The role of rheology, crustal structures and lithology in the seismicity distribution of the northern Apennines

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A R T I C L E   I N F O

Article history:
Received 14 July 2016
Received in revised form 4 November 2016
Accepted 13 November 2016
Available online 15 November 2016

Keywords:
Northern Apennines
Earthquakes
Seismicity cut-off
Brittle ductile-transition
Crustal rheology
Altotiberina fault

A B S T R A C T

The Northern Apennines of Italy is a unique area to study active crustal processes due to the availability of high-resolution subsurface geology (deep borehole and seismic profiles) and seismicity (back-ground and seismic sequences) data. In this work we have investigated the relationship between crustal structures and lithologies, rheological profiles and seismicity cut-off by constructing three integrated profiles across the Umbria-Marche Apennines.

At first approximation we observe a good correspondence between the background seismicity cut-off and the modelled brittle ductile transition (BDT): 90% of the seismic activity is located above the transition. In the area characterized by active extension, where the majority of the seismicity is occurring, most of the crustal earthquakes are confined within the brittle layer at depth < 12 km. In areas where the brittle layer is affected by regional structures, we observe an active role played by these structures in driving the seismicity distribution. One example is the region where the gently eastward dipping Altotiberina low-angle normal fault is present, where the seismicity cut-off is completely controlled by this detachment that separates an active hanging-wall from an almost aseismic foot-wall block. At smaller scale also the lithology plays a strong control on the seismicity distribution. We observe that the largest earthquakes of the area, 5.5 ≤ M ≤ 6.0, do not nucleate at the base of the BDT but they occur at the base of the sedimentary cover.

Our work suggests that rheology and therefore the position of the brittle ductile transition exerts a role at regional scale for the occurrence of crustal seismicity, however crustal structures and lithology play the major role at a more local scale and therefore they need to be considered for a better understanding of earthquake distribution within the seismogenic layer.

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1. Introduction

In the 70' and 80' the increasing deployment of high-density seismic networks with modern techniques for earthquake location demonstrated that seismic activity is concentrated in the upper part of the deforming continental crust (e.g. Archuleta, 1984; Cockerham and Eaton, 1987; Scholz, 2002). This observation allowed the development of the concept of the seismogenic regime defined by background micro-earthquakes (e.g. Sibson, 1983; Scholz, 2002). In the same period geological studies of exhumed faults exposed at the surface were used to produce a conceptual model for major faults where a localized fault zone of elasto-frictional behaviour generating random-fabric fault rocks (e.g. cataclasites) overlies regions where quasi-plastic processes occur along distributed shear zones with mylonitic fault rocks (Sibson, 1977). In the 80' Brace and Kohlstedt (1980) used laboratory data to develop a simple model for fault rheology. In the upper part of the model "the brittle part", the fault behaviour is predominantly frictional and Byerlee's friction rule (Byerlee, 1978) dictates the strength. In the lower part of the model "the ductile part", the fault behaviour, due to an increase in temperature, is plastic and the strength is controlled by the extrapolation of quartzite flow laws obtained during rock deformation experiments under controlled temperature (e.g. Brace and Kohlstedt, 1980; Hirth et al., 2001). From a mechanical viewpoint, the base of the brittle regime represents a transition zone from unstable frictional faulting to aseismic ductile slip (e.g. Sibson, 1983; Scholz, 2002). This brittle-ductile transition represents also the region of the peak shear resistance where the largest earthquakes tend to nucleate (e.g. Sibson, 1983; Scholz, 2002).

In the last fifteen years the improvements of earthquake detection and innovative earthquake location techniques (e.g. Waldhauser and Ellsworth, 2000) have shown the key-role played by fault zones in seismicity distribution within the seismogenic layer. Detachment faults separating active hanging-wall from aseismic footwall have been...
documented in compressional (e.g. Carena et al., 2002) and extensional (e.g. Chiarello et al., 2007) environments. Fault zone complexity has been illuminated for different structures (Powers and Jordan, 2010; Valoroso et al., 2014). At the same time the monitoring of active faults with geodetic and InSAR (interferometric synthetic aperture radar) techniques, has shown that within the brittle regime a significant amount of deformation can be aseismically accommodated (e.g. Perfettini et al., 2010). This is because faults within the seismogenic regime are frictionally heterogeneous with velocity weakening portions more prone to develop frictional instability resulting in earthquakes and velocity strengthening areas favouring stable sliding and fault creep (e.g. Marone, 1998; Scholz, 2002). Fault zone structure and lithology play an important role in the velocity weakening vs. velocity strengthening behaviour of the fault rocks (e.g. Collettini et al., 2011; Ikari et al., 2013; Carpenter et al., 2014; Tesi et al., 2014).

Within this framework the seismicity distribution has classically represented one of the main geophysical proxies to test hypothesis on the structure as well as the evolution and rheology of the continental crust. At first approximation the seismicity cut-off constrains the depth of the seismogenic layer and it should coincide with the brittle-ductile transition obtained from the rheology profiles. In detail, crustal structures and lithology play a key-role in seismicity distribution and in the asseismic component of the deformation of the seismogenic layer. However, for a single studied region the availability of a high-resolution and integrated dataset is not common.

Here we profit of a unique test site in the Northern Apennines of Italy characterized by the possibility of integrating data from subsurface geology (deep borehole and seismic profiles) and seismocity (back-ground and seismic sequences) with rheological profiles and frictional properties of the rocks, to discuss about: a) the relationship between the brittle-ductile transition and the seismicity cut-off; b) the role of crustal structures in seismicity distribution and c) the role of lithology for the nucleation of the major seismic sequences of the area. In detail, after modelling the brittle-ductile transition for the study region basing on heat flux data and kinematic information, we compare detailed structural information along 2D cross sections with the depth of the inferred strongest portion of the crust and rock composition, main geological structures and tectonic regime.

2. Geology of the area

The study area (Fig. 1) is centred in the eastern part of the Northern Apennines (Outer Northern Apennines, Barchi et al., 2001), corresponding to the Umbria and Marche regions. The Northern Apennines (hereafter NAP) of Italy is an arc-shaped, northeast-verging fold-thrust belt, formed in the framework of the convergence between the European continental crust (the Corsica-Sardinia block) and Adria (considered either as a promontory of the African Plate or as an independent microplate). Since Lower Miocene, the tectonic evolution of the NAP has been characterized by the contemporaneous activity and eastward migration of co-axial, coupled compression (in the foreland) and extension (in the hinterland), generating an overprint of contractional and extensional features in adjacent regions (Elter et al., 1975; Barchi, 2010). This process is still active, as demonstrated by both stress field data (Montone and Mariucci, 2016) and strain field data (Serpelloni et al., 2005; D’Agostino et al., 2009) data, generating compression in the present-day foreland (i.e. Adriatic coast and off-shore), while along the inner sector the compressional structures have been progressively dissected by the later extensional front. At present East-Northeast trending extension is active along the axial culmination of the Apenninic chain at a rate of 3 mm/yr (Serpelloni et al., 2005 among others) where the strongest instrumental (M6) earthquakes are located (Fig. 2).

The compressional structures of the NAP incorporate tectonic units derived from the Mesozoic Tethys Ocean and the adjacent continental passive margins. Top to bottom (West to East), they are: the Liguride units (mainly oceanic), the Tuscan units and the Umbria-Marche units (both deposited on the continental passive margin of Adria). In our study area, the Umbria-Marche Jurassic-Paleogene carbonate multilayer is exposed along the axial culmination of the NAP fold-thrust belt, showing maximum elevations around 2500 m. These pre-orogenic successions are unconformably overlain by syn-compressional units (mainly turbidites), deposited in foreland basins and becoming younger from west to east: the Late Messinian-Quaternary clastic sequence, deposited in the recentmost foreland basin, crops out in the eastern part of the study area (Marche Foothills). In the western part of the NAP both pre-orogenic and syn-orogenic successions are covered by continental and/or shallow marine clastic successions (syn-extensional units) deposited in recent extensional basins.

In order to characterize the main structures together with the distribution at depth of the exposed rocks that will be compared with the seismic activity and rheological properties, we have constructed three geological transects crossing the study area in WSW-ENE direction (T1, T2 and T3, see traces in Fig. 1).

The geological transects have been constructed basing on the integration of several different sets of both geological and geophysical (seismic reflection-refraction profiles and deep boreholes) data, which have been acquired during a wide time-window spanning from the late 70s to late 90s. The northernmost transect T1 is based on a re-interpretation of the eastern part of the deep seismic reflection profile CROP-03 (Barchi et al., 1998; Mirabella et al., 2011) and on section 1b of Baily et al. (1986). The western part of the central transect T2 is based on the integrated geological section by Mirabella et al. (2011) while the eastern part is based on the re-interpretation of the section 3b of Baily et al. (1986) and the seismic section proposed by Maesano et al. (2013). The southernmost transect T3 is the shortest in length (about 30 km less than the others). It is based for the western sector on the section proposed by Mirabella et al. (2008), while the eastern part is based on section 4 by Bally et al. (1986) and numerous boreholes drilled into the Meso-Cenozoic succession. For all the transects the information has been extrapolated down to the top of the lower crust located around 20–25 km of depth on the basis of seismic refraction data (Ponziani et al., 1995, 1998; De Franco et al., 1998) and of the CROP03 NVR profile (Pialli et al., 1998).

2.1. Geological transects

The uppermost part of the geological transects profiles down to the base of the sedimentary cover has been reconstructed integrating subsurface data (i.e. seismic profiles and deep boreholes, also reported in the Videpi database, http://umnig.sviluppoeconomico.gov.it/videpi/en/default.htm) and surface geology (Italian Geological Survey maps). The employed seismic reflection profiles consist of industrial profiles acquired by the oil industry in the 80s. On the basis of these sections, Bally et al. (1986) constructed a set of geological cross-sections (see traces in Fig. 1) providing a coherent reconstruction of the subsurface geological setting down to about a depth of about 12 km. The work by Bally et al. (1986) is the first regional study of the structural style of the Umbria-Marche thrust belt, based on extensive interpretation of seismic data. Their interpretation envisages a thin-skinned tectonic model in which the sedimentary cover is detached from the underlying basement constantly deepening from the Adriatic foreland toward the Apenninic chain, from about 7 km to 12 km (regional monocline). Later on other Authors (e.g. Sage et al., 1991; Ghisetti et al., 1993; Coward et al., 1999) proposed alternative thick-skinned models for the area. During the ’90s the CROP-03 profile (Pialli et al., 1998), crossing the whole Northern Apennines belt, provided further support to a thick-skinned model of deformation style of the Apennines (Barchi et al., 1998), also followed by more recent studies, providing detailed interpretations of the shallowest structures (e.g. Mirabella et al., 2008; Brozzetti et al., 2009; Maesano et al., 2013 and references within). The involvement of the basement, or at least of its upper part, in the compressional
structures, implies that the top basement is shallower (6 to 8 km beneath the axial ridge of the Apennines) compared to previous interpretations and the total shortening is lower (about 30%) than in the thin-skinned models.

Another outcome of the CROP-03 project, particularly relevant for the present study, is the recognition of a set of ENE-dipping low-angle normal faults (LANFs), dissecting the entire brittle upper crust till a depth of about 12 km (Barchi et al., 1998; Collettini et al., 2006). The western part of sections T1 and T2 confirm the presence of the easternmost of these LANFs (the Altotiberina fault; ATF; Boncio et al., 1998; 2000; Collettini and Barchi, 2002) that is still active releasing abundant micro seismicity (Chiaraluce et al., 2007, 2014).

2.2. Seismic refraction profiles and passive data

For the deepest part of the transects, extending from the top of the acoustic basement down to the top of the lower crust, we integrated the results of the CROP-03 deep seismic reflection profile (Pialli et al., 1998; Collettini et al., 2006) with the results of seismic refraction experiments and passive seismic studies, useful to constrain the crustal structure.

Within the study area two sets of refraction seismic refraction data are available (Fig. 1). The northern-most was acquired along the CROP-03 section (De Franco et al., 1998), while the southern-most crosses the study area from the Trasimeno Lake to the West toward the Mt. Conero area to the East (DSS-78, Ponziani et al., 1995). We complemented these data with the results of tomographic and receiver function studies performed on teleseismic events recordings (Piana Agostinetti et al., 2002; Mele and Sandvol, 2003; Piana Agostinetti and Amato, 2009). In general the association of these different data-sets (e.g. Barchi et al., 2006) convey in similar results, all highlighting the presence of a relatively thin Tyrrhenian crust, slightly deepening toward the east from 22 to 25 km, contrasting a thick Adriatic crust (30–35 km), separated by a sharp Moho step, approximately located underneath the Tiber Valley. While the main differences concern deeper portions such as the existence and the extent of a possible Moho doubling (i.e. a superposition of the Tyrrhenian Moho over the Adriatic Moho) beneath the Tiber Valley, along with the geometry and depth of the westernmost portion of the Adriatic Moho itself. Piana Agostinetti and Amato (2009) also hypothesise a different geometry of the Tyrrhenian crust, extending eastward beneath the Umbria-Marche Apennines. Most of these data are focused along a Perugia-Ancona transect, that grossly corresponds to the T2 section. For this reason, we reconstructed the crustal structure along the T2 section and projected the same deep structure to the other (T1 and T3) transects, parallel to the main structural axes. The assumption of a similar deep

Fig. 1. Geological sketch of the study area. Post-orogenic units: 1 – Alluvial, alluvial and debris (Holocene); 2 – Shallow marine and continental deposits, infilling the intermountain basins (Pliocene-lower Pleistocene); Syn-orogenic units: 3 – Liguride units, ophiolites, pelagites and turbidites (Jurassic-Eocene); 4 – Tuscan units, pelagites and siliciclastic turbidites (Cretaceous – Paleogene); 5 – Umbria-Marce turbidites, Marnoso-Arenacea and Lagca Flysch fms. (lower-upper Miocene); 6 – Eastern-Marche Molasse, Marine deposits, clays, sands and conglomerates (Messinian-Pleistocene); Pre-orogenic units: 7 – Umbria Marche carbonatic multilayer (Jurassic-lower Miocene-Paleogene). Within the map also the location of some data used in the work is shown; in particular the employed boreholes, the traces of sections after Bally et al. (1986); the trace of the DSS78 seismic refraction profile; the trace of the Crop-03 NVR seismic reflection profile. T1, T2 and T3 are the traces of the geological transects of fig. 3. SFL is the “Scheggia-Foligno line” (Barchi et al., 2012); IRT and ORT are the “Inner Ridge Thrust” and “Outer Ridge Thrust” respectively.
structure in the three transects is justified by the relatively short distance (<30 km) among the three transects.

2.3. The three transects across the Northern Apennines

Looking at the three (T1, T2 and T3) profiles we can divide the NAP into three main morpho-structural provinces bordered by two regional thrusts: the Scheggia-Foligno Line (SFL – Barchi et al., 2012; Fig. 3) and the Outer Ridge Thrust also named M. Sibillini thrust (ORT; Fig. 3). From west to east: the Umbria pre-Apennines (UPA in Fig. 3), the Umbria-Marche Ridge (UMR in Fig. 3), the Outer Marche Foothills (OMF in Fig. 3; Bally et al., 1986; Barchi et al., 2012). West of the SFL, the Umbrian pre-Apennines are characterized by extensive outcrops of Miocene Marnoso-Arenacea turbidites and by relatively low topographic elevation. Comprised between the SFL and the ORT, the Umbria-Marche Ridge is a belt composed by the major anticlines, with extensive exposures of Mesozoic-Paleogene carbonate multilayer, i.e. the Umbria-Marche succession (Cresta et al., 1989), forming the major mountain range of the region. East of the ORT, the Outer Marche Foothills represent the outer portion of the fold-thrust belt, characterized by extensive outcrops of recent (Pliocene-Quaternary), shallow marine and continental clastic sediments.

Along the three transects, these provinces show markedly different characters, that can be clearly recognised both at the surface and at depth.

In the western portion (UPA) of T1, we recognise the Umbrian turbidites of the Marnoso-Arenacea formation extensively cropping out with the top of the carbonates located at a depth of 2–3 km. The Umbria-Marche succession (composed by carbonates and evaporites) experienced a relatively low shortening and is affected only by a single W-dipping major thrust (MPT in Fig. 3), displacing the top of the acoustic basement, which shows an overall flat attitude at 5–6 km of depth while the Tiber Valley is bordered by the extensional system, driven by the E-dipping Altotiberina low angle normal fault (ATF).

In the central part of the T1 profile (UMR) the Mesozoic-Paleogene carbonates are exposed at the surface. The shortening suddenly increases, the top carbonates is shallower (it is actually exposed at the surface), while the top acoustic basement is much deeper (down to 10 km) and displaced by two major W-dipping thrusts (IRT and ORT).

The eastern part of profile T1 (i.e. Marche Foothills) is characterized by a lower shortening. The basement is still deep (8–10 km) and slightly W-dipping but the carbonates are covered by a relatively thick (up to 4 km) wedge of syntectonic clastic deposits. The top of the basement is affected by another major thrust (Coastal Anticline Thrust; CAT in Fig. 3) forming the coastal anticline (Bally et al., 1986).

Along the T1 profile the ATF, with its court of synthetic and antithetic splays, by far represents the main extensional feature of the region. It emerges in the Umbria pre-Apennines, generating the Tiber Valley basins, and dips beneath the Umbria-Marche Ridge till a depth of about 15 km.

Since the general structure of the mountain belt is longitudinally continuous, most of these features can be recognised also in the central (T2) and in the southern (T3) transects. However, they also show some peculiarities, mostly due to the fact that the total shortening of the thrust and fold belt increases southward; consequently, the structural elevation also tends to grow progressively in the same direction.
In the westernmost portion of transect T2 (UPA in Fig. 3), west of the Tiber Valley, the carbonates are exposed at the footwall of the ATF, where the regional culmination of the acoustic basement occurs at a depth of about 3 km. The Umbria-Marche Ridge is more shortened than along T1, but the general structure is quite similar, with the IRT and ORT stacking the basement units. In the Outer Marche Foothills, the profile is close to the northern termination of two major structural culminations (named Cingoli and M. Conero anticlines; see Fig. 1), which delimitate the Pliocene clastic depocentre, with a maximum thickness of about 4 km. Similarly to the T1 transect, at the surface all the extensional faults of the ATF system crop out in the Umbria pre-Apennines, while at depth the ATF dips beneath the Umbria-Marche Ridge till a maximum depth of about 18 km.

The southernmost transect (T3 in Fig. 3), being shorter, crosses only the easternmost portion of the UPA province where the Mt. Subasio anticline crops out. Consequently, the profile doesn’t cross the Tiber Valley and associated structures. In this sector the UMR becomes larger (about 40 km) and structurally more complex than along T1 and T2, at least at the surface. On the contrary, the subsurface structure again consists of the two major thrusts (IRT and ORT) displacing the top of the acoustic basement. Beneath the OMF, the syntectonic Neogene sediments reach their maximum thickness (up to 5 km).

3. Seismicity pattern

In order to characterize the seismicity pattern of the study area we have implemented the earthquake catalogue built by Carannante et al. (2013). The catalogue is created by locating the 2009–2011 seismicity in a 3-dimensional \( (V_p, V_s, V_p/V_s) \) tomographic model of the region with the non-linear earthquake location code NonLinLoc [Lomax et al., 2000], able to provide both hypocentre solutions in complex models and a complete description of the location uncertainties. After the
removal of the non-natural seismic events induced by the human activity (i.e. quarrying), the resulting catalogue of high-resolution earthquake hypocentres is composed by 14,597 events (grey points in the map in Fig. 2), with magnitudes ranging between −0.2 and 4.8 (events with $M_h > 3.5$ are reported as orange stars in Fig. 2).

In reason of the high seismic rate characterizing the region (Chiaraluce et al., 2009) and the low detection threshold reached by the Italian seismic network at the end of the 2008 for this sector of the peninsula ($M > 1.5$; Schorlemmer et al., 2010), three years of seismic activity can be considered meaningful of the on-going background activity. However, we added to this catalogue the largest seismic sequences occurred in the area in the last 30 years: Gubbio 1984, $M_s 5.3$ (red dots and focal mechanism in Fig. 2; Collettini et al., 2003), Colfurito 1997, $M_w 6.0$ (blue dots and 3 focal mechanisms in Fig. 2; Chiaraluce et al., 2003) and Gualdo Tadino 1998, $M_s 5.3$ (green dots and focal mechanism in Fig. 2; Caccio et al., 2005). We also reported the location and the focal mechanism solutions for the Conero earthquake ($M_s 4.4$) occurred in 2013 (light blue star in Fig. 2) and for all events occurred during 2009–2011 having $M_w > 3.5$ (grey focal mechanisms in Fig. 2; http://www.eas.slu.edu/eqeqc/eqc_mt/MECH.IT). The main reason for adding the seismic sequences to the background catalogue, that can be considered representative of the inter-seismic stage, is the need to include information coming from co- and early-post-seismic (e.g. sequence) phases. Despite the earthquakes catalogues we are exploiting in this study have different formal location errors, all of them have errors many times smaller than the size of the smaller geological structure (few kilometers) we are going to discuss and eventually interpret in this work. Thus they can be considered completely suitable for our purpose.

We used this combined catalogue to show in map view (Fig. 2) and along the three cross-sections (Fig. 4), the seismicity distribution of the study area. Along each cross-section, we have projected events located within an about 30 km wide strip, which is not always symmetric with respect to the cross-section trace, in order to minimize the overlap. In detail: in section T1 we have plotted the seismicity located within 30 and 10 km of distance respectively south and north from T1 trace, while in cross-sections T2 and T3 we have projected events nucleated plus or minus 20 km of distance from the traces.

Basing on the kinematic information such as focal mechanism distribution (this study), GPS data (Serpelloni et al., 2005; D’Agostino et al., 2009) and the seismotectonic subdivision proposed by Lavecchia et al. (2003), we have divided the area into three main different domains: extensional (light yellow box in Figs. 2 and 4), compressive (light blue box) and a transition one. While the extensional and compressional domains interest all the first 30 km of crust below the Umbria pre-Alpennines and Umbria Marche region and the eastern portion of the Outer Marche Foothills, respectively, the transition zone is instead characterized by extension within the upper crust (18–20 km of depth) and compression below.

Already from the map view (Fig. 2) we observe that the highest rate of seismic release occurs in the extensional domain, mainly along the axial zone of the Apennine chain and decreases toward the compressive area along the Outer Marche Foothills. The seismic activity striking the chain does not follow the arc-shaped geological structures characterizing the Northern Apennines and inherited from the compressional tectonic phase, but clusters along a quite straight NNW-trending longitudinal zone. This axial zone becomes wider from north to south: around 15–20 km and 25–30 km, respectively above and below latitude 43°.

The background seismic activity occurring in the extensional domain generally deepens eastward (Fig. 4). From north to south it goes from about 5 km, down to 11 km in section T1, from 8 to 14 km in section T2 and from 9 to 16 km in section T3. This implies a light deepening of the seismic activity also along the chain from north to south. We know that in cross-sections T1 and T2 there is some correlation between the presence of a well-known geologic extensional structure (ATF) and the seismicity distribution at depth (Chiaraluce et al., 2007 and reference therein). While for the sector imaged in section T3 there is no clear and direct link in between the known geologic structures with the seismic activity.

In correspondence of the transition domain, the T3 profile shows no seismicity at shallow crustal levels but a significant clustering at about 20 km of depth including the occurrence of a series of four extensional earthquakes with magnitude larger than 3.5 ($3.5 < M_s < 4.0$). Sections T1 and T2 do not show clustering of seismicity at depth $> 20$ km and also the shallow seismic activity ($depth < 10 \text{ km}$) is limited. The seismic activity is really sparse also within the easternmost compressional domain. Here the relatively more active sector is the one along the T2 profile with the events nucleating at almost all the depth range, similar to T1 sector. While in section T3 the occasional activity seems to be substantially deeper (below 10 km). All the largest events beneath the Adriatic coast show compressive kinematics, suggesting on-going orogenic accretion along the whole northern Apenninic arc.

4. Rheological profiles

We have constructed the trend of the brittle-ductile transition (hereinafter BDT) along the three studied profiles. The BDTs have been obtained by estimating the critical differential stress that is theoretically required to deform the lithosphere at a constant steady-state strain rate according to three main deformation mechanisms: brittle, high-pressure brittle and ductile. We describe the brittle behaviour by a frictional law (Sibson, 1983, Byerlee, 1978) and the ductile one as a power-law equation (after Ranalli, 1995) applied at different tectonic stress levels. Whereas the high-pressure brittle failure (after Zang et al., 2007) in the brittle field is reached when increasing pressure, the differential stress becomes nonlinearly dependent on pressure, temperature, and strain-rate. In any depth range, the predominant mechanism is the one requiring the smallest stress difference. The BDT is then a transition zone, which moves up and down in the crust in response to local conditions such as different stress state (i.e. compressional or extensional regime), pore fluid pressure, strain rate, rocks composition (i.e. their physical characteristics) and temperature.

The stress state has been obtained following the kinematic and seismotectonic subdivision described in the previous paragraph and showed in Figs. 2 and 4. Fluid pressure is assumed to be hydrostatic, while the strain rate is $10^{-15}$ s$^{-1}$ and $10^{-16}$ s$^{-1}$, for the extensional-transitional and compressional area, respectively (Guadagni et al., 1998; Dragoni et al., 1996).

As regards the rocks, we divided each transect into two layers: upper crust and lower crust following the indication coming from the geological sections in Fig. 3. For each of these layers we assigned two different mean lithological compositions with the corresponding rheological parameters coming from the literature: wet quartzite for the upper crust and dry granulite for the lower crust (see Table 1 for details) as proposed also by Chiarabba et al. (2009).

Temperature is a key parameter in determining the BDT. We obtained the temperature field at depth by numerically resolving the general equation of heat conduction over a 2D grid using the physical parameters summarized in Table 1. From the temperature distribution in depth we calculated the surface heat flow, which is compared with the measured one. In Fig. 5, the observed surface heat flow is reported. Finally, the temperature field has been obtained verifying the choice of the physical parameters assigned to the rock units through the constraints offered by the Curie’s isotherm (Pinna et al., 1991; Sperranzo and Chiappini, 2002) and the lithosphere-asthenosphere transition (e.g. Calcagnile and Panza, 1980; Lucente et al., 1999; Piromallo and Morelli, 2003). As a general remark, we point out that in our calculations variations of the order of 10% on the parameters, do not drastically influence the final results (Pauselli et al., 2010).

In Fig. 6A, B and C we show the attitude of the BDT and temperature profiles for cross-sections T1, T2 and T3 respectively. For the two northernmost sections (T1 and T2) the surface heat flow decreases quite
sharply from west to east, from a value of about 90–80 mW/m² to a value of about 40–20 mW/m². This trend is also reflected in the distribution of temperatures, whose estimated values at 30 km in depth vary from 800 to 600 °C, toward west, to 600–400 °C, toward east. A more constant value of the surface heat flow is instead observed for the southernmost profile, where it has remained equal to about 30 mW/m² and the temperature at a depth of 30 km is about 400 °C.

5. Data integration and discussion

In order to explore the relation between the seismicity distribution, crustal structures, lithology and rheology for this sector of the Northern Apennines, we show in Fig. 7 a comprehensive picture constructed with all the above-described data. We project along the three transects (T1, T2 and T3 in Fig. 7A, B and C respectively) seismicity (background and seismic sequences), geological structures, lithologies and the brittle-ductile transitions.

All the three profiles show the presence of a first (i.e. shallower) brittle-ductile transition, 1BDT, placed at a depth of about 10–15 km and a second brittle-ductile transition, 2BDT (Figs. 7A–C). A ≈ 5 km thick, discontinuous intra-crustal layer with ductile behaviour separates these brittle layers. To better correlate the effects of the BDT with the observed cut-off of the seismic activity, we have reported histograms describing the earthquake depth frequency distribution and the mean depth of the BD-transitions for each of the three domains (extensional, transitional and compressional). The seismicity cut-off is represented as the D90, the depth above which 90% of the earthquakes occur.

As previously documented, the transition and compressive domains are not affected by intense seismic activity or relevant sequences and therefore data integration is limited. Within these limitations, we note that in the transition domain of T1 and T2, the scarce background seismicity is confined within the brittle layer with a good correspondence between 1BDT and D90. While in T3 the transition domain shows the shallower brittle layer not affected by seismic activity but the earthquakes occur at deeper depth (> 16 km) below the second brittle-ductile transition (2BDT in Fig. 7C). This seismicity, including 4 relatively major events (3.5 < Ml < 4.0) with extensional kinematics, is occurring beneath the top of the lower crust and can be interpreted as the signature of the scraping portion of the Adriatic lithosphere at the delamination hinge, as proposed by Carannante et al. (2013).
In the offshore portion of the compressive domain the lack of geological data and the limited seismic activity do not allow for a discussion on an integrated dataset. On the contrary, in the extensional domains the presence of multidisciplinary observations provides a unique opportunity to test the role of the: 1) brittle-ductile transition, 2) geological structure, and 3) lithology in the seismicity distribution within the

Fig. 5. Heat flux isolines (from Della Vedova et al., 2001) and main measurements points (yellow boxes).

![Map showing transects and heat flux isolines](image)

Fig. 6. A, B, C: Top - calculated and observed surface heat flux along the three transects (T1, T2 and T3). Down - temperatures distribution for the same transects.
upper crust. In all the three transects we observe a good agreement between the seismicity cut-off and the location of 1BDT. This is confirmed also by the histograms showing a clear correspondence between 1BDT and D90 (Fig. 7A–C). In detail, for transects T1 and T2 the good correspondence between the seismicity cut-off and 1BDT is evident in the eastern part of the extension domain whereas the seismicity present in the western part is shallower than 1BDT. Here the seismicity cut-off becomes more and more shallower from east to west defining an eastward dipping geometry that corresponds with the east dipping Altotiberina low-angle normal fault (AAF: Boncio et al., 2000; Collettini...
and Barchi, 2002; Chiaraluce et al., 2007). Thus, data integration for this part of the extension domain highlights a structural control in the seismicity cut off. This cut-off is exerted by the ATF fault that separates an active hanging-wall block from an almost aseismic footwall block. In addition, the comparison between the ATF geometry, as inferred from geological and seismological data, and the 1BDT confirms that the ATF is not the brittle-ductile transition: the ATF is an active low-angle normal fault within the brittle crust that merges into the BDT transition only at the boundary between the extension and transition domains.

Our integrated dataset allows for a further discussion on the relationship between lithology, BDT and the occurrence of the major instrumental earthquakes of the area. The study area has been affected by 3 major seismic sequences in the last 30 years: Gubbio 1984, Ms 5.3 (Colletti et al., 2003), Colfiorito 1997, Mw 6.0 (Chiaraluce et al., 2003) and Gualdo Tadino 1998, Mw 5.3 (Ciaccio et al., 2005). Apart from the Gubbio (1984) sequence, where the mainshock location is not well constrained and the aftershock sequence is located in the hanging-wall block of the ATF (Colletti et al., 2003), for the Colfiorito and Gualdo Tadino sequences, the earthquake locations are excellent and therefore can be used for an integrated analysis. For both sequences the mainshocks nucleated at depth between 5 and 6 km within the Triassic Evaporites and the majority of the aftershock occurrence is within the overlying carbonates (Miller et al., 2004; Ciaccio et al., 2005; Mirabella et al., 2008). The occurrence of the major instrumental earthquakes within the sedimentary cover of the Umbria-Marche succession (mainly Carbonates and Evaporites), well above the brittle ductile transition and substantially avoiding the Verrucano formation and the crystalline basement (lithological control, e.g. Mirabella et al., 2008) is a remarkable observation. Standard rheological models in fact predict the occurrence of the major earthquakes at the brittle-ductile transition where the accumulation of larger differential stresses facilitates the nucleation of larger earthquakes (e.g. Sibson, 1983; Scholz, 2002). One possibility to overcome this disagreement is that the larger earthquakes of the area, nucleating near the brittle-ductile transition and cutting across the entire brittle layer, are not contained within the 30 years of instrumental observation. The other possibility is that the sedimentary rocks are lithologies more prone to deform in a seismic way while the underlying phyllic basement is more suitable for fault creep. Within the brittle layer a fault is prone to earthquake nucleation if friction decreases with increasing sliding velocity, i.e. velocity weakening behaviour; if on the contrary friction increases, i.e. velocity strengthening behaviour, the fault deforms by stable sliding and fault creep (e.g. Marone, 1998; Scholz, 2002). Friction experiments on carbonates, some of them collected in the Umbria-Marche Apennines, have shown that slip localization, increasing sliding velocity and temperature favour a velocity weakening behaviour (e.g. Carpenter et al., 2014; Tesei et al., 2014; Verberne et al., 2014, 2015).

For the anhydrites and dolostones of the Triassic evaporites, brittle deformation is favoured by fluid overpressure (De Paola et al., 2009; Colletti et al., 2009) and localization with temperature favour a velocity weakening behaviour (Scuderi et al., 2013; Pluymakers and Niemeyer, 2015). On the contrary, laboratory experiments on phyllosilicate-rich rocks document a velocity strengthening behaviour (e.g. Ikari et al., 2011) and data from exhumed faults outcropping in the Apennines confirm this behaviour (Colletti et al., 2011). This lithologial control for the occurrence of the major earthquakes in the Umbria-Marche Apennines, favoured by the velocity weakening behaviour of carbonate-bearing faults outlaying a creeping portion of phyllichal basement, is supported by the observation that during the Gualdo Tadino and Colfiorito seismic sequences the majority of the earthquakes nucleated within the sedimentary rocks with a few earthquakes occurring within the phyllichal basements (Mirabella et al., 2008).

6. Conclusions

Seismicity distribution has classically represented one of the main geophysical proxies to test hypothesis on the structure, evolution and rheology of the continental crust. At the same time earthquakes are expected to occur in distinctive areas in reason of the existing faults distribution as well as their diverse phases of the seismic cycle. Our work integrating seismological data (background and seismic sequences), geological and geophysical data together with rheological inferences for a large portion of the Northern Apennines, demonstrate that the occurrence of crustal seismicity is influenced by several factors acting at different scales.

We have shown that at first approximation there is a good correspondence in the upper crust between the brittle ductile transition and seismicity cut-off, as well as between the presence of brittle layers at greater depths and the seismic activity. This implies that one of the main elements controlling seismicity occurrence is fault rheology that is mainly dictated by heat flow. At the same time, large geological structures can strongly influence seismicity distribution. In this work, within the seismogenic layer, we have documented a regional low-angle normal fault that separates a seismic active hanging-wall from an almost aseismic foot-wall block.

The observation that the largest earthquakes of the area, $5.5 < M < 6.0$, do not nucleate at the base of the brittle layer but well within the seismogenic layer at the base of the sedimentary cover, indicates an important role played by the lithology in mainshock occurrence. In particular the carbonates and the dolostones-anhydrites of the sedimentary cover seem to be more prone to host the nucleation of the largest earthquakes of the area.

Our work suggests that at regional scale the brittle ductile transition and therefore crustal rheology controls the occurrence of crustal seismicity, however at a more local scale crustal structures and the propensity of fault rocks to develop frictional instabilities are extremely relevant for a comprehensive understanding of the earthquake occurrence.

Acknowledgements

We thank Matt Ikari for a constructive review. This research was supported by ERC grant No. 259256 GASS.

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