1. Introduction

Volcaniclastic systems dominate arc-related basins from trench through forearc to intra-arc and back-arc settings, but accompany active volcanoes only in intra-arc basins (Smith and Landis, 1995). Supply of sediment is controlled mainly by arc magmatism with eruption frequency and style considered the most significant factors controlling depositional systems and architecture. The most notable feature is voluminous supply of detritus from multiple, short-lived sources. Products include lava and pyroclastic flows, ash fall-out and hydroclastic debris, whereas compositions range from basaltic to silicic (Fackler-Adams and Busby, 1998). Differential subsidence and syndepositional faulting add to the complexity of basin filling, whereas the depositional architecture of most modern systems is poorly known because they are deep marine (Macdonald, 1993).

The complexity of volcaniclastic systems impacts on the approach with which their rock records are studied. Stratigraphic problems are compounded by the intense magmatism, thermal metamorphism and deformation inherent to arc environments (Ingersoll and Busby, 1995), problems exaggerated in ancient sequences by preservation in lithotectonic complexes. Indeed, the term arc-related is preferred for ancient sequences, although close spatial and temporal relationships between sedimentary, volcanic and plutonic rocks define an intra-arc setting (Ingersoll and Busby, 1995; Smith and Landis, 1995). The combination of deformation and chaotic stratigraphy precludes traditional litho-stratigraphic analysis. Rather, emphasis is placed on lithofacies analysis (Fackler-Adams and Busby, 1998), whereas sequence-stratigraphic

Stratigraphy, facies architecture, and palaeoenvironment of Neoproterozoic volcanics and volcaniclastic deposits in Fatira area, Central Eastern Desert, Egypt

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ABSTRACT

Fatira area in the Central Eastern Desert, Egypt, is a composite terrane consisting of Neoproterozoic volcanics and sediments laid down in submarine to subaerial environment, intruded by voluminous old to young granitic rocks. The various lithofacies of the study area can be grouped in three distinct lithostratigraphic sequences, which are described here in stratigraphic order, from base to top as the Fatira El Beida, Fatira El Zarqa and Gabal Fatira sequences. Each depositional sequence is intimately related to volcanic activity separated by time intervals of volcanic inactivity, such as marked hiatuses, reworked volcaniclastics, and or turbidite sedimentation.

Four submarine facies groups have been recognized within the oldest, folded eruption sequence of Fatira El Beida. The southern part of the study area is occupied by sheet lava (SL), pillow lavas (PL), pillow breccias (PB), and overlying Bouma turbiditic volcaniclastites (VC). The four facies groups of Fatira El Beida sequence occur in a predictable upward-deepening succession, essentially from base to top, an SL–PL–PB–VC stacking pattern. The coeval tholeiitic mafic and felsic volcaniclastic rocks of this sequence indicate an extensional back-arc tectonic setting. The El Beida depositional sequence appears to fit a submarine-fan and slope-apron environment in an intra-arc site.

The Fatira El Zarqa sequence involves a large volume of subaerial calc-alkaline intermediate to felsic volcanics and an unconformably overlying siliciclastic succession comprising clast-supported conglomerates (Gm), massive sandstone sheet floods (Sm) and mudstones (Fm), together with a lateritic argillite paleosol (P) top formed in an alluvial-fan system. The youngest rock of Gabal Fatira sequence comprises anorogenic trachydacites and rhyolites with locally emergent domes associated with autobrecciation and silt-dyke rock swarms that could be interpreted as feeders and subvolcanic intrusions.

Unconformity and lithofacies assemblages define seven events and three unconformity-bounded tectonic stages that record uplift-subsidence cycles in the study area. A proximal–distal relationship has been established within the depositional products, based on the relative dominance of erosional and depositional features.

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analysis is more appropriate than lithostratigraphy because it is event-oriented rather than lithology-oriented. However, sequence-stratigraphic analysis does not imply an ability to identify global eustatic events, and to distinguish them from more local tectonically driven changes and effects. Rather, as applied here, litho-stratigraphic analysis is used to define the depositional basins-filling architecture.

The Fatira area (460 km²) is located between latitudes 26°37′ and 26°52′N and longitudes 33°20′ and 33°30′E and is accessible by desert track from the Qena-Safaga asphalt road. Previous studies of wadi Fatira area (e.g. Abu El-Leil et al., 1991) have shown that the volcanic rocks there comprise both island arc Metavolcanics and Dokhan Volcanics. Although there has been much research on the volcanic geochemistry and tectonic setting of the volcanics (Abu El-Leil and El Gammal, 1991; Abu El-Leil et al., 1991; Moghazi, 2003), no detailed deciphering of the stratigraphy and the complex depositional system of the volcanic and volcaniclastic rocks has been carried out from a modern volcanological viewpoint. The goals of this study are: (a) to establish the stratigraphic architecture of volcanic bodies through the analysis of the strata relationship between sedimentary and volcanic units; (b) to infer the depositional environment of the volcanic units by means of the sedimentological analysis of coeval volcaniclastic deposits; and (c) to interpret the whole accumulation history in terms of edifice construction. Moreover, the facies sequence established in this work and the processes involved in its formations are compared with the sequences and processes described in other volcanic systems.

2. Geologic background

The exposed various volcanic and plutonic rocks in the Fatira area, belong to the northern part of the Arabian–Nubian Shield (ANS). In spite of four decades of study, ambiguity prevails in the stratigraphic and sedimentological status of several Proterozoic sedimentary successions of the ANS. The ANS (Fig. 1A) began to form ~870 Ma ago and was established ~620 Ma ago, when convergence between East and West Gondwana fragments closed the Mozambique ocean along the East Africa–Antarctic Orogen (EAAO, Stern, 1994; Jacobs and Thomas, 2004). Juvenile Neoproterozoic crust (1000–542 Ma) of the ANS is well exposed on the flanks of the Red Sea (Fig. 1B), making it an excellent place to study Neoproterozoic processes of crustal growth and obduction–accretion tectonics (Kröner, 1985). The ANS is characterized by four main lithologic components: juvenile arc supracrustal sequences, ophiolites, gneissic core complexes, and granitoid intrusions (Shalaby et al., 2005). These were intercalated by thrusting during accretion and by ~600 Ma left-lateral transcurrent movement along the Najd and other, NW-striking shear zones, particularly in the Central Eastern Desert of Egypt (CED, Sultan et al., 1988); a major unconformity formed about 600 Ma.

The basement complex in the CED displays strong ensimatic affinities and consists mainly of an island arc complex which includes abundant Banded Iron Formation (BIF) and is overthrust by dismembered ophiolitic sequences and intruded by older granitoids (Fig. 1C: Stern, 1981; Ries et al., 1983).

The island arc metavolcanics, about 610 m thick, lie below the unconformity, separating older basement units and ~600 Ma Dokhan volcanics and Hammamat sediments. They cover about 116 km² of the mapped area. These volcanics are overlain by tuffaceous metasediments which are intercalated with the BIF; the metasediments and metavolcanics are unconformably overlain by Dokhan volcanics and Hammamat sediments marking the southern flank of the Hammamat basin (Fritz and Messner, 1999). The metavolcanics are intruded by a range of plutons, ~700 Ma and

![Fig. 1.](image-url)
younger, in the region around Gebel Sibai in the CED (Bregar et al., 2002). Map-scale anticlinal folds and mesoscopic overturned and asymmetric open folds belonging to different deformational phases were recorded in these rocks.

The Dokhan Volcanics, about 680 m thick, occupy a vast area (~130 km²). They comprise both intermediate and felsic rock types. These rocks are overlain by local occurrences of molasse-type Hammamat sediments. However, other studies (Khalaf, 2004) proved the presence of interfingering relationship between the Hammamat Group and the Dokhan Volcanics, indicating that they are penecontemporaneous. In the North Eastern Desert (NED), no ophiolitic occurrences are recorded and also island arc rocks are rare. The important rock units are mainly granodiorites, Dokhan volcanics, sediments of the Hammamat Group, and younger pink granites. Accordingly, the Dokhan volcanics and the pink granites are considered as members of a bimodal suite and have been interpreted, with the sediments of the Hammamat Group, to have been formed during a period of strong crustal extension in northern Egypt (Stern et al., 1984).

3. Lithofacies terminology

The Fatira area is subdivided into three lithostratigraphic sequences on the basis of lithologies, lithological sequences, sedimentary structures and homogeneity of facies (Fig. 2). Despite the localized strong penetrative strain and regional greenschist metamorphism, delicate primary textures are commonly preserved and the prefix “meta” is omitted for simplicity. Sections and figures presented here are from the best exposed parts of the mapped area.

Volcanic rocks were formally classified based on texture, color index and phenocryst compositions. Whole rock and trace element geochemistry on reference samples supports this compositional classification (Khalaf et al., 2006) and is summarized in (Table 1). Deposits emplaced directly by volcanic processes are primary volcanoclastics; reworked units are secondary volcanogenic sediments. Primary volcanoclastic rocks are described following the terminology of Fisher and Schmincke (1984) and White and Houghton (2006), using standard granulometric classification (block >64 mm, lapilli 2–64 mm, ash <2 mm) and rock descriptions (tuff has >75% ash; lapilli tuff has >25% lapilli, >25% ash and <25% blocks; tuff breccia has >25% blocks and >25% ash and/or lapilli). The unified scheme for all primary volcaniclastic deposits proposed by White and Houghton (2006) is based on the initial depositional mechanism, and a refinement of grain size terminology. Volcanogenic sediments are redistributed immediately after volcanic eruption by various reworking processes, such as debris flows, hyperconcentrated flows, dilute fluvial flood flows, turbidity and storm currents (Kataoka and Nakajo, 2002).

4. Stratigraphy and lithofacies

The Fatira area comprises a deformed belt that consists of thick volcanics and volcanogenic sediments. The belt extends in a
Table 1
Chemical data of major and trace elements for Fatira volcanics, Central Eastern Desert, Egypt.

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<td>Extension back-arc</td>
<td>Fatira El Zarqa sequence</td>
<td>Subducted-related volcanic arc</td>
<td>Gabal Fatira sequence</td>
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Fe₂O₃: total iron as ferric oxide.
LOI: loss on ignition.
Major elements in wt.%. Trace elements in ppm. UDL: under detection limit.
Rock types: B = basalt; VC = volcanoclastics; BA = basaltic andesite; A = andesite; D = dacite; R = rhyolite.
NE–SW direction, displaying strong lateral variations in thickness and facies. Three volcanic sequences have been recognized in the investigated area (sequences I–III), which are here informally named, in ascending order, the Fatira El Beida, Fatira El Zarqa and Gabal Fatira (Fig. 3). The contacts between the rock types of these sequences and the intrusive rocks are sharp, imposing a noticeable contact metamorphism effect.

4.1. Fatira El Beida sequence

This volcanic sequence is of limited distribution, where it crops out along the flanks of Fatira El Beida wadi. The volcanics form highly to moderately dissected rocks and low hilly country, when compared with the topography of the surrounding terrane. These volcanics have obvious sharp contacts against all rock units. They are faulted along a N–S to NW–SE-trend against older granites and affected by abundant joints trending NE–SW, E–W and NNE–SSW. The lithofacies of the sequence include four groups of volcanic facies: sheet lavas (SL), pillow lavas (PL), pillow breccias (PB), and stratified volcanogenic sediments (VC). These facies groups, which are comparable to those identified in other basins of different ages, are the building blocks of the volcanic system (Castañares et al., 2001). The main characteristics of these groups and their facies are summarized in (Fig. 4).

4.1.1. Facies SL

These lavas are massive, and basaltic to doleritic. They display planar basal contacts and irregular tops, occasionally accompanied by surface folds, brecciated lava tops and millimeter-scale hyaloclastic deposits. They show various trends of fracture foliations. Internally, these flows tend to show three distinctive subdivisions: a lower vesiculated interval; a middle interval of massive to columnar lavas, capped by an upper interval of highly vesiculated lava.

The sheet lavas recorded in the study area represent non-channelized flows that form during the initial stages of subaqueous eruption. They are commonly deposited in the distal setting of

![Fig. 3. The stratigraphy of the Fatira area.](#)
the volcanic system, linked to extensive effusive episodes and associated with the deposits of low-density turbidity currents. These characteristics and the environment interpretation support the conclusion that these deposits reflect high magma discharge rates and high temperature compared to pillowed flows and preferential settling in low-relief areas, with source of lavas either from main feeding areas (proximal volcanic setting) or more local fissure feeding (Yamagishi, 1991; Gregg and Fink, 1995). Abundant vesicles at flow bases and tops are common features of subaqueous massive flows (Aubele et al., 1988).

4.1.2. Facies PL

This facies group consists of bulbous to spherical-shaped bodies, with a mean individual diameter of 30 cm. The surface structures diagnostic of pillow lavas are represented by irregular cracks and amoeboid vesicles (Fig. 5A). Internally, the pillows are strongly vesiculated (Fig. 5B). The pillow lavas are massive, highly compact and mostly basaltic in composition. Petrographically, these pillows display fine-to-medium-grained, subophitic to intergranular or intertextal textures and are composed of actinolite, chlorite, epidote and altered plagioclase with subordinate amounts of quartz. The margins are chilled, and occasionally fringed with globules or shards, while the core shows a variolitic texture.

These features of pillows (small size, presence of chilled margins and abundant vesicles) indicate their accumulation at high initial temperature, and hydrostatic pressure, moderate viscosity, and rapid eruption rate, when compared to modern mid-oceanic ridge pillows (Farahat et al., 2010). Pillow lava may resemble sub-

![Fig. 4. Stratigraphy of Fatira El Beida sequence.](image-url)
aerial pahoehoe flows (Walker, 1992), but pillowed flows are distinguished by radial joints and the association with hyaloclastite (Mcphie et al., 1993). Fractures in pillows develop from thermal contraction during cooling (Yamagishi, 1991). The bulk of pillow lava deposits are interpreted as having formed directly from central vents of proximal to medial volcanic settings, with development of pillow lava morphologies (cf. Dimroth et al., 1978).

4.1.3. Facies PB

This facies consists of highly chaotic, disorganized brecciated deposits of angular to subspherical lava clasts of various sizes. The pillow breccias are monogenetic volcanic breccias composed of broken fragments of pillowed lava (Fig. 5C), embedded in a matrix of fine-grained altered hyaloclastic materials. The breccias are clast-supported to matrix-supported structures. Such pillows are usually characterized by concentric joints and were therefore called concentric pillows (Yamagishi, 1985). The recorded concentric pillows are ellipsoidal or spherical in shape and are a few of tens of centimeters across (Fig. 5D). They contain abundant phenocrysts (30–40 vol.%) of plagioclase and pyroxene, set in a hyalopilitic to intersertal groundmass. Contorted large vesicles are sporadically distributed in their interior. They range in size from a few millimeters to 1 cm, and have a tendency to increase in size with the pillow size. The margins of the concentric pillows are characterized by quench crystals, such as fibrous and dendritic overgrowths on microlites.

These facies occur in the most proximal settings of the volcanic system and are intimately associated with the pillow lava facies. They may be the result of eruptions of relatively low temperature and high viscosity lava, associated with very low magma discharge rates (Bonatti and Harrison, 1988). Pillow breccia generally develops during quench fragmentation, resulting from lava–water interactions (Dimroth et al., 1978; Yamagishi, 1991). Alternatively, pillow fragment breccia forms during mechanical disintegration of pillow lava due to slumping (Fisher and Schmincke, 1984).

The transition from the pillow lavas to pillow breccias is produced by a decrease in effusion rate towards the end of an eruption (Dimroth et al., 1978). Both the texture and location of the pillow

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**Fig. 5.** (A) Basalt pillowed-shaped with amoeboid vesicles. (B) Sketch showing internal structure in pillow basalt. (C) Brecciated pillow fragments set in a fine-grained matrix. (D) Pillow breccias exhibiting ellipsoidal to spherical in shape. (E) Stacked volcaniclastic rocks suffering different fold styles and soft sediment deformation. (F) Bedded volcaniclastic rocks (VC) are rhythmically interstratified with carbonate-rich mesobands (CM).
breccias indicate that they formed as small, slope apron deposits produced by a combination of hydroclastic processes, autobrecciation at pillow lava, rock fall generated by gravitational unsteadiness, and perhaps other less perceptible phenomena, such as synvolcanic seismicity.

4.1.4. Facies VC

This facies consists of volcanogenic sediments stacked in strata, with angular clasts ranging up to boulder size, poorly to immediately vesiculated and generally involving hydroclastic processes. The volcanogenic sediments crop out mainly in the northwestern part of the investigated area. They form moderate to high relief, and are foliated and thrust sheared masses (Fig. 6) that exhibit soft sediment deformation with local-scale folds of different styles (Fig. 5E). Occasionally, these rocks show subparallel fracture arrays which are enriched in carbonate-rich mesobsand (Fig. 5F). These rocks, which make up the bulk of the Fatira El Beida volcanic sequence, are composed of volcanic breccias (VB), lapillituffs (LTm), clast-supported conglomerates (Gm), tuffaceous breccias (TB), Banded Iron Formation (BIF), and mudstones (FI) in sequences up to 100 m thick. The clast-supported conglomerates represent clastic sedimentation during a hiatus following cessation of basaltic volcanism.

4.1.4.1. Facies VB (diamictites). The volcanic breccias occur as massive or poorly structured units, 15 cm to many meters thick. They are composed of inequant, angular to subrounded monolithological clasts of basaltic rocks that are poorly sorted, ranging from few to several centimeters in diameter. The deposits are clast-supported and form beds that show flat, non-erosive bas, no internal stratification and disorganized clast fabrics (Fig. 7A).

This lithofacies displays features typical of debris-flow deposits: poor sorting, lack of stratification, and non-graded beds without erosive contact (Nemec and Steel, 1984). The large number of dense, angular to subangular rock clasts and broken minerals can be explained either by small-scale debris avalanche originating from rock slide, rock fall, or dome collapses, after or during basaltic eruptions.

4.1.4.2. Facies LTm. These deposits occur as discontinuous thin units, <3 m thick, between and along strike of individual mafic flows. Clasts in these rocks are typically of similar vesicularity or phenocryst content as adjacent flows and pillow lava. They are massive, light gray in color, non-graded, brecciated and comprise monogenetic angular plagioclase and basaltic clasts. All these clasts are set in platey, blocky pieces of sideromelane shards, showing dark marginal palagonite rims (Fig. 7B). Occasionally, the matrix contains local spherulitic and snowflake devitrification, indicating alteration of original glass. The contacts with underlying volcanics are locally gradational, with transitions over a meter from intact or partial pillows, to blocky or irregular breccias and lapillituffs.

The blocky to irregular morphology, clast textures and contact relationships of the lapillituffs and breccias are consistent with an interpretation as hyaloclastite breccias, derived by quench fragmentation of adjacent lava flows, rather than explosive fragmentation (Staudigel and Schmincke, 1984; Walker, 1999).

4.1.4.3. Facies Gm. This facies is characterized by crudely stratified beds, that are massive units 1–3 m thick and normally to symmetrically graded (Fig. 7C). The beds comprise unsorted, subrounded and disorganized heterolithic clasts, that are brown to dark gray and range from 2 mm to 10 cm in diameter. These clasts include microcrystalline basalts, vesicular sideromelane (Fig. 7D), bedded chert, mudstone, fine tuffs and undifferentiated volcanic clasts, set in a poorly sorted, feldspathic sandy matrix. All the clasts are invaded by extensive irregular carbonate and quartz veins. The multiple generations of these veins indicate erosion of a hydrothermally altered source. Their contacts to overlying and underlying beds are gradual. Horizontal stratification is distinguished by fine tuffaceous layers with developed stacking of pebbles.

The unsorted and commonly stratified structure of this facies reflects deposition from hyperconcentrated flows or high concentration turbidity currents fed by explosive eruption. Penetration of clasts into the underlying substrata indicates free fall of clasts (Bose et al., 2008). Hyperconcentrated flows are variously interpreted in the literature as cohesive and transformed cohesive flows (Mulder and Alexander, 2001) or cohesionless debris flows and sand flows (Allen, 1997). The clast-supported conglomerates with disorganized clasts, non-erosional base and coarse sandy matrix are most probably formed from subaerial pseudoplastic flows with high sediment concentration (sensu hyperconcentrated flow; Sohn et al., 1999). The sand/mud ratio >1:1 and presence of disorganized gravel to boulder-sized clasts bear telltale features in support of a high concentration flow character (Mulder and Alexander, 2001).

4.1.4.4. Facies TB. This facies occurs in the form of fining-upward units, ranging in thickness from 1.0 to more than 5.0 cm. It shows small-scale subdivisions into three facies: (i) massive with normal graded bedding (Ta); (ii) parallel laminated (Tb); and (iii) rippled or convolute laminated tuffaceous rocks (Tc) (Fig. 7E and F). The first unit is poorly sorted, coarse-grained, and encompasses angular to subangular crystal and lithic clasts. Altered plagioclase, chloritized hornblende with opacitic rims and monocrystalline quartz are the main crystals; the lithic clasts include basaltic and andesitic rock fragments set in a cryptocrystalline matrix. The lowermost facies unit is invariably sharp and locally fills shallow scours; whereas the middle unit is massive, fine-grained, moderately sorted and is composed of thin alternating laminated layers of fine ash and lithic-rich tuff. The third unit is fine-grained, poorly to moderately
sorted, wavey to convolute laminated and includes micrograded siltstone and mudstone.

The succession of structure intervals resembles Tabc of a “Bouma cycle” (Lowe, 1982). The normal graded bedding within the first unit represents deposition from low-density turbulent flow dispersion (Chakraborty and Pal, 2001). Incorporation of clasts from the substrate indicate strong shear at the base of the flows, a character of proximal turbidite deposition. The intervening massive or plane laminated unit two intervals lacking evidence of exposure and without current or wave features, resembles open ocean deposition below wave base (Bera et al., 2008). The third unit involves traction of grains on the beds, with division C representing a rippled or convolute laminated bedform, which may resemble hummocky cross-stratification (HCS) that attests to the occurrence of major storm events. The HCS units reflect the growth of vertically accreting bedforms with high sedimentation fall-out from upper flow regime currents (Cheel and Leckie, 1993). Lack of wave reworking on top of the inferred storm-laid beds indicates deposition between the fair weather and storm-wave bases. This is closely consistent with the non-erosional base of a HCS unit, which is more frequent in deeper shelves below a fair weather wave base (Bose et al., 1988). Absence of welding, lack of vesicles and rarity of glass shards indicate deposition of tuffaceous breccias in a cold state, under the direct influence of phreatic or phreatomagmatic phases (Pal et al., 2005).

4.1.4.5. Facies BIF. BIF vary in color from reddish-gray to steel gray or even black. It occurs as stratified bands, or separate lenses (Fig. 8A). They change in thickness from few centimeters up to 2 m. BIF bands have sharp contacts against the intercalated volcaniclastics and their deformational structures conform in the pattern and attitudes with those of the host rocks. Lamination, convolute and lenticular bedding, cross-lamination, slump and flaser structure are primary structures (Fig. 8B) inherited from the
parental precipitates and reflect the difference of the original components of the alternating bands. BIF bands exhibit also rhythmic banding, where iron-rich mesobands alternate with tuffaceous mudstone, chert or jasper. The BIF mesobands are commonly dissected by later quartz and carbonate veinlets. Relics of mesoscopic open folds of BIF are recorded in some of the deformed bands (Fig. 8C).

Plane-laminated, cross-laminated and rhythmically laminated BIF make up classic Tcde, Tde and Te (hemipelagic mud) turbidites (Walker, 1984). Massive to plane-laminated couplets in BIF indicate that the precursor sediments to BIF were density current deposits. For those sediments to have been draped across lowstand basin-floor fans, they must have been transported by bottom currents or gravity-driven turbidity currents (Krapez et al., 2003). The lack of hummocky cross-stratification (HCS) in these BIF bands indicates that they accumulated below storm-wave base. Occurrence of soft sediment deformation structures, however, indicates that the settlement of elutriated ash particles was associated with basin-scale instability (Chakraborty et al., 2009). The same conclusion was reached by El-Habaak (2001), who interpreted the BIF in Egypt as deposits accumulated below the wave base in a predominantly quiet-water shallow restricted basin, probably tidal flats or lagoons along the shores of island arcs during periods of relaxed volcanic activities. Many studies described the Egyptian BIF as submarine volcano-sedimentary Algoma-type deposits (Sims and James, 1984; El Gaby et al., 1988), but also suggested that the BIF evolved during photosynthesis by algae (El-Habaak and Mahmoud, 1995).

The ratio between BIF and clastic bands, fabric and deformational style in BIF are mainly controlled by concomitant changes in hydraulic regime and sea-water chemistry on a storm-dominated shelf (Pufahl and Fralick, 2004). Unlike other models, which rely on upwelling to transport Fe$^{2+}$-bearing deep ocean water over an O$_2$-enriched shelf, the most recent studies (Kirschvink, 1992; Stern et al., 2006) add another dimension by emphasizing the importance of storm-generated transport of oxygenated water following anoxia caused by a global ice sheets, to the deep sea where it mixes with Fe$^{2+}$-bearing ocean water to form insoluble Fe$^{3+}$, which precipitated out as BIF during the Sturtian Snowball Earth hypothesis.

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Fig. 8. (A) Conformable Banded Iron Formations (BIF) are rhythmically interstratified with volcaniclastic rocks (VC). (B) BIF showing convolute lamination, cross-lamination and flaser structure. (C) Small-scale open fold in BIF. (D) Load cast structure (arrow) of siltstone (SS) into mudstone (MS). (E) Subangular monolithological clasts in matrix-supported breccias. (F) Altered pumice fragments (PF) with wispy terminations embedded in a fine matrix.
BIF is part of a metasedimentary succession associated with ophiolites in the CED (Stern, 1981). The Fawakhir ophiolite in the CED has been dated at 736 ± 1.2 Ma (Andresen et al., 2009). Ophiolite, BIF and diamictite may preserve evidence of deep marine conditions during Sturtian glaciation in the CED of Egypt and in NW Saudi Arabia (Stern et al., 2006). The BIF also provides opportunities for studying the effect of Neoproterozoic climate change on the deep, open ocean. Analogy with the modern seafloor indicates that the ANS ophiolites formed at depths of 2–3 km below sea level and the overlying pelagic sediments should record chemical products of these deep waters (Stern et al., 2006).

4.1.4.6. Facies FI. This facies consists of massive to delicately laminated gray to green mudstone, which contains a variable proportion of discrete, interbedded siltstone. Petrographically, the mudstones are composed of laminated and wave-laminated layers, 1–4 cm thick, of mudstone and fine silt-sized debris intercalated with laminae of fine silt and probably clay-sized sediments. Mud-sized laminations are made of angular quartz grains, sericite, and actinolitic laths-rich laminae. Occasionally, fine-grained silt-rich laminae occur as lenticular laminae that swell laterally over a few centimeters into isolated, wavy form mud sets. The silt layers have a sharp base, may be massive or exhibit well developed normal graded bedding, together with the tracially sedimentary structures, which characterize the Bouma “B” and “C” divisions (Bull and Cas, 1991). The thicker layers of silt display loaded bases (Fig. 8D).

The rhythmic laminations, the grading in size, the intervals comprising complexly interbedded mudstone and siltstone, together with the sheet-like nature of this facies, indicate accumulation from turbidity currents, that transported mostly muddy sediments. The mudstone facies is interpreted to have formed mainly from suspension deposits in off-shore water and ambient sedimentation in a quiet, relatively deep, subaqueous environment below the influence of storm waves and bottom currents. The graded and loadstite layers may also represent low-density turbidity deposits and storm-generated sheet flow in a water depth of up to 200 m (Johnson and Baldwin, 1996). Periodic deposition of BIF and mudstone among the turbidites suggests varying oxygen conditions and periods of restricted sediment input in the deep-marine basin (Kimberley, 1989). The variation in color, texture and lamination style observed in the mudstone facies possibly resulted from changes in either water depth or sedimentation rate. The presence of mafic grain-rich laminae represents erosion of a mafic/ultramafic source, but also indicates slack water periods, possibly following a subordinate flood current stage (Eriksson and Simpson, 2000).

4.2. Fatira El Zarga sequence

This middle sequence occupies a large part of the mapped area, forming moderate to high mountainous ridges with uneven topography and crumpled peaks. This sequence includes voluminous lava flows and volcanioclastic (LF and VC) that are all unconformably overlain by alluvial fan volcanogenic sediments (AVFC) (Fig. 9). These rocks are faulted and jointed mainly along NE–SW, NNE–SW and NW–SE directions.

4.2.1. Facies LF and VC

The field observations determined two suites of massive vari-colored volcanics: (a) a lower volcanic suite of basaltic andesite and andesite lava flows and andesitic volcanioclastic intercalations; and (b) an upper volcanic suite of more silicic lava flows and pumice-rich tuffs.

4.2.2.1. Lower volcanic suite. This suite comprises three black, green and gray rock varieties, namely pyroxene andesites, hornblende andesites, phytic and amygdaloidal andesites. The pyroxene and hornblende andesites contain dominant zoned plagioclase (An50), colorless augite and green hornblende in an intergranular matrix dominated by plagioclase laths, mafic granules and iron oxides. The amygdaloidal andesites are mafics-free and encompass mainly plagioclase phenocrysts (An65) and chlorite-carbonate–quartz-rich amygdalae, in a hyalopilitic matrix dominated by plagioclase laths and secondary minerals. Flow and pilotaxitic textures are recognized. Amygdaloidal varieties of the andesites are common, where the filling minerals include chlorite, calcite and quartz.

The volcanioclastic deposits are coarse- to fine-grained, evenly bedded and poorly sorted; the beds are internally massive or plane laminated. Predominant lapillituffs and crystal-rich and crystal-free laminated tuffs constitute the main rock facies of these deposits. Occasionally, a 0.5– to 1.0-m-thick volcanioclastic breccia (Fig. 8E), which consists of disorganized subangular to angular, dominantly cogenetic lithic clasts up to 30 cm in size, is embedded in these deposits. The lithofacies of the volcanioclastics is composed of unsorted lapilli and tuff-sized crystal and lithic clasts with a fine-grained matrix. The lithic clasts are brown to dark gray, commonly oxidised, altered, millimeter to centimeter-sized subangular to angular and contain plagioclase, pyroxene, hornblende and fewer biotite phenocrysts. Block-sized lithic clasts with prismatic jointing occasionally occur. Stratification is not observed, but weak gradation is recorded.

Features of the volcanioclastic facies are characteristic of subaerially emplaced, non-welded plinian fall-out pyroclastic deposits during phreatomagmatic eruption. These deposits show little signs of hydrothermal alteration, corresponding to subaerial emplacement where hydrothermal alteration is limited, compared with subaqueously emplaced deposits (Cas and Wright, 1987).

4.2.2.2. Upper volcanic suite. The silicic volcanics are mainly dacites and rhyolites which show brown, pink and red colors. They are vesicle-free and have a conspicuous porphyrytic fabric. The main minerals of these rocks are K-feldspar, plagioclase (An15), quartz and occasionally opatic biotite. Zircon, apatite and opaques are accessory minerals, while chlorite and calcite are secondary products.

The pumice-rich felsic tuffs are fine-grained, homogenous, massive and rarely contain lithic fragments. Framework grains in these tuffs are dominantly broken feldspar clasts, volcanic lithic fragments and vitriclasts with minor volcanic quartz and opaque. The vitriclasts encompass felted aggregates of sericite, with minor carbonate, quartz and chloride, and commonly contain feldspar phenocrysts. Many have ragged, wispy terminations consistent with having originally been pumice clasts (Fig. 8F). Lithic fragments within these tuffs are subangular to subrounded and range in size from granules to large pebbles. The most common lithologies are porphyritic to aphanitic intermediate to felsic volcanics, which are non-vesicular and commonly have devitrification, spherulites and/or perlitic fractures.

The nearly ubiquitous occurrence of broken crystals, together with relict pumice and devitrified glassy matrix, lead to the interpretation of these tuffs as subaerial ignimbrite (Cas and Wright, 1987).

4.2.2. Facies AVFC

These sediments prevail in the NE part of the Fatira area and form relatively low hills compared with the topography of the lava flows. The beds strike in a N15°–N65°/dipping 20–55° to SE. The primary structures recorded include stratification and ill-defined graded bedding. These alluvial fan sediments encompass immature to submature interbedded sequences of clast-supported
conglomerates (Gm), massive sandstones (Sm), mudstones (Fl) and lateritic paleosols (P), which reflect an inter-eruption period. The contacts between these rock sequences are sharp and abrupt.

4.2.2.1. Facies Gm. The beds are sharply based, massive, 1–5 m thick and have erosional contacts. The deposits are dominantly crudely bedded channel-fills and sheet-like, clast-supported, disorganized to poorly organized conglomerates, with little indication of graded bedding. They are composed of oval, rounded to subangular rock clasts with weak imbrication (Fig. 10A), and measuring a few to 15 cm in diameter, set in a fine sandy matrix. The clasts are of various rock types, including andesitic and felsic volcanics, and reworked sedimentary fragments (Fig. 10B).

The conglomerates are interpreted in terms of longitudinal gravel bars associated with fluvial channels collectively forming a broad, low-relief bar complex (Olson and Larsen, 1993). A multipoint character of the source is also attested to by the local variation in clast composition (cf. Hwang and Chough, 2000). Low mud content in such conglomerates is suggestive of a subaerial origin.
4.2.2.2. Facies Sm. At the base of the upper interval, the rocks of this facies reveal local-scale ripple cross and horizontal lamination, and display a distinct fining-upwards trend (Fig. 10C). The sandstone rocks are massive, 3–7 m thick and dominantly composed of fine- to medium-grained, subangular to subrounded and moderately to poorly sorted grains of monocrystalline quartz, feldspar and rock fragments, embedded in a finer grained silty and clayey matrix. The feldspar grains are plagioclase which exhibit oscillatory zoning. The lithic fragments are mainly volcanic clasts which encompass twinned plagioclase and altered mafic minerals set in a matrix of fine quartz, chlorite and epidote. These clasts show trachytic, hyalopilitic, pilotaxitic and felsitic textures.
This facies association is similar to stacked deposits of sand-rich low sinuosity channel-fill, with only small-scale bar development, probably under flood conditions (Miall, 1996). The occurrence of cross-lamination sandstone at the base with horizontally laminated units suggests variable energy conditions, with transition from lower flow regime dune bedforms to upper flow regime plane bed conditions (Eriksson et al., 1995).

4.2.2.3. Facies Fl. The mudstone rocks are laminated, 1–3 m thick and encompass angular to subrounded chips of quartz, feldspar and opaques in a clay-size matrix. They display a rhythmic banding with alternating thin bands of silt-rich and mud-rich laminae. Most of the mineral constituents in each laminae are elongated and oriented parallel to the laminar planes. The silt and mud-rich laminae are continuously and distinctly graded. The mudstone facies association is interpreted as flood plain deposits, pointing to an inactive period of alluvial system during over-bank flooding that was later subjected to pedogenic processes (Jo et al., 1997). The above described three units point to waning flows and deposition from density underflow occurring during flood-related influxes and sediments (Massari et al., 1993).

4.2.2.4. Facies P. The paleosol deposits form a low-relief pediment of a very well-developed, relict ancient profile (2 m thick, Fig. 10D). These deposits are extremely porous, fine-grained, poorly sorted and laminated with brown to reddish-brown ochre. The weathering profile of these deposits consists of the following textural progression. The topmost part is a highly indurated quartz–carbonate layer, which passes downward into softer layers (clay-rich), in which the vermiciform/or tubes (hair-thin partings) become progressively well developed; then further downward into brownish-red horizons (authigenic iron compounds), in which individual pisoliths (spherical globules) structure may be recognized (Fig. 10E). The color of the clays varies from almost pure white, when free of iron, to bright red when they contain hematite or ochre-brown goethite. Steeplly dipping conjugate joints are also developed in the paleosol profile (Fig. 10F). Below the vermiciform and pisolithic horizons, undurated lithomarg layers occur, which eventually pass down into saprolitic and less weathered horizons, in which structure and crystal pseudomorphs of the protoliths may be recognized.

The development of a brownish-red profile, its intensive clays, and the presence of iron oxyhydroxides, suggest a lateritic environment, which was formed under very effective, deep chemical weathering processes. The presence of hematite or ochre-brown goethite, and the development of a brownish-red profile, suggest that the area was subjected to lateritic weathering, and the presence of iron oxyhydroxides, suggest a lateritic environment. The development of a brownish-red profile, its intensive clays, and the presence of iron oxyhydroxides, suggest a lateritic environment, which was formed under very effective, deep chemical weathering processes. The presence of hematite or ochre-brown goethite, and the development of a brownish-red profile, suggest that the area was subjected to lateritic weathering, and the presence of iron oxyhydroxides, suggest a lateritic environment. The most recent subdivision of deformation events in the CED includes four main events (Loizenbauer et al., 2001; Fritz et al., 2002; Shalaby et al., 2005). The oldest deformation phase in gneisses commenced by ~700 Ma (Fowler et al., 2007), followed by thrust-related structures associated with oblique convergence of the arc and back-arc assemblage onto the Saharan Metacraton around 620–640 Ma. Subsequently, a prolonged phase of transpression and lateral extrusion was linked to the formation of crust-scale NW-trending sinistral shear zones of the Najd Fault System, followed by exhumation of core complexes within orogen-parallel extension around 620–560 Ma. So, the three deformation events in the Fatira area, are in agreement with the general evolutionary model for the CED in the ANS that begins with NNW–SSW shortening, followed by ENE–WSW compression and subsequent deformation by the NNW–SSW striking Najd Fault System.
several studies often have difficulty in determining whether a particular ancient deposit formed subaerially or submarine (e.g. Martin et al., 2004). Even though the studied deposits are old, folded, metamorphosed and vertically dipping, primary sedimentary structures (e.g. pillows, graded bedding, cross-lamination) readily indicate younging, and hence permit stratigraphic reconstruction. Transition from subaqueous to emergent stages of volcanism has been documented by several studies.

The Fatira stratigraphy is characterized by complex vertical and horizontal volcanics and volcanogenic sediments. All available data indicate that the Fatira area involves a pile of subaqueous and continental deposits. The presence of homogeneous pillowed and massive sheet lavas in the preserved sections of the lower volcanic sequence is indicative of eruption on the flanks of a subaqueous mafic shield volcano or seamount (Walker, 1993). The widespread carbonate-rich mesobeds, interbedded with volcanogenic sediments (possibly from CO₂-bearing sea water) support a marine environment. Submarine eruption is further evidenced by: (1) the frequent features characteristic of traction transport such as bedding, stratification, rhythmic fine/coarse units, and hummocky cross-stratification (HCS), which suggest the possible role of wave or seismic activity; (2) rhythmic relations between volcanioclastic and BIF (storm deposits); (3) syn-sedimentary textures (e.g. graded bedding, slump structures) and inferred depositional processes (various types of turbidites, high and low-density deposits) are indicative of relatively deep-water flysch-like sediments (Lahtinen et al., 2010); and (4) occurrence of hyaloclastics and high vesicularity of the volcanic glass shards that indicate shallow subaqueous eruption (Staudigel and Schmincke, 1984). Concerning the depth of eruption and deposition of the older sequence, the abundance and relatively average size of vesicles and amygdules in the mafic lavas support moderate water depths (<1000 m) for theolithic volatile-poor basalts (Dolozi and Ayer, 1991). The paucity of mafic fragmental rocks relative to flow is also consistent with this interpretation. Taken together, and considering the lack of wave-influenced or tidal structures in the volcanogenic sediments, eruption and deposition must have been below storm-wave base (>200 m; Leckie, 1988).

This volcanogenic sedimentary sequence is generated by successive stages of deposition as a hyperconcentrated flow that evolved to a high-density turbidity current and finally, to a low-density turbidity current. These kinds of deposits have been closely studied in siliciclastic turbidite systems (Lowe, 1982) and resemble density turbidity current. These kinds of deposits have been closely

4.2. Depositional stages of the Fatira volcanics and volcanioclastics

In the preceding sections sedimentary facies associations have been outlined with the physical spectrum of depositional processes involved in sediment dispersal. The facies associations indicate an overall submarine to continental depositional system.

6.2.1. Submarine system

The association of mudstone with coarse-grained turbidites and debris flows implies a slope or basin-floor setting, whereas features consistent with a submarine fan are: (1) turbidites, which are most abundant in deep-marine basins (Walker, 1978) and (ii) packaged conglomerate and breccia bodies that represent channel and lobe deposits (Shanmugam et al., 1988). These depositional units become progressively thinner and fine-grained up-section, recording abandonment of the submarine fan and subsequent channel filling. In classic models of submarine fans (Walker, 1978), the proximal part, directly down-slope from the feeder channel or canyon, is characterized by a single deep, levee-bound channel. In a down-slope direction, the next elements are submarine-fan lobes, which develop as turbidity currents and debris flows dissipate after leaving the confines of the main channel. Classic fans have two lobe types. Suprafan lobes have proximal braided channels, whereas depositional lobes are not segmented by channels. The most-distal setting lacks channels, is relatively free of sand and is dominated by hemipelagites and mud–rich turbidites (Walker, 1978).

Taking into account the facies associations and their distributions, three depositional intervals were distinguished in El Beida sequence: proximal, middle and distal intervals (Fig. 11A), which are interpreted in terms of the classic model of submarine fan. The proximal interval which occurs in the innermost areas of the volcanic fan, is characterized by subaqueous effusive eruption with development of SL flows and amalgamation of pillow lavas, which occur in the front and marginal zones of sheet lavas. This interval is interpreted to be an area of abundant points of effusion, high magma discharge rate, and long-lived continuity (Castañer et al., 2001). The middle interval is characterized mainly by meter-scale packages of pillow lavas that pinch out over a distance of several meters. The pillow lavas may be underlain by thin sheet lavas or associated with pillow breccias and/or volcanioclastic deposits. The latter two intervals, coupled with debris–flow deposits (breccias and conglomerates) equate to the deposits of proximal submarine-fan channels, recording deposition in braided channels on sandy lobes at the mouths of main, levee-bound channels. Turbiditic tauffaeous breccias and mudstone deposits (distal interval) represent deposition in a lobe-fringe setting below storm–wave base, which is characterized by regularly interbedded tabular turbidites and hemipelagites (Walker, 1978). Lobe-fringe deposits interfinger with BIF. This interval (volcanic fringe) is characterized by a smooth relief and records the filling of a proximal canyon that fed a distal submarine fan. Volcaniclastic deposits are typically more common in remote parts of a volcanic edifice, away from the near-vent setting (Mcphie, 1995).

6.2.2. Continental system

This stage represents a complex interplay between volcanic activity and relatively long-lived epiclastic sediments. In the ancient rock record, in the absence of reliable paleoslope data and geomorphological attributes, the description and characterization of a fan– or a braid-delta relies considerably on sedimentological characterization and stratigraphic disposition. In sedimentological paralance, the dominance of sediment gravity flow and unconfined sheet flood deposits, along with the limited presence of channel flow products in the Fatira El Zarqa section, have been considered as earmark signatures in favor of an alluvial fan (cf. Blair and Mcpherson, 1994).

The sediments dispersal on the alluvial fan of El Zarqa section offers scope for interpreting the proximal–distal relationship within the alluvial-fan system (Fig. 11B). Landscape aggradation by catastrophic events, such as lava flows and stacking of volcanioclastic deposits, with an intervening period of fluvial incision, affected the northern part of Fatira area. Initially, the magmatic eruptions were rapidly succeeded by emplacement of mafic/felsic rocks and volcanioclastics situated near the volcanic vent. Transporting agents supplied sediments to fan bodies through conglomerate/sheetflow deposits (Gm and Sm) in a proximal-middle alluvial fan, sourced from the southern and northern parts of the study area. These
deposits were supplied by relatively higher gradient, lower order streams, that funneled over short distances, resulting in rapid transfer of overland flow and flash flooding (cf. Blair and McPherson, 1994). The presence of grain rounding, however, suggests that the water energy was only moderate. The gravity flow units (facies Gm) within this setting are products of occasional flood events in the feeder system with the supply of a denser sediment–water mixture. Sheetflood sandstones with poorly sorted grain populations and with ripple cross and horizontal lamination (facies Sm) represent a braid-plain system (cf. Eriksson et al., 1998). Continued accretion of fan sediments (distal facies) is characterized by the stacking sequence scale fining-upward succession involving mudstone deposits (facies FI), when the flows turned increasingly steady and weaker with loss of slope. The amounts of volcanics/volcaniclastics pinch out when moving far away from the volcanic vent. Substantial contemporary tectonic activity is evidenced by the switching from proximal to distal over-bank fluvial incision.

Finally, trachydacitic/rhyolitic domes and bimodal dyke swarms (Gabal Fatira sequence) intruded these rock types, causing gentle upwarping and the formation of dome-like structures. These structural features suggest that felsic volcanic and plutonic activity, related to the formation of a small caldera, contributed to the uplift of the Fatira area during this interval.

6.3. Geotectonic setting

Provenance, stratigraphic architecture and magmatic composition are powerful indicators of tectonic setting (Dickinson, 1988; Wilson, 1989). Petrogenetic discussions are not included here (see Khalaf et al., 2006).

The early facies crops out in a map-scale anticlinal fold core. Major element data confirm that the mafic rocks in El Beida sequence are mainly basalts, whereas volcaniclastics are dominantly dacitic in composition (Fig. 12A), recording bimodal products. Coeval mafic and felsic volcaniclastics support an extensional setting for the Fatira El Beida sequence. Upward-deepening sets of progradational depositional sequences (like SL → PL → PB → VC in wadi El Beida) characterize extensional basins, because subsidence increases exponentially to create accommodation that sediment flux cannot fill (Krapez, 1997). The Fatira El Beida sequence can be correlated with Shadli Older Metavolcanics (OMV) in the CED. The OMV are associated with CED ophiolites (Stern et al., 2004). These volcanics are low-K-tholeiitic basalts, often pillowed and overlain by immature wackes representing deep water turbidites. Some of the El Beida mafic rocks are quite primitive, with high Ni (up to 516 ppm) and Cr (up to 195 ppm; Table 1), which are inferred to have formed by melting mantle peridotite and erupted without significant fractionation, further implying short residence times in the crust. These rocks are tholeiitic with respect to the classification (Fig. 12B) of Miyashiro (1974), with MORB setting for mafic volcanics (Fig. 12C) and volcanic arc setting (VA) for felsic volcaniclastics (Fig. 12D). The bulk rock chemistry of these mafic and volcaniclastic rocks therefore testifies to generation of their protoliths in a tectonic environment of MOR and VA setting. This overlap points to a back-arc basin setting (Tadesse and Allen, 2004), where magma generation could involve subduction-modified mantle source materials (Wilson, 1989). A back-arc setting is further confirmed by the composition of the mafic rocks, which have Zr values in the range of 65–229 ppm, and Zr/Ti ratios of 30–84, matching well with that of back-arc basalts rather than that of island arc basalts (Woodhead et al., 1993). Moreover, the abundant occurrence of ophiolites in the CED supports a back-arc setting of the El Beida sequence, since many ophiolite complexes are now regarded as obducted floors of back-arc and forarc basins rather than as tectonically emplaced slivers of true oceanic litho-
sphere (Farahat et al., 2004). These old volcanics were generated by melting of depleted mantle as shown by epsilon Nd at 750 Ma of +5.1 to +9.0 and the fact that Nd model ages are similar to U–Pb zircon crystallization ages in the CED of Egypt (Ali et al., 2009a).

Most of the volcanogenic sediments are interpreted as the result of early resedimentation via sediment gravity flows of freshly erupted volcaniclastic materials. Recycling of older felsic supracrustal rocks in the CED is established from zircon provenance (Ali et al., 2009a), who emphasized that pre-Neoproterozoic zircons are abundant in ~750 Ma igneous rocks in Egypt and Saudi Arabia (Fig. 1A). This supports the observations of Wüst (1989) who studied detrital zircons from Neoproterozoic sediments in the CED and SED; wadi El Miyal metasediments yielded ages of 2410 Ma, while those along wadi El Allaqi yielded ages of 1460, 2400, and 2450 Ma. So, it is clear that significant magmatic digestion of these sediments could provide the pre-Neoproterozoic zircons seen in the Fatira area. The fairly rounded, compact to elongate clasts (Fig. 7C) also establish resedimentation from a subaerial environment. Hassan and Hashad (1990) argued that Mesoproterozoic to Archean crust may exist beneath the Eastern Desert of Egypt. However, pre-Neoproterozoic zircons may not be found in all ANS igneous rocks (Ali et al., 2009a). This presumably reflects different magma compositions and different amounts of time that magmas and xenocrystic zircons have to interact. The dominance of these recycled volcanicogenic sediments may be linked to fault-control on basin architecture, because most syneruptive products may have been stored in subaerial environments and only made available to the deep-marine environment following uplift-related erosion, so that it is not surprising that volcaniclastic systems dominate the Fatira El Beida basin fill. Because there is evidence for recycling of cratonic basement, the rift setting is highly likely to have been an intra-arc site.

Rates of uplift-subsidence, sedimentation and basin lengthening are high in strike-slip basins, with subsidence being controlled by a range of processes from pull-apart to fault-flank depression by listric rotation on normal faults or by tectonic loading on reverse oblique-slip faults (e.g. Christie-Blick and Biddle, 1985). Rates of subsidence and extension are certainly high enough to be accompanied by basaltic magmatism. The deformation event of the strike-slip shear zones is recognized and can be divided into an early and late transpression (660–560 Ma) during orogen-parallel extension in the CED (Abd El-Wahed, 2008). During the early transpression (660–645 Ma), older granite shear zones were deformed in a sinistral sense (Fig. 2). During the late transpression (645–560 Ma), an external sinistral set of NW–SE trending strike-slip shear zones was deformed in a sinistral sense, followed by reactivation of NE–SW oriented low angle normal faults as well as intrusion of the younger granites. Wadis El Beida and El Zarqa shear zones define the southwestern and northeastern margins of the Fatira area. During the early stage of thrusting and folding, the El Beida rock sequence was a NW-directed thrust, along which duplex imbricated slices were formed during D₁ and D₂. During the late stage of orogen-parallel extension, the thrust became overprinted by sinistral-strike-slip shear zones. These structures reflect

![Fig. 12. Major and trace element discrimination diagrams for volcanic rock sequences in the study area: (A) total alkalis–silica classification diagram of Le Maitre et al. (1989); (B) Fe(t)/MgO vs. SiO₂ classification diagram of Miyashiro (1974); (C) Ti–Zr diagram of Pearce and Cann (1973); (D) Nb–Y diagram of Pearce et al. (1984).](image-url)
a progressive ENE–WSW shortening, which eventually dominates deformation during late transpression. The ENE–WSW compression event was manifested by SW-dipping thrusts, which is usually related to the obduction of ophiolites (Abd El-Wahed, 2008). Moreover, the abundance of soft sediment deformation structures (such as pillows, slump folds) clearly suggests deposition of the El Beida marine facies in a tectonically active basin, for which a triggering mechanism other than earthquake is very unlikely (Mazumder, 2002). Bhattacharya and Bandyopadhyaya (1998) interpreted some of these structures as seismites. Finally, transgression in tectonically active areas is characterized by marine flooding and fine-grained clastic sediments, and may be cyclical in rift, pull-apart and foreland basins (Blair and Bilodeau, 1988).

Progressive shallowing of the palaeogeographic setting from an offshore deep marine to a terrestrial depositional setting, indicates that such gentle crustal uplift indeed took place prior to the commencement of subaerial volcanism (Mazumder, 2004). As a consequence, the continental freeboard (elevation of the continent above the mean sea level, cf. Eriksson, 1999) was greater already. This transition indicates relative sea-level fall and thus subaerial exposure of the depositional surface (i.e. higher continental freeboard). The subaerial rocks which were unconformably deposited upon the submarine rock facies (Fig. 3), developed as a distinct change in depositional style and sediment routing during the late phase of strike-slip shear zones accompanied by NE-trending linear zones, developing adjacent to normal faults and forming a graben basin (Fig. 2). Sinistral movement along shear zones and development of graben structure resulted in the formation of an intra-montane (or intra-arc) basin for the deposition and accumulation of the El Zarqa volcanics and siliciclastic sediments in an alluvial fan setting. The late extensional phase in the Fatira area culminated in the intrusion of felsic domes and mafic–felsic dykes (Gabal Fatira sequence) as well as younger granites. These dykes strike mainly NE–SW, orthogonal to the NW–thrusting and sinistral offsets in Neoproterozoic basement rocks.

The geochemical data for El Zarqa and Gabal Fatira sequence display a trend ranging from trachybasalt andesite through trachyandesite to trachydacite and rhyolite, perhaps reflecting fractionation of mafic mineral phases (Fig. 12A). These data also point to a calc–alkaline affinity with volcanic arc setting for the El Zarqa sequence (Fig. 12B–D); and to an alkaline affinity with anorogenic setting for the Gabal Fatira sequence (Fig. 12D). Overall, the latter two sequences incompatible elements are enriched relative to those of the Fatira El Beida samples. The Fatira El Zarqa sequence can be equated with the Dokhan volcanics and Hammamat sediments in the CED (Fig. 1C).

The gradual transition from an “orogenic” calc–alkaline magmatism to a continental intra-plate alkaline suite has been described in the ANS (Stern, 2008; Ali et al., 2009b). The late evolutionary stages of the northern ANS (650–545 Ma) document the growth and maturation of the continental crust from an orogen to a craton. During this period magmatism changed from calc-alkaline to alkaline and the tectonic settings in which these magmas were emplaced, have changed from collision to post-collision, to within-plate extension and finally to a stable platform setting.

7. Event stratigraphy and basin evolution of the Fatira area

Individual volcanic episodes may be thought of as relatively instantaneous events (i.e. typically of shorter duration than U/Pb SHRIMP errors) that may produce large thicknesses of volcanic rocks and volcaniclastic debris. These events separate erosional and depositional events that may be of long duration. Where precise geochronology is available, the volcanic events may be used to constrain the timing and duration of depositional and erosional events. Unconformity-bounded tectonic stages in the study area imply that subsidence, volcanism and sedimentation were not continuous (Fig. 3).

Within the Fatira area, the stages are uplift-subsidence cycles in which evidence of seven volcanic, depositional and erosional events are envisaged (Table 2). Volcanic, intrusive and depositional events broadly correlate with lithostratigraphic sequences in the CED (Fig. 1C), because each is compositionally distinct, and therefore mapped as a separate unit. The first event is marked by the eruption of submarine lava sheets, pillow basalts and pillow breccias. These mafic volcanics form the base of the succession in Fatira El Beida and are grouped together as tholeiitic basalts. Abundant hyaloclastics support subaqueous conditions, that along with the thickness of the basaltic materials indicate that event (1) was associated with transgression. Rocks of event (1) are unconformably overlain by turbiditic sedimentary deposits produced by event (2). Gradational contacts of these basalts and overlying volcanogenic sediments suggest that the mafic volcanic activity was apparently synchronised with the deposition of these turbiditic sediments. Consequently, event (1) and event (2) were approximately time-equivalent, albeit SHRIMP U–Pb detrital zircon supports the common presence of pre-Neoproterozoic reccylic siliciclastic rocks in the ANS (Stern et al., 2006; Kennedy et al., 2005; Ali et al., 2009a). Zircon provenances established that not all volcanioclastic sediments are resedimented syneruptive deposits, but most of them support a syneruptive origin. The upward-fining stratigraphy associated with event (2) is consistent with basin deepening/widening during transgression. The reported age for the pillowed tholeiitic basalts in the CED is ~750 Ma (Ali et al., 2009a). This age records the first tectono-metamorphic event (between ca. 736 ± 12 and 760 Ma), which has been considered to represent the age of primitive “arc”–like lavas terrane accretion and ophiolite obduction in the Neoproterozoic evolution of CED, Egypt (Stern, 2002; Ali et al., 2009a; Andresen et al., 2009). Intrusion of the older granites (Fig. 2) is the third event and their ages vary from 750 to 610 Ma (Stern and Hedge, 1985; Moghazy, 1999). The period of these granites marked the second tectono-metamorphic event (between ca. 620 and 650 Ma) and has been interpreted as a probable collisional stage during which the accreted juvenile terranes attached to the East Saharan Craton (Finger and Helmy, 1998). These rocks led to a shield-wide major uplift and substantial subaerial erosion.

The fourth event is marked by the eruption of the Fatira El Zarqa lava flows and volcaniclastics with angular unconformity, which suggests that the oceanic ophiolite basement was subaerial, immediately prior to formation of the Fatira El Zarga basin. The new SHRIMP U–Pb data established that andesitic volcanism in the type locality of the Dokhan volcanics occurred in the time span 593–602 Ma (Wildie and Youssef, 2000), which supports field evidence that the magmatic products of event (4) postdate the rock types of events (1–3). Evidence of a disconformity or low angle unconformity [event (5)] between the lava flows and alluvial fan sediments includes the recognition of epipelic materials. This unconformable surface records an erosional event and suggests that uplift and regression separated the Fatira El Beida and El Zarga basins. This surface represents a major unconformity event, which separates highly deformed older rock units and ~600 Ma Dokhan volcanics and Hammamat sediments in the Eastern Desert of Egypt (Sultan et al., 1988). The overlying volcaniclastic-dominated sedimentary rock facies was deposited during event (5). They show lateral thickness changes and occasionally preserves boulder-size clasts within conglomerates, indicating possible proximal deposition adjacent to palaeotopographic highs. A Rb/Sr date of 600–585 Ma has been obtained for these terrigeneous sediments (Willis et al., 1988). The geochronologic data indicate that volcanism [event (4)] and sediments deposition [event (5)] were contemporar-
neous and therefore, 600–585 Ma dates both magmatism and deposition. Emplacement of the felsic domes and bimodal sills–dykes are the sixth and seventh event; and the style of intrusion is constrained by a number of contact relationships. Stern and Hedge (1985) suggested that the minimum ages of rhyolitic rocks and dyke swarms are ca. 581 ± 7 and 558 ± 8 Ma, respectively, which lies within the reported age for the younger granites (610–550 Ma; Moghazi, 1999).

Based on a detailed structural study of the Eastern Desert of Egypt, Greiling et al. (1994) stated that collision ended at 615–600 Ma and extensional collapse occurred within the 600–575 Ma time span, followed by transpressional tectonism along major shear zones until 530 Ma. Accordingly, the events (4–7) and their emplacement ages (610–558 Ma), which postdate the main collisional stage (750–620 Ma), indicate that these rocks are post-orogenic.

8. Ancient and Phanerozoic arc-related basins: analogues the Fatira area

One of the most important principles of Precambrian sedimentation patterns is that Precambrian sedimentary lithologies contain structures, beside their inferred genesis, which have modern equivalents. The range of Precambrian basin types and their preserved basins-fills show no significant deviations from younger counterparts (Eriksson et al., 2004).

The volcanic facies of the Fatira El Beida sequence generally conform to the “standard” sequence of Dimroth et al. (1978), where massive parts of flows laterally and vertically become pillowed, overlain by pillow breccia, and capped by hyaloclastite or hyalotuffs. These rock assemblages resemble those of Phanerozoic seamounts or pillow lavas that develop on the ocean floor in mid-oceanic or back-arc rift zones. Mid-oceanic analogues include modern seamounts associated with the Mesozoic–Cenozoic seamounts in the Canary Islands and Cyprus (Staudigel and Schmincke, 1984). Further comparisons can be made to modern arc-back seamounts in the Mariana Trough and Lau Basin (Fryer, 1995), and to Cenozoic seamounts in southwest Japan and in the Japan Sea (Sohn, 1995).

A comparison with a modern, ensimatic back-arc basin such as the Lau Basin, south-west Pacific (e.g. Clift et al., 1995), shows some stratigraphic similarity. In the Lau back-arc basin, Clift et al. (1995) recorded basal volcanoclastics, dominantly dacite conglomerate and basalt hyaloclastite, abundant volcanogenic turbidite sandstone and fine pelagic sediments (nanofossil ooze, chalk and volcanic ash), overlying a basement of oceanic crust. The Neo-proterozoic El Beida sequence is best compared with the seamount at La Palma, Canary Islands, which is characterized by a deep water, basal, pillowed sequence intruded by dykes and silts, overlain by in situ hyaloclastite and pillow breccia and changing up-section into intermediate shallow water stratified, reworked deposits (Staudigel and Schmincke, 1984). Greater inferred heat production, sea floor spreading, and eruption rates during the Proterozoic relative to modern regimes produced more volcanic rocks with thicker tholeiitic basaltic sequences (Windley, 1995), suggesting that seamounts must have been prominent features on the Neo-proterozoic ocean floor.

Busby et al. (1998) proposed a three-stage model of arc-evolution based on studies of Baja California, Mexico. The model involves the development of a highly extensional intra-oceanic arc followed by mildly extensional fringing arc and finally a compressional continental arc system. An extensional setting for the Fatira El Beida basin is indicated by bimodal volcanism. There is evidence of an oceanic crust basement and, hence, an intra-oceanic arc model is considered likely for the El Beida basin. However, the phase 2 extensional fringing–arc system, that models the development of a back-arc basin over extended continental crust is consistent with the preserved Fatira El Beida sequence. The model of arc-development of Fackler-Adams and Busby (1998) involves intermediate to silicic explosive and effusive volcanism resulting in caldera-forming silicic ignimbrite eruption (stage I), followed by mafic, effusive hydroclastic rocks and dyke swarms (stage II). Voluminous, andesitic volcanics and debris, including volcanoclastics and epiclastics, may have formed in stage I (Fatira El Zarqa sequence) of Fackler-Adams and Busby’s (1998) model.

Hence, the Fatira area displays a temporal evolution from early back-arc tholeiitic mafic volcanics to late subduction-related magmas, which have evolved over a long time span in response to changing tectonic conditions. For example, in Fiji there was an evolution from early low-K arc tholeiites, to medium-K calc-alkaline lavas, to K-rich shoshonites, and finally to late stage back-arc alkaline rift-related basalts, which have erupted over a specific period

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<th>Table 2</th>
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<td>Suite</td>
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Neoproterozoic volcanic and siliciclastic strata in the Fatira area, provide an example of a subaerial and submarine depositional system. Detailed process-based facies and paleoenvironmental analysis led to the identification of 16 facies types, grouped under three rock sequences, which record paleoenvironmental settings ranging between continental alluvial fan and submarine shelf. The sequences investigated are: Fatira El Beida thololithic basalts and turbiditic volcanioclastics, which were deposited in an extensional back-arc basin; Fatira El Zarqa, which preserved the calc-alkaline volcanism and sedimentation in an alluvial fan basin; and the youngest domal alkali felsic volcanics and dyke-sill swarms of the Gabal Fatira sequence that formed in an anorogenic setting. The alluvial and submarine facies are arranged in a fining-upward sequence of intra-arc setting. The vertical and lateral facies distribution of the different sequences reveals the evolution, shallowing and restriction of the basins during the depositional history. These rock associations underwent three deformation phases, which coincide with the general evolutionary model for the ANS, that begins with NNW–WSW shortening, followed by ENE–WSW compression and subsequent deformation by the NNW–SSE-striking Najd Fault System. Stacking of lithoclasses is aggradational and is consistent with limited lateral migration of facies tracts in fault-confined basins. The gross vertical succession in these basins, from proximal up to distal is based on the relative dominance of erosional and depositional features. Unconformity and lithoclasses assemblages define seven events and three unconformity-bounded tectonic stages that record uplift-subsidence cycles in the study area. The stratigraphic architecture is typical of synorogenic basins but consistent with a strike-slip regime characterized by pulsed uplift-subsidence, steady-state basin lengthening and subsidence, and punctuating catastrophic subsidence. The strong similarity between Neoproterozoic volcanioclastic and modern arc-related volcanioclastic environments implies comparable tectonic settings and therefore comparable controlling tectonic factors.

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