TECTONICS, VOL. 23, TC4003, doi:10.1029/2004TC001623, 2004

Driving mechanism of tectonic activity in the northern Apennines: Quantitative insights from numerical modeling

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Received 7 January 2004; revised 16 April 2004; accepted 7 May 2004; published 9 July 2004.

[1] It is shown, by numerical modeling, that the recent deformation pattern observed in the northern Apennines, mainly characterized by progressive eastward migration and bowing of the belt, thrusting activity along its external front, and tensional tectonics in the internal area, can be reproduced, at a first approximation, by applying a belt-parallel (SE-NW) compression to the chain, which is simulated as a structural system characterized by a high mechanical strength and decoupled from the surrounding zones. The above compressional regime, obtained by imposing kinematic boundary conditions to the model, causes the outward escape of crustal wedges from the chain, in particular the northeastward displacement of the Romagna-Marche-Umbria Units and the counterclockwise rotation and northwestward displacement of the Ligurian Units. This kinematics produces compressional to transcompressional strain along the external front of the chain and tensional to transtensional strain in the internal area, in line with the observed features that concern both strain style and orientation of principal strains. Evidence and arguments supporting the kinematic boundary conditions and the model parameterization adopted in modeling are discussed. Numerical experiments have also been carried out to evaluate the influence of major features of the model parameterization and boundary conditions we have adopted in modeling and to provide insights into the possible influence of strong decoupling earthquakes in the central Apennines on tectonic and seismic activity in the northern Apennines. *INDEX TERMS:* 8110 Tectonophysics: Continental tectonics—general (0905); 8102 Tectonophysics: Continental contractional orogenic belts; 8107 Tectonophysics: Continental neotectonics; 8123 Tectonophysics: Dynamics, seismotectonics; 8164 Tectonophysics: Stresses—crust and lithosphere; *KEYWORDS:* northern Apennines, numerical modeling, extrusion tectonics, Mediterranean. **Citation:** Viti, M., J. De Luca, D. Babbucci, E. Mantovani, D. Albarello, and F. D'Onza (2004), Driving mechanism of tectonic activity in the northern Apennines: Quantitative insights from numerical modeling, *Tectonics*, *23*, TC4003, doi:10.1029/2004TC001623.

1. Introduction

[2] The northern Apenninic chain (Figure 1) is constituted by remnants of the Alpine belt and by younger accretionary units stacked since the Oligocene along the western consuming border of the Apulian/Adriatic domain [e.g., *Dercourt et al.*, 1986; *Castellarin and Vai*, 1986; *Boccaletti and Sani*, 1998; *Finetti et al.*, 2001]. The last tectonic phase, which has led to the present morphological setting of the northern Apenninic arc, started around the late Miocene. Since that time, this chain has experienced an eastward migration of several tens of kilometers and a progressive increase of bowing, as suggested by geological, geophysical and paleomagnetic observations. During this phase, thrusting has occurred along the external front of the belt, while several troughs have generated in the internal area [*Elter et al.*, 1975; *Castellarin and Vai*, 1986; *Speranza et al.*, 1997; *Costa*, 2003].

[3] Several driving mechanisms have been proposed for the above deformation pattern, such as gravitational collapse of an overthickened accretionary prism [e.g., *Carmignani and Kligfield*, 1990], a slab pull mechanism [e.g., *Patacca et al.*, 1993] and extrusion tectonics driven by the convergence of the confining blocks [e.g., *Mantovani et al.*, 1997a, 2002; *Mantovani*, 2004]. The last interpretation suggests that the deformation pattern of the northern Apenninic arc was produced by the lateral escape of crustal wedges, driven by the constrictional regime induced by the convergence of the confining plates. The physical plausibility of extrusion processes is now largely accepted [e.g., *Tapponnier*, 1977; *Mantovani et al.*, 2001b, 2002; *Hubert-Ferrari et al.*, 2003]. When long and narrow orogenic systems undergo a compression parallel to their main trend, the extrusion pattern involves the lateral escape of several small-scale wedges, decoupled by strike-slip faults, resulting in a wide-scale bowing of the belt [*Mantovani et al.*, 1997a, 2002; *Gelabert et al.*, 2002; *Mantovani*, 2004]. In these cases, crustal stretching occurs in the wake of the extruding wedges, with the opening of relatively large back arc basins. This mechanism has been proposed as responsible for the formation of the major trench-arc-back arc systems in the Mediterranean and circum-Pacific regions [*Mantovani et al.*, 2001b; *Mantovani*, 2004].

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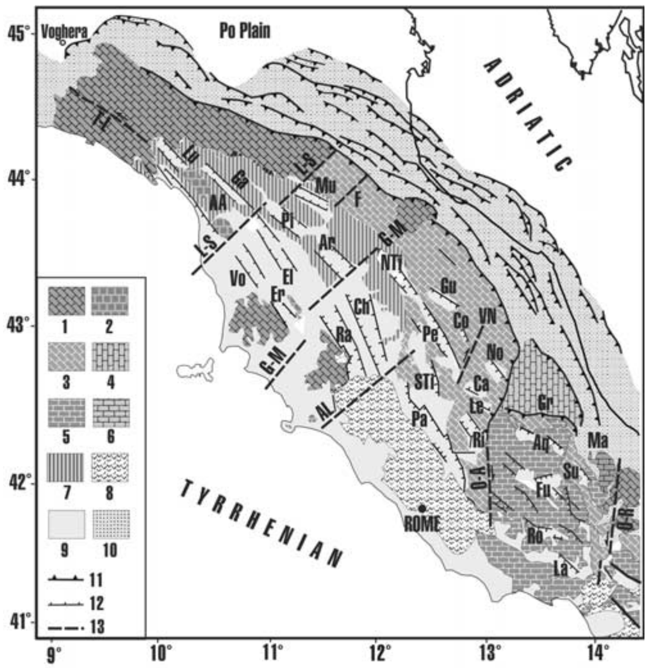

Figure 1. Tectonic sketch and main structural units of the central northern Apennines. Units numbered as follows: 1, Ligurian Units; 2, Tuscan metamorphic complex; 3, Romagna-Marche-Umbria Units; 4, Laga Units; 5, Latium-Abruzzi platform; 6, Maiella Mountains; 7, Tuscan Units; 8, magmatic products; 9, zones affected by Plio-Quaternary extension; 10, zones affected by Plio-Quaternary thrusting and folding, mostly buried beneath the foredeep sedimentary cover; 11, 12, and 13, major compressional, tensional and transcurrent features. Tectonic troughs are labeled as follows: Aq, L'Aquila; Ar, Arno; Ca, Cascia; Ch, Chiana; Co, Colfiorito; El, Elsa; Er, Era; Fu, Fucino; Ga, Garfagnana; Gu, Gubbio; La, Latina; Le, Leonessa; Lu, Lunigiana; NTi, Northern Tiber; Mu, Mugello; No, Norcia; Pa, Paglia-Tevere; Pe, Perugia-Foligno; Pi, Pistoia-Firenze; Ra, Radicofani; Ri, Rieti; Ro, Val Roveto; STi, Southern Tiber; Su, Sulmona; Vo, Volterra. Transversal lines are labeled as follows: AL, Albegna; F, Forlì; G-M, Grosseto-Val Marecchia; L-S, Livorno-Sillaro; O-A, Olevano-Antrodoco; O-R, Ortona-Roccamonfina; T-L, Tortona-Lunigiana; VN, Valnerina. AA, Alpi Apuane metamorphic complex; Gr, Gran Sasso thrust front; Ma, Maiella massif.

[4] Quantitative support to the above interpretation has been provided by numerical modeling [*Mantovani et al.*, 2001a], which shows that adopting kinematic boundary conditions compatible with the convergence of the confining plates allows quantitatively reproducing the major features of the recent/present strain field observed in the central eastern Mediterranean area.

[5] In this work, the same numerical approach is used at a smaller scale to quantitatively demonstrate that the major features of the recent/present strain field observed in the northern Apennines can be reproduced as an effect of a belt-parallel compressional stress field. A synthetic description of the recent deformation pattern and the post-Miocene evolutionary history of northern Apennines is provided in sections 2 and 3, as a necessary premise to numerical modeling. The observed strain to be simulated, the model parameterization and the boundary conditions adopted, along with the arguments and evidence that have led to their choice are described in section 4a. Numerical experiments are reported in sections 4b, 4c, and 5.

2. Recent Deformation Pattern of Northern Apennines

[6] The major aspects of the recent deformation pattern in the study area can be synthesized as follows:

[7] 1. Folding and thrusting up to the lower Pleistocene is well recognized along the external border of the belt by field and subsurface data [*Costa*, 2003; *Centamore and Nisio*, 2003, and references therein]. Seismicity [*Selvaggi et al.*, 2001; *Viti et al.*, 2001] and in situ stress measurements [*Montone and Mariucci*, 1999] suggest that tectonic activity, with a prevalent compressional regime, is still going on at the external front of the chain.

[8] 2. Since the late Miocene, several basins have been generated in the internal area (Figure 1), with an eastward decreasing age [e.g., *Bossio et al.*, 1998; *Bartolini*, 2003, and references therein]. Tensional tectonics in the internal area has been accompanied by widespread magmatic activity, whose age is decreasing eastward [*Serri et al.*, 1993]. NE-SW extension in the internal belt is also indicated by space geodesy measurements [e.g., *Caporali et al.*, 2003], although the reliability of this kind of information is quite uncertain, due to the coarse network of GPS permanent stations in central Italy, the relatively small period of observation and the uncertainty which may affect geodetic velocities [e.g., *Anzidei et al.*, 2001; *Nocquet and Calais*, 2003].

[9] 3. Paleomagnetic observations suggest that since the lower Pliocene the various sectors of the northern Apennines have undergone relative rotations, delineating an overall deformation pattern involving a progressive bowing of the belt [*Speranza et al.*, 1997; *Muttoni et al.*, 2000].

[10] 4. Several NE-SW transversal tectonic lines, interpreted as transfer zones or shear belts, have been recognized in the northern Apennines [e.g., *Boccaletti et al.*, 1985; *Liotta*, 1991; *Coltorti et al.*, 1996]. These features, some of which are shown in Figure 1, may have allowed the dissection of the orogenic belt in a number of crustal wedges and the lateral decoupling of adjacent wedges [e.g., *Sorgi et al.*, 1998].

[11] 5. Since the late Pliocene/early Pleistocene, the central northern Apennines have undergone a strong uplift, up to 2000 m, which has affected the internal area, the chain and the foredeep zone [*Centamore and Nisio*, 2003; *Argnani et al.*, 2003; *Bartolini*, 2003].

[12] 6. Most seismic energy (Figure 2a) is released in the series of tectonic troughs located in the internal side of the chain, the Lunigiana, Garfagnana, Pistoia-Firenze, Mugello, Casentino and Tiber Valley [*Mantovani et al.*, 1997b, and

references therein]. A roughly SW-NE alignment of relatively strong events is also recognizable in the Romagna Apennines (Forlì shear zone; see Figure 2b). Large earthquakes, reaching $M = 7$, have also occurred in the central Apennines (Latium-Abruzzi platform).

[13] Figure 2b reports the strain style for a number of zones, obtained by summing the seismic moment tensor of earthquakes [e.g., *Viti et al.*, 2001]. The strain regimes evidenced by earthquake focal mechanisms are obviously representative of the present deformation pattern. In our approach we have assumed that this pattern is not significantly different of that occurred in the Pliocene and Quaternary. Some remarks about this choice are reported in the next section, where the proposed evolutionary reconstruction is described. Such evolution is reported in order to explain the large-scale geodynamic framework that in our opinion determined the proposed belt-parallel compression in the northern Apennines.

3. Recent Evolution and Proposed Driving Mechanism of Tectonic Activity in the Northern Apennines

[14] The deformation pattern described in the previous section has been interpreted by *Mantovani et al.* [1997a, 2002] and *Mantovani* [2004] as an effect of the constrictional regime created in the central Mediterranean area by the convergence between Africa and the Anatolian-Aegean-Balkan (AAB) system. The geodynamic context and plate kinematics implied by the proposed interpretation are shown by the evolutionary reconstruction given in Figure 3. The key tectonic event, which determined the post late Miocene deformation pattern in the central Mediterranean region, was the clockwise rotation of the Adriatic plate, driven by the westward push of the Anatolian-Aegean-Balkan system, once that the pre-Apulian oceanic zone, which separated these two continental domains, was completely consumed [e.g., *Mercier et al.*, 1987].

[15] The rotation of the Adriatic began in the late Miocene (Figure 3b), after that this plate had decoupled from its Padanian protuberance, by the Schio-Vicenza fault system [*Castellarin and Cantelli*, 2000] and from Africa, by the fault system running from the Ionian Islands to the Sicily Channel, through the Victor Hensen and Medina faults [*Hieke and Wanninger*, 1985; *Della Vedova and Pellis*, 1989; *Hieke and Dehghani*, 1999]. The proposed clockwise rotation of the Adriatic, encompassing the northern part of the oceanic Ionian zone, can coherently account for the deformation occurred in the surrounding zones from the late Miocene to the early Pleistocene, as argued by *Mantovani et*

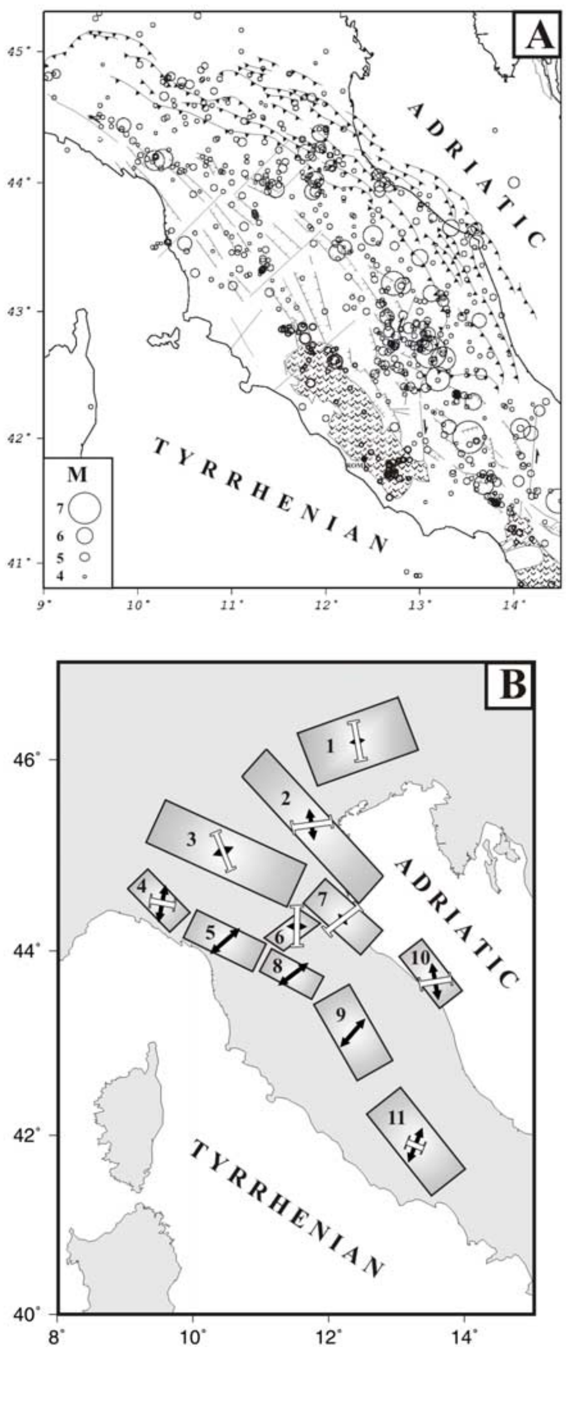

Figure 2. Main features of seismicity in the central northern Apennines. (a) Major earthquakes ($M > 4$) occurred since 1600 A.D. Data taken from *Working Group Parametric Catalog of Italian Earthquakes* [1999]. (b) Strain regimes (horizontal principal strain axes) obtained by summation of earthquake moment tensors in the main tectonic zones of the study area. Data used for computations have been taken from *Viti et al.* [2001]. Open bars and divergent arrows indicate shortening and lengthening, respectively. The length of symbols is normalized to the largest of the principal strains computed in each zone. Zones are numbered as follows: 1, Southeastern Alps; 2, Schio-Vicenza fault zone; 3, Padanian border of the northern Apennines; 4, Tortona-Lunigiana fault zone; 5, Garfagnana-Lunigiana troughs; 6, Forlì shear zone; 7, Romagna border of the northern Apennines; 8, Mugello trough; 9, Tiber Valley trough; 10, Ancona fault zone; 11, Avezzano fault zone in the Latium-Abruzzi platform.

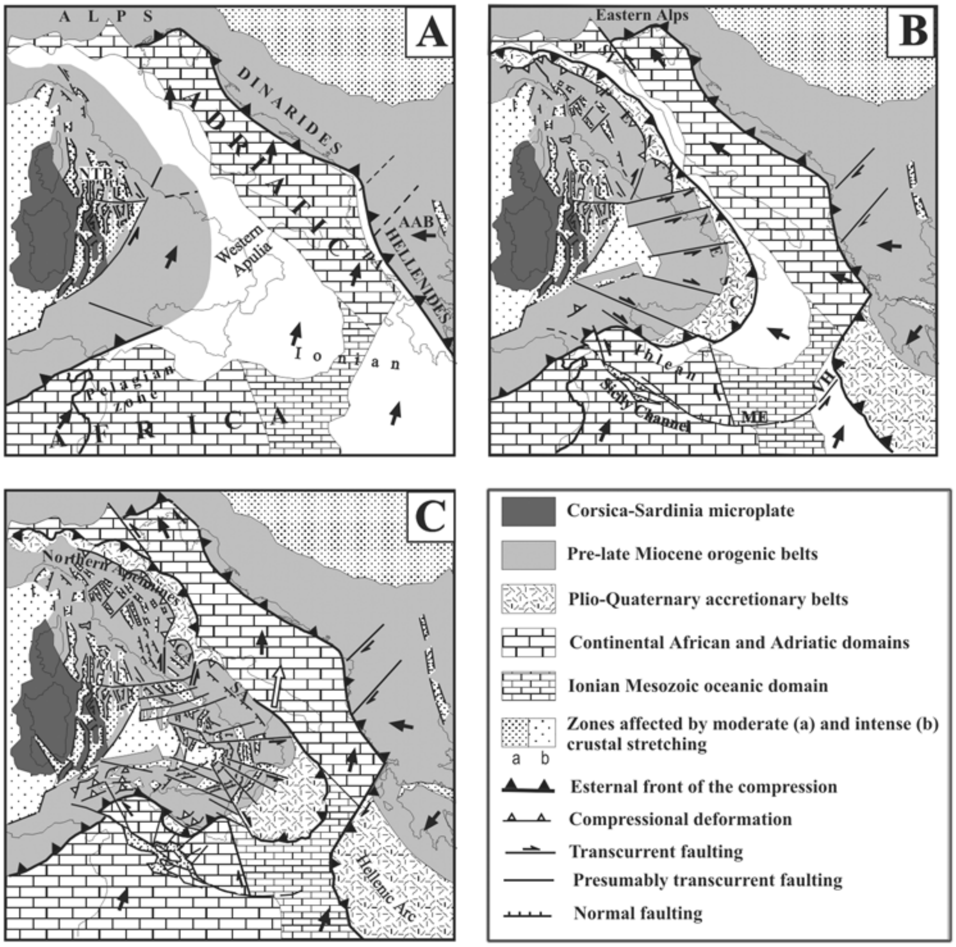

Figure 3. Evolutionary reconstruction of the central Mediterranean region (modified after *Mantovani et al.* [2002]). (a) Upper Miocene. The Adriatic behaves as a promontory of Africa, which moves roughly north-northeastward. This phase precedes the collision of the continental Adriatic domain with the Anatolian-Aegean-Balkan (AAB) system, which moves roughly westward. Open area indicates the lithospheric domains which have been consumed in the successive evolution. NTB, northern Tyrrhenian basin; pA, pre-Apulian zone. (b) Middle Pliocene. Since the late Miocene, the Adriatic, under the westward push exerted by the AAB system, undergoes a clockwise rotation. The Victor Hensen (VH) and Medina (ME) shear zones in the central Ionian allow decoupling the Adriatic plate from Africa. C, Calabrian Arc; P, Padanian zone; SV, Schio-Vicenza fault system. (c) Present. The collision of the southern Apennines (SA) with the continental Adriatic platform has stopped accretionary activity at this consuming boundary. After this event, the southern Apennines has begun to move roughly northwestward parallel to the Adriatic plate, inducing a belt-parallel push on the central northern Apennines. CA, central Apennines (Latium-Abruzzi platform). Kinematic patterns (solid arrows) are based on *Dercourt et al.* [1986] and *Mantovani et al.* [1997a, 2002, and references therein]. Motion rates are only indicative. The open arrow is southern Italy (about 5 mm/yr) indicates the geodetic velocity of the Matera station [e.g., *Anzidei et al.*, 2001; *Caporali et al.*, 2003; *Nocquet and Calais*, 2003]. Present coastline is reported for reference in each evolutionary phase. See text for explanation.

al. [1997a, 2002]. A major effect of this rotation was the roughly northwestward expulsion of a fragment of the northern continental margin of Africa, i.e., the Iblean microplate, which accommodated the convergence between the rotating Adriatic plate and the African continental margin (Tunisia).

[16] The roughly northwestward expulsion of the Iblean wedge and the simultaneous clockwise rotation of the Adriatic plate induced a strongly constrictional regime into the Alpine-Apenninic orogenic system which lay in between these blocks and the Corsica-Sardinia microplate (Figure 3b). The shortening required by this plate convergence was accommodated by another extrusion process, the roughly eastward escape of crustal wedges from the Alpine-Apenninic belt lying east of Sardinia, at the expense of the low-buoyancy lithosphere of the western Adriatic margin

and of the Ionian zone. This last consuming process was responsible for the reactivation, around the late Miocene, of accretionary activity in the whole Apenninic belt [e.g., *Ortolani and Pagliuca*, 1988; *Castellarin and Vai*, 1986; *Patacca et al.*, 1993]. The above constriction had its more evident effects in the southern Apenninic arc, but also caused longitudinal shortening of the northern Apennines, which was accommodated by the outward (roughly northeastward) escape of crustal wedges, at the expenses of the Adriatic lithosphere. In this work we try to demonstrate that this driving mechanism can account for the deformation pattern observed in the northern Apennines, that is thrusting along the external front of the arc, tensional deformation in the internal area, arc bending and generalized uplift.

[17] The clockwise rotation of the Adriatic plate lasted up to the late Pliocene-early Pleistocene, when the collision between the Apenninic belt and the continental Adriatic domain caused the suture of the consuming process in the southern Apennines [e.g., *Patacca et al.*, 1993]. This last event considerably reduced the mobility of the Adriatic plate, which was almost completely surrounded by continental domains (Figure 3c). In this tectonic context, the Adriatic plate started a slow clockwise rotation, as suggested by the occurrence of thrusting activity in the southern Dinarides, outer Hellenides [e.g., *Mercier et al.*, 1987] and Eastern Alps [e.g., *Castellarin and Vai*, 1986]. This kinematics is still going on, as clearly indicated by geophysical and geodetic data [e.g., *Anderson and Jackson*, 1987; *Nocquet and Calais*, 2003].

[18] After the suture of the subduction boundary in the southern Apennines, the push of Africa, not being anymore accommodated by that consuming process, caused an acceleration of other consuming processes, where more easily subductable lithosphere was present. In particular, this acceleration has involved the lateral expulsion of the Calabrian wedge (Figure 3c), at the expenses of the low-buoyancy Ionian lithosphere. This event is clearly and coherently documented by the distribution of compressional, tensional and transcurrent Quaternary deformation along the borders of this block and inside it [*Mantovani et al.*, 1997a, 2002].

[19] The suture of the southern Apennines also had significant effects in the central northern Apennines, since it emphasized the belt-parallel push exerted by the Calabrian arc and southern Apennines on the central northern part of the belt. This dynamics can account for the acceleration of uplift in the northern Apenninic arc since the late Pliocene/early Pleistocene.

[20] In the Plio-Quaternary deformation pattern of the Apenninic belt, the Latium-Abruzzi platform (central Apennines) played the role of a hinge zone, since it experienced a much lower outward migration with respect to the northern and southern arcs [e.g., *Sartori and Capozzi*, 1998, and references therein]. The transpressional decoupling of the Latium-Abruzzi platform from the outward escaping northern Apennines was achieved by dextral transpressional shear zones, such as the Ancona-Anzio, Valnerina and Gran Sasso tectonic belts [e.g., *Castellarin et al.*, 1982; *Decandia et al.*, 2002], while the decoupling from the southern Apennines was taken up by the sinistral Ortona-Roccamonfina transpressional border.

4. Numerical Modeling

4.1. Methodology and Choice of the Model

[21] This work aims at reproducing the active deformation of the northern Apennines by a two-dimensional (2-D) finite element (FE) model. The FE code here used is a commercial software (TWODEPEP) developed to solve various kinds of partial differential equations (details are available at http://www.pde2d.com). The study area is simulated by a heterogeneous elastic thin sheet, stressed by kinematic boundary conditions (displacements). The Navier-Cauchy equations of equilibrium have been numerically solved in the plane stress approximation, to obtain the horizontal displacement field within the sheet. The finite element mesh contains 1018 nodes and 1967 six-nodes, isoparametric quadratic triangular elements [*Sewell*, 1981]. The discretization of the domain has been optimized by using an automatic mesh generator (EASYMESH1.4, available at http://www-dinma.univ.trieste.it/nirftc/research/easymesh). The complex structural framework of the northern Apenninic arc and surrounding zones has been tentatively simulated by a heterogeneous distribution of elastic parameters within the sheet, in line with the suggestion of *Bird* [1996], that continental tectonics is better modeled by allowing localized deformation at major tectonic zones and minor distributed deformation in large crustal blocks. This choice is also supported by experimental evidence [e.g., *Talwani*, 1999], which indicates that the deformation rates may increase by several orders of magnitude from stable areas to active tectonic belts.

[22] In our modeling approach we assume that, at a first order approximation, all forces induced by belt deformations are proportional to the amount of relative displacements of surrounding blocks and tend to oppose them (e.g., gravitational spreading which opposes crustal thickening induced by block convergence). This approximation allows simulating the dynamic role of active belts by fictitious elastic elements whose parameters have no direct relationship with those of actual crustal rocks. The correct parameterization of such elements has to be obtained by a trial and error procedure, tending to minimize differences between computed and observed strain fields, as we have done in our experiments.

[23] The model has been divided into 189 subdomains (Figure 4), to simulate the poorly deformable zones and the tectonic belts, recognized on the basis of neotectonic and seismic deformation data. Each subdomain has been parameterized by assigning it two elastic constants (Young modulus, E, and Poisson ratio, ν). The Poisson ratio is assumed to be uniform over the whole model (ν = 0.25), whereas different values of E may be assumed for the various zones. The range adopted for E is 10^7–10^{11} Pa [*Mantovani et al.*, 2001a]. The highest values of E (10^{10}–10^{11} Pa) are assumed to simulate the poorly deformable zones, as the Adriatic plate, the European foreland and the Corsica-

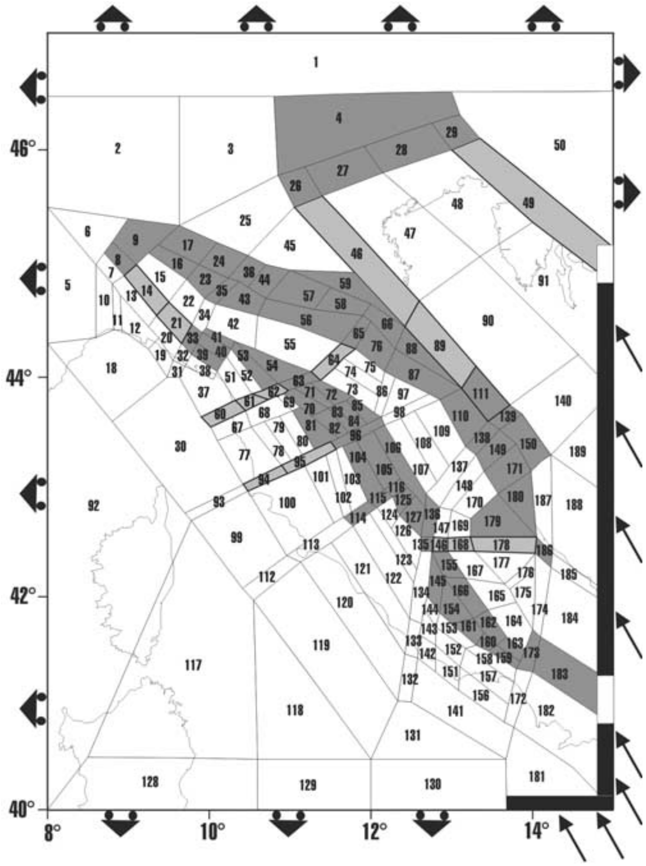

Figure 4. Model parameterization and boundary conditions adopted in numerical experiments. Each zone is identified by a number, in order to make easier any reference to it in the text. Open zones, simulating the most rigid structures, have the highest values of the Young modulus (E = $10^{10}-10^{11}$ Pa). Dark shaded zones, which simulate active extensional or compressional belts, have much lower E values (10^7-10^8 Pa). Light shaded boxes with thickened borders, simulating transcurrent, transtensional or transpressional belts, are anisotropic zones, characterized by low values of the shear modulus (μ = 10^8-10^9 Pa) in the direction of the long side of the zone and, in some cases, by low values of E. See caption of Figure 6. Kinematic boundary conditions, simulating the kinematics of the Adriatic plate and southern Apennines, are applied at the solid segments of the external border of the model. Empty segments indicate the sectors of the external border where a null stress condition is assumed. Along the other sectors of the external border, movements perpendicular to the border itself are prevented. See text for full explanation.

Sardinia block. Such choice is consistent with the mechanical properties of the crust-mantle system in the study area, suggested by the rheological profiles tentatively reconstructed in the Mediterranean area [*Viti et al.*, 1997]. These values of E have also been adopted for the axial part of the Apenninic belt. This last assumption may be justified by the fact that the stacking of relatively cold thrust sheets, mostly scraped off upper crustal levels, has cooled the whole crust of the Apennines, considerably increasing its mechanical strength. This hypothesis, suggested by thermal models of orogenic domains [e.g., *Midgley and Blundell*, 1997] is supported by the relatively low heat flow observed in the Apenninic chain [*Cataldi et al.*, 1995] and by estimates of crustal strength carried out across the belt [*Viti et al.*, 1997; *Gualteri et al.*, 1998]. The fact that few intense earthquakes have occurred in the axial part of the chain (Figure 2a) does not invalidate the choice of assigning this zone a high E value.

[24] Much lower values of E (10^7-10^8 Pa) have been assigned to the tectonic zones where most of deformation is expected to occur. Such values aim at simulating the reduced mechanical strength of tectonic structures, as, for example, major thrust-fold belts and extensional zones, which are periodically activated by large earthquakes and/or by aseismic slip [see *Zoback*, 2000; *Mantovani et al.*, 2001a]. In the model, such weak zones have been mainly adopted to simulate folding and thrusting observed along the external front of the Apenninic belt, corresponding to the Padanian and Adriatic foredeep (zones 8, 9, 16, 17, 23, 24, 35, 36, 43, 44, 56, 57, 58, 59, 65, 66, 76, 87, 88, 110, 138, 149, 171, and 180 in Figure 8) and the extensional activity in the internal area (zones 33, 39, 40, 41, 53, 54, 63, 70, 71, 72, 81, 82, 83, 84, 85, 96, 104, 105, 106, 115, 116, 125, 127, 135, and 136). A weak belt (zones 4, 26, 27, 28, and 29) simulates the thrusting processes observed along the northern border of the Adriatic plate, roughly corresponding to the Eastern Alps. Another weak belt (zones 145, 154, 155, 159, 160, 162, 163, and 166) has been assumed inside the Latium-Abruzzi (LA) platform, to simulate the presence in that area of SE-NW weak/shear zones, suggested by geological evidence and focal mechanisms of large earthquakes [*Amoruso et al.*, 1998; *Salvini*, 1993; *Piccardi et al.*, 1999]. The Laga complex has been simulated as a weak zone (zone 179) since it is constituted by very recent sedimentary deposits [e.g., *De Feyter and Delle Rose*, 2002]. A longitudinal weak belt has also been assumed in the southern Apennines (zones 173 and 183), to simulate the belt-parallel seismic zone recognized in this sector of the chain [e.g., *Cinque et al.*, 1993; *Vilardo et al.*, 2003, and references therein].

[25] Anisotropic belts in the model aim at simulating major transcurrent, transpressional, and transtensional faults. In this kind of zones, the elastic shear modulus is reduced by 3 orders of magnitude, with respect to poorly deformable zones, in the direction of anisotropy, while the Young modulus may be equal to or lower than that assumed in the poorly deformable zones, since the elastic properties in the direction of anisotropy do not depend on those related to any other direction [e.g., *Mantovani et al.*, 2001a].

[26] One of these anisotropic belts (zones 146,168, and 178) simulates the Gran Sasso transpressional border between the Laga Units (LU) and the LA platform, that is widely recognized as a lateral ramp through which the LU have moved eastward while overthrusting the LA platform [*Castellarin et al.*, 1982; *De Feyter and Delle Rose*, 2002].

The Forlì shear belt (zone 64) simulates a transversal NE-SW fault system suggested by geological information and by an alignment of earthquakes shown in Figure 2a [*Frepoli and Amato*, 1997; *Costa*, 2003]. The shear belt located in the Ligurian Apennines (zones 14 and 21) simulates the Tortona-Lunigiana fault system. The transcurrent regime along this fault is not clearly indicated in Figure 2 since no recent earthquakes have occurred in this zone, but can be tentatively recognized on the basis of seismotectonic features [*Eva et al.*, 1978; *Cattaneo et al.*, 1986; *Costa*, 2003]. Other transversal shear belts in the internal area (zones 60, 61, 62, 94, and 95) are included in the model, in order to explore the possible role of some major transversal fault systems suggested by a number of authors [*Liotta*, 1991; *Sorgi et al.*, 1998].

[27] The values of E adopted for isotropic and anisotropic weak belts are only tentative, since elastic properties cannot directly be linked with the rheological features of the lithosphere, which concern the resistance to permanent deformation. However, our choice is supported by the fact that it allows deformation to concentrate in the model zones corresponding to the actual deforming belts, thus reproducing the strongly heterogeneous deformation field observed in the northern Apennines. The role of mechanical anisotropy and weak zones in numerical modeling of continental deformation has been pointed out by *Vauchez et al.* [1998] and *Hubert-Ferrari et al.* [2003].

[28] The model is stressed by kinematic boundary conditions consistent with the convergence of the confining plates, as suggested by the evolutionary reconstruction illustrated in Figure 3 and by kinematic models proposed on the basis of neotectonic, seismological and geodetic data [*Anderson and Jackson*, 1987; *Anzidei et al.*, 2001; *Mantovani et al.*, 2002]. In particular, the evolution shown in Figure 3 implies that the northwestward motion of the Adriatic plate and southern Apennines has driven the deformation of the northern Apennines. The sectors of the external border, where displacement normal to the boundary is hampered (western, northern, and part of the southern side) simulate the lateral confinement of the modeled zone.

[29] Since we solved an elastic problem, boundary conditions have been imposed in the form of displacements proportional to the velocities of the Adriatic plate and southern Apennines. This allows interpreting the displacement fields given by the experiments as instantaneous velocities (see the scales on figures).

[30] Other details about the modeling approach here adopted and a discussion of its possible limitations are given by *Bada et al.* [1999] and *Mantovani et al.* [2001a]. We are aware that the modeling approach adopted in this work is oversimplified with respect to nature. The main shortcoming regards the use of an elastic 2-D model, since it is intrinsically unable to reproduce the finite deformation that occurs in real tectonic processes. However, stressing the model by small displacements, which implies infinitesimal strain only, should minimize the above problem.

[31] Another major simplification is the use of highly deformable zones at plate boundaries. This kind of parameterization aims at reproducing, by the deformation of these weak zones, the relative displacements of the confining poorly deformable domains, but may produce unrealistic strain values at boundary zones. However, this problem should not crucially affect the validity of the results obtained since the above effect does not influence the strain style at plate boundaries, i.e., the parameter that is compared with the observational constraints we use in our approach.

4.2. Numerical Experiments: Preferred Solution

[32] The purpose of our numerical experiments is reproducing the observed strain pattern (Figure 5), tentatively inferred from a wide collection of geological information, ranging from quantitative mesostructural analysis to qualitative geomorphological descriptions, and seismotectonic studies (see also Figure 2b). Since the assessment of strain magnitude may be affected by a large uncertainty, concerning both tectonic and seismological data, we have only considered, as constraints, the directions of horizontal principal strains and the horizontal strain style, i.e., the absolute value of the ratio between the amplitudes of the two principal strains. As regards tectonic data, the uncertainty affecting them is mainly due to difficulty of recognizing the activity of crustal faults and fractures from field observations, mostly represented by fault slips [*Marrett and Allmendinger*, 1990]. As regards focal mechanisms, the main source of uncertainty is due to the shortness of the available catalogues with respect to the long-term geological deformation [*Viti et al.*, 2001].

[33] In order to find the model parameterization and the boundary conditions that allows best reproducing the strain information described above, we carried out several experiments by varying the kinematics of the Adriatic plate and southern Apennines and the parameterization of isotropic and anisotropic weak zones. The solution we finally preferred is shown in Figure 6. The related displacement and strain fields are shown in Figures 6a and 6b, respectively. The model parameterization and the boundary conditions are illustrated in Figure 6c. The level of agreement between the observed and computed strains (Figure 6d) is evaluated separately for the strain style, i.e., the absolute value of the ratio between the amplitudes of lengthening and shortening (R), and the orientation of principal strains. To facilitate the evaluation of this agreement, we have defined seven strain styles: pure tensional ($R > 10$), transtensional closer to tensional ($5 \leq R < 10$), transtensional closer to transcurrent ($1.5 \leq R < 5$), pure transcurrent ($0.7 \leq R < 1.5$), transpressional closer to transcurrent ($0.5 \leq R < 0.7$), transpressional closer to compressional ($0.1 \leq R < 0.5$), and pure compressional ($R < 0.1$). A number has been assigned to each of these styles, ranging from 0 (pure tensional) to 6 (pure compressional), for both observed and computed strains. The difference between the observed and computed numbers, ranging from 0 (perfect agreement) to 6 (highest disagreement), is reported in each zone as indicator of the strain style agreement. The level of agreement between the azimuth of computed and observed strain axes is ranked in four classes (A, B, C, D). The first class (A) is assigned to the zones where the difference between the computed and observed

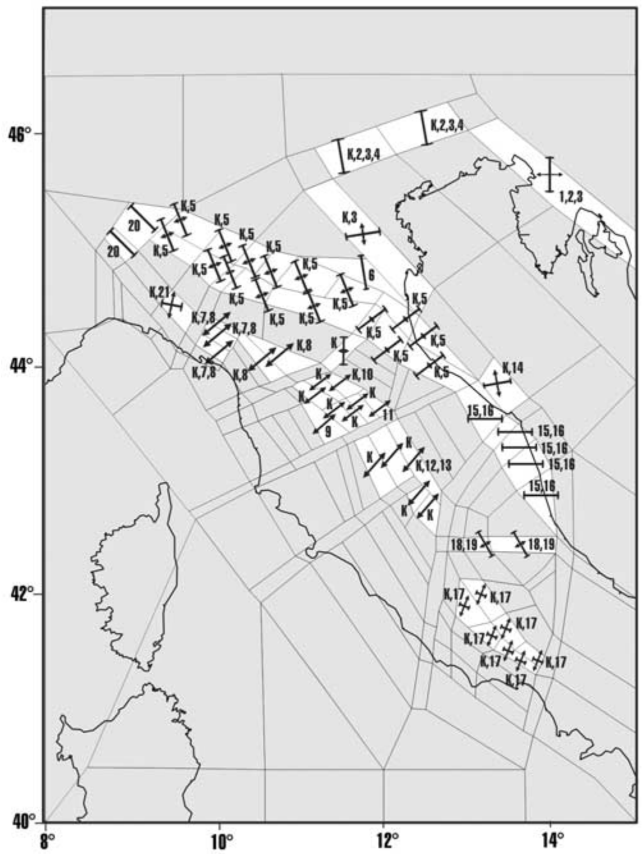

Figure 5. Principal axes of horizontal strain, inferred from seismological and neotectonic information in tectonic belts (weak zones in the model). Bars and divergent arrows indicate principal shortening and lengthening, respectively. When a K appears in first position, the strain pattern has been mainly obtained from summation of seismic moment tensor (Figure 2b). Numbers correspond to the papers from which information has been taken: 1 [*Carulli et al.*, 1990]; 2 *Bressan et al.*, 1998]; 3 [*Castellarin and Cantelli*, 2000]; 4 [*Slejko et al.*, 1999]; 5 [*Ghiselli and Martelli*, 1997]; 6 [*Montone and Mariucci*, 1999]; 7 [*Bernini*, 1988]; 8 [*Sorgi et al.*, 1998]; 9 [*Bonini and Sani*, 2002]; 10 [*Bonini et al.*, 2001]; 11 [*Bonini*, 1999]; 12 [*Bonini*, 1998]; 13 [*Boncio et al.*, 2000]; 14 [*Sagnotti et al.*, 1999]; 15 [*Centamore and Nisio*, 2003]; 16 [*Calamita and Deiana*, 1986]; 17 [*Piccardi et al.*, 1999]; 18 [*De Feyter and Delle Rose*, 2002]; 19 [*Castellarin et al.*, 1982]; 20 [*Costa*, 2003]; 21 [*Cattaneo et al.*, 1986].

azimuths ($\Delta\varphi$) is less than or equal to 15°, the second class (B) is assigned to zones where $15° < \Delta\varphi \leq 30°$, the third class is assigned to zones where $30° < \Delta\varphi \leq 60°$ and the fourth class is assigned to zones where $60° < \Delta\varphi \leq 90°$.

[34] It is worth noting that the preferred solution matches, at a first approximation, most of the observed strains, both concerning styles and orientations (Figure 6d). The significance of this result is emphasized by the fact that no particular conditions have been imposed to the behavior of weak zones in the model; that is, isotropic zones may accommodate either shortening or lengthening, while anisotropic zones may accommodate either sinistral or dextral shear.

[35] The displacement field (Figure 6a) shows that the longitudinal shortening induced by the roughly north-northwestward push of the Latium-Abruzzi platform, stressed on its turn by the southern Apennines, is mainly accommodated by the roughly northeastward expulsion of the RMU Units and by the northwestward minor translation and counterclockwise rotation of the Ligurian Units. The northern Apennines crustal wedge is decoupled from the LA platform by a shear zone, which simulates the Gran Sasso lateral ramp [e.g., *Castellarin et al.*, 1982; *Ghisetti and Vezzani*, 1991], and from the Ligurian Units by a transversal shear zone that simulates the Forlì fault system (Figure 1). In the real tectonic context, the roughly northeastward escape of the RMU Units occurs at the expense of the adjacent Adriatic foreland [e.g., *Castellarin and Vai*, 1986; *Mantovani et al.*, 1997a, 2002]. In the model, this underthrusting process is simulated by the shortening of the weak zones located between the belt and the Adriatic plate. In the internal side of the Apennines, the northeastward escape of the RMU Units is accommodated by lengthening of the weak zones located in this region, which simulates the tensional features observed from the Mugello trough to the southernmost sector of the Tiber Valley (Figures 1 and 2a).

[36] The SE-NW decoupling zone within the northernmost Ligurian Units (zones 14 and 21 in Figure 4 and "c" in Figure 6c) simulates the Tortona-Lunigiana fault zone (Figure 1). The minor west-northwestward displacement of the Ligurian Units is accommodated by shortening of the weak zone located at the northwestern front of this block, which simulates the Voghera thrust zone (Figures 1 and zones 8 and 9 in Figure 4).

[37] Shortening and lengthening of the weak zones located, respectively, NE and SW of the Ligurian Units accommodate the counterclockwise rotation of these units. This strain pattern simulates the compressional features observed along the Padanian margin of the Apenninic belt and the tensional tectonics recognized along the Lunigiana, Garfagnana, and Pistoia-Firenze troughs (Figure 1).

[38] It can finally be noted that the tensional regime observed in the southern Apennines, characterized by a SW-NE orientation of maximum lengthening [*Cinque et al.*, 1993; *Vilardo et al.*, 2003] is satisfactorily reproduced in the preferred solution (Figure 6b). This result provides further support to the reliability of the kinematic boundary conditions adopted in this experiment.

4.3. Sensitivity Tests

[39] The results reported in Figure 6 show that the proposed driving mechanism allows a fairly satisfactory reproduction of the observed features. However, to fully understand how much this achievement is significant it is necessary to have an idea of the sensitivity of the computed strain field to boundary conditions and model parameterization. In order to provide insights about this problem we report and discuss the results of some numerical experiments.

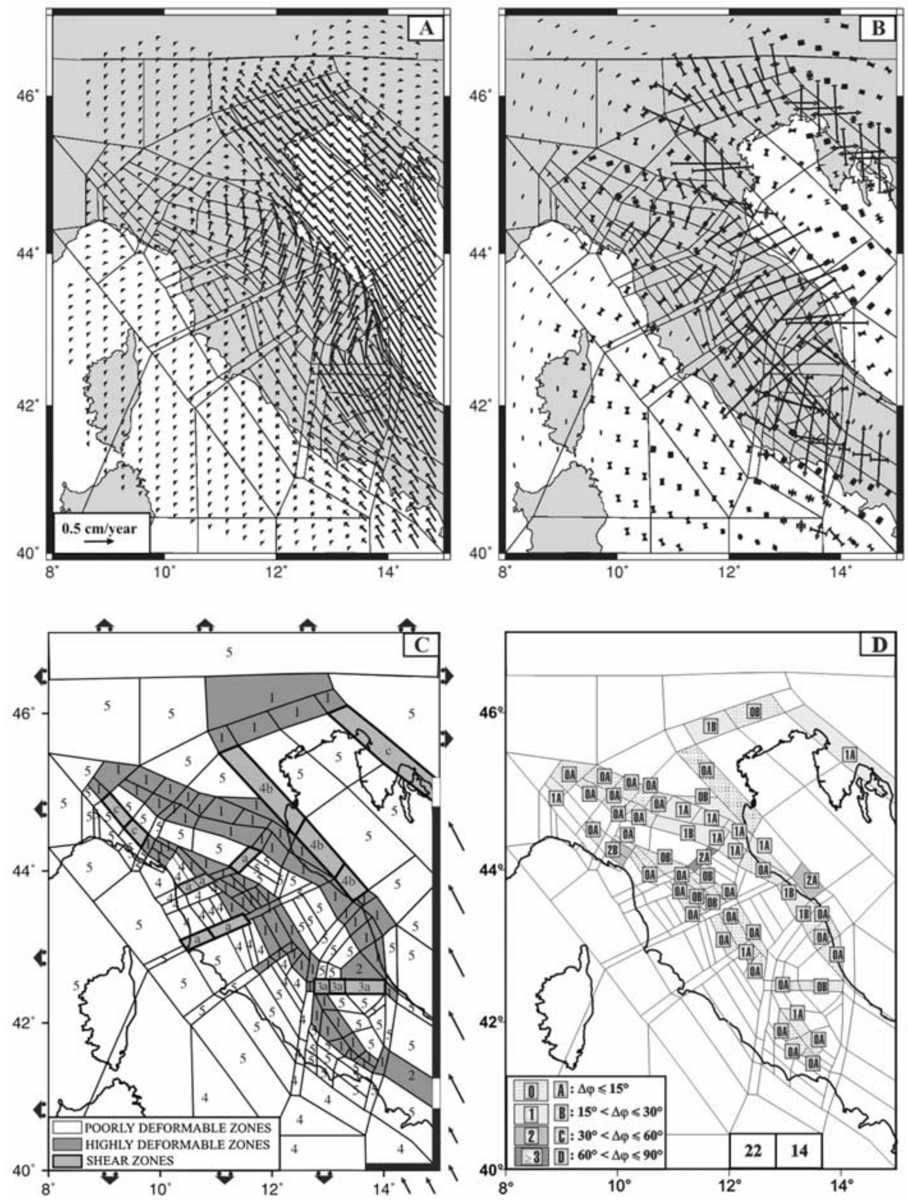

Figure 6. Preferred solution among the ones obtained by numerical experiments. (a) Velocity field. (b) Strain field. Bars and divergent arrows indicate the horizontal principal shortening and lengthening, respectively. The length of symbols is normalized to the largest principal strain computed in each point of the grid and thus only reflects the strain regime, i.e., the ratio between shortening and lengthening. (c) Parameterization of the model and boundary conditions adopted. Numbers 1 to 5 identify the isotropic zones (1, $E = 10^7$ Pa; 2, $E = 10^8$ Pa; 3, $E = 10^9$ Pa; 4, $E = 10^{10}$ Pa; 5, $E = 10^{11}$ Pa). A number plus a letter identifies the anisotropic zones; the number indicates the value of E, as described above, while the letter indicates the value of the shear modulus along the direction of weakness (a, $\mu = 10^8$ Pa; b, $\mu = 10^9$ Pa; c, $\mu = 10^{10}$ Pa). Arrows indicate boundary conditions, in accord with the scale shown in Figure 6a. In the segments of the border with white bars a null stress is adopted, in the segments with rolls displacement perpendicular to the border is hampered. (d) Level of agreement between computed and observed (Figure 5) principal strains. In each zone, the number indicates the level of agreement between strain styles (increasing from 1 to 3), while the letter (A, B, C, and D) indicates the level of agreement between principal strain directions (see the legend at the bottom of Figure 6d). The two big numbers reported at bottom right of Figure 6d provide a synthesis of the general misfit level; the number on the left (strain style agreement) is the sum of the first numbers in all zones, while the number on the right (azimuth agreement) is the sum of the numbers obtained by translating the letters into numbers (A = 0, B = 1, C = 2, D = 3) in all zones. See text for explanation.

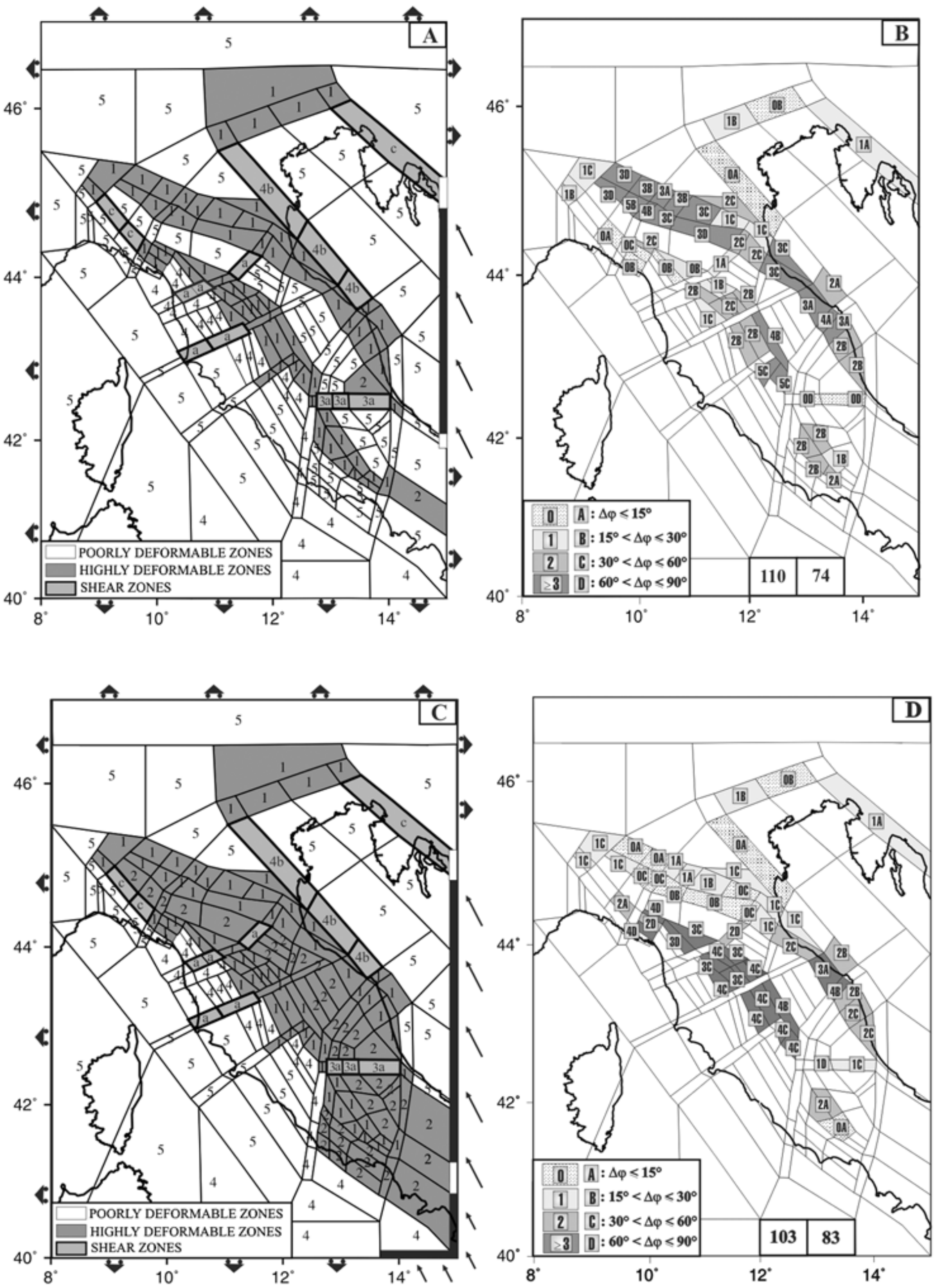

Figure 7. (a) Model parameterization and (b) misfit map related to a numerical experiment where kinematic boundary conditions are only applied to the Adriatic plate. (c and d) Same type of maps resulting from a numerical experiment where the axial part of the northern Apennines is parameterized as a weak zone ($E = 10^7 - 10^8$ Pa), like the adjacent belts. Symbols as in Figure 6. See text for comments.

[40] In the first test (Figures 7a and 7b), the best fit kinematic boundary conditions have been only applied to the Adriatic plate (Figure 7a). From the misfit distribution (Figure 7b) it is possible to note that the main features of the observed deformation pattern in the northern Apennines, that is, shortening along the external front and lengthening in the internal area, are not reproduced. Most of the weak zones surrounding the strong axial belt of the chain are dominated by strike-slip deformation, and strains in the internal side of the northern Apennines are much lower than those of the external zone. The result of this test clearly testifies that the push of southern Apennines on the central northern Apennines is a crucial condition for reproducing the observed deformation in the study area.

[41] The second test (Figures 7c and 7d) points out that another basic condition for reproducing the observed deformation pattern is the presence of a strong axial belt in the northern Apennines. When this condition is not fulfilled, as

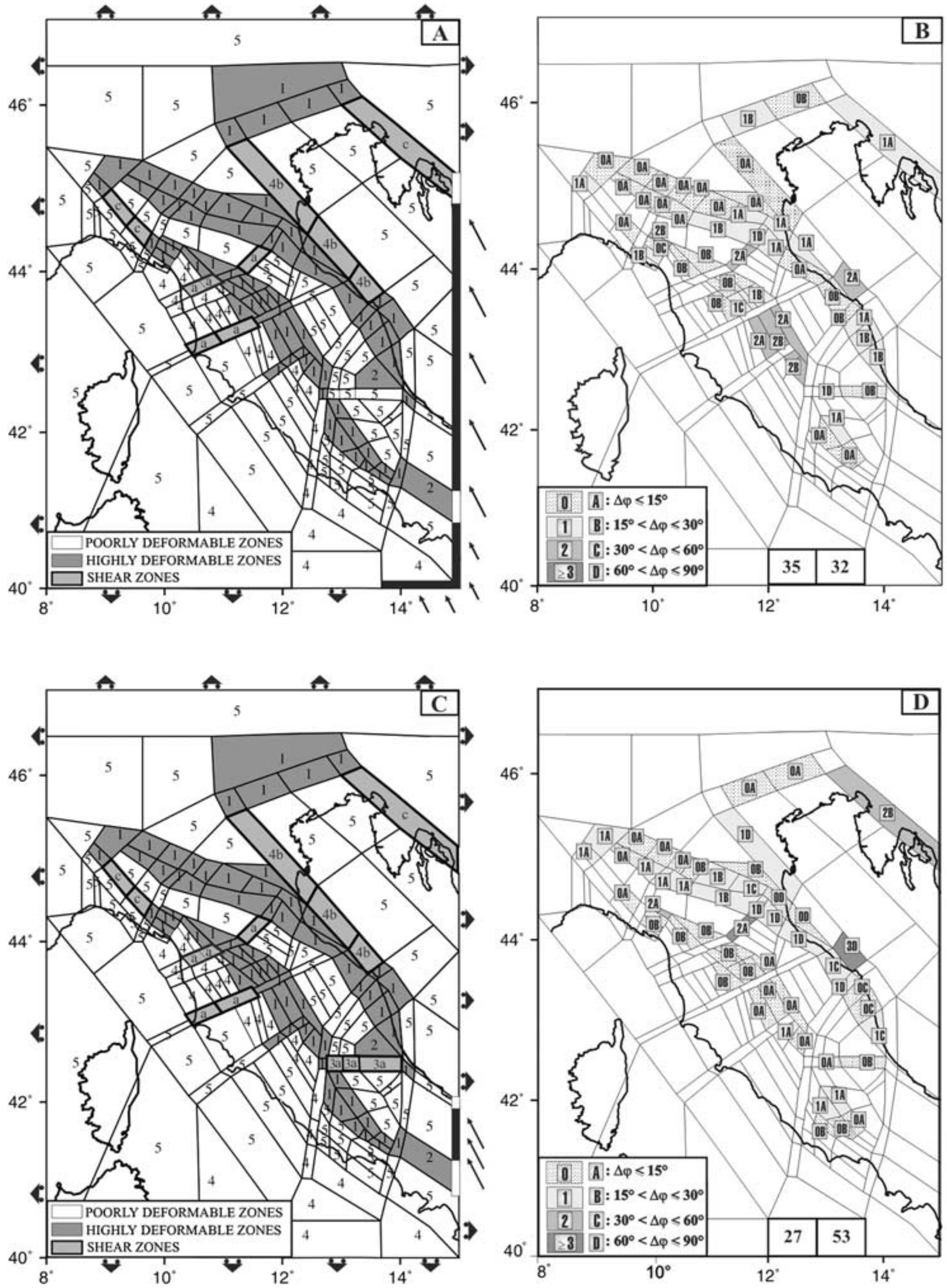

Figure 8. (a) Model parameterization and (b) misfit map resulting from a numerical experiment where the transversal shear zone located between the Lazio-Abruzzi platform and the Laga Units is not parameterized as a shear zone. (c and d) Same type of maps resulting from a numerical experiment where kinematic boundary conditions are only applied to the southern Apennines. Symbols as in Figure 6. See text for comments.

in this experiment (Figure 7c), the computed strains are very different from the observed ones (Figure 7d), both concerning strain styles and orientations, especially in the internal area.

[42] The third experiment (Figures 8a and 8b) underlines the important role of the Gran Sasso decoupling zone for generating the observed tensional strains in the internal area. When this border is not weakened (Figure 8a), concerning both east and shear modulus, to simulate the observed transpressional fault system, the internal area, with particular regard to the Tiber Valley zone, is dominated by a strike-slip strain regime instead of a tensional one (Figure 8b).

[43] In the fourth test (Figures 8c and 8d) the best fit kinematic boundary conditions have been only applied to the southern Apennines. The result of this experiment

(Figure 8d), not significantly worse than that of the preferred solution (Figure 6d), points out that the main feature of the observed deformation pattern, i.e., compression in the external front and extension in the internal side of the northern Apennines, is mainly related to the push of the central southern part of the belt. Worsening of agreement with respect to the best fit run, mainly concerning orientation of principal strains, can be noted at the external border of the northern Apennines. The better agreement that the preferred solution presents in this zone is due to the fact that the northwestward motion assumed for the Adriatic plate, combining with the extruding direction of crustal wedges in the northern Apennines, allows a better match of the orientations of strains along the external front of the belt (Figure 6).

5. Tectonics and Seismicity: The Case of the 1915 Avezzano Earthquake

[44] Numerical simulation can also be used to obtain insights into the connection between tectonic processes and seismic activity in the northern Apennines. To this purpose, we have investigated the tectonic/kinematic conditions that might have determined the peculiar seismic behavior of the northern Apenninic arc that followed the strong Avezzano (central Abruzzi) earthquake (13 January 1915, M_s = 7.0) [e.g., *Amoruso et al.*, 1998]. During the 1915–1920 time interval, this arc was affected by a level of seismic activity much higher than average, with 10 shocks of 4.5 < M < 5.5 and 6 of magnitude greater than 5.5 (Figure 9a). The seismic history of this zone since the XI century [*Working Group Parametric Catalog of Italian Earthquakes*, 1999] indicates that such amount of seismic energy is generally released over several tens of years or even more. To find a possible explanation for the sudden acceleration of seismic activity after 1915, we tried to simulate, by numerical experiments, the presumed effects of the Avezzano event on the kinematic behavior and related strain field in the northern Apenninic arc, in the framework of the geodynamic model described earlier. To tentatively reproduce the effects of the Avezzano earthquake, we have assumed that this strong event has activated the main longitudinal decoupling shear zone in the LA platform [e.g., *Amoruso et al.*, 1998; *Piccardi et al.*, 1999], emphasizing the mobility of the eastern part of this platform and, consequently, its northwestward push on the northern Apenninic arc. In the model, the "activation" of the Avezzano fault is simulated by a drastic lowering of elastic parameters (from 10^{11} to 10^{7}) in the fault zone.

[45] The effects of the Avezzano decoupling event, in terms of difference between the strain fields computed with and without the modeled shear zone, are shown in Figure 9b. The residual strain field reported in Figure 9b shows that the activation of the above shear zone in the LA platform causes a significant increase of compressional and tensional strain in the zones where the most intense earthquakes have occurred in the period 1915–1920 (Figure 9a).

[46] The fact that the distances of the presumably triggered north Apenninic earthquakes from the LA platform

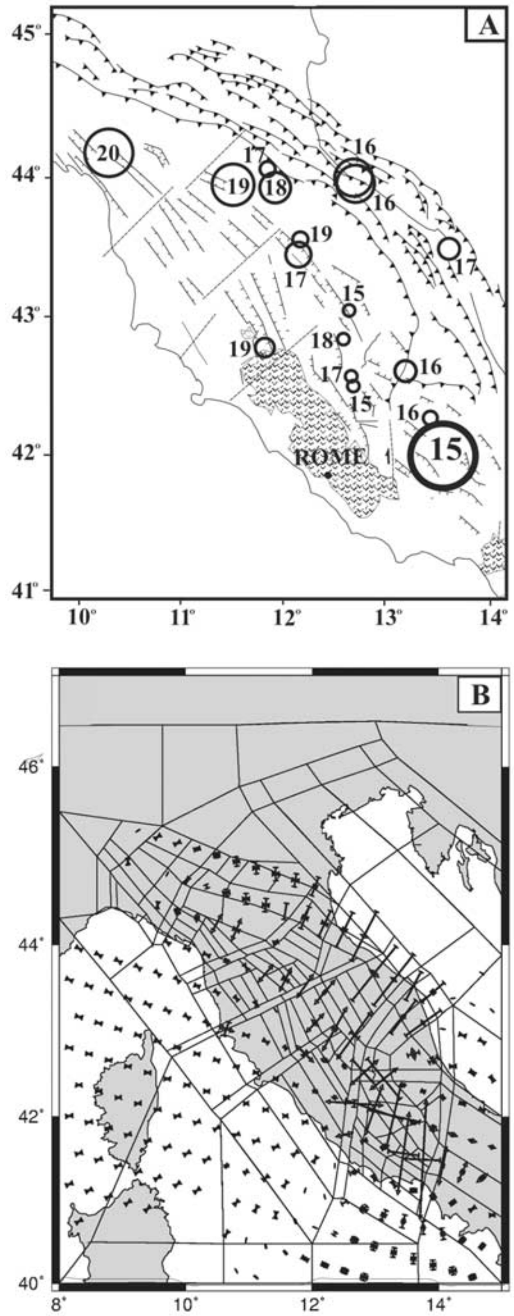

Figure 9. Strain perturbation induced by the 1915 Avezzano earthquake. (a) Distribution of major earthquakes (M > 4.5) occurred in the northern Apennines in the period 1915–1920 [*Working Group Parametric Catalog of Italian Earthquake*, 1999]. Numbers inside or close to circles indicate the year of earthquake occurrence. (b) Tentative modeling of the strain perturbation induced by the Avezzano earthquake (13 January 1915, M = 7) in the northern Apennines. This event is simulated by a drastic lowering of elastic parameters (from 10^{11} to 10^{7}) in the zones 164, 165 and 167 of the model (Figure 4). Strain symbols indicate, in term of principal axes, the difference between the strain obtained by "activating" and "not activating" the Avezzano shear zone, respectively.

progressively increase with time, at rates of some tens of kilometers per year, might be explained by considering that the propagation of the strain/stress perturbation induced by the Avezzano earthquake is controlled by the rheological properties of the crust-upper mantle system. The quantitative insights provided by elastic-viscous modeling of stress diffusion in the central Mediterranean area [e.g., *Mantovani et al.*, 2001c; *Viti et al.*, 2003] could explain the space-time distribution of major shocks in the example here considered.

6. Discussion and Conclusions

[47] Numerical experiments show that the Plio-Quaternary deformation pattern in the northern Apennines can be reproduced, at a first approximation, as an effect of a belt-parallel compression induced by plate convergence, provided that the chain has a high mechanical strength and is decoupled from the surrounding structures by weak zones. The northeastward lateral escape of the Romagna-Marche-Umbria Units and the counterclockwise rotation, with minor northwestward translation, of the Ligurian Units mainly accommodate the longitudinal shortening of the belt. In the northern and eastern margins of these crustal wedges, shortening of the weak decoupling zones simulates the effects (thrusting) of the convergent motion between the extruding wedges and the Adriatic plate, observed along this border. The lengthening of the weak zones located along the internal side of the extruding wedges simulates the extensional tectonics observed in Tuscany and Latium.

[48] Sensitivity tests suggest that the proposed extrusion mechanism can only occur if two basic conditions are fulfilled. The first is a roughly northward motion of the southern Apennines, which induces longitudinal compression in the central and northern parts of the belt. A parallel motion of the Adriatic plate improves the match of the trend of the maximum shortening axis in the interaction zones between the extruding Apenninic wedges and the Adriatic foreland. The second condition is that the mechanical strength of the axial part of the belt must be higher than the one of the adjacent zones. The evidence and arguments which led us to believe that such conditions are fulfilled in reality are reported in the text.

[49] While the main features of extrusion processes are well recognized in the shallow structures, the deeper implications of this phenomenon, in particular the thickness of the wedges involved in the lateral escape and the driving mechanism of vertical movements in the crust, are still surrounded by considerable uncertainty. As regards this problem, *Meissner and Mooney* [1998] have suggested that the ductile lower crust beneath orogens may laterally flow, dragged by the overlying brittle wedge, or it may be shortened and thickened by the convergence of the colder and stronger lower crust of the confining plates, which acts as a piston. In this last case, the whole crust undergoes a strong uplift, like the one that led to the formation of the Tibet plateau (hydraulic pump mechanism by *Zhao and Morgan* [1987]). The interconnection between extrusion and uplift depends upon the configuration of the lithospheric bodies confining the stressed orogenic system and by the rheological properties of the ductile layers involved [*Meissner and Mooney*, 1998]. The strong uplift that the northern Apennines has undergone in the Pleistocene [e.g., *Bartolini*, 2003; *Centamore and Nisio*, 2003] might be interpreted as an effect of shortening and thickening of the ductile layers induced by the proposed belt-parallel compression. In particular, the combination of brittle extrusion in the upper crust and ductile thickening of the lower crust might explain a quite peculiar feature of the inner Apenninic belt, i.e., the contemporaneous occurrence of extensional tectonics and uplift. A similar interpretation has been proposed for other tectonic contexts, such as the Eastern Alps [*Mancktelow and Pavlis*, 1994] and the Aegean system [*Avigad et al.*, 2001], to explain the synchronous occurrence of brittle extension, recognized at upper crustal levels as detachment faults, and orthogonal shortening of the underlying ductile lower crust. In this regard, recent structural studies [*Carosi et al.*, 2002], carried out in an exhumed sector of the Tuscan metamorphic complex (Alpi Apuane), suggest that belt-parallel compression has affected the deep crust in the latest tectonic phases.

[50] The quantitative arguments given in this work show that the proposed driving mechanism (belt-parallel compression driven by convergence of confining plates) can coherently account for the deformation pattern observed in the northern Apennines, but cannot exclude that other driving mechanisms, such as the ones so far proposed, can achieve the same result. To try to mitigate this ambiguity and to better justify our choice of quantifying the implications of the extrusion model, we report here a discussion on how the implications of the alternative driving mechanisms so far proposed can be reconciled with the major observed features in the study area.

6.1. Slab Pull

[51] It is widely recognized that slab rollback, driven by gravitational forces, can only occur when a long slab (at least 300 km; see *Hassani et al.* [1997]) is present under the subduction boundary, a condition which seems to be highly improbable under the northern Apennines, at the light of basic evidence and arguments. In this regard, it is widely recognized that the Apulian/Adriatic foreland, which has been consumed under the eastward migrating Apenninic/Alpine belt, at least since the late Miocene, had a continental-like character [e.g., *Serri et al.*, 1993; *Finetti et al.*, 2001]. Because of the high buoyancy of such type of lithosphere, its penetration into the mantle would encounter very strong resistance and thus the formation of a continuous long slab hundreds of kilometers long is very unlikely [e.g., *Shemenda*, 1993]. This hypothesis finds support in the reconstruction, by deep seismic surveys, of the crustal and upper mantle structure beneath the northern Apennines [*Finetti et al.*, 2001], which clearly shows that the consumption of the Adriatic foreland has developed through the formation of a number of slivers, each one reaching a moderate depth (less than 100 km). This evidence can explain the distribution of subcrustal earthquakes under the northern Apennines, few events reaching a maximum depth of 90 km [*Selvaggi and Amato*, 1992].

[52] The hypothesis that a long slab exists beneath the northern Apennines is only based on the results of tomographic investigations. However, this kind of information seems to be rather ambiguous in the study area. In fact, some authors [e.g., *Lucente and Speranza*, 2001] interpreted these data as an evidence of a continuous slab reaching a depth of 200 km or even 670 km, while others [e.g., *Van der Meulen et al.*, 1999] suggested a discontinuous structure of the slab under the same zone.

[53] Moreover, as argued by *Mantovani et al.* [2001b], the feasibility of the slab pull mechanism is questionable even when a long dense slab is present beneath the involved subduction boundary. This is mainly because modeling of subduction processes [e.g., *Shemenda*, 1993; *Hassani et al.*, 1997] indicates that the forces induced by slab pull in the shallow lithosphere are largely insufficient to produce back arc opening and thus slab rollback.

6.2. Slab Detachment

[54] This interpretation suggests that the deformation pattern in the northern Apennines can be explained as an effect of slab breaking under the study area [e.g., *Wortel and Spakman*, 2000; *Di Bucci and Mazzoli*, 2002]. This hypothesis would account for the fast uplift that has affected the Apenninic belt since the late Pliocene [e.g., *Bartolini*, 2003; *Centamore and Nisio*, 2003], which is interpreted as a sort of rebound of the shallow Adriatic foreland in response to its detachment from the subducted part.

[55] The feasibility of this interpretation is obviously based on the hypothesis that a long slab existed beneath the northern Apennines before the presumed detachment phase (middle-late Pleistocene) [e.g., *Di Bucci and Mazzoli*, 2002, and reference therein]. However, as argued in the previous point, the above condition is far to be demonstrated; in particular, it is not consistent with the results of deep seismic soundings in the northern Apennines [*Finetti et al.*, 2001]. Furthermore, slab detachment cannot easily explain the progressive bowing of the belt and the evidence of belt parallel compression in the whole Apenninic chain, described in the discussion of the previous model. Thus interpretation can only be used to explain the Pleistocene deformation, i.e., the one following the presumed slab detachment. Another driving mechanism must be identified to account for the Pliocene deformation pattern in the northern Apennines.

6.3. Gravitational Collapse

[56] The spreading force arising from lateral variations of crustal thickness has been invoked to explain the occurrence of tensional tectonics within major orogenic belts, such as the Andes and Himalayas. This mechanism has also been tentatively proposed as responsible for the deformation pattern in the northern Apennines [e.g., *Carmignani and Kligfield*, 1990]. However, the main implications of this model [e.g., *Rey et al.*, 2001] are not consistent with the asymmetric deformation pattern in the northern Apennines. The collapse model predicts subsidence and tensional deformation in the axial sector of the chain, where the crust is thickest, and compressional tectonics at both sides of the collapsing belt. Thus this model could account for the thrusting observed in the external front of the chain and for the tensional tectonics in the internal area, but it cannot easily explain why compressional deformation is not recorded to the west of the extensional zone. In addition, this model cannot easily explain the progressive bowing and uplift of the Apenninic arc. Crustal collapse occurs at the expense of the gravitational potential energy stored in the thickened crust, which consequently must thin, causing subsidence, not uplift.

[57] **Acknowledgments.** We are grateful to K. Stüwe and to two anonymous reviewers for their suggestions, which have allowed a considerable improvement of the paper. Figures have been drawn by using the GMT shareware software [*Wessel and Smith*, 1995]. This work has been carried out by the financial support of the Ministry of Public Education and Research (MIUR) and of National Space Agency (ASI).

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