

# Tectonic evolution of the Zagros Orogen in the realm of the Neotethys between the Central Iran and Arabian Plates: An ophiolite perspective

**Majid Sudi Ajirlu\*, Mohssen Moazzen, Robab Hajialioghli**

Department of Geology, University of Tabriz, Tabriz, Iran

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The Zagros Orogenic Belt includes the Fold and Thrust Belt, the High Zagros Belt, the Outer Zagros Ophiolitic Belt, the Sanandaj–Sirjan Metamorphic Belt, the Inner Zagros Ophiolitic Belt, and the Urumieh–Dokhtar Magmatic Belt. We divide the High Zagros evolutionary history into five stages: (1) triple junction formation, (2) continental lithosphere rifting, (3) generation, spreading, and maturation of the Neotethys Ocean, (4) subduction of the oceanic lithosphere, and (5) collision. The Neotethys triple junction, located at the southeastern corner of the Arabian Plate, formed during the Late Silurian–Early Carboniferous. Subsequently, this triple junction became a rift basin due to normal faulting and basalt eruption. The rifting stage occurred during the Late Carboniferous–Early Permian. Thereafter, extension of the basin continued, leading to spreading and maturation of the Neotethys oceanic basin during the Late Permian–Late Triassic. Probably at the end of the Late Triassic, closure of the Paleotethys Basin caused the initiation of two northeastward subductions: (1) oceanic–oceanic and (2) oceanic–continental. Oceanic–oceanic subduction continued until the Late Cretaceous and was terminated by the emplacement of the Outer Zagros Ophiolites, whereas oceanic–continental subduction continued until the Middle Miocene. Subduction in the southern Neotethys Basin between the Arabian and Central Iran Plates caused a tensional regime between Sanandaj–Sirjan and Central Iran, and the formation of a back-arc basin that by its closing led to the emplacement of the Inner Zagros Ophiolites during the Late Cretaceous.

**Keywords:** Zagros Orogenic Belt, Neotethys, Arabian Plate, Central Iran, Sanandaj–Sirjan, ophiolite

\*Corresponding author: Majid Sudi Ajirlu; Department of Geology, University of Tabriz, 51664 Tabriz, Iran  
E-mail: [majid.sudi@gmail.com](mailto:majid.sudi@gmail.com)

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## Introduction

The Alpine–Himalayan Orogenic Belt consists of two branches in Iran, the Alborz Mountains with an E–W trend and the Zagros Orogen with an NW–SE trend (Alavi 1994; Agard et al. 2005). The Zagros Orogen extends from eastern Turkey through northern Iraq and northwest of Iran to the Hormuz Strait and Oman (Alavi 1994; McQuarrie 2004; Agard et al. 2005; Homke et al. 2010). Because of the key situation of the Zagros Orogen within the Alpine–Himalayan Orogenic Belt and its importance in revealing the nature of Neotethys Ocean evolution and closure, numerous studies have been carried out on its magmatic, sedimentary, metamorphic, and structural characteristics (e.g., Takin 1972; Stöcklin 1977; Berberian and King 1981; Şengör 1984; Dercourt et al. 1986; Şengör et al. 1988; Glennie 1992; Alavi 1994; Stampfli and Borel 2002; Mohajjel et al. 2003; Golonka 2004; Shahabpour 2005; Agard et al. 2005; Kazmin and Tikhonova 2006; Robertson 2007; Sheikholeslami et al. 2008; De Vera et al. 2009; Allahyari et al. 2010; Homke et al. 2010; Saura et al. 2011; Ghazi et al. 2012; Saccani et al. 2013; Chiu et al. 2013; McQuarrie and van Hinsbergen 2013, among many others). On the basis of tectonomagmatic and structural features, the Zagros Orogen is subdivided into nine subzones (Falcon 1969; Berberian 1995; Emami et al. 2010; Homke et al. 2010), which are the Mesopotamian–Persian Gulf Foreland, the Dezful Embayment, the Fold and Thrust Belt, the Crush Zone (High Zagros or Imbricated Zone), the Outer Zagros Ophiolite Belt, the Sanandaj–Sirjan First Metamorphic Belt, the Mesozoic Magmatic Belt, the Sanandaj–Sirjan Second Metamorphic Belt, and the Urumieh–Dokhtar Magmatic Belt (UDMB) (Fig. 1). The formation of the Zagros Basin is related to the rifting and separation of Central Iran from the northern margin of Gondwana, which formed the Neotethys Basin (Agard et al. 2005; Yousefirad 2011; Chiu et al. 2013). Closure of the Neotethys Ocean and the collision between the Central Iran and Arabian Plates formed the Zagros Orogen (Yousefirad 2011; Chiu et al. 2013). To reconstruct the Zagros Orogen evolution, several models have been presented, but none can explain all of its geologic aspects completely. In this contribution, we try to present a comprehensive reconstruction model for the Zagros Orogen to better explain its characteristics, by reviewing and discussing all hypotheses.

## Geologic context

The Zagros Orogen is a result of the closure of the southern Neotethys, located between the Iranian and Arabian Plates (Alavi 1994; Talbot and Alavi 1996; Stampfli and Borel 2002; Casini et al. 2011). Ophiolites, which are Neotethys oceanic lithosphere remnants, are emplaced along the Zagros Orogen. These ophiolites are emplaced along two main belts (Figs 1 and 2; Stöcklin 1977; Homke et al. 2010; Ghazi et al. 2012). The first one is the Khoy–Nain–Shahr Babak–Dehshir–Baft Ophiolitic Belt (Inner Zagros Ophiolitic Belt or IZOB; Shafaii Moghadam and Stern

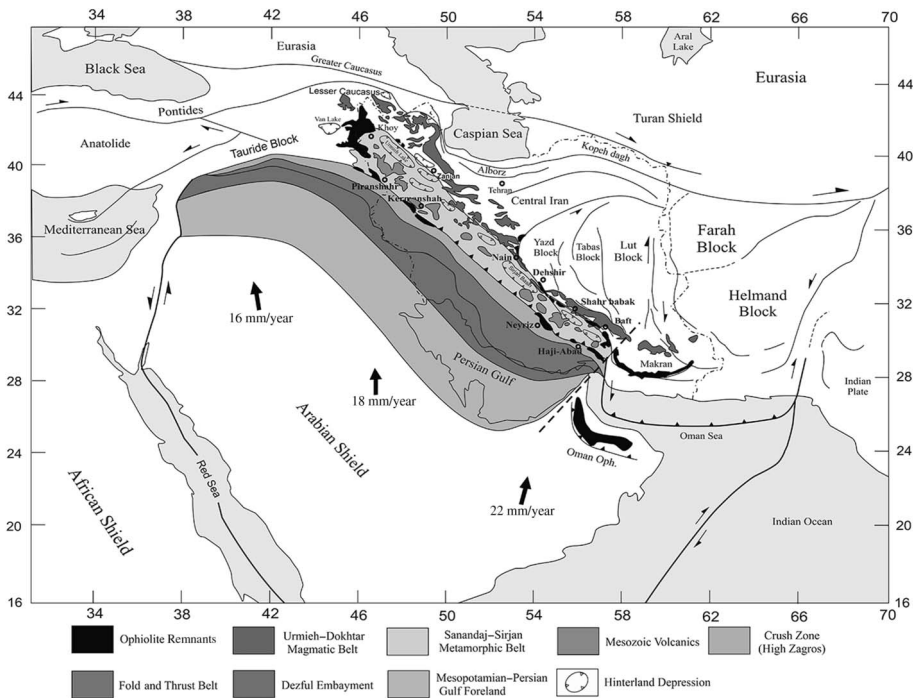


Fig. 1

Structural map of the Iran Plateau and the Zagros Orogen. Only Neotethys-related ophiolites are shown; dashed lines are concealed faults and solid lines are main faults

2011), located between the Sanandaj–Sirjan and Central Iran; the second one is the main ophiolitic belt of Piranshahr–Kermanshah–Neyriz–Haji-Abad [Outer Zagros Ophiolitic Belt (OZOB); [Shafaii Moghadam and Stern 2011](#)], located between the Sanandaj–Sirjan and Zagros (Fig. 1). The IZOB resulted from the closure of the southern Neotethys Basin, while the OZOB resulted from the closure of the Neotethyan back-arc basin ([Alavi 1994](#); [Ghasemi and Talbot 2006](#); [Allahyari et al. 2010](#); [Casini et al. 2011](#); [Ghazi et al. 2012](#); [Saccani et al. 2013](#)). In a schematic cross-section across the Zagros Orogen, the main structural elements are shown. These structures are cover sediments, the Cenozoic Magmatic Belt, the Inner Ophiolite Belt, the Second Metamorphic Belt, the Mesozoic Magmatic Belt, the First Metamorphic Belt, the Outer Ophiolite Belt, the Crush Zone, the Fold and Thrust Belt, the Dezful Embayment, and the Persian Gulf–Mesopotamian Foreland (Fig. 2). In the following, we will explain why it is necessary to divide the Sanandaj–Sirjan metamorphic zone into two subzones (the First Metamorphic Belt and the Second Metamorphic Belt).

The Fold and Thrust Belt on the Arabian Plate is a result of deformation of the Zagros Orogenic Belt (ZOB) passive margin sediments that were caused by continental

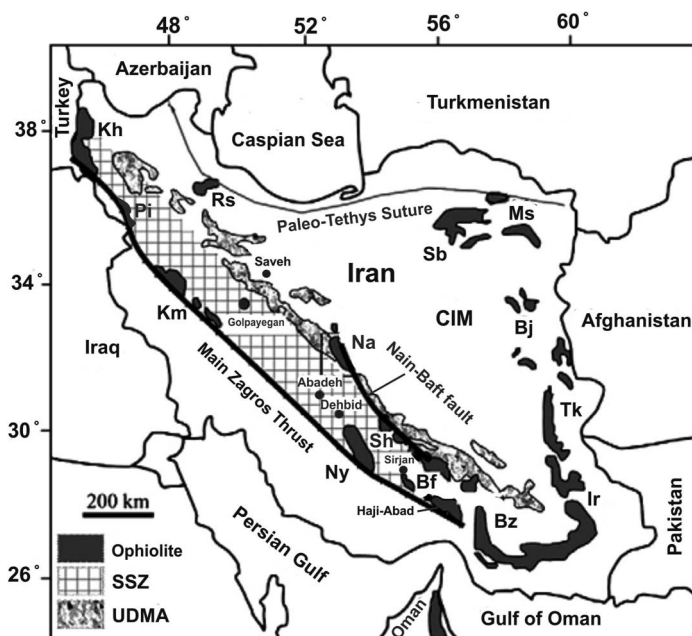


Fig. 2

Schematic cross-section from structural elements of the ZOB (not to scale). These structural elements from NE to SW include cover sediments, the Cenozoic Magmatic Belt, the Inner Ophiolite Belt (Khoy–Nain–Shahr Babak–Dehshir–Baft Ophiolites), the Second Metamorphic Belt, the Mesozoic Magmatic Belt (island arc), the First Metamorphic Belt, the Outer Ophiolite Belt (Neyriz–Kermanshah–Haji-Abad Ophiolites), the Crush Zone (or High Zagros), the Fold and Thrust Belt, the Dezful Embayment, and the Persian Gulf–Mesopotamian Foreland

collision (Alavi 1994; Ghasemi and Talbot 2006; Horton et al. 2008; Allahyari et al. 2010; Saccani et al. 2013). In the Sanandaj–Sirjan Metamorphic Belt (SSMB), two metamorphic belts are distinguished, which are separated by an andesitic–granitic Mesozoic island arc (Berberian and Berberian 1981; Desmon and Beccaluva 1983; Malekizadeh 1999; Mohajjel et al. 2003; Chiu et al. 2013). In the ZOB, the metamorphic belt is located in the northeast of the ophiolitic belt (Figs 1 and 2). Therefore, the direction of subduction was northeastward, and because two metamorphic belts were formed, two northeastward subductions must have occurred (Fig. 2). The presence of a volcanic arc and an island arc indicates that two subductions, one is oceanic–oceanic and the other oceanic–continental, were active (Ghasemi and Talbot 2006; Agard et al. 2007; Ghazi et al. 2012; Saccani et al. 2013). The oceanic–oceanic subduction arc-related calc-alkaline granitoid rocks of Siah Kuh of Early Jurassic age (199 Ma; Arvin et al. 2007) and other similar rocks are reported from the SSMB (Berberian and Berberian 1981; Malekizadeh 1999; Shahabpour 2007). The Chah Dozdan granite and

Chah Ghand gabbro are Middle Jurassic age (159–167 Ma; [Sheikholeslami 2002](#)). The other Mesozoic rocks in the SSMB are the Middle Jurassic Boroujerd Plutonic Complex (169 Ma; [Khalaji et al. 2007](#); [Mahmoudi et al. 2011](#)), the Astaneh Pluton (168 Ma), the Alvand Pluton (165 Ma; [Mahmoudi et al. 2011](#)), the Late Jurassic (157–149 Ma) Ghorveh Pluton, the mid-Cretaceous (109 Ma) I-type Hasan Salarn Pluton near the town of Saqqez ([Mahmoudi et al. 2011](#)), and the Middle Jurassic (165 Ma) Aligoodarz granitoid complex ([Esna-Ashari et al. 2012](#)). Likewise, Chiu et al. ([2013](#)) distinguished I-type granitoids of Middle to Late Jurassic ages (176–144 Ma) in the SSMB and granodiorite–diorite of Middle Jurassic age (166–167 Ma) in the UDMB.

In the ZOB, the Neyriz, Kermanshah, and Haji-Abad Ophiolites are unconformably covered by Late Cretaceous limestone and Paleocene volcanic and volcanoclastic rocks ([Ricou 1974](#); [Ghasemi and Talbot 2006](#); [Saccani et al. 2013](#)). The Nain, Shahr Babak, Dehshir, and Baft Ophiolites are also covered by Upper Cretaceous pelagic limestone ([Babaei et al. 2005](#); [Shahabpour 2005](#); [Ghasemi and Talbot 2006](#); [Azizi et al. 2006](#); [Shafaii Moghadam and Stern 2011](#); [Ghazi et al. 2011](#); [Saccani et al. 2013](#)). Therefore, the emplacement time of these ophiolites is Upper Cretaceous.

### **The Inner Zagros Ophiolitic Belt (IZOB)**

This ophiolitic belt includes the Khoy, Nain, Shahr Babak, Dehshir, and Baft ophiolitic complexes and is located between the Central Iran and Sanandaj–Sirjan continents, with an NW–SE trend (Figs 1 and 2). A back-arc basin of the southern Neotethys Basin existed between the Central Iran and Sanandaj–Sirjan ([Glennie 1992](#); [Ghazi and Hassanipak 1999](#); [Golonka 2004](#); [Ghasemi and Talbot 2006](#); [Shafaii Moghadam et al. 2009](#); [Ghazi et al. 2011, 2012](#)). The IZOB is a result of this back-arc basin closure.

#### *The Khoy Ophiolite complex*

The Khoy ophiolitic complex is located in the northwest of Iran. The ophiolite mantle sequence consists of serpentinized lherzolite and harzburgite with minor wherlites, pegmatitic gabbro, meta-gabbro, layered gabbro, pyroxenite dikes/veins, dunite veins, amphibolite, and rodingite veins (Fig. 3; [Ghazi and Hassanipak 1999](#); [Ghazi et al. 2003](#); [Juteau 2004](#); [Azizi et al. 2006](#); [Moazzen and Oberhänsli 2008](#)). The Khoy Ophiolite crustal sequence includes gabbro, massive, and flow basalt, and Late Cretaceous–Early Paleocene turbiditic sediments intercalated with pillow lavas and pelagic limestone (Fig. 3). Based on the studies by [Khalatbari et al. \(2003\)](#) and [Pessagno et al. \(2005\)](#), there are two ophiolitic complexes in the Khoy area: (1) an old polymetamorphic ophiolite, whose oldest metamorphic amphibole yielded a Late Jurassic apparent  $^{40}\text{K}$ – $^{40}\text{Ar}$  age; therefore, its primary magmatic age should logically be pre-Jurassic (Upper Triassic) and (2) a younger non-metamorphic Upper

Cretaceous ophiolite. Based on radiometric dating of the gabbro, the maximum age for the formation of this ophiolite complex is Late Albian (108–109 Ma) and its emplacement age is 110–111 Ma (Ghazi et al. 2003). In some areas, the ophiolitic rocks are unconformably overlain by the Paleocene–Eocene Fajan conglomerate that contains ophiolitic mafic and ultramafic blocks (Ghazi and Hassanipak 1999; Azizi et al. 2006; Pessagno et al. 2005).

### *The Nain Ophiolite*

The Nain Ophiolite is related to the closure of the Neotethys back-arc basin (Ghasemi and Talbot 2006; Rahmani et al. 2007; Ghazi et al. 2012). The Nain Ophiolite mantle sequence includes serpentized harzburgite, with some lherzolite and pyroxenite sills, and its crustal sequence includes plutonic and volcanic rocks, such as pegmatitic gabbro, diorite, granodiorite, diabase, basaltic andesite, andesite, and more silicic rocks (tonalite, trondhjemite, and aplite) (Fig. 3; Ghasemi and Talbot 2006; Rahmani et al. 2007; Shafaii Moghadam and Stern 2011; Ghazi et al. 2011, 2012). The compositional diversity of extrusive rocks (from basaltic andesite to andesite), with various Upper Cretaceous sedimentary rocks, indicates that the Nain Ophiolite has island-arc tholeiite (IAT) signatures (Hassanipak and Ghazi 2000; Ghasemi and Talbot 2006; Shafaii Moghadam et al. 2009; Ghazi et al. 2012).  $^{40}\text{Ar}/^{39}\text{Ar}$  dating on hornblende gabbro indicated the ages of  $101.2 \pm 0.9$  Ma,  $99.7 \pm 0.9$  Ma, and  $99 \pm 1.2$  Ma, and suggested that the generation age of this ophiolite is Upper Albian; zircon U-Pb data show that the age of Nain Ophiolite is 103–101 Ma (Hassanipak and Ghazi 2000; Azizi et al. 2011). The basement of the Nain Ophiolite complex is garnet amphibolite with low pitch lineation (Rahmani et al. 2007; Shafaii Moghadam et al. 2009). The Nain Ophiolite was covered by Coniacian–Maastrichtian pelagic limestone and chert (Rahmani et al. 2007; Shafaii Moghadam and Stern 2011).

### *The Shahr Babak Ophiolite*

The Shahr Babak Ophiolite is a part of the IZOB; its mantle sequence includes lherzolite, foliated harzburgite, dunite, serpentinite, and pyroxenite. The foliated harzburgite is crosscut by diabasic dikes and small-sized lenses of isotropic gabbro (Fig. 3; Ghazi and Hassanipak 1999; Ghasemi and Talbot 2006; Shafaii Moghadam et al. 2009; Shafaii Moghadam and Stern 2011). The Shahr Babak Ophiolite crustal sequence includes gabbro, pillow and massive lavas, diorite, and plagiogranite. Plagiogranitic veins commonly crosscut the isotropic gabbro (Fig. 3). Extrusive rocks of the Shahr Babak Ophiolite are basaltic andesite to rhyodacite–rhyolite and trachyandesite (Shafaii Moghadam and Stern 2011). Shafaii Moghadam et al. (2010) determined that the basaltic andesite and rhyodacite were generated in an island-arc environment and that the trachyandesite has IAT signatures. The Shahr Babak Ophiolite is covered by Coniacian–Maastrichtian pelagic limestone (Ghasemi and Talbot 2006).

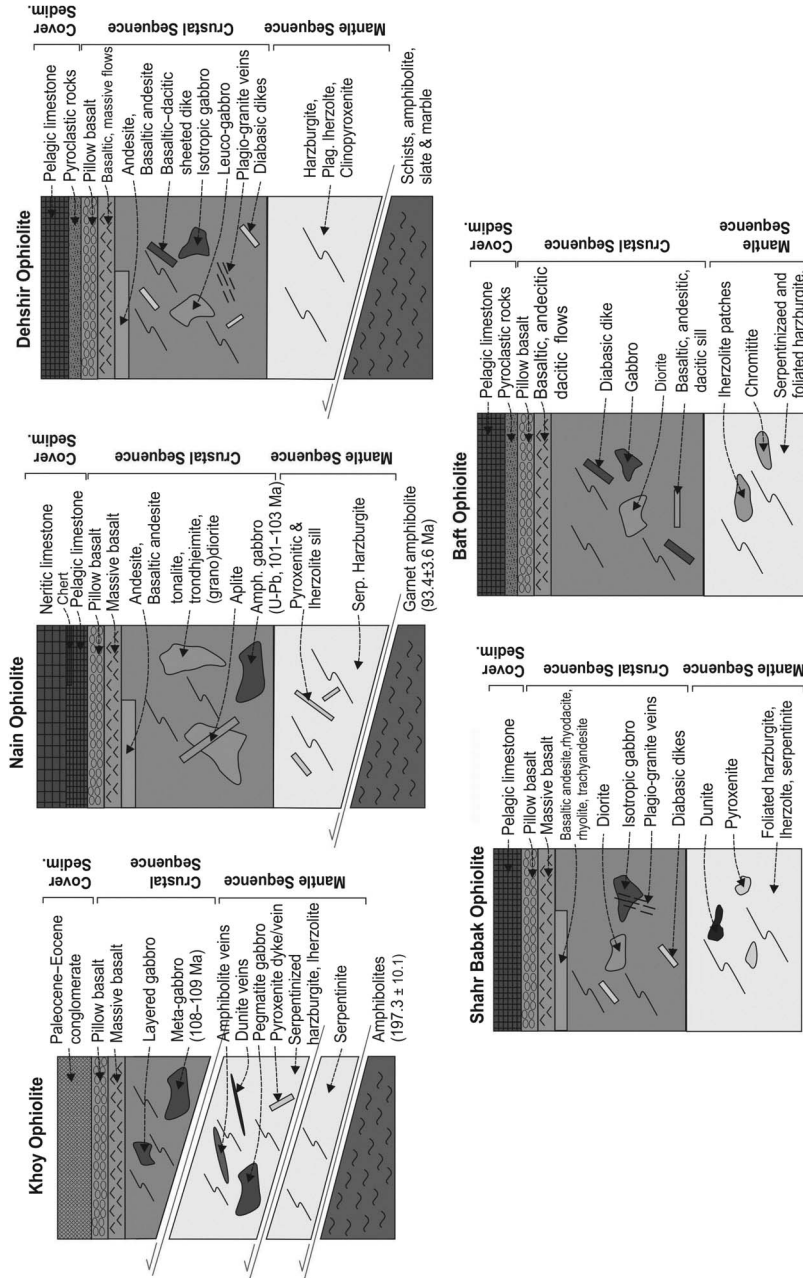


Fig. 3 Stratigraphic sequences of Khoy, Nain, Shahr Babak, Dehshir, and Baft Ophiolites (not to scale). The analyzed samples (whole rock and K-Ar age) are shown (modified after Shafaii Moghadam et al. 2009)

### *The Dehshir Ophiolite*

The Dehshir Ophiolite crops out at the intersection of two faults (the Nain–Dehshir and the Dehshir–Baft Faults). The mantle sequence of this ophiolite includes clinopyroxene-bearing harzburgite, plagioclase lherzolite, and clinopyroxene (Fig. 3; [Shafaii Moghadam and Stern 2011](#)). The Dehshir Ophiolite crustal sequence includes pillow lavas, basaltic and basaltic–andesitic massive flows, basaltic–dacitic sheeted dikes, leucogabbro, and isotropic gabbro (Fig. 3; [Shafaii Moghadam and Stern 2011](#)). Gabbroic and diabasic dikes crosscut the peridotite. Metamorphic rocks in the Dehshir Ophiolite complex are actinolite schist, amphibolite, slate, and marble. This ophiolite complex is covered by Turonian–Maastrichtian pelagic limestone; therefore, its emplacement age is Upper Cretaceous ([Shafaii Moghadam and Stern 2011](#); [Shafaii Moghadam et al. 2012](#)).

### *The Baft Ophiolite*

The Baft Ophiolite crops out along the Dehshir–Baft Fault. The Baft Ophiolite mantle sequence includes serpentinized and foliated harzburgite with patches of chromitite and minor lherzolite with diabasic dikes (Fig. 3; [Ghasemi and Talbot 2006](#); [Shafaii Moghadam and Stern 2011](#)). The Baft Ophiolite crustal plutonic and volcanic rocks consist of gabbro; diorite; pillow basalts; basaltic, andesitic, and dacitic lava flows; and basaltic, andesitic, and dacitic sills in pyroclastic rocks. Basalts have both features of mid-ocean ridge [tholeiites similar to normal mid-ocean ridge basalt (N-MORB)] and intra-plate affinities [intermediate between enriched mid-ocean ridge basalt (E-MORB) and IAT] ([Alavi 1994](#); [Arvin and Robinson 1994](#)). Based on lithostratigraphic relationships, the Baft Ophiolite was formed during the Cretaceous and was emplaced before the Paleocene ([Lippard et al. 1986](#); [Shafaii Moghadam and Stern 2011](#)).

## **The Outer Zagros Ophiolitic Belt (OZOB)**

The OZOB crops out along the southern Neotethys suture. The OZOB, which includes the Piranshahr, Kermanshah, Neyriz, and Haji-Abad Ophiolites, is the result of the Late Cretaceous closure and collision between the Sanandaj–Sirjan and the Arabian Shield (Fig. 1). Stratigraphic columns for these ophiolites are shown in Fig. 4.

### *The Piranshahr Ophiolite*

The Piranshahr Ophiolite is located in the northwest of Iran close to the town of Piranshahr and extends to the Rayat Ophiolites in NE Iraq and to the Cilo Ophiolite in SE Turkey ([Okay et al. 2010](#); [Hajjalioghli and Moazzen 2014](#)). The Piranshahr Ophiolite mantle sequence consists of serpentinite, tectonized harzburgite and minor lherzolite, dunite with cumulitic texture, wherlite, and troctolite with Cr-spinel



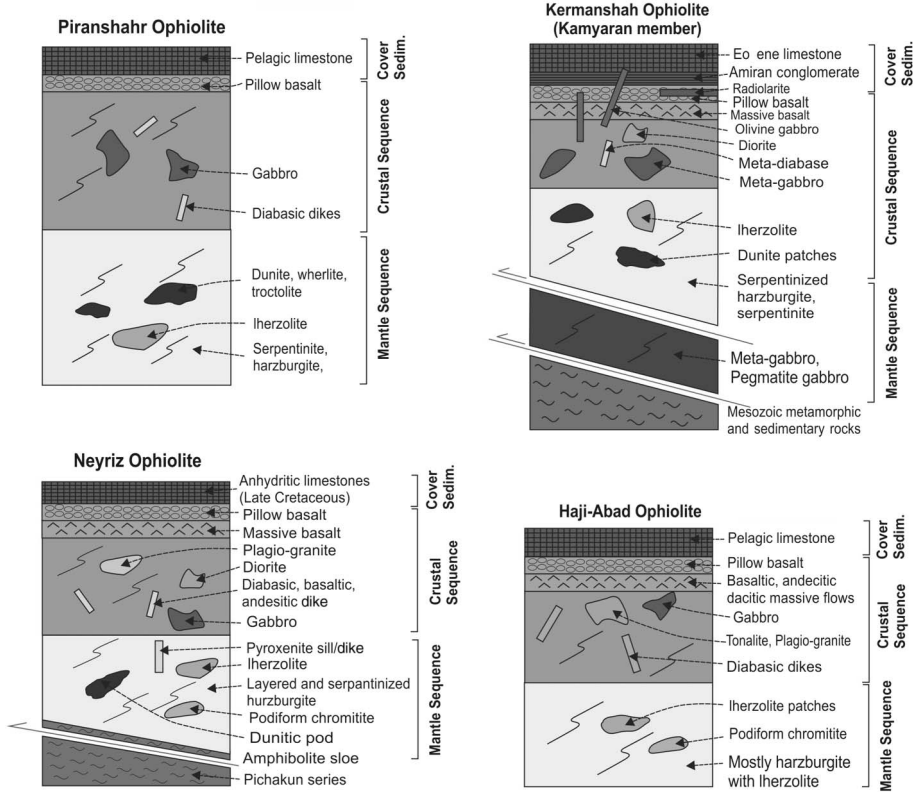


Fig. 4 Stratigraphic sequence of Piranshahr, Kermanshah, Neyriz, and Haji-Abad Ophiolites (not to scale)

(Fig. 4; [Khodabandeh and Soltani 2004](#)). They are mantle residues with distinct geochemical signatures of both mid-ocean ridge and suprasubduction zone affinities ([Hajjalioghli and Moazzen 2014](#)). The crustal sequence includes gabbro, dibasic dikes, rare pillowed basalts, and radiolarite overlain by Late Cretaceous pelagic limestone (Fig. 4; [Khodabandeh and Soltani 2004](#); [Hajjalioghli and Moazzen 2014](#)).

*The Kermanshah Ophiolite*

The Kermanshah ophiolitic complex, located along the main Zagros reverse fault, is an ophiolitic mélange, which includes dismembered ophiolitic sequences. The mantle sequence predominately includes lherzolite, dunite, and harzburgite, and its crustal sequence includes cumulative gabbro, sheeted dikes, pillow lavas, and diorites, and a volcanic sequence that ranges in composition from sub-alkaline basalt through alkaline basalt to trachyte (Fig. 4). Ghazi and Hassanipak (1999) recognized both

island-arc tholeiitic and alkaline lavas. Tholeiitic basalt is a normal mid-ocean ridge to enriched mid-ocean ridge basalt (Allahyari et al. 2010). The E-MORB characteristics of middle ocean ridge basalt and the lherzolite composition of peridotite indicate that partial melting and the sea-floor spreading rate was low. Ghazi and Hassanipak (1999), using geochemical data, clearly distinguished two distinct types of sub-alkaline basalt, suggesting an island-arc affinity, as well as alkaline basalt, suggesting a typical oceanic island. During the Maastrichtian, the Kermanshah Ophiolite was thrust into the Lower Triassic–Upper Cretaceous Bisoton limestone (Lippard et al. 1986; Berberian 1995; Ghasemi and Talbot 2006). The Amiran Formation conglomerate of Maastrichtian–Paleocene age contains ophiolite clasts. Subsequently, the initial thrusting of the Kermanshah Ophiolite occurred during Maastrichtian–Paleocene time (Alavi 1994; Hooper et al. 1995; Ghasemi and Talbot 2006). The Paleocene volcanic rocks and Eocene shallow-water limestone unconformably cover the Kermanshah Ophiolite (Braud 1987; Ghasemi and Talbot 2006).

### *The Neyriz Ophiolite*

The Neyriz Ophiolite is a part of the OZOB that preserved the oceanic lithospheric units. There are three imbricated sheets in the Neyriz Ophiolite that, from bottom to top, consists of the Pichakun Series (including Upper Triassic limestone, Middle Jurassic oolitic limestone, and Lower Cretaceous conglomeratic limestone), a *mélange* unit, and an ophiolite unit (Ricou 1968; Ghasemi and Talbot 2006; Shafaii Moghadam and Stern 2011). The Neyriz Ophiolite sequence includes layered and serpentinized harzburgite with minor lherzolite, residual dunite, scattered podiform chromitite, pyroxenitic sills or dikes, gabbro, diabasic–basaltic–andesitic dikes, diorite, plagiogranite, mafic and silicic volcanic differentiates (including MORB), and a sheeted dike complex in the Tang-e-Hana area (Fig. 4; Arvin 1982; Sarkarinejad 1994; Nadimi 2002; Robertson 2002; Babaei et al. 2005; Ghasemi and Talbot 2006). The Neyriz Ophiolite complex is tectonically juxtaposed beneath cataclastically deformed island-arc volcanic and volcanoclastic rocks called the Hassanabad unit by Babaie et al. (2000). Geochemical analyses reveal that the Hassanabad unit is of predominantly calc-alkaline island-arc composition (Babaie et al. 2000). Based on  $^{40}\text{Ar}/^{39}\text{Ar}$  dating, the Neyriz Ophiolite complex was formed at 96–98 Ma (Haynes and Reynolds 1980) and was emplaced at 89 Ma (Lanphere and Pamić 1983; Ghasemi and Talbot 2006). The Neyriz Ophiolite complex was thrust into Late Triassic to Middle Cretaceous sediments of the Pichakun Series and is unconformably overlain by Late Cretaceous anhydritic limestone of the Tarbur Formation (Ricou 1968, 1974).

### *The Haji-Abad Ophiolites*

The Haji-Abad (or Esfandagheh) Ophiolite, as a part of the OZOB, is located in the northwest of the town of Esfandagheh. The Haji-Abad Ophiolite mantle sequence

includes depleted harzburgite and residual dunite with aligned spinel and podiform chromitites (Fig. 4; Babaei et al. 2005; Ghasemi and Talbot 2006; Shafaii Moghadam et al. 2012). Leucogabbroic and gabbroic dikes and metamorphosed diabasic dikes crosscut the harzburgite. The Haji-Abad Ophiolite crustal sequence includes pillowed and massive lava flow, which is covered by Upper Cretaceous pelagic limestone (Fig. 4; Shafaii Moghadam and Stern 2011). Small tonalitic intrusions into the crustal sequence occur, and a fine-grained plagiogranitic sill crosscuts the tonalitic intrusions (Fig. 4). On the basis of their ages, the Haji-Abad Ophiolite is divided into two complexes: (1) Late Triassic–Cretaceous and (2) Upper Cretaceous in age (Ghasemi et al. 2002; Ahmadipour et al. 2003; Shafaii Moghadam et al. 2012). The older complex, which has suprasubduction zone geochemical signatures, is covered by marble and amphibolite of the Upper Triassic–Lower Jurassic Sargaz–Abshur complex (Shafaii Moghadam and Stern 2011; Ghasemi et al. 2002). The younger one was formed in the Late Cretaceous and is covered by volcanoclastic sandstone and siltstone turbidites (Shafaii Moghadam et al. 2012).

### **Evolution of the ophiolite suites**

Using the spatial distribution of the major ophiolitic complexes of the Zagros Orogen and other features, we describe the tectonomagmatic and tectonostratigraphic evolution of this Orogen in five stages: (1) triple junction formation, (2) continental lithosphere rifting, (3) generation, spreading, and maturation of the Neotethys Ocean, (4) subduction of the oceanic lithosphere, and (5) collision.

### **Triple junction formation stage (Late Silurian–Early Carboniferous)**

The formation of the Neotethyan triple junction began when a thermal bulge (plume) formed in the Gondwana continent through hot spots (Lapierre et al. 2004; Chauvet et al. 2009, 2011; Saccani et al. 2013). This plume rose toward the upper levels and lower depths of the mantle. By underplating beneath the Gondwana lithosphere, this lithosphere thinned and swelled (Fig. 6). Thereafter, by initiation of three radial fractures, three arms of the Neotethys Basin formed (Numan 2001; Lapierre et al. 2004; Chauvet et al. 2009). The joining of these three arms formed the Neotethys triple junction. This triple junction is located at the southeast corner of the Arabian Plate (Lapierre et al. 2004; Jassim and Goff 2006; Chauvet et al. 2009, 2011; Saccani et al. 2013). Extension into the continental lithosphere occurred by normal faulting. In many references, the rifting time of Neotethys is stated as Early Permian (e.g., Dercourt et al. 1986; Şengör et al. 1988; Glennie 1992; Ziegler 2001; Stampfli and Borel 2002; Shahabpour 2005; Almutury and Al-Asadi 2008; Muttoni et al. 2009; Allahyari et al. 2010; Homke et al. 2010; Saccani et al. 2013). However, comparing the lithostratigraphic columns of the ZOB and the

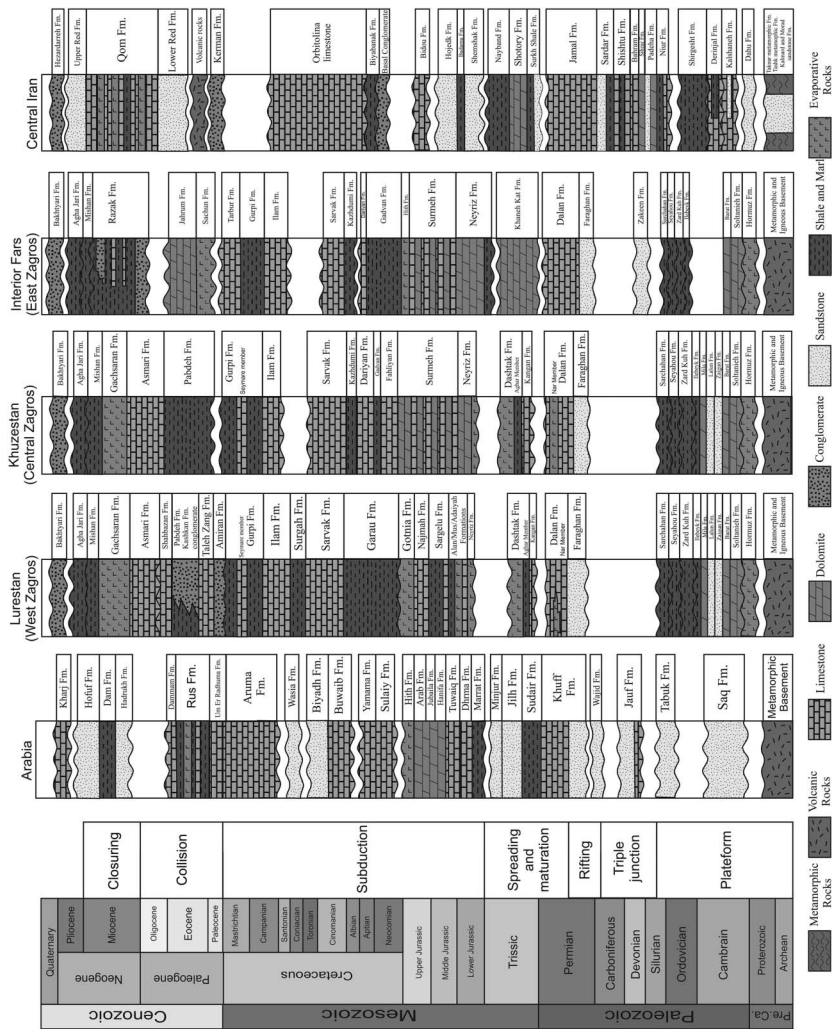


Fig. 5 Stratigraphic column of Arabia, Lurestan, Khuzestan, Interior Fars, and Central Iran (thickness not to scale)

Arabia (Fig. 5), it can be observed that from the Pre-Cambrian to the Late Ordovician, the two blocks of the Central Iran and Arabian Plates together formed a stable shield (Fig. 6; Gondwana Shield). In the Early Silurian (Caledonian Orogeny), this shield was divided into two segments of the Central Iran and Arabian Plates (Fig. 6). The uprising magmatic plumes and their under-plating caused swelling and bulging of the lithosphere (Fig. 7; [Lapierre et al. 2004](#); [Chauvet et al. 2009, 2011](#)). Subsequently, this swelling led to the termination of sedimentation and oceanic regression. Therefore, the lack of Ordovician and Silurian to Carboniferous strata along the Central Zagros and the existence of terrigenous sediments (Sare-Chahan and Zakeen Formations) in the Zagros Crush Zone indicate that during this time, the Neotethys plume was emplaced beneath the lithosphere of Gondwana and led to epeirogeny between the Central Iran and Arabian Plates (Fig. 5).

### **Continental rifting stage (Late Carboniferous–Early Permian)**

From the Late Silurian to the Early Carboniferous, the three arms of the Neotethys and its triple junction were probably formed (triple junction formation stage); then, by the eruption of plume magmas, the alkaline basalt of Dehbid and other extrusive rocks were generated (Fig. 6; [Ghasemi and Talbot 2006](#)). Magma eruption may have occurred during the Carboniferous and have led to subsidence along normal faults. This subsidence also created a shallow-water sedimentary basin. This basin is determined by the deposition of evaporates and terrigenous sediments (Faraghan and Dalan Formations). With continuing activity of normal faults, depth increased and the Neotethys Basin was formed ([Almutury and Al-Asadi 2008](#)). During the early stages of rifting, the basin was almost dry or contained fresh-water lakes (Fig. 7; [Almutury and Al-Asadi 2008](#)). Eventually, the floor of the Neotethys rift subsided and formed a shallow sea. This subsidence and spreading occurred from the Late Carboniferous to Liassic time ([Koop and Stoneley 1982](#); [Shahabpour 2005](#); [Sheikholeslami et al. 2008](#); [Homke et al. 2010](#); [Emami et al. 2010](#); [Saccani et al. 2013](#)). Continental sediments accumulated in the depressions of the down-faulted blocks, and basaltic magma was injected into the rift system (Fig. 7; [Almutury and Al-Asadi 2008](#)). Flood basalt could have extended over large areas of the rift zone during this phase ([Van der Pluijm and Marshak 1997](#); [Ghasemi and Talbot 2006](#)). Subsidence of the basin floor led to the formation of a passive margin along the northern and southern margins of the Arabian Plate (Figs 6–9; [Koop and Stoneley 1982](#); [Agard et al. 2005](#); [Jassim and Goff 2006](#); [Arfania and Shahriari 2009](#)). During the Late Permian to Late Triassic, the deep-water continental slope was formed along the northwestern and northeastern margins of the Arabian Plate (Fig. 6; [Faugères and Pujol 1991](#); [Shahabpour 2005](#); [Ghasemi and Talbot 2006](#); [Allahyari et al. 2010](#); [Yousefirad 2011](#); [Saccani et al. 2013](#)).

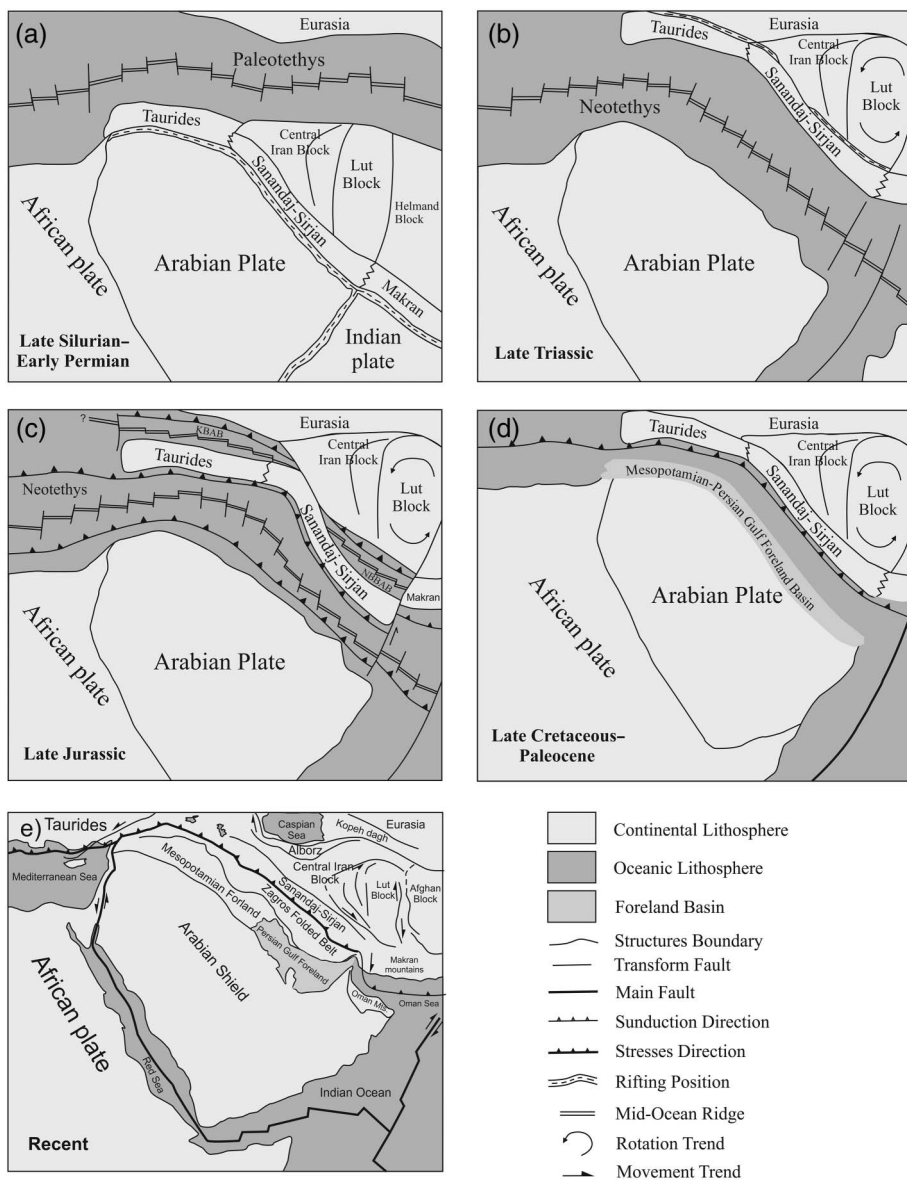


Fig. 6 Reconstruction history of the ZOB. (a) Formation of the Neotethys Basin triple junction. (b) Spreading and maturation in the Late Triassic. (c) Beginning of subduction of the southern Neotethys Basin and formation of the Khoy Back-arc Basin and Nain–Baft Back-arc Basin in the Early Jurassic, and then the beginning of subduction of the Neotethys back-arc basin in the Late Jurassic. (d) Closure of the Neotethys Back-arc Basin and the Neotethys southwestern subduction in the Late Cretaceous–Paleocene. (e) Recent structural elements of the ZOB

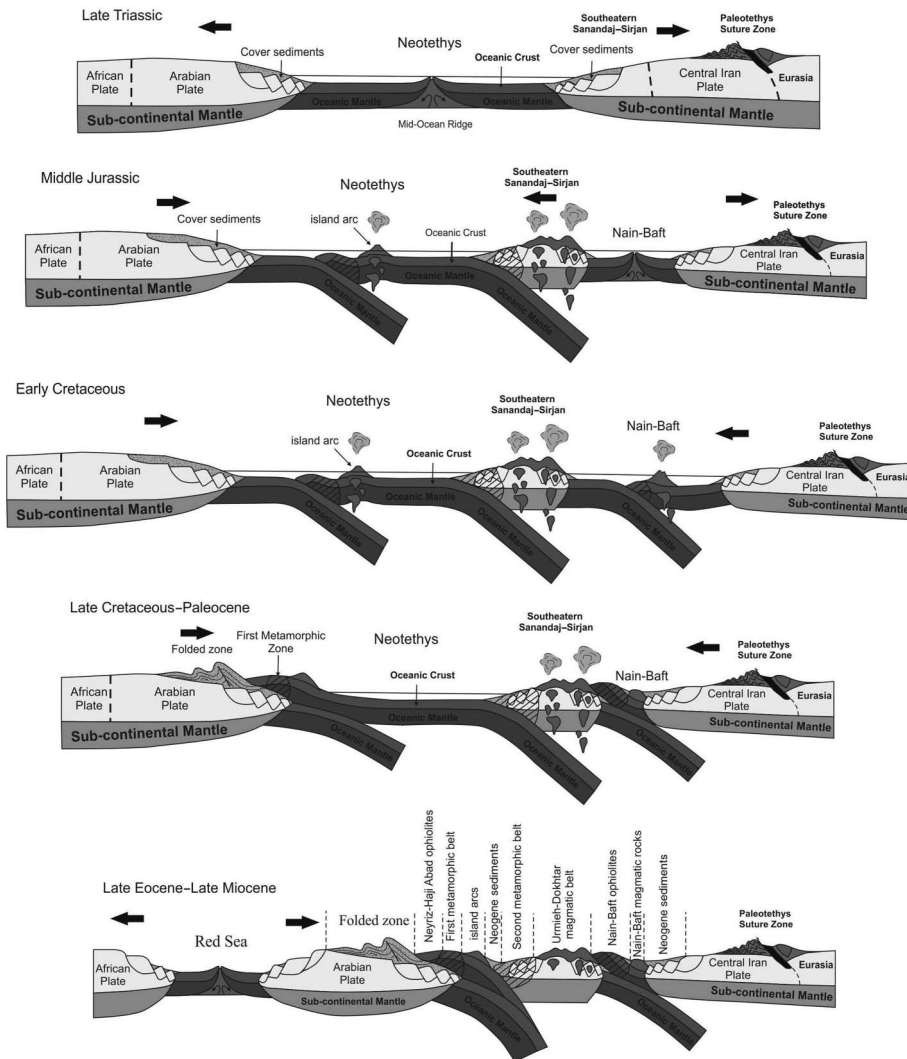


Fig. 7  
Reconstruction history of the SE ZOB

### Generation, spreading, and maturation of the oceanic lithosphere (Late Permian–Late Triassic)

At this stage, the Neotethys Ocean attained maximum width, and the shallow-water rift basin spread and led to the separation of the Central Iran continent and the Arabian Plate. This caused the extension of the narrow oceanic basin ([Şengör 1984](#); [Stampfli](#)

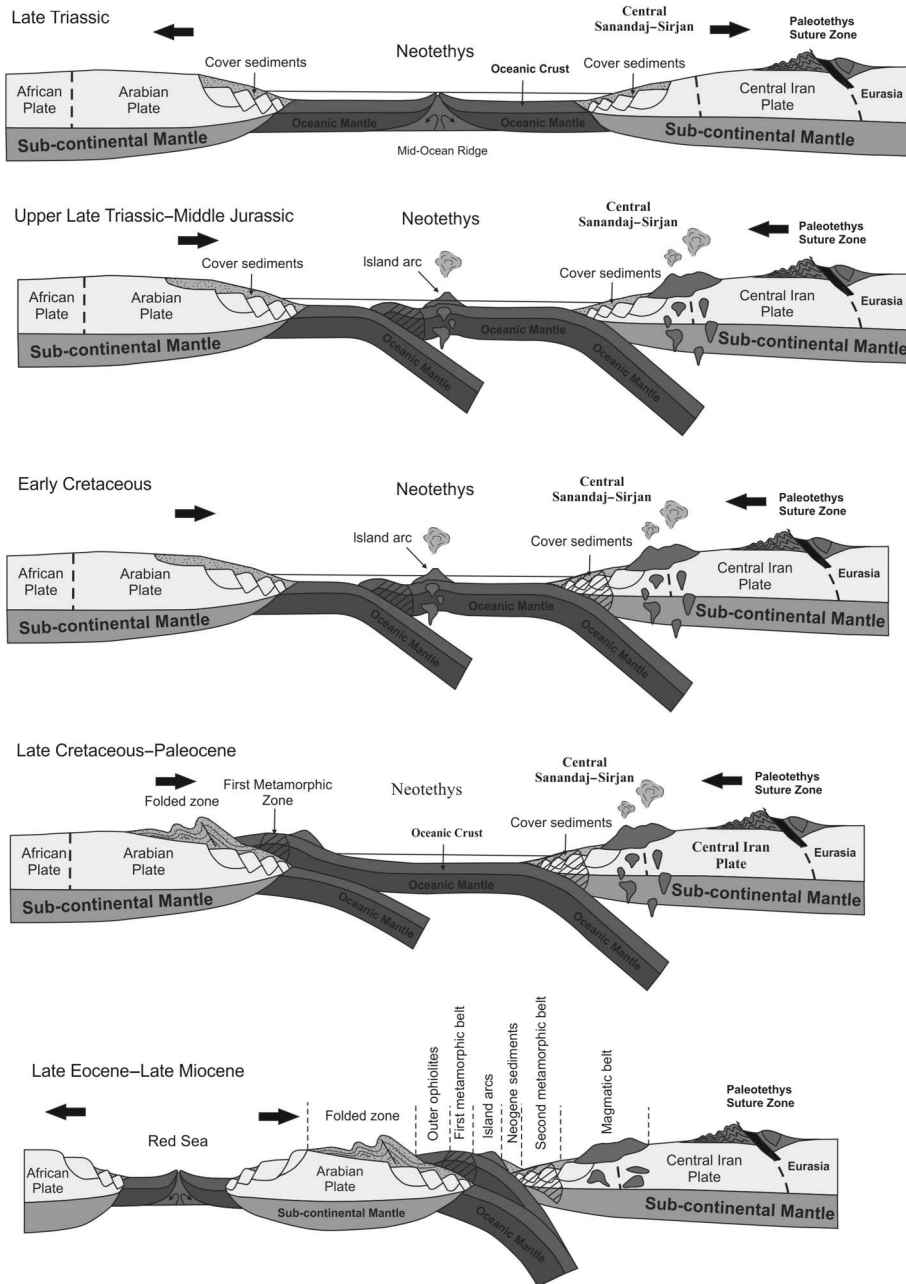


Fig. 8  
Reconstruction history of the central ZOB



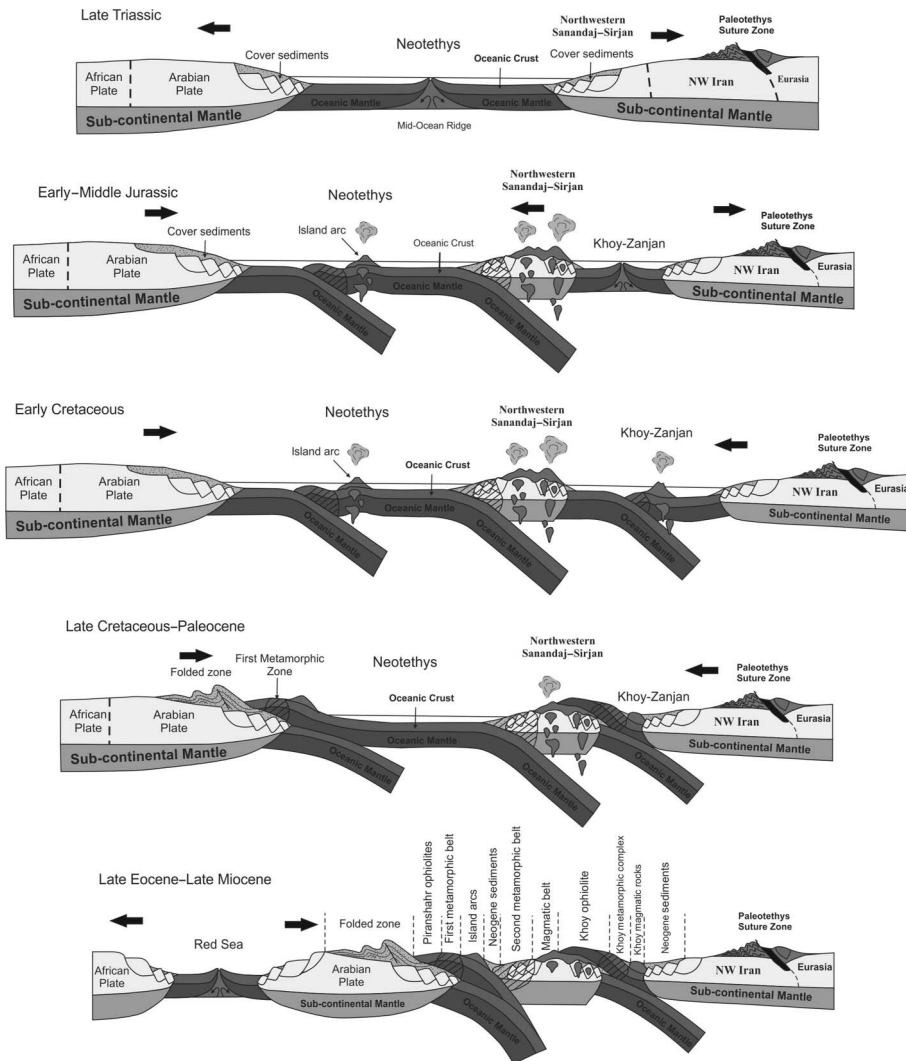


Fig. 9  
Reconstruction history of the NW ZOB

2000; Stampfli and Borel 2002; Robertson 2007). Through this spreading, the basin became deeper and wider (Fig. 6).

Due to the shallow nature of the basin during the rifting stages, evaporation was considerable; salt and sandstone was deposited on the floor of the rift as the Faraghan and Dalan Formations (Fig. 7). Thick salt and sandstone deposits of these formations lie at the base of the passive margin basin (Figs 5 and 6). At the center of the basin, where it

was deeper and wider, limestone of Khaneh Kat Formation was deposited (Figs 5 and 6). The spreading stage was terminated by the beginning of subduction related to the closure of the Paleotethys Basin in the northern Iran in the Late Triassic (Fig. 7).

### **Subduction of the oceanic lithosphere stage (Triassic–Late Cretaceous)**

At this stage, the Neotethys Basin, on account of the Arabian Plate movements caused by the opening of the Atlantic Ocean rift (Stampfli 2000; Bird et al. 2007), changed the divergence régime to one of the convergence (Fig. 6). The observed absence of sedimentation from Late Triassic to Lower Early Jurassic (Fig. 5) was a result of the closure of a Paleotethys Basin branch along North Iran (Berberian and King 1981; Bagheri and Stampfli 2008; Zanchi et al. 2009; Buchs et al. 2013). From the Late Permian to the Late Triassic, forces from the spreading of the Neotethys mid-ocean ridge caused the subduction of the Paleotethys oceanic lithosphere; following closure of the Paleotethys Ocean in the Late Triassic, tectonic forces caused epeirogeny in the Neotethys Basin. Then, during the Upper Late Triassic to the beginning of the Lower Jurassic, the high slab pushing forces of the Neotethys mid-ocean ridge caused a break in the Neotethyan oceanic lithosphere, which led to the onset of the subduction in the lower Early Jurassic (Figs 7–9). New studies by Chiu et al. (2013), which indicate a Middle Jurassic age for the oldest magmatic arc rocks of the SSMB and the UDMB, confirm this. The subduction led to the transformation of two passive margins into a passive margin and an active margin. The subduction of the Neotethys oceanic lithosphere beneath the Iranian and Anatolian continental lithosphere (McQuarrie and van Hinsbergen 2013) led to the eruption of magmas formed as volcanic arcs (Figs 7–9; Alavi 1994; Shahabpour 2007; Chiu et al. 2013). Likewise, it led to the formation of the Neotethyan back-arc rift in the Middle Jurassic age (Figs 6–9; Azizi et al. 2006; Ghasemi and Talbot 2006; Shafaii Moghadam et al. 2009; Ghazi et al. 2012).

The subduction period extended from the Upper Early Jurassic age to the Lower Miocene age (Mohajjel et al. 2003; Golonka 2004; Yousefirad 2011; Ghazi et al. 2012; Chiu et al. 2013). Therefore, the UDMB, with a Middle Jurassic age to Late Miocene age, can be considered as a volcanic arc of the Neotethys subduction system (Figs 7–9; Berberian and Berberian 1981; Berberian and King 1981; Berberian et al. 1982; Alavi 1994; Shahabpour 2007; Chiu et al. 2013). The UDMB consists predominately of intrusive and extrusive calc-alkaline igneous rocks and a low volume of shoshonitic and alkaline rocks (Chiu et al. 2013). The subduction of the Neotethys oceanic lithosphere created a tensional back-arc basin (Fig. 6; Kazmin and Tikhonova 2006; Azizi et al. 2006). The Khoy, Nain, Shahr Babak, Dehshir, and Baft Ophiolites mark the site of the Nain–Shahr Babak–Dehshir–Baft and of the Khoy Back-arc Basins, which opened between the SSMB and Central Iran (Fig. 6). Ghasemi and Talbot (2006) proposed an Upper Triassic–Lower Cretaceous age for the formation of the back-arc basin, whereas Ghazi et al. (2012) suggest an Early Jurassic time for its opening. Given that these ophiolites and their related igneous rocks have back-arc

basin features and that the Neotethys subduction began in the lower Early Jurassic age, rifting in this basin more likely occurred in the Middle Jurassic age. Likewise, Shafaii Moghadam et al. (2009) and Arfania and Shahriari (2009) believe that the lithosphere of this basin crumpled without subduction, whereas Ghazi et al. (2012), based on geochemical evidence from the related volcanic rocks with back-arc basin features, suggested that subduction in the back-arc basin began in the Late Jurassic age. Ghasemi and Talbot (2006) suggested the Late Cretaceous–Paleocene and Ghazi et al. (2012) suggested the Late Cretaceous (Maastrichtian) as closure time of this basin. Deposition of Late Cretaceous limestone over the Inner Zagros Ophiolite remnants confirms the Late Cretaceous as the closing time.

Concerning subduction in the southern Neotethys Basin, many researchers suggested that two subduction systems existed (Shahabpour 2005; Agard et al. 2005; Ghasemi and Talbot 2006; Afrania and Shahriari 2009; Allahyari et al. 2010; Yousefirad 2011; Ghazi et al. 2012; Saccani et al. 2013; Chiu et al. 2013), which is confirmed by the existence of both island-arc and volcanic-arc remnants (Figs 1 and 2). Agard et al. (2005) believe that the first subduction was oceanic–continental subduction. Following this, intra-oceanic fracturing formed an oceanic–oceanic subduction (second subduction). The position of the second subduction was close to the northern margin of Arabia (Kazmin and Tikhonova 2006; Agard et al. 2007; Allahyari et al. 2010; Saccani et al. 2013). There is no general agreement on when the second subduction was initiated. Desmons and Beccaluva (1983) and Dercourt et al. (1986) believe that the second subduction initiation time was Late Cretaceous. Ghasemi and Talbot (2006) declared mid-Cretaceous time at the start of the second subduction. Likewise, Agard et al. (2005) and Yousefirad (2011) suggested an Early Cretaceous age, whereas Saccani et al. (2013) suggested an early mid-Cretaceous age for this event. Agard et al. (2007) believe that stress accumulation in the weakest zone of the oceanic lithosphere led to buckling, rupture, and formation of a new subduction zone close to the Arabian Plate margin in the Lower Cretaceous that subsequently caused obduction of the oceanic lithosphere at the Arabian Plate margin. However, the presence of I-type arc-related calc-alkaline granitoid rocks of Siah Kuh of Early Jurassic age (199 Ma) from the SSMB (Berberian and Berberian 1981; Malekizadeh 1999; Arvin et al. 2007), of the Chah Dozdan granitic and Chah Ghand gabbroic rocks of Middle Jurassic age (Sheikholeslami 2002), of the lower Late Cretaceous Kermanshah boninitic intra-oceanic island arc (Allahyari et al. 2010; Saccani et al. 2013), and of low-to-medium-pressure metamorphism accompanied by coeval Jurassic andesitic intrusions and volcanism ranging from gabbro to granite with K/Ar ages between  $118 \pm 10$  Ma and  $164 \pm 4$  Ma along the South SSMB (Berberian and Berberian 1981; Afrania and Shahriari 2009) indicate that the beginning of the second subduction occurred prior to the Cretaceous, and was probably of Upper Late Triassic–Lower Early Jurassic timing. Therefore, both subductions simultaneously began in the Early Jurassic (Figs 7–9). Chiu et al. (2013), using U–Pb dating of 50 samples from intrusive and extrusive igneous rocks of the UDMB and the SSMB, distinguished five subduction-related magmatic episodes that include: (1) Middle to Late Jurassic

(I-type, 176–144 Ma), (2) Late Cretaceous (72–81 Ma), (3) Early Eocene to Late Oligocene (25–55 Ma), (4) Early to Late Miocene (22–6 Ma), and (5) Late Miocene to Late Pliocene (11–0.4 Ma) magmatism. The first episode was the result of two subductions, one is oceanic–oceanic and the other oceanic–continental, whereas the second and third episodes were the results of collision and continuing subduction.

The fourth episode represents the termination of the subduction-related magmatic activities, which ceased progressively from northwest to southeast, with magmatic activities ending in the Early Miocene (ca. 22 Ma) in Meghri (Armenia), the Middle Miocene (ca. 16 Ma) in Kashan, and the Late Miocene (ca. 10–6 Ma) in Anar, respectively. The fifth episode indicates post-collisional volcanism that began ca. 11 Ma in the Saray (eastern side of Lake Urmia), along with later eruptions in the Sahand (6.5–4.2 Ma) and the Sabalan ( $\leq 0.4$  Ma) volcanoes (Chiu et al. 2013). Based on the mentioned ages and episodes, two main magmatic gaps (between the Late Jurassic to Late Cretaceous and the Late Cretaceous to Early Eocene – Azizi and Moinevaziri 2009; Chiu et al. 2013) can be distinguished, which probably resulted from subduction of a cold oceanic lithosphere and/or low-angle subduction (Chiu et al. 2013).

### **Collision stage (Late Cretaceous–Late Miocene)**

At this stage, the Neotethys oceanic lithosphere was totally consumed. Following this, passive continental margin and island-arc collision occurred in the Late Cretaceous (Golonka 2004; Kazmin and Tikhonova 2006; Allahyari et al. 2010; Saccani et al. 2013). This collision caused obduction of the Neotethys oceanic lithosphere remnants into the Arabian Plate margin, but in the Zagros Crush Zone, there remained a shallow-water oceanic basin, where Upper Cretaceous limestone was deposited (Fig. 5; Alavi 1994; Ricou 1974; Haynes and McQuillan 1974; Babaei et al. 2005; Ghasemi and Talbot 2006; Yousefirad 2011). Erosion of continental collisional highs carried an enormous volume of detrital materials into the remaining shallow-water basin of the Neotethys. The Amiran Formation was formed in this way (Fig. 5; Braud 1987; Ghasemi and Talbot 2006; Yousefirad 2011). The foreland basin formed following the collision is the Mesopotamian–Persian Gulf of NW–SE trend; the hinterland basins are the depressions that are usually located between Zagros and Central Iran (Fig. 1; Homke et al., 2010; Emami et al. 2010; Saura et al. 2011). After the closure of the oceanic lithosphere between the Arabian Plate margin and the Neotethys island arcs, and the emplacement of the Outer Zagros Ophiolites, the subduction of Neotethys oceanic lithosphere between the island arcs and the Sanandaj–Sirjan Zone continued until the Late Miocene (Figs 6–9). Based on the southern Neotethys Basin structural elements that include Foreland Basin Sediments, Fold and Thrust Belt, Crush Zone, ophiolite remnants, the First Metamorphic Belt, island-arc magmatic rocks, the Second Metamorphic Belt, and the magmatic belt (Fig. 1), two subduction systems and two collisions can be considered for evolution of the Tethysides along the Zagros Orogen. The beginning of both subductions occurred in the Upper Late Triassic, but collision occurred at different times.

The first collision was due to the closure of the basin between the passive continental margin and the forearc in the Late Cretaceous; the second collision was caused by basin closure between the back-arc and the active continental margin, which occurred in the Late Miocene (Figs 6–9). A large amount of evidence indicates that the first collision occurred in the Upper Cretaceous. The Neyriz Ophiolite is unconformably covered by the Late Cretaceous anhydritic limestone of the Tarbur Formation (Fig. 4; Ricou 1968, 1974). The Amiran Formation conglomerates of Maastrichtian–Paleocene age (Fig. 5) have ophiolite clasts, showing that the emplacement time of these ophiolites is pre-Maastrichtian. Paleocene volcanism and Eocene shallow-water limestone deposition unconformably cover the Kermanshah Ophiolite (Braud 1987; Ghasemi and Talbot 2006). The Haji-Abad Ophiolite is overlain by Upper Cretaceous pelagic limestone. Based on  $^{40}\text{Ar}/^{39}\text{Ar}$  dating, the Neyriz Ophiolite complex formed at 96–98 Ma (Haynes and Reynolds 1980) and was emplaced at 89 Ma (Lanphere and Pamić 1983; Ghasemi and Talbot 2006). The Nain–Shahr Babak–Dehshir–Baft Ophiolites are also covered by Coniacian–Maastrichtian pelagic limestone (Babaei et al. 2005; Ghasemi and Talbot 2006; Shafaii Moghadam and Stern 2011). Therefore, the presence of an Upper Cretaceous unconformity, coverage by Upper Cretaceous limestone and volcanic rocks, and ophiolitic clasts in the Amiran conglomerate all indicate that in the Upper Cretaceous, the Neotethys Back-arc Basin between the Central Iran and Sanandaj–Sirjan was closed, and likewise that the oceanic–oceanic subduction of the southern Neotethys Basin was terminated (Figs 6–9). The ophiolite remnants obducted onto the Arabian Plate margin and Sanandaj–Sirjan, and the Inner and Outer ophiolites were emplaced. By emplacement of these ophiolites, a foreland basin was formed in front of the outer ophiolitic belt; ophiolite–radiolarite provided the detritic materials supplying the Upper Maastrichtian–Paleocene Amiran flysch deposit (Berberian and King 1981; Alavi 1994; Hooper et al. 1995; Homke et al. 2010). The Upper Maastrichtian–Paleocene Amiran flysch accumulated along a linear trough in the High Zagros area.

The final closure of Neotethys and the collision between the Arabian and Central Iran Plates took place in the Cenozoic. Agard et al. (2005) suggested the Late Oligocene for collision between Iranian and Arabian margins, whereas Navabpour et al. (2007) believe that collision occurred between 20 and 10 Ma age (Late to Middle Miocene). Chiu et al. (2013), based on the termination of calc-alkaline magmatism in the UDMB, showed that the collision age is Early to Late Miocene. This timing is consistent with the geologic evidence of the final closure of the Neotethys Ocean between Arabia and Eurasia at ~20 Ma in the Bitlis Suture Zone, to the NW of the Zagros (Okay et al. 2010).

## Conclusions

The purpose of this contribution is to present a comprehensive reconstruction model for the evolutionary history of the Zagros Orogen that justifies most petrologic,

tectonomagmatic, and stratigraphic characteristics. The formation of the Neotethyan triple junction began during the Late Silurian to Early Carboniferous. One of the three arms of the triple junction subsided, and began spreading, which resulted in a shallow-water basin during the Late Carboniferous–Early Permian. Subsiding and spreading continued from the Late Carboniferous to the Late Triassic. In the Late Triassic, the Neotethys Basin reached its maximum extension. From the Late Permian to the Late Triassic, the force of the Neotethys mid-ocean ridge spreading was consumed by subduction of the Paleotethys oceanic lithosphere. After the closure of the Paleotethys Ocean during the Late Triassic, tectonic forces caused epeirogeny in the Neotethys Basin that eventually led to the simultaneous initiation of two northeastward subductions during Upper Late Triassic. Oceanic–oceanic subduction continued until the Late Cretaceous and was terminated at this time, but oceanic–continental subduction continued to the Middle–Late Miocene.

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