# A GEOCHEMICAL ATTEMPT TO DISTINGUISH FOREARC AND BACK ARC OPHIOLITES FROM THE "SUPRA-SUBDUCTION" CENTRAL ANATOLIAN OPHIOLITES (TURKEY) BY COMPARISON WITH MODERN OCEANIC ANALOGUES

**M. Kenan Yalınız**

*Celal Bayar University, Manisa, Turkey (e-mail: mkenan@bayar.edu.tr).*

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## ABSTRACT

The Central Anatolian Ophiolite (CAO) includes oceanic crust and mantle fragments and contains all the components of a typical ophiolitic sequence: metamorphic tectonites, cumulates, isotropic gabbros, plagiogranites, dolerite sheeted dykes, basaltic lavas and sedimentary cover. They are found as dismembered but partially preserved allochthonous bodies in the Central Anatolian Crystalline Complex (CACC) representing the metamorphosed passive northern edge of the Tauride-Anatolide Platform (TAP), central Turkey. Geochemically, the magmatic rock units of the CAO display part of a dominant co-magmatic differentiated series of island-arc tholeiites (IAT). In addition, IAT are overlain by a subordinate group of boninite-like basalts which are chemically and mineralogically intermediate between IAT and more depleted boninites. The variations in lava chemistry of the CAO reflect eruption of progressively more depleted magmas through time and point to diverse mantle source compositions and partial melting.

Detailed chemical analyses of the magmatic units of the CAO revealed typical supra-subduction zone (SSZ) features with depleted high field strength elements (HFSE) and light rare earth elements (LREE: $La_N/Yb_N$: < 1) and enriched large-ion lithophile elements relative to normal mid-oceanic ridge (N-MORB) and back-arc basin basalts (BABB). In this respect the CAO is similar to oceanic crust generated in the Izu-Bonin and Mariana fore arcs. A N-MOR or BAB spreading seems unlikely. However, progressive depletion in the lava sequence and absence of calc-alkali basalts and their differentiates indicate that the CAO formed at an initial stage of subduction from previously depleted MORB mantle (DMM) and oceanic lithosphere, prior to development of any island arc within the Vardar-İzmir-Ankara-Erzincan (VIAE) ocean segment of Neotethys. Accordingly, a forearc setting proposed for the genesis of the CAO is inappropriate and misleading. It was generated above a short-lived north-dipping nascent intra-oceanic subduction zone during early-middle Turonian-early Santonian, then, it was rapidly emplaced southwards onto the CACC, soon after formation between post-early Santonian and pre-middle Campanian.

## INTRODUCTION

Ophiolites are preserved fragments of oceanic crust that have been tectonically emplaced onto continental crust during the closure of their oceans. In the past, ophiolites were generally believed to have formed at N-MOR along divergent plate margins. However, by the early 1980s, it was recognized that most ophiolites have geochemical characteristics of subduction zones that could not be explained by spreading at a N-MOR (e.g. Malpas and Langdon, 1984; Pearce et al., 1984; Saunders and Tarney, 1984; Rautenschlein et al., 1985). There is now widespread acceptance that many ophiolites were formed in SSZ settings. The increasing number of studies on ophiolites that contain rocks interpreted to have SSZ affinities along tectonic suture belts, have led some authors to conclude that essentially all ophiolites are of SSZ origin and that ophiolites representative of N-MOR are scarce or non-existent (e.g., Dick and Bullen, 1984; Moores et al., 1984; Searle and Stevens, 1984; Hawkins, 2003). SSZ ophiolites were generated within intra-oceanic subduction zone systems, prior to collisions, during ocean closure at convergent plate margins. The new oceanic crust formed in forearc or back-arc basins, as well as in the nascent stage of arc volcanism (pre-arc; Pearce et al., 1984) above a subducting slab. This is the so called SSZ tectonic setting, a term first proposed by Pearce et al. (1984). SSZ ophiolites formed by melting of the wedge-shaped asthenosphere and/or lithospheric mantle above a subducting slab. However, recent studies on geology and maturity of modern SSZ systems of the Western Pacific Ocean, (e.g., Izu-Bonin-Mariana -IBM- arc trench systems) indicate that these environments are highly complex and can exhibit elements of all three of the possible SSZ tectonic settings. Thus, it is difficult to distinguish whether SSZ-type ophiolite was formed in a forearc or back-arc basin and before or after the development of an arc (Pearce et al., 1984). Most of the available literature does not discuss in detail the precise tectonic setting of SSZ ophiolites. As SSZ processes have a key role in ocean closure, distinction among these tectonic environments can lead to better definition of regional tectonic models. Ophiolitic exposures in Central Anatolia (Central Anatolian Ophiolites, CAO of Göncüoğlu et al., 1991) play a key role not only for a better understanding of Neotethys evolution but also for the forearc, back-arc and pre-arc settings distinction. The present work presents chemical constraints for recognition of these SSZ type environments.

Regarding the regional geologic setting, it is now frequently accepted that the CAO is made of fragments of oceanic lithosphere generated within the Vardar-İzmir-Ankara-Erzincan (VIAE) ocean segment of the Neotethys (e.g., Özgül, 1976; Şengör and Yılmaz, 1981; Göncüoğlu and Türeli, 1993; Koçak and Leake, 1994; Yalınız et al, 1996; 1999; Floyd et al., 1998; 2000; Robertson, 2002; 2004). Detailed studies of the pseudo-stratigraphic relationships and geochemistry of magmatic units show that the CAO exhibits SSZ type affinity (e.g.,Yalınız et al., 1996; 2000a). However, generation of the CAO in a SSZ tectonic setting such as a nascent volcanic arc (pre-arc), back-arc or forearc basins is unclear. Thus, to better understand the evolution of the VIAE Ocean, it is critical to specify the paleotectonic setting of the CAO.

The following section describes geological and geochemical features of the CAO that we interpret to strongly support their SSZ origin within the VIAE Ocean. Interpretation is mainly based on much more detailed description in Göncüoğlu et al. (1991; 1992; 1993), Göncüoğlu and Türeli (1993), Yalınız (1996), Yalınız et al. (1996; 1999; 2000a; 2000b), Yalınız and Göncüoğlu (1998; 2000) Floyd et al. (1998; 2000) to which the reader is referred. In this study, together with the data from the better known SSZ-type Çiçekdağ Ophiolite (CO) and Sarıkaraman Ophiolite (SO), new incoming trace and REE data leading to reinterpretation of the petrogenesis of the CAO have also been used from the Bozkır Ophiolite (BO) (Table 1).

The aim of this paper is to re-examine the evolution of SSZ environments within the VIAE Ocean by discussing literature and new geochemical data recovered from the CAO, and to constrain their possible specific paleoenvironment based on correlation with well known SSZ type modern oceanic analogues. This paper also aims to discuss the differences between forearc, backarc and pre-arc tectonic environments in terms of magmatic affinities and magma source characteristics.

## CENTRAL ANATOLIAN OPHIOLITE (CAO)

The ophiolitic units cropping out in the Central Anatolia were initially defined by Özgül (1976) and later by Şengör

Table 1. Representative chemical analyses of basaltic lavas from the Central Anatolian Ophiolites.

| | Sarıkaraman (+) | | | Çiçekdağ (□;○) | | | Bozkır (◊) | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Sample | P-11 | M-82 | 157-D | C-58 | C-68 | C-79 | BD-1 | BD-2 | BD-3 | BD-9 | BD-10 | BD-12 | BD-23 |
| SiO2 | 52,4 | 48,6 | 53,2 | 52,3 | 54,6 | 50,5 | 50,8 | 51,4 | 50,4 | 51,1 | 48,8 | 53,7 | 51,6 |
| TiO2 | 0.7 | 0,54 | 0,87 | 0,77 | 0,55 | 0,65 | 0,42 | 0,45 | 0,68 | 0,59 | 0,5 | 0,45 | 0,61 |
| Al2O3 | 15,1 | 15,8 | 16 | 15 | 14,8 | 15,2 | 15 | 15,3 | 16,5 | 17,6 | 17,1 | 15,7 | 17,1 |
| Fe2O3 | 8,71 | 10,8 | 10,6 | 9,99 | 9 | 10 | 8,96 | 8,66 | 9,1 | 8,58 | 10,7 | 9,86 | 8,31 |
| MnO | 0,21 | 0,15 | 0,23 | 0,14 | 0,16 | 0,15 | 0,14 | 0,14 | 0,19 | 0,15 | 0,04 | 0,18 | 0,16 |
| MgO | 12,1 | 8,15 | 3,51 | 4,66 | 5,99 | 7,57 | 9,66 | 8,84 | 7,62 | 6,44 | 8,37 | 5,72 | 6,4 |
| CaO | 1,68 | 7,3 | 6,36 | 4,69 | 7,49 | 8,53 | 10,4 | 11,7 | 11,4 | 11,1 | 19,1 | 9,25 | 11 |
| Na2O | 2,53 | 1,71 | 4,29 | 6,19 | 4,22 | 3,09 | 1,24 | 1,71 | 1,99 | 2,36 | 2 | 3 | 2,63 |
| K2O | 0,01 | 0,13 | 0,26 | 0,12 | 1,5 | 1,38 | 0,11 | 0,12 | 0,34 | 0,19 | 0,08 | 0,44 | 0,19 |
| P2O5 | 0,06 | 0,02 | 0,04 | 0,07 | 0,05 | 0,06 | 0,01 | 0,04 | 0,06 | 0,03 | <0,01 | 0,05 | 0,09 |
| Cr2O3 | | | | | | | 0,06 | 0,06 | 0,05 | 0,04 | 0,05 | 0,01 | 0,22 |
| LOI | 6,82 | 6,71 | 4,89 | 4,9 | 1,57 | 2,88 | 3 | 1,6 | 1,7 | 1,7 | 1,4 | 1,6 | 1,4 |
| Total | 100 | 99,9 | 100 | 99,8 | 99,9 | 100 | 99,8 | 101 | 100 | 100 | 100 | 100 | 100 |
| Ba | 19 | 21 | 30 | 39 | 262 | 87 | 30 | 77 | 27 | 52 | 112 | 60 | 72 |
| Ga | 12 | 16 | 18 | 12 | 13 | 13 | 10,7 | 11,9 | 13,7 | 17 | 14,6 | 13,4 | 16,5 |
| Nb | 1 | 1 | 3 | 0,41 | 0,41 | 0,5 | <0,5 | <0,5 | <0,5 | 2,3 | <0,5 | <0,5 | 2,7 |
| Pb | 11 | 6 | 9 | 25 | 8 | 6 | | | | | | | |
| Rb | 5 | 7 | 13 | 4 | 27 | 24 | 3,3 | 2,4 | 17,5 | 5,8 | 0,8 | 11,3 | 5,8 |
| Sr | 34 | 105 | 175 | 106 | 84 | 164 | 92,8 | 129 | 127 | 153 | 77,9 | 75,6 | 152 |
| V | 252 | 347 | 155 | 443 | 286 | 293 | 204 | 229 | 245 | 244 | 276 | 260 | 241 |
| Y | 11 | 11 | 21 | 19 | 19 | 15 | 10,9 | 12,2 | 16,5 | 13,1 | 10,3 | 15,3 | 14,4 |
| Zn | 171 | 67 | 116 | 95 | 72 | 109 | | | | | | | |
| Zr | 54 | 26 | 39 | 32 | 39 | 33 | 17,1 | 19,7 | 26,2 | 40,8 | 11,3 | 21,2 | 42,7 |
| Hf | 1 | 0,4 | 0,9 | 0,73 | 0,83 | 0,73 | 0,6 | 0,7 | 0,9 | 1,5 | <0,5 | 0,9 | 1,5 |
| Sc | 32 | 41 | 41 | 38,8 | 38,3 | 46,7 | 45 | 43 | 37 | 38 | 45 | 45 | 38 |
| Ta | 0,07 | 0,06 | 0,1 | 0,03 | 0,04 | 0,04 | <0,1 | <0,1 | <0,1 | <0,1 | <0,1 | <0,1 | 0,1 |
| Th | 0,2 | 0,3 | 0,4 | 0,06 | 0,11 | 0,07 | 0,2 | 0,1 | 0,1 | 2,6 | <0,1 | <0,1 | 1,8 |
| U | 0,2 | 0,1 | 0,1 | 0,11 | 0,2 | 0,12 | 0,1 | 0,2 | <0,1 | 0,9 | <0,1 | <0,1 | 0,7 |
| La | 1,3 | 1,2 | 1,7 | 1,59 | 1,47 | 1,18 | 0,9 | 0,7 | 2,4 | 8,2 | 0,5 | 1,5 | 7,3 |
| Ce | 3 | 4 | 5 | 3,63 | 3,94 | 3,39 | 3,5 | 3,8 | 5,2 | 15 | 2,5 | 4,4 | 15,1 |
| Pr | | | | 0,65 | 0,67 | 0,57 | 0,37 | 0,48 | 0,67 | 1,66 | 0,25 | 0,62 | 1,66 |
| Nd | 3 | 3 | 4 | 3,64 | 3,73 | 3,17 | 2,5 | 3,1 | 4,3 | 8,4 | 2 | 3,9 | 8,6 |
| Sm | 0,86 | 0,88 | 1,51 | 1,27 | 1,34 | 1,13 | 0,9 | 1,2 | 1,6 | 2,2 | 0,7 | 1,2 | 2,1 |
| Eu | 0,29 | 0,36 | 0,82 | 0,5 | 0,56 | 0,45 | 0,35 | 0,45 | 0,67 | 0,57 | 0,42 | 0,49 | 0,67 |
| Gd | 1,5 | 1,6 | 2,3 | 1,83 | 1,97 | 1,64 | 1,51 | 1,79 | 2,5 | 2,18 | 1,36 | 1,9 | 2,03 |
| Tb | 0,3 | 0,3 | 0,5 | 0,34 | 0,37 | 0,31 | 0,24 | 0,32 | 0,42 | 0,35 | 0,24 | 0,38 | 0,38 |
| Dy | | | | 2,22 | 2,41 | 2,05 | 1,94 | 2,13 | 2,74 | 2,36 | 1,7 | 2,68 | 2,57 |
| Ho | | | | 0,49 | 0,54 | 0,45 | 0,39 | 0,48 | 0,6 | 0,5 | 0,4 | 0,55 | 0,53 |
| Er | | | | 1,46 | 1,6 | 1,32 | 1,18 | 1,25 | 1,81 | 1,43 | 1,12 | 1,62 | 1,54 |
| Tm | | | | 0,25 | 0,26 | 0,22 | 0,18 | 0,22 | 0,29 | 0,24 | 0,17 | 0,28 | 0,27 |
| Yb | 1,25 | 1,26 | 2,19 | 1,5 | 1,61 | 1,35 | 1,28 | 1,4 | 1,85 | 1,59 | 1,17 | 1,59 | 1,68 |
| Lu | 0,19 | 0,18 | 0,32 | 0,24 | 0,26 | 0,22 | 0,18 | 0,23 | 0,3 | 0,25 | 0,18 | 0,28 | 0,23 |

and Yılmaz (1981) as "Bozkır Units". They did not indicate any specific tectonic setting for their evolution; however, they proposed that ophiolites were derived from the VIAE and obducted towards south onto the Central Anatolian Crystalline Complex (CACC). Detailed field and laboratory work in the 1990's (Göncüoğlu et al. 1991, 1992, 1993; Yalınız 1996; Yalınız et al. 1996; 1999; 2000a; 2000b; Köksal-Toksoy et al. 2001; Floyd et al., 1998; 2000), however, showed that the CAO represents SSZ type oceanic crustal and mantle fragments and contains all of the components of a typical ophiolitic sequence: metamorphic tectonites, cumulates, isotropic gabbros, plagiogranites, dolerite sheeted dykes, basaltic lavas and sedimentary cover. They are found as dismembered but partially preserved allochthonous bodies in the CACC (Fig. 1). The CACC represents the metamorphosed passive northern edge of the Tauride-Anatolide platform (for a brief review see Yalınız and Göncüoğlu, 1998; Yalınız et al., 1999; 2000a). The CAO was generated above a short-lived north-dipping intra-oceanic subduction within the northern branch of the Neotethyan Ocean (VIAE Ocean) in early-middle Turonian to early Santonian (Yalınız et al., 1996; 2000b) and emplaced southwards onto the passive northern edge of the Tauride-Anatolide platform, between post-early Santonian and pre-middle Campanian (Yalınız et al., 1996; 1999; 2000a; 2000b). After emplacement, CAO was intruded by upper Campanian-lower Maastrichtian post-collisional (post-COLG) granitoids (Yalınız et al. 1999; Köksal and Göncüoğlu, 2008) (Fig. 1).

### Rock type and stratigraphy

The CO, SO and BO, though somewhat dismembered, represent the best documented examples of SSZ-type CAO. Ultramafic rocks are not observed in direct contact with them in these areas but in the NE Niğde area (Göncüoğlu et al, 1991). Their limited occurrence suggests that they mainly represent the upper part of the oceanic lithosphere. In the following sections, the CAO rock units and stratigraphic relationships will be summarized (Table 2), considering earlier studies (Göncüoğlu et al., 1991; 1992; 1993; Yalınız, 1996; Yalınız et al., 1996; 1999; 2000; 2000b; Yalınız and Göncüoğlu, 1998; 2000; Floyd et al., 1998; 2000).

### Sarıkaraman Ophiolite (SO)

The SO is one of the best exposed bodies of the CAO, it consists of a co-magmatic differentiated series of tholeiites in which the following units were recognized, starting from the bottom: low-Ti isotropic gabbros, ocean ridge (ORG)-type plagiogranites, dolerite dykes, basaltic volcanics (pillows, massive lavas and breccias) and ophiolitic sedimentary cover of early-middle Turonian to early Santonian age (Table 2, Fig. 1). Isotropic gabbros, coarse to fine grained, are frequently amphibolized and easily distinguished in the field by the absence of cumulate layering. They include random or en-echelon plagiogranitic veins and pods. The veins and pods, gradually pass upwards into plagiogranitic dykes. Plagiogranitic bodies display simple narrow fracture infillings to a wide complex zone of net veining or agmatites having numerous gabbro enclaves. Geochemically, plagiogranites are ORG-type trondhjemites in composition. Dolerite dykes are entirely subvertical to vertical. They display typical 'dyke-in-dyke' structure and 'asymmetrical chilling'. The volcanic section of the SO mainly includes pillow lavas interbedded with

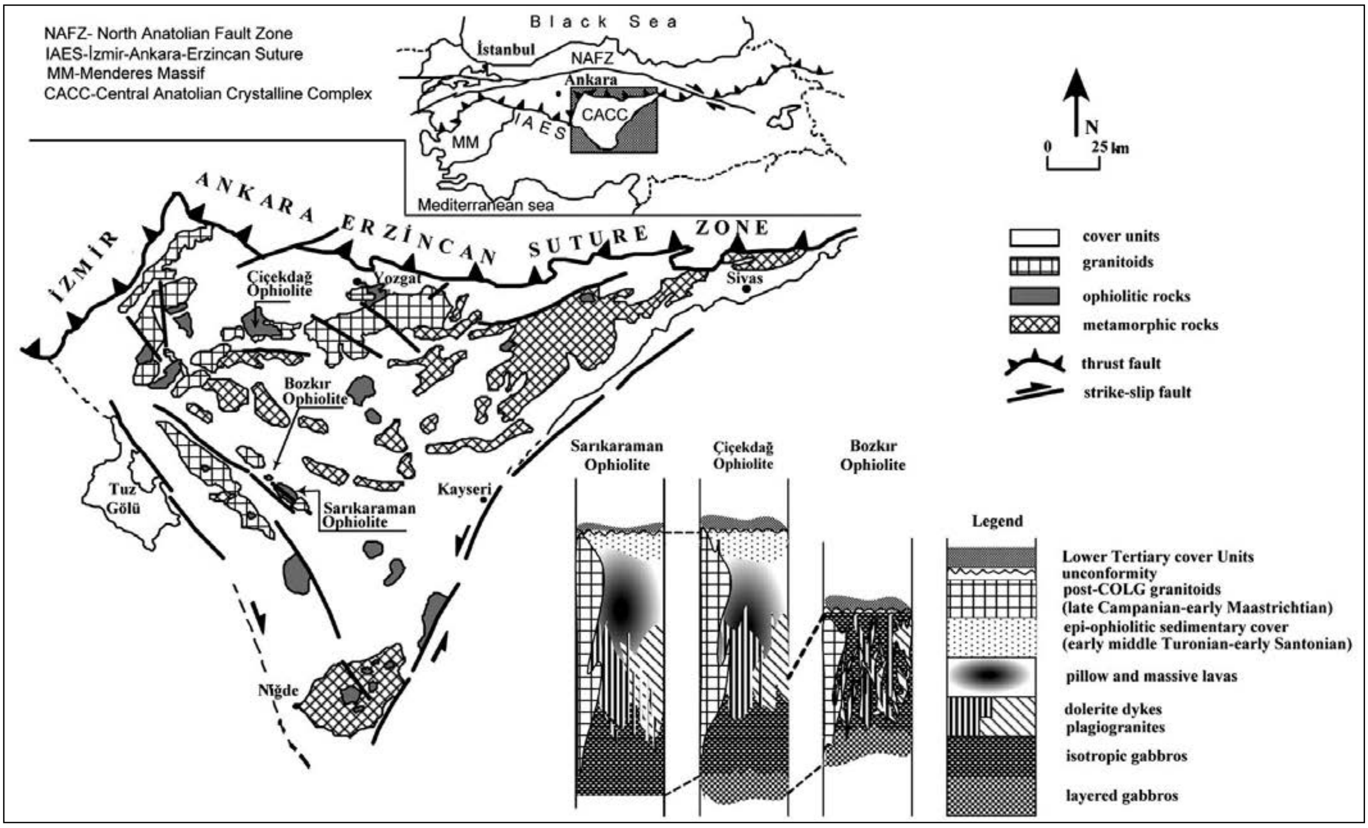


Fig. 1 - Simplified geological map of the Central Anatolian Crystalline Complex, showing locations and generalized columnar sections of the Bozkır, Çiçekdağ and Sarıkaraman Ophiolites.

massive lava flows and occasionally pillow breccias. It is also characterized by various high-level rhyolitic dykes and sills that traverse the upper volcanic section. In the uppermost section, the SO is characterized by the presence of lower-middle Turonian to lower Santonian sedimentary cover intercalated with the uppermost basaltic lavas. The SO is intruded by upper Campanian-lower Maastrichtian post-COLG-type quartz-monzonite.

### Çiçekdağ Ophiolite (CO)

The typical ophiolitic igneous lithologies are also present in the Çiçekdağ region (Table 2, Fig. 1). Well preserved and unfaulted sequences are rare, however, they locally exhibit much of the original structure. The CO comprises from bottom to top: (i) Low-Ti layered to isotropic gabbro; (ii) ORG-type plagiogranite (trondhjemite), (iii) sheeted dyke complex; (iv) basaltic lavas (mainly pillow lavas, with subordinate amounts of massive, rare basaltic breccias); (v) Turonian-Santonian ophiolitic sedimentary cover intercalated with the basaltic volcanics (e.g., pink pelagic cherty limestone). The magmatic rock units, like the SO, represent co-magmatic series of differentiated tholeiites. The gabbroic sequence is mainly composed of isotropic gabbro and subordinate layered gabbro. The sheeted dyke complex is markedly less extensive than the gabbros and locally has plagiogranitic dykes. The lavas normally exhibit pillow structure which is better developed at the top of the volcanic sequence. Massive basalts are also present. The CO was successively intruded by post-COLG granitoids (Akıman et al., 1993) (Table 2).

Table 2. Comparison among the Central Anatolian Ophiolites (Sarıkaraman, Çiçekdağ and Bozkır Ophiolites).

| | Lithology | Geochemistry | Age Constraints | Tectonic Setting |
|---|---|---|---|---|
| SARIKARAMAN OPHIOLITE | cover units | | early Palaeocene late Maastrichtian | |
| | Terlemez quartz monzonite | H-type granite | early Maastrichtian late Campanian- (K/Ar: 81-65 Ma) | post-COLG |
| | epi-ophiolitic sedimentary cover | | to early Santonian early-middle Turonian | |
| | basalt units | IAT | | SSZ |
| | dyke coplex | IAT | | SSZ |
| | plagiogranites | ORG-type granite | | SSZ |
| | isotropic gabbros | low-Ti IAT | | SSZ |
| ÇİÇEKDAĞ OPHIOLITE | cover units | | Lutetian | |
| | granitoids | | Late Cretaceous | post-COLG |
| | epi-ophiolitic sedimentary cover | | Turonian-Santonian | |
| | upper basalt units | transitional boninites (TiO2<0.5) | | SSZ |
| | lower basalt units | IAT (TiO2>0.5) | | SSZ |
| | dyke coplex | IAT | | SSZ |
| | plagiogranites | ORG type granite | | SSZ |
| | isotropic gabbros layered gabbros | low-Ti IAT | | SSZ |
| BOZKIR OPHIOLITE | cover units | | Quaternary | |
| | granitoids | | Late Cretaceous | post-COLG |
| | dyke complex | IAT | | SSZ |
| | plagiogranites | ORG type granite | | SSZ |
| | isotropic gabbros | | | SSZ |
| | rare layered gabbros | low-Ti IAT | | |

### Bozkır Ophiolite (BO)

The Bozkır Ophiolite is one of the isolated members of the CAO (Table 2, Fig. 1). It is particularly significant in the study of rocks formed from basic-acid fractional crystallization at high-level intrusives of supra-subduction zone type CAO.

In most of the complete ophiolite complexes, between the underlying layered cumulate sequence and the overlying sheeted dyke complex, there is generally a thin and irregular unit of "high-level" gabbros. They are characterised by variable textures and by the absence of well-defined igneous layering and include small bodies of fractionated intermediate to acid plagiogranites.

BO comprises a basic-acid fractionation sequence containing dominant isotropic gabbro, basaltic-andesitic dykes and plagiogranites (Table 2, Fig. 1). Isotropic gabbros are non-layered but occasionally laminated with planar orientation of the constituent minerals. Gabbros are characterised by the presence of abundant plagiogranitic patches, dykes up to 30 cm across and net veins which display gradational to sharp contacts with the gabbro host and with numerous partly assimilated gabbro enclaves. These plagiogranitic veins and dykes in the gabbro are usually characterized by well defined irregular to angular margins without marginal chilling but there are gradational to diffused margin types. The veins are commonly associated with coarse to acicular amphiboles and gabbro pegmatites. Since they appear to have no feeder veins and marginal chilling, they seem to have originated in situ by fractionation from the gabbro. Plagiogranitic veins and dykes throughout the gabbros are cut by later basaltic-andesitic dykes. Basic dykes are nearly parallel, ranging from a few mm to 1.5 cm. Some dykes are seen to branch or die out vertically or laterally within the gabbro. They locally display complex anastomising shape and blind-off shoots. BO is also intruded by post-COLG granitoids.

### Chemical summary

Magmas erupted in modern SSZ settings have a distinctive chemistry that help to distinguish them from N-MOR. The geochemistry of SSZ magmas indicates that their mantle source has been depleted in high field-strength elements (HFSE) and enriched in low field- strength elements (LFSE) relative to N-MORB type sources. These differences are related to the main source components of the SSZ petrogenesis and reflect the interaction of different components: (i) partial melting of a depleted N-MORB mantle (DMM) source, (ii) a subducted oceanic slab, and (iii) mantle counter flows (asthenospheric mantle penetrating into the mantle wedge above a subducted slab).

The CAO magmatic rocks are similar to those of SSZ settings rather than to those from N-MOR, as argued by Yalınız et al. *(*1996) (Table 2). Several diagrams can be used to discriminate the type of SSZ setting. Some authors use immobile trace element ratios such as Ti, Zr, Y, V, Th, Ta, and Cr to differentiate the fields for the mafic magma source (depleted to enriched source) and the subduction effect from various environments, such as IAT, MORB, and WPB (Pearce and Cann, 1971; 1973; Winchester and Floyd, 1977; Pearce and Norry, 1979; Wood, 1980; Pearce, 1982; Shervais, 1982; Fig. 2). The subalkaline tholeiitic CAO basalts (Fig. 2a) mostly plot in the IAT field on a number of plots which can be used to discriminate the provenance of the CAO basalts in the previous studies (Fig. 2b-e). Especially, in the Ti/V diagram of Shervais (1982), nearly all samples of CAO basalts clearly plot in the IAT field, having Ti/V values between 10-20, and a number of samples plot lower than 10, where boninites would be expected (upper lavas of CO in Fig. 2b). Boninitic affinities of some samples are also depicted in Fig. 2c (Pearce and Norry, 1979) having very depleted HFSE content relative to IAT. Boninite-like (not true boninite but more depleted relative to IAT) samples of the CAO were mainly determined on the CO studies (e.g. Yalınız et al., 2000a). The CO displays a change in lava chemistry along its stratigraphy represented by a progressive depletion from bottom to top in two sections sampled from the Mezargediği and Çökelik area (Fig. 5 in Yalınız et al., 2000a). In both sections, IAT basalts are overlain by younger basalts which are transitional from IAT to more depleted boninites. The upper basaltic volcanic section of the CO reflects the eruption of progressively more depleted magmas through time. Those rocks are chemically and mineralogically intermediate between boninites and IAT. Several authors have pointed out the presence of such rocks from the Mariana Trench and named them as transitional volcanics, in many ways transitional from boninites to arc tholeiites in their chemical characteristics (e.g. transitional volcanics; D51 samples from the Mariana Trench; Bloomer, 1987; Bloomer and Hawkins, 1987). On the basis of $TiO_2$ content, such rocks were defined for the first time as 'transitional boninites' by Beccaluva and Serri (1988), later, Taylor et al. (1994) classified the rocks having $TiO_2 < 0.6$ wt% as boninites. Finally, Bedard (1999) proposed a term 'intermediate-Ti Boninites' for the rocks chemically between IAT and boninites (0.3 to ~ 0.6 wt% $TiO_2$). Recently, intermediate volcanics having $TiO_2 < 0.5$ wt% were included in the definition of boninite by IUGS.

In terms of their N-MORB normalized multi element profiles, the CAO is characterized by depleted to highly depleted HFSE (Ta and Nb, are the most depleted elements among the HFSE) and enriched LILE (especially Th) relative to N-MORB (Fig. 3d-f). Within the mobile LILE group, Th is used as a representative of LILE because of its relatively immobile behaviour during alteration, compared to the other LILE. Thus, it is a relatively stable and reliable indicator, whose enrichment relative to other HFSE (especially Nb-Ta) is taken as an important indicator of a subduction zone component (e.g., Wood et al., 1979; Pearce et al., 1984). Th enrichment in the CAO magma source is also depicted from the plots of Th/Yb vs Ta/Yb (Pearce, 1982; Fig. 2d). Yb is used as a normalizing factor because of its effectiveness in largely eliminating variations due to partial melting and fractional crystallization. Addition of a subduction component by slab-derived fluids/melts results in an increase in Th/Yb in the mantle source, whereas Ta/Yb ratios are not affected, or little affected. Fig. 2d clearly displays stronger The enrichment in CAO lavas and may indicate enrichment of the mantle source by fluids/melts derived from the subducting slab.

The REE patterns for the CAO basaltic rocks are shown in Fig. 3a-c. Samples show very distinctive chondrite-normalized REE patterns. They are characterized by highly or slightly depleted LREE patterns relative to the HREE and mostly below the level of average N-MORB ($La_N/Yb_N$ ratios for BO: 0.36-0.83; SO: 0.55-0.68 and CO: 0.62-076).

### Evaluation of mantle sources and melting processes

The CAO lavas are peculiar in exhibiting greater depletion in HFSE and REE relative to N-MORB. This implies that they may be generated from a mantle more depleted in HFSE and REE than a N-MORB source. A more suitable model for

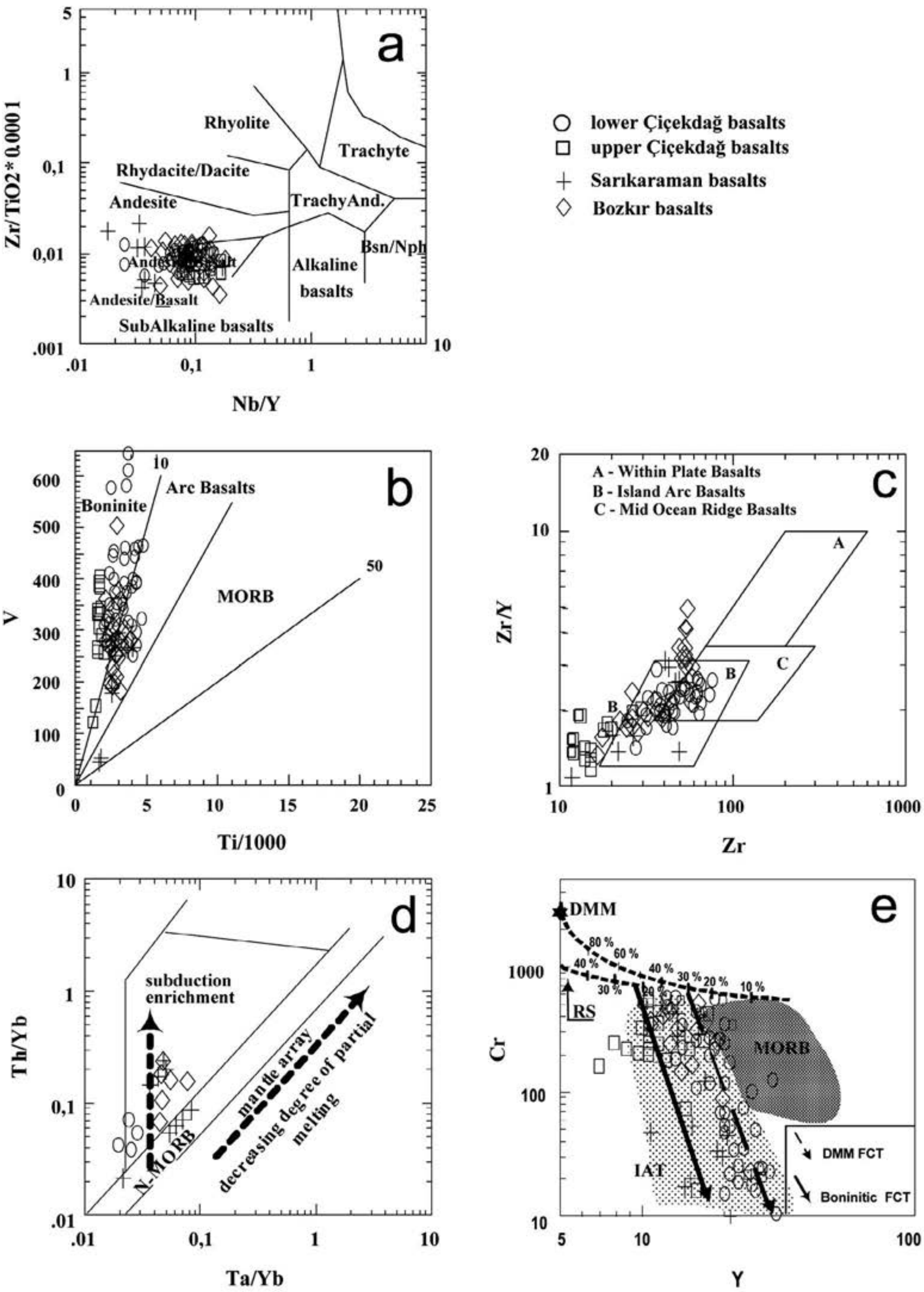


Fig. 2 - Tectonic discrimination diagrams of the basaltic rocks of the Central Anatolian Ophiolite. Data from Yalınız et al. (1996 and 2000): **a-** Zr/TiO2 vs Nb/Y (Winchester and Floyd, 1977); **b-** V vs TiO2 (Shervais, 1982); **c-** Zr/Y vs Zr (Pearce and Norry, 1979); **d-** Th/Yb vs Ta/Yb (Pearce, 1982); **e-** Cr vs Y (Pearce, 1982). DMM- partial melting trend of a depleted N-MORB mantle source; DMM FCT- fractional crystallization trend of the DMM source; RS: partial melting trend of the residue following N-MORB extraction (at 10% melting); Boninitic FCT- fractional crystallization trend of the RS.

the genesis of the CAO lavas involves re-melting of the HFSE and REE depleted source after extraction of MORB, which would lead to low concentrations of these elements in the second-stage melts. The depletion and disparity in HFSE and REE of the CAO lavas must be ascribed to source differences and different degree of partial melting relative to the N-MORB. Pearce (1982) and Pearce et al. (1984) demonstrated that plots of an incompatible against a compatible element provide useful information on the process responsible for the observed geochemical variations in a particular lava suite. For example, because neither Cr nor Y are affected to a significant extent by the processes that cause mantle heterogeneity, the compositions of primary magmas derived from the upper mantle can be modelled and presented as partial melting trends corresponding to different degrees of melting (Pearce, 1982; Pearce et al., 1984). Fig. 2e presents simple petrogenetic modelling of the CAO. Y-poor basalts (IAT) from the CAO display striking feature of Cr and Y values and appear to have been derived from higher degrees of partial of melting (20-40%) of the DMM source. However, it is also apparent in Fig. 2e that some lavas having lower Y (boninite-like lavas) appear to have been produced from more than 40% partial melting of the DMM source. Such degrees of partial melting seem unlikely to produce such boninite-like lavas. This implies that some CAO lavas may be generated from a mantle more depleted in HFSE than a DMM source. A more suitable model for the genesis of these boninite-like lavas involves re-melting the HFSE depleted source after extraction of the DMM, which would lead to low concentrations of these elements in the second-stage melts (sub-oceanic lithosphere; Pearce et al., 1984). Thus, with the assumption that a N-MORB-source was the predecessor of highly depleted CAO mantle (for boninite-like lavas), the residual source compositions generated from 10% partial re-melting of the DMM source are also indicated on Fig. 2e. From this figure, it is clear that approximately 25-30% partial melting of the residual source that experienced previously MORB melt extraction at about 10% seem more likely to produce such boninite-like lavas of the CAO. Summarizing, the melting history, it is thus necessary to propose that the CAO basalts represent two magmatic groups that were derived from partial melting of 20-40% of DMM source, followed by a subsequent 25-30% partially re-melting of an already depleted DMM source (probably oceanic lithosphere) modified by the addition of LILE-bearing aqueous fluids/melts derived from the underlying subduction zone.

## COMPARISON WITH MODERN OCEANIC ANALOGUES

Detailed petrological and geochemical investigations of the CAO reveal broad chemical characteristics of a SSZ setting. However, the results of the Deep Sea Drilling Projects (DSDP) and Ocean Drill Programs (ODP) carried on the IBM intra-oceanic SSZ systems from the Western Pacific indicate that petrogenesis of the oceanic crust formation is highly complex and controlled by many factors. These latter include (i) the mantle source, either depleted oceanic lithosphere or mixed with an enriched source (N-MORB, even OIB source; e.g., Ishizuka et al., 2003; Livermore, 2003; Martinez and Taylor, 2003), (ii) the subduction effect (based on the irregular collision boundaries of the microplates affecting arc-trench proximity, oblique subduction, and depth of Benioff zone below spreading centres) controls the water content in the mantle wedge above the subducted slab leading the degree of partial melting (e.g. Hawkesworth et al., 1977; Harris, 1992). For example, in Lau BAB, there are three active spreading centres: (i) East Lau Spreading Centre (ELSC), (ii) Intermediate Lau Spreading Centre (ILSC), and (iii) Central Lau Spreading Centre (CLSC) (e.g., Pearce et al., 1994). The petrogenesis of the oceanic crust generated in these three spreading centres differs from each other, depending on the proximity of the spreading centre to the arc-trench system (Tonga arc-trench system) situated east of these basins and controlled by the factors listed above. Spreading centres closer to the arc-trench system exhibit a chemistry transitional between island arc and N-MORB (ELCS) while others, far from the arc-trench system display N-MORB chemistry without any effect from the subducting slab (CLSC) (e.g., Pearce et al., 1994).

Concluding, to specify the precise setting of a SSZ type ophiolite, its modern oceanic analogues have to be analyzed and compared in detail. Although a complete discussion on the evolution of these basins is beyond the scope of this paper, the CAO will be compared shortly with its modern oceanic analogues to better understand its precise setting, genesis and evolution, within the Neotethys in the following chapter.

### Back-arc basin (BAB) origin

Recent studies have shown that BABB are erupted in distinct environments. For example, typical BABB from IBM intra-oceanic SSZ systems are being generated in an intra-oceanic BAB (mainly begins with rifting of an island-arc), whereas those from the Ryukyu arc system are being formed in an intra-continental BAB (e.g. Auzende et al., 1995; Gribble et al., 1996, 1998; Shinjo et al., 1999). It is generally accepted that BAB form by interaction between an upwelling asthenospheric MORB-like upper mantle, and a subduction component beneath an ocean ridge in a SSZ environment (e.g., Gill, 1976; Hawkins, 1977; Saunders et al., 1980; Fryer et al., 1981; Hawkins and Melchior, 1985; Jenner et al., 1987; Volpe et al., 1987; 1988; 1990; Woodhead, 1988; Price et al., 1990; Stern et al., 1990; Hawkins et al., 1990; 2000; Stern and Bloomer, 1992; Bloomer et al., 1995; Gamble et al., 1995; Taylor and Martinez, 2003). Controversies on the evolution of BAB mantle in an intra-oceanic BAB system still exist. Current studies on SSZ mechanisms provide data for understanding rifting of an island-arc, followed by spreading which gives rise to opening of a BAB for the IBM BAB. Research from the IBM BAB shows that BABB either exhibit N-MORB compositions or a transition from SSZ to N-MORB signatures and this chemical range in BABB chemistry is related to the progressive evolution (maturity) of the BAB's (e.g., Weaver et al., 1979; Sinton and Fryer, 1987; Woodhead et al., 1993; Pearce et al., 1994). In the early stage of BAB opening, the vast majority of BABBs display transitional features from SSZ to N-MORB (even SSZ) with variable depletion in HFSE and enrichment in LFSE relative to N-MORB (e.g., Gill, 1976; Weaver et al., 1979; Jenner et al., 1987; Sinton and Fryer, 1987; Woodhead, 1988; Woodhead et al., 1993; Wharton et al., 1994). Volcanism in this stage is dominated by tholeiitic magmas formed by adiabatic melting of depleted MORB mantle (DMM), modified by processes associated with dehydration of the subducting slab (e.g., Hawkins et al., 1984; Taylor et al., 1992a). However, as the BAB widens, the influence of the subducting slab on magma composition diminishes and the main magma becomes basaltic, similar to N-MORB which may be geochemically indistinguishable from the basalts that spread at N-MOR (e.g., Hart et al., 1972; Saunders and Tarney, 1978; 1984; Taylor and Karney, 1983; Hawkins et al., 1990; Evans et al., 1991; Pearce et al., 1994; Gribble et al., 1996; 1998; Encarnacion et al., 1999). In summary, BABB may have a range in magma compositions from SSZ to N-MORB, depending on the position of the subduction component and on undepleted asthenosphere diapirism into the mantle wedge.

BABB's exhibiting both dual geochemical nature (SSZ to N-MORB) and N-MORB chemical affinities are very common in other well-known modern oceanic systems, in the IBM and other intra-oceanic SSZs (Fig. 3g). In Fig. 3 the geochemical characteristics of BABB is summarized as follows:

1- Spider diagrams show that BABBs display pattern subparallel to the N-MORB trend but they are displaced to higher LFSE (Fig. 3g). They exhibit a range of compositions from SSZ to N-MORB, with variable depletion in HFSE and enrichment in LFSE relative to N-MORB (e.g., Saunders and Tarney, 1978; Natland and Tarney, 1982; McCulloch and Gamble, 1991; Woodhead et al., 1993; Hamilton, 1994; Taylor and Nesbitt, 1994; Pearce, 2003; Fig. 3g).

2- The REE patterns of the BABB differ from those of N-MORB in having flat to slightly enriched LREE patterns. They are mainly above 10X chondrite level and enriched in LREE ($La_N/Yb_N > 1$; e.g, Saunders and Tarney, 1978). However, in more evolved back-arcs, BABBs have a chondrite normalized pattern similar to N-MORB showing profiles that vary from flat to slight LREE depletion (e.g., Hawkins and Melchior, 1985; Hawkins, 2003).

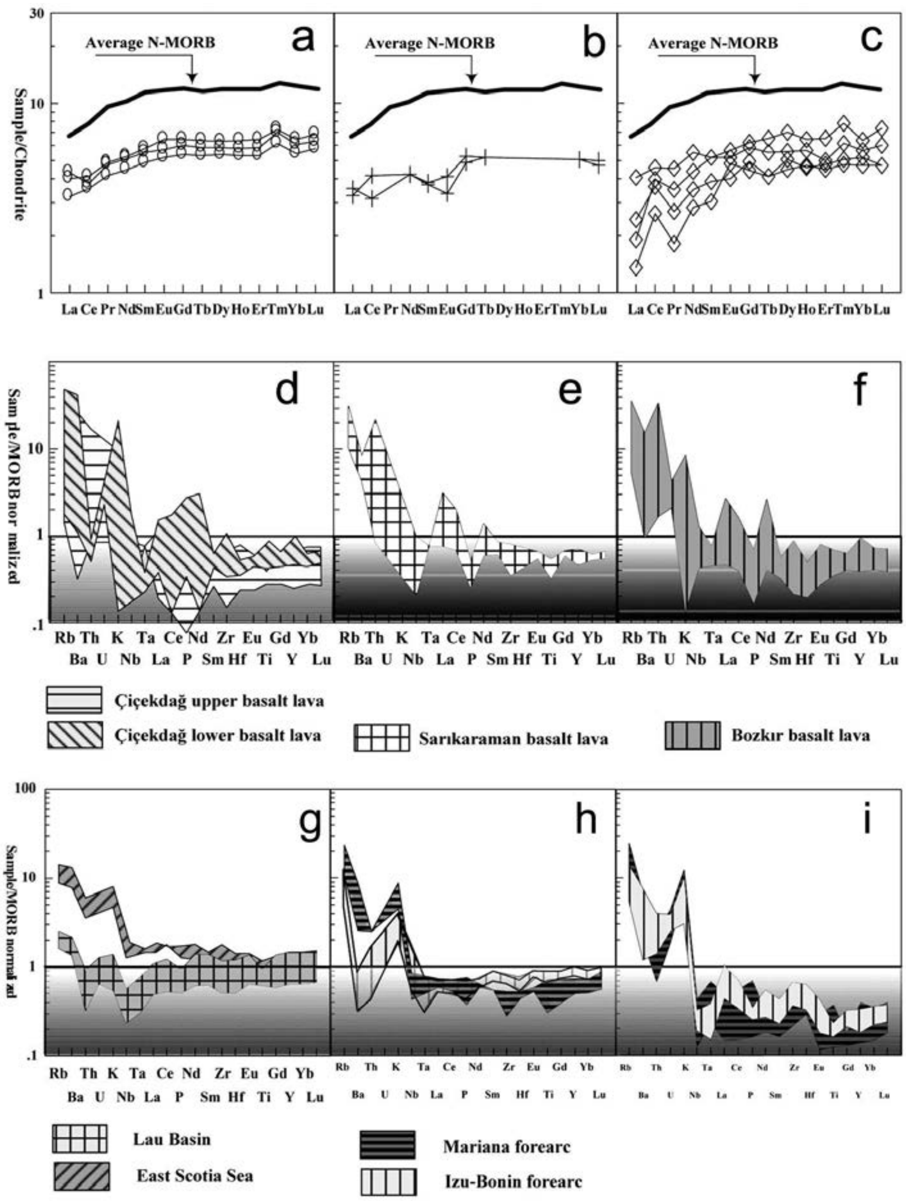


Fig. 3 - Chondrite-normalized REE patterns of: **a)** Çiçekdağ; **b)** Sarıkaraman and **c)** Bozkır basalts (symbols as in Fig. 2). Representative N-MORB normalized multi-element diagrams for: **d)** Çiçekdağ, **e)** Sarıkaraman and **f)** Bozkır basalts. N-MORB normalized multi-element diagrams for some typical: **g)** back-arc basalts, **h)** island arc tholeiites and **i)** boninites. N-MORB normalizing values are from Sun and McDonough (1989). Comparison data selected from the literature: Lau Basin (Pearce et al., 1994); East Scotia Sea (Saunders and Tarney, 1978); Mariana forearc (Hickey and Frey, 1981; Wood et al., 1981; Bloomer, 1987 and Pearce et al., 1992); Izu-Bonin forearc (Murton et al., 1992; Stern and Bloomer, 1992; Taylor and Nesbitt, 1992 and Taylor et al., 1992b).

The chemical characteristics of the CAO samples differ from those of back-arc lavas in that:

1- Their trace element patterns are characterized by enrichment of LILE (especially Th) relative to highly depleted HFSE, lower than those in BABB with distinctive Nb-Ta anomaly (Fig. 3d-f).

2- The CAO mafic rocks exhibit slightly or highly depleted LREE patterns ($La_N/Yb_N < 1$; Fig. 3a-c) and they are different from those of BABB having slightly enriched LREE patterns ($La_N/Yb_N > 1$)

3- The absence of N-MORB basalts and the presence of low-K tholeiitic and boninite-like basalts and related silicic differentiates, which are unlikely to be generated in a BAB.

In summary, depleted and very depleted HFSE geochemical signatures lower than those in BAB, indicating a depleted mantle source and presence of intense subduction component in the magmatic rocks of the CAO are inconsistent with their BAB origin.

## Forearc or Pre-arc origin

In most modern IBM intra-oceanic arc-trench systems from the Western Pacific, forearc is defined as "oceanic crust extending from intra-oceanic arc and/or remnant arc to trench and represents the initial stage of opening of SSZ systems" (e.g., Johnson et al., 1987). Forearc settings are mainly characterized by a distinctive rock assemblage including boninites, IAT and their low-K silicic differentiates in their nascent stages, BAB are unlikely to be generated there. Forearc basins are also characterized by the absence of calc-alkali and N-MOR basalts in their nascent stages (e.g., Hawkins, 2003). The magmatic rocks are characterized by enrichment of LILE coupled with lower HFSE and HREE relative to BABB. The simplest hypothesis to explain the low abundances of HREE and HFSE and high abundances of LILE is that their source region is characterized by re-melting of an already DMM sourcemetasomatized by slab-derived fluids above a subduction zone (e.g., Jenner, 1981; Saunders and Tarney, 1984).

Forearc basins are known to have very thin and extended oceanic crust formed above a nascent dehydrating subducting slab. They mostly represent oceanic crust formed at the initial stage of subduction prior to development of an arc. However, it is important to discuss the definition of "forearc". Oceanic crust, generated in such a basin without an arc edifice may have been closed and obducted soon after formation. Ophiolites generated in such a tectonic setting are generally characterized by: (i) depleted mantle source, (ii) subduction component, (iii) incipient arc-type (boninite-IAT) geochemical and petrological features, (iv) presence of a not well developed sheeted dyke complex, (v) ophiolitic sedimentary cover characterized mainly by deep-sea sediments and by the absence of arc related volcanics; (vi) thin sequences, compared to MORB type oceanic crust and island-arcs; (vii) immediate emplacement during or shortly after their formation (for references, see Shervais, 2001). Thus, a new term is needed that does not include the word "arc", as proposed earlier by Wallin and Metcalf (1998) and Pearce (2003).

IBM SSZ settings testify that forearc basin magmatism is characterized by low-K tholeiites and boninites and related silicic differentiates which are unlikely to form in an evolved active island arc and in BAB systems (Taylor et al., 1992a). Among the IBM SSZ settings, the forearc rocks are the most highly depleted in HFSE and enriched in LFSE.

In Fig. 3h-i, the geochemical characteristics of IAT and boninites, including the forearc basin basalts from the Mariana and Izu-Bonin forearcs are shown. They can be summarized as:

1- Both IAT and boninites show a selective enrichment of LILE and depletion of HFSE (strong Nb-Ta anomalies relative to Th) relative to N-MORB and even to BABB. In addition, boninites are more depleted in HFSE than IAT and are greatly enriched in LFSE relative to HFSE (Fig. 3i). However, IAT are similarly enriched in LFSE, and though not as strongly, depleted in HFSE (Fig. 3h). The more depleted HFSE contents of boninites with respect to IAT indicate the extremely depleted character of the boninites mantle source relative to the IAT mantle source. The enrichment of LFSE (LIL) reflects the addition of a 'subduction' component to their depleted source.

2- The REE patterns of boninites are greatly depleted relative to MORB and even to IAT. The chondrite normalized patterns have either a negative slope (transitional boninites) or a U-shaped pattern (e.g., Hickey and Frey, 1981; 1982; Beccaluva et al., 1986; Crawford et al., 1986; 1989; Arculus et al., 1992; Murton et al., 1992; Pearce et al., 1992; Hawkins, 2003). However, the REE patterns of some IAT display a large degree of overlap with N-MORB ones, although the former can exhibit varying degrees of LREE enrichment and depletion. The wide range of LREE variation observed in IAT samples is mainly assumed to reflect different degrees of partial melting in their mantle source (e.g., Green, 1973; Ewart et al., 1977; Kay, 1977; Hickey and Frey, 1981; 1982; White and Patchett, 1984; Crawford et al., 1986; Beccaluva and Serri, 1988; Brouxel et al., 1989). However, the most LREE depleted volcanics have also been found in forearc settings, suggesting that the first stage of island-arc building was accompanied by high degrees of partial melting of residual oceanic lithosphere (e.g., Hickey and Reagan, 1987; Brouxel et al., 1989). In addition, strong REE depletion of IAT relative to N-MORB are generally interpreted as a product of immature island arc stages (Crawford et al., 1986; Brouxel et al., 1989).

Regarding the CAO, a forearc or pre-arc setting is favoured by:

1- Their trace element patterns characterized by enrichment of LILE relative to highly depleted HFSE (with distinctive Nb-Ta anomaly relative to Th) which are lower than those in BABB (Fig. 3d-f).

2- The REE patterns of the CAO samples are different from those of BABB in having slightly depleted LREE patterns, similar to those of forearcs such as the Mariana (site 458-459) (Hickey and Frey, 1981; 1982; Beccaluva et al., 1986; Crawford et al., 1986; 1989; Pearce et al., 1992; Hawkins, 2003) and Izu Bonin (site 780-786-793; e.g. Arculus et al., 1992; Murton et al., 1992; Pearce et al., 1992) (Fig. 3a-c). However, the REE patterns of CAO do not generally make good discriminants, because similar patterns can be generated in N-MORB and evolved BABB. However, the REE patterns are characterized by a marked decrease in the total REE content relative to N-MORB ones and can best be matched with patterns displaying a negative slope similar to those found in immature island-arc sequences such as the Izu-Bonin Arc one.

3- A review of the literature for the IBM arc-trench system shows that, with few exceptions, all of the published data for the IBM forearc fall into either the IAT or boninite fields and failed to find the rocks with N-MORB characteristics, e.g.: site 458-459 (Hickey and Frey, 1981; 1982; Beccaluva et al., 1986; Crawford et al., 1986; 1989; Pearce et al., 1992; Hawkins, 2003); site 780-786-793 (e.g., Arculus et al., 1992; Murton et al., 1992; Pearce et al., 1992). However, BABB samples from Mariana, Lau and North Fiji back arc basins plot essentially in the same location as MORB from true ocean basins (e.g., Saunders and Tarney, 1978; McCulloch and Gamble, 1991; Woodhead et al., 1993; Hamilton, 1994; Taylor and Nesbitt, 1994;). Detailed geochemical studies of the CAO clearly show that some samples plot in the IAT field and part plot in the field of transitional boninites in a virtually complete overlay with data from forearc basins (Fig. 2b-e).

4- They are also characterized by the absence of calc-alkali basalts and their differentiates (absence of basaltic andesites, andesites, dacites and rhyolites).

5- The presence of boninite-like basalts ($TiO_2 < 0.5$) intercalated with island arc tholeiites and low-K felsic differentiates (plagiogranites) within the CAO allows a detailed analysis to be made of its tectonic evolution.

Their association is commonly assumed to represent the nascent stage of intra-oceanic subduction-related basins, and it is unlikely to occur in BAB (Hawkins, 2003). Boninites reflect hydrous mantle melting in a SSZ environment and form by high degrees of partial melting of a depleted source (e.g., Cameron et al., 1979). Although many models have been suggested for the source of depleted mantle contributing to boninite magmas, most models suggest that boninites are restricted to the forearc (e.g., Green, 1973; Crawford et al., 1981; 1986; 1989; Hickey and Frey, 1981; 1982; Hawkins et al., 1984; Hawkins, 1995; Bedart and Hebert, 1996; Kim and Jakobi, 2002;). Transitional volcanics (intermediate between boninites and IAT) observed in the Mariana forearc region more likely coincide with the boninite-like basalts of CAO (e.g., site 458; D51 samples from the Mariana Trench; Wood et al., 1981; Bloomer, 1987; Bloomer and Hawkins, 1987;). According to Wood et al. (1981), the Site 458 boninite-like volcanics overlie IAT which are more depleted in light REE; these authors suggest that "the arc tholeiites could be generated during the initial stage of hydrous metasomatism of the previously depleted mantle wedge by small degrees of partial melting. More extensive metasomatism would enrich the source in a number of incompatible elements as well as depress its solidus and increase the degree of partial melting so as to produce boninite-like magmas…". This interpretation can also be suggested for the boninite-like magma occurrence of the Çiçekdağ Ophiolite.

## CONCLUSIONS

1- The CAO is representative of a somewhat dismembered oceanic crust and mantle fragments retaining a recognizable and well preserved magmatic pseudostratigraphy. The CAO contains all of the components of a typical ophiolitic sequence: metamorphic tectonites, cumulates, isotropic gabbros, plagiogranites, dolerite sheeted dykes, basaltic lavas and middle Turonian-early Santonian epi-ophiolitic sedimentary cover. They are found as allochthonous bodies in the Central Anatolian Crystalline Complex (CACC), which represents the metamorphosed passive northern edge of the Tauride-Anatolide Platform.

2- The CAO represents cogenetic tholeiites and displays the eruption of progressively more depleted magmas through time. They are also characterized by the absence of arc related calc-alkali basalts and their differentiates (absence of basaltic andesites, andesites, dacites and rhyolites).

3- The CAO represents well exposed examples of SSZ ophiolites formed by short-lived extension and magmatism during the earliest stage of intra-oceanic subduction. In particular, the geochemistry of the igneous rocks shows: an enrichment of LFSE and depletion of HFSE and REE relative to N-MORB and BAB.

4- The IAT of the CAO probably formed through partial melting of a 20-40% of DMM source. Approximately 25-30% partial melting of the residual source that previously experienced MORB melt extraction at about 10% seems more likely to produce such boninite-like lavas of the CAO.

5- The CAO was generated above a short-lived, north-dipping, nascent intra-oceanic subduction zone during early-middle Turonian to early Santonian within the İzmir-Ankara branch of Neotethys rather than in the forearc basin (prior to development of any arc edifice) as it was suggested by previous authors (e.g., Yalınız et al., 1996; 1999; 2000a; Floyd et al., 2000; Fig. 4a, b). Then, it was rapidly emplaced southwards onto the passive margin of the CACC, soon after its formation between post-early Santonian to pre-late Campanian (Fig. 4c). Subsequently, it was intruded by late Campanian - early Maastrichtian post-collisional granitoids (Fig. 4d).

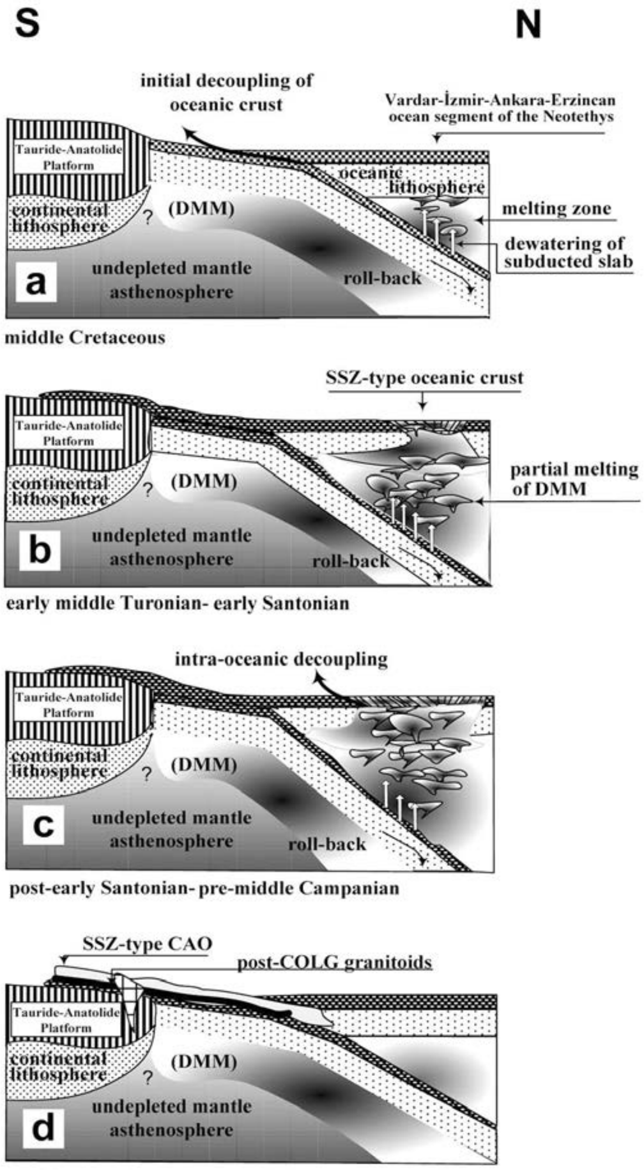


Fig. 4 - Cartoon cross sections showing the evolution of the supra-subduction zone Central Anatolian Ophiolite in the Vardar-İzmir-Ankara-Erzincan ocean segment of the Neotethys from early-middle Turonian to early Maastrichtian.

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