2 Geology and Geomorphological Landscapes of Eritrea

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Abstract

The landscape of Eritrea is highly variable and reflects the complex geological history of the area, which is only partially shared with the other regions of the Horn of Africa. The structural geomorphology of Eritrea was investigated through field surveys, literature reviews and a few topographic profiles oriented east–west across the country. The information collected led to the production of a new schematic geological map. The older crustal deformations controlled the orientation of the main fluvial systems, whereas later tectonic events affected their upstream drainage networks. The geological history of Eritrea is very long as it started in the Neoproterozoic, though it was punctuated by a few, more or less long intervals of quiescence. The modern landforms derive from the combined effects of the powerful uplift, to which the whole Horn of Africa was subjected throughout the Cenozoic, and the present arid climate. Fluvial erosion resulted in the accumulation of clastic deposits, a few thousands of meters thick, which gave rise to the coastal belt. In spite of an average denudation rate of 30 mm/ka (a value similar to those inferred for the upper Blue Nile and other areas with a similar structural setting), hard rocks such as the laterites, formed on top of old peneplain surfaces, are still preserved and their present-day elevation along east–west profiles witnesses an impressive upwards dislocation of about 2500 m associated with the impingement of the Afar Plume. In Eritrea, the emplacement of Trap basalts is spatially rather limited, especially if compared with the impressive expansion all across the neighboring Ethiopia. Most of volcanic activity of Eritrea is recent (Quaternary) and associated with the very last phases of the Danakil depression formation. Presently, arid conditions and a volcanic morphology provide the Eritrean Danakil with a unique and fascinating landscape.

Keywords

Crustal deformation • Structural morphology • Peneplain • Uplifting • Horn of Africa

2.1 Introduction

The coastal portion of the Horn of Africa is a wide belt (narrower in Eritrea and wider in Somalia) that includes the eastern side of the vast Ethiopian highlands, forming a horn-shaped peninsula protruding into the Indian Ocean. The many, large-scale geologic events that occurred since the Proterozoic to present have shaped this area into its major landforms. The doming that uplifted the Ethiopian highlands, the rifting and sea floor spreading of the Red Sea and the Gulf of Aden and the formation of the Afar triple junction depression are certainly the main tectonic events that produced a distinctive and variegated physiography on the area, characterized by marked relief contrasts (e.g., in the nearest southern highlands of Eritrea, from the Gulf of Zula to the highest peak of 2250 m asl, the horizontal distance is only 29 km), vast and gently inclined areas, narrow and elongated structural basins and new crust formation with the emplacement of a huge amount of volcanic products. At the same time, the endless action of erosion and deposition processes modified the landscape gifting us with spectacular deep valleys, high escarpments and weathering landforms, basin filling, alluvial fans and deltas, desert and coastal wind-blown dunes, sebkhas and a number of volcanoes that are too recent to be substantially affected by erosion.

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While in Eritrea and Djibouti the major landforms are associated with the uplift of Proterozoic and Paleozoic metamorphic basement, emplacement of Trap basalt series, normal faulting and recent volcanism, Somalia is mainly a tableland, inclined to south-east (going down to the sea level from an altitude of 500–600 m asl) and underlain by sedimentary rocks, resultant from a long-lasting (Mesozoic to Early Cenozoic) transgression and regression cycle. Uplifted terranes and escarpments are present only in the northern part of Somalia and coincide with the border of the Gulf of Aden trough. In Somalia, volcanism is limited to the northern part, whereas the basement crops out only in the area between the lower reaches of the Wabe Shebelle and Juba rivers, called the *Bur* region, and a much smaller area near Hargeisa and west of Bosaso.

Such a complex and intense geodynamic activity was interrupted by phases of tectonic stability that resulted in periods of extensive peneplanations, the most important of which affected the crystalline basement at the end of the Paleozoic and the sedimentary rocks during and after the Lower Cretaceous to Oligocene regression.

The main landscapes and landforms of Eritrea, Djibouti and Somalia are the results of very many complex, long-lasting, overlapping and interacting geologic events and processes. This chapter will focus mainly on Eritrea, and an overview of the major landforms and their relation with its geological structure is given.

2.2 General Physiography: From Eritrea to Somalia

The triangular backbone of the Horn of Africa, formed by the Eritrean–Ethiopian–Somali plateau, is bordered on two sides for over 4000 km by coastal belts of different widths (Fig. 2.1). Those of northern Eritrea and northern Somalia are only a few tens of kilometers wide and are located at the foot of the plateau. Those, equally narrow, of southern Eritrea and Djibouti are characterized by more or less (Djibouti) defined mountain chains and alignments secluded from the plateau structure.

Large coastal plains of central and southern Somalia separate the extreme southern slopes of the Somali plateau from the Indian Ocean. They are some hundred kilometers wide and often covered by bush vegetation. Within the variegated morphology of the Horn of Africa, each of these coastal regions is associated with a characteristic hinterland. In northern Eritrea and northern Somalia, a steep slope links the narrow coastal plain to the plateau with an altitude difference of more than 2000 m. In central Eritrea, the elevation difference of 2500 m between Asmara on the plateau rim and the Massawa coastal plain is gained in about 50 km of horizontal distance The same rim, which is NNW/SSE oriented, i.e., parallel to the adjacent Red Sea trend, attains the highest elevation in Eritrea with the Amba Soira (3018 m asl). From this rim, the plateau, characterized by a rugged morphology, goes down toward the Sudanese lowlands with a low-elevation (500–600 m asl), c. 170 km wide, belt punctuated by scattered isolated hills and groups of small mountains (around 1000 m asl). Near Asmara, the plateau becomes narrow, down to 70 km, bulges up northward in the Nakfa area and shrinks close to the northern boundary with Sudan (Fig. 2.2).

South of the Gulf of Zula, a c. 70 km wide, NW–SE-oriented belt extends for about 300 km parallel to the southernmost Red Sea, from Zula to the Ali Sabieh Mts. (south of Djibouti). From a general flat morphology close to sea level, three mountain regions rise from north to south: the Danakil block in Eritrea with Mt. Aibaba (1350 m asl), the Randa-Obock region and the Ali Sabieh Mts., both in Djibouti. In northern Djibouti, Eritrea, Ethiopia and Djibouti political boundaries meet as in a triple point close to the prominent Mussa Ali volcano (2028 m asl).

In Somalia, a wide triangular region with the corners in Zeila, Cape Guardafui and Garoe/Eil has a northern belt overlooking the Gulf of Aden and extending for 800 km in a ENE direction (Fig. 2.1). The seaward portion is bordered by a narrow coastal plain (the Guban region from Seylac to Berbera), which is connected to an imposing ridge further inland by a steep slope. The Shiick High, the Ahl Madow, Al Mescat and Al Bahari mountains follow each other in a row from the Gulf of Berbera to Cape Guardafui at the Gulf of Aden/Indian Ocean junction. This alignment, constantly above 2000 m in elevation, is the rim of the Somali plateau which gently decreases southward (the elevation of 500 m is reached 500 km from the rim) toward the plains of central Somalia. The plateau is dissected by two transverse ESE trending valleys (Darror and Nogal) with homonymous ephemeral rivers reaching the Indian Ocean.

The rather harsh landscapes of northern Somalia turn into the wide plains of central and southern Somalia where the rivers from the arcuate structure of the Somali plateau in the Harar region flow toward the Indian Ocean over a length of about one thousand km. West of Mogadishu, in the Bur region, spectacular inselbergs breach the monotonous flat morphology of southern Somalia.

2.3 Geological Outline of Eritrea

Though with a variable areal distribution, the Proterozoic basement makes up the largest part of relief of Eritrea and forms an equilateral triangle with a north-pointing vertex across the Sudan border. Basement rocks are the backbone of northern and central Eritrea covering more than the 70% of the region north of the Danakil depression (Fig. 2.3).

Fig. 2.1 Physiographic map of the Horn of Africa

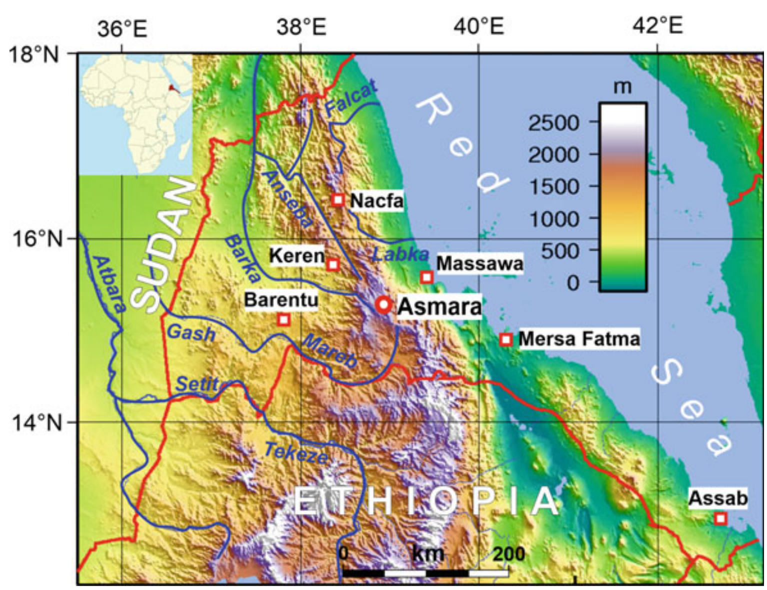

Fig. 2.2 Digital elevation map of Eritrea with the main physiographic elements

Investigations on the basement played an important role in the advancement of knowledge about the Eritrean territory, both for its undoubted geodynamic significance in the framework of the geology of East Africa and for its possible mining value. Starting from the pioneering geological observations by Baldacci (1891) and Bibolini (1920) in northern Eritrea, the petrographic studies on the basement rocks by Manasse (1909), Aloisi (1931) and Andreatta (1941) and the identification of the Barca suture zone by Francaviglia (1938), the attempts to recognize lithological succession and establish the chronology of the deformational and metamorphic events were discussed extensively in the first fundamental work on Eritrea's geology by Dainelli and Marinelli (1912) and in the second volume of the Geology of Eastern Africa by Dainelli (1943). In the 1:2,000,000 geologic map of the Horn attached to the volume, Dainelli (1943) made a first distinction between areas with different metamorphic grades. Subsequently, Mohr (1962), in Fig. 2.3 of his Geology of Ethiopia, proposed a schematic anticipatory map which takes into account the distinctions of metamorphic degree and the main foliations of the basement throughout East Africa and with particular detail in Eritrea. Later, authors of both papers (Gherardi 1951a, b) and regional maps (Kazmin 1973; Merla et al. 1979, as well as modern authors), maintained this distinction (Ghebreab et al. 2005; discussion in Andersson et al. 2006) (see later).

After 30 years of warfare, research on the basement geology resumed intensively in the nineties through the activity of the University of Asmara and the Eritrean Ministry of Mines with the collaboration of foreign universities, particularly from Sweden and the UK. Numerous publications span two decades of joint international studies, and a short list includes, among others, Berhe (1990), Drury and Berhe (1993), Drury et al. (1994), De Souza Filho (1995), Woldehaimanot (1995), Tadesse (1996), Ghebreab (1996), Talbot and Ghebreab (1997), Teklay (1997), De Souza Filho and Drury (1998), Drury and De Souza Filho (1998), Ghebreab (1998, 1999), Ghebreab and Talbot (2000), Beyth (2001), Drury et al. (2001), Ghebreab et al. (2002, 2005, 2009), Solomon and Ghebreab (2006) and Andersson et al. (2006). These studies confirmed that in the Eastern African Proterozoic continental framework, the Eritrean basement rocks were part of the southern Arabian-Nubian Shield, an accretion orogen (East African Orogen, Stern 1994) elongated from Sinai to Kenya (Fritz et al. 2013).

2.3.1 The Neoproterozoic Basement

Within the Neoproterozoic supercontinent Rodinia, Eritrea was in the central part of the Arabian-Nubian Shield (ANS). About one billion years ago, Rodinia began to rift apart, the Mozambique ocean started opening, and new margins (East and West Gondwana) accreted through ocean spreading, crustal growth, island arc- and plume-related magmatic activities (Teklay 2006). Various phases of convergence, which eventually gave rise to the East African Orogen (Pan-African Orogeny Kröner and Stern 2004; Johnson and Woldehaimanot 2003), followed from 850 to 550 Ma ago affecting the newly formed island arc, oceanic plateaus and ridges, old and new granitoid intrusions and volcano-sedimentary successions.

Although these lithostructural elements were intensively deformed and metamorphosed in various degrees, it has been possible to distinguish single tectonostratigraphic rock assemblages (terranes) with characteristic lithology, metamorphic grade, bounding tectonic lineaments and geodynamic histories. The terranes cover the whole northern and half of central Eritrea and are almost absent in the eastern/southern part of the country. They align NNE–SSW with a predominant west steeply dipping tectonic fabric and are separated by c. north–south trending transpressional suture belts. The width of the terranes is up to 100 km, and their length is up to 500 km. Some terranes and sutures continue along strike in north Ethiopia, Sudan and Saudi Arabia.

From the Sudan border to the east, the more prominent terranes are Barka, Hagar, Adhoba Abi, Nakfa, Arag and Ghedem (Fig. 2.4) (Drury and Berhe 1993; Drury and De Souza Filho 1998; Ghebreab et al. 2009). In particular, the Barka terrane is an intensely folded assemblage of amphibolite to granulite facies, with mafic gneisses, quartzites, marbles and a major olistostrome (De Souza Filho and Drury 1998). Initially, due to its location close to the stable Africa Sudanese crust, the Barka terrane was considered as a possible relic of an old continental crust, but more recent research proved that it was, like all the other northern Eritrean terranes, a product of the Pan-African event.

To the east of the Barka terrane and beyond the Barka river is the Hagar terrane dominantly composed by calc-alkaline metavolcanic and metavolcanoclastic rocks in greenschist facies, discontinuous marbles, cherts, oceanic basalts and pelagic sediments, basic–ultra-basic rocks and a major olistostrome. All these features point at accretionary wedge and fore-arc settings (Drury et al. 1994).

Further to the east, the Adobha Abi terrane exposes in a highly dismembered shear zone of calc-alkaline greenschist facies rocks and dispersed basic and ultrabasic ophiolites with oceanic island arc character (Woldehaimanot 2000).

Next is the Nakfa terrane which underlies most of the northern highlands of Eritrea with its c. fifty thousand square kilometers. It consists of greenschist facies calc-alkaline volcanic and volcanoclastic rocks mainly prevailing on tholeiitic volcanites, phyllites, slates, sandstones and conglomerates. It is interpreted as a relict island arc assemblage.

Peculiar in the Nakfa terrane is the great amount of syn- and post-tectonic granite and granitoid intrusions, particularly in its western half, commonly occurring as inselbergs

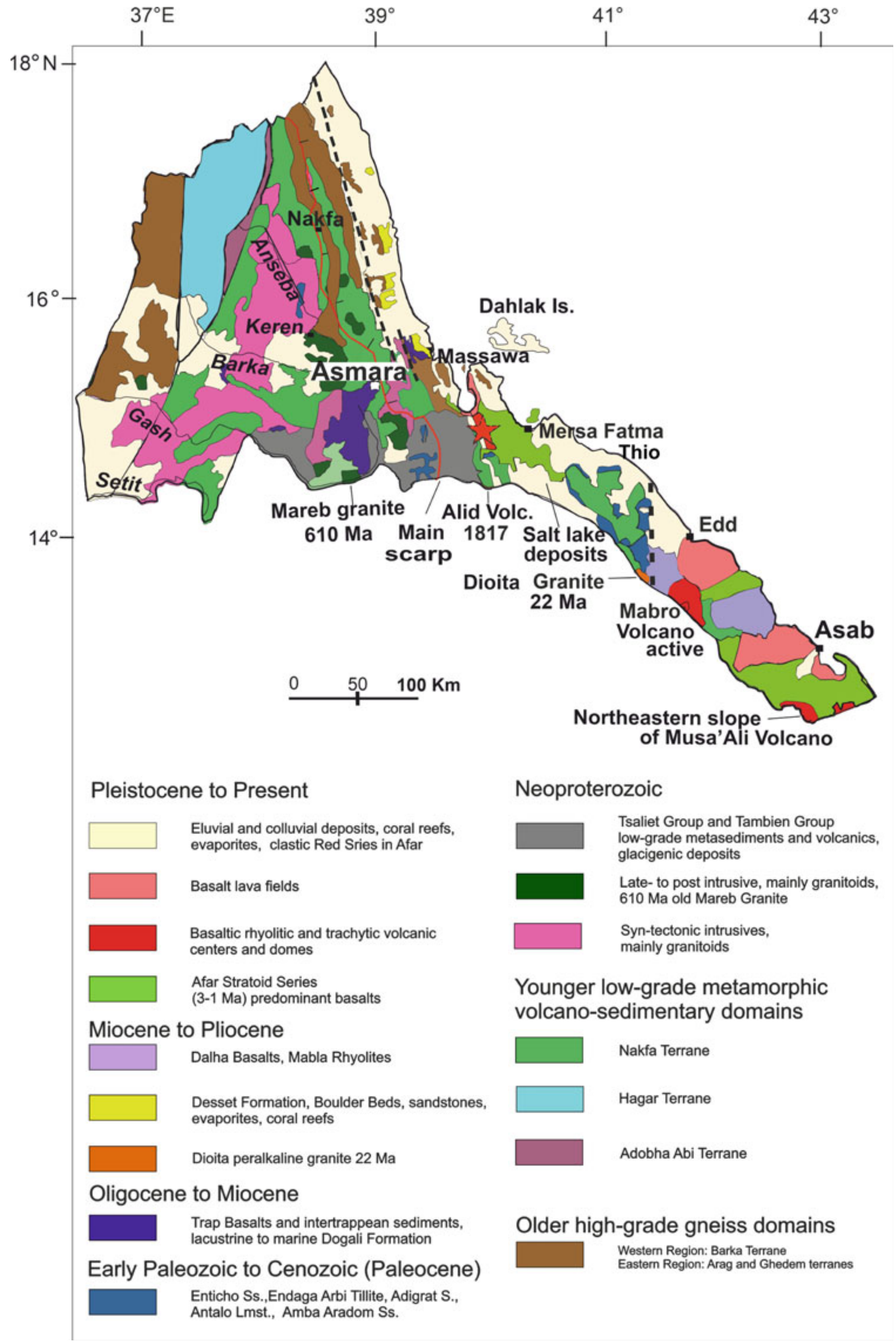

Fig. 2.3 Schematic geological map of Eritrea. The red solid line is the margin of the main escarpment; the black solid lines are the main faults, and the dash-dotted lines are the major lowland Tertiary faults. Red star: Buia *Homo erectus/ergaster* site and Alid volcano location. Sources of compilation: Brinkmann and Kürsten (1970), Kazmin (1973), CNR-CNRS (1971, 1975), Merla et al. (1979), Drury and Berhe (1993), Drury et al. (1994), Teklay (1997), Drury and De Souza Filho (1998), Sagri et al. (1998), Ghebreab and Talbot (2000), Beyth et al. (2003), Ghebreab et al. (2005, 2009), Kumpulainen et al. (2006, 2007), Avigad et al. (2007), Eritrea Department of Mines (2009), Bussert (2010, 2014)

(Fig. 2.5). Moreover, a N–S trending, tens of kilometers wide belt of "early" granitoids, has been reported by Teklay (1997, map in Drury and De Souza Filho 1998) from Nakfa to Barentu. They were connected with the first stages of the accretion processes dated at 811–846 Ma. Additionally, huge masses of "late granites" (650–630 Ma old—Kröner et al. 1991; Teklay 1997) mark the eastern part of the Nakfa terrane. One of these post-tectonic granites is Mt. Bizen (2435 m asl), famous for its precipitous slopes and the monastery on its top. It is located near Nefasit, on the Asmara-Massawa road (Fig. 2.6).

The area between Nakfa and Keren offers panoramic views of another remarkable landscape in the Rora range, where the weathered and dissected rocks of the basement

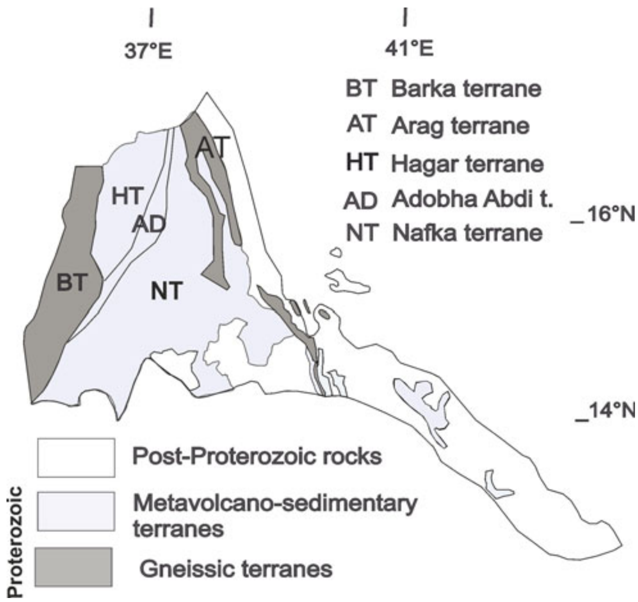

Fig. 2.4 Schematic map of major Neoproterozoic terranes of Eritrea (modified after Ghebreab et al. 2005; Avigad et al. 2007)

Fig. 2.5 Syn-tectonic granite near Keren. Notice the rounded morphology of the peak due to exfoliation weathering processes

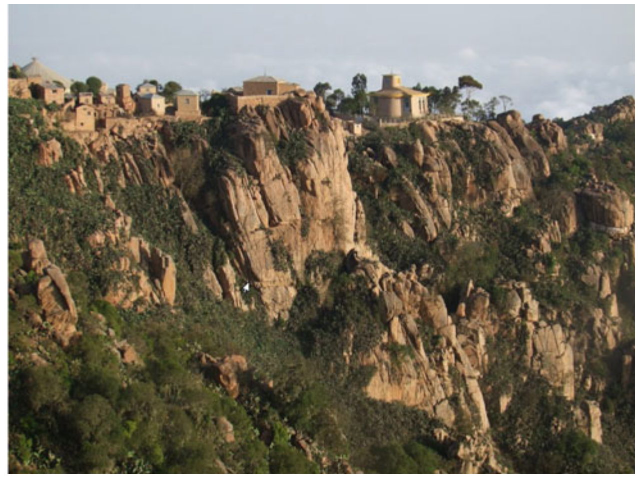

Fig. 2.6 Granitoids rocks (Ghebreab 1999) of the Nakfa terrane and the Bizen Monastry on the homonymous mountain (about 30 km East of Asmara)

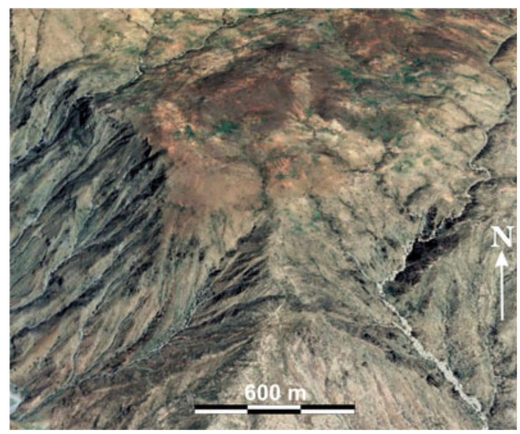

Fig. 2.7 Roras landscape 15 km SW of Nacfa. An unmetamorphosed horizontal sequence mostly composed of reddish Cretaceous laterites, but likely including older Phanerozoic successions, rests on a deeply eroded Neoproterozoic Nakfa terrane, mainly composed of greenschists facies metavolcanics. The flat tops of the Roras, now intensely cultivated, in the past were used as fortification places and host historical and archeological sites

(mainly metavolcanics) are a stand for remnants of a flat-lying sedimentary sequence (Fig. 2.7) (see more detail in the next sections).

The Nakfa terrane has an extension in the Danakil block, which is a continental fragment detached and rotated since the Miocene from the eastern margin of northern Ethiopia (Burek 1970) (Fig. 2.8). In the core of the Danakil block, a greenschist facies basement of limited extent with chlorite and sericite schists and granitoids is surrounded by Paleozoic to Mesozoic sediments (Vinassa de Regny 1921–1924; Bannert et al. 1970; Kazmin 1973; Bunter et al. 1998; Sagri et al. 1998).

In central Eritrea to the east of the Nakfa terrane, the Arag and Ghedem terranes sporadically expose their high-grade gneisses, metavolcanics and pelitic metasediments along a belt from the Takora region, at the northernmost Eritrea/Sudan border, to the Ghedem promontory 20 km south of Massawa. In the map of Fig. 2.3, it is assumed that along the coastal belt, rocks of the Arag/Ghedem terranes are overlain, from the northernmost outcrops, by the synrift Miocene Desset Formation.

The afore described multi-terrane classification for the Eritrean Neoproterozoic has been reconsidered in the last

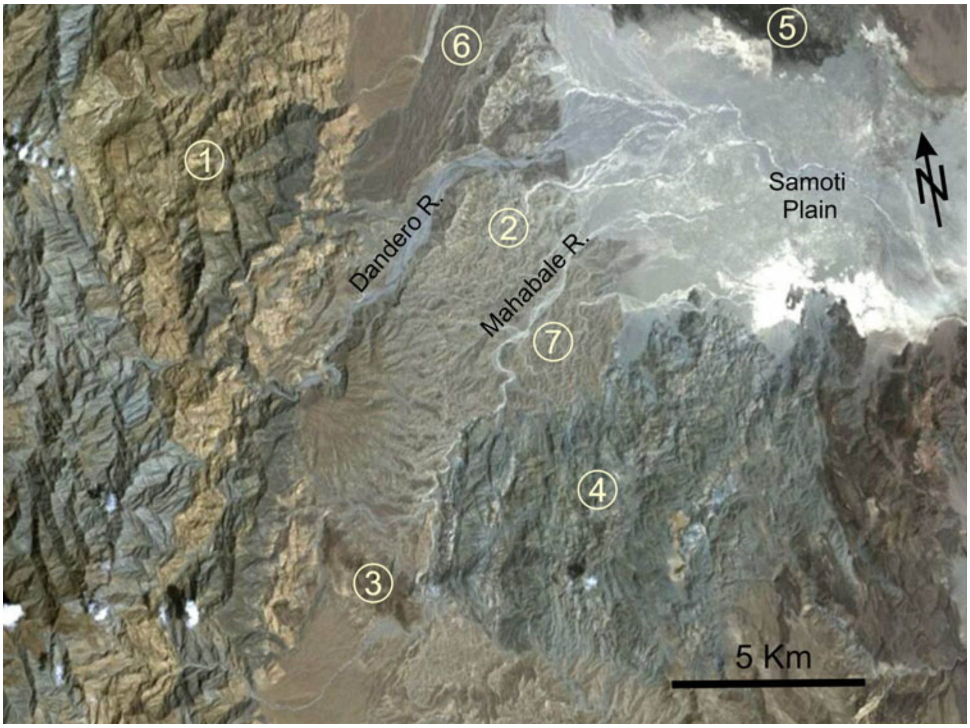

Fig. 2.8 Nakfa terrane along the western side of the Gulf of Zula/Alid volcano/Samoti plain corridor heading to the northern Afar depression. Starting from west: 1—on the left of the Dandero river Nakfa terrane with NNE-trending folded metamorphic limestone, sandstones, shales; 2—fluvial deposits of the Dandero and Mahabale rivers; 3—Adigrat sandstones brown outcrop across the valley; 4—on the high right reach of the Mahabale large greenish blue outcrop of the Nakfa terrane with metavolcanics and phyllites covered eastward by Neogene brownish rhyolitic domes and tuffs; 5—to the north of the whitish Samoti alluvial plain, basaltic lavas of the active Alid volcano limited to the west by a fault black; 6—in the northernmost central portion brownish boulder bed of the Dandero terraces; 7—Pleistocene sediments of the Dandero basin (Google Earth coordinates: 14°45′18″ N–39°57′29″ E)

twenty years and replaced by an alternative proposal of assembling the basement units within two comprehensive groups (“domains”): a high-grade gneiss domain, including the Barka, Arag and Ghedem terranes structurally overlain by a low-grade metavolcano-sedimentary domain, including the Hagar, Adoba Abi and Nakfa terranes, each domain comprising shear zone-bounded subdomains (Ghebreab 1996, 1999; Ghebreab et al. 2005). The low-grade domain largely widens in the central portion of northern Eritrea, and it is east–west longitudinally bounded by units of the high-grade domain (Fig. 2.3).

In the context of west/east Gondwana convergence, both domains are considered representatives of the same juvenile crust of accreted island arcs and granitoids (Fig. 2.9) (Ghebreab et al. 2005; Andersson et al. 2006; Avigad et al. 2007) with magmatism beginning at 800–850 Ma (Teklay 1997). Deformation patterns, metamorphic grades and crustal shortenings were predominantly acquired through suturing and escape tectonics, whereas large-scale extensional orogenic collapse brought high-grade lower crust slivers exposed to the surface (Fig. 2.10) (Ghebreab et al. 2005; Andersson et al. 2006). For instance, geochemical 40Ar/39Ar and field data indicate that some mylonitized Ghedem rocks were exhumed from a depth of c. 45 km to the upper crust (c. 15 km) between 593 and 567 Ma (Fig. 2.11) (Ghebreab et al. 2005).

In summary, from the early stages of rifting at ca. 870–840 Ma ago, the whole cycle of the East African Orogen amalgamation with the birth of the Gondwana (Stern and Johnson 2010) was achieved through distinct phases of exhumation at c. 550 Ma.

2.3.2 The Neoproterozoic and Paleozoic Glacial Episodes and Long Sediment Vacancies

The Neoproterozoic to Late Paleozoic history of the Eritrean crust records intervals during which wide peneplained areas were covered by continental and/or shallow marine sediments with traces of glacial activity. The discovery of these glacial deposits ignited a long debate, which started in the beginning of the last century when Italian prospectors and geologists active in northern Eritrea were reporting, within the Proterozoic metamorphic basement, conglomerates with coarse, predominantly quartzitic pebbles, some striated, in a hard, purple argillaceous-siliceous metamorphic matrix. Verri

Fig. 2.9 Tsada Amba carved in the granitoids of the Nakfa terrane with its monastery on top of the peak near Agaz SW of Keren

(1909) was the first to assume glacial origin for these coarse clastics. Later on, these conglomerates, named Rora Bagla conglomerates (Bibolini 1920, 1921) after the Bagla table-topped mountain (Rora means "mountain" in Tigre language) 25 km WSW from Nakfa, were commonly found associated with stromatolithic carbonates and interleaved within the basement schists in wide areas of the Hababland between Nakfa and the Haggar Mts., close to the Eritrea/Sudan border (Cecioni 1981), and more to the south in the Bizen Domain of the Heddas river 60 km SE of Asmara (Beyth et al. 2003), along the Mareb river near Adi Qwala, 80 km south of Asmara along the Eritrea/Ethiopia border, and in western Eritrea along the Setit river south of Barentu. Correlations were established with similar lithological successions belonging to the Tambien Group (Beyth 1972a, b) extensively outcropping in the Adwa-Adigrat area in northern Ethiopia. The Tambien Group is a 2–3 km thick, greenschists-grade metavolcanic and metasedimentary sequence deposited unconformably on the Nakfa terrane in an intra-oceanic arc platform setting during the last phases of the east/west Gondwana convergence (Avigad et al. 2007) (Fig. 2.12).

Unconformably above the Tsaliet metavolcanics indicating an igneous activity between 850 and 750 Ma (Tadesse 1996) lies the Tambien Group (Beyth 1972a, b). It comprises continental greywackes, slates- and shales-supported polymictic conglomerates, sometimes with striated pebbles, sporadic marine stromatolithic carbonates, and slates progressively becoming diamictitic with striated cobbles as large as 150 mm (Miller et al. 2003). These lithologies, sedimentary structures, paleogeographic reconstructions and chemostratigraphies ($\delta^{13}C$ and $^{87}Sr/^{86}Sr$ signatures of the carbonate rocks and $\delta^{13}C$ signature of the organic matter removed from the carbonates) are diagnostic of one or, possibly, two glacial events (Beyth 2001; Beyth et al. 2003; Miller et al. 2003, 2011; Alene et al. 1999, 2006; Avigad et al. 2007). Chronological constraints from syn-tectonic and post-tectonic granitoid intrusions, detrital zircon ages and correlations with global isotope stratigraphy provide for these glaciations, invading even the marine domain, a certain time interval bracketed between 720 and 660 Ma during the Cryogenian (Stern and Miller 2019). This interval recorded in the Eritrean Neoproterozoic coincides with that of many Neoproterozoic "Snow Ball Earth" deposits from other continents (Hoffman and Scharag 2002; Hoffman 2009) and is referred to as the Sturtian glacial episode (Stern and Miller 2019; Park et al. 2020). In addition, paleomagnetic

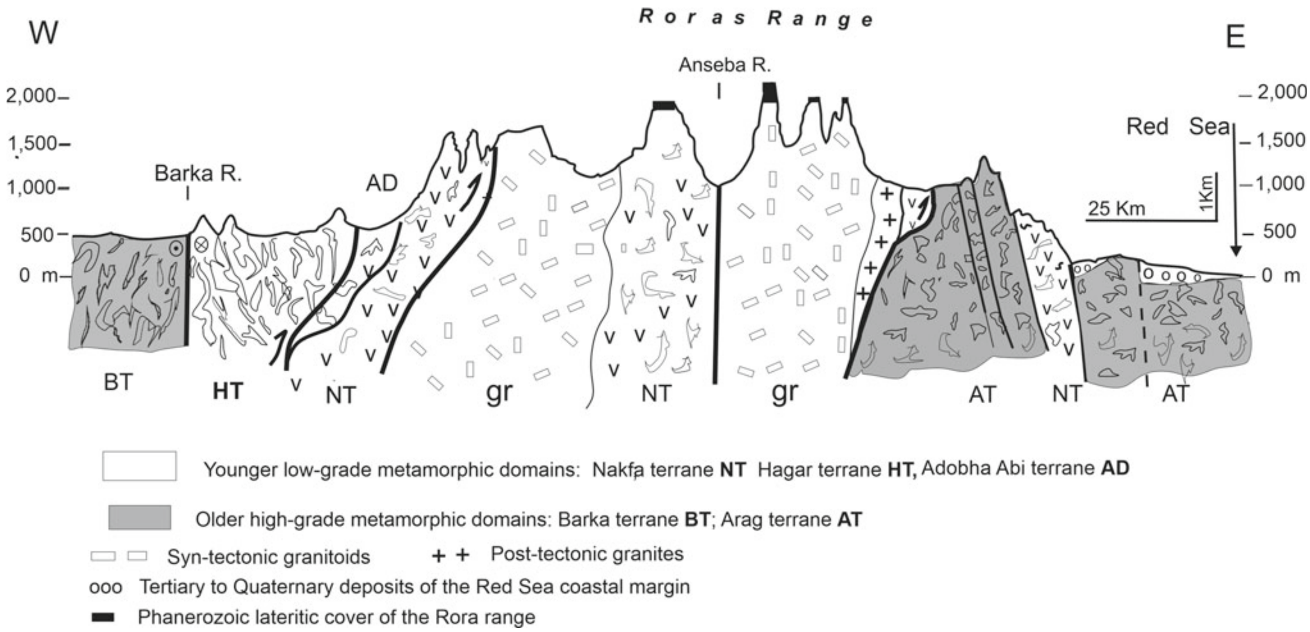

Fig. 2.10 General west to east, along the 16°9′ N parallel, geological section across the major Neoproterozoic structural units of the northern Eritrean plateau and adjoining Red Sea margin. For trace, see Fig. 2.20. *Legend* Older high-grade metamorphic domains: Barka terrane (BT) and Arag terrane (AT); younger low-grade metavolcanic and sedimentary domains: Nakfa terrane (NT) and Hagar terrane (HT). Main sources: Drury and Berhe (1993) and Ghebreab et al. (2005). Additional sources: Drury et al. (1994); De Souza Filho and Drury (1998)

Fig. 2.11 Folded gneissic rocks of the high-grade Ghedem Domain affected by Tertiary faulting along the western side of the Gulf of Zula-Samoti

reconstructions provide evidence that the East African portion of the Arabian-Nubian Shield was then drifting at low latitudes in a pan-glacial Earth (Hoffman et al. 1998; Stern and Miller 2019).

The Paleozoic deposits are exposed in sparse outcrops in central Eritrea from Mendefera southward to the Eritrean/Ethiopian border (Kumpulainen 2007), in the northern rim of the Danakil block (Kumpulainen 2007) and

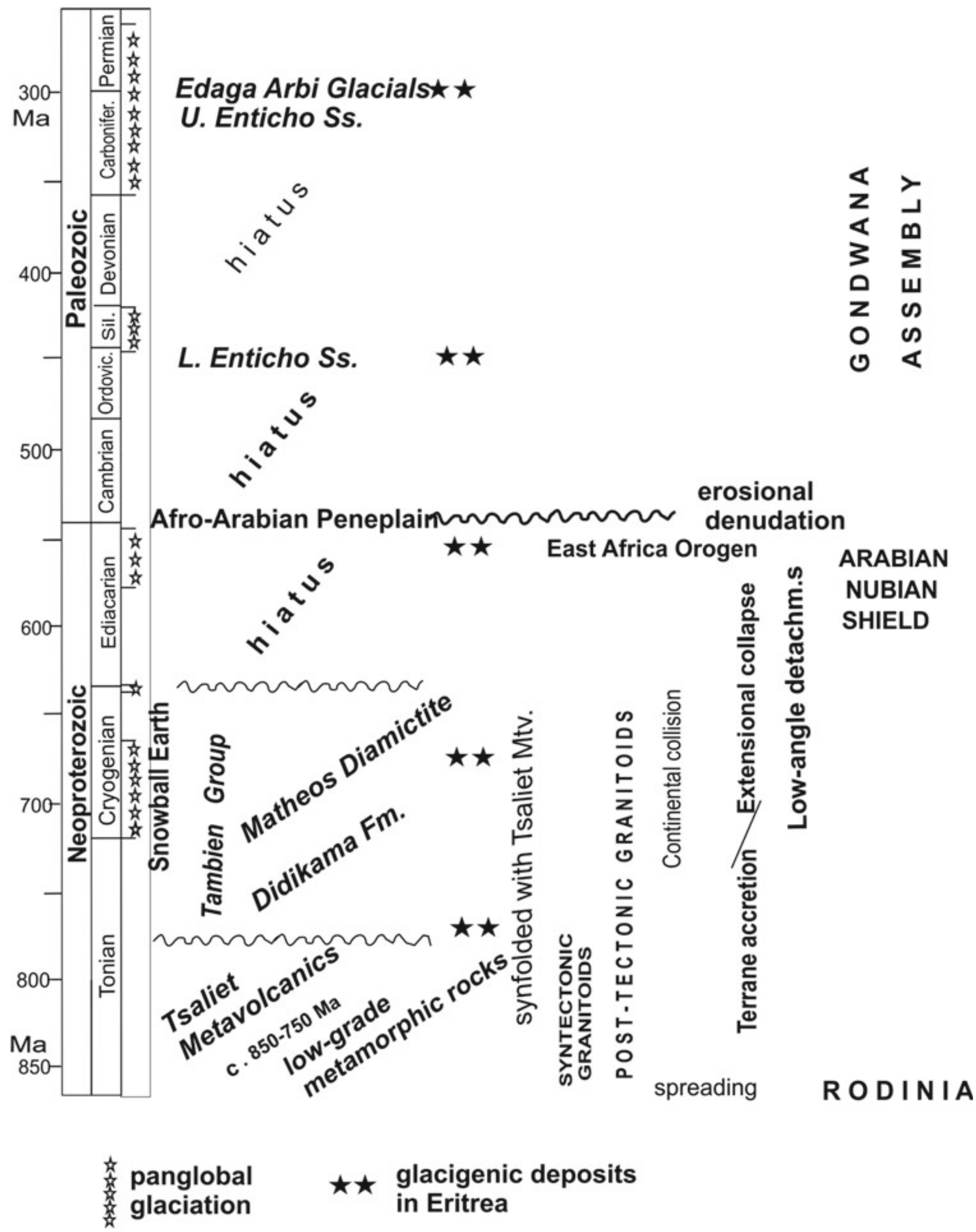

Fig. 2.12 Schematic stratigraphic chart for the Neoproterozoic and Paleozoic of Eritrea with pan-global glaciations and glacigenic deposits in Eritrea. Structural events of the East Africa Orogen in Eritrea

in a northernmost outcrop 20 km SSE of Asmara (Ghebreab and Talbot 2000). Those in central Eritrea are contiguous to the type area of Paleozoic deposits in the Adwa-Adigrat region of northern Ethiopia where Dow et al. (1971) and Beyth (1972a, b), drawing inspiration from Blanford (1869, 1870) and Merla and Minucci (1938), described the Enticho Sandstone. This is a c. 200-m-thick succession with a basal tillite including clasts up to boulders, a lower glacigenic sandstone unit and an upper shallow marine sandstone unit (Bussert and Dawit 2009) (Figs. 2.13 and 2.14). The Enticho Sandstone makes lateral and, in places, vertical upward transition to the Endaga Arbi Tillite, a glacial/periglacial deposit with shales and tillites with striated clasts and, at places, an upper shallow marine sandstone.

The age of both Paleozoic glacigenic formations is still uncertain due to complex local stratigraphy characterized by several facies changes, lack of detailed mapping and difficulty of finding diagnostic fossils. Based on correlation with glacial deposits in northern Africa and Arabia, using palynomorphs, hydrozoa and trace fossils determinations, some authors (Dow et al. 1971; Merla et al. 1979; Saxena and Getaneh 1983; Kumpulainen 2007) have assumed an Ordovician-Silurian age for the glacigenic event of the Endaga Arbi Tillite. Alternatively, a latest Carboniferous–Early Permian age of the Endaga Arbi Tillite has been documented through palynomorphs and trace fossils by Bussert and Schrank (2007). The same authors have tentatively suggested a younger Enticho Sandstone, heteropic of the

Fig. 2.13 Glacial grooves on the Early Paleozoic Enticho Sandstone near Eubucao pass (Amba Soira)

Fig. 2.14 Early Paleozoic Enticho Sandstone unconformably above the slates of the Neoproterozoic Nakfa terrane in the Dahan Dahan range (2800 m elev.), 12 km ENE of the Amba Soira

Carboniferous-Permian Endaga Arbi Tillite, and an older Enticho Sandstone of Ordovician-Silurian age. This would only in part agree with the Late Carboniferous–Early Permian paleomagnetic age estimation by Kidane (2013, 2014) for the Enticho Sandstone and for the Endaga Arbi glacials. Sacchi et al. (2007) also assumed that the occurrence of clastic volcanic material in some tillites could be better explained if referred to the diffuse Permian volcanic activity in Africa.

Although these uncertain litostratigraphic allocations are under investigation in Eritrea and Ethiopia, the two sets of dating (Ordovician-Silurian and Late Carboniferous–Early Permian) correspond chronologically to Paleozoic global glacial episodes (Bussert and Schrank 2007; Bussert and Dawit 2009; Bussert 2014) (Fig. 2.12). The petrography and geochemistry of the arenaceous glacial and periglacial material (Lewin et al. 2018) indicate a big difference in size between the large ice sheet covering Gondwana during the Ordovician-Silurian and the scattered local glaciers during the Carboniferous-Permian.

On a regional scale, the Paleozoic deposits, recently investigated in detail by Bussert and Schrank (2007), Hambrey (2007), Bussert and Dawit (2009), Bussert (2010), Dawit (2010) and Lewin et al. (2018), rest unconformably on a striated Proterozoic basement (Arag terrane and Bizen Domain). This unconformity is the Afro-Arabian Peneplain (Miller et al. 2003), dated c. 550 Ma and traceable from Morocco to Oman on the beveled Arabian and Nubian Shields that already merged into the Pan-African crust at the end of the East African orogeny.

Only a few hundred meters of sediments, resting on the Afro-Arabian Peneplain, attest a c. 330 million years interval (from c. 550 Ma—terminal phases of the East Africa Orogen to c. 220 Ma—the base of the Adigrat Sandstone) with very low sedimentation rates. The sediments were ice-proximal assemblages deposited on continental platform, in some places bordered by the sea (Kumpulainen 2007). It is likely that regional uplift events caused only episodic denudation in a low relief landscape, with no thick molasse deposition.

For its relevance, the Afro-Arabian Peneplain can be qualified as a *paleogeological hiatus surface* (Friedrich 2019; Friedrich et al. 2018), i.e., a continental-scale unconformable contact.

2.3.3 A Rifted Gondwana and the Jurassic Sea

The prolonged tectonic quiescence of the Eastern Africa was interrupted at c. 250 Ma ago by the Permo-Triassic rifting of Gondwana which, starting from South Africa (Karroo), produced the separation of Tanzania, Kenya and Madagascar through a dense network of branching rifts.

The unconformity, at the base of this event more or less at the transition Paleozoic/Phanerozoic, was called in Ethiopia and in Eritrea “First erosional cycle” by Merla and Minucci (1938) and PS2 planation surface by Coltorti et al. (2007) (Fig. 2.15).

The clastic supply, provided by the diffuse rift-shoulder uplift (Purcell 2018) and designated as “Adigrat Sandstone” (Blanford 1870), was dispersed over very large areas by braided or low-sinuosity rivers. They were crossing a vast ramp gently sloping southeast toward the rim of the Nubian plate and the Indian Ocean. The Adigrat Sandstone, a medium-grained cross-bedded fluvial arenite, thick up to 700 m, with interbeds of conglomerates, and its Late Triassic to Middle Jurassic correlative arenites, are widespread in the whole Horn of Africa and Arabia, including the Mazera Sandstone of Kenya, the Negrenegre Sandstone of Tanzania, the Minjur Sandstone of Saudi Arabia and the Kohan Series of Yemen (Mohr 1962; Kamen-Kaye 1978; Bosellini 1989; Bosellini et al. 1997). In Eritrea, the Adigrat Sandstone is present in the Danakil block (Vinassa de Regny 1930; Bannert et al. 1970; Sagri et al. 1998).

The Adigrat Sandstone, which in northern Ethiopia unconformably covers the Enticho Sandstone and the Endaga Arbi Tillite, has an uncertain lithostratigraphic characterization in Ethiopia and in Eritrea as well. Recent papers (Merla et al. 1979; Garland 1980; Kumpulainen 2007; Bussert and Schrank 2007), in fact, assigned its classical type section, located close to Adigrat in the northernmost Ethiopia, to the Enticho Sandstones rather than to the Adigrat Sandstone. A thorough discussion on this matter, mainly due to the commonly irrelevant fossiliferous content of the Adigrat Sandstone and meager distinctive features, is given by Merla et al. (1979) and Kumpulainen et al. (2006). Yet, Kumpulainen et al. (2006) addressed some distinctive features between the Adigrat Sandstone and the Enticho Sandstone, the former first being an immature fluvial type of sandstone, and the second latter a cross-bedded friable, predominantly fluvial sandstone, but probably, also partly, also of probable marine origin.

Uncertainties also involve the age of the Adigrat Sandstones regarded as Triassic–Liassic (Kazmin 1973) or Trias to Dogger (Merla et al. 1979), Triassic–Middle Jurassic (Tefera et al. 1997) and Early Jurassic (Hankel 2013). Recently, through pollen assemblages, Dawit (2010) obtained for the classical sections near Mekele and at the Blue Nile gorge a range in age from Late Triassic to Middle Jurassic. Through accurate facies analysis, the same author has documented some shallow marine character (lagoon and shoreface transition) of the Adigrat Sandstone, thus confirming what assumed by previous researchers (Merla and Minucci 1938; Dainelli 1943; Dow et al. 1971; Kumpulainen et al. 2006). Likely, the marine inflows were coming from the Neotethys across Arabia.

The general subaerial conditions of the Adigrat Sandstones deposition (Getaneh 2002) ended during the Middle to Late Jurassic when the vast platform on the East Africa craton was progressively inundated from SE due to further rifting events connected with the Gondwana dismembering and to the Mesozoic eustatic sea-level rise (Haq et al. 1988).

In Eritrea, evidence of this ingression is preserved in the northern Danakil block (regional maps by Brinkmann and Kürsten 1970) and further north in the Kezan area c. 100 km NW of Massawa (Bunter et al. 1998). At Kezan Anadarco and Eritrean Ministry of Mines, geologists identified 100 m of Antalo Limestones with dark and pure limestones, dolomitized and cherty limestones. At Kazen, the Antalo Limestone represents the northernmost evidence of the time-transgressive flooding of the Jurassic sea onto Eastern Africa covering a distance of more than 1500 km from the present-day Indian ocean during 30 Ma (from Toarcian to Callovian).

After the pioneer approaches by Vinassa de Regny (1930) who stressed the intense deformation of the block, the outcrops in the northern Danakil block were recently studied by Sagri et al. (1998). Unlike the classical succession, in which a full-fledged marine deposition is preceded by some hundreds of meters of a peritidal and lagoonal unit with

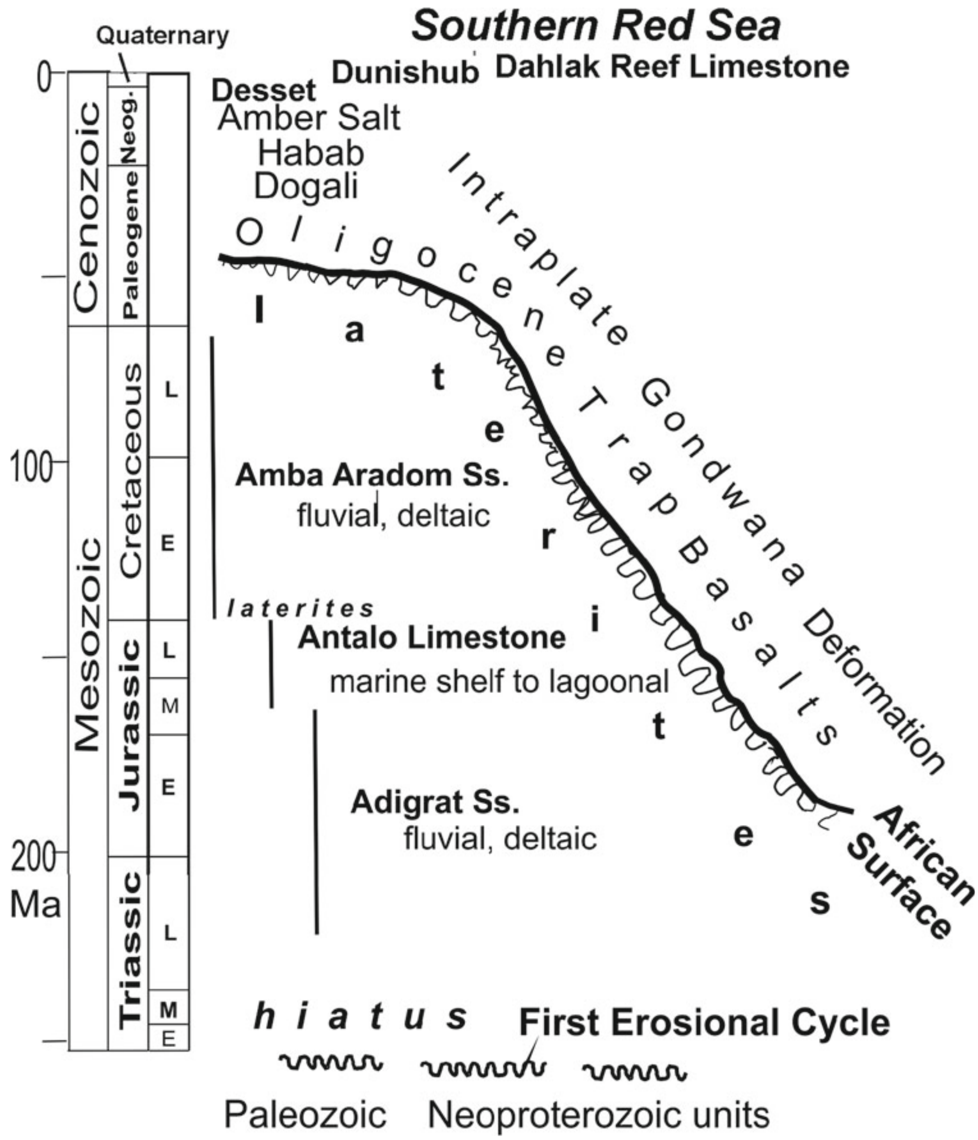

Fig. 2.15 Schematic stratigraphic chart for the Mesozoic and Cenozoic successions in Eritrea mainly relative to central Eritrea

characteristic shales and gypsum (e.g., Goatzion Formation—Assefa 1981; Russo et al. 1994), the succession near Adeilo includes just a few tens of meters of tectonized gypsum beds between the 700 m of Adigrat Sandstone s.l. (also including Paleozoic sediments) and the overlying marine Antalo Limestone (unit designated in Tigray by Blanford 1870). It comprises gray limestones, dark fetid marly limestones, marlstones, biocalcarenites, tempestite lumachellas and terrigenous calcareous sandy beds.

The Antalo Limestone is highly fossiliferous with nannoplancton, radiolarians, benthic foraminifers, brachiopods, pelicypods and ammonites ranging in age, in the Adeilo succession, from late Callovian to early Kimmeridgian (162–155 Ma) (Fig. 2.16) (see detailed biostratigraphy in Sagri et al. 1998).

The depositional environment of the Antalo Limestone varies repeatedly within shelf and open marine, in particular from inner to mid and outer ramp and, more rarely, to basin. Seven large-scale cycles from several tens to some hundreds of meters are framed between the major surfaces and make up seven depositional sequence due to the global fluctuations of the sea level over a period of time of about ten million years (Haq et al. 1988, Sagri et al. 1998).

In the Tigray type area, Bosellini et al. (1997) have defined as Antalo Supersequence a second-order depositional sequence including the Antalo Limestone and overlying Agula Shales with a basal gradual transition to the Adigrat Sandstone and an upper marked unconformity with the Aptian-Albian Amba Aradom Sandstone.

The Antalo Limestone in the Danakil Alps is at least one thousand meters thick (Hutchinson and Engels 1970; Sagri et al. 1998). This thickness, never recorded for this unit in the whole Horn of Africa, has prompted the idea of an ancestral Red Sea subsiding basin sub-parallel with and

Fig. 2.16 Antalo Limestones in the NW rim of the Danakil block, near Adeilo, 30 km SSW of Tiho. In this middle portion of the formation, thick beds of dark gray, sulfide-rich calcilutites and dark gray marly interbeds are partially replaced upward by yellowish calcilutites, marlstones and calcarenites. This succession, deposited in a subtidal environment, is particularly rich in ammonites and other molluscs, echinoderms, brachiopods, macro- and micro-foramifers and calcareous nannoplancton of Oxfordian-Kimmeridgian age (further details in Sagri et al. 1998). General dipping toward N

underlying the present-day Red Sea (Merla et al. 1979; Bosellini et al. 1995). A similar outstanding thickness (1700 m—Hutchinson and Engels 1970) for the Adigrat Sandstone at Ras Andabba has been disputed by Merla et al. (1979, p. 27).

In the southeastern part of the Danakil block, marly levels (Fig. 2.17) (possible correlative of the Agula Shales of the Mekele type section) mark the unconformable transition from the Antalo Limestone to the overlying short sections of the Amba Aradom Sandstone (Assefa 1981, formerly Upper Sandstone in Merla et al. 1979) composed of an alternation of fine to coarse-grained sandstones, siltstones and shales of braided fluvial environment. The succession Adigrat-Antalo-Amba Aradom was interpreted by Dainelli (1943) as an ideal complete transgressive-regressive cycle "with a chronostratigraphic and lithologic symmetry of ingression and regression deposits" (Merla et al. 1979) on a continental scale. The recognition of a distinct regional angular unconformity, marked by laterites, between an eroded Antalo Limestone and the Amba Aradom Sandstone in the type area of Mekele (Bosellini et al. 1997) has introduced a prominent element of discontinuity in the sedimentary cycle.

Toward the end of the Jurassic, there was a regional basin inversion with a progressive shallowing of the Antalo sea up to emersion, possibly due to a deepening of coastal basins along the embryonic Indian Ocean accompanied by basement highs rising. Because of its widespread occurrences, the Amba Aradom formation has many putative correlatives in the Horn of Africa, ranging in age from post-Kimmeridgian to pre-Middle Eocene. The most commonly accepted age for the Amba Aradom sandstone is Early Cretaceous. In the Harar region, Aptian to Albian marine macroforaminifer orbitolinids testify a short Early Aptian sea transgression during the Amba Aradom fluviatile deposition (Gortani 1973; Bosellini et al. 1999a; Bosellini et al. 2001). The similarity between Adigrat and Amba Aradom sandstones in the field and the difficulty of biostratigraphic or petrographic discriminations make uneasy to distinguish them when the interposed Antalo Limestone has been eroded or not deposited. In this case, the whole arenaceous body has been given different names, such as Tekeze Sandstone, Yesomma Sandstone (Figs. 2.18 and 2.19), Nubian Sandstone or other local names.

2.3.4 A Long Waiting Before the Volcanic Outburst: The Laterites

A layer of lateritic weathering typically covers truncated structures of the Eritrean and Ethiopian Proterozoic basement (among others, Mohr 1962; Davidson 1983). In the Ethiopian northern Tigray region close to central Eritrea, a laterite layer rests on mildly folded and eroded Mesozoic sedimentary sequences (Abul-Haggag 1961; Bosellini et al. 1997). From the Mareb river to the northern border, laterites are found in Eritrea conformably beneath the Trap basalts (Dainelli 1943). Moreover, a very informative map by Drury et al. (1994) shows that the laterites are diffusely present in northern Eritrea north of the line Massawa-Akordet representing the only pre-Oligocene sedimentary deposit.

The pre-Trappean laterite pedogenesis, firstly pointed out by Blanford (1870) and accurately described and discussed by Dainelli and Marinelli (1912) and Dainelli (1943), is commonly a few meters thick, but it can be over 100 m in

Fig. 2.17 Cycles of predominant yellowish marls and darker calcilutites make up the highest levels (Oxfordian-Kimmeridgian) of the Antalo Limestone plunging beneath the Edd-Thio coastal plain (Danakil block). In the background on the right the lower units of the Jurassic succession above the basement

places (Fig. 2.20). It can be red (more limonitic clay) above the crystalline schists or, more rarely, light colored (more kaolinitic clay) above a felsic substratum (Dainelli 1943). Laterite can appear as a massive clay, or as a densely brecciated clay with diffusely slickensided angular fragments (Figs. 2.21 and 2.22). Nice outcrops are in Asmara near the airport, on the hills downtown and near Kudus Geourgis Church 4 Km away on the road to Massawa (South of Asmara the contact with the overlaying Trap basalts is exposed along the Mareb river and near Shiket on the way to Adi Qwala).

In northern Eritrea, the pre-Trappean laterites compose the outstanding flat top of small, steep-sided plateaus forming the Roras Mts., a spectacular segmented range of Nakfa terrane granitoids and metavolcanics, elongated NNW for c. 100 km west of Nakfa with a constant elevation around 2500 m asl (Figs. 2.23 and 2.24).

Laterites are also present in Yemen and southern Sudan between the Early Oligocene basalts and the basement or Mesozoic strata (Davison et al. 1994; Al-Subbary et al. 1998; Kenea et al. 2001).

Laterite occurrences are particularly significant since they are considered indicative of the peneplanation of an area with extremely low relief and low elevation. In Eritrea and Ethiopia, the erosional event connected with the laterites has been variously labeled as “2° ciclo erosivo” (second erosive cycle) (“erosione prevulcanica”, Merla and Minucci 1938); “pre-Trappean peneplanation” (Mohr 1962) and “PS4” (the last out of a sequence of PSs the oldest being pre-Paleozoic, Coltorti et al. 2007) and recognized on a continental scale as the African Erosion Surface (Burke and Gunnell 2008).

Andrews Deller (2006) has mapped the different mineralogical facies of the laterites in a test area SW of Asmara and has determined a minimum Ar/Ar laterization age of c. 40 Ma for a laterite beneath the Trap basalts of Asmara (Andrews Deller 2003). Also Perelló et al. (2020) obtained a K/Ar Paleocene, Thanetian (59 Ma) age from a supergene alunite in a laterite close to Asmara.

Typical of regions with hot and humid tropical climate, the laterites suggest a long tectonic and morphological stability of a territory with modest clastic inflows. Moreover, their altimetric significance (they developed close to sea

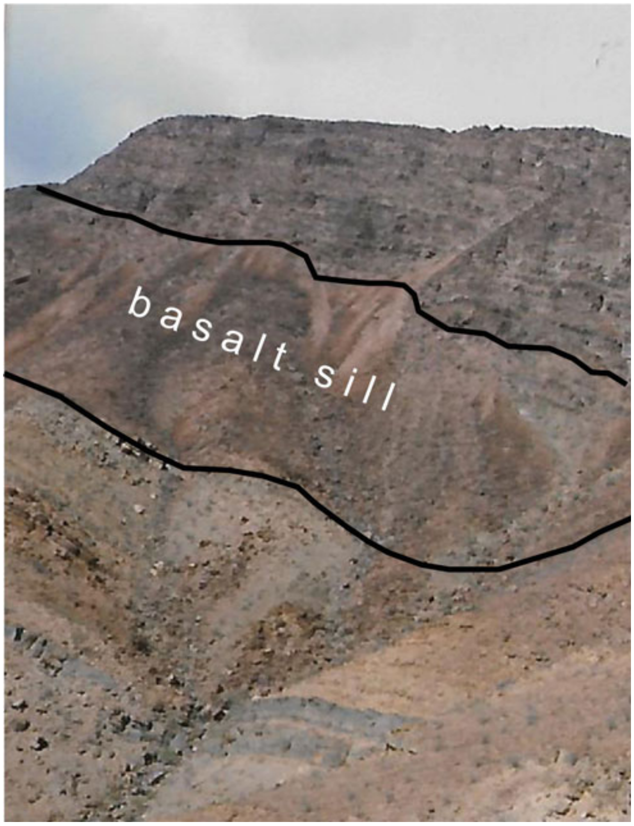

Fig. 2.18 Transition from the Kimmeridgian portion of the Antalo Limestones to the Yesomma Sandstone in the SE rim Danakil block, Belo Buye/Dahan region, 90 km SE of Tiho. The Yesomma Sandstone at the top of the mountain overlies a c. 20 m thick Tertiary basalt sill (brownish) with interposed varicolored shales and marlstones. Downsection, dark fetid limestones and yellowish marlstones are intruded by further basalt sills (light brown). General dipping toward NE

level) allows to assess the amount of uplift to which these laterite layers (and the basement if present) have been subjected. This possibility was put forward by Dainelli (1943) and applied to a vast area of Eritrea from the Mareb river to Nakfa by Drury et al. (1994). In the Asmara-Massawa transect, the elevation difference between laterite occurrences on the plateau and their correlative close to Massawa is c. 2.2 km driven by faulting in the Red Sea escarpment (Drury et al. 1994). Along the same sector, a similar magnitude of displacement is confirmed by the occurrence of large fossil mammals on the plateau and on the lowlands (see Sect. 2.4.1).

2.4 Syn-Rift Volcanic and Sedimentary Rocks on the Plateau and on the Lowland

2.4.1 Syn-Rift Successions on the Plateau

The Traps: flows and dikes

Since the Middle Eocene (about 40 Ma ago) in Kenya and southern Ethiopia and with a peak of activity in the Early Oligocene (about 30 Ma ago) in the remainder of Ethiopia, in Eritrea, Sudan and Yemen, a vast volcanism with predominant flood basalts resulted in the Ethiopian Large Igneous Province (ELIP). This volcanic assemblage is traditionally referred to as the Traps from Blanford's (1870) Trap Series. Following the classical single-stage Southern Red Sea evolution reconstruction with rifting since 30 Ma, more or less in connection with the arrival of the Afar plume,

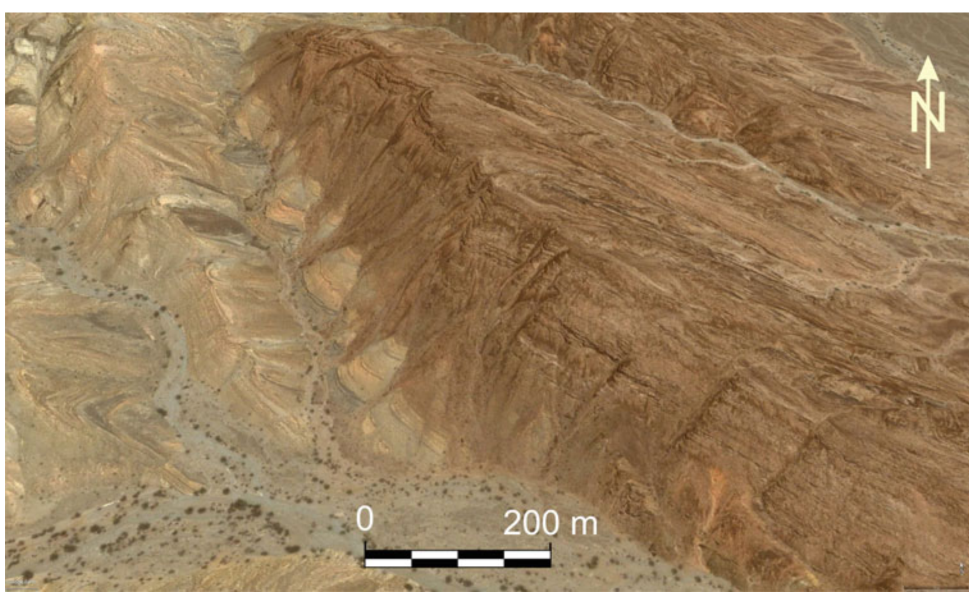

Fig. 2.19 Details of the Antalo Limestones/Yesomma Sandstone transition in a Google Earth image (site to the north of the outcrop in Fig. 2.18). Pinkish, yellowish and whitish marlstones and shales with rare black-gray fetid calcilutites are followed by a 20-m-thick Tertiary basalt sill (see also Fig. 2.18). General dipping toward NE. (Google Earth coordinates: 13°59′12″ N–49°18′46″ E)

Fig. 2.20 Poorly indurated and thin ferricrete horizon (a) separates the Neoproterozoic basement (b) from the Asmara Traps (c) (courtesy Sara Passerini)

and spreading since 5 Ma (e.g., Bosworth and Burke 2005), the Traps and associated sediments are considered syn-rift. A different, more complex tectonic history could be derived if the double-spreading model, resumed through new data by Almalki et al. (2014) for the conjugate Arabian margin, could be documented and adopted also for the Eritrea margin (see Sect. 2.4.3).

The Trap volcanic cover is widespread in Ethiopia with sections some 2000 m thick (Abbate et al. 2015), whereas there are only few isolated outliers of some hundreds of meters thick in Eritrea. They are located south of the Barentu/Massawa line and north of the Mareb river. One of the largest and better exposed outcrop is the Asmara outlier (the Asmara plateau) which is 100 km long in a NS direction and 40 km across (Fig. 2.25). With the constant interposition of laterites, the Asmara outlier rests on the basement in its northern sector and on the Adigrat Sandstone in the southern sector.

In the Asmara outlier, Zanettin et al. (1999, 2006a, b) not only recognized one formation of the typical Ethiopian Trap succession (the transitional tholeiitic Aiba/Alaji formation with a K/Ar age of c. 29 Ma), but also established two new formations: the tholeiitic Asmara Basalt, coeval with the Aiba/Alaji Basalt and youngest in the succession, and the mildly alkaline Adi Ugri Basalt with an Ar/Ar age of c. 22 Ma in the upper portion. Four basalts above the laterite or immediately over gave Ar/Ar ages from c. 33 Ma to c. 28 Ma (Drury et al. 1994). Ignimbritic events intervene in the basaltic successions, such as the 24 Ma old Serae Rhyolite (Zanettin et al. 2006a) intercalated in the lower Adi Ugri Basalt and correlative of the alkaline trachyte-rhyolite lavas and domes of the Senafe and of the Adwa-Axum (Ethiopia) areas and of the somewhat older Alaji Rhyolites of the typical northern Ethiopia Traps. At Senafe, a spectacular dome impends over the town, and its related lavas rest on both the Enticho Sandstones and the basement with interposed laterites (Fig. 2.26).

With reference to the distinction of the Ethiopian basalts based on the TiO_2 content (Pik et al. 1998), the NW Ethiopian basalts and the Asmara and Sudanese outliers are included in the low-Ti province (Rooney 2017). In Eritrea, Teklay et al. (2005) found a temporal transition from low to high TiO_2 tholeiitic basalts at the Amba Tekera, 40 km NW of Asmara. The upper basalt levels of the Amba Tekera gave an Early Miocene whole-rock Ar/Ar age of ca. 20 Ma (Teklay et al. 2003).

Connected to the Trap volcanism is an impressive, 7-km-wide and at least 50-km-long, sub-vertical basalt dike swarm that forms small parallel ridges, a few km long, in the landscape between Debaroa and Asmara (Mohr 1999). They impressed Dainelli and Marinelli since “They look like gigantic walls with horizontal basalt prism” (Dainelli and Marinelli 1912). The dike swarm cuts across the Asmara Basalt in a NNE direction. By contrast, many basalt dike swarms, scattered along the escarpment between Asmara and the Gulf of Zula and considered as Trap feeders, are aligned with the Red Sea spreading axis. They are subvertical with an average NNW strike, often pervasively sheared with slickensides mainly along the chilled margins. The slickensides are often dike-parallel with mainly sinistral polarity. Dikes with dip-slip (subvertical) shearing are also observed, and the best evidence for strike slip is on the road Asmara-Massawa between 25 and 28 km downhill Nefasit

Fig. 2.21 Typical laterite-facies profile above the Neoproterozoic basement and beneath the Trap basalts of the Asmara plateau. In the foreground, the grayish laterite envelopes relics of weathered basement. At distance, it becomes mottled with an irregular mixture of bright whitish kaolinitic and red iron-mineral lumps. In the higher slopes, a brownish ferricrete horizon develops beneath the basalt flows. Mereb river valley 30 km south of Asmara (courtesy Sara Passerini)

(1993 campaign with P. Passerini). Strike slip movements parallel to, or at small angles with, the Afar rifts have been considered significant in stretching the continental crust (Abbate et al. 1995) (Fig. 2.27).

The Mendefera intertrappean bed

In the volcanic story of the Asmara plateau, the deposition of a continuous sedimentary level at the transition between the Asmara Basalt and the Adi Ugri Basalt denotes a long period of volcanic and tectonic quiescence (Abbate et al. 2012, 2014). This level, quite evident in the landscape (e.g., along the Mendefera-Barentu road), is a few tens of meters thick and consists of fluviolacustrine silts and mudstones, commonly red and subordinately green and grayish, with lignite seams and tree trunks (Fig. 2.28). Basalt flows and dikes and rhyolitic layers are also present. In a section near Mendefera, the time constrains of this intertrappean intercalation are an Ar/Ar age of c. 29.0 Ma for the top of the Asmara Basalt beneath the sediments and an Ar/Ar age of c. 23.6 Ma for the base of the overlying Adi Ugri basalts (Abbate et al. 2014). The same site has yielded the fossil remains of two Late Oligocene/Early Miocene proboscideans (teeth and maxillary bone of a deinothere and a gomphoteriid) (Abbate et al. 2012), confirming a previous finding by Vialli (1966). The presence of *Gomphotherium sp*. signals the oldest occurrence of this proboscidean before its dispersal toward Asia and Europe.

Further proboscideans remains were found in the lowland intertrappean beds of Dogali near Massawa (Sagri et al.

Fig. 2.22 Extremely fine-grained pinkish kaolinitic laterite horizon wasted by an enigmatic blueish fluid

1998; Shoshani et al. 2006), i.e., two thousand meters below the Mendefera fossils.

2.4.2 Syn-Rift Successions on the Lowland

The Dogali section

The variety of hard rocks and sediments found on the plateau also crop out, with very good exposures, in the classical section near Dogali (20 km east of Massawa) (Porro 1936; Kazmin 1975; Drury et al. 1994; Antonielli et al. 2009) (Fig. 2.29). From bottom, orthogneiss and metasediments of the Ghedem Domain and a laterite layer (Ghebreab and Talbot 2000) with some basement pebbles heralding a precursory plateau rifting (Ghebreab 1998) are overlain by c. 200 m of Trap basalts with the lowermost flow Ar/Ar dated 28.0 Ma (Drury et al. 1994). They are followed by the c. 500-m-thick Dogali Formation (Kazmin 1973) consisting, in the lower half, of lacustrine yellow, brown laminated

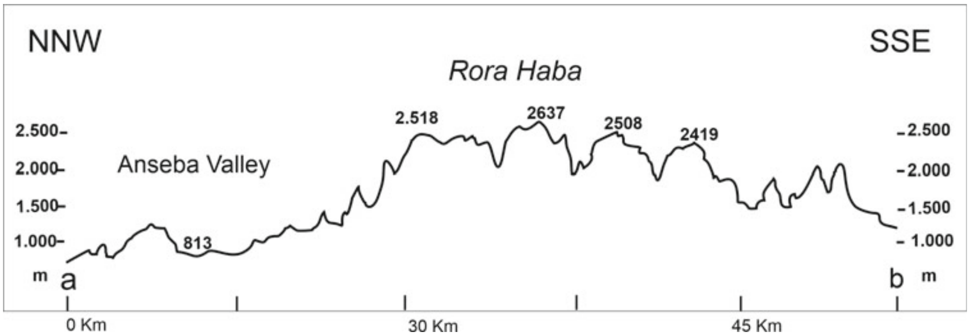

Fig. 2.23 NNW–SSE topographic section along the Roras Range in the Rora Haba area. Remnants of an original wide, faulted and weathered plateau with an average elevation of 2500 m are clearly visible in the central part of the profile. The location of the a–b section is reported in Fig. 2.24. Google Earth coordinates: a: 17°11′35.43″ N, 38°10′25.26″ E; b: 16°07′24.37″ N, 39°30′28.58″ E

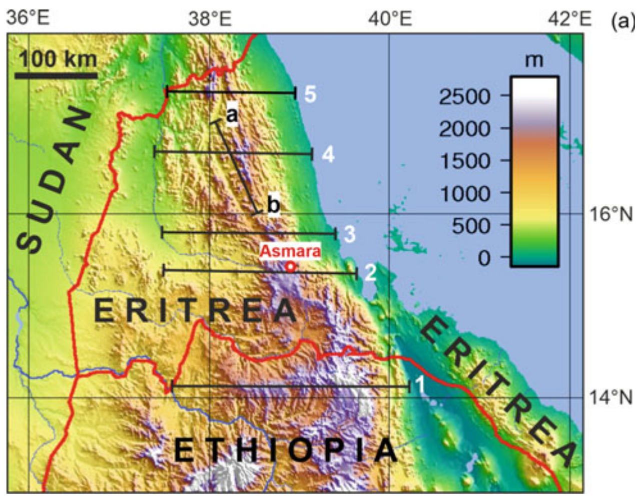

Fig. 2.24 Location of representative topographic profiles; profile a–b Fig. 2.23; profiles 1–5 Fig. 2.37

diatomite clay and silts alternating with basalt flows and tuffs. A channelized conglomerate up to 10 m thick with basalt pebbles and silicified tree trunks yielded Late Oligocene/Early Miocene primitive proboscidean remains (Sagri et al. 1998; Shoshani et al. 2006) and a basalt flow, 15 m above this fossiliferous level, has an Ar/Ar age of 26.8 Ma. In the upper portion of the formation, after 200 m of basalt, the succession includes an alluvial facies with sands and boulders up to 40 cm derived from basement rocks, basalts and felsic volcanites. The occurrence of basement rocks in the clastics indicates that faulting on the plateau and on the plateau shoulders was ongoing. The Dogali Formation terminates with a transition to cyclic arrangements of fluvial to coastal plain associations and episodic marine occurrences with coralline patch reefs, gypsum layers and mollusk shells (*Pecten*, *Ostrea*). Corals and mollusks give a Middle to Late (?) Miocene age (about 10 Ma) for the top of the formation (Zuffardi Comerci 1936; Montanaro Gallitelli 1939, 1973; Selli 1973).

Late Oligocene/Early Miocene paleoenvironmental reconstruction of central and northern Eritrea

The syn-rift Mendefera and Dogali sections with their plateau and lowland development can provide good hints for a Late Oligocene/Early Miocene paleoenvironmental reconstruction of central and northern Eritrea. Initially, the lowland and plateau regions of today were a unique, low-elevated wide area characterized by a relatively thin, volcanic cover (400–500 m as indicated by the Dogali section and the Asmara plateau region), underlain by laterites resting on basement rocks planed by older denudation cycles and dislocated by minor, local rifting. It is difficult to figure out the areal extension of the volcanic cover in Eritrea since the Trap volcanites are missing in northern Eritrea north of the Asmara plateau (Teklay et al. 2005) and resurface with a few tenths of meters thick outcrop only in the Red Sea Hills in Sudan (Kenea et al. 2001).

The basalt thicknesses on the plateau and on the lowland suggest that the volcanic outpourings were rather meager if compared with those of the nearby northern Ethiopian plateau. Moreover, they were intermittent (see intertrappean episodes), distributed in a rather long time interval between 29 and 20 Ma and possibly related to "narrow tongues" (Ebinger and Sleep 1998) of Afar plume material. As regards the environmental conditions, on the basis of similar lithostratigraphical and faunistical content, we can assume that the whole region was covered by a vegetation of moist tropical forest favorable to host proboscideans (see the Ethiopian Chilga lacustrine deposits, Tana Lake, Yemane et al. 1987).

The biophysical environment inferred for the northern and central Eritrea hinterland can also be assumed for the Sahamar-Sahel coastal region, the SSE/NNW elongated, narrow (250 by 25 km) coastal belt at the foot of the escarpment from Massawa to the Sudan border. For this poorly studied region, the available literature reports numerous patches of Neoproterozoic terranes, probably Arag, sporadic outcrops of laterites and Oligocene Dogali

Fig. 2.25 View of the Mendefera–Adi Quala plateau (50 km south of Asmara). The evident break in slope is due to the Trap basalts overlying the basement through a discontinuous interposed unit that in different near localities can be the Oligocene rhyolite body or Paleozoic sediments. An upper less clear break can be appreciated within the Traps and corresponds to an intertrappean sedimentary level

Fig. 2.26 View from the Mendefera-Adi Quala plateau toward SE across the Mareb valley with Miocene trachyrhyolitic volcanic plugs interspersed on the plain. In the near background, well evident is the triangular shape of the Metera plug incumbent on the Metera village and on the town of Senafe. Metera was an important Axumitic town (first centuries AD). In the extreme background, the ridges of the Eritrean plateau rim include the Amba Soira—the highest mountain peak (3250 m) in Eritrea

Fig. 2.27 Tertiary basalt swarm near Nefasit 40 km East of Asmara (see text)

Fig. 2.28 Red silts, mudstones and conglomerates of the intertrappean episode in the Mendefera Trap section. This intercalation yielded Oligocene big mammal faunas (Vialli 1966; Abbate et al. 2014). Despite their usual limited thickness, these intercalations cover a 1–2 Ma of volcanic quiescence

deposits, widespread fossiliferous Neogene to Pleistocene shallow marine and alluvial sediments (Porro 1936; Kazmin 1973; Merla et al. 1979; Drury et al. 1994). As shown in ENI (1971) and Kazmin's (1973) maps, high-grade basement chunks and Paleozoic metasediments can be found aligned in a 15-km-wide NNW-striking belt close to the foot of the escarpment.

Early to Late Miocene syn-rift deposition and volcanism

The syn-rift deposition continued during the Early to Middle Miocene with the deltaic to marine sandy and shaly Habab Formation. It can be seen in sparse exposures among the coast and widespread and very thick (up to 2500 m) covers in offshore drillings (Savoyat et al. 1989).

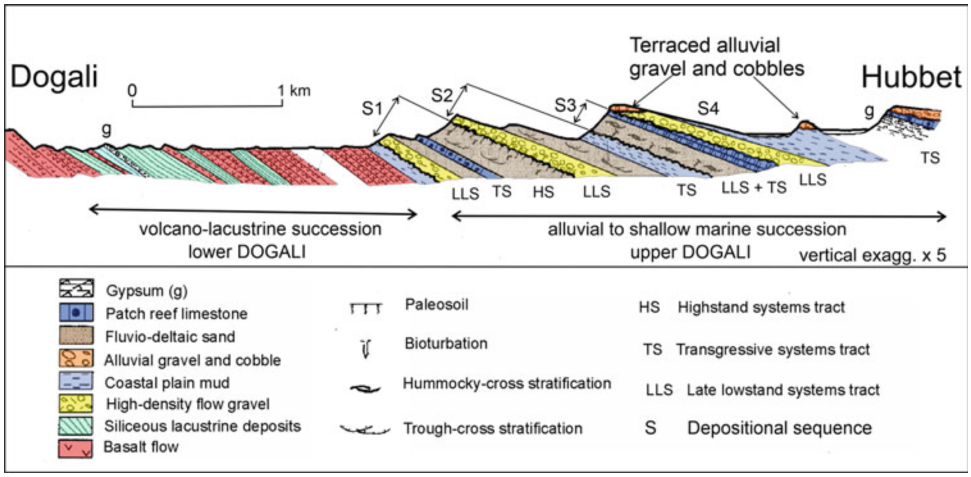

Fig. 2.29 Dogali succession restored in a SSW/NNE section between Dogali and Hibbet 15 km WNW of Asmara. In the upper—Dogali four depositional sequences (S1–S4). At the base of section—Paleocene to Eocene laterites with basalt clastics (Sagri et al. 1998)

The Late Miocene syn-rift rock units document an event that is well present in the Southern Red Sea drillings and have a notable effect on the whole Red Sea sedimentation: the restriction, up to stalling, of water circulation between the Red Sea and the Gulf of Aden. This was due to re-arrangements at the border of the Arabian/Nubian plates in the Bab El Mandeb and in northern Egypt (Bosworth and Burke 2005). This event prompted the deposition of the Amber Salt, three thousand meters of salt that become five thousand when involved in diapirism, as found in offshore drillings and geophysical prospecting.

A phase of intense rifting affected the plateau and the continental platform providing coarse clastic supply, also from the basement, to adjacent offshore and onshore basins (Desset Formation). In the Dogali area and along the whole Sahamar-Sahel, this formation is a prevailing sandy shallow marine unit with many pebbly inflows in the lower portion, sporadic coral reefs and evaporites and a few basalt flows (Bosworth et al. 2012). In the upper portion, it becomes deltaic and alluvial. According to Porro (1936), the Desset boulder beds have built a continuous hilly chain from the Sudanese border to the Gulf of Zula. The Desset Fm., more than 2000 m thick, is Late Miocene in age with a possible limited extension to the Early Pliocene (c. 5 Ma). In this case, the top of the Desset at the end of the syn-rift rock units was likely coeval with the first stages of the ocean spreading in the Southern Red Sea at 17° N lat. (Röser 1975).

The syn-rift Trap volcanism in central Eritrea can be matched in southern Eritrea with stages of the Afar plume activity represented by the Mabla Rhyolites and Dalha Basalts of Miocene to Pliocene age (CNR-CNRS 1971, 1975) widely outcropping in Djibouti but also present in the southernmost Eritrea near Edd. In the same area, there are outcrops of the Djioita granite (Varet, this volume) with an Ar/Ar age of 22 Ma. This body is correlative of similar granites of sialic crustal origin, such as the Affara-Dara, predominantly outcropping on the border of the old rifted Nubian continental crust and adjacent to new Afar units (Barberi et al. 1972).

Structural significance of the syn-rift successions during the Oligo-Miocene

The magmatic and sedimentary rock units and the structural setting of the syn-rift successions provide meaningful data for tracing and interpreting the paleogeographic events that marked the development of central and northern Eritrea during the Oligo-Miocene. As already outlined, between 29 and 20 Ma, this area, from the present-day continental platform to Sudan, was intermittently intruded and covered by the northernmost sprouts of the Afar plume, namely the Trap effusions (Ebinger and Casey 2001). In the open debate on what was the pre-Trap topography (e.g., Pik et al. 2003; Abbate et al. 2014; Coltorti et al. 2015; Sembroni et al. 2016), central and northern Eritrea is presumed to have been consisting of a wide low-elevation laterized peneplain covered by Trap basalts. After an early modest rifting event with basement coarse clastics emplaced before the Trap effusion, further sparse basalt basement coarse clastics supply is found along the Dogali succession. These clastics are high density flow gravels and alternate with much more common fine-grained lacustrine, fluvio-deltaic sediments. Two main coarse events are known: one toward the top of the Dogali Formation (c. 25 Ma) and a younger one, at the base of the Late Miocene Desset Formation (c. 10 Ma), deposited after fifteen million years without or with scarce coarse clastic sedimentation. We assume that the source area of the scarce clastic materials was the faulted eastern shoulder of a rising NNW/SSE elongated swell produced by the Afar plume impinging under the Eritrean crust (see later). Rifting, concomitant with the progressive uplift of the swell, produced a tectonic and erosional unroofing of the new highlands with sporadic clastic releases. The impingement of the Afar plume, the plate boundary forces producing the overall separation of western Arabia from Nubian Africa and, to a lesser degree, the crustal lateral extension induced by dike injections triggered and maintained these intermittent rifting processes with coarse sediments discharge.

Some datings are available from fission tracks analyses (Abbate et al. 2002; Ghebreab et al. 2002), which report a crustal cooling event with accelerated phase of denudation in the Early/Middle Miocene. Erosion resumed at the expenses of the Trap volcanites, which were covering the plateau. They were almost completely eroded and patches of laterites, probably lateritic ironstones or duricrusts such as those on the flat tops of the Roras (Dainelli 1943; Drury et al. 1994) were left above the basement (Fig. 2.30) (see also additional details in Sect. 2.3.4).

2.4.3 Post-rift Successions and Events

The Dahlak islands and a Danakil Homo

At the beginning of the Pliocene, when the Southern Red Sea began its active spreading offshore Eritrea, the Desset Fm. was overlain by the shallow marine sandy/shaly Dhunishub Fm. filling up as much as 2–3 km deep basins, adjacent to the Red Sea axial trough. Its littoral depositional environment promoted the growth of numerous carbonate platforms. The widest are the Eritrean Dahlak islands and the Saudi Farasan Islands (MacFadyen et al. 1930; Angelucci et al. 1981; Khalil 2012).

The Dahlak archipelago, with more than 200 islands (the largest island, Dahlak Khebir, is 760 km^2) and important trade center since the first centuries AD, sits on the Dahlak platform, a salient morphological modification in the offshore basins of central Eritrea (Fig. 2.31). Starting from 150 km north of Massawa (at c. 17° N), the continental platform widens abruptly up to 250 km and the conjugate shelf of the Saudi Farasan (the Roman garrison *Portus ferresanus*, II century AD) islands becomes similarly large. The interposed 5 Ma to present oceanic crust is c. 100 km wide.

The notable width of the two conjugate platforms was associated with a discontinuous spreading of the Southern Red Sea. According to Sultan et al. (1992) and Almalki et al. (2014), during the whole Miocene, there was a stalling in the ridge activity and the two platforms prograded and enlarged on the older oceanic crust until the spreading resumed during the Plio-Pleistocene. This still debated hypothesis is supported by magnetics and gravity data from oil exploration in the Farasan platform.

The vast shelf of the Dahlaks is less than 200 m deep and, in general, the islands rise 10–20 m asl. Only Dahlak Kebir attains a length exceeding 50 km. The islands are circular or semicircular and those with irregular edges also exhibit rounded bays and arcuate promontories and tongues. A peculiar feature is the domal structures ranging from a few meters to tens of meters, in places with the occurrence of diapiric salt. In the late thirties of the last century, deep oil drillings on the islands by AGIP encountered two thousand meters of Miocene salt after 100 m of Dahlak Reef Limestone and 200 m of clay, marl and sand (Carbone et al. 1998). The salt is correlated to the Amber Salt and the overlying unit to the Desset Formation. Diapirs rising from this impressive saline body generated the all-size domal structures in the Dahlak Reef Limestones (Fig. 2.32).

The substratum on which the present-day reefs grow is the Dahlak Reef Limestone of Pleistocene age (Montanaro Gallitelli 1973). In the nice map of Angelucci et al. (1981a, b), the Dahlak Reef Limestone is block faulted and gently folded. Yet, the strongest evidence of an ongoing active

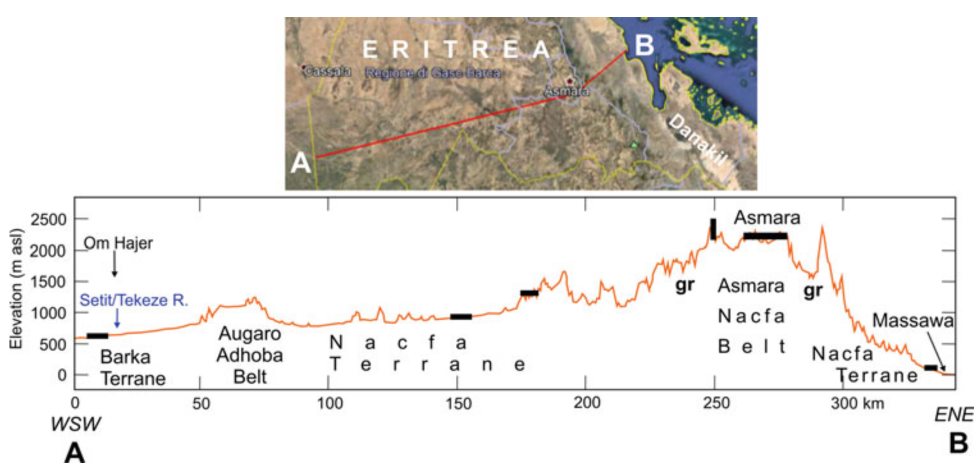

Fig. 2.30 This plateau cross section depicts the laterites associated with the pre-trappean surface recognized by Drury et al. (1994) in central Eritrea. From their original position, not far from the sea level, the laterites have been subjected to differential vertical movements described in text (Dainelli 1943; Drury et al. 1994). Terranes attribution based on Ghebreab and Talbot (2000) and Ghebreab et al. (2005)

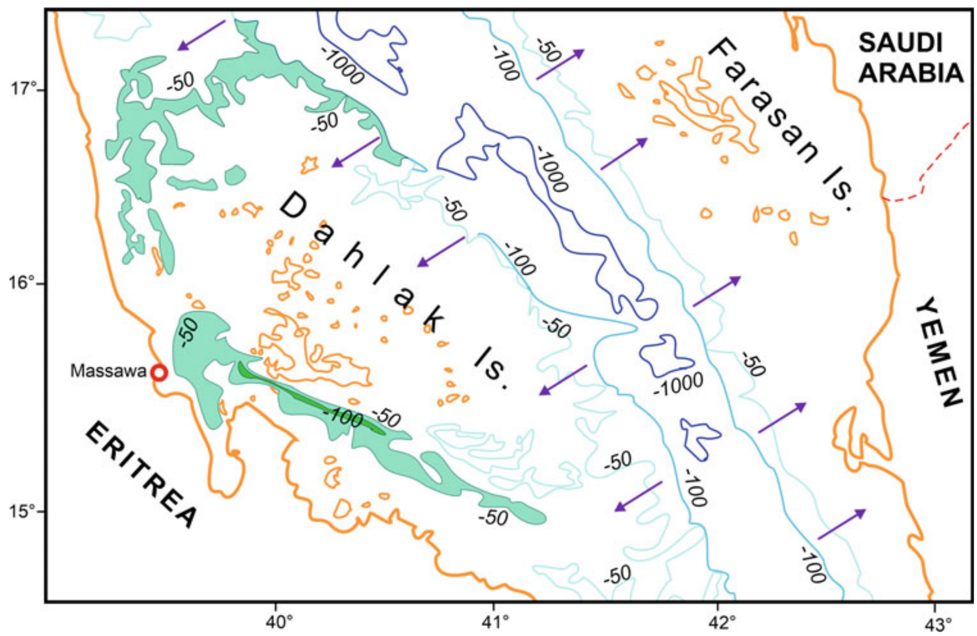

Fig. 2.31 The Dalhak and the Farasan Islands facing each other on conjugate margins separated by an active spreading axis since the Pliocene. The larger number of islets in both archipelagos is accounted for by a thinner (less than 100 m) present-day carbonate platform that can be easily pierced or, more generally, deformed by the upward rising salt domes originated from the extremely thick (more than 2000 m) evaporite level beneath the platform. The 50–100 deep narrow trough Massawa Channel (green in the map) separates the Dhalak islands from the coast (Massawa and Buri Peninsula). It does not find a similar feature in the Farasan Islands (modified after GPS Nautical Maps I-Boating Free Marine Navigation, 2022). The light blue lines are the −50 m isobaths; the blue lines are the −100 m isobaths; the dark blue lines are the −1000 m isobaths. The purple arrows indicate the Red Sea floor spreading

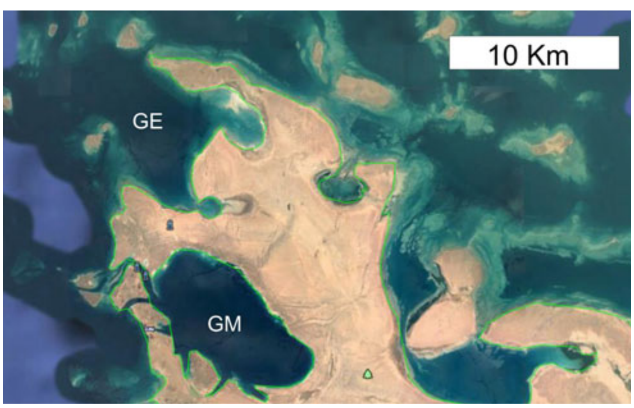

Fig. 2.32 Ghubbet Entatu (GE) and Ghubbet Mus Mefit (GM) are two bights in the NW portion of Dahlak Khebir, the largest island of the Dahlak archipelago. They are hollows as deep as ca. 50 m and with the longest axis of about 20 km. Ghubbet Mus Mefit is connected to the open sea through a tide channel. The ovoidal (as in Ghubbet Mus Mefit) or circular shapes of the hollows are common in many islands and islets of the archipelago. They are deep holes left by the leaching of the salt domes. See also Fig. 2.26

tectonics is the Massawa Channel, a NW/SE elongated rift which is 250 km long, 20 km wide, no more than 200 m deep. The channel starts from Massawa and, parallel to coast, reaches Tiho at a more or less constant distance of some tens of kilometers from the coast (Frazier 1970). Its occurrence can be related to the stretching of the Nubian plate margin in the process of separation of Arabia from Africa, and it can be seen as the first stage of a marginal basin (compare with eastern Ethiopian plateau margin adjacent to Afar, Zwaan et al. 2020). A rifting activity since the Late Oligocene is reported in the Southern Red Sea from oil prospecting by Hughes et al. (1991).

Close to Massawa, the regional NNW structural trend, marked by the alignment of the coastal belt, splits into two directions. One continues with almost the same NNW strike through the Gulf of Zula corridor and penetrates the hinterland toward the Afar axial range. The other points to SE, keeps parallel to the anticlockwise rotated Danakil block and goes on southward as far as Assab.

South of the Gulf of Zula, the NNW trend runs at the foot of the escarpment through small, clastic and post-rift basins. One of these is the Buia basin (Abbate et al. 2004; Ghinassi et al. 2015) (see also Chaps. 6 and 7 of this book for details) at the foot of the Alid volcano along the corridor Zula/Samoti/Badda that was a possible Pleistocene seaway connection between the Red Sea and the Danakil depression (Fig. 2.33). The Buia basin hosts sediments of alluvial plain (Fig. 2.34), ponds, playas and lacustrine delta environment. In a deltaic system, a *Homo erectus/ergaster* skull (Abbate et al. 1998; Macchiarelli et al. 1998), dated one million years

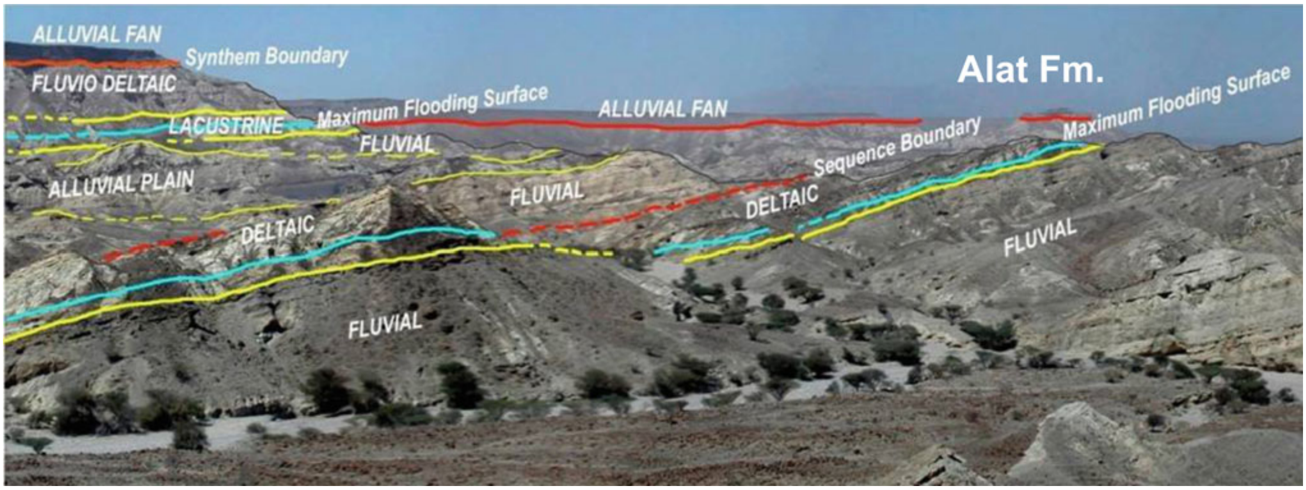

Fig. 2.33 Post-rift Neogene sediments at the apex of the Afar depression 20 km south of the Gulf of Zula. In the lowlands, at the foot of the highest rim of the Eritrean plateau between Senafe and Adi Keih, heavy sediment fluxes promote the birth of many alluvial basins, such as that of the Dandero river (in the foreground). The Dandero Group, a 500-m-thick sedimentary pile at the foot of the Mt. Aro, includes fluvial, alluvial plain, fluviodeltaic, deltaic and lacustrine deposits, irregularly repeated and intersected through time (Abbate et al. 1998). In a lacustrine to deltaic, one-million-year-old, unit (the Alat Formation, see notation on the right of the figure) an *erectus/ergaster Homo* cranium has been found (Macchiarelli et al. 1998)

Fig. 2.34 Middle Pleistocene deposits of the Dandero river filling a valley cut in the Proterozoic Nakfa green schists (20 km south of the Alid volcano)

through paleomagnetic (Albianelli and Napoleone 2004) and fission track analyses (Bigazzi et al. 2004), was found. The Buia *Homo* fills an important gap between early African *Homo erectus* (2.2–1.4 Ma) and the later African *Homo heidelbergensis* (0.65 Ma) (Fig. 2.35).

Fossils of big mammals and reptiles (Martínez-Navarro et al. 2004) and a rich dough of Acheulean stone tools (Martini et al. 2004) were also found together with the human remains (Fig. 2.35) South of Zula there are some northernmost sparse outcrops of the Red Series, a typical

Fig. 2.35 *Erectus/ergaster Homo* cranium found in the Alat Fm. (Dandero Group) together with abundant Acheulean stone tools and a rich large mammal fauna

marker of the Afar realm (Barberi et al. 1970) with red conglomerates, sand and mudstones, sometimes gypsiferous and rare freshwater lumachella limestones, basalt flows and acidic tuff. With an estimated thickness of one thousand meters, the Red Series was deposited by alluvial fans, high-energy streams and in small ephemeral lakes. The radiometric dating of the basalts spans from the Early Miocene to the Pliocene (Bannert et al. 1970) sharing both syn- and post-rift tectonic regimes.

The second trend (NW/SE) is apparent on the coastal belt and its adjacent hinterland stretching from the Buri peninsula, through the elongated Danakil block, to the volcanic bodies of the Assab region. The post-rift volcanic activity is witnessed by stratoid basaltic lavas and ignimbritic sheets. They are representatives of the post-rift Afar Stratoid Series (Plio-Pleistocene, Barberi and Santacroce 1980) and by recent volcanic centers. The latter include, from north to south, the Jalua (Mesfin and Yohannes 2014), Alid (rhyolite dome with basaltic lava fields, active 1817; Marini 1938, 1998; Clynne et al. 1996), Dubbi, Nabro (eruption June 2011; Hamiel and Baer 2016) and Mallahle stratovolcanoes.

According to Ghebreab and Talbot (2000), a c. 60-km-long and E–W trending lineament would connect the Alid volcano with the volcanic plugs of the Senafe region.

Moreover, a spectacular range of numerous basaltic cinder cones and lava flows makes up the Assab volcanic field. It covers an area of c. 5000 km^2 and its lava flows reach the Red Sea.

The Neogene syn- and post-structural events: research history and a new proposal

The times of drift events had a significant role in the Neogene structural evolution of the Eritrean continental margin. The stretching during the separation of Arabia from Nubian caused progressive rifting in the hinterland and new structural discontinuities followed the c. NNW trend of the Precambrian basement, mainly determined by terrane boundaries. It is worth noticing that this direction parallels the Red Sea ridge axis and many dikes swarms on the escarpment.

Many suggestions have been put forward on the main processes originating the scarp and causing its dismantling or modification since the Miocene.

Abul-Haggag (1961) recognized that the eastern margin of the Eritrean plateau does not represent “an original fault-face” but just a retreating scarp and that the scarp still preserves a profile with successive benches due to step faulting.

According to Drury et al. (1994), a possible main lineament is the major normal fault or extensional system which from south of Asmara heads northward retracing Precambrian lineations. From the western border fault of the Damas graben, it proceeds to NNW at the foot of the northern plateau in the Samar-Sahel region. An eastward dipping detachment from the Damas western border is supposed to have activated a series of ramps and horst across the continental margin.

Talbot and Ghebreab (1997) and Ghebreab and Talbot (2000) assume a low-angle detachment across the continental margin with associated top-to-basin domino faults starting from the western border of the Damas graben. The lithosphere was extended and thinned, and core complexes of high-grade crustal rocks were exposed to surface. The structure evolved into a new deeper detachment producing an east-directed ramp system and a faulted monoclinal flexure considered as the Red Sea escarpment in Eritrea. Although some authors (e.g., Beyth et al. 2003) claim that the low-angle detachments and associated structures are the products of Neoproterozoic collapse events and not Tertiary deformations related to the opening of the Red Sea, Talbot and Ghebreab (1997) and Ghebreab and Talbot (2000) reconstructions are the most commonly accepted.

The discontinuous row of marginal basins from the Damas graben in central Eritrea to the Robit graben in central Ethiopia (Abbate et al. 2014; Zwaan et al. 2020) has been framed in the Southern Red Sea rifting by Tesfaye and Ghebreab (2013). They consider the western margin of these marginal basins as the breakaway zone of a ramp-flat detachment in which the Danakil horst was the flat.

The denudation, whether tectonical or erosional, persists to be a key question in the interpretation of the Eritrean high-elevation passive margin. In addition to classical regional geology investigations, this issue has been approached with apatite fission track (AFT) dating and (U/Th)He thermochronometry.

Abbate et al. (2002) and Balestrieri et al. (2005) sampled the Eritrean margin along four transects coast-perpendicular and latitudinally distributed from Keren to Buia (south of Massawa) for AFT and (U/Th)He analyses. These authors found a positive correlation between FT ages and sample elevation indicative of an erosive denudation of the margin starting from a main border lineament. They also assumed that during the Miocene, a remarkable uplift affected the margin, so that by Middle Miocene the material removed by erosion was c. a couple of kilometers thick. Moreover, the (U/Th)He ages of the transects typify a denudation of the escarpment predominantly evolved by downwearing with a main phase of post-break up erosion starting at about 15 Ma from a regional NNW-trending border lineament parallel to the coast and stretched from the Labka river (50 km NE of Keren) through Damas and Ghedem to Buia in the northernmost Danakil depression.

Ghebreab et al. (2002) reported that the ages of four samples collected in the Asmara-Massawa area were not positively correlated with sample elevation. This result would suggest a process of tectonic denudation through low-angle detachments rather than erosion (Ghebreab et al. 2002). Three AFT profiles carried out across the Ghedem Mt. a Proterozoic feature 20 km south of Massawa lead to a similar conclusion.

Drury et al. (2005, Comment) objected that Balestrieri et al. (2005) and Abbate et al. (2002) assumptions in favor of an erosional denudation of the Eritrean margin were founded on a correlation between FT ages and sample elevation erroneously considered positive. From this incorrect attribution derive further disputable aspects connected with the margin degradation, such as the position of the border faults constrained by the occurrence of the youngest FT ages, the FT age and interpretation of some inliers (Ghedem) and the time of the onset of post-break up erosion.

From the previous short review, it is evident that we are confronted with different conceptual models for the Eritrean margin, and we think that a crucial area subject to various interpretations is from Asmara to the sea. For this margin, a stratigraphic and structural detailed mapping of the Proterozoic units of the escarpment and of the Tertiary to Pleistocene units of the lowland would consistently help in understanding the Eritrean margin formation and evolution. Important issues to be carried out or further investigated are as follows: (1) a denser re-sampling for FT and (U/Th)He dating after discordant interpretations; (2) new datings of the mylonites related to low-angle detachments (Talbot and Ghebreab 1997; Ghebreab et al. 2005); (3) detection of possible breakaway structures or border fault; (4) the conceivable occurrence of post-Proterozoic cover shreds on the escarpment mirroring Drury et al. (1994) Oligocene datum on the western slope of the plateau.

Trying to clarify this uncertain picture, we propose a section (Fig. 2.36) from Asmara to the sea drawing particular attention to the relationships between the post-Proterozoic cover on the plateau (Mendefera section) and on the lowland (Dogali section). Although presently 50 km apart and separated by a difference of two thousand meters in elevation, we assume that during the Oligocene these two low-elevated successions were contiguous above the basement units. Their correlability is based on common features, such as the same stratigraphic succession with coeval Trap basalts and intertrappean sediments, the same fossiliferous content (*Deinotherium* and *Gomphotherium*) and common Late Oligocene radiometric dating of the volcanites. The local occurrence of some basement clastics below the basalt distinguishes the lowland section and indicates the inception of some rifting action. Traces of this event are missing in all the numerous exposures of the plateau outcrops.

In our reconstruction, the plume-generated uplifted region in its progressive rising got faulted along a main, steeply east-dipping extensional feature running NNW/SSE, located halfway between Asmara and Massawa, a few kilometers east of the Damas region. The probably upflexured footwall constituted the edge of the Eritrean plateau with its more than two thousand meters. On the hanging wall to the east of the fault, the lowland units were only scarcely affected by the upheaval or by the main fault, while the diffusely flexure and faulted footwall rocks, crowned by a retrogressing rim, were undergoing uplift and slope erosion becoming a morphologically rough escarpment.

As shown in Fig. 2.36, we assume that the new fault was located and developed along the Neoproterozoic “transition zone”, a lithologically heterogeneous, intensely sheared, up to 6–7 km wide, NNW/SSE trending tectonic zone marking

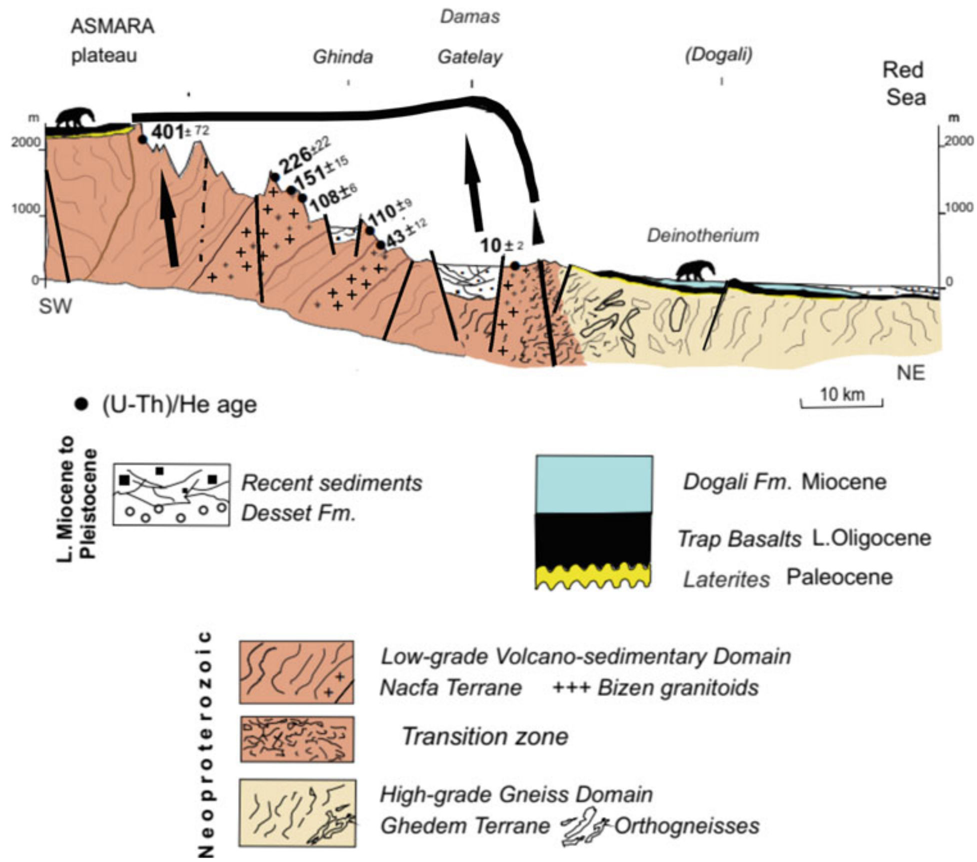

Fig. 2.36 General geological section across the Eritrean margin between Asmara and the Red Sea (20 km north of Massawa). Black dots: apatite fission track ages as in Abbate et al. (2002). The Afar plume impinging on the low-elevation Central Eritrea continental margin of the Nubian plate resulted in a major fault that disjointed by almost two thousand meters' difference in height the Asmara region from the Red Sea lowland. The footwall becomes the dissected margin of the Asmara plateau. Modified after Drury et al. (1994), Ghebreab (1998), Ghebreab and Talbot (2000), Ghebreab et al. (2005, 2007), Andersson et al. (2006)

the Neoproterozoic juxtaposition of the younger Nakfa terrane against the older Ghedem terrane (Ghebreab and Talbot 2000; Andersson et al. 2006; Ghebreab et al. 2007). This fault can be traced for more than one hundred kilometers from the Damas region to south of the Gulf of Zula. On its Neoproterozoic prominent low-angle deformations, younger steep normal faults were superimposed during the Tertiary extension related to the Red Sea rifting (Drury et al. 1994; Talbot and Ghebreab 1997).

The onset of the footwall exhumation, possibly eased by an isostatic response to extension, can be roughly time constrained by the age of the youngest common sedimentary, biostratigraphic or volcanic record on the plateau and on the lowland and/or by the thermal history of the footwall rocks. Based on the first approaches, the onset of the exhumation should postdate the Late Oligocene/Early Miocene age of the volcanites and of a common low-land environment suggested by stratigraphic considerations (see before). Moreover, in the Bizen granitoids footwall, close to the transition zone and at 300 m asl, the thermal history records a FT age of 10 ± 2 Ma and He ages of 5.5 and 9.2. A geothermal gradient of 25 °C/km and mean annual surface temperature of 25 °C requires 1.5–2.0 km of erosion removal since rifting began 15 Ma ago (Balestrieri et al. 2005), and this could approximate the difference in elevation between the plateau rim and the lowland units.

2.5 The Physiographic Evolution of the Eritrean Plateau and Its Denudation Rate

Physiographic variations through five profiles across central and northern Eritrea and northern Ethiopia

As anticipated in Sect. 2.4.2, at about the Late Oligocene the impinging Afar plume started to affect central and northern Eritrea that were a vast, low-elevated area since at least 50 Ma. Concurrently, the Eritrea continental margin was progressively rifted and pulled apart toward south Arabia.

The interaction between the plume-driven uplift and denudation starting from and almost even surface poses interesting questions about their effects on landscape morphology and dynamics.

Some hints are provided by a morphological appraisal of the present-day features of the Eritrean plateau through five WE topographical profiles with vertical exaggeration of c. 1:12 distributed along 300 km from the northernmost Ethiopia close to the Eritrea/Ethiopia border to the northern Eritrea close to the Eritrea/Sudan border (Fig. 2.37).

Section 1 runs all in the northernmost Ethiopia starting where the Setit (Tekeze) river enters Eritrea and terminates near the western border of the northern Afar depression. We

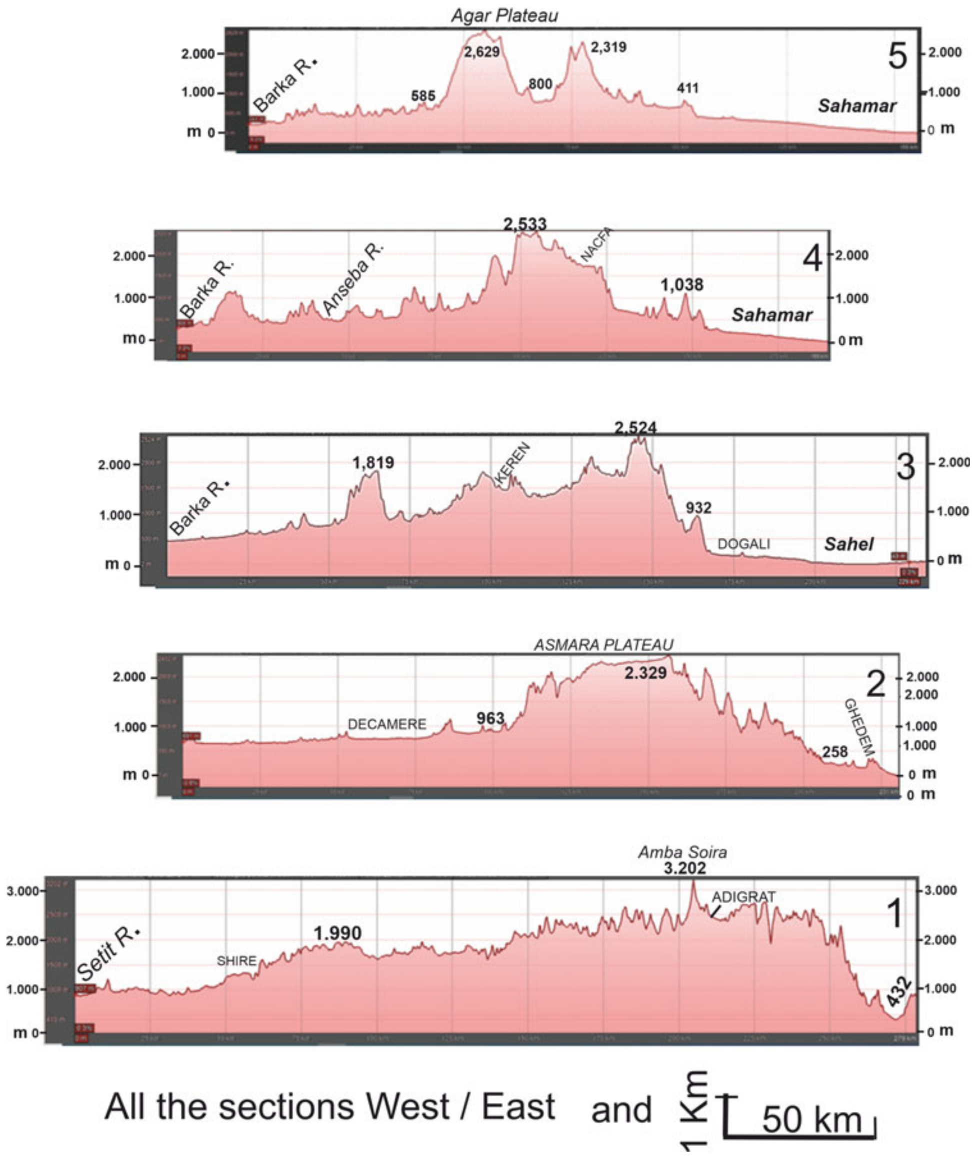

Fig. 2.37 Starting from Section 1 close to the Eritrea/Ethiopia border and heading north to Sudan the five sections record different physiographic features at growing latitudes along the Eritrean plateau showing similarities and differences of the crossed areas due to distinct responses to tectonic and magmatic events and erosive action

assume it keeps in some way the original physiographic conditions that appear more or less modified in Sections 2–5 across the Eritrean plateau. From the Adigrat-Axum region to the Shire region Section 1 shows a gentle westward slope across a WSW/ENE ridge transversal to the margin of the Eritrean/Ethiopian plateau with basement rocks covered by sub-horizontal Paleozoic sediments, large Paleogene Trap outcrops above laterites and, locally, more recent trachytes (see Garland 1980, section A–B). Large areas in the eastern half of the section have heights exceeding 2500 m, and north of the section the Amba Soira with its 3018 m is the highest elevation in Eritrea. It should be taken into account that without the large vertical scale exaggeration, the real inclination of the entire western flank of the swell (i.e., from Amba Soira to Shire) turns out less than one degree, and it is likely to envisage similar topographical conditions also for pre-erosion times with the diffuse, synchronous actions of plume-generated uplift and erosive processes.

As regards the altitude of this peneplain, we allow for a laterite level covered by a few hundred meters of Traps (see the Mendefera area) and obtain an original flat surface with a west/east extension of 250–300 km raising progressively for at least 2000 m while gradually eroded. The age of this plume-promoted upheaval is poorly constrained. A Late Oligocene/Early Miocene initiation was indicated in the previous paragraphs and the characteristics of the modern drainage system, in which several examples of ongoing river captures are present (see Chap. 4) would imply a persisting upheaval.

At contrast with most of the section, an abrupt descent is found at the extreme east of Section 1. It is marked by the Neogene Garsat graben, one of the northernmost marginal basins at the foot of the Eritrean/Ethiopian plateau (Zwaan et al. 2020).

Sections 2–5 are intended to show possible stages of the geomorphological evolution starting from a profile similar to that of Section 1.

The eastern half of Section 2 dominated by more or less eroded Trap basalts and post-tectonic granitoids reminds the corresponding portion in Section 1. In the western portion near Barentu, the essentially flat-lying surface with Trap basalts and laterites (see Fig. 2.37) can be correlated with similar occurrences at the top of the Asmara plateau.

Sections 2 and 3 provide a glimpse of some peculiarities relative to the coastal belt. Between these sections, almost in correspondence of an ideal Massawa-Keren line, there is a marked change in the coastal belt physiography. North of Massawa the Red Sea coast opens into the 30–40-km-wide and 300-km-long Sahamar-Sahel coastal plain. This plain maintains a constant width and a rectilinear seaward profile proceeding northward as far as the Tokar delta in Sudan. It is predominantly covered by Quaternary deposits, but outcrops of older units, those dated for the Proterozoic included, are also present. At the foot of the highlands, the plain is bound by an extended NS trending normal fault system connected to the rifting of the continental margin (Drury et al. 1994).

South of the Massawa-Keren line, for about 200 km toward SE as far as the Anfile Bay, the coastal lowland becomes narrow and indented with bays and cliffs. This change in the coastal physiography is matched inland by the reduction of width of the plateau, which rapidly decreases from the ca. 150–200 km south of Asmara to less than 50 km northward. Concomitantly, the N–S trending main escarpment makes an eastward shift (see Fig. 2.2). To the west, a distinctive morphological feature, trending approximately W–E, is the upper course of the Barka river which, after a long N–S path from the Eritrean border, turns eastward toward Asmara. Its active headwater erosion and the occurrence in the same region of WNW–ESE trending Tertiary dilatation fractures (Drury et al. 1994) resulted in the opening of a wide alluvial valley floor at the expense of the adjacent highland, which also reaches the foot of the Asmara plateau.

The occurrence of these west–east tectonic lines cutting across the central Eritrean plateau and the northern Ethiopian plateau can be matched with similar volcano-tectonic lineaments more to the south, such as the Tullu Wellel (Nekempti)-Yerer (Addis Ababa) and Bonga-Goba (Abbate and Sagri 1980; Mohr 1987; Wolde 1989; Abebe et al. 1998) interpreted to result from Cenozoic reactivation of pre-existing Neoproterozoic structures.

Sections 3, 4 and 5 cut across an area traditionally known as the Eritrean Plateau, but rightly named “highland” by Dainelli (1943) because of its rugged topography with reduced plateau remnants. Unlike Sections 4 and 5, Section 3 still shows a rough declining westward slope.

Section 4 intersects the valleys of the Anseba and Barka rivers, two main rivers of Eritrea flowing to the north, i.e., perpendicularly to the cross section. Their erosive action at the expenses of the western slope of the Rora range is evident. A faulted step, hosting the town of Nakfa, separates the highest (2523 m asl) central peak from the seaward basal hills.

Sections 3 and 4 have rather similar profiles with their eastern flank markedly down faulted.

Section 5 exhibits a topographic profile rather different from the previous ones. The uplifted portion has been severely eroded, and only two prominent, 2500 m high, mountains made of Neoproterozoic massive rhyolites of the Agar terrane survived substantial erosion.

Moreover, the highest elevations in the five sections are found with repeated values around 2500–3000 m generally close to the topographic step toward the lowlands. This suggests that the eastern border of a general, unique surface has been raised with the same intensity along a distance of at least 500 km from central to northern Eritrea. Moreover, if the size of the swell in the profiles 1–5 is compared, it is evident that, on the whole, it gets less prominent northward. Was the plume losing its capacity of building highlands in this direction? Yet, Zeyen et al. (1997) traced the path of the Afar plume across Sudan and as far as southern Egypt through heat flow analyses and pointed out that, in Eritrea, the volcanic activity connected in some way to the plume had already substantially diminished before entering Sudan. This suggests that northern Eritrea was less and less affected by plume activity without neglecting a northward, more effective denudation.

Denudational rates of elevated regions

It is possible to infer long-term denudational rates of elevated regions such as the Eritrean plateau. If one assumes the existence of a post-Proterozoic cover no more than 500 m thick (now completely eroded) on the plateau and erosion processes starting 15 Ma ago (Balestrieri et al. 2005), a denudation rate (DR) of 33 mm/ka is obtained, i.e., one-third of what was found by apatite/He data (Balestrieri et al. 2005) in the margin south of Asmara. Even lower DR (around 15 mm) is reported by McDougall et al. (1975) for the Blue Nile and Tekeze river basins since the Miocene. Pik et al. (2003) used the same thermochronometry for the Tekeze and upper Blue Nile basins and obtained DR values of 30–35 and 29–34 mm/ka, respectively. The average of all these data is 32 mm/ka, which is almost coincident with DR of 33 mm/ka calculated in this study for Eritrea. This slow denudation rate in a tectonic active region as the Eritrean plateau was perceived by Dainelli (1943) and Abul-Haggag (1961) and the latter author referred to it as "the slow tempo of erosion". The $^{40}Ar/^{39}Ar$ dating of low-Ti tholeiitic and high-Ti alkaline flood basalts in Eritrea of Teklay et al. (2003) suggests that erosion started about 20 Ma, leading to a lower value for the DR. On the base of these data, the DR of the Eritrean plateau can be approximate at 30 mm/ka. Further studies and dating are however necessary to fix this value with greater accuracy.

In the literature, denudation rate (DR) varies widely, commonly across two–three orders of magnitude. Such a variability depends on a number of factors including the method used to calculate the DR (Bishop 1985); the age and the kind of rocks considered; the tectonic history; the prevailing dislocation mechanism and climate. In spite of the many studies, there are still considerable uncertainties about the dominant factors controlling the DR.

Notwithstanding the impressive uplifting, the DR calculated for Eritrea is relatively low if compared with the mean estimate of 30 mm/ka obtained by fission tracks during the Cenozoic in regions of subdued relief, but it is not far from the estimated present-day global mean of 43 mm/ka (Summerfield 1991). Moreover, the DR of the Eritrean is an average value as it was likely lower at the beginning of the uplifting and higher when it reached its maximum elevation.

Ruxton and McDougall (1967) and Ahnert (1970) have shown that DR tends to increase with elevation though their values were obtained from modern erosion data and are one–two orders of magnitude higher than in Eritrea. Local relief (i.e., the difference between the maximum and minimum elevation in an area) is considered to be even more important than elevation in propelling the DR as found by Ahnert (1970) for 20 middle latitude basins. The current physiography of the Eritrea highlands is characterized by a large relief ratio (Fig. 2.30), and the DR of the recent past is expected to be much higher than the average of 33 mm/ka obtained across a time span of 15 Ma, especially in the river systems draining the escarpment as suggested by their steeper channels.

For river basin located in tectonically active areas, Summerfield (1991) reported present-day DRs higher than long-term average of in Eritrea, e.g., Indus 124 mm/ka, Ganges 271 mm/ka, Brahmaputra 677 mm/ka, Colorado 84 mm/ka and Nile 15 mm/ka. This latter value is very low because the sediment contribution of the White Nile (which is the largest upstream basin) is close to zero, whereas the most of the sediment supply to the Main Nile comes from the Blue Nile and the Atbara, whose catchments are much smaller (they account for about 18% of the upstream Nile basin) and whose sediment comes from the Ethiopian highlands (Sutcliffe and Parks 1999; Billi and Badri 2010). On the base of these considerations, we can therefore calculate the current DR of the northern Ethiopia highlands as 82 mm/ka, i.e., a DR value close to that measured for the Colorado or the Indus. Unfortunately, there is no sediment transport data for the Eritrean rivers, but it is reasonable to assume for the Eritrean highlands a current DR similar to that of the Ethiopian highlands. For the passive margin of south-east Australia, Persano et al. (2002) calculated an average DR of 45 mm/ka across the last 90–100 Ma. The south-eastern Australian margin is a typical high elevated passive margins characterized by a steep escarpment that separates the coastal plain from a low relief inland plateau. Such a physiographic setting and the structural events that generated it are similar to those of eastern Eritrea (Figs. 2.30 and 2.36), so it is not surprising that the active average DR of these areas is very similar.

Average DRs of the same order of magnitude of those calculated for Eritrea are reported by Zhang et al. (2003) for the crystalline rock region of the Huangling (DR = 17 mm/ka). This brief review of DR data reported in the literature for comparable settings indicates that the average DR calculated for Eritrea is not particularly low though phases with different intensity of erosion may have occurred in the past 15 Ma in relation to the variability of uplift rate and climate conditions. Further studies are, however, necessary to verify the control of denudation models on denudation rates.

In a tectonic active area, the river network development plays an important role in landmass denudation and, more specifically, in the export of the sediment delivered by erosion processes on slopes. Bankfull flow is a significant geomorphological factor since it maximizes the sediment transport of a river and, in the long term, transports the greatest cumulative volume of sediment (Wolman and Miller 1960). Bankfull flow is, therefore, the geomorphologically most active discharge contributing to landmass denudation and to the river network development.

The stress necessary to move the bed particles during bankfull floods is approximated by the bankfull Shields stress ($\tau*_{bf}$), defined by Pfeiffer et al. (2017) as follows:

$$\tau*_{bf} = \rho R_{bf} S / (\rho_s - \rho) D_{50} \tag{2.1}$$

in which ρ is the density of water, ρ_s is the density of sediment, R_{bf} is the hydraulic radius at bankfull stage, S is the channel slope, and D_{50} is the median size of bed material.

According to Sklar and Dietrich (2008), in channels of tectonically active areas, the stress needed to transport bed material is typically much higher than the stress to incise the rock. This is because in tectonically active areas, the sediment supply is generally high and rivers tend to maintain bankfull stresses much higher than the critical flow (i.e., the flow stress required to initiate the entrainment of the bed particles) (Pfeiffer et al. 2017). Under these conditions, the attainment of an equilibrium profile is delayed as the most of the river flow energy is dissipated to move the large quantities of sediment entering the channel and produced by active erosion processes.

Even a rapid glance at the Eritrean rivers reveals that those draining the escarpment are smaller, steeper in their headwaters, choked with sediment and probably have not yet reached an equilibrium profile (see also Chap. 4), whereas those draining the gently westward inclined plateau are much larger, have lower gradients (especially downstream of their headwaters) and their channels are not chocked with sediment as much as their eastern counterparts. Equation (2.1) indicates that channel slope and sediment size are important factors in controlling the bankfull stress. In the uplifting of Eritrea, associated with the Afro-Arabian doming, on the westward sloping monocline of the plateau, gentler channel gradients are expected than in the escarpment channels. On the plateau, the river systems developed on low gradients and the reduced sediment supply may have allowed them to develop an equilibrium profile. This geomorphological setting may provide a complementary explanation of the low denudation rates obtained with different methods reported in this chapter. On the escarpment, characterized by a faulted stepped morphology, the river headwater slopes were steeper since the beginning of the dome faulting and the escarpment formation. Here, the steep embryo channels started to develop their catchments and, given their steeper gradient, used a large proportion of their flow stress to remove the sediment supplied from slopes. The continuous downstream conveyance of the supplied material favored the backward migration of the headwaters and contributed to the shaping of the very steep escarpment as we know it today. This interpretation is confirmed by Pfeiffer et al. (2017) who found a decreasing trend of sediment transport capacity and erosion rates moving from tectonically active areas of North America west coast to the eastern, more stable inland.

2.6 Conclusions

The present, large-scale landscape of Eritrea is the result of a long and complex sequence of tectonic events interrupted by long periods of quiescence during which erosion prevailed with the formation of peneplains covered by lateritic duricrusts. These events are recorded in the uplifted basement which makes the backbone of northern and central Eritrea. The following Jurassic sea transgression that covered the largest part of the Horn of Africa and gave way to thick continental and marine sedimentary sequences is recorded only in small areas of southern Eritrea which mark the northernmost inland ingression of the Jurassic sea. After a long time of relative quiescence, the sea regression progressively affected the whole Horn of Africa and led to the onset of the Afro-Arabian doming and rifting, which are the most important tectonic events that caused the remarkable uplifting of the Eritrean plateau and the formation of the Red Sea. These large-scale tectonic events were preeminent in providing Eritrea with its present physiography characterized by a gently westward inclined plateau and a sharp, 2000 m high escarpment along the Red Sea coastline.

The sequence of paleoenviroments and events for the northern and central Eritrea since the Neoproterozoic can be outlined as follows:

1. Rifting apart of the Neoproterozoic supercontinent Rodinia about one billion years ago and opening of the Mozambique ocean.
2. Accretion of the new margins (East and West Gondwana) through ocean spreading, crustal growth, island arc- and plume-related magmatic activities.
3. Various phases of continental collisions from c. 650 Ma to c. 550 Ma ago and formation of the East African Orogen within the Gondwana supercontinent.
4. Two tectonically superimposed terranes in the Eritrean East African Orogen: the lower one with amphibolite facies metamorphic rocks, in the upper one greenschist facies metamorphic rocks.
5. Deformation and/or exhumation through low-angle detachments, tectonic escape, suturing, orogenic collapse and unroofing.
6. Denudation of mountain belts and rift basins filled by continental and/or shallow marine sediments with traces of glacial activity (Snow Ball Earth).
7. c. 550 Ma Afro-Arabian Peneplain over the Gondwana Assembly; relative quiescence with very scarce clastic sedimentation occasionally glacigenic till c. 260 Ma.
8. Diffuse intraplate rifting of Gondwana with alluvial, fluvial and deltaic sediments followed at c. 250 Ma by the sea ingression from the Indian Ocean with marine sedimentation lasting 15 Ma.
9. 15 Ma of marine deposition ensued by continental clastic deposition until the end of the Cretaceous.
10. Intense erosion down to the basement marked by the African Surface with notable laterite pedogenesis particularly evident in central Eritrea.
11. Wide areal extension with low relief and low elevation inherited from the Late Cretaceous.
12. Afar plume volcanism (Oligocene Traps) locally overlying basement clastics derived from precursory rifting.
13. Since 20–15 Ma, a regional fault zone linked to enhanced Afar plume impingements generates, from west to east, a wide Eritrean plateau subject to progressive uplift, a diffusely faulted Eritrean escarpment and a coastal lowland.

14. Mio-Pliocene episodes of intense rifting and conglomerate production (Desset Fm.).
15. Late Miocene water circulation restraint up to stalling between the Red Sea and the Gulf of Aden.
16. Extensional phases connected to continuous stretching of the Nubian margin and active spreading since the Pliocene in the Southern Red Sea.
17. Continuous upheaval and slow erosion (c. 30 mm/ka) of the whole plateau and escarpment as confirmed by the occurrence of several ongoing river captures, especially in the easternmost part of the escarpment.

The contrasting physiography and bedrock lithological characteristics produced a variety of modern landscape and landforms. Both old (Proterozoic) and more recent faults exert a substantial control on the river network of Eritrea, whereas the lateritic duricrust acts like hard discontinuity contrasting the landscape incision of the plateau.

The denudation rates calculated with different methods revolve around the value of 30 mm/ka, which may seem a relatively low value for a region that experienced such an impressive uplifting. This value, however, is comparable to that estimated for the present-day global mean (43 mm/ka). The denudation rate of Eritrea is an average value calculated across a long time interval during which river erosion was likely lower at the beginning of the uplifting, when the regional slope was gentle, and progressively increased as the plateau reached its maximum elevation.

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