Tectono-stratigraphic evolution of the intermontane Tarom Basin (NW sectors of the Arabia-Eurasia collision zone): insights into the vertical growth of the Iranian Plateau margin

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Abstract

The intermontane Tarom Basin of NW Iran (Arabia-Eurasia collision zone) is located at the transition between the Iranian Plateau (IP) to the SW and the Alborz Mountains to the NE. This basin was filled by Late Cenozoic synorogenic red beds that retain first-order information on the erosional history of adjacent topography, the vertical growth of the plateau margin and its lateral (orogen perpendicular) expansion. Here, we perform a multidisciplinary study including magnetostratigraphy, sedimentology, geochronology and sandstone petrography on these red beds. Our data show that widespread Eocene arc volcanism in NW Iran terminated at ˜ 38-36 Ma, while intrabasinal synorogenic sedimentation occurred between ˜ 16.5 and < 7.6 Ma, implying that the red beds are stratigraphically equivalent to the Upper Red Formation. After 7.6 Ma, the basin experienced intrabasinal deformation, uplift and erosion in association with the establishment of external drainage. Fluvial connectivity with the Caspian Sea, however, was interrupted by at least four episodes of basin aggradation. During endorheic conditions the basin fill did not reach the elevation of the plateau interior and hence the Tarom Basin was never integrated into the plateau realm. Furthermore, our provenance data indicate that the northern margin of the basin experienced a greater magnitude of deformation and exhumation than the southern one (IP margin). This agrees with recent Moho depth estimates, suggesting that crustal shortening and thickening cannot be responsible for the vertical growth of the northern margin of the IP, and hence surface uplift must have been driven by deep-seated processes.

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Key Points:

- In the Tarom Basin arc volcanism terminated at ~38-36 Ma, while intermontane synorogenic deposition occurred from ~16.5 to < 7.6 Ma
- The Iranian Plateau formed in the broken retroforeland of the Arabia-Eurasia collision zone
- Crustal shortening and thickening cannot be responsible for the vertical growth of the

Iranian Plateau margin

Abstract

The intermontane Tarom Basin of NW Iran (Arabia-Eurasia collision zone) is located at the transition between the Iranian Plateau (IP) to the SW and the Alborz Mountains to the NE. This basin was filled by Late Cenozoic synorogenic red beds that retain first-order information on the erosional history of adjacent topography, the vertical growth of the plateau margin and its lateral (orogen perpendicular) expansion. Here, we perform a multidisciplinary study including magnetostratigraphy, sedimentology, geochronology and sandstone petrography on these red beds. Our data show that widespread Eocene arc volcanism in NW Iran terminated at ~ 38-36 Ma, while intrabasinal synorogenic sedimentation occurred between ~ 16.5 and < 7.6 Ma, implying that the red beds are stratigraphically equivalent to the Upper Red Formation. After 7.6 Ma, the basin experienced intrabasinal deformation, uplift and erosion in association with the establishment of external drainage. Fluvial connectivity with the Caspian Sea, however, was interrupted by at least four episodes of basin aggradation. During endorheic conditions the basin fill did not reach the elevation of the plateau interior and hence the Tarom Basin was never integrated into the plateau realm. Furthermore, our provenance data indicate that the northern margin of the basin experienced a greater magnitude of deformation and exhumation than the southern one (IP margin). This agrees with recent Moho depth estimates, suggesting that crustal shortening and thickening cannot be responsible for the vertical growth of the northern margin of the IP, and hence surface uplift must have been driven by deep-seated processes.

KEYWORDS: Iranian Plateau, plateau margin uplift, deep seated processes, magnetostratigraphy, depositional settings, intermontane sedimentation.

1. Introduction

Orogenic plateaus are vast and elevated morphotectonic provinces, which provide the unique opportunity to decipher the interplay between shallow, deep-seated and surface processes, and their influences on Earth’s landscape at various timescales (e.g., Dewey et al., 1988; Isacks, 1988; Molnar et al., 1993). They contain internally drained basins that have coalesced and have been filled with thick sedimentary deposits and hence retain insights into orogenic, erosional and geodynamic processes (e.g., Alonso et al., 1990; Meyer et al., 1998; Sobel et al., 2003;, Strecker et al., 2009; Carrol et al., 2010; Horton et al., 2012; Pingel et al., 2019). Plateau’s building models predict that reduced fluvial connectivity promotes basin filling, inhibits intrabasinal faulting, and triggers the outward propagation of the deformation fronts. Combined, these processes are thought to be responsible for the lateral (orogen perpendicular) plateau expansion through the integration of new sectors of the foreland into the plateau realm. (Sobel et al., 2003; Garcia Castellanos et al., 2007). The application of these models, however, is not straightforward mostly because the interplay between tectonic and surface processes may trigger different scenarios. This includes basin excavation and erosion with the destruction of the typical plateau morphology (e.g., Strecker et al., 2009; Heidarzadeh et al., 2017). Therefore, while the sedimentary basins in the plateau interior are tectonically stable up to time scales of few 10^7 years (e.g., Alonso et al., 1990; Bush et al., 2016), intermontane basins at the transition with the foreland may experience a more complex evolution including several episodes of basin filling and plateau integration, fluvial incision and tectonic deformation at shorter time scales (10^5 to few 10^6 years; e.g., Streit et al., 2015; Schildgen et al., 2016; Tofelde et al., 2017; Ballato et al., 2019; Pingel et al., 2019). Thus, these transitional basins hold precious information on the growth of the plateau margin, the evolution of adjacent mountain ranges, the sediment routing systems and the connectivity history among different sedimentary basins.

The NW-SE-oriented Iranian Plateau (IP) is located on the upper plate of the Arabia-Eurasia collision zone and represents the second collisional plateau in elevation and size after Tibet (see Hatzfeld & Molnar, 2010 for a comparison). The IP is parallel to the Zagros orogenic belt and is characterized by high elevation (average elevation is ~1800 m), low internal topographic relief (few hundred of meters), dry climatic conditions, endorheic sedimentary basins in its interior (four out of six basins are internally drained), and steep and dissected flanks bounded by major reverse faults (Ballato et al., 2013, 2017 Heidarzadeh et al., 2017). In central Iran, the northern margin of the IP is marked by a sharp boundary with the adjacent foreland, which comprises the rigid Central Iranian Block (Figure 1). In NW Iran, the IP approaches the Caspian Sea and it is separated from the intracontinental Alborz and Talesh mountains by an elongated, NW-SE oriented intermontane basin called Tarom Basin. Currently, this basin is drained by the Qezel-Owzan River, the second largest river in Iran that flows from the interior of the IP to the Caspian Sea. The basin is composed of post Eocene, synorogenic red beds that offer the opportunity to investigate puzzling aspects of this collision zone, such as: the timing and mechanisms of plateau margin uplift, its lateral expansion (i.e., the possible incorporation of the intermontane Tarom Basin in the plateau realm) and the link with the adjacent growing Alborz Mountains. For this purpose, we have performed a multidisciplinary study including the characterization of the depositional environments, the sediment provenance areas and the depositional age of the post Eocene synorogenic red beds. Our magnetostratigraphic analysis and new zircon U-Pb ages, document that the widespread Eocene arc volcanism terminated at ∼ 38-36 Ma, while the deposition of the red beds occurred from ∼16.5 Ma to at least ∼7.6 Ma during the growth of the adjacent basin margins. Further, we document the occurrence of alternating periods of efficient and limited fluvial connectivity and we discuss the mechanisms that may have led to the growth of the IP margin in this sector of the Arabia-Eurasia collision zone.

1.1. Geological setting

The Tarom Basin is a NW-SE oriented, elongated, intermontane basin located along the northern margin of the Iranian Plateau between the western Alborz Mountains to the NE and the Tarom range to the SW (Arabia-Eurasia collision zone; Figure 1).

The western Alborz Mountains consist of Pre-Cambrian crystalline basement rocks, Paleozoic and Mesozoic marine deposits, Eocene volcanics, volcaniclastics and intrusives of variable age (Figure 1). This assemblage indicates a complex history of deformation, exhumation, metamorphism, magmatism, subsidence and sedimentation that includes: development of a metamorphic basement during the Neoproterozoic Pan-Africa Orogeny (e.g., Guest et al., 2006; Hassanzadeh et al., 2008), deposition of unconformable carbonate and clastic marine deposits of Pre-Cambrian and Paleozoic age associated with the opening the Paleo-Tethys Ocean (e.g., Horton et al., 2008), occurrence of the Triassic Cimmerian Orogeny (e.g., Zanchi et al., 2009; Omrani et al., 2013), renewed Mesozoic subsidence with the sedimentation of post-orogenic clastic sediments of the Shemshak Formation (e.g., Zanchi et al., 2009; Wilmesen et al., 2009), deposition of shallow- to deep-marine Middle to Late Jurassic sediments during the opening of the South Caspian Basin (e.g., Brunet et al., 2003), Cretaceous thermal subsidence and marine sedimentation (Brunet et al., 2003), Late Cretaceous to Paleocene deformation and exhumation during a regional compressional event (e.g., Guest et al., 2006; Yassaghi & Madanipour, 2008; Madanipour et al., 2017), deposition of Eocene volcaniclastics in a backarc system associated with the rollback of the Neo-Tethyan oceanic slab (Guest et al., 2006; Ballato et al., 2011, 2013; Verdel et al., 2011; Rezaeian et al., 2012) and finally, contractional deformation and exhumation during the closure of the Neo-Tethys ocean and the collision between Eurasia and Arabia starting from the latest Eocene-earliest Oligocene (e.g., Guest et al., 2006; Ballato et al., 2011, 2013, Rezaeian et al., 2012; Mouthereau et al., 2012; Madanipour et al., 2017, 2018; Pirouz et al., 2017; Koshnaw, et al., 2018). This final event led to development of a narrow, double-verging mountain belt with over 3 km of topographic relief that represents an effective orographic barrier to moist air masses sourced from the Caspian Sea (Figure 1; Ballato et al., 2015). Available low-temperature thermochronology data document slow exhumation from the Early Oligocene followed by an acceleration during the last 12 Ma (Madanipour et al., 2017). Currently, the range accommodates left-lateral shearing between the Caspian Sea and Central Iran (Djamour et al., 2010) and is characterized by the occurrence of few seismogenic faults including the Rudbar Fault, which ruptured in 1990 leading to the catastrophic Mw 7.3 earthquake (Berberian & Walker, 2010). The Tarom range consists of a ~ 4-km-thick pile of Eocene volcanic and volcanoclastic rocks of the Karaj Formation (Figures 1 and 2; Stocklin, J., Eftekharnezhad, J., 1969) that were deposited in the backarc of the Neo-Tethys subduction zone between ~ 55 and 38-36 Ma (Guest et al., 2006; Ballato et al., 2011, 2013; Verdel et al., 2011; Rezaeian et al., 2012). This was associated with the emplacement of Late Eocene (~ 41 to 37 Ma) shallow intrusive rocks (Nabatian et al., 2014). In the Tarom range these deposits form a broad, south-verging anticline (Heidarzadeh et al., 2017) with smaller scales anticline-syncline pairs (Figure 2), cut by minor high angle (both south and north dipping) reverse faults, locally with a lateral component. Available low-temperature thermochronology data indicate that uplift and exhumation of the Tarom range could have started around the latest Eocene-earliest Oligocene and resumed during the last ~ 10 Ma (Rezaeian et al., 2012).

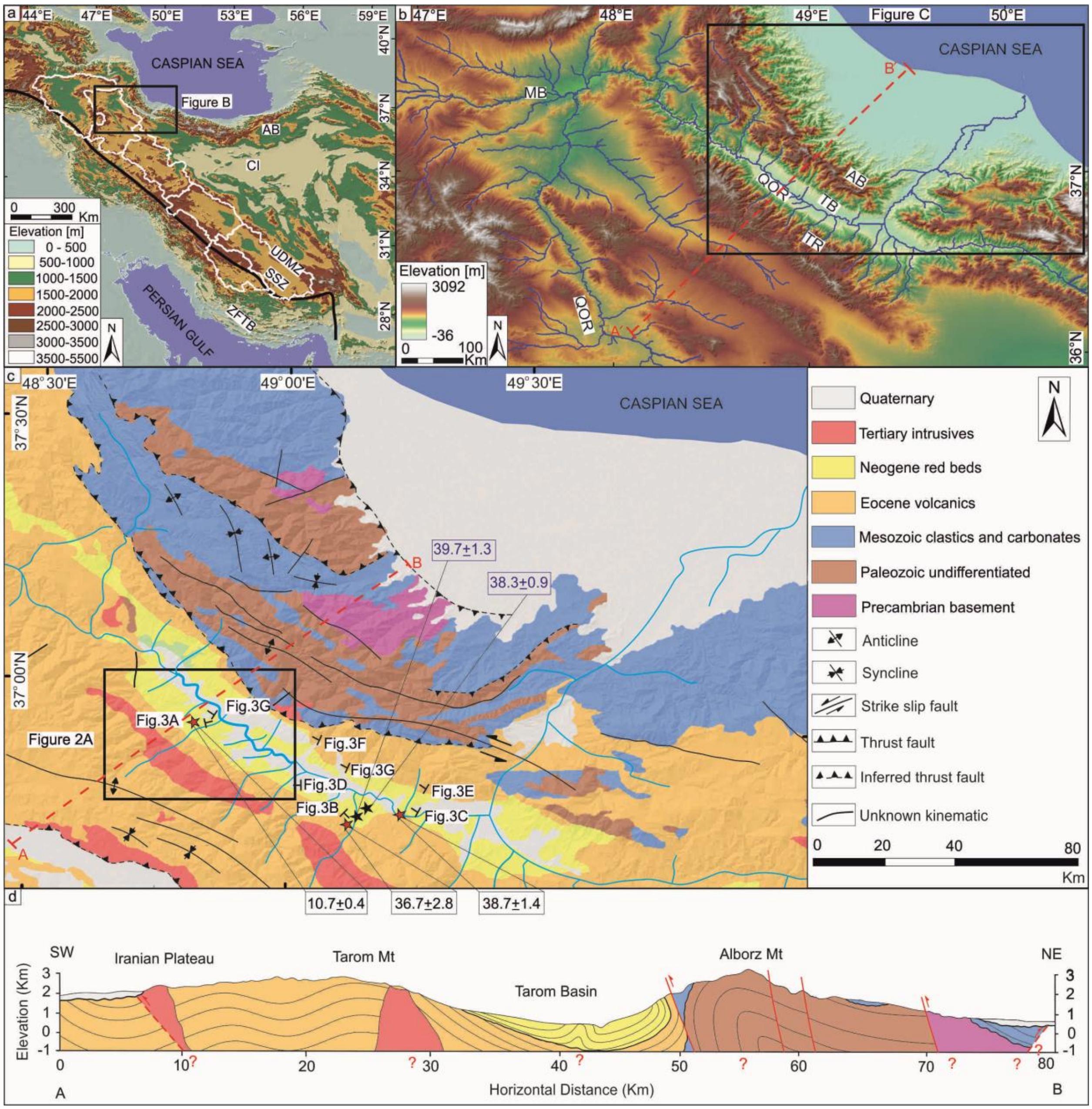

Figure 1. (a) Shuttle Radar Topographic Mission Digital Elevation Model (SRTM DEM) of Iran showing the Iranian Plateau; the white polygons indicate six main drainage basins forming the Iranian Plateau while the black line shows the approximate location of the suture zone, which separates the lower Arabian plate (and the Zagros Fold and Thrust Belt; ZFTB) from the upper Eurasian plate (Ballato et al., 2017). The Urumieh Doktar Magmatic Zone

(UDMZ) and the Sanandaj Sirjan Zone (SSZ) represent the backbones and the margins of the plateau, respectively. (b) DEM of NW Iran showing the Mianeh Basin (MB), Tarom Basin (TB) and its bounding Tarom range (TR) and Alborz Mountains (AB), along the southern and the northern margins of the basin, respectively. Note the Qezel-Owzan River (QOR) drainage system (~ 55000 km^2) connect the Iranian Plateau and the Caspian Sea through the Tarom Basin. A-A′ line shows the approximate location of the crustal scale section shown in figure 14c. (c) Simplified geologic map of NW Iran (Stocklin and Eftekharnezhad, 1969; Davies, 1977) showing the location of the panoramic field photographs of figure 3. The red stars show the location of our new zircon U-Pb ages (expressed in Ma); the black stars (and blue ages) represent reworked Eocene volcanic material within red beds that do not provide information on their depositional age. (d) Regional geological cross section (modified after Stocklin et. al, 1969).

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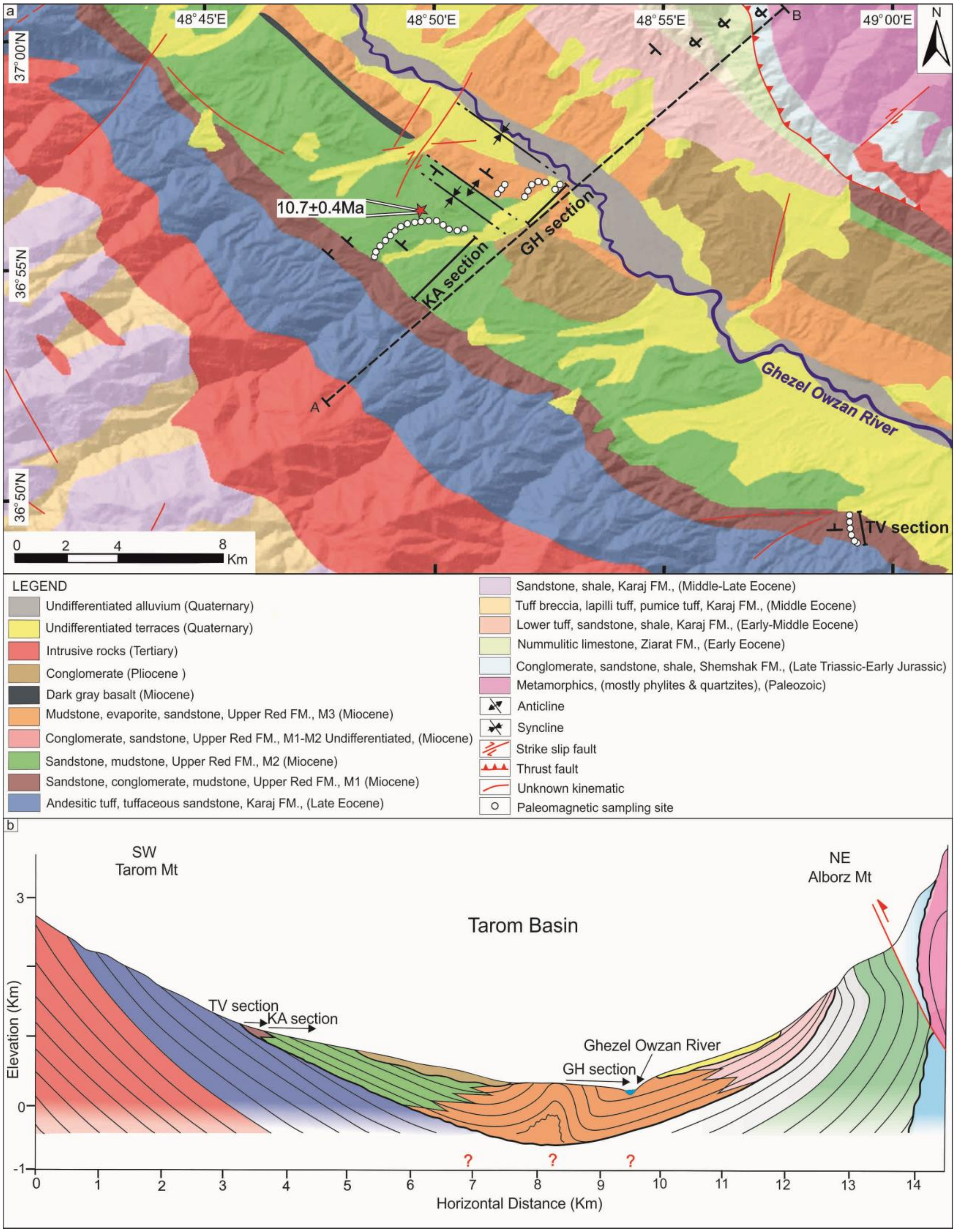


a
48°45'E
48°50'E
48°55'E
49°00'E
37°00'N
36°55'N
36°50'N
N
B
A
10.7±0.4Ma
KA section
GH section
TV section
Ghezel Owzan River
0 2 4 8 Km
LEGEND
Undifferentiated alluvium (Quaternary)
Undifferentiated terraces (Quaternary)
Intrusive rocks (Tertiary)
Conglomerate (Pliocene)
Dark gray basalt (Miocene)
Mudstone, evaporite, sandstone, Upper Red FM., M3 (Miocene)
Conglomerate, sandstone, Upper Red FM., M1-M2 Undifferentiated, (Miocene)
Sandstone, mudstone, Upper Red FM., M2 (Miocene)
Sandstone, conglomerate, mudstone, Upper Red FM., M1 (Miocene)
Andesitic tuff, tuffaceous sandstone, Karaj FM., (Late Eocene)
Sandstone, shale, Karaj FM., (Middle-Late Eocene)
Tuff breccia, lapilli tuff, pumice tuff, Karaj FM., (Middle Eocene)
Lower tuff, sandstone, shale, Karaj FM., (Early-Middle Eocene)
Nummulitic limestone, Ziarat FM., (Early Eocene)
Conglomerate, sandstone, shale, Shemshak FM., (Late Triassic-Early Jurassic)
Metamorphics, (mostly phylites & quartzites), (Paleozoic)
Anticline
Syncline
Strike slip fault
Thrust fault
Unknown kinematic
Paleomagnetic sampling site
b
SW
Tarom Mt
NE
Alborz Mt
Tarom Basin
TV section
KA section
Ghezel Owzan River
GH section
Elevation (Km)
3
0
-1
? ? ?
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14
Horizontal Distance (Km)

Figure 2. (a) Geologic map (Amini, 1969) superimposed on a SRTM hillshade model of the study area (TB). The white circles show the location of the three sections sampled for magnetostratigraphy named TV, KA and GH. The base of section G is also visible in figure 3h. (b) Geologic cross section across the Tarom Basin.

1.3. Regional stratigraphy

The Tarom Basin was filled by post Eocene red beds that rest in angular unconformity onto Eocene volcanics and volcaniclastics of the Karaj Formation (Figure 2). The stratigraphic position of the red beds is unknown, mostly because the Late Oligocene-Early Miocene marine transgression that led to the widespread deposition of the shallow-water marine limestones of the Qom Formation (Reuter et al., 2009) did not reach the Tarom Basin. These marine deposits are sandwiched between the clastic deposits of the Lower Red (LRF; Oligocene) and Upper Red (URF, Miocene) formations and represent a regional marker that can be followed along the southern margin of the Eurasian plate. Therefore, their absence, does not allow differentiating the stratigraphic position of the red beds exposed in the Tarom Basin, which have been considered either Neogene (Stocklin and Eftekharnezhad, 1969; Davies, 1977) or Miocene in age (Figures 1 and 2; Amini, 1969).

The LRF and the URF are exposed virtually everywhere along the southern margin of the Eurasian plate, where they have a thickness varying from few hundreds to few thousands of meters. These red beds are characterized by a variable amount of sandstones, conglomerates, mudstones, evaporites and locally volcanics, and are mostly considered synorogenic sediments associated with collisional deformation (e.g., Morley et al., 2009; Ballato et al., 2008, 2011, 2017; Rezaeian, et al., 2012; Madaniopour et al., 2017). Lithologically, the LRF is rather heterogeneous, while the URF seems to have more uniform characteristics, and hence has been differentiated into 3 Units (M1, M2 and M3; e.g., Davoudzadeh et al., 1997). Units M1 and M3 are generally dominated by mudstones and evaporites with a variable amount of sandstones and conglomerates while Unit M2 is characterized by abundant sandstones. The URF is superseded by supposed Pliocene conglomerates (Hezadarreh Formation; Rieben et al., 1955) that are generally thought to mark an intensification of collisional deformation (e.g., Rezaeian, et al., 2012; Madaniopour et al., 2017). These conglomerates, however, are diachronous and their age depends on their position with respect to the coeval active mountain fronts. For example, in the southern Alborz Mountains (Ballato et al., 2008) and in the interior of the Iranian plateau (Tavaq Conglomerates, Great Pari Sedimentary Basin; Ballato et al., 2017) conglomeratic deposition started at ~ 7.5 and ~ 10.7 Ma, respectively.

1.4. Stratigraphic and structural setting of the Tarom Basin

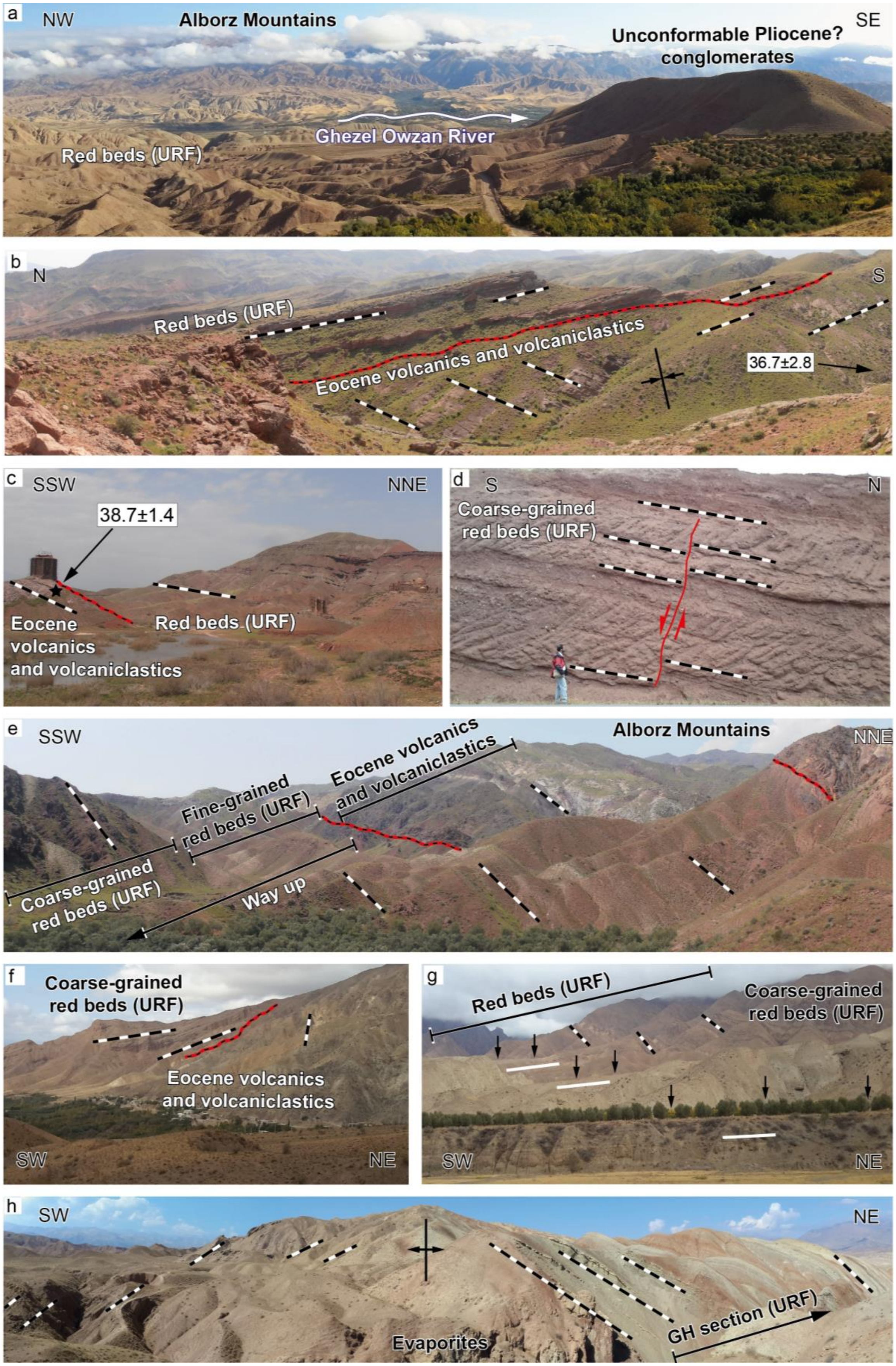
The red beds of the Tarom Basin consist of coarse- to medium-grained clastic deposits passing laterally toward the basin axis to finer grained sediments and evaporites (Figure 3b). The minimum thickness of the basin-fill sediments observable in the field in the central sectors of the basin is about 1185 m, while the lack of major intrabasinal unconformities within the red beds suggests that sedimentation was rather continuous. In some parts, the red beds are unconformably covered by gently deformed, conglomerates of supposed Pliocene age (Figure 3a). Furthermore, at least three generations of terrace conglomerates can be observed in the field, suggesting the occurrence of recent phases of sediment aggradation and fluvial incision (Figure 3g).

Along the southern margin of the basin, the red beds dip few degrees toward the NE (up to 20°), while the underlying volcanics are generally steeper (Figure 3c) and can be locally folded (Figure 3b). In addition, the southern margin of the basin is characterized by several subvertical synsedimentary normal faults (Figure 3d), mostly parallel to the strike of the basin, that provide evidences for localized extension sub-parallel to the regional shortening direction (NE-SW; Madanipour et al., 2017). These faults are not linked to major extensional events and hence did not control the basin-scale subsidence pattern (Paknia, 2019; PhD thesis; see chapter III).

Along the northern side of the basin, the setting is more variable and complex, and the Eocene deposits of the Karaj Formation are either sub-vertical or overturned. In the central-southern sectors of the basin, the unconformable red beds are also subvertical to overturned (Figure 3e) and exhibit a rapid shallowing upward trend suggesting the occurrence of growth strata. Conversely, in the central-northern sectors of the basin the angular unconformity is more pronounced, and the red beds dip less than 30° to the south-west (Figure 3f). There, we do not have evidences for syndepositional contractional deformation.

The central sectors of the basin are also characterized by several upright syncline-anticlines pairs, subparallel to the strike of the basin with a lateral extent of few kilometers (Figure 2). Figure 3h shows the core of one of these anticlines which is characterized by evaporites layers that have been deformed in a disharmonic manner and may have acted as local decollement horizon.

Currently, the basin is drained by the ~800 km long Qezel-Owzan River (QOR), which is flowing from the elevated Iranian Plateau to the Caspian Sea (Figure 1). The connection between the interior of the Iranian Plateau, the Tarom Basin and the Caspian Sea occurs through a serious narrow bedrock gorges suggesting a protracted history of internal drainage conditions followed by fluvial captures (Heidarzadeh et al., 2017). In particular, the connectivity between the Tarom Basin and the Iranian Plateau must have been established during the last 4 Ma through lake overspill as suggested by the stratigraphic record of a sedimentary basin in the plateau interior (Mianeh Basin, Figure 1b; Heidarzadeh et al., 2017).



a
NW
Alborz Mountains
SE
Unconformable Pliocene? conglomerates
Ghezel Owzan River
Red beds (URF)
b
N
S
Red beds (URF)
Eocene volcanics and volcaniclastics
36.7±2.8
c
SSW
NNE
38.7±1.4
Eocene volcanics and volcaniclastics
Red beds (URF)
d
S
N
Coarse-grained red beds (URF)
e
SSW
Alborz Mountains
NNE
Eocene volcanics and volcaniclastics
Fine-grained red beds (URF)
Coarse-grained red beds (URF)
Way up
f
Coarse-grained red beds (URF)
Eocene volcanics and volcaniclastics
SW
NE
g
Red beds (URF)
Coarse-grained red beds (URF)
SW
NE
h
SW
NE
Evaporites
GH section (URF)

233

Figure 3. Panoramic field photographs (see figure 1 for location) highlighting the main geometrical relationships among the units and formations exposed in the Tarom Basin. (a) Northeast-facing photo showing conglomerates supposed Pliocene age in unconformity onto deformed red beds; the conglomerates are tilted to the NNE and have a dip angle of ca. 25°. On the foreground the mountain front of the Alborz Mountains with several generation of terraces is visible (see figure g for details). (b and c) Southeast- and northwest-facing photos documenting the unconformity (red and black line) between the Karaj Formation and the red beds in the southern margin of the basin. Black and white dashed lines show the bedding while the zircon U-Pb ages reported are in Ma (see Table 3 and figure 1). (d) Synsedimentary normal fault exposed along the TV sections (Paknia, 2019; PhD thesis; see chapter III). (e and f) Northwest-facing photos documenting the unconformity (red and black line) between the Karaj Formation and the red beds in the southern margin of the basin. Note that in figure e the red beds are overturned. (g) West-facing photo displaying three major terrace conglomerates (see black arrows); these deposits are virtually undeformed (white lines) and cover in unconformity steeply dipping red beds (black and white dashed lines). (h) Northwest-facing photo showing the core of the anticline that represents the base of the stratigraphic section GH.

2. Material and methods

To unravel the basin-fill history of the Tarom Basin and its tectono-stratigraphic evolution in the framework of collisional deformation and plateau building processes, we performed a multidisciplinary study including:

1) A detailed sedimentologic analysis that provides the basis for an assessment of the depositional environments (Tables 1, 2 and 3; see section 3)

2) A geochronologic study (U-Pb on zircons) of the uppermost volcanic of the Karaj Formation and the red beds that combined with (see section 4)

3) A paleomagnetic and magnetostratigraphic analyses provides a chronostratigraphic framework for the Late Cenozoic basin-fill sediments (see section 4)

4) A provenance study (sandstone petrography and paleocurrent analysis; see section 4), which allows identifying compositional variations related to the exposure of new sediment sources and/or drainage-pattern reorganizations in the sediment source area (Detailed information about the analytical methods are provided in the Appendix section). This approach was employed on two stratigraphic sections exposed along the southern margin of the basin (TV and KA sections; Figure 2) and on a third one located in the northern limb of a north-vergent anticline in the central sectors of the basin (GH section; Figure 2). These sections are stratigraphically continuous and are not affected by major faults, therefore they represent an ideal setting for magnetostratigraphic sampling. Furthermore, recent papers from Central and Northern Iran have shown that the Late Cenozoic red beds have good magnetic properties and hence are suitable for paleomagnetic analysis (Ballato et al., 2008, 2017; Cifelli et al., 2015; Mattei et al., 2015, 2017, 2020). The red beds exposed along the southern basin margin (TV and KA section) are tilted northward with a dip angle of 14 to 30^{o}, whereas in the central sectors of the basin (GH section) strata are steeply dipping to the north (and occasionally overturned) with a dip angle of 40 to 88^{o}. The stratigraphic sections along the southern margin cover the lowermost stratigraphic interval of the basin fill and consist mainly of reddish or light brownish conglomerates with intercalations of mudstone and fine-grained sandstone layers evolving up section into channelized sandstones with conglomerate lenses (fluvial channels, see next section) and finer-grained sediments with tabular geometries (flood plain deposits, see next section). The stratigraphic section in the central sectors of the basin consists mainly of reddish, greyish and brownish mudstones, thin bedded sandstones and evaporates layers, locally with intercalations of conglomerates lenses, which become more abundant toward the top of the section.

3. Depositional systems in the Tarom Basin

Based on our field observations (lithological characteristics, lateral and vertical grain size variations, sedimentary structures and geometry of the sedimentary bodies) and according to the classification scheme of Miall (1985; 1996), we established a total of eighteen lithofacies types (Table 1 and Figure 4) and recognized eight facies associations (Table 2 and Figure 5). The combination of the facies associations led to the reconstruction of four depositional environments (alluvial fan, braided river, playa-lake and lacustrine settings; Figure 5). In the following, we describe the main characteristics of these depositional settings.

3.1. Alluvial fan system

Alluvial fan deposits (Figures 5a and 5b) are located along both margins of the Tarom Basin and include two facies associations: (1) disorganized granule-boulder conglomerate (G1; Figures 4a and 5a), and (2) moderately to well organized granule-boulder conglomerate (G2; Figures 4b and 5b). We interpret the G1 facies association with weakly developed clast imbrications and erosive basal contacts as high-energy stream-floods equivalent to those produced by gravel-laden streams or sediment gravity flow deposits (hyperconcentrated and turbulent flow) in poorly confined channels (Figure 4a and 5a; e.g., Maizels, 1989; Stanistreet & McCarthy, 1993; Ridgway & DeCelles, 1993; Miall, 1996; Blair, 1999). The beds geometry suggests the occurrence of sheet flows (Hein, 1982) with limited development of longitudinal bars (Boothroyd & Ashley, 1975; Todd, 1989). The G2 facies association is interpreted as traction-current deposits in poorly confined channels under conditions of higher bed shear stress (Figures 4b and 5b; e.g., Stanistreet and McCarthy, 1993; Miall, 1996; Blair, 1999; Ballato et al., 2011).

3.2. Braided fluvial system

The braided river deposits (Figures 5c, 5d and 5e) are characterized by four facies associations: (1) well-organized granule-pebble conglomerate (G3), (2) sandstone (S), (3) interbedded fine-grained sandstone and mudstone (SM), and (4) evaporite (E). The G3 facies association is interpreted to reflect traction-current deposits (longitudinal bars or lag deposits) related to the waning stage of high-energy flow in a laterally confined system (e.g., Stanistreet & McCarthy, 1993; Miall, 1996; Blair, 1999). The erosive basal contact, together with the lens geometry and the interfingering with stratified sandstones suggests deposition in a braided channel with a variable proportion of gravel and sand (Figures 4c and 5c; e.g., Miall, 1996). The S facies association is interpreted to represent deposition in lower and upper plane-bed flow regimes in a confined flow (e.g., Miall, 1996). Planar (Sp) and trough cross-stratified (St), medium to coarse-grained, pebbly sandstones are interpreted as migrating bedforms (fluvial dunes) in a confined flow in an upper to lower flow regime (Figure 4c; Uba et. al, 2005; Siks & Horton, 2011). Overall, these observations indicate deposition in fluvial channel. The SM facies association (Figure 5d) includes sandstones with cross (Sr; Figure 4d) and planar lamination (Sh and Sl; Figure 5d) that are interpreted as sheet-flow deposits in a poorly confined to unconfined flow evolving from the upper flow regime to a waning flow stage. The SM facies association includes also massive to parallel laminated mudstones (Fm and Fl; Figure 4f), which can be locally dominant and are interpreted to represent suspension fallout deposits (e.g., Ghibaudo, 1992) from standing or slowly moving waters in the floodplain (e.g., Miall, 1977 and 1978). Locally, the SM facies association are characterized by the development of carbonate nodules and rizholithes indicating paleosols formation (Figure 4g) during lengthy pauses in sedimentation or slow sedimentation rates (e.g., Kraus, 1999). The occasional occurrence of E facies association (Ev;

Figure 4h) is interpreted to represent precipitation of salt minerals from concentrated water solution after evaporation of standing water in the floodplain. Complete desiccation of standing water is also documented by mud cracks (e.g., Lowenstein & Hardie, 1985).

Finally, in the KA stratigraphic section in proximity of the southwestern basin margin we found, embedded in the fluvial deposits, the BD facies association. This disorganized package of blocks with different size and sediments of variable grain size is interpreted as landslide deposits (sturzstrom) caused by gravitational collapse of the adjacent mountain front (e.g., Hermanns & Strecker 1999; Paknia, 2019; PhD thesis; see chapter III, see also Table 2 this work). This interpretation is further supported by the occurrence of a clay-reach sheared basal contact and the presence of a dense and irregular network of fractures (jigsaw cracks).

3.3. Lacustrine system

The lacustrine system is located along the central sectors of the basin (Figures 4e, 4f, 5f and 5f; section GH) and is characterized by two facies associations: (1) mudstone (M) and (2) interbedded fine-grained sandstone and mudstone (SM). Tabular bodies of laminated mudstone of the M facies association are typical of suspension deposits in a lacustrine offshore setting and indicate a deepening of the system (Figure 4f). Lenses of fine grained-sandstone with symmetrical ripple marks interbedded with mudstone (lenticular and waving bedding Figures 4e and 5f) in the SM facies association indicate deposition in the lacustrine shoreface-offshore transition. In few sectors of the GH stratigraphic section, the tabular sandstones with symmetric ripples become dominant suggesting sedimentation in the lacustrine shoreface (e.g., Horton & Schmitt, 1996; llgar & Nemec, 2005; Chakraborty & Sarkar, 2005; Keighley, 2008; Ghinassi et al., 2012). These intervals, however, are relatively rare and generally have a limited thickness (<

1 m), therefore most of the lacustrine sediments exposed in the section were deposited either in the offshore or in the shoreface-offshore transition setting.

3.4. Playa lake system

The playa lake system is also located in the central sectors of the basin where it alternates with the lacustrine setting (GH stratigraphic section, Figures 4h and 5h). These deposits include two facies associations such as (1) mudstone (M) and (2) evaporite salt minerals (E). The first facies association (mudstone; M) is interpreted to represent deposits settled from suspension in arid to semiarid, oxidizing conditions as documented by the presence of red coloured sediments and the occurrence of desiccation cracks (e.g., Lowenstein & Hardie, 1985). The second facies association (E) is interpreted to represent evaporite layers (mostly gypsum) precipitated during short-lived rain episodes followed by desiccation. Overall, these observations suggest that sedimentation occurred in a shallow playa lake setting.

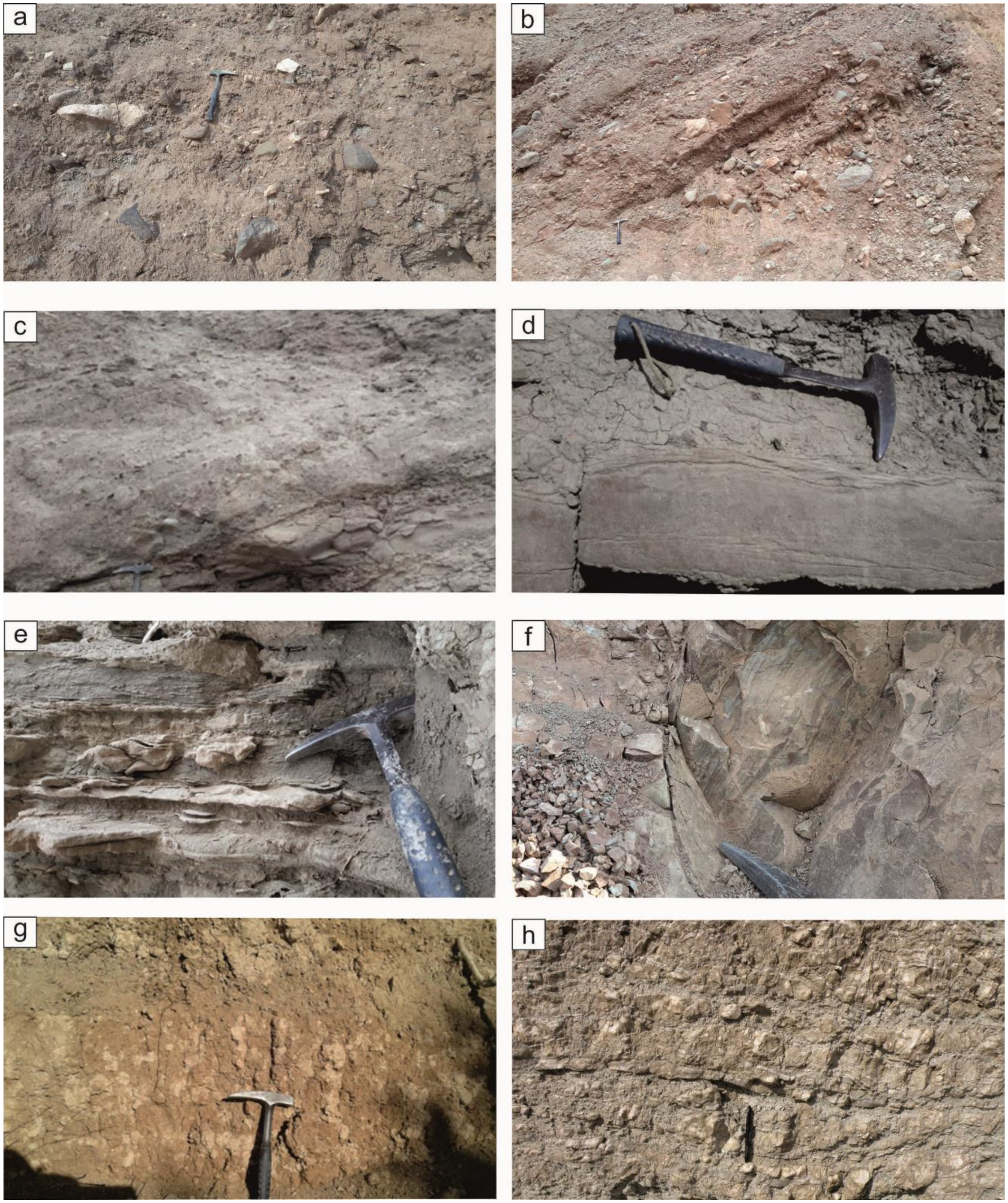


Figure 4. Close up view photographs of lithofacies characteristics. (a) Disorganised, structureless, matrix-supported, mostly monomictic (clasts are Eocene volcanics) conglomerate with subangular to angular clasts reflecting mass flow deposits (Facies code Gmd). (b) Disorganised, structureless, clast-supported, mostly monomictic conglomerate

with crude bedding and subangular to moderately rounded clasts (stream-flood deposits; Gcd). (c) Conglomerates and coarse-grained sandstones with planar cross bedding representing traction current bedforms (Gp and Sp, respectively). (d) Horizontally laminated sandstone (Sl) and rippled sandstone (Sr) indicating traction currents of variable energy in sandy dominated system. (e) Lenticular bedding with symmetrical rippled sandstone (Smw) alternated with laminated mudstone (Fl) reflecting an alternation of current (bidirectional) and suspension deposits. (f) Massive structureless (Fm) to finely laminated (Fl) calcareous mudstone (suspension deposits). (g) Mudstone with carbonate nodules (P) indicating paleosol formation. (h) Evaporate deposits (Ev) reflecting evaporation from standing water.

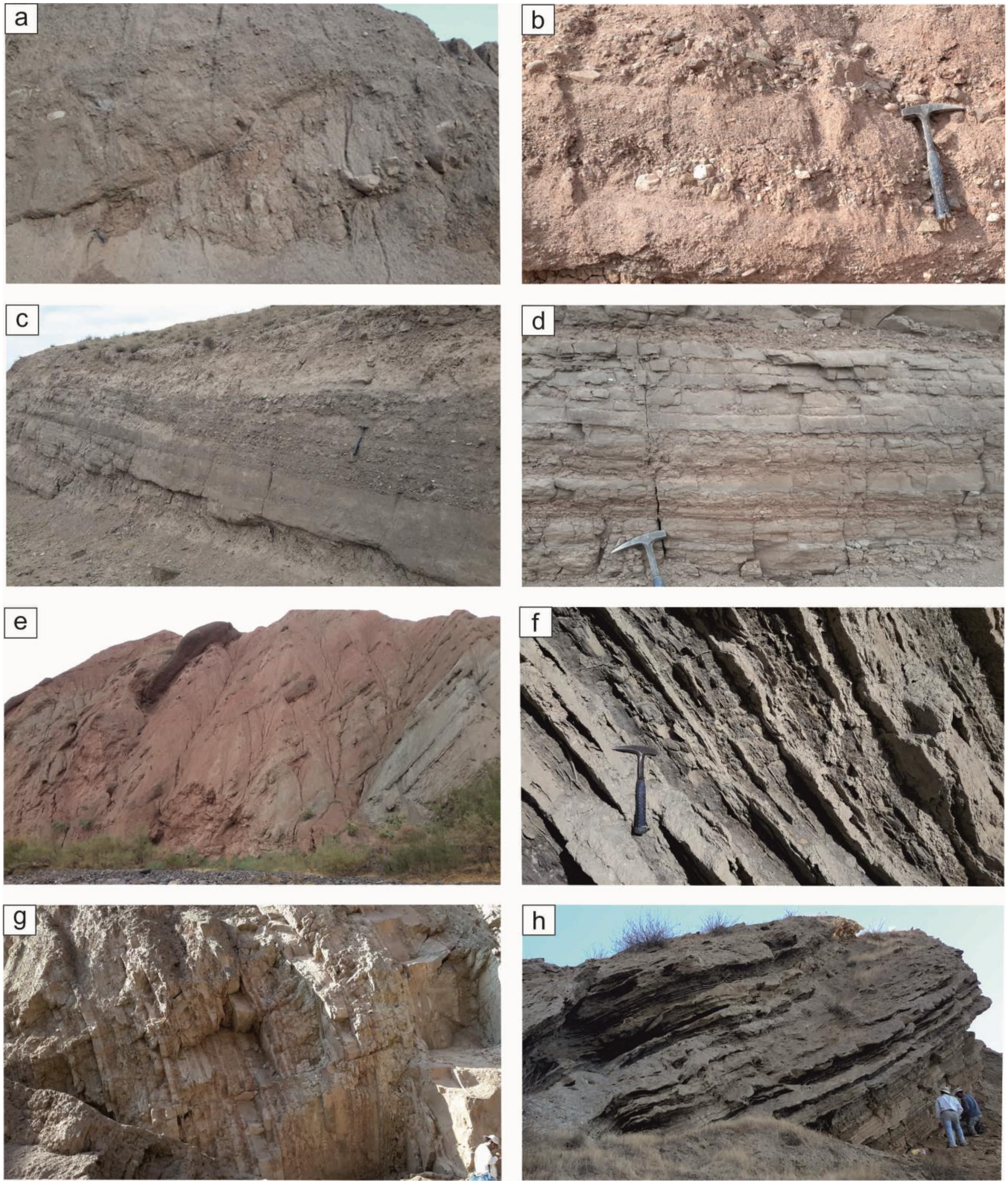


Figure 5. Representative views of different depositional systems in the Tarom Basin. (a) Disorganized granule-boulder conglomerate (facies association G1; base of KA stratigraphic section) and (b) moderately to well organized granule-boulder conglomerate (facies association G2; KA stratigraphic section) representing an alluvial fan setting.

(c) Horizontally to trough cross-stratified pebbly sandstone and conglomerate in a fluvial channel (facies association S; KA stratigraphic section), of a braided river system. (d) Horizontally, thin bedded, fine grained sandstone and laminated mudstone sheets (facies associations SM; KA stratigraphic section) representing flood plain deposits of the braided river system. (e) Overview of the braided river system with lenses of conglomerate and coarse-grained sandstone (facies association S and G3) embedded in flood plain deposits (facies associations SM; top of GH stratigraphic section). (f) Fine grained sandstone and mudstone deposits with flat geometry (facies association SM; GH stratigraphic section) reflecting deposition in the shoreface-offshore transition in a lacustrine depositional setting; the sandstone layers indicate distal storm beds. (g) Alternation of mudstone and fine-grained sandstone deposit with flat to tabular geometry (facies association SM; base of GH stratigraphic section; lacustrine depositional setting); when the mudstone dominates deposition occurred in the offshore setting, otherwise the alternation of mudstone and sandstone indicates deposition in the shoreface-offshore transition. (h) Gypsum layers (Evaporite deposits) precipitated during short-lived desiccation episodes (facies association E, GH stratigraphic section), representing a playa lake depositional setting.

Table 1

Description and Interpretation of Lithofacies

Facies code	Characteristics	Interpretation
Gmd	Disorganised, structureless, matrix-supported, mostly monomictic conglomerate. Granules to boulders, subangular to angular clasts. Maximum clast diameter 40 cm	Mass flows deposits from hyperconcentrated or turbulent flow
Gcd	Disorganised, structureless, clast-supported, mostly monomictic conglomerate with crude bedding. Granules to boulders, subangular to moderately rounded clasts. Maximum clast diameter 40 cm	Stream-floods deposits with concentrated clasts
Gco	Moderately organized, clast supported, monomictic to polymictic conglomerate. Granules to cobbles, subangular to rounded clasts, normal grading, and weak imbrication. Maximum clast diameter 20 cm	Traction bedload deposits
Gh	Clast-supported, horizontally bedded, monomictic to polymictic conglomerate. Granules to pebbles, subrounded to well-rounded clasts, normal to inverse grading with imbrication. Maximum clast diameter 5 cm	Traction current bedforms (bars)

Gt	Clast-supported, trough cross-stratified, monomictic to polymictic conglomerate. Granules to pebbles, subrounded to well-rounded clasts, normal grading. Maximum clast diameter 5 cm	Traction current bedforms (bars)
Gp	Clast-supported planar cross-stratified monomictic to polymictic conglomerate. Granules to pebbles, subrounded to rounded, normal grading. Maximum clast diameter 5 cm	Traction current bedforms (bars)
Br	Matrix supported, structureless monomictic breccia. Granules to boulders, very angular clasts, inverse grading. Maximum clast diameter 1 m	Rock avalanche deposits (sturzstrom)
Sp	Planer cross-stratified sandstone. Medium to coarse grain size, moderately to well sorted occasionally with pebbles	Dune migration during upper to lower flow regime
Sl	Horizontally laminated sandstone. Very fine to medium grain size, well sorted occasionally with pebbles	Bedforms deposited under upper to lower flow regime
Sr	Rippled sandstone (asymmetric ripples). Very fine to medium grain size, well sorted	Ripples under lower flow regime
Sh	Horizontally stratified sandstone. Very fine to coarse grain size, moderately to well sorted, occasionally with pebbles	Planar bed flow during upper flow regime
St	Trough cross-stratified sandstone. Medium to coarse grain size moderately to well sorted, occasionally with pebbles	Dune migration during upper to lower flow regime
Smw	Rippled sandstone (symmetrical ripples). Fine to medium-grain size well sorted	Wave (bidirectional current) deposits
Fm	Massive structureless calcareous mudstone	Suspension deposits, overbank or abandoned channel
Fl	Finely laminated calcareous mudstone. Flat parallel lamination, small-scale ripples, locally with mud cracks	Suspension deposits, overbank or abandoned channel
Mr	Sheared reddish clay with unsorted angular clasts	Shearing stress at the base of a rock avalanche
P	Mudstone to fine-grained sandstone with carbonate nodules	Paleosol formation
Ev	Evaporites, locally associated with gypsum-filling fractures	In situ accumulation during evaporation of standing water

Table 2

Description, Lithofacies, Architectural Elements, and Interpretation of Depositional Processes and Environments of Facies Association

Facies association	Description	Lithofacies	Architectural elements	Interpretation of depositional process	Depositional setting
G1 (disorganized granule-boulder conglomerate)	Structureless to poorly organized, matrix- to clast-supported conglomerate. Beds 0.2 to1 m thick with lateral extent of few tens of meters and a planar to slightly erosive basal contacts. Interbedded with facies associations G2 and G3	Gmd, Gcd	Gravel sheets and poorly confined channels	Sediment gravity-flow deposits	Alluvial-fan system
G2 (moderately to well organized granule-boulder conglomerate)	Moderately to well-organized, clast-supported, ungraded to normally graded, moderately to poorly sorted, poorly imbricated conglomerate. Moderate to poor horizontal and trough cross-stratification. Beds 0.2- to 1-m-thick with a lateral extent of few tens of meters and a slightly erosive basal contact. Interbedded with facies associations G1, G3, S and SM	Gco, Gh, Gt	Gravel sheets, and gravel downstream accretion macroforms (bars)	Traction bedload deposits in a gravel-dominated, poorly confined channel or in a gravel sheet	Alluvial-fan system
G3 (well organized granule-cobble conglomerate)	Well organized, clast-supported, channelized, horizontally, planar and trough cross-bedded, moderately to well sorted, conglomerate with slightly erosional contacts and a lateral extent of up to tens of meters. Interbedded with facies associations S, G2, SM, and rarely M	Gco, Gp, Gh, Gt, Sh, St, Sp	Channel-fill complex and gravel bedforms (gravel bars and lenses)	Traction bed load deposits in a gravel-dominated, well-confined channel	Alluvial-fan and proximal fluvial system
DB (Disorganized, granules to boulder breccia)	Chaotic, matrix supported, poorly sorted breccia with a sheared clay basal contact and few tens of meters lateral extent	Br, Mr	Probably lobate (full geometry not exposed)	Gravitational collapse from the adjacent mountain front	Landslide deposits (sturzstrom)
S (sandstone)	Channelized, fine to medium-grained, locally coarse-grained to pebbly, normally grained, fining upward sandstone. Sedimentary structures include horizontal, planar and trough cross-bedding and towards the top of the sandstone body ripples and parallal lamination. Beds 0.3-	Sh, St, Sp, Sl, Sr, Gh, Gt, Gp	Channel-fill complex, sandy bedforms and sandy downstream accretionary	Channel fill deposits in a well-confined sand-dominated fluvial channel	Fluvial system (channel complex)

	to 1.5-m-thick with lateral extent of few tens of meters. Erosive concave-up base contacts. Interbedded with facies G3, SM, M, and rarely E		macroforms		
SM (interbedded fine-grained sandstone and mudstone)	Fine-grained sandstone and siltstone with a tabular geometry. Sedimentary structures include parallel lamination symmetrical and asymmetrical ripples locally climbing. Beds 0.1- to 0.5-m-thick, and a lateral extent up to several tens of meters. Basal contacts are flat, non-erosive, and rarely slightly concave up. Proportion between mudstone and sandstone variable. Locally, palaeosol horizons consisting of mottled mudstone and calcite nodules, developed. Interbedded with facies S, G3, M and locally E (in this case they are associated with gypsum-filled fractures)	Sh, Sl, Sr, Smw Fl, Fm, P	Sheet-like and wedging deposits	Sheet-flow deposits in poorly confined to unconfined flow, evolving from upper flow regime to waning flow stage and suspension from standing water, and lacustrine sediments deposited either above the mean fair-weather wave base (sandstone dominating) or above the mean storm wave base (mudstone dominating)	Fluvial (floodplain), playa-lake and lacustrine system (beach to nearshore and offshore environme nt)
M (mudstone)	Massive to laminated grey to light red mudstone. Locally, poorly developed calcrete as well as gypcrete. Beds with a flat non-erosive contact typically 0.02- to 0.5-m-thick and a lateral extent up to several tens of meters. Interbedded with facies S, SM, and locally E (in this case they are associated with gypsum-filled fractures)	Fl, Fm, P	Sheet-like	Suspension deposits in standing water	Fluvial (floodplain), playa-lake and lacustrine system (offshore)
E (evaporite)	Evaporite deposits, 0.05 to 0.3 m thick with a lateral extent of several tens of meters. Generally associated with gypsum-filled fractures. They can form packages of	Ev	Sheet-like	Evaporation deposits from standing water	Play-lake or fluvial (highly

	up to 20 m. Interbedded with facies M, rarely with SM and S				evaporative flood plain) system

4. Results

4.1. Zircon U-Pb geochronology

Five samples were collected for Zircon U-Pb dating in the Eocene volcanics and the Neogene red clastics to constrain the top age of the Karaj Formation and provide independent age constrains on the depositional age of the synorogenic red beds. Results are shown in table 3 and in the Appendix A1.

The contact between the Karaj Formation and the overlying red beds is well exposed along both margins of the basin. Considering that the northern margin has experienced a greater degree of deformation and erosion (compare Figures 3b and 3c with Figures 3e and 3f) we sampled the contact along the southern margin of the basin in two different locations (Figure 1). Sample GH-15-03 represents a > 20-m-thick white tuff that can be followed along strike for about five kilometers. This lithotype is stratigraphically located below a thick package (several tens of meters) of coarse-grained volcaniclastic deposits that are less suitable for zircon U-Pb dating and represent the top of the Karaj Formation in this area (Figure 3b). These units are characterized by a system of open syncline-anticline pairs with a wavelength of several tens of meters (Figures 3b and 6). Our tuff sample (GH-15-03) yielded only few zircon grains with a weighted average age of 36.7 ± 2.6 Ma (Table 3). We collected another sample (GH-15-01) along strike to the SE from a rhyolite exposed on top the Karaj Formation (Figure 3c). In this area the angular unconformity with the overlaying red beds has a low angle (< 10°). This sample yielded a weighted average

age of 38.7 ± 1.4 Ma. This age overlaps with the previous sample (within a two-sigma error) suggesting that the termination of widespread arc volcanism should have occurred sometime between 38 and 36 Ma. This age agrees with those obtained by previous studies (~ 36 Ma, Ballato et al., 2011; ~ 37 Ma, Verdel et al., 2011) in central and northern Iran.

An additional, few cm-thick, ash layer (TM-16-01) was collected within the red beds in proximity of the top of the KA stratigraphic section. This sample is fundamental for pinpointing the magnetostratigraphic correlation (see next sections) and yielded a weighted average age over 13 grains of 10.7 ± 0.4 Ma (Table 3). This value does not include nine grains that clustered around 13-12 Ma. If we include these grains the weighted average age over 22 grains will be 11.3 ± 0.5 Ma (Table 3). Considering that a ~ 10.7-My-old tuff has been dated about 120 km to the NW in three different locations (Ballato et al., 2017), we prefer to consider the 10.7 Ma option as more reliable that the 11.3 Ma. Accordingly, the 13-12-My-old zircon grains should represent crystals that spent 2-3 million of years in the magmatic chamber before the eruption.

Finally, two more samples were collected in the red beds, directly upsection of sample GH-15-03. These two samples are located right above the unconformity (GH-15-02, resampled in a second stage as GH-17-02) and about 400 m (stratigraphically) above it (GH-17-04; Figure 6). The first sample is a weathered, reworked white tuff, while the second one is a light green tuffaceous sandstone with very pristine biotite crystals. These samples gave very similar ages (39.7 ± 1.3 and 38.3 ± 0.9 Ma, respectively; Table 3), which look almost identical to those obtained for the top of the Karaj Formation. Therefore, based on the stratigraphic separation between them we consider these two samples as reworked volcanic material from the eroding Karaj Formation that does not provide indication about the depositional age of the red beds.

Combined, our new zircon ages indicate that arc volcanism in this area must have lasted until 38-36 Ma, while the deposition of the red beds appears to have occurred during the Miocene.

Table 3

Zircon U-Pb Dating Results

Sample code	Age (Ma)	Error 2s (Ma)	N of grains analyzed	N of grains used	MSWD	Rock type	Formation / Unit	Lat (Dec°)	Long (Dec°)	Elevation (m)
GH-15-01	38.7	1.8	11	10	0.4	Rhyolite	Karaj F	36.74525	49.23086	375
GH-15-02/ GH-17-02	39.7	1.3	18	16	1.8	Reworked tuff	Red Beds	36.70804	49.14391	752
GH-15-03	36.7	2.8	6	4	0.8	White tuff	Karaj F	36.70342	49.14172	840
GH-17-04	38.3	0.9	10	10	1.0	Tuffaceous sandstone	Red Beds	36.72139	49.14806	576
TM-16-01	10.7	0.4	24	13	1.3	Ash	Red Beds	36.91298	48.83748	600
TM-16-01 alternative	11.3.	0.5	24	22	3.8					

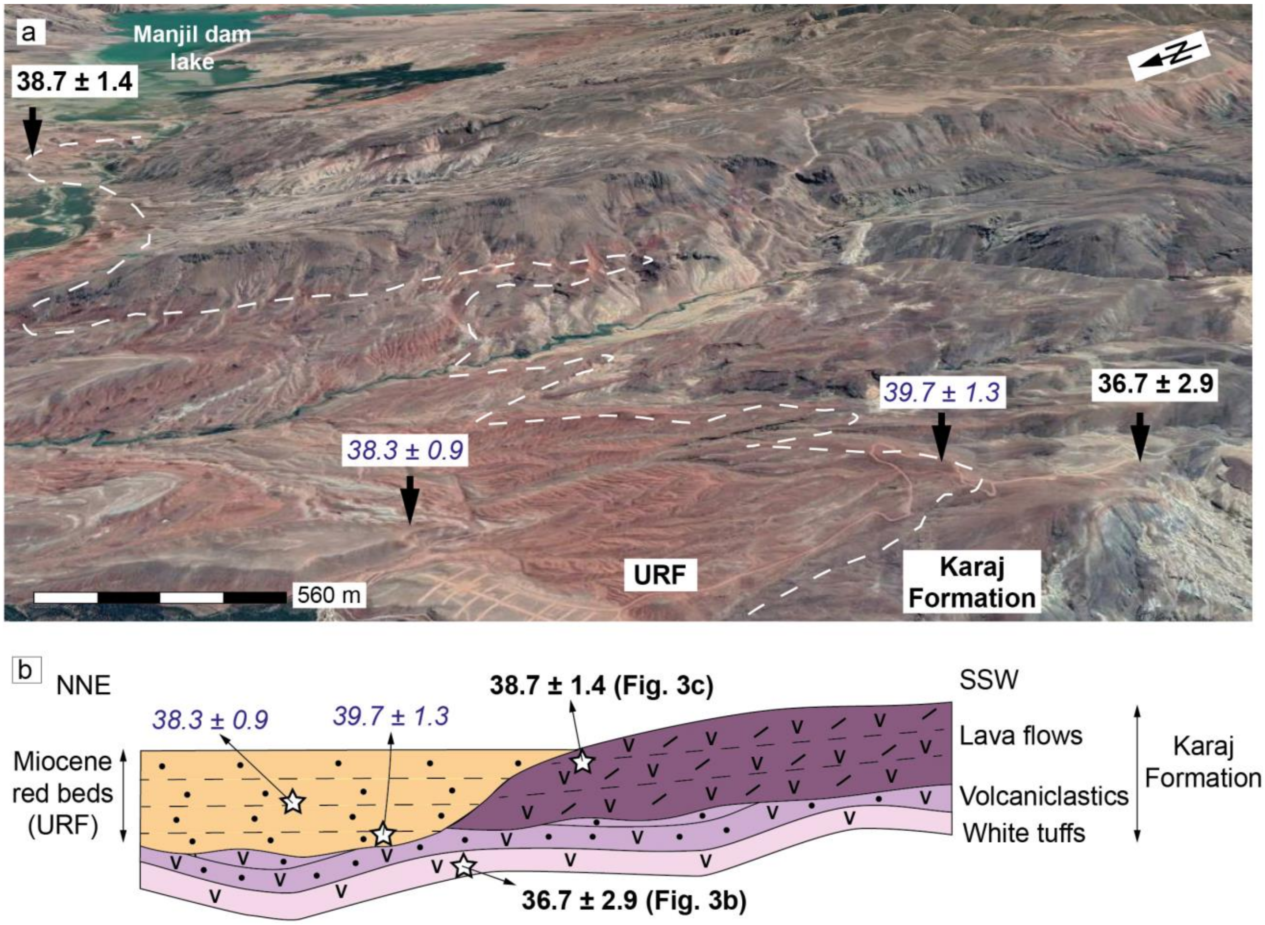


Figure 6. (a) Google Satellite Imagery showing the relationship between the Karaj Formation and the red beds along the southern margin of the basin in proximity of the Manjil dam lake (see the same ages reported figure 1 for location). (b) Schematic cartoon showing the geometrical relationships between the top of the Karaj Formation and the red beds along the southern margin of the Tarom Basin.

4.2. Paleomagnetic results

Seventy-two samples were collected along the 153-m-thick TV stratigraphic section (M1 member), while 143 and 321 samples were collected from the 565-m-thick KA (M2 member) and the 1185-m-thick GH Section (M3 member), respectively. Paleomagnetic sampling was carried out using an ASC 280E petrol-powered transportable drill with a water-cooled diamond bit. Cores were oriented in situ using a magnetic compass. Five hundred thirty-four samples were measured at the Alpine Laboratory of Paleomagnetism (ALP) at Peveragno (Turin) and at the INGV Laboratory of Paleomagnetism (Rome, Italy) shielded room, using a 2G Enterprises DC-SQUID (superconducting quantum interference device) cryogenic magnetometer. Data were analysed using the software Remasoft 3.0 (Chadima & Hrouda, 2006). The NRM of one specimen per core was measured by means of progressive stepwise demagnetization using thermal (384 specimens) or alternating field (AF) (150 specimens) procedures. Thermal demagnetization was carried out using temperature increments (80-100°C up to 430°C and 30-50°C above 430°C) until the NRM decreased below the limit of instrument sensitivity or random changes appeared in the paleomagnetic directions. Stepwise AF demagnetization was carried out using a set of three orthogonal AF coils mounted in-line with the Superconducting Rock Magnetometers (SRM) system, with 5–10 mT increments up to 20 mT, followed by 20 mT steps up to 120 mT.

One hundred sixty-two samples were either too weakly magnetized to allow reliable complete stepwise demagnetization or gave unstable directions during stepwise demagnetization. Such samples were discarded from further analyses. In most of the remaining samples, after the removal of a viscous low temperature/low coercivity normal polarity component at 180°/250° C or 10-30 mT, the NRM vectors aligned along a single linear path toward the origin of the orthogonal diagrams for both normal and reverse polarities (Figure 7a-f). In these samples ChRM directions were calculated by principal component analysis (PCA) (Kirschvink, 1980) of the linear component between 250/320°C and 530/660°C.

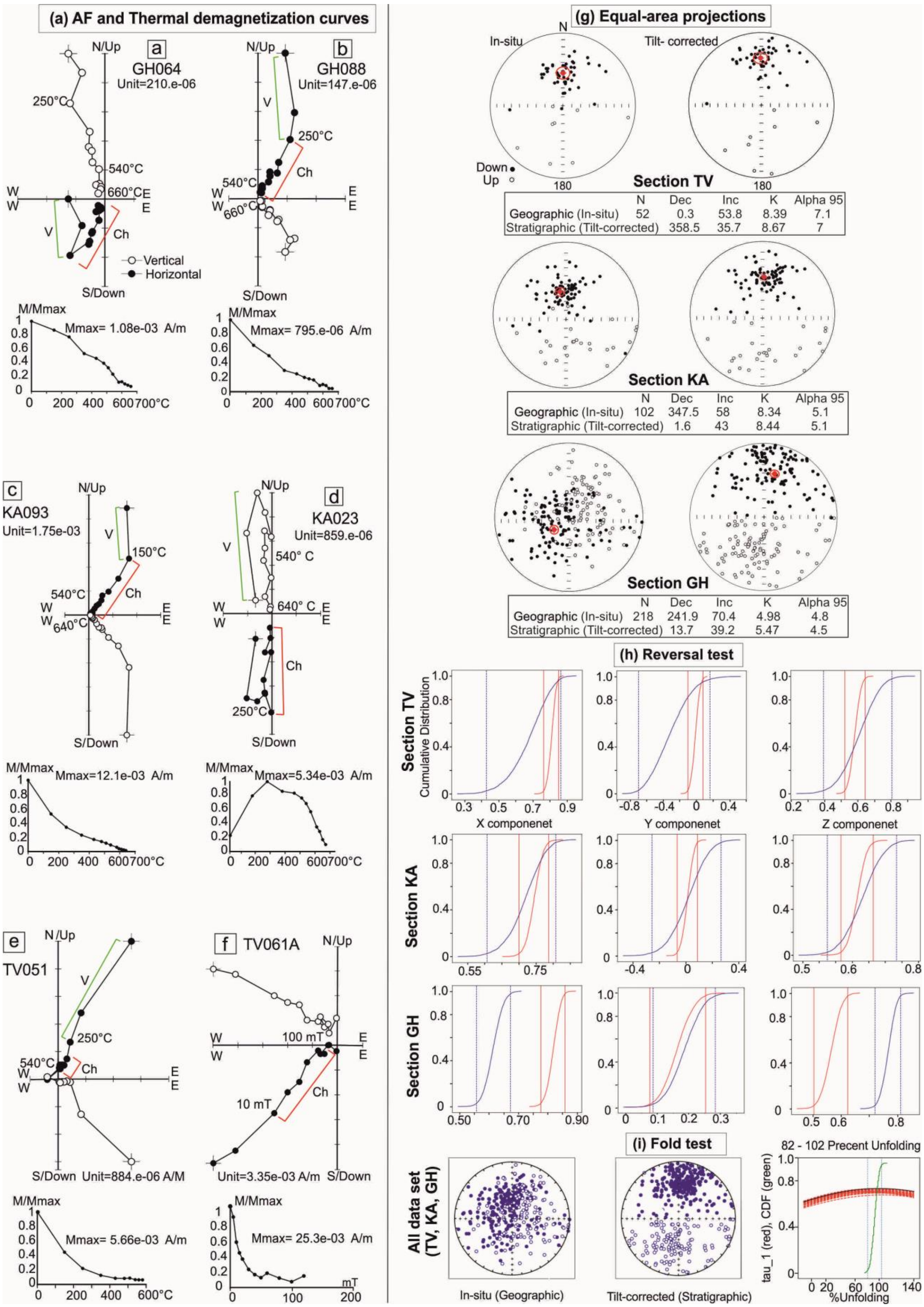

(a) AF and Thermal demagnetization curves
a GH064 Unit=210.e-06
b GH088 Unit=147.e-06
c KA093 Unit=1.75e-03
d KA023 Unit=859.e-06
e TV051 Unit=884.e-06 A/M
f TV061A Unit=3.35e-03 A/m
N/Up
S/Down
W
E
V
Ch
Vertical
Horizontal
M/Mmax
Mmax= 1.08e-03 A/m
Mmax= 795.e-06 A/m
Mmax=12.1e-03 A/m
Mmax=5.34e-03 A/m
Mmax= 5.66e-03 A/m
Mmax= 25.3e-03 A/m
(g) Equal-area projections
In-situ
Tilt- corrected
Down
Up
Section TV
N Dec Inc K Alpha 95
Geographic (In-situ) 52 0.3 53.8 8.39 7.1
Stratigraphic (Tilt-corrected) 358.5 35.7 8.67 7
Section KA
N Dec Inc K Alpha 95
Geographic (In-situ) 102 347.5 58 8.34 5.1
Stratigraphic (Tilt-corrected) 1.6 43 8.44 5.1
Section GH
N Dec Inc K Alpha 95
Geographic (In-situ) 218 241.9 70.4 4.98 4.8
Stratigraphic (Tilt-corrected) 13.7 39.2 5.47 4.5
(h) Reversal test
Section TV
Section KA
Section GH
Cumulative Distribution
X componenet
Y componenet
Z componenet
(i) Fold test
All data set (TV, KA, GH)
In-situ (Geographic)
Tilt-corrected (Stratigraphic)
82 - 102 Precent Unfolding
tau_1 (red), CDF (green)
%Unfolding

Figure 7. (a) Tilt corrected diagrams of Thermal and AF demagnetization analysis of representative samples. Demagnet ization diagrams and intensity decay curves are shown to the left. The black and white circles represent projections onto the horizontal and vertical plane, respectively (Zijderveld, 1967), while numbers at each demagnetization step denote temperatures in °C (150 to 680) and magnetic field values in mT (5 to 120). (b) Mean normal and reverse polarity of ChRM components for the three investigated stratigraphic sections on equal-area stereographic projection in geographic and tilt-corrected coordinates (Dec = declination; Inc = inclination; K = precision parameter, α95 = semi-angle of the cone of 95% confidence). (c) Bootstrap reversal test results for the three stratigraphic sections and (d) fold test results for the entire dataset (Tauxe et al., 1991). The reversal test on TV and KA samples is positive, while GH samples show a negative reversal test. The fold test (all samples from the three studied sections) is positive.

4.2.1. TV stratigraphic section

In the TV Section the initial Natural Remnant Magnetization (NRM) intensities vary between 8.59×10^{-4} and 1.01×10^{-2} A/M (Figure 8). The highest NRM values (average of 4.34×10^{-2} A/M) were obtained in the alluvial fan deposits at the base of the section (first ~15 m; Figure 8). The bulk susceptibility (k) values range from 170 to 10970×10^{-6} SI (Figure 8). High k values are most probably related to the significant contribution of the volcanoclastic Karaj Formation which is particularly rich in magnetite (Ballato et al., 2008). In the TV Section a reliable ChRM has been obtained in 54 samples, 8 with a reverse polarity and 45 with a normal polarity. The maximum angular deviation (MAD) of the recognized magnetic components is lower than 10° (52 samples) except for two samples where it is 11.2 and 14.9°.

4.2.2. KA stratigraphic section

In the KA Section the initial Natural Remnant Magnetization (NRM) intensities vary between 9.91×10^{-4} and 1.01×10^{-2}A/M, whereas the bulk susceptibility (k) values range between 460 and 26570 × 10^{-6} SI (Figure 9). As for the TV Section these high values are probably related to the presence of detrital magnetite from the Karaj Fm. In the KA Section a reliable ChRM has been obtained in 102 samples, 25 with a reverse polarity and 77 with a normal polarity. The maximum angular deviation (MAD) of the recognized magnetic components is lower than 10° in 85 samples and it varies between 10.2 and 14.8° in 18 samples.

4.2.3. GH stratigraphic section

NRM intensities for the GH samples are about one order of magnitude lower than the other two sections, and vary between 9.89 × 10^{-5} and 1.01 × 10^{-3} A/M (Figure 10). Magnetic susceptibility (k) values are also lower than those recorded in the other sections, and range from 70 to 3650 × 10^{-6} SI, possibly reflecting a more composite sediment source area (Figure 10). In the GH Section a reliable ChRM has been obtained in 218 samples, 98 with a reverse polarity and 120 with a normal polarity. The maximum angular deviation (MAD) of the recognized magnetic components is lower than < 10° in 200 samples and is comprised between 10.1 and 14.8° in 18 samples.

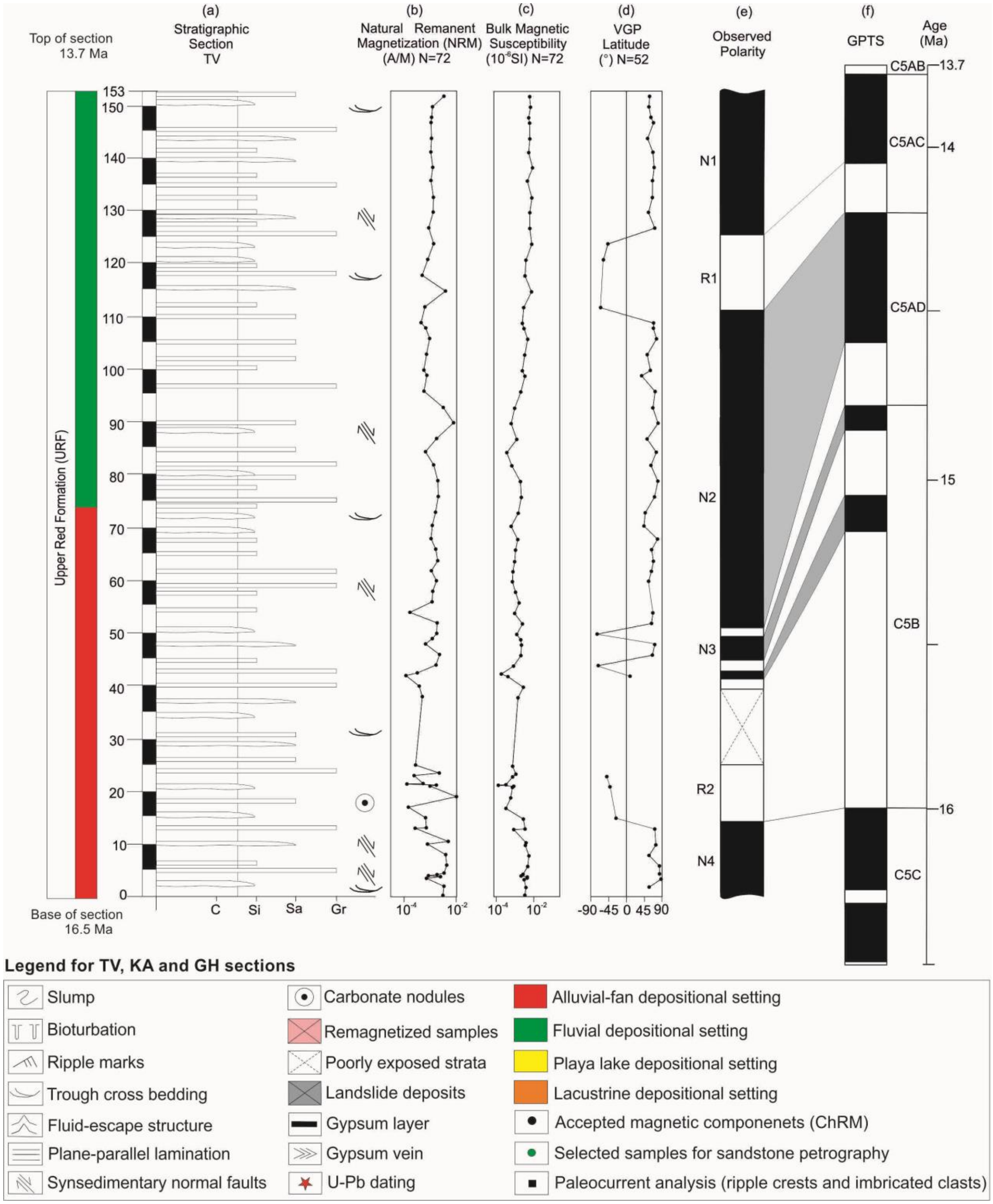


Figure 8. (a) Stratigraphic sections TV including (b) NRM (Natural Remnant Magnetization), (c) Bulk magnetic susceptibility, and (d) VGP latitude (Virtual Geomagnetic Pole). The VGP latitudes were used for constructing (e)

observed polarity scales, which were subsequently correlated each stratigraphic section with (f), the reference GPTS (geomagnetic polarity time scale) of Gradstein et al. (2012). Grey magnetozones of observed polarity scale were detected by means of only one sample

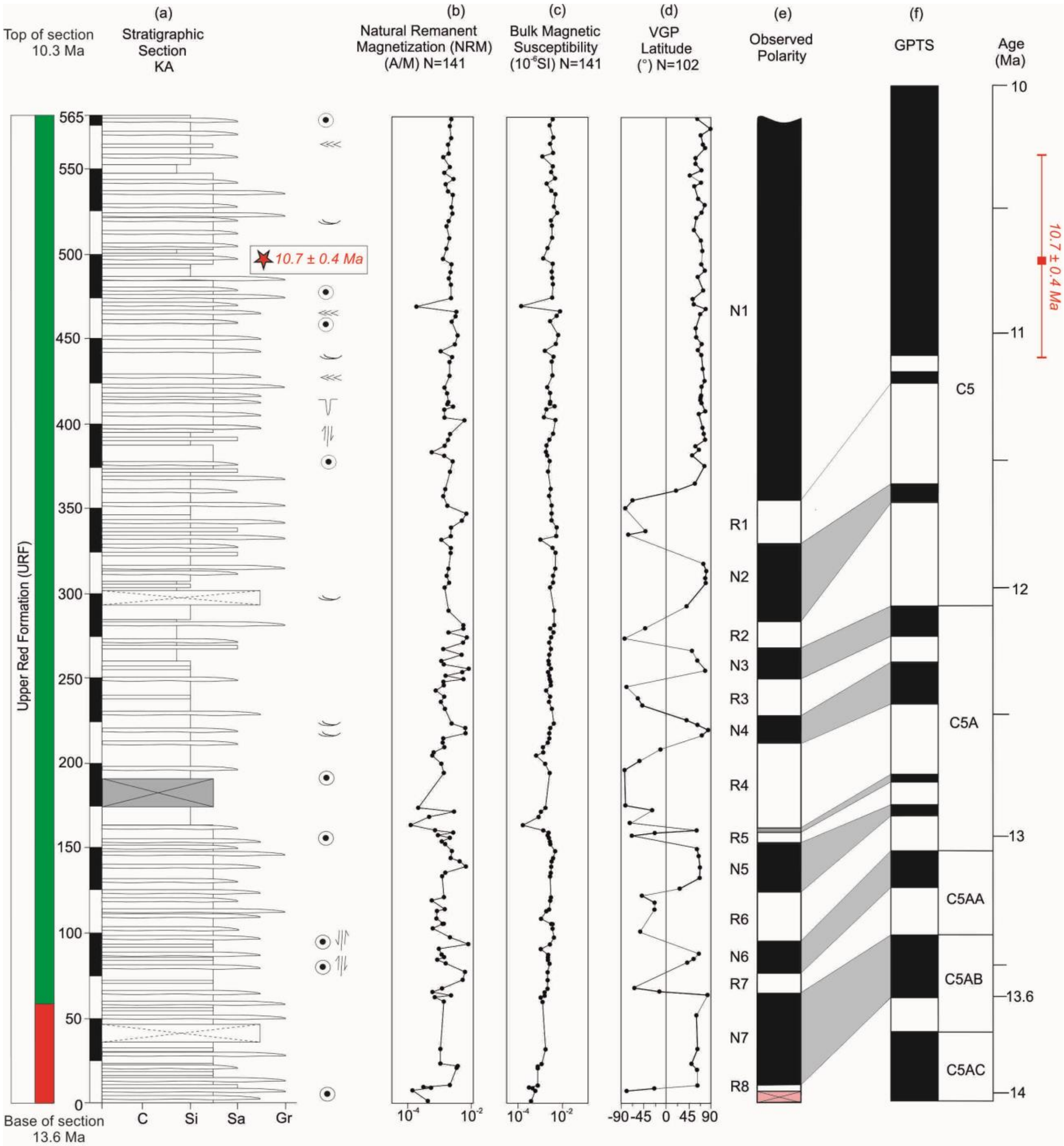


Figure 9. (a) Stratigraphic sections KA including (b) NRM (natural remnant magnetization), (c) Bulk magnetic susceptibility, and (d) VGP latitude (virtual geomagnetic pole). The VGP latitudes were used for constructing (e) observed polarity scales, which were subsequently correlated each stratigraphic section with (f), the reference GPTS

(geomagnetic polarity time scale) of Gradstein et al. (2012). Grey magnetozones of observed polarity scale were detected by means of only one sample.

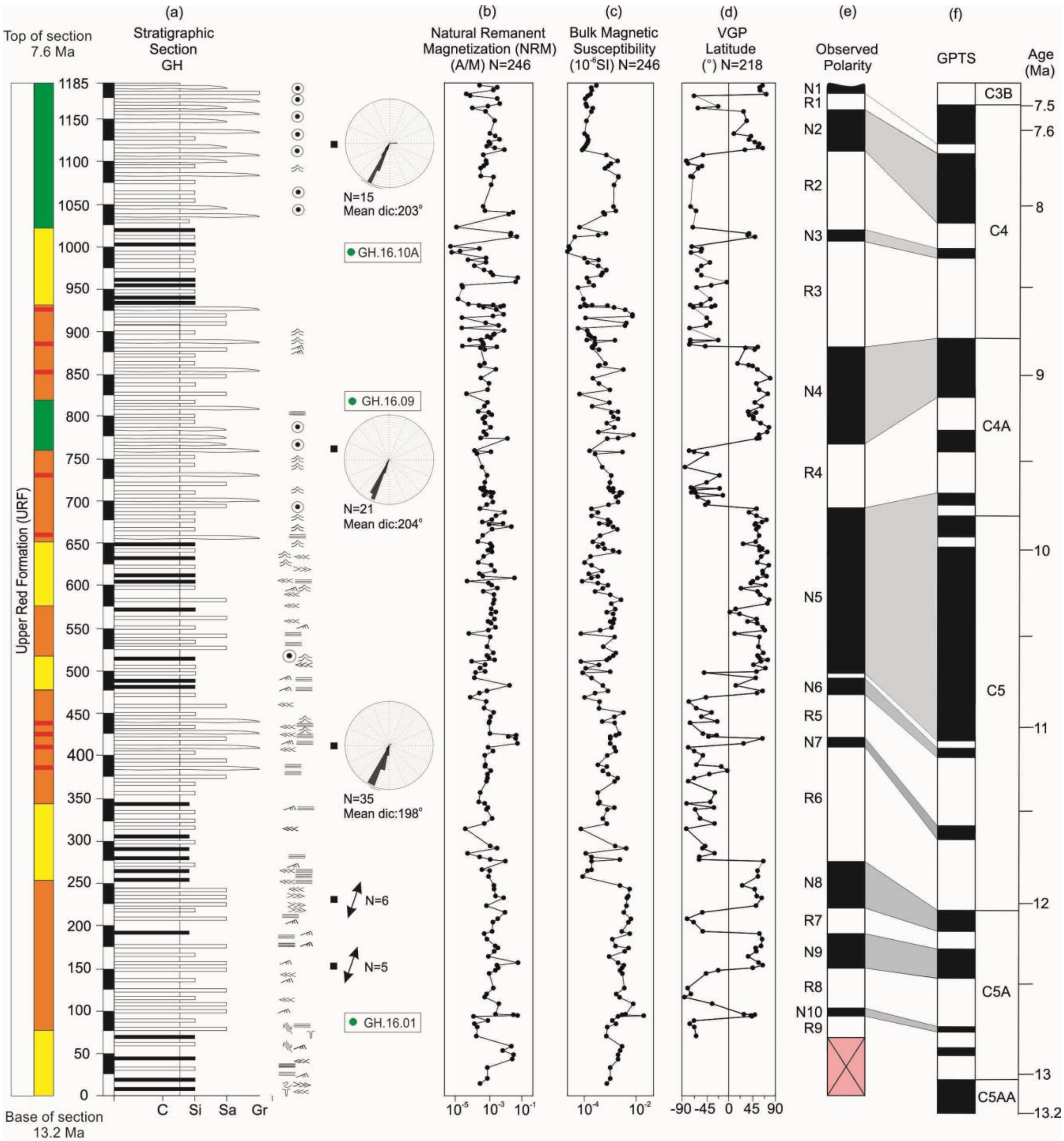


Figure 10. (a) Stratigraphic sections GH including (b) NRM (natural remnant magnetization), (c) Bulk magnetic susceptibility, and (d) VGP latitude (virtual geomagnetic pole). The VGP latitudes were used for constructing (e)

observed polarity scales, which were subsequently correlated each stratigraphic section with (f), the reference GPTS (geomagnetic polarity time scale) of Gradstein et al. (2012). Grey magnetozones of observed polarity scale were detected by means of only one sample

4.2.4. Paleomagnetic tests

To assess the primary nature of the isolated ChRM directions the reversal and fold tests were performed using a Python script, based on the orientation matrix method of Tauxe & Watson (1994). For each magnetostratigraphic section the bootstrap reversal test (Tauxe et al., 1991) has been carried out separately. In the TV and KA sections the normal and reverse polarities directions are antipodal and the reversal test is positive (Figure 7h). On the contrary in the GH section the normal and reverse polarities are not antipodal and the bootstrap reversal test is negative, suggesting that data population could be partially affected by a recent magnetic overprint that was not completely removed during stepwise demagnetization (Figure 7h). The fold test was carried out for all the ChRM directions from the three stratigraphic sections (in total 373 direction) in order to have significant differences in the bedding attitudes. The mean direction of the entire dataset is better grouped after tectonic correction (D = 7.5°; I = 40.0°, k = 6.3, $\alpha_{95\%}$ = 3.2°) rather than before (D = 308.8°, I = 74.2°, K = 4.4, $\alpha_{95\%}$ = 3.9) (Figure 7g). At the same time, the bootstrap fold test (Tauxe et al., 1991) is positive showing that the degree of unfolding to produce the maximum $\tau 1$ is between 86 and 106 % (Figure 7i). These results demonstrate that the ChRM directions from the three stratigraphic sections were most likely acquired before folding. Finally, it is worth to note that the mean ChRM direction obtained from the three stratigraphic sections (D = 7.5°; I = 40.0°) is very similar to the one obtained from 14 sites from the same basin (D = 10.2°; I = 40.6°) with a positive reversal and fold tests (Mattei et al., 2017). These data further support the primary origin of the ChRM in red beds of the URF as also demonstrated by a recent paleomagnetic study in NE Iran (Mattei et al., 2019). On this basis we are confident that our data allow determining correct polarities (latitude of the Virtual Geomagnetic Poles, VGP) and hence to build up a reliable local magnetic polarity stratigraphy.

4.3. Magnetostratigraphy

The VGP latitudes from the new paleomagnetic data set define normal and reverse polarity magnetozones (Figures 8e, 9e and 10e) and hence allow us to construct for each section a magnetic polarity stratigraphy to be correlated with the Geomagnetic Polarity Time Scale (GPTS) (Gradstein et al., 2012). In the following, we first correlate the KA section based on an independent radiometric age, and then we correlate the underlying TV and the overlying GH stratigraphic sections.

4.3.1. KA stratigraphic section

In the KA stratigraphic section 7 normal (N1-N7) and 8 reverse (R1-R8) polarity zones were defined. A Zircon U-Pb age of 10.7 ± 0.4 Ma (Table 3) from an ash layer in the upper part of the section at ~ 500 m suggests that the long-lasting normal polarity zone N1 should be correlated with chron C5n1n. Consequently, the two short reverse polarity zones R1 and R2 and the longer normal polarity zone N2 should belong to the same C5 chron. According to these correlations, the polarity zones N3, N4, N5 as well as the reverse polarity zones R3, R4, R5 and R6 should correspond to chron C5A. In the lower part of the section, the normal and reverse polarity zones N6 and R7 can be correlated with chron C5AA, while the long lasting normal polarity (N7) and the short reverse polarity zone at the base of the section can be correlated to chron C5AB. Based on this correlation the most likely depositional age for the KA stratigraphic section will be between ~ 13.6 to 10.3 Ma (Figure 9).

4.3.2. TV stratigraphic section

Patterns of VGP latitudes in section TV define 4 normal and 2 reverse polarity zones denoted as N1-N4 and R1-R2, respectively. Stratigraphically, the TV section lies underneath the KA stratigraphic section (Figure 2), thus we correlate the uppermost long normal polarity zone N1 and the reverse polarity zone R1 with chron C5AC. Consequently, the long normal polarity zone N2 in the middle part of the section is correlated with chron C5AD and the short normal polarity zone N3 with chron C5B. One reverse polarity zone in chron C5AD, one short normal as well as a reverse polarity zone in the upper part of chron C5B in the GPTS are missing in our records. Besides these three incompatibilities, which represent the time period between ca. 14.6 to 15.1 Ma, we successfully matched up each chron with the GPTS. We note that the missing chrons come from the lower part of the section where the sedimentation rate is lower (~ 0.025 mm/yr) (Figure 11) and the probability to miss a chron greater. The reverse polarity zone R2 in the lowermost part of the section should correspond to chron C5B, while the long normal polarity zone N4 at the base of the section should correlate with chron C5C. Accordingly, a depositional age of ~ 16.5 to 13.7 Ma is proposed for the TV stratigraphic section (Figure 8).

4.3.3. GH stratigraphic section

Patterns of VGP latitudes in section GH define 10 normal and 9 reverse polarity zones, denoted as N1-N10 and R1-R9, respectively. Stratigraphic sections KA and GH overlap, hence, in our tentative correlation we associate the long-lasting, distinctive normal polarity zone N1 of section

KA with the normal zone N5 in the middle part of section GH. The uppermost normal polarity zones N1, N2 and N3 as well as the short reverse polarity zone R1 and long-lasting reverse polarity zones R2 and R3 at the top of the section can be correlated with chron C4. Consequently, the normal and reverse polarity zones N4 and R4 correlate with chron C4A. The long-lasting normal polarity zone N5 in the middle part of the section as well as the two short normal polarity zones N6 and N7 and two long reverse polarity zones R5 and R6 correspond to chron C5. Finally, the normal polarity zones N8, N9 and N10 and the reverse polarity zones R7, R8 and R9 in the lowermost part of the section should correlate with chron C5A. Based on this correlation the depositional age of section GH should range from ~13.2 to 7.6 Ma (Figure 10). Combined our data document a depositional age for the red beds in Tarom Basin from ~ 16.5 to at least 7.6 Ma. Importantly, this implies that these red clastics belong to the Upper Red Formation.

4.4. Sediment accumulations rates

The sediment accumulation rates for each stratigraphic section were calculated based on the magnetostratigraphic correlations and the stratigraphic thickness measured in the field (Figure 11). The oldest record (from ~ 16.5 Ma) is from the TV section where rates are relatively low (0.025 mm/yr) until ~ 14.6 Ma when an increase up to ~ 0.1 mm/yr occurs. From ~ 13.6 Ma the record includes both the GH and KA sections with similar rates of ~ 0.21 mm/yr at least until ~ 12.1 Ma. By ~ 12.1 Ma, sediment accumulation rates for the GH section increase up to ~ 0.29 mm/yr and remain higher than those in the KA section (at least until the top of the KA section at ~ 10.3 Ma). At the top the section, sediment accumulation rates decrease down to 0.15 mm/yr. Overall, the sediment accumulation rates from the intermontane Tarom Basin are slightly lower than those recorded in the Miocene foreland basins of N Iran (0.3 to 2.2 and 0.3 to 0.5 mm/yr for the southern Alborz Mountains and the Great Pari Basin, respectively; Ballato et al., 2008, 2017) but they are still comparable with rates observed in tectonically active regions of the Alpine-Himalayan orogenic belt (e.g., Charreau et al., 2005; Huang et al., 2006; Zhu et al., 2008; Chang et al., 2012).

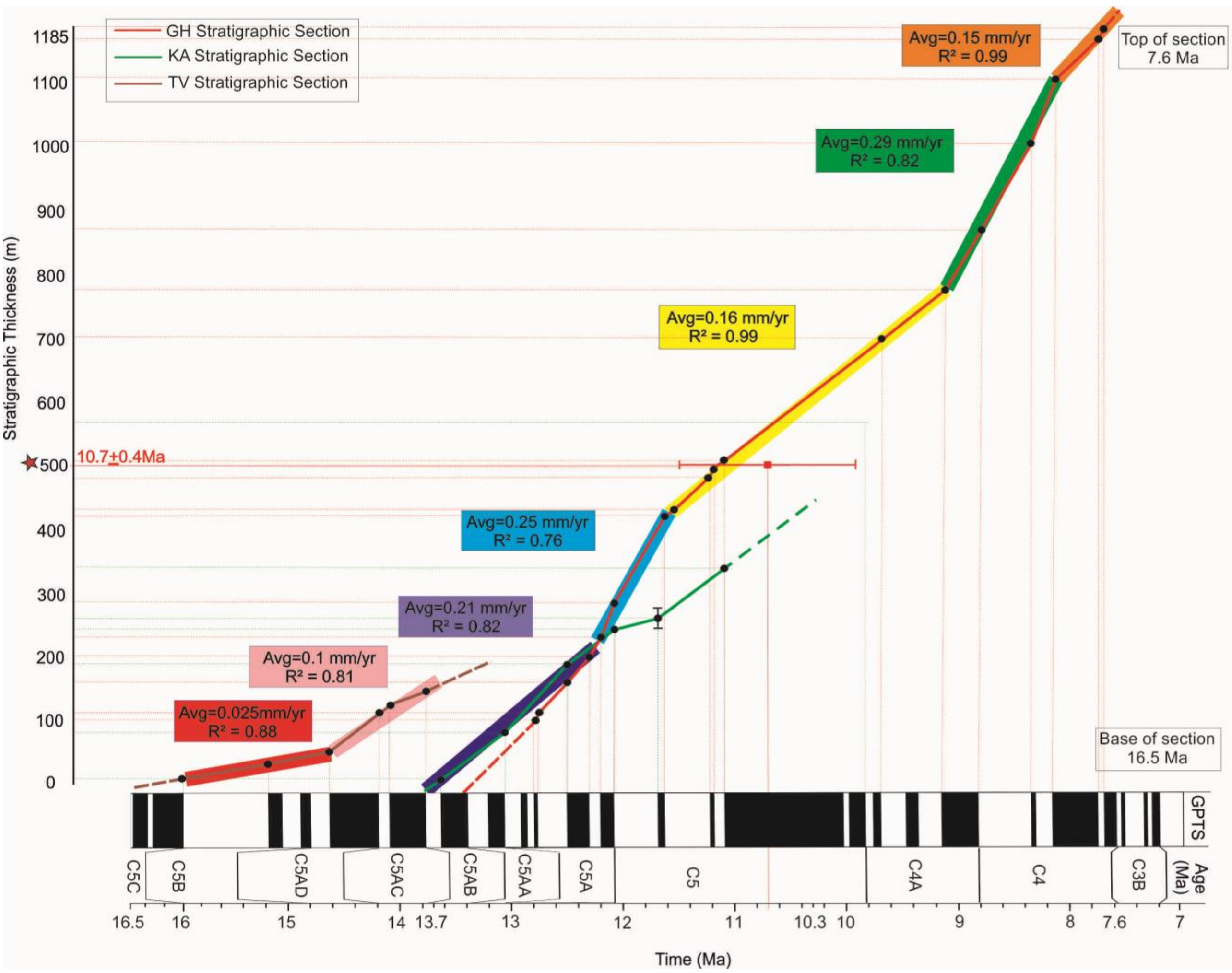


Figure 11. Long-term sediment accumulation rates for the Miocene synorogenic sediments of the three investigated stratigraphic sections. Rates have been obtained by using a linear best fit model (see correlation coefficient R^2) according to the different segments shown with the colourful boxe

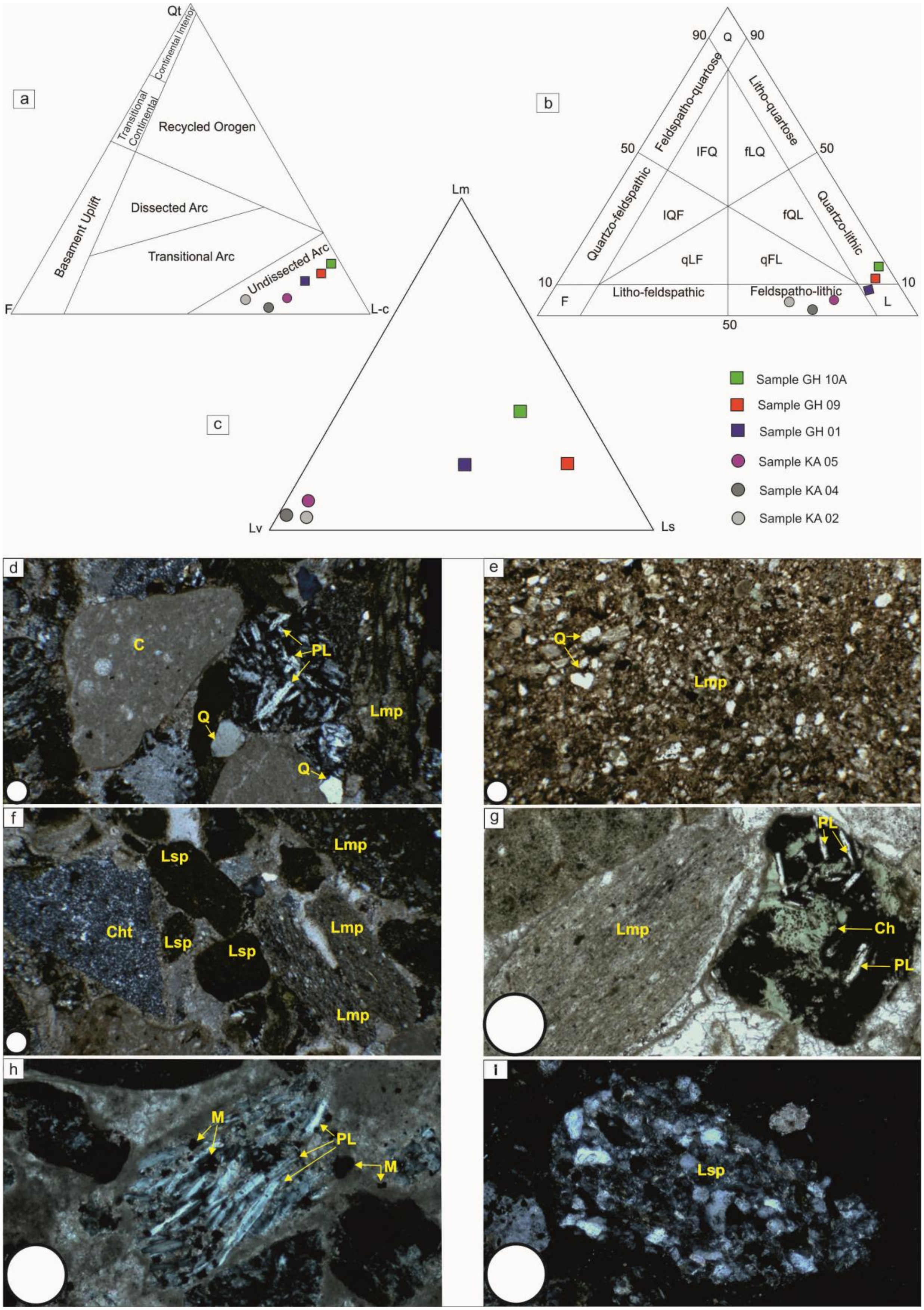
4.5. Sandstone petrography

Petrographic analyses were performed on 6 thin sections collected along the KA and GH stratigraphic sections according to the Gazzi-Dickinson method (Ingersoll, et al., 1984). Results are plotted on QFL-c, QFL and Lm-Lv-Ls ternary diagrams (Figures 12a-c, respectively, Dickinson et al., 1985; Garzanti, 2019). A detailed table can be found in the Appendix (Tables A3.1 and A3.2). The KA sandstones are rather homogenous and mainly composed of volcanic mafic clasts (Lvm, 50 and 58%) and plagioclase (Pl) grains (Figures 12a, 12c, 12d, 12g and 12h). These are more abundant in the lower part of the section (30 vs 19%). A few lithic meta felsic particles (Lmv; 6 to 9%) as well as a small amount (less than 5%) of quartz and heavy minerals (epidote) are the other constituents observed in the KA samples. Finally, a minor amount (≤ 3%) of lithic fragments such as lithic volcanic felsic (Lvf), lithic limestone (Lcc), lithic terrigenous (Lp), lithic metasedimentary (Lms) and metabasalt lithic fragment (Lmb) were also observed. Conversely, the GH sandstone samples contain a lower proportion of volcanic lithics, and a higher proportion of low-grade metamorphic particles (Figures 12b, 12c, 12e, 12f and Table A3.1).

The most abundant constituent of the framework components is represented by lithic metasedimentary (Lms) clasts, which range upsection from 14 to 37% (Table 1). The second most abundant constituents are lithic terrigenous (Lp; 8-25%). Other particles that are much more abundant than in the KA samples are meta felsic (Lmv) and lithic limestone (Lcc) clasts (4 to 17% and 9 to 16%, respectively). Volcanic mafic clasts (Lvm) are less abundant than in the KA samples and show a significant upsection decrease from 21 to 3%. Quartz (Figures 12e and 12i) and feldspar particles were also observed in GH sandstones (Figures 12d and 12h). Feldspar grains are less abundant than in the KA samples, with plagioclase particles ranging from 3 to

10%, while the alkali feldspars display also a very small amount (1%). Instead, Quartz grains are more abundant (9 to 13%). A minor amount (≤ 3%) of other lithic fragments (Lvf, Lch, Lmf) and heavy minerals (such as epidote) were also observed.

Overall, the abundance of volcanic clasts in the KA samples indicates that the main sediment source along the southern margin of the basin must have been from the Eocene volcanics (Karaj Formation) of the Tarom range. It should also be noted that while the thin sections from the KA do not present any clasts of intrusive rocks, the unconformable conglomerates of supposed Pliocene contains abound clasts of granitoides, which are currently exposed along the southern slope of the range (Figure 2). This indicates post 7.6 Ma exposure of the granitoides of the Tarom range. Concerning the central sectors of the basin, the occurrence of metamorphic and sedimentary lithics, as well as the progressive decrease in volcanic grains suggests that the central sectors of the basin (GH samples) where mostly sourced from the northern basin margin (Alborz Mountains). This agrees with paleocurrent directions obtained in different sectors of the GH stratigraphic section (Figure 10).



a
Qt
Continental Interior
Transitional Continental
Recycled Orogen
Dissected Arc
Transitional Arc
Undissected Arc
Basement Uplift
F
L-c
b
Q
90
90
Feldspatho-quartose
Litho-quartose
50
50
IFQ
fLQ
IQF
fQL
Quartzo-feldspathic
Quartzo-lithic
qLF
qFL
10
10
Litho-feldspathic
Feldspatho-lithic
F
L
50
c
Lm
Lv
Ls
Sample GH 10A
Sample GH 09
Sample GH 01
Sample KA 05
Sample KA 04
Sample KA 02
d
C
PL
Q
Q
Lmp
e
Q
Lmp
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Lsp
Lmp
Cht
Lsp
Lsp
Lmp
Lmp
g
PL
Lmp
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Lsp

695

Figure 12. QFL triangular diagrams with tectonic zones defined by (a) Dickinson, (1985) and (b) Garzanti, (2019). Q

represents total quartz grains (Qm = monocrystalline and Qp = polycrystalline), F represents total feldspar grains (P = plagioclase and K-feldspars), L total lithic clasts and L-c: total lithic clasts excluding carbonates. (c) Lm-Lv-Ls ternary plot for the Tarom Basin (Lm = metamorphic; Lv = volcanic; Ls = sedimentary). (D to I) Representative photomicrographs of sandstone samples. (d) Sample GH-16-05 (stratigraphic position of ~ 410 m) showing a large calcareous grain (c), a volcanic mafic grain with plagioclases (PL), a slate fragment with rough cleavage (Lmp) and quartz grains. (e) Sample GH-16-04 (at ~ 370 m) with metamorphic clasts and quartz (Q) grains in a terrigenous-carbonatic matrix. (f) Sample GH-16-05 (at ~ 410 m) with chert (Cht), pelitic lithic (Lsp) and metamorphic fragments (Lmp). (g) Sample GH-16-10B (~ 990) showing a volcanic mafic grain (Lvm) with Pl altered in green Chlorite (Ch), and Lmp. (h) Sample KA-16-05 (~ 450) displaying a volcanic mafic grain with Pl and magnetite (M) crystals. (I) Sample GH-16-01 (~ 75 m) showing a sandy siltstone lithic fragment with detrital micas (Lsp). Note that all photos are under cross polarized light except figure f. Small and large white circles show scales of 4 and 10 microns, respectively.

5. Discussion and Conclusions

Based on our new age determinations and the reconstruction of the depositional systems and sediment dispersal patterns we propose a four-stage evolutionary model for the Tarom Basin for the last ~38-36 Ma (Figure 13a-d) and we discuss the main implications of our findings for the lateral (orogen perpendicular) evolution of the IP, including the mechanisms that led to the growth of its northern margin (Figure 14).

5.1. ~38-36-16.5 Ma: topographic growth of the southern margin, formation of angular unconformities and development of external drainage conditions

The geometrical relationships among the strata of the Karaj Formation exposed along the southern sectors of the Tarom Basin suggest that minor folding must have occurred during the latest stages of Eocene arc volcanism around 38-36 Ma (Figures 3b and 6). This could represent the earliest event of Late Eocene-Early Oligocene collisional deformation recorded across the entire Arabia-Eurasia collision zone from the Zagros to the Caucasus, Talesh, Alborz and Koph Dagh mountains (Vincent et al., 2007; Morley et al., 2009; Ballato et al., 2011, 2015; Mouthereau et al., 2012; Rezaeian et al., 2012; Roberts et al., 2014; Tadayon et al., 2018). Furthermore, our Middle-Late Miocene age of the overlying red beds indicates that the topographic growth of the Tarom range prevented the Late Oligocene-Early Miocene marine transgression that led to the deposition of the shallow-marine sediments of the Qom Formation (Figure 14a; e.g., Reuter et al., 2009). Therefore, between 38-36 Ma and ~ 16.5 Ma (initiation of red beds sedimentation) the Tarom Basin must have experienced external drainage conditions. This implies that the eroded sediments were delivered directly to the Caspian Sea and hence a connection between the Tarom Basin and Caspian Sea must have been established after the end of arc volcanism (Figure 14a). Sometime during this ~ 20-My-long period both basin margins experienced tilting that led to the development of an angular unconformity between the Karaj Formation and the overlying red beds (Figure 3). Prior to that, the Alborz Mountains represented a topographic barrier between central Iran and the Caspian Sea as suggested by the lack of Eocene volcanics along the northern slope of the Alborz (Figure 14a; Guest et al., 2006a).

5.2. ~16.5 to < 7.6 Ma: intermontane basin development and internal drainage conditions

Sedimentation of continental red beds in the Tarom Basin started at ~ 16.5 Ma and lasted at least until 7.6 Ma. This indicates that these sediments are stratigraphically equivalent to the Upper Red Formation (e.g., Ballato et al., 2008, 2017). During that time interval sedimentation occurred in an intermontane basin developed most likely as flexural response to tectonic loading from the adjacent uplifting mountain ranges (Alborz Mountains to the N and Tarom range to the S;

Figures 13b and 14a). Basin development was associated with a sharp increase in sediment accumulation rates (one order of magnitude, from 0.025 to 0.21 mm/yr) along the TV section at ~14.6 Ma (Figure 11). Furthermore, the occurrence of lacustrine and playa lake deposits in the basin depocenter implies the development of internally drained conditions associated with the topographic growth of the Alborz Mountains, which must have disconnected the former drainage system from the Caspian Sea. Such a topographic growth was triggered by widespread regional deformation related to a more advanced stage of the Arabia-Eurasia collision (e.g., Ballato et al., 2011; Mouthereau et al., 2012) in agreement with available low-temperature thermochronology data in NW Iran (Guest et al., 2006b; Rezaeian et al., 2012; Ballato et al., 2013, 2015; Madanipour et al., 2013, 2017). This is further corroborated by the presence of growth strata along the north margin of the basin indicating syndepositional contractional deformation (Figures 2, 3e and 13b-d).

Our sediment provenance data provide additional information on to the evolution of the sediment source area. The southern side of the basin received sediments from the growing Tarom range. There, exhumation has been limited to less than 3-4 km as documented by available 41-32-My-old apatite fission track ages that may still record magmatic cooling (Rezaeian et al., 2012). This is also shown by the sandstone petrography data from the KA section, that have a rather constant composition dominated by volcanic lithics and feldspars (feldspatho-lithic arenite; QFL plot; Figure 12b), as expected for undissected arc regions (QtFL-c ternary diagram; Figure 12a). Instead, the central part of the basin received a greater amount of sediments from the Alborz Mountains as documented by the higher proportion of metamorphic lithics and quartz grains (quartzo-lithic arenite; Figure 12b). Although these sample plot also in the undissected arc (Figure 12a), the upsection increase in metamorphic grains and the relative decrease in volcanic lithics suggests erosional unroofing with the progressive exposure of the metamorphic basement. This agrees with a fully reset Miocene apatite fission track age (Rezaeian et al., 2012) indicating that exhumation along the Alborz Mountains was greater than in the Tarom range.

5.3. <7.6 Ma to Pliocene? drainage reintegration, basin uplift, deformation and erosion

Sometime after ~7.6 Ma, the Tarom Basin was reintegrated into an external drainage system and a new fluvial connection with the Caspian Sea developed. One possible cause could be fluvial headward erosion triggered by the km-scale, base level drop of the Caspian Sea between ~ 5.5 and 3 Ma (Forte & Cowgill, 2013;). Alternatively, basin capture may have occurred through overspill from the Tarom Basin into the Caspian Sea. In any case, after 4 Ma, the Tarom Basin must have been integrated into the drainage system of the Qezwl-Owzan as documented by overflow processes from the adjacent and more elevated Mianeh Basin of the Iranian Plateau that led to the development of ~1-km-deep Amardos gorge (Figure 1; Heidarzadeh et al., 2017). The establishment of an external drainage system appears to coincide with intrabasinal deformation, basin uplift and erosion, as recorded by several post 7.6 Ma anticline-syncline pairs, in the central sectors of the basin (Figures 2 and 3h). This is well visible in the central sectors of the study area (GH section) where the occurrence of subvertical to overturned red beds suggests the development of a north verging anticline most likely associated with a detachment horizon within gypsum layers at the base of the red beds.

5.4. Pliocene? to Present: alternating episodes of basin aggradation, incisions and excavation

Following intrabasinal deformation, the Tarom Basin experienced at least one major episode of (supposed) Pliocene conglomerate deposition (Stocklin, 1969; Figure 3a) as well as three main phases of basin aggradation and incision, as documented by distinct levels of Quaternary terrace conglomerates (Figures 2, 3a and 13d). These unconformable deposits suggest the occurrence of alternating phases of limited (or absent) and efficient fluvial connectivity with the Caspian Sea. A similar configuration has been described in the intermontane basins of arid to semiarid climatic regions like those forming the Eastern Cordillera and the broken foreland of NW Argentina. There, the landscape response to Quaternary climate changes is thought to be the main driver of short-term cycles (10^5 years) of basin filling and excavation, while tectonics plays a major role in controlling the long-term filling history (10^6 years; Strecker et al., 2009; Streit et al., 2015; Schildgen et al., 2016; Tofelde et al., 2017; Ballato et al., 2019; Pingel et al., 2019). Here, the lack of chronological constraints does not allow unravelling the role of different forcing mechanisms. In any case, it should be noted that, the supposed Pliocene conglomerates are slightly folded into a broad syncline suggesting a possible interplay between intrabasinal deformation and sedimentary loading/unloading cycles, which can hinder/promote intrabasinal deformation (Ballato et al., 2019). For example, these conglomerates are in unconformity onto folded Miocene red beds, therefore, their deformation must have occurred after their deposition either during or after their removal through fluvial erosion (i.e., during sedimentary unloading).

Finally, it should be noted that a similar long-term, tectono-stratigraphic history has been proposed for the intermontane Taleghan-Alamut basin of the central-western Alborz Mountains (Guest et al., 2007). There, the deposition of Middle-Late Miocene red beds was followed by Late Miocene-Pliocene intrabasinal deformation, Pliocene aggradation with conglomerate deposition and Quaternary fluvial incision. This common evolution suggests that the orogen may have responded along strike in a similar way to (either tectonic or climatic) forcing mechanisms (Ballato et al., 2015).

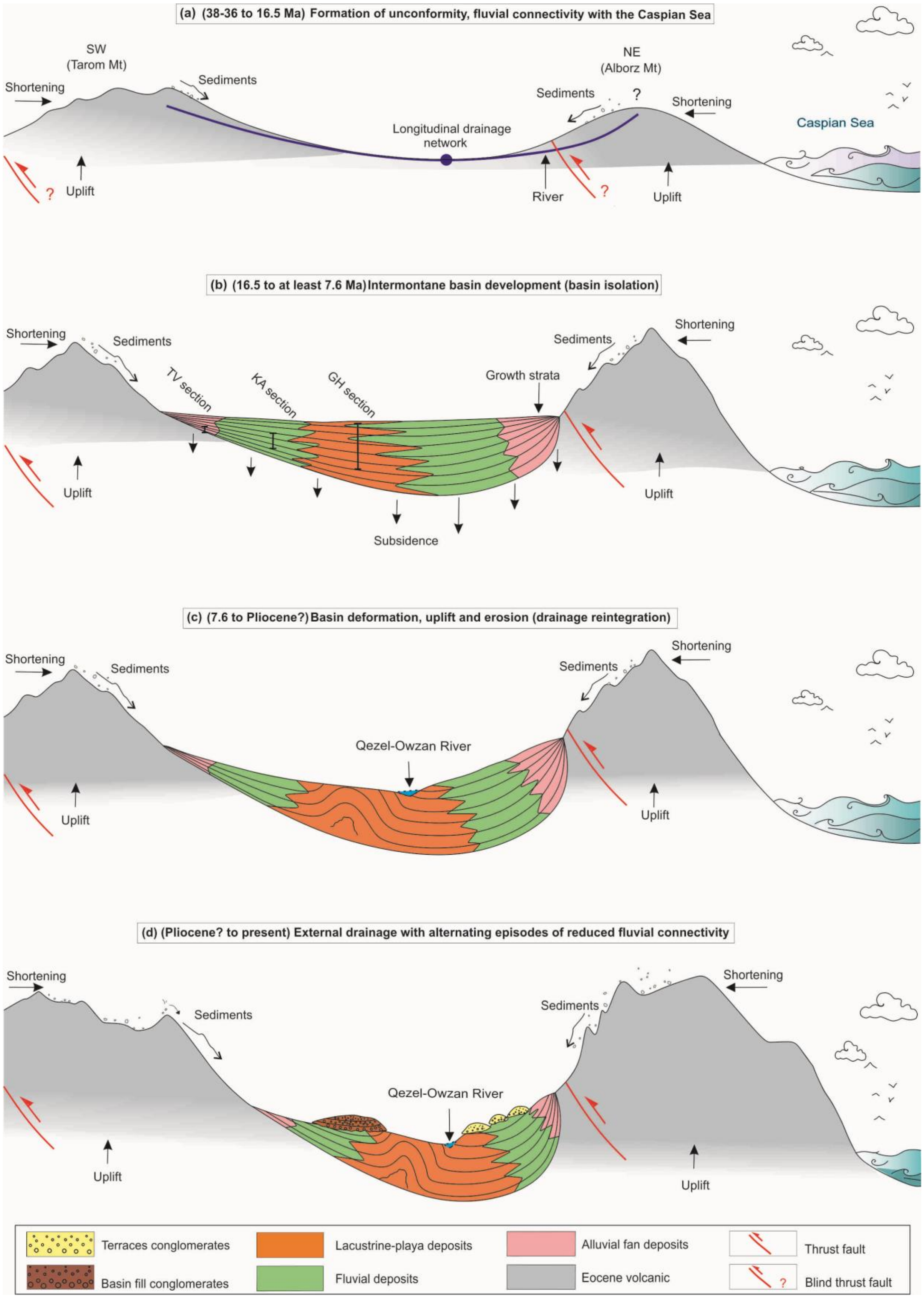
(a) (38-36 to 16.5 Ma) Formation of unconformity, fluvial connectivity with the Caspian Sea
SW
(Tarom Mt)
NE
(Alborz Mt)
Shortening
Sediments
Sediments
?
Shortening
Caspian Sea
Longitudinal drainage
network
?
Uplift
River
?
Uplift
(b) (16.5 to at least 7.6 Ma) Intermontane basin development (basin isolation)
Shortening
Sediments
Sediments
Shortening
TV section
KA section
GH section
Growth strata
Uplift
Uplift
Subsidence
(c) (7.6 to Pliocene?) Basin deformation, uplift and erosion (drainage reintegration)
Shortening
Sediments
Sediments
Shortening
Qezel-Owzan River
Uplift
Uplift
(d) (Pliocene? to present) External drainage with alternating episodes of reduced fluvial connectivity
Shortening
Sediments
Sediments
Shortening
Qezel-Owzan River
Uplift
Uplift
Terraces conglomerates
Basin fill conglomerates
Lacustrine-playa deposits
Fluvial deposits
Alluvial fan deposits
Eocene volcanic
Thrust fault
Blind thrust fault

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Figure 13. Schematic diagram showing the Late Cenozoic evolution of the Tarom Basin (a) ~38-36-16.5 Ma, uplift and tilting, formation of angular unconformities, and development of an external drainage system flowing into the Caspian Sea. (b) ~ 16.5-7.6 Ma, basin isolation and internal drainage conditions, development of an intermountain basin, uplift of the basin-bounding mountain ranges (Tarom and Alborz ranges). The red bars show the location of three measured stratigraphic sections (c) ~7.6 Ma-Pliocene? drainage reintegration with renewed fluvial connectivity with the Caspian Sea, intrabasinal deformation, basin uplift and erosion. (d) Pliocene? to present, cycles of incision and aggradation, folding of basin fill conglomerates.

5.5. Implications on plateau building processes

Our multidisciplinary dataset provides new insights into the lateral (orogen perpendicular) development of the Iranian Plateau and the vertical growth of its northern margin (Tarom range). The hinterland of IP recorded foreland sedimentation starting from ~ 16.5 Ma, shortly after the Late Oligocene-Early Miocene marine transgression that led to the deposition of the Qom Formation (Ballato et al., 2017; Figure 14a). This implies that plateau uplift must be younger than ~ 16.5 Ma. Flexural subsidence was triggered by mountain building processes along the plate suture zone as documented by early Miocene low-temperature thermochronology data from the Sanandaj-Sirjan Zone (Francois et al., 2014; Barber et al., 2018). Foreland basin initiation in the plateau interior coincided with the development of the endorheic Tarom Basin and hence with Middle Miocene topographic growth along the northern sectors of the Arabia-Eurasia collision zone (Ballato et al., 2011, 2013, 2015; Rezaeian et al., 2012). Such a configuration indicates that the retroforeland basin of the Arabia-Eurasia collision zone was partitioned into a broken foreland, like in the North American Cordillera and the South American Andes (e.g., Jordan & Allmendinger 1986; Strecker et al., 2012). The retroforeland was compartmentalized after ~ 11 Ma (Ballato et al., 2017) through the growth of few, orogen parallel, mountain ranges in the plateau interior, which appear to have a regular wavelength of 40-50 km (Figures 14b and

14c). This led to the development of few internally drained intermontane basins and eventually of a typical low-relief plateau morphology (Sobel et al., 2003; Garcia Castellanos et al., 2007), that is still preserved in the sectors of the plateau that are internally drained (Figure 1).

Interestingly, while uplift in the broken foreland of the Andes occurred through the reactivation of steep basement faults (Sierra Pampeanas; e.g., Jordan & Allmendinger 1986) or listric reverse faults (Santa Barbara System; e.g., Kley & Monaldi, 2002) that extend up to at least ~ 25 km of depth (Alvarado et al., 2007; Richardson et al., 2012), the IP presents a more complex pattern of deformation and a shallow seismicity (maximum depth of 20 km, with the majority of the hypocenters around 10 km; Maggi et al., 2002). Although a clear structural model for the IP is currently missing, there are no evidences for a dominant vergence toward the upper plate with a lower crust décollement rooted into the plate boundary as documented in the Altiplano and Puna plateaus (e.g, Horton et al., 2018). A possible reason could be that Iran represents a mobile orogenic belt (e.g., Faccenna et al., 2010) where different microplates were accreted and sutured from the early Triassic (Zanchi et al., 2009; Wilmesen et al., 2009). This has produced some peculiar characteristics such as: 1) the occurrence of orogenic sutures and several crustal scales anisotropies that were repeatedly reactivated under extensional (Late Jurassic and Eocene; e.g., Brunet et al., 2003; Zanchi et al., 2006; Verdel et al., 2011) and compressional (Late Cretaceous to Paleocene and latest Eocene to Oligocene; Guest et al., 2006; Zanchi et al., 2006; Yassaghi & Madanipour, 2008; Rezaeian et al., 2012; Madanipour et al., 2017) regimes before widespread Miocene collisional deformation (e.g., Ballato et al., 2011, 2013; Mouthereau et al., 2012); 2) the presence of a composite stratigraphy (Figure 14b) with few episodes of accelerated subsidence along different depocenters that led to the deposition of several km-thick clastic (the Late Triassic, Shemshak Formation; e.g., Wilmsen et al., 2009; the Miocene, Upper Red Formation, e.g., Ballato et al 2017) and volcaniclastic (the Eocene Karaj Formation; Verdel et al., 2011) sedimentary sequences; 3) the occurrence of a warm lithosphere associated with Eocene magmatism that continued in several sectors of the IP until the present (e.g., Chiu et al., 2013; Rabiee et al., 2020).

During the growth of the IP margin, the Tarom Basin recorded continues syntectonic sedimentation at least until ~ 7.6 Ma with the accumulation of more than ~1.2 km of red clastics. Low-temperature thermochronology data document an acceleration in fault-related exhumation along both margins of the Tarom Basin starting from 12-10 Ma (Rezaeian et al., 2012; Madanipour et al., 2017), in agreement with our sediment accumulation rates. At the same time, our sandstone petrography data suggest that the Alborz Mountains experienced a greater magnitude of exhumation than the Tarom range. This implies that topographic growth in the Tarom range was associated with limited erosional exhumation, as also documented by the occurrence of subdued topography onlapped by basin-fill units in the plateau interior (Heidarzadeh et al., 2017). This suggests that most of the Miocene convergence within the upper plate must have been absorbed via crustal shortening and thickening in the western Alborz Mountains and in the plateau interior rather than along its northern margin. This agrees with a recent seismological study indicating a Moho depth of at least 45 km in the plateau interior that tapers northward to ~ 35 km underneath the northern plateau margin and the Tarom Basin, and increase up to 40-45 km beneath the western Alborz (Figure 14c; Motaghi et al., 2018). Importantly, the occurrence of a ~ 35 km-deep Moho beneath the Tarom range, which is more elevated than the thickened plateau interior, suggests that crustal shortening and thickening cannot be responsible for the topographic growth of the plateau margin. Therefore, surface uplift along the Tarom, must have been triggered by deep-seated, mantle driven processes (e.g., Hatzfeld & Molnar, 2010) rather than crustal/lithospheric shortening and thickening (e.g., Sobel et al., 2003). One possible cause could be the removal of a thickened lithospheric mantle sometimes between 12 and 10 Ma, when deformation processes appears to have accelerated across Northern Iran (Hatzfeld & Molnar, 2010; Francois et al., 2014), and widespread uplift seems to have occurred in the plateau interior (Figure 14b; Ballato et al., 2017). This agrees with the occurrence of a thin lithospheric mantle across most of the upper plate (Rahmani et al., 2019, and references therein), from the suture zone to the Caspian Basin. In any case, although paleoaltimetric data are not yet available and therefore there are not constraints on the vertical growth of the plateau, our reconstruction shows that: 1) the lateral (orogen perpendicular) expansion of the plateau must have occurred over the last 11 Ma, and 2) by 11 Ma the IP must have reached a lateral size similar to present-day one.

Finally, the reconstruction of the basin fills history of the Tarom Basin and our field observations do not indicate the presence of elevations like those attained by the intermontane basins of the plateau interior. This shows that the Tarom Basin was never incorporated into the IP during its phases of internal drainage or limited connectivity with the Caspian Sea. Such a conclusion agrees with a shallow Moho beneath the Tarom Basin (Figure 14c) and corroborates the idea that topographic ponding

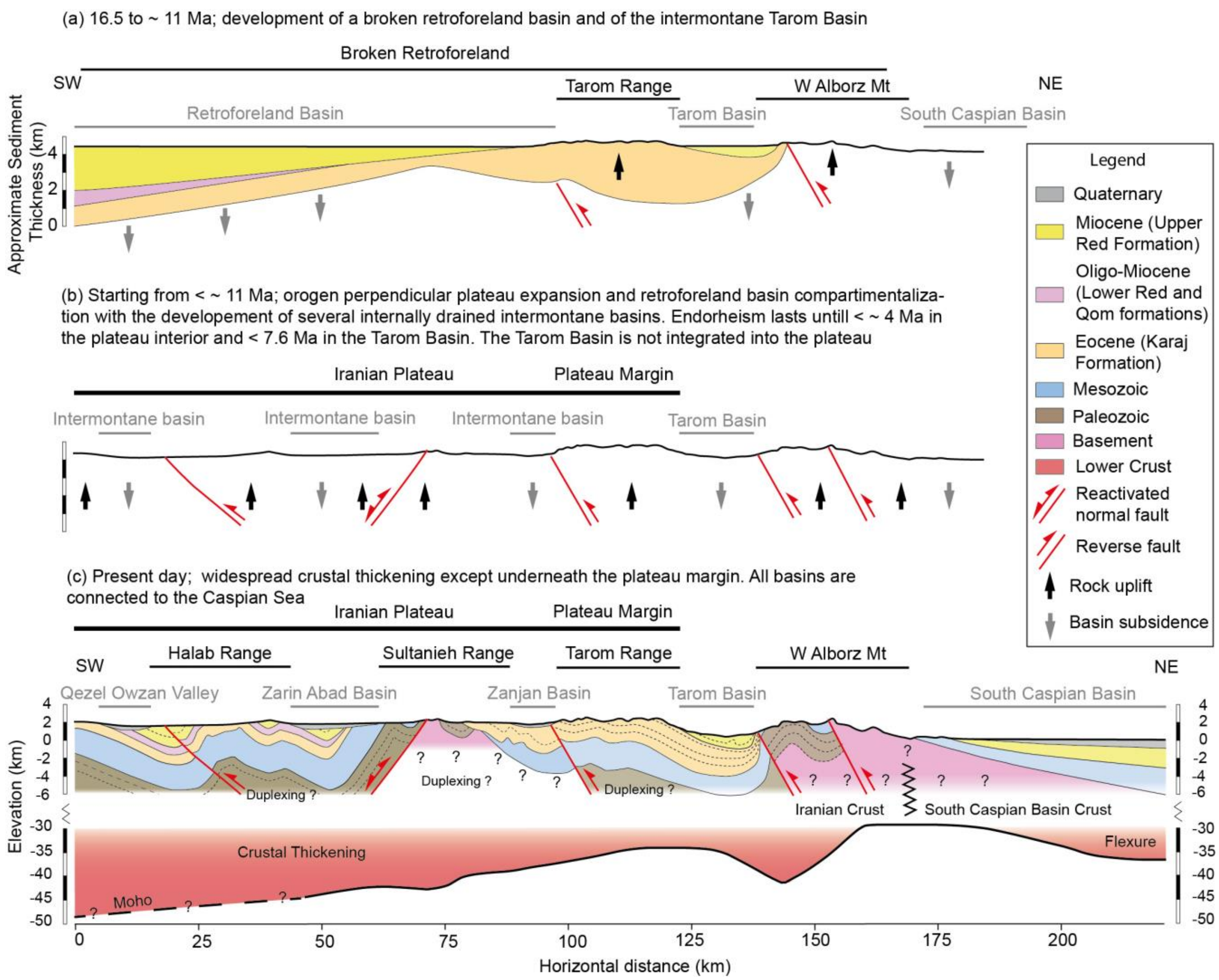

Figure 14. (a and b) Schematic reconstruction of the Late Cenozoic, broken, retroforeland basin of the Arabia Eurasia collision zone during the orogen perpendicular expansion of the Iranian Plateau (see text for details). (c) Geologic cross section (see figure 1 for location) based on Stocklin & Eftekharnezhad, (1969), Davies (1977) and our field observations, and Moho depth (solid line) from Motaghi et al., (2018). The dashed line is extrapolated from the trend in crustal thickness across the IP shown in Rahmani et al., (2019).

5.6. Conclusions

Our work represents the first detailed study in the Tarom Basin, an intermontane basin at the transition between the Iranian Plateau and the Alborz Mountains. Combined, our data show that the regional, Eocene arc volcanism in this area ended at ~ 38-36 Ma in association with the onset of low-magnitude compressional deformation. This was followed by a prolonged phase of erosion with development of angular unconformities. By ~16.5 Ma, the topographic growth on the northern side of the basin (western Alborz Mountains) must have disconnected the Tarom Basin from the Caspian Sea, leading to the formation of an internally drained intermontane basin. Our new ages document that the synorogenic deposits of the Tarom range are stratigraphically equivalent to the Miocene Upper Red Formation. The accommodation space available for sedimentation was most likely controlled by lithospheric flexural in response to tectonic loading of the adjacent mountain ranges. Internal drainage conditions lasted at least until ~7.6 Ma, when basin incision and excavation occurred in association with intrabasinal deformation. Subsequently the occurrence of supposed Pliocene conglomerates and at least three Quaternary terrace conglomerates indicate multiple phases of aggradation and incision. This cyclic behaviour occurred during alternating episodes of reduced and renewed fluvial connectivity with the Caspian Sea. The lack of a detailed chronology, however, does not allow understanding the forcing mechanisms for these cycles. In any case, the elevation of the Tarom Basin during endorheic conditions did not reach those one of the plateau interiors, therefore, the basin was not morphologically integrated into the IP. Furthermore, our reconstruction indicates that the plateau was built on the broken retroforeland of the Arabia-Eurasia collision zone. Specifically, a retroforeland basin developed starting from ~16.5 Ma during tectonic loading and topographic building along the plate suture zone. This coincided with topographic growth along the northern sectors of the collision zone and the development of the intermontane Tarom Basin. Starting from ~11 Ma, intraforeland uplift led to the compartmentalization of the basin with the growth of several mountain ranges over a typical wavelength ~40-50 km and intervening endorheic intermontane basins. During this process the plateau reached a lateral size (orogen perpendicular) like the present one. The northern margin of the IP (Tarom Range) experienced limited erosional exhumation and crustal thickening, suggesting that the vertical growth of the plateau must have been triggered by deep-seated processes (delamination of thickened lithospheric mantle?) rather than crustal shortening and thickening, possibly by 12-10 Ma when upper plate deformation accelerated.

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Appendix

In the following we provide a detailed description of the analytical procedures for each methodology used in this thesis. The raw data can be found in form of tables and figures.

A1. Zircon U-Pb-dating

A2. Zircon U-Pb-dating

A3. Sandstone petrography

A1. Zircon U-Pb-dating

Mineral separation was performed according to standard techniques (crushing, sieving, water table, magnetic separation and heavy liquids as needed) at the Institute of Earth and Environmental Science of the University of Potsdam. Zircons grains where sent to the the Geochronology Laboratory in the Department of Earth and Space Sciences, University of California Los Angeles for the sample preparation and the laboratory measurements. Epoxy grain mounts of hand-selected zircons were gently ground to expose grain interiors and were given final polish with 1 µm diamond. After ultrasonic cleaning, grains were surveyed for internal compositional zonations and/or inclusions via cathode luminescence (CL) imaging. Mounts were then coated with ~100Å of Au. U-Pb ages were determined based on U, Pb, and Th isotopic spot measurements using the UCLA CAMECA ims 1270 ionprobe following the analytical procedure explained in Schmitt et al. (2003). Each analytical run collected data for ten cycles, and age calculations were performed by means of ISOPLOT (Ludwig, 2003). The final ages listed in Table 3 of chapter 2 represent the weighted mean at the 95% confidence level for a given number of aliquots ranging from two to seven (Figure A1.1; Mahon, 1996).

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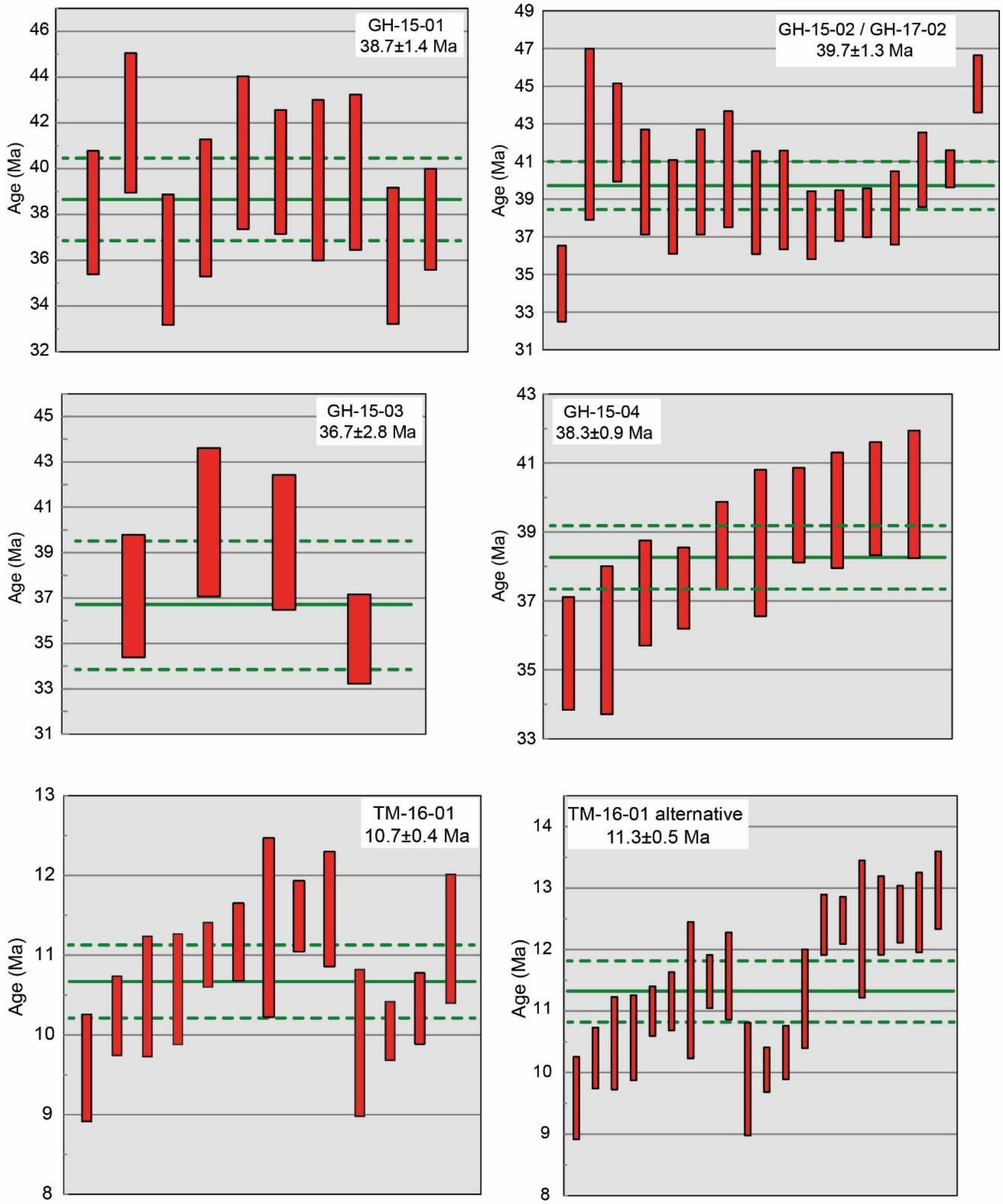

Figure A1: Weighted averages for the analyzed samples shown with a green lines and associated error (in two sigmas) in a dashed green line. The red boxes display the raw data of selected grains (2 sigma error). For sample TM-16-01, two possible solutions are shown (see section 4.1; geochronology for details).

A2. Zircon U-Pb-dating

A total of 536 oriented samples were collected from the three investigated stratigraphic section (TV, GH and KA section) for a combined stratigraphic thickness of 1185 m. The mean sampling interval is typically ~ 3m with at least two cores at each site. In case of poor outcrop conditions or in sectors composed mostly of coarse-grained sediments the sampling intervals was as large as ~ 5-6 m. All the samples were cored with a portable gasoline-powered drill. The orientations of the cores were measured by using a magnetic compass to determine both azimuth of core axis (declination) and dip of the core axis (inclination) and also corrected for ~ 5º E present day declination using magnetic field calculators (www.ngdc.noaa).

Magnetic measurements were then performed using a 2-G Enterprises superconducting rock magnetometer equipped with DC-SQUID coils within a magnetically shielded room at the Alpine Laboratory of Paleomagnetism (ALP) at Peveragno (Turin) and at the INGV Laboratory of Paleomagnetism (Rome, Italy) shielded room in Rome, both in Italy. After measuring the Normal Remanent Magnetization (NRM), samples were subjected to stepwise (up to 15 steps) thermal demagnetization, using heating routine increments (150°C up to a temperature of 480°C and 30–50°C increments above 480°C) until the signal decreased below the instrumental detection limit or random changes of the paleomagnetic directions occurred. A set of sister specimens were chosen for AF demagnetization. Stepwise alternating field (AF) demagnetizations were done using a three-axis demagnetizer with a maximum field of up to 100/120 mT, coupled with a 2G–DCSQUID magnetometer. Data processing was conducted by means of Rema soft program and led to the isolating the stable polarity directions of the characteristic remanent magnetization (ChRM) by using the principal component analysis (Kirschvink, 1980), data statistical analysis by means of Fisher statistics (Fisher, 1953), and finally the calculation of the Virtual Geomagnetic Pole (VGP) from the ChRM vectors.

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A3. Sandstone petrography

Six sandstone samples collected from the KA and GH sections in the Tarom basin were analyzed under a polarized microscope in transmitted light (Table A3.1). In each sample, 400 points were counted by using the Gazzi-Dickinson method (Ingersoll et al. 1984) the results of the modal analysis are plotted in the ternary diagrams of Garzanti (2019) and Dickinson (1985) in order to identify the local tectonic setting and the sediment provenance area (Table A3.1 and Table A3.2).

Table A3.1

Sandstone composition of the KA and GH stratigraphic studied sections in the Tarom Basin

	QFL; Garzanti (2019)			QtFL-c; Dickinson (1985)		
Sample Number	Q	F	L	Qt	F	L-c
KA-16-02	5	31	64	4	32	64
KA-16-04	1	27	72	1	27	72
KA-16-05	5	20	75	5	20	75
GH-16-01	9	10	81	13	12	75
GH-16-09	11	5	83	15	7	78

GH-16-10A	13	4	83	15	4	81

Note. (1) QFL by Garzanti (2019); (Q) Total quartz grains (Qm = monocrystalline + Qp = polycrystalline), (F): Total feldspar grains (P = plagioclase + K-feldspars), (L) Total lithic fragments. (2) QtFL- by Dickinson (1985); (Qt) Total quartzose grains (Qm + Qp), (F) Total feldspar grains (P + K), L-c: Total lithic fragments (excluding carbonates).

Table A3.2

Lm-Lv-Ls ternary plot for the Tarom basin

Sample Number	Lm	Lv	Ls
KA-16-02	5	89	6
KA-16-04	6	93	1
KA-16-05	7	88	5
GH-16-01	20	39	41
GH-16-09	20	13	67
GH-16-10A	36	17	47

Note. (Lm = metamorphic; Lv = volcanic; Ls = sedimentary)

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