Interactions between mantle upwelling, drainage evolution and active normal faulting: an example from the central Apennines (Italy)

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SUMMARY

In this paper we show that the processes that have shaped the Quaternary surface development of the Apennines in central Italy are all consequences of a single subcrustal process, the upwelling of the mantle. The relationship between gravity and topography shows that mantle convection is responsible for a long-wavelength (150–200 km) topographic bulge over the central Apennines, and stratigraphic evidence suggests this bulge developed in the Quaternary. Active normal faulting is localized at the crest of this bulge and produces internally-draining fault-bounded basins. These basins have been progressively captured by the aggressive headward erosion of major streams that cut down to the sea on the flanks of the regional bulge. The only surviving closed basins are those on the Apennine watershed most distant from the marine base level, where continued normal faulting is still able to provide local subsidence that defeats their capture by the regional drainage network. Understanding the competition between regional capture and local, fault-related subsidence of intermontane basins is crucial for recognizing potentially hazardous active faults in the landscape and also for interpreting the sediment supply to adjacent offshore regions. Central Italy provides a good modern analogue for processes that are probably common in the geological record, particularly on rifted margins and intracontinental rifts, but may not have been fully appreciated.

Key words: crustal deformation, Italy, mantle upwelling, normal faulting, Quaternary, uplift.

1 INTRODUCTION

This paper is concerned with the evolution of drainage and sedimentation in areas of continental extension that are also affected by regional uplift. Uplift and extension are commonly associated today, and understanding their combined effect on surface processes is important, not just to help identify and assess the risk from active faulting in places where earthquakes occur, but also to help interpret the older geological record in areas of economic importance, such as rifted continental margins. We illustrate the principles involved with examples from the Apennines of central Italy, where the effects of Pleistocene uplift and normal faulting dominate the active structure and geomorphology.

Extension of the upper continental crust is accommodated by normal faulting and there are many places where this process is occurring today. From studies of the earthquakes and surface faulting in such active regions, a lot is now known about the normal faults that move in large ($M_s > \sim 5.5$) earthquakes and various generalizations are possible. In particular: (1) such faults are roughly planar in cross-section, cutting right through the seismogenic upper crust with a relatively restricted dip range of about 30° – 65° ; (2) they typically occur in en echelon systems having maximum segment lengths that are probably related to the regional thickness of the seismogenic upper crust; (3) they form half-grabens with a characteristic maximum width that again depends on the thickness of the seismogenic layer (see e.g. Jackson & White 1989; Jackson & Blenkinsop 1997; Scholz & Contreras 1998; Ebinger et al. 1999). In extensional regions that are active today, the lateral continuity of the normal faulting and its associated tilting are major controls on the drainage patterns and distribution of sedimentary facies in half-graben basins (e.g. Leeder & Gawthorpe 1987; Leeder & Jackson 1993). These controls are particularly obvious where the normal faults occur near sea level, which provides a reference against which the vertical motions of footwalls and hanging walls are easily identified (e.g. Jackson et al. 1988; Goldsworthy & Jackson 2000). Although some areas of active continental extension are near sea level, such as the Aegean Sea and Gulf of Suez, others are in continental interiors and associated with elevated regions, such as Tibet, parts of East Africa, the northern Basin and Range province in Nevada and Lake Baikal. The association of extension with elevation is not surprising. The interiors of many plates are in compression because of the relative elevation of oceanic ridges (called 'ridge push': e.g. Forsyth & Uyeda 1975; Sykes 1978) and uplift is one way in which such stresses can be overcome to allow extension to occur. In continental interiors that are far from the sea the effects of regional uplift on the drainage and sedimentation patterns can be minimal, with the dominant process being the creation of internallydraining basins often occupied by lakes, such as in Tibet (e.g. Armijo et al. 1986). In the Apennines of central Italy regional uplift and extension both occur within about 100 km of the coastline, so that two competing processes are recognizable in the drainage and sedimentation: (1) the creation of internallydraining, fault-bounded half-graben, and (2) the capture and emptying of such basins by the aggressive headward erosion of major rivers that are cutting down to sea level in response to regional uplift. As a result, the geomorphology in central Italy is quite different from other active extensional provinces that are either already at sea level (like Greece) or in continental interiors (like Nevada). We suspect that it is partly for this reason that many active (and hazardous) tectonic structures in central Italy were unrecognized until relatively recently; they are more difficult to see unless the more subtle geomorphology is understood.

We begin with a brief review of the geological and tectonic setting of the central Apennines and of the present-day distribution of active faulting. We then summarize the geological and geomorphological evidence for regional uplift in the Quaternary. Then we use an analysis of the gravity and topography data in the frequency domain to show that the long wavelength topography is supported by convective motions in the mantle, and that this is the likely origin of the regional uplift. The main part of the paper then focuses on the regional topography and drainage networks and their evolution during the Quaternary at different scales. This allows us to discuss (1) the geological and geomorphological consequences of the regional uplift and normal faulting in the intermontane basins, and (2) the effect of mantle upwelling on the distribution of active normal faulting. Our aim is to show that the related processes of mantle upwelling, crustal extension and regional drainage evolution can be observed acting together in the central Apennines and that their interactions are clearly expressed in the geological and geomorphological record.

2 GEOLOGICAL AND TECTONIC SETTING OF THE APENNINES

Most reviews of the late-Tertiary evolution of the Tyrrhenian-Apennines system emphasize the eastward migration during the Neogene of paired extensional (in the west) and compressional (in the east) belts, together with flexural subsidence of the Adriatic foredeep and volcanism, all of which are envisaged as responses to the 'roll-back' of the subducting Adriatic-Ionian lithosphere (Elter *et al.* 1975; Malinverno & Ryan 1986; Royden, 1993; Serri *et al.* 1993; Faccenna *et al.* 1996; Jolivet *et al.* 1998). Thus the progressively eastward-younging foredeep and synrift deposits in the Apennines record the coeval activity and migration of the paired compressional–extensional belts in the Neogene (Patacca, *et al.* 1990). However, during the Quaternary the flexural subsidence, compressional deformation and east-

ward retreat of the subduction hinge all decreased dramatically (Patacca et al. 1990; Kruse & Royden 1994; Cinque et al. 1993) and the Apennines became dominated by crustal extension and vertical movements. Seismic reflection profiles in the Adriatic Sea show that the Mesozoic-Cenozoic sequence which is deformed by thrust anticlines is in turn overlain by prograding Quaternary deltaic sequences fed by streams draining the eastern flank of the Apennines, with little evidence of compressional deformation after the Early Pleistocene (Dondi, et al. 1985; Ori et al. 1993; Argnani et al. 1997). This Quaternary depositional pattern marks a dramatic change in subsidence rate and sediment supply from the Pliocene, during which up to 7000 m of sediment accumulated in a flexural trough close to the thrust front (Bigi et al. 1992). The Quaternary evolution thus involves the final infilling and extinction of the Mio-Pliocene Adriatic foredeep (Ori et al. 1993) and a regional NE tilting of the whole Adriatic coastal belt of central Italy (Dufaure, et al. 1989; Dramis, 1992; Kruse & Royden 1994). Evidence of present-day thrusting is contained in weakly deformed and tilted Quaternary deposits (Bigi et al. 1997) and moderate compressional seismicity on the NE side of the Northern Apennines (Frepoli & Amato 1997; Selvaggi et al. 2001) where intermediate seismicity down to a depth of 90 km may suggest that subduction is still active (Selvaggi & Amato 1992).

In the area of the central Apennines that is currently undergoing extension, normal faulting has been active since the Upper Pliocene–Early Pleistocene (Patacca *et al.* 1990; Bosi & Messina 1991). Normal faults cut a bedrock sequence dominated by resistant Mesozoic limestones and less resistant Upper Miocene flysch that were previously deformed by NW–SE striking Neogene thrust faults (Parotto & Praturlon 1975; Bigi *et al.* 1992). This extension is responsible for Pleistocene intermontane basins that are partially filled with alluvial, fluvial and lacustrine deposits and coarse conglomerates or breccias (Cavinato *et al.* 1993b; Cavinato & DeCelles 1999).

3 DISTRIBUTION OF ACTIVE DEFORMATION

Earthquake focal mechanisms show that the central Apennines are undergoing NW-SE extension, with seismicity concentrated along the main topographic ridge (Anderson & Jackson 1987; Amato et al. 1997) on active normal faults that overprint earlier compressional structures (D'Agostino et al. 1998). The total rate of extension across the Apennines is not yet well constrained. Seismic moment summations yield estimates between 0.9 and 3.5 mm yr^{-1} (Jackson & McKenzie 1988; Pondrelli et al. 1995) but are likely to be imprecise as the seismicity is relatively low and dominated by a few large events. Analysis of the deformation revealed by the Italian first-order triangulation networks in the interval 1865–1963 yields estimates of 3 mm yr⁻¹ as an upper bound for the extension rate accommodated in the central Apennines (Houseman & England, 1999). Both earthquake slip vectors and VLBI measurements suggest the motion of the Adriatic relative to Europe can be described by rotation about a pole in northern Italy (Anderson & Jackson 1987; Ward 1994), with the VLBI data predicting an extension rate in the Apennines of about 3 mm yr^{-1} at the 43°N increasing southward to 6 mm yr^{-1} at the latitude of Matera (Fig. 1). First GPS estimates of active crustal extension (D'Agostino et al. 2001) show that strain accumulation in the interval 1994–1999 is concentrated in a ~ 40 km wide belt extending at a rate of $6 \pm 2 \text{ mm yr}^{-1}$ (1 σ). The low levels of internal deformation



Figure 1. Summary map of the geodynamic framework of central Italy. The velocity of Matera relative to stable Eurasia was determined by VLBI (Ward 1994); the error ellipse represents the 95% confidence level. Thick dashed grey lines are small circles around the pole of rotation of the Adriatic region with respect to Europe (Ward 1994). Thin dashed lines enclose the boxes used for the calculation of the swath topographic profiles in Fig. 4. Dark shading represents area above elevation 500 m. Focal mechanisms are from the Harvard CMT catalogue. Active faults (thick lines) are from Galadini & Galli (2000) and Valensise & Pantosti (2000).

in the Adriatic Sea suggest that much of the extension in the Apennines is absorbed by shortening in the Dinarides (Anderson & Jackson 1987) with some also accommodated in the external part of the Northern Apennines (Frepoli & Amato 1997).

The pattern of active deformation in the central Apennines (Fig. 2) is revealed by the historical and instrumental seismicity and by the distribution of faults that have been active in the Late Pleistocene and Holocene. The historical and instrumental seismicity is generally confined to a belt whose width varies along the length of the Italian peninsula. In the northern and southern Apennines the active belt is only 20-30 km wide and may correspond to a single active fault system (Valensise & Pantosti 2000). However, in the central Apennines (approximately between the latitudes of 41.5°N and 42.5°N) the largest earthquakes (Selvaggi 1998) and the faults active in Late Pleistocene and Holocene time seem to be distributed over a broader width of at least 50 km and involve at least two major sub-parallel fault systems. The increase in width of the actively extending belt in the central Apennines is also associated with an increase in average elevation and the across-strike width of the Apennine topographic belt itself. In this central area the eastern normal

fault system bounds the Laga and Gran Sasso massifs (Blumetti et al. 1993; Giraudi & Frezzotti 1995) and the Sulmona basin (Vittori et al. 1995), while the western one starts from the northern end of the l'Aquila basin, crosses the Velino-Sirente massif, bounds the Fucino basin, and continues southward through the Marsica region (Fig. 2). The western fault system has produced several substantial earthquakes in the last thousand years (Fig. 2). The last shock ($I_{max} = X$ MCS; Boschi et al. 1999) of a sequence that occurred in 1703 is thought to have ruptured the fault system in the upper Aterno valley (Blumetti 1995; Cello, Mazzoli & Tondi 1998a). Trenching studies along the Ovindoli fault in the Velino-Sirente massif have revealed a previously unknown earthquake that occurred between 860 AD and 1300 AD (Pantosti et al. 1996). The normal fault system bounding the Fucino basin is known to have ruptured in 1915 $(M_s \sim 6.9)$ when co-seismic surface faulting was described at the time by Oddone (1915) and later re-investigated by others (Serva et al. 1986). In 1984 the southern Latium earthquakes $(M_s \sim 5.8 \text{ and } \sim 5.2)$ occurred in the Marsica region SE of the Fucino basin (Fig. 1, Westaway et al. 1989). The eastern fault system shows evidence of Late-Pleistocene to Holocene

Figure 2. Shaded relief map with faults active in the Late Pleistocene-Holocene from Galadini & Galli (2000), major Neogene thrust fronts and historical earthquakes. Light shading on the Adriatic side of the Apennines represents Lower Pleistocene transitional conglomerates closing the Quaternary sequence. On the Tyrrhenian side the trace of the Early Pleistocene shoreline is also shown.

activity (Galadini & Galli 2000) but can not be associated with any known historical earthquakes with the possible exception of the Aremogna fault which may have been activated in the 1349 earthquake (Valensise & Pantosti 2000). This apparent quiescence may suggest that the eastern system is now inactive and that extension is taken up only by the western fault system. Alternatively, it may indicate that seismicity is clustered episodically on to a single fault system with cycles whose time scale (perhaps 10^4-10^5 yr) is longer than the historical or paleoseismological catalogue; a suggestion that has also been inferred from geomorphological or historical data in Nevada, Turkey and Greece (Wallace 1987; Ambraseys 1989; Jackson & Leeder 1994; Jackson 1999) and which has some support from numerical models (Cowie 1998). Further insights are provided by GPS measurements in the interval 1994-1999 (D'Agostino *et al.* 2001) showing significant active strain accumulation across the western fault system suggesting also the possible existence of another undetected active fault system more to the SW.

This summary of the active deformation in the central Apennines highlights two main points: (1) that the active extension is concentrated along the main topographic ridge of the Apennines, and (2) that the increase in width of the actively extending belt between 41.5°N and 42.5°N correlates with the higher elevation and increased width of the topographic belt (see also Fig. 3). We argue below that the correlation between regional topography and active normal faulting reflects a dynamic link between sub-crustal processes and crustal extension. We first review the geological and geomorphological

Figure 3. (a, b, c) Swath topographic profiles across the Apennines. Elevation points from the 40 km-wide swaths in Fig. 1 are projected along the trace of the sections, and maximum, minimum and mean elevations calculated at regularly spaced intervals of 2 km. Regional topography (thick dashed lines) is obtained by fitting a 4th order polynomial fit to the mean elevations. The sections show that the longwavelength topographic bulge forming the Apennine belt is higher in the central Apennines (swath profile b). In this area the drainage divide, defined by the internally-drained and suspended Fucino basin, is shifted west of the highest elevations. Insets show the observed mean topographic elevations and the topography (dotted lines) obtained by dividing the filtered EGM96 gravity field by 50 mGal km^{-1} (see text for explanations). (d) shows the free-air gravity anomaly (mGal), derived from the EGM96 spherical harmonic coefficients, used for the calculation of topography expected from dynamic support (insets in a, b, c). See text for details on the applied filter.

evidence of regional uplift to define the approximate wavelength and amplitude of the Quaternary vertical movements. We then analyse the relationship between gravity and topography in the frequency domain to provide insights into the support of longwavelength topography and the origin of the regional uplift. We can then look at the effects of the interactions between normal faulting and regional uplift on the drainage network and topography.

4 QUATERNARY REGIONAL UPLIFT

4.1 Geological and geomorphological evidence

Various geological and geomorphological data indicate that a widespread surface uplift occurred during the Quaternary (Demangeot 1965; Ambrosetti et al. 1982; Dufaure et al. 1989; Dramis 1992), when the tectonics of the central Apennines was already dominated by crustal extension. Scattered outcrops of highly dissected marine Messinian-Lower Pliocene conglomerates, weakly deformed by later compressional structures and frequently made up of clasts (granites, metamorphic rocks) whose source area is hundreds of kilometres distant (Accordi & Carbone 1988), are found in the Apennines at high elevations (>1500 m in the Gran Sasso range), indicating that profound changes in elevation and morphology occurred in the Quaternary. One of the most significant effects of the uplift is found in the Tiber river valley close to the Tyrrhenian coast (Fig. 2). Here an Early Pleistocene shoreline that is continuously exposed for almost 100 km has been uplifted to an elevation of 200-400 m (Ambrosetti et al. 1987; Alfonsi et al. 1991; Girotti & Piccardi 1994). This ancient shoreline follows the long-axis of the Apennines and provides an important reference for the evaluation of vertical movements. It shows almost no shortwavelength deformation over its nearly continuous exposure, but smoothly decreases in elevation to the south, suggesting that it was raised by a large-wavelength regional uplift rather than by localized fault activity, which is apparently responsible only for minor local effects (Alfonsi et al. 1991). Palaeontological analyses (Gliozzi & Mazzini 1998) in the more western parts of the Rieti and Terni basin fills showed the existence of brackish marshes influenced by the contiguous Early Pleistocene Tyrrhenian Sea showing that the basin floors were approximately at sea-level and were significantly uplifted after the Early Pleistocene. Remnant Neogene-Pleistocene marine deposits found by Marinelli et al. (1993) in the Latium and Tuscany areas on the Tyrrhenian side of the Apennines increase in elevation to the NE, also supporting the suggestion of regional uplift. On the NE side of the Apennines the Adriatic foothills are characterized by a NE-dipping Pleistocene sequence made up of fine-grained marine deposits passing upward into sandy and conglomeratic deltaic deposits at the top of the Quaternary sequence (Cantalamessa et al. 1986; Ori et al. 1993). From the geomorphological point of view the whole Adriatic coastal region of central Italy displays a homogeneous evolution in the Quaternary, characterized by a regional NE tilting that is clearly reflected in the NE-trending parallel drainage network (Demangeot 1965; Dufaure et al. 1989; Dramis 1992). Furthermore, transitional fluvial to marine Lower-Middle Pleistocene deposits at the top of the Quaternary sequence are uplifted up to 500 m some kilometres inland of the coastline (Fig. 2; Cantalamessa et al. 1986; Bigi et al. 1995), giving an indication of the amount of uplift on the Adriatic side. A precise evaluation of regional uplift rates in the Middle–Late Pleistocene is hampered by the relative scarcity of remnant shoreline deposits and uncertainties in their ages, especially on the Adriatic coast. Nevertheless, approximate long-term uplift rates have been derived by Bigi *et al.* (1995) obtaining values ranging between 0.3 and 0.5 mm yr⁻¹ over the last ~1 Ma. Similar values (0.1–0.26 mm yr⁻¹) can be estimated for the long-term uplift on the Tyrrhenian side on the basis of the elevation (~200–400 m) of the Early Pleistocene shoreline (~1.5–2 Ma). These rates are likely to be higher along the central axis of the Apennines itself.

The geological and geomorphological data summarized above suggest a general doming of the Central Apennines on a wavelength larger than 150–200 km during the Quaternary. A post-700 kyr regional bulging of the Southern Apennines is envisaged by Bordoni & Valensise (1998), based on the elevations of marine deposits of 'Tyrrhenian' age (corresponding to isotopic stage 5e of ~ 125 ka). A similar Quaternary evolutionary pattern has been also described by Argnani *et al.* (1997) for the Northern Apennines, showing that the regional Pleistocene doming is a process that affected the whole Italian peninsula.

4.2 The origin of regional uplift and analysis of gravity and topography

The review of geological and geomorphological evidence for a regional doming of the Apennines in the Quaternary, listed in the previous section, requires a major phase of uplift whose timing coincides with or post-dates the change from shortening to extension. This observation argues against crustal thickening as the main cause for regional uplift. Instead attenuated uppermantle seismic velocities (Mele et al. 1996; 1997) beneath the Apennines, Quaternary mantle-derived magmatism Serri et al. 1993, Beccaluva et al. 1989), and mantle-derived helium in ground waters and natural gases (Hooker et al. 1985; Italiano et al. 2000) suggest that mantle processes contribute to the regional uplift of the Apennines. Processes occurring in the mantle can affect the regional elevation in various ways. An increase in regional elevation can be determined by thermal expansion caused by lithospheric thinning. This process requires (1) a thermal anomaly at the base of the lithosphere to thin the thermal boundary layer conductively or (2) the mechanical removal of the bottom part of the lithosphere by development of convective instabilities (Houseman et al. 1981). These processes are characterized by different time scales; while the first mechanism relies on the slow conductive propagation of heat through the lithosphere, the convective removal of the boundary layer can occur rapidly, in times less than the thermal time constant of the overlying rigid plate. Dynamic support caused by normal stresses at the base of the lithosphere that are related to convective flow is another way in which the mantle may cause uplift or subsidence of the surface. The resulting vertical motions are time-dependent and have time scales consistent with velocities of flow in the convecting mantle (McKenzie 1977). In this section we will compare these alternative models (1) with the constraints imposed by the results provided by the analysis of gravity and topography in the frequency domain and (2) with the timing, amplitude and wavelength of the uplift and the geological and geophysical data on crustal and lithospheric structure.

Short wavelength topographic relief is supported by flexural stresses, involving vertical motions of density contrasts in the lithosphere such as the crust-mantle boundary. Complete isostatic (Airy) compensation occurs as the elastic strength of the lithosphere tends to zero or the topographic load wavelength tends to infinity. Short wavelength topographic relief is thus essentially uncompensated and supported by the elastic strength of the lithosphere. The transitional wavelength between uncompensated short-wavelength and isostatically compensated longerwavelength topographic relief is diagnostic of the long-term $(>10^{6}$ years) elastic strength of the lithosphere defined by its effective elastic thickness T_e (Turcotte & Schubert 1982). For this process the linear transfer function (Dorman & Lewis 1970) between free-air gravity anomalies and topography in the frequency domain (admittance) progressively decays to zero from short to long-wavelengths, reflecting the absence of positive free-air anomalies for isostatic compensation. The decay of the admittance function at transitional wavelengths can be modelled by the theoretical admittance of an elastic plate, from which we can derive an estimate of T_e . Since flexurally supported topography is confined to short wavelength this mechanism is more readily isolated than other compensation mechanisms. Long wavelength gravity anomalies occur on the Earth and other planets such Mars and Venus, in regions where topographic swells, vulcanism and rifting suggest upwelling of the astenospheric mantle (Crough 1983; McKenzie 1994). When analysed in terms of isostatic compensation (McNutt & Shure 1986) these anomalies require depths of compensation (defined as the depth below which there are no horizontal pressure gradients) that imply the existence of horizontal density variations located below the lithosphere where viscosity is low and where horizontal density variations cannot be maintained for geological timescales and must therefore drive material flow (Parsons & Daly, 1983). Since the likely origin of these density variations is differential thermal expansion, this flow is by definition convective flow, irrespective of whether the density difference arise from hot upwelling or cold downwelling. Numerical studies of mantle convection show that admittance values of ~50 mGal km⁻¹ over the continents and ~30 mGal km⁻¹ over the oceans are typical for vigorous convection at high Rayleigh numbers (McKenzie 1994). These values arise from the joint contribution of a topographic bulge in the surface, which gives a positive gravity anomaly, and the anomalously low density of the hot asthenosphere material beneath, which gives a negative, but smaller, anomaly (Parsons & Daly 1983). In practice, the viscosity structure of the mantle determines the relative magnitudes of these two effects. Long wavelength topographic swells associated with positive gravity anomalies, apparent depths of compensation below or at the bottom of the lithosphere, and admittance values of $\sim 50 \text{ mGal km}^{-1}$ thus result from density variations beneath the lithosphere and therefore must be dynamically supported.

Fig. 4 shows the admittance as a function of wavenumber calculated from two different data sets (Fig. 4a,b). Detailed descriptions of the data and analytical methods are given in D'Agostino & McKenzie (1999) and McKenzie & Fairhead (1997). The observed admittance of 110 mGal km⁻¹ at large wavenumber shows that short-wavelength topography is essentially uncompensated. The decay of admittance of the SGI dataset (box in Fig. 4a) between wavelengths of 80 and 150 km (wavenumbers 0.0125-0.007 km⁻¹) reflects the increasing isostatic compensation of topography with increasing wavelength

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Figure 4. Free-air admittance analysis of the Apennines (modified from D'Agostino & McKenzie 1999). (a) Map of the SGI free-air gravity data set. The box encloses the area used for the corresponding spectral analysis. (b) Map of the free-air gravity constructed from the potential coefficients EGM96 complete to degree and order 360 (Lemoine *et al.* 1996). (c) Plot of the free-air admittance (error bars are 1 standard deviation). SGI data are plotted for short wavelengths (<150 km) and EGM96 data are plotted for long wavelengths. Corresponding wavelengths in kilometres are marked vertically above the admittance values. The dashed curve shows the best fitting elastic plate model for the SGI dataset in the wavelength range 80–200 km using T_e =3.7 km. Continuous line curves represent the admittance of long wavelength topographic swell supported by a low-density anomaly in the mantle for various depths of compensation (McNutt & Shure, 1986). The observed admittance is reproduced with a range of apparent compensation depths ranging between 75 and 150 km. These depths of compensation are not consistent with the existence of static horizontal density variations in the viscous mantle and are therefore best interpreted to be dynamically supported by mantle convection (see text for discussion). (d) Plot of coherence between free-air gravity and topography. These values represent the phase correlation between gravity and topography and show that about 80% and 50% of the gravity signal is correlated with topography for the SGI and EGM96 datasets respectively.

and can be used to obtain an estimate of the effective elastic thickness (T_e). In this wavelength range the admittance is best modelled with a T_e of ~4 km. At longer wavelengths we can use the box gridded with the EGM96 dataset (box in Fig. 4b) whose larger dimensions provide more reliable estimates for long wavelength anomalies than does the SGI dataset. In this wavelength range the admittance cannot be modelled by flexure caused by crustal loading but tends instead to a constant value of about 50 mGal km⁻¹. The existence of these gravity anomalies may suggest that the long wavelength part of the topography is isostatically compensated by a subcrustal density contrast. We thus calculate the range of depths of compensation

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required to support the observed gravity anomalies. These depths of compensation can be formulated both in terms of Airy or Pratt compensation. In the former case the depth represents the average depth of the compensating density contrast, in the latter the mid-depth of the thermal anomaly. The resulting values are thus considered as apparent depths of compensation. We estimated the apparent depth of compensation using the linear filter technique proposed by McNutt & Shure (1986). This filter predicts the relationship between gravity and topography for long wavelength topographic swells caused by a low-density mantle layer, and includes also the contribution of surficial crustal loading. Using the elastic thickness derived from the short wavelength analysis, the observed long-wavelength EGM96 gravity anomalies can only be reproduced with this model if we assume an apparent depth of compensation ranging between 75 and 150 km. These large values of compensation depth imply the existence of horizontal density variations in regions of the mantle where the viscosity is sufficiently low to drive convective flow. These considerations together with the an approximately constant value of 50 mGal km⁻¹, typical of convective flow from numerical experiments, suggest that isostatic compensation in the mantle is not consistent with the thermal and viscosity structure of the mantle, and that the observed long wavelength gravity anomalies and topography are more simply explained by a dynamic support maintained by flow in the convecting mantle. Numerical experiments also show that the effect of a low-viscosity layer beneath the lithosphere in a convecting mantle is to decrease the apparent depth of compensation that can appear to be well within the mechanical boundary layer (Robinson et al. 1987).

Other arguments point against the possibility of a significant thinning of the thermal boundary layer or crustal thickness variations for the origin of regional uplift. Regional uplift of 0.5-0.1 km requires 25-50 km of lithospheric mantle thinning, assuming an average density contrast of 0.02 g/cm³ between the lithosphere and asthenosphere. The resulting uplift could be produced (1) if the horizontal stretching is limited to the mantle portion of the lithosphere. Alternatively (2) the uplift could be produced by the conductive thinning of the thermal boundary layer. These mechanisms seem unrealistic when compared to geological and geophysical observables because they require complete mechanical decoupling between crust and mantle and a large time interval to conductively thin the lithosphere. Crustal refraction profiles (Scarascia et al. 1994) show that the crustal thickness increases from the Tyrrhenian coast to the Apennines. Crustal thickness variations can thus explain part of the long-wavelength Apennines topography, but cannot account for the timing of uplift (see previous Section) and the presence of large wavelength gravity anomalies.

The main results of this section can thus be briefly summarized as follows:

(i) The effective elastic thickness (T_e) of the lithosphere in the Apennines is ~4 km. This value is within the range of typical T_e estimates obtained in actively extending regions where the uplift of the isotherms strongly contributes to the decrease in the strength of continental lithosphere. It is this low value of T_e that allows the surface effect of mantle upwelling on topography (normally filtered by higher values of T_e) to be observed in a relatively small area such as the Apennines.

(ii) The observed long-wavelength admittance strongly supports the hypothesis that the long-wavelength (>150 km) topography in the Apennines is dynamically supported by mantle convection and not isostatically compensated by thickened crust or thinning of the lithospheric mantle. We thus suggest that the onset of the regional Quaternary uplift in the Apennines is also likely to have been triggered by mantle upwelling, not by crustal thickening, in agreement with the geological and geomorphological evidence summarized in the previous section. Such convective support requires temperature contrasts to exist below the lithosphere, which could result from either hot upwelling material from a plume or cold downwelling from subduction.

4.3 Physiography of the central Apennines

We have summarized the evidence for an important regional Quaternary uplift of the central Apennines that is related to mantle upwelling, crustal extension and the distribution of active normal faulting. These processes are likely to have had a significant impact on the geomorphological evolution of the area. In this section we use a regional analysis of the topography and drainage to provide information on the wavelength and amplitude of the uplift and its interaction with geomorphic processes.

The main topographic features of the Apennines are revealed by NE–SW trending swath profiles perpendicular to the belt (Fig. 3). The profiles were made by projecting elevation points from a 40 km-wide swath centred on the line of the section. The minimum, maximum and mean elevations of the projected points were then determined in 2 km segments along the profile. For this analysis we used a digital elevation model containing average elevations in cells of 10 and 7.5 longitude and latitude arcseconds (~230 m) respectively. For the more detailed studies of local topographic features we describe later we used digital topography data originally derived from contour lines on 1:25,000 topographic sheets.

Fig. 3b shows that the topographic relief of the central Apennines is dominated by the superposition of two main wavelengths. The first is an approximately 30 km spacing of ranges whose maximum elevations increase progressively towards the NE. These ranges are spatially confined to the extended terrain, with the highest elevations on the eastern margin of the extensional domain in the footwall of the easternmost normal faults. The basic structure of the 30 km-spaced ranges is of Mesozoic limestones deformed by Neogene thrusting and bounded by Quaternary normal faults on their SW side that have, in some cases, evidence of Holocene activity (Galadini & Galli 2000). A semi-regular spacing of normal-fault bounded ranges is a common feature in areas of distributed extension such as Nevada and Greece, and may be related to the maximum shear stress that is sustainable on major fault systems (e.g. Jackson & White 1989; Foster & Nimmo 1996). In this case the regularly-spaced ranges occur where the actively extending region is wider than it is to the north or south and involves at least two major normal fault systems (Fig. 2). The minimum elevations in the swath profile in Fig. 3b roughly correspond to the regional level of fluvial incision and reveal a shift between the drainage divide and the highest elevations. The Apennine watershed is located near the flat area in the middle part of the minimum elevation profile in Fig. 3b, which corresponds to the internally-draining Fucino basin, whose basin floor is suspended above the regional levels of fluvial incision to the NE and SW.

The second dominant wavelength in the swath profile corresponds to broad topographic bulge ~200 km wide, shown by the polynomial fit to the mean elevations in Fig. 3. As already discussed, the existence of raised Pleistocene shorelines on the flanks of this topographic bulge and the gravity–topography signature in the frequency domain suggest that mantle upwelling has driven the recent regional uplift and that the long-wavelength topographic bulge is the direct expression of mantle upwelling beneath the Apennines. The 50 mGal km⁻¹ convective signal at long-wavelength (Fig. 4c) and the high coherence between the gravity and topography (Fig. 4d) suggest a direct comparison between the observed

topography and the topography expected by mantle upwelling. This expected topography could be calculated by simply dividing the long-wavelength free-air gravity by the 50 mGal km^{-1} convective signal. The free-air gravity (Fig. 3d) has been derived from the EGM96 potential coefficients applying a filter to remove the effects of the short-wavelength flexurally supported gravity anomalies, and the long-wavelength field associated with deeper mantle processes. The filter has been applied in the frequency domain by cosine-tapering the EGM96 potential coefficients between 150 and 250 km and between 300 and 1000 km. Coefficients corresponding to wavelengths larger than 1000 km and smaller than 150 have been set to zero. Wavelengths between 250 and 300 have been band-passed. For spherical harmonics, $2\pi a/(l+0.5)$ was used as the equivalent wavelength, where a is the Earth radius and l is the harmonic degree. The comparison between the observed and expected topography (insets in Fig. 3a,b,c) shows that the topography expected from convection has similar amplitude, wavelength and position to the observed long-wavelength regional topography, confirming that a large part of the gravity signal is coherent with topography and is likely to be related to dynamic support in the mantle.

The age of the Pleistocene marine deposits suggests that the bulge itself is quite recent and probably younger than 1 Ma. A map view (Fig. 5) of the long-wavelength regional topography can be obtained by low-pass filtering the observed topography in the space domain with a Gaussian filter ($6\sigma = 100$ km). Fig. 5, and a comparison with Fig. 2, confirm that the closed basins

Figure 5. Map of the long-wavelength topography (contours every 200 m obtained by Gaussian filtering in the space domain; $6\sigma = 100$ km) and regional drainage. Between points A and B the main drainage divide (grey dashed line) is shifted west of the line of highest elevations (black dotted line). The distribution of closed extensional basins (Colfiorito and Fucino) and active deformation (*cf* Figs. 1 and 2) closely follows the top of the long-wavelength topographic bulge.

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of Colfiorito and Fucino, the active normal faulting, and the main Apennine drainage divide are all located at the top of the long-wavelength topographic bulge.

4.4 The regional drainage network

The drainage network of the central Apennines is markedly asymmetric (Fig.5). The Adriatic foothills are characterized by a parallel drainage pattern perpendicular to the NW-SE Apennine topographic trend (Mazzanti & Trevisan 1978). In this area the Quaternary evolution shows a reversal in the sense of the vertical movements with respect to the previous Neogene history, with the foredeep of the earlier thrust belt now uplifted (Dufaure et al. 1989; Dramis 1992; Kruse & Royden 1994). Geological data and seismic reflection profiles both show that relatively undeformed Quaternary deposits regionally cap and seal earlier anticlines (Ori et al. 1993). The Quaternary drainage evolution on the Adriatic side thus seems to have been controlled by regional NE tilting after the end of the main thrusting. The drainage network has been deeply entrenched in the pre-existing depositional or erosional surfaces with the formation of three orders of terraces after the Early-Middle Pleistocene (Coltorti et al. 1991). In many places the Early Pleistocene fluvial-alluvial and coastal sedimentary sequence (Fig. 2) has been stripped by erosion and only occasional remnants, with a marked NE dip, are preserved on interfluves and near the coast (Bigi et al. 1995).

On the western flank of the Apennines drainage evolution has been deeply influenced by normal faulting. NW-trending axial drainage systems along the main syn-rift basins, combined with shorter NE-trending transverse sections that frequently cut through Mesozoic-Cenozoic bedrock, have together established a through-going fluvial network connecting the major intermontane basins. The majority of the Quaternary extensional basins are now part of this regional drainage system while the only remaining closed basins (Fucino and Colfiorito) are located along the Apennines watershed. The swath-profiles (Fig. 3b) and the filtered map (Fig. 5) show that the drainage divide in the central Apennines is located approximately in the middle of the Italian peninsula along the crest of the longwavelength topographic bulge and is shifted west of the line of highest elevations. Thus in the region of the bulge itself, some of the fluvial systems draining to the Adriatic Sea (e.g. Aterno-Pescara Vomano, Tronto) have their headwaters in the actively extending area west of the highest elevations, flowing through the easternmost normal fault footwall blocks in deeply incised gorges to enter the Adriatic foothills where they assume the characteristic NE trend. Whether this pattern is caused by stream antecedence on the highest elevations or from regressive erosion by the Adriatic rivers is an old debate in the geomorphological literature (Demangeot 1965; Mazzanti & Trevisan 1978). A clue is contained in the observation that the region where the drainage divide is west of the line of highest elevations (roughly between points A and B on Fig. 5) is also where the line of highest elevations is closer to the Adriatic coast than it is further north or south (see also Fig. 1). To us, this suggests that the local drainage divide was defeated, allowing streams flowing into the Adriatic to enlarge their catchments westwards by capturing previously internally-drained, fault-bounded intermontane basins. The evolution of the Aterno-Pescara catchment, discussed in more detail later (Fig. 7), is a good example of this process. A similar evolution is also suggested

Figure 6. The Terni basin. (a) Geological map (modified from Ambrosetti *et al.* 1987. Points A and B respectively represent the lowest outlet of the Early Pleistocene Terni basin and the gorges formed by the Nera River in the Narnese–Amerina ridge after the Early Pleistocene. (b) Topographic cross-section showing the amount of sediments removed from the Terni basin. The top of the Terni basin fill roughly corresponds with the elevation of the Early Pleistocene shoreline in the Tiber valley showing the important erosional dissection of the Terni basin fill occurred after the Early Pleistocene. (c) Perspective view of topography of the Terni basin and Tiber valley. The grey arrow in this and following block diagrams points North.

Figure 7. (a) Schematic map of the Aterno–Pescara river system. The Popoli gorges are the outlet of the Aterno–Pescara catchment and their evolution has controlled the erosional/aggradational processes upstream during the Pleistocene. (b, c) Geological cross-sections from the l'Aquila and Sulmona basins showing a clear transition from lacustrine or low-gradient fluvial Lower–Middle Pleistocene deposits to transversely-sourced Middle Pleistocene alluvial fan deposits.

by the drainage network in the southern Apennines, where the pattern is reversed: there the highest elevations are closer to the Tyrrhenian Sea and the drainage divide is shifted further east (Amato, Cinque & Santangelo 1995). The regional trend is thus to locate the drainage divide in the centre of the Italian peninsula, overcoming local complexities given by the closer proximity of the highest elevations to the Adriatic or Tyrrhenian coast (Fig. 5). The trace of the divide is characterized by sinuosities that locally reflect regressive erosion from one side of the watershed, and by large areas of internal drainage corresponding to the actively extending Colfiorito and Fucino basins, which have not yet been captured. Thus the present-day drainage pattern represents the incomplete evolution from an earlier system that was probably dominated by internallydrained, fault-bounded fluvial basins towards a regional throughgoing drainage network, so that internally-drained basins are now preserved only along the watershed. This evolution is driven by (1) the regional uplift induced by mantle upwelling beneath the long-wavelength topographic bulge and (2) the proximity of the coast, providing a base level to which the regional drainage must incise.

The intermontane basins of the central Apennines contain a record of the depositional and erosional events that occurred during the Quaternary. Their stratigraphy is likely to reflect how the regional drainage network evolved on the flanks of the developing long-wavelength topographic bulge and also how it interacted with active normal faulting. We summarize this evidence in the next section.

5 QUATERNARY STRATIGRAPHY OF THE INTERMONTANE BASINS

The stratigraphy of the Quaternary intermontane basins in the central Apennines basins has been studied closely and synthesized in the studies by Cavinato *et al.* (1993b) and Bosi & Messina (1991) among others. Lithological and palaeontological correlations throughout these basins highlight common regional tectonic and climatic controls in addition to the influence of the basin-bounding normal faults.

The first deposits associated with the formation of the intermontane basins consist of coarse-grained breccias and conglomerates generally assigned to the Upper Pliocene–Early Pleistocene (Demangeot 1965; Cavinato *et al.* 1993b; Bosi & Messina 1991; Bagnaia *et al.* 1989). Low-relief surfaces, locally preserved above the major calcareous ridges (Demangeot 1965; Bosi & Messina 1991; Dramis 1992) and occasionally corresponding to outcrops of these early breccia deposits, are significantly dissected and fragmented by erosion and tectonic deformation. A progressive younging in the age of the oldest deposits of the intermontane basins is observed from west to

east, with Middle-Late Pliocene sediments at the bottom of the westernmost basins (Terni and Rieti) and Early Pleistocene sediments at the bottom of the more eastern ones (Sulmona and Colfiorito; see Fig. 2 for location). The Early Pleistocene is characterized by lacustrine environments in most of the intermontane depressions, recorded by widespread lake beds that are revealed within the incised basin fills (Michetti & Serva 1990; Bosi & Messina 1991; Cavinato et al. 1993b) or have been drilled (GE.MI.NA. 1962). The Early-Middle Pleistocene lacustrine deposits are generally overlain by units that are transitional from lacustrine and low-gradient fluvial environments to coarser deposits representative of alluvial fans (Blumetti & Dramis 1992; Miccadei et al. 1998). This transition is frequently marked by erosion and incision of the lake beds so that the Middle Pleistocene deposits are often entrenched and unconformably overlie the fluvial-lacustrine units. In some cases depositional surfaces are preserved within the lacustrine and fluvial deposits while in others it is more difficult to estimate the amount of incision of the fluvial-lacustrine sequences. After the Middle Pleistocene, deposition of lacustrine sediments in the intermontane basins was drastically reduced and continued only in basins that maintained internal drainage. During the Pleistocene abundant pyroclastic materials derived from alkaline-potassic volcanic centres on the Tyrrhenian coast provide radiometrically datable tephra layers interbedded within the continental sequences (Cavinato et al. 1993b; Miccadei et al. 1998).

In summary, the intermontane basins generally record a history in which the Early–Middle Pleistocene continental fluvio-lacustrine environments were later incised and covered by alluvial fans. This succession is consistent with the capture of internally-draining, fault-bounded basins by the regressive headward erosion of major regional streams cutting down to a lower base level. The changes in facies and environment are unlikely to be climatically induced, since some closed lake basins, such as Fucino, have survived to modern times. Over the Quaternary as a whole, this evolution is also typical, with slightly different timing, of the intermontane basins of the Northern (Argnani *et al.* 1997) and Southern Apennines (Capaldi, Cinque & Romano 1988).

6 SUMMARY OF THE PREVIOUS SECTIONS

At this point it is sensible to summarize the main points so far;

(i) The regional topography of the central Apennines is dominated by a NW–SE trending bulge ~ 200 km wide, the top of which correlates with the location of the currently active normal faulting, the Tyrrhenian-Adriatic watershed, and the survival of closed fault-bounded basins.

(ii) Admittance analysis of the gravity and topography data shows that the topographic relief at wavelengths longer than 150 km is supported dynamically by mantle convection, suggesting that the ~ 200 km topographic bulge has been formed above upwelling mantle beneath the Apennines.

(iii) A rough estimate of the uplift rate, based on uplifted Pleistocene marine deposits on the flanks of the Apennines, is $\sim 0.2-0.5$ mm yr⁻¹ and is likely to be higher along the axis of the Apennines.

(iv) Preserved stratigraphy in the intermontane basins shows a regional transition from Early–Middle Pleistocene lacustrine and low-gradient fluvial deposits to Middle Pleistocene alluvial fan sequences, consistent with a major reorganization of the regional drainage network from one dominated by internallydraining fault-bounded basins to the formation of a throughgoing river system.

Regional uplift was coeval with the development of synextensional intermontane basins whose geological and geomorphological evolution should therefore have been influenced by the interactions between normal faulting and regional uplift. In the next sections we analyse in greater detail the relationships between the drainage network, regional uplift and normal faulting to obtain some understanding of how drainage is captured and integrated into a regional network, how surviving closed, actively-extending basins have maintained their internal drainage, and finally, the present-day balance between regional incision and tectonic subsidence in the extensional basins. The aim of this approach is to provide insights in to how the drainage network responds to mantle upwelling and interacts with normal faulting.

7 EVOLUTION OF THE DRAINAGE NETWORK

The drainage system of central Italy consists of an integrated stream network on the flanks of the Apennines and a few closed, internally-draining basins located along the main watershed. We look first at examples of basins that have been captured and integrated into the regional network and then at the surviving closed basins of Fucino and Colfiorito.

7.1 Examples from the captured basins

7.1.1 The Terni basin

The Terni basin, occupied by the Nera river, constitutes the southwestern part of the large Tiber basin (Ambrosetti et al. 1987), one of the Plio-Pleistocene syn-rift basins of the Apennines (Fig. 6). Its evolution has been controlled by a N-S to NW-SE trending normal fault system bounding the Martani mountains (Brozzetti & Lavecchia 1995). The fluvial and lacustrine deposits infilling the basin consist of sands, clays and gravels of Pliocene to Early Pleistocene age (Ambrosetti et al. 1987). During the Early Pleistocene a continental fluviolacustrine environment in the Terni basin was laterally continuous with a shallow-marine environment in the nearby Tiber river valley to the SW. Transitional marine-continental Pleistocene deposits are preserved above the Narnese-Amerina ridge showing that these two environments were in communication through the former outlet of the Terni basin marked by point A in Fig. 6a,c. Palaeontological evidence also shows the existence of brackish marshes in the Terni basin documenting the periodic influx of marine waters into the continental basin. Early Pleistocene marine sediments onlap the SW side of the Narnese–Amerina ridge covering the normal fault system that bounds the ridge on the Tiber river valley (SW) side, indicating that this fault had probably become inactive by Lower Pleistocene time. Holes bored by Lithophaga (a marine mollusc) can be found on the Narnese-Amerina ridge (Fig. 6a) marking the position and uplift (350-400 m above present sea level) of the early Pleistocene shoreline in the Tiber river valley (see section 4.1). These observations suggest that during the Early

Pleistocene the fault-controlled Terni basin had its floor approximately at sea level and had its lowest outlet on the Narnese-Amerina ridge. The sediment available to fill the basin exceeded the potential accommodation space created by local tectonic subsidence, so that the basin was filled to its lowest outlet, which was probably point A in Fig. 6a. Since the Early Pleistocene the drainage network has deeply incised the basin fill to a depth of about 250 m and the Nera river has formed a deep gorge crossing the Narnese-Amerina ridge (point B in Fig. 6a,c). The topographic section in Fig. 6b shows the incision and erosion of the Terni basin fill caused by the downcutting Nera river. Thus local tectonic subsidence caused by the fault system bounding the Martani mountains has been outpaced by fluvial downcutting since the formation of the Early Pleistocene shoreline features. This fluvial downcutting is not related to structural position with respect to the active faulting, being in the hanging wall of the Martani mountains fault, but is associated with the regional uplift that has raised the Early Pleistocene Tiber river valley shoreline.

7.1.2 The Aterno–Pescara river system

The Aterno–Pescara river system (Fig. 7) upstream of the Popoli gorges drains a large area (2200 km²) of the central Apennines, and lies west of the line of highest elevations that also marks the easternmost limit of late Quaternary normal faulting. We argued earlier that the catchment enlargement of the Adriatic-flowing rivers in this part of the Apennines was achieved by defeat of local drainage divides along the line of highest elevations. If this is true, then the effects of the capture should be recorded in the Pleistocene continental sequences that are preserved in the area. Quaternary deposits along the Aterno river are found in three main sedimentary basins, from NW to SE: the L'Aquila–Fossa basin, the Subaequan basin and the Sulmona basin (Fig. 7a).

In the upper reaches of the Aterno river drainage is mostly axial (NW–SE) and in the hanging walls of the main normal fault systems. After flowing through the Pleistocene deposits of the L'Aquila–Fossa basin the river is then deeply incised in

the hanging wall of the Subequan normal fault system (see also Fig. 8) before entering the S. Venanzio gorges, which occupy a dextral en echelon step in this fault system, to emerge into the Sulmona basin. The Pleistocene evolution of all these basins is characterized by a sequence of erosional and aggradational phases (Bertini & Bosi 1993; Bagnaia et al. 1989; Miccadei et al. 1998) with a common trend in which Lower to Middle Pleistocene lacustrine deposits are incised and unconformably overlain by Middle Pleistocene fluvial or alluvial fan deposits (e.g. Fig. 7b,c). In the Sulmona basin tephra layers interbedded within the continental deposits provide dates (Miccadei et al. 1998) for the lacustrine deposits (0.7-0.44 Ma BP) and the overlying alluvial fan deposits (0.35-0.135 Ma BP). The common evolution of all the basins along the Aterno-Pescara system upstream of the Popoli gorge suggests an external control due to a major lowering of the base level, rather than faultcontrolled erosional processes. The incision of the lacustrine deposits in the Sulmona basin represents the entrance of the external drainage network into the basin following the defeat of the local divide that was located in the Popoli gorges (Miccadei et al. 1998). The elevation of the Sulmona basin now acts to some extent as a buffer that controls the erosional-aggradational processes further upstream in the Aterno valley. Thus we believe the capture of the basin by the external drainage through the Popoli gorge triggered a wave of regressive erosion in the basins located upstream, causing a major reorganization of the drainage network, a change from lacustrine to alluvial-fluvial environments, and the incision of the lacustrine deposits.

The geological and geomorphological data therefore support the previous hypothesis that the Pleistocene drainage has evolved by integration and capture of pre-existing low-energy internal drainage systems by an incising regional network on the flanks of the evolving long-wavelength topographic bulge. This evolution seems to be broadly symmetrical on the two flanks of the Apennines.

Where the regional drainage has entered the fault-bounded intermontane basins, the present drainage pattern reflects the balance between the regional downcutting and the local subsidence induced by active normal faulting. This relationship is

Figure 8. (a) Perspective view and (b) topographic cross-sections in the Middle Aterno valley. The river course is deeply incised in the hangingwall of the NW–SE active normal faults and in the Pleistocene deposits of the Subaequan basin.

clear in the hanging wall of the Subequan normal fault system (Fig. 8a, 10a), where the almost-flat syn-rift deposits of the basin are suspended as high terrace-like features well above the bed of the Aterno river which has incised deeply into them.

7.1.3 Drainage and fault evolution and migration

The previous examples have emphasized how the establishment of a new, lower base level following the capture and integration of the extensional basins into the regional drainage network has led to erosion and incision of basin fills. In general, the fluvial downcutting by the regional network is faster than local tectonic subsidence in the hanging walls of the active faults. In many places in the Apennines this situation has produced a landscape that is quite different from that where internallydraining fault-bounded basins are unaffected by regionallyincising external drainage networks. Under favourable conditions, relatively subtle geomorphological features such as deviated drainage and wind gaps can provide evidence of fault evolution or propagation (e.g. Leeder & Jackson 1993). Not all such features have been obliterated in the captured Apennine basins, and we offer one example below.

The Marine Mountain and Pettino Mountain faults (Fig. 9) are located in the upper Aterno river valley (see Fig. 7 for location) and form a right-stepping en echelon pair bounding ~ 1000 m-high mountain fronts of Mesozoic limestone. Recent activity on the faults is revealed by fault scarps in Upper Pleistocene slope deposits at the base of both mountain fronts (Blumetti 1995). The topographic relief associated with the

Pettino Mountain front dies out towards the NW suggesting a similar decrease in the throw on the fault (Fig. 9). At its NW end the decreasing offset on the Pettino Mountain fault is probably absorbed by the Marine Mountain fault, whose associated relief increases to the NW. The area where the two faults overlap exposes terraced gravel deposits that are incised and suspended about 50 m above the present alluvial plain of the Aterno river (Fig. 9). The incision of these deposits progressively increases southward towards the Pettino Mountain fault. At the NW termination of this fault, near the village of S. Vittorino, a dry valley suspended about 50 m above the Aterno river trends perpendicular to the fault strike and contains gravels and conglomerates. This wind gap represents the former course of a stream that once drained south from the base of the Marine Mountain fault scarp. The observations that the dry valley has been beheaded by the development of a stream channel flowing sub-parallel to the fault and has also been uplifted above the present alluvial plain by the Pettino Mountain fault, suggest that the present-day morphology and drainage has resulted from the northwest propagation of the Pettino Mountain fault tip.

Fig. 9 shows another feature common in en echelon fault steps, which is the development of a major stream catchment flowing sub- parallel to the faults down the relay ramp between the fault tips (Jackson and Leeder 1994). These catchments are always likely to be bigger than the small footwall-sourced stream channels flowing perpendicular to the faults (such as the earlier S. Vittorino valley), and are likely to replace them as the frontal fault tip propagates, as we suspect has happened at S. Vittorino.

Figure 9. (a) Shaded relief map of the Marine Mt–Pettino Mt fault system. Light shading represents alluvial and fluvial Quaternary deposits. The western propagation of the NW tip of the Pettino fault has left a wind gap near the S. Vittorino village formerly occupied by a southflowing river successively beheaded by the development of a major stream catchment flowing sub-parallel to the faults down the relay ramp between the fault tips. (b) Cross-sections drawn perpendicular and along the strike of the Pettino Mt fault. (c) Perspective view of topography.

This example shows that the more subtle geomorphological consequences of fault evolution can sometimes be recognized in the captured basins, perhaps more easily in cases like S. Vittorino which are quite distant from the regional base level. Further downstream, where the effects of regional downcutting are more dramatic, it is probably more difficult to see them (see Fig. 8 for a comparison).

7.2 Examples from the closed basins

Where the base level is regionally controlled the general downcutting tends to obscure geomorphological features that directly reflect the interaction between normal faulting and drainage. It should be easier to observe such features where internally-draining fault-bounded basins have survived along the Tyrrhenian-Adriatic watershed, such as at Fucino and Colfiorito, in both of which the local base levels are suspended above the regional one (Figs. 3 and 5). These basins can also provide some understanding of the competition between local subsidence (induced by active faulting) and regional regressive

erosion (induced by regional uplift). The recent 1915 January 13 Fucino and 1997 September Colfiorito seismic events also allow us to examine the consistency between the present-day deformation and the longer-term Quaternary evolution.

7.2.1 The Fucino basin

The Fucino plain is a closed basin in the middle of the Apennines belt (Fig. 11) whose local base level (650 m) is well above the regional levels of fluvial incision in the nearby Liri valley and Sulmona basin (Fig. 3b). The central part of the basin once contained the third largest lake in Italy ($\sim 150 \text{ km}^2$), the last part of which was artificially drained at the end of the 19th century. Seismic reflection profiles and surface geology indicate that the Fucino basin is a half-graben controlled by a SW-dipping fault system consisting of three parallel subsidiary faults (Blumetti *et al.* 1993; Cavinato *et al.* 1993a; Messina 1996; Piccardi *et al.* 1999). The fault furthest to the NE juxtaposes Mesozoic bedrock and Plio–Pleistocene continental deposits. The two southwestern faults produced coseismic

Figure 10. (a) Panoramic view over the Aterno River valley and the Subaequan basin. Thick arrows mark the trace of the range-bounding normal fault. Note the fluvial incision in the hanging-wall of the normal fault by the Aterno River (see also Fig. 8). (b) Panoramic view over the closed Fucino basin. White arrows mark the approximate trace of the 1915 earthquake surface ruptures. The basin still preserves internal drainage and its base level is controlled by the local fault-related subsidence.

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Figure 11. (a) Landsat TM (190/31) image of the closed Fucino basin. The Fucino basin local base level (corresponding to the flat cultivated area in the centre image, bounded to NE by the active normal fault system that moved during the 1915 M_s =6.9 earthquake) stands at an altitude of ~650 m, markedly suspended above the average level of fluvial incision in the nearby Liri valley and Sulmona basin. (b) Simplified map of the Fucino basin. (c) Perspective view of topography.

surface ruptures during the 1915 January 13 Fucino (M_s 6.9) earthquake that were initially described by Oddone (1915) and later rediscovered by Serva *et al.* (1986). The Plio–Pleistocene basin fill is exposed only in the footwalls of the faults bounding the basin to the NE and NW, and in the hanging walls is deeply buried below the present basin floor. This outcrop pattern emphasizes that depositional and erosional processes in the Fucino basin are controlled by faulting rather than by the influence of the regional drainage network.

More recent evidence for vertical displacements is provided by elevation changes in benchmarks around the former Fucino lake after the 1915 earthquake (Loperfido 1919). These were used by Ward and Valensise (1989) and Amoruso *et al.* (1998) to estimate source parameters of the 1915 earthquake and show up to 50 cm of local subsidence in the Fucino basin, consistent with the longer-term Quaternary pattern of sedimentation (see also Michetti *et al.* 1996 for a critical discussion). Estimates of the extension rate across the Fucino basin obtained from trenching studies are in the range 0.6–1 mm yr⁻¹ (Galadini & Galli 1999).

The lowest potential outlet of the internally-drained area is located on a broad saddle (point A in Fig. 11b,c) at the southwestern end of the NE-SW trending Tre Monti fault in the hangingwall of the active fault system. In this area the river network suggests a capture by the external drainage from the Salto River valley. In addition, the drainage pattern and preserved terraces in the upper Liri valley have been interpreted by Giraudi (1996) to imply capture of the upper course of the Liri river (which originally drained the Fucino basins through the Capistrello passage) by regressive headward erosion coming from further downstream, leaving behind the wind gap near Capistrello. Thus the drainage area of the wider Fucino basin was reduced in part during the Pleistocene by the external capture of its outer catchments. Yet the innermost part of the Fucino basin has retained its internal drainage. We attribute its survival to two factors: (1) its greater distance from the regional (marine) base level compared with the other captured extensional basins, and (2) the continued local tectonic subsidence which, in spite of a reduction of the internal catchment during the Quaternary, has managed to attract and keep the local drainage.

The Fucino basin thus provides a striking example of the competition between the regional river downcutting, which tends to capture and integrate the basin into the regional drainage network, and movement on the normal fault system, whose continued activity has managed to promote local tectonic subsidence, accumulation of sediments and preservation of the internal drainage.

7.2.2 The Colfiorito basin

The Colfiorito area is another closed extensional basin system located on the Apennines watershed and surrounded by valleys that are deeply incised into the Mesozoic-Cenozoic bedrock (Fig. 12). The basin floor is at an elevation of ~ 800 m whereas the level of fluvial incision in nearby valleys rapidly decreases to ~ 400 m within a few kilometres of the basin. Lacustrine and alluvial deposits containing remains of Lower Pleistocene mammal fauna (Coltorti et al. 1998) are exposed for ~ 100 m in thickness in the central and southern parts of the basin. The southern part has been already been partially captured by the southward-flowing external drainage (point D in Fig. 12). The area of internal drainage is bounded to the NE by an active normal fault system (Cello et al. 1997) that ruptured during the 1997 Umbria-Marche earthquake sequence (Amato et al. 1998). Despite the different interpretations given to the surface ruptures observed after the main shocks, the fault geometry revealed by seismological, geodetic, and field data (Amato et al. 1998; Stramondo et al. 1999; Basili et al. 1998; Cello et al. 1998b; Cinti et al. 1999) is consistent with the longer-term Quaternary evolution of the area. In particular, radar interferometry and GPS data showed several tens of centimetres of seismic subsidence in the hangingwall of the fault system bounding the basin (Stramondo et al. 1999), and surface ruptures of a few cm in amplitude were found after the earthquake along previously-mapped fault scarps at the foot of mountain fronts bounding the Quaternary basin.

In the NE part of the basin a broad saddle in the mountain front (point A in Fig. 12) is the lowest potential outlet of the internally-drained area and separates the upper catchment of the NE-Adriatic-draining Chienti river from the main Colfiorito basin. On the side facing the basin this saddle is cut by a fault scarp belonging to the main basin-bounding fault system. SAR interferometry and GPS data show that it was the area immediately SW of this saddle that subsided several cm relative to the footwall in the 1997 earthquakes (Stramondo *et al.* 1999). The saddle itself is now a wind gap, which we believe once contained the Chienti river and drained the Colfiorito basin. It was subsequently interrupted and abandoned by the movement on the basin-bounding fault. Exposed basin fill in the central part of the basin near Cesi (Fig. 12) suggests that the drainage has even switched episodically between internal to external in the past, depending on the durability of the fault-related dam.

Faulting in the 1997 earthquakes is thus consistent with the longer-term Quaternary evolution of the Colfiorito basin, in which motion on the basin-bounding fault system has competed with regional incision of the surrounding rivers to control whether the basin is dominated by external or internal drainage. The general setting is similar to the Fucino basin, but whereas the lowest potential outlet of the Fucino basin is in the hanging wall of the basin-bounding fault system, it is in the footwall at Colfiorito. This difference does not appear to be particularly significant as there are other places within the Colfiorito basin, in the hanging wall of the active fault system, where the basin is very close to being captured by the external drainage (points B, C and D in Fig. 12). Whether the lowest outlet is actually in the footwall or hanging wall probably depends only on the relative effectiveness of the regressive erosion by the regional network on either side of the watershed.

Figure 12. (a) Schematic map of the Colfiorito area. Thick barbed lines are major Quaternary faults (modified from Cello *et al.* 1997). The Quaternary basin is in grey. The grey dashed line encloses the area of internal drainage. The lowest potential outlet of the basin (point A) corresponds to a wind gap of the Chienti River along the mountain front formed by the basin-bounding active normal fault. (b) Perspective view of topography.

8 EVOLUTION OF THE DRAINAGE NETWORK

Our interpretation of the interaction between regional uplift, active faulting and drainage evolution is summarized in Fig. 13. The stratigraphy and geomorphology of the intermontane basins in the Apennines show that the Quaternary drainage evolved from an internally-drained system to a through-going river network which frequently cuts through the Mesozoic-Cenozoic ridges that form inter-basin highs. This evolution was accompanied by a change from low-gradient fluvial and lacustrine deposits to steeper-gradient braided fluvial and alluvial fan deposits that are often incised into the earlier sediments. Our interpretation is that these changes were induced by a fall in the regional base level that caused increased incision of rivers, the regressive erosion and headward enlargement of catchments, and the progressive capture and emptying of the internallydrained basins and their sedimentary fills. This evolution was ultimately related to the development of a long-wavelength topographic bulge and regional uplift induced by mantle upwelling beneath the Apennines. In most places today the effects of downcutting by the regional network dominate over the local subsidence caused by slip on the active faults. For this reason, except in some of the higher and more distant parts of the regional network, the more subtle geomorphological effects of fault growth and interaction are usually obscured.

The surviving closed basins of Fucino and Colfiorito allow us to see the interaction between the regional downcutting and normal faulting. The preservation of these two basins on the watershed is related to two factors:

(i) They are both far from the base levels of the streams that are trying to capture them. The wave of regressive erosion triggered by the base level fall has effected these internal regions later than those basins closer to the coasts, such as the Tiber, Sulmona and Campo Imperatore basins. Once captured, the increased discharge through these well-integrated basins makes it likely that fluvial incision rates at their outlets will always dominate over local vertical motions related to faulting.

(ii) The second factor is that normal faulting is most active along the crest of the topographic bulge. Only in this location can faulting efficiently delay the capture of closed basins by causing subsidence that preserves internal drainage. There is even evidence that some recovery of internal drainage is possible after partial loss to the external network, where slip on the basin-bounding fault is rapid enough to produce a dam (at Colfiorito) or sufficient subsidence (at Fucino).

9 DISCUSSION

Our conclusion from the previous sections is that mantle upwelling beneath the central Apennines has been the dominant geodynamical process during the Quaternary, controlling both the geomorphological evolution and the distribution of active deformation. This control has been exerted through the formation of a long-wavelength topographic bulge and the localization of active extension along its crest. The interactions between the regional uplift and active normal faulting are then clearly expressed in the evolution of the drainage network and the Quaternary sedimentation in the intermontane basins.

The Apennines provide an ideal place to observe this type of interaction because of their short distance from the marine base level, so that effects of base-level variations are rapidly propagated upstream, and because the active faulting is localized on top of the long-wavelength bulge. On a catchment scale, we can see that the along-stream distance from the base level can reduce the predominance of regional fluvial downcutting over the localized subsidence related to normal faulting, and is one of the reasons that internally-drained basins are preserved along the main watershed. Thus distance from the base level seems to be a critical factor that allows the geomorphological effects of mantle upwelling to be recognized in places where it accompanies active normal faulting. In places far from the influence of regional base levels, such as parts of the East African Rift, Nevada or Tibet, the effect of any regional uplift on the depositional regimes in the extensional basins is likely to be minor and less clear. Indeed, the morphology of these classic, internally-drained, basin-and-range-style provinces is quite different from, and not always a useful guide to, the morphology in the Apennines, where the hanging-wall basins may be deeply dissected and preserved only as terraces suspended above deep gorges, as in the Subaequan basin in Fig. 10. We suspect this is partly the reason why many active normal faults in the Apennines remained unidentified until relatively recently. Yet the drainage evolution we describe in

Figure 13. General scheme of the interactions between regional mantle-controlled uplift, active faulting and drainage evolution in the central Apennines (see Fig. 4 caption for explanation of the topographic profiles).

the Apennines is likely to have been quite common in the past. In all continental margins that formed by rifting, the internallydrained fault-bounded grabens will eventually be captured by headward erosion from streams draining to the sea, whether or not the rifting was initiated by uplift in the first place. The consequences for stratigraphic development on the margin are therefore important.

Local tectonic subsidence in the intermontane extensional basins provides accommodation space for syn-rift sediments. Whether the fluvial system then remains internal and spatially confined or externally open is an important controlling factor in the subsequent mode of sediment dispersal. In the Apennines, the change from an internally-drained, fluvio-lacustrine system to an open through-going river system in the Quaternary should have corresponded to a marked increase in the sediment supply to the adjacent Adriatic and Tyrrhenian seas. There is some evidence for this in the Adriatic, where well-developed progradational sequences in the Quaternary that mark an increase in sediment flux and the termination of flexural subsidence are coeval with the large-wavelength bulging of the Apennines and the formation of a through-going fluvial network. In detail, the response of sediment flux to tectonic uplift is likely to be nonlinear, because the internally-draining intermontane basins represent stores of already-worked, easily removable sediments. Thus the defeat of a local drainage divide, such as the Popoli gorges in Fig. 7, by the external network is likely to result in pulses of sediment delivered to offshore regions as each internal basin is progressively captured. The competing processes we describe in the Apennines are of more general interest because they may have been common in the geological record. The history of many rifted continental margins may involve internally-draining basins that were originally created above sea level by intracontinental extension but which were then captured by headward-eroding streams cutting down to sealevel before finally being drowned by later subsidence. The capture and emptying of those earlier basins provides a source of re-worked sediments to the offshore marine environment, as we describe in the central Apennines. The influence of these processes in determining the supply and character of potential reservoir rocks on continental margins gives them an importance beyond their regional significance in Italy.

In terms of the deformation processes in the upper crust, the main effect of the regional uplift is to localize the active normal faulting at the crest of the long-wavelength topographic bulge. Only here can the continued activity of basin-bounding faults preserve internally-draining basins from the efficient downcutting by the regional fluvial network on the flanks of the developing bulge.

The significant deviations of the long-wavelength gravity anomalies from the values expected from crustal loading of an elastic plate require the existence of sublithospheric density variations. Since these density contrasts cannot be statically maintained the simplest explanation is that they drive convective flow in the mantle which dynamically support the long wavelength topography. Whether the mantle convection responsible for the support of long-wavelength topography is related to the latest stage of the subduction of the Adriatic–Ionian lithosphere, that has dominated the Neogene history of the Tyrrhenian–Apennines system or from a mantle plume is still unclear. The main point here is that the observed gravity anomalies can only be explained with sublithospheric horizontal density variations that ultimately require a convective

flow that affects surficial processes. The existence of sublithospheric horizontal density variations has been already suggested by Carminati et al. (1999), who imposed a buoyant upwelling asthenosphere beneath the Apennines to reproduce the observed pattern of Plio-Quaternary vertical motions with their finite element model. The type of mantle flow suggested by the positive admittance signature of 50 mGal km⁻¹ is not that expected from the corner flow driven by a subducted slab since this flow should produce subsidence not uplift. Our opinion is thus that the long wavelength topographic bulge is not a direct expression of a subduction-related process such as corner flow (Cavinato & DeCelles 1999), as geological evidence does not provide compelling evidence for a vigorously active slab during the Quaternary. Yet, at the same time, the convectivelysupported topographic bulge is elongated parallel to the former Neogene subduction zone, which may suggest some connection between these processes. The horizontal variations of seismic velocities imaged by tomography beneath the Apennines (Lucente et al. 1999; Di Stefano et al. 1999) and interpreted in terms of remnants of subducted lithospheric slabs and astenospheric upwelling, may correspond to the density variations required by the long wavelength gravity anomalies.

The active extension in the Apennines may now be related more closely to ongoing sub-crustal processes than to the relative motions of the surrounding lithospheric blocks within the Mediterranean collision zone. Thus it may be the extension induced by the mantle-supported uplift that forces the Adriatic lithosphere to move to the NE with respect to Eurasia. Nevertheless, the extension in the Apennines must be accommodated somewhere, and seismicity, focal mechanisms and geodetic data all suggest it is absorbed principally in the Dinaric belts of former Yugoslavia and Albania, with some taken up also in the external part of the Apennines further north (Anderson & Jackson 1987; Ward *et al.* 1994; Frepoli & Amato 1997).

10 CONCLUSIONS

We have attempted to show how topography, gravity, active faulting and geomorphology can be used to provide a unified picture of the Quaternary development and tectonics of central Italy, in which the surface processes are all consequences of upwelling in the mantle. In this picture, mantle upwelling in the Quaternary caused a long-wavelength topographic bulge which localized active normal faulting at its crest to produce internallydraining, fault-bounded, intermontane basins. These basins have subsequently been captured by aggressive headward erosion by rivers draining to the marine base level, emptying the basin fills and delivering the sediment to the offshore regions. The two remaining closed basins are those on the watershed, furthest from the marine base level, where active normal faulting has produced sufficient local subsidence to enable them to survive.

A series of fortuitous circumstances allow these processes to be recognized in central Italy quite easily. The effective elastic thickness is sufficiently small for the relatively limited land area to be large enough to recognize the long-wavelength asymptote in the admittance that is characteristic of mantle convection. The Apennines are close enough to the sea for the uplift to be reflected in the incision and headward regression of the regional drainage network. Evolution of the drainage has been going on long enough for capture of the internal basins by the external drainage to be advanced, yet has not been complete, so that surviving basins on the watershed allow us to study how that process occurs. The fact that the survived closed basins are bounded by active normal faults indicates also the role of local tectonic subsidence in delaying the capture by the external drainage network.

Although the conditions that make these processes obvious in Italy are unusual, the processes themselves are not, nor are those processes likely to be influenced by whether or not they are convenient for us to study. The capture of closed basins by regional drainage is probably very common in the geological record, especially on rifted margins. The consequences for the stratigraphic development of such places are profound, and Italy provides a good modern analogue in which to study them.

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