Partitioning the ongoing extension of the central Apennines (Italy): fault slip rates and bulk deformation rates from geodetic and stress data

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Key Points:

- We perform a joint inversion of GNSS velocities and stress data to determine fault slip rates and bulk deformation rates in the central Apennines;
- Complex modeling can extract reasonable estimates of long-term fault slip rates in satisfactory agreement with available geological estimates;
- Model fault slip rates may complement and constrain geological slip rates for any application in seismic hazard assessment.

Abstract

We investigated whether the joint inversion of geodetic and stress-direction data can constrain long-term fault slip rates in the central Apennines, and ultimately how extension is partitioned among fault slip and bulk-lithosphere permanent strain. Geodetic velocities are collected in the interseismic fault stage with steady secular deformation, thus long-term estimates can be derived with a model of elastically-unloading seismogenic faults within a viscously-deforming lithosphere.

As the average spacing of permanent GNSS stations is similar to the average length of seismogenic faults (25-35 km), if not larger, we decided to merge permanent and temporary GNSS measurements, resulting in a denser geodetic dataset. Given that most normal faults in the Apennines have slip-rates around or below 1 mm/a, and most campaign GNSS velocities carry similar uncertainties, simple local back-slip models cannot be applied. Hence, sophisticated modeling is required to extract reasonable bulk deformation rates and long-term fault slip rates even at signal/noise ratio of order unity.

Given the spatial distribution of the GNSS network, we estimated the long-term slip rate of seven major fault systems, that are in satisfactory agreement with available geological slip rates. The resulting spatial distribution of bulk deformation rates locally fits short-term transients; in other cases, they represent the currently-unclear signature of tectonic processes like upper-crustal visco-plastic deformation, aseismic slip or missing faults in the adopted database.

We conclude that the time is ripe for determining fault slip rates using geodetic and stress-direction data, particularly where fault activity rates are hard to determine geologically.

1. Introduction

Regional tectonics has shown that the crust between the Tyrrhenian and Adriatic coastlines is undergoing extension at 3.0-3.5 mm/a: in the past this rate was assumed to be as large as 5-6 mm/a (Westaway, 1992; D'Agostino et al., 2001; Hunstad et al., 2003), but the currently accepted estimate is now rather stable thanks to over 20 years of reliable GNSS time-histories (e.g. Carafa & Bird, 2016; Devoti et al., 2017). This broad-scale extension is largely accommodated by seismogenic faulting in the brittle portion of the Apennines crust, by viscous deformation at larger depth (>10-15 km), and by distributed plastic deformation in the shallow crust between fault tips. The partitioning of this long-term extension among active faults and bulk lithosphere deformation, however, remains a critical aspect to be solved prior to using GNNS data for seismic hazard assessment and for addressing broader geodynamics issues.

In the past different causes of Apennines extension have been suggested, including from diffuse basal shear tractions, lateral variations of topography, and plate interactions (see Carafa et al., 2015 and references therein). Regardless of the methodological choices, the aim of these forward models was to determine the relationships between extension and its causes in the central Apennines; yet, so far they have not been used to explore the partitioning of extension among different faults. This limitation is mainly related to the inadequate knowledge of input parameters, such as, but not limited to, lateral and vertical variations of density and temperature, appropriate flow laws for unfaulted lithosphere, and fault rheology. Additional information on the partitioning of Apennines extension can be gleaned from field geology analyses, stress-direction data, and geodetic measurements, although these observational constraints bring little or no information on fundamental dynamic causes. Thus, achieving the limited goal of determining how extension is partitioned in central Apennines by using existing kinematic information (and related uncertainties) is more realistic than inspecting the relationships between alternative causes and their gross surface effects. In other words, for determining how extension is partitioned among faults and bulk lithosphere deformation, a kinematic inverse model has to be preferred over a dynamic forward one.

So far, data on fault activity in the central Apennines have been collected using different geological and/or geochemical constraints, ranging from the offsets of stratigraphic markers detected in paleoseismological trenches, to variations in cosmogenic isotope concentrations along exhumed fault planes (e.g. Zreda & Noller, 1998; McCalpin, 2009). Although the Apennines landscape is still largely dominated by post-Miocene compressional tectonics, during most of the Quaternary a youthful phase of horizontal extension has overprinted the preexisting fabric, allowing the development of new seismogenic active faults. Nevertheless, several critical methodological questions must be faced prior to using geological fault slip rates, especially for constraining any long-term inverse model. For instance, fault slip rates obtained by cosmogenic dating of limestone or by ^{14}C- and AMS-dating (Accelerator Mass Spectrometry) of offset horizons seen in paleoseismological trenches bring information about a handful specific sites – if not just a single site – along the entire length of the active fault. As such, these estimates may not be representative of the slip rate of the entire fault because of the inevitable along-trace variability. A further issue concerning the use of these methods lies in the assumption that all the observed surface strains are exclusively associated with long-term tectonics and seismogenic processes. In active fault zones, and especially those occurring in steep terrains and in areas of strong lithological heterogeneity, tectonic deformation is often intertwined with non-tectonic surface processes such as mass wasting and erosion. In the specific case of central Apennines, Kastelic et al. (2017) and Stemberk et al. (2019) have shown that fault scarps that separate footwall carbonate bedrock from hanging-wall carbonate talus, possibly representing the surface extension of upper crustal seismogenic faults, are exhumed also by non-tectonic processes, making it hard to detect and quantify the long-term tectonic deformation. Moreover, even surface coseismic ruptures reflect the influence of gravity and rheology, as shown for the 24 August 2016, Amatrice earthquake, and for the 30 October 2016, M_w 6.5 Norcia earthquake (Huang et al., 2017; Di Naccio et al., 2019). It follows that discarding as marginal these surficial processes, that is to say, not removing their characteristic signature from the total deformation, may lead to a gross overestimation of the slip-per-event, and ultimately of the long-term slip rate.

Conversely, geodetic measurements have been proven to provide robust high-resolution observations for coseismic and postseismic fault slip in the central Apennines. Generally such measurements are not used to quantify the relative amount of strain rate due to the fault interseismic stage or to bulk deformation, but have been relegated to large-scale geodynamics issues in the context of the broader Mediterranean region (e.g. Barba et al., 2008; Devoti et al., 2008; Carafa et al., 2018). Because the geodetic sampling interval encompasses only a small fraction of the earthquake cycle (see Figure 1), some assumptions are required to use GNSS measurements for long-term estimates such as fault slip-rates.

The integration of GNSS measurements into long-term deformation estimates has been performed based on two categories of models: 1) dynamic earthquake-cycle models with deformation rates continuously varying throughout the earthquake cycle due to the complex interaction between the elastic upper crust and the viscoelastic lower crust (Johnson, 2013), and 2) kinematic models whose interseismic deformation is assumed (or proved) to be time invariant (Bird, 2009; Zeng & Shen, 2016, Hussain et al. 2018), without resolving the cause of the motion (see Figure 1b). Dynamic modeling for each fault would require the earthquake history consisting of a sequence of periodic, identical events ending with the most recent one (Meade et al., 2013). In this case, any corrections to geodetic time series are related to the stage of each fault in its own earthquake-cycle. Unfortunately, determining the "characteristic" earthquake of each fault of the central Apennines and its current stage in the current earthquake cycle is unrealistic; a comparison of the earthquake record length (Stucchi et al., 2004) with the expected long-term slip rates (Galadini & Galli, 2000; Basili et al., 2008) suggests that the activity of two out of three sources of potentially damaging events (Mw 5.5+) may have gone undetected so far. This implies that not even one complete "seismic cycle" could be modeled based on the available earthquake sample. This point, ~~to all ,~~ together with its simplicity and low computational costs, make kinematic modeling more practical than dynamic modeling for determining fault slip rates from interseismic velocities.

Under the realistic assumption of constant interseismic strain rates, these kinematic models have the main drawback of being inappropriate for reproducing any time-variable deformation due to earthquakes, such as afterslip or viscoelastic rebound. Nevertheless, after inspecting the geodetic observations along several strike-slip faults, Meade et al. (2013) reported that the variation of the velocity field in interseismic period is small enough to be reasonably well represented by a constant trend. This is well evident also in time series of GNSS stations of the central Apennines (e.g. Figure 1c). This implies that the GNSS time series of several benchmarks placed around an active fault provide little information on its earthquake-cycle stage, but the interseismic secular trends, after an adequate elastic correction (the "yearly-averaged displacement" in Figure 1b), may fit sufficiently well the long-term slip rate of that fault. It follows that interseismic velocities can be used to constrain the average long-term velocity field as well as the offset of stratigraphic markers deriving from field data analyses.

In a recent work, Carafa and Bird (2016) proved that introducing the SH_{max} orientations as soft constraint in joint inversions with GNSS measurements greatly improves the model performance with respect to available focal mechanisms. This implies that the spatial variations of borehole breakout azimuth and pressure-axes directions from fault plane solutions give additional constraints on the stress orientations in the brittle part of the crust with respect to GNSS measurements. It is important to recall that each focal mechanism describes the deformation tensor orientations, not the stress; minor misalignments between the P-axes and the maximum horizontal stress direction may hence be expected. The long-term extension of the Apennines is also consistent with the orientation of borehole breakouts, which in their turn are coherent with the average deformation axes of available fault plane solutions (see Carafa and Barba, 2013; Montone and Mariucci, 2016). This empirical observation implies that the angular differences between P-axes and SHmax are expected to be small. As a consequence, a robust long-term deformation model of central Apennines should necessarily consider the SHmax data records.

In summary, no reliable information can be obtained from any of these observations (fault geometries, GNSS measurements and stress directions) on the earthquake-cycle stage of the fault, on the absolute stress, and on the causative forces. However, they still provide strong constraints for inverse modeling of long-term kinematic estimates such as fault slip rates or bulk deformation rates of the lithosphere. Inverse models merging offsets of stratigraphic marker rates, GNSS velocities and stress orientations have already been proved to be reliable enough to be implemented in official seismic hazard models for the United States (e.g., Bird, 2009; Field et al., 2014). Replicating this experience in Italy, and particularly in the central Apennines, requires the development of an exploratory model for testing the reciprocal reliability of available datasets in determining fault slip rates and bulk deformation rates. Creating this exploratory model and discussing its merits, limitations and implications is exactly the goal of this study. More specifically, we present a long-term model with a twofold purpose. On the one hand, we determined fault slip rates and bulk deformation rates of the central Apennines using an innovative approach that is completely free from any dynamic assumptions; as a main consequence, our results are independent from any geological estimates as well as from any constraints on continuum flow laws or fault rheology. On the other hand, we performed a thorough consistency check by comparing our slip rates with the geologic estimates, in search of a desirable overlap or, alternatively, of possible systematic biases. Encouragingly, we found a gross compatibility between our results and the long-term estimates derived from the offset of stratigraphic markers in paleoseimological trenches, or from hundred-meter vertical offsets of Late Quaternary breccias and tephra.

2. Active tectonics data

In this section we present the input data used in our inversion model: the active faults, GNSS measurements and stress direction records.

2.1. Active fault database

The central Apennines are a NE-verging thrust and fold belt consisting of Mesozoic-Cenozoic carbonate rocks and syn-orogenic flysch deposits. In the current stress regime they have been undergoing regional-scale uplift and NE-SW extension since Late Neogene-Early Quaternary times (Patacca et al., 1990; Lavecchia et al., 1994; Galadini et al., 2003). More specifically, the central Apennines are bounded to the west by the Tyrrhenian extensional domain and to the east by the youngest Pliocene-Pleistocene compressional structures of the Apennines and External Dinarides fronts (Patacca et al., 1990; Ghisetti & Vezzani, 1997; Scisciani et al., 2002; Kastelic & Carafa, 2012), and comprise the area that experiences the largest and most frequent earthquakes of the entire chain. The identification and characterization of seismogenic active faults has been a major research target over the past few decades. This larger effort has resulted in a number of compilations of active faults, some of which are supplied as fully georeferenced computer databases. These databases combine intensity data reported for large earthquakes of the past with geological and geophysical evidence and with the results of geomorphological and/or paleoseismological investigations (Galadini & Galli, 2000; Meletti et al., 2000; Valensise & Pantosti, 2001; Basili et al., 2008; Valentini et al., 2017 among others). In some instances, these databases disagree on the exact locations or even on the Holocene activity of some faults (for a review see Vannoli et al., 2012, with references), but in general there exists a satisfactory agreement on fault geometries among different compilers, and complete coherence on fault rake. For a few active faults (e.g. the Fucino, Paganica and Matese faults, Figure 2 and Supporting Figure S1) the joint use of historical earthquakes or recent seismicity and of geological and geomorphological evidence yielded a rather robust description of the entire tectonic system and of its earthquake potential. For the majority of faults, however, clues on the state of activity are derived from faulted or displaced Late Quaternary continental deposits and landforms, or from scant historical evidence for ancient earthquakes, only occasionally supported by paleoseismological observations and by geophysical and geological investigations on basin evolution.

Our fault dataset includes 18 active normal faults, all presumed to be seismogenic in the current literature, occurring over a region that extends for more than 150 km from the Gran Sasso Massif in the north, to the Bojano basin in the south (Figure 2 and Supporting Figure S1). Other more external, i.e. more easterly, active (contractional) faults are not included in the dataset. Fault data were collected using only peer-reviewed sources of structural, geological and seismological data. We presume a fault to be active if there exists evidence of tectonic activity in the Late Pleistocene-Holocene deriving from:

- seismicity investigations (Esposito et al., 1987; Westaway et al., 1989; Cucci et al., 1996; Gasperini et al., 1999; Pace et al., 2002; Fracassi & Milano, 2014; Bonini et al., 2016; Guidoboni et al., 2019, among several others);
- investigations of continental deposits and landforms (Bagnaia et al., 1992; Carraro & Giardino, 1992; Bertini & Bosi, 1993; Galadini & Messina, 1993; 1994; Giraudi, 1994; Vittori et al., 1995; Brancaccio et al., 1997; Corrado et al., 1997; Galadini et al., 1998; Cavinato et al., 2002; Di Bucci et al., 2002; Di Bucci et al., 2005; Falcucci et al., 2009; Boncio et al., 2010; Pizzi et al., 2010; Di Bucci et al., 2011; Gori et al., 2011; Saroli & Moro, 2012; Saroli et al., 2014, among several others); or
- paleoseismological investigations (Frezzotti & Giraudi, 1989; Giraudi, 1989; Giraudi & Frezzotti, 1995; Michetti et al., 1996; Pantosti et al., 1996; Galadini & Galli, 1999; D'Addezio et al., 2001; Galadini & Galli, 2003; Galli & Galadini, 2003; Salvi et al., 2003; Ceccaroni et al., 2009; Galli et al., 2011; Moro et al., 2013; Galli et al., 2015).

In accordance with the current seismicity and the associated focal mechanisms, all active seismogenic faults strike from NW-SE to NNW-SSE and dip toward the SW, i.e. towards the Tyrrhenian Sea, with the only exception being the Matese fault, which dips to the NE.

Our main effort was to inspect the overall coherence and completeness of all data. In view of this goal, in a few instances we discarded faults described as active in the older literature because more detailed investigations revealed their current state of inactivity (for example, the Fiamignano fault; Bosi, 1975; Galadini, 1999). Thus, the mapped traces shown in Figure 2 represent the surface expressions of seismogenic sources that, according to the current literature, are capable of M_w 5.5+ earthquakes. As several of these faults have scarce or no seismicity associated, we could not determine instrumentally their dips and their seismicity cutoff depths. We hence adopted a fixed dip of 50°, as suggested by recent central Apennines earthquakes. We also adopted a maximum seismogenic faulting depth of 15 km for our entire study area, as found for different parts of the central Apennines by Valoroso et al. (2013) and Frepoli et al. (2017). All faults are briefly described in Supporting Text S1, and shown in Figure 2 and Supporting Figure S1.

2.2. Interseismic GNSS data and processing

To gather the largest possible number of GNSS measurements we complemented the observations from the permanent stations with data from temporary geodetic networks: first, because these additional measurements are closely-spaced and purposely located around active faults; and second, because given the long time elapsed since these networks were created (nearly 20 years), they are possibly less perturbed by transients affecting the shorter time series of most permanent stations. Thus, the analyzed dataset consists of continuous data provided by permanent stations and by observations that were collected using survey-style benchmarks belonging to the CAGeoNet and SAGNet temporary networks (Anzidei et al., 2005; Anzidei et al., 2008; Sepe et al., 2009; Galvani et al., 2013).

The permanent stations belong to different networks: the International Global Navigation Satellite Systems Service (GNSS) run by the IGS [http://www.igs.org]; the Integrated National GPS Network (Rete Integrata Nazionale GPS-RING; Avallone et al., 2010); the Italian Space Agency (Agenzia Spaziale Italiana; ASI) [http://geodaf.mt.asi.it/]; Leica Geosystems (Italian Positioning Service [ItalPos] network) [https://hxgnsmartnet.com/it-it]; the Regione Abruzzo network; and the University of Perugia and University of L'Aquila networks. INGV also collects and processes GPS data from networks that do not belong to the GNSS (IGS), or to the EUREF Permanent Network (EPN), or are not in partnership with the international scientific community, thus expanding and supplementing the GPS database with crucial regional data whose historical records are often lost (Devoti et al., 2017).

The CAGeoNet and SAGNet networks have provided repeated measurements over the time interval 1999 to 2017. Whenever logistics allowed, measurements were carried out in approximately the same period of the year, so as to minimize the bias possibly arising from seasonal variations. Each station site was occupied for an average observation window of 48 h, for at least three survey sessions per station, with a sampling interval of 30 s. The measurements carried out after 2007 are generally characterized by longer data collection periods, reaching up to one or two years of continuous observations for selected stations.

GNSS measurements were arranged into 17 CGPS (continuous Italian and European GPS stations) clusters, and 2 surveys-style GPS stations clusters. Nine IGS common fiducial stations were shared among all clusters and used as anchor stations in the subsequent combinations. Each cluster was independently processed, and then least-squares combined into a single daily solution.

For the processing we used Bernese 5.0 software (Beutler, 2007), following the scheme discussed in Galvani et al. (2012). More specifically, the processing was based on the Bernese Processing Engine (BPE) procedure, that follows the standard analysis guidelines devised for all EUREF Analysis Centers [http://www.epncb.oma.be]. The daily station coordinates, along with the hourly troposphere parameters, were solved using the *a priori Dry Niell* troposphere model, with the corrections estimated by the *Wet Neill* mapping function. The ionosphere was neither estimated nor modeled as we used the L3 (ionosphere free) linear combination of L1 and L2. The *a priori* GPS orbits and the Earth orientation parameters were fixed using the precise IGS products. We applied the ocean-loading Finite Element Solution model FES2004 provided by the Ocean Tide Loading web service [http://holt.oso.chalmers.se/loading], and used the IGS absolute antenna phase-center corrections. The daily solutions were obtained in a loosely constrained reference frame; that is, all of the *a priori* station coordinates were left free to 10 m *a priori* sigma. The IGS08 absolute antenna phase center correction was applied to each station receiver antenna. These procedures generated daily, loosely-constrained solutions that are free from any *a priori* reference frame datum. The coordinates and the complete associated covariance matrices were saved in the standard Solution Independent Exchange (SINEX) format.

In order to express the GPS time series in the same reference frame, the daily solutions were first projected imposing tight internal constraints (at millimeter level); then the coordinates were transformed into the IGS realization of the ITRF2008 frame (i.e., IGb08) by a 4-parameter Helmert transformation (translations and scale factor). The regional reference frame transformation uses 45 IGb08 anchor sites located in central Europe. To avoid common translations of the entire network, the time series were readjusted through a common mode filtering procedure as proposed by Wdowinski et al. (1997).

Our inversion scheme assumes that in all interseismic stages each GNSS benchmark has a constant velocity resulting from steady tectonic strain accumulation (both elastic and permanent). Therefore, in our estimation we took a series of precautions to avoid non-tectonic signals in the time series. First, we inspected all available datasets and technical reports on the distribution of landslides and other mass-wasting phenomena, that may be a typical source of non-tectonic strain. We found no conclusive indication that any benchmark occurs on unstable slopes, Then, coseismic offsets of the 6 April 2009, L'Aquila (M_w 6.3), 24 August 2016, Amatrice (M_w 6.0), 26 October 2016, Visso (M_w 5.9), 30 October 2016, Norcia (M_w 6.5) and 18 January 2017, Campotosto (M_w 5.5) earthquakes were discarded manually from our time series. As several GNSS benchmarks included in our network have been proved to record afterslip due to L'Aquila earthquake (Devoti et al., 2012), we removed a threemonth portion of their respectiev time series. In contrast, we verified that postseismic motions from the 2016-2017, central Apennines earthquakes did not affect significantly the GNSS benchmarks comprising our network.

All GPS velocities and related uncertainties were estimated with a linear weighted least-squares fit of all the coordinate time series using the full daily covariance matrix. Each time series was modeled by removing the constant drift (velocity), the annual sinusoidal components, and any occasional offsets due to changes in the station equipment between subsequent measurement windows (Figure 1c). We kept the Eurasia plate fixed by minimizing the horizontal velocities of 24 stations located on its stable portion. The selection of fully Eurasian stations was statistically inferred using the $\chi 2$ test-statistic to select the subset of stations that defined the stable plate in ITRF2008 (Nocquet et al., 2001). The estimated Euler pole coordinates and rotation rate for this motion of stable Eurasia plate were fixed at 55.85°N, 95.72°W and 0.266° ± 0.003° Ma^{-1}, respectively. The resulting horizontal velocities for 227 stations (152 in our study area) with respect to an Eurasia-fixed reference frame are shown in Figure 3 and reported in Supplementary Table S1.

2.3. Stress direction data

Carafa and Bird (2016) found that joint inversions of GNSS measurements and SH_{max} orientations return long-term deformation models that fit better the available focal mechanisms of Apennines earthquakes. This means that stress data contain information on the brittle part of the crust which may be used to complement geodetic observations. For this reason, we built our dataset interpolating the Italian Present-day Stress Indicators database (IPSI: Montone & Mariucci, 2016; Mariucci & Montone, 2019; Figure 3), available at http://ipsi.rm.ingv.it/ (last access 31st May 2019). This georeferenced database is constantly updated and contains maximum (SH_{max}, most-compressive) and minimum (Sh_{min}, least-compressive) horizontal stress directions collected in the broader central-Mediterranean area. We selected data ranked with quality from A to C, neglecting D- and E- quality data. The resulting dataset for central Apennines is formed by borehole breakouts and formal inversion of earthquake data. The breakouts are enlargements of the walls of the wellbore, free surfaces that are unable to sustain shear traction, caused by the stress concentration due to far-field forces. The seismicity data are obtained from inversions of P-, B-, and T- axes of diffuse seismicity (Figure 4).

Given the uneven sampling and scatter in SH_{max} orientations, an interpolation was required to know the stress direction and its uncertainty for any point of the study area. We followed the well-established interpolation scheme proposed by Bird and Li (1996), expanded in Carafa and Barba (2013) and Carafa et al. (2015b), and implemented in SHINE, a publicly accessible web-based application http://shine.rm.ingv.it/index.phtml (last accessed on 29 March 2020). The resulting dataset is presented in Supporting Figure S2. The reciprocal distance and large scatter of available data did not allow us to solve for stress directions along the Tyrrhenian (western) side of our model; notwithstanding this limitation, we have been able to estimate the SH_{max} azimuth for the 75% of our study area (Supporting Figure S3).

3. Determining fault slip rates from geodetic velocities, SH_{max} orientations and off-fault stiffness.

Most normal faults in the Apennines are expected slip at rates around or below 1 mm/a, and most campaign GPS velocities exhibit comparable uncertainties; thus the signal/noise ratio is expected to be rather large, such that complex modeling is required to extract reasonable estimates of long-term fault slip rates even at signal/noise ratio of order unity.

In the following we focus on an inversion scheme that optimizes the available datasets, whereas all the algebra needed for recovering the long-term strain rates and fault slip rates by formal inversion is fully explained in Supporting Text S2. This inversion scheme has been implemented in Version 5.3 of NeoKinema software (Bird, 2009; Bird & Carafa, 2016), and its numerical stability has been recently tested in a similar geographic setting by Carafa et al. (2018). To keep the algebra of our inverse method simple, we assumed that the lateral variation of long-term horizontal surface velocities adequately describes the partitioning among bulk strain rates and fault slip rates. This implies that, excluding active fault zones, any unfaulted volume of the lithosphere has a null vertical gradient of horizontal velocities. This approximation has been repeatedly shown to be adequate for reproducing the observed Holocene geodynamics of the broader Mediterranean region (Barba et al., 2008; Howe & Bird, 2010; Carafa & Barba, 2011; Carafa et al., 2015a; Splendore et al., 2015).

Assuming that it is only necessary to estimate the horizontal components of the long-term average velocity, we can achieve a solution for our study area by finite element modeling, subdividing the whole study area into contiguous, non-overlapping spherical triangles (elements) delimited by three vertices (nodes) [see Supporting Figure S3]. We discretized our study area with 2,567 nodes (and 4,956 triangular elements) whose southward v and eastward w velocity components are the unknowns to be determined.

In our model we consider:

1) the active-fault database consisting of Z=18 faults (fully described in Section 2.1 in terms of location and geometry) and their r_z (z=1,…,18) fault offset rates, with the associated uncertainty ;

2) the r_k (k= 1,…,304) horizontal component of the long-term velocity derived from the elastic correction of the interseismic velocities at the 152 GNSS benchmarks (see Section 2.2);

3) the related geodetic covariance matrix $\tilde{C}$;

4) N=3 assumptions on the stiffness (n=1) and direction (n=2,3) of the long-term deformation of unfaulted lithosphere (explained in Section 3.3).

To determine the long-term horizontal velocities at the 2,567 nodes we must solve a 5,134x5,134 linear system, that can be written as

$$\begin{bmatrix} A_{ij} & B_{ij} \\ C_{ij} & D_{ij} \end{bmatrix} \begin{bmatrix} v_j \\ w_j \end{bmatrix} = \begin{bmatrix} E_i \\ F_i \end{bmatrix} \quad (1)$$

where

$$A_{ij} = \frac{1}{2}\sum_{h=1}^{304}\sum_{l=1}^{304} N_{hl} f_{hi} f_{lj} + \frac{1}{L_0}\sum_{z=1}^{18}\frac{f_{zi} f_{zj}}{\sigma_z^2}dl + \frac{1}{A_0}\sum_{n=1}^{3}\iint_{area}\frac{f_{ni} f_{nj}}{\sigma_n^2}da \quad (2a)$$

$$B_{ij} = \frac{1}{2}\sum_{h=1}^{304}\sum_{l=1}^{304} N_{hl} f_{hi} g_{lj} + \frac{1}{L_0}\sum_{z=1}^{18}\frac{f_{zi} g_{zj}}{\sigma_z^2}dl + \frac{1}{A_0}\sum_{n=1}^{3}\iint_{area}\frac{f_{ni} g_{nj}}{\sigma_n^2}da \quad (2b)$$

$$C_{ij} = \frac{1}{2}\sum_{h=1}^{304}\sum_{l=1}^{304} N_{hl} g_{hi} f_{lj} + \frac{1}{L_0}\sum_{z=1}^{18}\frac{g_{zi} f_{zj}}{\sigma_z^2}dl + \frac{1}{A_0}\sum_{n=1}^{3}\iint_{area}\frac{g_{ni} f_{nj}}{\sigma_n^2}da \quad (2c)$$

$$D_{ij} = \frac{1}{2}\sum_{h=1}^{304}\sum_{l=1}^{304} N_{hl} g_{hi} g_{lj} + \frac{1}{L_0}\sum_{z=1}^{18}\frac{g_{zi} g_{zj}}{\sigma_z^2}dl + \frac{1}{A_0}\sum_{n=1}^{3}\iint_{area}\frac{g_{ni} g_{nj}}{\sigma_n^2}da \quad (2d)$$

and the forcing-vector components are

$$E_i = \frac{1}{2}\sum_{h=1}^{304}\frac{f_{hi}r_k}{\sigma_k^2} + \frac{1}{L_0}\sum_{z=1}^{18}\frac{f_{zi}r_z}{\sigma_z^2}dl + \frac{1}{A_0}\sum_{n=1}^{3}\iint_{area}\frac{f_{ni}r_n}{\sigma_n^2}da \qquad (3a)$$

$$F_i = \frac{1}{2}\sum_{h=1}^{304}\frac{g_{hi}r_k}{\sigma_k^2} + \frac{1}{L_0}\sum_{z=1}^{18}\frac{g_{zi}r_z}{\sigma_z^2}dl + \frac{1}{A_0}\sum_{n=1}^{3}\iint_{area}\frac{g_{ni}r_n}{\sigma_n^2}da \qquad (3b)$$

where j,i=1,…,2,567 nodes; $\tilde{N} \equiv \tilde{C}^{-1}$ is the normal matrix. The spatial nodal coefficients f_{zi}, g_{zi} link the long-term nodal velocity components to the fault position and dip-direction are, f_{hi}, g_{hi}, f_{lj}, g_{lj} to the two long-term components of all GNSS benchmarks, and f_{ni}, g_{ni} to the model area. Finally, L_0 and A_0 are user-selected dimensional tuning parameters for equalizing the fit to all available datasets.

3.1. Constraints from active fault geometries

As stated in the Introduction, we argue that some critical issues must be faced before using field-derived geological offset-rate estimates r_z; therefore, we neglected them in our inversions. In practice, we determined that setting σ_z =3.0 mm/a and L_0=800,000 m in Equations 2 and 3 for all faults guarantees that any prior geologic slip-rate estimate r_z has been effectively ignored in the inversion. This means that the only tuning parameter to be explored in our inversion is A_0; larger values of this reference area imply that the GNSS measurements (corrected with the quasistatic elastic assumption) were closely fit. Conversely, a low A_0 means that the fit to GNSS measurements has been partly compromised to achieve a stiff-microplate, stress direction-fitting solution. Thus, in our inversion scheme, fault geometries and positions were used to determine fault slip rates from final long-term nodal velocities with the Lagrange multiplier method.

3.2. From the interseismic to the long-term time scale: geodetic constraints and quasistatic elastic correction

The targets of our modeling approach are the slip rates and bulk deformation rates in the long-term-average (or “tectonic”) timescale. Instead, in an interseismic time interval, which is when most GPS velocities are collected, we regard the strain rates as a sum of a permanent-strain process (possibly long-term-averaged), and a temporary elastic strain rate caused by temporary locking of nearby seismogenic fault patches in the upper crust. Thus, interseismic velocities at the GNSS benchmarks need to be adjusted to long-term-average velocities, which are the target of Equation (1). In our code this adjustment is done iteratively with the classical quasistatic elastic approach (see Figure 1), assuming a Poisson ratio of 0.25 and neglecting gravity, because it deals with stress changes as perturbations to the absolute state of stress. The main assumption of this approach is that the seismic cycle is divided into two main stages: the first includes the coseismic displacement and the rapid postseismic deformation; the second is the secular interseismic deformation. In this latter stage the deformation is time-invariant but spatially variable with respect to the position of the fully-locked active fault (see Figure 1). Thus, in the quasistatic elastic approach, the yearly-averaged displacement is the long-term slip rate, determined by iteration for each fault with common starting values for our experiment being r_z=1.0 mm/a (σ_z=3.0 mm/a). To be consistent with the quasistatic elastic approach, as explained in Section 2.2, our estimation of the secular trend at each GNSS benchmark consists of a linear weighted least-squares fit of all the coordinate time series; this trend was determined by removing the constant drift velocity, any annual sinusoidal components, and any seismic and occasional non-seismic offsets (See Figure 1).

3.3. Microplate assumption and horizontal-stress principal-axes constraints

Although our inversion does not explicitly contain any flow laws, the stiffness of the continuum and its long-term deformation must be somehow taken into account and modeled. The unfaulted portions of the lithosphere are those where the geologists did not recognize any evidence of active faulting, implying that minimal plastic deformation is expected to occur in these volumes. However, some young or blind active faults may still remain undetected, but their interseismic stages could be eventually recorded in GNSS time series. Additionally, some kind of long-term viscous deformation, though possibly slow, is expected in the lithosphere. Thus, in all unfaulted elements we approximate the lithosphere as an isotropic Newtonian viscous body with no internal deformation, but with an intrinsic standard deviation μ of the nominally zero long-term strain rate to be determined. This formalism is usually referred to as *the microplate assumption*. In detail, we determined the parameter μ self-consistently by iteration, setting it to be the regional RMS average of the modeled long-term strain rate in unfaulted elements. The microplate assumption is implemented in our inversion scheme as the n=1 in Equations 2 and 3; it can be seen as a realistic departure from the rigid-microplate assumption in regions of diffuse deformation or blind faulting.

In the long-term-average (or “tectonic”) timescale, which relates to long-term fault slip rates and other permanent strain markers (such as stylolites), we regard the accumulation of strain as an irreversible process in an isotropic crustal medium. Thus, the principal axes of the long-term-average strain rate should align with the principal axes of stress. For this reason, using the SHmax azimuths is desirable in the estimation of consistent long-term horizontal velocities. Under the previously stated assumption of an isotropic and Newtonian-viscous lithosphere, the SH_{max} dataset allows constraining the orientations of long-term (non-elastic) horizontal strain rates. In our inversion we used these constraints in each unfaulted element, imposing the most-compressive principal value of the 2D strain rate tensor to be approximately parallel to the SH_{max}. This target implies that the shear strain rate (off-diagonal entries in the 2D strain rate tensor) in principal-axis coordinates has to be minimized. In Supporting Text S2 we show how we implemented numerically this constraint in our code.

4. Results

Tested values

We generated a series of kinematic models with varying weight-factor A_0 (see Equations 2 and 3) in the range $10\text{-}40\cdot10^7$ m^2, at $1\cdot10^7$ m^2 steps. From the resulting long-term average horizontal velocities we can determine the bulk deformation rates for unfaulted lithosphere and the slip rates of active faults. Recalling Equation (1), models with larger values of A_0 fit GPS data better, while the microplate stiffness and any stress-direction constraints are secondary. However, models with a smaller A_0 approximate a stiff-microplate, thus the model fits the stress data better than the GPS measurements. As we assumed that the model parameter μ is set to be the regional RMS average of the modeled long-term strain-rate in unfaulted elements, no other subjective choices were needed. For models with A_0 <$21\cdot10^7$ m^2 we could not satisfy the assumption that μ equals the regional RMS and obtain a stable solution. These models put too much weight on stress data and continuum stiffness, such that they are unable to give a sufficiently-good fit to short-term velocity constraints. For this reason, these models were discarded from all following analyses.

Figure 5 shows the performance of remaining models with respect to GPS measurements and SH_{max} data by plotting the two dimensionless prediction misfits (stress direction error and geodetic error) of each model. In detail, the dimensionless scalar stress direction error is defined as

$$N_2^{\text{stress}} = \sqrt{\sum_{i=1}^{E} A_i \frac{(p_i - r_i)^2}{\sigma_i^2} \Big/ \sum_{i=1}^{E} A_i} \qquad (4)$$

where p_i is the azimuth of the most-compressive horizontal principal strain rate axes in each of the E finite elements, r_i is the corresponding target azimuth of the most-compressive horizontal principal stress derived by interpolating the SH_{max} from the IPSI dataset, A_i is the area of those elements, and σ_i is the corresponding standard deviation from stress-direction interpolation. Figure S4 shows the misfit between observed and modelled SH_{max}.

The second scalar misfit is the geodetic error, defined as:

$$N_2^{\text{GPS}} = \sqrt{\frac{1}{M} \sum_{i=1}^{M} (p_i - r_i) \sum_{j=1}^{M} \left[\tilde{N}\right]_{ij} (p_j - r_j)}\,. \qquad (5)$$

where p_i and p_j are the short-term horizontal velocities derived from long-term velocities by subtracting the elastic solutions driven by model fault slip rates, r_i and r_j are the geodetic horizontal velocities, and $\tilde{N} \equiv \tilde{C}^{-1}$ is the normal matrix, with $\tilde{C}$ being the geodetic velocity-covariance matrix. Figure S5 shows the unfit portion of geodetic measurements.

Figure 5 confirms that the selection of A_0 has a strong influence on both N_2^{GPS} and N_2^{stress}. Models that exhibit a very large N_2^{stress} or N_2^{GPS} must be discarded, even if a numerical solution is available, because they do not match closely enough the available geodetic and stress constraints. Thus, we selected models with A_0 =$23\cdot10^7$ m^2 and A_0 =$24\cdot10^7$ m^2 as those reproducing most effectively the models for long-term velocities in the central Apennines. These two models show similar misfits, but the first exhibits a slight preference for SH_{max}, while the second performs better with geodetic data.

Figure 6 illustrates the model performance along the four selected profiles shown in Figure 2. Each panel shows the amount of bulk deformation in green and fault slip rate in red (top diagram), the short-term horizontal velocities with respect to our GPS solution (middle diagram), and the most-compressive horizontal stress azimuth with respect to the IPSI database (bottom diagram).

4.1. Bulk permanent deformation

Some diffuse, possibly aseismic, permanent long-term deformation is expected in the lithosphere, especially in a region of widespread deformation such as the central Apennines. However, the resulting long-term strain rate map, shown in Figure S6, is a mix of this possibly dominant component of long-term viscous/aseismic deformation with artifacts due to inaccurate fault connectivity at depth, faults missing in the database, and unmodeled short-term geodetic transients.

First-order structural and geometrical complexities such as sharp bends, intersections with cross structures and overlapping en-echelon arrangements are invoked to describe properly each tip of an active fault; nevertheless, their long-term interaction remains much more complex, as clearly shown by the Amatrice and Norcia earthquakes. Thus, it is not surprising that the largest modeled strain rates concentrate at fault tips or in gaps between faults. So far, it is unclear whether these strain rate peaks represent viscous/aseismic long-term deformation, or imply continuity at depth of adjacent faults that do not appear to be connected the surface, because at their tips faults exhibit only a few short, discontinuous earthquake ruptures. However, they do represent long-term features.

Large long-term strain rates may also suggest a weakness in our fault database, likely a missing fault. In this specific case, not applying the quasistatic elastic correction may results in an incomplete correction of short-term interseismic deformation rates to long-term ones. This could be the case of the under-investigated area surrounded by the Barrea, Aremogna-Cinque Miglia and Matese faults (see the large green area in Profile 3, Figure 6), currently undergoing rather fast permanent bulk deformation. Geomorphic conditions such as limited bedrock erodability and lack of youthful faulted continental deposits may hinder the identification and characterization of possible active faults in this region, if any.

In two areas, shown in Figure S6 with the letter “T”, the bull-eye patterns of strain rate peaks are spatially correlated with recently-documented short-term transients. According to Di Luccio et al. (2018) and Esposito et al. (2020), the southernmost peak is spatially consistent with the intrusion of a dike-like body surrounded by deep earthquakes showing tensile-shear ruptures triggered by a fluid overpressure episode. The northernmost peak is likely related to the horizontal oscillation of GNSS station positions in the L'Aquila intermountain basin due to water level changes (Devoti et al, 2018) possibly coupled with residual postseismic effects of 2009 L’Aquila earthquake. While the first peak does not dramatically affect any active fault estimation, the second one affects an area crossed by the Paganica and Campo Felice faults; for this reason we decided not to report their long-term estimates. This task is postponed to a time when geodetic observations will be more likely to be free from any short-term contamination, or when a coherent physical model can be implemented in the covariance matrix, as suggested in Bird and Carafa (2016).

4.2 Fault slip rate estimates

In each model we used the whole fault database (for completeness all results are reported in Supporting Table S2), but we show and discuss only the slip rate estimates we obtained for the seven fault zones that are adequately illuminated by the GNSS network (Fucino, Aquae Iuliae, Matese, Middle Aterno Valley, Upper Sangro Valley- Mt. Marsicano, Barrea, Liri Valley). Only for these faults we confident that the interseismic signal is adequately sampled by a sufficient number of benchmarks, both in the hanging-wall and in the footwall.

Figure 5 shows the slip rate variability for these seven faults as a function of the selected A_0. The fastest-slipping fault is the Fucino, which shows also the largest variability in the tested range of A_0. Keeping in mind that the slip rates resulting from models with A_0 > $24{\cdot}10^7$ m^2 are essentially unconstrained by SH_{max} data, our preferred slip rate interval is 1.8-1.9 mm/a. The Matese fault also shows large slip rates, the preferred range being 1.1-1.2 mm/a. All remaining faults exhibit smaller slip rates - consistently below 1.0 mm/a - taht remain rather stable over the whole interval of tested A_0.

As stated earlier, in all our inversions we fixed the fault dip (50°) and locking depth (15 km). Some variability needs to be considered, however, as we expect that different dips and locking depths have an impact on slip rate estimates. This may happen in two ways: (1) a different fault dip, or locking depth, affects the conversion of interseismic to long-term geodetic velocities around the fault, possibly causing the global solution to assign a different fault heave-rate; and (2) the conversion of fault heave rate to slip rate is a function of the dip angle. To this end we selected a preferred model ($A_0 = 23 \cdot 10^7$ m^2) and perturbed it by testing different dip angles (from 40° to 60° at 5° steps) and locking depths (13-17 km at 1 km step). Supporting Figure S7 shows that the slip rate is most sensitive to changes in fault dip, with effects up to 0.5 mm/a for Fucino fault (less for other faults). Conversely, the influence of the locking depth on slip rates is negligible.

The estimates of fault rake we obtained show a consistent pattern: with the exception of the Middle Aterno Valley and Liri Valley fault, the remaining five faults exhibit a small, but consistent, left-lateral component (Table 1). This pattern seems more pronounced for the southern faults (Matese and Aquae Iuliae) and for the Fucino fault.

Table 1 shows the preferred range of slip rates we obtained by combining perturbations on dip and locking depth with the results of our two preferred models ($A_0 = 23 \cdot 10^7$ m^2 and $A_0 = 24 \cdot 10^7$ m^2).

5. Model fault slip rates vs. geologic/geochemical slip rates

For the seven faults that are adequately illuminated by the GNSS network we performed a cross-check between two datasets that are fully independent of each other: model-based slip rates, and slip rates (often throw rates) determined with standard geological methods (analyses of geological piercing points, paleoseismology, topographic profiles, cosmogenic nuclides). We knew that the check could give rise to two opposite scenarios: (a) the range of slip rates determined in our work overlaps the range of geologic rates, suggesting that the two sets of estimates are compatible and overall reliable; or (b) some geological rates are incompatible with the corresponding model-based estimates, suggesting that one of the two rates is biased, at least for the specific fault under scrutiny. In the following we discuss the results of the cross-check on a fault-by-fault basis.

5.1. Fucino fault system

In our model we assume that all surface expressions of this large fault system (Magnola, Marsicana Highway, S. Benedetto dei Marsi-Gioia dei Marsi, and Trasacco faults), that were identified and studied by different research groups (Galadini & Galli, 1999; Roberts & Michetti, 2004; Saroli et al., 2008; Gori et al., 2017), root into a single large seismogenic fault, as suggested by other investigators (Roberts and Michetti 2004; Roberts & Michetti, 2004; Lavecchia et al., 2006; DISS Working Group, 2015 among others). As such they could be modeled as a single system referred to as the Fucino fault system. This system is responsible for the 13 January 1915 earthquake (M_w 7.1, Rovida et al., 2016), the largest reported earthquake in central Italy, and its evolution is reshaping the Fucino Plain, a pre-existing intermountain basin, the largest in the central Apennines. In our model this fault system exhibits the largest slip rate, although the associated uncertainty is rather large (1.5-2.4 mm/a, see Profile 4 in Figure 5), with a left-lateral component of 0.3-0.5 mm/a. Such pronounced uncertainty is partly due to the possible variability of the dip angle and of the locking depth (as shown in Supporting Figure S7), and partly to the large uncertainties of geodetic measurements in the Fucino Plain itself (see Profile 2 in Figure 5).

As for the existing geological slip rate estimates, different analyses have been performed along all surface expressions of the Fucino fault system. Galadini and Galli (1999) determined throw rates of 0.4-0.5 mm/a for the Marsicana Highway fault, 0.24-0.29 mm/a (minimum throw rate) for the San Benedetto dei Marsi-Gioia dei Marsi fault, and 0.27-0.29 mm/a (minimum throw rate) for the Trasacco fault. The estimates sum up to a cumulative minimum throw rate of 1.0-1.3 mm/a and a total slip rate of 1.3-1.7 mm/a (assuming no strike-slip component). Morewood and Roberts (2000) estimated a maximum slip rate of 2.13 mm/a for the whole Fucino fault system, with a maximum horizontal extension rate of 1.60 mm/a. Similarly, Roberts and Michetti (2004) reported a number of throw rates up to 2.00 mm/a. In both these latter works the largest throw rates were obtained by summing the offset rates obtained from trenches dug across the various splays (or based on the thickness of Quaternary sediments in the hanging wall) with the rates obtained from fault scarps. All these geological slip rate estimates fall within the interval determined in our model, with a slight preference for the lower part of the range (see profile 2 of Figure 5, Table 1 and Figure 7).

5.2. Aquae Iuliae and Matese fault systems

For the Aquae Iuliae fault our model yields a slip rate of 0.2-0.3 mm/a. This estimate is lower and rather inconsistent with respect to the short-term slip rate of 1.5-1.9 mm/a proposed by Galli and Naso (2009) for the past 2,000 years based on archeoseismological analyses. Nevertheless, our range is almost compatible with the lower end of the medium-term slip rate reported by the same investigators (0.45-1.00 mm/a) and similar to their long-term slip rate of 0.50 mm/a, determined by analyzing the dislocation of the Volturno river terraces. On the basis of similar data, Cinque et al. (2000) reported a long-term throw rate of 0.2-0.4 mm/a, which fits our estimates (see Table 1 and Figure 7). Given all of this evidence, the main issue affecting the slip rate estimates obtained for the Aquae Iuliae faults is their large variability for different time spans. Hence, quantifying the difference between geodetic and geologic estimates remains pointless until new and more constrained geological analyses are performed.

The model-based slip rates we obtained for the Matese fault system are larger, making it the second-fastest fault system of the entire set, after Fucino. Various investigators reported rather large slip rates for it: active extension at rates above 1.0 mm/a has been reported by Ferranti et al. (2014), whereas Ferrarini et al. (2017) obtained post-Last Glacial Maximum (LGM) throw rates in the range 0.9-1.8 mm/a by analyzing the topographic profiles of fault scarps. Similar analyses were performed in the southern portion of the Matese fault system by Guerrieri et al. (1999), who found slip rates in the range 0.7-1.0 mm/a. These rates differ significantly from the 0.1-0.5 mm/a range reported by Cinque et al. (2000) for the same time interval; a similarly low range was inferred from scarp profiles reported in Faure Walker et al. (2012). Such large variability of slip rate estimates among different studies using topographic profiles (but focusing on different localities) clearly reveals the limitations of this approach. Our own slip rate estimates fall in the range 1.0-1.6 mm/a, with a significant left-lateral component of ca. 0.4 mm/a. Thus, model slip rates for the Matese fault system are comparable with the estimates proposed by Guerrieri et al. (1999) and Ferrarini et al. (2017). Other geological estimates were obtained by other methods: for instance, Galli and Galadini (2003) reported a slip rate of 0.9 mm/a by means of paleoseismic investigations, and a throw rate of 1.0 mm/a constrained by the 300 m vertical offset of 300 ka-old tephra (see Table 1 and Figure 7). Also in this case the geological and geodetic rates are compatible, and hence presumably reliable: this conclusion casts serious doubts on the estimates supplied by Cinque et al. (2000) and Faure Walker et al. (2012), who are likely to have sampled and measured non-tectonic dislocations.

5.3. Middle Aterno Valley, Upper Sangro Valley –Mt. Marsicano, Barrea, Liri Valley fault system

Different geological slip rate estimates are available for the Middle Aterno Valley fault zone. Based on the vertical offset of Quaternary units reported by Bertini and Bosi (1993), Galadini and Galli (2000) calculated a long-term throw-rate of 0.33-0.43 mm/a. In their turn, Falcucci et al. (2015) proposed a slip rate in the range 0.23-0.34 mm/a based on the displacement of 1.0 Ma-old breccias. Messina et al. (2009) inferred a slip rate of 0.4 mm/a from the different elevation of the Middle Pleistocene alluvial sequence with respect to the present valley floor. The slip rates obtained by our models fall in the range 0.2-0.4 mm/a with almost pure dip-slip rakes (see Table 1 and Figure 7). Thus, also for the Middle Aterno Valley fault system, geodetic data and long-term slip rate estimates seem compatible, and hence most likely reliable.

No geologic slip rates are available for the Barrea fault system: neither from the dislocation of surface features, nor from trenching. Thus, there is no opportunity to compare our model-based estimates (0.3-0.4 mm/a, see Table 1) with actual geological observations. A careful inspection of the geodetic network, however, shows that the number of benchmarks falling in the eastern portion of this area is rather limited (see Profile 3 in Figure 5), implying that we need caution in considering these geodetic estimates fully reliable. We conclude that more efforts are required by both geophysicists and geodesists to increase the knowledge on the active tectonics of this area.

For the Upper Sangro Valley-Mt. Marsicano fault we found a slip rate in the range 0.4-0.7 mm/a, with a minor left-lateral component (0.1 mm/a). These values are in good agreement with the 0.16-0.83 mm/a estimate reported in Roberts and Michetti (2004), and with the 0.51 estimate mm/a reported in Papanikolaou et al. (2005). According to Galadini and Galli (2000), this fault system is responsible for the displacement of alluvial deposits related to the later part of the Early Pleistocene (0.9-1.0 Ma), whereas Galadini and Messina (1993) highlighted the existence of both dip-slip and strike-slip displacements, probably as a result of strain-partitioning phenomena. As for the slip rates, Galadini et al. (1998) calculated 0.17-0.21 mm/a and 0.14-0.18 mm/a for the dip-slip and left-lateral components, respectively (see Table 1 and Figure 7). Our estimates are larger than those reported in Galadini et al. (1998); however, these investigators stated that their estimates are minimum figures because other fault branches are thought to be active within the same zone. Also in this case, and similarly to the Matese fault, the large variability that exists among different estimates makes it hard to quantify the misfit between model-based and geological slip rates. Nevertheless, we remark that our model-based slip rates consistently fall within the range of geological slip rates.

Interestingly, we found a really low slip rate for the Liri Valley fault (perhaps <0.2 mm/a, if we consider the variability arising from the choice of a different dip and locking depth). This finding is in strong disagreement with rates up to 1.11 mm/a rate reported by Roberts and Michetti (2004), who used the characteristics of bedrock fault scarps to infer post-LGM activity. We recall that this fault places the Mesozoic carbonate bedrock in contact with Miocene clayey-arenaceus flysch without involving Late Quaternary continental deposits; for this reason, the recent activity of this fault is largely debated (see Supporting Text S1). In fact, according to Galadini (1999) the formation of these bedrock scarps may have been controlled by erosional processes, overemphasizing a feeble tectonic signal or simply mimicking it. Given our results, we definitively opt for this interpretation.

5.4. Left-lateral component

An interesting result of our model is the marginal but consistent left-lateral component that we found for five out of seven faults. The perpendicular component of the heave rate (pure extension) is always much greater than the along-strike component, implying that this left-lateral component has a minor effect on rake, which never exceeds -70° (Figure 7 and Table 1). Such a small left-lateral component is not a novelty in the broader extensional geodynamics of central Apennines , having been reported both for the Upper Sangro Valley-Mt. Marsicano fault (Galadini & Messina, 1993) and for the Fucino fault (Giraudi, 1989; Galadini & Giuliani, 1995). The focal mechanism of the 7 May 1984 earthquake (M_W 5.9), that is reported in the literature to have been caused by the Barrea fault, shows a slight left-lateral component, an observation confirmed by geological data collected by Pace et al. (2002). Ferrarini et al. (2017) observed fault striae for the Matese fault with frequency peaks aligned N030°-N045° and N005°-N015°; altogether these findings support the hypothesis of dip-slip to oblique (normal-left lateral) kinematics for the Matese fault. We carefully inspected the rake variability for all these faults with respect to the tuning parameter A_0 (Figure 5), and found that the detected left-lateral component is not an artifact due to numerical issues of our code or to a subjective choice of the best A_0. Rather, it arises from present-day stress-directions and regional tectonic setting, affected as they are by a combination of lithostatic pressure (gravitational potential energy), lithosphere-asthenosphere interaction (basal tractions), and side strength resistance to plate movements (Carafa et al., 2015a).

6. Conclusions

In this experiment we inverted simultaneously both horizontal interseismic geodetic velocities and SH_{max} azimuths to obtain long-term velocities and slip rates for a set of 18 large central Apennines active faults. As the uncertainties associated with fault slip rates are comparable to those characterizing most campaign GPS velocities, simple 2D or 3D back slip models cannot be used. We have proved that reasonable estimates of long-term fault slip rates can be extracted even at signal/noise ratios of order unity using a more sophisticated modeling approach. For well sampled faults, the slip rate estimates fit convincingly well the corresponding geological estimates, leading us to conclude that they can be considered for seismic hazard models also in regions such as the Apennines. We remark that geodetic and geological data can be used together to highlight (and possibly model) both the likely occurrence of short-term transients in GPS time series, or the existence of non-tectonic processes contributing to the progressive exposure of presumed active faults. Differently, we report large permanent bulk strain rates for several parts of the study area. Given the sparseness of GNSS stations and the difficulties in mapping active faults, we cannot be conclusive in assigning these permanent strain rates to missing, possibly blind faults, especially in the easternmost part of the central Apennines in the understudied region surrounded by Barrea, Aremogna-Cinque Miglia and Matese faults.

Finally, we wish to point out that ours is the first fully-reproducible experiment investigating how extension is partitioned among slip on different faults and bulk lithosphere strain. A denser geodetic network and longer time series will definitely improve the overall performance of our modeling technique; nevertheless, significant improvements are required also in the understanding of the surface manifestation of active faulting. We are well aware of several ongoing disputes (briefly reported in Supporting Text S1) on the active faults of the central Apennines and, as a consequence, we are forced to regard our fault database as neither exclusive nor exhaustive, a characteristic clearly shared with any other available database. As such, our fault slip rate estimates are burdened by several limitations affecting both geodetic and geologic data. Nevertheless, the strongest point of our technique rests in its ability to perform a complete seismogenic characterization which preserves the overall compatibility and coherence between regional kinematics (as described by geodetic velocities) and local geometries (as illustrated by field geology).

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References

Anzidei, M., Baldi, P., & Serpelloni, E. (2008), The coseismic ground deformations of the 1997 Umbria-Marche earthquakes: a lesson for the development of new GPS networks, 2008, 51(2-3), doi: 10.4401/ag-3029.

Anzidei, M., Baldi, P., Pesci, A., Esposito, A., Galvani, A., Loddo, F., & Cristofoletti, P. (2005), Geodetic deformation Across the Central Apennines from GPS Data in the time span 1999-2003, 2005, 48(2), 13, doi: 10.4401/ag-4405.

Avallone, A., et al. (2010), The RING network: improvement of a GPS velocity field in the central Mediterranean, 2010, 53(2), 16, doi: 10.4401/ag-4549.

Bagnaia, R., D'Epifanio, A., & Sylos Labini, S. (1992), Aquila and Subequan basins: an example of Quaternary evolution in central apennines, Italy, Quaternaria Nova, 2, 187-209.

Barba, S., Carafa, M. M. C., & Boschi, E. (2008), Experimental evidence for mantle drag in the Mediterranean, Geophysical Research Letters, 35(6), 6, doi: 10.1029/2008gl033281.

Basili, R., Valensise, G., Vannoli, P., Burrato, P., Fracassi, U., Mariano, S., Tiberti, M. M., & Boschi, E. (2008), The Database of Individual Seismogenic Sources (DISS), version 3: Summarizing 20 years of research on Italy's earthquake geology, Tectonophysics, 453(1-4), 20-43, doi: 10.1016/j.tecto.2007.04.014.

Bertini, T., & Bosi, C. (1993), La tettonica quaternaria della conca di Fossa (L'Aquila), Il Quaternario, 6, 293-314.

Beutler, G. (2007), Bernese GPS Software*Rep.*, Astronomical Institute, University of Bern, Bern.

Bird, P. (2009), Long-term fault slip rates, distributed deformation rates, and forecast of seismicity in the western United States from joint fitting of community geologic, geodetic, and stress direction data sets, Journal of Geophysical Research: Solid Earth, 114(B11), doi: 10.1029/2009jb006317.

Bird, P., & Carafa, M. M. C. (2016), Improving deformation models by discounting transient signals in geodetic data: 1. Concept and synthetic examples, Journal of Geophysical Research: Solid Earth, 121(7), 5538-5556, doi: 10.1002/2016jb013056.

Bird, P., & Li, Y. (1996), Interpolation of principal stress directions by nonparametric statistics: Global maps with confidence limits, J Geophys Res-Sol Ea, 101(B3), 5435-5443, doi: Doi 10.1029/95jb03731.

Boncio, P., Pizzi, A., & Brozzetti, F. (2010), Coseismic ground deformation of the 6 April 2009 L'Aquila earthquake (central Italy, Mw6. 3), Geophysical Research …, 37, 2-7, doi: 10.1029/2010GL042807.

Bonini, L., et al. (2016), Imaging the tectonic framework of the 24 August 2016, Amatrice (central Italy) earthquake sequence: new roles for old players?, Annals of Geophysics, 59(Fast Track 5), 1-10, doi: 10.4401/ag-7229.

Bosi, C. (1975), Osservazioni preliminari su faglie probabilmente attive nell'Appennino Centrale, Bollettino Societa' Geologica Italiana, 94(3), 827-859.

Brancaccio, L., Cinque, A., Di Crescenzo, G., Santangelo, N., & Scarciglia, F. (1997), Alcune osservazioni sulla tettonica quaternaria nell'Alta Valle del F. Volturno, Il Quaternario. Italian Journal of Quaternary Sciences, 10(2), 321-328.

Carafa, M. M. C., & Barba, S. (2011), Determining rheology from deformation data: The case of central Italy, Tectonics, 30(2), doi: 10.1029/2010tc002680.

Carafa, M. M. C., & Barba, S. (2013), The stress field in Europe: optimal orientations with confidence limits, Geophysical Journal International, 193(2), 531-548, doi: 10.1093/gji/ggt024.

Carafa, M. M. C., & Bird, P. (2016), Improving deformation models by discounting transient signals in geodetic data: 2. Geodetic data, stress directions, and long-term strain rates in Italy, Journal of Geophysical Research: Solid Earth, 121(7), 5557-5575, doi: 10.1002/2016jb013038.

Carafa, M. M. C., Kastelic, V., Bird, P., Maesano, F. E., & Valensise, G. (2018), A "Geodetic Gap" in the Calabrian Arc: Evidence for a Locked Subduction Megathrust?, Geophysical Research Letters, 45(4), 1794-1804, doi: 10.1002/2017gl076554.

Carafa, M. M. C., Tarabusi, G., & Kastelic, V. (2015b), SHINE: Web application for determining the horizontal stress orientation, Computers & Geosciences, 74, 39-49, doi: 10.1016/j.cageo.2014.10.001.

Carafa, M., Barba, S., & Bird, P. (2015a), Neotectonics and long-term seismicity in Europe and the Mediterranean region, J Geophys Res-Sol Ea, 120(7), 5311-5342, doi: 10.1002/2014jb011751.

Carraro, F., & Giardino, M. (1992), Geological evidence of recent fault evolution. Examples from Campo Imperatore (L'Aquila-central Apennines),, Il Quaternario, 5(2), 181-200.

Cavinato, G. P., Carusi, C., Dall'Asta, M., Miccadei, E., & Piacentini, T. (2002), Sedimentary and tectonic evolution of Plio–Pleistocene alluvial and lacustrine deposits of Fucino Basin (central Italy), Sedimentary Geology, 148, 29-59, doi: 10.1016/S0037-0738(01)00209-3.

Ceccaroni, E., Ameri, G., Gómez Capera, A. A., & Galadini, F. (2009), The 2nd century AD earthquake in central Italy: archaeoseismological data and seismotectonic implications, Natural Hazards, 50(2), 335-359, doi: 10.1007/s11069-009-9343-x.

Cinque, A., Ascione, A., & Caiazzo, C. (2000), Distribuzione spazio-temporale e caratterizzazione della fagliazione quaternaria in Appennino meridionale, in Le ricerche del GNDT nel campo della pericolosità sismica (1996-1999), CNR-Gruppo Nazionale per la Difesa dai Terremoti, edited by F. Galadini, C. Meletti and A. Rebez, p. 397, Roma.

Corrado, S., Di Bucci, D., Leschiutta, I., Naso, G., & Trigari, A. (1997), La tettonica Quaternaria della Piana di Isernia nell'evoluzione strutturale del settore molisano, Il Quaternario. Italian Journal of Quaternary Sciences, 10(2), 609-614.

Cucci, L., D'Addezio, G., Valensise, G., & Burrato, P. (1996), Investigating seismogenic faults in Central and Southern Apennines (Italy): modelling of fault-related landscape features, Annals of Geophysics, 39, 603-618, doi: https://doi.org/10.4401/ag-3995.

D'Addezio, G., Masana, E., & Pantosti, D. (2001), The Holocene paleoseismicity of the Aremogna-Cinque Miglia Fault (Central Italy), Journal of Seismology, 5(2), 181-205, doi: 10.1023/a:1011403408568.

D'Agostino, N., Giuliani, R., Mattone, M., & Bonci, L. (2001), Active crustal extension in the Central Apennines (Italy) inferred from GPS measurements in the interval 1994-1999, Geophysical Research Letters, 28(10), 2121-2124, doi: 10.1029/2000gl012462.

Devoti, R., Anderlini, L., Anzidei, M., Esposito, A., Galvani, A., Pietrantonio, G., Pisani, A. R., Riguzzi, F., Sepe, V., & Serpelloni, E. (2012), The coseismic and postseismic deformation of the L'Aquila, 2009 earthquake from repeated GPS measurements, italian Journal of Geosciences, 131, 348-358, doi: https://doi.org/10.3301/IJG.2012.15.

Devoti, R., et al. (2017), A Combined Velocity Field of the Mediterranean Region, Annals of Geophysics, 60(2), doi: 10.4401/ag-7059.

Devoti, R., Riguzzi, F., Cuffaro, M., & Doglioni, C. (2008), New GPS constraints on the kinematics of the Apennines subduction, Earth and Planetary Science Letters, 273(1-2), 163-174, doi: 10.1016/j.epsl.2008.06.031.

Di Bucci, D., Corrado, S., & Naso, G. (2002), Active faults at the boundary between Central and Southern Apennines (Isernia, Italy), Tectonophysics, 359(1), 47-63, doi: https://doi.org/10.1016/S0040-1951(02)00414-6.

Di Bucci, D., Naso, G., Corrado, S., & Villa, I. M. (2005), Growth, interaction and seismogenic potential of coupled active normal faults (Isernia Basin, central-southern Italy), Terra Nova, 17(1), 44-55, doi: 10.1111/j.1365-3121.2004.00582.x.

Di Bucci, D., Vannoli, P., Burrato, P., Fracassi, U., & Valensise, G. (2011), Insights from the Mw 6.3, 2009 L’Aquila earthquake (Central Apennines) – unveiling new seismogenic sources through their surface signatures: the adjacent San Pio Fault, Terra Nova, 23(2), 108-115, doi: 10.1111/j.1365-3121.2011.00990.x.

Di Luccio, F., Chiodini, G., Caliro, S., Cardellini, C., Convertito, V., Pino, N. A., Tolomei, C., & Ventura, G. (2018), Seismic signature of active intrusions in mountain chains, Science Advances, 4(1), e1701825, doi: 10.1126/sciadv.1701825.

Di Naccio, D., Kastelic, V., Carafa, M. M. C., Esposito, C., Milillo, P., & Di Lorenzo, C. (2019), Gravity Versus Tectonics: The Case of 2016 Amatrice and Norcia (Central Italy) Earthquakes Surface Coseismic Fractures, Journal of Geophysical Research: Earth Surface, 124(4), 994-1017, doi: 10.1029/2018jf004762.

DISS Working Group (2015), Database of Individual Seismogenic Sources (DISS), Version 3.2.0: A compilation of potential sources for earthquakes larger than M 5.5 in Italy and surrounding areas, doi: 10.6092/INGV.IT-DISS3.2.0.

Esposito, A., et al. (2020), Concurrent deformation processes in the Matese massif area (Central-Southern Apennines, Italy), Tectonophysics, 774, 228234, doi: 10.1016/j.tecto.2019.228234.

Esposito, A., Luongo, G., Marturano, A., & Porfido, S. (1987), Il terremoto di Sant' Anna del 26 Luglio 1805, edited, pp. 171-191.

Falcucci, E., et al. (2009), The Paganica Fault and Surface Coseismic Ruptures Caused by the 6 April 2009 Earthquake (L'Aquila, Central Italy), Seismological Research Letters, 80(6), 940-950, doi: 10.1785/gssrl.80.6.940.

Falcucci, E., Gori, S., Moro, M., Fubelli, G., Saroli, M., Chiarabba, C., & Galadini, F. (2015), Deep reaching versus vertically restricted Quaternary normal faults: Implications on seismic potential assessment in tectonically active regions: Lessons from the middle Aterno valley fault system, central Italy, Tectonophysics, 651-652, 186-198, doi: 10.1016/j.tecto.2015.03.021.

Faure Walker, J. P., Roberts, G. P., Cowie, P. A., Papanikolaou, I., Michetti, A. M., Sammonds, P., Wilkinson, M., McCaffrey, K. J. W., & Phillips, R. J. (2012), Relationship between topography, rates of extension and mantle dynamics in the actively-extending Italian Apennines, Earth and Planetary Science Letters, 325-326, 76-84, doi: 10.1016/j.epsl.2012.01.028.

Ferranti, L., Palano, M., Cannavò, F., Mazzella, M. E., Oldow, J. S., Gueguen, E., Mattia, M., & Monaco, C. (2014), Rates of geodetic deformation across active faults in southern Italy, Tectonophysics, 621, 101-122, doi: 10.1016/j.tecto.2014.02.007.

Ferrarini, F., Boncio, P., de Nardis, R., Pappone, G., Cesarano, M., Aucelli, P. P. C., & Lavecchia, G. (2017), Segmentation pattern and structural complexities in seismogenic extensional settings: The North Matese Fault System (Central Italy), Journal of Structural Geology, 95, 93-112, doi: doi:10.1016/j.jsg.2016.11.006.

Field, E. H., et al. (2014), Uniform California Earthquake Rupture Forecast, Version 3 (UCERF3) -The Time-Independent Model, B Seismol Soc Am, 104(3), 1122-1180, doi: 10.1785/0120130164.

Fracassi, U., & Milano, G. (2014), A soft linkage between major seismogenic fault systems in the central-southern Apennines (Italy): Evidence from low-magnitude seismicity, Tectonophysics, 636, 18-31, doi: https://doi.org/10.1016/j.tecto.2014.08.002.

Frepoli, A., Cimini, G. B., De Gori, P., De Luca, G., Marchetti, A., Monna, S., Montuori, C., & Pagliuca, N. M. (2017), Seismic sequences and swarms in the Latium-Abruzzo-Molise Apennines (central Italy): New observations and analysis from a dense monitoring of the recent activity, Tectonophysics, 712-713, 312-329, doi: 10.1016/j.tecto.2017.05.026.

Frezzotti, M., & Giraudi, C. (1989), Evoluzione geologica tardo-pleistocenica ed olocenica del Piano di Aremogna (Roccaraso-Abruzzo): implicazioni climatiche e tettoniche, Memorie della Società Geologica Italiana, 42, 5-19.

Galadini, F. (1999), Pleistocene changes in the central Apennine fault kinematics: A key to decipher active tectonics in central Italy, Tectonics, 18(5), 877-894, doi: 10.1029/1999tc900020.

Galadini, F., & Galli, P. (1999), The Holocene paleoearthquakes on the 1915 Avezzano earthquake faults (central Italy): implications for active tectonics in the central Apennines, Tectonophysics, 308, 143-170, doi: 10.1016/S0040-1951(99)00091-8.

Galadini, F., & Galli, P. (2000), Active tectonics in the central Apennines (Italy) - Input data for seismic hazard assessment, Natural Hazards, 22(3), 225-270, doi: Doi 10.1023/A:1008149531980.

Galadini, F., & Galli, P. (2003), Paleoseismology of silent faults in the central Apennines (Italy): the Mt. Vettore and Laga Mts. Faults, Annals of Geophysics, 46(5), 815-836, doi: 10.4401/ag-3457.

Galadini, F., & Giuliani, R. (1995), Elementi per una valutazione della cinematica quaternaria della Piana del Fucino (Italia centrale): l'analisi delle deformazioni dei ciottoli delle unità plio-pleistoceniche, Il Quaternario, 8(1), 183-192.

Galadini, F., & Messina, P. (1993), Characterization of the recent tectonics of the Upper Sangro River Valley (Abruzzi Apennines, Central Italy), Annals of Geophysics, 37, 277-285.

Galadini, F., & Messina, P. (1994), Plio-Quaternary tectonics of the Fucino basin and surrounding areas (central Italy), Giornale di Geologia, 56, 73-99.

Galadini, F., Giraudi, C., & Messina, P. (1998), Nuovi dati sulla tettonica tardopleistocenica dell'alta valle del Sangro (Appennino centrale): implicazioni sismotettoniche, Il Quaternario, 11, 347-356.

Galadini, F., Messina, P., Giaccio, B., & Sposato, A. (2003), Early uplift history of the Abruzzi Apennines (central Italy): available geomorphological constraints, Quaternary International, 101-102, 125-135, doi: 10.1016/s1040-6182(02)00095-2.

Galli, P. A. C., & Naso, J. A. (2009), Unmasking the 1349 earthquake source (southern Italy): paleoseismological and archaeoseismological indications from the Aquae Iuliae fault, Journal of Structural Geology, 31(2), 128-149, doi: 10.1016/j.jsg.2008.09.007.

Galli, P. A. C., Giaccio, B., Messina, P., Peronace, E., & Zuppi, G. M. (2011), Palaeoseismology of the L'Aquila faults (central Italy, 2009, Mw 6.3 earthquake): implications for active fault linkage, Geophysical Journal International, 187(3), 1119-1134, doi: 10.1111/j.1365-246X.2011.05233.x.

Galli, P., & Galadini, F. (2003), Disruptive earthquakes revealed by faulted archaeological relics in Samnium (Molise, southern Italy), Geophysical Research Letters, 30(5), doi: 10.1029/2002gl016456.

Galli, P., Giaccio, B., Peronace, E., & Messina, P. (2015), Holocene Paleoearthquakes and Early–Late Pleistocene Slip Rate on the Sulmona Fault (Central Apeninnes, Italy)Holocene Paleoearthquakes and Early–Late Pleistocene Slip Rate on the Sulmona Fault (Central Apeninnes, Italy), B Seismol Soc Am, 105(1), 1-13, doi: 10.1785/0120140029.

Galvani, A., Anzidei, M., Devoti, R., Esposito, A., Pietrantonio, G., Pisani, A. R., Riguzzi, F., & Serpelloni, E. (2013), The interseismic velocity field of the central Apennines from a dense GPS network, 2013, 55(5), doi: 10.4401/ag-5634.

Gasperini, P., Bernardini, F., Valensise, G., & Boschi, E. (1999), Defining seismogenic sources from historical earthquake felt reports, B Seismol Soc Am, 89(1), 94-110.

Ghisetti, F., & Vezzani, L. (1997), Interfering paths of deformation and development of arcs in the fold-and-thrust belt of the central Apennines (Italy), Tectonics, 16(3), 523-536, doi: 10.1029/97tc00117.

Giraudi, C. (1989), Datazione di un evento sismico preistorico con metodi geologici e radiometrici: Piani di Aremogna e delle Cinque Miglia, in I terremoti prima del Mille in Italia e nell'area mediterranea, edited by E. Guidoboni, pp. 53-64, ING and SGA, Bologna.

Giraudi, C. (1994), Elementi di geologia del Quaternario della Piana di Campo Imperatore (Gran Sasso d'Italia), Atti Tic. Sci. Terra, spec. ser., 2, 137-143.

Giraudi, C., & Frezzotti, M. (1995), Palaeoseismicity in the Gran Sasso Massif (Abruzzo, Central Italy), Quaternary International, 25, 81-93, doi: https://doi.org/10.1016/1040-6182(94)P3716-L.

Gori, S., Falcucci, E., Galadini, F., Moro, M., Saroli, M., & Ceccaroni, E. (2017), Geoarchaeology and paleoseismology blend to define the Fucino active normal fault slip history, central Italy, Quaternary International, 451, 114-128, doi: 10.1016/j.quaint.2017.01.028.

Gori, S., Giaccio, B., Galadini, F., Falcucci, E., Messina, P., Sposato, A., & Dramis, F. (2011), Active normal faulting along the Mt. Morrone south-western slopes (central Apennines, Italy), International Journal of Earth Sciences, 100(1), 157-171, doi: 10.1007/s00531-009-0505-6.

Guerrieri, L., Scarascia Mugnozza, G., & Vittori, E. (1999), Analisi stratigrafica e geomorfologia della conoide tardo-quaternaria di Campochiaro ed implicazioni per la conca di Bojano in Molise, Il Quaternario. Italian Journal of Quaternary Sciences, 12(2), 237-247.

Guidoboni, E., Ferrari, G., Tarabusi, G., Sgattoni, G., Comastri, A., Mariotti, D., Ciuccarelli, C., Bianchi, M. G., & Valensise, G. (2019), CFTI5Med, the new release of the catalogue of strong earthquakes in Italy and in the Mediterranean area, Sci Data, 6(1), 80, doi: 10.1038/s41597-019-0091-9.

Howe, T. M., & Bird, P. (2010), Exploratory models of long-term crustal flow and resulting seismicity across the Alpine-Aegean orogen, Tectonics, 29(4), doi: 10.1029/2009tc002565.

Huang, M.-H., Fielding, E. J., Liang, C., Milillo, P., Bekaert, D., Dreger, D., & Salzer, J. (2017), Coseismic deformation and triggered landslides of the 2016 Mw 6.2 Amatrice earthquake in Italy, Geophysical Research Letters, 44(3), 1266-1274, doi: 10.1002/2016gl071687.

Hunstad, I., Selvaggi, G., & D'Agostino, N. (2003), Geodetic strain in peninsular Italy between 1875 and 2001, Geophysical Research …, 30(4), 1-4, doi: 10.1029/2002GL016447.

Hussain, E., Wright, T. J., Walters, R. J., Bekaert, D. P. S., Lloyd, R., & Hooper, A. (2018), Constant strain accumulation rate between major earthquakes on the North Anatolian Fault, Nature Communications, 9(1), 1392, doi: 10.1038/s41467-018-03739-2.

Johnson, K. M. (2013), Slip rates and off-fault deformation in Southern California inferred from GPS data and models, Journal of Geophysical Research: Solid Earth, 118(10), 5643-5664, doi: 10.1002/jgrb.50365.

Kastelic, V., & Carafa, M. M. C. (2012), Fault slip rates for the active External Dinarides thrust-and-fold belt, Tectonics, 31, 18, doi: 10.1029/2011tc003022.

Kastelic, V., Burrato, P., Carafa, M. M. C., & Basili, R. (2017), Repeated surveys reveal nontectonic exposure of supposedly active normal faults in the central Apennines, Italy, J. Geophys. Res.-Earth Surf., 122(1), 114-129, doi: 10.1002/2016jf003953.

Lavecchia, G., Boncio, P., Brozzetti, F., de nardis, R., Pace, B., & Visini, F. (2006), Studio della pericolosità sismica della regione Abruzzo.*Rep.*, 31pp pp, Dipartimento di Scienze della Terra, Università di Chieti.

Lavecchia, G., Brozzetti, F., Barchi, M., Menichetti, M., & Keller, J. V. A. (1994), Seismotectonic zoning in east-central Italy deduced from an analysis of the Neogene to present deformations and related stress fields, Geological Society of America Bulletin, 106(9), 1107-1120, doi: 10.1130/0016-7606(1994)106<1107:Szieci>2.3.Co;2.

Mariucci, M. T., & Montone, P. (2019), IPSI 1.3, Database of Italian Present-day Stress Indicators, edited by INGV.

McCalpin, J. (2009), Paleoseismology (2nd ed.), Academic Press, Elsevier Publishing, San Diego, USA.

Meade, B. J., Klinger, Y., & Hetland, E. A. (2013), Inference of Multiple Earthquake-Cycle Relaxation Timescales from Irregular Geodetic Sampling of Interseismic Deformation, B Seismol Soc Am, 103(5), 2824-2835, doi: 10.1785/0120130006.

Meletti, C., Patacca, E., & Scandone, P. (2000), Construction of a Seismotectonic Model: The Case of Italy, pure and applied geophysics, 157(1), 11-35, doi: 10.1007/pl00001089.

Messina, P., Galli, P., Giaccio, B., & Peronace, E. (2009), Quaternary tectonic evolution of the area affected by the Paganica fault (2009 L'Aquila earthquake), in *28◦ GNGTS National Conference*, edited, pp. 45-50, Trieste 16-19 November 2009.

Michetti, A. M., Brunamonte, F., Serva, L., & Vittori, E. (1996), Trench investigations of the 1915 Fucino earthquake fault scarps (Abruzzo, central Italy): Geological evidence of large historical events, Journal of Geophysical Research: Solid Earth, 101(B3), 5921-5936, doi: 10.1029/95jb02852.

Montone, P., & Mariucci, M. T. (2016), The new release of the Italian contemporary stress map, Geophysical Journal International, 205(3), 1525-1531, doi: 10.1093/gji/ggw100.

Morewood, N. C., & Roberts, G. P. (2000), The geometry, kinematics and rates of deformation within an en échelon normal fault segment boundary, central Italy, Journal of Structural Geology, 22, 1027-1047, doi: 10.1016/S0191-8141(00)00030-4.

Moro, M., Gori, S., Falcucci, E., Saroli, M., Galadini, F., & Salvi, S. (2013), Historical earthquakes and variable kinematic behaviour of the 2009 L'Aquila seismic event (central Italy) causative fault, revealed by paleoseismological investigations, Tectonophysics, 583, 131-144, doi: https://doi.org/10.1016/j.tecto.2012.10.036.

Pace, B., Boncio, P., & Lavecchia, G. (2002), The 1984 Abruzzo earthquake (Italy): an example of seismogenic process controlled by interaction between differently oriented synkinematic faults, Tectonophysics, 350(3), 237-254, doi: 10.1016/s0040-1951(02)00118-x.

Pantosti, D., D'Addezio, G., & Cinti, F. R. (1996), Paleoseismicity of the Ovindoli-Pezza fault, central Apennines, Italy: A history including a large, previously unrecorded earthquake in the Middle Ages (860-1300 A.D.), Journal of Geophysical Research: Solid Earth, 101(B3), 5937-5959, doi: 10.1029/95jb03213.

Papanikolaou, I. D., Roberts, G. P., & Michetti, A. M. (2005), Fault scarps and deformation rates in Lazio–Abruzzo, Central Italy: Comparison between geological fault slip-rate and GPS data, Tectonophysics, 408, 147-176, doi: 10.1016/j.tecto.2005.05.043.

Patacca, E., Sartori, R., & Scandone, P. (1990), Tyrrhenian basin and Apenninic arcs: kinematic relations since Late Tortonian times, Memorie della Società Geologica Italiana, 45, 425-451.

Pizzi, A., Falcucci, E., Gori, S., Galadini, F., Messina, P., Di Vincenzo, M., Esestime, P., Giaccio, B., Pomposo, G., & Sposato, A. (2010), Active faulting in the Maiella Massif (central Apennines, Italy), GeoActa, Special Publication, 3, 57-73.

Pondrelli, S. (2002), European-Mediterranean Regional Centroid-Moment Tensors Catalog (RCMT), edited by I. N. d. G. e. Vulcanologia.

Roberts, G. P., & Michetti, A. M. (2004), Spatial and temporal variations in growth rates along active normal fault systems : an example from The Lazio – Abruzzo Apennines , central Italy, 26, 339-376, doi: 10.1016/S0191-8141(03)00103-2.

Rovida, A., Locati, M., Camassi, R., Lolli, B., & Gasperini, P. (2016), CPTI15, the 2015 version of the Parametric Catalogue of Italian Earthquakes, doi: 10.6092/INGV.IT-CPTI15.

Salvi, S., Cinti, F. R., Colini, L., D'Addezio, G., Doumaz, F., & Pettinelli, E. (2003), Investigation of the active Celano–L'Aquila fault system, Abruzzi (central Apennines, Italy) with combined ground-penetrating radar and palaeoseismic trenching, Geophysical Journal International, 155(3), 805-818, doi: 10.1111/j.1365-246X.2003.02078.x.

Saroli, M., & Moro, M. (2012), San Pietro Infine basin, in Field-trip guidebook- 16th joint Geomorphological meeting. July 1-5, 2012, edited by AIGeo, Rome.

Saroli, M., Lancia, M., Albano, M., Modoni, G., Moro, M., & Scarascia Mugnozza, G. (2014), New geological data on the Cassino intermontane basin, central Apennines, Italy, Rendiconti Lincei, 25(2), 189-196, doi: 10.1007/s12210-014-0338-5.

Saroli, M., Moro, M., Borghesi, H., Dell'Acqua, D., Galadini, F., & Galli, P. (2008), Nuovi dati paleoseismologici del settore orientale del bacino del Fucino (Italia centrale), Il Quaternario, 21, 383-394.

Scisciani, V., Tavarnelli, E., & Calamita, F. (2002), The interaction of extensional and contractional deformations in the outer zones of the Central Apennines, Italy, Journal of Structural Geology, 24(10), 1647-1658, doi: 10.1016/s0191-8141(01)00164-x.

Sepe, V., Brandi, G., De Martino, P., Dolce, M., Obrizzo, F., Pingue, F., & Tammaro, U. (2009), S.A.G.NET: Rete GPS dell’Appennino meridionale*Rep.*

Splendore, R., Marotta, A. M., & Barzaghi, R. (2015), Tectonic deformation in the Tyrrhenian: A novel statistical approach to infer the role of the Calabrian Arc complex, Journal of Geophysical Research: Solid Earth, 120(11), 7917-7936, doi: 10.1002/2015jb012313.

Stemberk, J., Moro, G. D., Stemberk, J., Blahůt, J., Coubal, M., Košťák, B., Zambrano, M., & Tondi, E. (2019), Strain monitoring of active faults in the central Apennines (Italy) during the period 2002–2017, Tectonophysics, 750, 22-35, doi: 10.1016/j.tecto.2018.10.033.

Stucchi, M., Albini, P., Mirto, M., & Rebez, A. (2004), Assessing the completeness of Italian historical earthquake data, 2004, 47(2-3), doi: 10.4401/ag-3330.

Valensise, G., & Pantosti, D. (2001), The investigation of potential earthquake sources in peninsular Italy: A review, Journal of Seismology, 5(3), 287-306, doi: 10.1023/a:1011463223440.

Valentini, A., Visini, F., & Pace, B. (2017), Integrating faults and past earthquakes into a probabilistic seismic hazard model for peninsular Italy, Nat Hazard Earth Sys, 17, 2017-2039, doi: 10.5194/nhess-17-2017-2017.

Valoroso, L., Chiaraluce, L., Piccinini, D., Di Stefano, R., Schaff, D., & Waldhauser, F. (2013), Radiography of a normal fault system by 64,000 high-precision earthquake locations: The 2009 L'Aquila (central Italy) case study, Journal of Geophysical Research: Solid Earth, 118(3), 1156-1176, doi: 10.1002/jgrb.50130.

Vannoli, P., Burrato, P., Fracassi, U., & Valensise, G. (2012), A fresh look at the seismotectonics of the Abruzzi (Central Apennines) following the 6 April 2009 L’Aquila earthquake (Mw 6.3), Italian Journal of Geosciences, 131(3), 309-329, doi: 10.3301/ijg.2012.03.

Vittori, E., Cavinato, G. P., & Miccadei, E. (1995), Active faulting along the northeastern edge of the Sulmona basin, central Apennines, Italy, in Perspective in Paleoseismology, edited by L. Serva and B. D. Slemmons, pp. 115-126, Special Publication, Association of Engineering Geologists, Sudbury, Washington.

Westaway, R. (1992), Seismic Moment Summation for Historical Earthquakes in Italy - Tectonic Implications, J Geophys Res-Sol Ea, 97(B11), 15437-15464, doi: Doi 10.1029/92jb00946.

Westaway, R., Gawthorpe, R., & Tozzi, M. (1989), Seismological and field observations of the 1984 Lazio-Abruzzo earthquakes: implications for the active tectonics of Italy, Geophysical Journal International, 98(3), 489-514, doi: 10.1111/j.1365-246X.1989.tb02285.x.

Zeng, Y., & Shen, Z. K. (2016), A Fault-Based Model for Crustal Deformation, Fault Slip Rates, and Off-Fault Strain Rate in California, B Seismol Soc Am, 106(2), 766-784, doi: 10.1785/0120140250.

Zreda, M., & Noller, J. S. (1998), Ages of Prehistoric Earthquakes Revealed by Cosmogenic Chlorine-36 in a Bedrock Fault Scarp at Hebgen Lake, Science, 282(5391), 1097, doi: 10.1126/science.282.5391.1097.

Figure captions

Table 1. Geodetic slip rates and rakes resulting from our preferred models and comparison with estimates available in the geological literature.

Figure 1. a) Kinematic (quasi-static) model of the seismic cycle of a fault embedded in a visco-plastic deforming lithosphere. Long-term fault slip rate: 1.2 mm/a; total diffuse deformation from left to right: 0.25 mm/a in 210 km. The velocity variation along x-axis due to diffuse deformation is modelled with a cumulative normal distribution function between -105 km and 105 km, mean = 0 km and sigma =25 km. b) Time evolution of horizontal distance between point A and point B for different time-scales. c) Two examples of geodetic measurements in the central Apennines including interseismic and coseismic/postseismic stages.

Figure 2. Instrumental seismicity (M_w>4.5, Pondrelli, 2002) and active faults of the study region from the relevant literature. Active fault planes that are adequately illuminated by the GPS networks are shown in red. Active faults:1) Ailano-Piedimonte; 2) Matese; 3) Aquae Iuliae; 4) Cassino - San Pietro Infine; 5) Barrea; 6) Porrara; 7) Aremogna-Cinque Miglia; 8) Upper Sangro Valley- Marsicano; 9) Sora; 10) Morrone; 11) Middle Aterno Valley; 12) Fucino; 13) Liri Valley; 14) Cappucciata; 15) S. Pio delle Camere; 16) Paganica; 17) Campo Felice; 18) Campo Imperatore. In cyan profiles shown in Figure 6.

Figure 3. GPS coverage obtained by merging permanent GNSS and temporary benchmarks (see text for further details), and trace of the profiles shown in Figure 6.

Figure 4. Stress data records from the IPSI database (Montone & Mariucci, 2016; Mariucci & Montone, 2019) White lines: profiles shown in Figure 6.

Figure 5. Left: stress and geodetic scalar misfits for different values of the tuning parameter A_0. The brown symbols represent the scalar misfits of the preferred models. Right: fault slip rates variability for different values of A_0. Best estimates are enclosed in the rectangle.

Figure 6. Model performance along four selected profiles (for profile position and fault names see Figure 2) compared with GPS (black circles) and stress data (grey circles). Model predictions at the center of the grid finite elements are shown with blue dots. Blue lines show the long-term and interseismic horizontal velocities. Green rectangles refer to areas undergoing bulk permanent deformation (see Figure S6 for the map of the total bulk deformation rate); red rectangles indicate fault slip rates along the selected profiles (see Table 1 for along-strike averaged slip rates). $A_0=23\cdot10^7$ m^2 for the selected model.

Figure 7. Predicted focal mechanism for active faults with adequate coverage by GPS benchmarks. Histograms: comparison among geodetic slip rates resulting from our preferred models (grey stripes) with estimates available in the geological literature. The letters refer to Table 1 (column “Reference ID”) and link each fault offset estimate to the respective reference.