Extensional basins in the tectonically bimodal central Apennines fold-thrust belt, Italy: Response to corner flow above a subducting slab in retrograde motion

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ABSTRACT

The Apennines fold-thrust belt has developed over the past ~20 m.y. in response to southeastward retrograde migration of the Adriatic trench in a region of ongoing subduction but little plate convergence. The orogenic belt is characterized by a gently sloping southwestern flank in a state of regional extensional collapse, and a steeply sloping northeastern flank in regional contraction. In any given area in the central Apennines, the chronologies of thrusting and subsequent extension provide an estimate of the time required for attainment of high elevation (2.5–3.0 km). The time between initial extension and eruption of local ultramafic lavas provides an estimate of the time required to collapse by regional crustal thinning. These estimates suggest that the northeastern, contractional flank of the range is rising at least twice as fast as the southwestern flank is collapsing. We propose that the bimodal state of stress in the Apennines is maintained by corner flow in the mantle wedge beneath the crest of the range.

INTRODUCTION

Many of Earth's contractional mountain belts exhibit regional late-stage extension (Constenius, 1996) and/or extension contemporaneous with regional contraction (Burchfiel et al., 1992). The causes of extension in fold-thrust belts are generally attributed to overthickening of the orogenic wedge (Burchfiel et al., 1992) or changes in plate motion directions and descent angles of subducting slabs (Coney and Harms, 1984). These mechanisms operate in collisional and retroarc orogenic belts that involve converging plates (e.g., the Himalayas and Andes; Royden, 1993). The Apennines, however, are in a separate class of mountain belts because, although they have been constructed by folding and thrusting, they have developed in the absence of major plate convergence, as the subduction of Tethyan lithosphere has continued while Africa-Eurasia convergence has decreased (Dewey et al., 1989; Doglioni, 1991). One of the results of this process of mountain building is that the Apennines are collapsing nearly as rapidly as they are rising. The northeastern (Adriatic) flank of the range is an active fold-thrust belt (Fig. 1), and the southwestern (Tyrrenian) flank is dominated by Pliocene-Quaternary extension (Royden et al., 1987; Patacca et al., 1992; Lavecchia et al., 1994). Ongoing extension in the Apennines cannot be ascribed to gravitational collapse of overthickened crust because the crust is not overly thick (25–30 km; Scarascia et al., 1994). Neither can changes in plate vectors be called upon, because no such changes have occurred recently (Dewey et al., 1989; Mazzoli and Helman, 1994). What is the process by which regional extension and contraction are simultaneously accommodated in the Apennines? Our approach is to synthesize the kinematic histories of contraction and extension as recorded by development of the Apennine foreland basin system and the numerous extensional basins that are superimposed on the range. We then consider the larger question of how this history has developed and been maintained.

TECTONIC AND STRUCTURAL SETTING

The Apennines are a northeast-verging imbricate fold-thrust belt that has developed since early Miocene time along the margin of the Adriatic microplate in response to east-southeastward retreat of the generally northwestward dipping subduction zone between the southern margin of central Europe and a remnant of probable Neotethyan oceanic lithosphere (Fig. 1; Malinverno and Ryan, 1986; Royden et al., 1987; Doglioni et al., 1998). In the process, several blocks of continental crust (Sardinia, Corsica, and Calabria) have rifted from the European margin and the Ligurian and Tyrrhenian extensional oceanic basins have opened. Geophysical (Malinverno and Ryan, 1986) and plate motion (Dewey et al., 1989) studies indicate that southeastward migration of the subduction zone has been driven mainly by the weight of the downgoing slab (as discussed by Malinverno and Ryan, 1986; Royden et al., 1987).

Recent deep seismic profiling across the northern Italian peninsula shows that crustal thickness increases from ~22–24 km beneath the western part of the peninsula to 35–40 km beneath the axis of the Apennines (Barchi et al., 1998). Deep earthquakes and seismic tomography image a subducting slab of Adriatic lithosphere beneath the axis of the fold-thrust belt, and shear-wave splitting analysis suggests pronounced northeast-southwest anisotropy in the mantle wedge above

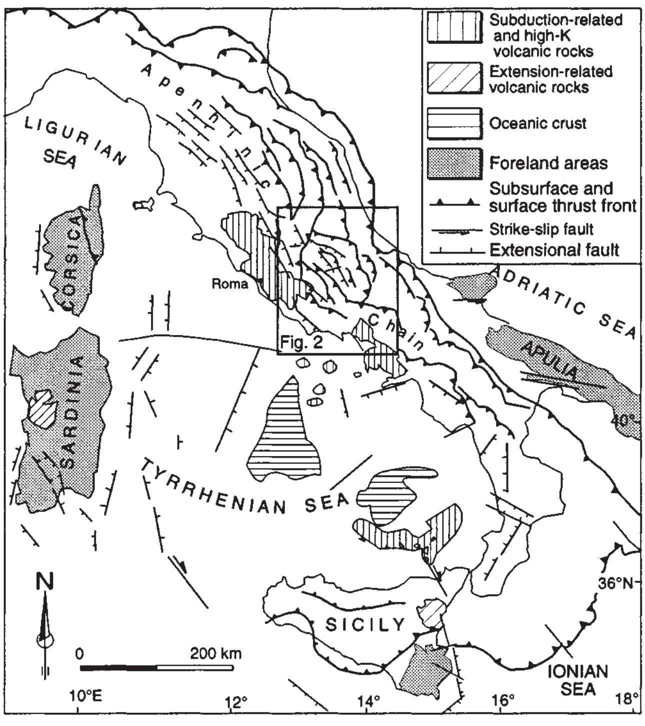

Figure 1. Generalized tectonic map of central Mediterranean region. Box in center outlines area shown in Figure 2.

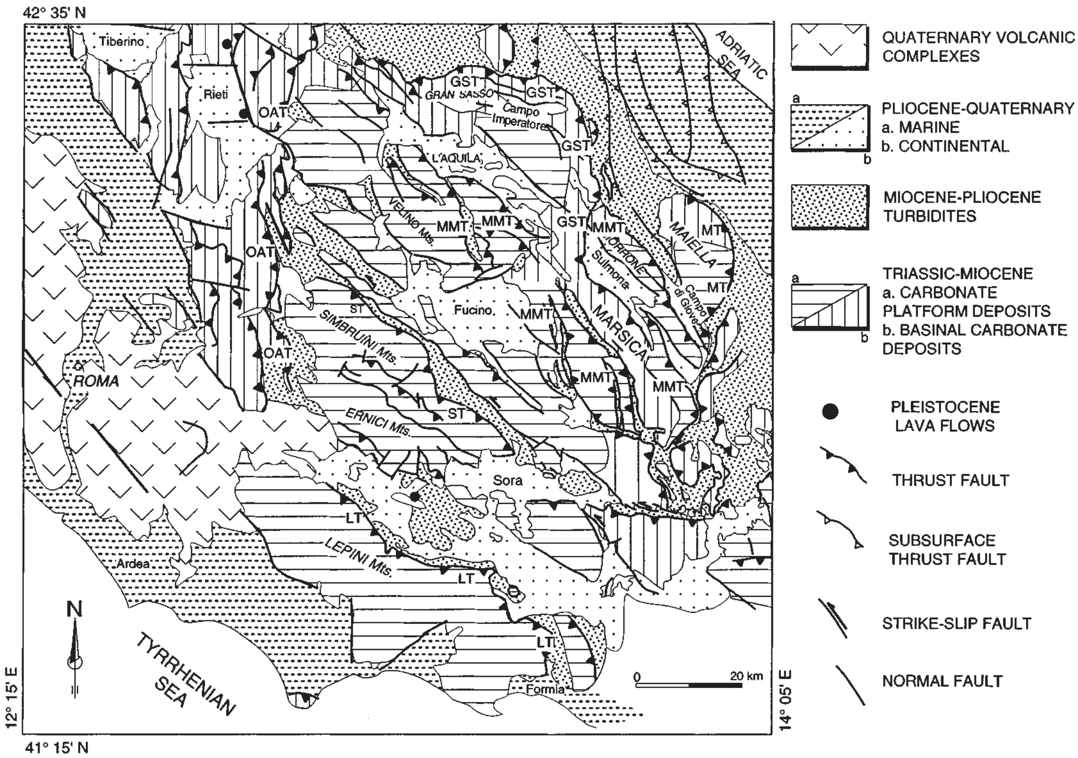

Figure 2. Generalized geologic and tectonic map of central Apennines. Thrust systems referred to in text: MT, Maiella thrust; GST, Gran Sasso thrust; MMT, Marsica-Morrone thrust; ST, Simbruini thrust; OAT, Olevono-Antrodoco thrust; LT, Lepini thrust.

and west of the slab (Scarascia et al., 1994; Amato, 1998; Ciaccio et al., 1998; Margheriti et al., 1998; Mariucci et al., 1999). The transition from a shallow to subvertically dipping slab takes place below the transition from active compressional to dominantly extensional strain in the overlying fold-thrust belt.

Major thrust systems in the central Apennines strike northwest-southeast and dip gently southwestward. These include the Lepini, Simbruini, Marsica, Morrone, and Maiella thrust systems; several more thrusts and related folds are concealed beneath several kilometers of Pliocene-Quaternary synorogenic sedimentary rocks to the northeast of the topographic front of the range (Fig. 2; Patacca et al., 1992; Cavinato et al., 1995). The Olevano-Antrodoco and Gran Sasso thrust systems are significant out-of-sequence thrusts that truncate some of the north- to northwest-striking systems (Fig. 2; Cipollari and Cosentino, 1997; Ghisetti and Vezzani, 1997). The exposed thrust sheets consist of Triassic–middle Miocene carbonate rocks (Bally et al., 1986; Cavinato et al., 1995). The blind thrusts in the northeastern part of the thrust belt carry Messinian-Quaternary clastic sediments (Patacca et al., 1992).

High-angle normal faults and extensional basins of generally post-Messinian age are widespread in the central Apennines (Keller et al., 1994; Lavecchia et al., 1994), and the present Tyrrenian margin is extensional (Fig. 1). The most important extensional basins are the Campo Imperatore, L'Aquila, Sulmona, Fucino, Rieti, Tiberino, and Ardea basins (Fig. 2). Several of these basins contain small volumes of middle Pleistocene ultramafic lavas (Laurenzi et al., 1994).

CHRONOLOGY OF THRUSTING AND EXTENSION

Miocene through Pliocene synorogenic sediments record the kinematic history of thrusting in the central Apennines. These deposits are marine turbiditic and olistostromal facies in the Burdigalian through late Messinian, and shallow marine to nonmarine in the post-late Messinian part of the record (Ricci-Lucchi, 1986; Patacca et al., 1992; Cipollari and Cosentino, 1997; Cipollari et al., 1997). The main phases of late Miocene–recent thrust-sheet emplacement migrated toward the Adriatic foreland as follows: Lepini thrust, late Tortonian; Simbruini and Olevano-Antrodoco thrusts, early Messinian; Marsica, Morrone, and Gran Sasso thrusts, late Messinian–early Pliocene; and the Maiella thrust, late-early Pliocene (Fig. 3). The Olevano-Antrodoco and Gran Sasso thrusts were reactivated during the late Messinian and late-early Pliocene, respectively (Cipollari et al., 1997). Each of the main thrust systems produced a large ramp anticline and significant topographic relief that provided successive accumulations of synorogenic wedge-top and foredeep sediments.

The extensional history of the central Apennines is directly correlated with opening of the Tyrrhenian basin since late Tortonian time. Lavecchia et al. (1994) showed that the timing of extensional deformation migrated eastward ~2 m.y. after and 75–100 km behind the compressional front (Fig. 3). In the central Apennines regional extension has produced a network of high-angle, generally west dipping normal faults and extensional basins across the southwestern slope of the range. Many of the normal faults join and reactivate thrust fault ramps in the subsurface, perching the previously elevated frontal parts of thrust sheets at high elevations and dropping their trailing parts into the subsurface. The extensional basins include the Campo Imperatore, Campo di Giove, L'Aquila, Sulmona, Leonessa, Subequo, Campo Felice, Rieti, Fucino, Tiberino, Sora, Formia, and Ardea basins (Fig. 1).

Central Apennines extensional basins can be classified into three types in an evolutionary continuum on the basis of age, size, morphology, elevation, and internal lithofacies. Typical of the youthful stage of extensional basin evolution are the Campo Imperatore and Campo Felice basins (Fig. 2). These basins are relatively high (>1.3 km), small (10–30 km^2), and shallow (0.2–0.3 km). They are filled with coarse-grained alluvial fan, colluvial, glaciogenic, and intermittent lacustrine and braided-stream facies of middle Pleistocene to recent age. The youthful stage basins lack large source-area drainage basins and through-flowing river systems. Topographic relief along their flanks ranges from ~0.5 to 1.0 km.

Intermediate-stage extensional basins in the central Apennines are represented by the Leonessa, Sulmona, L'Aquila, Subequo, and Tirino basins (Fig. 2). These basins occupy elevations ranging from ~0.4 to ~0.8 km, have areas of 30–120 km^2, and are filled by lower-upper Pleistocene alluvial, fluvial, and lacustrine deposits. Moderate-sized rivers flow through the L'Aquila, Subequo, and Sulmona basins, depositing mixed bedload and suspended-load sediment. Shallow lacustrine and palustrine deposits are also present. Modern alluvial and colluvial fans exist along the flanks of these basins, and topographic relief along basin margins is 1–2 km. The main boundary faults are high-angle normal faults that strike northwest-southeast and north-south and are locally cut by east-west transfer faults.

The mature stage of extensional basin development is represented by the Tiberino, Fucino, Rieti, and Ardea basins (Fig. 2). These basins are at elevations of 0–0.7 km, exceed 250 km^2 in area, have well-integrated catchment areas, and are filled by 0.5–2.0 km of lower Pliocene to recent, alluvial, fluvial, shallow-marine, and volcanogenic deposits. These are the lowest elevation extensional basins in the Apennines, and represent the transition from fully nonmarine to shallow-marine (Tyrrhenian) settings. In addition to their larger sizes and partially marine fills, the mature basins contain late Pleistocene ultramafic volcanic rocks that were erupted along some of the major boundary faults.

The general evolutionary pattern of the central Apennines extensional basins includes gradual enlargement, deepening, and a concomitant change from alluvial to fluvial to lacustrine to shallow-marine environments through time. We suggest that basin evolution was strongly controlled by decreasing elevation and ongoing extensional subsidence.

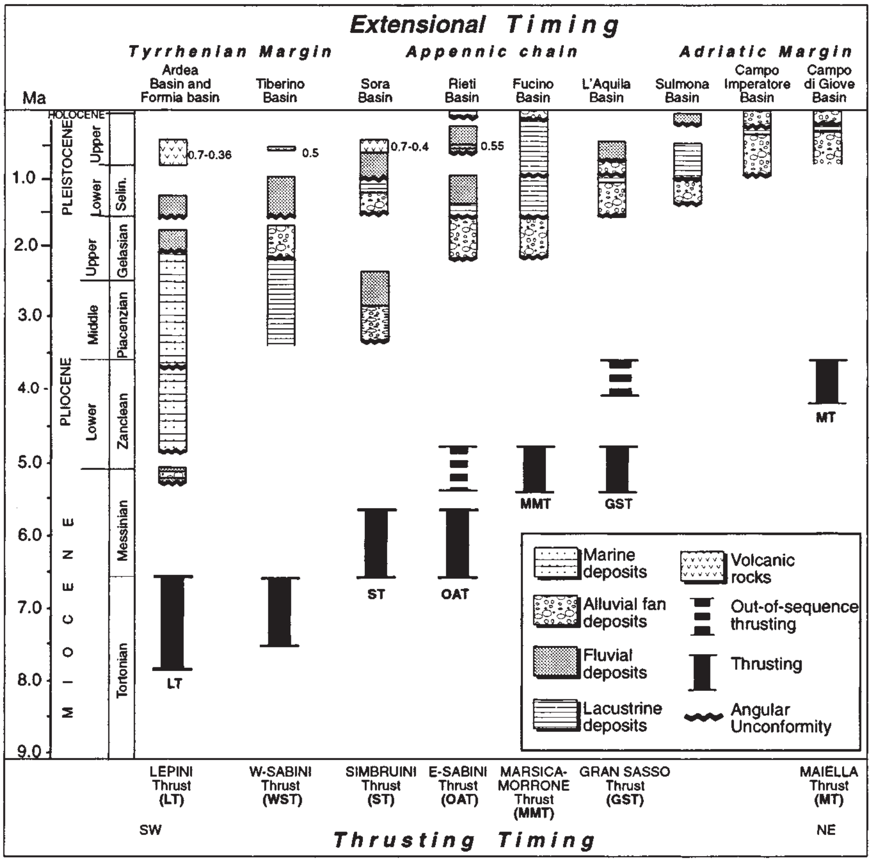

Figure 3. Time-space migration of thrusting and subsequent development of extensional basins across central Apennines, after Patacca et al. (1992), Cipollari and Cosentino (1997), and Cipollari et al. (1997).

DISCUSSION

Modeling and geophysical studies indicate that the mantle wedge above a subducting slab should contain a corner-flow cell in which mantle material is viscously coupled with the subducting slab (Forsyth and Uyeda, 1975; Turcotte and Schubert, 1982; Royden, 1993; Fig. 4). In the upper limb of the corner-flow cell, material is pulled toward the trench, setting up a significant trenchward stress at the base of the overlying lithosphere (Chapple and Tullis, 1977). This stress maintains a local contractional environment in the frontal part of the overlying fold-thrust belt or accretionary prism. Where the rate of slab subduction (i.e., the rate of underthrusting) exceeds the rate of convergence of the two plates involved, the trench will migrate toward the subducting plate and the interior of the overriding plate will thin and extend as relatively hot mantle wells upward to replace material that is dragged trenchward (Royden, 1993; Waschbusch and Beaumont, 1996). Thus, in zones of subduction-zone retreat, such as the Apennines, it is expected that the front of the orogenic belt will be contractional (due to trenchward flow in the mantle wedge) while its trailing part will be in a state of regional extensional collapse (due to crustal thinning). The zone of transition between contractional and extensional strain might mark the lateral limit of significant trenchward flow in the upper part of the mantle wedge (Fig. 4). In the Apennines, the southwestern boundary of the corner-flow cell is predicted to be approximately beneath the topographic crest of the range, as confirmed by deep earthquakes and seismic tomography (Ciaccio et al., 1998). The northeast-southwest anisotropy in the mantle wedge beneath the Apennines is also consistent with northeastward flow in the upper limb of the corner-flow cell (Margheriti et al., 1998).

If the 2–4 m.y. lag time between the onset of thrusting and initial extension at any given locality in the central Apennines (Fig. 3) represents the amount of time required for rocks to be initially incorporated into the orogenic wedge and carried up to maximum elevations, then the lag can be used to estimate the rate of rock uplift in the contractional part of the Apennines. Mesozoic rocks currently exposed at the highest elevations (2.7–3.0 km) in the range were incorporated into the orogenic wedge from depths of 2–6 km beneath the Adriatic Sea (Argnani and Frugoni, 1997). This yields a range of rock uplift rates of 1.2–4.5 mm/yr, the upper end of which is approximately half the maximum rates of rock uplift in the most rapidly rising mountain belts on Earth. Thick foredeep and wedge-top sediments in the Adriatic foreland basin system indicate that large amounts of erosion have taken place since the late Miocene, which sug-

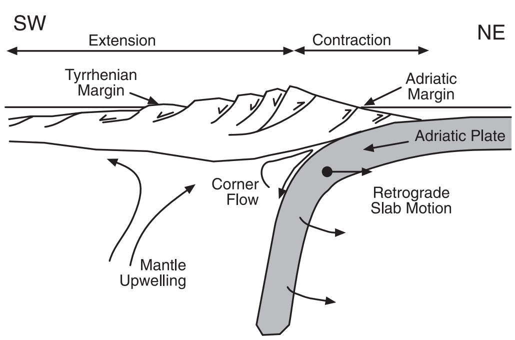

Figure 4. Schematic cross section of central Apennines showing subducting Adriatic slab, hypothetical corner-flow cell in mantle wedge, and distribution of contraction and extension.

gests that surface uplift rates have lagged significantly behind rock uplift rates.

Crustal thinning sufficient to permit the eruption of mantle-derived melts has taken place beneath the western, mature extensional basins (Ardea, Tiberino, Sora, and Rieti basins). The ~1.7–5.0 m.y. lag time between initial extension and volcanism may represent the range of time required for crustal thinning from ~40 km beneath the range crest to ~24 km beneath the western flank of the range, at a rate of ~3–9 mm/yr. Assuming Airy isostatic balance, crustal density of 2700 kg/m^3, and mantle density of 3300 kg/m^3, this rate of crustal thinning would have driven a coeval elevation collapse of ~0.6–1.7 mm/yr during late Miocene–Pliocene time. These simple calculations suggest that the northeastern flank of the range has risen at least twice as fast as its southwestern flank has fallen, and may help to explain the present steeply asymmetric topographic profile of the central Apennines, with a slope of 0.084 on the northeast flank versus a slope of 0.019 on the southwest flank. Although orographic precipitation patterns and regional drainage patterns undoubtedly affect the regional topography, the asymmetry must be fundamentally controlled by differing rates of tectonically driven uplift and subsidence.

The histories of contraction and extension in the Apennines fold-thrust belt suggest that the range is behaving as a crustal wave propagating northeastward. The front of the wave is the actively contracting part of the fold-thrust belt, the crest of the wave is the topographic culmination of the range, and the trailing trough of the wave is the collapsing southwestern slope of the range. Published models to explain the bimodal tectonic behavior of the Apennines call for mantle upwelling in response to foundering of the thickened Alpine lithospheric root without any active subduction (Decandia et al., 1998), mantle upwelling above the subducting slab (Keller et al., 1994), and general eastward flow of upper mantle beneath the Apennines (Doglioni, 1991). The first of these models is at odds with geophysical and geological data that clearly demonstrate a prolonged history of subduction along the Apennines, and in any case it lacks an explanation for the maintenance of contraction along the Adriatic margin. Our model differs from that of Doglioni (1991) in calling for the localization of contraction in the eastern Apennines by corner flow in the mantle wedge (trench suction), and regional extension in the western Apennines in response to mantle upwelling. Because these processes probably operate in any subduction system undergoing retrograde migration, it is likely that similar bimodal orogens characterized by paired contractional and extensional domains are typical of zones of incomplete continental collision, such as the Mediterranean.

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