Cenozoic metallogeny of Greece and potential for precious, critical and rare metals exploration

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Abstract

The Cenozoic metallogeny in Greece includes numerous major and minor hydrothermal mineral deposits, associated with the closure of the Western Tethyan Ocean and the collision with the Eurasian continental plate in the Aegean Sea, which started in the Cretaceous and is still ongoing. Mineral deposits formed in four main periods: Oligocene (33–25 Ma), early Miocene (22–19 Ma), middle to late Miocene (14–7 Ma), and Pliocene-Pleistocene (3–1.5 Ma). These metallogenic periods occurred in response to slab-rollback and migration of post-collisional calc-alkaline to shoshonitic magmatism in a back-arc extensional regime from the Rhodopes through the Cyclades, and to arc-related magmatism along the active south Aegean volcanic arc. Invasion of asthenospheric melts into the lower crust occurred due to slab retreat, and were responsible for partial melting of metasomatized lithosphere and lower crustal cumulates. These geodynamic events took place during the collapse of the Hellenic orogen along large detachment faults, which exhumed extensive metamorphic core complexes in mainly two regions, the Rhodopes and the Cyclades. The detachment faults and supra-detachment basins controlled magma emplacement, fluid circulation, and mineralization.

The most significant mineralization styles comprise porphyry, epithermal, carbonate-replacement, reduced intrusion-related gold, intrusion-related Mo-W and polyvalentic veins. Porphyry and epithermal deposits are commonly associated with extensive hydrothermal alteration halos, whereas in other cases alteration is of restricted development and mainly structurally controlled. Porphyry deposits include Cu-Au-, Cu-Mo-Au-Re, Mo-Re, and Mo-W variants. Epithermal deposits include mostly high- and intermediate-sulfidation (HS and IS) types hosted in volcanic rocks, although sedimentary and metamorphic rock hosted mineralized veins, breccias, and disseminations are also present. The main metal associations are Cu-Au, Cu-Mo-Au-Re, Mo-Re, and Mo-W variants. Epithermal deposits include mostly high- and intermediate-sulfidation (HS and IS) types hosted in volcanic rocks, although sedimentary and metamorphic rock hosted mineralized veins, breccias, and disseminations are also present. The main metal associations are Cu-Au, Cu-Mo-Au-Re, Mo-Re, and Mo-W variants.

Finally hundreds of polymetalliferous veins hosted by metamorphic rocks in the Rhodopes and Cyclades significantly add to the metal endowment of Greece.
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1. Introduction

Greece is historically an important country as a metal producer especially during antiquity. Mining dates back to prehistoric times, focusing initially on gold, copper and iron. Later, in the Classical (5th to 4th centuries BC) and Hellenistic (4th to 2nd centuries BC) times, ore exploitation became more intense, especially for Au, Ag, Cu and Fe. Major mining centers operated around the Aegean, where a large number of underground galleries and metallurgical furnaces from that time are still preserved. Several metals, such as Ag, Au, Fe and Cu, were also produced during the Byzantine (4th to 5th centuries AD) and Ottoman periods (15th to 19th centuries AD). Since the beginning of the 20th century, especially after the Second World War (post-1945) several mines have operated for the exploitation of Cr, Ni, Al, Mn, Fe, Pb, Zn, Cu, Au and Ag from chromite, laterite, bauxite, iron-manganese, and Pb-Zn-(Cu) deposits. Carbonate-replacement deposits (Tsirtambides and Filippidis, 2012), however, the majority of these operations has been suspended in recent years for economic, political, social, and environmental reasons (Melfos and Voudouris, 2012). Nevertheless, Greece remains the most important country in the European Union (EU) for the production of Ni-Fe from laterites (2014 production: 2.4 Mt containing 18,500 t Ni) and Al from bauxites (2014 production: 1.8 Mt containing 170,000 t Al). The production of Mg from magnesite is also significant (2014 production: 270,000 t magnesite). More recent exploration has discovered deposits of gold, silver, copper, molybdenum, rhenium, antimony, tungsten, and tellurium, mainly in northern Greece (Melfos et al., 2002; Voudouris, 2006; Voudouris et al., 2009; Fornadel et al., 2011; Melfos and Voudouris, 2012; Voudouris et al., 2013a,b,c; Bristol et al., 2015; Stergiou et al., 2016).

The geological setting of the Hellenides (Fig. 1), part of the Alpine-Himalayan orogeny, comprises numerous geotectonic terranes formed in the complicated geodynamic environment of the Paleozoic and Neotethys (Fig. 1: Pe-Piper and Piper, 2002; Schmid et al., 2008; Jolivet and Brun, 2010; Ring et al., 2010; Jolivet et al., 2013; Menant et al., 2016). During post-orogenic episodes, after the closure of the Vardar and Pindos (Neotethys) oceanic basins, large scale detaching events formed that enabled metamorphic core complexes. Meanwhile, Tertiary to Quaternary calc-alkaline to alkaline magmatism in the Aegean region occurred behind the active Hellenic subduction zone (Jolivet et al., 2013).

In this geotectonic regime several significant ore deposits and hundreds of smaller occurrences formed in association with magmatism and magmatic-hydrothermal systems. Deposits include porphyry, carbonate-hosted replacement Pb-Zn-Ag-Au, high-, intermediate- and low-sulfidation (IS and LS) epithermal polymetallic and precious metal vein, skarn, and intrusion-related gold deposits. They are mainly concentrated in two broad regions, the Rhodopes and the Cyclades (Figs. 1 and 2), which underwent similar tectono-metamorphic and magmatic evolution since the Cretaceous. These regions are the most promising targets for future exploration of precious, rare, and critical metals in Greece.

This paper focuses on the metallic mineral deposits of these two metallogenic provinces. The geology and ore types involved in each magmatic belt are reviewed, along with some specific features of these deposits including ore mineralogy, hydrothermal alteration, geochemistry, and age determinations and resource estimates where available. These data are used to evaluate the potential of these regions for future exploration and discovery of precious, rare, and critical metals.

2. Geological setting

2.1. Geodynamic overview of pre-Cenozoic to Cenozoic evolution in the Aegean region

The Hellenide orogen is a discrete terrane of the Alpine-Himalayan deformational belt and represents a geotectonic link between the southern Balkan Peninsula (e.g., the Dinarides/Albanides) and Turkey (e.g., Pontides, Anatolides). It consists, from north to south, of three continental blocks (Rhodopes, Pelagonia, and Adria-External Hellenides) and two intervening oceanic domains (Vardar and Pindos Suture Zones) (Kydonakis et al., 2015a,b). The Hellenides formed as a result of the ongoing Alpine collision between the African and Eurasian plates since the Late Jurassic to the present above the north-dipping Hellenic subduction zone. This convergence caused thrusting and SW-verging nappe-stacking of the Rhodopes, Pelagonia and Adria continental
blocks, and closure of the Vardar and Pindos oceanic domains of the Neotethys (Robertson, 2002; Schmid et al., 2008; Jolivet and Brun, 2010; Kydonakis et al., 2015a,b; Menant et al., 2016; Brun et al., 2016).

Late Precambrian to Late Jurassic evolution of the region was marked by the opening and closure of ocean basins and continental crust formation above the accompanying subduction zones (Anders et al., 2006; Himmerkus et al., 2006; Reischmann and Kostopoulos, 2007). A Permo–Carboniferous igneous event related to northward subduction of Paleotethys beneath the southern margin of Europe is recognized from the Pelagonia, Rhodope, and Attico-Cycladic areas, and records an active continental margin evolution in the Precambrian-Silurian basement of the Hellenides (Anders et al., 2006; Reischmann and Kostopoulos, 2007).

The Vardar Ocean opened during the Late Triassic–Early Jurassic, separating the Pelagonia and Rhodope continents (Robertson et al., 2013). Northeastward subduction created a Late Jurassic magmatic arc along the southern margin of the Rhodope...
continent, while future ophiolites formed by supra-subduction zone spreading within the Vardar Ocean. Final collision between Europe and Pelagonia at the end of the Cretaceous closed the Northern Neotethys Ocean along the Vardar Suture Zone, as evidenced by obducted Jurassic ophiolites on the Pelagonia continental block (Robertson et al., 2013).

Accretion of the Pelagonia continental block to the Eurasian margin was followed by the onset of subduction of the Pindos oceanic basin in the Paleocene (~60–55 Ma) and then by its closure at ~35 Ma when the Adria and Pelagonia microcontinents collided (Brun and Faccenna, 2008; Ring et al., 2010; Jolivet et al., 2013; Menant et al., 2016). In the center of the Aegean, the Cyclades partly expose the buried parts of the Pindos Ocean and its continental margins (Menant et al., 2016). A single subduction zone was active from the Early Cretaceous, which consumed both oceanic and continental lithospheric mantle beneath the Aegean Sea (van Hinsbergen et al., 2005; Brun and Faccenna, 2008; Ring et al., 2010; Jolivet et al., 2013; Menant et al., 2016; Brun et al., 2016). The geodynamic evolution of the Aegean area includes slab roll-back since the Eocene, and slab tearing below western Anatolia during the Miocene. As a result of slab roll-back, collapse of the accretionary wedge and the opening of back-arc basins allowed

Fig. 2. Distribution of Cenozoic mineralization in Greece (references as in Fig. 1).
for the post-orogenic exhumation of the lower parts of the stretched crust as metamorphic core complexes, such as in the Rhodopes and the Cyclades, with voluminous post-collisional magmatism (Fig. 1; Menant et al., 2016, and references therein).

2.2. The Rhodope Massif

The Rhodope Massif, together with other major units of the Balkan orogen in Serbia, Former Yugoslavian Republic of Macedonia (FYROM), Bulgaria and Greece (e.g., the Dacia-Serbo-Macedonian Mega-unit and the Strandzha-Circum-Rhodope Belt), and the Sakarya block in western Turkey, are considered to have been part of the European continental margin (Schmid et al., 2008; Himmerkus et al., 2009a; Burg, 2012; Kydonakis et al., 2015a,b). The Rhodope Massif is bordered to the north by the Maritza shear zone separating the Rhodopes from the Srednogorie in Bulgaria, and to the south by the Vardar Suture Zone. The Rhodope Massif is a heterogeneous crustal body composed of three sub-domains (Fig. 3), namely the Northern Rhodope Domain, the Southern Rhodope Core Complex (both separated by the Nestos thrust fault), and the Chalkidiki Block, which coincides with the so-called Serbo-Macedonian Massif (excluding the Kerdylion Unit, which belongs to and has a common tectono-metamorphic history with the Southern Rhodope Core Complex; Kydonakis et al., 2015a,b). The Chalkidiki block and the Northern Rhodope Domain share a similar tectono-metamorphic history and participated in the same...
Mesozoic crustal-scale southwestward accretionary event before development of the Southern Rhodope Core Complex (Kydonakis et al., 2015a,b, 2016).

The Northern Rhodope Domain consists of the following main metamorphic units: (i) a lower unit of high-grade basement including orthogneisses derived from Permo-Carboniferous protoliths; this unit includes four metamorphic core complexes (the Arda, Biala Reka–Kechros, and Kesebir-Kardamos migmatitic domes); (ii) an intermediate unit of high-grade basement rocks that have both continental and oceanic affinities, and with protholiths ranging in age from Neoproterozoic through Ordovician, and Permo-Carboniferous to Early Cretaceous; and (iii) an overlying uppermost Mesozoic low-grade unit of the Circum-Rhodope Belt and the Evros ophiolite (Turpault and Reischmann, 2010; Kirchenbaur et al., 2012; Meinhold and Kostopoulos, 2013; Bonev et al., 2015). The rocks of the intermediate unit experienced high to ultra-high pressure metamorphism with subsequent high-grade amphibolite-facies overprint (Mposkos and Kostopoulos, 2001).

The Chalkidiki Block (Fig. 3) represents a thrust system composed of four NW-trending units: the Vertiskos Unit including the Thermes-Volvi-Gomati (TVG) complex, the Circum-Rhodope Belt, the Chortiatis Magmatic Suite, and the eastern vardar ophiolites. The latter two units comprise a Middle to Late Jurassic arc/back-arc spreading system (Bonev et al., 2015; Kydonakis et al., 2015a,b; Siron et al., 2016). The Vertiskos Unit is a Gondwana-derived basement fragment made of Silurian–Ordovician peraluminous orthogneiss with intercalated paragneiss, marble, amphibolite, eclogite, and serpentinite, and was incorporated into the Southern European Arcs by the end of the Palaeozoic (Himmerkus et al., 2009a). The Vertiskos Unit was affected by Palaeozoic deformation and a Late Jurassic–Cretaceous amphibolite-facies overprint (e.g., Kilias et al., 1999; Kydonakis et al., 2015a,b, 2016). A-type granitoids of the Arnea–Kerkini Magmatic Complex intruded during the Early Triassic (Himmerkus et al., 2009b). The Thermes-Volvi-Gomati (TVG) complex is a tectonic mélangé separating the Vertiskos Unit from the southern Rhodope Kerdylion Unit (see below) and consists of the Athos-Volvi Suture Zone and mafic and ultramafic rocks (Siron et al., 2016). The Circum–Rhodope Belt comprises low-grade metasedimentary rocks of Triassic–Jurassic protolith age and minor rhylolites, fringing the crystalline basement of the Vertiskos Unit (Meinhold and Kostopoulos, 2013).

The Southern Rhodope Core Complex (SRCC), which is tectonically juxtaposed against the Vertiskos Unit, is composed of Permo-Carboniferous orthogneiss and massive Triassic marble intercalated with amphibolitic and metapelitic rocks (Dinter et al., 1995; Brun and Sokoutis, 2007; Turpault and Reischmann, 2010). The SRCC displays intense, penetrative, top-to-the-southwest shearing under amphibolite and greenschist facies conditions (Burg et al., 1996). Parts of the SRCC have experienced partial melting and formation of migmatites at Thasos Island and the Kerdylion Unit. These Permo-Carboniferous orthogneisses in the SRCC may be equivalent to those in the Arda, Biala Reka–Kechros, and Kesebir-Kardamos migmatitic domes in the Northern Rhodope Domain (Fig. 3; Brun and Sokoutis, 2007; Kydonakis et al., 2015b).

2.3. The Cyclades

In the center of the Aegean Sea the Cyclades consist of three main units (Bonneau, 1984): the lowermost Pre-Alpidic Cycladic Basement Unit, the intermediate Cycladic Blueschist Unit, and the uppermost Pelagonian Unit (Fig. 4). The Blueschist Unit represents a polymetamorphic terrane which tectonically overlies the basement gneiss and consists of a metamorphosed volcano-sedimentary sequence of clastic metasedimentary rocks, marbles, calc-schists, and mafic and felsic meta-igneous rocks (Parra et al., 2002; Bröcker and Pidgeon, 2007; Katzir et al., 2007; Ring et al., 2010; Jolivet et al., 2013; Scheffer et al., 2016).

The Cyclades have been exhumed since the Eocene as metamorphic core complexes formed in low, medium, and/or high temperature environments (Menant et al., 2016, and references therein). The Cycladic Blueschist Unit experienced two stages of metamorphism during the Tertiary: the first occurred during the Eocene (~52–53 Ma), which was marked by high-pressure eclogite- to blueschist-facies metamorphism at 15–20 kbar, and ~500 °C. Exhumation of the Cycladic Blueschists first occurred between 45 and 35 Ma in a syn-orogenic context, following a cold retrograde path (Menant et al., 2016; Laurent et al., 2017). The Cycladic Blueschist Unit underwent an early Oligocene medium temperature-medium pressure metamorphic overprint (~9 kbar, 550–570 °C) followed by greenschist-facies retrograde metamorphism. Finally, in the early Miocene, high-temperature medium-pressure metamorphism (~5–8.5 kbar, 500–700 °C) and associated migmatites developed in the central Cyclades (Jolivet and Brun, 2010; Grasemann et al., 2012; Menant et al., 2016, and references therein; Scheffer et al., 2016). The amphibolite to greenschist metamorphic event occurred during extension-related exhumation and was coeval with back-arc extension at the Rhodopes in northern Greece (Altiherr et al., 1982; Lister et al., 1984; Gautier and Brun, 1994; Jolivet et al., 2010; Ring et al., 2010).

The upper Pelagonian Unit did not experience Eocene high-pressure metamorphism, and consists of various klippen of unmetamorphosed Late Permian to Jurassic volcaniclastic rocks, ophiolites, and carbonates, greenschist facies rocks of Cretaceous age, and granitoids (Reinecke et al., 1982; Bröcker and Pidgeon, 2007). It was thrust onto the Blueschist Unit at ~25–20 Ma (Boronkay and Doutsos, 1994).

Exhumation of the Cycladic rocks as metamorphic core complexes was accommodated during the Oligocene–Miocene by several ductile to brittle detachment systems (see below). The extensional event also allowed for various granitoids (granite, granodiorite, leucogranite) to be intruded throughout the Cyclades between 15 and 7 Ma (Altiherr et al., 1982; Pe-Piper and Piper, 2002; Skarpelis et al., 2008).

2.4. Back-arc extension and metamorphic core complexes in the Aegean

In the Rhodope region, Paleogene-Neogene exhumation of the metamorphic pile and core complexes followed the Cretaceous syn-metamorphic SW-directed thrusting (Burg et al., 1996; Kilias et al., 1999; Bonev et al., 2006a,b; Brun and Sokoutis, 2007; Burg, 2012; Kydonakis et al., 2015a,b; Brun et al., 2016). In the Northern Rhodope Domain, early (syn-orogenic) exhumation was initiated in the early Eocene (~55 Ma) and lasted up to ~42 Ma (middle Eocene), and was coeval with the accretion of Pelagonia following the closure of the Vardar Ocean (Wüthrich, 2009; Menant et al., 2016). Two later stages of extension and metamorphic core complex formation occurred in the Rhodope Massif from ~42–35 Ma and ~24–12 Ma (Wüthrich, 2009; Jolivet and Brun, 2010; Márton et al., 2010; Moritz et al., 2010; Burg, 2012; Jolivet et al., 2013; Kounov et al., 2015). The first phase of core complex formation in the Rhodopes (40–35 Ma) was coeval with the subduction of the Pindos Ocean following accretion of the Pelagonian continental block to the Eurasian margin (Brun and Favennac, 2008; Wüthrich, 2009). The second phase of core complex-related extension took place after suturing of the Pindos oceanic domain (~35 Ma; Ring et al., 2010), and was coeval with the accretion and subduction of continental blocks of the external Hellenides, and back-arc
extension initiated by the onset of the subduction of the Mediterranean Ocean.

In the central and eastern Rhodope area of Bulgaria and Greece (e.g., the Northern Rhodope Domain core complexes), the Arda, Biala Reka-Kechros, and Kesebir-Kardamos migmatitic domes (Fig. 3) were progressively exhumed along several ductile to brittle shear zones, all active from $/C24_{42}^{35}$ Ma (Bonev et al., 2006a,b, 2010; 2013; Wüthrich, 2009; Márton et al., 2010; Moritz et al., 2010; Kaiser-Rohrmeier et al., 2013). In the South Rhodope Core Complex, cooling/exhumation of the metamorphic rocks occurred along two major detachment faults (Brun and Sokoutis, 2007; Wüthrich, 2009; Márton et al., 2010; Moritz et al., 2010; Kaiser-Rohrmeier et al., 2013). In the South Rhodope Core Complex, cooling/exhumation of the metamorphic rocks occurred along two major detachment faults (Brun and Sokoutis, 2007; Wüthrich, 2009; Márton et al., 2010; Moritz et al., 2010; Kaiser-Rohrmeier et al., 2013). In the South Rhodope Core Complex, cooling/exhumation of the metamorphic rocks occurred along two major detachment faults (Brun and Sokoutis, 2007; Wüthrich, 2009; Márton et al., 2010; Moritz et al., 2010; Kaiser-Rohrmeier et al., 2013).

In the Aegean back-arc domain at $/C24_{35}^{30}$ Ma ago, an increase of the rate of slab retreat led to the initiation of post-orogenic extension, largely accommodated by large-scale structures such as the North Cycladic Detachment System (NCDS), active from $/C24_{35}^{10}$ Ma (Jolivet et al., 2013)(Fig. 4). The North Cycladic detachment system dips toward SW with top-to-the-SSW kinematics. Two more detachment systems, the South Cycladic Detachment System and the Central Cycladic Detachment (Naxos and Paros), have been recognized in the Cyclades (Iglseder et al., 2011, and references therein). New $^{40}$Ar/$^{39}$Ar and U-Th/He thermochronological data suggest that the West Cycladic Detachment System accommodated extension throughout the Miocene (Grasemann et al., 2012). The evolution of the western Cyclades can be resolved into a coherent and uniform tectonic progression involving SSW-directed ductile to brittle extension, localized plutonism, and rapid cooling of the footwall between 9 and 6 Ma. Because both the North and the West Cycladic Detachment Systems were active until the late Miocene but exhibit opposing shear sense, Grasemann et al. (2012) proposed that a large part of the stretching of the Aegean crust was accommodated by these two bivergent crustal-scale detachment systems.

In the Rhodope Massif the exhumation of metamorphic core complexes along detachment faults resulted in the formation of Palaeocene to early Eocene, late Eocene–Oligocene, and Miocene supra-detachment sedimentary basins (Bonev et al., 2006a,b; Márton et al., 2010; Killas et al., 2013a). Sediments transgressively or tectonically overlie the metamorphic units of the domes in fault-bounded half-grabens located along the hanging wall of the low-angle detachment faults. Continental extension in Biga and the northeastern Aegean islands of Limnos and Lesvos, as well as in the Cyclades, also resulted in a system of back-arc E–W-
NNW–SSE-oriented extensional sedimentary basins (Bonev and Beccalotto, 2007; Burchfiel et al., 2008; Brun and Sokoutis, 2007). Since ~16 Ma, extension in the Aegean region has been controlled by the westward extrusion of Anatolia along the North Anatolian Fault (Jolivet et al., 2013).

2.5. Cenozoic magmatic evolution

Several possible causes for the post-collisional volcanic and plutonic activity in the Aegean–west Anatolian region have been proposed, including slab roll-back, post-orogenic collapse and lithospheric thinning, delamination of crustal/lithospheric slices, and slab break-off (e.g., Fytikas et al., 1984; de Boorder et al., 1998; Pe-Piper et al., 1998; Ersoy and Palmer, 2013, and references therein). However, mantle tomographic images indicate that a single slab was subducted since at least the latest Cretaceous, and was responsible for the geodynamic events in the Aegean domain (Jolivet and Brun, 2010). The relationships between subduction dynamics and magmatic evolution in the Aegean region has been studied by Pe-Piper and Piper (2006), Ersoy and Palmer (2013), Jolivet et al. (2013, 2015), and Menant et al. (2016), and their work is summarized below.

Initial “Andean-style” subduction of the Vardar-Izmir-Ankara Ocean and associated Late Cretaceous (~92–67 Ma) arc magmatism resulted in the Apuseni–Banat–Timok–Srednogorie–Pontides magmatic belt (Gallhofer et al., 2015). This was followed by post-collisional Paleocene–Eocene (56–40 Ma) adakite-like magmatism derived from slab melts, in the Rhodope and the west Srednogorie regions, and is part of a 250 km-long NW-trending belt which continues through northern Turkey (Marchev et al., 2013). This magmatic belt includes, among others, the Pirin, Rila, Barutın–Elatiá–Sklotoı̈, Sithonia, and Jerissos plutons in Bulgaria and Greece (Fig. 3). Adakite-like magmatism was related to a deep slab break-off (Vardar Ocean) and was followed by asthenospheric upwelling, fast exhumation, and the first period of core complex formation in the Rhodopes from 42 to 35 Ma (Marchev et al., 2013). A slab break-off model has also been proposed for the generation of the adakite-like Eocene magmatic rocks in NW Anatolia (Ersoy and Palmer, 2013).

Asthenospheric upwelling and the first stage of core complex formation in the Rhodopes was followed by orogenic collapse, steep faulting, and extensional late Eocene to Miocene calc-alkaline to shoshonitic and calc-alkaline plutonic, subvolcanic and volcanic rocks, which include the Tertiary basins in Rhodopi and Evros counties (Arikas and Voudouris, 1998; Christofides et al., 2004). Several plutons (e.g., Vrondou, Xanthi, Maronia-Kirkı-Leptokarya) intruded contemporaneously with detachment faulting (e.g., in the footwall of detachments) and are partly mylonitized. In the Kassandra mining district, Oligocene intrusives were emplaced at ~27–25 Ma, contemporaneous with movement of the Kerdyon detachment (Brun and Sokoutis, 2007; Hahn et al., 2012; Siron et al., 2016). The late Eocene–Oligocene magmatism in Bulgaria and Greece, Serbia, Kosovo, and FYROM shows a decreasing influence of crustal contamination with time and an increasing input from the mantle, until the eruption of purely asthenospheric magmas. The magmatism evolved from K-rich trachybasalts (34 Ma) via shoshonites, calc-alkaline and high-K calc-alkaline basalts (33–31 Ma), to alkaline basalts (28–26 Ma), and is considered to have resulted from the melting of deformed mantle, metasomatized by earlier subduction processes (Cvetković et al., 2004; Marchev et al., 2005, 2013; Prelević et al., 2005; Schefer et al., 2011; Bonev et al., 2013; Lehmann et al., 2013). It is further suggested that the late Eocene-Oligocene (~35–25 Ma) magmatism in the Rhodope Massif was caused by convective removal of the lithospheric mantle (lithospheric delamination) and subsequent upwelling of the asthenosphere (Christofides et al., 2004; Marchev et al., 2005; Pe-Piper and Piper, 2006).

Early Miocene magmatism in the northeastern Aegean islands of Limnos and Lesvos (Figs. 1 and 3) and in western Turkey occurred mostly to the south of the Oligocene igneous activity of the Rhodope block and was coeval with the second phase of core complex-related extension in the Rhodope Massif and with the entrance of the Mediterranean Ocean into the Hellenic subduction zone (Dinter and Royden, 1993; Pe-Piper and Piper, 2002; Pe-Piper et al., 2009). The early Miocene igneous rocks form a belt of shoshonitic and calc-alkaline plutonic, subvolcanic and volcanic rocks on the Greek islands of Lesvos, Limnos, and Samothraki (Fig. 1), and in western Turkey (Pe-Piper and Piper, 2002; Pe-Piper et al., 2009; Ersoy and Palmer, 2013).

Early Miocene (22–17 Ma) calc-alkaline to shoshonitic magmatism was also present in the Rhodope Massif (e.g., Skouries monzonite porphyry; Kroll et al., 2002; and Kavala and Pangeon granitoids; Dinter et al., 1995), with the latter being emplaced contemporaneously to movement along the Strymon detachment (Dinter and Royden, 1993; Brun and Sokoutis, 2007).

During the Miocene (from ~17 to ~7 Ma) several intrusions, partly associated with coeval volcanic rocks, were emplaced during the exhumation of metamorphic core complexes in the footwall of major detachments in the Menderes Massif and Cyclades (Altherr and Siebel, 2002; Pe-Piper and Piper, 2006; Iglser et al., 2009; Denèle et al., 2011; Menant et al., 2016). Magmatic rocks are of high-K calc-alkaline to shoshonitic affinity, and granitoids are of both I- and S-type (Altherr and Siebel, 2002; Ersoy and Palmer, 2013). For these magmas, a subduction-modified metasomatized mantle source with the contribution of crustal sources has been proposed (Altherr and Siebel, 2002; Stouraiti et al., 2010; Ersoy and Palmer, 2013). Typical arc-related volcanism with medium-K calc-alkaline composition has been erupted since the Pliocene along the active South Aegean Volcanic Arc (Fytikas et al., 1984).

3. Types and styles of magmatic-hydrothermal ore deposits

The Oligocene-Miocene and Pliocene-Pleistocene magmatic belts in the Rhodopes and the Cyclades are considered among the most endowed areas within the Alpine-Balkan-Pontide-Anatolide segment of the Western Tethyan belt, which extends...
from Serbia to Turkey, Armenia, and Iran (Janković, 1997; Arikas and Voudouris, 1998; Heinrich and Neubauer, 2002; Marchev et al., 2005; Voudouris and Aliferis, 2005; Voudouris, 2006; Arvanitidis, 2010; Melfos and Voudouris, 2012, 2016; Vigit, 2012; Siron et al., 2016; Voudouris et al., 2016a,b). These belts are rich in precious, rare, and critical metal deposits including Cu-Au, Cu-Mo-Au-Re, and Mo porphyries, Cu-Au-Te high-sulfidation (HS), Pb-Zn-Ag-Au intermediate-sulfidation (IS), and Au-Ag low-sulfidation (LS) epithermal deposits, as well as Fe-Cu-W-Au skarns, Pb-Zn-Ag-Au carbonate-replacement deposits, and reduced intrusion-related gold systems with proximal to distal mineralization (e.g., intrusion-hosted Au-Ag-Bi-TexW sheeted veins, structurally controlled Au-As rich lodes, and distal-disseminated Carlin-style Au mineralization). The location of Cenozoic deposits in Greece, as well as their characteristics, is presented in Figs. 2–4 and in Tables 1–4. All these deposits are related to post-collisional magmatic rocks, which are in part controlled by detachment faults and exhumation of metamorphic core complexes in a back-arc setting. Arc-magnetism along the active South Aegean Volcanic Arc is related to sub-marine epithermal deposits at Milos (Aliferis et al., 2013) and to active VMS-epithermal ore deposition at Kolumbo volcano, near Santorini Island (Killas et al., 2013b).

3.1. Porphyry deposits

In northern Greece, porphyry deposits are mainly concentrated at both in the eastern (e.g., Maronia, Pagoni Rachi, Konos, Kas, siteres, Myli, Melitena, Figs. 3) and western parts of the Rhodope Massif (e.g., Skouries, Fisoka, Tsikara, Vathi and Gerakario; Fig. 3). The porphyry deposits in the eastern part of the Rhodopes (e.g., the Northern Rhodope Domain) are associated with felsic to intermediate subvolcanic rocks of calc-alcaline to high-K calc-alcaline affinity with subalkaline to alkaline character, and of Oligocene age (Ortelli et al., 2009, 2010; Moritz et al., 2010; M. Ortelli, pers. commun., 2010). Most of these porphyry systems show epithermal overprinting evidenced by Au-Ag-rich polymetallic veins. The porphyry systems in the Northern Rhodope Domain are not well explored and generally contain low Cu with variable Au, Mo and Re concentrations (Voudouris et al., 2013c).

3.1.1. Porphyry deposits in the eastern Rhodope Massif

The Maronia porphyry Cu-Mo-Au mineralization, exposed at Kitsmata Hill, is related to a microgranite porphyry emplaced within the Maronia pluton, which consists mainly of monzonite with an age of 29.8 ± 1.3 to 28.4 ± 0.9 Ma (Rb-Sr on whole rock biotite, and fission-track dating of apatite; Kyriakopoulos 1987; Del Moro et al., 1988; Biggazzi et al., 1989). Sodic-potassic, propylitic, sericitic, and argillic alteration styles predominate, and mineralization is mainly associated with three highly silicified zones within the sericitic zone (Melfos et al., 2002). Molybdenite contains very high Re concentrations of up to 2.88 wt.%. Surface samples show concentrations of up to 7600 ppm Mo, 5460 ppm Cu, and 1 ppm Au. Drilling yielded a 10 m intercept grading up to 1.74 wt.%. Bulk ore chemical analyses of porphyry-mineralized samples revealed 400 ppm Cu, 6000 ppm Mo and up to 0.3 ppm Au. High-sulfidation epithermal style alteration is superimposed upon the porphyry quartz stockworks and is characterized by the development of advanced argillic- and silicic alteration and pyrite mineralization.

3.1.2. Porphyry deposits in the western Rhodope Massif

In the western part of the Rhodope Massif, major porphyry deposits are clustered within the Kerdylion Unit of the Southern Rhodope Core Complex (Siron et al., 2016), as well as within the Serbo-Macedonian Vertiskos Unit (Kockel et al., 1975) (Figs. 2 and 3). Skouries is a Cu-Au porphyry deposit in the Kassandra mining district, NE Chalkidiki, with proven and probable reserves of 152.7 Mt grading 0.8 g/t Au, and 0.5% Cu, for 3.8 Moz Au and 776 Mt Cu (Eldorado Gold Corp., 2017). Economou-Eliopoulos and Eliopoulos (2000) and Eliopoulos et al. (2014) also reported up to 41 ppb Ru, up to 150 ppb Pt, and up to 610 ppb Pd in drill-hole samples. The mineralization is associated with a Miocene monzonite porphyry (Kroll et al., 2002; Siron et al., 2016) and occurs in the form of stockwork veins and disseminations (Frei, 1995). The main ore minerals are pyrite, chalcopyrite, bornite, magnetite, minor galena and tetradhrite, and traces of molybdenite (Fig. 5b). Gold grains occur commonly as inclusions within chalcopyrite. Rare Pd-Pt-Au-Ag tellurides also occur (McFall et al.,
Table 1
Characteristics and reserves or resources data of most significant Tertiary porphyry type mineralizations in Greece. HR: Host Rock; ALT: Alteration; MIN: Mineralization.

<table>
<thead>
<tr>
<th>Deposit name</th>
<th>Ore district</th>
<th>Geotectonic belt or unit</th>
<th>Commodities</th>
<th>Deposit style</th>
<th>Morphology of ore bodies</th>
<th>Main host rocks</th>
<th>Age: HR, ALT, MIN</th>
<th>Tonnage and grades</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Eastern Rhodope Massif</strong></td>
<td>Maronia</td>
<td>Circum-Rhodope belt</td>
<td>Cu Mo Fe Pb Zn</td>
<td>High-K calc-alk Cu-Mo</td>
<td>Stockwork, disseminated, veins</td>
<td>Microgranite porphyry</td>
<td>Oligocene</td>
<td></td>
<td>Melfos et al., 2002</td>
</tr>
<tr>
<td><strong>Pagoni Rachi</strong></td>
<td>Kirki-Sapes-Kassiteres-Esimi</td>
<td>Circum-Rhodope belt</td>
<td>Cu Mo Fe Re</td>
<td>High-K calc-alk Cu-Mo</td>
<td>Stockwork, disseminated, veins</td>
<td>Granodiorite-tonalite porphyry</td>
<td>Oligocene</td>
<td></td>
<td>Voudouris et al., 2013b</td>
</tr>
<tr>
<td><strong>Kassiteres</strong></td>
<td>Kirki-Sapes-Kassiteres-Esimi</td>
<td>Circum-Rhodope belt</td>
<td>Cu Mo Pb Zn Ag Bi Te</td>
<td>High-K calc-alk Cu-Mo</td>
<td>Disseminated, vein, stockwork</td>
<td>Microdiortiorite porphyry</td>
<td>Oligocene (ALT: 33.1–31.2 Ma)</td>
<td></td>
<td>Voudouris et al., 2006; Ortelli et al., 2010</td>
</tr>
<tr>
<td><strong>Myli</strong></td>
<td>Kirki-Sapes-Kassiteres-Esimi</td>
<td>Circum-Rhodope belt</td>
<td>Cu Mo Fe Re Pb Zn Ag As</td>
<td>High-K calc-alk Cu-Mo</td>
<td>Vein, stockwork, disseminated</td>
<td>Granodiorite porphyry</td>
<td>Oligocene (ALT: 32.0 ± 0.5 Ma)</td>
<td></td>
<td>Ortelli et al., 2010</td>
</tr>
<tr>
<td><strong>Melitena</strong></td>
<td>Kassiteres-Esimi belt</td>
<td>Circum-Rhodope</td>
<td>Cu-Mo Fe Re Pb Zn Ag As</td>
<td>High-K calc-alk Cu-Mo</td>
<td>Vein, stockwork, disseminated</td>
<td>Granodiorite porphyry</td>
<td>Oligocene (ALT: 32.0 ± 0.5 Ma)</td>
<td></td>
<td>Ortelli et al., 2010</td>
</tr>
<tr>
<td><strong>Western Rhodope Massif</strong></td>
<td>Skouries</td>
<td>Kerdylion Unit</td>
<td>Cu Au Pb Ru Te Mo</td>
<td>High-K calc-alk Cu-Au</td>
<td>Breccia, disseminated, veins, stockwork</td>
<td>Monzonite porphyry</td>
<td>Miocene (HR: 20.56 ± 0.48 to 19.59 ± 0.17 Ma; ALT: 19.9 ± 0.9 Ma)</td>
<td>Reserves: 152,736 Mt at 0.8 g/t Au, and 0.3 % Cu, for 3.8 Moz Au and 776 Mt Cu</td>
<td>Frei, 1995; Hahn et al., 2012; Siron et al., 2016; Eldorado Gold Corp., 2017</td>
</tr>
<tr>
<td><strong>Fisoka</strong></td>
<td>Kassandra mining district</td>
<td>Kerdylion Unit</td>
<td>Cu Au Pb Zn</td>
<td>High-K calc-alk Cu-Au</td>
<td>Breccia, disseminated, veins, stockwork</td>
<td>Diorite and granodiorite porphyries</td>
<td>Oligocene (HR: 24.7 ± 0.14 Ma; ALT: 24.5 ± 1.2 to 23.0 ± 1.2 Ma)</td>
<td></td>
<td>Gilg and Frei, 1994; Tompouloglou 1981; Siron et al., 2016</td>
</tr>
<tr>
<td><strong>Alatina</strong></td>
<td>Kassandra mining district</td>
<td>Kerdylion Unit</td>
<td>Cu Au</td>
<td>High-K calc-alk Cu</td>
<td>Breccia, disseminated, veins, stockwork</td>
<td>Diorite and granodiorite porphyries</td>
<td>Oligocene (HR: 27.00 ± 0.19 to 26.65 ± 0.31 Ma; ALT: 21.2 ± 1.03 Ma)</td>
<td></td>
<td>Gilg and Frei, 1994; Tompouloglou, 1981; Gilg and Frei, 1994; Siron et al. 2016</td>
</tr>
<tr>
<td><strong>Tsiikara</strong></td>
<td>Kassandra mining district</td>
<td>Kerdylion Unit</td>
<td>Cu Au Pb Zn</td>
<td>High-K calc-alk Cu-Au</td>
<td>Breccia, disseminated, veins, stockwork</td>
<td>Diorite and granodiorite porphyries</td>
<td>Oligocene (HR: 0.9 ± 0.1 Ma)</td>
<td></td>
<td>Gilg and Frei, 1994</td>
</tr>
<tr>
<td><strong>Dilofon</strong></td>
<td>Kassandra mining district</td>
<td>Kerdylion Unit</td>
<td>Cu Au Pb Zn</td>
<td>High-K calc-alk Cu-Au</td>
<td>Breccia, disseminated, veins, stockwork</td>
<td>Diorite and granodiorite porphyries</td>
<td>Oligocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td></td>
<td>Gilg, 1993</td>
</tr>
<tr>
<td><strong>Vathi</strong></td>
<td>Kilkis</td>
<td>Vertiskos Unit</td>
<td>Cu Au Ag Fe Mo U</td>
<td>High-K calc-alk Cu-Au</td>
<td>Breccia, disseminated, veins, stockwork</td>
<td>Diorite and granodiorite porphyries</td>
<td>Oligocene (HR: 10.9 ± 0.9 Ma)</td>
<td>Resources: 15 Mt at 0.8 g/t Au, and 0.3 % Cu</td>
<td>Veranis and Tsamantouridis, 1991; Stergiou et al., 2016; Frei, 1992; Tsirambides and Filippidis, 2012</td>
</tr>
<tr>
<td><strong>Gerakario</strong></td>
<td>Kilkis</td>
<td>Vertiskos Unit</td>
<td>Cu Au</td>
<td>Sub-alk Cu-Au</td>
<td>Vein, disseminated</td>
<td>Syenite and granodiorite porphyries</td>
<td>Oligocene (HR: 18 ± 0.5 to 17 ± 1 Ma)</td>
<td>Probable reserves: 28 Mt at 0.9 g/t Au and 0.4% Cu</td>
<td>Voudouris and Alfieris, 2005; Voudouris and Alfieris, 2005</td>
</tr>
<tr>
<td><strong>Northeastern Aegean region</strong></td>
<td>Fakos Limnos</td>
<td>Rhodope massif</td>
<td>Cu Mo Ag Bi Te</td>
<td>Sub-alk Cu-Mo porphyry</td>
<td>Stockwork, breccia, disseminated, vein</td>
<td>Quartz monzonite porphyry</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td></td>
<td>Voudouris and Alfieris, 2005; Fornadel et al., 2012</td>
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<tr>
<td><strong>Sardes</strong></td>
<td>Limnos</td>
<td>Rhodope massif</td>
<td>Cu Mo Au As Zn Pb</td>
<td>Sub-alk Cu-Mo porphyry</td>
<td>Veinlets, stockworks, disseminated</td>
<td>Quartz monzonite porphyry, sandstones, marls</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td></td>
<td>Voudouris and Alfieris, 2005</td>
</tr>
<tr>
<td><strong>Stipsi Lesvos</strong></td>
<td>Rhodope massif</td>
<td>Cu Mo Re Bi Pb Se Ag Au</td>
<td>Sub-alk Mo porphyry</td>
<td>Veinlets, stockwork, disseminated</td>
<td>Dacite porphyry</td>
<td>Miocene (HR: 18.4 ± 0.5 Ma)</td>
<td></td>
<td></td>
<td>Voudouris and Alfieris, 2005</td>
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<tr>
<td><strong>Cyclades</strong></td>
<td>Plaka Lavrion</td>
<td>Attic-Cycladic crystalline Belt</td>
<td>Cu Mo W</td>
<td>Sub-alk Cu-Mo porphyry</td>
<td>Sheeted quartz veins, stockwork</td>
<td>Granodiorite porphyry</td>
<td>Miocene (HR: 9.4 to 7.1 ± 0.6 Ma)</td>
<td></td>
<td>Altherr and Siebel, 2002; Voudouris et al., 2008a</td>
</tr>
</tbody>
</table>
Table 2
Characteristics and reserves or resources data of most significant Tertiary epithermal type mineralizations in Greece. HR: Host Rock; ALT: Alteration; MIN: Mineralization.

<table>
<thead>
<tr>
<th>Deposit name</th>
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<th>Deposit style</th>
<th>Morphology of ore bodies</th>
<th>Main host rocks</th>
<th>Age: HR, ALT, MIN</th>
<th>Tonnage and grades</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eastern Rhodope Massif</td>
<td>Perama Hill</td>
<td>Petrosa Graben</td>
<td>Circum-Rhodope Belt</td>
<td>Au Ag Cu Bi Pb Te Se</td>
<td>High and intermediate sulfidation</td>
<td>Upper parts: oxidized; deeper parts: disseminated, vein, massive sulfide lodes</td>
<td>Oligocene</td>
<td>Reserves: 9.697 Mt at 3.13 g/t Au and 4 g/t Ag, with a total of 0.975 Moz Au and 1.151 Moz Ag</td>
<td>Voudouris et al., 2011a; Eldorado Gold Corp., 2017</td>
</tr>
<tr>
<td>Mavrokoryfi</td>
<td>Petrosa Graben</td>
<td>Circum-Rhodope Belt</td>
<td>Ag Au Cu Te</td>
<td>High sulfidation</td>
<td>Vein, massive sulfate lodes, breccia</td>
<td>Brecia, disseminated, veinlets, vugs fillings</td>
<td>Oligocene</td>
<td>Resources: 0.28 Mt at 19.5 g/t Au, 9 g/t Ag, and 0.4% Cu</td>
<td>Voudouris et al., 2006; Glory Resources, 2012</td>
</tr>
<tr>
<td>Viper</td>
<td>Kirkos-Sapes-Kassitites</td>
<td>Circum-Rhodope Belt</td>
<td>Cu As Sb Ag Pb Zn Bi Te Se</td>
<td>High and intermediate sulfidation</td>
<td>High and intermediate sulfidation</td>
<td>High and intermediate sulfidation</td>
<td>Oligocene</td>
<td>Resources: 0.87 Mt at 2.2 g/t Au, and 1.5 g/t Ag</td>
<td>Ortelii et al., 2010; Glory Resources, 2012</td>
</tr>
<tr>
<td>Scarp</td>
<td>Kirkos-Sapes-Kassitites</td>
<td>Circum-Rhodope Belt</td>
<td>Cu Au</td>
<td>High and intermediate sulfidation</td>
<td>Breccia, disseminated</td>
<td>Granodiorite-tonalite porphyry, tuff, rhyodacite</td>
<td>Oligocene</td>
<td>Resources: 0.21 Mt at 3.5 g/t Au, and 5.2 g/t Ag</td>
<td>Ortelii et al., 2006; Voudouris et al., 2014</td>
</tr>
<tr>
<td>St. Demetrios</td>
<td>Kirkos-Sapes-Kassitites</td>
<td>Circum-Rhodope Belt</td>
<td>Cu As Sb Ag Pb Zn Bi Te Se</td>
<td>High and intermediate sulfidation</td>
<td>Breccia, disseminated, veinlets, vugs fillings</td>
<td>Granodiorite-tonalite porphyry, tuff, rhyodacite</td>
<td>Oligocene</td>
<td>Oligocene (ALT: 31.9 ± 0.6 Ma)</td>
<td>Voudouris et al., 2011b</td>
</tr>
<tr>
<td>St. Barbara</td>
<td>Kirkos-Sapes-Kassitites</td>
<td>Circum-Rhodope Belt</td>
<td>Cu Au Sb Ag Pb Zn Bi Te Se</td>
<td>High and intermediate sulfidation</td>
<td>Veins, breccia</td>
<td>Andesite, tuff, monzodiorite</td>
<td>Oligocene</td>
<td>Oligocene (ALT: 31.2 ± 0.4 Ma)</td>
<td>Voudouris et al., 2013b</td>
</tr>
<tr>
<td>St. Philippos</td>
<td>Kirkos-Sapes-Kassitites</td>
<td>Circum-Rhodope Belt</td>
<td>Pb Zn Ag As Cu Bi Sn Cd In Ga Ge Au</td>
<td>Intermediate and high sulfidation</td>
<td>Vein, breccia, disseminated</td>
<td>SANDstone, conglomerate, marl, tuff, microgranite porphyry</td>
<td>Oligocene</td>
<td>Past mining produced 0.2 Mt at 7.5 % Pb + Zn</td>
<td>Voudouris et al., 2014; Melfos and Voudouris, 2015</td>
</tr>
<tr>
<td>Pefka</td>
<td>Evros</td>
<td>Circum-Rhodope Belt</td>
<td>Cu Pb Ag Zn Bi Sn Ge Ga In Mo V As Hg Te Se</td>
<td>High and intermediate sulfidation</td>
<td>Veins</td>
<td>Tuffs, andesite, rhyodacite, latite</td>
<td>Oligocene (HR: 30.7 Ma)</td>
<td>Oligocene (HR: 19.53 ± 0.75 Ma)</td>
<td>Voudouris et al., 2016; Innocenti et al., 1984; Voudouris and Melfos, 2012</td>
</tr>
<tr>
<td>Loutrons</td>
<td>Evros</td>
<td>Circum-Rhodope Belt</td>
<td>Fe Pb As Ag</td>
<td>High and intermediate sulfidation</td>
<td>Veins</td>
<td>Rhyolite</td>
<td>Miocene (HR: 33 ± 1.2 to 24.6 ± 0.6 Ma)</td>
<td>Miocene (HR: 15.93 ± 0.75 Ma)</td>
<td>Miocene (HR: 19.5 ± 0.7 to 20.2 ± 0.2 Ma)</td>
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<tr>
<td>Kalotycho</td>
<td>Kalotycho-Melitena</td>
<td>Rhodope Belt</td>
<td>Fe</td>
<td>High and intermediate sulfidation</td>
<td>Veins</td>
<td>Tuffs, andesite, trachyte, dacite, rhyolite</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td>Oligocene (HR: 19.53 ± 0.75 Ma)</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
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<tr>
<td>Northeastern Aegean region</td>
<td>Fakos (Lesvos)</td>
<td>Rhodope massif-Pontide Unit</td>
<td>Cu Mo Au Ag Bi Te As Sb Zn Pb</td>
<td>High and intermediate sulfidation</td>
<td>Stockwork, breccia, disseminated</td>
<td>Quartz monzonite porphyry</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td>Oligocene (HR: 19.53 ± 0.75 Ma)</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
</tr>
<tr>
<td>Megala Thermia</td>
<td>(Lesvos)</td>
<td>Rhodope-Pontide Unit</td>
<td>Au Pb Zn Fe Cu Mo</td>
<td>High and intermediate sulfidation</td>
<td>Veinlets, veins, breccia</td>
<td>Andesite, latite</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td>Oligocene (HR: 19.53 ± 0.75 Ma)</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
</tr>
<tr>
<td>Cyclades</td>
<td>Profitis Ilias</td>
<td>Attic-Cycladic ore belt</td>
<td>Pb Zn Ag Cu Bi Sb Te</td>
<td>Intermediate sulfidation</td>
<td>Massive to semi-massive ore, veins</td>
<td>Rhyolite, pyroclastic rocks</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td>Oligocene (HR: 19.53 ± 0.75 Ma)</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
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<tr>
<td>Cyclades</td>
<td>Chondro Vouno</td>
<td>Attic-Cycladic ore belt</td>
<td>Pb Zn Ag Cu Bi Sb Te</td>
<td>Intermediate sulfidation</td>
<td>Veins</td>
<td>Pyroclastic rocks</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td>Oligocene (HR: 19.53 ± 0.75 Ma)</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
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<tr>
<td>Cyclades</td>
<td>Triades-Galana</td>
<td>Attic-Cycladic ore belt</td>
<td>Ag Au As Bi W Mo</td>
<td>Intermediate sulfidation</td>
<td>Veins</td>
<td>Dacites, andesites, pyroclastic and volcanosedimentary rocks</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td>Oligocene (HR: 19.53 ± 0.75 Ma)</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
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<td>Cyclades</td>
<td>Kondaras-Katsimouti-Vani</td>
<td>Attic-Cycladic ore belt</td>
<td>Ag Au As Bi W Mo</td>
<td>Intermediate sulfidation</td>
<td>Massive to semi-massive ore, veins, stratabound and stratiform layers</td>
<td>Dacite, andesite, pyroclastic and volcanosedimentary rocks</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
<td>Oligocene (HR: 19.53 ± 0.75 Ma)</td>
<td>Miocene (HR: 21.3 ± 0.7 to 20.2 ± 0.2 Ma)</td>
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<td>Deposit name</td>
<td>Ore district</td>
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<td><strong>Rhodope Massif</strong></td>
<td>Kassandra mining district</td>
<td>Rhodope massif</td>
<td>Au Ag Pb Zn Cu Sb</td>
<td>Carbonate replacement, vein</td>
<td>Mantos, veins, disseminated</td>
<td>Marble</td>
<td>Oligocene (MIN, apy: 26.1 ± 5.3 Ma)</td>
<td>Reserves: 16.1 Mt at 7.9 g/t Au, 128 g/t Ag, 4.3% Pb and 5.7% Zn</td>
<td>Kalogeropoulos et al., 1989; Gilg, 1993; Forward et al., 2011; Hahn et al., 2012; Eldorado Gold, 2017</td>
</tr>
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<td><strong>Madem Lakkos</strong></td>
<td>Kassandra mining district</td>
<td>Rhodope massif</td>
<td>Au Ag Pb Zn Cu</td>
<td>Carbonate replacement, vein</td>
<td>Mantos, veins, disseminated</td>
<td>Marble, pegmatite, amphibolite, gneiss</td>
<td>Oligocene</td>
<td>Past mining produced 13.5 Mt of Ag-Pb-Zn ore</td>
<td>Gilg, 1993; Eldorado Gold, 2017</td>
</tr>
<tr>
<td><strong>Mavres Petres</strong></td>
<td>Kassandra mining district</td>
<td>Rhodope massif</td>
<td>Au Ag Pb Zn Cu</td>
<td>Carbonate replacement, vein</td>
<td>Mantos, veins, disseminated</td>
<td>Marble, amphibolite, gneiss</td>
<td>Oligocene</td>
<td>Reserves: 1.87 Mt at 160 g/t Ag, 6% Pb and 8.8% Zn</td>
<td>Gilg, 1993; Forward et al., 2010; Eldorado Gold Corp., 2017</td>
</tr>
<tr>
<td><strong>Plavithsa</strong></td>
<td>Kassandra mining district</td>
<td>Rhodope massif</td>
<td>Pb Zn Mn As Au</td>
<td>Carbonate replacement, vein</td>
<td>Mantos, breccia, disseminated, banded epithermal veins</td>
<td>Marble, amphibolite, gneiss</td>
<td>Miocene</td>
<td>Resources: 10.54 Mt at 5.7 g/t Au</td>
<td>Siron et al., 2016; Eldorado Gold Corp., 2017</td>
</tr>
<tr>
<td><strong>Thermes</strong></td>
<td>Rhodope massif</td>
<td></td>
<td>Pb Zn Fe Cu Mn As Sb Cd Te Ag Au</td>
<td>Carbonate replacement, vein</td>
<td>Veins, manto</td>
<td>Marble, amphibolite, orthogneiss</td>
<td>Oligocene</td>
<td></td>
<td>Kalogeropoulos et al., 1996</td>
</tr>
<tr>
<td><strong>Kimmeria</strong></td>
<td>Rhodope massif</td>
<td></td>
<td>Fe Cu Zn W Bi Au</td>
<td>Skarn</td>
<td>Massive, disseminated</td>
<td>Marble, gneiss, amphibolite</td>
<td>Oligocene (HR: 28.8 ± 0.7 to 26.3 ± 0.1 Ma)</td>
<td>Resources: 2 Mt at 20 g/t Ag and 1.98 % Cu</td>
<td>Liat, 1986; Kyriakopoulos, 1987; Vavelidis and Amstutz, 1983</td>
</tr>
<tr>
<td><strong>Thasos, E- and S-part</strong></td>
<td>Thasos</td>
<td>Rhodope massif</td>
<td>Pb Zn Fe Cu As Cd Ag</td>
<td>Carbonate replacement, vein</td>
<td>Mantos, veins</td>
<td>Marble, schist, amphibolite, gneiss</td>
<td>Miocene?</td>
<td>Past mining produced: 2 Mt with 12% Zn + Pb, and 3 Mt with 44% Fe</td>
<td>Vavelidis and Amstutz, 1983</td>
</tr>
<tr>
<td><strong>Thasos, W- and N-part</strong></td>
<td>Thasos</td>
<td>Rhodope massif</td>
<td>Cu Fe Mn As Ni Bi Te Ag Au</td>
<td>Carbonate replacement, vein</td>
<td>Mantos, veins, breccia</td>
<td>Marble, schist, amphibolite, gneiss</td>
<td>Miocene?</td>
<td>Past mining produced: 2 Mt with 12% Zn + Pb, and 3 Mt with 44% Fe</td>
<td>Vavelidis and Amstutz, 1983</td>
</tr>
<tr>
<td><strong>Attic-Cycladic Belt</strong></td>
<td>Lavrion (Kamariza-Sounio-Plaka)</td>
<td>Attic-Cycladic crystalline belt</td>
<td>Pb Zn Ag Fe Cu As Sb Bi Te Ni Cd Ge ln Sn Au</td>
<td>Carbonate replacement, vein</td>
<td>Massive, disseminated, layers, veins</td>
<td>Marble, schist</td>
<td>Miocene (HR: 9.4 ± 0.3 to 7.3)</td>
<td>Past mining produced: Antiquity 6th-1st c. BC: 13 Mt at 400 g/t Ag and 20% Pb – 1864–1978: 1.5 Mt ancient slags at 50 g/t Ag and 10% Pb and 30 Mt ore at 140 g/t Ag and 10% Pb; Resources: 4 Mt with 7% Pb + Zn</td>
<td>Marinos and Petrachek 1956; Conophagos, 1980; Voudouris et al., 2008b; Bonsall et al., 2011</td>
</tr>
<tr>
<td><strong>Serifos</strong></td>
<td>Attic-Cycladic ore belt</td>
<td>Attic-Cycladic crystalline belt</td>
<td>Fe Cu As Pb Zn</td>
<td>Skarn</td>
<td>Massive ore, veins</td>
<td>Marble, schist, gneiss</td>
<td>Miocene (HR: 11.6 ± 0.1 to 9.5 ± 0.1)</td>
<td>Past mining produced: 6.6 Mt for Fe</td>
<td>Salesmin, 1985; Ducoux et al. 2017; Vavelidis, 1997; Neubauer, 2005</td>
</tr>
<tr>
<td><strong>Sifnos</strong></td>
<td>Attic-Cycladic ore belt</td>
<td>Attic-Cycladic crystalline belt</td>
<td>Fe Cu As Pb Zn Mn Ag</td>
<td>Shear zone and detachment fault related Carbonate replacement, vein</td>
<td>Veins, massive, lenses</td>
<td>Gneiss, schist, marble</td>
<td>Miocene</td>
<td>Past mining produced 21–26 Kt of Fe-Pb-Zn ore; Resources: 0.1 Mt of Fe-Pb-Zn</td>
<td>Melidonis and Constantinides, 1983; Voudouris et al., 2014</td>
</tr>
<tr>
<td><strong>Syros</strong></td>
<td>Attic-Cycladic ore belt</td>
<td>Attic-Cycladic crystalline belt</td>
<td>Pb Zn Fe Cu Mn As Sb Bi Te Cd Ag Au</td>
<td>Carbonate replacement, vein</td>
<td>Dissemination, massive bodies, veins</td>
<td>Marble, schist, gneiss</td>
<td>Miocene</td>
<td>Past mining produced 21–26 Kt of Fe-Pb-Zn ore; Resources: 0.1 Mt of Fe-Pb-Zn</td>
<td>Melidonis and Constantinides, 1983; Voudouris et al., 2014</td>
</tr>
</tbody>
</table>
Table 4
Characteristics of most significant Tertiary intrusion related Au & Mo-Cu-W mineralizations in Greece. HR: Host Rock; ALT: Alteration; MIN: Mineralization; py: pyrite.

<table>
<thead>
<tr>
<th>Deposit name Ore district</th>
<th>Geotectonic belt or unit</th>
<th>Commodities</th>
<th>Deposit style</th>
<th>Morphology of ore bodies</th>
<th>Main host rocks</th>
<th>Age: HR, ALT, MIN</th>
<th>Tonnage and grades</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Reduced intrusion-related gold systems in the Rhodope Massif</strong></td>
<td></td>
<td></td>
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<tr>
<td>Palea Kavala Palea Kavala</td>
<td>Rhodope Massif</td>
<td>Fe Cu Pb Zn Mn Cd As Sb Bi Te Ag Au</td>
<td>Intrusion hosted veins, carbonate replacement, vein</td>
<td>Veins, disseminated, massive fissures (pods), mantos, breccia</td>
<td>Granodiorite, marble, schist</td>
<td>Miocene (HR-granodiorite: 21.1 ± 0.8 to 19.7 ± 0.3 Ma)</td>
<td>Resources: 1.5 Mt containing up to 34.5 g/t Au, up to 180 g/t Ag, up to 13% Pb + Zn, up to 40% Fe and up to 42% Mn</td>
<td>Fornadel et al., 2011</td>
</tr>
<tr>
<td>Pangeon</td>
<td>Rhodope Massif</td>
<td>Fe Pb Zn Cu Mn As Sb Cd Bi W Te Ag Au</td>
<td>Intrusion hosted veins, carbonate replacement, vein</td>
<td>Mantos, veins</td>
<td>Granodiorite, marble, schist</td>
<td>Miocene (HR-granodiorite: 21.7 ± 0.5 to 18.8 ± 0.6 Ma)</td>
<td></td>
<td>Eliopoulos and Kilias, 2011; Vaxevanopoulos, 2017</td>
</tr>
<tr>
<td><strong>Intrusion-related Mo-Cu-W deposits</strong></td>
<td></td>
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<tr>
<td>Kimmeria</td>
<td>Rhodope Massif</td>
<td>Cu Mo W Bi</td>
<td>Vein</td>
<td>Veins</td>
<td>Granodiorite</td>
<td>Oligocene (HR-granodiorite: 28.8 ± 0.7 to 26.3 ± 0.1 Ma)</td>
<td></td>
<td>Walenta and Pantzartzis, 1969; Voudouris et al., 2010; Theodoridou et al., 2016</td>
</tr>
<tr>
<td><strong>Other intrusion-related (?) polymetallic veins</strong></td>
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<tr>
<td>Rhodope Massif</td>
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</tr>
<tr>
<td>Kallintiri</td>
<td>Circum-Rhodope Belt Rhodope Massif</td>
<td>Sb Pb Zn Ag Au Te Hg Ti Cu Ag Fe As Bi Te Ni Au</td>
<td>Detachment fault related Carbonate replacement, vein</td>
<td>Veins</td>
<td>Marble, schist, sediments Schist</td>
<td>Oligocene?</td>
<td></td>
<td>Kanellopoulos et al., 2014</td>
</tr>
<tr>
<td>Thasos</td>
<td>Thasos</td>
<td>Cu Ag Pb Te Bi</td>
<td>Shear-zone hosted</td>
<td>Massive, veins, disseminated</td>
<td>Orthogneiss, amphibolite, metagabbro, schist</td>
<td>Miocene (MIN-py: 19.2 ± 2.1 Ma)</td>
<td></td>
<td>Vavelidis and Melfos, 2004</td>
</tr>
<tr>
<td>Stamos</td>
<td>Thermos-Volvi-Gomati complex Rhodope Massif</td>
<td>Cu Au Ag Bi Te</td>
<td>Shear-zone hosted</td>
<td>Massive, veins, disseminated</td>
<td>Gneiss, amphibolite</td>
<td>Oligocene-Miocene?</td>
<td></td>
<td>Vavelidis and Tarkian, 1995</td>
</tr>
<tr>
<td>Nea Madytos</td>
<td>Thermos-Volvi-Gomati complex</td>
<td>Cu Au Ag Pb Te Bi</td>
<td>Shear-zone hosted</td>
<td>Massive, veins, disseminated</td>
<td>Gneiss, amphibolite</td>
<td>Oligocene-Miocene?</td>
<td></td>
<td>Vavelidis et al., 1999</td>
</tr>
<tr>
<td>Drakontio</td>
<td>Kilkis Vertiskos Unit</td>
<td>Cu Pb Zn Ag Au</td>
<td>Shear-zone hosted</td>
<td>Massive, veins, disseminated</td>
<td>Gneiss, schist, amphibolite</td>
<td>Oligocene-Miocene?</td>
<td></td>
<td>Vavelidis et al., 1996</td>
</tr>
<tr>
<td>Koronouda</td>
<td>Kilkis Vertiskos Unit</td>
<td>Cu Au Ag Zn Pb Fe As Ni Co Sh Bi Te Bi</td>
<td>Shear-zone hosted</td>
<td>Massive, veins, disseminated</td>
<td>Gneiss, pegmatoids</td>
<td>Oligocene-Miocene?</td>
<td></td>
<td>Vavelidis et al., 1996</td>
</tr>
<tr>
<td>Stefania</td>
<td>Kilkis Vertiskos Unit</td>
<td>Cu Ag Au Bi Te Co Ni As</td>
<td>Shear-zone hosted</td>
<td>Massive, veins, disseminated</td>
<td>Gneiss, amphibolite</td>
<td>Oligocene-Miocene?</td>
<td></td>
<td>Melfos et al., 2001</td>
</tr>
<tr>
<td>Laodikino</td>
<td>Kilkis Vertiskos Unit</td>
<td>Cu Au Fe As Zn Pb Te Co Ni Sb Bi</td>
<td>Shear-zone hosted</td>
<td>Massive, veins, disseminated</td>
<td>Gneiss, amphibolite</td>
<td>Oligocene-Miocene?</td>
<td></td>
<td>Vaxevanopoulos, 2017</td>
</tr>
<tr>
<td><strong>Attic-Cycladic Belt</strong></td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kallianou</td>
<td>Attic-Cycladic ore belt</td>
<td>Fe As Pb Cu Zn Sb Te Cd Ag Au</td>
<td>Detachment fault related</td>
<td>Veins, dissemination</td>
<td>Schist, marble</td>
<td>Miocene</td>
<td></td>
<td>Voudouris et al., 2011b</td>
</tr>
<tr>
<td>Tinos</td>
<td>Attic-Cycladic ore belt</td>
<td>Fe As Pb Cu Zn Sb Te Cd Ag Au</td>
<td>Detachment fault related</td>
<td>Epithermal veins</td>
<td>Veins</td>
<td>Marble, schist</td>
<td>Miocene (HR: ~18–15 Ma)</td>
<td></td>
</tr>
<tr>
<td>Sifnos</td>
<td>Attic-Cycladic ore belt</td>
<td>W Fe Cu Au Pb Zn Mn Mn</td>
<td>Shear zone and detachment fault related</td>
<td>Veins, massive, lenses</td>
<td>Gneiss, schist, marble</td>
<td>Oligocene-Miocene</td>
<td></td>
<td>Vavelidis 1997; Neubauer 2005</td>
</tr>
<tr>
<td>Antiparos</td>
<td>Attic-Cycladic ore belt</td>
<td>Ag</td>
<td>Veins</td>
<td>Veins</td>
<td>Marble, schist</td>
<td>Oligocene-Miocene</td>
<td></td>
<td>Kevrekidis et al., 2015</td>
</tr>
</tbody>
</table>
A 40 m-thick oxidation zone with malachite, azurite and limonite is underlain by a 2–3 m-thick enrichment zone, where covellite, chalcocite, chalcopyrite, pyrite and magnetite coexist with malachite and azurite. At least three phases of monzonite porphyry are recognized, which are related to intense potassic and propylitic alteration.

Other porphyry deposits in the Kassandra mining district occur at Fisoksara, Alatina, Tsikara, Aspra Chomata, and Dilolo where extensive alteration zones are associated with porphyritic intrusions of granodioritic, syenitic, monzodioritic, and dioritic compositions. Whole-rock K-Ar ages have been obtained at the Tsikara and Fisoksara porphyry systems by Tompouloglou (1981). In Tsikara, the monzodiorite porphyry was dated at 21.2 ± 1.03 Ma and in Fisoksara the diorite porphyry has an age of 24.55 ± 1.22 Ma; a K-Ar date for sericite gave an age of 23.01 ± 1.15 Ma. However, recent zircon U-Pb geochronology indicates late Oligocene ages (27–25 Ma) for both the Tsikara and Fisoksara intrusive rocks (Siron et al., 2016). Hydrothermal alteration is dominated by sericitic alteration, which is locally superimposed on propylitic and a potassic alteration.

In the Kilkis district, the most important porphyry deposit is located at Vathi, which is estimated to contain 15 Mt ore with grades of 0.30% Cu and 0.8 g/t Au (Veranis and Tsamantouridis, 1991; Stergiou et al., 2016). The mineralization is associated with quartz monzonite porphyry dikes with ages of 18 ± 0.5 to 17 ± 1 Ma (U-Pb zircon; Frei, 1992). The mineralization is associated with potassic and propylitic alteration and contains pyrite, chalcopyrite, galena, bornite, sphalerite, molybdenite, magnetite, and hematite. A subsequent mineralization stage is related to quartz-tourmaline-sericite veins and breccias and is composed mostly of disseminated pyrite, arsenopyrite, and molybdenite (Fig. 5c). Porphyry-style mineralization was overprinted by high-to-intermediate-sulfidation epithermal quartz veins, which are spatially associated with sericitic and argillic alteration. Bulk chemical analyses of surface samples from quartz stockworks yielded up to 780 ppm Cu, 335 ppm Mo, 0.73 ppm Au, up to 330 ppm U, up to 500 ppm La, and up to 715 ppm Ce (Stergiou et al., 2016).

The Gerakari porphyry-epithermal system is located very close to Vathi, with probable reserves of 28 Mt grading 0.4% Cu and 0.9 g/t Au (Tsirambides and Filippidis, 2012). Copper mineralization is associated with potassically altered syenite porphyry and a sericitically and propylitically altered granodiorite porphyry. U-Pb ages obtained on zircons from the syenite porphyry and granodiorite porphyry yielded ages of 22 ± 0.8 Ma and 34 ± 0.5 Ma, respectively (Frei, 1992). The mineralization occurs as pyrite-chalcopyrite disseminations and magnetite-pyrite-chalcopyrite veins. Stibnite bearing quartz veins in the adjacent gneiss are attributed to a subsequent epithermal stage.

3.1.3. Porphyry deposits in the northeastern Aegean region

In the northeastern Aegean region, the Limnos and Lesvos Islands host at least three Miocene porphyry-style deposits at Fakos, Sardeas and Stipsi respectively (Figs. 2 and 3; Voudouris and Alfieris, 2005; Fornadel et al., 2012). The Fakos deposit is hosted by a ~20 Ma quartz monzonite porphyry stock (Fornadel et al., 2012). Early stage A- and B-type quartz veinlets are associated with potassic and propylitic alteration and contain pyrite, chalcopyrite, galena, bornite, sphalerite, molybdenite, magnetite, and hematite. A subsequent mineralization stage is related to quartz-tourmaline-sericite veins and breccias and is composed mostly of disseminated pyrite, arsenopyrite, and molybdenite (Fig. 5c). Porphyry-style mineralization was overprinted by high-to-intermediate-sulfidation epithermal quartz veins, which are spatially associated with sericitic and argillic alteration. Bulk chemical analyses of surface samples from quartz stockworks yielded up to 780 ppm Cu and up to 83 ppm Mo.

The Sardeas porphyry system in the northwestern part of Limnos Island comprises a stockwork quartz veinlets containing pyrite and molybdenite, hosted within a Miocene monzonite porphyry and
quartz-sericite-tourmaline-altered sedimentary rocks (Voudouris and Alfieri, 2005).

The Stipsi porphyry Cu-Mo prospect in the north-central part of Lesvos Island is hosted by a high-K calc-alkaline granodiorite porphyry and the surrounding volcanic rocks. It is characterized by intense propyritic, sericite-carbonate, argillic, and at the upper levels, advanced argillic and silicic alteration (Fig. 5d; Voudouris and Alfieri, 2005). Maximum concentrations based on bulk chemical analyses from surface samples of the Stipsi prospect were 0.48 ppm Au, 1330 ppm Cu, 170 ppm Mo, and 1.7 ppm Ag. The porphyry-style mineralization consists of dense quartz stockworks containing molybdenite, bismuthinite, and galena (Fig. 5e), overprinted by intermediate-sulfidation epithermal quartz-carbonate veins (Voudouris and Alfieri, 2005).

3.1.4. Porphyry deposits in the Cyclades

In the Cyclades, porphyry Mo-W mineralization occurs as sheeted quartz veins and stockworks cutting a late Miocene granodiorite stock in the Plaka area of the Lavrion mining district (Figs. 2, 4, and 5f; Altherr and Siebel, 2002; Voudouris et al., 2008a). The sheeted veins strike generally NW-SE and their thickness reaches 40 cm. The mineralization consists of pyrite, molybdenite, chalcopyrite, pyrrhotite and minor scheelite. Quartz, hydrothermal biotite, K-feldspar and sericite are gangue minerals. The granodiorite was intensely altered to potassic, sodic, propylitic, and sericitic assemblages, locally with silicification. Chemical analyses of surface ore samples revealed up to 1200 ppm Mo, 760 ppm W, and 3 wt.% Fe. The granodiorite intruded the footwall of a detachment fault. Hydrothermally altered and locally mineralized variably deformed felsic dikes and sills crosscut both the granodiorite and the metamorphic footwall and hanging wall rocks, and occur along the detachment fault, indicating that the intrusion of the magma took place during the detachment movement. Geochronological dating of the granodiorite and porphyritic dikes and sills based on K-Ar ages on biotite, feldspar and whole rock, fission track age from apatite, and U-Pb and (U-Th)/He ages on zircon, provided an age of 9.4–7.1 Ma for the igneous activity in Lavrion (Altherr et al., 1982; Skarpelis et al., 2008). Zircon (U-Th)/He thermochronometry from footwall granitoids at Plaka area between 7.9 ± 0.6 and 3.2. Epithermal deposits

3.2.1. Epithermal deposits in the eastern Rhodope Massif

Several polymetallic epithermal deposits and prospects containing Au, Ag, Cu, Te, Se, Bi, In, and Ga occur in the Greek part of the eastern Rhodope Massif (Figs. 2 and 3). These deposits are mostly high-sulfidation Cu-Au-Ag systems (Perama Hill, Mavrokoryfi, Sapes-Kassiteres, Peftka, Kalotycho) which evolved to Pb-Zn-Ag-Au-Te high-sulfidation epithermal mineralization is superimposed upon and also developed in the high- and the intermediate-sulfidation environment (Ortelli et al., 2009, 2010). The St. Demetrios high-to-intermediate sulfidation epithermal system is superimposed upon and also developed in the periphery of porphyry-style mineralization (Voudouris et al., 2006; Voudouris, 2014; this study). Preliminary \(^{40}\)Ar/\(^{39}\)Ar data of magmatic-steam alunite and adularia from the Sapan and St Barbara deposits suggest almost identical ages (e.g. 31.2 ± 0.4 to 31.9 ± 0.6 Ma) for the ore formation in the high- and the intermediate-sulfidation environment (Ortelli et al., 2009, 2010). The St. Philippos intermediate-to-high sulfidation epithermal Pb-Zn-Ag-Bi-Sn-In deposit (Fig. 3) is located in the Kirki ore field and is genetically related to a microgranite porphyry (Voudouris et al., 2013b). It contains an unusual ore mineralogy consisting of several Pb-As-Cu-Ag-Bi-Sn sulfosalts. Dickite, alunite, pyrophyllite, barite, and calcite are the main alteration minerals. Wurtzite and sphalerite contain considerable amounts of In (up to 3.5 wt.%), and Ga (up to 1.6 wt.%), whereas Ge is present in a smaller amounts (up to 0.3 wt.%). Driesner and Pintea, 1994). Past production has been 0.2 Mt of ore at 7.5% Pb + Zn.

The Peftka high-to-intermediate sulfidation epithermal Cu-Au-Ag-Te deposit located in the Evros mineralization district (Fig. 3) is hosted in andesitic to rhyolitic volcanic rocks. A whole rock K-Ar age of 30.7 ± 1.2 Ma was reported for a trachyandesite (Christofides et al., 2004). The main alteration styles are silicification, sericitization, and advanced argillic alteration, which are crosscut by late carbonate-bearing veins related to E- and
NNW-trending faults. Two distinct mineralization styles are observed in Pefka (Voudouris, 2006; Repstock et al., 2015): (1) early high-sulfidation veins with enargite, Bi-sulfosalts, and gold; and (2) late intermediate-sulfidation veins with tennantite/tetrahedrite and Au-Ag tellurides. Bulk analyses of mineralized samples from both epithermal mineralization styles contain up to 10 ppm Au, up to 23.5 ppm Mo, up to 105 ppm Bi, up to 468 ppm Te, up to 675 ppm In, 17 ppm Ga, 6 ppm Ge, >100 ppm Ag, >1 wt.% Cu and >1 wt.% As (Melfos and Voudouris, 2012; this study). Past mining operations produced 3 Kt ore at 7% Cu.

The Loutros area, close to the Greek-Turkish border (Fig. 3), hosts an intermediate-sulfidation epithermal deposit associated with a zeolite-altered rhyolitic lava dome, which yielded which yielded a whole rock K-Ar age of 19.53 ± 0.75 Ma (Christofides et al., 2004). The mineralization consists mainly of early NW-trending massive pyrite and marcasite veins and breccias, and late-stage barite-galena veins with sphalerite and minor chalcopyrite. The barite-galena veins contain up to 31 ppm Ag.

The Kalotycho high-sulfidation epithermal system covers a large area along the Greek-Bulgarian border (Fig. 3), and is hosted by Oligocene calc-alkaline and shoshonitic volcanic rocks. The volcanic rocks include andesites, dacites, trachytes, rhyodacitic ignimbrites and rhyolites, which were emplaced in a NE-SW trending sedimentary basin containing Eocene-Oligocene basal conglomerates and sandstones (Eleftheriadis, 1995). Whole-rock K-Ar ages of the volcanic rocks range from 33.0 ± 1.2 to 24.6 ± 0.6 Ma (Innocenti et al., 1984). The high-sulfidation epithermal alteration and mineralization are developed along NW-trending faults and are characterized by massive pyrite mineralization and advanced argillic, silicic-argillic, and propylitic alteration (Voudouris and Melfos, 2012).

3.2.2. Epithermal deposits in the northeastern Aegean region

In the northeastern Aegean islands of Limnos and Lesvos, epithermal mineralization is superimposed on or occurs peripherally to porphyry systems (Voudouris and Alfieris, 2005). The Fakos Au-Ag-Te high- to intermediate-sulfidation epithermal deposit on Limnos Island is associated with ~20 Ma quartz monzonite and shoshonitic subvolcanic rocks which intruded Eocene to Miocene sedimentary rocks (Figs. 2 and 3; Voudouris and Alfieris, 2005; Fornadel et al., 2012). The epithermal mineralization is characterized by polymetallic veins containing pyrite, chalcopyrite, sphalerite, galena, enargite, bournonite, tetrahedrite-tennantite, hessite, petzite, altaite, an unknown cervelleite-like Ag-telluride,
Native Au, and Au-Ag alloy (Voudouris and Alfieris, 2005). Bulk chemical analyses of surface samples returned up to 11 ppm Au.

The Megala Therva Pb-Zn-Cu-Ag-Au intermediate-sulfidation epithermal deposit on Lesvos Island, is related to NNE–SSW- and NW–SE-trending quartz veins hosted in 21–18 Ma old andesite-latite lavas (Fig. 3; Kontis et al., 1994). Silicic, propylitic, argillic, and adularia-sericite are the main alteration types. The veins are enriched in Au (up to 21 ppm), Ag, Pb, Zn, Cu and Mo (Kontis et al., 1994). The Megala Therva epithermal system extends further to the south, where similar intermediate-sulfidation quartz-carbonate veins with platy calcite (Fig. 6b) overprint the porphyry Cu-Mo mineralization at Stipsi (Voudouris and Alfieris, 2005).

3.2.3. Epithermal deposits in the Cyclades

The island of Milos is located in the central part of the early Pliocene to Recent South Aegean Active Volcanic Arc, and belongs to the Cycladic Blueschist Unit of the Cyclades (Fytikas et al., 1986). Calc-alkaline volcanic activity in western Milos spans a period from ~3.5 to 0.9 Ma and originated from several emergent eruptive centers that produced submarine felsic pyroclastic deposits, pumice flows, dacitic-andesitic flow domes, lava flows, and felsic subvolcanic rocks (Fytikas et al., 1986; Stewart and McPhie, 2006; Alfieris et al., 2013). Milos Island is one of the most densely mineralized areas in Greece, characterized by intermediate-sulfidation epithermal Au-Ag-Te and base metal deposits (Fig. 6c) under transitional shallow submarine to subaerial conditions (Vavelidis and Melfos, 1998; Kilias et al., 2001; Naden et al., 2005; Alfieris et al., 2013; Papavassiliou et al., 2017). Epithermal-type mineralization rich in Au-Ag-Te occurs at Profitis Ilias (reserves: 5 Mt at 4.4 g/t Au) and Chondro Vouno (reserves: 3.3 Mt at 4.2 g/t Au), whereas Pb-Zn-Cu-Ag-Mn-Ba-rich mineralization occurs at in the Triades-Galana (resources: 1.2 Mt at 1 g/t Au and 124 g/t Ag) and Kondaros-Katsimouti-Vani districts (Fig. 4). The deposits are closely associated with active geothermal systems that are characterized by mixing of seawater, meteoric and, minor magmatic water (Naden et al., 2005; Alfieris et al., 2013; Papavassiliou et al., 2017).

3.3. Carbonate-hosted replacement and skarn deposits

3.3.1. Carbonate-hosted replacement and skarn deposits in the Rhodope Massif

The Olympias Au-Ag-Pb-Zn carbonate-replacement polymetallic deposit (Fig. 3; Siron et al., 2016) is hosted mainly within the lower calcitic marble horizon of the Kerdylion Unit and along the contact with the upper biotite gneiss. It forms mainly massive stratabound (manto), disseminated and cavity- or fracture-filling ore bodies consisting of various proportions of pyrite, asarsonopyrite, chalcopyrite, marcasite, sphalerite, galena, bournonite, boulangerite, (Kalogeropoulos et al., 1989). Deformed sulfide ore is also present and is characterized by brecciation, mylonitization, folding and shearing. Mineralogical and geochemical features of the Olympias mineralization in both deformed and undeformed ores styles suggest that they were derived from the same system and were deposited after the onset of regional metamorphism (Kalogeropoulos et al., 1989; Gilg, 1993). Two major parts of the ore body are distinguished: the western section is approximately 250 m wide and extends 1500 m to the southwest dipping 30–35° to the east, whereas the eastern ore body is ~75 m wide and dips 25–30° to the southeast (Forward et al., 2011). The proven and probable reserves are 16.1 Mt containing 4.3% Pb, 5.7% Zn, 128 g/t Ag and 7.9 g/t Au, for a total of 66.3 Moz Ag and 4.1 Moz Au (Eldorado Gold Corp., 2017). The Olympias deposit is currently in development with a planned annual production of approximately 190,000 oz Au per year by 2018, and with an expected mine life of over 25 years. The deposit is open at depth and the mineralization extends approximately 790 m below sea level. The majority of gold is “invisible”, incorporated in arsenian pyrite and arsenopyrite (average ~50 ppm for both minerals; Chrysosouli and Cabri, 1990). Re-Os dating of arsenopyrite yielded an age of 26.1 ± 5.3 Ma (Hahn et al., 2012).

Two major carbonate-replacement deposits occur in the Kerdylion Unit along the Stratoni fault, at Madem Lakkos and Mavres Petres (Fig. 3). The Madem Lakkos has been exhausted, but past mining produced 13.5 Mt of Ag-Pb-Zn ore. Reserves at Mavres Petres are estimated to be 1.87 Mt at 160 g/t Ag, 6% Pb and 8.8% Zn (Eldorado Gold Corp., 2017). Two ore stages occur at the two deposits: early replacement-style massive sulfide forming undeformed and formed (folded, sheared, brecciated) ore bodies consisting dominantly of galena and sphalerite with lesser pyrite and arsenopyrite, is overprinted by pyrite-rich disseminations that form the matrix to breccias and surrounds previous base metal-rich sulfide mineralization (Siron et al., 2016). Quartz, calcite and minor rhodochrosite form the gangue minerals. The average gold grade is ~5 g/t Au and is related to the arsenian pyrite and arsenopyrite as “invisible” gold (Forward et al., 2010).

The Piatvista prospect is a siliceous-manganese carbonate-replacement deposit with associated Au-rich veins with a typical epithermal affiliation, located at the Stratoni area (Fig. 3; Siron et al., 2016). Piatvista was the focus of exploration drilling by Eldorado Gold in 2012, which resulted in the discovery of a new resource estimated to be 10.54 Mt at 5.7 g/t Au and 57 g/t Ag (Eldorado Gold Corp., 2017). The Piatvista ore deposit occurs along and adjacent to the Stratoni fault and was explored and exploited in the 1960s for manganiferous at the surface. The mineralization styles at Piatvista vary and include brecciated semi-massive to massive sulfide lenses, banded epithermal quartz-rhodochrosite-gold veins, and hydrothermal breccias containing clasts of altered marble and massive sulfide (Siron et al., 2016).

The Thermes prospect is a polymetallic Zn-Pb-Fe-Cu-As-Ag-Au-Te carbonate-replacement deposit located in Xanthisi area, NE Greece, at the southernmost part of the Arda dome (Fig. 3; Arvanitidis and Dimou, 1990; Kalogeropoulos et al., 1996). Veins and breccias containing Pb-Zn-(Fe-Cu) mineralization also occur, associated with NWW- and NNE-trending faults. The deposit contains a resource of ~1 Mt with grades of <4 g/t Au, 10–340 g/t Ag, 1.2–14.5% Pb, 2.1–16.7% Zn and 0.06–0.10% Cu (Gialoglou and Drynniotsi, 1983). Thermes represents the southern extension of the Madan ore field in Bulgaria (Kaiser-Rohrmeier et al., 2013).

At Kimeria a massive Cu-W-Mo-Au-bearing skarn-type mineralization consisting of chalcopyrite, magnetite, pyrrhotite, scheelite, and minor molybdenite and gold, is genetically related to the Xanthi I-type pluton (Fig. 3; Walenta and Pantzartzis, 1969; Vavledis et al., 1990; Voudouris et al., 2010). The Xanthi pluton ranges in composition from gabbro through monzonite to granodiorite, and has an Oligocene age (28.8 ± 0.7 and 26.3 ± 0.1 Ma; Rb-Sr in whole rock and biotite; Kyriakopoulos, 1987). The emplacement of the pluton is controlled by two major structural systems: the Kavala-Xanthi-Komotini normal fault and the Nestos thrust fault (Fig. 3). The pluton intrudes gneisses, mica schists, amphibolites and marbles of the Northern Rhodope Domain. Mining operations in the 1930s, including underground galleries as well as surface excavations, extracted the magnetite-pyrrhotite-rich skarn mineralization and the Cu-Mo ore. The remaining resource is estimated to be ~2 Mt with 20 g/t Au and 1.98% Cu (Gialoglou and Drynniotsi, 1983).

Thasos Island contains silver-rich Pb-Zn carbonate-replacement deposits hosted in rocks of the Southern Rhodope Core Complex (Fig. 3; Vavledis and Amstutz, 1983). The deposits were mined for silver and gold during antiquity, and for Fe, Pb and Zn between 1905 and 1964. These past mining operations produced a total of 2 Mt of ore containing 12% Zn + Pb, and 3 Mt of ore containing
44% Fe (Gialoglou and Drymniotis, 1983). Vavelidis and Amstutz (1983) distinguished four Pb-Zn-Ag-bearing horizons in carbonate rocks (marbles and dolomites) and schists, located on the eastern and the southern parts of the island. The mineralization occurs as veins and mantos of massive to semi-massive sulfides in the carbonate rocks, and veins crosscutting the schists. The main ore minerals are galena, sphalerite, pyrite, marcasite, arsenopyrite, chalcopyrite, cerussite, smithsonite, and Fe-Mn oxides; gangue minerals include calcite, ankerite, siderite, barite, and quartz.

3.3.2. Carbonate-hosted replacement and skarn deposits in the Attic-Cycladic belt

In the Attic-Cycladic Belt, the Lavrion ore district (Figs. 2 and 4) is a world-class carbonate-replacement deposit. It covers an area of ~150 km², and is famous for the exploitation of Pb-Ag-rich ore during ancient times, mainly during the Classical period, from 6th to 4th century B.C. (Conophagos, 1980). According to ancient writers (e.g., Aeschylus, Herodotos, Xenophon, Strabo) the Lavrion mines contributed significantly to the power of ancient Athens. After the decline of Athens, the mines were closed and then operated again in the 19th century. The exploitation of sulfide ores continued until the 1970s. Conophagos (1980) estimated that ~2.3 million metric tons of Pb and 7800 metric tons of Ag were extracted during the antiquity era. Remaining resources are estimated to be 4 Mt of ore containing 7% Pb + Zn. The carbonate-replacement Pb-Zn-Ag mineralization is the most economically important in the Lavrion district and occurs in the form of strata-bound lenses; bedded replacements (mantos), and chimneys up to tens of meters in length, in the Kamariza and Soumin areas (Marinos and Petrascheck, 1956; Voudouris et al., 2008a,b; Bonsall et al., 2011). According to Voudouris et al. (2008a,b), Bonsall et al. (2011), and Berger et al. (2013) the ore deposits in Lavrion are genetically related to the Miocene Lavrion granitoids, and ore formation occurred under extensional kinematic conditions. The carbonate-replacement and vein mineralogy is dominated by pyrite, arsenopyrite, sphalerite, chalcopyrite and galena (Fig. 6d), with various silver sulfosalts and native gold (Voudouris et al., 2008b), and quartz, fluorite, calcite and sericite gangue. Skarn deposits also occur around the Lavrion granodioritic body, and consist of early magnetite followed by pyrrhotite, pyrite, arsenopyrite, sphalerite, chalcopyrite, galena, bismuthinite, tetradymite, and native bismuth (Voudouris et al., 2008a,b; Bonsall et al., 2011). Extensive zones of supergene mineralization occur above the primary mineralization zones (Marinos and Petrascheck, 1956; Skarpelis and Argyraki, 2009). Various critical and rare elements occur in the mineralization, including Ag, Au, Bi, Sn, Ga, Ge, In, Sb, W, Ni, Te, and Se (Voudouris et al., 2008a,b; Zaimis et al., 2016). The sulfur isotope compositions of sulfides ($\delta^{34}S = -4.9$ to $+5.3\%$) in carbonate-replacement and vein style mineralization, along with the presence of elevated Ag, Bi, As, Sn, In, and Sb contents, point to a magmatic source for those elements, but is unclear whether the source was the Plaka granodiorite, the mafic to felsic dikes, or a granitoid at depth (Bonsall et al., 2011).

On Serifos Island (Fig. 4), a skarn deposit was mined for iron in antiquity and from 1869 to 1940, producing 6.6 Mt of ore. The mineralization is associated with a shallow intrusion of I-type granodiorite with an age of 11.6 ± 0.1 to 9.5 ± 0.1 Ma, based on zircon U–Pb dating (Iglsereder et al., 2009). The granodiorite was intruded along the Serifos detachment fault (Ducoux et al., 2017). The ore consists of pyrite, chalcopyrite, arsenopyrite, galena, sphalerite, tennantite, and is extensively oxidized. Calcite, barite, garnet, pyroxene, epidote, fluorite, talc, chlorite, adularia are the main gangue minerals.

The Sifnos Island carbonate-replacement Pb-Zn-Ag-Au deposits occur within marbles of the Cycladic Blueschist Unit (Fig. 4). Three main types of mineralization are distinguished in Sifnos (Vavelidis et al., 1985; Vavelidis, 1997; Neubauer, 2005): (1) Subconcordant ore bodies along shear zones composed mainly of pyrite and chalcopyrite; (2) Au-bearing (6 ppm on average) quartz veins in the footwall of the detachment faults with pyrite, chalcopyrite, pyrrhotite, gold, and various undefined Ag-minerals; (3) Fe-Mn-Pb-Zn-Ag mineralization related to a major normal fault forming subhorizontal layers and lenses or irregular oxidized ore bodies in the marble. Sulfur isotope analyses in pyrites from the quartz veins ($\delta^{34}S = +1.59$ to $+1.73\%$) show a magmatic origin of the sulfur. On Sifnos Island, the various lithological units (marbles, schists, etc.) are separated by several shear zones and detachment faults, dominated by the Sifnos detachment (Avgad, 1993; Ring et al., 2011). Sifnos has been undergoing normal faulting from 30 to 23 Ma, and top-to-the-SSW extension along the Sifnos detachment was nucleated in the brittle-ductile transition zone at 13 Ma or slightly earlier, and continued to ~10 Ma (constrained by zircon fission track ages; Brichau et al., 2010). Structural relationships suggest that ore formation occurred under extensional kinematic conditions when the Sifnos metamorphic core complex reached a near-surface level during the late Miocene (~11 Ma; Neubauer, 2005, 2007).

Polymetallic Pb-Zn-Ag-Au mineralization in the southern part of Syros Island occurs as disseminations and massive sulfide bodies replacing marbles, along schist foliation planes and in veins crosscutting the marbles and schists in the Cycladic Blueschist Unit. The mineralization consists of base metal sulfides (pyrite, arsenopyrite, Fe-rich sphalerite, galena, pyrrhotite, stibnite, molybdenite), sulfoсалts (Ag-rich tetrahedrite, famatinite, bournonite, bismuthinite, cosalite, stannite), tellurides (bessite), and Fe-oxides (magnetite, hematite). Bulk ore analyses show elevated contents of Mo, Sn, Te, Se, and Hg, in addition to Pb, Zn, Ag, and Au. Voudouris et al. (2014) suggested an intrusion-related origin for the Syros mineralization, assuming the presence of a Miocene granitoid at depth that fed the system with volatiles and metals. Deformation and mineralization at Syros were contemporaneous, occurring under brittle-ductile conditions probably related to detachment faulting. The Syros mineralization resembles other intrusion-related systems in the Attic-Cycladic Belt such as the Lavrion deposit.

3.4. Reduced intrusion-related gold systems in the Rhodope Massif

Various ore deposits in the Rhodope Massif are suggested here to belong to the reduced intrusion-related gold deposit classification (sensu Hart et al., 2002; Baker et al., 2005), although they have previously been classified as different deposit types, e.g. carbonate-replacement type, shear zone hosted mineralization or metamorphic rock hosted veins. The most important reduced intrusion-related systems occur in the Southern Rhodope Core Complex (Fig. 3;Palea Kavala and Pangeon).

In the Palea Kavala area (Fig. 3), metamorphic rocks of the lower tectonic unit of the Southern Rhodope Core Complex were intruded by the ~22–19 Ma Kavala (or Symvolon) pluton (Christofides et al., 1998) along the trend of the Kavala-Komotini fault zone. The Kavala pluton is an I-type intrusion and is dominantly composed of amphibole-biotite granodiorite with subordinate amounts of diorite, tonalite, monzogranite, and monzodiorite. The Palea Kavala region contains ~150 base- and precious-metal occurrences within the Kavala pluton and the surrounding metamorphic rocks. They were exploited in ancient times for gold or lead and silver, and in modern times (early 20th century) for iron and manganese, with considerable exploitation by both underground and surface mining techniques. These occurrences have variable metal contents and are mostly weathered and oxidized (particularly those that contain Mn). The estimated total resources are 1.5 Mt containing up to
34.5 g/t Au, up to 180 g/t Ag, up to 13% Pb + Zn, up to 40% Fe, and up to 42% Mn (Chatzipanagis and Dimitroula, 1996). Oxidized Fe-Mn-Au and Fe-Mn (Pb ± Zn ± Ag) bodies are localized in marbles, whereas Fe-As-Au, Fe-Cu-Au, and Bi-Te-Au deposits occur in gneisses and granitoids, as well as along the gneiss-marble contact of the lower tectonic unit in milky quartz veins. The Miocene Kavala pluton is crosscut by an approximately 4 km-long sheeted-quartz vein system (Kavala vein) that contains Bi-Te-Pb-Sb-Au mineralization (Melfos et al., 2008; Fornadel et al., 2011) and the 30 m-long Chalkero quartz vein (Chalkero vein). The Kavala quartz vein system is characterized by several parallel to subparallel quartz veins ~1 m thick and tens of meters apart. The SW–NE orientation of the vein system reflects both large- and small-scale regional structures, such as the Kavala–Komotini fault. The Bi-Te-Au mineralization is dominantly hosted in the Kavala vein, though, rare pods of massive pyrite, up to 30 cm long, are found in the granodiorite (Fig. 6e).

On Pangeon mountain (Fig. 3), a reduced intrusion-related gold system contains Au-Ag ± Bi ± Z ± W mineralization, which is genetically associated with the Pangeon granitoids (Vaxevanopoulos, 2017). These granitoids (Nikisiani, Mesoropi, Podochori, Messolakkia bodies) are the surface exposures of the same pluton, which occupies the core of a large anticline and extends ~25 km in a SW-NE direction, intruded into the Southern Rhodope Core Complex along the Strymon detachment fault (Eletheriadis and Karones, 2003). The Pangeon granitoids consist of tonalite, granodiorite, and granite, and have high-K calc-alkaline gangue minerals consist of quartz, calcite, ankerite, and barite. These granitoids are associated with quartz boudins, veins and segregations, which show a slight zonation centered on the granitoids. The majority of this mineralization is oxidized and only small proportions contain primary ore minerals. Bulk sample analyses from Pangeon yielded up to 315 ppm W, up to 60 ppm Te, >2000 ppm Bi, >2000 ppm Sb, >100 ppm Ag, >100 ppm Au (Vaxevanopoulos, 2017).

3.5. Intrusion-hosted Mo-Cu-W deposits

The Oligocene Kimmernia intrusion-hosted Mo-Cu-W-Bi-Au deposit is located close to the Nestos thrust in the Rhodopes and is associated with the Oligocene Xanthi pluton (Figs. 2 and 3), described above in terms of its skarn mineralization. The Kimmernia ore district includes sheeted quartz veins crosscutting sericite-carbonate-altered granodiorite and contains pyrite, molybdenite, scheelite, wolfframite, and chalcopyrite (Fig. 6f; Walenta and Pantartzis, 1969; Voudouris et al., 2013; Theodoridou et al., 2015). Vein hosted mineralization contains 1 wt.% Cu, >0.2 wt.% Mo, up to 2.7 g/t Au, up to 79.5 g/t W, up to 456.6 g/t Bi, and traces of Te (up to 4 g/t) (Theodoridou et al., 2016). Primary fluid inclusion studies indicate CO$_2$ in the sheeted veins. Underground galleries as well as surface excavations aimed at the extraction of the Cu-Mo ore date back to the 1930s, but no production data are available. Similar Mo-Cu-W-Bi-Au veins occur also at the plutons of Samothraki and Tinos islands.

3.6. Other intrusion-related polymetallic vein deposits

Polymetallic quartz veins cutting metamorphic rocks in the Rhodope massif (e.g. Kallintiri, Thasos Island, Stanos, Nea Madytos, Drakontio, Koronouda, Stefania, and Laodikio) (Fig. 3) and Cyclades (e.g. Kallianou, Tinos Island, Sifnos Island, Antiparos Island, and “Vein 80” in Lavrion; Fig. 4), are thought to have formed in response to Tertiary extensional tectonics in the Aegean.

3.6.1. Intrusion-related polymetallic vein deposits in the Rhodope Massif

The Kallintiri deposit on the southwestern edge of the Biala Reka-Kechros metamorphic dome (Fig. 2) is characterized as a detachment-related, epithermal deposit, hosted in sillified marble and argillic-sericitic-altered schists of the Mesozoic Makri Unit, as well as in Paleogene supra-detachment sedimentary rocks (Kanellopoulos et al., 2014). This deposit shares some similarities to low-sulfidation detachment-related deposits in Bulgaria, but Kallintiri is Sb-rich with additional Pb-Zn-Ag-Au-Te-Hg-Tl. The mineralization was deposited at the brittle-to-ductile transition within and above a detachment fault, and occurs in the form of disseminations, quartz-barite-carbonate veins, and breccias (Kanellopoulos et al., 2014).

Thasos Island (Fig. 3), besides the previously mentioned Pb-Zn-Ag carbonate-replacement deposits, also hosts several Au-rich Cu-Bi-Au ± Z ± W deposits hosted within schist, gneiss, and amphibolite of the Southern Rhodope Core Complex in the northern and western parts of the island (Vavelidis and Amstutz, 1983; Vavelidis et al., 1995; Vavelidis and Melfos, 2004). Where veins crosscut marble, small massive sulfide bodies also form. The mineralization contains chalcopyrite, pyrite, magnetite, pyrrhotite, gersdorffite, bis-muthinite, tetrahedrite, tennantite, tetradyrmite, hessite, Fe- and Mn-oxides, and gold. Gold-grain size reaches up to 500 μm. The gange minerals consist of quartz, calcite, ankerite, and barite. The relatively high homogenization temperatures of fluid inclusions and the presence of Te in the Cu-Au-Bi-Ag mineralization (Vavelidis et al., 1995; Vavelidis and Melfos, 2004) may indicate a magmatic source at depth.

Shear zone-hosted Cu-Au-Pb-Bi-Au-Te deposits (e.g., Stanos, Nea Madytos, Drakontio, Koronouda, Stefania, and Laodikio) are found in metamorphic rocks of the Vertiskos Unit at Chalkidiki and Kilkis (Fig. 2). All these deposits share common mineralogical composition, fluid characteristics, and host rocks, but exploration projects are very limited or are lacking (Bristol et al., 2015).

The Stanos Cu-Au-Bi-Pb-Ag-Au-Te deposit in the NE Chalkidiki area (Fig. 3) is hosted in marble, amphibolite, gneiss, and metabasalt, and schist of the Vertiskos Unit, and is spatially related to regional NW–SE-trending shear zones (Voudouris et al., 2013a; Bristol et al., 2015). The mineralization forms disseminated to massive ore bodies along foliation planes and in boudinaged quartz veins (Fig. 6g). A Re-Os isochron age of 19.2 ± 1.1 Ma was obtained for pyrite in Stanos (Bristol et al., 2015).

Shear-related Au-Ag-Cu-Te-Bi mineralization in the NE Chalkidiki (Nea Madytos) and Kilkis (Drakontio, Koronouda, Stefania, Laodikio) areas (Fig. 3; Table 4) hosted by the Vertiskos Unit is considered to have a genetic affiliation to magmatic rocks of a possible Oligocene-Miocene age, although these intrusions are not exposed in the area (Bristol et al., 2015). These mineralizations are enriched in Au (up to 283.3 ppm) and also contain Ag (up to 765 ppm) and Cu (up to 4.3 wt.%), consisting mainly of chalcopyrite and pyrite, with traces of galena, sphalerite, Cu-Au-Ag-Bi sulfides, and gold (Thymiatis, 1995; Vavelidis and Tarkian, 1995; Vavelidis et al., 1996, 1999; Melfos et al., 2001). The ore bodies are associated with quartz boudins, veins and segregations, which were emplaced along shear and strike-slip zones, operated under ductile-to-brittle conditions.

3.6.2. Intrusion-related polymetallic vein deposits in the Attic-Cycladic belt

The Kallianou mining district in south Evia Island is famous for the exploitation of gold-silver-rich ore during ancient and recent times. Indicated mineral resources are 500,000 tons at an average
grade of 2–2.4% Pb, 0.7% Zn, 0.5–0.8% Cu, 35–60 ppm Ag and 5 ppm Au (Voudouris et al., 2011b and references therein). However, the sulfide-bearing quartz veins contain up to 9 ppm Au and 202 ppm Ag and the oxidized ore up to 30 ppm Au. Mineralization is hosted in mica schists and marbles of the Cycladic Blueschists and occurs either as dissemination within breciated marbles or as quartz veins that crosscut the foliation of the schists (Fig. 6h). The quartz veins (up to 3 m thick and 100 m long) generally strike NW-SE, are discordant to metamorphic structures and were formed close to the brittle-ductile transition (Voudouris et al., 2011b). Ore minerals in the quartz veins occur in masses to disseminations, filling fractures, or cementing breciated quartz fragments. There is a district-scale zonation at Kallianou with Au and Te enriched in the lower topographic levels and Ag at higher elevations (Voudouris, Unpublished data). This may suggest the presence of a buried granitoid at depth. The host rocks of the Kallianou deposit represent the footwall of the North Cycladic Detachment System (Ring et al., 2007; jolivet et al., 2010). High-P conditions of the Cycladic Blueschists at south Evia Island persisted until ~33 Ma, when the rocks started to be exhumed and finally re-equilibrated under greenschist-facies conditions at ~21 Ma (Ring et al., 2007). The veins were probably formed during the Miocene.

The vein hosted Au-Ag-Te rich mineralization at Tinos Island consists of quartz veins crosscutting Triassic marbles and schists of the Cycladic Blueschist Unit (Tombros et al., 2007). The mineralization contains electrum and an extremely rich suite of tellurides, including Au-Ag tellurides and silver sulfotellurides. Mineralization occurs in the footwall of the North Cycladic Detachment (Fig. 4). The intrusion of a granodiorite of ~18 to 15 Ma age was followed by the intrusion of a boron and fluorine-rich leucogranite peripheral to the granodiorite at ~14 Ma (Altherr et al., 1982). The mineralization at Panomos Bay is considered to be genetically related to the 14 Ma old, peraluminous leucogranite (Tombros et al., 2007).

On Sifnos Island, a vein hosted mineralization occurs in the southern part of the island, hosted in greenschists and gneisses of the Cycladic Blueschist Unit (Vavelidis, 1997; Vavelidis et al., 1985; Neubauer, 2005). Mineralization occurs as subcordant and subvertical pyrite-chalcopyrite-pyrrotite ore bodies with Fe-rich carbonates and quartz. These ores occur in the footwall of low-angle normal faults, which separate greenschists and gneisses from overlying marbles. The quartz veins formed during subhorizontal extension and their gold contents range from 1.5 to 12.2 ppm (Neubauer, 2005). E-W extension during movement of a NNE-directed semi-ductile normal fault in the hanging wall of the greenschist-gneiss unit was responsible for the formation of Au-bearing quartz veins.

Steeply dipping quartz veins containing galena with 800–2000 ppm Ag crosscut biotite-muscovite schists and marbles of the Cycladic Blueschist Unit at Agios Georgios of Antiparos Island (Kevrekidis et al., 2015). The mineralized veins were deposited from fluids in the epithermal stage with a significant magmatic contribution mixed with meteoric water. Kevrekidis et al. (2015) suggested that the mineralization is genetically related to the neighboring Miocene Paros leucogranite.

A major Pb-As-Sb-Cu-Ag rich banded vein with epithermal affinities (Fig. 6i), known as “Vein 80” or “Filoni 80”, crosses pyrrhotite bearing hornfelses and carbonate-replacement mineralization at Plaka area, Lavrion deposit (Voudouris et al., 2008a). The Filoni 80 vein trending ESE-WNW, is up to 2 m thick and up to 1 km long. This vein includes early deposition of pyrrhotite followed by arsenopyrite, löllingite, pyrite, marcasite, a Cu-Bi bearing assemblage including lillianite homologues, pyrrargyrite, chalcopyrite, Bi-bearing tetrahedrite-tennantite, bouronite, lead sulfantimonides and finally by galena and native arsenic. Quartz, siderite, fluorite and calcite are the main gangue minerals. An estimated total production of this vein is 90,000 t with an average Pb + Zn content of 14–15 wt.% and 500 ppm Ag (Conophagos 1980).

4. Discussion

4.1. Time-space relationships between detachment systems, magmatism and ore deposition

In the northern Rhodope core complexes of Bulgaria and Greece the Arda, Biała Reka-Kechros and Kesebir-Kardamos domes were exhumed from ~42 to 35 Ma along the ductile- to brittle Madan-Xanthi, Kechros and Tokachka-Kardamos low-angle detachment faults respectively (Boney et al., 2006a,b, 2010, 2013; Wüthrich, 2009; Moritz et al., 2010; Márton et al., 2010; Kaiser-Rohrmeier et al., 2013). Exhumation of the metamorphic rocks in the southern Rhodope core complex occurred along the ductile Kerdyilon detachment, active from ~42 Ma until ~24 Ma, and the ductile-to-brittle Strymon detachment from 24 Ma until 12 Ma (Brun and Sokoutis, 2007; Wüthrich, 2009; Kounov et al., 2015). As a consequence of the metamorphic core complex exhumation along large detachment faults, magmas were emplaced along the transition from ductile to brittle deformation related to shear zones attributed to post-orogenic exhumation (Jones et al., 1992; Kiliás et al., 2013a; Siron et al., 2016).

Around the Arda, Biała Reka-Kechros and Kesebir-Kardamos domes, the Oligocene porphyry Cu-Mo-Au deposits are associated with plutonic and subvolcanic intrusions either in the footwall of detachment fault systems that accommodated their exhumation (e.g., Maronia deposit, M. Sanchez, pers. commun., 2013), or located in supra-detachment grabens (Konus Hill, Pagoni Rachi, Myli, Melitena, etc.). In association with the porphyry deposits, the high-sulfidation style epithermal gold deposits (e.g., Sapes, Perama Hill etc) occur within the supra-detachment basins, where mineralization is controlled by steeply-dipping extensional faults and associated fractures. An interesting aspect for the Oligocene plutonic intrusions of Maronia, Kassiteres, Kirki and Leptokarya is their alignment along a NE-SW direction (Fig. 2), which parallels to the NE-trending axis of Oligo-Miocene intrusive centers in the Kassandra mining district and also to the NE-elongated Early Miocene Kavala pluton (Siron et al., 2016). This direction parallels the NE-SW back-arc extension and according to Siron et al. (2016), this indicates a northeast-oriented principal axis of extension evidenced by crustal-scale ruptures developed prior to Early Miocene and believed to have localized magmatism. At Kassandra mining district in Chalkidiki, the exhumation of the southern Rhodope core complex may have influenced ascent of magmas to shallow crustal levels, preferentially exploiting favorable oblique structural intersections between NE-trending transtensional faults and ENW-trending structures (Siron et al., 2016).

A subsequent metallogenic episode is associated with the magmatism of Early Miocene at ~22–19 Ma and is restricted: (a) in southern Rhodope core complex, either along the Strymon detachment fault (e.g., Pangeon and Palea Kavala reduced intrusion-related gold systems, and Thasos Pb-Zn carbonate-replacement deposits and intrusion-related (?) polymetallic veins), (b) along a NW-trending corridor through Chalkidiki and Kilkis regions including the Skouries, Vathi and Gerakario porphyry Cu-Au deposits, as well as the Stanos, Koronouda and Laodikino shear-zone intrusion-related(?) polymetallic deposits, (c) along a NE-trending direction at Loutrós north of the island of Samothraki, and (d) at Limnos and Lesvos Islands occurring in the western extension of Biga Peninsula.

The ductile shear zone at Stanos area occurs west of the Kerdyilon detachment and within the Athos-Volvi Suture Zone and coincides with Cu-Au ore deposition at ~19 Ma (Bristol et al., 2015).
This may indicate operation of a previously unrecognized NW-trending ductile detachment zone, which accommodate emplacement of granitoids and ore deposition in Chalkidiki and Kilkis during the Early Miocene. During the same period (~19 Ma) the Loutros rhyolites and the Samothraki granite were emplaced and exhumed during NE directed extension, like the coeval Kavala granodiorite beneath the Strymon Detachment (Dinter and Royden, 1993; Lips et al., 2000). Mineral deposits are classified as intrusion-hosted Cu-W with peripheral intrusion- and metamorphic rock-hosted Pb-Zn-Ag sulfide veins (Samothraki) or as epithermal styles at Loutros.

The porphyry-epithermal systems at Fakos and Sardes in Limnos Island, and Megala Therma and Stipsi in Lesvos Island occur along major E-W and NNE-trending steep-dipping extensional faults and associated fractures, like those considered to control porphyry and epithermal ore formation at Biga Peninsula (Sanchez et al., 2016).

During the Miocene back-arc extension of the Cyclades, stretching was accommodated mainly by the North Cycladic Detachment System, which is exposed at Mykonos, Tinos and further to Evia offshore Kallianou ore district (Jolivet et al., 2010) and the West Cycladic Detachment System which is exposed on Lavrion, Kea, Kythnos, and Serifos (Grasemann et al., 2012; Scheffer et al., 2016; Ducoux et al., 2017). The evolution of the Cyclades can be resolved into SSW- and NE directed ductile to a brittle extension, localized plutonism, and rapid cooling of the footwall of both detachment faults during Middle and Late Miocene. At Lavrion, the West Cycladic Detachment Fault operated under ductile-to-brittle conditions (Grasemann et al., 2012; Scheffer et al., 2016) and, based on U-Th/He dating of titanite, zircon andapatite (Seman et al., 2013a,b), accommodated exhumation of the metamorphic rocks in at least two major periods: Middle Miocene (16–12 Ma) and Late Miocene (6–9 Ma), the latter also coinciding with mineralization. Similarly to Lavrion, other mineral occurrences in the Cyclades are in part spatially associated with Miocene plutonic rocks, which intrudes the footwall units along detachment faults under ductile to brittle extensional kinematic conditions (e.g., Mykonos, Menant et al., 2013), in part by extensional structures in the footwall of detachment faults without any adjacent granitoids (e.g., Tinos, Sifnos, Syros, Kallianou) (Neubauer, 2005; Tombros et al., 2007; Voudouris et al., 2014; Kevrekidis et al., 2015). On Mykonos Island (Cyclades), epithermal style barite-base metal mineralization occurred when the pluton crossed the ductile-to-brittle transition during its exhumation below the North Cycladic Detachment System at ~11–10 Ma (Menant et al., 2013; Tombros et al., 2015).

In this broad geodynamic regime, the Oligocene-Miocene to Pliocene-Pleistocene magmatic-hydrothermal ore deposits in Greece were mostly formed in four periods: Oligocene (33–25 Ma), Early Miocene (22–19 Ma), Middle to Late Miocene (14–7 Ma) and Pliocene-Pleistocene (3 to 1.5 Ma). A compilation of magmatic and mineralization ages and metal content of known systems, is demonstrated in the time-space plot of Fig. 7.

4.2. Comparison to Cenozoic deposits in southern Balkans and Western Anatolia

The Biga Peninsula in northwestern Turkey shares many similarities in tectono-magmatic and metallogenic history to the Oligocene-Miocene Rhodope and Serbo-Macedonian Massifs (Yigit, 2012). Several porphyry, HS-IS and LS epithermal and skarn/carbonate-replacement deposits in Biga are associated with Late Eocene-Oligocene magmatism (38–25 Ma; Yigit, 2012), mostly contemporaneous to the major Oligocene magmatic-metallogenic event (~33–25 Ma) that took place in the western part of the belt in Greece and FYROM.

In addition, the Cu-Au belt of the Greek part of the Rhodopes extends further to the southeast, at the Biga Peninsula in Turkey. Yigit (2012) distinguished at least three phases of porphyry Cu–Au–Mo mineralization and two phases of high sulfidation epithermal gold mineralization in Biga Peninsula, where porphyry deposits as well as causative intrusions have a younger age from north to south. The oldest phase (52–47 Ma) of porphyry Cu–Au–Mo mineralization is related to the Dikmen and Karabiga stocks and belongs to the adakitic-magmatism in NW Anatolia described by Ersoy and Palmer (2013). The second phase is associated with the Early Oligocene Alankoy granodiorite (~28 Ma), and coincides with the second phase of high sulfidation epithermal mineralization (~28 Ma). The third late Oligocene porphyry phase (~25 Ma) is related to the Tepeoba and Kestane granodiorite porphyries. A late Eocene (~38–39 Ma) age for the early phase in high sulfidation elliptical systems at Kartaldag and Kuscati prospects and an Early Oligocene age (~25–31 Ma) for the second high sulfidation phase at Agi Dagi, Alankoy, Kirazli, Kucukdag, are considered by Yigit (2012), Sanchez et al. (2016) and Leroux (2016).

Finally, major carbonate-hosted deposits, such as the Papazilik Pb–Zn (Au–Ag) deposit and base-metal skarns in the Yenisei district, are considered as a part of Oligocene magmatic-hydrothermal systems at Biga and comparable to the polymetallic Pb–Zn–Ag–Au deposits in the Rhodope Massif (e.g., Olympias, Chalkidiki, Agi Dagi, Anankoy, Kirazli, Kucukdag) by Yigit (2012), Sanchez et al. (2016) and Leroux (2016).
4.3. Metal enrichment in Cenozoic deposits in Greece

Cenozoic magmatic–hydrothermal deposits in Greece demonstrate various enrichment in precious, critical, and energy critical metals (e.g., Melfo and Voudouris, 2012; Tsrambides and Filippidis, 2012), as defined by John and Taylor (2016) and Kelley and Spry (2016). Tellurium is a common constituent throughout Cenozoic to Miocene–Pleistocene Greek deposits (Voudouris et al., 2007) and PGE and Re enrichment has been reported for Skouries (Economou-Eliopoulos and Eliopoulos, 2000) and the porphyry Cu–Mo–Au deposits in the Rhodope Massif (Voudouris et al., 2013c), respectively.

The Re, Te (and Au) enrichment in the high-K calc-alkaline porphyry Cu–Mo–Au mineralization at Pagoni Rachi, Konos Hill, Maronia, Meliteni, and the associated epithermal Au-Ag–Te deposits at Sapes, Perama Hill, and Pefka, expanding over an area of ~5400 km², is a case study attributed to anomalous concentration of these elements in source areas of magma generation (Voudouris et al., 2013c). The Au, Re and Te enrichment in the eastern Rhodope Oligocene–Miocene magmatic-hydrothermal systems is also present at the shoshonitic Miocene porphyry deposits of Limnos and Lesvos Islands (Voudouris et al., 2013c). This metal enrichment belt seems to continue further to the east in the Biga Peninsula, where Au-Ag tellurides were described from the high-sulfidation Kızılkükdağ deposit (Smith et al., 2014) and also to the west where, besides the Skouries deposit, also the Oligocene-Miocene Vathi and Bucim porphyry Cu–Au deposits are enriched in Au, Ag, and Pd-tellurides (Eliopoulos and Economou-Eliopoulos, 1991; Tarkian et al., 1991; Economou-Eliopoulos and Eliopoulos, 2000; Serafimovski et al., 2013).

This enrichment is common in porphyry and epithermal style ore deposits in post-subduction settings elsewhere, where partial melting of the residues of previous cycles of calc-alkaline arc magmatism, may generate high-K calc-alkaline to shoshonitic post-collisional magmas (Richards, 2009, 2011, 2015). Delamination of sub-continental mantle lithosphere and slab retreat, associated with the ingress of asthenospheric melts into the lower crust, can cause small volume partial melting of arc-metasomatized lithosphere and/or hydrous, amphibolitic, lower crustal cumulates (Richards, 2011). Small amounts of chalcophile and siderophile element-rich sulfides left in these cumulates may be redissolved during partial melting thus giving rise to Au-rich as well as normal Cu ± Mo porphyry and epithermal Au systems. Potential source regions of lithospheric contamination include subduction-modified lower crust, sub-continental lithospheric mantle, or juvenile lower crust formed by underplated arc basaits (Richards, 2015 and references therein).

Ersoy and Palmer (2013) suggested for the Rhodope-Aegean west Anatolian Oligocene–Miocene high-K calc-alkaline magmatic rocks crustal hybridization processes involving lower crustal melting, and mixing of these magmas with mantle-derived mafic melts, and subsequent AFC (assimilation and fractional crystallization) processes. Throughout the Oligocene and Miocene, the magmatic activity was derived from subcontinental mantle lithosphere that had been intensely contaminated by the Late Eocene and onwards by oceanic and continental subduction (Ersoy and Palmer, 2013). A slab retreat after ~35 Ma (Jolivet et al., 2013) associated with a southward fast migration of the magmatic centers, may have caused upwelling of asthenospheric mantle, melting of metabasaltic amphibolites that underplated subducted continental crust and high-K calc-alkaline to shoshonitic magmatism (Pe-Piper et al., 2009).

Melting of mafic or ultramafic rocks, as well as involvement of mantle metasomatism in the source rocks, could contribute to Re enrichment as proposed elsewhere (Stein et al., 2001). Previous subduction magmatism may be responsible for the enrichment in Au, Te as well as other fluid–mobile elements such as Pb, As, Sb, Cu and the platinum group elements (PGE) in the mantle lithosphere, probably due to fluids released from the oceanic and continental subducted slab and associated sediments (Sun et al., 2003; Tessalina et al., 2008; Cook et al., 2009; Grabbezev and Voudouris, 2014). Tellurium enrichment is also present in epithermal deposits at Milos Island in the South Aegean Volcanic Arc, which are typical of subduction-related magmatic activity and their mantle source was metasomatized by subduction of oceanic assemblages (Ersoy and Palmer, 2013).

Less fertile magma sources beneath the Bulgarian part of the Rhodope may be the reason for the rarity of Au-bearing tellurides from the Oligocene volcanic-hosted intermediate sulfidation deposits in the eastern Rhodope, Bulgaria, compared to their abundance in northeastern Greek high sulfidation epithermal
Au-Ag-Cu-Te deposits. At these deposits Au-Ag-tellurides were introduced by fluids of intermediate sulfidation states and contribute to their total gold potential (Voudouris, 2006). Marchev et al. (2005) have already pointed out significant geographical variations in the style and composition of the Early Oligocene Rhodopean deposits, from Pb-Zn-Ag veins, skarns and carbonate-replacement base-metal deposits in the central Rhodope (e.g., Madan ore field), through intermediate sulfidation epithermal Pb-Zn-Ag-Au deposits (Madjarovo, Zvezdel, Spahievo) in volcanosedimentary basins at the Bulgarian part of the eastern Rhodope, to high-intermediate sulfidation epithermal Au-Ag-Cu-Te polymetallic deposits in the Greek part of the Rhodope (Sapes, Perama Hill). Based on Sr and Pb isotope systematics on igneous rocks, sulfides and gangue minerals from the southeastern Bulgarian ore deposits, Marchev et al. (2005) demonstrated a decreasing input to the hydrothermal systems of Palaeozoic or older crustal material, from the much thicker continental crust in the central Rhodope and an increase of mantle contributions towards the eastern Rhodope, which is underlain by a thinner crust. The increasing amount of crustal components, from the SE Rhodope towards the central Rhodope, also coincides with an increasing proportion of acid rocks in the central Rhodope, where crustal contamination could have diluted original mantle magmas containing Cu and Au with an increasing contribution of Pb from a continental basement source.

However, further to the north, the Cretaceous arc-related porphyry-epithermal deposits in the Srednogorie-Pontides Belt are quite similar to those in northern Greece in respect to Au, Te, Re and PGE enrichment (Zimmerman et al., 2008; Eliopoulos et al., 2014). These deposits, as for example, Majdanpek, Elatsite, and Assarel, are related to subduction of the Vardar Ocean (Gallhofer et al., 2015), and a melt-metal source in fertile mantle and/or juvenile lower crust was responsible for the relatively high Re concentrations (up to 0.35 wt.%) in molybdenite from the Majdanpek (Serbia) and Elatsite (Bulgaria) Cu-Mo porphyry deposits (Zimmerman et al., 2008).

4.4. Perspectives for future exploration

The Greek part of the Western Tethyan metallogenic belt has been recognized as a favorable metallogenic province representing the link between Balkans and Anatoloides-Taurides, where intensive precious metal exploitation currently takes place. Although mineral exploration and mining activity for precious and base metals in Greece date back to ancient times, only a few major magmatic-hydrothermal prospects have been studied with modern exploration methods. With the exception of drilled projects/deposits at Chalkidiki (e.g., Skouries, Olympias) and western Thrace (Sapes, Perama Hill) in northern Greece, by Eldorado Gold Corp., in recent years, the rest of mineralization in the Rhodope, the eastern Aegean islands and the Cyclades, are insufficient explored and only limited information exists in addition to the evidence from historic exploitation.

Several questions remain open and are related to issues about gold, the age of magmatism and associated mineralization and metal enrichment in the Cenozoic metallogenic provinces of Greece. Future investigations should focus on exploring gold potential at the porphyry deposits, like in Pagoni Rachi, Maronia, Konos Hill, Vathi, Fakos. Exploration targets for porphyry Au-ore should not only be the potassic- but also the sodic-calcic-potassic alteration, which usually is barren and occurs in the periphery of other porphyry systems (Halley et al., 2015). The enrichment in rare and critical metals, like Re, Te, Se, and PGE, in several northern Greek porphyry systems, may increase their economic potential since they could be extracted by-products in metallurgy processing.

Although the Oligocene-Miocene Western Tethyan metallogenic belt in Balkan-Western Turkey has prospects mainly for porphyry, epithermal, skarn, and carbonate-replacement deposits, there is increasing evidence that there may also be a province for Reduced Intrusion-Related Gold, Carlin-style Sb-Au, as well as low sulfidation epithermal Au deposits. These three styles of ore deposits have been discovered in various regions along the Western Tethyan belt and it is only a matter of future exploration for additional discoveries.

Typical end-member low-sulfidation gold-silver deposits are not yet discovered in Greece, but present both in western Turkey (e.g., Kucukdere and Kisicik deposits) where they are related to Miocene volcanic rocks (Yigit, 2012) and also in SE Bulgaria, (e.g., Ada Tepe, Rosino, Stremsti, etc.) where they are older than the adjacent magmatic-related deposits and are mainly hosted in Maastrichtian-Paleocene sedimentary rocks, above detachment faults contacts with the underlying Palaeozoic metamorphic rocks (Marchev et al., 2005; Márton et al., 2010).

Carlin-style gold deposits are also missing in Greece, but present in western Turkey, at the Kızıldaml and Findikli prospects in the eastern Biga Peninsula and also at several deposits at Menderes Massif (Yigit, 2012). Carlin-style gold mineralization also occurs in Allich deposit FYROM (Percival and Radtke, 1994). The Kalinintiri Sb-As-Ag deposit is a polymeric ore, sharing some affinities to Carlin-style mineralization, however, its exact classification is a matter of further investigations (Kanellopoulos et al., 2014).

Reduced intrusion-related gold deposits, like those present in the Greek part of Rhodope (e.g., Miocene age), might be present in Biga, at the Mo rich porphyry Au Diken prospect which is related to quartz-feldspar porphyry, granodiorite and aplite dikes.

The Rhodope and Cyclades region also include ore mineralization enriched in Au, Ag, Cu, Bi, and Te, which is associated with detachment faults, shear zones and metamorphic rock-hosted veins in Stanos, Nea Madytos, Drakontio, Koronouda, Stefania, Lao-dikino, Lavrion, Kallianou, Sifnos, Tinos, Syros, and Antiparos. They are considered to have a genetic affiliation to magmatic rocks of a possible Miocene age, although these intrusions are not exposed and are deduced to be buried at shallow depths. It is suggested that ore fluids derived from the magmas, circulated at low depths along extension-related shear zones and interacted with the country rocks of the exhumed metamorphic core complexes. These large tectonic structures should also be exploration targets.

Acknowledgments

The authors are thankful for the comments and suggestions given from the two reviewers, Thomas Bissig and Ladislav Palinkas. Tim Baker and Jeremy Richards are sincerely thanked for their reviews and their constructive and valuable comments on an earlier version of this manuscript. Editor Franco Pirajno is greatly acknowledged for editorial handling.

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