Rezaeian M., 2008, Coupled tectonics, erosion and climate in the Alborz Mountains, Iran. PhD thesis, University of Cambridge; 219 p.

1-1 Mountain Building and the Role of Erosion

Mountain belts are regions where strain accumulation produces lithospheric shortening, thickening, consequent uplift of rocks and Earth's surface. Mountain building at convergent plate boundaries creates new continental crust and high topography (4-5km), it interferes with patterns of air flow and precipitation, drives sediment fluxes into basins, and draws carbon dioxide from the atmosphere by erosion-driven weathering. First and foremost, mountain building is associated with continent collision, and the Alpine-Himalayan System is the principal modern example.

The evolution of active continental collision zones is described on the basis of a series of kinematic and dynamic assumptions using the concept of a critically tapered Coulomb wedge. The observation that the architecture of a fold-and-thrust belt resembles a wedge of noncohesive material, deforming ahead of an advancing backstop has prompted evaluation of the mechanics of deforming wedges at convergent margins according to Coulomb theory. This theory predicts the accumulation of crustal mass which maintains its shape and critical taper geometry during orogenesis, bounded by a basal detachment and a free surface. In this paradigm, orogenic wedges are characterized by a gentler side called pro-wedge and a steeper side called retro-wedge, over a down going (pro-) plate with a finite velocity (Fig.1.1). The theory states that wedge deformation is ultimately controlled by the balance of internal and basal friction, and that the wedge must be on the verge of Coulomb failure everywhere and at every time. This treatment prescribes a constant taper angle during wedge growth; hence, the wedge preserves a fixed length to height ratio throughout deformation. The surface taper of deforming wedges is set by the rate and location of material displacements within the wedge, but also by the removal of material along the wedge surface. This is done by erosion. Erosion at the toe of a wedge leads to an increase of the topographic taper. Erosion within the wedge leads to a decrease of the topographic taper. Coulomb theory predicts that the critical taper of an eroding wedge will be maintained by adjustments of the pattern of internal deformation (*e.g.*, Chapple, 1978; Davis *et al.*, 1983; Dahlen, 1984; 1990; Dahlen & Barr, 1989; Koons, 1989; 1990; Beaumont *et al.*, 1992; 2000; Willett *et al.*, 1993; Kooi & Beaumont, 1996; Naylor *et al.*, 2005; Burbank, 2005).

The total work due to tectonic forcing in the fold-and-thrust system is the sum of work against gravity, and work against frictional resistance (Molnar & Lyon-Caen, 1988; Dahlen & Barr, 1989; Masek & Duncan, 1998), But, the temporal evolution of mountain ranges is influenced by factors other than plate motion alone. When the crustal thickening process has reached a critical point, it begins to slow down convergence via viscous coupling of the deep continental root and the subducting plate; also the increase in elevation is a process by which an orogen progressively stocks up potential energy that can be dissipated through time (Del Castello *et al.*, 2004; Iaffaldano *et al.*, 2006).

Moreover, analogue and numerical modelling of mountain building in corroboration with natural examples suggest that, on a finer time scale, wedge development is characterized by discontinuous, alternating phases of wedge thickening, accommodated by slip along thrust faults within the wedge (underthrusting), and wedge lengthening, accommodated by nucleation of new thrust faults in front of the wedge (accretion) (*e.g.*, Mulugeta & Koyi, 1992; Storti & McClay, 1995; Kostantinovskaya & Malavieille, 2005; Del Castello & Cooke, 2007). A critical wedge grows either in width by frontal accretion, with material incorporated at the leading edge, or in height by underplating (basal accretion), where material is incorporated into the wedge at depth. The balance between the processes of frontal and basal accretion permits the wedge to maintain a critical angle of taper at all times (Davis *et al.*, 1983) (Fig. 1.2). Switching from underthrusting to accretion mode is predicted by critical taper theory to be cyclic rather than monotonic. These transitions reflect adjustments to the balance between work against frictional slip along the basal and inner wedge faults, work of deformation within the host material and uplift work against gravity (Del Castello & Cooke, 2007).

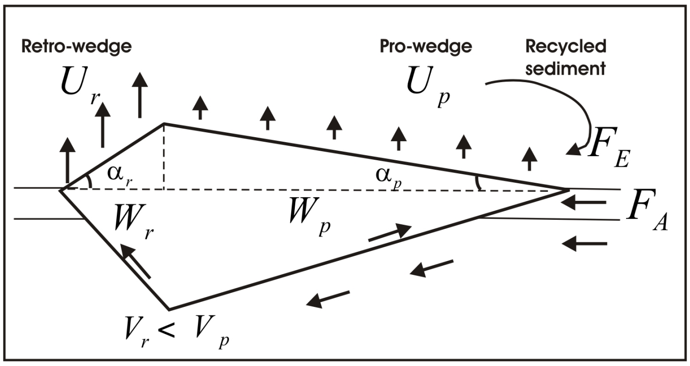


Fig. 1.1: Schematic representation of a critical wedge, defining the self-similar geometry of the wedge and key variables (after: Beaumont *et al.*, 2000; Willett *et al.*, 2001; Wipple & Meade, 2006). (*U* is the rock uplift rate, *V* is the horizontal plate velocity, *W* is the width, and *F* is a flux. Subscripts p and r denote pro- and retro-, respectively, and subscripts A and E denote accretion and erosion, respectively. α is the surface taper of the wedge).

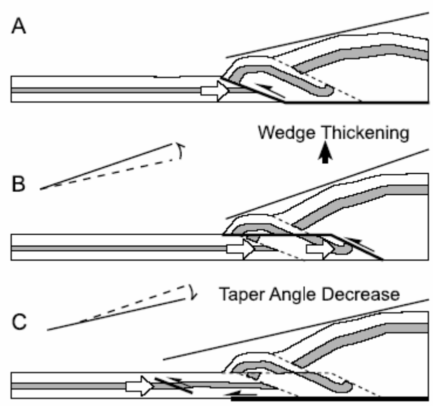


Fig. 1.2: Schematic representation of underthrusting and accretion. (a) Initial accretion. (b) Deactivation of the basal dècollement causes underthrusting under the roof thrust (white arrows). (c) Nucleation of a new fore-thrust allows the wedge toe to step forward. Continuous thick lines indicate active fault segments. Insets in Figures 1.2b and 1.2c show transitory increase/decrease of taper angle during underthrusting and accretion phases, respectively (Del Castello & Cooke, 2007).

Modelling results suggest that switches between underthrusting and accretion are driven by the interaction between gravitational forces and shear resistance along fault surfaces such that a trade-off exists between the amount of energy dissipated as internal strain and that absorbed via frictional heating on the faults. Within the accretionary system of a fold-and-thrust belt, faults are, therefore, constantly initiating, growing, and ceasing up in a complex interplay, as the system persistently seeks to minimize the total work.

Mountain belts may reach a form of topographic steady state where the gross morphology of the system remains the same and simply downwears vertically as rock mass is advected through the system (Davis *et al*., 1983; Willett *et al*., 2001; Willett & Brandon, 2002). To maintain topographic steady state under a variable erosional flux, the field of internal deformation must adjust continuously to replenish material where it has been removed. Therefore, a direct coupling between erosion, critical wedge geometry and tectonic deformation is expected (Whipple & Meade, 2006; Stolar *et al*., 2006; 2007) (Fig. 1.1).

In this concept, the rate of change in the cross-sectional area of the critical wedge (*A* in the equation 1.1), is defined by the difference between total mass influx (F_A), and the total erosional efulx (F_E):

$$\frac{dA}{dt} = F_A - F_E \tag{1.1}$$

If the mass-flux is in a steady state condition, F_E equals F_A and the rock uplift rate is determined by the local erosion rate. In a condition away from steady state, erosion rate and rock uplift are strongly coupled, even though they exhibit different transient behaviours (Whipple & Meade, 2006). Topography contains information about this coupling.

It is generally thought that the erosion of a mountain belt is determined by the runoff of water along its surface, and the slope of the topography over which it flows. The product of these two is termed the stream power. For the purpose of simplicity, other controls on erosion are often held constant in conceptual, numerical or physical evaluations of mountain belt dynamics. This has resulted in the notion that climate, through precipitation, is a first order control on the topographic and tectonic evolution of mountain belts. However, although it is clear that precipitation drives erosion, the role of climate in shaping mountain belts is difficult to pin down in geological examples. Of course, there are clear cut cases, such as the Southern Alps of New Zealand, where an extreme asymmetry of precipitation across the mountain belt has resulted in a focusing of rock uplift and exhumation almost entirely within one flank of the topographic edifice. But, in other cases, the role of climate is more ambiguous. Recent work in Taiwan, for example, has found no relation between the pattern of precipitation and erosion by conventional means. There, erosion does not track stream power in a simple way, indicating that other controls over erosion take precedence. The controls on patterns and rates of erosion on orogenic scales remain poorly constrained (*e.g.*, Koons, 1990; Dadson, 2004). Another preconception that has hampered progress of understanding is that mountain belts are, generally, at or close to steady state. Modelling work (*e.g.*, Willett, 1999) has shown that for a mountain belt to achieve any form of steady state, the required shortening far exceeds ten times the undeformed thickness of the (brittle part of the) incoming plates. In many active mountain belts this is clearly not the case. Many mountain belts are probably not at steady state. Moreover, the final shape of a mountain belt is determined by the interplay of deformation and erosion before steady state is reached. In these early stages of mountain building, pre-existing structures and the properties of the rock mass presented for erosion must help determine the pattern of erosion. Few studies have documented this in detail.

Numerical and analogue models, useful though they are in defining the roles of certain processes in mountain building, remain incomplete and underconstrained. This is true for erosion as well as for processes within the deforming wedge. In this thesis, I shall seek to document the erosion and exhumational evolution of a young mountain belt, exposed to a strong climatic gradient and not yet at steady state. This mountain belt is the Alborz Mountains, located centrally within the Alpine-Himalayan system.

1-2 Tectonic Setting

1-2-1 The Alpine-Himalayan Chain

The Alpine-Himalayan chain (Fig. 1.3), which includes (from west to east) the Pyrenees, European Alps, Apennines, Dinarides, Carpathians, Anatolian Plateau, Caucasus, Alborz, Zagros, Kopeh Dagh, Makran, Hindu Kush, Karakorum, Tien Shan, Tibet, and the Himalayas stretches from Spain to Indonesia and is the result of a step wise closure the Mesozoic Tethyan (Neo-Tethys) sea way. This gave rise to detachment of cratonic blocks from the southern landmass of Gondwana, their transfer and accretion to the southern margin of the northern landmass of Eurasia and eventually, collision of Africa-Arabia-India with Eurasia. After collision shortening continued and in the central Neo-Tethyan zone the progressive northwards movement of Arabia built the Turkish-Iranian Plateau, which is active today (*e.g.*, Ziegler, 1997; Brunet & Cloetingh, 2003).

A compressional stress regime prevailed along the northern and southern margin of the Neo Tethyan seaway from the Late Cretaceous. It resulted in uplift and inversion of basins, and the construction of pre-Alpine-Himalayan ranges. This was followed, during the Eocene, by a major extension phase, located in the SE Black Sea through the southern Lesser Caucasus and into the Alborz area. Extension was caused by the development of a back-arc basin to the north of the Neo-Tethys subduction zone (Berberian & King, 1981; Kley & Eisbacher, 1999; Brunet & Cloetingh, 2003; Nikishin *et al.*, 2003).

Since Miocene times, the most prominent expression of ongoing shortening has been the emplacement of nappes and tectonic inversion of rifted basins (Brunet & Cloetingh, 2003).

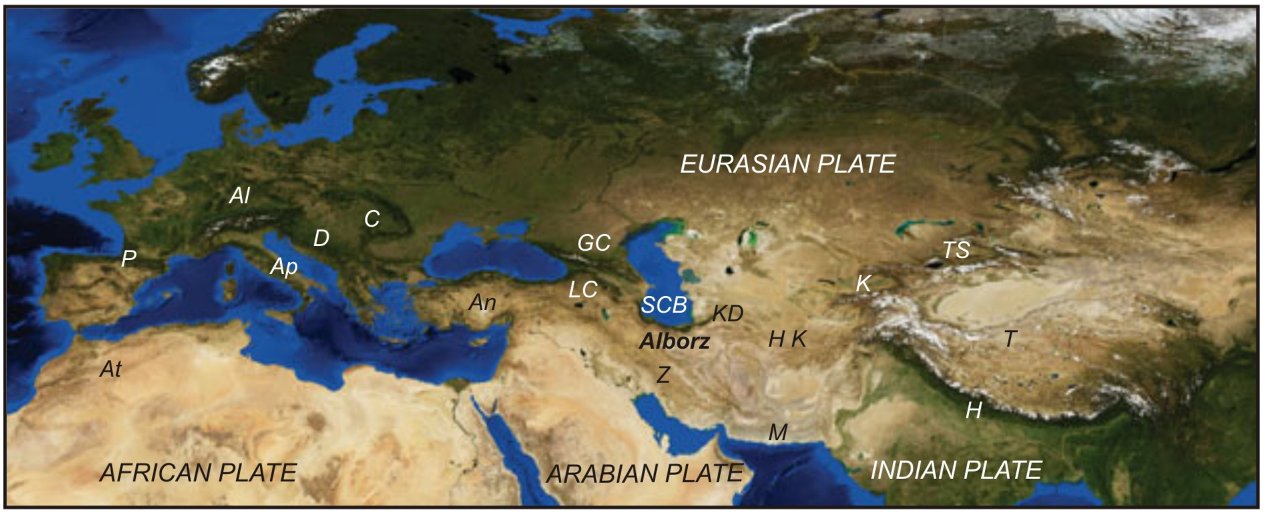


Fig. 1.3: The Alpine-Himalayan orogen including mountain belts of the Alborz, Alps (Al), Anatolia (An), Apennines (Ap), Atlas (At), Carpathians (C), Dinarides (D), Great Caucasus (GC), Himalaya (H), Hindu Kush (HK), Karakorum (K), Kopeh Dagh (KD), Lesser Caucasus (LC), Makran (M), Pyrenees (P), South Caspian Basin (SCB), Tibet (T), Tien Shan (TS), and Zagros (Z) (image from: NASA).

The central position of the Iranian Plateau between Alps and Himalayas highlights the importance of the Alborz Mountains. However, in contrast to the Himalayas and Alps, which have a long and rich tradition of geophysical and geological investigations, the geodynamic evolution of the Alborz is poorly constrained. This study is the first systematic examination of the exhumational and topographic history of the Alborz Mountains.

1-2-2 The Iranian Plateau

The Iranian plateau has accommodated convergence between the Arabian and Eurasian plates since the Latest Cretaceous-Early Tertiary. The onset of substantial compression has been marked by a major erosional unconformity throughout the greater part of Iran (Berberian & King, 1981; Berberian, 1983; Sengor, 1990; Ziegler, 2001). Convergence has driven a steady, slow subduction along the Neo-Tethyan suture, located north of the High Zagros, at a rate of 32±2mm/yr over 56 My (McQuarrie *et al.*, 2003), and 22-25 mm/yr at the present time (Vernant *et al.*, 2004a). Since initial collision, there has been ~300-500 km of convergence between Arabia and Eurasia (Allen *et al.*, 2004), of which 6-17% has been accommodated in the Alborz (Fig. 1.4).

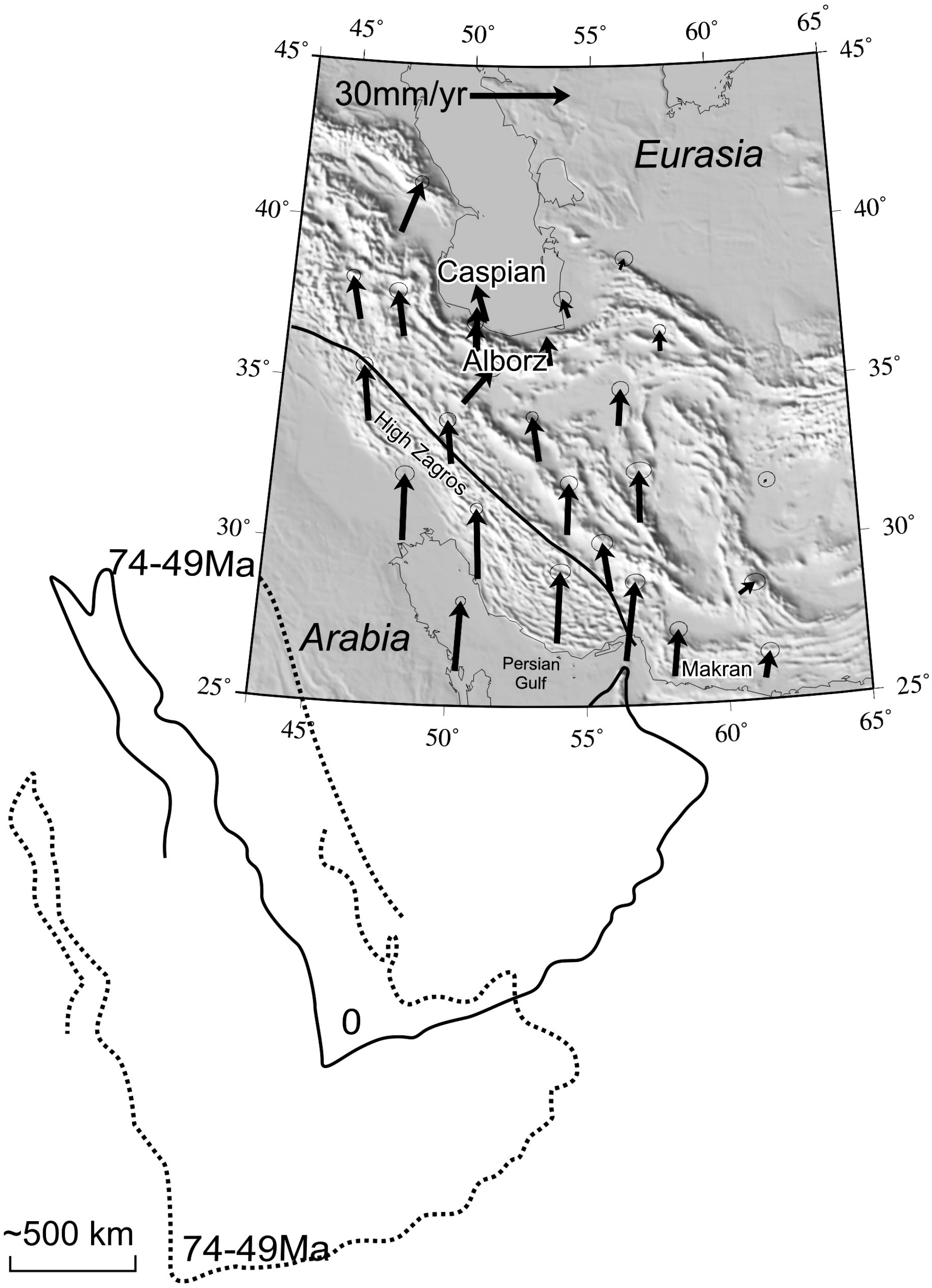


Fig. 1.4: Geotectonic setting of the Iranian Plateau in the Arabia-Eurasia collision zone and its plate tectonic reconstruction since the Late Cretaceous (Allen *et al.*, 2004). The present location within the Zagros Mountains of the Neo-Tethys suture and the Arabian Plate margin has been shown by a thick line, and the position of the Arabian continent around 74-49 Ma by a dotted line. The present, GPS-derived velocity field of the collision zone has been displayed, with respect to stable Eurasia, after Vernant *et al.* (2004a).

In the Iranian Plateau, several tectono-stratigraphic domains surround the Alborz Mountains; understanding of the geological evolution of these neighbouring domains is required to interpret the evolution of the Alborz Mountains over geological time. They are: Kopeh-Dagh (Russian: Kopet-dagh) to the east, Central Iran to the south and southeast, the Urumiyeh-Dokhtar magmatic assemblage to the southwest, the

Talesh (Russian: Talysh) to the northwest, and the South Caspian Basin to the north (Fig. 1.5). These domains have tectono-stratigraphic histories that are related to, and influenced by the evolution of the Alborz Mountains.

Fig. 1.5: Tectono-stratigraphic setting of the Iranian Plateau, located between the Arabia and Eurasia plates. Major tectono-stratigraphic domains are indicated on the topographic map. The Alborz Mountains are surrounded by Central Iran, Kopeh-Dagh, South Caspian Basin, Talesh and Urumiyeh-Dokhtar magmatic assemblage.

The Kopeh-Dagh in the northeast is formed in the Hercynian metamorphosed basement of the southern margin of Eurasia (the Turan Platform; Ruttner, 1984, 1993). It is characterized by relatively continuous carbonate-siliciclastic sedimentation in a marine-fluvial environment, as a result of frequent regressions and transgressions from the Middle Jurassic to the Miocene (Afshar-Harb, 1979; Alavi, 1996; Mahboubi *et al.*, 2001). As with the Zagros, the Kopeh-Dagh was folded into elongate, NW-SE trending folds during the last phase of the Alpine orogenesis. It is distinguished from the Alborz by the lack of any igneous activity in the Tertiary and Quaternary (Stocklin, 1974; Afshar-Harb, 1979; National Iranian Oil Company, 1982a; b; Lyberis *et al.*, 1998; Mahboubi *et al.*, 2001).

Central Iran, to the south of the Alborz, is characterized by Gondwana-derived blocks of Precambrian basement and a nearly complete Paleozoic, platform-type sedimentary cover similar to that of the Alborz. This region had a close tectono-stratigraphic tie with the southern Alborz during the Tertiary, but it was more mobile during the Mesozoic (Stocklin, 1974; Berberian & King, 1981).

The Urumiyeh-Dokhtar magmatic assemblage to the southwest, which mechanically speaking is a part of Central Iran, represents an Andean-type magmatic-arc formed in response to subduction of Neo-Tethyan oceanic crust beneath Central Iran (Berberian *et al.*, 1982; Alavi, 1996). Subduction may have started in the Mesozoic and persisted to the Oligocene-Miocene. During this time, the locus of magmatic activity migrated inland from SW to NE (Berberian *et al.*, 1982; Berberian, 1983).

The Talesh Mountains form the northwestern continuation of the Alborz. The trend of this arcuate mountain belt changes from N-S adjacent to the South Caspian, to WNW-ESE towards the Lesser Caucasus. Its stratigraphy appears to be similar to that of the southern Alborz in the Mesozoic and the Paleogene, with a mixed clastic-carbonate succession in the Mesozoic; volcanic, volcaniclastic, and clastic rocks, with minor hypabyssal intrusions in the Paleogene, and clastic sediments during the Miocene-Pliocene (Allen *et al.*, 2003b; Vincent *et al.*, 2005).

At the northern edge of the Arabia-Eurasia collision zone, the South Caspian Basin forms an intracontinental depression floored by probable oceanic crust which is overthrust by surrounding mountain belts. To its south, the arcuate Alborz Mountains follow the shape of the South Caspian Basin aseismic, rigid block (Berberian, 1983; Jackson *et al.*, 2002). Apparently, the South Caspian Basin, which opened in the Jurassic (Berberian, 1983; Brunet *et al.*, 2003; Golonka, 2004), was connected to the early Black Sea during the Late Cretaceous-Paleogene, and was isolated from the latter in the Early Pliocene (Zonenshain & Le Pichon, 1986; Golonka, 2004; Smith-Rouch, 2006). The South Caspian Basin is considered as either (i) a remnant of the Paleo-Tethys oceanic crust being trapped since the Late Triassic along a major suture (Stocklin, 1974); (ii) a Mesozoic-Tertiary Neo-Tethyan back-arc basin (Berberian, 1983); or (iii) a Middle-Late Jurassic Neo-Tethyan back-arc basin with possible reactivation during the Cretaceous (Brunet *et al.*, 2003).

During the Eocene the South Caspian Basin was part of a vast epi-continental system that developed as a series of brackish seaways, lakes and wetlands within the interiors of central Eastern Europe and western Asia. Major relics of this ancient aquatic realm are called Para-Tethys (Laskarev, 1924; Zonenshain & Le Pichon, 1986). In the northern part of the Tethys ocean, the marginal sea of Para-Tethys underwent major subsidence from the Eocene peaking in the Pliocene-Pleistocene (Berberian, 1983). It was separated from the Tethys ocean in the Oligocene (Rogl, 1999) following collision of relict fragments of Gondwana and Eurasia (Aliyeva *et al.*, 2004). The Para-Tethys was divided into the western and eastern Para-Tethys by Carpathian Mountains. The present day Black Sea and the Caspian Sea are remains of the eastern Para-Tethys (Vasiliev *et al.*, 2004) and may have existed as two distinct basins from the Eocene-Oligocene (Smith-Rouch, 2006). They are thought to have been closed off as a marine realm in the Late Burdigalian (Zonenshain & Le Pichon, 1986; Rogl 1999), and have existed from then as trapped, relatively aseismic blocks within the Eurasian part of the Arabia-Eurasia collision zone, collecting thick piles of sediment (Jackson *et al.*, 2002; Allen *et al.*, 2003a).

The previously mentioned tectono-stratigraphic domains are easily identified looking at the map showing teleseismically recorded earthquakes in last 40 years (Fig. 1.6). The central Iranian Plateau and South Caspian Basin are the rigid blocks surrounded by active zones. Seismicity is scattered in the Alborz Mountains.

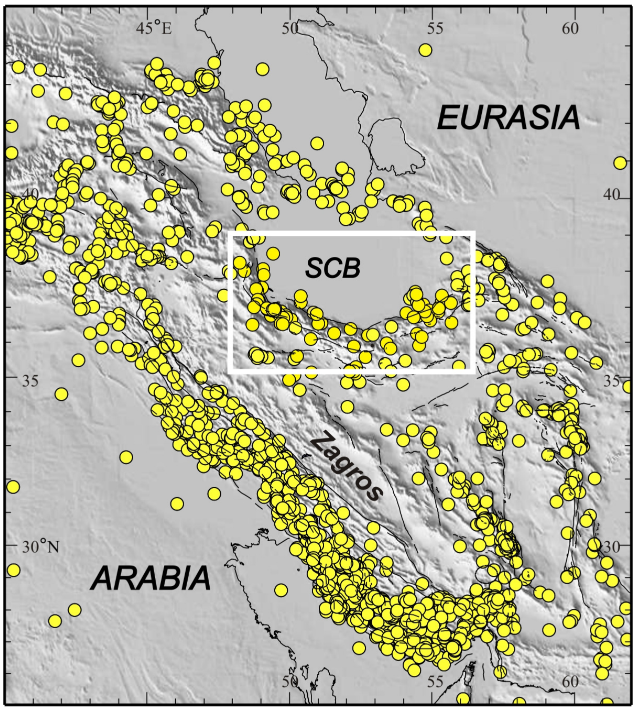


Fig. 1.6: Seismic setting of the Alborz Mountains (marked by white box), in the collision zone between Arabia and Eurasia. Yellow circles are teleseismically recorded earthquakes in a period of 1964-2004 from Tatar *et al.*(2007). Subduction of the Arabian plate beneath Eurasia is expressed in localized seismicity within the Zagros fold-and-thrust.

1-2-3 The Alborz Mountains

The Alborz Mountains, located along the northern edge of the collision zone between the Eurasian and Arabian plates, preserves a unique record of convergence since the Paleocene. It constitutes an important link in the Alpine-Himalayan chain, wrapping around the South Caspian Sea (a trapped remnant oceanic crust; Berberian, 1983), and forming a gently curved mountain belt in northern Iran (Fig. 1.4 & 1.5).

Stratigraphy and pre-existing geological structures have had a significant impact on mountain building in the Alborz. This is evident, in the first instance, in the asymmetric spatial pattern of geological formations in the mountain (Fig.1.7a,b) with Paleocene-Eocene formations dominating the southern Alborz, and Mesozoic and older rocks building the northern Alborz. Previous work has not been able to determine whether Paleogene formations ever reached important thickness in the northern Alborz, or were thick but have been eroded.

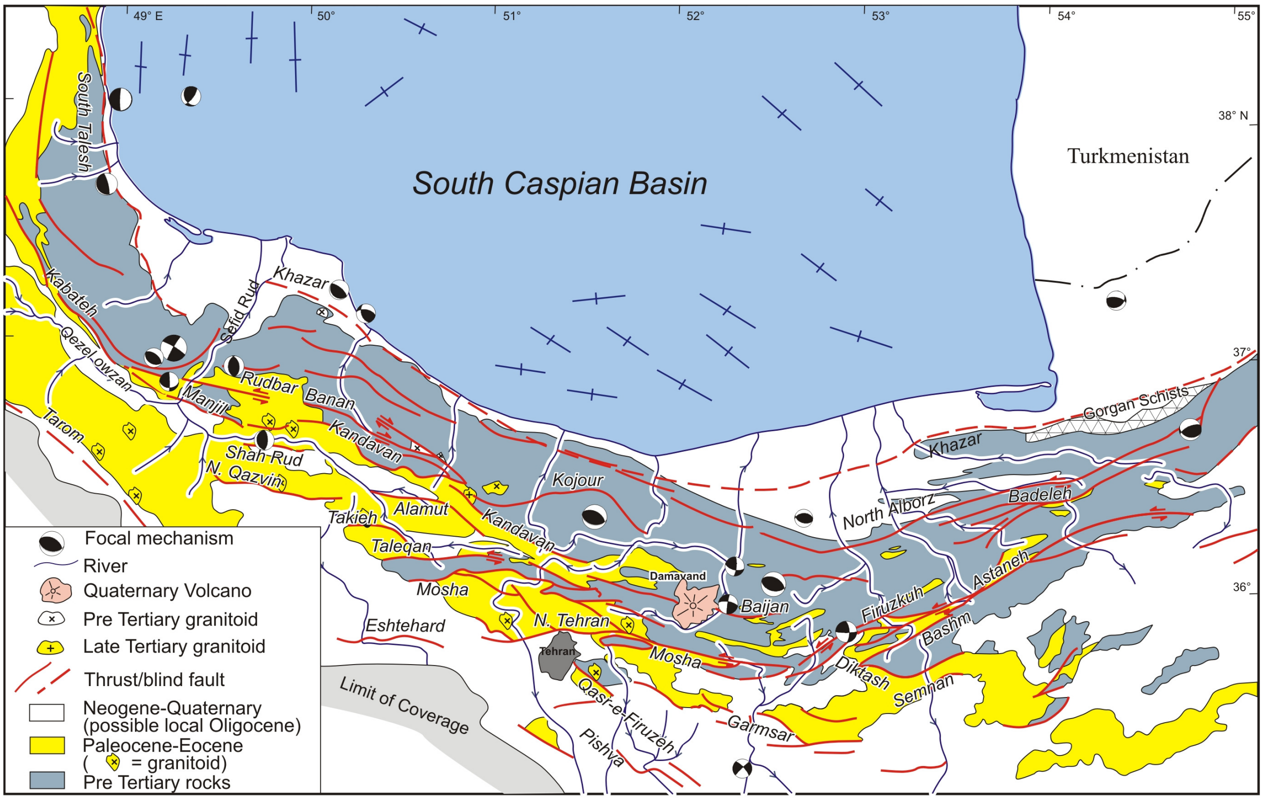


Fig. 1.7a: Generalised geological-structural map of the Alborz (After Allen *et al.*, 2003a); some geological features are added after Berberian *et al.* (1985, 1996), Berberian & Yeats (2001), Jackson *et al.*, 2002, Guest *et al.* (2006b), and Nazari (2006). The Focal mechanisms of earthquakes are from Jackson *et al.* (2002) and Tatar *et al.*(2007).

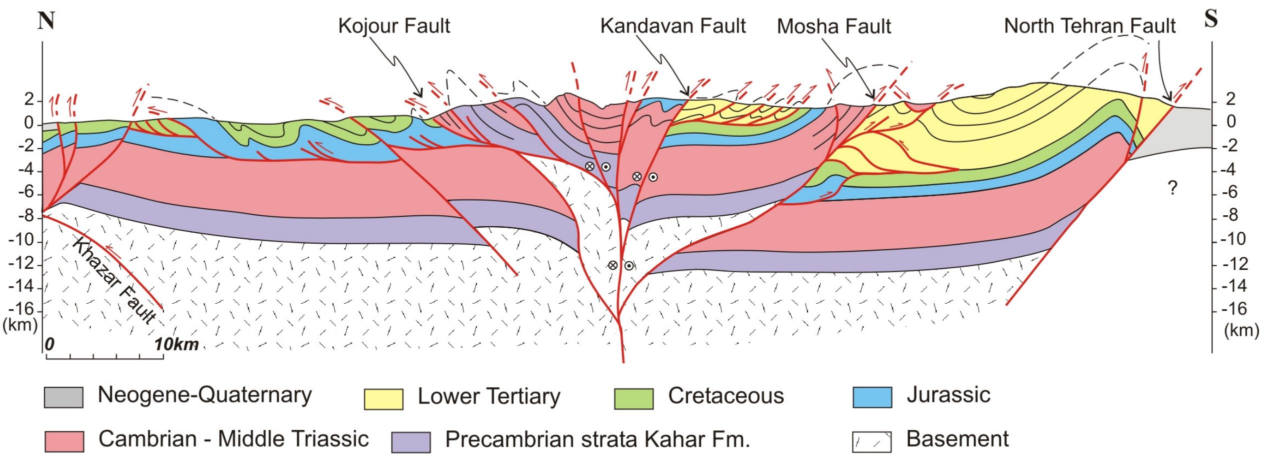


Fig. 1.7b: Cross-section through the Alborz at the longtidue of Tehran (51° 30' E) (Allen *et al.*, 2003a).

Major vertical movements have been recorded by stratigraphy through several tectonic perturbations in Tertiary time. About 4 km, and locally up to 10 km, of submarine volcanic and tuffaceous rocks were deposited during the Eocene in a rapidly subsiding trough now located within the southern Alborz, and covered by a thick sequence of Oligocene-Pliocene molasses. The latter started with terrestrial deposits or subaerial volcanics, overlain by marine limestones, which later changed into gypsum and salt-bearing deposits and finally into conglomerates. The total thickness of sediments shed off the mountain belt since Oligocene has been estimated at 4000 to >7400 m (Stocklin, 1974; Geological Survey of Iran, 1987; 2003; Ballato *et al*., 2008). According to stratigraphy (Geological Survey of Iran, 2001) the mountain belt has been uplifted from sea level to > 3000m since the Early Miocene.

On the basis of geological observations, several episodes of deformation of the Alborz Mountains have been documented. They are the Cretaceous-Paleocene, Eocene-Oligocene, Middle Miocene and Miocene-Pliocene (Berberian & King, 1981; Hempton, 1987; Axen *et al.*, 2001a; Allen *et al.* 2003a; Guest *et al.*, 2006b). Moreover, Axen *et al.* (2001a) has revealed a major exhumation phase within the western Alborz since the late Miocene. Miocene N-S compression has resulted in right- and left-lateral slip, and since the early Pliocene left lateral slip driven by NNE-SSW transpression has dominated in the Alborz (Axen *et al.*, 2001a; Allen *et al.*, 2003a). From the middle Pleistocene onward WNW-ESE transtension has occurred within the internal domain of the range (Ritz *et al.*, 2006; Landgraf *et al.*, 2008).

Geodetic measurements pin the modern rate of convergence of Arabia and Eurasia at 22-25 mm/yr (Vernant *et al.*, 2004a). Of this, ~ 5 mm/yr is absorbed by the Alborz Mountains. This makes up ~40% of the total shortening within the central Iranian Block (Vernant *et al.*, 2004b). The total Late Cenozoic shortening across the Alborz Mountains is poorly constrained, but minimum estimates are around 30 km; obliquity of the convergence has caused a similar amount of left lateral slip in the mountain belt (Allen *et al.*, 2003a; Allen *et al.*, 2006).

Given the present deformation rates, the minimum total shortening and strike slip offset in the Alborz can have been achieved in as little as 6-7 My. Similarly, Allen *et al.* (2004) have extrapolated the present deformation rates on major active fault systems in the entire Arabia-Eurasia collision zone, finding that at present deformation rates the total estimated shortening across the Arabia-Eurasia collision zone could have been achieved in ~3-7 Ma. This implies that the present kinematics of the Arabia-Eurasia collision are unlikely to be representative of the entire period since onset of collision, and that deformation rates have been anomalously high in the recent past.

The Alborz Mountains represent a crustal weak zone with evidence for repeated reactivation of inherited crustal fabrics of Mesozoic age (Ballato *et al*., 2008); structurally, it can be divided into four zones each with their distinct structural grain: N-S in Talesh, NW-SE in western Alborz, E-W in central and north-east Alborz, and NE-SW in south-east Alborz, which are lined up by the shape of rigid South Caspian Block. The main NW-SE and NE-SW trends with major thrust and strike slip components, respectively, are accommodating the recent stress field which is oriented NNE-SSW. The strike slip component allows extrusion of the South Caspian Basin, which is moving northwestward with respect to Eurasia (Copley & Jackson, 2006; Hollingsworth *et al*., 2006).

Deformation in the Alborz is governed by main frontal thrust faults at its borders and strike slip faults in relatively high elevation within the range; however they do not necessarily meet at deeper levels of the crust to form a flower structure (Tatar *et al*., 2007). The Khazar fault (Fig. 1.7a) is interpreted to define at depth the structural boundary between the Central Iranian and South Caspian blocks. If so, the seismic profile obtained from the aftershock pattern following the Baladeh earthquake ($M_W = 6.2$) (Tatar *et al*., 2007) illustrates underthrusting of the South Caspian plate.

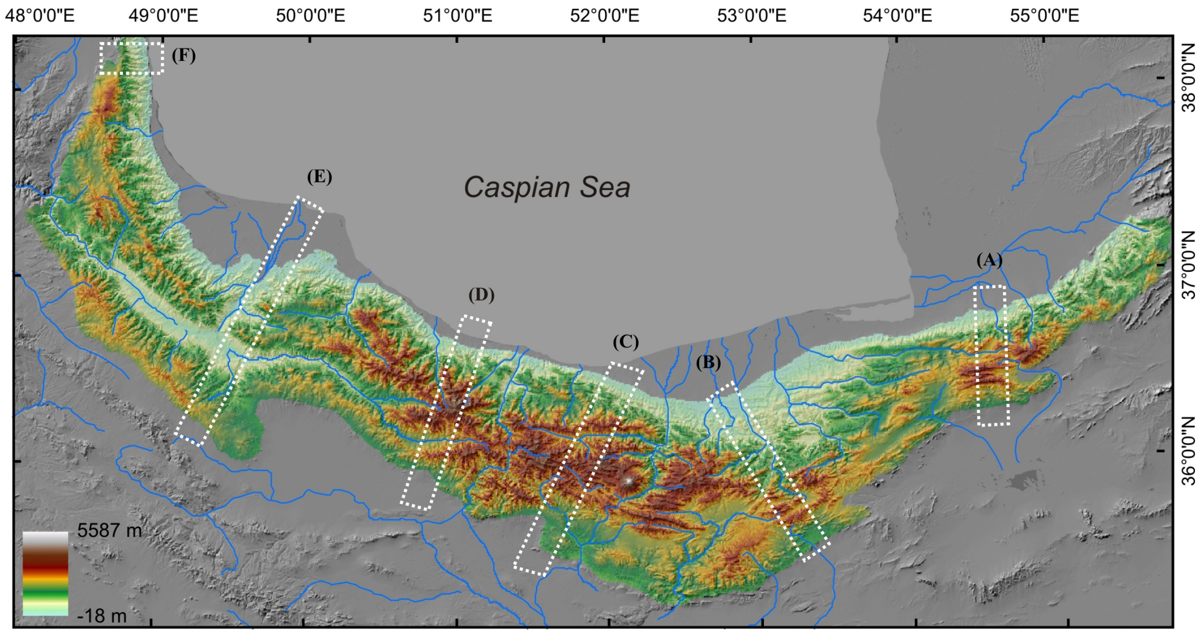
The Alborz Mountains accommodate ~ 40% of the deformation between Eurasia and the Central Iranian Plateau, according to GPS data (Vernant *et al*., 2004b). It is estimated that the mountain belt has been shortened to about 20-50% of pre-shortening width at the longitude of Tehran (Stocklin, 1974; Berberian, 1983; Allen *et al*., 2003a; Guest *et al*., 2006b). Across Iran as a whole, aseismic creep on faults or by folds plays a leading role in the deformation of the region. Jackson *et al.* (1995) have found that the sum of earthquakes between 1909 and 1992 can account for only a small part (~10-20%) of the total deformation required by the convergence between the Arabia and Eurasia plates. Nevertheless, in the Alborz, the seismic deformation is probably 100% (Masson *et al*., 2005).

Apart from a few exceptions, well-constrained earthquakes in the Alborz have occurred at depths <15 km, with either reverse focal mechanisms or left lateral strike slip on longitudinal faults (Jackson *et al*., 2002). A minor normal component has been recognised via seismic data and geological observations mainly in the south central Alborz (Jackson *et al*., 2002; Ashtari *et al*., 2005; Guest *et al*., 2006b; Ritz *et al*., 2006; Zanchi *et al*., 2006). Extension has been attributed either to the partitioning of deformation in the Central Alborz or to gravitational collapse (Alavi, 1996; Guest *et al*., 2006b; Ritz *et al*., 2006; Zanchi *et al*., 2006). This extension maybe coincide with a decrease of relief in the internal Alborz, a diminishing of sedimentation rates in South Caspian Basin, and intraplate magmatism of Damavand since 1.8 Ma (Brunet *et al*., 2003; Davidson *et al*., 2004; Ritz *et al*., 2006). Whether this extension is regional in significance or a local effect is not clear, but existing GPS measurements indicating shortening across the range as a whole.

Teleseismic waveform modelling of two moderate-sized earthquakes in north-central Alborz (Baladeh earthquake) and north-west Alborz (Talesh) has revealed greater centroid depths of 27-35 km. Both these earthquakes occurred on the contact between continental (Alborz) and suspected oceanic (South Caspian) crust, suggesting subduction of the latter west-south-westwards under the Alborz Mountains (Jackson *et al*., 2002; Allen *et al*., 2003b; Tatar *et al*., 2007).

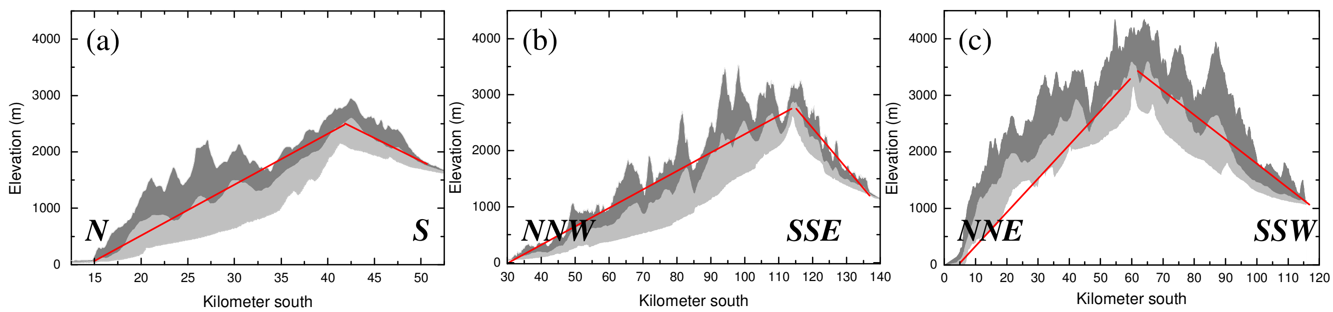
The topography of the Alborz Mountains shows a systematic variation in taper angle from east to west, coincident with structural trends (Fig. 1.8; 1.9).

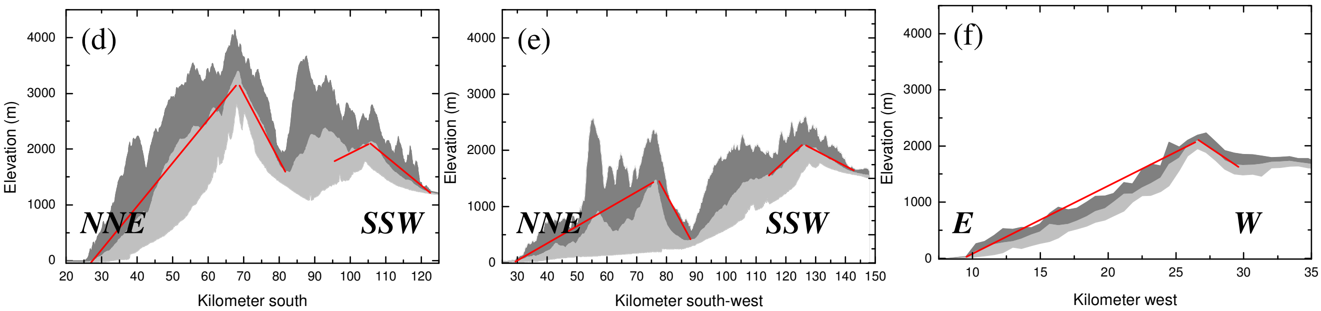
Fig. 1.8: Topographic map of the Alborz Mountains, computed using a 90 m digital elevation model (DEM) acquired by the Shuttle Radar Topography Mission (SRTM). The highest peak of Damavand volcano (5586m) is located in the central Alborz. Boxes A to F are the locations of topographic cross sections (a) to (f) in Figure 1.8.



The geometry of the Alborz Mountains changes from an asymmetric double-sided wedge in east (Fig. 1.9a & b) to a more or less symmetric double-sided wedge in the centre (Fig. 1.9c), and an asymmetric double sided wedge with a second, smaller range to the south, in the centre-west, and single asymmetric double-sided wedge in the west.

Fig.1.9: Topographic cross sections of the Alborz Mountains illustrating the typical two-sided geometry of the mountain belt. (a) & (b) are in the eastern Alborz with NE-SW structural trend; (c) is in the central Alborz with E-W structural trend; (d) & (e) are in the west-central Alborz with NW-SE and E-W structural trends; and (f) is in the western Alborz with N-S structural trends. Plotted at 500 m intervals are the maximum (dark grey), mean (medium grey), and minimum (light grey) elevations along 20-km-wide swath profiles taken from the SRTM DEM. Straight red lines are linear eye-fits to the mean elevations, shown only to demonstrate the close approximation to a taper form. Horizontal axis is the distance from Caspian Sea coastline in km.





The Alborz Mountains have a NE-SW trend east of ~52º 30΄E where the Khazar and North Alborz faults are documented as separate south dipping structures along the northern range front (Fig. 1.7a). Here the north flank of the mountain belt (possibly the pro-side of the wedge) is >4.5 times wider than the south flank (Fig. 1.9a & b). Thrust faults with duplex geometry verging SSE are characteristic of the eastern part of this zone (~54-54º 30΄E) (Shahriari *et al.*, 2003). This part of the mountain belt may represent an early stage of wedge development, with a minimal taper angle, relatively low topography, and a pronounced asymmetry.

West of ~52º 30΄E, the structural trend changes from NE-SW to the NW-SE, and the main structures of Khazar and North Alborz faults join. Here, the structures are documented as blind faults, with shallower foreland dipping panels of imbricated Mesozoic and younger strata. These north-dipping structures and pre-Quaternary strata are observed to the west until ~50º E (Allen *et al.*, 2003a). In this sector, the structural profile of the mountain belt displays a symmetry with major structures verging south in south flank and north in north flank (Allen *et al.*, 2003a), although some north verging local structures have been described in marginal south flank (Geological Survey of Iran, 1987; 2003; Ballato *et al.*, 2008). Here the topography of the Alborz range is less asymmetric, with steep topography and high ridge pole elevations.

West of 52ºE, a major tectonic vergence toward the interior of the mountain belt emerges in the south flank of the Alborz (Guest *et al.*, 2006b; 2007). Here the mountain belt appears to be made of two adjacent asymmetric ranges with shallower topography to the west (Fig. 1.9d & b).

Finally, in the westernmost Alborz (Fig. 1.9f) a main SW tectonic vergence is implied in structural cross sections (Geological Survey of Iran, 1999); the topography again is highly asymmetric, and the east flank of the mountain belt is up to >6.8 times wider than the west flank.

Tectonic forcing in the Alborz is low compared to other active areas; shortening rates are 7-15% to 30% of those in Taiwan, Andes, and Himalaya, respectively (Bilham *et al.*, 1997; Yu *et al*, 1997; Liu *et al*, 2000). Similarly, climatic forcing, approximated in the annual precipitation is relatively mild.

The Alborz Mountains form a barrier across the prevailing NE-SW air-flow from the Caspian Sea. Precipitation in the northern Alborz and Caspian coast is the result of thermodynamic destabilization of Siberian cold air as it crosses the warmer sea surface (Khalili, 1973). Air flow from the Caspian basin is blocked by the high topography of the central Alborz, and water vapour is advected westward along the mountain front as well as southward across the range during much of the year. Total annual rainfall ranges from 85 mm in SE Alborz up to 1800 mm in the NW (Fig.1.10).

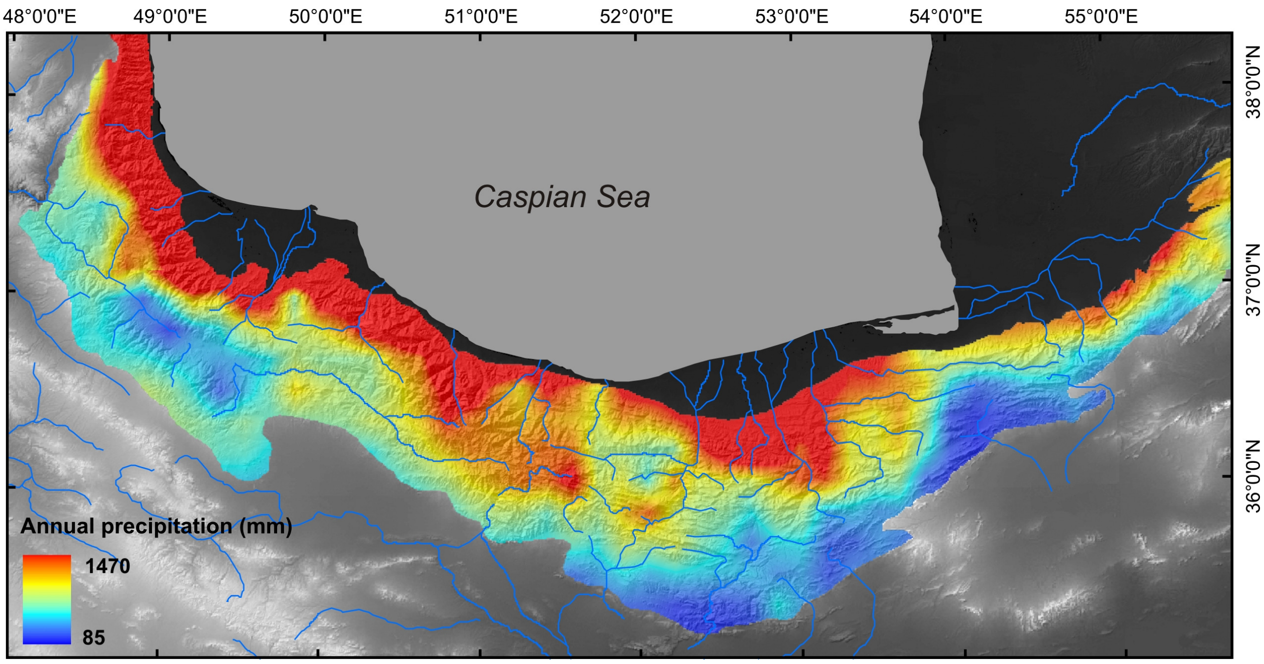


Fig.1.10: Total annual precipitation in the Alborz, derived from 877 meteorological stations (data from TAMAB and IRIMO). The intervals are estimated on the basis of the quarter of standard deviation of the precipitation.

In summary, the Alborz sustain a strong asymmetry in climate, vegetation, lithology, geological structure and topography between the north and south flanks of the mountain belt. This makes the Alborz an excellent complement to better known places such as the Himalayas, the Andes, the Southern Alps and Taiwan. The central questions investigated in this thesis are: How has this asymmetry arisen? How has it affected the exhumation of the mountain belt? And how is modern erosion responding to it? Answers to these questions would shed light on the relative importance of controls on erosion and the resulting exhumation of rock mass, as well as on the origins of a mountain belt that forms a substantial link in the Alpine-Himalayan chain. They require a range of tools and techniques applied on geomorphological as well as geological time scales.

I have used hydrometric, geotechnical, geological and satellite data to explore and explain the short erosion of the Alborz Mountains, and stratigraphic and low-temperature thermochronometric data to investigate the long term exhumation of the mountain belt. The two ensemble approaches are highly complementary, and together yield a reliable insight into the structural and exhumational history of the

Alborz Mountains, as well as the reasons for the localization of exhumation and structural activity. The structure of the thesis is described in the following section.

1-3 Outlines

In chapter 2, the history of cooling of the Alborz fold-and-thrust belt is constrained by means of apatite fission track analysis and apatite (U-Th)/He thermochronometry. These techniques are well suited to the study of recent and relatively shallow exhumation of mountain belts. They are found to cover the entire constructional history of the Alborz Mountains, revealing four main phases of cooling (magmatic and exhumational). This is the first thermochronometric study of the Alborz Mountains on the scale of the mountain belt.

In chapter 3, the stratigraphic record of geological processes in the Alborz region since the Late Cretaceous is reviewed. This is done largely by synthesising a range of published local and regional studies. The stratigraphic record is the sedimentary counterpart to the thermochronometric record presented in Chapter 2. It can therefore be used to test the validity of findings in Chapter 2, and to add a level of spatial and temporal detail.

In Chapter 4, thermochronometric and stratigraphic data are combined in a reconstruct the history of the Alborz region from the very onset of mountain building. Three main phases of compressional deformation and concomitant erosion are identified, dated and compared. These phases are then evaluated in the broader context of deformation and exhumation along the southern margin of Eurasia, and global synchronicity of exhumation is explored.

In Chapter 5, a switch is made to shorter, geomorphic time scales. This is motivated by the finding that long-term exhumation patterns in the Alborz do not neatly match the strong climatic gradient across the mountain belt. To address this problem, the current pattern of erosion is constrained using hydrometric data from stations throughout the mountain belt. The result is a detailed map of decadal average erosion rates which has few precedents elsewhere.

In Chapter 6, precipitation, runoff, vegetation, lithology and topography are invoked as possible controls over erosion. Each variable is quantified on the scale of the mountain belt, and its correlation with erosion is tested. The surprising and important finding of this chapter is that the pattern of erosion of the Alborz Mountains is set by the resistance of the surface and substrate against erosion, rather than by the intensity of the classic drivers of erosion.

In Chapter 7, this finding is used to explain the long-term exhumation pattern in the Alborz Mountains. It is concluded that erodibility is a primary factor that determines the location and intensity of erosional exhumation of rock mass, and couples into the structural evolution of mountain belts. Erodibility is, to an important degree, seeded by the geological history of a deformed area prior to mountain building. It can therefore be included in geodynamic models with a good degree of confidence.