-TITLE PAGE-

Did Adria rotate relative to Africa?

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For: Solid Earth

Abstract

The first and foremost boundary condition for kinematic reconstructions of the Mediterranean region is the relative motion between Africa and Eurasia, constrained through reconstructions of the Atlantic Ocean. The Adria continental block is in a downgoing plate position relative to the strongly curved Central Mediterranean subduction-related orogens, and forms the foreland of the Apennines, Alps, Dinarides, and Albanides-Hellenides. It is connected to the African plate through the Ionian Basin, likely with lower Mesozoic oceanic lithosphere. If the relative motion of Adria versus Africa is known, its position relative to Eurasia can be constrained through a plate circuit, thus allowing robust boundary conditions for the reconstruction of the complex kinematic history of the Mediterranean region. Based on kinematic reconstructions for the Neogene motion of Adria versus Africa, as interpreted from the Alps and from the Ionian Basin and its surrounding areas, it has been suggested that Adria underwent counterclockwise vertical axis rotations ranging from ~0 to 20°. Here, we provide six new paleomagnetic poles from Adria, derived from the Lower Cretaceous to Upper Miocene carbonatic units of the Apulian peninsula (southern Italy). These, in combination with published poles from the Po Plain (Italy), the Istria peninsula (Croatia), and the Gargano promontory (Italy), document a post-Eocene 9.5±8.7° counterclockwise vertical axis rotation of Adria. Our results do not show evidence for significant Africa-Adria rotation between the Early Cretaceous and Eocene. The Alpine and Ionian Basin end-member kinematic models are both permitted within the documented rotation range, yet are mutually exclusive. This apparent enigma could possibly be solved if one or more of the following conditions are satisfied: (i) Neogene shortening in the western Alps has been significantly underestimated (by as much as 150 km); (ii) Neogene extension in the Ionian Basin has been significantly underestimated (by as much as 420 km); and/or (iii) a major sinistral strike-slip zone has decoupled North and South Adria in Neogene time. Here we present five alternative reconstructions of Adria at 20 Ma, highlighting the kinematic uncertainties, and satisfying the inferred rotation pattern from this study and/or from previously proposed kinematic reconstructions.

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1. Introduction

The complex geodynamic evolution of the central Mediterranean region has been dominated by convergent motion between the African and European plates. Rather than being accommodated along a discrete plate boundary, the complex paleogeography of the region led to convergence being accommodated along segmented subduction zones, and to distributed overriding plate shortening. In addition, subduction roll-back since the late Eocene has formed a series of extensional back-arc basins and strongly curved subduction zones and associated mountain belts (e.g., Dewey et al., 1989; Doglioni et al., 1997; Gueguen et al., 1998; Jolivet et al., 2009; Rosenbaum and Lister, 2004; Stampfli and Hochard, 2009; Wortel and Spakman, 2000). It is this complex evolution that has made the Mediterranean region instrumental in the development of fundamental concepts that link surface deformation to deep mantle processes (Carminati et al., 2012; Cavazza et al., 2004; Doglioni, 1991; Faccenna and Becker, 2010; Govers and Wortel, 2005; Jolivet et al., 2009; Malinverno and Ryan, 1986; Wortel and Spakman, 2000).

Detailed kinematic reconstructions constitute a fundamental tool for advancing our understanding of the complex geodynamics of the Mediterranean region. A common boundary condition adopted by all reconstructions is represented by the relative motions summarized in the Eurasia-North America-Africa plate circuit based on marine magnetic anomalies of the Atlantic Ocean (e.g., Capitanio and Goes, 2006; Dewey and Sengör, 1979; Dewey et al., 1989; Gaina et al., 2013; Rosenbaum et al., 2002; Savostin et al., 1986; Seton et al., 2012; Torsvik et al., 2012; Vissers et al., 2013), which defined the area generated and consumed between Africa and Europe since the break-up of Pangea.

A critical element in Mediterranean reconstructions is the continental domain of Adria (Figure 1). Adria is a fragment of continental crust intervening the European and African plates composed of essentially undeformed platform carbonates currently exposed on the Apulia peninsula and Gargano promontory of southern Italy, the Istria peninsula of Croatia, and the Adige Embayment of the southern Alps (Figure 1). Adria is in a downgoing plate position relative to all surrounding mountain belts: it is overthrust by the Apennines in the west and the Dinarides-Albanides-Hellenides in the east, and although it was originally in an

Formatted: Indent: First line: 1.27 cm overriding plate position in the Alps, it became overthrust since Neogene time. Tectonic slices of the Adriatic upper crust are currently exposed in all circum-Adriatic mountain ranges (Bernoulli and Jenkyns, 2009; Faccenna et al., 2001; Gaina et al., 2013; Handy et al., 2010; Schmid et al., 2008; Stampfli and Hochard, 2009; Stampfli and Mosar, 1999; Ustaszewski et al., 2008; Vai and Martini, 2001; van Hinsbergen and Schmid, 2012) (Figure 1). To the south, Adria is separated from the North African passive continental margin by oceanic lithosphere of the Ionian Basin (Catalano et al., 2001; Frizon de Lamotte et al., 2011; Gallais et al., 2011; Speranza et al., 2012).

There is no zone of intense compression between Adria and Africa, and Adria has been paleolatitudinally stable relative to Africa within paleomagnetic error bars (of typically several hundreds of kilometres) (e.g., Channell et al., 1979; Rosenbaum et al., 2004). Because the motion of Adria relative to Europe would be the best boundary condition to reconstruct the central Mediterranean kinematic history since the Mesozoic, it is crucial to reconstruct any past relative motions between Adria and Africa. Different approaches to this end, however, led to contrasting results. The Ionian Basin's sea floor is widely regarded as Mesozoic (e.g., Catalano et al., 2001; Frizon de Lamotte et al., 2011; Gallais et al., 2011; Schettino and Turco, 2011; Speranza et al., 2012), implying a semi-rigid connection between Adria and Africa since that time. Eastward increasing Neogene shortening in the Alps (Schmid et al., 2013; Schönborn, 1999), however, has been used to infer a Neogene ~20° counterclockwise (ccw) rotation of Adria relative to Eurasia (Ustaszewski et al., 2008), but only ~2° of which can be accounted for by African-Europe plate motion. Based on this kinematic model, therefore, Adria must have been decoupled from Africa during the Neogene. GPS measurements suggest that at present, Adria moves NE-ward motion relative to Africa (D'Agostino et al., 2008), consistently with a NE-ward motion of Adria versus Africa of 40 km over the past 4 Myr inferred from kinematic reconstruction of the Aegean region (van Hinsbergen and Schmid, 2012). Conversely, Wortmann et al. (2007) argued for a Cenozoic 8° clockwise (cw) rotation of Adria versus Africa to avoid overlaps of Adria with Eurasia in pre-Cenozoic reconstructions, and Dercourt et al. (1986) postulated a 30° ccw rotation

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Paleomagnetic data can provide useful quantitative constraints on the vertical axis rotation history of Adria. However, published results from Adria's sedimentary cover yielded contrasting interpretations involving (i) no rotation (Channell, 1977; Channell and Tarling, 1975), (ii) 20° cw rotation since 30 Ma (Tozzi et al., 1988), (iii) 20° ccw rotation since the late Cretaceous (Marton and Nardi, 1994), or (iv) more complex models where a 20° ccw early-late Cretaceous rotation was followed by a late Cretaceous-Eocene 20° cw rotation and a post-Eocene 30° ccw rotation (Márton et al., 2010).

In this paper, we present a new paleomagnetic study of the Lower Cretaceous to Upper Miocene stratigraphy of the Apulian carbonate platform (southern Italy). We compare our results to, and integrate these with published datasets, and evaluate the range of paleomagnetically permissible rotations values in terms of their kinematic consequences for Central Mediterranean reconstructions.

2. Geological Setting

Prior to the onset of Africa-Europe convergence in the mid-Mesozoic, Adria was much larger continent stretching from the Italian Alps to Turkey (Vlahović et al., 2005). Gaina et al. (2013) introduced the term 'Greater Adria' for all continental lithosphere including many Mesozoic intracontinental rift basins and platforms that are now incorporated in the surrounding fold-thrust belts and that existed between the Vardar ocean (or Neotethys) and the Ionian Basin.

Greater Adria was separated from Eurasia in the northeast by the Triassic Vardar, or Neo-Tethys Ocean (Gaina et al., 2013; Schmid et al., 2008;) and in the north and west by the Jurassic Piemonte Ligurian, or Alpine Tethys Ocean (e.g., Favre and Stampfli, 1992; Frisch, 1979; Handy et al., 2010; Rosenbaum and Lister, 2005; Vissers et al., 2013). To the south the Ionian Basin separated Adria from Africa (Figure 1). Adria's conjugate margin across the Ionian Basin is likely the Hyblean Plateau of Sicily bounded to the east by the Malta escarpment (Catalano et al., 2001; Chamot-Rooke and Rangin, 2005; Speranza et al., 2012).

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Before the Calabrian subduction zone retreated away from Sardinia in the late Miocene (Cifelli et al., 2007; Faccenna et al., 2001; 2004; Rosenbaum et al., 2008), the Ionian Basin extended farther to the north-west. This oceanic lithosphere was at least Jurassic in age, as evidenced by off-scraped sediments now exposed in Calabria (Bonardi et al., 1988). The modern Ionian Basin is floored by a >5 km thick sequence of sediments, which in the west have been thrust in response to subduction below Calabria (the Calabrian accretionary prism), and in the east in response to subduction below the Aegean region (the 'Mediterranean ridge') (e.g., Finetti, 1985; Gallais et al., 2011; Minelli and Faccenna, 2010; Reston et al., 2002; Speranza et al., 2012). The Ionian abyssal plain is the only relatively undeformed portion that serves as the foreland of the central Mediterranean subduction systems (Gallais et al., 2011; Hieke et al., 2006; Speranza et al., 2012). Given the crustal thickness of 7-9 km (Chamot-Rooke and Rangin, 2005) and very low heatflow (Pasquale et al., 2005), this ocean floor is likely an old remnant of the Neotethys Ocean (e.g., Gallais et al., 2011; Speranza et al., 2012). The age of the Ionian Basin has been estimated to range from late Paleozoic to Cretaceous (Dercourt et al., 1986; Frizon de Lamotte et al., 2011; Gallais et al., 2011; Golonka, 2004; Robertson et al., 1991; Schettino and Turco, 2011; Sengör et al., 1984; Stampfli and Borel, 2002), with the most recent suggestion giving a late Triassic age (Speranza et al., 2012).

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Despite the uncertainties on the opening age and direction (NE-SW according to Chamot-Rooke and Rangin (2005) and Speranza et al. (2012) or NW-SE according to Frizon de Lamotte et al. (2011) and Gallais et al. (2011), there is general agreement that the Ionian abyssal plane has not been strongly deformed since the middle Mesozoic. Minor late Miocene inversion was associated with only a few kilometers of shortening (Gallais et al., 2011). Also the Malta escarpment has not been reactivated after the Mesozoic, and gently dips towards the basin floor, except to the northwest where the margin is reactivated in Pliocene and younger times, likely as a STEP fault accommodating Calabrian trench retreat (Argnani and Bonazzi, 2005). Late Miocene and younger NE-SW extension, however, has been documented within the African passive margin, forming the ~140 km wide Sicily Channel rift zone (Argnani, 2009) between Sicily and the Tunisian coast (Figure 1). This rift system is

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associated with up to ~40% crustal thinning and contains active rift-related volcanoes (Civile et al., 2008). The dimension of the rifted zone and the crustal attenuation may indicate some tens of kilometers of extension. This system connects to the SE to the Sirte and Tripolitana basins of Lybia (Capitanio et al., 2011) and was interpreted to reflect renewed late Miocene and younger NE-SW extension between Adria (and the Ionian Basin), and Africa, likely caused by slab-pull forces of the subducting African plate (Argnani, 1990; Belguith et al., 2013; Capitanio et al., 2011; Civile et al., 2010; Goes et al., 2004).

Our study area, the Apulian carbonate platform, hereafter called 'Apulia' (Figure 2), is part of Adria and lies in the Plio-Pleistocene foreland of the Apennine fold-thrust belt to the west (D'Argenio et al., 1973). Recent NE-SW, low-magnitude extension evident from Apulia (Figure 2) is interpreted to result from flexural bending of the downgoing Adriatic lithosphere into the Apennine subduction zone (Argnani et al., 2001; Doglioni et al., 1994). To the northeast, Apulia borders the Adriatic Sea, which represents the Late-Miocene-Quaternary foredeep of the Dinarides-Albanides-Hellenides belt (Argnani, 2013; Argnani et al., 1996; Bertotti et al., 2001; de Alteriis, 1995). The southwestern margin of Apulia appears to constitute a passive margin of the Ionian Basin in a narrow segment between the Calabrian prism and the Mediterranean ridge along the Apulian Escarpment (Finetti, 1985), where accumulation of sediment since the Mesozoic has compensated the thermal subsidence of the oceanic lithosphere (Channell et al., 1979; Ricchetti et al., 1998).

The northern margin of the platform is exposed on the Gargano promontory that was located close to the northestern transition of Apulia toward the adjacent Adriatic Basin (Bosellini et al., 1999b; Graziano et al., 2013; Santantonio et al., 2013). The Adriatic Basin, from which the present-day Adriatic Sea roughly inherited the location, was a Jurassic deep-water continental rift basin that continued northwestward into the Umbria-Marche basin, now incorporated in the Apennine fold-thrust belt, and southeastward into the Ionian Zone which is now part of the Hellenides-Albanides and should not be confused with the previously mentioned oceanic Ionian Basin, located on the opposite side of Apulia (Fantoni and Franciosi, 2010; Flores et al., 1991; Grandic et al., 2002; Mattavelli et al., 1991; Picha,

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2002; Zappaterra, 1990; 1994). Basin-transition units of Apulia have in Pliocene and younger times become incorporated in the Pre-Apulian zone of western Greece, exposed on the Ionian Islands which became separated from Apulia along the Kefallonia Fault Zone (Kokkalas et al., 2012; Royden and Papanikolaou, 2011; Underhill, 1989; van Hinsbergen et al., 2006).

To the north of Apulia, in the central Adriatic Sea, the fronts of the external Dinarides and Apennines converge, producing the Mid-Adriatic Ridge (Figure 1). Along this structure, the Adriatic Basin is cut by Neogene NW-SE striking thrusts, some of which invert Mesozoic extensional structures (Fantoni and Franciosi, 2010; Grandic et al., 2002; Kastelic et al., 2013; Scisciani and Calamita, 2009; Scrocca, 2006). South of the Mid-Adriatic Ridge, it is has been suggested that W-E or SW-NE strike-slip faults dissect the Adriatic Sea unlikely, The exact location and kinematics of these structures is controversial, but is primarily considered dextral in origin based on seismicity, low-resolution seismic lines and GPS velocities. As a result, three alternative zones have been suggested as the decoupling zones between North and South Adria. The first one is the Pescara-Dubrovnik line, whose presence was hypothesized by Gambini and Tozzi (1996), and that roughly corresponds to a segment of the boundary that, according to Oldow et al. (2002), borders two fragments of Adria with different GPS-measured velocity. The second one is the Tremiti Line of Finetti (1982) or the Tremiti Structure of Andre and Doulcet (1991), whose presence is evident from seismicity (Favali et al., 1993; 1990) and sea-floor deformation (Argnani et al, 1993). According to Doglioni et al. (1994) and Scrocca (2006), this dextral lithospheric structure segments Adria in order to accommodate a differential slab retreat, and, according to Festa et al. (2014), its subsurface expressions were enhanced by salt tectonics. Finally, also the Mid-Adriatic Ridge was interpreted to be a boundary between two different sectors of Adria (Scisciani and Calamita, 2009), assuming that some structural highs of the external Dinarides (i.e. the Palagruza High of Grandic et al., 2002) represent the southward prosecution of the same ridge in the eastern Adriatic Sea.

Apulia was considered an isolated carbonate platform that developed away from emerged continents (D'Argenio et al., 1973) until the discovery of dinosaur

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footprints that suggested the presence of some continental bridges between Apulia and other coeval exposed regions in Late Jurassic to Early Cretaceous time (e.g. Bosellini, 2002). During the Mesozoic, shallow water carbonate deposition was able to compensate the regional subsidence, and led to the accumulation of a stratigraphic succession up to 6000 m thick (Ricchetti et al., 1998). The succession, whose Cretaceous interval is widely exposed, consists mainly of dolomitic and calcareous rocks (Ricchetti, 1975). In the Murge area (Figure 2), where its age has been best constrained (Spalluto, 2011; Spalluto and Caffau, 2010; Spalluto et al., 2005), the succession forms a monocline dipping gently towards the SSW, thus exposing younger rocks from NNE to SSW (Ciaranfi et al., 1988) (Figure 2). This monoclinal succession is deformed by gentle undulations and steep normal and transtensional faults with an overall NW-SE orientation (Festa, 2003). The southernmost part of the exposed Apulia (i.e. the edge of the Salento Peninsula facing the Otranto Channel, Figure 2) represents the position of the Mesozoic platform margin (Bosellini et al., 1999b). It probably sharply passed to a southern intraplatform pelagic basin, recognized in the subsurface of the submerged Apulia (Del Ben et al., 2010). Post Cretaceous carbonate rocks cropping out along this Salento margin show well-preserved tens of meters thick clinoforms, i.e., slope deposits that formed along and rework rocks of the old Apulia margin (Bosellini et al., 1999b). These slope deposits reach up to 25/30° of primary non-tectonic dip (Bosellini, 2006; Tropeano et al., 2004).

3. Paleomagnetic sampling, analysis and results

3.1. Sampling and laboratory treatment

We collected 456 samples from nine localities covering the Cretaceous and Cenozoic carbonate stratigraphy of Apulia. Cores samples were collected with a gasoline powered motor drill and their orientation was measured with a magnetic compass.

The samples were measured at the Paleomagnetic Laboratory Fort Hoofddijk of Utrecht University, the Netherlands. The nature of the magnetic carriers was investigated for representative samples using an in-house developed horizontal translation type Curie balance, with a sensitivity of 5 x 10^{-9} Am^2 (Mullender et al., 1993). Approximately 60 mg of powder obtained from each sample was subjected to stepwise heating-cooling cycles up to 700°C.

For each locality, eight to ten samples were selected as pilot samples, and of each sample two specimens were retrieved for both thermal (TH) and alternating field (AF) demagnetization. AF demagnetization and measurement of the remanence were carried out using an in-house developed robotized sample handler coupled to a horizontal pass-through 2G Enterprises DC SQUID cryogenic magnetometer (noise level 1×10^{-12} Am^2) located in a magnetically shielded room (residual field <200 nT). Samples were demagnetized by stepwise AF treatment (alternating field steps: 5, 8, 12, 15, 20, 25, 30, 35, 40, 45, 50, 60, 70, 80 and 100 mT). Thermal demagnetizations were performed in a magnetically shielded oven using variable temperature increments up to 500°C. After each heating step the remanence was measured with a 2G Enterprises horizontal 2G DC SQUID cryogenic magnetometer (noise level $3x10^{-12}$ Am^2).

Thermal demagnetization treatment demonstrated to be more effective for the sampled rocks as it provided more stable demagnetization diagrams than the AF technique. The remaining samples of each locality were therefore thermally demagnetized.

Demagnetization diagrams were plotted on orthogonal vector diagrams (Zijderveld, 1967) and the characteristic remanent magnetizations (ChRMs) were isolated via principal component analysis (Kirschvink, 1980). Samples with a maximum angular deviation (MAD) larger than 15° were rejected from further analysis. Because secular variation of the geomagnetic field induces scatter in paleomagnetic directions whose distribution gradually becomes more ellipsoidal towards equatorial latitudes (Creer et al., 1959; Tauxe and Kent, 2004), we calculated site mean directions using Fisher (1953) statistics on virtual geomagnetic poles (VGPs) following procedures described in Deenen et al. (2011). At each locality a 45° cut-off was applied to the VGPs (Johnson et al., 2008). The results were then filtered by the paleomagnetic quality criteria of the N-dependent reliability envelope of Deenen et al. (2011). Mean values and statistical parameters are listed in Table 1.

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3.2 Results

Curie balance results are noisy because of the very low intensities of these carbonates, and do no not reveal meaningful information about the carriers of the remanence. Upon close inspection it can be seen that some new magnetic mineral is created upon heating, just above 400°C. This points to the presence of minor amounts of pyrite converted to magnetite. The cooling curves are higher than the heating curves, confirming that new magnetic minerals were created that were not fully removed upon heating to 700°C (Figure 3).

The very low NRM intensities of these limestones also cause nearly 30% of the demagnetized specimens (167) to show an erratic demagnetization pattern and many samples yielded no interpretable directions. Nevertheless, a total of 298 demagnetized specimens show a weak but stable and measurable remanence. In general, the lowest temperature steps (or AF steps) show a viscous or present-day overprint (Figure 4). After removing this overprint, the characteristic remanent magnetization (ChRM) directions can be interpreted. Most specimens show interpretable results up to temperatures of approximately 400-450°C. Above this temperature, intensities become too low or spurious magnetizations occur that hamper any further interpretation (e.g. Figure 4g). Of the more successful demagnetization diagrams, we use 8-10 successive temperature steps for the ChRM directions determined by principal component analysis.

<u>*Locality Petraro quarry (PA)*</u>

The Petraro quarry (PA) is located in NE Murge close to the town of Barletta (Figure 2). This section shows the oldest part of the Calcare di Bari Formation cropping out in the Murge area and consists of a well-bedded, 55 m-thick, shallow-water carbonate succession in which few dm-thick carbonate beds are irregularly alternated with a few m-thick dolomitic beds (Luperto-Sinni and Masse, 1984). Carbonate lithofacies are made up of biopeloidal wackestones/packstones and microbial bindstones with rare intercalations of biopeloidal and oolitic grainstones interpreted as formed in inner shelf peritidal environments. Dolomites consist of an anhedral or subhedral mosaic of dolomitic crystals, which totally or partly replaced

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the carbonate precursor. Based on the study of the microfossiliferous assemblage of PA (mostly benthic foraminifers and calcareous algae), Luperto-Sinni and Masse (1984) refer this succession to the Valanginian (~140-136 Ma; according to the geological time scale of Gradstein et al. (2012)). We sampled a 10 m-thick interval of this section and avoided dolomitic beds.

The NRM intensity of these samples is very low (30-300 µA/m) and stable ChRMs were isolated for only 39 specimens at temperature steps between 220 and 500 °C (Figure 4a-c). The ChRMs show both normal and reverse polarities, and yield a positive reversal test (Johnson et al., 2008; McFadden and McElhinny, 1990) (classification C; $\gamma = 15.9 < \gamma_c = 19.5$). The distribution of the ChRMs satisfies the quality criteria of representing PSV (i.e. $A95_{min}<A95<A95_{max}$; Deenen et al., 2011). The tilt corrected mean ChRM direction for this locality after a fixed 45° cut-off is D±ΔD=130.8±8.5°, I±ΔI=-23.4±14.6° (N=29, K=11.5, A95=10.9°) (Table 1 and Figure 5).

Locality Casa Rossa quarry (CR)

The Casa Rossa quarry (CR) is a large limestone quarry in the NE Murge area (Figure 2), located SW of Trani. The outcropping section consists of a well-bedded, more than 40 m-thick, shallow-water carbonate succession. Similarly to the Petraro quarry, carbonate beds consist of biopeloidal wackestones/packstones and microbial bindstones showing evidence of desiccation features (mud cracks and birdeyes) interpreted as formed in inner shelf peritidal environments. Interbedded with the carbonate lithofacies, there are few-mm thick green shale intercalations interpreted as palaeosols. Based on the study of the microfossiliferous assemblage of CR, Luperto-Sinni and Masse (1984) refer this succession to the Barremian to lower Aptian (~129-121 Ma).

We sampled a stratigraphic thickness of 20 meters in the lower part of the outcropping succession. The low intensity of these rocks (5-100 µA/m) did not allow to obtain high quality remanence components because of high MAD values) and the distribution of the isolated ChRMs is highly scattered, failing all the adopted

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quality criteria (Figure 5). The locality is therefore not considered for further analyses.

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Locality Cavallerizza quarry (CU)

The Cavallerizza quarry (CU) is located in the western Murge area (Figure 2), close to the town of Ruvo di Puglia. The outcropping section shows rudist biogenic beds, late Cenomanian in age (~98-94 Ma), belonging to the uppermost part of the Calcare di Bari Fm (Iannone and Laviano, 1980). Rudist beds are topped by a horizon of green clays, 1 m thick, interpreted as a palaeosol, which marks a regional unconformity covering the whole Turonian (~94-90 Ma). Peritidal limestones of the Calcare di Altamura Fm, Coniacian-Santonian in age (~90-83.5 Ma), overlie green shales and mark the recovery of carbonate marine sedimentation after the Turonian subaerial exposure. A total of 43 samples were collected from the lower, 15 m thick, grey-brown rudist limestones of the Calcare di Bari Fm. According to Laviano et al. (1998), upper Cenomanian rudist beds cropping out in the Ruvo area record the progradation of a rudist-inhabited margin into a shallow intraplatform basin. Samples are characterized by generally low intensities (10-290 μA/m), but show interpretable demagnetization diagrams (Figure 4d-e). The mean tilt corrected direction after applying a 45° cut-off to the ChRMs distribution is D±ΔD= 333.2±7.1°, I±ΔI= 44.9±8.0° (N=32, K=16.8, A95=10.2°) (Table 1 and Figure 5). The VGP scatter for this site is consistent with that expected from PSV ($A95_{min}$<A95<$A95_{max}$).

Locality Caranna quarry (CN)

The Caranna quarry (CN) is located in SE Murge (Figure 2), close to the town of Cisternino. The outcropping section consists of an about 20 m-thick succession of thin-bedded pelagic chalky limestones (microbioclastic mudstones to wackestones) containing planktonic foraminifers and calcispheres. According to Pieri and Laviano (1989) and Luperto-Sinni and Borgomano (1989), these deposits formed in relatively deep-water, distal slope environments in late Campanian to early Maastrichtian times (~78-69 Ma). All 45 samples were collected from the lower part of the outcropping succession. Only 30% of the analyzed specimens yielded interpretable demagnetization diagrams because of the low intensity of the NRM (8-34 μA/m). Stable ChRMs were isolated at low temperatures commonly not exceeding 280°C (Figure 4f-g) and their distribution provided a mean value of D±ΔD = 2.3±11.8°, I±ΔI = 51.7±10.7° (N=15, K=15.7, A95=9.9°) (Table 1 and Figure 5). Although the distribution of the ChRMs reflects a PSV-induced scatter, the obtained mean direction is not statistically different from the present day field direction (PDF; Figure 5) and is inconsistent with the expected Cretaceous inclinations. It is very likely that a recent magnetic overprint affected this site, and the obtained results are not considered further.

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Locality Porto Selvaggio cove (PS)

The succession of the Porto Selvaggio cove (PS) crops out in western Salento. It mostly consists of upper Campanian chalky limestones (~78-72 Ma), slightly dipping to the SE, overlying sub-horizontal shallow marine limestones and dolostones (Reina and Luperto-Sinni, 1994a). According to Mastrogiacomo et al. (2012) chalky limestones sampled in this study formed in an intraplatform basin and record the evidence of a syn-sedimentary tectonic activity, as shown by the occurrence of two horizons of soft-sediment deformation structures (slumps). Out of the 52 demagnetized specimens, 48 yielded interpretable diagrams for the calculation of the ChRMs (Figure 4h-i). The NRM of those samples is characterized by relatively low intensities (10-2000 μA/m) and both normal and reversed ChRM's that did not pass the reversal test (γ=29 > γ_c= 14.7) (McFadden and McElhinny, 1990). The mean normal polarity ChRM shows, after a fixed 45° cut-off, a D±ΔD = 357.7±10.3°, I±ΔI = 45.4±11.4° (N=23, K=10.1, A95=9.2°) (Table 1, Figure 5), very close to the present-day field, and likely the result of a recent overprint. The reverse polarity ChRMs yield a mean value that is statistically different from the present-day field direction (D±ΔD = 165.0±8.9°, I±ΔI = -18.4±16.2°, N=14, K=21.6, A95=8.8°; see *Table 1*). The distribution of the reverse polarity ChRMs satisfies our criteria. Accordingly, only the reversed polarity ChRM is used for further analyses.

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Locality Massafra (MA)

This locality was sampled from a road cut close to the town of Massafra in the south of Murge (Figure 2). We sampled a 15 m-thick stratigraphic interval mostly comprising well-bedded white to light-brown shallow-water limestones with a Maastrichtian age (72-66 Ma) (Reina and Luperto-Sinni, 1994b). Sampled limestones mostly consist of peritidal, mud-supported, biopeloidal mudstones and wackestones showing a benthic microfossiliferous assemblage (mostly benthic foraminifers and ostracodes). The NRM intensity in those samples is relatively low (0.08-6 mA/m) and only 18 samples yielded interpretable demagnetization diagrams (Figure 4). The mean direction of the isolated ChRMs in tilt-corrected coordinates is D±ΔD = 8.8±10.7°, I±ΔI = 46.6±11.4° (N=17, K=15.2, A95=9.4°) (Table 1 and Figure 5). Before tilt correction, this direction is not statistically different from the present-day field (D±ΔD=359.8±12.6°, I±ΔI=52.2±11.2°, N=17, K=12.4, A95=10.5°) and is probably the effect of a recent overprint. Accordingly, this site is not considered for further analyses.

Locality Torre Specchialaguardia (TS)

An about 10 m-thick succession of clinostratified breccias and bioclastic deposits was sampled at the Torre Specchialaguarda locality (TS) in E Salento (Figure 2). This succession belongs to the Upper Eocene (Priabonian, 44-38 Ma) Torre Specchialaguardia Limestone Fm (Parente, 1994), which formed in a steep forereef slope onlapping a rocky Cretaceous to Eocene paleocliff (Bosellini et al., 1999b). According to Parente (1994) and Bosellini et al. (1999b), this formation is the oldest non-deformed unit in eastern Salento and its current dip of ~30° to the ESE is a primary, non-tectonic orientation. A total of 56 samples yielded NRM intensities ranging between 0.15 and 3 mA/m and usually gave stable demagnetization diagrams characterized by curie temperatures around 420°C (Figure 4l-n). The remanence displays both normal and reverse polarities that pass the reversal test (McFadden and McElhinny, 1990) (classification C, $\gamma = 8.2 < \gamma_c = 11.7$). After a fixed 45° cut-off, the mean in situ ChRM direction is D±ΔD = 356.0±5.6°, I±ΔI = 44.8±6.3° (N=47, K=17.9, A95=5.1°) (Table 1, Figure 5) and the ChRMs distribution satisfies our criteria.

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Locality Castro (OC)

An about 10 m-thick section was sampled close to the village of Castro (OC) in E Salento (Figure 2). The outcropping succession consists of Upper Oligocene (Chattian, 28-23 Ma) limestones belonging to the Castro Limestone Fm (Bosellini and Russo, 1992; Parente, 1994). This unit represents a fringing reef complex and shows a very well-preserved lateral zonation of the reef subenvironments (Bosellini and Russo, 1992; Parente, 1994). The sampled section shows clinostratified bioclastic deposits belonging to the reef slope subenvironment showing no evidence of tectonic deformation (Bosellini and Russo, 1992). Very low NRM intensities characterize these rocks (15-180 μA/m) and stable ChRM components with maximum unblocking temperatures between 220-500°C were isolated from 31 specimens (Figure 4o-p). The mean ChRM direction after a fixed 45° cut-off is D±ΔD = 180.5±3.2°, I±ΔI = -44.2±3.7° (N=29, K=85.8, A95=2.9°) (Table 1 and Figure 5). The VGP distribution does not entirely satisfy our criteria, since the A95 value is lower than A95min, indicating that PSV is underrepresented. The reverse polarity of the ChRMs and their low inclinations excludes a present-day (or recent) overprint, and the underrepresentation of PSV may be the result of some averaging PSV within each limestone sample.

Locality Novaglie (MN1-3)

Three different sites belonging to the Lower Messinian succession of the Novaglie Fm were sampled within three km of each other, close to the eastern Salento coast (Figure 2). The outcropping successions consist of in situ coral reef bioconstructions, clinostratified breccias and associated bioclastic and lithoclastic prograding slope deposits and fine-grained, bioclastic base-of-slope calcarenites. Similarly to the previous two localities, the bedding attitude in the sampled sites is most likely primary (Bosellini et al., 1999b; 1999a; 2001; Vescogni, 2000). At each sub-site 20 samples were collected from a 10 m-thick interval. NRM intensities range between 9 and 5000 μA/m. A total of 16, 13, and 6 ChRMs were successfully isolated from sub-site MN1, MN2, and MN3, respectively (Figure 4q-s). Overall, the direction of the isolated ChRMs is substantially scattered, with both normal and reverse polarities. The reversal test yielded a negative result (McFadden and McElhinny, 1990), therefore separate mean values were calculated at each sub-site. After a fixed 45° cut-off, site MN1 yielded a mean paleomagnetic direction of D±ΔD = 355.1±12.5°, I±ΔI = 61.1±7.9° (N=14, K=19.5, A95=9.2°) (Table 1, Figure 5). The VGP distribution passes our quality criteria. Only 6 specimens of MN 2 yielded a poorly-defined ChRM, with a dispersion well beyond our quality criteria (Figure 5). This sub-site was discarded. The large scatter of the ChRMs of sub-site MN3 yields, after the 45° cut-off, eight samples with a mean ChRM of D±ΔD= 2 22.5±13.7°, I±ΔI = -33.2±20.2° (N=8, K=19.0, A95=13.0°) (Table 1, Figure 5). Despite the low number of specimens, the A95 envelope passes the Deenen et al. (2011) criteria (Table 1).

4. Discussion

4.1 Paleomagnetic constraints on the rotation of Adria

Reliable paleomagnetic poles were obtained from six localities (out of nine) sampled throughout Apulia (Figure 2). The results from three localities were discarded because the distribution of the isolated ChRMs did not match the adopted quality criteria or because of a present-day overprint. One more site (MN3), although passing the quality criteria, yielded an anomalous declination (042.5±13.7°) indicating a strong clockwise rotation, not seen in the rest of the reliable sites. The anomalous direction at site MN3 may be explained considering that the samples, collected in a forereef breccia, could represent a large fallen block within the Messinian slope deposits. Regardless of the cause of this local rotation, we consider this direction not meaningful for the analysis of the regional rotation of Adria.
The rotation of Adria and its relationship with the African plate has always been a moot point (Márton et al., 2003, 2008; Caporali et al., 2000). Our new data provide new constraints for the rotation of Adria during the Cenozoic and, more importantly, can test the robustness and reliability of the available dataset.

The results of the Oligocene site OC (Figure 2) can be compared with those obtained by Tozzi et al. (1988) from the same area. These authors interpreted the

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Formatted: Indent: First line: 1.27 cm local ~30° ESE-ward bedding dip as a result of tectonic tilting, inconsistent with sedimentological studies (e.g., Bosellini, 2006), and calculated a post-Paleogene ~25° cw rotation of Adria by restoring this bedding to the horizontal. The paleomagnetic direction should be interpreted in *in situ* coordinates, and our results as well as those of Tozzi et al. (1988) are coincident and indicate no, or a minor counterclockwise post-Oligocene rotation of Adria with respect to Africa (Figure 6).

To assess whether and when Adria rotated relative to Africa, we combine our results with published data from Apulia, Gargano, Istria and the Adige Embayment, and compare them to the expected directions for the European and African plates calculated from the Global APWP of Torsvik et al. (2012) using a reference location of 40.7°N, 17.2°E (Table 1, Figure 7). Mean paleomagnetic directions and statistical parameters from the existing database were re-calculated at each site by averaging VGPs obtained through parametric bootstrap sampling using the provided mean values and statistical parameters (Table 1). This procedure overcomes the loss of information on the original data scatter that occurs when only the mean paleomagnetic direction at a given locality is computed by averaging site averages. In addition, sites with different numbers of samples should weigh differently, since large datasets provide a better representation of PSV than small data sets (see Deenen et al., 2011).

The updated paleomagnetic database is composed of twelve poles from Apulia (Marton and Nardi, 1994; Scheepers, 1992; Tozzi et al., 1988), five from the Gargano promontory (Channell, 1977; Channell and Tarling, 1975; Speranza and Kissel, 1993; Vandenberg, 1983), twelve poles from the Adige Embayment in the foreland of the Southern Alps (Márton et al., 2010; 2011), and eight poles from the Istria peninsula of Croatia (Marton et al., 2008; Márton et al., 2003) (Table 1). At six out of twelve localities from the Adige Embayment PSV is underrepresented ($A95<A95_{min}$; Table 1). We assume that this is a result of within-sample averaging due to low sedimentation rates and have included these sites in our analysis.

Figure 7a shows all declinations vs. age, from all four sectors of Adria. Approximately 40% of the poles are not statistically different from the expected African declinations. The remaining poles, representing the majority of the dataset, consistently show small counterclockwise deviations from the African APWP. The data provide no support for significant rotations between the northern and southern sectors of Adria.

To calculate the magnitude of rotation of Adria with respect to Africa we combine the data sets from the different regions. We used two approaches. One approach is to calculate a full-vector (six-point sliding window) moving average at every data point, from which we determined the D values and a ΔD error envelope. The other approach is to calculate a (fourth order) polynomial best-fit based on declination values only (Figure 7b). Both approaches show a remarkably coincident pattern that display a systematic ccw deviation of the mean declination of Adria relative to Africa from the entire Early Cretaceous to Late Cenozoic time interval. We interpolated the declination curve of the APWP of Africa (Torsvik et al., 2012) to obtain the declination at the ages corresponding to our moving average, and determined the difference at each data point. This yields an average deviation of all data of 9.5 ± 8.7° ccw.

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This obtained magnitude is accidentally comparable to the total rotation of Adria calculated from the upper Cretaceous of the Adige Embayment and Istria by Márton et al. (2010). These authors, however, interpreted their total rotation as the result of two distinct phases of cw and ccw rotation. In particular, an average of Eocene rocks was interpreted by Márton et al. (2010) to show 30° ccw rotation of Adria versus Africa. They suggested a ~20° cw rotation of Adria between the Cretaceous and Eocene, followed by a post-Eocene ~30° ccw rotation. These Eocene poles are included in our analysis, but taking all available data into account, we see no solid ground for interpreting significant rotation phases between the early Cretaceous and the late Cenozoic.

In summary, paleomagnetic data allow for a counterclockwise rotation of Adria relative to Africa anywhere between negligible (1°) and quite significant (18°) values, but with a very consistent average of 9.5°. The timing of this rotation is ill constrained, but can be estimated from the average declination (Figure 7) since roughly 20 ± 10 Ma.

4.2 Regional kinematic implications

The rotation pattern of Adria as emerging from this study can now be interpreted in the wider context of the central Mediterranean region. Our compilation of new and published paleomagnetic data do not lend support to models that infer either large Cretaceous vertical axis rotations (Dercourt et al., 1986; Márton et al., 2010) or a small cw rotation (Wortmann et al., 2007). We observe that two major types of scenarios can be accommodated within the range of rotation documented in this study (i.e. 1-18° ccw). One type of scenario is put forward from an Alpine point of view (post-20 Ma, ~20° ccw rotation of Adria relative to Europe around an Euler pole in the western Alps, corresponding to a ~17° ccw rotation of Adria relative to Africa). The other type derives from an Ionian Basin point of view (assuming near-rigidity between Africa and Adria and hence no differential rotation, according to Rosenbaum et al. (2004)). The paleomagnetically permissible rotation range derived here, can therefore not discriminate the two end-member kinematic scenarios for Adria. Accordingly, we will show the kinematic consequences of the permitted minimum and maximum rotation of Adria as a function of the location of its Euler pole.

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An Euler pole for the relative motion between Adria and Eurasia located at 45.0°N, 6°4E, near the city of Torino was computed by Ustaszewski et al. (2008) based on westward decreasing Neogene shortening in the Alps, and northward underthrusting of Adria below the southern Alps. Their inferred 20° ccw rotation relative to Eurasia translates to a paleomagnetically permitted ~17° ccw rotation of Adria relative to Africa. Assuming internal rigidity of Adria, a rotation around this pole by 17° would require up to 420 km of ENE-WSW extension in the Ionian Basin measured at the modern southeasternmost tip of stable Adria along the Kefallonia Fault (Figure 8a). This scenario would require that the entire Ionian Basin is Miocene in age, inconsistent with any of the inferred ages that range from Permian to Cretaceous. Similarly, a 9.5° rotation of Adria (average rotation constrained by our paleomagnetic analysis) would yield ~230 km of ENE-WSW extension, still much higher than what is geologically documented (Figure 8b).

Adria could rotate ccw without extension in the Ionian Basin if the Adria-Africa Euler pole is located in the far southeast of Adria (Figure 8c). Assuming Adriatic rigidity, and applying the 17°ccw rotation derived from reconstructions of the Alps, this would, however, lead to a reconstructed overlap of (i.e. major Neogene extension between) Adria and the Dinarides and Hellenides, and predicts >400 km of E-W convergence in the western Alps. In contrast to this scenario, geological data is consistent with continuous convergence of Adria relative to the Dinarides and Hellenides. In addition, the amount of westward retreat of in the western Alps is much smaller than required by this scenario (e.g., Ustaszewski et al., 2008; van Hinsbergen and Schmid, 2012) (Figure 8c).

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Alternatively, we may explore the maximum amount of rotation around the Euler pole constrained by Ustaszewski et al. (2008) that is permitted by the available structural evidence from the Sicily Channel and kinematic reconstructions of the Aegean region, and that is comparable with the minimum permissible rotation documented by this study. The ~40% crustal attenuation in the 140 km wide Sicily Channel (Civile et al., 2008) would suggest a (maximum) amount of NE-SW latest Miocene to Plio-Quaternary extension between Adria and Africa of ~40 km. A similar amount of Adria-Africa relative motion since the Early Pliocene was inferred to have occurred from a kinematic reconstruction of the Aegean region by van Hinsbergen and Schmid (2012) to avoid overlaps between Adria and the west-Aegean fold-thrust belt. This corresponds to a 1.7° ccw rotation of Adria. This reconstruction is consistent with the geological record of the circum-Ionian Basin, but it would require ~150 km NW-SE directed convergence between Adria and Europe since 20 Ma, to be accommodated in the Western Alps, in contrast with widely accepted lower values of no more than some tens of km (Ustaszewski et al., 2008; Handy et al., 2010) (Figure 8d).

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Our discussion above identifies an inconsistency between the kinematic interpretations from the geological record of the Alps and the Ionian Basin. Paleomagnetic data permit the scenarios that fulfill the constraints from both regions, but these scenarios cannot be reconciled with each other. Moreover, the

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average rotation suggested by the paleomagnetic data violates both end-member scenarios.

Since a key assumption in the above analysis is the rigidity of Adria, we explore a final scenario whereby we decouple north and south Adria, e.g. along the Mid-Adriatic Ridge, or along the Tremiti fault (Figure 2). Applying the reconstruction of Ustaszewski et al. (2008) for North Adria (17° ccw rotation), and the reconstruction of van Hinsbergen and Schmid (2012) for South Adria (1.7° ccw rotation). This would require as much as 160 km of left-lateral strike-slip between North and South Adria, and none of the identified structures appear likely candidates to accommodate such major displacements (Figure 8e).

The discussion above indicates that, although scenarios based on kinematic interpretations from both the Alps and the Ionian basin infer Neogene Adria-Africa relative rotations that are within the range documented in this study, these scenarios are mutually exclusive. Paleomagnetic data alone - with the error envelope calculated here - cannot solve this “Adriatic enigma”, but calls for a reassessment of the kinematics of three areas centered around three questions: (i) Since shortening reconstructions may underestimate the true amount of convergence: is the amount of Neogene shortening in the western Alps significantly underestimated? (ii) Is it possible to quantify the timing and amount of potential strike-slip zones separating a North and South Adria block? (iii) Is it possible that there is a large amount of Neogene extension along the Apulian escarpment, perhaps hidden below the advancing Calabrian prism and the Mediterranean ridge?

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5. Conclusions

We provide six new paleomagnetic poles from the Lower Cretaceous to Upper Miocene of the Murge and Salento areas of the Apulian Platform, southern Italy. These new data, combined with recalculated published poles from the Gargano promontory, the Istria peninsula of Croatia, and the Adige Embayment of the southern Alps, constrain a counterclockwise rotation of Adria relative to Africa at 9.5±8.7°, occurring sometime after 20 ± 10 Ma. Our revised paleomagnetic database for Adria discards significant rotations of Adria versus Africa between the Early

Cretaceous and the Eocene, as invoked by several studies. The permissible rotation magnitude (1-18° counterclockwise) is consistent with two end-member models for the Central Mediterranean region requiring (i) a Neogene ~18° counterclockwise rotation of Adria relative to Africa (based on kinematic reconstruction of the Alps), or (ii) negligible rotation of Adria based on kinematic reconstruction of the Ionian Basin. Although paleomagnetic data from Adria are not in disagreement with both models, we establish that these scenarios are mutually exclusive. We cannot solve this enigma, but call for kinematic studies that will address the following three key questions: (i) Was Neogene shortening in the western Alps significantly underestimated? (ii) Was Neogene extension in the Ionian Basin significantly underestimated? (iii) Was a North Adria block decoupled from a South Adria block along a large-offset sinistral strike-slip fault? Resolving these questions may lead to a solution of the conundrum associated with the kinematics of Adria.

Acknowledgements

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Figure 1. Regional tectonic map of the Mediterranean region. AE = Adige Embayment; AEs = Apulian Escarpment; Ga = Gargano; HP = Hyblean Plateau; IAP = Ionian Abyssal Plain; II = Ionian Islands; Is = Istria; KFZ = Kefallonia Fault Zone; KP = Karaburun Peninsula; MAR = Mid-Adriatic Ridge; ME = Malta Escarpment; TF = Tremiti Fault.

Figure 2. A simplified geological map of the exposed Apulian Foreland (Apulia, southern Italy) indicating our new, and previously published paleomagnetic sampling sites (modified from Pieri et al., 1997). Numbers and codes correspond to sites listed in Table 1.

Figure 3: Thermomagnetic curves measured on a Curie balance (Mullender et al., 1993) for representative samples. Arrows indicate heating (red) and cooling (blue).

Figure 4: Orthogonal vector diagrams (Zijderveld, 1967), showing representative demagnetization diagrams for all sampled sites. Except for OC, TS and MN all sites are in tilt corrected coordinates. Closed (open) circles indicate the projection on the horizontal (vertical) plane.

Figure 5: Equal area projections of the VGP (left) and ChRM directions (right) of all sites in both in situ and tilt corrected coordinates. Open (closed) symbols corresponds to the projection on the upper (lower) hemisphere. Large dots in the ChRM plots indicate the mean direction and relative cone of confidence ($\alpha 95$). Red (small) dots indicate the individual directions rejected after applying a 45° cut-off. Geen asterisk (*) indicates the present-day geocentric axial dipole (GAD) field direction at the sampled location. Sites TS, OC, and MN were sampled in sediments with a primary bedding attitudes, and should be considered in *in situ* coordinates.

Figure 6: Equal area projections of both in situ and tilt corrected ChRMs from our site OC (left) and from the same locality of Tozzi et al. (1988) (right), illustrating the apparent clockwise rotation that would result from a tilt correction of the bedding at this locality. The strata here have a primary dip (Figure 7) and should be considered in *in situ* coordinates. Symbols are as in Figure 5.

Figure 7: a) Age (Ma, following the timescale of Gradstein et al. (2012)) vs. declination plot for our new, and published data from Adria. The error envelope for the African and Eurasian APWPs are from Torsvik et al. (2012). Vertical error bars correspond to the ΔD calculated at each site (Table 1); horizontal error bars correspond to age uncertainty. All data are recalculated to a reference location (40.7°N, 17.2°E). Numbers and site abbreviations correspond to data entries in Table 1. b) Same data, with a polynomial 4th order trend line (purple dashed line) and the declination component of a full-vector 6-point moving average with error bars (ΔD) (red line).

Figure 8: Reconstructions at 20 Ma for rotation scenarios of Adria versus Africa that are permitted by paleomagnetic data, using (a) the rotation pole of Adria versus Eurasia of Ustaszewski et al. (2008), and a 20° ccw rotation of Adria versus Europe, corresponding to 18° ccw rotation of Adria versus Africa, constrained from shortening reconstructions in the Alps by the same authors; (b) a 9.5° ccw rotation of Adria versus Africa around the same pole corresponding to the paleomagnetically constrained mean rotation in this paper; (c) a 9.5° ccw rotation of Adria versus Africa corresponding to the paleomagnetically constrained mean rotation in this paper, around a pole located in the SE of Adria and allowing minimum relative latitudinal motion between Africa and Adria; (d) a 1.7° ccw rotation of Adria versus Africa around the pole of Ustaszewski et al. (2008), corresponding to the maximum amount of Neogene extension documented in the Sicily Channel; (e) Decoupling north and south Adria along the Mid-Adriatic Ridge of Scisciani and Calamita (2009), with North Adria following a scenario as in (a), and South Adria following a scenario as in (d). Reconstructions of: i) the Adriatic front of the Alps, Carpathians and

Dinarides follow Ustaszewski et al. (2008), ii) the western Mediterranean region follow Faccenna et al. (2004), and van Hinsbergen et al. (2014), and iii) the Aegean-Albanian region follow van Hinsbergen and Schmid (2012). Reconstruction is given in a Europe-fixed frame, with the position of Africa determined using rotation poles of Gaina et al. (2002) and Müller et al. (1999) for the North Atlantic and Central Atlantic Ocean, respectively.

Table 1: Table showing all the data of our, and published paleomagnetic results for Adria. Areas are indicated in Figure 1. Lat = Latitude of the site; Lon = Longitude of the site; Ages are assigned based on the biostratigraphy of the sites, and translated into numerical ages using Gradstein et al. (2012); Nd = number of demagnetized specimens; Ni = number of interpreted specimens; N45 = total number of specimens that fall within the 45° cut-off; D = declination; ΔDx = error on declination sensu Butler (1992); I = Inclination; ΔIx = error in inclination sensu Butler (1992). Statistical parameters are given by a cone confidence using Fisher (1953) statistics on virtual geomagnetic poles (A95, K) and directions (α95, k). $A95_{min}$ and $A95_{max}$ correspond to the confidence envelope of Deenen et al. (2011). Right-hand panel: directions and errors at a reference location 40.7°N, 17.2°E.

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