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# Joint Inversion of Coseismic and Early Postseismic Slip to Optimize the Information Content in Geodetic Data: Application to the 2009 M w 6.3 L'Aquila Earthquake, Central Italy

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1 **Joint inversion of co-seismic and early post-seismic slip**  
2 **to optimize the information content in geodetic data:**  
3 **Application to the 2009  $M_w$ 6.3 L'Aquila earthquake,**  
4 **Central Italy**

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9 **Key Points:**

- 10 • Simultaneous inversion of geodetic data with different spatio-temporal resolution  
11 maximizes the information content
- 12 • If the early afterslip signal is neglected, coseismic slip is overestimated by 30% and  
13 postseismic slip is underestimated by a factor of 3
- 14 • Early afterslip (6 days after mainshock) is preferentially located at the transition  
15 zone between coseismic rupture and longer-term afterslip

16 **Reference:** Ragon, T., Sladen, A., Bletery, Q., Vergnolle, M., Cavali, O., Aval-  
17 lone, A., et al. (2019). Joint inversion of coseismic and early postseismic slip to optimize  
18 the information content in geodetic data: Application to the 2009  $M_w$ 6.3 L'Aquila earth-  
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## 21 **Abstract**

22 When analyzing the rupture of a large earthquake, geodetic data are often critical. These  
23 data are generally characterized by either a good temporal or a good spatial resolution,  
24 but rarely both. As a consequence, many studies analyze the co-seismic rupture with data  
25 that also include one or more days of early post-seismic deformation. Here, we invert si-  
26 multaneously for the co- and post-seismic slip with the condition that the sum of the two  
27 models remains compatible with data covering the two slip episodes. We validate the ben-  
28 efits of this approach with a toy model and an application to the 2009  $M_w$ 6.3 L'Aquila  
29 earthquake, using a Bayesian approach and accounting for epistemic uncertainties. For  
30 the L'Aquila earthquake, we find that if early post-seismic deformation is not explicitly  
31 acknowledged co-seismic signal, co-seismic slip models may overestimate the peak am-  
32 plitude while long-term post-seismic models may largely underestimate the total post-  
33 seismic slip amplitude. This example illustrates how the proposed approach could im-  
34 prove our comprehension of the seismic cycle, of fault frictional properties, and the spa-  
35 tial and temporal relationship between seismic rupture, afterslip and aftershocks.

## 36 **1 Introduction**

37 The occurrence of earthquakes and seismic sequences is mainly controlled by the  
38 spatial and temporal evolution of crustal stresses. The co-seismic stress changes and the  
39 redistribution of stress following an earthquake thus both play an important role in the  
40 seismic cycle and the mechanical behavior of faults, including the generation of new seis-  
41 mic sequences. To understand both co-seismic and post-seismic processes, and their re-  
42 lationship, is thus a crucial step to propose realistic earthquakes scenarios and reliable  
43 hazard estimates.

44 While earthquakes can last for a few seconds to minutes, the subsequent period of  
45 post-seismic stress relaxation can last months to years. Post-seismic relaxation is gen-  
46 erally modeled by several interacting mechanisms, such as localized shear on the fault  
47 (a.k.a. afterslip) [e.g. *Marone et al.*, 1991; *Freed*, 2007; *Johnson et al.*, 2012], visco-elastic  
48 deformation in the lower crust or mantle [e.g. *Nur and Mavko*, 1974; *Pollitz et al.*, 1998;  
49 *Freed and Burgmann*, 2004] or poroelastic rebound [e.g. *Peltzer et al.*, 1998; *Jonsson et al.*,  
50 2003]. The interactions between co-seismic stress changes, aftershocks and post-seismic  
51 deformation are still poorly understood [e.g. *Perfettini and Avouac*, 2007]. Slip on the  
52 fault may be governed by two brittle deformation modes following rate and state fric-

53 tion laws [*Rice and L. Ruina*, 1983]: seismic rupture may occur in velocity weakening  
54 areas, whereas afterslip may develop in velocity strengthening zones [e.g. *Marone et al.*,  
55 1991]. In contrast, *Helmstetter and Shaw* [2009] also show that afterslip processes may  
56 be primarily driven by stress heterogeneities, independently of the rate and state fric-  
57 tion behavior. Aftershocks may be triggered by co-seismic stress changes, without di-  
58 rect relation to post-seismic deformation [*Dieterich*, 1994]. Or, aftershocks may also be  
59 primarily triggered by the post-seismic reloading due to afterslip [e.g. *Perfettini and Avouac*,  
60 2004; *Hsu et al.*, 2006; *Peng and Zhao*, 2009; *Ross et al.*, 2017]. The variability of these  
61 theories emphasizes the need to refine our comprehension and description of the co-seismic  
62 and post-seismic phases and their transition.

63 Our understanding of the co-seismic processes mainly derives from modeling of seis-  
64 mic, geodetic and tsunami data, and our understanding of post-seismic behavior is mainly  
65 based on the modeling of geodetic observations [e.g. *Burgmann et al.*, 1997; *Wang et al.*,  
66 2012; *Perfettini and Avouac*, 2014; *Gualandi et al.*, 2017] or simulation [e.g. *Smith and*  
67 *Sandwell*, 2004; *Barbot and Fialko*, 2010; *Cubas et al.*, 2015]. Geodetic observations thus  
68 remain a cornerstone to identify and characterize the co- and post-seismic processes. GNSS  
69 time series are commonly used and can provide a good temporal resolution (seconds to  
70 days). But the spatial resolution of such observations is usually limited, as GNSS sta-  
71 tions are usually several tens or hundreds of kilometers apart. In contrast, synthetic aper-  
72 ture radar interferometry (InSAR) can provide extensive spatial coverage (in the order  
73 of a meter) but with a limited temporal resolution. Indeed, while earthquakes last for  
74 a few seconds, very often satellites have a revisit time of more than a few days. If earth-  
75 quakes do not nucleate just before the visit of a satellite, which is generally the case, the  
76 measured deformation is the co-seismic signal plus a fraction of the post-seismic defor-  
77 mation. As a consequence, most earthquake models based on geodetic observations are  
78 biased by an unwanted post-seismic deformation signal. In practice, both co-seismic in-  
79 terferograms or campaign GNSS offsets generally cover time periods extending at least  
80 a few days before and after the mainshock. Pre-earthquake signals, when evidenced, are  
81 usually related to small slip episodes at depth near the hypocenter [e.g. *Kato et al.*, 2012;  
82 *Ruiz et al.*, 2014]. The associated surface deformation signals are usually hard to detect  
83 and neglected in co-seismic studies. The post-seismic deformation happening in the first  
84 few days after the mainshock is usually detectable in the geodetic data but incorporated  
85 in source estimation problems as if it was part of the co-seismic signal [e.g. *Elliott et al.*,

2013; *Lin et al.*, 2013; *Cheloni et al.*, 2014; *Bletery et al.*, 2016; *He et al.*, 2017; *Salman et al.*, 2017; *Barnhart et al.*, 2018], with the justification that it is comparatively small. Similarly, post-seismic models generally do not account for observations related to the early post-seismic deformation because they are often contaminated by co-seismic signal [e.g. *D'Agostino et al.*, 2012; *Cheloni et al.*, 2014]. What we name here the early post-seismic phase corresponds to the overlooked part of the post-seismic deformation, and can last from a few hours after the mainshock in the best case, to a few days in most studies. Yet, the largest post-seismic deformation rate is expected during the first few days after the mainshock, considering that its main trend is to decrease exponentially with time after an earthquake.

This early post-seismic phase remain largely unexplored, because of the limited temporal and spatial resolution of geodetic data [e.g., *Twardzik et al.*, 2019]. Neglecting this phase may also bias our understanding of both co-seismic and post-seismic processes. And this bias is probably persisting if seismic data (i.e. purely co-seismic) are added to the inverse problem, since geodetic data tend to have a stronger control on the inferred distribution of slip, at least in the first 10 km below the Earth surface [e.g. *Delouis et al.*, 2002].

The recent advent of high-frequency GNSS has allowed the recording of the strictly co-seismic signal (10 to 30 seconds after the earthquake time occurrence) without any contamination by early post-seismic deformation [e.g., *Twardzik et al.*, 2019]. Well instrumented earthquakes are thus now characterized by at least two geodetic datasets, one being strictly co-seismic and the other which also includes some days of early after-slip. In this study, we propose a generic inversion methodology to jointly infer co-seismic and early post-seismic slip models, taking advantage of the complementary spatial and temporal resolutions of different geodetic observations (typically InSAR and image correlation data). A comparable approach has been employed by *Floyd et al.* [2016] and *Walters et al.* [2018]. To further explore the potential of the approach, we first explicitly detail the governing equations of the implementation and then evaluate the approach through a toy model. We also analyze and illustrate the benefits of the methodology with a real event, our analyses being performed in a Bayesian framework. In detail, we investigate the impact of accounting for early afterslip on co-seismic models. We consider the 2009  $M_w$ 6.3 L'Aquila earthquake, Central Italy, which has been intensively studied but whose very early post-seismic phase has not been imaged. The choice of the L'Aquila event is

119 also motivated by the large density of near field observations and the overall quality of  
 120 the instrumentation. Additionally, this event ruptured a relatively well known and sim-  
 121 ple fault geometry, in an area where crustal properties have been investigated in detail:  
 122 this will ensure the forward physics and its uncertainties can be estimated.

## 123 2 Inversion Framework

### 124 2.1 Dual time inversion of co-seismic and early post-seismic data

125 For a number of earthquakes, we have the opportunity to use two static datasets:  
 126 one which is strictly co-seismic ("co") and the other which contains co-seismic and early  
 127 post-seismic signal ("co+post"). To infer the strictly co-seismic and early post-seismic  
 128 slip distributions, one approach could be to invert separately for the two datasets, and  
 129 assume that the strictly post-seismic ("post") solution is the difference between the "co"  
 130 and "co+post" models. However, in this case, the model "co" would be constrained by  
 131 fewer observations (only few GNSS offsets), most of the co-seismic information being in  
 132 the "co+post" dataset (dense map of InSAR offsets). An alternative approach is to as-  
 133 sume that the "co+post" slip model is the sum of the "co" and "post" slip distributions  
 134 [e.g., *Floyd et al.*, 2016; *Walters et al.*, 2018]. We then have:

$$\begin{cases} \mathbf{d}_{\text{co}} = \mathbf{G}_{\text{co}}^{\text{co}} \cdot \mathbf{m}_{\text{co}} \\ \mathbf{d}_{\text{co+post}} = \mathbf{G}_{\text{co+post}}^{\text{co}} \cdot \mathbf{m}_{\text{co}} + \mathbf{G}_{\text{co+post}}^{\text{post}} \cdot \mathbf{m}_{\text{post}} \end{cases} \quad (1)$$

where matrices of the Green's functions  $\mathbf{G}_{\text{data}}^{\text{model}}$  have been calculated for the correspond-  
 ing dataset and model. For instance,  $\mathbf{G}_{\text{co+post}}^{\text{co}}$  is the matrix of the Green's functions cal-  
 culated from the model "co" for the data "co+post". The Eq. 1 can also be represented  
 in the following matrix form:

$$\begin{pmatrix} \mathbf{d}_{\text{co}} \\ \mathbf{d}_{\text{co+post}} \end{pmatrix} = \begin{pmatrix} \mathbf{G}_{\text{co}}^{\text{co}} & 0 \\ \mathbf{G}_{\text{co+post}}^{\text{co}} & \mathbf{G}_{\text{co+post}}^{\text{post}} \end{pmatrix} \cdot \begin{pmatrix} \mathbf{m}_{\text{co}} \\ \mathbf{m}_{\text{post}} \end{pmatrix}. \quad (2)$$

The redesigned Green's functions matrix is now composed of 3 sub-matrices. As we fo-  
 cus on the early post-seismic phase, we can make the assumption that both  $\mathbf{G}_{\text{co+post}}^{\text{co}}$  and  
 $\mathbf{G}_{\text{co+post}}^{\text{post}}$  matrices are identical because we suppose both co-seismic and early post-seismic

deformations are elastic. We can thus write

$$\begin{pmatrix} \mathbf{d}_{\text{co}} \\ \mathbf{d}_{\text{co+post}} \end{pmatrix} = \begin{pmatrix} \mathbf{G}_{\text{co}} & 0 \\ \mathbf{G}_{\text{co+post}} & \mathbf{G}_{\text{co+post}} \end{pmatrix} \cdot \begin{pmatrix} \mathbf{m}_{\text{co}} \\ \mathbf{m}_{\text{post}} \end{pmatrix}. \quad (3)$$

If strictly post-seismic observations are available, we could also incorporate these data into our equation to help constrain the "post" model:

$$\begin{pmatrix} \mathbf{d}_{\text{co}} \\ \mathbf{d}_{\text{co+post}} \\ \mathbf{d}_{\text{post}} \end{pmatrix} = \begin{pmatrix} \mathbf{G}_{\text{co}}^{\text{co}} & 0 \\ \mathbf{G}_{\text{co+post}}^{\text{co}} & \mathbf{G}_{\text{co+post}}^{\text{post}} \\ 0 & \mathbf{G}_{\text{post}}^{\text{post}} \end{pmatrix} \cdot \begin{pmatrix} \mathbf{m}_{\text{co}} \\ \mathbf{m}_{\text{post}} \end{pmatrix}, \quad (4)$$

135 with  $\mathbf{G}_{\text{post}}^{\text{post}}$  reflecting the response of the Earth for the strictly post-seismic data. The  
 136 "post" dataset then corresponds to the same post-seismic time window as that covered  
 137 by the "co+post" dataset.

The off-diagonal terms of the redesigned Green's function matrix allow us to make use of the "co+post" dataset to constrain both "co" and "post" models. In the following, we refer to this approach as Combined Time Windows (CTW) approach. The CTW approach can be generalized to cover various intervals of post-seismic deformation. Indeed, while for many earthquakes strictly co-seismic data are now available, non-strictly co-seismic datasets usually cover variable time intervals. If, for instance, two intervals of post-seismic deformation contaminate the co-seismic signal, with only one of these intervals observed independently, our equation 3 can be adapted as

$$\begin{pmatrix} \mathbf{d}_{\text{co}} \\ \mathbf{d}_{\text{co+post1}} \\ \mathbf{d}_{\text{co+post2}} \\ \mathbf{d}_{\text{post2}} \end{pmatrix} = \begin{pmatrix} \mathbf{G}_{\text{co}}^{\text{co}} & 0 & 0 \\ \mathbf{G}_{\text{co+post1}}^{\text{co}} & \mathbf{G}_{\text{co+post1}}^{\text{post1}} & 0 \\ \mathbf{G}_{\text{co+post2}}^{\text{co}} & \mathbf{G}_{\text{co+post2}}^{\text{post1}} & \mathbf{G}_{\text{co+post2}}^{\text{post2}} \\ 0 & 0 & \mathbf{G}_{\text{post2}}^{\text{post2}} \end{pmatrix} \cdot \begin{pmatrix} \mathbf{m}_{\text{co}} \\ \mathbf{m}_{\text{post1}} \\ \mathbf{m}_{\text{post2}} \end{pmatrix}, \quad (5)$$

138 with  $\mathbf{d}_{\text{post2}}$  reflecting the surface displacement for the time interval between times 1 and  
 139 2, and  $\mathbf{G}_{\text{post2}}^{\text{post2}}$  and  $\mathbf{m}_{\text{post2}}$  associated Green's functions and slip model. Indeed, this ap-  
 140 proach could be used to investigate as many time windows of post-seismic deformation  
 141 as needed.

142 To refine co-seismic models and investigate early post-seismic deformation of the  
 143 L'Aquila earthquake, we follow here the approach described by Eqs 1 and 3. We do not  
 144 incorporate any information on the strictly post-seismic phase to investigate the very sim-

145 ple case where only co-seismic data (contaminated or not by early post-seismic defor-  
 146 mation) are available.

## 147 **2.2 Accounting for Epistemic Uncertainties**

148 When imaging a slip distribution on a fault, the physics of the forward model is  
 149 usually assumed of minimum complexity to simplify the computation and also often be-  
 150 cause we don't know well the Earth's interior rheological structure. For instance, this  
 151 structure is frequently approximated as a homogenous elastic half-space and the causative  
 152 fault geometry is usually reduced to a flat rectangular plane. The uncertainties related  
 153 to our approximations of the physics of the Earth affect the inferred source models [*Ragon*  
 154 *et al.*, 2018]. As the early post-seismic slip is of limited amplitude, it may be particu-  
 155 larly impacted by uncertainties of the forward model. We thus account for epistemic un-  
 156 certainties following the approach developed by *Duputel et al.* [2014] for the Earth elas-  
 157 tic properties and *Ragon et al.* [2018] for the fault geometry. The epistemic uncertain-  
 158 ties are calculated from the sensitivity of the Green's Functions and are included in a  
 159 covariance matrix  $\mathbf{C}_p$ .

## 160 **2.3 Bayesian approach**

161 Our inverse problem solves for both co-seismic and early post-seismic slip param-  
 162 eters, the latter being of limited amplitude. While the co-seismic parameters will be rea-  
 163 sonably well constrained, multiple and probably different early post-seismic models will  
 164 probably be equally realistic, as they will equally explain the observations. To get a ro-  
 165 bust image of the early post-seismic phase, we thus solve our problem with a Bayesian  
 166 sampling approach which relies on the AlTar package (<https://github.com/AlTarFramework/altar>),  
 167 which is a rewrite of the code CATMIP [*Minson et al.*, 2013]. AlTar combines the Metropo-  
 168 lis algorithm with a tempering process to realize an iterative sampling of the solution  
 169 space of the source models. A large number of samples are tested in parallel at each tran-  
 170 sitional step. Additionally, a resampling is performed at the end of each step to replace  
 171 less probable models. The probability of each sample to be selected depends on its abil-  
 172 ity to fit the observations  $\mathbf{d}_{\text{obs}}$  within the uncertainties  $\mathbf{C}_\chi = \mathbf{C}_d + \mathbf{C}_p$ , with  $\mathbf{C}_d$  the  
 173 observational errors and  $\mathbf{C}_p$  the epistemic uncertainties.



174 The ability of each model parameter to solve the source problem is evaluated through  
 175 repeated updates of the Probability Density Functions (PDFs)

$$f(\mathbf{m}, \beta_i) \propto p(\mathbf{m}) \cdot \exp[-\beta_i \cdot \chi(\mathbf{m})], \quad (6)$$

where  $\mathbf{m}$  is the current sample,  $p(\mathbf{m})$  is the prior information on this sample,  $i$  corresponds to each iteration and  $\beta$  evolves dynamically from 0 to 1 to ensure an exhaustive exploration of the solution space [Minson *et al.*, 2013].  $\chi(\mathbf{m})$  is the misfit function:

$$\chi(\mathbf{m}) = \frac{1}{2}[\mathbf{d}_{\text{obs}} - \mathbf{G} \cdot \mathbf{m}]^T \cdot \mathbf{C}_\chi^{-1} \cdot [\mathbf{d}_{\text{obs}} - \mathbf{G} \cdot \mathbf{m}]. \quad (7)$$

176 The use of AlTar with the CTW approach allows us to specify prior information on each  
 177 model, and thus to ensure the quasi-positivity of both co-seismic and post-seismic slip  
 178 models (or of any time window model).

179 The impact of different approaches to regularization, along with the ad-hoc choice  
 180 of smoothing parameters, may bias or induce unwanted artefacts in inferred slip mod-  
 181 els [e.g., Du *et al.*, 1992; Beresnev, 2003; Aster *et al.*, 2005; Causse *et al.*, 2010; Gallovic  
 182 *et al.*, 2015; Gombert *et al.*, 2017]. Because our Bayesian approach aims at sampling all  
 183 possible families of models compatible with the observations, it doesn't include any ad-  
 184 hoc regularization parameter [e.g., Minson *et al.*, 2013].

### 185 **3 Application to a simplified 2D model**

186 To ensure that our methodology allows us to reliably infer the slip distribution of  
 187 different time windows, we first analyze a synthetic 2D case where the slip is imaged ei-  
 188 ther independently or with the CTW approach. For this case, we assume two time win-  
 189 dows named co-seismic and post-seismic for simplicity.

#### 190 **3.1 Forward Model**

191 We assume a fault extending infinitely along strike and which is 20 km wide along  
 192 dip. The fault is discretized along dip into sub-faults of 1 km width and is dipping  $55^\circ$ .  
 193 We assume the co-seismic slip on this fault to be purely dip-slip and to vary gradually  
 194 with depth between 0 m and 1.5 m, positive values indicating normal slip, with maxi-  
 195 mum slip between 9 and 14 km depth. We also assume that there is post-seismic slip on  
 196 the same fault, with a similar location and direction and an amplitude equal to a tenth

197 of the co-seismic slip amplitude. We compute the corresponding "co" and "co+post" syn-  
 198 thetic observations using the expressions of surface displacement in an homogeneous elas-  
 199 tic half-space [Segall, 2010]. These synthetic observations are computed for a profile line  
 200 of 100 points, centred on the fault at the surface and spaced every kilometer. The toy  
 201 model and synthetic data setup are illustrated in Figure S1. For each data point, the syn-  
 202 thetic observations describe the surface displacement along the profile line and the ver-  
 203 tical (the along strike displacement being null). A correlated Gaussian noise of 5 mm  
 204 (i.e., noise with a Gaussian amplitude distribution with 5 mm of standard deviation, char-  
 205 acterized by an exponential covariance) is added to the synthetic data to simulate mea-  
 206 surement errors. Note that, for this toy model, the number of "co" data is the same as  
 207 the number of "co+post" observations.

208 Using these 100 synthetic observation points, we then estimate the depth distri-  
 209 bution of slip still assuming a homogeneous elastic half-space. We use a uniform prior  
 210 distribution  $p(\mathbf{m}) = \mathcal{U}(-1 \text{ m}, 5 \text{ m})$  for the dip slip component (uniform implies that all  
 211 values are considered equally likely with no a priori knowledge), a zero-mean Gaussian  
 212 prior  $p(\mathbf{m}) = \mathcal{N}(-0.1 \text{ m}, 0.1 \text{ m})$  on the strike-slip component and include 5 mm of ob-  
 213 servational uncertainty in  $\mathbf{C}_d$ . We do not account for epistemic uncertainties as our for-  
 214 ward model is identical to the one used to generate the data. We first solve for the "co"  
 215 and "post" slip following the CTW approach (Figures 1a and 1c). Then, we run inde-  
 216 pendent inversions, one to solve for the "co" slip (Figure 1b) and the other one to in-  
 217 fer the "co+post" slip. We calculate the post-seismic solution as the difference between  
 218 the "co" and "co+post" models (Figure 1d).

## 219 3.2 Results

220 Both independent and CTW inversion approaches allow to correctly infer the "co"  
 221 slip, as the median of the PDFs is very close to the target model (Figures 1a and 1b).  
 222 As expected from the inversion of surface data, the resolution is very good on shallow  
 223 parts of the fault but quickly decreases with depth. The posterior uncertainty on the deep-  
 224 est parameters is slightly decreased in both approaches because the lower tip of the fault  
 225 acts as an additional constraint. In contrast, the inversion methodology has a larger im-  
 226 pact on the inferred "post" slip distributions. When jointly inverting "co" and "co+post"  
 227 observations, the true "post" slip is estimated accurately for almost all subfaults (Fig-  
 228 ure 1c). When solving the two slip stages separately, the mean of the models is not as

229 good at estimating the target model (Figure 1d). The reduced posterior uncertainty of  
 230 the "post" model for the independent inversion is an artifact resulting from the subtrac-  
 231 tion of two gaussian-shaped curves.

232 In summary, both of the inversion approaches allow to reliably infer the "co" slip  
 233 distributions, probably because its signal is dominating in the observations. But the CTW  
 234 approach provides a more robust estimation of the "post" slip distribution. In this 2D  
 235 case, co-seismic and co+post signals have been observed by the same number of stations.  
 236 However, for most earthquakes, the number of "co" data points available (usually GNSS)  
 237 will be very limited compared to the quantity of "co+post" observations (usually InSAR).  
 238 We thus expect that if performing independent inversions for a real event, the inferred  
 239 "co" slip distribution will be less reliable than in the case of a CTW inversion, where the  
 240 whole "co+post" dataset is used to guide the choice of co-seismic parameters. We now  
 241 compare these two approaches on a real earthquake.

#### 242 **4 Application to the 2009 $M_w$ 6.3 L'Aquila earthquake, Central Italy**

243 The L'Aquila earthquake nucleated within the Apennines orogenic system (Figure 2),  
 244 where the current seismic activity results from the ongoing extensional tectonics of the  
 245 area. The mainshock nucleated on the Paganica fault [Figure 2, *Atzori et al.*, 2009; *Fal-*  
 246 *cucci et al.*, 2009; *Chiaraluce et al.*, 2011; *Vittori et al.*, 2011; *Lavecchia et al.*, 2012; *Ch-*  
 247 *eloni et al.*, 2014], southwest of the city of L'Aquila, and has been followed by at least  
 248 4 aftershocks of  $M_w > 5$  [*Scognamiglio et al.*, 2009; *Chiarabba et al.*, 2009; *Pondrelli et al.*,  
 249 2010]. Although the L'Aquila earthquake has been intensively studied, most co- and post-  
 250 seismic models have considered the first days of post-seismic deformation as if they were  
 251 part of the co-seismic phase [e.g. *Anzidei et al.*, 2009; *Atzori et al.*, 2009; *Cheloni et al.*,  
 252 2010; *Trasatti et al.*, 2011; *Cirella et al.*, 2012; *D'Agostino et al.*, 2012; *Cheloni et al.*,  
 253 2014; *Balestra and Delouis*, 2015; *Volpe et al.*, 2015]. To avoid the contamination of co-  
 254 seismic signal by early afterslip, *Yano et al.* [2014] proposed to explore independently  
 255 the early post-seismic deformation, yet with datasets covering different time intervals (1  
 256 day after the mainshock for GNSS, 6 days for InSAR). Significant post-seismic displace-  
 257 ment is recorded up to several months after the mainshock [e.g. *D'Agostino et al.*, 2012;  
 258 *Gualandi et al.*, 2014; *Cheloni et al.*, 2014; *Albano et al.*, 2015], yet most studies of long-  
 259 term post-seismic signal did not analyze the first few days of post-seismic deformation.

#### 260 4.1 Data, Forward Model and Prior Information

261 From a geodetic perspective, this event has been particularly well documented. We  
 262 can distinguish two main static datasets: one which is strictly co-seismic ("co", using con-  
 263 tinuous GNSS data), and the other which also includes several days of post-seismic slip  
 264 ("co+post", using cGNSS and InSAR). Two SAR images were acquired 6 days after the  
 265 mainshock rupture, making the L'Aquila earthquake a perfect case study for our pro-  
 266 posed approach. The "co" dataset corresponds to surface displacements measured be-  
 267 tween the earthquake time occurrence ( $t_0$ ) and 25-30 s after  $t_0$ , and includes the static  
 268 offsets of 41 (including high-rates) GPS stations surrounding the earthquake area pro-  
 269 cessed by *Avallone et al.* [2011]. The "co+post" dataset covers the co-seismic phase plus  
 270 6 days of post-seismic slip, documented by 40 static GPS offsets and 2 InSAR frames:  
 271 an ascending COSMO-SkyMed frame and a descending Envisat frame (Tab. S1). The  
 272 observations and their processing are detailed in Supplementary Material [Section S1,  
 273 *Rosen et al.*, 2004; *Lohman and Simons*, 2005; *Jolivet et al.*, 2012].

274 The Paganica fault is generally thought to be responsible for the co-seismic rup-  
 275 ture of the L'Aquila earthquake, and also for most of its post-seismic deformation [*D'Agostino*  
 276 *et al.*, 2012; *Cheloni et al.*, 2014; *Yano et al.*, 2014] along with the northernmost Cam-  
 277 potosto fault [Figure 2, *Gualandi et al.*, 2014]. Although the distribution of relocalized  
 278 aftershocks and surface rupture suggest that the Paganica fault system is possibly seg-  
 279 mented [*Boncio et al.*, 2010; *Lavecchia et al.*, 2012] and/or curved at depth [*Chiaraluce*  
 280 *et al.*, 2011; *Lavecchia et al.*, 2012; *Valoroso et al.*, 2013], its geometry remains poorly  
 281 constrained below the surface. The variability of published morphologies for the causative  
 282 fault [*Lavecchia et al.*, 2012] suggests that even with a large amount of observations and  
 283 a great seismotectonic knowledge of the area, it is not possible to determine a unique fault  
 284 geometry. This is why we approximate the Paganica fault geometry as a planar surface.  
 285 We determine strike and position from the trace of the co-seismic surface rupture [*EMER-*  
 286 *GEO Working Group*, 2010; *Boncio et al.*, 2010] and formerly identified seismogenic faults  
 287 [e.g. *Boncio et al.*, 2004b]. We select the dip and width based on aftershock relocations  
 288 and focal mechanisms [e.g. *Chiaraluce et al.*, 2011; *Chiaraluce*, 2012; *Valoroso et al.*, 2013].  
 289 Hence, our preferred geometry extends over 25 km south of coordinates ( $13.386^\circ$  E,  $42.445^\circ$   
 290 N) with a strike of  $N142^\circ$ . We set fault dip at  $54^\circ$  and width at 18 km, such that the  
 291 fault is reaching the ground surface. This geometry is in agreement with already pro-  
 292 posed causative structures [e.g. *Lavecchia et al.*, 2012]. The fault is divided into 154 sub-

293 faults of 1.8 km length and 1.6 km width. As our fault geometry does not reflect the re-  
 294 ality and is poorly constrained, we account for its uncertainties [*Ragon et al.*, 2018, 2019]  
 295 and assume a standard deviation on the fault dip of  $5^\circ$  and on the fault position (the  
 296 fault position varying perpendicularly from the fault trace) of 1.5 km, regarding the dis-  
 297 crepancies between published fault models [e.g. *Lavecchia et al.*, 2012]. We do not ac-  
 298 count for uncertainties in the fault strike for simplicity, and because its influence might  
 299 be small when uncertainty in the fault position is already acknowledged [*Ragon et al.*,  
 300 2019].

301 We perform the static slip inversion assuming a 1-D layered elastic structure de-  
 302 rived from the CIA velocity model [*Herrmann et al.*, 2011], and calculate Green’s func-  
 303 tions with the EDKS software [*Zhu and Rivera*, 2002]. We precompute Green’s functions  
 304 at depths intervals of 500 m down to 15 km depth and every 5 km below. Laterally, the  
 305 Green’s functions are computed every kilometer to reach the maximum epicentral dis-  
 306 tance of 100 km. Then, we interpolate and sum pre-computed Green’s functions given  
 307 our fault geometry and data locations. The strong variability in published elastic mod-  
 308 els for the central Italy [*Herrmann et al.*, 2011] can have a strong influence on co-seismic  
 309 slip estimates [e.g. *Trasatti et al.*, 2011; *Volpe et al.*, 2012; *Gallovic et al.*, 2015]. We thus  
 310 account for the uncertainties in our Earth model [*Duputel et al.*, 2014] assuming a stan-  
 311 dard deviation on shear modulus of 4 % at depths greater than 15 km and 13 % above.  
 312 These values have been chosen a priori considering the variability between layered mod-  
 313 els and the horizontal variability of 3D crustal models for several depth intervals [*Magnoni*  
 314 *et al.*, 2014].

315 We perform our static slip inversion as previously detailed in Section 2. We spec-  
 316 ify prior distributions for each model parameter: a zero-mean Gaussian prior  $p(\mathbf{m}) =$   
 317  $\mathcal{N}(-10 \text{ cm}, 10 \text{ cm})$  on the strike-slip component (we assume that, on average, the slip di-  
 318 rection is along dip) and we consider each possible value of dip-slip displacement equally  
 319 likely if it does not exceed 20 cm of reverse slip and 5 m of normal slip:  $p(\mathbf{m}) = \mathcal{U}(-$   
 320 20 cm, 500 cm).

## 321 4.2 Co-seismic and early post-seismic slip models

322 We will start by analyzing models inferred by the independent approach as applied  
 323 to the two datasets. The first model is inferred from "co" data (model COgps) and the

324 second is estimated from the "co+post" dataset (model COPOST). The "co+post" dataset  
325 is similar to what has been used in several previous studies to infer the co-seismic slip  
326 [*Cirella et al.*, 2012; *D'Agostino et al.*, 2012; *Cheloni et al.*, 2014; *Yano et al.*, 2014; *Volpe*  
327 *et al.*, 2015]. The results of these two inversions will then be compared to those of the  
328 CTW inversion. For the sake of comparison, these inversions are performed without ac-  
329 counting for epistemic uncertainties. This refinement will only be added in a final inver-  
330 sion.

331 For each approach, we estimate 300,000 samples among the most plausible mod-  
332 els of the full solution space. This set of samples provides information on the possible  
333 parameter values and on their uncertainty. Mean and median models are basic proba-  
334 bilistic values but can give a good overview of the range of most likely solutions. More  
335 detailed quantities, such as the marginal posterior distribution of a given parameter or  
336 the variability of slip between neighbor subfaults, will inform on the uncertainty and trade-  
337 off of the inversion. Yet, the mere interpretation of average and median samples should  
338 be taken with caution: the mean model does not reflect the uncertainty of inferred pa-  
339 rameters, and it does not reproduce either the covariance between several parameters.

340 As we are tied to the need of presenting our results with 2D figures when the ex-  
341 ploration is done on a parameter space with tenth of dimensions, we choose to represent  
342 our results in 3 different ways. For the first representation (e.g., Figures 3a-d, 5a-b), we  
343 divide our set of 300,000 samples into 25 families, and represent the median model of each  
344 family as subdivisions of subfaults (each subfault is thus divided in 25 pixels, more in-  
345 formation in Figure S3). The first family gathers samples with parameters of less than  
346 50 cm offset from the mean model parameters. Other families are built iteratively around  
347 a randomly selected model that has not fitted within antecedent families, except for the  
348 last family which regroups orphan samples). This first representation illustrates the vari-  
349 ability in slip amplitude for a particular parameter. Also, it allows for a visual estima-  
350 tion of the covariance between neighbor parameters: if samples that infer large dip-slip  
351 on a subfault infer small dip-slip on its neighbor subfault, and vice versa, then there is  
352 a trade-off between these two parameters.

353 A second representation shows the posterior marginal Probability Density Func-  
354 tions of the dip-slip parameters which, for a particular parameter, will inform on the amount  
355 of slip uncertainty associated to each subfault (e.g. Figures 3e-g, 5c-e). In particular, the

356 shape and width of the posterior PDF can be considered as a proxy of the model res-  
 357 olution for the inferred parameter.

358 The third representation is a classic map view of the dip-slip amplitude and rake  
 359 of the mean model (e.g. Figures a and b in S4, S8, S12).

#### 360 ***4.2.1 Approaches assuming independent datasets***

361 When solving for the model COgps, we find that most of the slip is concentrated  
 362 in the shallow parts of the fault (Figures 3a, S4a,c,e). Slip amplitudes reach 230 cm in  
 363 the first two kilometers below the Earth surface. These values largely contradict field  
 364 observations hardly reporting more than 15 cm of surface offset [*Faluccci et al.*, 2009;  
 365 *Vittori et al.*, 2011]. This contradiction probably derives from our limited set of obser-  
 366 vations, with only 4 GPS stations documenting the rupture in the near field (Figure 4a).  
 367 The COgps model is thus largely under-determined and unlikely to represent a reliable  
 368 image of the co-seismic deformation. In contrast, the COPOST slip model is inferred from  
 369 a more populated dataset extending over a large part of the Paganica fault (Fig. 3b, S4b).  
 370 The patch of highest slip amplitude, reaching more than 150 cm, is well constrained and  
 371 located between 5 to 7 km depth (Figure 3b). Up to 100 cm slip is also inferred below  
 372 the hypocenter. The inferred slip is almost purely dip-slip (Fig. S4b,d,f). The scalar seis-  
 373 mic moment of model COPOST, calculated with  $\mu = 3.5 \cdot 10^{10}$ , is  $M_0 = 4.9 \pm 0.67 \cdot 10^{25}$   
 374 dyne.m. This value corresponds to a  $M_w$  6.4 earthquake rather than a  $M_w$  6.3.

375 The comparison between observations and predicted surface displacement is shown  
 376 in Figure 4 for the GPS datasets and in Figure S5 for the interferograms. As expected,  
 377 the COgps model well explains the "co" dataset (Figure 4b), but its predictions hardly  
 378 fit the interferograms of the "co+post" dataset (Table S2). In contrast, the predicted  
 379 surface displacement of the COPOST model well approaches the "co+post" observations  
 380 (Figures 4b and S5), with limited residuals (Tab. S2). The surface displacement that hap-  
 381 pened up to 6 days after the mainshock is similar when estimated with GPS or with the  
 382 ENVISAT interferogram (Figure S6). Yet, the displacement measured with the COSMO-  
 383 SkyMed interferogram differs from the one measured with GPS, particularly for the high-  
 384 rate GPS station CADO (Figure S7). This misfit could be related to unwrapping errors  
 385 or effects that has not been completely corrected, explaining why the COSMO-SkyMed  
 386 data are systematically less well fitted by predictions than other observations.

#### 387 **4.2.2 Dual time approach, without epistemic uncertainties**

388 With the CTW approach, we infer two slip models: the strictly co-seismic model  
 389 sCO (see Figures 3c, S8a,c,e and S9 for an animated compilation of probable models)  
 390 and the model sPOST which reflects the 6 days displacement following the mainshock  
 391 (Figures 3d and S8b,d,f). The model sCO, exploiting information from both the "co"  
 392 and "co+post" datasets, is in agreement with the main characteristics of the COPOST  
 393 model (Figure 3b): the location, rake and amplitude of the maximum slip patch are com-  
 394 parable (Figures S8a,c,e), and a large amount of slip is also inferred at depth, up to 75  
 395 cm on average and exceeding 150 cm for some models (Figure 3c). However, unlike the  
 396 COPOST model, the two main slip patches of the sCO model are delimited by an un-  
 397 ruptured area (Figure 3c). Overall, the two models differ on average by 44% and by up  
 398 to 75% for some subfaults characterized by high slip amplitudes (Figures 3e-g, S10), mainly  
 399 because of the variability of the amount of slip inferred below 5 km depth. For the model  
 400 sCO,  $M_0 = 3.50 \pm 0.63 \cdot 10^{20}$  N.m and is equivalent to a moment magnitude of 6.3, cor-  
 401 responding to the GCMT value of 6.3.

402 With the CTW approach, we also find that a large portion of the fault slipped dur-  
 403 ing a 6 days time window after the mainshock (Figure 3d), with maximum amplitude  
 404 of 30 cm in the dip-slip direction (Figure S8b). The largest post-seismic slip ( $\approx 45$  cm)  
 405 is located between the co-seismically ruptured patches (Figure 3c), and is well constrained  
 406 with only 15 cm of posterior uncertainty (Figure S8f). Overall, post-seismic slip (30 cm  
 407 and below) tends to locate around the highest co-seismic slip patch and the hypocen-  
 408 ter, but also overlaps the deepest co-seismic slip patch. Yet, below 10 km depth, the pos-  
 409 terior uncertainty can reach 100% of the median slip amplitude, meaning that the res-  
 410 olution is poor and that it is difficult to interpret anything at that level of detail (Fig-  
 411 ures S8d,f). The seismic moment of model sPOST is  $M_0 = 1.58 \pm 0.63 \cdot 10^{20}$  N.m. The  
 412 predicted surface displacements fit well the observations (Figures 4c,d and Figure S11)  
 413 with residuals similar to the ones of the COPOST model (Tab. S2).

414 As expected, the areas of largest post-seismic slip in the sPOST model correspond  
 415 to the locations of largest divergence between COPOST and sCO models (3b-g). In sum-  
 416 mary, usual approaches using independent datasets do not allow us to infer reliable im-  
 417 ages of the strictly co-seismic and early post-seismic phases. Whereas the "co+post" slip  
 418 model is reliable, the "co" model is not robust enough to retrieve the early afterslip from



419 the subtraction of these two slip distributions. Additionally, the scalar seismic moment  
 420 of model "co+post" corresponds to a moment magnitude greater than the GCMT  $M_w$   
 421 of 6.3. In contrast, the CTW approach allows us to infer robust estimates of both co-  
 422 seismic and post-seismic slip, to exploit all the information collected within our geode-  
 423 tic observations, and to correctly estimate the seismic moment. However, the reliabil-  
 424 ity of these models can be questioned as they do not account for uncertainties in the for-  
 425 ward model.

#### 426 *4.2.3 Dual time approach, accounting for epistemic uncertainties*

427 Here, we present the results of the CTW approach, as in the previous section, but  
 428 accounting for epistemic uncertainties. We will name the resulting models COpref and  
 429 POSTpref since they correspond to our preferred approach providing the most complete  
 430 and objective evaluation of the problem (see Figures 5, 6, S12 and S14 for an animated  
 431 compilation of probable co-seismic models). The distribution of the co-seismic slip dif-  
 432 fers by 42% on average and by up to 88% locally from the models inferred without ac-  
 433 counting for uncertainties (Figures 5c, S13). The co-seismic slip is now limited (on av-  
 434 erage) to a single 10 km long patch located between 5 and 10 km depth, reaching more  
 435 than 150 cm amplitude to the south (right-hand side of Figure 5a), with a rake of  $90^\circ$   
 436 that does not vary along the patch (left-hand side of Fig. 6). The corresponding scalar  
 437 seismic moment  $M_0 = 3.03 \pm 0.64 \cdot 10^{20}$  N.m is slightly lower than what was estimated  
 438 for the model sCO but is still very close ( $M_w = 6.28 \pm 0.06$ ) to the (GCMT) value of  
 439  $M_w$  6.3.

440 Compared to the COpref model, the main characteristics of the POSTpref model  
 441 are not strongly affected by the inclusion of uncertainties. Overall, post-seismic slip (20  
 442 cm and below) occurs mostly below the co-seismic high slip patch, where almost no (less  
 443 than 20 cm) co-seismic slip is imaged. Subfaults with the largest post-seismic slip (more  
 444 than 40 cm, almost purely dip-slip as shown in the right-hand side of Fig. 6) tend to be  
 445 located around or on the edges of the co-seismic high slip patch (Figure 5b). The pres-  
 446 ence of large post-seismic slip below 10 km depth is unlikely as the posterior uncertainty  
 447 reaches 150% of the median slip (the resolution is poor at depth, Figure S12). Thus, only  
 448 3 narrow zones most probably slipped post-seismically (see Figure 5b, and a compari-  
 449 son of median and maximum a posteriori models in Figure S15).  $M_0$  is similar to model  
 450 sPOST with a value of  $1.60 \pm 0.63 \cdot 10^{20}$  N.m. The addition of epistemic uncertainties

451 has increased the residuals between observations and predictions (see Tab. S2 and Fig-  
 452 ure S16). This behavior was expected as the inclusion of  $C_p$  allows the inversion to tol-  
 453 erate for larger misfits at data points where the forward model predictions are less re-  
 454 liable [Ragon *et al.*, 2018, 2019].

455 In summary, the CTW approach shows that if early post-seismic is not acknowl-  
 456 edged as post-seismic signal, co-seismic models may be biased by more than 40% on av-  
 457 erage and of up to 75% locally. But we also learn from these different tests that adding  
 458 more data into the problem is not sufficient, and epistemic uncertainties remain criti-  
 459 cal for the inference of a reliable model. Altogether, our results emphasize the need to  
 460 account for two types of bias in the slip models: the contamination of co-seismic obser-  
 461 vations by some early post-seismic signal, and not acknowledging the uncertainties as-  
 462 sociated with the forward problem.

## 463 5 Discussion

### 464 5.1 Discussion of the CTW approach

465 Observations of co-seismic or post-seismic processes are often contaminated by other  
 466 sources of deformation (mainly post-seismic or co-seismic, respectively) and are widely  
 467 used, as non-contaminated data are rare and scarcely distributed. Optimizing the use  
 468 of the information content in each dataset is thus critical to improve the robustness of  
 469 both co-seismic and post-seismic slip models. A first approach would be to account for  
 470 potential uncertainties in the co-seismic model due to early afterslip in the form of a co-  
 471 variance matrix, as already proposed in *Bletery et al.* [2016]. While this approach helps  
 472 inferring more reliable co-seismic models at a low computational cost, it does not allow  
 473 us to estimate the early afterslip and needs a prior evaluation of the amount of afterslip  
 474 considered as co-seismic signal. Another strategy would be to jointly infer "co" and "co+post"  
 475 data as if they were strictly co-seismic, and to select models that better explain the "co"  
 476 observations, as in [Chlieh *et al.*, 2007]. In this case, the computational cost is increased  
 477 because several models have to be tested. Additionally, with these approaches the early  
 478 post-seismic slip is not estimated. In contrast, the CTW approach we use in this study  
 479 allows us to discriminate co-seismic from early post-seismic slip and to reliably estimate  
 480 corresponding slip models, as suggested also by the results of *Floyd et al.* [2016] and *Wal-*  
 481 *ters et al.* [2018]. The CTW approach takes advantage of the InSAR data that recorded

482 both co- and post-seismic deformation to help constrain both strictly co- and early post-  
483 seismic models.

484 Our results on the L'Aquila event show that the early afterslip, here correspond-  
485 ing to 6 days after the co-seismic rupture, can reach a fourth to a third of the amplitude  
486 of the co-seismic slip. If the early afterslip is acknowledged as co-seismic signal, co-seismic  
487 models of the L'Aquila event are biased. The impact of early afterslip on the co-seismic  
488 models is particularly large in the case of the L'Aquila event and questions the generic  
489 nature of this result. Overall, early afterslip remains poorly studied but has been shown  
490 to range from 0.6% to more than 8% of the co-seismic peak slip in the first 3-4 hours fol-  
491 lowing an earthquake [respectively for the 2009 great Tohoku-Oki earthquake and the  
492 2012  $M_w$  7.6 Nicoya earthquake, *Munekane, 2012; Malservisi et al., 2015*]. *Twardzik et al.*  
493 [2019] also conclude that co-seismic models may be biased by more than 10 % if the early  
494 post-seismic deformation is neglected. Thus, that the post-seismic deformation ongoing  
495 6 days after the mainshock reaches up to 20 % of the co-seismic slip of the L'Aquila earth-  
496 quake might not be an extreme case.

497 Our tests also demonstrate that models are largely impacted by the introduction  
498 of epistemic uncertainties (Figure 7). This impact could mean the assumed fault and Earth  
499 properties are not realistic enough to capture the real seismic rupture, and/or that small  
500 variations of the fault geometry (slight curvature, roughness) or of the Earth model (3D  
501 heterogeneities) largely affect our slip models. The influence of epistemic uncertainties  
502 is greater on the co-seismic model, as expected from the fact that these uncertainties scale  
503 with the amount of slip [*Duputel et al., 2014; Ragon et al., 2018*]. Accounting for uncer-  
504 tainties of the forward model allowed us to exclude the possibility of deep slip for the  
505 co-seismic models, but not totally for the post-seismic models probably because of the  
506 much lower slip amplitudes. Additionally, accounting for  $C_p$  prevented the most prob-  
507 able co- and post-seismic slips to overlap in deeper parts of the fault. The inclusion of  
508 epistemic uncertainties acts like a smoothing constraint on the slip distribution, but with  
509 a smoothing factor being controlled by the inaccuracies of the forward problem.

## 5.2 Non-uniqueness of co-seismic and afterslip models of the L'Aquila earthquake

Our results on the L'Aquila event indicate that the strictly co-seismic slip is concentrated in a thin horizontal band located between 5 and 7 km depth and reaching more than 150 cm in amplitude at its southern end, with no large slip amplitudes inferred below 8 km depth (Figures 6 and 8). The highest amplitude is reached at about 6 km depth south west of the hypocenter, a rupture area also imaged in the co-seismic models of *Gualandi et al.* [2014], *Gallovic et al.* [2015] and *Cirella et al.* [2012] (inferred respectively from GPS only, from accelerometric and high rate GPS data, and from GPS, InSAR and strong motion, see Figure 8) and most of other authors [*Atzori et al.*, 2009; *Trasatti et al.*, 2011; *D'Agostino et al.*, 2012; *Serpelloni et al.*, 2012; *Cheloni et al.*, 2014; *Balestra and Delouis*, 2015; *Volpe et al.*, 2015]. It is the only recurrent pattern we can notice between the 4 slip models of Figure 8. Indeed, while we do not image any shallow slip, other published slip models do with up to 1.5 m in amplitude [Figure 8, *Cirella et al.*, 2012; *Volpe et al.*, 2015]. Our results suggest that this imaged shallow co-seismic slip may rather be early post-seismic slip (Figure 8). At greater depths, most authors infer large slip amplitudes while our preferred model shows no slip below 8 km depth.

The imaged patches of post-seismic slip ( $\geq 15$  cm) are located around our co-seismic slip, near its hypocenter and southern end. Interestingly, our inferred post-seismic slip is also located near areas that ruptured co-seismically as inferred by other studies (Figure 8). The post-seismic slip that occurred several days to months after the mainshock is characterized by 3 wide slip areas, located SW of the main co-seismic slip patch, above the hypocenter close to the surface, and around the hypocenter [*D'Agostino et al.*, 2012; *Cheloni et al.*, 2014; *Gualandi et al.*, 2014]. Most of these post-seismic models incorrectly treat the first days of post-seismic signal as a co-seismic deformation. While we infer likely afterslip in similar locations, the afterslip patches are limited to narrower areas near the co-seismic rupture (Figure 9). Most of these longer-term post-seismic models cover time periods ranging from 6 days to 9 months after the mainshock, they overlook a large part of the early post-seismic deformation. Thus, the peak amplitude of the early afterslip is up to 3 times larger than what was imaged for several months by *D'Agostino et al.* [2012] and *Cheloni et al.* [2014].

541 Our results show that the amplitude and distribution of long-term afterslip may  
 542 be largely underestimated (here by a factor of 3) if the deformation occurring the first  
 543 few hours to days after the mainshock is not accounted in the post-seismic budget. Thus,  
 544 overlooking the early part of the postseismic phase measured in geodetic data may not  
 545 only bias the estimates of the coseismic slip, but also our estimates of the postseismic  
 546 phase.

### 547 **5.3 Fault frictional properties and relationship between afterslip and af-** 548 **tershocks**

549 The comparison between our early post-seismic model and images of longer-term  
 550 post-seismic slip suggest that afterslip may nucleate preferentially around the co-seismic  
 551 rupture in the days following the mainshock (Figure 9b). Afterwards, the afterslip prop-  
 552 agates and extends, both along-dip and laterally, away from the co-seismic slip [*D'Agostino*  
 553 *et al.*, 2012; *Cheloni et al.*, 2014; *Gualandi et al.*, 2014]. This behavior agrees with mod-  
 554 els explaining afterslip as a result of rate dependent friction behavior. Indeed, in these  
 555 models the afterslip relaxes the stress increment induced in velocity-strengthening area  
 556 by the co-seismic rupture [*Marone et al.*, 1991; *Perfettini and Avouac*, 2004]. The post-  
 557 seismic sliding thus nucleates close to the mainshock asperity and propagates with time  
 558 outward from the rupture zone. That early afterslip relates to the stress changes induced  
 559 by the co-seismic rupture has also been modeled for other events [e.g. the  $M_w$ 8.0 Tokachi-  
 560 oki,  $M_w$ 7.6 Chi-Chi, and the  $M_w$ 6.0 Parkfield, *Miyazaki et al.*, 2004; *Chan and Stein*, 2009;  
 561 *Wang et al.*, 2012, respectively].

562 Additionally, our results show that early afterslip nucleates within narrow areas (1-  
 563 2 km wide), and does not happen everywhere around the co-seismic rupture. Areas slid-  
 564 ing aseismically just after the mainshock are thus limited in size around the co-seismic  
 565 rupture, suggesting that frictional properties vary at a small-scale around the rupture  
 566 zone. It may also suggest that the regions adjacent to co-seismic rupture are potentially  
 567 unstable (i.e. are steady-state velocity weakening). This interpretation agrees with the  
 568 results of *Gualandi et al.* [2014] suggesting the longer-term afterslip regions, that are also  
 569 located farther away from the co-seismic ruptured zone, are characterized by a transi-  
 570 tion between velocity weakening and velocity strengthening behavior. This implies that  
 571 co-seismic rupture occurs and triggers early afterslip in velocity weakening regions; while

572 afterslip propagates away from the ruptured zone in fault regions that progressively be-  
573 come stable with the distance to the mainshock.

574 In Figure 10, we compare the slip distributions imaged for the mainshock and 6  
575 days after, with the distribution of aftershocks detected over 6 days and 9 months af-  
576 ter the mainshock [Valoroso *et al.*, 2013]. As for many earthquakes, aftershocks are dis-  
577 tributed mainly at the ends of the fault [Das and Henry, 2003] with few events located  
578 near the co-seismic rupture. Six days after the mainshock (Figure 10a), our results show  
579 no clear correlation between the location of early afterslip and aftershocks. Months af-  
580 ter the mainshock, the areas with a high density of aftershocks are similar to 6 days af-  
581 ter the mainshock [as suggested by Henry and Das, 2001, whereas the cumulated num-  
582 ber of aftershocks is 8 times larger, see Figure S17] and the post-seismic slip has extended  
583 farther away from the co-seismic rupture. This is why we can observe a spatial corre-  
584 lation between some areas of long-term post-seismic slip and aftershocks [D’Agostino *et al.*,  
585 2012; Cheloni *et al.*, 2014]. The spatial correlation is particularly striking for the south-  
586 ern afterslip patch, for which few early aftershocks are located within the early afterslip  
587 area (Figure 10a) while the aftershock cluster overlies the monthly afterslip that prop-  
588 agated outward from the co-seismically ruptured zone (Figure 10b).

589 From this analysis, we can thus draw only one conclusion: there is no correlation  
590 between the area of large ( $\geq 15$  cm) early afterslip and the location of aftershocks for the  
591 first few days after the mainshock. This conclusion contradicts the observations made  
592 for some other earthquakes although mainly at longer time scales [e.g. Hsu *et al.*, 2006;  
593 Perfettini and Avouac, 2007; Wang *et al.*, 2012; Ross *et al.*, 2017, for time periods span-  
594 ning respectively 11 months, 3.5 years, 5 days and 2.5 months]. Our results could also  
595 suggest that, for some parts of the fault, aftershocks nucleation precedes aseismic slip  
596 which is going to occur for months after the mainshock; aftershocks could thus be partly  
597 explained by stress changes due to the co-seismic rupture. But these aftershocks could  
598 also be triggered by early afterslip with an amplitude so low that it is not inferred by  
599 our model. The absence of clear correlation between early afterslip and aftershocks may  
600 also be related to the presence of high pressure fluids in the seismogenic zone of the L’Aquila  
601 event, and of Central Italy in general, with the widespread emissions of CO<sub>2</sub> rich fluids  
602 for deep origin [Chiodini *et al.*, 2000; Frezzotti *et al.*, 2009; Chiodini *et al.*, 2011]. Already,  
603 Miller *et al.* [2004] and Antonioli *et al.* [2005] proposed that the aftershocks and spatio-  
604 temporal migration of the seismicity of the 1997 Umbria-Marche seismic sequence (80

605 km NE of the L'Aquila event) were driven by the co-seismically induced fluid pressure  
 606 migration. Similarly, the increase in seismicity rate of the L'Aquila earthquake and the  
 607 occurrence of some aftershocks may have been driven by fluid flows [Luccio *et al.*, 2010;  
 608 Terakawa *et al.*, 2010; Malagnini *et al.*, 2012], as for the aftershocks migration follow-  
 609 ing the 2016 Amatrice earthquake [Tung and Masterlark, 2018; Walters *et al.*, 2018]. High  
 610 pressure fluids have been observed before the co-seismic rupture, and may have impacted  
 611 the nucleation phase of the L'Aquila earthquake [Lucente *et al.*, 2010]. Finally, Malagnini  
 612 *et al.* [2012] show that the strength of the Campotosto fault, just north of the main rup-  
 613 ture (see Figure 2), has been controlled by fluid migration for at least 6 days after the  
 614 mainshock, a time window corresponding to our study of early afterslip. The perturba-  
 615 tions in pore fluid pressure induced by the co-seismic rupture may have triggered the first  
 616 aftershocks of the L'Aquila earthquake. Fluid migration may have prevented aftershocks  
 617 and early afterslip from affecting the same areas of the fault, especially if the increase  
 618 in fluid pressure first produced aseismic slip, followed by triggered seismicity around the  
 619 pore pressure front [Miller *et al.*, 2004]. Finally, if early aftershocks were triggered by  
 620 changes in fluid pressure, it may justify the possibility that some of these early aftershocks  
 621 nucleated before the occurrence of long-term afterslip in similar regions of the fault.

## 622 **6 Conclusion**

623 In this study, we use a simple and efficient approach to account for the differences  
 624 in temporal resolution of various geodetic datasets. A redesign of the Green's Functions  
 625 matrix allows us to optimize the use of the information content of datasets covering dif-  
 626 ferent time periods. In particular, we propose here generic equations governing the Com-  
 627 bined Time Window (CTW) approach, and detail the benefits of this approach with toy  
 628 models. A comparable strategy has already been proposed by Floyd *et al.* [2016] and used  
 629 in an optimization framework. With the CTW approach, we image simultaneously the  
 630 strictly co-seismic slip and the early afterslip (6 days after the mainshock) of the 2009  
 631  $M_w 6.3$  L'Aquila earthquake using two datasets: one covers the two slip episodes (e.g. In-  
 632 SAR) while the other records the co-seismic signal only (e.g. continuous GNSS). We show  
 633 that when the two phases are inverted independently, as is usually the case, the estimated  
 634 slip distributions are not reliable because strictly co-seismic observations are usually of  
 635 poor spatial resolution. Additionally, overlooking the early post-seismic deformation re-  
 636 sults in models that overestimates the co-seismic slip, and underestimates the total post-

637 seismic slip budget. In contrast, the CTW approach allows us to accurately estimate both  
638 co-seismic and early post-seismic slip models. More generally, the CTW approach will  
639 also be useful to discriminate geodetic signals from earthquakes that cluster closely in  
640 space and time, or from different phases of the post-seismic deformation.

641 Our results show that neglecting the contribution of the early post-seismic defor-  
642 mation will likely bias estimates of the co-seismic and/or the post-seismic slip. For our  
643 test case of the L'Aquila earthquake, the peak co-seismic slip is likely 30% greater when  
644 early post-seismic signal is recorded as co-seismic deformation. The long-term afterslip  
645 estimates are underestimated by a factor of 3 when the first 6 days of post-seismic de-  
646 formation are not acknowledged. Our investigation of the L'Aquila event also stressed  
647 the strong influence of uncertainties in the forward model, mainly stemming from our  
648 imperfect knowledge of the fault geometry and the Earth structure, on the imaged slip  
649 distributions. These uncertainties alone are sufficient to cause contradictory interpre-  
650 tations on the slip history on the fault (e.g. with the existence of shallow or dip slip).

651 Our preferred slip model for the L'Aquila earthquake tends to be simpler than many  
652 previous models, with one thin horizontal band of slip located around 7km depth, reach-  
653 ing 150cm in amplitude near its southern end. Our model thus excludes the possibility  
654 of major shallow or deep co-seismic slip patches (less than a few km or deeper than 10).  
655 The early post-seismic slip (6 days after the mainshock) mostly occur in the same in-  
656 termediate depth range (7 km +/- 3 km), initiating on the edges of the co-seismic slip,  
657 with possibly some overlap. Some afterslip may also have occurred at greater depths.  
658 A comparison with longer term afterslip models suggests that the early afterslip patches  
659 might have simply expanded over time from their initial position. Aftershocks are more  
660 spatially distributed (7 km +/- 5 km) but still concentrated at intermediate depth. Sev-  
661 eral studies suggest that aftershocks might be driven by afterslip [e.g., *Perfettini and Avouac,*  
662 *2007; Hsu et al., 2006; Sladen et al., 2010; Ross et al., 2017; Perfettini et al., 2018*] but  
663 here aftershocks are only partially overlapping. This result suggests that post-seismic  
664 reloading may be influenced by fluids as advocated in several previous studies [e.g., *Luc-*  
665 *cio et al., 2010; Terakawa et al., 2010; Malagnini et al., 2012; Guglielmi et al., 2015; Scud-*  
666 *eri and Collettini, 2016*].



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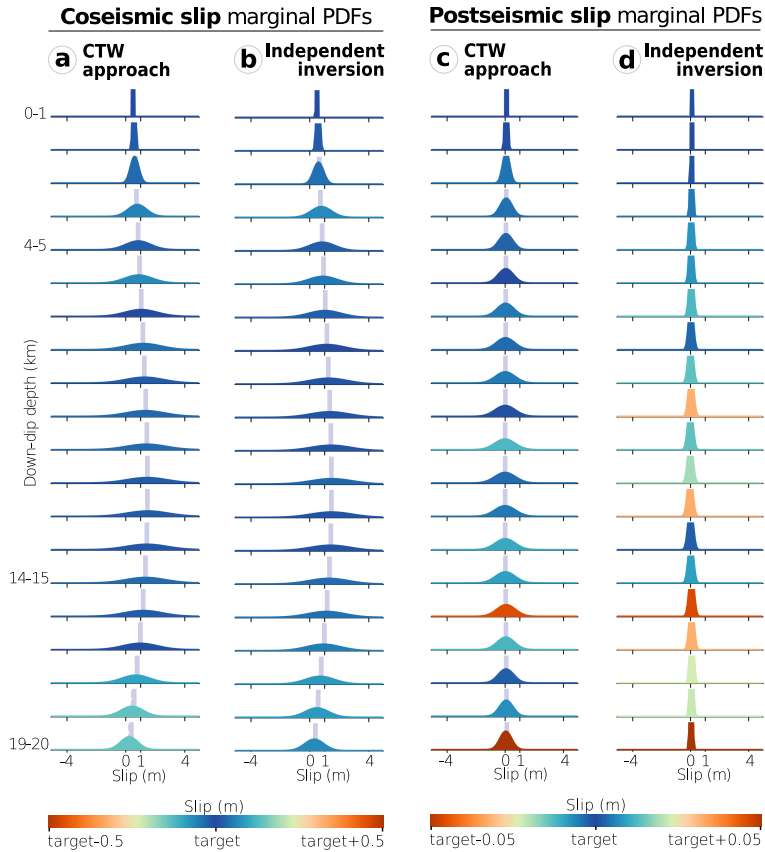
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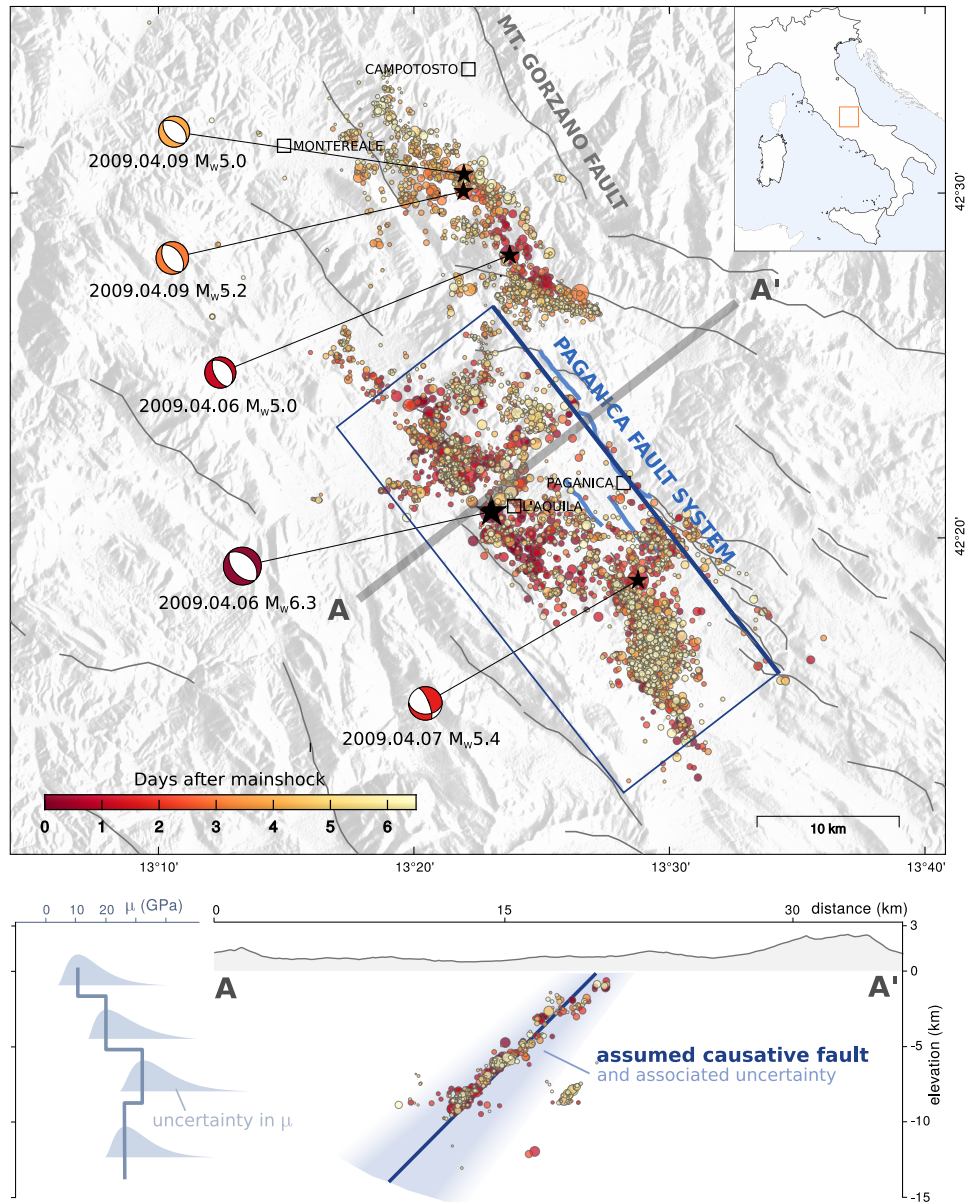
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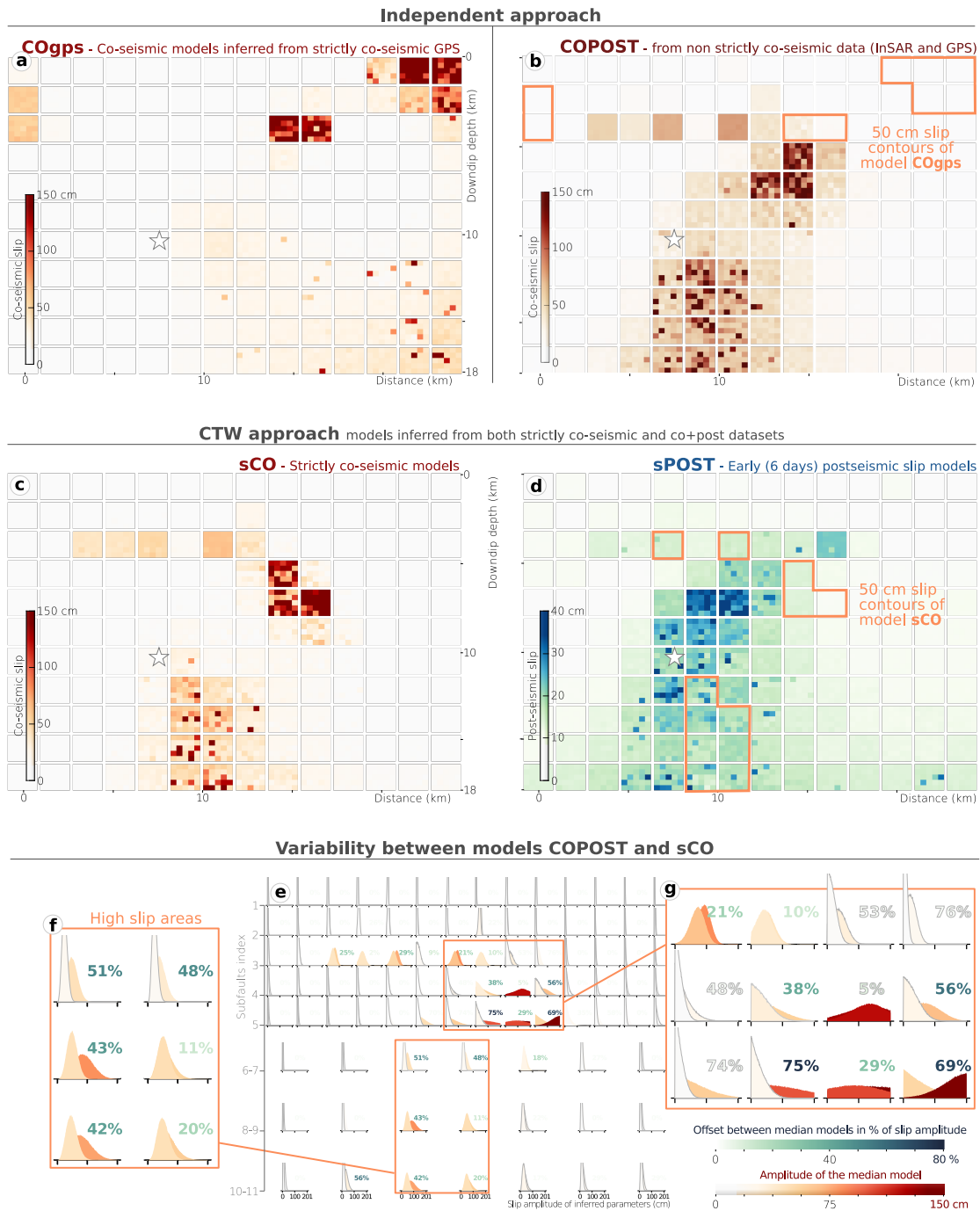
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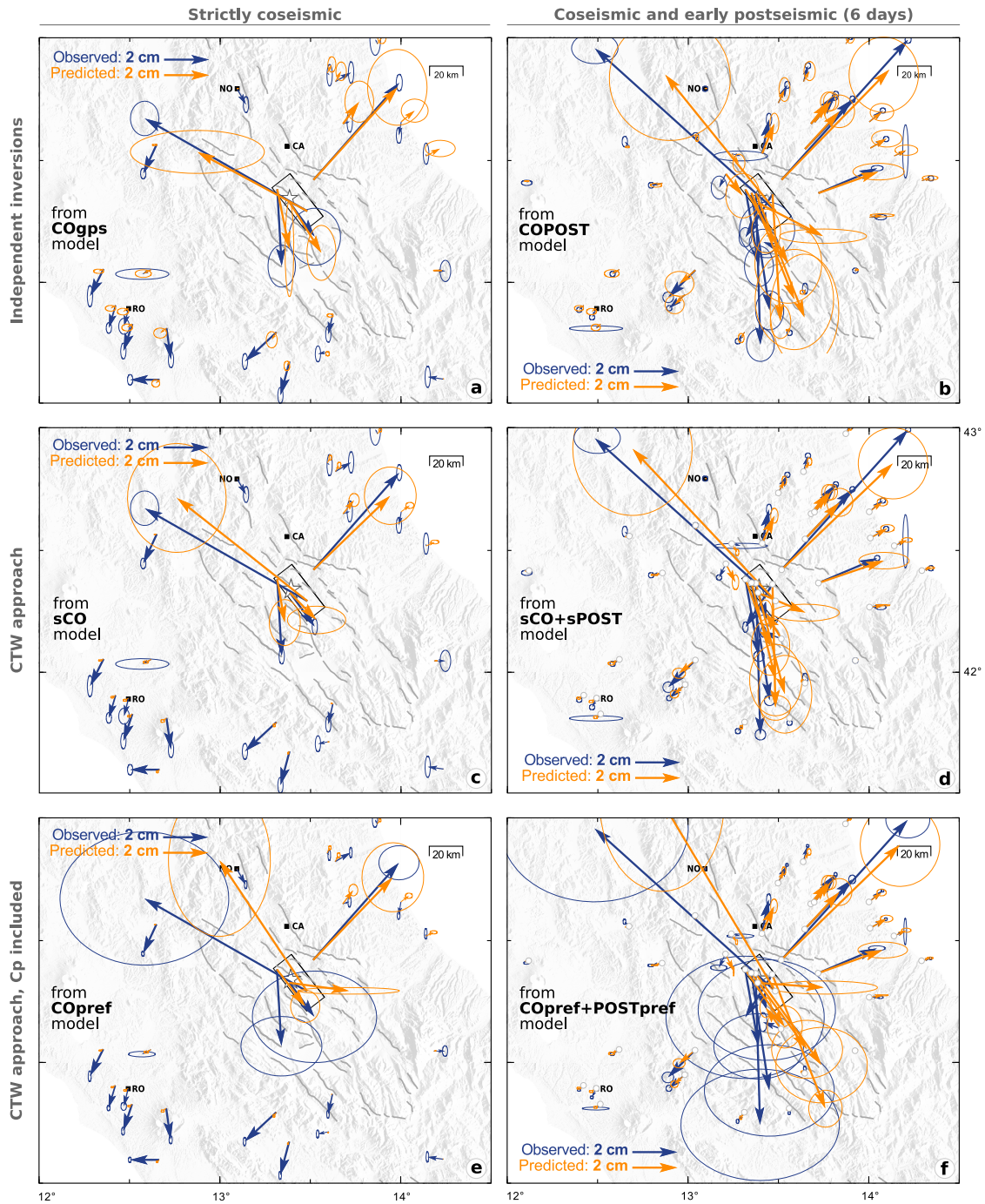
**Figure 1.** Co-seismic and post-seismic slip inferred for the simplified case of a fault that extends infinitely along strike. The co- and post-seismic slip models inferred from the CTW approach are shown in (a) and (c), and can be compared to the slip inferred from the inversion of co-seismic data only (b) and the post-seismic slip distribution (d) resulting from the difference between slip inferred from the inversion of co+post data and slip of (b). The fault is discretized along dip in 20 subfaults, and the input slip (i.e. the target) for each subfault is represented as a gray vertical line. For each subfault, the posterior PDFs of co-seismic (a and b) and post-seismic (c and d) slip is colored according to the offset between the target parameter and the posterior mean, with a colorscale saturated at 50 cm for the co-seismic slip and at 5 cm for the post-seismic slip. The target slip is well inferred if the PDF of a particular parameter is colored in dark blue, while it is not if the PDF is colored in red.



**Figure 2.** Seismotectonic framework of the area involved in the 2009 seismic sequence (top) and assumed forward model and associated uncertainties (bottom). In the map, coloured circles are the aftershocks from 2009 April 6 at 01:32 UTC to 2009 April 12, from the catalog of *Valoroso et al.* [2013]. The aftershocks are coloured from their occurrence time after the mainshock. Beach balls are the focal mechanisms of the mainshock and four main aftershocks, with their respective epicenters located by black stars. Solid gray lines are the major seismogenic faults of the area [Boncio *et al.*, 2004a; Lavecchia *et al.*, 2012]. The observed co-seismic surface rupture is indicated with continuous blue lines [Boncio *et al.*, 2010]. Our assumed fault geometry is shown with a dark blue rectangle. In the elevation profile (bottom), uncertainty in the fault geometry is illustrated in blue. The assumed elastic modulus  $\mu$  and associated uncertainties are also illustrated for the 12 first kilometers below the Earth surface.

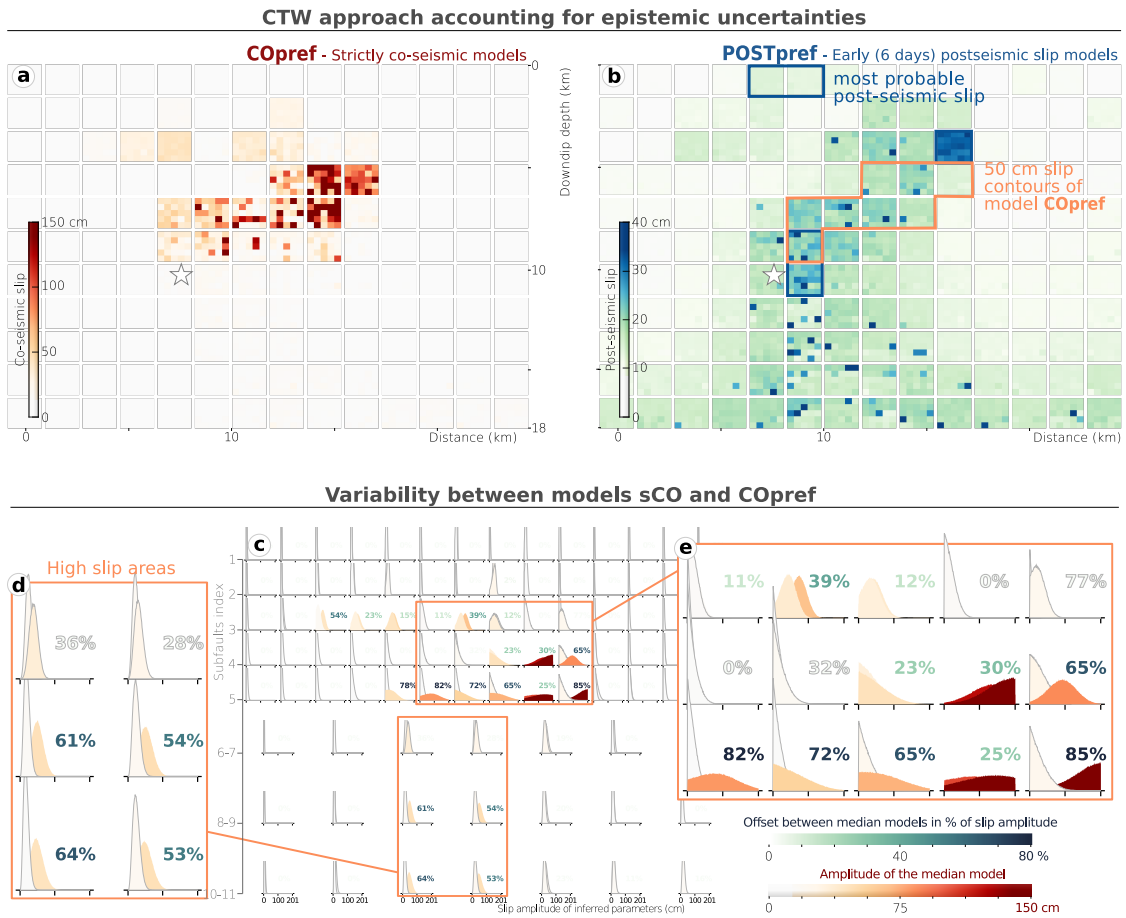


**Figure 3.** Comparison between finite-fault models inferred with the independent or the CTW approach. (a) Strictly co-seismic (30 s after the mainshock) slip model, named COgps, inferred from the strictly co-seismic dataset (GPS only). (b) Non-strictly co-seismic model COPOST inferred from the co-seismic dataset contaminated with some post-seismic deformation. (c) and (d) Strictly co-seismic sCO and early postseismic sPOST (6 days after the mainshock) slip models inferred jointly with the CTW approach. (a) to (d) illustrate the slip amplitude of the median models (sum of strike- and dip-slip component) of 25 families of inferred models (more information in the text and Figure S3). Each subfault (large square) is divided into 25 pixels colored from the slip amplitude of the corresponding median model. (e) Comparison between the posterior Probability Density Functions of the dip-slip parameters of models COPOST (b) and sCO (c), colored from the amplitude of their median model. In the last four rows, the PDFs have been merged, covering 2 subfaults along strike and 2 subfaults along dip (i.e. patches two times bigger

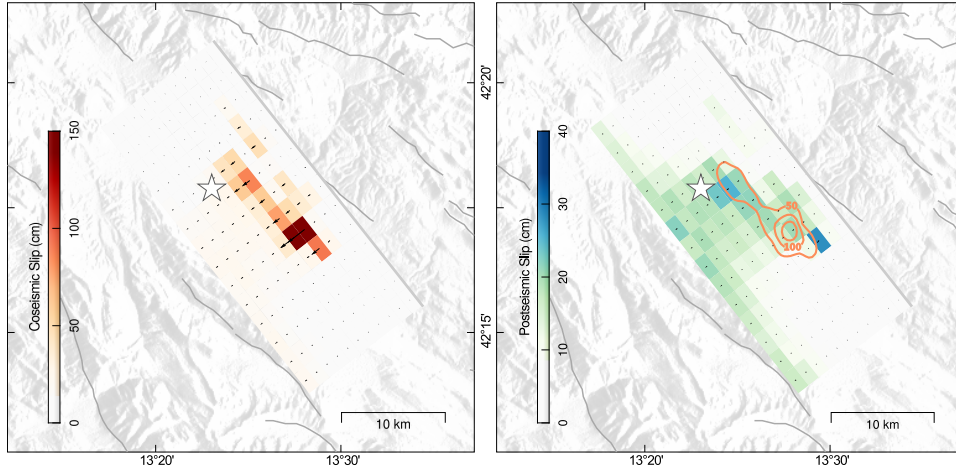


**Figure 4.** Comparison of horizontal surface displacement at GPS stations. Strictly co-seismic displacement is shown on the left while "co+post" displacement is shown on the right. Observed surface displacement is in blue with 95% confidence ellipses. Predictions are in orange with 95% confidence ellipses. In the top and middle rows, observational confidence ellipses (in blue) include only data errors. (a) and (b) The predictions have been calculated independently: using "co" data (a) and the "co+post" dataset (b). In (c) and (d), predictions are derived from the CTW approach. (e) and (f) show the predictions for a similar inversion setup, except epistemic uncertainties have been added to the data errors, enlarging the confidence ellipses. Our fault geometry is shown with a black rectangular box. The cities of Norcia (NO), Campotosto (CA) and Roma (RO) are indicated with black squares. Major seismogenic faults are shown in gray solid lines and the epicenter is the white star.

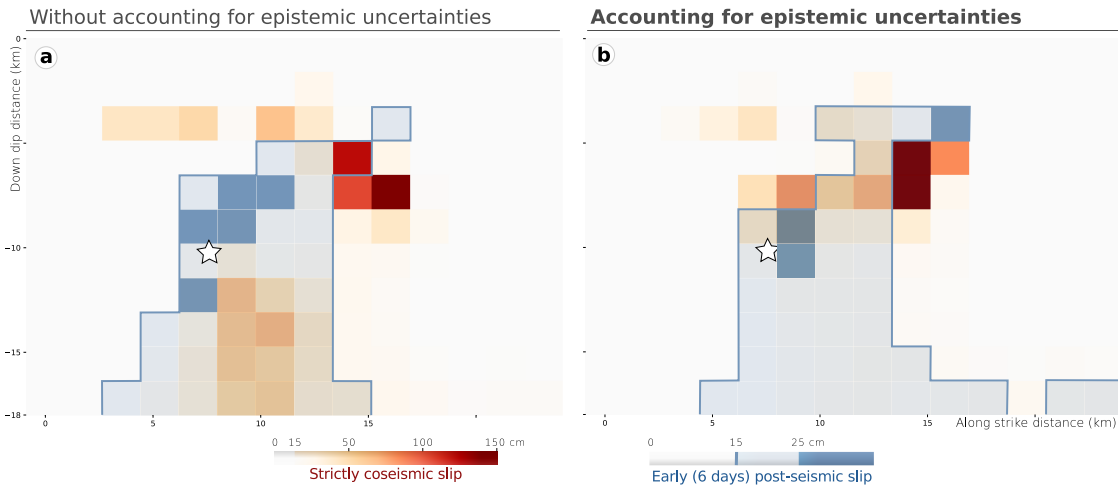




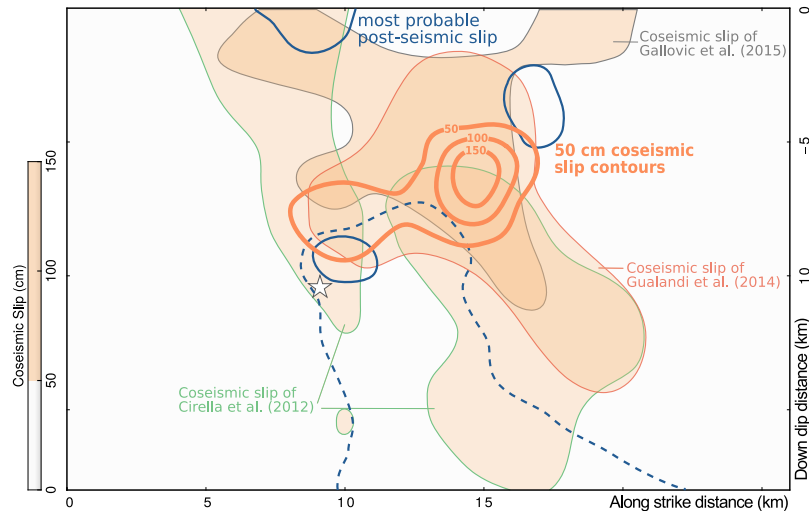
**Figure 5.** (a) and (b) Strictly co-seismic COpref and early postseismic POSTpref (6 days after the mainshock) preferred slip models (sum of strike- and dip-slip components), inferred with the CTW approach and accounting for epistemic uncertainties. (c) Comparison between the posterior Probability Density Functions of the dip-slip parameters of models COpref (a) and sCO (Figure 3c), colored from the amplitude of their median model. In the last four rows, the PDFs have been merged, covering 2 subfaults along strike and 2 subfaults along dip (i.e. patches two times bigger than for the first four rows). The COpref model PDFs are in the foreground while the sCO PDFs are in the background. The offset between the median models is shown as percentage with a different color scale. Two high slip areas are illustrated in detail: the highest slip patch (e) and the deep slip patch (d). The plots of the posterior PDFs are truncated between 0 and 200 cm to simplify the visualization.



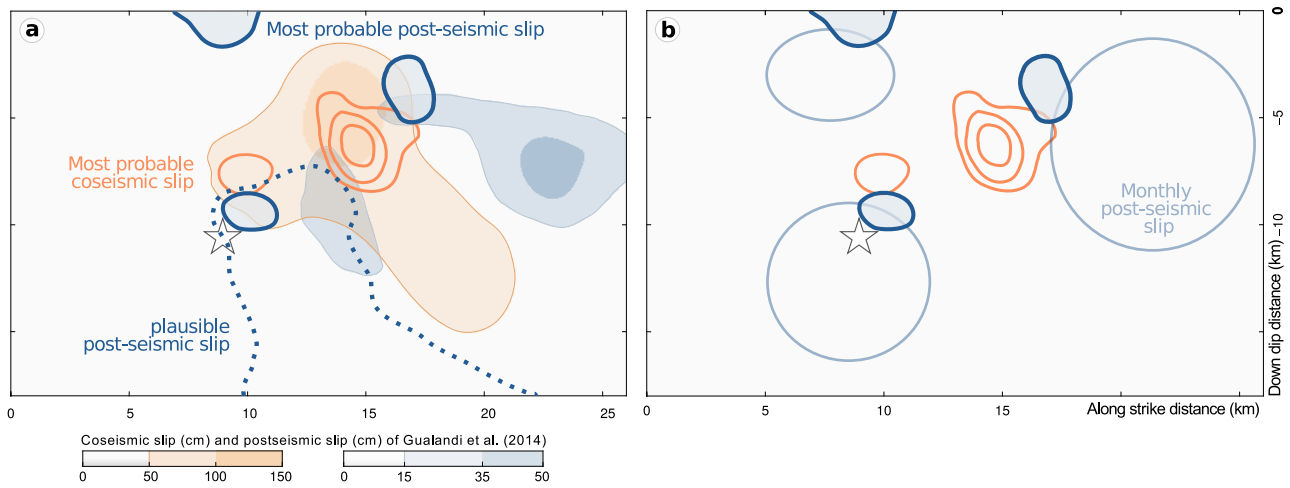
**Figure 6.** Our preferred slip models of the L'Aquila earthquake, inferred with the CTW approach and accounting for epistemic uncertainties. (a) and (b) show the dip-slip amplitude and rake of the average model in map view, the hypocenter being the white star. In (b), orange lines also show the 50 cm co-seismic slip contours.



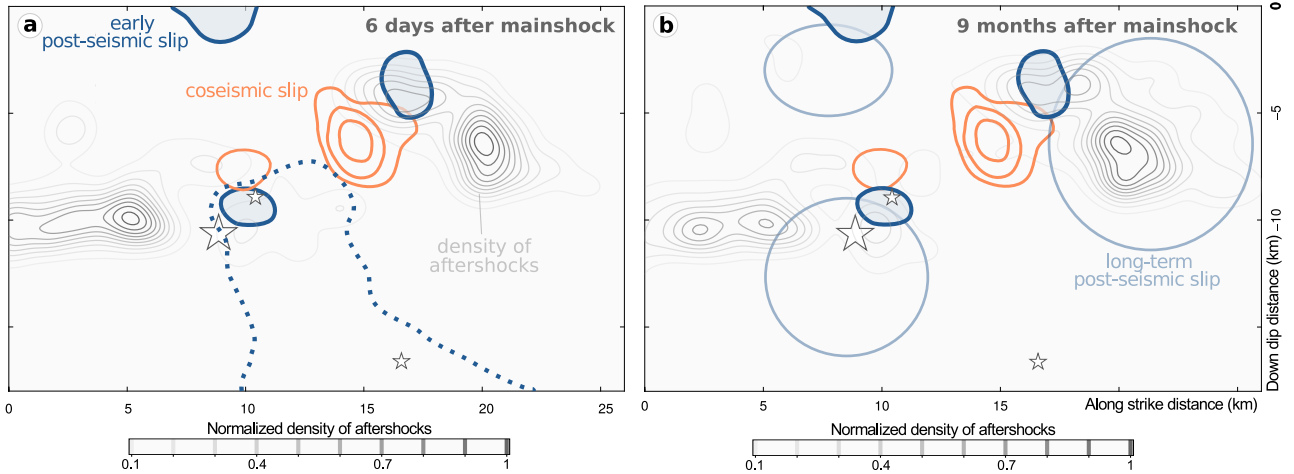
**Figure 7.** Effect of epistemic uncertainties ( $C_p$ ) on the distribution of strictly co-seismic slip and afterslip. The slip models (sum of median strike- and dip-slip components) have been inferred accounting for epistemic uncertainties (b) or not (a). The strictly co-seismic slip median model is in light gray to dark orange colorscale. The subfaults that slipped of more than 15 and 25 cm up to 6 days after the mainshock, according to our median model, are in transparent light and medium blue respectively. The afterslip does not overlap the co-seismic slip when  $C_p$  is accounted for (b), whereas the two slip distributions overlap at depth when no  $C_p$  is included (a).



**Figure 8.** Comparison between the slip distributions inferred with the CTW approach and co-seismic slip distributions of other studies. The strictly co-seismic slip of *Gualandi et al.* [2014] inferred from GPS only, the strictly co-seismic slip of *Gallovic et al.* [2015] inferred from accelerometric and high rate GPS data, the co-seismic slip of *Cirella et al.* [2012] inferred from GPS, InSAR and strong motion, are projected in our fault plane in transparent light orange when slip exceeds 50 cm. The 50 cm contours of our strictly co-seismic slip distribution and the 15 cm contours of our afterslip inferred accounting for epistemic uncertainties are in bold lines, respectively orange and dark blue. The area of afterslip delimited with a dashed blue line is considered as less plausible as inferred with large uncertainties. The hypocenter is the white star.



**Figure 9.** Comparison between our most probable strictly co- and post-seismic slip distribution 6 days after the mainshock and the post-seismic slip up to 306 days after the mainshock. (a) Our most probable slip distributions are represented with bold orange and dark blue lines, respectively for co-seismic (50 cm contours) and post-seismic slip (slip  $\geq 10$  cm). The area of afterslip delimited with a dotted blue line is considered as less plausible as inferred with large uncertainties. The co-seismic slip and afterslip 306 days after the mainshock inferred by *Gualandi et al.* [2014] are plotted with the same color codes but as color swaths. (b) Our results are compared to the area that slipped post-seismically during about 6 months (176 and 194 days respectively) after the mainshock as modeled by both *D’Agostino et al.* [2012] and *Cheloni et al.* [2014]. The hypocenter is the white star.



**Figure 10.** Distribution of most probable co-seismic slip and afterslip models, and the normalized density of aftershocks that occurred (a) within 6 days after the mainshock or (b) up to 9 months after the mainshock [ $M_C = 0.88$ , catalog of *Valoroso et al.*, 2013]. The strictly co-seismic 50 cm slip contours are in orange, while the contours of most probable afterslip (slip  $\geq 10$  cm) are in blue. The area delimited by dotted blue lines has plausibly hosted some afterslip, but not as probably as the other regions. The areas that slipped post-seismically during about 6 months after the mainshock as modeled by *D'Agostino et al.* [2012], *Cheloni et al.* [2014] and *Gualandi et al.* [2014] are the blue circles. The density of aftershocks located within 3 km of the fault (to account for potential uncertainty of the fault geometry) is calculated with a kernel density estimation method [*Parzen*, 1962] with a smoothing factor of 0.6. The cumulated number of aftershocks of  $M_c = 0.88$  is of  $\sim 6000$  6 days after the mainshock and 8 times larger 9 months after the mainshock (Figure S17). The hypocenter and aftershocks of  $M_w \geq 4.4$  are the white stars.