

**COMMENT ON "TECTONIC EVOLUTION OF THE TROODOS OPHIOLITE WITHIN THE TETHYAN FRAMEWORK" BY Y. DILEK, P. THY, E. M. MOORES, AND T. W. RAMSDEN**

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**INTRODUCTION**

Dilek et al. [1990] (hereinafter referred to as Dilek et al.) have presented a wide-ranging model which they claim (p.820) "reconciles conflicting models for the evolution of the Troodos ophiolite by placing the geological and geochemical data within the context of regional geology." They express (p.812) the hope that their model "will stimulate much needed discussions and observations leading to better understanding of ophiolites." We propose to take them at their word!

The essence of Dilek et al.'s model is that the Troodos ophiolite formed in the vicinity of a sinistrally slipping transform fault in a marginal basin, behind a southward dipping subduction zone that was consuming crust from a remnant Palaeotethyan ocean to the north. A graben structure (the Mitsero graben) that lies within the ophiolite they claim is an abandoned spreading axis fossilized by ridge jumping. The Troodos lower and upper pillow lavas are seen as forming in distinct geotectonic events, the former relating to seafloor spreading in the marginal basin in the Late Jurassic to Early Cretaceous and the latter to non seafloor spreading-related residual magmatism in the Late Cretaceous prior to the inception of northward dipping subduction to the south of Cyprus.

In this comment it is our intention to show (1) that examination of the internal structure of Troodos and its relationship to the transform fault leads to the conclusion that the transform was a dextrally slipping structure and that the Mitsero graben was not a fossil spreading axis but instead was an off-axis feature formed by amagmatic stretching; (2) that field, petrological, and geochemical evidence from Cyprus does

not support genesis of different parts of the ophiolite in distinct geotectonic settings; and (3) that, while the geotectonic evolution of Turkey is complex and still in many respects an open question, there is as yet little or no firm evidence for southward subduction in the Jurassic to Cretaceous.

**SPREADING STRUCTURE**

We start by considering the section concerned with the spreading structure of Troodos. Dilek et al. base their model on the work of Varga and Moores [1985], who identified within the Troodos massif three ocean-floor graben structures defined by the opposing tilt of faulted blocks of sheeted dikes. These three structures, from west to east the Solea, Ayios Epiphaniios (subsequently renamed Mitsero) and Larnaca grabens, respectively, they suggested might represent abandoned ridge axial valleys fossilized by the successive eastward migration of the active spreading axis. This novel and interesting hypothesis was tested and critically reviewed by Allerton and Vine [1987], who concluded from palaeomagnetic evidence that the grabens could indeed represent fossil spreading centers, but that one or more of them could equally well have been off-axis grabens that had accommodated extension in periods of reduced magma supply. One of the great attractions of this alternative model is that it explains the graben structures in Cyprus in terms of fundamental processes of spreading at seafloor spreading axes and is very much in accord with recent observations from the modern oceans. Karson [1990], for example, has modeled rifting and crustal extension at spreading axes in terms of periods of undersupply of magma during continued crustal extension; Harper [1985] and Norrell and Harper [1988] have interpreted ocean-floor block rotations and detachment faults in the Josephine ophiolite in much the same way.

Allerton and Vine [1987] argued that for any of the grabens to represent spreading axes abandoned by ridge jumping, rather than simply off-axis grabens, it was necessary at the very least to demonstrate cross cutting relationships at the boundaries between each graben. They looked for, but were unable to find, such cross cutting relationships at the junction between the Solea and Mitsero grabens, although we understand (E. M. Moores, personal communication, 1987) that the western margin of the Solea graben is apparently intrusive into older crust further to the west. Subsequently, Allerton [1988, 1989] has recognized a similar intrusive relationship at the western margin of the Larnaca structure. These observations lend support to Varga and Moores' original hypothesis that both the Solea and Larnaca grabens might potentially be fossilized spreading centers abandoned after ridge jumping (note that more recently MacLeod et al. [1990] have offered independent evidence in support of the Solea graben as a fossilized axial valley). However, no intrusive relationships have yet been demonstrated at the margins of the Mitsero graben, whether by Dilek et al. or any independent authors.

Notwithstanding the above uncertainties, Dilek et al. present the Mitsero graben as a full-fledged ridge axis, without any discussion of Allerton and Vine's alternative model, or attempt to justify their own. They then go on to compare the intersection of the Mitsero graben and southern Troodos transform fault zone (STTFZ, or "Arakapas transform fault" of Dilek et al.) with the Vema ridge-transform intersection (RTI) on the Mid-Atlantic Ridge and conclude (p.815, Figure 3

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caption) from “compatibility of first-order geometric features in both regions...that the Mitsero-Arakapas region in the Troodos ophiolite may represent a slow-spreading ridge and transform fault intersection similar to the [Vema] RTI, which slips at a rate of 24 mm/yr.”

The principal “first-order geometric feature” to which they refer is the curvature of the tip of the Mid-Atlantic Ridge toward the active domain of the Vema transform fault. This phenomenon has frequently been observed adjacent to oceanic transform faults with a variety of slip rates [Fox and Gallo, 1984] and is interpreted by most authors, including Dilek et al., to indicate partial transmission of transform shear stresses into the spreading center, and consequent rotation of the minimum principal horizontal stress (i.e. extension) direction. With this we concur.

In Cyprus the sheeted dike complex north of the STTFZ deviates from the regular N-S strike typical of the rest of the Troodos massif and swings through NE-SW to near E-W, over a distance of some 5-10 km, as the transform is approached [Simonian and Gass, 1978]. The origin of this swing has been controversial: either it is an original feature of the spreading fabric due to dike intrusion in a sigmoidal stress field, as described above [Varga and Moores, 1985; Murton 1986; Dilek et al.], or else it results from the physical block rotation of originally N-S dikes by fault drag, as originally favored by Simonian [1975]. In the former case the STTFZ must have been a sinistrally slipping transform, in the latter case dextral.

It was realized by several research groups that these two hypotheses could be tested by a palaeomagnetic study of the dikes and lavas in the vicinity of the STTFZ: if the primary magnetization vector of the dikes rotates clockwise as their strikes rotate clockwise, the swing was impressed by fault drag; however, if the magnetization vector remains unchanged, then either remagnetization has occurred or else the sigmoidal stress field hypothesis is valid. The issue was addressed in precisely this fashion by Bonhommet et al. [1988], who showed conclusively that the magnetization vectors changed as the dike strike changed and therefore that the fault drag mechanism was valid. This confirmed earlier data from Clube and Robertson [1986], who had measured substantial clockwise block rotations in lava flows and sediments from the Arakapas fault belt itself, and concluded in favor of dextral strike-slip motion. Bonhommet et al.'s study was corroborated independently by Allerton [1988, 1989], who used a more sophisticated geometrical analysis (described by Allerton and Vine [1987]) to calculate rotation axes and magnitudes. He documented clockwise rotations, often in excess of 90°, about steeply plunging axes in dikes in and adjacent to the transform zone; he also documented cross cutting relationships with younger dikes having similar axes but smaller magnitudes of rotation, proving that the rotation was an ocean-floor phenomenon rather than a product of some later emplacement-related deformation [MacLeod 1990b].

Three independent studies have therefore shown that significant clockwise block rotations have occurred in and adjacent to the STTFZ and that these rotations can be interpreted only in terms of drag induced by dextral strike-slip faulting during transform fault activity. This has been further corroborated by a fourth independent study [Morris et al., 1990], which has been published since the submission of Dilek et al.'s paper. We therefore conclude that the curvature of the Mitsero “spreading axis” (sic) into the transform zone shown by Dilek et al. (their Figure 3) is an artifact of the

tectonic rotation of dikes originally intruded at a near N-S spreading axis.

Dilek et al.'s use of a comparison between “first-order geometric features” of ophiolite and ocean-floor structures alone to deduce aspects of their tectonic evolution, at the expense of detailed field observation, we thus believe to be misguided. Note that the field data they do cite [from Ramsden, 1987], of strike-slip slickensides on NE to ENE striking dike-parallel faults near the Arakapas fault belt but dip- or oblique-slip slickensides on N-S faults in the graben away from the transform zone, are in fact consistent with the fault drag hypothesis. In the sigmoidal stress field model it is the extension direction itself that would have been rotated, such that dip-slip and not strike-slip slickensides should be observed on NE and ENE trending faults parallel to the sheeted dikes in the zone of dike swing north of the STTFZ.

Furthermore, the fact that the dike swing and thus block rotation is observed on both sides of the Mitsero graben axis/STTFZ intersection implies that both sides must have experienced the effects of transform deformation. Since ridge-transform intersections, by definition, partition domains of strike-slip (active transform zone) from non-strike-slip (inactive transform zone) deformation, this argues strongly against the Mitsero graben representing a fossil spreading axis. In summary, we suggest that the available data are more consistent (1) with the Mitsero graben being an off-axis graben formed by stretching in a period of low magma supply, rather than an abandoned axial valley; and (2) with slip on the STTFZ being (at least predominantly) right-lateral rather than left-lateral.

#### PETROLOGY/GEOCHEMISTRY

Dilek et al. incorporate into their model petrological and geochemical data from previous studies of sheeted dikes [Baragar et al., 1988] and plutonics [Thy et al., 1989], principally from the International Crustal Research Drilling Group borehole CY-4. Apparently their main purpose in this section of their paper is to show that the lower pillow lavas (LPL) and upper pillow lavas (UPL) and related plutonics formed in separate geotectonic settings (namely, seafloor spreading-related vs non spreading-related), as required by their regional plate tectonic model. On the basis of the conclusions of Thy et al. [1989], they suggest that the CY-4 borehole sampled two distinct plutonic sequences, separated by an intrusive boundary at a depth of about 1330 m. In their view there is petrological and chemical evidence to support a correlation between the upper (older) plutonic suite and the sheeted dike complex/LPL; and another between the lower (younger) suite and the UPL. Thy et al. cited cryptic mineral variation that they observed in the upper plutonic sequence as evidence for a dynamic, steady state, open-system magma chamber that was undergoing frequent replenishment of basaltic magma in a seafloor-spreading regime. However, in the lower plutonic sequence they found evidence for only one such replenishment event and concluded that this must instead indicate crystallization in a relatively closed-system magma chamber, reflecting a slow or non seafloor spreading environment. This is reminiscent of the much earlier model of Gass and Smewing [1973], who believed that the two lava sequences were erupted in different geotectonic settings: the LPL at a ridge axis and the UPL from off-axis foci.

It cannot escape notice, however, [MacLeod, 1990a] that Thy et al.'s conclusions are fundamentally at odds with other

studies of the CY-4 plutonic section published in the same volume [Gibson et al., 1989]. In particular, Browning et al. [1989] studied some parts of the borehole in much greater detail and were able to demonstrate convincingly that cryptic variation does indeed occur in the lowermost part of the plutonic section (between 1740 and 2263 m), that is, within Thy et al.'s lower plutonic sequence. The cyclicity of variation discovered by Browning et al. is on a much finer scale (on average about 3 m) than that recognized in the upper plutonics by Thy et al. and would only have been apparent with much more detailed sampling than that carried out by Thy et al. (who took only 0.14 samples per meter in the lower part of the core) [Browning et al., 1989, MacLeod, 1990a]. The available evidence therefore suggests that open-system fractionation processes were operative at all levels of the cumulate portion of the CY-4 borehole. If open-system fractionation processes are taken as "evidence" for seafloor spreading, then it is not possible to distinguish different geotectonic environments for the LPL and UPL simply on the strength of the evidence presented from the CY-4 plutonic section.

In discussing the geochemical characteristics of the Troodos sheeted dike complex, Dilek et al., using data presented by Baragar et al. [1988], take great pains to emphasize the geochemical consanguinity between the LPL and sheeted dike complex and suggest that these dikes are distinct from the UPL ("The majority of dikes...[show] relatively higher  $\text{TiO}_2$  (Figure 4b) and FeO contents than UPL with similar MgO contents"). They use this conclusion to support their hypothesis (see their Figures 7 and 8) of successive, unrelated phases of magmatism in different geotectonic settings. Although their statement (p.815) that "the majority of the sheeted dikes are geochemically related to the LPL" is strictly correct (as has been stated many times before from field evidence), Dilek et al. fail to point out Baragar et al.'s [1988, 1989] principal conclusion, that the observed spread in sheeted dike compositions spans the entire range of observed lava compositions, both LPL and UPL. Baragar et al. show, on the basis of a variety of criteria including REE profiles, that 23% of the dikes analyzed (95 out of 418) have depleted chemistries comparable to the UPL. Our view of the available data is that the Troodos sheeted dike REE patterns do in fact span the entire range of compositions of the UPL as well as the LPL from northern Troodos. This is particularly evident when viewing Baragar et al.'s [1989] expanded data set. Note that it is inappropriate to compare the Baragar et al. sheeted dike, Rautenschlein et al. [1985] Akaki lava, and Cameron [1985] type I lava REE profiles from northern Troodos on the one hand with Cameron's [1985] type II/type III profiles from southern Troodos on the other, as both of the latter have been modified by an episode of light REE enrichment to give elevated La/Yb ratios [Taylor and Nesbitt, 1988; MacLeod, 1988; Rogers et al., 1989]. This enrichment is spatially restricted, affecting only the STTFZ, Anti-Troodos plate area, and the westernmost part of the Troodos massif [Taylor and Nesbitt 1988, MacLeod 1988].

We also note that Dilek et al. (their Figure 5) appear either to have misplotted the Rautenschlein et al. [1985] and Cameron [1985] lava data or else to have used an unusual chondrite for normalization: using, for example, Nakamura [1974] the maximum LPL heavy REE concentration is approximately 20x chondrite rather than about 28x chondrite as they indicated.

## REGIONAL SETTING

In their Figures 7 and 8 and accompanying discussion Dilek et al. imply a Late Jurassic to Early Cretaceous age for the Troodos upper plutonics/sheeted dike complex/LPL crust, and early to middle Late Cretaceous age for the lower plutonics/UPL. Both magmatic suites, they suggest, inherited their low field strength element enrichment from fluids liberated from an active or recently active southward dipping subduction zone to the north that was related to the consumption of Palaeotethys. They propose that a change in the relative convergence direction between Afro-Arabia and Eurasia (to a more N-S orientation) put the Troodos (LPL) ridge axis under compression and terminated spreading along it, while continued melting formed large magma chambers which underplated the crust and gave rise to the UPL magmatism. Continuing convergence led to the development of one or more north dipping subduction zone(s) to the south of the Troodos and Hatay (Kizildag) ophiolites in Campanian times, with the consequent formation of metamorphic soles, and ophiolite emplacement soon afterward in the Late Cretaceous (Maastrichtian). However, there are a number of problems with this model, both on local and more regional grounds.

1. U-Pb ages from zircons extracted from Troodos plagiogranites give ages of  $91.6 \pm 1.4$  Ma (i.e. mid-Cretaceous: Cenomanian to Turonian on the Harland et al. [1990] time scale) for the formation of the Axis Sequence crust (i.e., upper plutonics/sheeted dikes/LPL) [Mukasa and Ludden 1987] and not Late Jurassic to Early Cretaceous as suggested by Dilek et al.

2. There is no evidence of a Troodos-wide stratigraphical or structural break (marked, for example, by erosion, or deposition of hydrogenous pelagic sediment [Robertson, 1975]) between the LPL and UPL. The ferromanganous umbers overlying UPL of the Limassol Forest complex at the southern margin of the Troodos ophiolite are of Turonian age [Blome and Irwin, 1985]. These umbers are considered to be rapidly deposited hydrothermal precipitates and, together with the massive sulphides and ferruginous ochres, can all be related to the construction of a spreading axis. Accumulation of massive sulphides took place at black smokers within the axial spreading zone, accompanied by precipitation of the oxidized umbers within and above nearby (flank) eruptions represented by the UPL [Boyle and Robertson, 1984]. Spatial and temporal separation of the LPL and UPL is therefore not likely.

3. As accepted by Dilek et al., the non depleted (LPL) and depleted (UPL) geochemical types are locally interbedded on the northern flank of Troodos [Desmet, 1976; Desmons et al., 1980; Rautenschlein et al., 1985] (and, more recently, Taylor [1987,1990]). Taylor's work has shown that volcanism in the grabens of northern Troodos was continuous, closely related chemically and stratigraphically randomly interlayered. Interbedding of lava types has also been demonstrated from the southern flank of Troodos by Rogers et al. [1989], following the Ph.D. theses of Bechon [1982] and MacLeod [1988]. The coexistence of the LPL and UPL is confirmed by Baragar et al.'s [1988] observation that there is no preferred temporal relationship between LPL and UPL magma types in the sheeted dike complex (i.e., depleted dikes cut non-depleted dikes as frequently as the inverse relationship). LPL and UPL magmas must therefore have been available for eruption at the same time.

4. As discussed above, the petrological/geochemical data from the CY-4 plutonics and the sheeted dike complex that Dilek et al. cite as evidence for the LPL magmatism being related to seafloor spreading and the UPL being unrelated to spreading do not stand up to detailed examination. Open-system fractionation processes were occurring at all levels within the Troodos magma chamber(s). Also, the presence of 23% of depleted dikes within the sheeted complex implies significant extension and therefore that some seafloor spreading was continuing during UPL eruption.

5. Livermore and Smith's [1985] relative motion path of Afro-Arabia with respect to Eurasia shows a smooth change in convergence vector between 119 Ma (Aptian) and 70 Ma (early Maastrichtian). There was no "rapid change in relative plate motion" in the period spanning the eruption of Troodos LPL and UPL (circa 91 Ma).

6. Dilek et al. relate the formation of subophiolite metamorphic soles in the eastern Mediterranean area to the onset of compression within a preexisting marginal basin immediately following eruption of the UPL in Cyprus and their counterparts in Hatay (Kizildag). The base of the Troodos massif is, however, not exposed, and although amphibolite facies metamorphic rocks of the Mamonia Complex of SW Cyprus may relate to fault (transform?) juxtaposition of hot Troodos-type crust and mantle against older, cold crust [Clube and Robertson, 1986], no metamorphic sole is known to underlie the Troodos ophiolite. While radiometric ages of the Mamonia amphibolites are Turonian to Santonian (ranging from 90-83 Ma [Spray and Roddick, 1981]), on a wider scale most metamorphic sole ages from the southern Turkish ophiolites are significantly older than the formation age of Troodos (up to 104 Ma [Whitechurch et al., 1984]). These ages suggest that intra-oceanic convergence took place in different areas of Neotethys at different times and show that the change to intra-oceanic convergence cannot thus be simply related to a rapid change in relative plate motion in the Late Cretaceous.

7. Dilek et al. suggest that marginal basins in the eastern Mediterranean opened above a subduction zone dipping southward from Palaeotethys during Middle to Late Jurassic times. In fact, the Late Triassic to mid Cretaceous was a time of passive margin sedimentation and hence stability throughout the southern Turkey and Cyprus areas [e.g. Robertson and Woodcock, 1980]. It is generally accepted that rifting of the Neotethyan basins instead took place in the Early to Late Triassic, from evidence in Cyprus [Lapierre and Rocci, 1977; Robertson and Woodcock, 1979], Antalya [Gutnic et al., 1979; Robertson and Woodcock, 1982] and Baër-Bassit [Delaune-Mayer, 1984]; although opinions differ as to whether wide [Sengor and Yilmaz, 1981] or narrow [Robertson and Dixon, 1984] ocean basins or merely rifts [Dercourt et al., 1986] opened at this time. This contrasts with the Hellenides to the west, where ophiolite genesis and emplacement took place in late Middle Jurassic to Late Jurassic times (e.g. Smith, 1977).

Geochemistry of lavas erupted within these Neotethyan basins, which are supposed to have been sited above Dilek et al.'s subduction zone, do not indicate a subduction influence, for example, Late Triassic lavas in the Mamonia Complex [Pearce, 1975; Malpas et al., 1987] and Antalya [Waldron, 1984; Robertson, 1990]. Late Jurassic to Early Cretaceous extrusives in Baër-Bassit are alkaline to strongly alkaline [Delaune-Mayer, 1984]. The only definitely dated oceanic

extrusives of Lower Cretaceous age in southern Turkey (Egridir area) are of Mid-Ocean Ridge Basalt type [Waldron 1984]. Triassic to Early Cretaceous sediments in Cyprus and southern Turkey are of terrigenous origin and contain no volcanic arc detritus [Robertson, 1981]. There is as yet no geochemical evidence, therefore, to support the existence of any southward-dipping subduction zone in the early Mesozoic in this region.

8. Dilek et al.'s Mesozoic subduction zone dips southward from the northern margin of microcontinental units in central to southern Turkey. They follow Sengor and Yilmaz [1981] and Robertson and Dixon [1984] in postulating a number of microcontinents separated by small ocean basins (unlike those who envisage only a single large microcontinent [Dercourt et al., 1986]). Thus their inferred arc-trench gap spanned a number of microcontinents and small ocean basins, with a width possibly in excess of 600-800 km. These intervening margins and basins remained passive from Late Triassic to Maastrichtian time, with no evidence of regional arc magmatism (e.g., Antalya [Robertson and Woodcock, 1982; Waldron 1984]). In eastern Turkey, southward subduction and marginal basin formation has, in fact, previously been postulated by Perincek and Ozkaya [1981], who explained the Maden Complex in this area as a Late Cretaceous to Early Tertiary marginal basin. However, more recent studies by a number of independent groups have shown that tectonic accretion, arc magmatism and collision in this area can all be related to northward subduction in Late Cretaceous to Early Tertiary times [e.g. Yazgan, 1984; Aktas and Robertson, 1984]. Arc magmatism to which Dilek et al. refer in the Eastern Pontides of Turkey is not considered relevant here as this area lay on the northern margin of the Neotethys in late Mesozoic to early Tertiary times, very far from the area under discussion. In summary, we feel that there is no positive evidence of development of a Jurassic to Cretaceous southward dipping subduction zone in central to southern Turkey.

## SUMMARY

Our discussion of the various points raised by Dilek et al. leads us to the following conclusions:

1. There is no identifiable evidence for the existence of a subduction zone dipping southward from central southern Turkey, at least during Middle Jurassic to Upper Cretaceous times.
2. Field, petrological and geochemical data do not support genesis of the lower and upper pillow lavas and related intrusives in two fundamentally different geotectonic settings, namely seafloor spreading-related versus non seafloor spreading-related environments.
3. The Mitsero graben does not represent a fossilized spreading axis comparable with the Vema ridge-transform intersection on the Mid-Atlantic Ridge but is probably instead an off-axis graben. The curvature of the axis of the Mitsero graben is not a primary feature of the spreading fabric but is a secondary feature created by later block rotation due to slip along the transform. The southern Troodos transform fault zone was, at least for the greater part of its history, a dextrally slipping structure.

To conclude on a more positive note, there are several other features of Dilek et al.'s model with which we concur entirely: for example, the recognition that the Troodos ophiolite was not regionally emplaced southward over Arabia in the Upper

Cretaceous (in common with ophiolites further to the east such as Baër-Bassit and Oman) but remained within a remnant Neotethyan ocean basin west of the Arabian promontory until undergoing collision with a microcontinent in Pliocene to Pleistocene times; and that Palaeotethys was not necessarily sutured by the early Mesozoic but that the southern Turkey

area instead existed as a mosaic of microcontinents and small ocean basins rather than as a single continental area. We are glad to see that Dilek and his coauthors are now essentially in agreement with our models for the driving mechanism for the rotation of the Cyprus microplate [Clube and Robertson, 1986; MacLeod, 1987, 1988, 1990b].

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