A new active tectonic model for the construction of the Northern Apennines mountain front near Bologna (Italy)

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Received 30 July 2007; revised 26 March 2008; accepted 22 April 2008; published 22 August 2008.

[1] We integrate existing and new geologic data [REtreating TRench, Extension, and Accretion Tectonics (RETREAT project)], particularly on the origin, growth, and activity of the mountain front at Bologna, Italy, into a new model that explains Apennine orogenesis in the context of a slab rollback - upper plate retreat process. The Bologna mountain front is an actively growing structure driving rock uplift ~1 mm/year, cored by a midcrustal flat-ramp structure that accommodates ongoing shortening driven by Adria subduction at a rate of ~2.5 mm/year. The data we use are assembled from river terraces and associated Pleistocene growth strata, geodesy including releveling surveys, reinterpretation of published reflection lines, and a new high-resolution reflection line. These data constrain a simple trishear model that inverts for blind thrust ramp depth, dip, and slip. Apennine extension is recognized both in the foreland, as high-angle normal faults and modest stretching in the carapace of the growing mountain front, and in the hinterland, with larger normal faults that accomplish some crustal thinning as the upper plate retreats. This coevolution of extension and shortening shares some notable characteristics with other basement-involved collisional orogens including the early Tertiary Laramide orogeny in the American West and the Oligocene to Miocene evolution of the Alps. We propose a possible relationship between underplating and the development of the Po as a sag basin as a Quaternary phenomenon that may also apply to past periods of Apennine deformation (Tortonian). Continued shortening on the structure beneath the Bologna mountain front represents by far the most important and underappreciated seismogenic source in the front of the northern Apennines.

Citation: Picotti, V., and F. J. Pazzaglia (2008), A new active tectonic model for the construction of the Northern Apennines mountain front near Bologna (Italy), *J. Geophys. Res.*, *113*, B08412, doi:10.1029/2007JB005307.

1. Introduction

[2] The northern Apennines of Italy offer an opportunity to study orogenic growth associated with slab rollback and the degree to which this process is coupled to retreat of the upper plate because, unlike other generally accepted roll-back – retreat settings that lie in deep ocean basins, the entire orogen from the forearc through the back-arc has been recently uplifted and subaerially exposed. Crustal deformation, rock uplift and topographic growth of the Apennines (Italy) (Figure 1) are tied to the subduction of Adria, a process that began in the Oligocene during Alpine orogenesis and a period of major Mediterranean plate reorganization [*Rosenbaum and Lister*, 2005]. It is the observation of contemporaneous crustal shortening and extension, long recognized in the Apennines [*Elter et al.*, 1975], that has lead several studies to argue for slab rollback [e.g., *Malinverno and Ryan*, 1986; *Royden*, 1988] coupled by mantle flow to retreat of the upper plate. Seismic tomographic images of an intact slab below the northern Apennines [*Piromallo and Morelli*, 2003], the lack of appreciable orogen-parallel stretching distal to the euler pole, the presence of an internal, extensional earthquake belt that follows the crest of the range [*Pondrelli et al.*, 2002], an external, compressional earthquake belt close to the mountain front [e.g., *Frepoli and Amato*, 1997], and geodynamic models [e.g., *Doglioni et al.*, 1999; *Carminati et al.*, 2003a, 2003b] are all permissive of a slab rollback – retreat model.

[3] In this context, a broad collaboration of North American, Italian, and Czech scientists have conducted new geologic, geophysical, and geodynamic experiments as part of the NSF Continental Dynamics REtreating TRench, Extension, and Accretion Tectonics (RETREAT) project to further test and better understand the slab rollback – upper plate retreat process with the goal of elucidating it as a more general process in Mediterranean orogenesis. This paper provides a synthesis of existing and new geologic observations that both support and reject parts of the classic rollback – retreat model. In particular, we focus our

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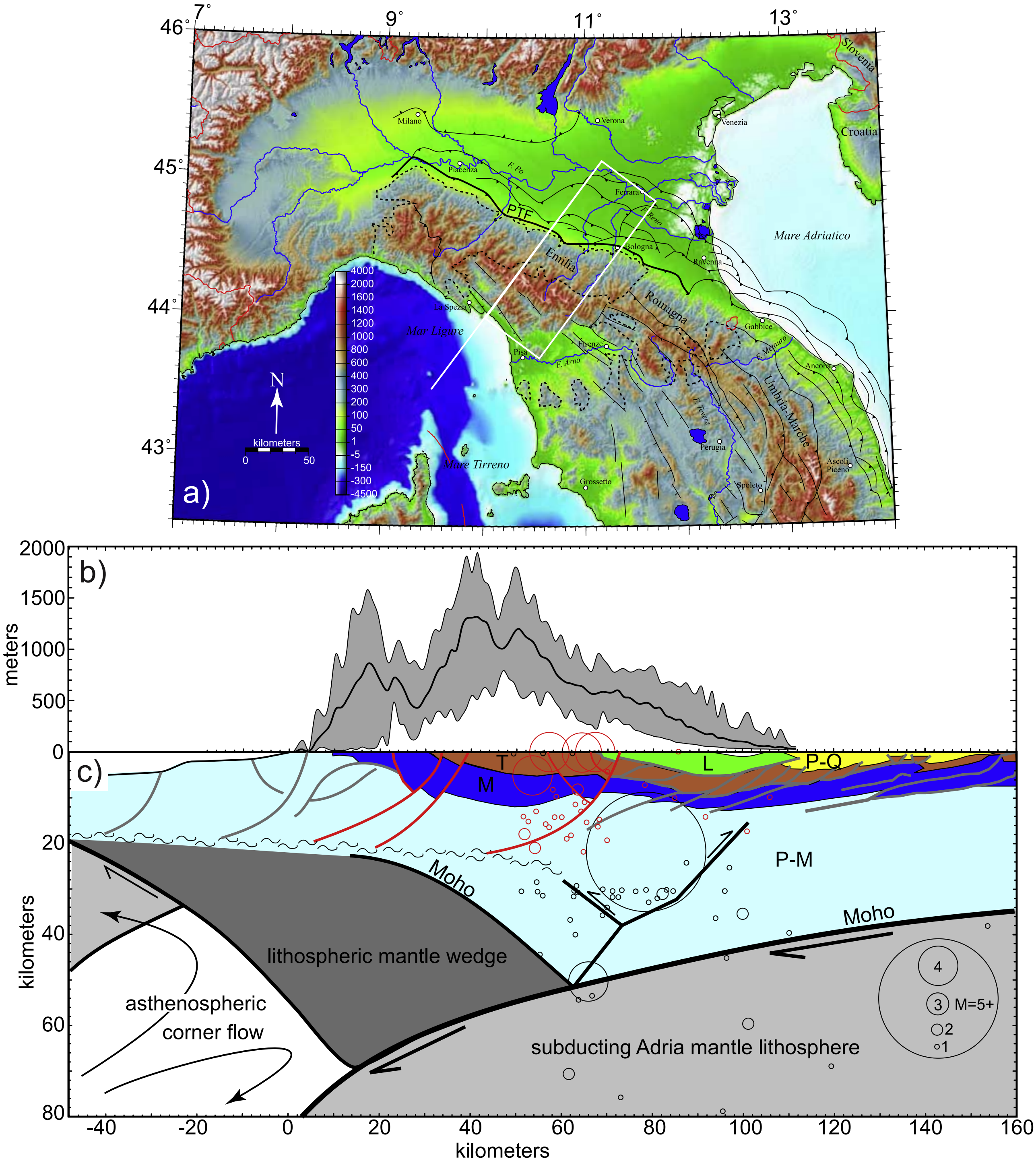

Figure 1. (a) Shaded topography (Shuttle Radar Topography Mission, SRTM 90 m) and bathymetry (General Bathymetric Chart of the Oceans, GEBCO 1 km) of northern and central Italy and the study location (white box). Thrust faults have triangles on the upper plate, and normal faults have tics in the direction of the hanging wall. PTF, Pedeapenninic Thrust Fault, according to *Boccaletti et al.* [2004]. The thick dotted lines encircle the approximate location of Ligurian and epi-Ligurian rocks. (b) Topographic swath profile across the northern Apennines representing the topography of the white box all projected to a cross-section line defined by the box's western edge. (c) Generalized crustal and upper mantle geologic cross-section (in progress, RETREAT working group) and earthquake epicenter data [*Boccaletti et al.*, 2004]. Line of cross-section is shown by the white line on the western side of the white box, extending to the offshore. P-Q, Pliocene-Quaternary deposits; L, Ligurian nappe; T, imbricated Tertiary siliciclastic foredeep turbidites; M, Mesozoic limestones; P-M, deformed Paleozoic-Mesozoic Tethyan crust, undifferentiated. Emilia, Romagna, and Umbria-Marche refer to segments of the Apenninic chain.

analysis on the origin, growth, and active deformation of the northern Apennines mountain front, a geomorphic feature that we argue is critical in understanding the geodynamics of the Apennines and the rollback – retreat process. Our goal is to address the Quaternary and modern tectonics of the northern Apennines and develop an active tectonic model consistent with diverse geologic, geomorphologic, geophysical, and geodynamic setting of the mountain front. Our framing of the problem below illustrates the equivocal nature of much of the data that have been used so far to develop a consistent geodynamic model of the northern Apennines. We propose that the geology and geomorphology of the range and adjacent Po Plain have integrated tectonic deformation over the appropriate length scales pertinent to the appropriate tectonic and geodynamic interpretation.

[4] Our approach centers on delivery of four new lines of evidence: (1) reinterpreted, balanced cross-sections spanning the mountain front and extending into the Po foreland, (2) a shallow, high-resolution reflection seismic line (P. P. Bruno et al., High-resolution shallow imaging of the northern Apennines mountain front near Bologna, Italy, using "wide-aperture" shallow seismic reflection data, submitted to *Geophysical Research Letter*, 2008), (3) uplifted and deformed Quaternary terraces, and (4) a synthesis of recently published GPS and releveling geodetic data (a model for the surficial deformation). These four data sources are placed in the context of existing knowledge of the geometries and sedimentary infill of the Po foreland through time, which provides a first-order constraint on the Neogene-Quaternary tectonic setting. We conclude that Apennine orogenesis, and particularly the Quaternary tectonics that have built the northern Apennine mountain front are consistent with unsteady rollback of Adria, the hard coupling of extending Apennine crust to the retreating Tyrrhenian plate, and long-wavelength deformation of the plate boundary perhaps in response to lithospheric horizontal stresses (buckling) or to rollback-retreat dynamic topography [*Shaw and Pysklywec*, 2007]. Our results add to the emerging view that the northern Apennine mountain front remains tectonically active and may be the nascent phase of an Alpine-style collision.

2. Background

2.1. Northern Apennines Geology, Tectonics, Crustal Structure, and Geodesy

[5] Subduction of Adria beneath what is now southern France during Alpine orogenesis built a large, south-facing, accretionary mélange including ophiolite slivers called the Liguran accretionary complex. Beginning 30 Ma [*Le Pichon et al.*, 1971], back arc spreading in response to rollback of the Adriatic slab pulled the Corsica/Sardina block south and east along with the Ligurian complex and began making a new Apennine accretionary wedge. The Ligurian accretionary complex rode passively on top of the wedge as an intact forearc structural lid [*Reutter*, 1981; *Treves*, 1984; *Patacca et al.*, 1990; *Pini*, 1999; *Zattin et al.*, 2002]. At about 8 Ma a new back-arc started to develop, rifting off part of the old accretionary wedge and creating the Tyrrhenian sea. From 30 to 4 Ma, the Apennine wedge remained submerged as it overrode highly thinned, Adriatic continental lithosphere capped predominantly by Tethyan carbonates. The wedge grew by underplating of deep-water turbidites, olistostromes, and crustal rocks. The turbidites were deposited in a series of foredeeps (or trench basins), developed successively in front of the advancing wedge, from Late Oligocene to Late Miocene [Macigno, Cervarola, Marnoso-Arenacea basins; *Ricci Lucchi*, 1986], all of which are overlain by the Ligurian lid. At about 4 Ma, the wedge became emergent as it advanced over the thicker west-facing passive margin of the Adriatic continental platform. Today, the northern Apennines expose a largely intact, ~4 km-thick Ligurian nappe that carries Late Eocene to Pliocene epi-Ligurian marine sediments in wedge-top basins [piggyback basins of *Ori and Friend*, 1984, Figure 1b]. Balanced cross-sections for the northern Apennines [*Bally et al.*, 1986; *Hill and Hayward*, 1988] indicate ~130 to 150 km of subduction since 30 Ma, which indicates relatively slow long-term rates at ~4 to 5 km/m.y. (4–5 mm/year). This geologic estimate agrees with plate tectonic reconstructions of *Dewey et al.* [1989, see their Figure 6], but is approximately half the overall convergence velocity estimated for the northeastward propagation of the paired deformation fronts over the past 15 m.y. [*Bartole*, 1995; *Cavinato and DeCelles*, 1999; *Basili and Barba*, 2007] at ~8–10 km/m.y. corrected for crustal extension [*Makris et al.*, 1999].

[6] Elevated forearcs similar to the Apennines include Barbados, Japan, the north island of New Zealand, the Andean and Central American margins, Cascadia and Alaska (Kodiak Island), and are a characteristic feature of subduction zones world-wide. Some do show surface extension (New Zealand, southern Andes), but nowhere is it as prevalent as in the case of the Apennines where it has been linked to slab rollback [*Dewey*, 1980; *Reutter*, 1981; *Royden*, 1993]. As such, the Apennines subduction may represent the continental equivalent of Marianas-style behavior where extension occurs within the volcanic arc or immediately behind the arc, resulting in the opening and spreading of a backarc basin. The key difference in this subduction is that it exhibits evidence for crustal-scale extension in front of the arc. Highly attenuated Tethyan continental lithosphere, rather than true oceanic lithosphere characterizes many Mediterranean subducting slabs.

[7] The notion that there are two Mohos beneath the Apennines, a shallow one ~25 km deep south and west of the topographic crest and a deep one ~50 km deep north and east of the topographic crest has persisted since the first seismic refraction experiments [e.g., *Ponziani et al.*, 1995], and still is popular, given its reviviscence in the CROP 3 line interpretation [*Barchi et al.*, 1998; *Morgante et al.*, 1998], new refraction [*Makris et al.*, 1999], and recent high-resolution tomography (R. Di Stefano and J. Park, personal communication, 2007; Figures 1b and 1c). These observations are consistent with the rollback-retreat model where crust is being thickened along the leading edge of Adria-Tyrrhenian convergence, but then being thinned as the retreating Tyrrhenian plate is stretched.

[8] A growing database of GPS geodesy constrains the modern movements of the northern Apennines, Europe, and Adria [*Serpelloni et al.*, 2005; *Grenerczy et al.*, 2005; *Stein and Sella*, 2006; R. Bennett, personal communication of RETREAT preliminary geodetic results, 2008]. The GPS deformation field is the product of near-field elastic effects associated with locked upper crustal faults, as well as regional permanent deformation associated with shortening

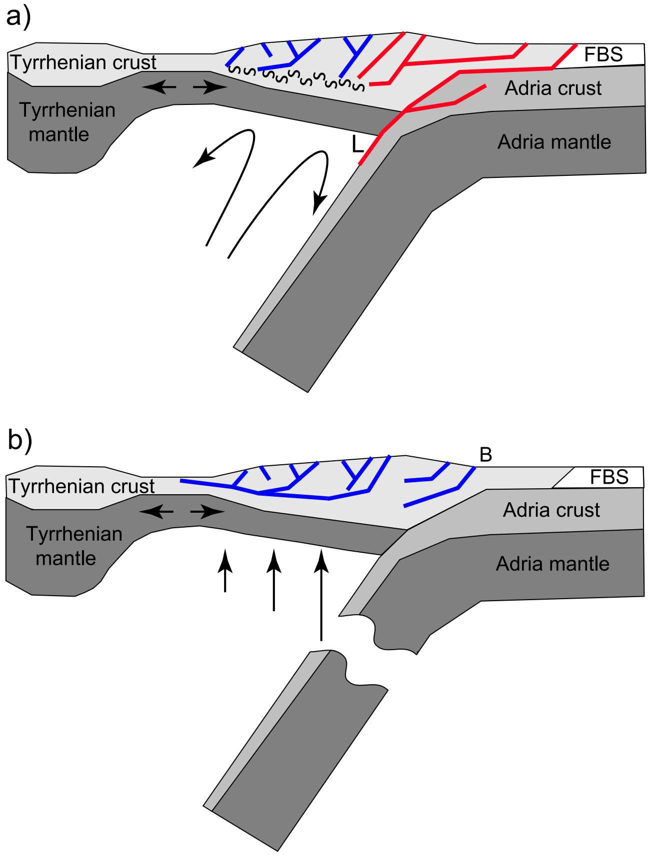

Figure 2. The two end-members for the Quaternary tectonics and geodynamics of the Apennines, defined by the degree of crust-mantle coupling that drives the extensional process. (a) Decoupled or partially coupled rollback-retreat (a generalization of a model being tested by the RETREAT working group) with the conceptual notion of a lithospheric-scale thrust defining the mountain front [L; *Lavecchia et al.*, 2003a]. FBS, foreland basin sediments. (b) Fully coupled upper plate retreat and slab detachment model with the conceptual notion of a normal fault defining the mountain front [B; see *Bertotti et al.*, 1997]. Blue and red are normal and thrust faults, respectively.

within the crustal wedge. The results from these studies agree that Corsica and Elba (i.e., the Tyrrhenian plate) move with and are parts of Europe, but they disagree in the relative movements of the Apennines and Adria with respect to the central Alps. Horizontal velocities projected on a north-east oriented line from Corsica through Bologna and ending in the eastern Alps argue for active stretching in the backarc and range crest, and shortening in the foreland focused on the current Apennine mountain front and eastern Alps mountain front [*Serpelloni et al.*, 2005; R. Bennett, personal communication of preliminary RETREAT results]. The Apennines apparently are moving northeast a little faster than Adria such that the stretching at and west of the range crest is mostly recovered across the foreland. In contrast, *Stein and Sella* [2006] report that the Apennines and Adria are not converging, but rather moving together toward the northeast as Nubia (Africa) continues to move northwest. This latter convergence direction has been documented in the northern Apennines by both deformation of late Cenozoic and Quaternary deposits [*Picotti et al.*, 2007] and seismicity [*Selvaggi et al.*, 2001: *Piccinini et al.*, 2006: *Santini*, 2003, *Santini and Martellini*, 2004] as well as transtensional crustal extension at and west of the range crest [*Cello et al.*, 1997; *Amoruso et al.*, 1998; *Piccardi et al.*, 1999; *Viti et al.*, 2004]. As shortly summarized above, the GPS geodesy results, particularly those of the RETREAT experiment, cannot be uniquely interpreted as an abrupt change in the northeast-directed horizontal velocity field in the vicinity of the Apennine mountain front alone; however, they are permissive of our general argument for the origin of this topographic feature.

2.2. The Slab Rollback: Upper Plate Retreat Model and Statement of the Problem

[9] The northern and eastern front of the northern Apennines (Figure 1a) is a striking, first-order topographic feature that has escaped a simple genetic explanation despite numerous studies. It remains unclear if the mountain front is an active tectonic feature, what structures, if any, underlie it, and in what large-scale geodynamic context it was constructed. Generally speaking, the answers to these questions are obvious from the well mapped geology that argues for the mountain front as being one expression of the Adria-Tyrrhenian plate boundary. Nevertheless, clear evidence of active tectonics consistent with plate boundaries such as historic great earthquakes, growing folds, and contemporary geodetic deformation has proven to be elusive. Particularly difficult to reconcile has been the disparity in individual data sets. For example, the results of earthquake seismology, which seem to support continued tectonic activity for the mountain front and the thrust front buried beneath the Po Plain [compiled in *Boccaletti et al.*, 2004, Figure 1c], are at odds with the large database of high-resolution reflection seismology that is equivocal at best in terms of identifying active folds and faults [*Pieri*, 1987; *Bertotti et al.*, 1997; *Picotti et al.*, 1997; *Argnani et al.*, 2003]. What is currently lacking is a synoptic overview of all of the pertinent geologic, geomorphologic, geophysic, and geodetic data of the northern Apennines arranged in the context of the rollback – retreat or competing models.

[10] We begin by framing the crustal deformation, rock uplift, and topographic growth of the northern Apennines in a conceptual geodynamic model where Adria is subducting and rolling back [*Dewey*, 1980; *Reutter*, 1981; *Royden*, 1988] beneath a Tyrrhenian upper plate that is retreating and extending (Figure 2). The long-noted paired compression – extension fronts [*Elter et al.*, 1975; *Malinverno and Ryan*, 1986] in the Apennines are explained in this model by two end-member variations. The first variation considers the paired fronts as responses of a steady-state or near steady-state, Platt-style orogenic wedge [*Platt*, 1986, Figure 2a]. The Platt-style wedge predicts no coupling between the Apennine crust and the retreating Tyrrhenian plate where the mass flux into the wedge by convergence must be balanced by the mass flux out of the wedge by extension and erosion, such that the wedge dimensions and partitioning of extensional and compressions regimes remain more or less uniform, and compressional deformation propagates into the foreland. Partial coupling between the upper (Tyrrhenian) plate and Apennine crust would predict that some of the observed extension is driven by the retreating

plate motion and some of it is driven by continued frontal accretion or underplating.

[11] The second variation considers the paired deformation fronts as relict features in a zone of crustal extension fully coupled to the retreating upper plate and eating into formerly shortened crust atop the Adria slab (Figure 2b). This variation does not require ongoing convergence, a balance of mass fluxes into and out of the Apennines, uniform wedge dimensions, partitioning of deformation, or foreland propagation of the thrust front. It does predict that a northeastward zone of active extension [e.g., *Frepoli and Amato*, 1997] marks the location of crustal coupling to the upper plate and that there should be a long-wavelength lithospheric deformation driven by the former coupled rollback – retreat mantle flow field or detached slab [*Wortel and Spackman*, 2000].

[12] The conceptual geodynamic model above must accommodate three key observations. First, over the past 2 million years, the Apennines have risen above sea level to lofty heights and high relief at the same time that the Po foreland has undergone broad subsidence. Quaternary deformation can be generalized as broad uplift of the range, and broad subsidence of the Po foreland. For example, south and west of the Apennines topographic crest normal faults are well known to have accommodated some thinning of the crust forming basins filled by Plio-Pleistocene marine and continental deposits [e.g., *Sartori*, 1990; *Spadini et al.*, 1995; *Martini and Sagri*, 1993]. By the middle Pleistocene, regional uplift brought the entire Apennines emergent [*Calamita et al.*, 1999; *Carminati et al.*, 1999; *Bartolini*, 2003] at a time when evidence for regional compression became established from Corsica [*Fellin et al.*, 2005] through the Po foreland. We treat this deformation as being consistent with some process operating at the lithospheric scale, directly attributable to dynamic topography generated by rollback – retreat coupling.

[13] Second, Quaternary marine and fluvial deposits preserved along the mountain front dip more steeply toward the foreland than can be easily explained by the broad uplift and subsidence pattern [*Bertotti et al.*, 1997; *Simoni et al.*, 2003]. Ever since the publication of reflection seismology lines beneath the Po Plain [*Pieri and Groppi*, 1981; *Bally et al.*, 1986; *Pieri*, 1987; *Hill and Hayward*, 1988], the model of the Apennines as a fold and thrust belt, and specifically as a thin-skinned fold and thrust belt has become solidified in many minds. As a result, the most commonly accepted interpretation on the origin of the mountain front and the foreland dipping strata is that of a blind thrust fault [*Vann et al.*, 1986; *Tozer et al.*, 2006], formed during the Quaternary as an out-of-sequence structure behind the recognized thrust front buried beneath the Po Plain. Around Bologna there are numerous papers that clearly identify the mountain front as being cored by a thrust fault [e.g., *Lipparini*, 1966; *Castellarin et al.*, 1985; *Amorosi et al.*, 1996; *Benedetti et al.*, 2003; *Vannoli et al.*, 2004; *Alvarez*, 1999; *Ghisetti and Vezzani*, 2002], with one study giving the structure the formal name of the Pedeapenninic Thrust Front (PTF) [*Boccaletti et al.*, 1985]. The recent seismotectonic map of the Emilia-Romagna Region of Bologna [*Boccaletti et al.*, 2004] shows the PTF as an emergent to blind thrust fault (Figures 1a and 1b), striking beneath Bologna and other large mountain front cities with obvious important seismic hazards. Some authors [e.g., *Doglioni et al.*, 1999; *Montone and Mariucci*, 1999; *Carminati et al.*, 2003a, 2003b; *Pierdominici et al.*, 2005; *Scrocca*, 2006; *Scrocca et al.*, 2007] propose that the main activity is still occurring on the thrust front, buried in the Po Plain (the Ferrara arc) rather than the mountain front, but there is no or little topographic expression of this activity or commonly deformed young reflectors in the seismic lines.

[14] Third, thrust faults and folds in the Ferrara and Emilia arcs and beneath the Bologna mountain front are mostly sealed by middle and upper Pleistocene deposits [e.g., *Toscani et al.*, 2006]; however, slow fold growth [*Ferranti et al.*, 2006] and river channel perturbations [*Burrato et al.*, 2003] suggests locally active blind structures or differential compaction [*Scrocca et al.*, 2007] of Po Plain sediments. Particularly along the mountain front northwest of Bologna [*Picotti et al.*, 2007; cf. *Benedetti et al.*, 2003] and along the northern Marche mountain front [*Di Bucci et al.*, 2003] there is clear evidence for northwest-directed shortening from geologic, seismic, and reflection seismology data. Similarly, the M5.3 2003 Monghidoro earthquake, located some 20 km south of the Bologna mountain front at a depth of approximately 15 km resulted from slip on a northwest verging reverse fault [*Piccinini et al.*, 2006]. The same orientation of main compressional axis has been reported for the M5.4 1996 Reggio Emilia earthquake, 60 km northwest of Bologna [*Selvaggi et al.*, 2001]. Similarly, small, but active high-angle faults with strikes oblique to orogen strike are mapped at the Bologna mountain front [*Bertotti et al.*, 1997; *Di Bucci and Mazzoli*, 2002; *Simoni et al.*, 2003]. These faults are not accompanied by actively subsiding basins trapping Quaternary sediment. We consider this type of deformation to represent shorter-wavelength processes acting at the crustal scale.

[15] Two other mountain front origin models have been proposed. One interprets the mountain front as lying in the footwall of a zone of high-angle normal faults [*Bertotti et al.*, 1997, Figure 2b, B]. This model notes that the mountain front segments mimic in scale and location the normal fault segments at and west of the topographic crest; however, *Bertotti et al.* [1997] interpretation falls short of explaining the much larger scale uplift and oroclinal bending of the mountain front over hundreds of kilometers in strike length. The other view for the origin of the mountain front involves a lithospheric-scale reverse fault that duplicates the Moho at depth and projects toward the surface as a thrust fault [*Lavecchia et al.*, 2003a, Figure 2a, L]. This large thrust represents lithospheric deformation that may be driven by a mantle plume [*Lavecchia et al.*, 2003b], viewed as ultimately responsible for the long history of syn-convergence and extension in the Apennines. Embedded in the plume-lithospheric thrust model is the assumption that the rollback – retreat process has stopped, or at the least slowed considerably in the Quaternary; however, seismic tomography beneath the northern Apennines supports an intact slab [*Piromallo and Morelli*, 2003; *Lucente et al.*, 2006].

3. Results

3.1. Po Foreland

[16] The sedimentary record of the Po – Adriatic foreland indicates deposition in one of two basin types, an asym-

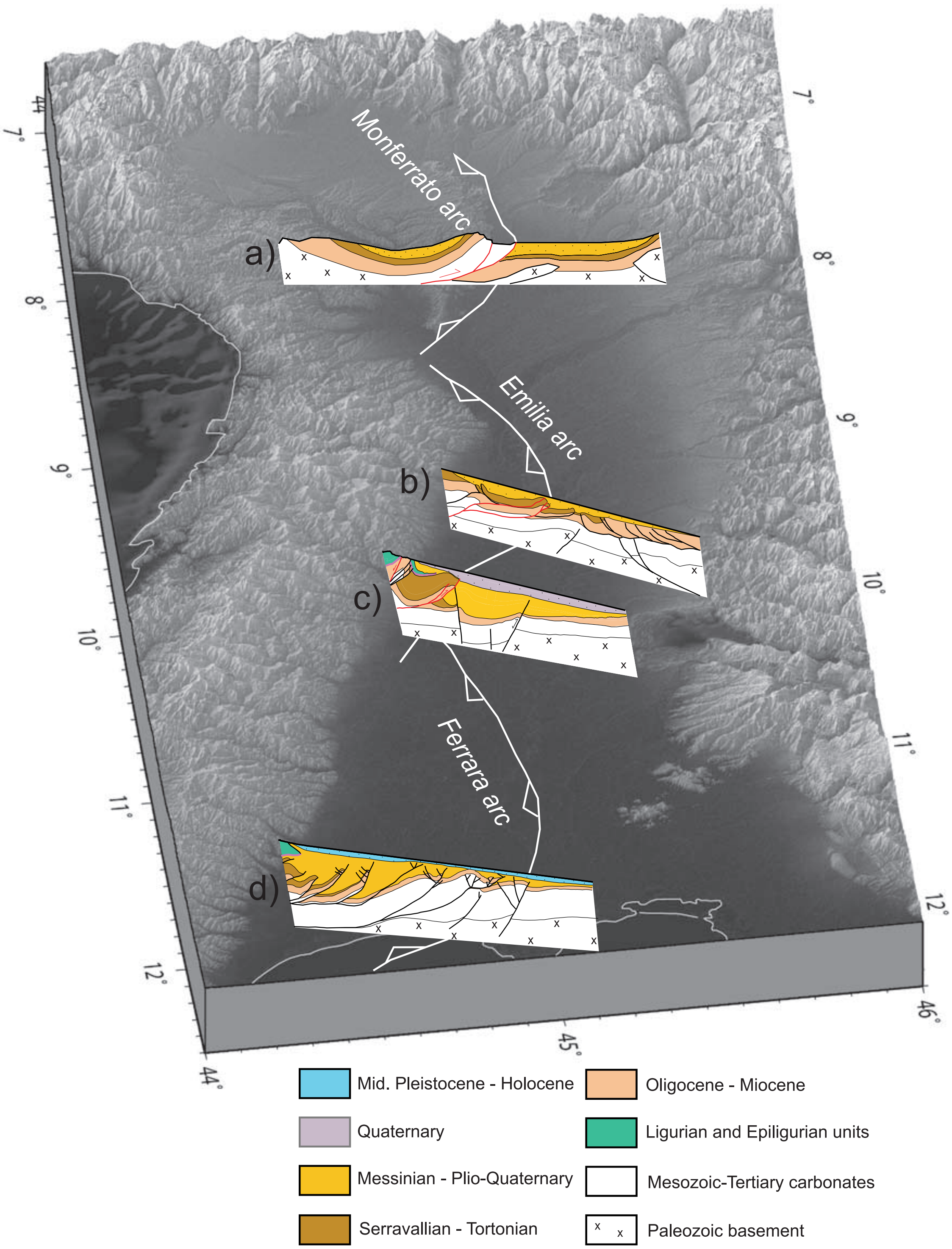

Figure 3. The geometry of the Po Plain basins. Transition from the Messinian-Plio-Quaternary wedge-shaped basin to the Pleistocene-Holocene sag basin (dotted patterns) began in the west (up on the figure) and progressed to the east (down on the figure). (a) From *Bertotti et al.* [2006]; (b) from *Bello and Fantoni* [2002]; (c) from *Picotti et al.* [2007]; (d) our data (see Figure 8).

metric, wedge-shaped basin with the deep part of the wedge adjacent to the thrust front, and a broad, symmetric sag basin where the deepest part does not necessarily coincide with the thrust front. The modern Po foreland geometry was established in the middle Pleistocene and is of the broad, symmetric sag-type [*Picotti et al.*, 1997]. Complete reconstruction of these basin geometries for the past is not possible for all parts of the foreland because of subsequent deformation; however, enough intact parts of both basin types are preserved to confirm the alternating geometries over 20 Ma (Figure 3).

[17] The wedge-shaped basins are foreland basins sensu strictu (and sensu *DeCelles and Giles* [1996]) often called foredeep basins in the Apennine literature, because of their remarkable bathymetry [*Ricci Lucchi*, 1986]. The wedge shape of these basins clearly thickens toward the thrust front, but is also variously compartmented by the coeval growth of thrusts [e.g., *Ricci Lucchi*, 1986; *Zoetemeijer et al.*, 1993]. It is this geometry, based largely on subsurface Po Plain data that argues for lithospheric flexure by the orogenic wedge load, plus other "hidden loads" [*Beaumont*, 1981; *Royden and Karner*, 1984; *Royden*, 1988], presumably related to the subducting slab.

[18] More recently, *Bertotti et al.* [2001] documented for the southern part of the Adriatic plate the presence of the other basin geometry, whose origin in the context of Apennine and Dinarides wedge loading is unclear. The main characteristics of the sag basins is that they share a comparable subsidence rate to the wedge basins, but the depocenter is somehow maintained in their center, with fixed margins, rather than a depocenter migrating with the orogenic load [see Figure 12 of *Bertotti et al.*, 2001]. Sag basins of this type have been interpreted as the downwarped portion of crust or lithosphere responding to long-wavelength buckle folds responding to intraplate stresses [*Cloetingh et al.*, 1999; *Bertotti et al.*, 2001; *Sokoutis et al.*, 2005; *Leever et al.*, 2006], or dynamic mantle topography [*Burgess et al.*, 1997].

[19] In the subsurface of the western Po Plain (around the Monferrato arc, see Figure 3) the Tertiary to Quaternary deposits [*Bertotti et al.*, 2006] are folded into a pair of anticlines and synclines with a wavelength of $\sim$40 to 45 km. The synclinal shape is typical of the sag basins (Figure 3), and it is well defined by the Messinian to present deposits that preserve the basin margins. Here in the westernmost portion of the Po foreland, only the sag-basin geometry is displayed.

[20] Further east in the Po foreland, both wedge and sag basin geometries are preserved and the transition from wedge-shaped to sag-shaped began in the west in the Messinian, propagated to the central Po foreland where at the Emilia arc made the transition from wedge to sag-shape in the early Pleistocene [Figures 1 and 3; Figure 3 of *Picotti et al.*, 2007], and then propagated further east, where, by the middle Pleistocene, the entire foreland had made the transition to the sag shape basin [*Pieri and Groppi*, 1981; *Pieri*, 1987; *Bello and Fantoni*, 2002]. For example, at the Ferrara arc, a Messinian to Early Pleistocene wedge basin developed with a maximum thickness of about ten kilometers. The transition to a sag basin occurred in the middle Pleistocene with the subsidence of the Ferrara arc and its subsequent burial in 100–500 m of middle and late Pleistocene deposits (Figure 3). Stratal geometries visible in seismic lines across the Ferrara arc clearly show how this formerly growing anticlinal fold was an island in the early Pleistocene arm of the Adriatic Sea, until being buried completely in the late Pleistocene.

[21] Formation of the wedge basins temporally coincides with rapid foreland propagation of the thrust front. There are at least two times when wedge basins dominated the Apennine foreland, in the early Miocene and in the Messinian-Pliocene [*Fellin et al.*, 2005], although the precise reconstruction of the older of these two is difficult to reconstruct, given subsequent deformation. In contrast, formation of the sag basins seems to coincide with the stalling of the thrust front and out of sequence trusting in the hinterland of the orogenic wedge. The Tortonian and, locally, the Serravallian deposits in the Emilia and Ferrara arc appear as sag basins in the best published seismic sections [e.g., *Pieri and Groppi*, 1981; *Bello and Fantoni*, 2002; *Picotti et al.*, 2007, see Figure 3] and sedimentological studies [e.g., *Roveri et al.*, 2003] confirmed this geometry with respect to the early Miocene wedge stage. This model of alternating wedge and sag basins linked to the propagation and backstepping of the thrust front makes a prediction that the westernmost Po foreland should be dominated by sag basins because of the less overall convergence. The study of *Bertotti et al.* [2006] and our simple reconstruction both show that the western Po foreland is dominated by sag basins, the central Po foreland (Emilia arc) records an intermediate behavior, with a Messinian to Quaternary sag basin superimposed on an older Miocene wedge basin, and the eastern Po foreland recording at least two phases of wedge to sag transitions since the early Miocene. The details of the Messinian to Quaternary transition for the Ferrara arc is more fully explored by the data presented below.

3.2. High-Resolution Reflection Seismology

[22] A high-resolution reflection seismic line was acquired across the mountain front in November of 2005 near the village of Ponte Ronca (Figure 4) with the aim of imaging the first 300 to 400 m of the subsurface. The line is 1930 m long, and used a vibratory energy source (Ivi minivib), sweeping linearly at frequences ranging 10 to 200 Hz. The tight spacing of receivers (5 m) allowed a good cover and a large number of traces. The vertical resolving power is about 12 m (details are forthcoming in Bruno et al., submitted manuscript, 2008). The main target we expected to image in this line, was the PTF, which has been mapped by several authors as a northeast-verging emergent thrust, forming the mountain front and offsetting Pleistocene and locally Holocene deposits [e.g., *Benedetti et al.*, 2003; *Boccaletti et al.*, 2004; *Lavecchia et al.*, 2004]. Two migrations, a low-resolution preliminary and high-resolution detailed migration have been completed, the low-resolution one of which we report here. The seismic line shows excellent reflectors and images of the subsurface notwithstanding two short interruptions where the line crosses the railway and main mountain front-parallel road (Figure 4b). The first 50 to 30 m of deposits have not been imaged, because of the absence of useful offsets in the reflected signal.

[23] Interpretation of the line is anchored by the excellent imaging of the Sabbie Gialle deposits. The line illuminates the whole interval of the Sabbie Gialle (Sabbie di Imola, IMO 1, IMO 2, and IMO 3 in the official cartography), the uppermost Lower Pleistocene near-shore deposits representing the final phase of marine conditions for this portion of the Po foreland. This interval, calibrated by means of four wells (Figure 4) is characterized by high-amplitude reflectors and good continuity at its base, located around 200 m below sea-level at the northern end (beginning) of the line. As a whole, the thickness of this package decreases southward toward the mountain front from around 100 m to 40.

[24] Excellent exposures of IMO 1–3 in quarries along the foothills [see *Amorosi et al.*, 1998] as well as natural

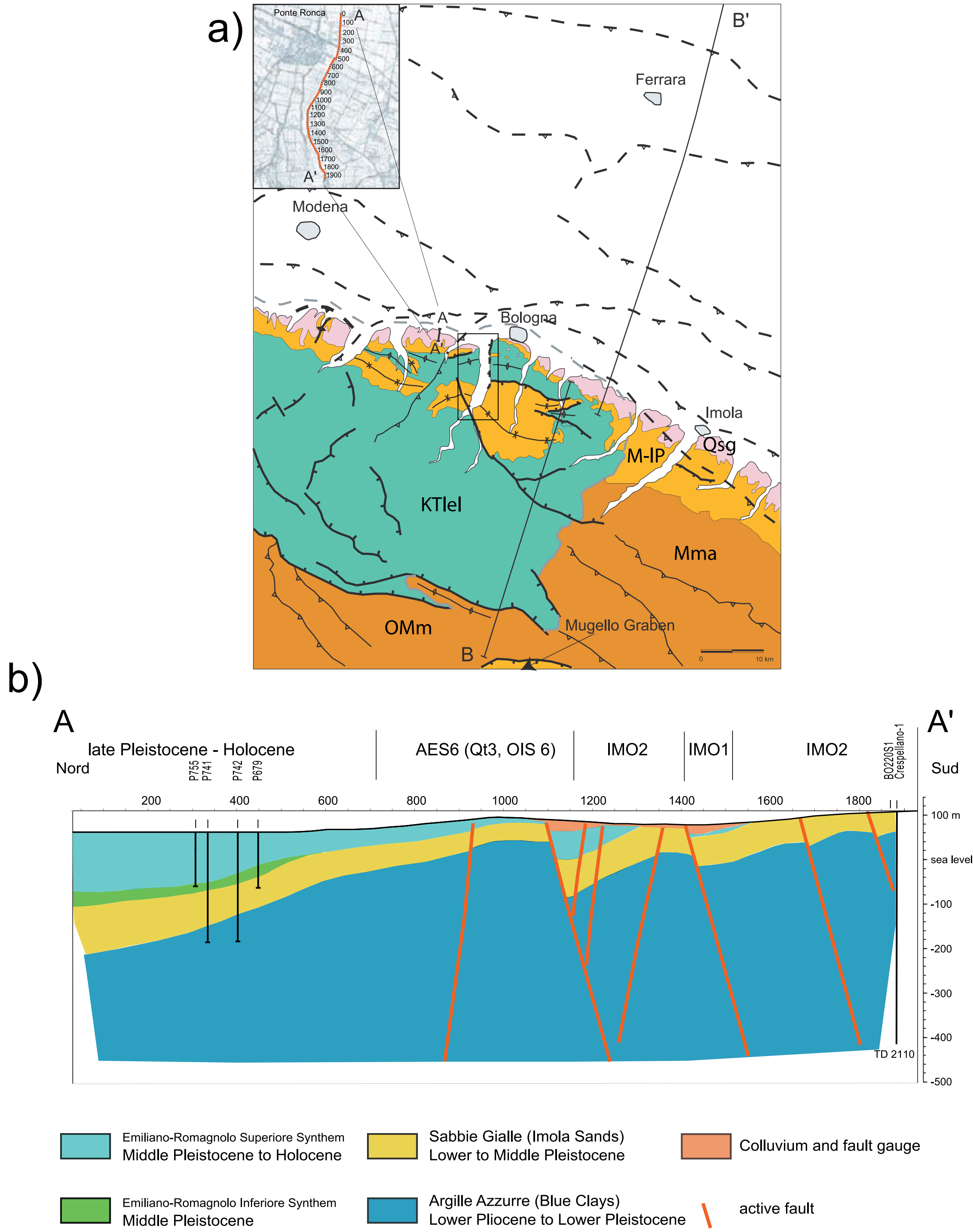

Figure 4. (a) Structural sketch of the northern Apennines around Bologna, with track of the cross-sections of Figure 4b (A–A′) and Figure 8 (B–B′). (b) Preliminary geologic interpretation of the Ponte Ronca seismic line from a preliminary coarse depth migration (not shown) and scaled for no vertical exaggeration. The stratigraphic nomenclature above the interpretation is from the 1:10,000 scale maps of Servizio Geologico Emilia-Romagna and *Martelli et al.* [2008] and defined in Table 1. Note the local mismatch between surface geology and line interpretation. Vertical lines between stratigraphic units indicate contacts on the 1:10,000 scale geologic map. Well logs are on file with the Servizio Geologico Emilia-Romagna and are courtesy of Paolo Severi. The box indicates the extent of the mapped area in Figure 5. Qsg, Sabbie Gialle (IMO1 and IMO2); M-lP, Miocene and lower Pliocene marine sediments including Pliocene Argille Azzure; Mma, Miocene Marnoso-arenacea; OMm, Oligocene-Miocene Macigno Fm.; KTlel, Cretaceous and Tertiary Ligurian and epi-Ligurian rocks.

Table 1. Reno Terrace and Po Plain Stratigraphy and Ages

Terrace	Regione Map[a]	Po Plain Aquifer Group[b]	MOIS[c]	Age (Ky)	Source[d]
Qt9	AES 8a	A1	1	1.5 ± 0.5[d]	^{14}C
Qt8	AES 8a	A1	1	6 ± 1[d]	^{14}C
Qt7	AES 8	A1	1	9 ± 1[d]	^{14}C
Qt6	AES 8	A2	2	13 ± 2[d]	^{14}C
Qt5	AES 8	A2	2	22 ± 2[d]	^{14}C
Qt4	AES 7b	A2	3 (4?)	30 ± 5[d]	^{14}C, (K. W. Wegmann and F. J. Pazzaglia, Fluvial terrace straths as growth strata markers of late Quaternary tectonic deformation in the Emilia-Romagna and Marche Apennines, submitted to *Basin Research*, 2008)
Qt3b	AES 7a	A3	6	140 ± 10	*Amorosi et al.* [1996]; *Di Dio* [1998]; *Martelli et al.* [2008]
Qt3a	AES 7-6	A4	8	250 ± 25	*Di Dio* [1998]; *Martelli et al.* [2008]
Qt2	AEI; AES 5-4-3	B3-B2-B1	14-12-10	440 ± 25	*Di Dio* [1998]; *Martelli et al.* [2008]
Qt1	AEI; AES 2-1	B4	16	620 ± 25	*Amorosi et al.* [1998]; *Martelli et al.* [2008]
Qt0	IMO 3	C1	17	700 ± 25	
Qt0/Qsg	IMO 2	C2	18	750 ± 25	*Martelli et al.* [2008]
Qt0/Qsg	IMO 1	C3	19	780 ± 25, 900 ± 400	^{10}Be, Pazzaglia et al. [in review]

[a]AEI, Emiliano-Romagnolo Inferiore Synthem; AES, Emiliano-Romagnolo Superiore Synthem, IMO, Imola Sand Formation (Sabbie Gialle) [*Martelli et al.*, 2008].
[b]*Di Dio* [1998].
[c]Marine oxygen isotope stage.
[d]Average calibrated ^{14}C ages from multiple dates reported.

exposures in the river valleys south of Ponte Ronca provide outcrop analogues for these seismic facies; but we note the lack of perfect correspondence between our interpretation of the seismic reflectors and the mapped geology (Figure 4b). Direct correspondence to the surface geology is afforded less weight, given the overall poor exposure of these units and the uncertainty in precise locations of unit contacts, but it remains an important guiding constraint. The basal continuous reflector represent the unconformable base of the unit, abruptly capping the Lower Pleistocene marine mudstones (Argille Azzurre), and the base of the lower phase of Sabbie Gialle deposition called Imola Sand 1 (IMO 1).

[25] Approximately 1 km south of the end of the seismic line, these units are truncated by erosion, forming the triangular heads of erosional flat-irons. The younger Sabbie Gialle units, (IMO 2–3) are less deformed and more continuous than IMO 1. It forms a more or less continuous cap on the mountain front, rising from the plain just south of Ponte Ronca, and projecting southward at a dip of less than 10° to the flat iron head with the only interruption coming at the normal fault. The variable geometries of the internal reflectors are particularly well-imaged in this unit and represent the various sub- environments in this shore deposits, spanning from foreshore to shoreface to backshore facies with a general northward progradational pattern.

[26] The Sabbie Gialle units are unconformably overlain by alluvial fan gravels of the lower and upper Emiliano-Romagnolo synthems (AEI and AES, Table 1). These units are not well imaged except for a few bright reflectors under the Po Plain; however, their presence, general geometry, and thickness are confirmed by both surface exposures and the wells. Deposits of AEI are middle Pleistocene sand and gravel deposits correlative updip to river terraces mapped as Qt1 and Qt2. Similarly, deposits of AES are late middle to late Pleistocene sand and gravel deposits correlative updip to river terraces Qt3-Qt9. Detailed descriptions of these Pleistocene terrace deposits follows below.

[27] The main target of the seismic line, the PTF as an emergent or nearly emergent blind thrust, was not imaged. These results are consistent with numerous industry lines, include those used to make Figures 1d and 8. Rather, the general structure revealed in the seismic line is the foreland-dipping limb of a large anticline, with clear growth strata beneath the Po Plain and unconformities separating uplifted Sabbie Gialle, alluvial fan, and terrace deposits on the mountain front. The unconformity at the base of IMO 1 is rotated 18° toward the foreland indicating a forelimb tilt rate of ~23°/Ma. The anticline carapace is stretched by normal faults, the largest of which drops the base of IMO 1 ~140 m near shot point 1400. The geometry and scale of this normal fault is the same as those described for the mountain front 20 km to the southeast [*Bertotti et al.*, 1997], and within the Reno valley.

3.3. Reno Terraces and Rates of River Incision

[28] The northern Apennines are drained toward the foreland by numerous parallel rivers, the largest of which is the Reno River (~750 km^2) with headwaters rising on the range crest and mouths emptying at the mountain front onto large, low-gradient alluvial fans that spill out into the Po Plain. The Reno alluvial fan is also among the largest on the Po Plain and its subsurface stratigraphy is known in detail [*Di Dio*, 1998] because of the numerous water supply wells that have been drilled for Bologna and surrounding communities [e.g., *Amorosi et al.*, 1996]. The Reno River is flanked by a flight of nine fluvial terraces that record the incision of the channel into the bedrock of the Bologna mountain front (Figure 5). The terraces are well preserved for nearly 20 km south of the mountain front, and a particularly continuous reach of terraces lies between Sasso Marconi and Casalecchio di Reno on the west side of the valley. The Reno channel is steep and broadly convex as it approaches the mountain front. It lies at an elevation of ~50 m at the mountain front and remains a mixed bedrock-alluvial channel upstream, allowing for numerous anthropogenic features and activities such as dams and gravel bed mining. The channel has little room for non-tectonically driven incision into bedrock, especially at the mountain

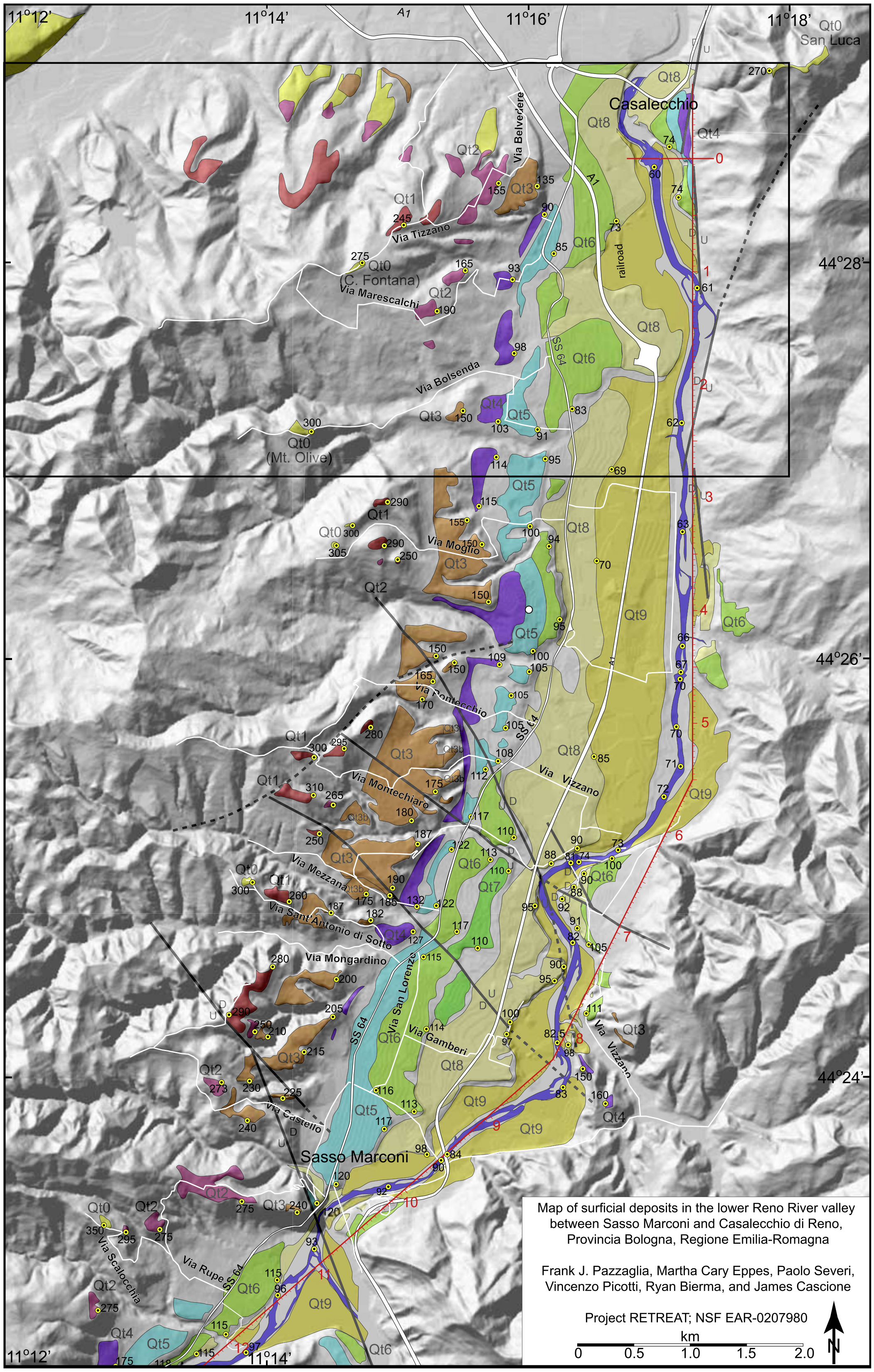

Figure 5. Map of terraces in the lower Reno valley between Sasso Marconi (south) and Casalecchio di Reno (north). Yellow points are elevations in meters. Topography in the black outline is used for the swath profile of Figure 6a. The red line is the valley plane profile to which all longitudinal profile data are projected in Figure 7. Terrace stratigraphy is illustrated in Figure 6 and Table 1.

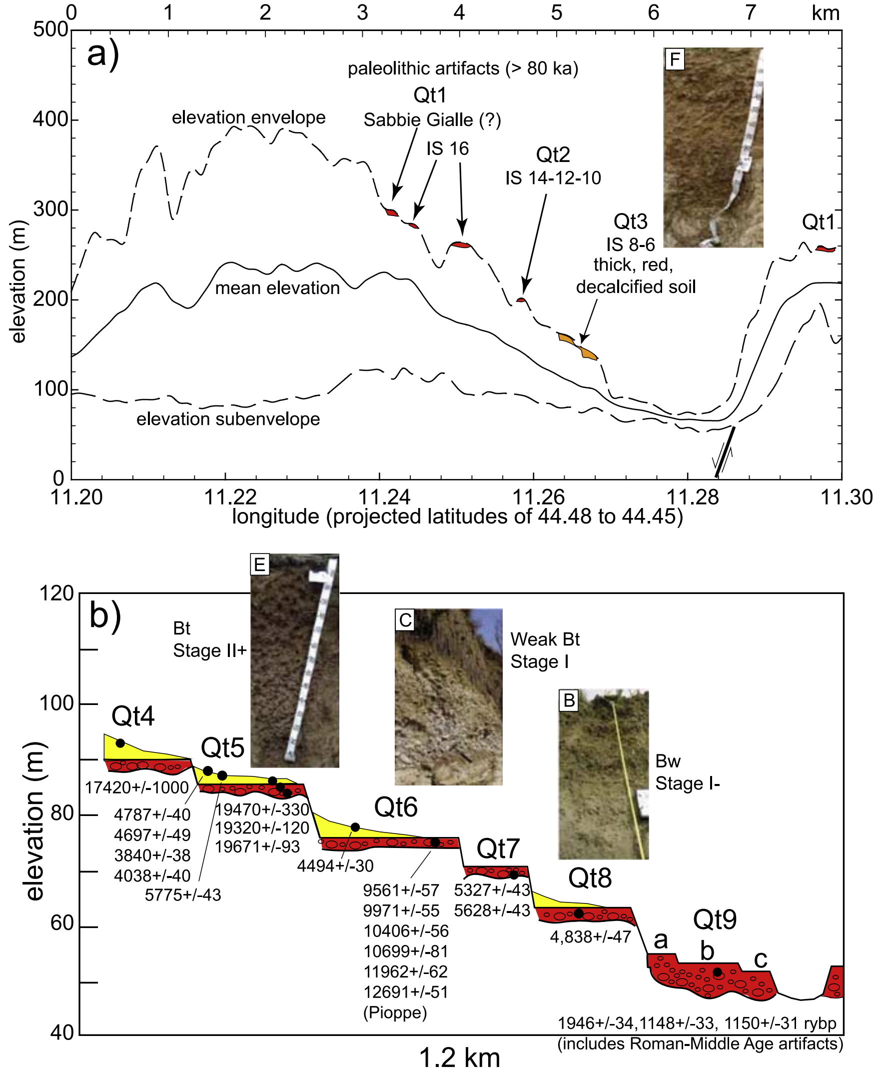

Figure 6. (a) Topographic swath profile of the lower Reno valley between latitudes 44.48 and 44.45 with terrace locations and inferred ages. (b) Composite cross-section of the inner Reno valley showing the late Pleistocene terraces and numeric ages. Red: alluvium; yellow: colluvium. Inset photographs of soils F, E, C, and B illustrate changes in relative weathering characteristics of the terrace deposits and are described in detail in *Epps et al.* [2008].

front, given its proximity to sea level. Pleistocene glacio-eustatic drawdowns did generate knickpoints on the Po River and some of its major tributaries, but none of these knickpoints had migrated to Bologna or beyond on the Reno channel before being drowned and arrested by the ensuing eustatic rise and transgression [*Amorosi et al.*, 2004].

[29] The terraces of the Reno and other northern Apennine streams are mostly strath terraces with thin alluvial fills atop broad, low relief unconformities carved into bedrock (Figure 6). The Reno terraces are composed of two major facies, both of which thicken where in proximity to a side valley sediment source such as a tributary stream. The basal facies is well sorted and stratified axial stream gravel and sand, ~1–3 m thick with a provenance including the hard sandstone of the Macigno and Cervarola formations as well as carbonate and chert from Ligurian units that outcrop in the Reno headwaters. The top facies is an overbank sand and silt 1–4 m thick that exhibits evidence of pedogenesis and bioturbation. The overbank deposits are commonly overlain by colluvial, alluvial fan, and/or eolian deposits that can locally be up to 10 m thick and obscure the original terrace tread (top). The problem of younger colluvial and alluvial fan deposits burying terraces, such that the elevation of the basal strath is obscured, is more common for the older terraces.

[30] The Reno terraces are organized into two major groups, an upper suite of four terraces and gravel remnants that start at the basin divide and step down to a broad, prominent, continuous bench etched into the valley side, and a lower suite of five continuous terraces that step down to the active floodplain (Figure 6). Upper suite terrace Qt0, Qt1, Qt2, and Qt3 begin as scattered sand and gravel remnants on isolated hilltops (Qt0) at elevations typically above 250 m and step down to one or more stratified deposits that mantle the prominent mid-valley bench (Qt3). These deposits all show signs of intense pedogenesis, erosion, or post-depositional burial by younger colluvial

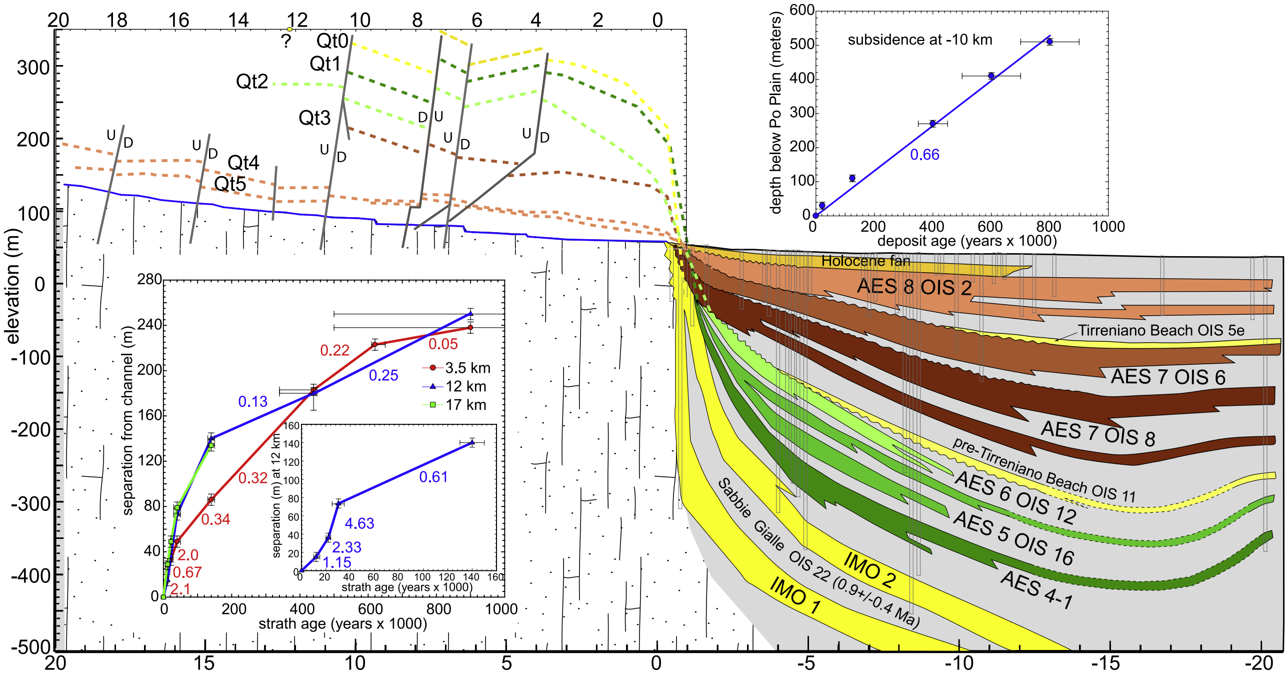

Figure 7. Projection and correlation of Reno terrace data across the mountain front to subsurface alluvial stratigraphy in the Po Plain. Numbers on inset graphs are incision or subsidence rates in millimeters per year. OIS, oxygen isotope stage, is used for stratigraphic correlation of deposits and terraces for which there exists no independent numeric ages. Age of IMO (Sabbie Gialle Qsg; cosmogenic; 0.9 ± 0.4 Ma) is from *Cyr and Granger* [2008].

and alluvial fan deposits. There are no numeric ages from these deposits, but geomorphic, stratigraphic, pedogenic [*Epps et al.*, 2008], and anthropologic [*Amorosi et al.*, 1996] observables are used to constrain relative ages and provide a basis for long-valley correlation. In contrast, the lower suite of terraces mostly lie within radiocarbon age and have been dated in several places (Table 1). Furthermore, the soils of these lower terraces are well preserved and have been assembled into a calibrated chronosequence which allows for a robust extrapolation of correlated terraces throughout the basin [*Epps et al.*, 2008].

[31] Ages of Qt5 and Qt6, the widest and best-preserved pre-Holocene terraces of the lower terrace suite, have calibrated calendar radiocarbon ages of ~23 ka and 11 ka respectively. These ages argue for wide strath carving and terrace deposition during cold phases of the Pleistocene when delivery of coarse sediment in the headwaters was enhanced. The Reno alluvial fan is known to have prograded and deposited sand and gravel during this time, a response consistent with an increased sediment load [cf. *Amorosi et al.*, 1996]. We appeal to this model of Reno strath carving and deposition of strath terraces during times of alluvial fan aggradation of coarse material, both of which are driven by cold climates in the Pleistocene (Figures 6 and 7). The intervening warm periods in the Pleistocene and Holocene are times of alluvial fan entrenchment, deposition of fine grained material on the Po Plain, and Reno River incision. In this way, incision of the Reno valley has been unsteady for time spans less than 50–100 k.y.; however, the Reno terrace record likely exceeds 100 k.y. so the long-term incision record is directly related to the long-term rate of rock uplift.

[32] Our model for strath terrace genesis, supported by the numeric ages of the lower terrace suite, allows us to use the marine oxygen isotope record of Pleistocene glacial-interglacial climate change to estimate the age of the upper terrace suite as well as suggest correlations between the terraces and their down-dip correlative deposits in the Reno alluvial fan (Figure 7; Table 1). We rely on several studies that detail the rationale and procedures for this type of terrace to basin deposit correlation [*Pazzaglia*, 1993; *Pazzaglia and Gardner*, 1994; *Pazzaglia and Brandon*, 2001] as well as excellent local mountain front exposures along strike that clearly show a marker horizon such as IMO 1 rising out of the Po Plain, coarsening up dip, and becoming a mapped terrace (Qt0) at the apex of a mountain front flatiron [*Martelli et al.*, 2008]. The highest and oldest terrace remnants (Qt0) are likely fluvial equivalents to the Sabbie Gialle (IMO) deposits, Qt1 and Qt2 represent AEI and older AES units, and Qt3, represented by at least two distinct treads (Qt3a and Qt3b), but only one indistinct strath, is a late middle Pleistocene (~140–250 ka) terrace. The results, when plotted on a common vertical reference profile oriented down the centerline of the Reno valley shows the separation of the terrace from the modern channel and the variations in that separation presumably driven by variable rates of incision in response to non-uniform rock uplift.

[33] Two key patterns emerge in the incision data (Figure 7). The first is that the terraces rise very steeply from their Po alluvial fan equivalent deposits at the mountain front, and bend tightly within the first 3 km of the river valley to project more or less parallel to the modern channel further upstream. The warped terrace profiles are reminiscent of one limb of a growing fold where the straths

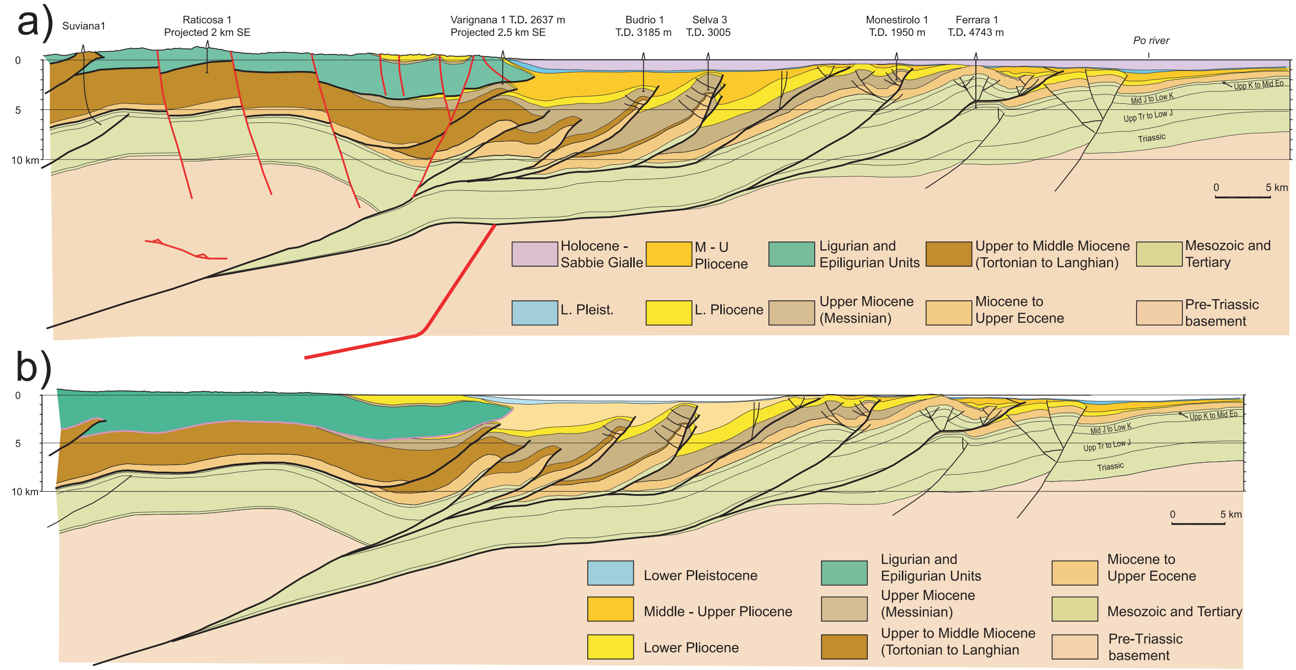

Figure 8. Geological cross-sections for (a) modern and (b) restored for the early Pleistocene. See Figure 4 for cross-section locations. The Po Plain portion comes from a reinterpretation of the seismic line C of *Pieri* [1987]; the Apennines transect comes from a reinterpretation of the cross-section IV of *Martelli and Rogledi* [1998]. The red lines in Figure 8a are active faults. The line denoting a fault in and out of the plane of the cross-section represents the Monghidoro earthquake [M 5.3, 2003; *Piccinini et al.*, 2006]. The deep ramp structure is inferred from modeling results of observed deformation and river incision (see Figure 9).

represent progressively deformed unconformities. The second observation is that the terraces are offset by high-angle faults, some of which may show a growth pattern. These faults are mapped (F. J. Pazzaglia, unpublished data; V. Picotti et al., Topographic expression of active faults in the foothills of the Northern Apennines, in *TOPOEUROPE*, edited by S. Cloetingh et al., submitted to *Tectonophysics*, 2008) as striking for 1–10 km with several 10s to 100s m of offset. Although the amount of slip on an individual fault is not great, collectively, they have down dropped the highest terrace remnants from the mountain front so that these deposits are still locally preserved upstream.

[34] Rates of Reno incision are calculated near the mountain front, at Sasso Marconi some 12 km from the mountain front, and at 17 km from the mountain front where terrace preservation begins to decline. The results show that from ~800 ka to ~150 ka, the Reno River incised slowly at about 0.2 – 0.3 mm/year. Since ~150 ka, the rate of incision has increased to ~1mm/year. The precise incision rate depends upon the location with respect to an individual fault or the frontal warping, but the result is the same, slower incision rates early, and an acceleration of the rates since the late middle Pleistocene. The long-term subsidence of the Reno alluvial fan over the same 800 k.y. time span has been steady at ~0.67 mm/year (Figure 7).

4. An Apennine Cross-section, Trishear and Flexural Modeling, and Quaternary Tectonics

[35] We produce a new crustal cross-section of the northern Apennines orogenic wedge from the water divide into the Po Plain, up to the buried structures of the Po Plain of the Ferrara arc (Figure 8, see location in Figure 4). The section has been pieced together from merging and reinterpreting the *Martelli and Rogledi* [1998] mountain section and the depth-converted C-C'' seismic line of *Pieri* [1987] in the Po Plain. We interpreted this section into depth to obtain a first-order balance. Well-control has been updated with respect to *Pieri*'s [1987] interpretation. The complexity of the section reflects a long structural history best recorded by growth strata. The general geometry is of the oft-cited fold and thrust belt [*Bally et al.*, 1986; *Pieri*, 1987; *Hill and Hayward*, 1988], with a surficial gravity nappe and a lower thrust belt, but the total amount of shortening is relatively small and probably in the range of 75–100 km for the whole belt [*Hill and Hayward*, 1988].

[36] The mountain front is well imaged in this section, the most important feature being the change in strata geometry represented by Early to Middle Pleistocene deposits. The Lower Pleistocene (blue interval in Figure 8) still records the wedge shape of the underlying basin, with maximum thickness close to the present day mountain front, covering and sealing the tip of the Ligurian units. The Middle Pleistocene Sabbie Gialle shows an opposite trend, with a divergent pattern thickening toward the Po Plain, thinning toward the mountain front, with the same geometry seen in detail in the high-resolution seismic line (Figure 4). Sabbie Gialle clinoforms abruptly change from a sigmoidal to a shingle pattern north of the mountain front and generally show a northward progradational trend. Separated by a pronounced couplet of reflectors, the Sabbie Gialle is

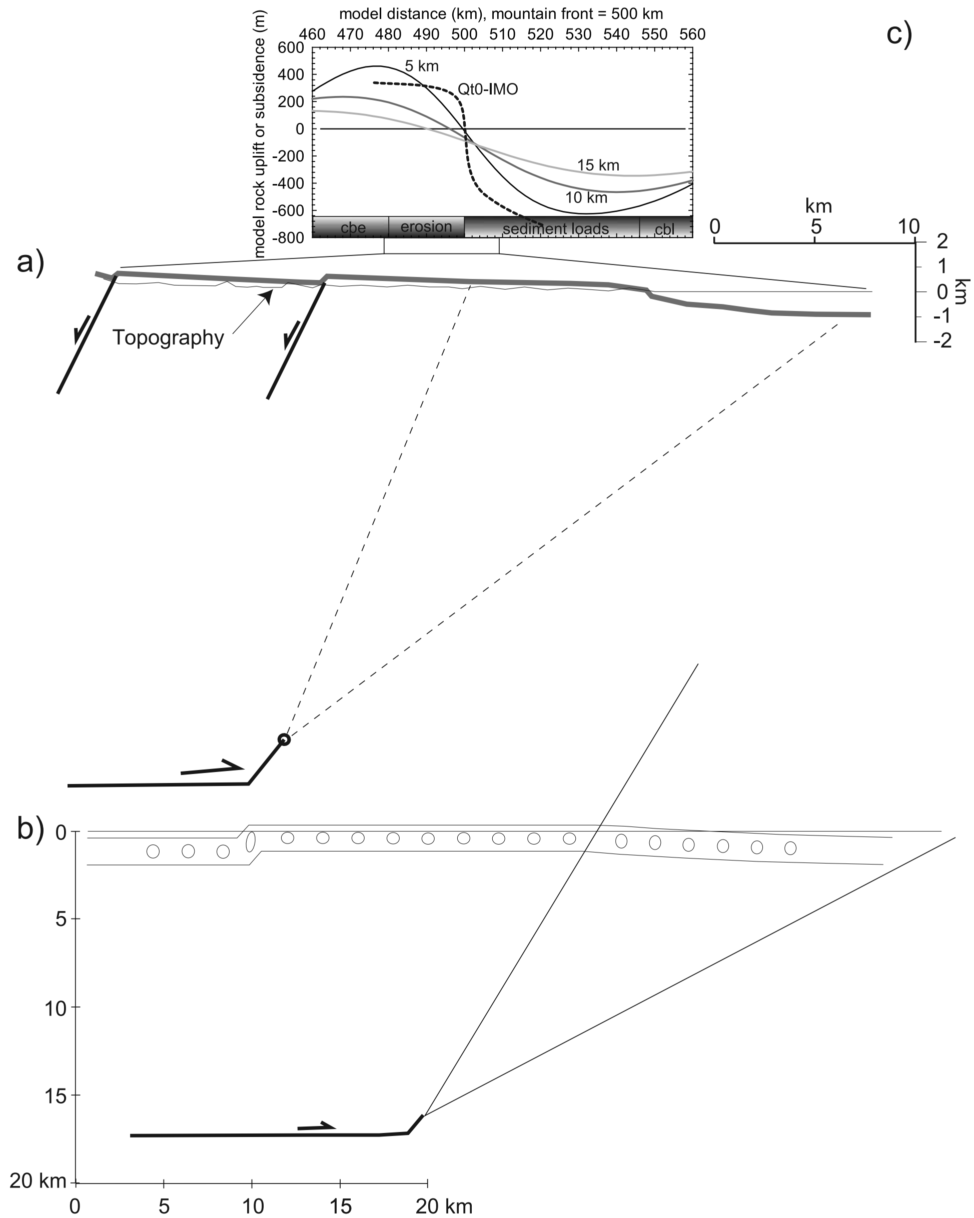

Figure 9. Results of trishear modeling inverting for the depth, dip, and slip rate on a buried thrust. (a) Best fit results to the IMO stratigraphic horizon (thick gray line) are a fault tip at a depth of 16 km, with a ramp angle of 55°, a slip to propagation ratio of 1.5, a trishear angle of 30°, and total slip of 2 km. (b) Predicted strain at the surface given the trishear parameters in Figure 9a. Note the correct prediction of the location of the observed main normal faults. (c, inset) Deformation across the mountain front considering flexure of regionally compensated vertical loads applied to an elastic plate of finite thickness and an erosion rate of 0.8 mm/year. Labeled lines correspond to different modeled elastic thicknesses (5–15 km). Qt0-IMO, real deformation of the terrace-Sabbie Gialle stratigraphic horizon; cbe, constant model boundary erosion; cbl, constant model boundary loading. See Table 3 for flexural parameters used in the modeling.

covered by AEI and AES. Between the mountain front and the Ferrara arc, the Middle Pleistocene to Holocene section is more than 1 km thick (Figure 8a). Toward the Monestirolo structure on the Ferrara arc, the Sabbie Gialle disappear, showing a parallel onlap, apparently because Monestirolo and Ferrara were previously emergent and eroded highs that were subsequently drowned and buried by the Sabbie Gialle and younger alluvium. The parallel reflectors seal the inactive Monestirolo and Ferrara anticlines. Even accounting for the fact that rapid and thick accumulation of sediment over growing structures tends to minimize the discordance in growth strata, any continued growth of these anticlines since the middle Pleistocene would have to be slow and subordinate to the fact that the entire arc is subsiding and all capping reflectors are virtually parallel (Figure 8a).

[37] We depict two important, deep, active compressional structures (shown in red around 15–20 km of depth in Figure 8a). One is the fault plane that slipped in the M 5.4

Table 2a. Predicted Total Fault Displacement of a Modeled Blind Thrust Fault With a Trishear Geometry Fit, Fixed P/S ratio[a] = 1 and 800 m of Qt0-IMO Stratigraphic Displacement[b]

Fault Depth[a]	Fault Dip[a]	Trishear Angle[a]	Fold Half Wavelength[a]	Total Displacement[a]
5	30	10	2010	1650
5	30	30	10,720	1650
5	30	50	> 25,000	1650
5	40	10	1150	1300
5	40	30	5460	1300
5	40	50	13,210	1300
5	50	10	750	1050
5	50	30	3890	1050
5	50	50	8020	1050
10	30	10	5500	1700
10	30	30	> 20,000	1700
10	30	50	> 25,000	1700
10	40	10	3170	1400
10	40	30	12,500	1400
10	40	50	> 25,000	1400
10	50	10	2100	1100
10	50	30	8830	1100
10[c]	**50**	**50**	**17,130**	**1100**
15	30	10	8900	1700
15	30	30	> 35,000	1700
15	30	50	> 73,000	1700
15	40	10	5200	1300
15[c]	**40**	**30**	**20,000**	**1300**
15	40	50	> 45,000	1200
15	50	10	3720	1050
15	50	30	13,200	1100
15[c]	**50**	**50**	**26,300**	**1050**
20	30	10	12,500	1650
20	30	30	> 50,000	1650
20	30	50	> 50,000	1650
20	40	10	7390	1300
20[c]	**40**	**30**	**27,000**	**1300**
20	40	50	> 50,000	1300
20	50	10	5130	1050
20[c]	**50**	**30**	**18,030**	**1050**
20	50	50	38,700	1050

[a]*Allmendinger* [1998], using FaultFold 4.5.4.
[b]Highlighted rows = fold half wavelength ~22 km.
[c]Best fit model parameters to field data.

Monghidoro 2003 earthquake [*Piccinini et al.*, 2006]. This structure is a lateral ramp at a depth of ~20 km that slipped with a top toward the southeast during the 2003 earthquake. Deep thrust-sense earthquakes like the 2003 event are the most common kind of seismicity documented for the northern Apennines, including the wedge tip buried beneath the Po Plain [*Boccaletti et al.*, 2004]. Second, we show a blind reverse fault as part of a flat-ramp structure below the mountain front that we argue to be the key structure responsible for broad deformation of the mountain front, the foreland tilting of young strata there, and the development of active normal faults. Note that movement on this structure would currently be restricted to the steeply-dipping ramp portion.

[38] The reverse fault is modeled using the inverse, trishear approach [e.g., *Erslev*, 1991; *Hardy and Ford*, 1997; *Allmendinger*, 1998] using the publicly available FaultFold program (http://www.geo.cornell.edu/geology/faculty/RWA/programs.html; Figure 9). The trishear approach enjoys relatively simple kinematics and seemed closest to the geometry of our case history. Trishear kinematics are well described and based on the constancy of the triangular area symmetric to and located at the tip of a fault undergoing hanging wall translation. The model accepted as input the observed geometry of the foreland-dipping strata at the mountain front including the Reno River terraces and their down-dip alluvial fan equivalents in the Po foreland.

[39] We applied a crude Monte Carlo approach to modeling a family of different fault geometries keeping ramp angle, trishear angle, and fault tip depth as adjustable parameters (Table 2a). Sensitivity experiments demonstrated that other parameters such as the slip to propagation ratio (S/P) are not important in our analysis, particularly for any value ≥ 1. We can easily distinguish the best fit of a deforming marker bed embedded in the model and the Qt0 – IMO horizon visually given the sensitive range of model responses to modest variations in ramp angle, trishear angle, and fault tip depth shown in Table 2a. The key parameters we look to fit are the 800 m of stratigraphic offset and the 22 km half wavelength of the Qt0-IMO fold as it drapes the mountain front (Figure 9). Exploration of fault tip depth, fault dip, and trishear angle indicates that deeper faults generally reproduce the requisite observed fold half wavelength (Table 2a). When considering just the emergent portion of the fold, defined by the Qt0 terrace and the exposed part of IMO, the best model predicts a ramp angle of 55°, a trishear angle of 30°, and a fault tip depth of 16–15 km (Table 2b). This rather deep location of the fault tip is required in order to obtain an almost equally wide uplift in the hanging wall block, to match with the observation of a broad anticline limb with the maximum uplift located at around 12 km from the mountain front (see Figure 7). The predicted location places the active reverse fault in the middle crust and in the basement of the thrust sheets depicted in Figure 8. The best fit total slip is ~2 km, which accounts for ~2.5 mm/year slip rate over the 800 ky time span since deposition of the Sabbie Gialle. The trishear model makes some basic predictions about how the Qt0 stratigraphic horizon should be deformed across the mountain front (Figure 9). Note that the maximum strain, marked by ellipses, is located at the end of the backlimb of the frontal anticline, which is the site of the observed south-dipping normal faults (Figures 6, 7, and 8).

[40] Flexural deformation of the crust in response to vertically applied loads, such as sediment loading in the Po Plain [*Scardia et al.*, 2006] represents an alternative mechanism with the potential for the long-wavelength deformation of the Qt0-IMO horizon (Figure 9c). We model the effect of vertical loads in a simple, 2D infinite elastic plate model following the general procedures described by *Turcotte and Schubert* [2002] using coarse finite element cells where the loads are constrained by the known thickness and densities of sediment of eroded rock (Table 3). The reference horizon in the modeling is "regional" defined here as 50 m above sea level, the elevation of the mouth of the Reno River at the mountain front. Sensitivity analysis shows that the model is particularly sensitive to the erosion rate so we applied both the minimum and maximum Reno River incision rates of 0.2 and 0.8 mm/year as the basin-wide average erosion rates. The 0.8 mm/year is almost certainly an overestimation of the basin-wide rate, but it serves to amplify the flexural effect in illustrating it as a viable mechanism. The effective elastic thickness of the flexing plate was left as the lone free parameter and a simple least squares fit analysis was employed to see which

Table 2b. Results of Trishear Modeling Solutions Considering Different Propagation-to-Slip Ratios and Fit Only to the Emergent Part of the Fold Defined By the Qt0 Terrace and Outcropping Portion of the IMO Deposits (Partial Fold Wavelength of 12 km) and Stratigraphic Separation of 800 m Across the Bologna Mountain Front

Ramp Angle (degrees)	Trishear Angle (degrees)	Depth to Ramp Tip (km) P/S[a] = 1.5	Depth to Ramp Tip (km) P/S[a] = 0.5	Total Slip (km) P/S[a] = 1.5	Total Slip (km) P/S[a] = 0.5
35	30	7	6	1	0.8
	45	3	3	0.5	0.35
	60	1	1	<0.5	0.2
45	30	10	10	1.6	1.3
	45	5	5	0.8	0.8
	60	3	2	0.5	0.35
55[b]	30	16	15	2.5	2
	45	9	7	1.5	1.1
	60	6	4	0.8	0.85

[a]*Allmendinger* [1998], using FaultFold 4.5.4.
[b]Best fit model parameters to field data.

thickness produced the best fit to the deformed Qt0-IMO horizon. Using an erosion rate of 0.2 mm/year, elastic thicknesses ranging from 5 to 30 km generates flexural profiles that poorly match both the wavelength and amplitude of the Qt0-IMO deformation. Using an erosion rate of 0.8 mm/year produced flexural bending at a wavelength similar to (5 km) or longer than (10, 15 km) observed Qt0-IMO deformation (Figure 9c), but poorly matches observed fold amplitude. In either case, there is no single model that matches observed fold wavelength or amplitude.

[41] Superimposed on the uplifting and deforming emergent part of the Apennines are high-angle normal faults that we show cutting the entire nappe and thrust package (Figure 8a). Tectonic geomorphology and mapping of late Pleistocene deposits indicate that these faults are active [*Bertotti et al.*, 1997; *Simoni et al.*, 2003; *Pazzaglia et al.*, in review; Bruno et al., submitted manuscript, 2008; Picotti et al., submitted manuscript, 2008]. The depth reached by some of these faults is poorly constrained and many of them not depicted in the cross-section probably do not continue for more than few kilometers given their short (<20 km) strike lengths; however, there are some data that argue for the depths depicted in the figure. For example, fluids leaking at the surface along normal faults are known to have been generated at 6 to 8 km depth [*Capozzi and Picotti*, 2002, 2006]. The cumulated vertical throw depicted is 3.5 ± .5 km

Table 3. Parameters Used for Flexural Modeling Shown in Figure 9c

Cell Type	Cell Number	Cell ρ (kg/m^3)/Cell Porosity (%)	Erosion (mm/yr)/ Duration (my)	Cell Area (m^2)	Model Cell Location (m)
Boundary load	1	1612/38	. . .	3,640,000	566,000
Boundary load	2	1612/38	. . .	3,640,000	562,000
Boundary load	3	1612/38	. . .	3,640,000	558,000
Boundary load	4	1612/38	. . .	3,640,000	554,000
Boundary load	5	1612/38	. . .	3,640,000	550,000
Boundary load	6	1612/38	. . .	3,640,000	546,000
Load	1	1612/38	. . .	3,640,000	542,000
Load	2	1612/38	. . .	3,700,000	538,000
Load	3	1612/38	. . .	3,600,000	534,000
Load	4	1612/38	. . .	3,500,000	530,000
Load	5	1612/38	. . .	3,700,000	526,000
Load	6	1612/38	. . .	4,400,000	522,000
Load	7	1612/38	. . .	4,800,000	518,000
Load	8	1612/38	. . .	4,900,000	514,000
Load	9	1612/38	. . .	5,200,000	510,000
Load	11	1612/38	. . .	2,100,000	502,000
Load	12	1612/38	. . .	1,500,000	498,000
Erosion	1	2600/0	0.8/1	3,200,000	494,000
Erosion	2	2600/0	0.8/1	3,200,000	490,000
Erosion	3	2600/0	0.8/1	3,200,000	486,000
Erosion	4	2600/0	0.8/1	3,200,000	482,000
Erosion	5	2600/0	0.8/1	3,200,000	478,000
Boundary erosion	1	2600/0	0.8/1	3,200,000	474,000
Boundary erosion	2	2600/0	0.8/1	3,200,000	470,000
Boundary erosion	3	2600/0	0.8/1	3,200,000	466,000
Boundary erosion	4	2600/0	0.8/1	3,200,000	462,000
Elastic thicknesses (km)	D (N m)	α (m)			
5	7.8E + 20	17,476			
10	6.2E + 21	29,391			
15	2.1E + 22	39,837			

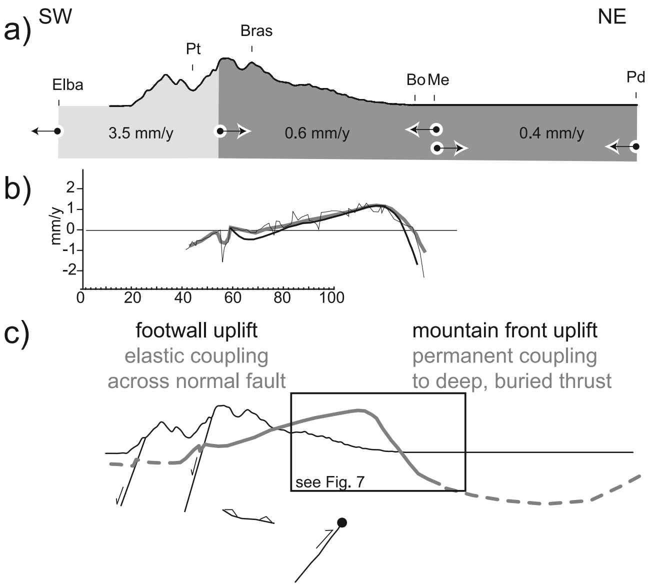

Figure 10. (a) Geodetic horizontal strain from the work of *Serpelloni et al.* [2005]. Pt, Prato; Bras, Brasimone; Bo, Bologna; Me, Medicina; Pd, Padua. (b) Geodetic vertical strain from releveling data down the axis of the Reno valley from the wok of *D'Anastasio et al.* [2006]. Thin black line is the raw data, thick black line is the smooth data, and thick gray line is the smooth data corrected for ground water withdrawal subsidence in the Po Plain. (c) Summary diagram combining short- and long-term deformation in the northern Apennines expressed in topography and geodetic data, respectively.

with the error being associated to the scale of the section and imprecision in the stratigraphic separation. The dip of the faults, again not well constrained into depth, is considered to range between 75° and 65°. The resulting total stretching associated with these faults is therefore low, ranging between 0.8 and 1.8 km. The onset of the fault activity is similarly poorly constrained, but it is likely that it youngs northward from the water divide, where could be as old as about 2 My, the same age of the onset of faulting in the Mugello graben [*Benvenuti*, 1997, Figure 4]. Close to the foothills, the faults cut IMO and younger alluvial deposits so here they are younger than 800 k.y. Therefore the minimal values for a stretching rate between the crest and the mountain front ranges from ~0.4 to 1 mm year^{-1} with most of that stretching probably focused near the range crest.

[42] We can restore the cross-section accounting for the stretching and rock uplift associated with the normal faults and the deformation focused at the mountain front which generates a picture of the Apennines in the early Pleistocene (Figure 8b). The Ligurian units are depicted as a continuous body, extending to the Apennine topographic crest that was located more or less at the modern position following activation of the Mugello border faults [*Benvenuti*, 1997]. The movement of the Ligurian nappe toward the foreland decelerated and stopped at the end of the Pliocene after rapid advancement during Messinian to Early Pliocene [*Zattin et al.*, 2002]. The thrust belt beneath the Po Plain is interpreted as the evolution of a thick basement ramp throughout the Messinian and Pliocene, propagating a fold- and thrust belt toward the foreland, with the in-sequence activation of structures. The main detachments were located at the Upper Cretaceous to Paleocene (Fucoidi Marls and Scaglia units) and bottom Upper Triassic intervals, respectively. The thrust top of the Ferrara structures had emerged as islands within the paleo- Adriatic sea. The two most external structures are rooted in the basement, and possibly associated to inversion of older Tethyan margin faults. A similar interpretation was proposed by *Coward et al.* [1999], *Mazzoli et al.* [2001], and *Butler et al.* [2004] for the most external structures in the Northern Marche. As a whole, these structures account for the telescoping of the wedge-shaped foredeep basin and its sedimentary basement, with a divergent pattern over the growing structures [*Pieri*, 1987]. The shortening along the studied transect, representing the Messinian to Pleistocene deformation, in our interpretation is ~27 km, with the maximum rate of shortening occurring during the Messinian and Lower Pliocene, and a deceleration into the Pleistocene. Our calculated mean long-term rate of shortening is ~4.5 mm year^{-1}, comparable with that of *Bartole* [1995], *Cavinato and DeCelles* [1999], and *Basili and Barba* [2007] rates.

5. Geodetic Strain, GPS, and Releveling

[43] Recent geodetic data from GPS surveys, releveling, and DInSAR model the contemporary deformation field of the northern Apennines (Figure 10). Vertical surface velocities have been obtained from a releveling line spanning 50 years that follows the railway line in the Reno River valley, corrected for anthropogenic ground water withdrawal in the Po Plain [*D'Anastasio et al.*, 2006, Figure 10b]. This railway is oriented on top of and parallel to the river terraces, so the releveling line is favorably oriented with respect to the terrace correlation and incision shown in Figure 7. A direct comparison of the GPS and releveling data shows subsidence coinciding with the zone of crustal stretching and uplift coincident with the mountain front, which is located in the zone of crustal shortening. A more recent study using DInSAR data, also considering for ground water withdrawal in the Po Plain similarly argues for surface uplift rates of the Bologna mountain front of ~1 mm/year [*Stramondo et al.*, 2007].

[44] In detail, the releveling line argues that the entire crest of the range, even that portion in the zone of crustal shortening, is presently subsiding, whereas the region where topography is the lowest, that is the mountain front, has the highest rate of surface uplift. Subsidence at and around the range crest is most easily explained as the near-field effects of a locked normal fault. In contrast, uplift at the mountain front is most easily explained as regional, permanent deformation associated with anticlinal growth cored by the modeled, deep reverse fault. The amplitude of the warp in the releveling line is similar to the inferred amplitude of the growing anticline and the surface uplift rate of 1.3 mm/year compares well to the DInSAR data and late Pleistocene rate of Reno river incision.

6. Discussion

6.1. Origin of the Bologna Mountain Front

[45] Deformed river terraces and Sabbie Gialle stratigraphic horizon, the high-resolution reflection line, geodetic

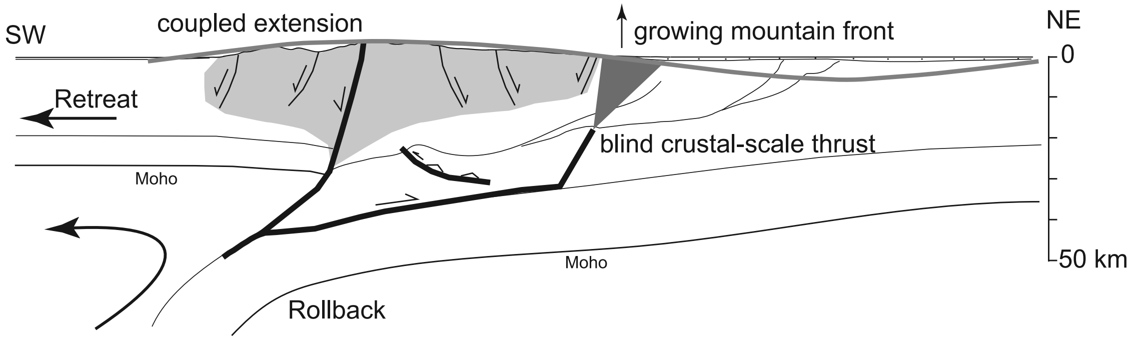

Figure 11. Active deformation of the northern Apennines in the context of a partially coupled rollback-retreat geodynamic model.

data, and trishear modeling collectively indicate that the Bologna mountain front is a broad, anticline growing vertically at about ~1 mm/year, associated with an out-of-sequence deep reverse fault, in a slowly, but actively shortening part of the Apennine wedge (Figure 11). The inferred location of the reverse fault is not imaged on seismic profiles, but it does correspond to the general location of deep compressive seismicity [*Chiarabba et al.*, 2005; *Pondrelli et al.*, 2006; R. Di Stefano and J. Park, personal communication, 2007]. This feature should not be viewed as the reactivation of a shallow thrust-cored anticline near or beneath the mountain front on Figure 8, but rather a deeper structure, embedded in the crust that is uplifting and deforming the entire previously shortened section (red faults in Figure 8a). The architecture of the mountain front therefore is similar to the conceptual models of a deep blind thrust or a triangle zone anticlinal stack involving shortening of the entire crust [*Vann et al.*, 1986].

[46] A crustal-scale reverse fault as part of a larger duplex of thickened crust is appealing because the uplift that drives it has an amplitude of several tens of kilometers, and it has the additional effect of stretching the carapace, consistent with the recent formation of active normal faults [*Morley*, 2007]. We recognize that active normal faults described in our study, and those of related studies [*Bertotti et al.*, 1997] locally shape the topographic details of the mountain front, but are not ultimately the main reason for its origin. In this respect, the northern Apennine mountain front is similar to the Brooks Range example discussed in *Vann et al.* [1986]. The Brooks Range is one of several Laramide-style ranges that were deformed in the American West during the late Mesozoic and early Cenozoic. The similarity in the scale of deformation, the foreland tilting synorogenic strata, the rate of shortening, the small total amount of shortening, and the formation of normal faults in the carapace typical of many Laramide Ranges such as the Bighorns and Owl Creek ranges in Wyoming [*Wise*, 1963; *Blackstone*, 1988] are striking. Such crustal- or whole-lithospheric-scale faults, as described for the Laramide Rockies [e.g., *Magnani et al.*, 2005] or as proposed for the northern Apennines, have been argued in the past [*Lavecchia et al.*, 2003a], but not as well supported by the synthesis of geomorphic, geologic, geodetic, and structural modeling data we present.

[47] Continued subduction of Adria beneath the Apennines and eastern Alps has been proposed to result in shortening accommodated by crustal-scale faults [*Lavecchia et al.*, 2004]. The shorter-wavelength, thrust fault cored anticlines in Figure 8 are modeled to represent smaller, second-order folds that form as a result of the mechanical properties of a weak horizon in the sedimentary package [*Massoli et al.*, 2006].

[48] The important geologic and geomorphic observation that middle and late Pleistocene littoral and alluvial deposits that originally deposited nearly horizontal, have now been deformed such that the hinterland portion has been uplifted above regional, and the foreland portion has subsided below regional is clearly evident for the Bologna mountain front. If the deformation is caused by a deep, crustal-scale reverse fault, we would predict that the fault should cut down section and sole ultimately into the master decollement that defines the base of the Apennine wedge, in effect, the subduction interface (Figure 1d). This model predicts that evidence for similar deformation of Pleistocene deposits should mark the position of the mountain front elsewhere along strike of the range. South and east of Bologna, the Sabbie Gialle, its equivalent littoral units, and younger alluvial deposits are everywhere found to have been lifted up above regional, an observation in line with our prediction. Similar triangular headed flatiron landforms capped by the Sabbie Gialle or equivalent deposits at locations as widely separate as Bologna, east along the Po Plain, and further south along the Adriatic coast, argue for a similar Quaternary deformation style along the strike length of the range.

6.2. Quaternary Tectonics of the Northern Apennines and the Rollback-Retreat Model

[49] An actively growing mountain front provides some first-order insights into the geodynamics of the rollback-retreat model. First and foremost, crustal shortening continues in the northern Apennines, which means that Adria is continuing to subduct. Currently, this shortening is occurring mainly at depth, whereas frontal accretion has stopped [*Pieri*, 1987 and Figure 8; *Di Bucci and Mazzoli*, 2002; *Di Bucci et al.*, 2003]. As a result, we can quickly dispense with the fully coupled model of Figure 2b. The reasons for why shortening has moved back from the late Pliocene – early Pleistocene thrust front beneath the Ferrara arc to being building the current mountain front are not clear, but we know it is coincident with the transformation of the foreland from wedge-shape to sag-shaped and suspect that both are linked to unsteadiness in the rollback process.

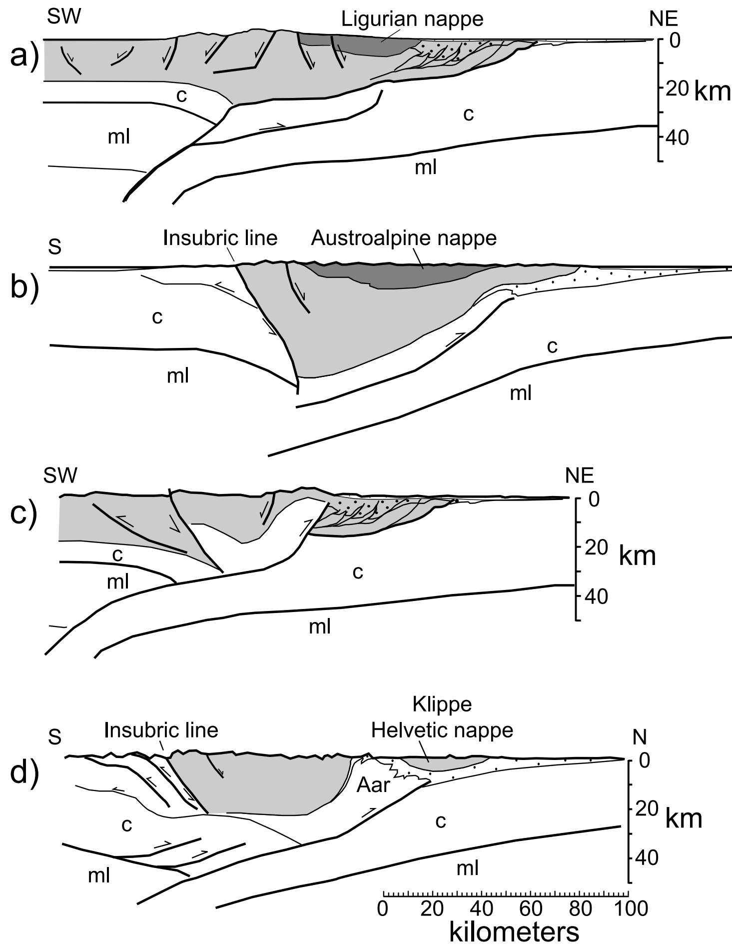

Figure 12. A comparison of the major tectonic features of the northern Apennines to the central Alps. (a) Northern Apennines at present; (b) central Alps 19 Ma; (c) northern Apennines 20 My into the future, with continued movement of the mountain front blind thrust fault; (d) central Alps at present, with the Aar Massif splitting the upper crustal orogenic wedge following shortening on a crustal-scale blind thrust. c, crust; ml, mantle lithosphere. Dark lines indicate active faults. Dots indicate proforeland deposits.

[50] A fully decoupled rollback-retreat model with a Platt-style crustal wedge (Figure 2a) would predict that active shortening at the mountain front, and the presence of normal faults particularly in the rear of the wedge would be the result of whole wedge adjustments to maintain critical taper. Burial of the thrust front with Po Plain alluvium would have the effect of reducing taper on the thrust front and the hinterland stepping of shortening to increase wedge taper, but this mechanism fails to account for the foreland subsidence that created the accommodation space for the sediment in the first place. Most of the middle-late Pleistocene sediment filling the Po foreland is derived from the Alps. Isostatically-corrected loading of the foreland from the backstripped sedimentary section shows that increased erosion and sediment yield from the Alps alone could not have driven the new accommodation space [*Scardia et al.*, 2006]. Again, the normal faults documented at the mountain front and foreland-portion of the wedge are too small, too few, and too high angle to accommodate very much crustal thinning. The same is true for the larger normal faults on the back side of the wedge in Toscana. Collectively, the focused deformation growing the mountain front, the recent transition to a sag-type, rather than wedge-type foreland, and insignificant crustal stretching leads us to reject a fully decoupled rollback-retreat model.

[51] A partially coupled rollback-retreat model (Figure 2b) predicts that subduction and rollback of Adria generates sub-slab mantle flow that drives retreat of the upper slab and a wave of extension that cuts into and dismantles the crustal wedge built by subduction-driven shortening. Preliminary results from the RETREAT seismic experiments that document a complicated sub-slab mantle flow field [*Plomerova et al.*, 2006] and our observations for long-wavelength foreland subsidence and orogen uplift are consistent with dynamic mantle flow in a partially-coupled model.

[52] The trishear model of a deep, steep crustal ramp remains as the best explanation for the mountain front tilt, but still cannot account for both longer-wavelength subsidence in the Po Plain and uplift in the Apennine range, which are best explained by other processes. We step back from our detailed discussion of deformation at the mountain front to point out that it is embedded in a longer-wavelength deformation that likely affects the entire lithosphere and the mantle flow in the coupled model is a potential process driving the deformation (Figure 11). Lithosphere buckling has been proposed as an important intraplate deformation

[*Cloetingh et al.*, 1999] and has been described in its interaction with the foredeeps on the southern Adriatic plate [*Bertotti et al.*, 2001]. We explored the kinematics of the buckling, in terms of linear variations of a bent elastic plate (i.e., without any internal deformation), aimed to assess the amount of stretching and shortening in the uplifting and subsiding part of the system, respectively. We found that only meters of horizontal shortening at a rate of 0.004 mm/year would be necessary to account for the Middle Pleistocene to Holocene sagging of the Po Plain. Similarly, upward bending of the Apennines south of the mountain front to account for 4 km of uplift in the crest in the last 2 Myr [*Balestrieri et al.*, 2003] predicts ∼170 to 200 m of stretching with an average stretching rate of ∼0.1 mm/year. These values do not agree well with the GPS data, the 2.5 mm/year of shortening predicted for the blind reverse fault proposed in our analysis, or the ∼1 to 1.8 km of stretching due to normal faulting, so we reject a pure horizontal buckling process. There may remain a component of horizontal buckling in the northern Apennines in that it predicts failure in the crust at the inflection point [*Cloetingh et al.*, 1999; *Sokoutis et al.*, 2005] which for our case is the mountain front, in good agreement with the observed location of the deep active structure.

[53] Similarly, we reject the effects of shallow vertical loads on a pure elastic plate as a unique solution for explaining the warping of Pleistocene strata at the mountain front, but recognize that it may contribute to the long-wavelength lithospheric deformation (Figure 9c, Table 3). There remain other, related dynamic mantle processes and vertical loads that may contribute to the long-wavelength lithospheric deformation. These include upwelling of warm asthenosphere beneath the thinning lithosphere of the retreating plate, a process similar to that proposed by *D'Agostino et al.* [2001] and the rise and spreading of warm asthenosphere related to the head of a hotspot plume [*Lavecchia et al.*, 2003b], both of which may be related to the modeled dynamic topography described by *Shaw and Pysklywec* [2007]. Whatever the cause of dynamic support of Apennine topography, we argue that it does contribute to the overall observed deformation and that deformation is more consistent with a partially coupled rollback-retreat model, than a fully decoupled one, where crustal deformation is restricted to Platt-type critical tapered wedge processes only.

6.3. Seismotectonic Consequences of the Described Tectonic Model

[54] The above described tectonic model bears some consequences on the seismotectonics of the region around Bologna. With respect to the models and maps available in literature [e.g., *Lavecchia et al.*, 1994; *Boccaletti et al.*, 2004], the main difference is the depth of the compressional structure that has been established at 15 to 20 km under the mountain front rather than at the surface. This model appears consistent with the hypocenters of the compressional earthquakes as described by *Pondrelli et al.* [2002, 2006]. The focusing of the deformation documented by the increasing strain throughout the Pleistocene along the mountain front allows interpreting this structure as an important potential seismogenic source. The distribution of the compressional earthquakes within the middle crust (15 to 25 km) both internally [e.g., *Piccinini et al.*, 2006] and under the Po Plain [*Pondrelli et al.*, 2006] suggests, furthermore, a diffuse compressional state of stress, deeper than the inactive thrust wedge, possibly acting through reactivation of basement structures. Furthermore, the evidence of activity in the upper crust of the uplifting belt coherently indicate that the extensional stresses stored in the upper crust are released by a dense network of faults, few of which display a significant strain. This observation should match the absence of important extensional earthquakes in the Po Plain side of the northern Apennines [*Chiarabba et al.*, 2005; *Pondrelli et al.*, 2002, 2006]. The subsidence recorded by the relevelling toward the crest (Figure 10), however, suggests that some major SW dipping normal faults of the Tyrrhenian side of the Apennines, are in a phase of interseismic locking, an observation that bears important consequences for the seismicity of the crestal range of the belt, whose topography results from the coseismic uplift of the footwall blocks.

6.4. Comparison of Apennine and Alpine Orogen Evolution

[55] Formation and growth of the Bologna mountain front as a surficial response to the extinction of shallow thrust faults, and nucleation of shortening on a deep crustal ramp is a process bearing analogies with the widely accepted evolution of the central Alps [e.g., *Pfiffner and Heitzmann*, 1997; *Schmid et al.*, 1997]. The modern northern Apennines bear many similarities in gross cross-section to the reconstruction of the central Alps 20 Ma (Figure 12) [*Schmid et al.*, 1997]. These similarities include the end of rapid propagation of the prowedge into the foreland, and the initiation of a major crustal ramp in the prowedge hinterland. There are also notable differences including that the Alpine accretionary wedge was much formerly thicker than the modern Apennine wedge which is wider, almost certainly the effect of crustal thinning in the Apennines because of a retreating upper plate and none in the Alps which remained a dually-vergent orogen.

[56] If the nascent crustal ramp beneath Bologna continues to develop at the modern slip rate of ∼2.5 mm/year, in 20 m.y. middle and lower crustal rocks will have been transported to the surface, splitting the Apennines prowedge and producing an Aar-type massif in a sea of folded and deformed nappes. A possible analogue of this process, but at a smaller scale was the emplacement of the Alpi-Apuane core [*Carmignani and Kligfield*, 1990; *Molli and Tribuzio*, 2004] in the middle Serravallian and Tortonian (∼13–8 Ma) as indicated by rapid cooling of those rocks at that time [*Fellin et al.*, 2007]. We are intrigued by the temporal coincidence in the formation of the Alpi Apuane and the pre-modern sag-type Po foreland, a coincidence that suggests the two processes are related. Perhaps shortening in the Apennines alternates between relatively thin frontal accretion and foreland propagation of the trust front or thick underplating and duplexing of the entire crust, an interpretation similar to that proposed by *Fellin et al.* [2005]. The former corresponds to times of wedge-type forelands and extensional dismemberment of Alpi-Apuane or Aar-type massifs as the upper plate retreats. The latter corresponds

to times of sag-type forelands and emplacement of Alpi-Apuane or Aar-type massifs which temporarily outpace the extensional thinning in the retreating plate. In fact, it is the retreat of the upper plate that remains the largest difference in our comparison between the Alps and Apennines, suggesting that, while similar in gross cross-section, the Apennines orogenic evolution will ultimately diverge from what is observed in the modern Alps.

7. Conclusions

[57] We summarize existing and new geologic data, particularly on the origin, growth, and active deformation of the northern Apennines mountain front, a prominent feature which developed in the middle to late Quaternary in the Apennine pro-wedge, with the aim to test and better understand the slab rollback – upper plate retreat process as a more general process in Mediterranean orogenesis (NSF Continental Dynamics RETREAT project). The Quaternary and modern tectonics of the northern Apennines are used to develop an active tectonic model consistent with diverse geologic, geomorphologic, geophysical, and geodynamic setting of the Bologna mountain front. The main finding of our investigation is that the Bologna mountain front is an actively growing structure, cored by a mid-crustal flat-ramp structure that accommodates ongoing shortening driven by Adria subduction. The deep blind fault has a slip rate of ~2.5 mm/year and is driving surface uplift of the mountain front at ~1 mm/year. The specific style of shortening is consistent with the general uplift of the entire Apennine wedge in the Pleistocene and the transition from frontal accretion and foreland propagation of the fold and thrust belt in the Tertiary to underplating of increasingly normal continental Adriatic lithosphere in the Quaternary.

[58] We document that the Po Plain around Bologna is a subsiding sag basin, superposed on top of the former proforeland basin, where shallow thrust-cored folds appear to be mostly inactive since the middle Pleistocene. This situation is not universally true for the Apennines as in the Emilia arc to the west there is local evidence of reactivation of upper crustal thrusts that have been reported close to the mountain front [e.g., *Picotti et al.*, 2007]. Long-wavelength Po subsidence (80–100 km) has a symmetrical half in what appears to be long-wavelength hinterland uplift (Figure 11). Superimposed on this long-wavelength deformation at Bologna is shortening focused at and building the mountain front. This shortening is accommodated by a reverse fault, located at the middle crust (15–20 km depth) that is creating a fold (half wavelength ~22 km, emergent half wavelength of ~12 km) that has been documented and modeled using the geometry of deformed middle to upper Quaternary deposits including river terraces. Mid-crustal shortening is well supported by a growing database of active seismicity [*Pondrelli et al.*, 2002, 2006; *Chiarabba et al.*, 2005; *Piccinini et al.*, 2006]. Active normal faults, clearly evident offsetting young river terraces, strike across the deforming mountain front (Picotti et al., submitted manuscript, 2008). We interpret these structures as the effect of a diffuse extensional state of stress acting on the upper crust of the uplifting belt, due primarily to the inflation and bending associated with the thrust ramp [*Morley*, 2007]. The high-angle nature and small overall throw on these normal faults at the mountain front argues that they do not accommodate significant crustal thinning.

[59] The middle crustal structure shares strong similarities with the analogue structure of the Alps, where the external massifs thrust system, cutting the lower plate middle crust under the deformed wedge, developed since the Early Miocene. We propose a possible relationship between underplating and the development of a sag basin as a Quaternary phenomena that may also apply to past periods of Apennine deformation, such as the Tortonian when the Po foreland was deep, wide, and symmetrical and there was rapid erosional unroofing of mid-crustal rocks in the Alpi Apuane [*Fellin et al.*, 2007].

[60] As a whole, this geologic model for the active tectonics of the mountain front of the northern Apennines documents upper crustal extension and middle crustal compression as overlapping processes, both in space and time [cf. *Picotti et al.*, 1997; *Lavecchia et al.*, 2003a]. These tectonic features are associated to a change from frontal accretion to underplating that produced the deepening and the backstepping of the compressional deformation, in a context of ongoing subduction and effective retreat of the upper plate. We favor a geodynamic model where crustal extension in the Apennines is partially coupled to the retreating upper plate (Figure 2b). Normal faults at the mountain front speak to a fully decoupled foreland where Platt-style crustal adjustments to underplating are dominant. In contrast, large normal faults in the hinterland speak to an increased coupling of the range there to a middle crust decollement that we envisioned tied to the highly stretched Tyrrhenian lithosphere further west.

[61] The deformational history of the middle crustal fault responsible for the growth of the Bologna mountain front is increasing strain through the late Pleistocene which suggests that this structure is focusing most of the subduction component and is by far the most important seismogenic source. The distribution of the compressional seismicity document that, beside this structure, other middle crustal sources are patchily distributed both internally, within the underplating volume, and in the Po Plain, possibly associated to reactivation of basement structures. Finally, the number of the normal fault active at the upper crust of the uplifting belt suggests a diffusion of the strain that possibly inhibits the development of important extensional seismogenic sources in this sector of the chain.

[62] **Acknowledgments.** This paper has been worked out during a visiting program of V.P. at the Earth and Environmental Department of the Lehigh University, whose staff is warmly thanked for the hospitality. A "Marco Polo" grant of the University of Bologna favored the V.P. mobility. Numerous discussions and interactions with RETREAT group leaders Mark Brandon and Sean Willett helped us cast our geologic results and trishear approach in the context of the broader, still-evolving geodynamic models being tested by the RETREAT group. Discussions with RETREAT group members J. Park, D. Cowan, R. Bennett, P. Reiners, D. Granger, M. C. Eppes, S. Pondrelli, N. D'Agostino, P. Severi, and E. Serpelloni and with G. Bertotti, P. P. Bruno, R. Capozzi, K. Wegmann, and J. Cascione are kindly acknowledged. The Geological Survey of the Regione Emilia Romagna kindly provided published and unpublished data. The Zola Predosa Municipality is acknowledged for the logistic support during seismic acquisition. Special thanks to Claudio Berti for independent modeling and confirmation of our trishear approach. The journal reviewers, S. Mazzoli and anonymous, and the associate editor, G. Rosenbaum, are thanked for their constructive criticism and many helpful suggestions. This

work has been funded by GEO-EAR-0207980 (Project RETREAT) and RFO, University of Bologna.

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