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Late Neogene and active orogenic uplift in the Central Pontides associated with the North Anatolian Fault: Implications for the northern margin of the Central Anatolian Plateau, Turkey

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[1] Surface uplift at the northern margin of the Central Anatolian Plateau (CAP) is integrally tied to the evolution of the Central Pontides (CP), between the North Anatolian Fault (NAF) and the Black Sea. Our regional morphometric and plate kinematic analyses reveal topographic anomalies, steep channel gradients, and local high relief areas as indicators of ongoing differential surface uplift, which is higher in the western CP compared to the eastern CP and fault-normal components of geodetic slip vectors and the character of tectonic activity of the NAF suggest that stress is accumulated in its broad restraining bend. Seismic reflection and structural field data show evidence for a deep structural detachment horizon responsible for the formation of an actively northward growing orogenic wedge with a positive flower-structure geometry across the CP and the NAF. Taken together, the tectonic, plate kinematic, and geomorphic observations imply that the NAF is the main driving mechanism for wedge tectonics and uplift in the CP. In addition, the NAF Zone defines the boundary between the extensional CAP and the contractional CP. The syntectonic deposits within inverted intermontane basins and deeply incised gorges suggest that the formation of relief, changes in sedimentary dynamics, and >1 km fluvial incision resulted from accelerated uplift starting in the early Pliocene. The Central Pontides thus provide an example of an accretionary wedge with surface-breaking faults that play a critical role in mountain building processes, sedimentary basin development, and ensuing lateral growth of a continental plateau since the end of the Miocene.


1. Introduction

[2] Orogenic plateaus are first-order morphotectonic features of many Cenozoic orogens, including the India-Eurasia collision zone, the central Andes, and the North American Cordillera. Common surface characteristics include subdued interior topography in an arid to semi-arid environment, contrasting with steep, rugged topography, and often humid conditions at their flanks [Allmendinger et al., 1997; Strecker et al., 2007]. Additional characteristics commonly include crustal thickening, spatially disparate late-stage volcanism often associated with extensional tectonism, and a variety of geophysical anomalies, possibly linked to lithospheric-scale processes [e.g., Isacks, 1988; Kay and Kay, 1993; Allmendinger et al., 1997]. Despite many unifying topographic, geologic, geophysical, and regional climatic characteristics, an intense debate continues regarding possible deep-seated processes [Isacks, 1988; Molnar et al., 1993; Kay et al., 1994; Allmendinger et al., 1997; Clark and Royden, 2000; Şengör et al., 2003; Garzione et al., 2006; Schildgen et al., 2007, 2009] and climatically influenced processes [Molnar and England, 1990; Masek et al., 1994; Horton, 1999; Montgomery et al., 2001; Barnes and Ehlers, 2009; Strecker et al., 2007, 2009; Hilley and Coutand, 2009] driving plateau development. For example, structural models [e.g., Allmendinger et al., 1997] explain plateau evolution through consecutive stages of crustal shortening and thickening, while thermomechanical models are based on magmatic underplating, lower crustal flow [e.g., Royden et al., 1997; Clark and Royden, 2000; Shen et al., 2001], and/or mantle delamination [Kay and Kay, 1993; Molnar et al., 1993; Kay et al., 1994; Lamb and Hoke, 1997; Keskin,
2003; Faccenna et al., 2006; Göğüş and Pysklywec, 2008]. Removal of mantle lithosphere and subsequent, isostatically driven wholesale plateau uplift represents one end-member scenario for plateau evolution that could explain the inferred rapid rise of some orogenic plateaus [e.g., Garzione et al., 2006; Molnar and Garzione, 2007]. Other models link the filling of basins and their successive incorporation along the margins of plateaus to arid climatic conditions. [Mašek et al., 1994; Horton, 1999; Sobel et al., 2003; Strecke, 2007, 2009]. In such a scenario, plateau formation and sustenance might be governed by the conspiring effects of localized uplift, low precipitation, low rock erodibility, and low fluvial efficiency. This will cause minor erosion on the leeward flanks of an orogen, leading to the loss of fluvial connectivity, enhanced sediment storage within the orogen, and reduced relief contrasts between basins and ranges. Such processes might eventually cause the incorporation of basins into the plateau realm and piecemeal lateral plateau growth [e.g., Métiñier et al., 1998; Sobel et al., 2003].

[5] How these models or combinations thereof may apply to different stages of plateau evolution may be recorded in the structural and sedimentologic characteristics of plateau margins. A critical test of the different models is to decipher whether the plateau margins have grown successively laterally and vertically or if the plateau resulted from wholesale regional uplift related to mantle processes.

[4] The Central Anatolian Plateau (Figure 1) meets many surface [Erol, 1981] and subsurface characteristics [Keskí, 2003; Şengör et al., 2003; Faccenna et al., 2006; Gans et al., 2009] of other orogenic plateaus, but on a smaller, more accessible scale. Elevations of the interior between 1200 and 1500 m extend over a region of 400 × 750 km, and contrast with higher, deeply incised flanks. Ranges at the plateau flanks have elevations in excess of 2000 m, with the Taurides to the south and the Pontides to the north (Figure 1). The plateau is characterized by crustal thicknesses between 37 and 42 km [Toksöz et al., 2003] and heat flow values of 107 ± 45 mW/m² [ilkılık, 1995], which is compatible with a hot and thinned mantle lithosphere [Toksöz et al., 2003; Gans et al., 2009].

[5] Structurally, the central part of the plateau is in an extensional tectonic regime, characterized by normal faults [Şengör et al., 1985; Dhont et al., 1998a; Genç and Yürür, 2010] and alkaline volcanic activity [Deniel et al., 1998; Dhont et al., 1998b; Şen et al., 2004]. The Central Anatolian Plateau is transitional between the Aegean extensional province to the west and the Bitlis-Zagros continental collision zone to the east (Figure 1) [Şengör et al., 1985]. Those boundaries define the relatively rigid Anatolian plate, which has been extruding westward since the Miocene, as a result of Arabia-Eurasia collision [e.g., McKenzie, 1970; Şengör and Yıldız, 1981; Jackson and McKenzie, 1984; Şengör et al., 1985; McClusky et al., 2000]. Extrusion is accommodated by right-lateral motion along the North Anatolian Fault and left-lateral motion along the East Anatolian Fault [McKenzie, 1970; Şengör and Yıldız, 1981; McClusky et al., 2000].

[6] The Central Anatolian Plateau thus constitutes a premier research target to unravel neotectonic processes associated with orogenic plateau formation and ongoing plateau margin deformation. However, despite its small size, good accessibility, rich fossiliferous carbonates, and datable volcanic units, the uplift history of this region is not well known and possible links to deep-seated processes have remained enigmatic.

[7] In this study, we focus on the northern margin of the Central Anatolian Plateau, which coincides with the central Pontides (CP) between the North Anatolian Fault (NAF) and the Black Sea (Figure 1). We use new morphometric and
structural field data together with GPS-derived velocity data to evaluate the mode, mechanism, and pattern of deformation and tectonics along the northern plateau margin. We present (1) an analysis of the topography and drainage networks in the CP; (2) a characterization of the geometries and kinematics of active tectonic structures; (3) an evaluation of the role of these structures in the regional tectonic framework and uplift history of the plateau; and (4) an evaluation of the role of the NAF in the evolution of the northern plateau margin.

Our data suggest that the restraining force along the NAF is the primary mechanism that activated faults, folds, and differential uplift along the flank of the plateau, ultimately responsible for the formation of a northward-growing orogenic wedge and piecemeal topographic surface uplift of the northern plateau margin.

2. Geologic Setting

The Pontides are a 1100 km long and E-W sinuoidal-trending orogenic belt parallel to the Turkish Black Sea coast. Topography of the belt increases from an average of 150 m in the west to 2000 m in the east (Figure 1). The Pontides are an important orographic barrier for northerly moisture-bearing winds, resulting in a maximum of 1200 mm/yr of precipitation along the northern flanks, while the leeward sectors receive ~400 mm/yr (Figure 2a) [Türkės and Erlat, 2005; Akçar and Schlüchter, 2005]. The Central Pontides (CP) span the northward arched part of the belt between the Izmir-Ankara-Erzincan suture to the south and the Black Sea to the north [Okay et al., 2006] (Figure 1). Here, we primarily focus on the northern part of the CP between the NAF and the Black Sea. This area forms an integral part of the northern margin of the Central Anatolian Plateau, a region which has been oroclinally bent since earliest Paleocene time [Meijers et al., 2010]. This sector of the mountain range is 350 km long and 90–120 km wide with an average elevation of 960 m.

North of the Pontides, the Black Sea oceanic basin developed from back-arc extension associated with the northward subduction of the Tethyan plate during the Mesozoic [Okay et al., 1994; Okay and Tüysüz, 1999]. Currently, the southern Black Sea margin is under compression due to the Africa-Arabia and Eurasia collision (Figure 1) [Şengör and Kidd, 1979; Kocyiğit, 1987; Finetti et al., 1988; Barka, 1996; Barka and Reilinger, 1997; Tüysüz, 1999; Cloetingh et al., 2003]. South of the Pontides, the NAF is the most prominent active member of the dextral North Anatolian Fault Zone (NAFZ) (Figure 1) [Ketin, 1948; Barka, 1996; Şaroğlu et al., 1992; Şengör et al., 2005]. The NAF extends for 1200 km with an overall E-W strike and northward concave arc-shaped geometry parallel to Black Sea coast between the northern Aegean and eastern Turkey. Fault motion was initiated between 13 and 11 Ma in eastern Anatolia following the onset of continental collision between Arabia and Eurasia (Figure 1) [Şengör and Kidd, 1979; Barka and Kadinsky-Cade, 1988], and propagated westward between 8.5 and 5 Ma in the Central Pontides [Hubert-Ferrari et al., 2002], and between 5 and 3 Ma in the Marmara Region [McKenzie, 1972; Dewey and Şengör, 1979; Şengör et al., 1985; Armijo et al., 1999, 2004].

[11] The NAF changes strike along its length to form a broad and asymmetric restraining bend along the middle portion of the fault (Figure 1) [Enure et al., 2009]. The splays of the NAF veer off from the main trunk to the east of this bend and turn southwest toward the interior part of the Central Anatolian Plateau [Şaroğlu et al., 1992; Chorowicz et al., 1999] (Figure 1). In the north, the CP are located at the apex and western continuation of this broad bend, where higher contractional strain is likely to be accumulated.

[12] A sequence of earthquakes with surface ruptures along the NAF has migrated from east to west between 1939 and 1999 [Şaroğlu et al., 1992; Barka, 1996]. The focal mechanism solutions of these earthquakes suggest an anticlockwise rotation of the shortening axes from N-S on the eastern flank and NW-SE on the western flank of the broad bend of the NAF [Barka and Reilinger, 1997; McClusky et al., 2000; Reilinger et al., 2006]. The rupture segments of these earthquakes cut through the CP from east to west (Figure 1b). The present geodetic slip rate of the NAF is 24 ± 1 mm/yr [Reilinger et al., 2006].

[13] Paleoseismological studies reveal that active tectonism of the NAF is characterized by pure dextral strike-slip faulting without any dip-slip component, except within local km-scale bends or stepovers [Okamura et al., 1993; Barka and Kadinsky-Cade, 1988; Hubert-Ferrari et al., 2002; Kondo et al., 2005; Dor et al., 2008; Kozaci et al., 2009; Kondo et al., 2010]. However, Quaternary thrusts to the north and the focal mechanism solution of the Bartın earthquake (03.09.1968; Ms: 6.6) [McKenzie, 1972], offshore of the northwestern CP (Figure 1b) document ongoing shortening in the CP. The Bartın event is the largest recorded earthquake in the northern sector of the study area, with coseismic deformation that uplifted the Black Sea coastline by 0.5 to 0.7 m (Figure 1c) [Wedding, 1969; Ketin and Abdüsselamoglu, 1970; McKenzie, 1972]. The fault plane solutions of the Bartın earthquake and the main shocks of the 26 November 1943 Tosya and 1 February 1944 Bolu-Gerede earthquakes along the NAF have similar horizontal stress-axis orientations, suggesting ongoing NW-SE crustal shortening along the western flank of the CP (Figure 1b) [McKenzie, 1972; Kiratzi, 1993; Yüür, 2003].

The remaining active tectonic structures between the Black Sea and the NAF are the Cide, Balıfakı, Erikkı, Derbent, Ekinveren, and Karabük thrusts, which are mostly inherited structures (Figure 2c) [Tüysüz, 1999, Sunal and Tüysüz, 2002; Şenel, 2002]. The Balıfakı and Erikkı thrusts, south of the Sinop Peninsula (Figure 2c), and Derbent thrust, south of the Kızılırmak Delta dip to the south and strike approximately E-W, parallel to the orocline. The Ekinveren Fault is the principal intermontane active tectonic structure within the CP (Figure 2c). This north dipping, E-W striking thrust fault comprises multiple en échelon segments along the northern margin of the Kastamonu Basin (Figure 2c) [Tüysüz, 1999]. Additional thrust faults are located south of the Kastamonu Basin, along the northern boundary of the Ilgaz Range. Here, south-dipping thrusts may represent a transpressional splay of the NAF to the south (Figure 2c). In the easternmost sector of the study area, within the Vezirköprü Basin, an abandoned dextral splay of the NAF delimits the northern margin of the Vezirköprü Basin (Figure 2c) [Dirik, 1993]. In the western sector of the CP, the Karabük and Cide faults are the most prominent active tectonic structures. The Cide
faults are reverse faults striking parallel to the NE-SW oriented shoreline [Sunal and Tüysüz, 2002; Şenel, 2002], while the Karabük Fault is a reverse fault striking NNESSW along the eastern margin of the Karabük Range (Figure 2c) [Koçyiğit, 1987; Yürür, 2003]. [15] The main morphotectonic units in the study area include the Sinop, Karabük, and Ilgaz ranges, the Devrekani, Karabük, and Kastamonu intermontane basins, and the Gerede, Karabük, Çerkeş, Ilgaz, Tosya, and Vezirköprü basins along the NAF (Figure 2c). The Sinop, Karabük, and Ilgaz ranges form topographic barriers between the Black Sea and the interior parts that have been deeply incised by several rivers. These include the Kızılırmak and Filyos rivers, their major tributaries, as well as the Devrekani and shorter coastal rivers (Figure 2c). The Devrekani and several other coastal rivers have short, steep bedrock channels, with headwater regions within the Sinop Range. In contrast, the Filyos River is sourced from the Çerkeş basin and the Kızılırmak River is sourced from the Central Anatolian plateau, both of which are south of the NAF. The Filyos and Kızılırmak rivers flow through gorges more than 1000 m
deep that traverse the CP orocline and the NAFZ from south to north (Figure 2c).

[16] The CP comprise rocks from the Precambrian to the Quaternary [Tüysüz, 1999, Figure 3], which were amalgamated during the Variscan and Alpidic orogenies before Neogene time. The orogenic structures formed during the closure of the Neotethys and collision of an island arc (Pontides) with the Sakarya continent to the south mainly during late Campanian–Maastrichtian time [Okay and Tüysüz, 1999; Okay et al., 2006]. Following collision, the CP were shortened further, uplifted along a N-vergent detachment [Şenel et al., 1985], and developed a thin-skinned foreland fold-and-thrust-belt during the Paleocene–Eocene [Okay and Tüysüz, 1999; Tüysüz, 1999; Sunal and Tüysüz, 2002]. The basement consists of Triassic to Cretaceous sequences in metamorphic and rare magmatic rocks as well as Eocene flysch units (Figure 3) [Yılmaz et al., 1997; Tüysüz, 1999; Okay et al., 2006].

[17] The Neogene deposits relevant to the post-orogenic and consequent plateau-margin evolution are distributed (1) along the coastal zone at the Kızılrmak Delta and the Sinop Peninsula; (2) in intermontane basins (Karabük, Devrekani, Kastamonu, and Vezirköprü); and (3) in the Çerkeş, Ilgaz, Tosya, Kargi, Havza, and Merzifon basins along the NAF (Figure 3). Irrlitz [1972] defined the Pontus Formation in the basins along the NAF, which is divided into a lower and an upper series. The lower series mostly comprises lacustrine, gypsum, sandstone, siltstone, clays-tone, and marl in the basin center, while the upper series contains fluvial sand, gravel, and colluvium [Irrlitz, 1971, 1972; Barka and Hancock, 1984; Över et al., 1993; Andrieux et al., 1995; Dhont et al., 1998b]. These units are folded [Barka and Hancock, 1984; Andrieux et al., 1995] due to activity of the NAF. The age of the Pontus formation is given as Plio-Pleistocene (circa 4 to 2 Ma) based on biostratigraphic constraints from chariophystes and ostracods [Över et al., 1993] and mammals in the Tosya Basin [Ünay and de Bruijn, 1998]. The age of the volcanic rocks at the bottom of the Pontus Formation is 8.5 Ma [Adıyaman, 2000], which suggests that the folding related to NAF started between 5 and 8.5 Ma in the Tosya Basin. [Hubert-Ferrari et al., 2002].

### 3. Methods

[18] To understand the mode and pattern of active deformation, and possible effects of the NAF in the study area, we integrated morphometric and plate kinematic analyses with structural and geomorphic field observations. The resulting data allow us to assess the role of both recent and cumulative long-term fault activity on deformation of the northern plateau margin. Our regional morphometric analysis employs topographic and precipitation swath profiles and topographic residual analysis to identify regional topographic anomalies, as well as steepness and concavity values of...
longitudinal river profiles that may correlate with ongoing rock uplift [e.g., Snyder et al., 2000; Kirby et al., 2003; Molin et al., 2004].

3.1. Digital Topography
[19] We generated digital topography of the Central Pontides (CP) from 3 arc-sec SRTM digital elevation models projected into the local UTM projection and resampled at 90 m resolution. We used DEMs with 25 m resolution generated from digitized 1:25,000 scale topographic maps to perform basin-scale morphometry, including an analysis of steepness and concavity of river profiles in very narrow and deeply incised gorges, which are not visible in the SRTM data. We manually extracted the channel profile of the Kızılarmak River, which is currently dammed in several locations, by using 1:25,000 scale topographical sheets published in 1959 by the General Command of Mapping (Turkey). The remaining synthetic drainage network was extracted from the SRTM data using standard hydrology tools in ESRI’s ArcGIS to fill sinks and to calculate flow directions and flow accumulations. Using the stream profiler tool (available from http://www.geomorphertools.org/), we extracted longitudinal profiles, steepness values, and concavity values of rivers within the synthetic network. To assess the effects of variable bedrock and local climate patterns on the rivers, we examined geological contacts and formations from the 1:500,000 scale Sinop Quadrangle of Turkey [Senel, 2002] and mean annual precipitation data recorded between 1930 and 2001 by local meteorological stations (courtesy of M. Dirican).

3.2. Swath Profiles
[20] Topographical swath profiles show minimum, mean and maximum elevations across a given width along the length of the profile. In tectonically active areas, they typically display strong relief contrasts, rugged topography, and high relief. In recently incised regions with narrow gorges, the mean elevation curve is close to maximum elevation curve [Masek et al., 1994]. We extracted four swath profiles from SRTM topographic data and from gridded mean annual precipitation data. Three of the swath profiles were extracted orthogonal to the major ranges and one was extracted parallel to the general physiographic trends of the coastal zone along 40-km-wide sectors. Swath profiles of topography and annual precipitation data are plotted together to evaluate possible correlations between topography and precipitation.

3.3. Topographic Residuals and Incision
[21] Topographic residual maps subtract minimum and maximum elevations averaged over a given sampling window and illustrate relief variations in the landscape [Stearns, 1967]. Residual maps are derived from envelope surface maps, consisting of ridges/interfluves, and subenvelope surface maps, consisting of valley bottoms/channels. They depict fluvial incision of the landscape and provide complimentary information to swath profiles. We created a residual map by subtracting minimum and maximum elevations averaged over a sampling window of 4 × 4 km. The 4 km sampling window was chosen because it is large enough to include at least two major ridges and/or valleys along the major ranges within the study area, and therefore shows the characteristic valley-to-ridge relief pattern.

3.4. Longitudinal profiles, channel steepness and concavity
[22] Typical graded longitudinal profiles of bedrock channels are concave-up and smooth under steady tectonic and climatic conditions. Deviations from this equilibrium form may be a consequence of spatial variations in lithology, climate, or sediment flux, transient profile evolution, or non-uniform rock uplift across active structures [Kirby et al., 2003; Wobus et al., 2006, Cyr et al., 2010].

[23] The steepness index (or sometimes “normalized steepness index”) provides a simple matrix to compare slopes in rivers of different size and to detect deviations in longitudinal channel profiles from a graded form. In essence, it is the slope of a channel or channel segment that is normalized to its drainage area. Under steady state conditions, the slope of a channel decreases as a power law function of contributing drainage area [e.g., Hack, 1973; Flint, 1974; Howard and Kerby, 1983]. The local channel slope can be expressed as

\[ S = k_s A^{-\theta} \]

where \( S \) = local channel slope, \( A \) = upstream drainage area, \( k_s \) = channel steepness index and \( \theta \) = channel concavity index. The steepness index has been shown to correlate with rock uplift rate (\( U \)), lithology, and climate [e.g., Merritts and Vincent, 1989; Whipple and Tucker, 1999; Snyder et al., 2000; Kirby and Whipple, 2001; Wobus et al., 2003; Kirby et al., 2003; Duvall et al., 2004; Wobus et al., 2006]. The steepness index is typically calculated using a reference concavity (\( \theta_{ref} \)), because concavity tends to vary little across different landscapes [Snyder et al., 2000; Wobus et al., 2003, 2006]. In this study we used \( \theta_{ref} = 0.45 \), a value used in many other active tectonic settings [Wobus et al., 2003, 2006]. The steepness index and concavity are measured directly by regression analysis of channels from log-log plots of drainage area (\( A \)) and slope (\( S \)) [Montgomery and Buffington, 1997; Snyder et al., 2000; Kirby et al., 2003; Wobus et al., 2006]. We derived steepness and concavity values for all rivers in the study area with a drainage area exceeding 1 km². For display purposes, the threshold drainage area was set at 5 km² threshold. We focus our discussion on the results of four bedrock rivers that are representative of different lithologic and climatic areas along the coastal zone to assess the role of lithology and climate on the regional patterns. We also analyzed longitudinal profiles of the two major orogen-traversing rivers, the Filos and the Kızılarmak.

3.5. Plate Kinematics
[24] To better identify possible effects of the NAF on the structural development of CP, we analyzed previously published regional GPS-derived velocity data to determine components of slip parallel to and normal to the NAF. Unfortunately, geological studies have only provided estimates of fault-parallel strike-slip rates along the NAF, possibly due to the lack of preserved vertical offset markers on time scales of 10^7-10^8 years [Saroğlu et al., 1992; Barka, 1996]. We thus rely on an alternative approach,
which involves estimating the slip vector along the NAF due to the rotation of Anatolia about an Euler pole with respect to a fixed Eurasian reference frame. This provides us with the rotation of Anatolia with respect to the CP across the NAF. Any stress acting on structures across the CP is most likely generated by westward motion of the Anatolian microplate along the NAF, as no significant movement of Eurasia has been documented in the region [Tari et al., 2000; McClusky et al., 2000; Reilinger et al., 2006].

Several authors have estimated the Euler pole position and rotation rate between these two plates using different methods [McClusky et al., 2000; Jackson and McKenzie, 1984; Taymaz et al., 1991]. We use the pole reconstruction of Reilinger et al. [2006] derived from velocity vectors obtained from GPS measurements distributed over a wide region. We calculated the horizontal plate-motion slip vectors along the NAF and, by using the local fault strike, decomposed the absolute vector into fault-normal and fault-parallel components. Only errors in Euler pole position and rates of rotation provided by Reilinger et al. [2006] have been propagated into the calculation, although we also consider uncertainties in local fault orientations, which have been integrated from detailed structural mapping. This analysis can reveal the divergence of slip vectors along the apex of the northward-convex arc of the NAF, and in turn can be important for understanding local changes in the tectonic stress field, as well as the potential response of faults with respect to the regional stress orientation and distribution. The errors involved in this calculation were found to be constant and below 5% and are thus not included in the results.

3.6. Structural Analysis

Our structural observations are based on field work in the Sinop Peninsula and the Kastamonu Basin, where deformed Neogene and Quaternary deposits are well exposed. In the Sinop area we measured bedding plane orientations at 12 sites within Neogene marine strata and plotted them using the program Stereonet Cylindrical Best Fit (R. Allmendinger, Stereonet, 2002, available at http://www.geo.cornell.edu/geology/faculty/RWA/programs.html). From the Kastamonu area we measured meso- and macroscale faults at 7 sites and plotted bedding orientations from 18 sites within Neogene continental strata. The bedding plane orientations allow for a determination of shortening and extension directions using the FaultKin Linked Bingham distribution (R. Allmendinger, FaultKinWin, 2001, available at http://www.geo.cornell.edu/geology/faculty/RWA/programs.html). Collected fault slip data included strike and dip of the fault, and trend and plunge of the slickenlines. Geomorphological field observations focused on the identification of uplifted marine deposits and terraces, deformed colluvial deposits, paleosoils, fluvial terraces, deeply incised gorges, and deflected stream channels.

4. Results

4.1. Swath Profiles

Topography is a first-order manifestation of surface and sub-surface processes, and topographic signals are fundamental to understanding the morphotectonic development of a region. Topographic swath profiles orthogonal to general trend of the northern margin demonstrate strong relief contrasts, rugged topography, and high relief. They also show northward tilting of the top surfaces on the hanging wall blocks of the Karabük, Ekinveren, and Erikli faults, which delimit the major ranges within the study area (Figures 4b, 4c, 4d, and 4e). The highest relief contrast is observed within the Ilgaz Range in the southernmost part of the area, located immediately west of the northward bend in the NAF (Figures 4c). Topographic swath profiles generally show a marked cross-range asymmetry, with steeper southern slopes compared to northern slopes, except for the central part of the Sinop Range. In this zone, where the southern slopes are delimited by the western segments of the Ekinveren Fault, the northern slopes are steeper than southern slopes. This pattern persists despite the mean annual precipitation on the southern slopes being less than one half of the mean annual precipitation on the northern slopes (Figures 2a and 4c).

The parallel swath profile along the crest of the Sinop Range reveals a splayed pyramidal top-surface topography (Figure 4e). The highest peaks are located in the central part of the profile and descend toward west and east as low-relief (<300 m) erosional remnants of a relict upland-surface. The mean and maximum elevation curves are close along the most of the profiles across the Karabük, Sinop and Ilgaz ranges, indicating recently incised regions (Figure 4).

The swath profiles of mean annual precipitation also reveal a cross-range asymmetry, generally showing higher precipitation in coastal regions compared to the interior parts of the CP (Figure 2a). Precipitation averages 1000–1200 mm/yr in the northwestern coastal regions and drops to 500 mm/yr toward the high-relief interior. This decrease is less pronounced along the profile across the northeastern part of study area, between the Sinop Peninsula and the NAF (Figures 2a and 4d). In this relatively low-relief region, coastal precipitation of 700 mm/yr drops to 500 mm/yr in the interior. The highest precipitation variations are located central part of the Sinop Range where the highest peaks of the range are also located (Figure 4e). These data demonstrate that precipitation is orographically controlled, especially along the northern and northwestern flanks of the northern margin.

4.2. Topographic Residuals and Incision

The sub-envelope, envelope, and residual maps of the study area show the relationship between relief and tectonic structures. The sub-envelope surfaces are base levels of the fluvial network within the study area. Their elevations vary between 0 to 1800 m (Figure 5a). The low base levels extend in a narrow zone along the coast and farther inland along the gorges of the Filyos and Kızılrmak rivers. The base levels along the coastal zone increase over a short distance from 0 to 900 m in the Sinop Range. High accurate base levels extend across the Sinop Range at 900 to 1100 m and across the Ilgaz Range at 1400 to 1600 m (Figure 5a).

The envelope surfaces are ridge levels of the fluvial network. They vary between 0 and 2500 m (Figure 5b). The highest surfaces extend over the Ilgaz Range along the NAF. The other high surfaces extend over the Karabük and central part of the Sinop ranges. High accurate summit surfaces
are revealed at 1600–1800 m across the Ilgaz Range, 1200–1400 m across the Karabük Range, and 1000–1200 m across the Sinop Range (Figure 5b). The relatively low-relief, accordant surfaces appear to be indicators of an eroded and deformed relict landscape. The envelope surfaces also highlight escarpments coincident with faults along the range boundaries (Figure 5b).

Residual surfaces derived from sub-envelope and envelope surfaces demonstrate strong relief/incision contrasts between ranges and intermontane basins. Residuals vary between 0 and 1600 m across the ranges and between 0 and 300 m across the intermontane basins (Figure 5c). The Karabük and Devrekani basins, near the western half of the Kastamonu Basin, are low-relief areas similar to basins located in the central part of the plateau (e.g., Çankırı Basin). In contrast, the Ilgaz Range and Karabük ranges and the west and central parts of the Sinop Range are high-relief areas surrounding the basins (Figure 5c). In tectonically active areas higher topographic residuals correspond with active stream incision in response to higher bedrock uplift (e.g., Bürgmann et al., 1994; Blythe et al., 2000; Molin et al., 2004; Gomez et al., 2006; Wobus et al., 2006).

4.3. Longitudinal Profiles, Normalized Steepness and Concavity

The Filyos and Kızılirmak rivers integrate intermontane basins throughout the CP as they traverse the plateau margin from south of the NAF to the Black Sea (Figure 2c). The longitudinal profile of the Filyos River has a smooth, convex-up shape between the Gerede Basin and the Black Sea (Figure 6a). The Gerede tributary of the river is disturbed where it is ∼70 km offset by the NAF (Figure 2c), close to its headwater area, and then flows toward the Karabük Basin with increasing channel gradients (Figure 6a). The composite plot of the tributaries also demonstrates deviations from the concave-up, smooth graded channel profiles within the Filyos Basin.

Figure 4. (a) Swath profiles of topography and precipitation across the field area showing a shaded relief map of study area and locations of the 40 km wide swath profiles. Letters in the swath bands indicate swath profiles in Figures 4b, 4c, 4d and 4e. (b, c, and d) Swath profiles perpendicular to general trend of the margin and (e) a swath profile parallel to general trend of the coastal zone. In each swath profile, minimum and maximum values are indicated by the upper and lower bounds of the colored regions, while the black line in the center of the topographic swath profile indicates the mean elevation.
These deviations indicate knickpoints at 800 m elevation along the Bolu, 900 m along the Araç, and 1200 m along the Gerede tributaries of the Filyos River (Figure 6a). Those are in the Bolu and Gerede tributaries are related to active strand of the NAF.

Further east, the shape of the Kızılrmak River longitudinal profile is nearly flat, with only a small decrease in elevation with distance (Figure 6b). This minimum change in slope is likely related to the minor changes in drainage area, as we analyze only the lower 350 km of the total 1350 km channel length of the Kızılrmak. Along this lower portion of the channel, the relief is only 500 m, while the relief of the full channel length is much higher. The river is ~30 km offset by the NAF to the east of the Tosya Basin (Figure 6c) and then flows toward the Black Sea with increasing channel gradients (Figure 6b). The main channel of the Kızılrmak River and its tributaries (Devrez and Gökmak rivers) have knickpoints along their courses. These knickpoints are more evident along the tributaries that flow parallel to contractional and inverted basins of the study area (Figure 6c).

The map of the channel steepness demonstrates marked variations between coastal and interior parts of the CP (Figure 7). The high-gradient channels are located on the coastal (northern) topographic fronts of the ranges (except southern topographic fronts of the Sinop Range along the Ekinveren Fault) and reveal distinct variations between hanging and footwall blocks of the faults. The low-gradient channels are located in the Devrekani, Karabük, and Kastamonu basins in the interior of the CP. The networks of the low-gradient channels within the Devrekani Basin at about 900–1000 m elevations imply a perched character of the basin above deeply incised gorges (Figure 7).

In order to assess other potential factors affecting channel gradients besides inferred uplift, such as lithology and precipitation, we compare the spatial distribution of steepness values with bedrock maps and annual precipitation data (Figures 7b and 7c). Some of the higher gradients along channels coincide directly with lithologic contacts (e.g., the...
knickpoint at ~750 m elevation in Figure 8a). However, some of the contacts are both lithologic and fault contacts (e.g., along the Erikli fault (EF) in Figure 7b, and the upstream Cide faults in Figure 8a). We compare steepness and concavity values together with lithologic variations of the four characteristic rivers along the coastal zone to show deviations of gradients related to lithology and to the trend of the coastal zone (Figure 8). In the western part of the coastal zone, basin 2 and basin 3 (Figure 2c) are characterized by step-like knickzones coincident with Upper Jurassic-Lower Cretaceous neritic limestones and Upper Triassic-Lower Jurassic metamorphic rocks (Figures 8a and 8b) and also coincident with the Cide faults in the downstream portion of basin 2. Those segments of the streams are characterized by very high steepness values. Steepness values along those segments are typically an order of magnitude higher than steepness values of the headwater areas (Figures 8a and 8b). The concavity values are also positive and very high, ranging from 3.3 to 5.0. Concavity values significantly higher than 0.3 to 1.2 indicate transient conditions [Tucker and Whipple, 2002; Whipple, 2004]. In contrast, very low steepness and concavity values in the headwater areas of the basins indicate the existence of a relict landscape, such as at the western part of the Sinop Range. Toward the eastern part of the study area, the lithologies along the channels of basins 4 and 5 are broadly similar (Figures 8c and 8d). There are knickzones on the hanging wall blocks of the Erikli and Derbent faults, but they are not as dramatic as the knickpoints along basins 2 and 3 and they have lower steepness values (Figures 8a and 8b). The steepness values of the channel segments are still relatively high in basins 4 and 5 (Figures 8c and 8d) and are more than twice as high than the headwater areas. In summary, despite strong lithologic control on channel steepness in some streams, the lithologic variations do not fully correspond with steepness variations, and thus other factors must explain the pattern of steepness values.

[37] Precipitation can also influence river gradients [Roe et al., 2002]. Greater precipitation increases erosivity and is typically accompanied by a reduction in channel gradients, because rivers with higher discharge are more effective at balancing rock uplift [e.g., Whipple and Tucker, 1999; Roe et al., 2002; Kirby et al., 2003; Whipple, 2009]. Mean annual precipitation along the coastal zone increases from 600 mm/yr in the east to 1200 mm/yr in the west (Figure 2a). However, the steepness values of the channels increase from the eastern to western part of the area despite

Figure 6. Longitudinal profiles of the two orogen traversing rivers and their major tributaries. See Figure 2 for locations. The Filyos River is basin 1 and Kızılirmak River is basin 7 6. Arrows indicate offset reaches of the rivers by North Anatolian Fault. NAFZ: North Anatolian Fault Zone.

Figure 7. (a) Active faults and channel steepness index values for major rivers and their tributaries larger than 5 km² overlain on shaded relief. (b) Channel steepness values and active faults overlain on bedrock map of the study area. Explanation for lithologic units and symbols can be found in Figure 3. (c) Channel steepness values overlain on precipitation map of the study area. In each panel, heavy black lines indicate major, characteristic drainage basins in the study area, which are numbered at their outlets. AF: Aracı Fault; BF: Bahfakı Fault; CF: Cide Fault; EF: Erikli Fault; EkF: Ekinveren Fault; KF: Karabük Fault; KyF: Kaya Fault; NAF: North Anatolian Fault.
Figure 7
Figure 8. Longitudinal topographic and geologic profiles with inset slope-area plots for four major, characteristic channels in the study area. Linear regressions of slope-area data are shown as blue lines, while regression with fixed concavity ($\theta_{ref} = 0.45$) are shown as cyan lines. Arrows above the longitudinal profiles indicate the segment of the channels represented by the regression. Locations shown on Figure 7 as follows: (a) basin 2, (b) basin 3, (c) basin 4, and (d) basin 5.
the twofold increase in precipitation (Figure 7c). The similar pattern of steeper channels and greater precipitation is also observed in the western flanks of the northern margin. The Bartın Basin and the western rims of the Karabük Range are exposed to similarly high precipitation, but there is still an increase of steepness immediately across the boundary between the two areas (Figure 7c). These mismatches between precipitation patterns and expected channel gradients suggest that precipitation has only a minor impact on channel steepness values, and that the observed steepness contrasts may be even more pronounced if the effects of variable precipitation could be removed.

Although sea level variations could play important role in the influencing channel forms, we expect such effects to induce similar responses in all of the coastal rivers. Given the wide range in channel forms, variable number of knickpoints, and variable elevations of knickpoints along the channels, a uniform forcing from changing sea level seems unlikely to explain the pattern of steepness values.

Overall, precipitation and sea level variations appear to have only minor influences on channel steepness and concavity values. In contrast, lithology and surface-uplift rate remain viable alternatives for explaining the variable form of river profiles throughout the region.

4.4. Plate Kinematics

Our regional analysis of available GPS-derived velocity data reveals significant variations of fault-parallel and fault-normal slip rates along the northward-convex restraining bend of the NAF, at the southern flank of the CP (Figures 9a, 9b, and 9c). Slip vectors diverge from the fault strike at the apex of the northward convex arc of the NAF (Figure 9a). The vectors are directed toward the west of the CP, at approximately 35°E longitude. This decreases the NAF-parallel slip rate from 24 to 20 mm/yr, and results in an increase of NAF-normal motion of up to 8 mm/yr at the apex of the bend between 33°E and 34°E longitude (Figure 9b). Interestingly, the highest point in the CP (Ilgaz), the highest point of the Sinop Range, and the narrowest portion of the Kastamonu Basin are longitudinally aligned with the highest components of NAF-normal motion (Figure 9a). Along the whole western flank, positive fault-normal rates demonstrate an important compressional component in the CP associated with the broad bend of the NAF.

5. Tectonic Implications

Although there are local lithologic and climatic effects, the strong relief contrasts, rugged topography, associated with deeply incised regions, and steep channel gradients demonstrate the immediate effects of rock uplift associated with active structures. Our plate kinematic analysis demonstrates that this rock uplift could be a result of contractional strain associated with the NAF. In order to elucidate geomorphic and structural results of this strain, we carried out additional studies in the coastal zone and the intermontane basins of the CP.

5.1. Deformation in the Coastal Zone

We combine our onshore observations with offshore seismic profiles from the Bartın area, Sinop Peninsula, and Kızılırmak Delta along the coastal zone related to extract regional active tectonic information. The Bartın area is the only place along the southern Black Sea coast where coseismic uplift has been reported [Wedding, 1969; Ketin and Abdüsselamoğlu, 1970; McKenzie, 1972]. The Sinop Peninsula is one of the few places along the Black Sea coast where Miocene marine sediments are preserved, and where flights of Quaternary coastal abrasion platforms have been eroded into folded Miocene and Quaternary deposits. Farther east, different uplifted levels of deposits related to the Kızılırmak Delta document recent tectonic activity along the coastal zone.

During the M6.6 earthquake coseismic deformation uplifted the Black Sea shoreline up to 0.7 m [Wedding, 1969; Ketin and Abdüsselamoğlu, 1970; McKenzie, 1972]. The location and focal mechanism solutions of the earthquake indicate a 4 km deep offshore hypocenter about 6 km away from the shoreline (Figure 2b) [Alptekin et al., 1986]. Offshore seismic profiles north of the Bartın Basin reveal imbricate thrust faults that deform Plio-Quaternary continental slope marine deposits (Figure 10a) [Damci et al., 2004]. The location and geometry of the imbricate thrust system appear to correlate with the location and focal mechanism solution of the Bartın earthquake, and also with suggested inherited onshore imbricate thrust systems [Sunal and Tüysüz, 2002] propagating toward the north on a low-angle detachment.

Toward the east, an offshore seismic profile located north of the Sinop Peninsula also shows imbricate thrusts on a low-angle detachment surface that Finetti et al. (1988) interpreted as deforming abyssal-plain sediments since the late Miocene (Figure 10b). The pattern of deformation in the overlying sediments indicates that thrusts get younger toward the south, indicating backstepping deformation toward the Sinop Peninsula. The onshore seismic profiles across the Sinop Peninsula and Sinop Range reveal thrust imbricates similar to the offshore structures (Figures 10c and 10d).

Deformation of late Cenozoic deposits on land furnish additional information concerning the nature and timing of regional deformation. The Miocene marine sediments of the Sinop Peninsula (Figures 3 and 11) include Langhian to Tortonian (early middle Miocene) platform limestones, sand-, silt-, and mudstones [Özsayar, 1977; Gedik and Korkmaz, 1984]. These units crop out at up to 200 m above sea level (asl) in the Sinop Peninsula. They are gently folded into an open, E-W oriented syncline (Figure 11) and unconformably overlie the pre-Neogene basement. This suggests general N-S directed regional shortening during post-Tortonian regional deformation. The top of these Neogene deposits is truncated by an erosional surface of unknown age indicating the termination of this folding episode.

The post-Tortonian deformation of the Sinop peninsula may be related to the Erikli and Balıfk file faults, which are the major structures within the peninsula (Figure 11). The Erikli Fault defines a prominent, straight mountain front associated with higher steepness values (from 60 to 90 to 90–120) along the rivers within the hanging wall block of the fault, suggesting active deformation related to activity along this structure (Figure 11). Unfortunately, the scarcity of Quaternary deposits along this structure limits evaluation of its very recent activity. The Balıfk file Fault is located near the southern boundary of the Neogene deposits in the Sinop
Peninsula (Figure 11). Although undeformed fluvial terraces of late Pleistocene and Holocene age are observed along the trace of this structure, the dip of bedding of Neogene deposits in the footwall of the Balıfakı Fault increases toward the fault. Despite these indications of reduced tectonic activity, marine terraces at elevations between 4 and 60 m asl document ongoing uplift of the peninsula and the coastal zone of the CP (Figures 11a, 11b, 11c, 11d, and 11e). This requires the existence of an active offshore structure accommodating uplift, which may indicate a northward migration of contractional deformation with respect to the Sinop Range. 

The Kızılırmak is the longest river in Turkey, extending over 1350 km from eastern Anatolia and crossing the CAP to the Black Sea (Figure 1). At the Black Sea coast it forms a large delta, which provides an archive for eustatic sea level fluctuations and tectonic uplift since the Pliocene [Akkan, 1970]. The oldest fluvial deposits within the delta

**Figure 9.** (a) Slip vectors along the North Anatolian Fault (NAF) arising from rotation of the Anatolian microplate with respect to a fixed Eurasian reference frame. The rectangle indicates the boundary of the study area. Heavy dashed line indicates the boundary between contraction (to west, indicated with white arrow) and extension (to east). Dots denote center points of Quaternary faults in the Central Pontides, Colors are explained in Figure 9c. (b) Variation of fault normal and fault parallel slip rates along the North Anatolian Fault. Positive values indicate contraction, negative values indicate extension. Balıfakı, Eriklı, Ekinveren, Karabük and Kayı faults have positive values (compression). (c) Variation in NAF azimuth and location of faults in the Central Pontides.
contain silt and sandstones covered by conglomerate, inferred to be Pliocene in age [İnandık, 1956; Akkan, 1970]. The more recent late Pleistocene–Holocene delta evolution of the Kızılirmak is reflected by the delta geometry with four different morphologic surfaces, including (1) the active delta; (2) a paleo-delta at 30–35 m asl; (3) an intermediate delta surface at 90–100 m asl, and (4) an upper abrasion platform at 140–150 m asl [Akkan, 1970]. Erinç and İnandık [1955], Akkan [1975], and Karabıyıkkoğlu [1984] in the Sinop Peninsula and Bilgin [1963] in the Ünye area near east of the Kızılirmak Delta identified marine terraces at ~8–10 and ~25–30 m include fossil assemblages characteristic of the Karangat stage of the Black Sea chronology. The Karangat stage is equivalent of OIS5 [e.g., Tschepalyga, 1997]. For these reasons we believe that marine terraces in the Sinop Peninsula and levels of the Kızılirmak Delta are of Pleistocene age.

5.2. Deformation in Intermontane Basins

[...] The Kastamonu Basin (Figures 2c and 3) records quasi-continuous and long-term deformation of the northernmost continental sedimentary succession with evaporites and fluvial sediments yielding direct links to the central parts of the plateau. The basin also hosts structures related to the onset of recent tectonic activity of the Pontides. For these reasons, we focused our field observations of deformation affecting the Kastamonu and Karabük basins.

Figure 10. (a) Offshore seismic profile located north of Bartın [Dumç et al., 2004]; location is shown with dashed line in Figure 2c. (b) Offshore seismic profile (SP 1-578) located north of the Sinop Peninsula [Finetti et al., 1988] showing tectonic inversion out of sequence thrusting in Neogene deposits. Location of line is shown in inset. (c) Onshore seismic profiles [Aydın et al., 1995] crossing the Kastamonu Basin and Sinop Peninsula (see Figure 2c for location) displaying a thin-skinned thrust belt which has a bivergent subsurface geometry. (d) Onshore seismic profile across the southern margin of the Sinop Range (see Figure 2c for location).
Figure 11
The Ekinveren Fault at the northern margin of the Kastamonu Basin is the principal structure bounding the mountain front of the Sinop Range (Figure 2c). A seismic profile across the eastern segment of the fault demonstrates that imbricated faults and folds have a south-vergent geometry (Figure 10c). Our geomorphic and structural field observations indicate recent activity of these faults. The unidirectional deflections of the streams toward the west on undeformed pediment surfaces and wind gaps over anticlines record active tectonic activity along splays of the Ekinveren Fault into the Kastamonu Basin (Figure 12). In addition, we identified south- and north-vergent thrust faults in the center of the Kastamonu Basin displacing gravels (Figure 13a), paleosoils (Figure 13b), Pleistocene terraces of the Gökirmak River, (Figures 13c, 13d, and 13e) reverse-fault bounded basement uplifts, and two anticlines in the eastern part of the basin. This suggests that these structures are active conjugate faults of the Ekinveren Fault.

Even more definitive evidence for surface uplift can be derived from the Devrekani Basin, which is a perched basin north of the Ekinveren Fault (Figure 12). In addition, we identified south- and north-vergent thrust faults in the center of the Kastamonu Basin displacing gravels (Figure 13a), paleosoils (Figure 13b), Pleistocene terraces of the Gökirmak River, (Figures 13c, 13d, and 13e) reverse-fault bounded basement uplifts, and two anticlines in the eastern part of the basin. This suggests that these structures are active conjugate faults of the Ekinveren Fault.

In summary, the structural data from the Neogene folding of Sinop Peninsula suggests a NNE-SSW shortening orientation (Figure 15a). However, the shortening orientation within the Kastamonu Basin deduced from the Neogene at 1040 m elevation give indisputable evidence for 1040 m of surface uplift of the basin since early to middle Miocene time.

In the western part of the Kastamonu Basin, the Karabük Fault is one of the prominent onshore thrust faults in the CP delimiting the eastern front of the Karabük Range (Figure 2c). The fault was initiated during the Lutetian [Kocyiğit, 1987], but its more recent activity is not well known. Yürür [2003] suggested probable tectonic activity associated with the 1944 Bolu-Gerede earthquake. Our observations corroborate this idea, as steepness values immediately increase on the hanging wall of the Karabük Fault (Figure 7a). Tectonic activity of these structures is furthermore supported by westward tilting of the relict upland surface on the hanging wall block of the fault (Figures 4b), and a series of hanging valleys and fluvial terraces at six different elevations within the Karabük Range (Figures 14a and 14b).

In summary, the structural data from the Neogene folding of Sinop Peninsula suggests a NNE-SSW shortening orientation (Figure 15a). However, the shortening orientation within the Kastamonu Basin deduced from the Neogene...
Figure 13. Field evidence for ongoing shortening in the Central Pontides. (a and b) Deformed Quaternary colluvium and gravels by the Ekinveren Fault in the north of Kastamonu. (c and d) Deformed Pleistocene fluvial terraces to the east (Figure 13c) and north (Figure 13d) of the Kastamonu. (e) Slickenlines showing pure thrust faulting of the fault in Figure 13d.
faults and folds is predominantly NW-SE (Figures 15b and 15c), which is similar to the orientation of regional shortening associated with the NAF. This difference between Sinop Peninsula and Kastamonu Basin may result from variation of NAF-normal component of strain along the restraining arc of the NAF (Figures 9a and 9b). Sinop Peninsula is located at the apex, where contraction is predicted to be NNE-SSW oriented, and Kastamonu Basin is located western flank of the bend, where contraction is predicted to be NW-SE oriented. The geomorphic and structural observations along the offshore and onshore coastal zone support an interpretation of uplift induced by contractional deformation within the study area.

6. Discussion

Our new data from structural observations, morphometric analysis, and the interpretation of previously published GPS-derived velocity data document how the regional evolving tectonic and topographic setting along the northern margin of the Central Anatolian Plateau is closely related to the activity of the restraining bend of the NAF and the spatiotemporal evolution of a northward-advancing orogenic wedge. The results of our investigations document that the landscapes of the northern margin of the plateau are actively deforming and uplifting. Active surface uplift without crustal processes could be attributed to an isostatic response to erosional unloading or to crustal thickening and flow within the lower crust [Molnar and England, 1990; Royden et al., 1997]. However, there is a general correspondence among topography, relief, river channel steepness, and active contractional structures within the study area. Furthermore, faults directly control the distribution of the mountain ranges in the Central Pontides. Taken together, the observations indicate that shortening and crustal processes induced surface and rock uplift between the NAF and the abyssal plain of the Black Sea.

6.1. Differential Uplift Within the Central Pontides

Our morphometry shows a low-relief (<300 m) relict upland surface (Figure 5c) across the northern margin suggesting regional surface uplift in the study area. The range-normal and range-parallel swath profiles (Figures 4b, 4c, 4d, and 4e) highlight regional topographic asymmetries as well as eastward and northward tilting of the relict upland surface resulting from spatial variations of surface uplift. Although the amount of surface uplift is still enigmatic, the 1400 m and 1100 m depth of the range-crossing gorges of the Filyos and Kızılırmak rivers provide a maximum value of incision in response to regional surface uplift (Figure 5). A minimum value of 1040 m is based on the uplifted early to middle Miocene marine sediments in the Devrekani Basin. The longitudinal profiles and steepness analysis of the river channels (Figures 6 and 9) are consistent with our topographic observations, and reveal the fluvial response to ongoing differential surface uplift. The western region of the CP (west of Sinop Peninsula) is characterized by transient topography. This region has higher relief and steeper channel gradients compared to the eastern region, despite receiving greater precipitation (Figures 2a, 4b, 4c, and 5). All observations are consistent with an interpretation of greater active tectonic activity and surface uplift of the western sector of the CP compared to the eastern sector. Considering the pattern of highest NAF-normal slip components bounding the western sector of the CP (Figure 11), the higher topographic relief and steeper channels appear to be regional geomorphic manifestations of strain concentration in the western CP due to the restraining bend geometry of the NAF.
6.2. The Role of the NAF in the Development of the Northern Plateau Margin

[55] The broad asymmetric bend of the NAF forms a regional-scale restraining bend along the southern boundary of the CP (Figure 1), which we consider to be an integral morphotectonic entity of the CAP. Activity along the NAF, tectonic activity in the CP, and lateral plateau growth are clearly related. The asymmetry and kinematics of the restraining bend give rise to a concentration of contractional strain, predominantly at the western flank of the broad bend. This is confirmed by the projection of the absolute geodetic slip vectors that show the highest magnitude of NAF-normal motion distributed in the western part of the CP (Figures 11a and 11b).

[56] The concentration of strain in the CP apparently resulted in the reactivation of inherited structures and formation of younger structures since late Neogene time. The inherited structures were reactivated in the same sense with respect to their former kinematics if they were compatible with the NW-SE orientation of the greatest horizontal stress in the CP, which is compatible with dextral simple shear of the NAF. This is demonstrated by the younger thrust faults and folds, and the focal mechanism of the 1968 Bartin earthquake, all of which have similar inferred principal stress axes (Figure 2b). The different morphometric and geologic indicators of recent surface uplift and geomorphic characteristics support this notion of activity along the NAF inducing deformation and surface uplift in the CP.

6.3. Mode and Mechanism of Uplift

[57] Along the coast, individual mountain ranges and intermontane basins within the CP are parallel to the northward-convex bend geometry of the NAF (Figure 2c). There is a strong topographic north-south asymmetry of mountain ranges in the orocline (Figures 4b, 4c, and 4d). For example, the southern flanks of the Ilgaz, Karabük and Sinop ranges are significantly steeper compared to the northern flanks (Figures 4b, 4c, and 4d). This asymmetric topographic pattern, coupled with the spatial distribution and geometry of faults, suggests that the CP constitute an orogenic wedge. Indeed, topographic characteristics, structural field observations, and seismic reflection profiles across the Balişfaki, Eriklı and Ekinveren faults, as well as the geometry of offshore faults (Figure 10) [Finetti et al., 1988; Aydin et al., 1995] reveal a bivergent wedge geometry associated with a shallow detachment (Figure 10). All these observations are compatible with a positive flower structure, which we infer to be rooted in the NAF.

6.4. A Conceptual Model for Deformation and Uplift of the Central Pontides

[58] Our morphometric and plate kinematic analyses document the fundamental influence of the NAF on the evolution of CP and the northern margin of the Central Anatolian Plateau. The NAF is instrumental in inducing regional deformation and decouples the northern plateau margin from the interior part of the plateau. Consequently, strain is broadly partitioned between the NAF and the abyssal plain of the Black Sea. In this actively deforming landscape, strain is accommodated by thrust faults, which are mostly inherited structures reactivated during the Cenozoic. These structures accommodate differential surface uplift and create asymmetric topography across the CP.

[59] Although topographic boundaries are controlled by active thrust faults, geomorphic and structural field observations indicate that no single fault has accommodated large magnitudes of upper crustal shortening since the Miocene. Our structural and geomorphic field observations, combined with onshore and offshore seismic reflection data, suggest that the northern margin of the plateau constitutes an active accretionary orogenic wedge with northward polarity between the NAF and the abyssal plain of the Black Sea (Figure 16). Two principal observations support this view: (1) the topographic asymmetry is compatible with the characteristics of landscapes in pro-and retro-wedges, respectively [e.g., Willet, 1999; Hoth et al., 2007]; and (2) the thrusts and reverse faults rooted in a basal detachment, which is imaged in seismic reflection profiles and deduced from field observations, clearly document the bivergent character of the Sinop Range in the CP and the NAFZ geometry (Figure 10).

[60] All our geological, structural, and morphometric observations document that the transpressional deformation that we relate to the NAF is distributed in the CP and extends to the abyssal plain of the Black Sea as a response to regional strain partitioning. On a regional scale, the geometry of the CP resembles a positive flower structure with a thin-
Acknowledgments.

Aras with 25. This work was carried out as an individual A conceptual model for the Central Pontides, which demonstrate an active orogenic wedge TC5005 Fault, EF: Erikli Fault, EkF: Ekinveren Fault, KyF: Kayı Fault, et al. [2010], the northern margin of the CAP is clearly related to positive flower structure developed over a shallow detachment surface linked to North Anatolian Fault at depth. Structures south of the NAF are based on interpretations from Kaymakçı et al. [2010]. Yellow areas are schematic representations of deformed Neogene deposits in the Sinop Peninsula and intermontane basins. BF: Bafıfakı Fault, EF: Erikli Fault, EkF: Ekinveren Fault, KyF: Kayı Fault, NAF: North Anatolian Fault. Vertical exaggeration = 7.

starting in the late Miocene to early Pliocene. The onset of dextral crustal shearing along the NAF was critical for the development of the northern plateau margin, controlling tectonic inversion and contractional deformation of CAP Neogene basins. Rather than wholesale plateau uplift, our observations suggest piecemeal development of the Central Anatolian Plateau. Our tectonic concept implies that the North Anatolian Fault is the main driving mechanism for wedge tectonics and uplift in the Central Pontides, and the North Anatolian Fault Zone defines the boundary between the extensional Central Anatolian Plateau and the contractional Central Pontides.

7. Conclusions

Structural and geomorphic results obtained from field observations and morphometry document a probable mechanism and timing of surface uplift of the northern margin of the Central Anatolian Plateau. Our analysis of NAF-normal components of geologic slip vectors indicate that regional strain is accumulated in the broad restraining bend of the NAF at the southern boundary of the Central Pontides (CP). Deformation potentially propagates along a deeper structural detachment horizon that creates an active orogenic wedge with a northward polarity and a positive flower-structure geometry across the CP. Our tectonic model implies that the NAF is the main driving mechanism for the contraction and uplift in the CP, and that it decouples the CP from the interior parts of the plateau, where extension has dominated since at least the beginning of the Quaternary.

Syn-tectonic deposits within the inverted basins, deeply incised gorges that straddle the northern margin of the plateau, and uplifted marine sediments and terraces imply that the formation of relief, including more than 1 km of fluvial incision, resulted from accelerated surface uplift starting in the late Miocene to early Pliocene. The onset of dextral crustal shearing along the NAF was critical for the development of the northern plateau margin, controlling tectonic inversion and contractional deformation of CAP Neogene basins. Rather than wholesale plateau uplift, our observations suggest piecemeal development of the Central Anatolian Plateau. Our tectonic concept implies that the North Anatolian Fault is the main driving mechanism for wedge tectonics and uplift in the Central Pontides, and the North Anatolian Fault Zone defines the boundary between the extensional Central Anatolian Plateau and the contractional Central Pontides.

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References

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