017 M_w 5.5
78 Campotosto event (Figure 1 and Table S1). This seismic sequence ruptured the complex Mt
79 Vettore fault system (in red in Figure 1) (Pizzi et al., 2017; Porreca et al., 2018; Villani, Pucci, et
80 al., 2018), and the adjacent Amatrice-Campotosto fault (in orange in Figure 1). During the
81 Norcia earthquake, the rupturing of an antithetic fault on the opposite side of the Castelluccio
82 basin seems necessary to fit geodetic data and is supported by alignments of relocated
83 aftershocks (Chiaraluce et al., 2017; Walters et al., 2018; Cheloni et al., 2019). In addition, the
84 role of an inherited west-dipping thrust called OAS (Olevano-Antrodoco-Sibillini thrust) in the
85 Norcia earthquake coseismic rupture geometry has been widely discussed. While some studies
86 suggested that only the Mt Vettore fault system was activated (Chiaraluce et al., 2017; Liu et
87 al., 2017; Papadopoulos et al., 2017; Pavlides et al., 2017; Pizzi et al., 2017; Wang et al., 2018;
88 Xu et al., 2017), others suggested that also the OAS thrust ruptured as a reactivated high-angle
89 normal fault during the event, as suggested by geodetic and seismological observations
90 (Cheloni et al., 2017; Scognamiglio et al., 2018; Walters et al., 2018). Although the reactivation
91 of the thrust is not clearly demonstrated (Cheloni et al., 2019), the OAS appears to have played

92 a role in the aftershock distribution (Chiarabba et al., 2018; Chiaraluca et al., 2017; Pizzi et al.,
93 2017). The earthquakes of the 2016 sequence appear to have nucleated near crosscutting
94 structures that seem to have been loaded by previous ruptures in the sequence (Chiaraluca et
95 al., 2017; Pino et al., 2019). This seismic sequence is thus an excellent case study to better
96 understand the link between structural segmentation, aseismic slip and frictional properties
97 that might control this rupture propagation.

98 Post-seismic processes during this sequence have been observed in the seismicity (e.g.,
99 Albano et al., 2018; Tung & Masterlark, 2018), but no aseismic slip has been detected with
100 geodetic data so far. Using Interferometric Synthetic Aperture Radar (InSAR) time-series from
101 Sentinel-1 data, we document a small but significant post-seismic deformation transient that
102 we find is best explained by aseismic slip on faults associated with this sequence. Surface
103 displacements are presented and analysed. We also explore simple modelling schemes that
104 provide a framework for our interpretation and discussion.

105 2 Geological setting

106 The Central Apennines were affected by an extensional phase during the Jurassic,
107 followed by a compressive phase during the Neogene (e.g., Calamita et al., 2011). The OAS
108 (Olevano-Antrodoco-Sibillini thrust) is one the main thrusts resulting from the Neogene
109 compressive phase (Calamita et al., 1994) and has been interpreted as a transpressive ramp
110 (Di Domenica et al., 2012). In the area affected by the 2016-2017 seismic sequence, the OAS
111 thrust is expressed by parallel splays associated with fault-bend folds characteristic of
112 structural ramps (Calamita et al., 2012) (Figure 1). The ongoing ENE oriented extension of 2
113 to 4 mm/yr (D'Agostino, 2014; Carafa & Bird, 2016; Devoti et al., 2017), which probably began
114 in the Early Pleistocene (e.g., Galadini & Galli, 2000), is currently accommodated through
115 normal fault systems such as the Monte Vettore fault system that hosted the 2016-2017
116 seismic sequence.

117

118 3 Surface displacements during the seismic sequence

119 3.1 InSAR processing

120 Synthetic radar interferometry (InSAR) is now systematically used to constrain
121 deformation fields (Elliott, Walters, et al., 2016) and can document centimetre to millimetre
122 scale slow aseismic ground deformation using an adapted processing chain (Hussain et al.,

123 2018; Aslan et al., 2019). We used C band (5.5 cm wavelength) images from Sentinel-1A/B
124 images (Figure 1-A) spanning almost two years (July 28, 2015, to June 11, 2017) for the
125 ascending track (A117, subswath IW3). To confirm the main observations made on the
126 ascending track, we processed descending track (D22, subswaths IW2 and IW3) images from
127 October 26, 2016 to February 11, 2017. SAR images were processed in VV polarization. We
128 used the NSBAS processing chain (Doin et al., 2011, 2015) modified for Sentinel data by
129 Grandin (2015) to generate differential interferograms. The interferogram network, and
130 examples of unfiltered and uncorrected interferograms, are provided in Figures S1 and S2. The
131 Shuttle Radar Topography Mission digital elevation model (DEM) at 3 arc sec resolution (Rabus
132 et al., 2003), resampled at 45m resolution, has been used to accurately coregister the focused
133 SAR images and to correct interferograms from the topographic contribution to the
134 interferometric phase. We removed a ramp in range and in azimuth for each interferogram
135 using the methodology of Cavalié et al., (2007) included in NSBAS. We de-noised the
136 interferograms before unwrapping using collinearity, a criterion to characterize at each pixel
137 the local spatial variability of the phase (Pinel-Puységur et al., 2012). The collinearity was used
138 to adapt the strength of the filter. We filtered for a window of 12 pixels (400m in range and
139 700m in azimuth); the filter is described in Doin et al. (2011) and is based on the collinearity
140 value which weights the complex phase in a sliding window. For the filter we had the option
141 of adapting the weighting of the phase within windows of different sizes. This size adaptation
142 depends on the collinearity within the windows. Unwrapping was performed in 2D with the
143 NSBAS chain (Grandin et al., 2012; Doin et al., 2015). After unwrapping, to account for errors
144 associated with stratified troposphere, we removed a quadratic cross-function of elevation (z)
145 and azimuth to ramps in azimuth (y) and in range (x) estimation following the function
146 $ax+by+c+ez+fz \times az+g \times (z \times az)^2$ using a least-square approach (Daout et al., 2019). Time-series
147 were then calculated following the NSBAS method (Doin et al., 2011; Daout et al., 2016) using
148 an approach based on the Small Baseline Subset time-series Analysis (SBAS) of López-Quiroz
149 et al., (2009)'s algorithm. The smoothing of the pixel time-series is performed by minimizing
150 the Laplacian of the temporal evolution of the deformation (Cavalié et al., 2007). The final pixel
151 size is 62 m in azimuth and 37 m in range. We removed pixels with an RMS value greater than
152 0.7. For the ascending track, we build a time series spanning the 2 years (from 28/07/2015 to
153 11/06/2017) (Figure S1 and see missing links in the time series in Figure S3). For this complete
154 2-year time-series, we encounter problems in unwrapping the co-seismic interferograms due

155 to the large deformation with respect to the Sentinel wavelength in near field (aliasing).
156 Fringes are too close in space and cannot be unwrapped. At pixels that are incoherent in the
157 co-seismic interferogram (i.e. in near field), this causes gaps in the complete (2 yrs) time-series
158 spanning the main earthquakes (Norcia, Amatrice and Campotosto earthquakes) (Figure S3),
159 or lead to an underestimation of the coseismic displacement (brown circles in Figure 2-D). In
160 addition, we also build three time-series in between earthquakes in order to avoid possible
161 bias: before the 24th August (Amatrice earthquake), between 27th August and 26th October
162 (between the Amatrice and Visso earthquake), and after the 30th October (Norcia earthquake).
163 We excluded the SAR data from January 18, 2017, which produced noised interferograms. For
164 the descending track, we build two time-series: between 27th August and 26th October (Visso
165 earthquake), and after the 30th October (Norcia earthquake).

166 3.2 Time-series results and description of the main features

167 3.2.1 Ascending Track

168 The ascending time-series built before the seismic sequence (from July 28th 2015 to
169 August 21th 2016) does not show significant nor localized surface displacements along the
170 main faults (Figure S4-A,B). After October 30th, the (post-Norcia) time-series shows
171 centimetre-scale displacements going away from the satellite in the LOS direction in three
172 areas (Figure 2-B):

- 173 • South of Amatrice: In the area affected by the Campotosto earthquake (January
174 18th, 2017 Mw 5.0 -5.5 EQ) the ground surface moved away from the satellite by
175 more than 60 mm in LOS (Figure 2-B). This coseismic displacement results in a step
176 function in the time series, and can therefore be easily separated from any gradual
177 post-seismic deformation.
- 178 • Near Arquata del Tronto: At the southern extremity of surface rupture of the Mw
179 6.5 October 30th Norcia earthquake (red faults in Figure 2), surface displacements
180 are detected over an area of $\sim 12 \text{ km}^2$, and follow logarithmic evolution (in Figure
181 2-D and Figure 3-A see time-series at point 1 where the cumulative post-Norcia
182 displacements in LOS on January 06, 2017 is in average $\sim 35.5 \pm 1.7 \text{ mm}$ and reach
183 $50.5 \pm 2.1 \text{ mm}$ on 30 April 2017)
- 184 • Castelluccio Basin: On the hanging wall of the Mt Vettore fault, slow deformation
185 affects an area of $\sim 50 \text{ km}^2$ (Figure 2), and is associated with displacements in the

186 LOS direction that are 13.2 ± 1.4 mm in average at Point 2 on 6 January and reach
187 37.9 ± 1.3 mm on 30 April 2017 (Figure 2-D and Figure 3-A).

188 To rule out possible bias due to the Campotosto earthquakes affecting the area near
189 Castelluccio and near Arquata del Tronto, we confirmed these previous observations with a
190 shorter time-series calculated between the Norcia and Campotosto earthquakes (November
191 1st – January 12th) (Figure 4). We prefer to use, for the rest of the manuscript, the longer post-
192 Norcia time-series (November 1st – June 11th) that shows a better signal-noise ratio.

193 In the post Amatrice earthquake time series (August 27 to October 26) we did not observe
194 any localized surface displacements similar to the pattern observed after October 30 (Norcia
195 earthquake) (Figure S4-B). Yet near the town of Amatrice, we observe diffuse surface
196 displacements ($< \sim 2.5$ cm) moving away from the satellite. However, this time-series is
197 constrained by only 9 scenes and 20 interferograms, which prevents from properly (i)
198 constraining a low amplitude signal and (ii) correcting for atmosphere and topography. The
199 surface displacements here have a low signal to noise ratio. The variance of the uncorrelated
200 noise is ~ 40 mm² (see Supplementary Text S1 and Figure S5-A), and the standard deviation
201 sigma of the noise is thus 6.3 mm. We take $3 \cdot \text{sigma} = 20$ mm to set our limit of detection. The
202 characteristic length scale of correlated noise is 5.0 km and there is autocovariance for
203 distances smaller than ~ 15 km (see Figure S5-A). Observed patterns cannot be differentiated
204 from noise, analyses on SAR images from other satellites should be carried out to confirm or
205 not the post-Amatrice (August 24 earthquake) surface displacements.

206 3.2.2 Descending Track

207 To confirm the post-October 30 (Norcia earthquake) observations we processed
208 descending interferograms (Figure 2-C). We used a 3-month dataset for the descending track
209 (November 1st, 2016 to February 11, 2017). Time-series calculated for the descending track
210 also indicate slow deformation after Norcia earthquake (October, 30), that reaches on average
211 20.5 ± 2.7 mm on January 24 in the LOS direction near Arquata del Tronto (point 1). The
212 deformation reached on average 10.8 ± 1.5 mm in the Castelluccio basin (in point 2). Assuming
213 negligible north-south displacements, by combining the results from ascending and
214 descending tracks, the displacement is dominated by subsidence in this area (Figure S6).

215 In the first order, deformation observed in the descending track is compatible with
216 results inferred from the ascending track. Noise and unwrapping issues led us to mask noisy
217 areas and resulted in more blank pixels in the descending track picture which should be used

218 with caution. The relief projected in the descending LOS geometry masked a ridge (Mt Bove -
219 Mt Vettore – Mt Gorzano high massifs) (Figure S7). In addition to noise, this resulted in an
220 incoherent area and made unwrapping difficult in that area. The temporal evolution of the
221 features slightly differs from the one in the ascending track. The removal of the 7th November
222 SAR images (affected by strong atmosphere conditions) and the poor constraint on the 19th
223 November SAR images (see the network in Figure S1) could affect the time-series in November.
224 Falcucci et al. (2018) had similar difficulties in using descending SAR images to survey the
225 January 2017 Campotosto seismic event. This might be due to the early morning acquisition
226 time, which amplifies decorrelation due to change of moisture level or thawing.

227 In addition, we produced time-series calculated between the Amatrice and Norcia
228 earthquakes. This time-series is also affected by a short-wavelength atmospheric turbulence
229 that makes it difficult to interpret (Figure S5-B-C). The variance of noise is $\sim 21 \text{ mm}^2$. The
230 characteristic length scale of correlated noise is 5.0 km and there is autocovariance for
231 distances smaller than $\sim 15 \text{ km}$ (see Figure S5). Again, displacement less than 19 mm for the
232 ascending track and under 14 mm for the descending track, cannot be considered as
233 detectable signal, therefore constraining smaller surface displacements between the Amatrice
234 and Norcia events demands complementary time-series with SAR images from other satellite.

235 3.2.3 Comparison with GNSS

236 Few GNSS stations exist near the studied area (ARQT, LNSS see station locations in Figure
237 2-A). Those GNSS time-series agree with InSAR time-series sampled at the same locations
238 (Figure S8).

239 3.3 Pattern temporal evolution

240 By sampling the time-series, the temporal evolution is shown by averaging pixel
241 displacement in area. To enhance the precision of the displacement temporal evolution
242 without using time-series inversion, we temporally track a stable pattern in our unwrapped
243 interferograms (Grandin, 2009). The method is detailed in Supplementary Text S2 and is
244 illustrated by Figures S9-S10.

245 This method can only be used on areas where the displacement pattern is well defined,
246 and works well in our case to track the evolution of the area near Arquata del Tronto (pattern
247 in Figure 4 which encompasses the point 1). After the Norcia earthquake, the amplitude ratio
248 is higher, and a logarithmic decay is clearly identified there (Figure 3-B). The pattern tracking
249 shows a temporal evolution similar to the averaging surface displacement method. We are

250 therefore confident that pixel averaging adequately captures the evolution of displacement
251 that we want to describe. For the descending interferograms, the pattern is not precisely
252 trackable and its time evolution is noisy (grey curve in Figure S10-C). This could be due to a
253 poor correlation for the pattern in some interferograms. The evolution of the pattern around
254 Castelluccio Basin (pattern in Figure 4) is poorly resolved too.

255 4 Modelling 3D deformation field

256 4.1 Comparison with topography and geology

257 To interpret the displacement pattern near Arquata del Tronto we compared it with the
258 topography and the geology. There is no correlation with either the slope or the relief (Figure
259 S11), thus we can exclude a gravitational process (i.e. landslide) as an explanation for the
260 observed displacements. Concerning geology, the OAS thrusts fractured Meso-Cenozoic
261 carbonate rocks over Neogene pelagic sediments; the carbonates host some of the largest
262 aquifers in the Apennines while the Neogene sediments are thought to be an aquiclude (Figure
263 S12) (Boni et al., 2010). The OAS delimiting these two units therefore acts as an impermeable
264 boundary (Boni et al., 2010). Petitta et al., (2018) and Valigi et al., (2019) show that the 2016-
265 2017 Italian seismic sequence had effects on spring discharge, water-table levels, and
266 streamflow of those aquifers. In our results, the extent of the Maiolica unit, which hosts a
267 shallow aquifer (Boni et al., 2010), corresponds well with the deformation area near Arquata
268 del Tronto (Figure S11-A-B). The long pre-seismic InSAR time-series did not show any seasonal
269 deformation in this aquifer, which means that groundwater seasonal processes can be
270 excluded (point 1 in Figure 2-D). In addition, it seems difficult to get more than 1 cm of
271 subsidence from winter rain in this area (i.e., Silverii et al., 2016).

272 4.2 Poro-elastic modelling

273 Poro-elastic effects have been studied during the 2016-2017 seismic sequence to explore
274 the aftershocks and earthquakes triggering. Tung & Masterlark (2018) suggest that fluid
275 migration caused by the Amatrice earthquake (August 24, 2016) could have triggered the Visso
276 earthquake (October 26, 2016), as well as some associated aftershocks. They inferred that this
277 poro-elastic triggering should have occurred in an intermediately fractured crust. They also
278 estimated that afterslip and viscoelastic-relaxation were quite negligible with respect to
279 poroelastic effects. Albano et al. (2018) also suggest that post-seismic fluid diffusion after the
280 Amatrice earthquake is related to aftershocks. According to their pore fluid diffusion model,
281 there could be some associated afterslip (~ 10 cm) after the Amatrice earthquake.

282 The Maiolica karst aquifer is shallow and unconfined but Roeloffs (1996) postulates that
283 every aquifer reacts as a confined aquifer to a disturbance at short timescale. Here, to compare
284 with the displacement pattern near Arquata del Tronto, we thus test a forward model of poro-
285 elastic rebound using Relax software (Barbot & Fialko, 2010). We perform a simple poro-elastic
286 model with no lateral variation of diffusivity, that does not account for the spatially
287 heterogeneous distribution of local aquifers. We use the slip distribution models of the Norcia
288 and Visso mainshocks (Maubant et al., 2017), (Figure S13 and Figure 5–A). We do not take into
289 account the Amatrice earthquake since poroelastic rebound from the Amatrice earthquake
290 would likely have stabilized by the time of the Visso and Norcia earthquakes (Figure 4-f in
291 Albano et al., (2018)). We tested several diffusivities for a shallow layer (0-5 km depth), from
292 1.5 m².s proposed by the Tung and Masterlark (2018)’s aftershocks analysis to 10⁴ m²/s
293 (maximum diffusivity value for karst) (Roeloffs, 1996). Other parameters are described in detail
294 in Figure 5. This model predicts uplift in the Castelluccio Basin and near Arquata del Tronto
295 (Figure 5-D) where we observe subsidence (Figure 5-C). This discrepancy allows us to rule out
296 poro-elastic rebound as the main driver of observed deformation. These poroelastic models
297 are not exhaustive, we only explore simple geometrical configurations using parameter values
298 informed by the local geology, however, other karstic configurations would likely predict uplift
299 as well.

300 4.3 Viscoelastic modelling

301 Viscoelastic relaxation in the crust and in the mantle is also an important aseismic
302 process, and has been inferred to be the driver of post-seismic deformation following many
303 earthquakes (e.g., Pollitz et al., 2001; Zhao et al., 2017). To test whether viscoelastic relaxation
304 is a plausible driving mechanism for post-Norcia deformation, we used Relax to perform
305 forward modelling in a simple layered framework. Our model uses the stress perturbation
306 from the three mainshocks: the Amatrice earthquake using the slip distribution of Ragon et
307 al., (2019) added to the Visso and Norcia earthquakes using the slip distribution model of
308 Maubant et al., (2017). We set the upper, middle and lower crust thickness to the values
309 proposed in Laske et al., (2013) and Verdecchia et., al. (2018) (see Table 1). We tested a
310 viscoelastic relaxation governed by a Newtonian rheology with $\dot{\gamma} = \frac{\tau}{\eta}$ where $\dot{\gamma}$ is the viscous
311 strain rate, τ is the deviatoric stress and η is the Newtonian viscosity. We used viscosity values
312 η inferred in studies of viscoelastic relaxation modelling using GPS measurements following
313 1997 Umbria-Marche earthquakes, or using levelling line measurements following the 1915

314 Fucino earthquake (Amoruso et al., 2005; Aoudia et al., 2003; Riva et al., 2007). These studies
315 inferred that the ductile structure of the Central Apennines consists of an elastic upper crust
316 overlaying a middle crust (10^{18} - 10^{19} Pa s), a viscous lower crust (10^{17} - 10^{18} Pa s) and the upper
317 mantle (10^{21} Pa s) (Table 1).

318 Although those models are geometrically simple and do not take into account
319 earthquakes prior 2016, they predict at a long-wavelength uplift (>50 km) in the area where
320 we observe subsidence (Figure 6). This discrepancy in both sign and wavelength allows us to
321 rule out a simple visco-elastic relaxation as the main driver of observed deformation. Visco-
322 elastic models are not exhaustive here: nonlinear viscoelastic rheologies (e.g., power law
323 creep) in lower crust (Freed & Bürgmann, 2004) could also be investigated; however finite
324 element models with creeping lower crust seem also to predict uplift in similar settings of
325 normal fault systems (e.g., Thompson & Parsons, 2016). It is worth also mentioning that the
326 observed logarithmic decay could also potentially be associated with Burgers rheology or shear
327 zone (e.g., Hetland & Zhang, 2014) and could also be investigated in future work. Although, as
328 with the poroelastic modeling, it seems unlikely that a different geometry or rheology could
329 completely reverse the sign of uplift and subsidence (for example, a different rheology would
330 only affect the temporal evolution/spatial distribution).

331 4.4 Modelling the temporal evolution of afterslip

332 The logarithmic-like temporal evolution of displacement (Figure 3-A), and the
333 disagreement between the observations and the predictions of poroelastic and viscoelastic
334 models, suggest that afterslip may have been the main driver of postseismic deformation
335 (Marone et al., 1991). Thus to characterize the temporal evolution of the deformation, we
336 fitted the deformation decay (of the raw time-series and of the pattern tracking evolution)
337 with the logarithmic function (Figure 3-A-B respectively) from the Marone et al. (1991) model
338 and Zhou et al. (2018) reformulation for rate-strengthening afterslip:

$$339 \quad U(t) \approx \alpha \times \ln(1 + c \times t)$$

$$340 \quad c = \frac{\beta \times V_i}{\alpha}$$

341 In which U: afterslip in time t, α : characteristic length scale, $\beta \times V_i$: initial rate at the
342 beginning of the post-seismic period, with β a scaling vector by which the sliding rate evolves
343 in response to the stress and V_i the pre-seismic slip rate.

344 At point 1 (near Arquata del Tronto), the good fit of the afterslip model suggests that the
345 main post-seismic process is likely an afterslip phenomenon. Since we found similar results for
346 c value (describing the temporal decay) using the raw pixel time series and using the pattern
347 tracking (independent for the time-series inversion), we are confident in our fitting (Table 2).
348 The first post-Norcia Sentinel image has been acquired 2.6 days after the Norcia earthquake;
349 therefore those 2.6 days of early afterslip are missing in our data. We compare the temporal
350 decay of the deformation with the cumulative moment and number of aftershocks (blue and
351 green curves in Figure S14). Although the completeness magnitude is high (ML 3 in Figure S5
352 in Chiaraluce et al., (2017)), those two curves obtained with less than 50 events are not
353 following the same trends, aftershocks do not seem to be induced by afterslip. At point 2 (near
354 Castelluccio di Norcia), the temporal decay does not fit with the afterslip law as the rate at the
355 beginning of the post-seismic period is close to zero (c value for the purple curve fit in Figure
356 3-A). Other post-seismic processes could be a complementary driver of the deformation here
357 and should be taken into account in further modelling (i.e. finite element forward model
358 accounting for afterslip superimposed with poro-elastic rebound and viscoelastic relaxation).

359 4.5 Afterslip modelling on faults

360 4.5.1 Inversion strategy and fit to the data

361 To obtain the afterslip distribution we use the Classic Slip Inversion (CSI) Python tools
362 (Elliott, Jolivet, et al., 2016; Jolivet et al., 2015) to invert for the slip. We use the constrained
363 least-squares formula of Tarantola (2005) to solve the inverse problem (see details in
364 Supplementary Text S3 and Figures S15 to S18). We chose a dip of 40° for the Mt Vettore fault
365 (as Cheloni et al., (2017) see Figure S15) and project the fault down dip from the mapped fault
366 at the surface (modeled fault geometry in yellow Figure 2-A). We discretize the fault into 88
367 rectangular patches. The smoothed cumulative surface displacement on February 11, 2017,
368 measured in both ascending and descending tracks are inverted to obtain the afterslip
369 distribution. The resolution is good for short-wavelength features at shallow depths (<5 km
370 depth) (Figure S16). As several geometries have been used to model the mainshock rupture
371 (OAS reactivation, antithetic fault), we explored three cases: (1) slip only on the Mt Vettore
372 Fault, (2) slip on the Mt Vettore Fault and on the OAS and (3) slip on the Mt Vettore Fault and
373 an antithetic fault with a geometry similar to the one proposed by Cheloni et al., (2019) and
374 Maubant et al., (2017) (dip 65°) (Figures S17 to S20). We also explored the rake: (1) dip-slip
375 only and (2) variable rake. A rigorous statistical comparison between the cases is difficult due

376 to variation in model parameters and number of degrees of freedom, as explained by Cheloni
377 et al., (2019). We thus used the RMS values to compare the different cases. Here the case that
378 reduces the residuals the most is the inversion with an antithetic fault. This case is in addition
379 consistent with geological and seismological observations (i.e. Cheloni et al., 2019). The
380 different cases were compared using the RMS (root mean square), and allows us to choose the
381 inversion with an antithetic fault (Table S2).

382 Both ascending and descending tracks are well modelled (Figure 7). It is worth noting
383 that the resolution is low at depth (Figure S16). The total geodetic moment released by the
384 afterslip model is equivalent to $M_w \sim 5.78$ (without taking into account the Campotosto
385 earthquake) which corresponds to $\sim 8.7\%$ of the geodetic moment released during the Norcia
386 earthquake ($M_w 6.5$).

387

388 4.5.2 Afterslip distribution

389 4.5.2.1 Near Arquata del Tronto

390 We obtain a maximum afterslip of ~ 100 mm below Arquata del Tronto at shallow depth
391 (0- 2 km depth) (Figure 7 Figure 8). The shallow part depth $< 3-4$ km toward the south is
392 associated with very low coseismic slip for both the Norcia and Amatrice earthquakes and is
393 located at the edge of the coseismic asperities. This is also consistent with an afterslip process:
394 the afterslip is located where the slip gradient increased the shear stress on the unruptured
395 portions. This feature has also been observed after the L'Aquila earthquake (D'Agostino et al.,
396 2012). We calculate a geodetic moment released equivalent to $M_w \sim 5.1$ in this area (patches
397 above 2km depth and south of Arquata del Tronto). The cumulative moment released by the
398 seismicity (Figure S14-D) at this date corresponds to $\sim 2.0\%$ of the moment released by the
399 afterslip model. The deformation is thus mainly aseismic.

400 Brozzetti et al., (2019) found some surface ruptures south of the 30th October Norcia
401 rupture ($42^\circ 47'N$ in their Fig. 1-C) mapped by Villani et al., (2018). The ruptures mapped by
402 Brozzetti et al., (2019) are subtle and according to us may correspond to post-seismic
403 deformation, as was observed after the L'Aquila earthquake (D'Agostino et al., 2012).

404 4.5.2.2 Castelluccio Basin

405 We observe a maximum slip of ~ 170 mm below Castelluccio at ~ 5 km depth. Some
406 afterslip overlaps with the coseismic rupture area, but not in the area of maximum coseismic
407 slip (Figure 7). As also the shape of the measured time-series did not fit with the afterslip's law

408 (Figure 3-A), a sole afterslip process is unlikely. It is possible that poroelastic and fluid flow
409 processes could be at work here since there is a large basal aquifer (Boni et al., 2010).
410 Modelling accounting for both processes could be performed, for example fully coupled
411 poroelastic finite element numerical modelling with spatially variable material properties (e.g.,
412 Albano et al., 2017).

413 5 Discussion

414 5.1 Afterslip mechanism near Arquata del Tronto

415 Based on Dietrich (1979), Ruina (1983) and Marone et al., (1991) equations, we
416 estimated α (the characteristic length scale over which the elastic stress changes by order of
417 the frictional stress) in Figure 3 and Table 2. We can now estimate the friction
418 parameter $(a - b)$ since according to Perfettini and Avouac (2004) and Zhou et al., (2018):

$$419 \alpha = \frac{(a - b) \times \sigma}{k}$$

420

421 with k : the effective stiffness, and σ the effective normal stress. This relation can be applied
422 to our case since the duration of our analysis (~ 200 days) is much shorter than the
423 characteristic time t_d (Gualandi et al., 2014). Where $t_d = \frac{\alpha}{V_{pl}}$ (> 12 years) where V_{pl} is the local
424 plate loading rate (< 2.1 mm/yr (Puliti et al., 2020)). The fit to the displacement of the time-
425 series sampled in point 1 by an afterslip law (Figure 3) leads to $\alpha = 30.9 \pm 4.0$ mm. To estimate
426 σ , we assume that the mean normal stress may vary from the hydrostatic to lithostatic
427 pressure, with a rock density of 2.5 kg.m^{-3} (Albano et al., 2018) and a depth of the slipping area
428 of 3-5 km. This leads to $\sigma = 44\text{-}124$ MPa. For $k = G/h$ with $G = 30$ GPa (the shear modulus near
429 the surface) and $h = 8$ km (rate-strengthening depth based on the coseismic slip models), we
430 obtain $(a-b) = 3 \times 10^{-3} - 7.8 \times 10^{-4}$. More complex models (e.g. finite element model,
431 heterogeneous $(a-b)$ values) could be performed by future studies in order to reproduce the
432 surface displacements and to propose a more precise value of $(a-b)$. However, our estimation
433 is in agreement with experiments on carbonates. Scuderi and Collettini (2016) found values of
434 $(a - b)$ evolve from velocity strengthening behaviour ($a - b \approx 0.005$) at fluid pressure condition
435 of sub-hydrostatic to a velocity neutral behaviour ($a - b$ approaching 0), when the fault is at
436 near lithostatic fluid pressure. Pluymakers et al., (2016) found that wet anhydrite and dolomite
437 gouges at depths < 6 km, exhibit $(a-b)$ values ranging from 10^{-2} to 10^{-4} .

438 In Arquata del Tronto, the obtained (a-b) positive value (0.0026) is in agreement with
439 a velocity-strengthening area. The obtained (a-b) value is also very low and corresponds to a
440 quasi-neutral rate-dependency of friction implying a high sensitivity to stress perturbations.
441 Potential stress perturbations needed to reach strengthening may involve either a decrease in
442 effective normal stress and / or an increase in shear stress (Scholz, 1998). Walters et al., (2018)
443 calculated the Coulomb stress change (CFF) after the Norcia earthquake. At shallow depths,
444 they found a positive CFF change below Arquata del Tronto, which might have triggered the
445 observed afterslip in this area. An increase in pore fluid pressure would have a similar effect.
446 The afterslip is located near to the Olevano-Antrodoco-Sibillini (OAS) thrust. Chiarabba et al.,
447 (2018) detected a high V_p/V_s near the OAS thrust (around 3 km depth), indicating high pore
448 pressure that could have been further increased by nearby major shocks (Amatrice and Norcia
449 earthquake). This high pore pressure could have induced the observed afterslip allowing for
450 the release of the residual accumulated stress. This scenario is also supported by the fact that
451 pore fluid pressure variation is common in this area, where the inherited structures seem to
452 control the rupture pattern and could also concentrate high shear stress. Future work could
453 examine detailed Coulomb Stress modelling with high spatial resolution taking all fault
454 complexity and pore fluid effects into account. These works could be useful, despite some
455 limitations related to the resolution of the coseismic slip and by the ambiguity about which
456 faults were involved.

457 If this overlapping of coseismic slip and afterslip is real below 4-5 km depth (below
458 Arquata del Tronto) could be an artefact caused by the smoothing used to regularize the
459 inversion of the slip distribution, combined with the fact that the resolution at depth is limited.
460 If this overlapping of co-seismic and slow slip was confirmed, a simple interpretation of rate
461 and state friction cannot be proposed here. Overlaps between coseismic slip and afterslip have
462 been observed after several earthquakes. For example after 2004 M_w 6 Parkfield earthquake,
463 afterslip was inferred to overlap the coseismic slip (Freed, 2007). Johnson et al., (2006)
464 proposed frictional spatial heterogeneity to explain the afterslip distribution, and proposed (a-
465 b) values on the order of $10^{-4} - 10^{-3}$. After the 2015 Ilaapel M_w 8 megathrust earthquake,
466 Barnhart et al. (2016) also observed afterslip and coseismic slip overlapping, and advocated
467 that stress heterogeneities likely provide the primary control on the afterslip distribution. After
468 the 1978 M_w 7.3 Tabas-e-Golshan earthquake, Zhou et al., (2018) also observe such
469 overlapping and found $(a-b) \sim 3 \cdot 10^{-3}$. They proposed that the fault is creeping during the whole

470 interseismic period and that the earthquake propagates through the rate-strengthening region
471 although they also stated that a change in frictional properties is also possible due to shear
472 heating. Thomas et al., (2017), after the 2003 M_w 6.8 Chengkung earthquake estimated that
473 (a-b) varies with depth from 0.018 near the surface to less than 0.001 at depth larger than 19
474 km, to explain areas showing both seismic and aseismic behavior they advocated shear heating
475 processes. Using numerical modelling, Noda and Lapusta (2013) argued that coseismic slip
476 could propagate into velocity-strengthening regions of a fault, and that these regions can
477 appear locked or creeping during the interseismic period. They observed this behavior due to
478 rapid shear heating. For our case it is difficult to pinpoint the particular fault zone mechanism
479 here, but we can propose a switch to (a-b) positive value below 4-5 km depth, due to coseismic
480 shear heating processes (Thomas et al., 2017) requiring temperature changes of several
481 hundred degrees (e.g., Rice, 2006). This could be a potential explanation since Smeraglia et al.,
482 (2017) suggested that nanostructures from the Mt Vettoreto Fault rocks were generated by
483 coseismic shear heating and grain comminution (reduction of particle sizes).

484

485 5.2 Barrier of rupture propagation

486 Near Arquata del Tronto, at shallow depth (< 3-4 km) we observe that the aseismic slip is
487 localized on an area associated with limited coseismic slip, for both the Amatrice and Norcia
488 earthquakes. The observed aseismic slip seems located in an area where coseismic ruptures
489 have not propagated during the Norcia or Amatrice earthquakes. This slowly slipping zone is
490 located near a structural complexity (inherited thrust OAS), that Chiaraluce et al., (2017), Pizzi
491 et al., (2017) and Puliti et al. (2020) interpreted as a barrier that concentrated stress after the
492 Amatrice earthquake. We propose that maybe as a result of this geometric complexity (that
493 may concentrate high stress and high porosity resulting from fracturation), this zone might be
494 also characterized by frictional properties at shallow depth that could have favoured the arrest
495 of the ruptures (e.g., Kaneko et al., 2010). Scholz (1998) shows that when an earthquake
496 propagates into a velocity strengthening field it will produce a negative stress drop, that rapidly
497 terminates propagation. Our simple calculations yield here a slightly positive (a-b) value,
498 consistent with this model.

499 6 Conclusion

500 The Sentinel-1 InSAR time-series show cm-scale post-seismic displacements after the
501 Norcia earthquake (October 30). The displacements are more obvious in the ascending track
502 than in the descending track, and correspond to ~1 to 5 cm of subsidence in the Castelluccio
503 basin and at the southern tip of the Norcia surface rupture. In the area of the Castelluccio
504 basin, whether the deformation can be explained by afterslip only is less clear and would
505 require additional modelling to include the effects of poro-elastic release or shear zone
506 deformation. At the southern edge of the Norcia coseismic rupture, deformation evolves with
507 time following a logarithmic decay consistent with afterslip triggered by the Norcia
508 earthquake. Although good observations of the 2.6 first days of afterslip are not available, after
509 3 months the slow deformation evidenced in this study released a geodetic moment of Mw
510 ~5.78, which corresponds to ~8.7 % of the geodetic moment released during the Norcia
511 earthquake. Cumulated moment released by aftershocks in the southern tip of the Norcia
512 rupture accounts for ~2.0 % of the moment released by the modelled deformation, the
513 deformation is thus here mainly aseismic. This afterslip takes place at the southern tip of the
514 Norcia rupturing patch, and seems to have acted as a barrier to the propagation of Norcia and
515 Amatrice ruptures at shallow depth (<3-4 km). Our results suggest that the observed post-
516 Norcia earthquake afterslip might have been triggered in response to heterogeneities of pore
517 fluid pressure, potentially facilitated by structural complexity and intense faulting in the area.
518 Such small and localized slow deformation due to afterslip may be best detected with InSAR
519 time-series in regions of sparse GNSS coverage. It might actually be quite common and could
520 be an underestimated phenomenon. This likely points toward a bias in the literature in favour
521 of high and wider afterslip, that is easier to detect. However detection of small afterslip
522 transient is crucial to further understand the physics of earthquakes, and the link between
523 slow slip and seismic rupture (e.g., Kaneko et al., 2010; Rousset et al., 2017).

524

525 7 Tables

526

527 7.1 Table 1

528

	Depths (km)	Model 1 η	Model 2 η	Model 3 η	Model 4 η
Upper crust	0-10.7	Elastic	Elastic	Elastic	Elastic
Middle crust	10.7-21.7	Elastic	10^{19} Pa s	10^{18} Pa s	10^{18} Pa s
Lower crust	21.7-33.1	10^{18} Pa s	10^{18} Pa s	10^{18} Pa s	10^{17} Pa s
Upper mantle	33.1-below	10^{21} Pa s	10^{21} Pa s	10^{21} Pa s	10^{21} Pa s

529

530 Table 1 : Crust and mantle viscosities (η) tested in Figure 6.

531

532

533 7.2 Table 2

534

Fit	α	C (days ⁻¹)	γ	R2
Point 1 near Arquata del Tronto	30.9 ± 4.0	0.26 ± 0.26	-6.4 ± 6.2	0.87
Point 2 near Castelluccio di Norcia	156.4 ± 230.7	0.003 ± 0.007	2.54 ± 3.5	0.81
Pattern tracking	1.6 ± 0.1	0.24 ± 0.1	-0.37 ± 0.21	0.95

535

536 Table 2: Parameters obtained from the fit of the afterslip model in Figure 3 using a non-linear

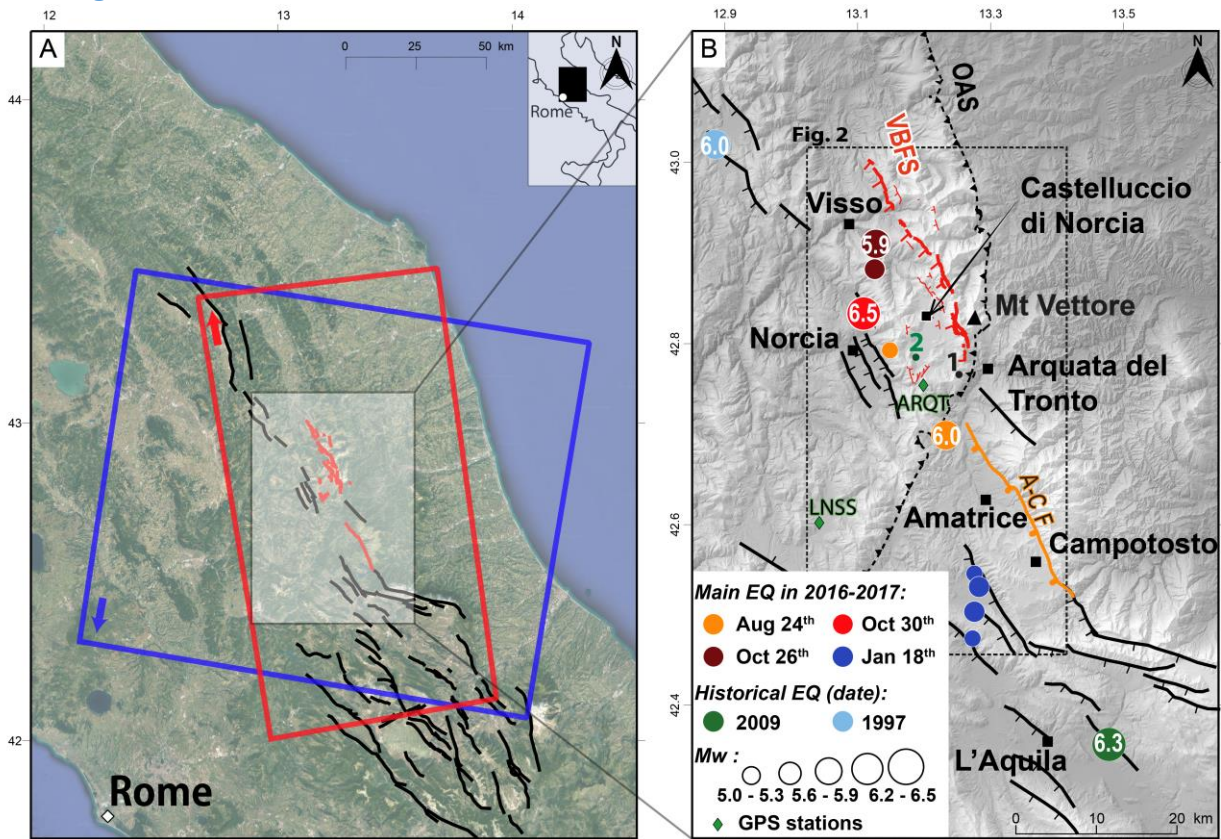
537 least squares method (Moré, 1978). R2 is the coefficient of determination of the regression.

538 The pattern encompasses the point 1 (see Figure 4)

539

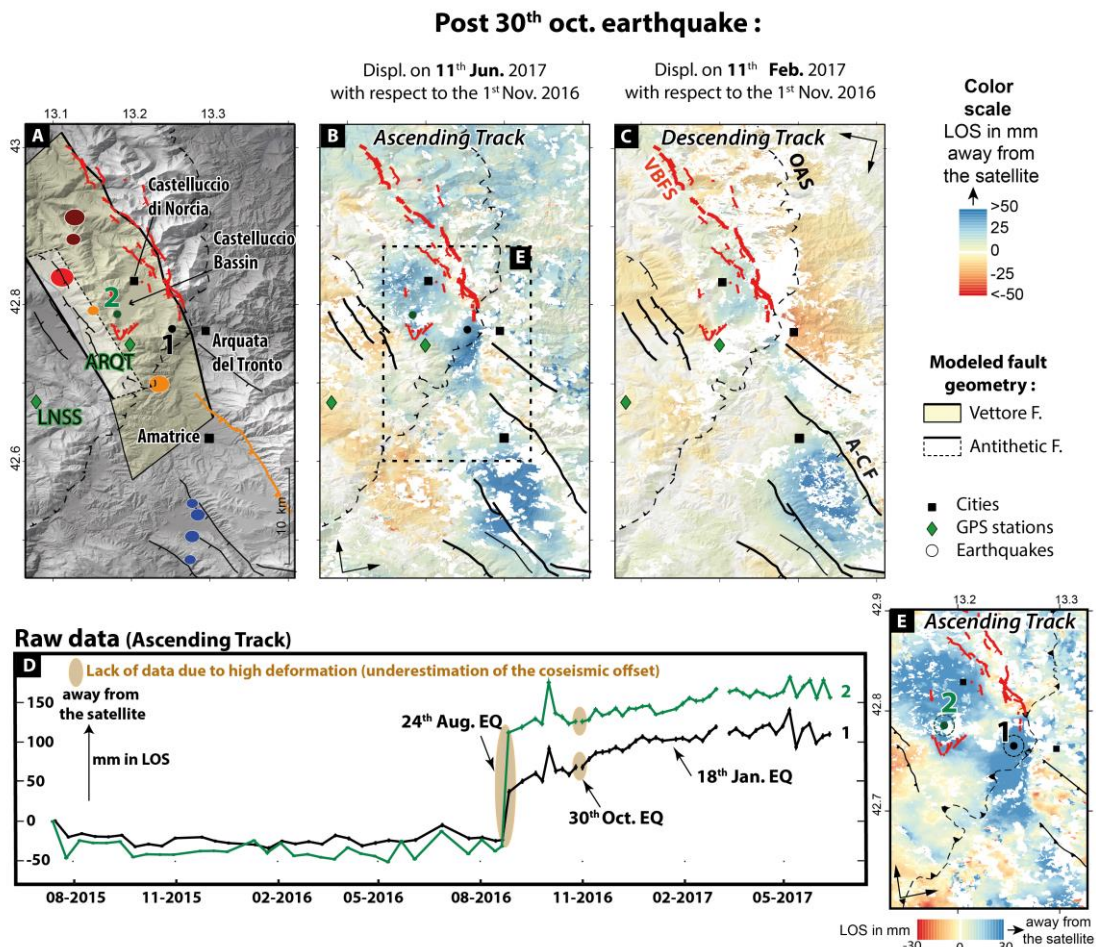
540 8 Figures

541 8.1 Figure 1



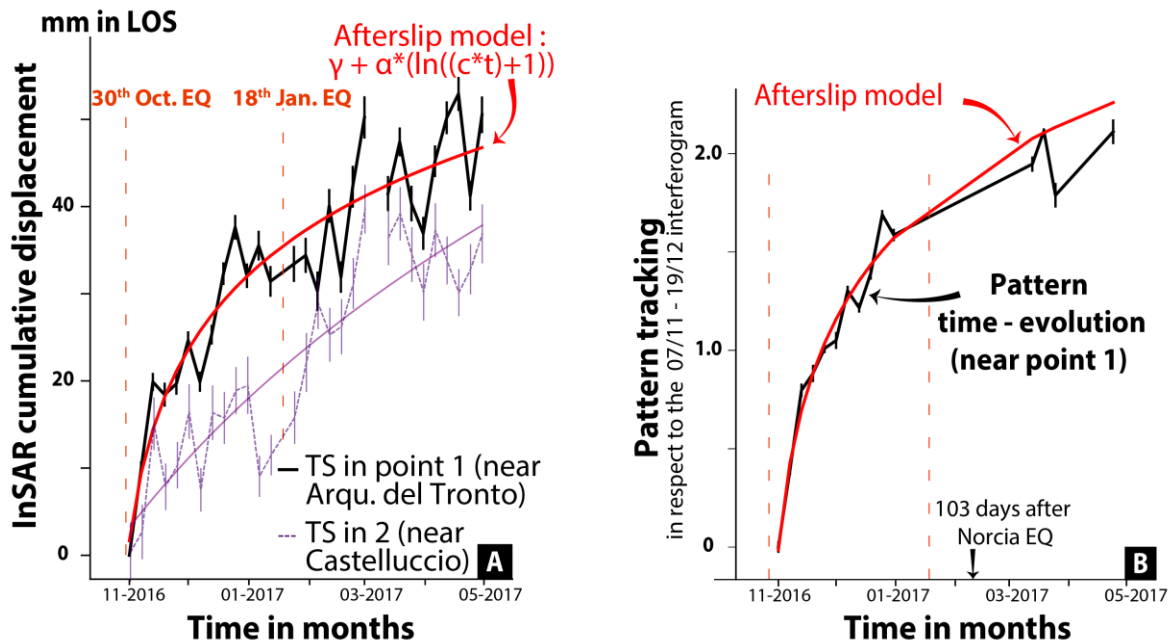
542
 543 Figure 1: Neotectonic framework of the Central Apennines. Black lines are main normal active
 544 faults compiled from Benedetti (1999), Tesson et al., (2016) and references therein (VBFS: Mt
 545 Vettore – Mt Bove Fault System, A-C F: Amatrice-Campotosto Fault). In red, coseismic surface
 546 ruptures observed in the field after the mainshocks of the Central Italy 2016-2017 seismic
 547 sequence (Villani, Civico, et al., 2018). (A) Blue and red rectangles are, respectively, the
 548 descending (D22 subswaths IW2 and IW3) and ascending (A117, subswath IW3) Sentinel tracks
 549 processed in this study. (B) Black dashed line represents the Neogene OAS thrust (Olevano-
 550 Antrodoco-Sibillini thrust) (Di Domenica et al., 2012). In orange the Amatrice-Campotosto fault
 551 which was activated during the 2016—2017 seismic sequence. ARQT and LNSS (in B) are GNSS
 552 stations available in this region from
 553 http://geodesy.unr.edu/NGLStationPages/gpsnetmap/GPSNetMap_MAG.html (Blewitt et al.,
 554 2018).

555



557
 558 Figure 2: (A) Map of the studied area, green squares are the GPS stations, points 1-2 are the
 559 points of sampling. The modeled geometries of the Mt Vettore and antithetic faults are plotted
 560 (see section 4.5). Circles are the earthquakes (see Figure 1). **(B-C)**: Time-series calculated
 561 cumulative post-Norcia (30 oct) displacements for ascending (B) and descending track (C),
 562 respectively. In panels A,B,C,E : Red lines are surface ruptures associated with 30th October
 563 2016 Norcia earthquake (Villani, Civico, et al., 2018), and black lines are active faults. Black
 564 dashed line is the OAS thrust. (VBFS: Mt Vettore – Mt Bove Fault System, A-C F: Amatrice-
 565 Campotosto Fault, OAS: Olevano-Anrodoco-Sibillini thrust). **(D)** Displacement through time
 566 averaged over a circle of 20 pixels (dashed circles in panel E) in diameter around points 1-2. **(E)**
 567 Cumulative displacement on June 11, 2017, with respect to the 1st November 2016 date.
 568

570



571

572

573 Figure 3: **(A)** Temporal variation of displacement averaged over a circle of 20 pixels in diameter
 574 through time of point 1 (black curve) and 2 (purple dashed curve) localized in Figure 2-E. We
 575 fitted the curve with afterslip model (see results in Table 2).

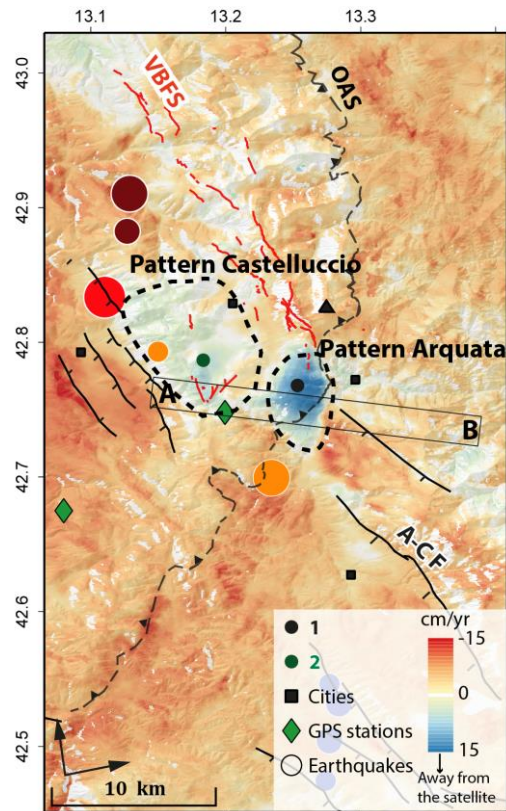
576 **(B)** Temporal variation of pattern amplitude. Pattern is indicated in Figure 4. In red, the fit with
 577 an afterslip model (see results in Table 2).

578

579

580

581 8.4 Figure 4

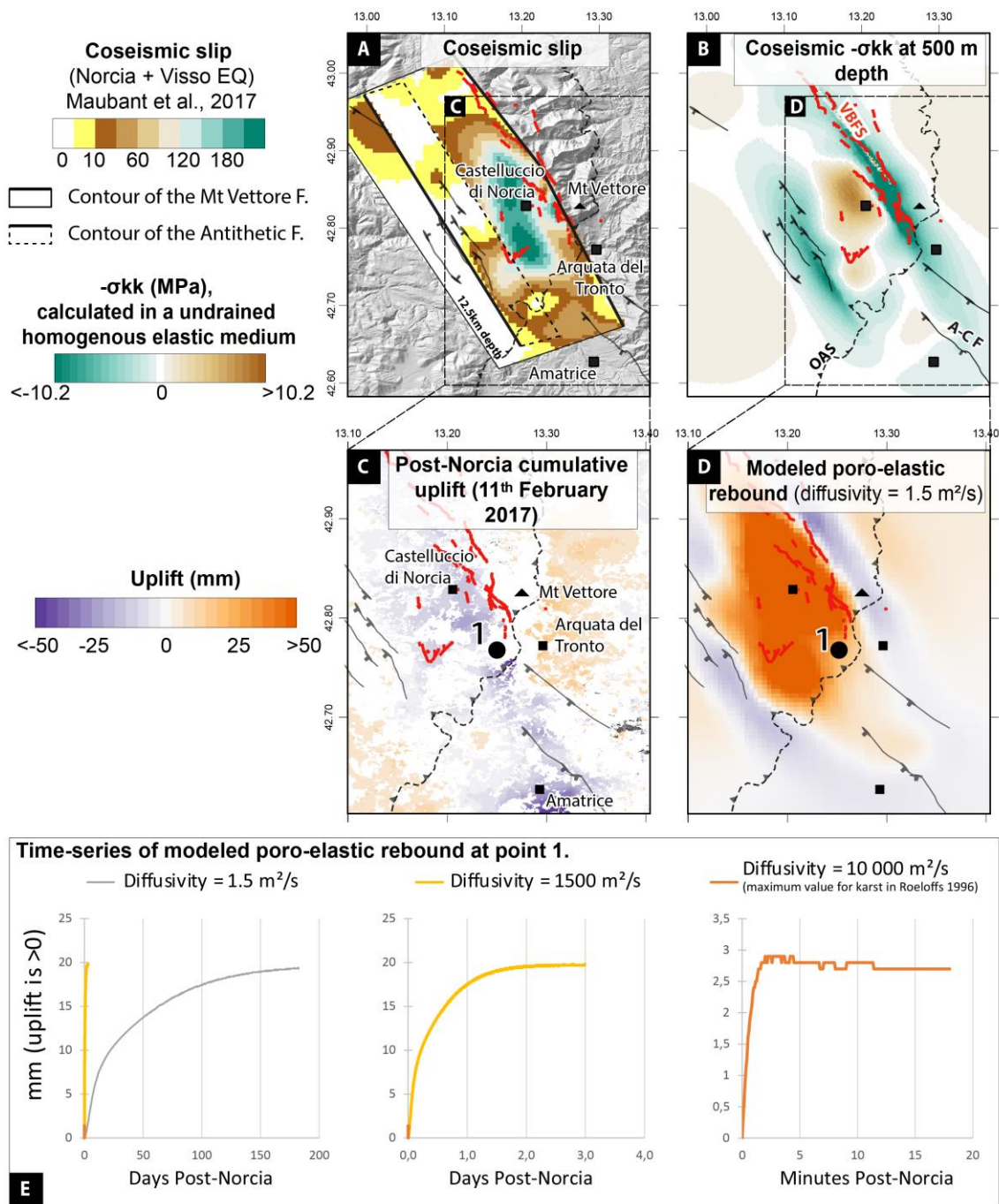


582

583 Figure 4: Mean velocity map between Mw 6.5 Norcia and Mw 5.7 Campotosto earthquakes
584 (November 1st – January 12th). This map is computed using 35 ascending interferograms. The
585 swath A-B is used in Figure S11 to compare InSAR displacements and topography. Circles
586 represent the earthquakes (see color code in Figure 1).

587

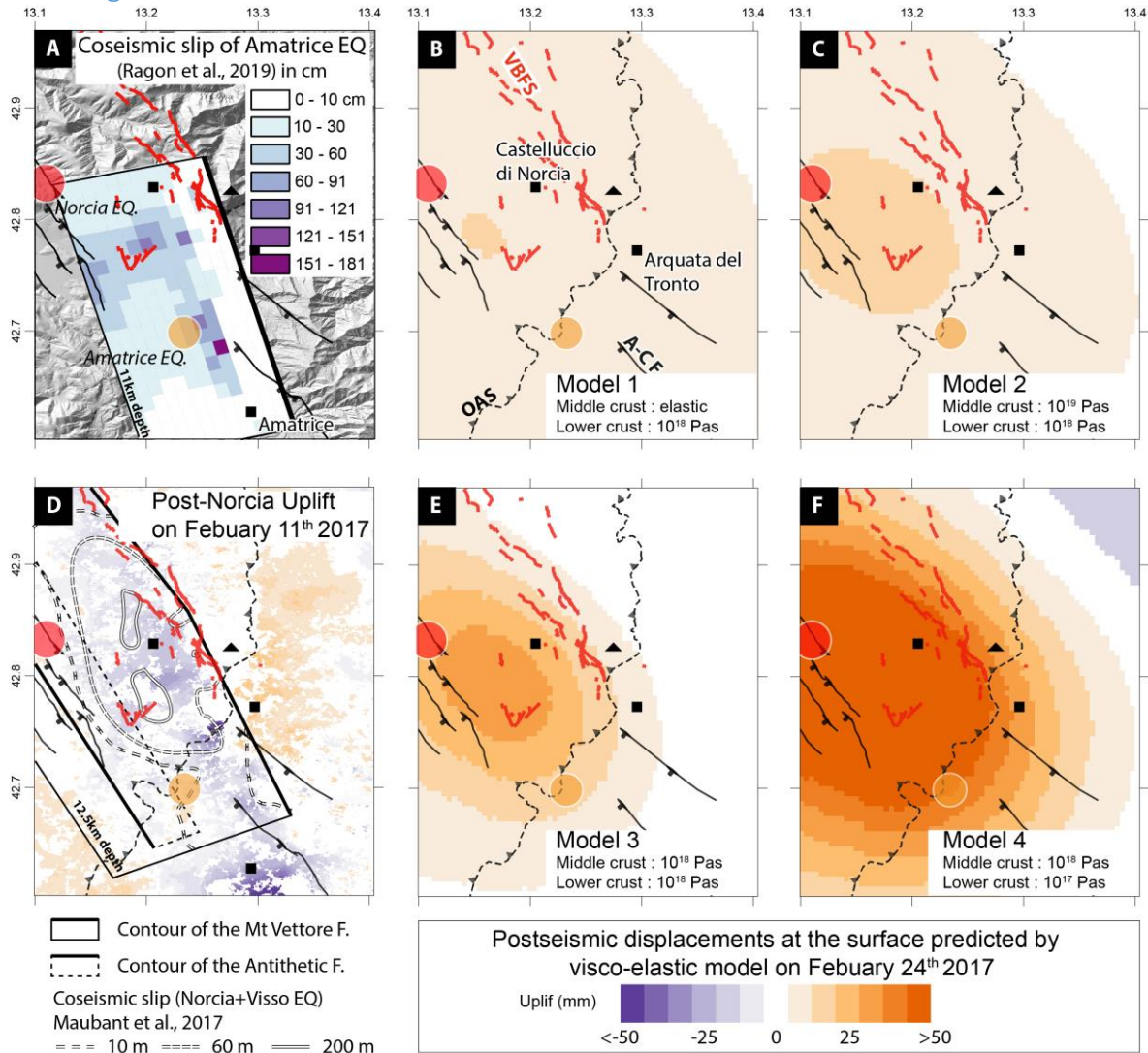
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590

591 Figure 5: Poro-elastic rebound modelled with Relax software (Barbot & Fialko, 2010). We used
 592 the coseismic slip models of the Norcia and Visso earthquakes from Maubant et al., (2017).
 593 **(A)** Mapview of the coseismic slip distribution. **(B)** Coseismic $-\sigma_{kk}$ modelled at 500m depth,
 594 calculated in an undrained homogenous elastic medium (Poisson $\nu=0.34$; Lamé $\lambda=6.36E+04$
 595 MPa, shear modulus $G=30$ GPa, Gravity wavelength $\gamma=5.39E-04$ km⁻¹). To remind, the pore
 596 pressure change (Δp) is equal to $-B \cdot \sigma_{kk} / 3$, with (B the Skempton's coefficient). **(C)** Post-Norcia
 597 cumulative displacement map (11th February 2017) observed by InSAR (uplift component). **(D)**

598 Model-predicted poro-elastic rebound as of 25th February 2017, obtained using the difference
599 between drained and undrained conditions. Poisson drained is 0.26 (cf. layer 1 in Albano et al.
600 2018), and for undrained conditions is 0.34 (from $\beta=0.4$ in the table 1 and equation 15 in
601 Barbot et al., (2010)). The layer between (0 and 5 km) is characterized by a diffusivity equals
602 to $1.5 \text{ m}^2/\text{s}$ (Tung & Masterlark, 2018). **(E)** Modelled poro-elastic rebound time-series at point
603 A using several diffusivities, from $1.5 \text{ m}^2/\text{s}$ (Tung & Masterlark, 2018) to $10^4 \text{ m}^2/\text{s}$ (Roeloffs,
604 1996). Because the delay time is inversely proportional to the assumed diffusivity, the
605 response for high diffusivity is shorter. (VBFS: Mt Vettore – Mt Bove Fault System, A-C F:
606 Amatrice-Campotosto Fault, OAS: Olevano-Antrodoco-Sibillini thrust).
607

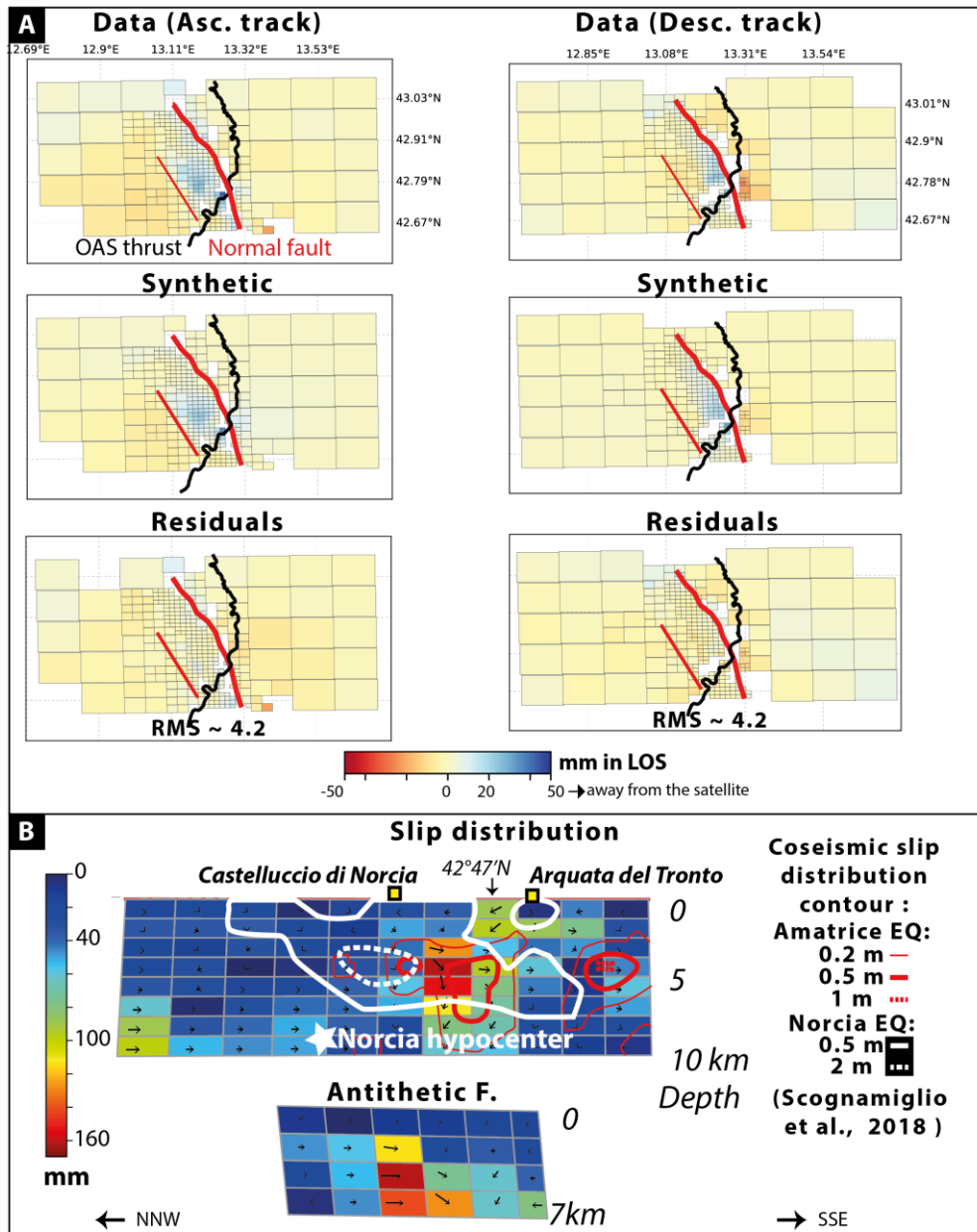


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610 Figure 6: Surface displacements predicted by visco-elastic models, we used the Amatrice,
 611 Norcia and Visso earthquakes coseismic slip distribution modelled by Ragon et al., (2019) and
 612 Maubant et al., (2017). **(A)** Coseismic slip distribution of the Amatrice earthquake from Ragon
 613 et al., (2019) **(B-C-E-F)** Surface displacements predicted by four visco-elastic models on 24th
 614 February 2017 using Relax software (Barbot & Fialko, 2010). Mantle viscosity is fixed for all
 615 models at 10^{21} Pas. **(D)** Post-Norcia earthquake (30th October) cumulative displacement map
 616 (11th February 2017) observed by InSAR (uplift component). (VBFS: Mt Vettore – Mt Bove Fault
 617 System, A-C F: Amatrice-Campotosto Fault, OAS: Olevano-Antrodoco-Sibillini thrust)

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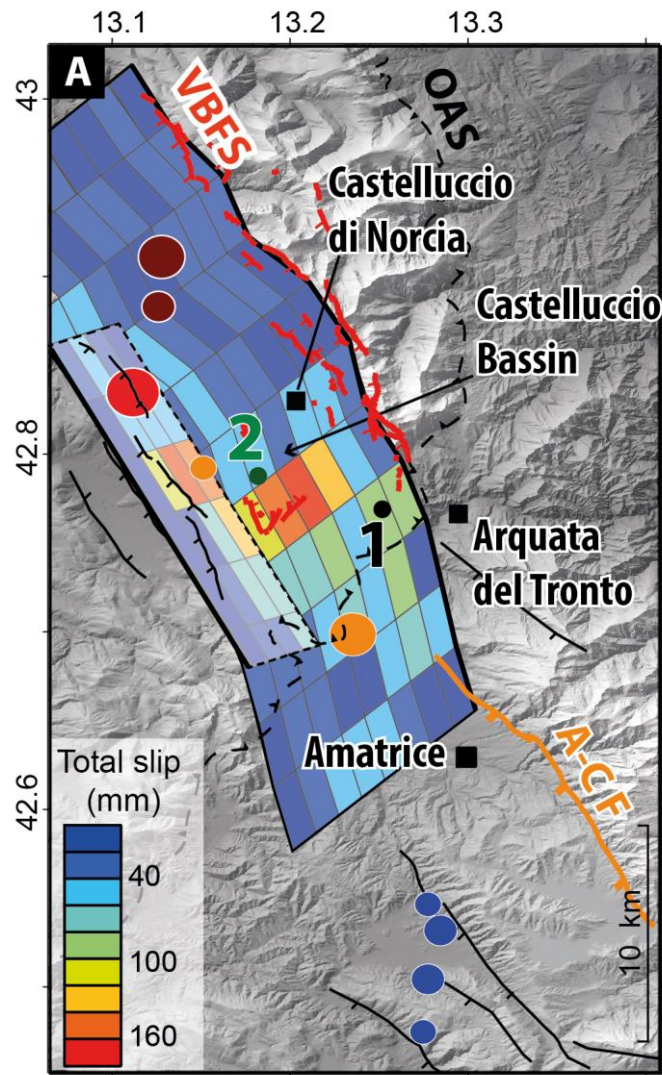


622
623 Figure 7: Result of afterslip inversion using the CSI software (see Supplementary text S3 for
624 details). **(A)** Subsampled displacement maps for the ascending (left column) and descending
625 (right column) tracks. Top row shows the input subsampled data, middle row shows the
626 synthetic displacements predicted by the model, and bottom row shows the residuals (i.e.
627 model - data). The modeled fault geometries used for the inversion are plotted in Figure 2-A.
628 The input maps are the smoothed cumulative displacement maps on February 11th (calculated
629 in the ascending and descending post-30th October time-series). **(B)** Afterslip distribution
630 inverted from the data on the main fault and the antithetic fault. Black arrows in patches show
631 the rake. Our model has low resolution at depth (Figure S16). Coseismic slip distributions for

632 the Amatrice and Norcia earthquakes (Scognamiglio et al., (2018) are superimposed in white
633 and red , respectively. Comparison with the Cheloni et al.,(2019) model in Figure S21 shows
634 similar patterns.

635

636



638
 639 Figure 8: Afterslip distribution model plotted on mapview. The slip distribution is inverted from
 640 the data on the main fault and the antithetic fault (see Figure 7). (VBFS: Mt Vettore – Mt Bove
 641 Fault System, A-C F: Amatrice-Campotosto Fault). Circles are the main earthquakes of the
 642 seismic sequence.
 643

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666 distribution model will be provided in [10.5281/zenodo.3755127](https://doi.org/10.5281/zenodo.3755127) . Seismicity used for this research is
667 included in (Chiaraluce et al., 2017) and in <http://cnt.rm.ingv.it/>. The GNSS stations are downloaded
668 from <http://geodesy.unr.edu/NGLStationPages/stations/>

669

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671

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