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Cryogenian ophiolite tectonics and metallogeny of the Central Eastern Desert of Egypt

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The Central Eastern Desert (CED) is characterized by the widespread distribution of Neoproterozoic intra-oceanic island arc ophiolitic assemblages. The ophiolitic units have both back-arc and forearc geochemical signatures. The forearc ophiolitic units lie to the west of the back-arc related ones, indicating formation of an intra-oceanic island arc system above an east-dipping subducted slab (present coordinates). Following final accretion of the Neoproterozoic island arc into the western Saharan Metacraton, cordilleran margin magmatism started above a new W-dipping subduction zone due to a plate polarity reversal. We identify two belts in the CED representing ancient arc–forearc and arc–back-arc assemblages. The western arc–forearc belt is delineated by major serpentinite bodies running ∼NNW–SSE, marking a suture zone. Ophiolitic units in the back-arc belt to the east show an increase in the subduction geochemical signature from north to south, culminating in the occurrence of bimodal volcanic rocks farther south. This progression in subduction magmatism resulted from diachronous opening of a back-arc basin from north to south, with a bimodal volcanic arc evolving farther to the south. The intra-oceanic island arc units in the CED include coeval Algoma-type banded iron formations (BIFs) and volcanogenic massive sulphide (VMS) deposits. Formation of the BIFs was related to opening of an ocean basin to the north, whereas development of the VMS was related to rifting of the island arc in the south. Gold occurs as vein-type mineral deposits, concentrated along the NNW–SSE arc–forearc belt. The formation of these vein-type gold ore bodies was controlled by the circulation of hydrothermal fluids through serpentinites that resulted in Au mobilization, as constrained by the close spatial association of auriferous quartz veins with serpentinites along the western arc–forearc belt.

Keywords: ophiolites; forearc; back-arc; Neoproterozoic; volcanogenic massive sulphides; banded iron formations; gold–quartz veins; Eastern Desert

Introduction

The Eastern Desert of Egypt constitutes the northwestern segment of the Arabian–Nubian Shield (ANS) (Figure 1A), which marks the northern extension of the East African Orogen. The ANS is a major orogenic system that formed near the end of the Proterozoic as a result of the closure of the Mozambique Ocean and the collision between East and West Gondwana (Stern 1994; Dilek and Ahmed 2003; Johnson *et al.* 2003; Kusky *et al.* 2003). It is one of the best examples of extensive crustal growth during the Neoproterozoic era (Stern 2008; Johnson *et al.* 2011). The ANS formed through amalgamation or accretion of intra-oceanic island arcs and collision of these arcs with a continental margin (Stoeser and Camp 1985; Abdelsalam and Stern 1996); and these accretion processes led to the formation of well-defined arc–arc and continent–arc suture zones (Stern 1994). The suture zones are marked by widespread ophiolitic rocks (Dilek and Ahmed 2003; Stern 2005, 2008).

Cryogenian ophiolitic assemblages are widely distributed in the Eastern Desert, especially in the central part (Stern and Hedge 1985) (Figure 1B). However, unlike the well-defined elongated sutures in other parts of the ANS, the Central Eastern Desert (CED) suture is more complicated due to its pervasive deformation (Abdelsalam and Stern 1996). Another obstacle to understanding the nature of the suture in the CED is the debate over the nature of the ophiolitic units and generally over the tectonic evolution of the CED. Although the ophiolitic assemblages in the CED show supra-subduction zone (SSZ) geochemical signatures, their specific tectonic setting of formation (i.e. forearc or back-arc tectonic settings) is not well defined (Stern *et al.* 2004; Azer and Stern 2007; Abd El-Rahman *et al*. 2009a, 2009b; Ali *et al.* 2009; Farahat 2010).

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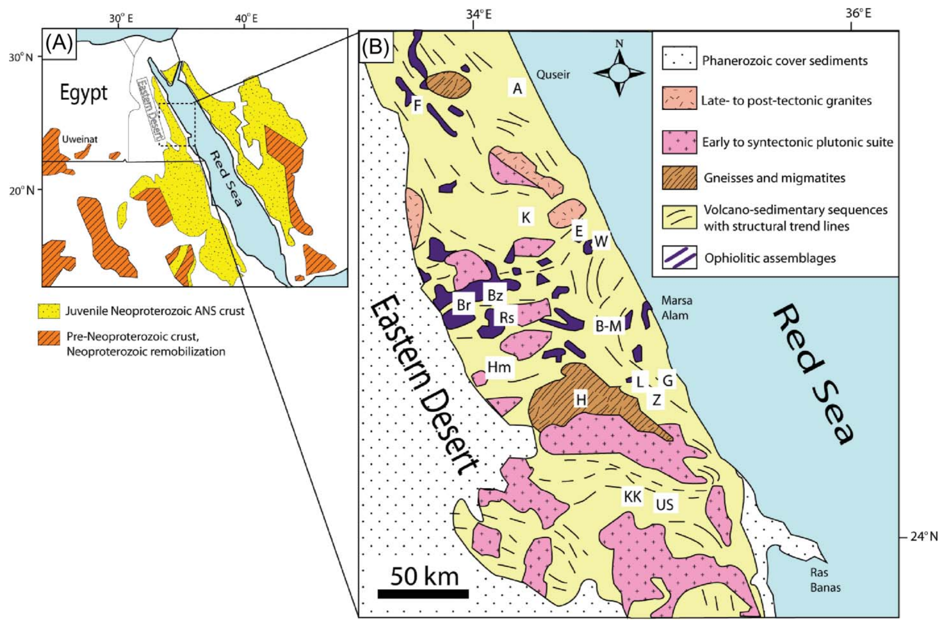

Figure 1. (A) Map of the Arabian–Nubian Shield (modified after Ali *et al.* 2009). (B) Simplified geological map of the Central Eastern Desert (CED) of Egypt showing the locations of Ambagui (A), Beririq (B), Barramiya (Br), Bizah (Bz), Essel (E), Fawakhir (F), Ghadir (G), Hafafit (H), Hamash (Hm), Kareim (K), Kolet Um Kharit (KK), El-Lawi (L), Mubarak (M), Ras Salatit (Rs), Um Samuiki (US), Wizer (W) and El-Zarka (Z) areas that will be mentioned in the text. (modified after Kröner *et al.* 1987).

The intra-oceanic island arc assemblages of the CED are characterized by the banded iron formation (BIF) and the volcanogenic massive sulphides (VMSs) (Garson and Shalaby 1976; Hussein and El Sharkawi 1990; Botros 2003). In addition and similar to the rest of the ANS (Goldfarb *et al.* 2001), the CED is characterized by the widespread distribution of gold deposits (Kochin and Bassyni 1968; Botros 2002, 2004). The gold mining sites of the CED date back to the Pharaonic periods (Klemm *et al.* 2001; Gabr *et al.* 2010). There have been several attempts to make a connection between the tectonic evolution and the metallogeny of the Eastern Desert, including the central segment (Garson and Shalaby 1976; Botros 2002, 2004). However, these studies were hindered by the lack of a well-constrained tectonic model for the evolution of the Eastern Desert. The importance of the spatial distribution of the ancient mining sites, especially in the CED, has also not been carefully explored until now.

In this article, we synthesize the tectonic evolution of the CED during the cryogenian time based on the available geochemical data from several ophiolitic and island arc assemblages. In the first part of this article, we evaluate the spatial variation of the extant geochemical data and delineate the suture zones in the CED. Then, we discuss the distribution of BIFs, VMSs, and gold deposits associated with the ophiolitic and island arc assemblages within the framework of the inferred tectonic model. The geochemical and tectonic evolution of the Proterozoic ophiolites in the ANS may have important implications for the Archean greenstone belts and ophiolites (Dilek and Polat 2008).

Rock assemblages of the CED

The rock assemblages of the CED are commonly grouped under two major tectonostratigraphic units (El-Gaby *et al.* 1988; Abdeen and Greiling 2005; Abd El-Wahed and Kamh 2010). The structurally lower one, the infrastructure, is composed of gneisses and migmatites that crop out in dome structures, such as the Meatiq, Sibai, and Hafafit domes (Figure 1B). The overlying unit, the suprastructure, includes the Neoproterozoic ophiolite complexes and island arc-related metavolcanic and metasedimentary rocks. The suprastructure is also known as the Pan-African Nappe Complex (Bregar *et al.* 2002).

There is a controversy over whether the infrastructure gneisses and migmatites represent a pre-Neoproterozoic crust or exhumed Neoproterozoic rocks. The pre-Neoproterozoic rocks occurred in the Uweinat area in the southwestern corner of Egypt (Schandelmeier *et al.* 1988) (Figure 1A) and constitute the Saharan Metacraton, Abdelsalam *et al.* (2002). El-Gaby *et al.* (1984, 1988) and Khudeir *et al.* (2008) have suggested that this older continental crust extends beneath the Pan-African Nappe Complex in the Eastern Desert. The gneissic domes in the CED have been interpreted as the deeper levels of the juvenile Neoproterozoic crust with no evidence for the involvement of older continental crust (Sturchio *et al.* 1983; Kröner *et al.* 1994; Bregar *et al.* 2002; Andresen *et al.* 2009; Liégeois and Stern 2010; Lundmark *et al.* 2012).

The ophiolitic and island arc assemblages cover a large area of the CED (Figure 1B). These rocks are regionally metamorphosed at lower grades, mainly greenschist facies, than the gneisses of the infrastructure. The ophiolite complexes occur as dismembered blocks and fragments that include serpentinite, metagabbro, and pillow metabasalt rocks within the mélange of the CED (Shackleton *et al.* 1980). The best preserved exposures of the Neoproterozoic ophiolites occurred in the Wadi Ghadir (El-Sharkawi and El Bayoumi 1979; El Bayoumi 1980) and Fawakhir areas (Nasseef *et al.* 1980) (Figure 3). The ophiolitic ultramafic blocks in the CED are almost completely serpentinized and are typically altered to talc-carbonate bodies along shear zones (El-Sharkawi and El Bayoumi 1979; Nasseef *et al.* 1980; Azer and Stern 2007; Farahat 2008; Ali-Bik *et al.* 2012). Both layered and isotropic metagabbros are common in the CED (El-Sharkawi and El Bayoumi 1979; Abd El-Rahman *et al.* 2009a, 2009b), and are generally associated with diorite, plagiogranite, and appinite veins as common in the Phanerozoic Tethyan ophiolites (Dilek and Eddy 1992; Dilek and Thy 2006). Sheeted dike complexes, which are characteristic features of Penrose-type ophiolites (Anon 1972; Dilek and Furnes 2011; Kusky *et al.* 2011), are only locally preserved in the CED, whereas pillowed metabasalts are widespread (El-Sharakawi and El Bayoumi 1979; Nasseef *et al.* 1980; El Bayoumi 1983).

In addition to the ophiolitic rocks, the intra-oceanic island arc assemblages in the CED are represented by weakly metamorphosed tholeiitic to calc-alkaline volcanic rocks and associated volcaniclastic rocks (Kröner *et al.* 1987). The island arc assemblages range from basic to felsic in composition. The tuffs and volcaniclastic graywackes are associated with both BIFs and massive sulphides, indicating their subaqueous nature (Sims and James 1984). Stern (1981) and Khalil (1997) classified the metavolcanic rocks in the CED into Older and Younger Metavolcanics that are equivalents of the ophiolitic, ocean-related, volcanic rocks and the island arc volcanic rocks, respectively. Ali *et al.* (2009) criticized such classification of the metavolcanic rocks, and considered that both the ophiolitic and island arc assemblages in the CED constitute an artifact of one (∼750 Ma) crust-forming event. The ages of the well-preserved ophiolitic rocks in Wadi Ghadir (746 ± 19 Ma, Kröner *et al.* 1992) and in Fawakhir (736.5 ± 1.2 Ma, Andresen *et al.* 2009) in the CED are compatible with the ∼750 Ma crustforming event proposed by Ali *et al.* (2009) (Figure 2).

The later stage of the crustal evolution of the CED is characterized by the eruption of the Dokhan volcanic suite (El-Gaby *et al.* 1988). The geochemistry of these volcanic rocks indicates medium-K to high-K calc-alkaline affinity and their tectonic setting is a matter of controversy. Abdel Rahman (1996) noted the geochemical similarity of the Dokhan volcanic suite to Andean-type magmatic units and suggested a continental arc setting for their origin. Stern *et al.* (1984) suggested a post-collisional origin in an extensional setting for the formation of the Dokhan volcanic suite. Resseltar and Monard (1983) and Mohamed *et al.* (2000) attributed the formation of the Dokhan Volcanics to a transitional stage between subduction and extension. Recently, Eliwa *et al.* (2006) recorded in the Dokhan

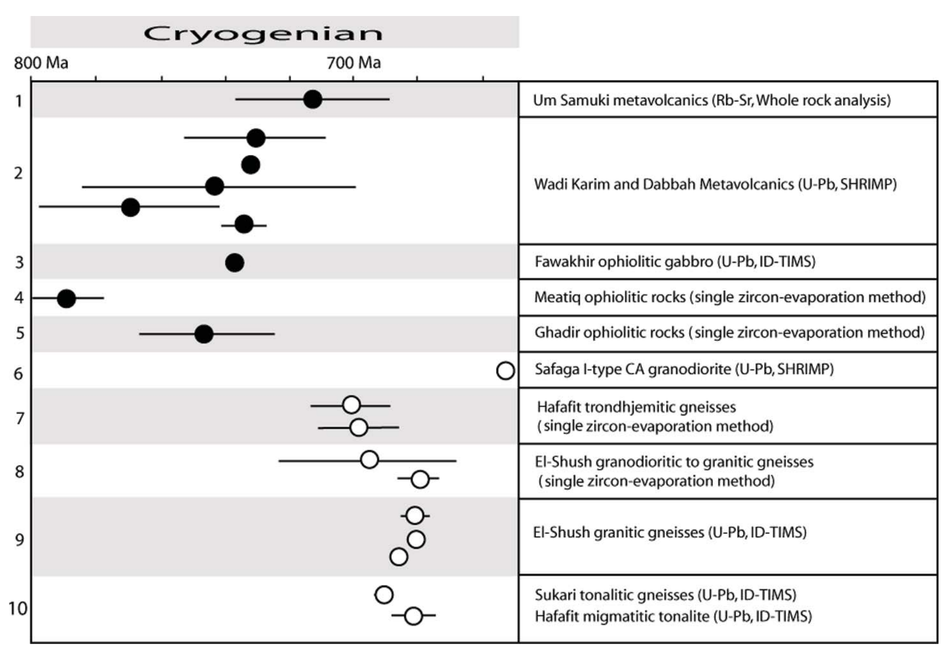

Figure 2. Compilation of ages of the intra-oceanic island arc assemblage (black circles) and later granitoids of Cordilleran stage (white circles) during Cryogenian time. Sources: 1 – Stern *et al.* (1991), 2 – Ali *et al.* (2009), 3 – Andresen *et al.* (2009), 4 – Loizenbauer *et al.* (2001), 5 – Kröner *et al.* (1992), 6 – Moussa *et al.* (2008), 7 – Kröner *et al.* (1994), 8 – Bregar *et al.* (2002), 9 – Augland *et al.* (2012), 10 – Lundmark *et al.* (2012).

Volcanics a unique occurrence of adakitic rocks, which they interpreted as a product of the subduction of an oceanic ridge before collision and later melting of the hot oceanic slab beneath the central part of the CED. The volcanic activity that produced the Dokhan suite was mainly subaerial and associated with the formation of molasse-type Hammamat sedimentary rocks that were deposited in non-marine, alluvial fan/river environments (Grothaus *et al.* 1979). Most of the Hammamt fragments were derived from the Dokhan Volcanics (Abd El-Rahman *et al.* 2010, and references therein).

These crustal rocks were intruded by a series of granitoid plutons, which have been classified as: subduction-related, suture-related, and intraplate anorogenic granites (El-Gaby 1975; Hussein *et al.* 1982; Bentor, 1985). Recent geochronological data on the granitic rocks and gneisses of granitic composition from the CED indicated the predominance of two granitic magmatic pulses. The earlier one resulted in the formation of calc-alkaline, synorogenic granitic rocks, developed at convergent plate boundaries (Figure 2), whereas the later one produced more alkaline varieties, characterizing within-plate intrusions (Kröner *et al.* 1994; Bregar *et al.* 2002; Shalaby *et al.* 2005; Moussa *et al.* 2008; Andresen *et al.* 2009; Pease *et al.* 2010; Augland *et al.* 2012; Lundmark *et al.* 2012).

Tectonic evolution of the CED suture

We examine the tectonic evolution of the CED during the Cryogenian time, which is characterized mainly by the formation of an earlier intra-oceanic island arc system and a later post-accretion system. We summarize these two systems below.

Intra-oceanic island arc system

Since the metamorphosed mafic-ultramafic associations of the CED were identified as remnants of the Neoproterozoic oceanic crust (El-Sharakawi and El Bayoumi 1979; El Bayoumi 1980; Shackleton *et al.* 1980), several models have been proposed to illustrate the evolution of the CED within the framework of the plate tectonic paradigm. Ries *et al.* (1983) suggested a SE-dipping subduction and the closure of a basin between the intra-oceanic island arc and a continental margin. In their model, the closure of the ocean basin led to NW-directed emplacement of the ophiolitic assemblages and finally resulted in the collision of the intra-oceanic island arc with a passive continental margin.

El Ramly *et al.* (1984) and El Bayoumi and Greiling (1984) proposed a SW-dipping subduction that resulted in the formation of intra-oceanic island arc system, which was separated from a passive continental margin by a back-arc basin. Subsequently, the marginal basin and the island arc system were imbricated, from east to the west, and were then thrust onto the continental margin. As this thrust tectonics evolved, a new convergent margin was established due to a post-collisional polarity reversal (Kröner 1985). Kröner *et al.* (1987) supported the inferred westward subduction model over the eastward subduction model of Ries *et al.* (1983) based on two points. The first is that the eastward subduction model cannot account for the supra-subduction origin, proposed by El Bayoumi (1983) and Kröner (1985), for the ophiolitic rocks of the CED. The second is the difficulty to create active continental margin magmatism from mantle reservoirs below a passive continental margin.

The debate over the subduction direction, above which the Neoproterozoic juvenile rocks were formed, continued with the work by other researchers (Ragab 1987; Ragab and El-Alfy 1996; Loizenbauer *et al.* 2001; El-Sayed *et al.* 2007), who suggested a generally eastward subduction for the evolution of the CED. However, some others (El-Gaby *et al.* 1988; Abd El-Naby and Frisch 2006; Abd El-Naby *et al.* 2008; El Bahariya 2008) proposed a generally west-dipping subduction zone for the formation of the juvenile island arc and ophiolitic rocks in the CED. Unfortunately, most of the above-mentioned arguments over the direction of subduction depend on the structural architecture of the CED, which is highly complicated. Shackleton (1994) pointed out the difficulty in deciphering the subduction direction without geochemical evidence on the arc–magmatic polarity. In a modern intra-oceanic arc system, the magma chemistry shows a systematic variation across the arc (Bloomer *et al.* 1995; Hochstaedter *et al.* 2001). A transect from the back-arc to the forearc setting shows that volcanic rocks become more depleted in the incompatible trace elements and more affected by slab-derived fluids (Hochstaedter *et al.* 2000; Dilek *et al.* 2008). Such correlation between the variation of the geochemistry and subduction polarity occurs regardless of the eruption age (Taylor *et al.* 1992).

Both the Ti-V diagram of Shervais (1982) and Cr-Y diagram of Pearce (1982) can be used to deduce the variability in the subduction signature of ophiolite rocks. We utilize tracing of the spatial distribution of the ophiolitic rocks in the CED along with the variations in their geochemical signature in order to infer the subduction polarity (Figures 1B and 3; Dilek *et al.* 2008). The Fawakhir ophiolite complex has a forearc trace element signature (Abd El-Rahman *et al.* 2009a), whereas the Wadi Ghadir ophiolite exhibits a back-arc trace element affinity (Farahat *et al.* 2004; Abd El-Rahman *et al.* 2009b; Basta *et al.* 2011). To the east of the Fawakhir ophiolite, the pillow lavas of the Wadi Ambagui area are more enriched in incompatible trace elements and show a back-arc geochemical signature (El-Mahallawi 1994; Abd El-Rahman *et al.* 2009a) (Figure 1B). The same scenario can be proposed for the south traverse extending from Wadi Ghadir to the east through Hafafit and towards Hamash and Ras Salatit in the west. The Wadi Ghadir ophiolite with a back-arc geochemical signature is located to the east of the Hafafit culmination (Abd El-Naby *et al.* 2008). The geochemical data on the Hafafit amphibolites from Abd El-Naby and Frisch (2006) were reinterpreted as forearc by Abd El-Rahman *et al.* (2009b) (Figure 3). The ophiolitic metagabbro of the Hamash area shows subduction trace element geochemical signatures reminiscent of a forearc association rather than a back-arc unit (Said 2010). Gahlan *et al.* (2012) measured the CPX trace element contents of the wehrlite rocks of the Ras Salatit ophiolite and inferred that these ultramafic rocks were in equilibrium with melt, the composition of which is largely similar to those of the lavas from the Izu-bonin forearc setting. The pillow lavas of Wadi Beririq, half the distance between Wadi Ghadir to the south and Wadi Ambagui to the north, show a back-arc geochemical signature (Farahat *et al.* 2004). This consistency of having ophiolitic crustal rocks with a forearc geochemical signature to the west and a back-arc geochemical affinity to the east may indicate the evolution of the Neoproterozoic intra-oceanic island arc assemblage of the CED above an E-dipping subduction zone (Figure 4).

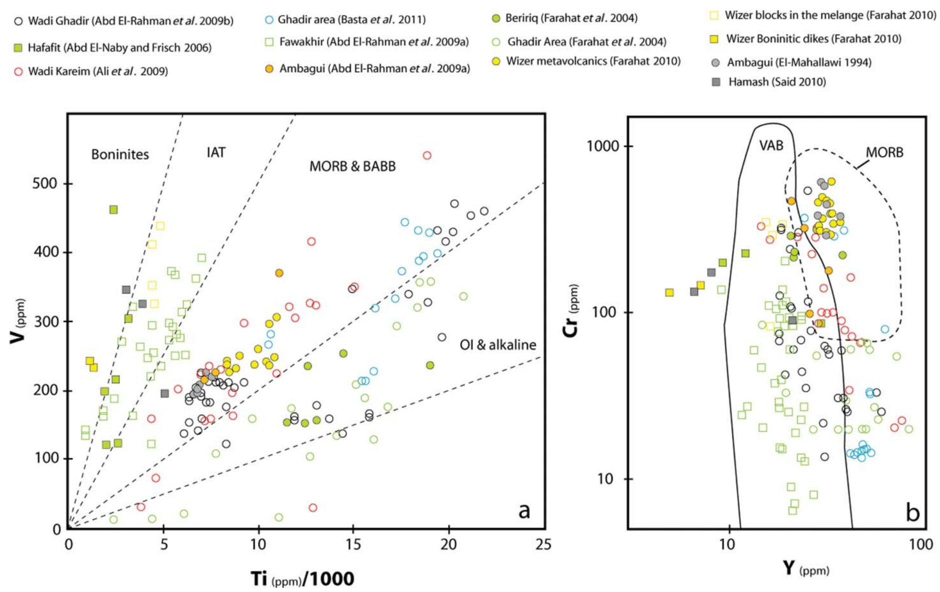

Figure 3. Plot of the ophiolitic rocks of the CED on (A) the Ti–V diagram of Shervais (1982) and (B) the Cr–Y diagram of Pearce (1982).

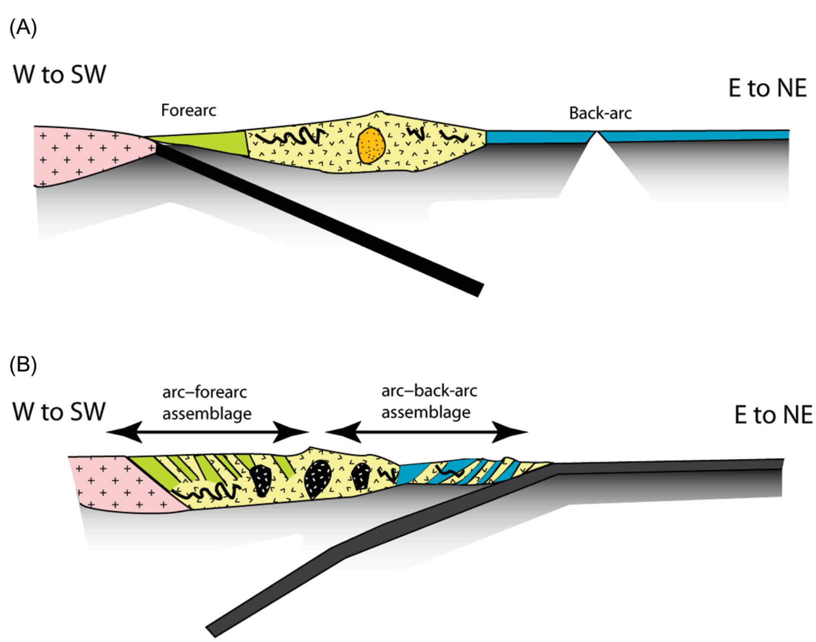

Figure 4. Tectonic evolution of the CED first in an intra-oceanic island arc stage (A) and then in the Cordilleran margin stage (B).

Ahmed *et al.* (2001) studied the chemistry of chrome spinel from the chromitite and dunite of the CED, and considered the high Cr# of the primary chrome-spinel as an indicator for a SSZ tectonic setting. The geochemical data from the ophiolitic mantle rocks in the CED vary systematically along an E–W traverse. The chrome-spinel of the western (Wadi Bezah and El-Barramiya, Figure 1B) chromitite and dunite are characterized by higher Cr# (0.68–0.8 and 0.69–0.78, respectively) than the chrome-spinel of the eastern (Wadi El-Lawi and Wadi El-Zarka) chromitite and dunite (0.55–0.71 and 0.54–0.69, respectively). Azer and Stern (2007) and Khalil and Azer (2007) considered the CED serpentinites as forearc mantle fragments based on their depleted nature. This conclusion raises the question of how the depleted forearc mantle rocks can be spatially associated with crustal rocks having a back-arc geochemical signature in the eastern part of the CED.

Recently, Farahat (2010) proposed the formation of the intra-oceanic island arc system above a SE-dipping subduction zone. His principal evidence was the exclusive occurrence of boninitic dikes in the Wizer ophiolitic rocks to the northeast of the CED (Figure 1B), indicating their development in a protoarc–forearc setting. The formation of boninitic rocks needs a hydrous fluid supply to a hot, shallow refractory mantle (Crawford *et al.* 1989; Dilek and Furnes 2011). Although such conditions are dominant in the protoarc–forearc setting (Casey and Dewey 1984; Stern and Bloomer 1992; Kim and Jacobi 2002; Dilek *et al.* 2007, 2008), they may also be encountered in back-arc tectonic settings (Ewart *et al.* 1994). The back-arc boninites can be produced by the ascent of MORB-source diapirs during the incipient rifting stage of an island arc (Crawford *et al.* 1981; Meffre *et al.* 1996) or along ridge–fracture zone intersections (Edward 1995).

Boninites in the Wizer area occur as minor dikes intruding the ophiolitic MORB-like metabasaltic rocks, which are thrust over the island-arc metavolcanic blocks in the mélange matrix (Farahat 2010). Since the boninitic dikes intrude the back-arc oceanic crust (the ophiolitic MORB-like metabasaltic rocks of Wizer area, Figure 3), they are younger in age than the main intra-oceanic island arc system of the CED. Thus, they might have been developed as a result of an intra-oceanic closure of the back-arc basin. The consumption of the back-arc basin oceanic crust likely produced an incipient island arc and permitted hydrous fluids from the subducted oceanic crust to interact with the hot shallow depleted mantle (cf. Figure 8 of Wang *et al.* 2002). This process led to the formation of boninitic magmas intruding the arc–back-arc assemblage at Wadi Wizer. This mechanism may also explain the depleted nature of some of the upper mantle peridotites associated with the typical back-arc crustal rocks of the Wadi Ghadir area.

The post-accretion system

With the consumption of the easterly dipping subducted plate, the intra-oceanic arc system approached and finally collided with a passive continental margin (Saharan Metacraton) to the west. Collision of an intra-oceanic volcanic arc with a continental margin is one of the most important accretion processes that result in continental crustal growth (Brown and Ryan 2011). An arc-passive margin collision is also a major mechanism for the initiation of a new active continental margin (Casey and Dewey 1984). The initiation of an active continental margin after an arc-passive margin collision may most likely be achieved through reversal in the subduction polarity (Clift *et al.* 2003).

The best preserved ophiolitic assemblages in the CED, the Wadi Ghadir and Fawakhir ophiolite complexes, are intruded by calc-alkaline dikes with an Andean margin geochemical affinity (Abd El-Rahman *et al.* 2009a, 2009b). These dikes could be part of the Cordilleran stage magmatism in the CED as proposed by El-Gaby *et al.* (1988). This stage was manifested in the production of calc-alkaline granitic rocks above a W-dipping subduction zone, following the arc-continental collision (Figure 4). This inferred reversal of the subduction polarity resolves the problem raised by Kröner *et al.* (1987) over the formation of the ophiolites in the CED above an E-dipping subduction zone to the east of a passive margin, while being intruded by Andean-type magmatism.

The subduction polarity reversal might have caused the collapse of some segments of the back-arc oceanic ridge, leading to ridge subduction and to the formation of adakitic magmas that produced the Dokhan Volcanics in the CED before the final collision of East and West Gondwana (Eliwa *et al.* 2006). Later, with the development of the Andean margin, the back-arc crust would have been incorporated into the newly developed accretionary prism (Figure 4). Thus, we can divide the intra-oceanic island arc assemblage in the CED into an arc–forearc assemblage in the west and an arc–back-arc assemblage in the east (Figure 4).

Arc–forearc assemblage

Along the E–W traverses running through the CED, we see the ophiolitic rocks with a forearc geochemical affinity in the west relative to the back-arc ophiolitic rocks in both Hafafit and Fawakhir areas. Neoproterozoic ophiolitic serpentinites are widespread in the CED. However, the large serpentinite bodies are concentrated on the western side of the CED extending in a NNW–SSE direction (Figure 5). Serpentinite bodies are found in the forearc regions of intra-oceanic island arc systems, such as the Izu–Bonin–Mariana system (Fryer 1992). Fryer *et al.* (1995) noticed the similarity between those serpentine seamounts

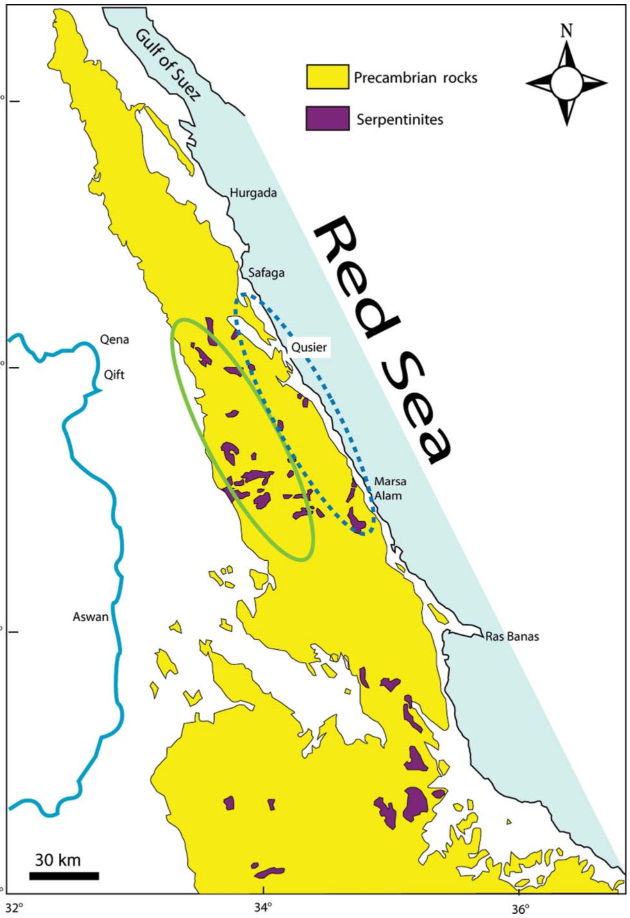

Figure 5. Distribution of serpentinite in the Eastern Desert of Egypt (from Said (2006)). The green ellipse represents the arc–forearc association, and the dashed purple ellipse represents the arc–back-arc association in the CED.

that formed in forearc regions and the serpentinite bodies preserved in accreted forearc–arc terranes during continental growth. Akaad (1997 and references therein) described diverse types of macroscopic fabrics in serpentinites ranging from block-in-matrix fabrics to flow-banded scaly serpentinites, which are localized within the serpentinite bodies. Some of these fabrics are similar to what were described by Fryer *et al.* (1995) in the serpentinite seamounts in a forearc environment. Moreover, the ophiolitic serpentinites in the CED show a SSZ geochemical signature (El-Sayed *et al.* 1999; Ahmed *et al.* 2001; El Bahariya and Arai 2003; Azer and Khalil 2005; Farahat 2008; Abu El Ela and Farahat 2010). Azer and Stern (2007) inferred that ophiolitic serpentinites in the CED originated in a forearc setting. The spatial distribution of the large ophiolitic serpentinite masses of a forearc mantle origin is consistent with the distribution of the ophiolitic crustal rocks that have an inferred forearc affinity (Figure 5). The distribution of the large ophiolitic serpentinite bodies in the CED can be traced as a line defining a probable suture zone along which the forearc front of an intra-oceanic arc system collided with older continental crust (Abd El-Rahman 2010).

Arc–back-arc assemblage

Ophiolites with a back-arc origin have been documented in many areas of the CED by Farahat *et al.* (2004), El Bahariya (2008), Abd El-Rahman *et al.* (2009a, 2009b), and Ali *et al.* (2009) (Figure 3). All these occurrences are located east of the inferred arc–forearc belt forming a NW–SE-tending arc–back-arc belt (Figures 1B and 5). As Th is known to be a non-conservative element (contributed from the subducted slab to the mantle wedge) relative to Nb (Pearce and Peate 1995), primitive mantle-normalized Th/Nb ratios (Th/Nb_{pm}) can be considered as a proxy for the subduction signature. Along the back-arc ophiolitic belt, variations of Th/Nb_{pm} ratios were detected from the north to the south. In the northern part of the belt, the Ambagui ophiolitic pillow lavas have less-pronounced trace element subduction signatures ($Th/Nb_{pm} = 2.87$; Abd El-Rahman *et al.* 2009a) than those of the ophiolitic crustal rocks at Wadi Ghadir in the south ($Th/Nb_{pm} = 4.26$; Abd El-Rahman *et al.* 2009b). Such distinction in the trace element subduction signature was also recorded by Farahat *et al.* (2004) between the ophiolitic pillow lavas in the Ghadir area to the south ($Th/Nb_{pm} = 2.94$) and the pillow lavas of the Wadi Beririq area to the north ($Th/Nb_{pm} = 1.07$). Moreover, the ophiolitic ultramafic rocks along the arc–back-arc belt exhibit a MORB affinity to the north in the Essel area and an increase in the subduction signature in the south in the Mubarak area (El Bahariya 2008). The Wadi Kareim back-arc rocks have a variable Th/Nb_{pm} (1.41–13.61) and the highest Th/Nb_{pm} average of the recently studied back-arc rocks (5.44) (Ali *et al.* 2009). These high values might be attributed to the spatial–temporal association between the back-arc rocks and the typical arc rocks in the Wadi Kareim area.

Considering an increase in the subduction signature towards the south, El Bahariya (2008) proposed a subduction direction towards the NW even though the subduction signature in the rocks becomes weaker generally towards the north. His inferred NW-directed subduction does not explain the E–W variation in the subduction signature as described by Abd El-Rahman *et al.* (2009a, 2009b) or the spatial distribution of the back-arc geochemical affinities of the ophiolitic rocks in the CED (Figures 1B and 3). However, the dilution in the subduction trace element signature from the south to the north can be better explained in the framework of an eastward subduction direction by having a tectonic set-up similar to that of the Lau Basin (Figure 6). The longitudinal transect of the Lau back-arc basin exhibits a southward-propagating system from a true oceanic spreading setting in the north to a rifted arc system in the south (Parson

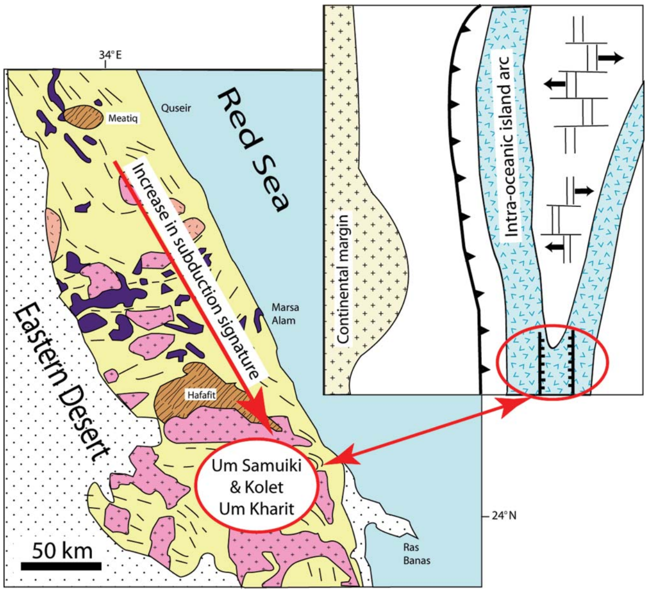

Figure 6. Distribution map of the ophiolitic assemblages in the CED (after Kröner *et al.* 1987) showing the location of the bimodal volcanic units in Um Samuki and Kolet um Kharit relative to the trend of increasing subduction signature in the arc–back-arc belt (left) and its palaeogeographic location during the intra-oceanic island arc stage (right).

and Wright 1996; Dilek *et al.* 1997). The same scenario is proposed for the evolution of the eastern ophiolitic back-arc belt in the CED where the spreading happened above an almost easterly dipping subduction zone as the back-arc spreading propagated southward. Thus, the intensity of the subduction zone geochemical signature of the back-arc ophiolitic rocks in the CED is expected to increase southward as the southern segment of the back-arc spreading ridge was closer to the subduction zone relative to the northern segment.

The southward-propagating arc rifting model for the evolution of the Neoproterozoic back-arc ophiolitic belt is supported by the occurrence of bimodal volcanic rocks farther to the south in the Um Samiuki and Kolet Um Kharit areas (Figure 6). These rocks are known as the Shadli Metavolcanics (after El Ramly 1972). The tectonic setting of these rocks is a matter of debate. Shukri and Mansour (1980), Hafez and Shalaby (1983), and Khudeir *et al.* (1988) identified the bimodal volcanic suite in the Um Samuki area as an island arc sequence. However, Stern *et al.* (1991) proposed a continental rift origin for these bimodal volcanic rocks. Farahat (2006) recognized both the subduction zone and the within-plate geochemical signatures from the bimodal volcanic suite of Kolet Um Kharit. Selim (2002) divided the volcanic rocks of the Um Samuki area into five units. The stratigraphically lower unit has the highest andesite content. Moving upward in the volcanic section, the andesite occurrence decreases and the bimodality becomes the main character for the younger volcanic units. The bimodal nature of the suite and its subduction zone geochemical signature collectively suggest the formation of these rocks in a rifted oceanic island arc setting. This interpretation is compatible with the southward-propagating arc rifting model for the early stages of the evolution of the CED (Figure 6).

Mineralization related to the intra-oceanic island arc stage

The mineralization types associated with the intra-oceanic island arc stage are BIFs and VMSs. The BIFs occur as well-defined stratigraphic units within the layered volcanic and volcaniclastic rocks of the arc–back-arc assemblages located within the western side of the CED (Figure 7A). Ali *et al.* (2009, 2010) demonstrated the spatial association between the ophiolitic rocks, the diamictite and the BIFs in the arc–back-arc assemblages of the Wadi Kareim area in the CED. Due to the temporal and spatial association with the volcanic and volcaniclastic rocks, Sims and James (1984) classified the BIFs in the CED as Algoma-type. The BIFs were deformed and regionally metamorphosed along with the associated rocks into greenschist facies mineral assemblage. The most dominant iron minerals are hematite

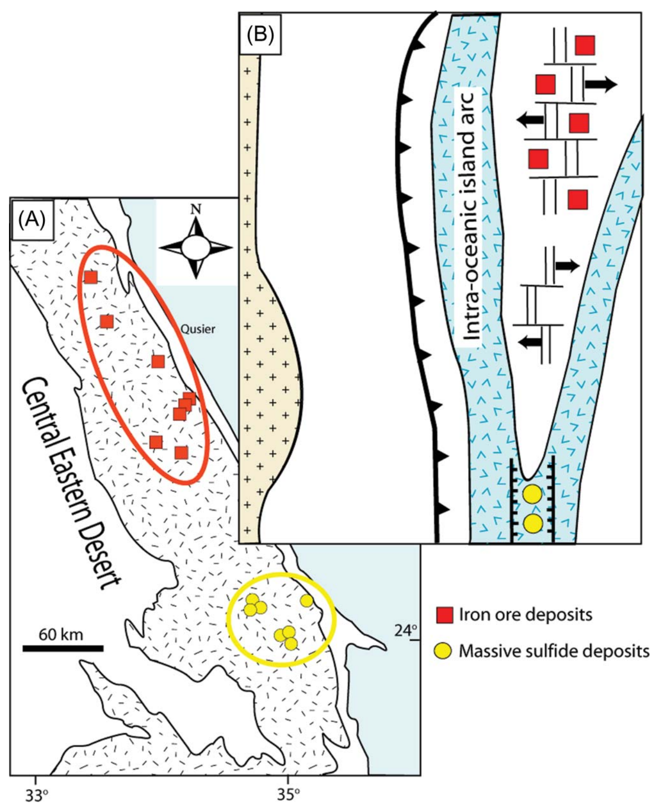

Figure 7. (A) Distribution of the iron formations and massive sulphide deposits in the Eastern Desert (from Botros 2004). (B) Palaeogeographic sketch showing the tectonic setting of auriferous iron deposits and massive sulphide deposits in relation to the evolution of the intra-oceanic island arc stage in the CED.

and magnetite, alternating with silica-rich bands, and thus oxide facies is the dominant facies of the BIFs (Sims and James 1984). These authors have suggested that the BIFs may represent chemical precipitates of exhalative origin that accumulated during periods of quiescence of the submarine volcanic activity. Stern *et al.* (2006) interpreted the deposition of the ANS BIFs in the framework of the 'Snowball Earth Hypothesis'. According to this hypothesis, Fe derived from seafloor hydrothermal vents remained in solution as Fe^{2+} as the basins were isolated from the atmosphere with ice, and then melting of ice re-oxygenated the sea water and iron oxidized to the insoluble Fe^{3+} state.

The VMS deposits are clustered to the south of the BIF suites (Figure 7A). The sulphide deposits are confined to specific felsic horizons of the bimodal Shadli Volcanic Group (Searle *et al.* 1976; El Aref *et al.* 1985). The deposits are characterized by sharp upper contacts and a keel zone with pervasive alteration minerals, mainly chlorite and talc with minor carbonate and tremolite (Abdel Kader and Shalaby 1982; Salem *et al.* 1999; Botros 2003). The sulphide deposits are polymetallic and include disseminated and massive chalcopyrite, sphalerite, pyrite, and galena that are associated with gold and silver anomalies (Searle *et al.* 1976; Takla *et al.* 1998; Botros 2003). The sulphides were mobilized during regional metamorphism and later deformational events (Takla *et al.* 1998). In the Um Samuiki area, the volcanic products show a temporal evolution from a basaltic–andesitic–rhyolitic stage to bimodal stages, and the massive sulphides are confined to the lower bimodal stage (Selim 2002). According to Barrett and Maclean (1999), the VMSs in bimodal sequences are less likely to be developed in mature back-arc basins than in rifted island arcs. The spatial transition from the BIF occurrences in the north, which could have formed in a wide back-arc basin, to the VMSs in the south that are associated with the felsic varieties of the bimodal assemblage supports the southward propagation of the back-arc spreading into a rifted island arc (Figure 7B). Both BIF and VMS deposits are contemporaneous as the ages of their host rocks in Wadi Kariem (Ali *et al.* 2009) and Um Samuiki (Stern *et al.* 1991) overlap.

Post-accretionary gold mineralization

The gold mineralization associated with the post-accretionary stage in the CED is mainly vein-type deposits and disseminations in alteration zones. The auriferous veins are hosted in the intra-oceanic island arc metavolcano-sedimentary assemblages, the sheared ophiolitic serpentinites and/or the granitic rocks intruding them (El-Bouseily *et al.* 1985; Klemm *et al.* 2001; Zoheir and Lehmann 2011). The auriferous veins are hosted in the island arc and ophiolitic assemblages that were metamorphosed at P–T conditions below amphibolite facies, dominantly within the greenschist facies (Botros 2004). As veins are structurally controlled, greenschist facies conditions are the favourite ones for the brittle–ductile deformation, which form the channel ways for gold mineralization (El-Gaby *et al.* 1988). The auriferous veins are composed mainly of quartz and carbonates with some sulphides (Botros 2002, 2004). Pyrite and to a lesser extent chalcopyrite are the main sulphide minerals, and they occur as disseminations and as fracture-filling in quartz. Arsenopyrite, sphalerite, tetrahedrite, galena, gersdorffite, and pyrrhotite are also recorded as trace quantities. Gold occurs as native, tiny specks within the quartz veins or as inclusions in sulphides, especially in pyrite and arsenopyrite.

The nature of the mineralizing fluids and the source of gold for the vein-type deposits related to the Cordilleran margin stage are not well understood (Zoheir 2008). From the available fluid inclusion and stable isotope data, which are limited compared with the distribution of the vein-type deposits in the CED, the mineralizing fluids show a considerable amount of CO_2 and wide salinity and temperature ranges (Botros 2002, 2004). The available data indicate that the mineralizing fluids, mainly metamorphic in origin (Hassaan and El Mezayen 1995), were mixed with fluids derived from different granitic phases (Harraz 2000; Klemm *et al.* 2001; Botros 2004). Zoheir *et al.* (2008) interpreted many of the vein-type gold deposits as orogenic gold deposits. There is a debate over the source of gold, whether it was mobilized from the island arc-related mafic metavolcanic rocks and their associated tuffaceous sedimentary rocks or from the ophiolitic serpentinites (Bakhit 2001; Klemm *et al.* 2001; Said 2006, 2010; Zoheir *et al.* 2008), or even derived from the Cordilleran margin magmatism (El-Gaby *et al.* 1988).

Most of the vein-type gold deposits in the CED are hosted within the western arc–forearc belt (Figures 5 and 8). The reason for this spatial distribution could be the preservation of the serpentinite bodies, which delineate the western forearc suture of Abd El-Rahman (2010) (Figure 5). Such serpentinites have high primary gold contents, which may have been mobilized by granitic intrusions during or after the build-up of the Cordilleran margin (Klemm *et al.* 2001; Said 2006, 2010; Zoheir and Lehmann 2011). Buisson and LeBlanc (1987) suggested the enrichment of the upper mantle rocks of other Neoproterozoic ophiolites in gold. Furthermore, some of the serpentinites of the CED are associated with graphite schist such as in the El-Sid mine in the Fawakhir area (El-Bouseily *et al.* 1985) and in the Barramiya area (Said 2006). These carbon-rich varieties may represent an important sink for gold from the circulating hydrothermal solutions (Cox *et al.* 1991; Goldfarb *et al.* 2005). The vein-type gold deposits in the CED are less commonly present in the eastern arc–back-arc belt (Sukari gold mine is an exception). Some of the vein-type gold occurrences are associated with the BIF deposits and their interlayered metavolcanic and volcaniclastic rocks (Zoheir and Akawy 2009). Generally, iron-rich rocks represent a good sink for

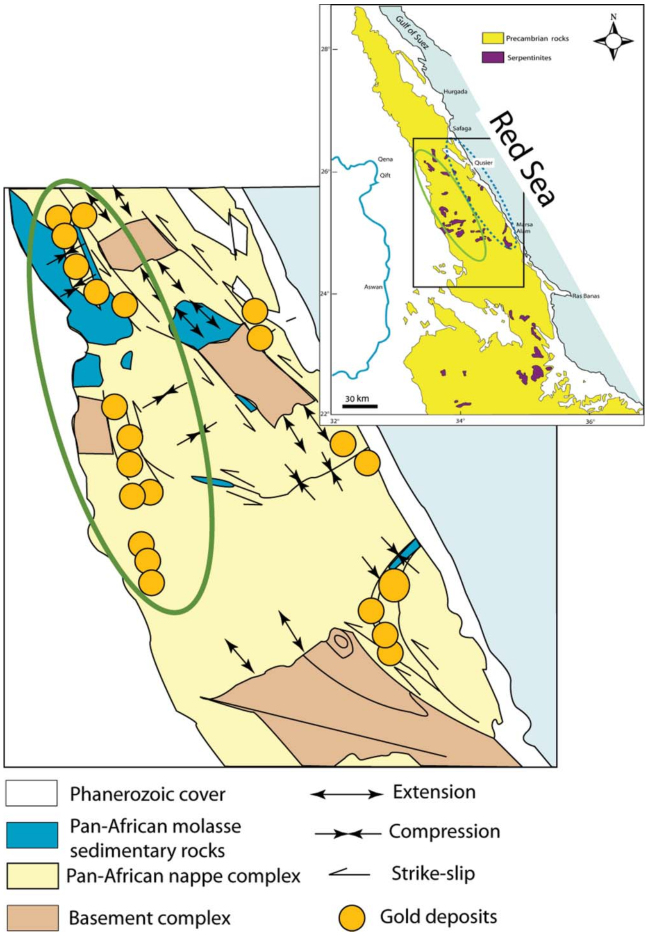

Figure 8. Simplified geological map showing the distribution of the main vein-type gold deposits in relation to the major structural features of the CED (from Helmy *et al.* 2004). The number of gold deposits increases along the western arc–forearc belt in harmony with the high density of serpentinite occurrences.

gold from the circulating hydrothermal deposits (Goldfarb *et al.* 2005). In addition to the suitability of the chemical environment, Helmy *et al.* (2004) linked the spatial distribution of the vein-type gold deposits in the CED to fractures induced by Najd Fault System (NFS) (Figure 8). The NFS is a NW-trending sinistral strike–slip fault system and can be considered as the first-order ore-controlling regional structure. However, the dominance of the vein-type gold deposits along the eastern side of the fault system can still be related to the gold enrichment of the arc–forearc belt due to the dominance of the serpentinite bodies.

Concluding remarks

This study reveals the following conclusions regarding the tectonic evolution of the CED based on the distribution of the ophiolitic rocks and associated mineralization generated mainly during Cryogenian time:

(1) Neoproterozoic ophiolitic rocks in the CED have both forearc and back-arc geochemical affinities.
(2) Arc–forearc assemblages are preserved to the west of the arc–back-arc assemblages, indicating that the intra-oceanic island arc system was established above an E-dipping subduction zone (present coordinates). Both assemblages extend as NW–SE belts defining a suture zone in the CED.
(3) Arc–back-arc assemblages show an increase in subduction geochemical signature towards the south. The occurrence of bimodal volcanism in the south indicates progressive opening of a back-arc basin from the north to the south, and the preservation of the rifted island arc assemblages farther southward.
(4) The transition of CED rock assemblages from an intra-oceanic island arc system to a cordilleran margin was caused by the accretion of the intra-oceanic island arc onto the older continental margin, the Saharan metacraton. This collisional accretion led to a flip in subduction polarity and to the initiation of cordilleran-type magmatism above a wester-dipping subduction zone.
(5) The spatial distribution and variations in styles of mineralization, which are related to the intra-oceanic island arc assemblages are consistent with the evolution of the CED.
    (a) BIF-associated tuffaceous metasedimentary rocks are associated with the arc–back-arc assemblages.
    (b) VMSs hosted in the bimodal volcanic rocks were erupted in the rifted island arc environment at the southern tip of the propagating back-arc basin to the south of the auriferous BIF occurrences.
(6) Vein-type gold mineralization is dominant along the western arc–forearc belt. We attribute the localization of vein-type deposits within the arc–forearc belt to the dominance of serpentinites, which show high primary gold contents. The cordilleran and/or later post-tectonic magmatism evidently may introduce a suitable condition for gold mobilization and concentration as vein deposits.

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