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The role of the Kyrenia Range Lineament, Cyprus, in the geological evolution of the eastern Mediterranean area

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Rewarding insights into major crustal lineaments come from the integrated study of well exposed examples. One is the Kyrenia Range, a narrow arcuate lineament of several hundred kilometres in length comprising northern Cyprus and its offshore extension. The Kyrenia Range consists mostly of Mesozoic and Tertiary sedimentary and subordinate volcanic and metamorphic rocks, disposed in four rock groups separated by unconformities recording deformation events. The lineament is dominated by a steeply dipping composite thrust pile located partly along, and partly straddling, the abrupt northward termination of crust similar to the Troodos Igneous Complex at depth.

The 200 Ma history of the lineament involved episodic rift, passive-margin, active-margin, strike-slip and uplift phases. The area was rifted off Gondwana in the late Triassic to form a southerly Turkish microcontinent capped by a gently subsiding carbonate platform. After formation of a small ocean basin to the south during the Cretaceous (Troodos ocean), northward subduction began (?Santonian). The first major deformation (D1) is attributed to pervasive (?dextral) strike-slip, which removed the Mesozoic passive margin and brecciated and metamorphosed the remaining platform. In the Maastrichtian and early Tertiary the area subsided and scree breccias were shed from scarps into pelagic carbonate-depositing seas, while bimodal within-plate-type lavas were erupted in an extensional setting influenced by strike-slip. By mid Eocene time, shortening, first evidence by flysch and olistostrome deposition, culminated in strong southward thrusting (D2) and localized metamorphism. Northward subduction south of Cyprus ensued and the range lay in an extensional fore-arc setting in late Eocene and Miocene time. The area then subsided dramatically and accumulated thick turbidite sequences derived from eroding Tauride Mountain areas to the northeast. Faulting and general uplift in the late Miocene was followed by renewed compressional deformation climaxing in mid Pliocene time (D3) with large-scale thrusting and tilting. Pulsed vertical uplift continued through the Quaternary.

Similar volcanic and metamorphic rocks formed along the Kyrenia Lineament at intervals. Sedimentary rocks emerge as the most sensitive tectonic setting indicators. Long-lived lineaments like the Kyrenia Range are inherently very complicated, and perceived simple solutions in other cases should be viewed with some scepticism.

1. Introduction

The understanding of crustal lineaments requires the study of processes on a global scale, but also integrated views of specific lineaments, which are necessarily complicated. To this end we present a synthesis of the most clearly expressed lineament in the eastern Mediterranean area.

The eastern Mediterranean Sea marks the site of one of a number of small ocean basins which were created and then largely destroyed along the northern margin of Gondwana in Mesozoic

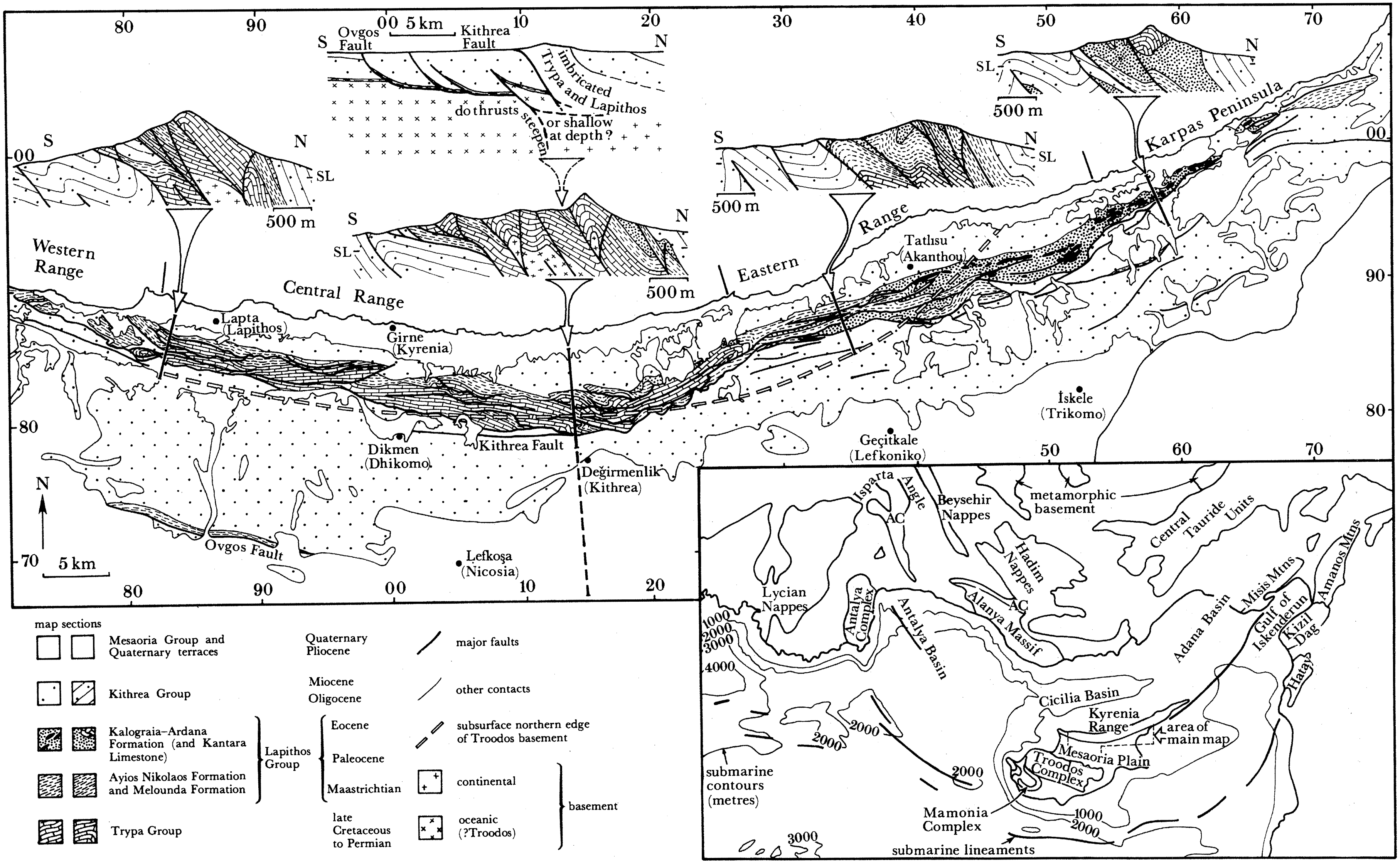

FIGURE 1. Geological map of the main part of the Kyrenia Range with (top centre) true-scale crustal section and four enlarged but true-scale geological sections (modified from Baroz 1979; Knup & Kluyver 1969; Ducloz 1972). Inset: location map of east Mediterranean area.

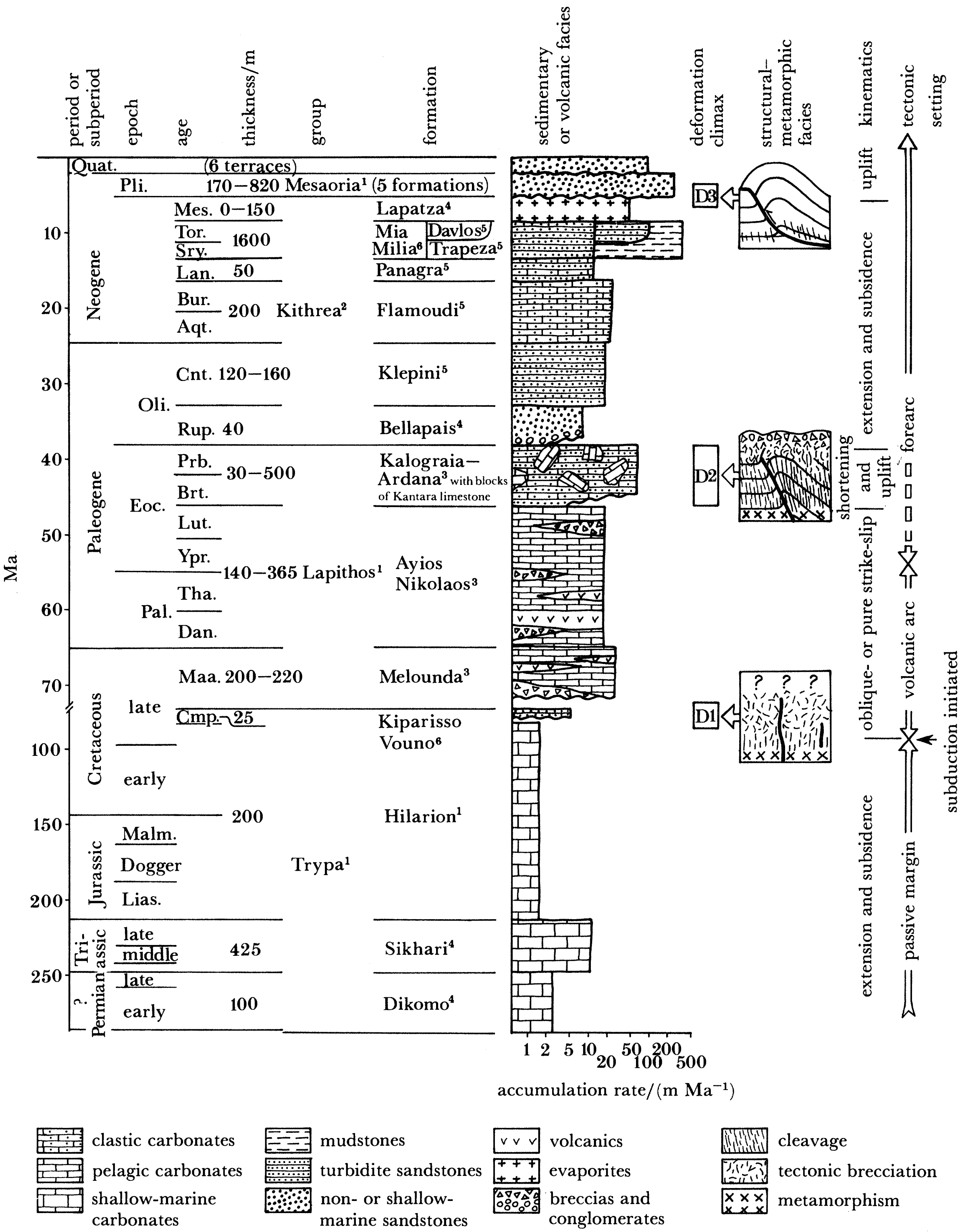

FIGURE 2. Stratigraphic summary of the Kyrenia Range. Vertical scale is true time but with a scale change at 75 Ma. Group and formation names defined by [1]Henson *et al.* 1949; [2]Weiler 1965; [3]Knup & Kluyver 1969; [4]Ducloz 1972; [5]Baroz & Bizon 1974; [6]Baroz 1979.

and Tertiary time (Smith 1971; Dewey *et al.* 1973; Robertson & Woodcock 1980; Sengör & Yilmaz 1981; Robertson & Dixon 1984). Geophysical results indicate that the eastern Mediterranean south of Cyprus comprises oceanic or highly attenuated continental crust (Makris *et al.* 1983), rather than the northward extension of the main African continent (see, for example, Ricou & Marcoux 1980; Ricou *et al.* 1984). The narrow arcuate lineament of the Kyrenia Range of northern Cyprus is critical in its location close to the inferred northern margin of this Mesozoic ocean basin (figure 1), of which the Troodos ophiolite is a preserved part (Gass 1980). In this paper we integrate the existing geological information on the Kyrenia Range with our own reconnaissance studies, and provide new trace-element chemical data on the igneous and metamorphic rocks.

The Kyrenia Range emerges as an extremely complex lineament preserving evidence of a Mesozoic passive-margin phase, followed by a complicated active-margin history involving subduction, strike-slip faulting, several phases of thrusting, metamorphism and vertical uplift.

To place these detailed events in context we first outline the geological setting of the range. A spatial summary is given in figure 1 and a temporal summary in figure 2. The detailed sedimentary and structural events are then described in time sequence. Salient data and their interpretations are given in reference tables (tables 1, 2, 4, 5, 6) and the text develops the palaeogeographic history of the lineament as summarized in figure 11.

2. Geological setting

(*a*) *Outline geological history*

A tectonostratigraphic synthesis of the Kyrenia Range (figure 2) shows four rock groups separated by major unconformities, each representing an important deformation event.

Shallow-water carbonates (Trypa Group) were deposited through most of Mesozoic time, but were strongly brecciated and metamorphosed in the early late Cretaceous (D1, figure 2). Major structures of this age are difficult to distinguish from those of later events. The deformed and deeply eroded Trypa Group was unconformably overlain, first by patchy volcaniclastic sediments (Kiparisso Vouno Formation) and then by an uppermost Cretaceous to early Eocene sequence of pelagic limestones, carbonate breccias and bimodal acid–basic volcanics (Lapithos Group).

A late Eocene south-directed thrust event (D2, figure 2) imbricated and folded the Lapithos Group along with its Trypa basement. Greenschist facies metamorphics were formed at depth and passed into lower-grade pervasively cleaved Lapithos rocks. At higher structural levels these grade into a carapace of tectonic breccias, and finally into breccias redeposited by sedimentary processes. These breccias, mostly carbonates, form one element of a varied late Eocene syntectonic sequence (Kalograia–Ardana Formation) that also includes olistostromes with large limestone olistoliths.

A thick clastic sequence (Kithrea Group) unconformably overlies the deformed Lapithos and Trypa Groups. Deposition was continuous through Oligocene and Miocene time. During latest Miocene and early Pliocene time the range was uplifted and deformed by south-directed thrusting (D3, figure 2). Kithrea Group rocks were folded in the thrust terrains but remained uncleaved. Rapidly deposited Pliocene clastics (Mesaoria Group) and a series of raised Quaternary terraces record continuing uplift of the range. This important lineament can clearly only be understood by taking a completely integrated view of its geological history.

(*b*) *Present geological structure*

The Kyrenia Lineament comprises a 160 km long southward-convex segmented arc up to 10 km wide, rising to a height of 1023 m. Geophysical surveys indicate that to the east the lineament sweeps under the Levant Sea to join the Misis Mountains of southern Turkey (Biju-Duval *et al.* 1974, 1977), while to the west the range extends offshore towards the Antalya Basin (figure 1).

The central spine of the range is composed of Trypa and Lapithos Group rocks in a complex set of thrust sheets, now vertical or dipping steeply northwards (figure 1). In the Central Range (figure 1) Trypa sheets are dominant, but elsewhere Mesozoic carbonates are only present as olistoliths. This spine of pre-Oligocene rocks is bordered by Kithrea Group sediments both to the north and south. The northern contact is usually a northward-dipping unconformity. The southern contact is usually a northward-dipping thrust, though an original unconformity is preserved in places. In the Western Range and the Karpas Peninsula (figure 1) the pre-Oligocene spine is partly covered by Kithrea Group sediments, again cut by Neogene (D3) thrusts. Important D3 faults also crop out south of the range in the Karpas Peninsula, and particularly west of Nicosia (Ovgos Fault, see figure 1).

The deep structure of northern Cyprus has been deduced particularly from magnetic anomalies (figure 1 of Aubert & Baroz 1974), which allow the extent of the Troodos Igneous Complex to be traced. Its upper surface appears to dip northward beneath the Mesaoria Plain more steeply than at outcrop, probably because of a series of north-downthrowing faults (Cleintaur *et al.* 1977). The depth of the Troodos basement is confirmed by deep boreholes south of Nicosia and at Geçitkale (Lefkoniko). The magnetics suggest that the Troodos slab is abruptly truncated at a steep fault. This underlies the southern edge of the Western and Central Range, but must cross obliquely beneath the Eastern Range (figure 1). The present topographic range therefore does not coincide with this important basement structure. The juxtaposition of Troodos basement with that beneath the Western and Central Range must pre-date at least the later part of the Kithrea Group, which blankets the two terranes uniformly. It probably occurred during late Eocene (D2) or late Cretaceous (D1) time, or both. The present topographic range must result from the Neogene events (D3).

3. Permian: epicontinental seas

The only definite Permian rocks identified in the Kyrenia Range are unmetamorphosed fusilinid-bearing neritic limestones forming detached blocks (Kantara Limestone) in the Kalograia–Ardana Formation of the Eastern Range (figure 2; Knup & Kluyver 1969*a*, *b*). Throughout the Tauride Mountains on the Turkish mainland, hundreds of metres of similar neritic limestones accumulated in widespread epicontinental seas which flooded the northern margin of Gondwana. Those in the Alanya Massif (figure 1) are *Mizzia*-bearing low-grade metamorphosed limestones, which are dark and hydrocarbon-rich (Argyriadis 1974; Özgül 1984), whereas those along the north edge of the Massif (Antalya Complex) are typically shelly and plant-rich (Okay & Özgül 1984). There are insufficient data on the Cyprus rocks for useful comparison at present.

formation	member	age	thickness m	lithofacies	interpretation
Kisparisso Vouno		late Campanian	25	graded sandstones with clasts of basic and acid volcanics, shallow-water carbonates, radiolarian chert, quartzite, mica schist inter-bedded mudstones contain smectite > illite > chlorite	turbidites shed over drowned, partly eroded, carbonate platform igneous, metamorphic and deep-sea sediments derived from strike-slip belt to the south
	unconformity now faulted				
Saint Hilarion		late Cretaceous Jurassic	200	highly recrystallized limestones and dolomites, mostly brecciated white marbles, plane and cross-laminated limestones, minor clastics local cycles from bioclastic → algal → burrowed limestone → dolomites	stable carbonate platform with tectonic or eustatic transgressive cycles ending with regressive intra- and supratidal deposition, lagoonal and storm deposits
Sikhari	Upper Middle	late Triassic	250	black laterally variable late diagenetic sugary textured dolostones minor quartz and rare diagenetic feldspar chert nodules no primary textures preserved breccias may pre-date dolomitization megalodonts locally abundant	subsiding shelf, partly lagoonal with transgressive–regressive cycles ending with supratidal environments
	Antifonitis	?middle or Early Triassic	25	pale grey plane-laminated limestone, thicker graded limestones and orange–red shales up to several metres thick. organic-rich	unstable shelf or incipient rift carbonate turbidites shed into restricted anoxic basins, probably fault controlled
	Lower	?early Triassic or Permian	150	fine early diagenetic dolomitic and bioclastic mudstone vadose calcite cements, fenestrae, pelecypods, ostracods and forams locally black or organic-rich	low-energy intertidal to supratidal carbonate carbonate platform
	fault contact				
Dikomo		?Permian	100	finely laminated calcilutite and silty marl mostly recrystallized to grey, blue and white schistose marble interbedded chlorite schist and minor clastics diminish upwards	stable shelf carbonates; shallow water, partly algal
	mostly fault contacts				
metamorphics		unknown	?	quartzites, metacherts, quartz-muscovite albite schists, metatuffs, amphibolites with relict lava and gabbro textures, serpentinites with relict periodotite textures	greenschist facies metamorphics of basic–ultrabasic intrusives, basic volcanics and cherts overburden of *ca.* 6 km at normal geothermal gradient

[a] Sources are specified in the text.

4. Triassic: initiation of subsiding carbonate platform

The lowest units of the main Kyrenia Range stratigraphy to be dated with certainty are the middle and upper members of the Sikhari Formation, which contain late Triassic megalodonts and ostracods (Ducloz 1972). These are stratigraphically underlain by the Antifonitis Member and the lower member of the Sikhari Formation, which is in turn structurally underlain by schistose limestones and marbles of the Dikomo Formation (table 1, figure 2; German Water Mission Report 1965).

The distinctive Antifonitis Member is about 25 m thick and comprises alternations of recrystallized, but still recognizably graded, medium- to thick-bedded limestones together with less recrystallized thinner-bedded calcilutites, and red, purple, and locally black papery calcareous shales with numerous thin stromatolitic horizons. The assemblage is similar to the widespread early Triassic (Skythian) facies exposed on mainland Turkey in both the Antalya Complex and the Alanya Massif (figure 1). Regionally, early Triassic time was marked by faulting and volcanism along the northern Gondwana margin (Marcoux 1974) during the initial phases of continental rifting. This correlation suggests that the Lower Sikhari and the Dikomo carbonates may be of basal Triassic or Permian age. We have noted recrystallized shelly limestones of possible Permian age in the Dikomo Formation (eastern Central Range, 168 805; Eastern Range, 486 916). The formation is sufficiently unrecrystallized to identify alternations of finely laminated calcilutites, silty marls and mudstones. It appears that low-grade schists separating the Dikomo and Sikhari carbones originated as muddy intervals which later became shear zones.

The Dikomo Formation was deposited on a shallow muddy shelf. The generally more dolomitic Sikhari Formation is interpreted as an intertidal to supratidal carbonate platform deposit (table 1; Baroz 1979). The Antifonitis Member probably marks an interval with carbonate clastics shed from horsts and anoxic organic-rich muds accumulating in depressions.

In the mainland Taurides, strong post-rifting subsidence allowed up to 1500 m of neritic carbonates to accumulate in late Triassic time (Poisson 1977). Subsidence of the Kyrenia Range shelf was more modest (250 m) and conditions remained largely tidal (figure 11) behind the platform edge.

5. Jurassic–early Cretaceous: stable carbonate platform

Throughout the Taurides, Jurassic to early Cretaceous palaeogeography saw subsiding Bahama-type carbonate platforms fringed by deep-water passive margin deposits (Robertson & Woodcock 1982, 1984; figure 11). In the Kyrenia Range the Hilarion Formation conformably overlies the Sikhari Formation. Despite moderate to strong recrystallization, hydrozoans and algae of at least partly Jurassic age have been identified (Henson *et al.* 1949; table 1). The limestones and dolomites have poorly preserved sedimentary structures and they probably accumulated on a stable, gently subsiding, platform. They show transgressive–regressive cycles that were either tectonically or eustatically controlled. These produced a range of lagoonal, intertidal to supratidal environments (Baroz 1979).

Although the upper levels of the Hilarion Formation may have been eroded in Campanian time (as we shall discuss), the sequence is much thinner (200 m) than those in the Taurides (up to several kilometres, such as Bey Dağları (Poisson 1977)). No Mesozoic platform cover to the Alanya Massif (figure 1) north of Cyprus has been recognized, and this was probably a raised area in the Mesozoic (Okay & Özgül 1984).

Summarizing, the Mesozoic Kyrenia Range carbonates document a gently subsiding carbonate platform. By early late Cretaceous time, renewed crustal stretching was probably producing new ocean crust to the south (figure 11), of which the Troodos and other eastern Mediterranean ophiolites are preserved remnants.

6. Late Cretaceous: initiation of an active margin

(a) *Regional setting*

During late late Cretaceous (*ca.* Santonian) time the mosaic of small 'Neotethyan' ocean basins and carbonate platform in the Tauride region was shortened and deformed for the first time (Şengör & Yilmaz 1981). In the southernmost oceanic strand (Troodos ocean), extending from the eastern Mediterranean into eastern Anatolia, regional arguments indicate that northward subduction of oceanic crust began at this time (figure 11; Robertson & Woodcock 1980; Şengör & Yilmaz 1981; Michard *et al.* 1984, Model 2; Aktaş & Robertson 1984; Robertson & Dixon 1984). This event is recorded in the Kyrenia Range by pervasive deformation (D1, figure 2), metamorphism, uplift and erosion, and the unconformable deposition of volcaniclastic sediments (Kiparisso Vouno Formation).

(b) *Late Cretaceous deformation* (D1) *and metamorphism*

The D1 event is reliably dated as pre-Maastrichtian by the unconformably overlying basal breccias of the Lapithos Group (figure 2). These are locally derived but contain abundant clasts of metamorphosed and deformed Trypa carbonates and other metasedimentary and meta-igneous rocks (Ducloz 1972; Baroz 1979). Deposition of the basal Lapithos directly over the Triassic Sikhari Formation in the Eastern Range implies major pre-Maastrichtian tectonic displacement and erosion. The Upper Campanian Kiparisso Vouno Formation intervenes between the Trypa and Lapithos Groups, but the nature of its lower contact is in doubt owing to poor exposure (Baroz 1979). Since the formation probably lies unconformably on the Trypa Group, its non-metamorphic and less-deformed state requires a pre-late Campanian age for D1. A less likely possibility is that the Kiparisso Vouno sheet was thrust in pre-Maastrichtian time from an area unaffected by D1 (Baroz 1979).

All the Trypa carbonates show recrystallization, some possibly representing appreciable metamorphic grades. The only reliable grade estimates come from minor non-calcareous metamorphics associated with the Trypa Group (table 1) and suggest temperatures up to 450 °C (persistence of stipnomelane), with a maximum overburden of 6 km at normal geothermal gradients (Baroz 1979). Enigmatic field relations make it difficult to decide whether these rocks are Palaeozoic metamorphics that formed a basement to the Trypa Group, low parts of the Trypa that were metamorphosed during the D1 event, or, as is more likely, mostly exotic to the Range. We have observed chlorite schists interbedded with schistose marbles of the Dikomo Formation, implying D1-age metamorphism to lower greenschist facies of at least some Range lithologies.

Coherent schistose fabrics are only seen in the Dikomo Formation. In higher levels of the Trypa Group this is replaced by an almost isotropic tectonic brecciation (figure 3). This is not restricted to obvious fault zones but is pervasive throughout much of the Trypa Group, leaving

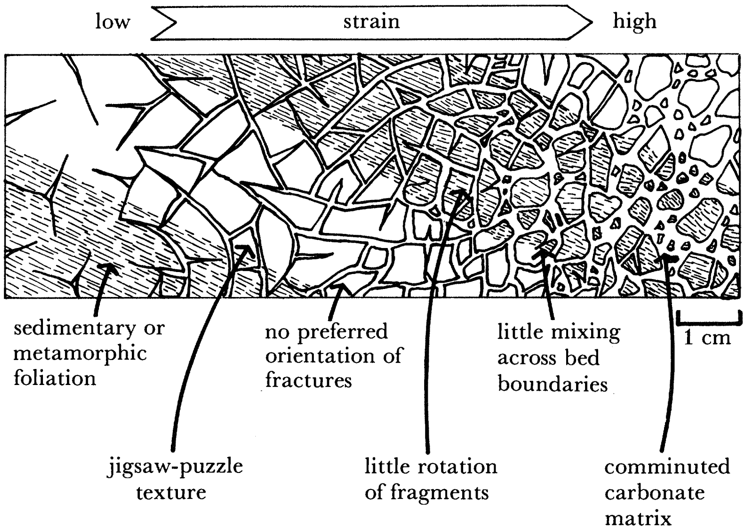

FIGURE 3. Progressive development of carbonate breccias during the D1 (late Cretaceous) event.

only small slivers of unbrecciated rock. It is not restricted to dolomites as is the primary or early diagenetic brecciation recognized by Baroz (1979). A tectonic origin was favoured by Moore (1960), Ducloz (1972) and Dreghorn (1978). This is confirmed by distinctive tectonic geometries (figure 3), such as the jigsaw-puzzle fit of clasts, parallel lamination in adjacent clasts and the low mixing across original lithological boundaries. The precise origin of the breccias is uncertain, but the presence of comminuted grains rather than new vein growth between clasts suggests formation under high effective confining stress rather than by hydraulic fracturing under high pore-fluid pressures. Such breccias have not been recorded from otherwise similarly metamorphosed limestones in the Alanya Massif on mainland Turkey (figure 1).

Later overprinting by Eocene (D2) and Neogene (D3) events makes it impossible to deduce the large-scale geometry and kinematics of the D1 structures.

(*c*) *Campanian clastic deposition*

Baroz (1979) described the few important localities in the Central Range where the Trypa and Lapithos Groups are separated by the thin (up to 25 m) poorly exposed Kiparisso Vouno Formation (Table 1). Our own observations at the type locality (858 855) confirm the presence of unmetamorphosed soft-weathering sandstones and mudstones, interpreted as relatively thin-bedded medium- to fine-grained turbidites. We confirm the extraordinarily heterogeneous petrography described by Baroz (1979). Grains of sedimentary origin comprise radiolarite, shells, planktonic and benthic foraminifera (for example, Orbitolines). Abundant igneous material comprises both basic and acidic undevitrified volcanic glass, mafic lithic extrusive rock fragments, feldspar, biotite, apatite and ferromagnesian crystals. Metamorphic components include quartzite, mica-schist, amphibole, tourmaline, sphene, marble, musciovite and chlorite.

The source area must have consisted of igneous extrusives, variable-grade metamorphics and deep- and shallow-water Mesozoic sediments (radiolarites, *Globotruncana*-bearing micrites, neritic limestones). Derivation from the present Central Range rocks seems unlikely in view of the abundant exotic grains and the distal turbiditic nature. However, metamorphic rocks may well have been available for erosion adjacent to the range by this time, even though this would have involved rapid exhumation. The nearest exposed metamorphic basement is 250 km to the north (figure 1) and none has been identified in the Western Tauride mountains. For example, unmetamorphosed successions range down to the Ordovician in the Antalya Complex (Tahtalı Dağ, Gutnic *et al.*; Robertson & Woodcock 1982) and to Cambrian in the Hadim Nappes (Özgül 1984).

(*d*) *Comparison with SW Cyprus*

The possible significance of the Kiparisso Vouno Formation and the metamorphic slivers is suggested by comparison with SW Cyprus. Here, rocks matching the Kiparisso Vouno grains accumulated in the late Cretaceous before being blanketed by Maastrichtian chalks (Lefkara Formation). A steep arcuate lineament of several kilometres in width separates the Troodos Complex from late Triassic extrusives of the Mamonia Complex (a continental margin assemblage) (figure 1; Robertson & Woodcock 1979) and includes screens of serpentinites and Troodos rocks with discontinuous pods of amphibolite, greenstone, marble and pelitic and psammitic schists (Swarbrick 1980). The Troodos sedimentary cover includes the 750 m thick Campanian–Lower-Maastrichtian volcaniclastic Kannaviou Formation (Robertson 1977*a*). This is strikingly similar to the Kiparisso Vouno Formation in containing grains of fresh basic and acid vocanics, metamorphics, and both deep- and shallow-water sedimentary rocks. The origin of the Kannaviou volcanism has remained obscure because no comparable rocks have been found on the adjacent Turkish mainland. The suggestion (Robertson 1977*a*) that the volcaniclastic material was derived directly from the late Cretaceous volcanics in the Kyrenia Range (Moore 1960) is now discounted after the demonstration (Baroz 1980) that these are slightly younger (Maastrichtian). However, all these volcanics could be related to a similar tectonic setting, one involving strike-slip faulting over an extended period.

The late Cretaceous metamorphism in SW Cyprus and the juxtaposition of contrasting late Triassic and late Cretaceous basement blocks has been attributed to strike-slip faulting in which still-hot Troodos oceanic crust was slid against an early Mesozoic rift and passive margin terrane (Robertson & Woodcock 1980; Spray & Roddick 1980; Swarbrick 1980). Strike-slip faulting along a possibly related lineament in the SW segment of the Antalya Complex (SW Turkey) has been reported both from late Cretaceous ophiolitic rocks (Reuber 1984) and from adjacent continental margin units (Woodcock & Robertson 1982). In the present context, large sub-vertical strike-slip faults provide high-heat-flow zones in which metamorphic rocks may form, while associated large dip-slip components can quickly exhume these rocks without the need to erode a huge overlying ophiolitic nappe as in sub-ophiolite sole rocks (Woodcock & Robertson 1977; Spray *et al.* 1984). Erosion of just such a transform lineament, associated with localized mafic to acid volcanism could have produced all the exotic Kiparisso Vouno Formation grains. The geophysical evidence for a major basement fault zone that is now beneath, but was originally south of, the Central Range (figure 1) suggests a possible source zone (figure 11).

(*e*) *Relation to Cyprus rotation*

Well known palaeomagnetic declination evidence indicates a 90° anticlockwise rotation of the Troodos ophiolite since its formation in late Cretaceous time. Preliminary palaeomagnetic work on the *in situ* Troodos sedimentary cover suggested Miocene rotation (Shelton & Gass

1980). More detailed study now demonstrates that most of the rotation had already taken place by early Eocene time (Clube *et al.* 1985). Was the Kyrenia Range also rotated? If so, it might have originated to the NW within the Isparta angle of SW Turkey (figure 1). Palaeomagnetic data from the mainland Taurides (Lauer 1981, 1984) do not yet provide a consistent answer, but the dissimilarity of the Kyrenia Range carbonate rocks to those of the Isparta angle area (Robertson, unpublished data) tends to oppose major rotation of the Kyrenia Range. In this case the Kyrenia Range formed the southern extension of the Alanya Massif on the mainland. The northern boundary of the rotating microplate must then have lain south of the Western and Central Ranges (figure 11), probably coincident with the northern edge of the Troodos basement slab overthrust by the range rocks in Neogene time (figure 1).

In summary, circumstantial arguments suggest the following model (figure 11). Regional northward subduction of oceanic crust began before Campanian time. By the Campanian an associated, probably right-lateral, strike-slip zone was initiated south of the Kyrenia Range carbonate platform. This began the tectonic removal of the pre-existing platform passive margin, and its replacement by Troodos ocean crust rotated in from the east. During early, presumably transpressional, phases (early tangential tectonics of Baroz 1980) the platform edge was sheared and compressed, leading to the pervasive brecciation of the carbonates and genesis of low-grade psammitic and pelitic rocks along active fault strands. The protoliths of some of these rocks were the originally argillaceous intervals in the Mesozoic succession, especially within the Sikhari and Dikomo Formations. These rocks were tectonically slivered upwards in the fault zones, as were amphibolite facies, metasedimentary, and meta-igneous rocks formed from former margin and ophiolitic rocks located along the fault zone south of the range. Bimodal basic–acid volcanism occurred along these lineaments. Local subsidence allowed the Kiparisso Vouno turbidites to flood over the new deformed, metamorphosed and deeply eroded platform.

7. Maastrichtian to early Eocene: submarine volcanic lineament

(*a*) *Maastrichtian sedimentation*

The Trypa Group is unconformably overlain, or in places faulted against laterally discontinuous basal breccias of the Melounda Formation (Lapithos Group, see table 2). There is an *in situ* pelagic matrix that contains *Globotruncanids* and other diagnostic Maastrichtian planktonic foraminifera (Ducloz 1972; Baroz 1979) and sponge spicules. The breccias are oligomict, usually clast-supported, with varied, mostly angular, clasts (table 2). Most were shed locally from active submarine fault scarps, cutting already deformed and metamorphosed Trypa Group rocks. The highly brecciated carbonates readily disaggregated to form local talus fans along fault scarps in a deep marine pelagic carbonate-depositing régime.

The basal breccias pass laterally and vertically into thin-bedded pelagic chalks composed of scattered planktonic foraminifera and calcite-replaced radiolaria in a relatively pure micrite background (table 2). Higher in the sequence (for example, 344 872, 261 837) are calciturbidites containing a redeposited microfauna with terrigenous and volcanic debris. The calciturbidites could have been shed from unstable highs within the range outcrop area. Although their psammitic and pelitic metamorphic clasts can be matched with local outcrops (Ducloz 1972), the radiolarite debris is apparently exotic to the Western and Central Range. This was possibly derived from the still active strike-slip lineament to the south, as postulated for Campanian time.

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Table 2. Geological summary of the lower Lapithos Group (*a*) sedimentary and (*b*) igneous rocks[a]

(*a*) sedimentary

formation	unit	age	thickness/m	lithofacies	interpretation
Ayios Nikolaos	redeposited limestones	Luretian, Ypresian, late Palaeocene	40–150	redeposited limestones, mostly turbidites, laterally continuous locally rests on Trypa Group has clasts of pelagic chalk, neritic limestone, dolomite, chert, radiolarite, lava, with reworked *Globotruncana*, benthic forams, calcareous algae, bryozoa, orbitolines, nummulites and pelecypods	emergence of carbonate platform to north, shedding neritic material mixed with clastics from active fault scarps
	limestone breccias	early Eocene,	0–15	laterally discontinuous limestone-dominated breccias, locally overlie Hilarion Formation oligomict, clasts up to 15 cm, similar to basal Melounda breccias	renewed faulting gave re-exposed sea-floor fault scarps shedding 'basement' clasts locally
	pelagic limestones and extrusives	late Palaeocene	100–200	pelagic carbonates (similar to Melounda Formation) intercalated with the breccias and redeposited limestones and extrusives	deep marine pelagic chalk deposition, interrupted by clastic and fissured volcanic episodes
Melounda	calciturbidites and extrusives	Maastrichtian	200	thin-bedded graded calciturbidites in pelagic carbonate abundant planktonic forms, also sponge spicules, radiolaria, quartz, muscovite, biotite, feldspar and volcanic grains mafic extrusive flows	tectonically unstable deep sea floor with redeposition of pelagic carbonates into fault controlled basins fissure eruptions
	pelagic limestones and extrusives	Maastrichtian	100	finely laminated pelagic chalks, often pink and Fe-rich abundant replacement chart little terrigenous debris intercalated mafic lavas	pelagic blanket over more stable segments of sea-floor with subdued topography, fissure eruptions
	limestone breccia	Maastrichtian	0–20	oligomict breccias, angular clasts, mostly Trypa Group marbles, also sandstone, radiolarian chert, lava, serpentinite, quartzites, chlorite schists disconformably overlies Trypa Group or Kisparisso Vouno Formation	mass-wasting off submarine fault scarps cutting metamorphosed Trypa Group

(*b*) igneous rocks

formation	unit	field relations	petrography	chemistry	interpretation
Ayios Nikolaos	volcanics (late Eocene, Palaeocene)	up to several tens of metres thick, mostly mafic flows and sills interbedded with pelagic chalks and calciturbidites mostly located further north than Upper Cretaceous volcanics	basalt + dolerite (olivine > augite, feldspar An_{40-60}), trachyte (An_{35}, K-feldspar, hornblende, aegirine augite rare), lamprophyre (An_{40} > hornblende augite > biotite)	major elements indicate undersaturated, bimodal, extensively altered high but variable Al Fe enrichment trend trace elements on m.o.r.b. normalized plots indicate strongly enriched within-plate origin	major element shoshonitic trends (Baroz 1980) may indicate intracontinental strike-slip régime trace elements suggest within-plate origin
		mafic and acid flows up to tens of metres thick, intercalated with pelagic chalk acidic extrusives, mostly tuffs, become more abundant upwards	basalt, dolerite, trachyandesite, dacite and rhyolite absence of andesite augite > olivine, plag. An_{45-65}, Ti-augite, biotite, bronzite and barkevitic amphibole	major elements indicate oversaturated, bimodal volcanics high but variable Al Fe, Mg, Mn, Ca and Ti inverse to Si trace elements on normalized plots show m.o.r.b. to strongly enriched within-plate	major elements suggest calc-alkaline (?)volcanism related to northward (?)subduction (Baroz 1980) trace elements indicate within-plate origin, compatible with strike-slip influence

[a] Sources are specified in the text.

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(*b*) *Maastrichtian volcanism*

Interbedded with the Maastrichtian chalks are numerous flows and sills (table 2*b*; Bear 1959; Moore 1960; Baroz 1975, 1980). The chalks are bright pink in places, reflecting hydrothermal iron discharge during volcanism. Baroz (1980) stressed the bimodal nature of the volcanic suite and took the major element chemistry to indicate a calc-alkaline composition. He related this to northward subduction before southward obduction of the Troodos and Kyrenia Range rocks onto the African continent, in line with the views of Ricou *et al.* (1984). New analyses of major and trace elements of relatively undeformed mafic extrusives (Melounda Formation) are given in table 3. Sufficiently unfractionated samples were plotted on the various immobile trace-element discrimination diagrams of Pearce (1980) (figure 4*a*, *b*). On the Cr–Y diagram (figure 4*b*), which distinguishes well between within-plate, m.o.r.b. and island-arc tholeiites, all the samples plot in the within-plate field. Representative samples plotted on m.o.r.b.-normalized 'spider' diagrams (figure 5*a*) show trends lying between m.o.r.b. and strongly enriched within-plate basalts. Note show the relative enrichment in large-ion-lithophile elements to high-field-strength elements characteristic of hydrous melting above a subduction zone (see, for example, Saunders *et al.* 1980). The trace elements thus provide no support for Baroz's view (Baroz 1980) that the volcanics belong to a calc-alkaline suite related to subduction.

(*c*) *A Maastrichtian strike-slip lineament*

The igneous and sedimentary data indicate that the Maastrichtian was a time of strong east–west faulting, basement erosion, and quiet fissural eruption of mostly mafic flows in a deeply submerged area undergoing pelagic calcareous deposition. Because the rotation of southern Cyprus continued into Tertiary time (Clube *et al.* 1985) this faulting and volcanism can be attributed to continuing strike-slip displacement, now dominantly transtensional, located south of the Kyrenia Range Lineament. The localized acidic volcanism of the Western Range could conceivably be related to local transpression, causing greater fractionation and crustal assimilation during magma ascent.

(*d*) *Palaeocene volcanism*

After a short break, volcanic rocks reappear throughout the Kyrenia Range, reaching their greatest exposed thickness and lateral extend in the Eastern Range and the Karpas Peninsula (Bear 1959).

In the Karpas Peninsula (688 028) we have observed flows up to 60 m thick of mafic pillow lavas with subordinate lava breccias and calcareous interstitial sediment. These are overlain by weakly calcareous radiolarian mudstones in beds up to 0.35 m thick and then by up to 20 m of pink and grey mudstones and claystones below the regional basal Kithrea unconformity. Thin sections of the red siltstones above the lavas reveal radiolarians in terrigenous silt. Elsewhere in the range Baroz (1979, 1980) identified basalt, dolerite, trachyte, and also lamprophyre flows and sills (table 2*b*). From the major element chemistry he diagnosed the lavas as an undersaturated bimodal suite with a shoshonitic trend. In his view the Kyrenia Range had already been emplaced over the African continental margin and therefore the volcanics were 'post-tectonic', possibly generated in an intracontinental strike-slip régime. New analyses of major and trace elements (table 3) show undifferentiated samples plotting in the

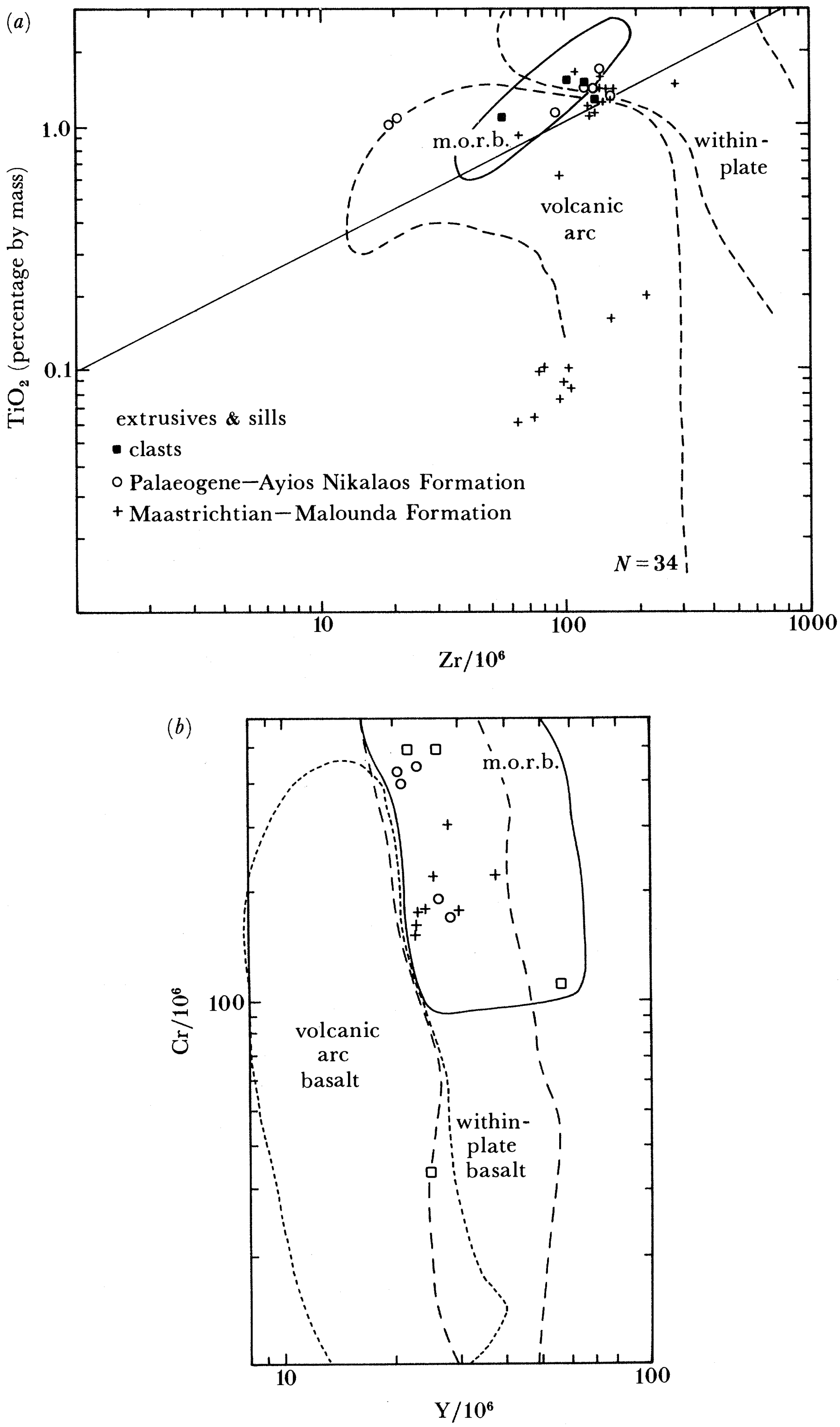

FIGURE 4. Discriminant diagrams of Kyrenia Range igneous rocks (*a*) TiO_2 against Zr. Samples plotting below the horizontal line are significantly fractionated (mostly rhyolites) and should not be used for tectonic discrimination. Of the remainder, most plot near the boundaries of m.o.r.b., within-plate and volcanic arc fields. There are no obvious systematic differences in the tectonic settings of the Maastrichtian (Malounda Formation) and the Palaeogene (Ayios Nicalaos Formation) igneous rocks. (*b*) Cr against Y plot of the basalts only. Volcanic arc basalts possess a lower content of immobile elements (such as Y) relative to m.o.r.b. and within-plate basalts. Cr is an index of fractionation (Pearce 1980). The Kyrenia basalts fall in the m.o.r.b. and within-plate fields. None should be classified as volcanic arc basalts (c.f. Baroz 1980). See table 1 for selected full analyses.

Table 3. Selected analyses of Kyrenia Range mafic igneous and schistose metamorphic rocks

	Maastrichtian			Palaeogene		clasts		schist clasts		mafic schist	
	1	2	3	4	5	6	7	8	9	10	11
SiO_2	46.8	45.2	85.0	46.8	50.2	51.4	46.7	44.4	52.9	56.0	51.8
Al_2O_3	18.1	18.1	8.1	17.8	16.7	15.2	19.3	15.0	14.3	8.0	12.2
Fe_2O_3	11.2	11.2	0.9	11.2	8.3	10.7	10.0	12.3	12.0	8.6	9.3
MgO	6.4	1.9	0.06	3.1	8.4	7.6	6.2	3.2	6.6	16.4	14.1
CaO	10.3	14.6	0.5	13.4	10.7	9.6	9.9	17.7	4.7	8.0	9.5
Na_2O	4.3	4.6	2.6	4.5	2.0	3.8	2.1	2.9	5.5	1.5	2.1
K_2O	0.8	2.0	2.9	1.6	2.1	0.4	3.3	2.1	1.1	0.7	0.6
TiO_2	1.4	1.6	1.06	1.5	1.1	1.1	1.5	1.7	1.5	0.2	0.2
MnO	0.2	0.1	0.01	0.1	0.1	0.1	0.1	0.2	0.2	0.2	0.02
P_2O_5	0.2	0.5	0.01	0.5	0.2	0.1	0.4	0.2	0.2	0.02	0.01
total	99.8	100.1	100.3	100.7	99.9	99.9	99.5	100.00	99.1	99.8	100.1
Ni	151	141	3	175	167	89	223	117	142	275	295
Zn	149	82	7	84	72	23	87	114	9	61	77
Pb	3	5	10	4	5	6	9	6	8	4	6
Rb	18	26	86	28	28	9	30	48	32	16	12
Sr	924	668	82	587	406	161	319	509	44	26	146
Y	30	25	14	24	23	21	28	27	30	8	11
Zr	140	118	63	123	91	58	133	116	110	9	10
Nb	5	18	11	19	13	3	22	21	4	2	1
Cr	180	212	4	215	434	110	499	266	466	705	1021
Ce	21	49	52	52	29	4	28	27	2	—	—
Nd	13	23	21	24	14	5	14	12	5	0	—
Sc	47	35	—	33	30	52	50	41	50	40	50
V	25	229	—	249	285	345	309	336	363	271	275
Cu	79	48	—	43	49	26	36	64	16	21	64
Ba	251	512	704	532	169	51	237	719	150	79	50

Major elements of ignited samples are in percentage by mass oxides; minor elements in p.p.m. (by mass). The analysis for major and trace elements used a Philips PW2450 X-ray fluorescence spectrophotometer. The major element analysis used fused discs made with the Johnson–Matthey 'Spectroflux 165'. Calibrations used U.S.G.A. and C.R.P.G. rock standards with corrections for secondary absorption. The trace elements were analysed by using pressed-powder discs. Absorption corrections were made with coefficients calculated from the major element composition. Calibration was achieved by using a wide range of international rock standards, with correction for interelement interference effects from use of spiked synthetic glass standards. Total Fe is expressed as Fe_2O_3. Sample details: (1) sheared mafic lava from flow interbedded with chalk, Panagra gorge (777 872); (2) from 8 m thick basaltic flow in chalk, Antifonitis area (274 847); (3) rhyolite, Panagra gorge (775 871); (4) pillowed lava interbedded with chalk, Antifonitis area (280 854); (5) thick pillow lava near Platinisso (668 028); (6) basalt clast in basal Kithrea flysch, Flamoudi area, Eastern Range (493916); (7) basalt clast from debris flow in basal Kithrea flysch, Lapithos area (866864); (9) epidiorite in basal Kithrea flysch, near Bellapais (777870); (10) mafic schist screen between Trypa Group carbonates and Lapithos chalk, Agios Antonios (086 804); (11) mafic schist, Fortomenos valley (170 806).

within-plate field (figure 4). The m.o.r.b.-normalized trends (figure 5*b*) are indistinguishable from the Maastrichtian Melounda Formation mafic rocks. The trace element chemistry also confirms that the thick flows in the Karpas Peninsula match those in the Lapithos Group rather than, for example, Troodos ophiolitic lavas.

(*e*) *Palaeocene to Eocene tectonic setting*

The Maastrichtian tectonic setting probably continued into early Eocene time. Mafic lavas continued to be erupted in a transtensional régime. The highly radiolarian but only weakly calcareous siltstones overlying Lapithos Group lavas in the Karpas Peninsula may have accumulated close to or below the carbonate compensation depth. Carbonate breccias continued to be shed from active sea-floor fault scarps.

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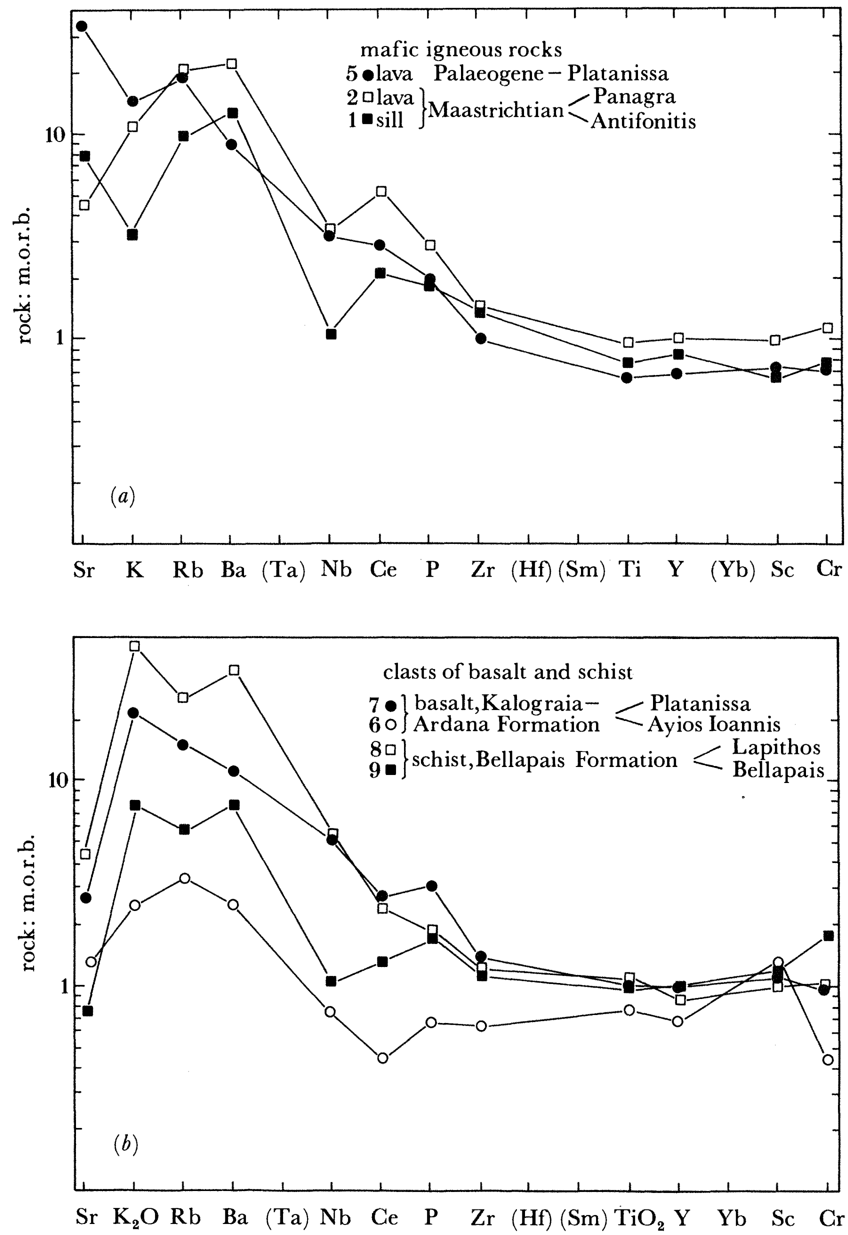

FIGURE 5. Geochemical patterns of representative Kyrenia Range mafic igneous rocks and schists. Selected elements are normalized against the m.o.r.b. values of Pearce (1980). Average m.o.r.b. plots as a horizontal line. (*a*) Fresh volcanic arc basalts are relatively enriched in the large-ion lithophile elements (l.i.l.e.), for example, Sr, K, Rb and Ba. Relative to m.o.r.b. and volcanic arc basalts, within-plate basalts contain much higher values of most of the elements plotted (with the exception of Y, Sc and Cr; Pearce 1980). Although somewhat scattered owing to alteration, the basalts are all of essentially within-plate basalt type. Relative enrichment in Nb, Ce and P particularly distinguish these rocks from volcanic arc basalts. There are no systematic differences in the patterns of the Maastrichtian (Malounda Formation) and the Palaeogene (Ayios Nikalaos Formation) basalts. (*b*) M.o.r.b. normalized plots of basaltic and schist clasts. Basalt clasts in the Eocene Kalograia–Ardana Formation, Eastern Range, are similar to the Maastrichtian and Palaeogene lavas, as are several samples of schist from the basal conglomerate of the Kithrea flysch (Bellapais Formation).

By late Palaeocene to early Eocene time, increasingly numerous turbidity currents carried mostly intrabasinal debris, probably derived from a tectonically unstable substratum within the Kyrenia Range. Exotic material becomes more abundant in the overlying Kalograia–Ardana Formation. Regionally, during early to mid Eocene time, the Alanya Massif north of Cyprus was finally tectonically emplaced northwards over both the Antalya Complex and the relatively *in situ* Tauride carbonate platform to the north (Akseki Platform (Okay & Özgül 1984)). From this time onwards large volumes of metamorphic and volcano–sedimentary material were shed southwards from strongly uplifting areas. Increasing southward migration of the compression front may have ended mafic volcanism and caused a shift from strike-slip to thrust kinematics within the Kyrenia Range.

8. Mid to late Eocene: uplifting thrust belt

(*a*) *Mid to late Eocene deformation* (D2)

Throughout the western Taurides north of Cyprus, the late Eocene was a time of pervasive, generally southward, thrusting, affecting both the carbonate platforms and the already tectonically emplaced Mesozoic continental margin and ophiolitic rocks (Blumenthal 1963). Regionally, the deformation has been related to the final closure of Neotethyan oceanic strands within the inner Taurides ('Inner Tauride Ocean' (Görür *et al.* 1984)), creating a series of concentric arcuate thrust fronts of which the Kyrenia Range is the most southerly.

In the Kyrenia Range the thrust event is broadly synchronous with the deposition of the heterogeneous Kalograia–Ardana Formation (Dixey 1972). At one end of a spectrum of field relations this unit overlies the Ayios Nicolaos Formation through a conformable sedimentary transition and bears the same strong cleavage (for example, 290 867, and apparently throughout the Eastern Range). At the other end of the spectrum the Kalograia–Ardana Formation is itself uncleaved, but contains clasts of already cleaved lower Lapithos lithologies (for example, the Central Range, 153 791). It is seen to unconformably overlie Trypa rocks as low as the Sikhari and probably the Dikomo Formation, locally mantling both late Cretaceous (D1) structures and imbricates of Trypa Group and Lapithos Group rocks (D2). Midway in the spectrum is a common relation (figure 6*a*) in which cleaved Melounda or Ayios Nicolaos formation sediments grade structurally upwards into tectonic breccias, becoming more disrupted upwards, passing then into redeposited breccias of the Kalograia–Ardana Formation (for example, the Eastern Range and northern flank of the Central Range). The scale of the phenomenon (over tens or hundreds of metres) and the tectonic nature of the lower brecciation suggest that the lower Lapithos rocks were still actively deforming as the Kalograia–Ardana Formation accumulated above.

An upper time limit to the D2 deformation is given by the early Oligocene age of the basal Kithrea Group, which overlies Lapithos rocks with a strong structural or sedimentary break.

The most obvious major D2 structures are the steep thrust intercalations of Trypa and Lapithos rocks in the Central Range (figure 1). The steep dip of the overlying Kithrea Group along the northern flank of the range shows that the thrust faults developed as lower angle D2 structures and were only tilted northward during the later D3 event. This northward steepening has affected D2 structures throughout much of the range. The D2 thrusts are marked by strongly brecciated limestones and scaly clays with a thrust-parallel fabric and minor folds indicating dip-slip displacements.

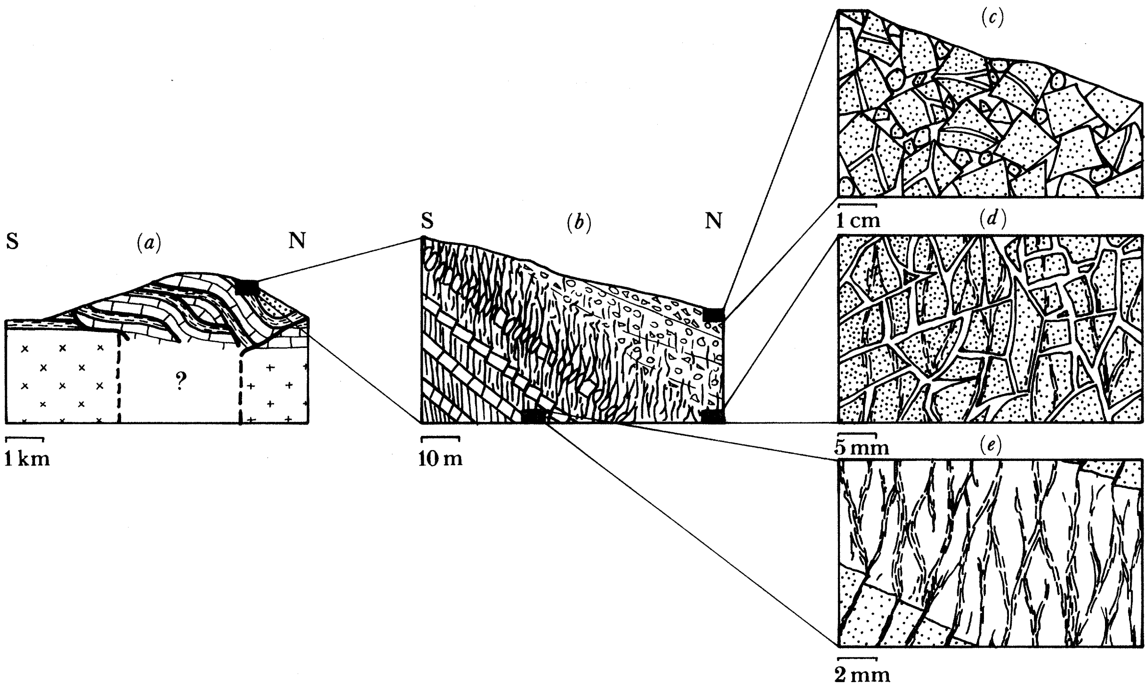

FIGURE 6. Structural style and textural variation during the D2 (mid to late Eocene) event. (*a*) gross structure (ornament as figure 1); (*b*) transition from pressure-solution cleaved Lapithos carbonates (*e*) through tectonically brecciated and veined carbonate (*d*) to Kalograia–Ardana Formation redeposited carbonates (*c*).

In places in the Central Range, particularly near the southern flank (such as 108 793), thin slices of metamorphic rock of up to greenschist facies occur along faults. These lithologies – marbles, schists and meta-lavas – might be regarded as rethrust products of the late Cretaceous (D1) metamorphism. However in places (such as 040 824), marbles and schistose metalavas pass laterally into lower-grade Lapithos chalks and lavas. Several samples of schistose greenstones from the Central Range (131 790) have been analysed (table 3) and, when plotted on the immobile trace element diagrams (figure 4) and normalized against m.o.r.b. (figure 5), are indistinguishable from the Lapithos Group mafic lavas. Some of the Kyrenia Range metamorphics were thus apparently derived from Tertiary Lapithos Group rocks. Others appear to have Cr-rich sedimentary protoliths, possibly the late Cretaceous Kiparisso Vouno Formation (table 3, nos. 10, 11). They thus appear to represent a localized metamorphism distinct from the late Cretaceous (D1) metamorphism.

The typical unmetamorphosed Lapithos rocks develop a strong cleavage that is approximately axial planar to sporadic mesoscopic folds. In the chalks, the cleavage comprises dark pressure-solution seams about 0.1 mm thick and spaced 0.1–5 mm apart (figure 6*b*). A more widely spaced pressure-solution cleavage cuts breccia and calciturbidite beds and, more rarely, lava units. The stronger cleavage in the chalks and the often strong cleavage refraction both suggest selectively higher strain in the chalk units as a result of a component of bed-parallel simple shear (figure 6*a*). Before late Miocene (D3) tilting, the D2 structures mostly faced upwards to the south and verged southwards (figure 6*a*), compatible with south-directed overthrusting (for example, eastern Central range and Eastern Range). However, there are smaller areas of north-facing and north-verging folds and thrusts (for example, the north flank of Central Range (figure 1)). In areas where minor folds are common (such as 261 837) they are markedly non-cylindrical. Baroz (1979) showed that the minor folds throughout the range tend to parallel the major folds and faults. His subdivision of folds on the basis of small variations in orientation seems, however, of doubtful significance given their variable initial geometry and later modification by Neogene (D3) deformation.

The important upward transition from cleaved Lapithos rocks to tectonically brecciated and then redeposited lithologies is displayed in figure 6*b*, *c*, *d*. The tectonic breccias (figure 6*c*) are diagnosed by the presence of carbonate vein-filling within a jigsaw-puzzle texture. Cleaved rocks passed up into this brecciation profile during still-active emplacement of host thrust sheets. It is impossible to estimate the amount of overburden removal, but cleavage was probably forming at depths of the order of the maximum preserved thickness of Ayios Nicolaos Formation, about 250 m.

Other preserved products of superficial deformation during late Eocene time are chaotic slump-sheets and homogenized sediments containing boudinaged metre-scale rafts. These mélange sheets form the highest structural units of the pre-Miocene sequence in the Western Range.

Although the Trypa Group rocks must have been strongly deformed during the D2 event, the specific effects are difficult to isolate in the already strongly brecciated carbonates. The typical D1 brecciation involved comminution of rock debris without much veining, and therefore the abundant vein-filled hydraulic fracture breccias are probably of D2 age.

(*b*) *Syntectonic sedimentation*: *Eastern Range*

Where exposed (such as 287 867), the Kalograia–Ardana Formation of early Middle Eocene age is transitional upwards from the calciturbidites of the Ayios Nicolaos Formation (Kluyver 1967; Baroz 1979). Here the basal unit (the Mavri Skala flysch of Baroz, see table 4 and figure 7 of Baroz (1979)) comprises alternation of pink chalky limestone with replacement chert, siltstones, pink mudstones and turbiditic sandstones in beds up to 1 m thick. Trypa-derived carbonate clasts decrease in abundance eastwards. The sandstones include a great diversity of sedimentary and igneous grains including plagioclase, picotite, pyroxene, serpentinite, lava, quartz, muscovite, amphibole, calcite, glauconite, radiolarite, chert, micrite, pelmicrite, oosparite, large foraminifera (nummulites), coeval and reworked planktonic foraminifera and *Lithothamnium*. Only about 10% of the debris could have been derived from the exposed Kyrenia range rocks (Baroz 1968, 1979).

Our observations show that along the southern flank of the Eastern Range (for example, 390 873) the Mavri Skala flysch is thinner bedded and finer grained than outcrops further north, consistent with a more distal origin. Chalky turbidites and fissile chalky mudstones pass into medium-bedded turbiditic sandstones and subordinate fine matrix-supported conglomerates, including much Trypa debris. South of the main range front (561 964) Neogene (D3) faulting has exhumed Lapithos Group rocks which pass up into a 30 m succession of thin-bedded turbiditic sandstones intercalated with grey marls. Here, and in the main range to the north, palaeocurrents are from the north (Weiler 1965; our observations).

Along the northern flank of the Eastern Range the turbiditic sequence coarsens upwards into 500 m of debris flows and olistostrome rudites (Armenian Monastery rudites (table 4, figure 7 of Baroz (1979)) with blocks of limestone up to several kilometres in diameter (Kantara Limestones). The smaller limestones blocks are interstratified with thick-bedded turbiditic

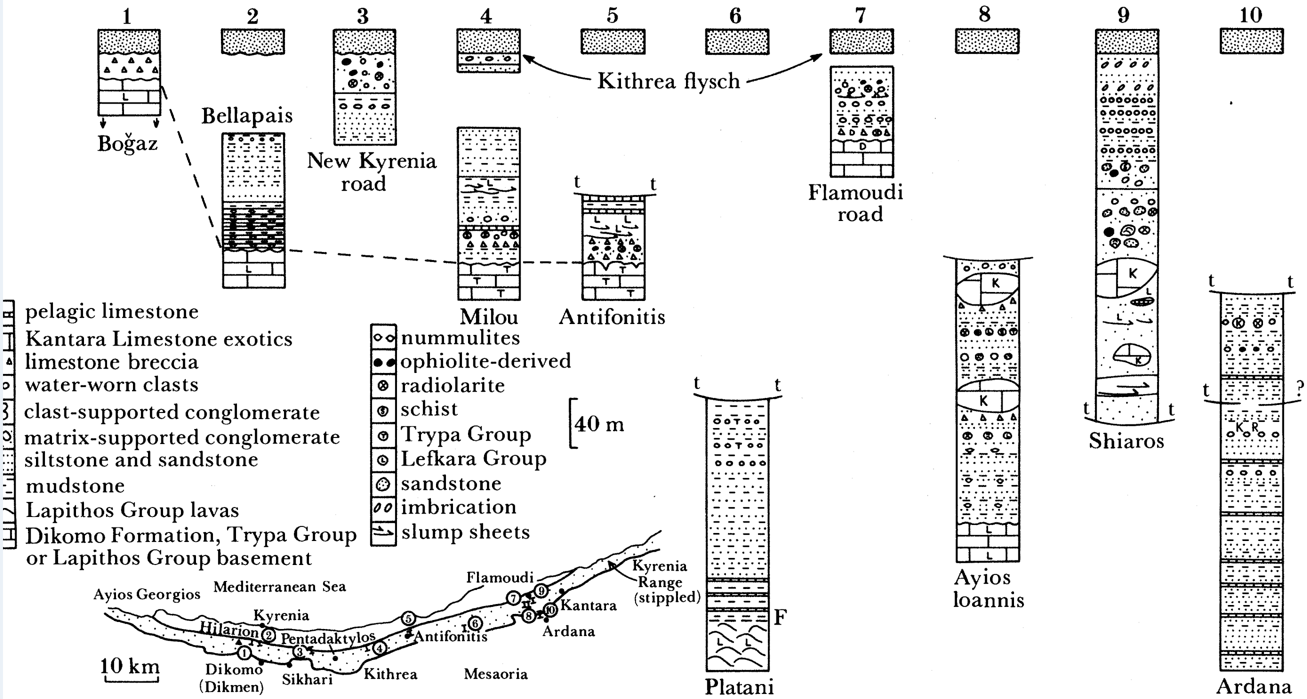

FIGURE 7. Lithological logs of the Kalograia–Ardana Formation.

sandstones, debris flows and massive clast-supported conglomerates. The clasts are moderately to well rounded and consist mostly of Kantara Limestones (especially close to the detached blocks), Trypa carbonates, mafic lava, chert, and all the exotic lithologies of the underlying turbidites (table 4). The larger limestone blocks occur in the highest levels of the succession, along the north flank of the Eastern Range. They are mantled by cemented screes, giving the misleading impression of laterally continuous lenses as mapped by Baroz (1979). The limestones are extremely heterogeneous, including calcilutites, thin-bedded calciturbidites and white crystalline limestones, with micritic limestones, oosparites, pelmicrites, algal limestone and shelly and stylolitic calcilutites, locally pink. Baroz (1979) determined ages ranging from Middle and late Permian through Jurassic, with most being Cretaceous. The derived breccias contain a matrix with planktonic foraminifera dating from the early–Middle Eocene boundary.

The largest Kantara Limestone blocks are set in distinctive debris flows containing rounded clasts of mafic lava, pelagic chalk, turbiditic quartzose sandstone, often plant-bearing and red radiolarian chert (for example, from 494 914 to 536 934). The debris flows contain occasional rafts of weathered mafic pillow lavas and serpentinite, the latter partly monomict sedimentary conglomerates. The lava clasts are chemically indistinguishable from the Lapithos igneous rocks (figures 4, 5).

The Kalograia–Ardana Formation in the Eastern Range can be interpreted as an upward-coarsening and -thickening wedge derived from the north in early Middle to late Eocene time. Distal turbidites first prograded over a deep sea floor south of the present Eastern Range front. Further north conglomerates, debris flows and slide blocks accumulated to about 1 km thickness, eventually building up to sea level. The clast-supported conglomerates were rapidly shed from prograding alluvial fan deltas to the north. The Kantara Limestone slide blocks were

Table 4. Geological summary of the Kalograia–Ardana Formation[a]

age	Western range: lithofacies	Western range: interpretation	Central range: lithofacies	Central range: interpretation	Eastern range: lithofacies	Eastern range: interpretation
	15–20 m undeformed crudely stratified limestone breccia clasts of Trypa and Lapithos Groups, schist and large forams	post-deformation ?sub-aerial screes unconformably over Lapithos rocks	matrix-supported rudites with serpentinite, gabbro, radiolarite and lava rounded clasts of Trypa and Lapithos Group	debris-flow deposition of ophiolitic and mature beach or alluvial derived clastics	clast-supported rudites with clasts of all the underlying exotic + Trypa Group rocks transitional to Kithrea Group	shallowing-up to beach with high-energy reworking rapid subsidence at beginning of Kithrea deposition
late Eocene	huge disorganized slide sheets of Lapithos carbonates, with some rounded clasts of Trypa and Lapithos matrix-supported rudites of Trypa and Lapithos locally 100 % monomict lava rudites	mass sliding and debris flow of still relatively soft Lapithos chalks and lavas related to southward thrusting probably equivalent to olistostrome mélange in Eastern Range			olistostrome melange with clasts of chert sandstone, calcilutite, mafic lava, serpentinite and large olistoliths of Kantara Limestone	proximal debris flows and slide blocks of Mesozoic platform margin rocks flow to south
			'Armenian monstery rudites': 200 m of matrix-supported rudites and rafts of mostly Lapithos lavas and limestones Trypa and exotic clastics	debris flow and slumping of locally derived underlying sequences, plus fine-grained exotic input (e.g. red chert)	'Armenian monastery rudites' up to 500 m of oligomict conglomerates and turbidites, with large blocks of Kantara limestone	rapidly deposited debris flows, from platform, margin ophiolitic and metamorphic areas, little Trypa
middle Eocene			'Milou and Bellapais breccias' up to 40 m of Trypa-derived breccias with benthic and planktonic forams	redeposition of platform debris off fault-bounded highs into pelagic basins	'Mavri Skala flysch' up to 400 m of thin to thick-bedded turbidites with 'exotic' neritic, deep-sea sediments ophiolitic debris	coarsening-up turbidite sequence derived from north

[a] Sources are specified in the text.

derived by disintegration of part of a Mesozoic carbonate platform or associated smaller build-ups. The mélange assemblage of sandstones, radiolarites and pelagic limestones was probably derived from one of the Mesozoic successions which fringed the passive margins of all the major Tauride carbonate platforms (such as the Antalya Complex (Robertson & Woodcock 1982)). Serpentinites could have come from an already deformed segment of such a margin. At the same time the local Eastern Range basement was faulted and provided rafts of Lapithos igneous and sedimentary rocks to the mélange.

(c) *Syn-tectonic sedimentation*: *Central Range*

In the Central Range the lowest units (the Milou and Bellapais breccias of Baroz; table 4; Baroz (1979)) comprise up to 40 m of Trypa-derived breccias with a mixed benthonic and planktonic foraminiferal fauna. As noted by Ducloz (1972), algal and foraminiferal limestones are not now exposed in the range. Overlapping these breccias and the lower parts of the Lapithos Group are debris flows and olistostromes up to 200 m thick (the Armenian Monastery rudites of Baroz (1979), table 4). In places (such as 236 838), fissured Sikhari Formation dolomites are unconformably overlain by debris flows that are dominated by reworked Lapithos and Trypa Group rocks, including metamorphics. The Trypa rocks are mostly reworked in crudely stratified carbonate breccias with sutured clasts infilled by pink pelagic carbonate. The derived Lapithos rocks form detached rafts up to tens of metres long and debris flows of still plastic sediments (such as 187 810). Exotic lithologies become increasingly common upwards (for example, 156 793), including gabbros, serpentinite, diabase and radiolarian chert.

Interpretation of the Kalograia–Ardana Formation in the Central Range suggests that in early Middle Eocene time, faulting of the Trypa Group basement sharply increased the supply of submarine scree breccias into the basin. Benthonic foraminiferal chalks accumulated on local highs only to be completely eroded and redeposited, while incompetent Lapithos sedimentary rocks and igneous units collapsed into debris flows and slump sheets. By this time debris flows were also being shed from a deeply dissected ophiolitic suite, while parts of the Central Range were already near to sea level, supplying well rounded Trypa clasts. In the absence of any palaeocurrent data the source of the ophiolitic debris is uncertain. A southerly source from undeformed Troodos basement seems unlikely because the exposed succession south of the range (near Ovgoros) still preserves Lapithos rocks and distal Kalograia–Ardana facies. A northerly derivation could imply an oceanic crustal strand between the Trypa Group platform and the Alanya Massif (figure 11), or even derivation from ophiolitic units located still further north (such as the Hadim Nappes). At present, derivation from the reactivated fault lineament south of the range front appears the most plausible option.

(d) *Late tectonic sedimentation*

The latest facies mapped as the Kalograia–Ardana Formation accumulated immediately after the climax of the D2 deformation but occur below the regionally continuous Kithrea Group unconformity. Along the northern flank of the Eastern Range (for example, 487 918) olistostromes are overlain by up to 30 m of almost undeformed clast-supported conglomerates with well rounded clasts up to 1 m in diameter derived from the north. Clasts lithologies are similar to those in the lower parts of the formation. The conglomerates pass *transitionally* up across the mapped Kithrea Group basal contact into turbidites.

Along the northern flank of the Central Range deformed Lapithos Group rocks are overlain by undeformed, crudely stratified clast-supported breccias (such as 972 829). Clasts are of Trypa limestones, Lapithos chalks and lavas, chert nodules, schist and notably nummulitic calcarenite, a facies not known in outcrop in the range. Further west (for example, 874 858) similar breccias below the Kithrea Group basal conglomerates include numerous mica-schist blocks up to 1 m long. In the Western Range (774 870) similar metamorphic clast-rich breccias up to 10 m thick pass laterally over 10 m into 1 m of coarse sandstone and fine conglomerate containing well rounded clasts of mostly Trypa Group rocks. Immobile trace elements (such as Cr, Y, Ti, Zr) suggest Lapithos Group protoliths for the mafic metamorphics (figures 4, 5 and table 3).

Along the Central Range southern flank (149 790) the Kithrea Group is underlain by turbidites with pelagic chalk partings and matrix-supported conglomerates up to several metres thick (table 4). Clasts include red and black chert, mafic lava, nummulitic limestone, serpentinite and pale orthoquartzite up to 0.25 m in diameter, an assemblage contrasting strongly with the overlying Kithrea Group.

On the southern flank of the Eastern Range (503 902) the typical Kalograia–Ardana rudites pass conformably up into turbidites with several debris-flow intercalations, and then into Kithrea Group turbidites.

The various sub-Kithrea facies document the palaeotopography immediately after the climax of late Eocene to early Oligocene deformation (D2). The northern flank of the Eastern Range was emergent and accumulated probable beach deposits around ephemeral islands. Deeper water persisted further south, with Kalograia–Ardana turbidites passing into Kithrea turbidites without a break. The Central Range was emergent by end-D2 time and shed Trypa material from steep scarps. However, its southern flank remained submerged until later. Nummulitic limestone probably accumulated during range uplift, but was then eroded. The source of orthoquartzite debris is unknown. Along both Western Range flanks, subaerial scree breccias accumulated and along the south flank pass into alluvial sandstones. Rejuvenation of the old fault lineament south of the range is the only obvious source of the coarse angular metamorphic debris.

(*e*) *Late Eocene tectonic setting*

By late Eocene time a series of diachronous continental collisions had closed any remaining oceanic strands within the Turkish mainland, welding the Tauride and Pontide belts with Eurasia (Şengör & Yilmaz 1981; Robertson & Dixon 1984). Yet a gap of several hundred kilometres still existing between Africa and Eurasia (Livermore & Smith 1984*a*, *b*) must have been represented by oceanic crust in the east Mediterranean near Cyprus. The Kyrenia Range D2 deformation is thus to be interpreted in terms of continuing active-margin evolution rather than by continental collision. Further east the Arabian promontary did not impinge on Eurasia until Miocene time (the Lice-Çüngüs flysch of Eastern Anatolia; Aktaş & Robertson 1984).

Two main models can be considered. In the first the Kyrenia Range deformation can be seen as the southernmost of the Tauride compressional arcs, thrust over already emplaced ophiolitic basement to the south. In the second, the ophiolitic basement was actively subducted northwards, in which case the Kalograia–Ardana Formation could be compared with a trench-accretionary complex. There is however no evidence of sequential slicing or age contrasts, and the region to the north was being strongly uplifted and eroded at this time in contrast to the subsidence of typical arc-trench gaps. Around the eastern Troodos margin the Upper Paleocene–Lower Eocene *in situ* sedimentary cover includes northerly derived calciturbidites containing large benthonic foraminifera like those of the range at this time (Middle Lefkara Formation (Robertson 1976)). An open ocean setting for the Troodos ophiolite far from a continental margin is unlikely at this time. An accretionary origin for the Kithrea flysch is thus not favoured.

The scenario we prefer (figure 11) is that by mid Eocene time the Troodos ocean crust had already been juxtaposed with the Kyrenia Range by strike-slip faulting along the suture now beneath the Central Range (figure 1). Southward compression was then taken up either by transpression along the suture (Western and Central Range) or by thrusting of the range over the ophiolitic crust to the south (Eastern Range). A basin was formed by loading ahead of the thrust front in the Eastern Range, progressively filled with Kalograia–Ardana clastic sediments and olistostromes, and was itself then involved in the thrusting. Slivers of metamorphic rocks already formed along deep thrusts or in the strike-slip were exhumed. The strongly differing Kalograia–Ardana facies and structural styles of the Central and Eastern Ranges probably reflect the sharply contrasting basement structure revealed by gravity and magnetic data.

9. Oligocene–Miocene: an extensional basin

(*a*) *Regional evidence*

As noted earlier, palaeomagnetic data indicate that in the Oligocene to early Miocene hundreds of kilometres still separated the African and Eurasian continental margins (Livermore & Smith 1984). Subsequent northward subduction has possibly taken place along the modern arcuate fault zone running south of Cyprus (figure 1; McKenzie 1972; Jackson & McKenzie 1984). In this view the Troodos Complex and the Kyrenia Range have both been broadly in a fore-arc setting since that time.

(*b*) *Transgressive flood plain deposits*

Throughout most of the range the base of the Kithrea Group comprises up to 40 m of clast-supported conglomerates (figure 8, Bellapais Formation, table 3), crudely stratified, channelled, or disposed in fining-upwards cycles of several metres in thickness. The clasts, which are mostly well rounded, comprise all the rocks of the adjacent Kyrenia Range (for example, metamorphics with Lapithos Group-type volcanic protoliths; see figures 4, 5 and table 3) and a variable content of exotic lithologies: diabase, gabbro, gabbro pegmatite and serpentinite. The limestone clasts include rare Permian fusilinid limestones not known in the range (Weiler 1965, 1969). In the Central Range (for example, 989 836, 148 790) the Kithrea conglomerates unconformably overlie Lapithos Group rocks, but according to Ducloz (1972) contacts with the Trypa Group are always faulted. Along the southern flank of the Eastern Range the conglomerates overlie deeply eroded and reworked Lapithos volcanic rocks (such as 398 870), while further east, conglomerates are virtually absent and marine Kalograia–Ardana sandstones pass transitionally into Kithrea Group turbidites.

The exotic clasts occur along both flanks of the Western and Central Ranges but decrease in abundance eastwards. In the Karpas Peninsula the basal conglomerates, several metres thick, comprise chert, limestone and lava, but without identifiable Trypa Group or ophiolite-derived rocks. Measurements of the well developed clast orientation (figure 9) suggest an eastern provenence with a component from the north in the Western Range.

Interpretation of the basal conglomerates suggests deposition on a broad flood plain with

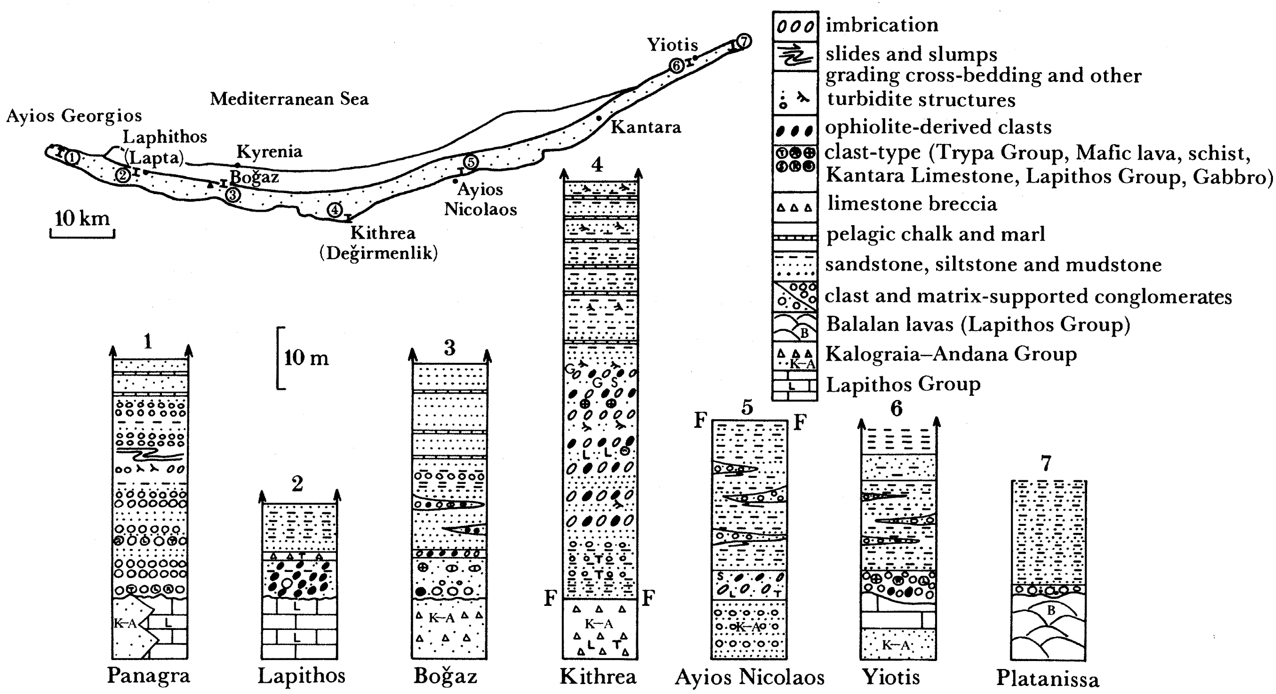

FIGURE 8. Lithological logs of the basal Kithrea Group.

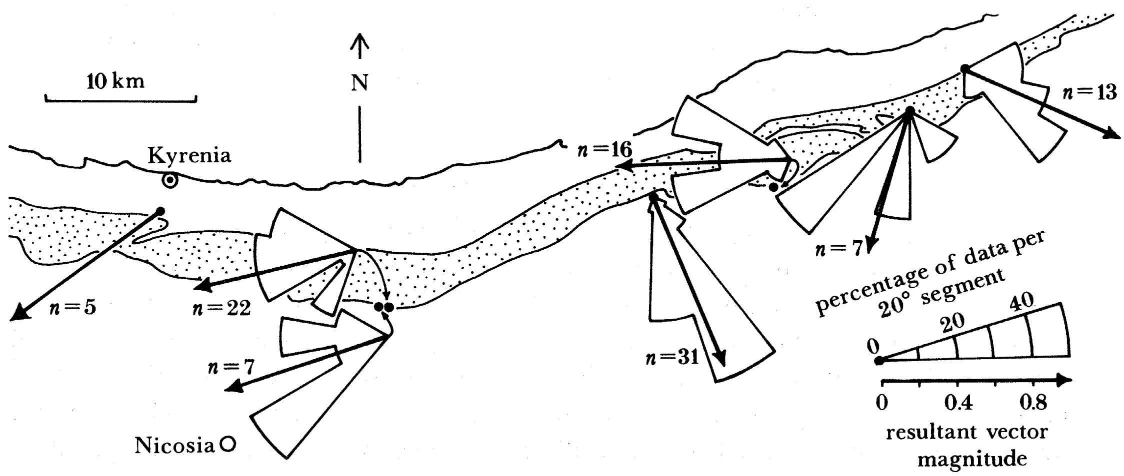

FIGURE 9. Palaeocurrents from clast imbrication in clast-supported conglomerates in the basal Kithrea Group flanking Trypa and Lapithos spin to the range (stippled); n = number of readings.

a hinterland generally to the north and a coastline running south of the present range front. This is suggested by the palaeocurrents and the relatively uniform small size (10–40 cm diameter) and rounding of the exotic clasts. These were probably derived from an already deeply eroded ophiolite sheet. During basal Kithrea Group time the Kyrenia Range apparently formed an impersistent range of low coastal hills, shedding the locally derived alluvial debris. An important implication is that the Cicilia–Ardana Basin (figure 1) would have formed by 4–5 km of post-Oligocene subsidence of a previous continental area between Cyprus and mainland Turkey.

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(c) *Oligocene–Miocene turbidite deposition*

A striking feature of Kyrenia Range geology is the rapid upward transition from the basal Kithrea alluvial conglomerates into turbidites (figure 8). Typically the conglomerates fine upwards over several metres, passing into medium-bedded, coarse-grained graded sandstones, then into turbidites with white chalky partings containing an indigenous planktonic fauna of late Oligocene age (such as 774 872, 989 836, 047 835; logs 1, 3, figure 7). Even more strikingly (148 790, figure 8), coarse conglomerates locally pass sharply into a pebbly sandstone, 6 m thick, then into medium-bedded sand turbidites; a transition so sharp as to have been mapped as a fault by Baroz (1979). On both flanks of the Eastern Range (504 901 and 488 918) the lowest Kithrea turbidites contain several metre-thick debris-flow rudites with associated slumping.

Evidence thus exists for a rapid collapse of the entire Kyrenia Range Lineament after early Oligocene time, ushering in a prolonged period of turbidite deposition that blanketed the entire range. These Kithrea Group turbidites reach a tectonic thickness of 2200 m in a borehole south of the Eastern Range front (Geçitkale (Lefkoniko), see figure 1), which bottomed in Troodos pillow lavas. The stratigraphy and micropalaeontology of the Kithrea Group has been described in detail by Baroz (1977) and Baroz & Bizon (1977). We restrict ourselves here to points relevant to the overall tectonic setting (Table 5). As far as can be judged from the patchy outcrop, no distinction is to be drawn between depositional environments across the range, but a persistent feature is that coarsening and thickening trends and abundant palaeocurrents indicate an easterly derivation (Weiler 1964, 1965, 1969, 1970). The lowest turbidites (Bellapais Formation, see table 5) are often plant-rich and consist of 80% extra-basinal material. They comprise abundant mafic igneous, metamorphic and sedimentary-derived material. The overlying late Oligocene Klepini Formation becomes more sandy eastwards, again rich in derived volcanic grains. By contrast the overlying Flamoudi Formation (Lower Miocene, see table 5) is much richer in carbonate grains and both benthonic and planktonic foraminifera. The Aquitanian in the Taurides was a time of major marine transgression and high shelf carbonate productivity. Much of the former source terrain of ophiolitic debris was submerged, leaving only deformed Mesozoic carbonate platform rocks upstanding. The overlying Panagra Formation (Langhian) is highly calcareous, with an abundance of epiclastic volcanic grains in turbiditic sandstones. Acid tuffs appear locally.

From Serravallian to Tortonian time onwards (late Miocene) an important distinction appears between facies north and south of the Kithrea Fault (figure 1, table 5). To the north the mostly fine-grained Trapeza Formation is overlain by Davlos Formation turbidites with dominant extrabasinal grains and many intercalations of acid tuffs and hyaloclastites. To the south the Mia Milia Formation is much thicker (1600 m) and comprises turbidites with a higher proportion of Kyrenia Range grains. Judging from the record of D.S.D.P. sites 375 and 376 (Leg 62) on the Florentine Rise SW of Cyprus (figure 1), clastic input reached its maximum during this late Miocene interval. There, the earlier Miocene facies consist mostly of deep-water nannofossil foraminiferal marlstones, with an incoming of turbiditic sandstones and siltstones in the Tortonian interval (Hsü *et al.* 1976; Baroz *et al.* 1978). In the D.S.D.P. cores the late Miocene culminates in deep-water evaporites including halites, equivalent to the Lapatza gypsum south of the Kyrenia Range front (table 5).

TABLE 5. GEOLOGICAL SUMMARY OF THE KITHREA GROUP[a]

formation	age	thickness m	lithofacies	interpretation
Lapatza	Messinian	0–150	organic-rich mudstones with reworked fossils pass up into alternating white pelagic chalk, mudstones and organic partings, then into laminated saccaroidal and selenitic gypsum	regional uplift of Cyprus diminishes clastic input high surface-water productivity sea level drawdown creates barred basin in Mesaoria which finally desiccates
Mia Milia	Tortonian Serravalian	1600	S of Kithrea fault only thin- to medium-bedded turbidites with Trypa- and Lapithos-derived clasts	turbidite fill of rapidly subsiding half-graben with Kithrea growth fault to N local exposure and erosion of Kyrenia Range rocks possible calc-alkaline volcanism
Davlos	Tortonian	250	N of Kithrea fault microconglomeratic base, thins and fines W on both flanks of range compositionally constant, acid tuffs	turbiditic overspill from Kithrea half-graben onto raised range area to N quite extensive ?calc-alkaline volcanism in basin
Trapeza	Tortonian Serravalian	140	N of Kithrea fault organic-rich mudstones becoming sandier and thicker to E diagnetic ironstone concretions and gypsum	relatively raised area N of Kithrea half-graven receives fine sediment only quiet setting of plant debris, turbidites deflected S
Panagra	Langhian	50	interbedded bioclastic limestones with large benthonic forams and terrigenous clastics with basic volcanic and sedimentary grains some acidic tuff	regression re-exposes Tauride volcano–sedimentary source area reworking of shelf carbonates onset of acid volcanism
Flamoudi	Burdigalian Aquitanian	200	turbiditic sandstones become microconglomeratic to E grains mostly platform carbonate-derived packstone interbeds have planktonic and benthic forams, glauconite	transgression of Tauride platform increases shelf carbonate productivity and cuts off other sources
Klepini	late Oligocene	120–160	turbiditic sandstone and mudstone, coarsens and thickens east basic volcanic-derived. Black organic-rich interbeds become pink upwards	east-derived turbidites, similar on both flanks of range relatively deoxygenated diagenesis; ample nutrients favoured high plankton productivity
Bellapais	early Oligocene	40 (max)	basal conglomerate rich in Kyrenia-derived clasts, with ophiolitic and metamorphic clasts passes up into turbidites with mainly exotic grains, particularly basic volcanic	alluvial and beach conglomerates shed into rapidly subsiding basin same facies on both flanks of range and in fault blocks within the range

[a] Sources are specified in the text.

(*d*) *Oligocene–Miocene tectonic setting*

From late Oligocene to late Miocene time the whole Kyrenia Range was a deeply submerged part of a huge submarine fan system issuing from the Ardana Basin or Gulf of Iskenderun area (figure 1; Weiler 1965). The source was an area of deformed Tauride units. By contrast, the Miocene facies on the Alanya Massif due north of Cyprus comprise much thinner coastal alluvial fan and fan-delta deposits (unpublished data). During late Oligocene and Miocene time the Troodos Massif was gradually emerging (Robertson 1977*b*), while the area to the north, including the Kyrenia Range and the Cicilia–Ardana Basin, must have subsided by several kilometres. This is attributed to strong crustal extension of the 'fore-arc' area. In late Miocene time the Kithrea Fault became active, bounding a half-graben in the Mesaoria Plain. The Kyrenia Range, though still submerged, subsided less rapidly than this basin, possibly a prelude to Pliocene uplift. Miocene deposition culminated in the widespread Messinian sea-level drawdown, resulting in gypsum deposition within several silled basins in the Mesaoria south of the range.

10. Pliocene and Quaternary: thrusting and uplift

(*a*) *Pliocene sedimentation*

During Neogene time, in common with other parts of Cyprus, the Kyrenia Range saw a switch to pulsed major uplift that we attribute to regional northward underthrusting of African crustal elements (Gass & Masson-Smith 1963; Robertson & Woodcock 1980).

Pliocene sediments are patchily exposed south of the Kyrenia Range. In the west, the lowest Pliocene (Mirtou Formation, see table 6) has a thin basal conglomerate consisting of clasts of late Miocene evaporites with minor Kyrenia Range Lapithos and Trypa lithologies (Baroz 1979). The overlying nannofossil-rich mudstones are comparable with the condensed deep-water nannofossil marls and sapropels recorded in D.S.D.P. sites 375 and 376 (leg 42) off SW Cyprus (Hsü *et al.* 1978; Baroz *et al.* 1978). The sparcity of Kyrenia Range debris suggests that at this time the range was not significantly elevated with respect to the Mesaoria basin to the south. However, the facies contrast with the deep-water D.S.D.P sites suggests that the whole of northern Cyprus had, since the late Miocene, been elevated above the surrounding eastern Mediterranean sea floor.

It was probably about this time, after stripping of the Kithrea Group but before emergence, that micritic limestones bearing both planktonic and benthonic foraminifera penetrated deep fissures in the Trypa Group carbonates in the Central Range (for example, 056 805 (Ducloz 1972)). Uplift of the Kyrenia Range became more apparent later in early Pliocene time as the Mirtou Marls are unconformably overlain by up to 20 m of conglomerates calcarenites and mudstones (Nicosia Formation, see table 6). Clasts, often well rounded, are mostly Kithrea Group-derived with few clasts of older Kyrenia Range lithologies. Calcarenite interbeds contain a rich, mostly shallow, marine fauna, including pelecypods, brachiopods, gastropods, benthonic and planktonic foraminifera and calcareous algae (Reed 1935; Henson *et al.* 1949). In the Karpas Peninsula extensive Pliocene deposits unconformably overlie the Kithrea group.

Throughout early Pliocene time, sharply contrasting much thicker (100–750 m) sequences accumulated south of the Ovgos Fault (figure 1) around the northern and eastern Troodos perimeter ('Potami Formation', see table 5). Successions comprise conglomerates, mudstones, lithic sandstones and limestones with a neritic fauna (Reed 1930; Bear 1960; Cockbain 1959;

Table 6. Geological summary of the Pliocene to Recent Units[a]

group	formation	age	thickness	lithofacies	interpretation
	six terraces	Quaternary (Flandrian–Emilian)	heights 0.6, 10, 30, 60, 140, 180 m	repeated regressive cycles from shallow marine to coastal then continental facies occur on N and S flank of range higher terraces tilted away from range axis	continuous tectonic uplift of range superimposed on general uplift of island and effects of climatic and eustatic changes
	Karka	Villafranchian	variable	breccias and lacustrine deposits on terraces at 300–820 m height marine facies to 300 m facies seal thrusts but cut by normal faults	pulsed uplift (more than 500 m) following Villafranchian erosion surface, gives terraces above cemented screes ephemeral intermontane lakes humid weathering gives Karka surfaces
	Athalassa	middle Pliocene	50–26 m	conformable on Nicosia except at basin margins marine conglomerate, well-rounded Troodos – and Kyrenia-derived clasts passes up into coastal and continental facies	Fan deltas shed from N and S into shallow regressing sea, pass into coastal, beach and dune environments and alluvial fans with caliche-clad floodplains
Mesaoria	Potami	early Pliocene	100–750 m	transgressive over Troodos periphery (equivalent to Mirtou and Nicosia Formation Troodos-derived mudstones, sandstones and conglomerates	subsiding basin bounded by Troodos to S and Ovgos Fault to N shallow open marine shelf, storm affected, with marginal clastic plant-rich facies
	Nicosia	early Pliocene (Piacenzian)	20 m	conformable on Mirtou Formation or (in Karpas) unconformable on Kithrea Group Kithrea-derived sandstones + thin conglomerates, with limestones (neritic + pelagic fauna)	shallow marine facies, little supply from Kyrenia Range high-energy coastal belt marked by uplift and erosion of Kithrea Group in Karpas
	Mirtou	early Pliocene (Zanclian)	50 m	basal conglomerates, mostly reworked Messinian plus Kyrenia carbonates mudstone above becomes sandier and more fossiliferous up and S of Ovgos Fault	shallow marine transgression over subdued topography to N and S well oxygenated raised well above surrounding east Mediterranean (cf. D.S.D.P. site 376)

[a] Sources are specified in the text.

Ducloz 1965). Derivation was from an eroding vegetated Troodos landmass to the south and west. A series of north-down steep faults stepped the Troodos basement at this time, allowing thick sediments to accumulate as far north as the Ovgos Fault. This fault may have then had a south-down component and substantial later displacements may also have enhanced the facies contrast across this line.

In mid Pliocene time, huge volumes of clastic material began shedding from both the Troodos Massif and the Kyrenia Range into a rapidly dwindling shallow Mesaoria sea. South of both the Western and the Eastern Range the Athalassa Formation (table 6) consists mostly of shallow marine conglomerates, limestones and mudstones, probably of Middle Pliocene age (Reed 1930, 1935; Henson *et al.* 1949; Cockbain 1959; Moshkowitz 1963; Baroz 1979). Clasts were derived from all the Kyrenia Range lithologies. In the type area further south, around the Troodos periphery, the Athalassa Formation is transitional from the underlying Nicosia Formation. The inferred palaeogeography is of the two rapidly rising landmasses shedding a series of braided alluvial fans into shallow regressing seas.

(*b*) *Neogene deformation* (D3)

A third major deformation event is most clearly indicated by strong southward thrusting and south-vergent folding of the Kithrea Group south of the Kyrenia Range (figure 1) and by southward thrust emplacement of the already deformed Troodos and Lapithos sheets onto Kithrea Group along the southern range front. The precise age of this deformation is uncertain. It affects all the Kithrea Group formations, including the Lapatza (Messinian), and was regarded by Baroz (1979) as end Miocene but pre-Mirtou Formation (lowest Pliocene). However the mostly fine-grained nature of this formation precludes major thrust uplift of the Kyrenia Range at this time, and the abrupt switch to conglomerate deposition (Athalassa Formation) in mid Pliocene time is more likely to represent the culmination of the D3 event. There is a major sub-Athalassa unconformity, but the earlier sub-Mirtou unconformity suggests that the D3 deformation was progressive through latest Miocene to mid Pliocene time (figure 2).

The geometry of the major structures in the Kithrea Group is unknown in detail. Lack of borehole and geophysical control and the fact that some of the major thrusts coincide with and controlled earlier facies changes prevent construction of balanced sections. Major thrusts are mapped (Baroz 1979) where low stratigraphic levels are exposed in the hanging wall. Lapithos Group rocks are upthrust 3 km south of the Eastern Range and along the Ovgos fault, 13 km south of the Central and Western Range. The open kilometre-scale folds in the Kithrea Group may be culminations above buried thrust ramps or fault tips (figure 10*d*). The positions of some deep footwall ramps may have been controlled by the pre-existing north-facing steps in the Troodos basement.

Close to the southern flank of the range, folds become tighter and thrusts more common. Relations are commonly obscured by Plio–Pleistocene deposits. However, clear imbrication of Lapithos and Kithrea Groups occurs on the south flank of the Eastern Range (see, for example, figure 10*a*), and faults south of the Central Range obliquely transect Trypa and Lapithos sheets and thrust them over Kithrea Group. Along this southern flank faults dip northward, with rapid changes from steep to shallow dips which may indicate a ramp and flat geometry. However, megascopic geometries are complex because of the interaction of D2 and D3 structures.

On a mesoscopic scale D3 structures in the Kithrea Group comprise sporadic folds, mostly

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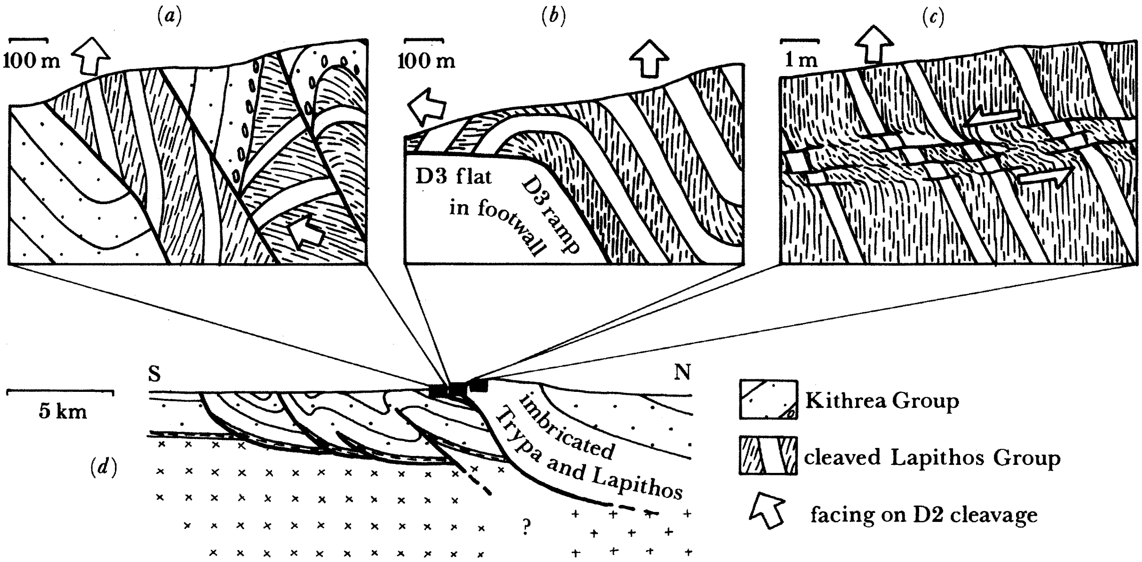

FIGURE 10. Schematic geometry of D3 (Neogene) structures in N–S sections. See text for explanation.

sub-horizontal and south-facing, and an uncommon weak cleavage in mudstones. The strong D2 cleavage in the Lapithos Group is sometimes crenulated or cut by small low-dip south-verging shear zones, probably of D3 age (figure 10*c*). The generally south-verging cleavage-bedding relation imposed during D2 is reoriented by D3 structures on a variety of scales. Throughout most of the range D2 structures have been tilted northwards by up to 90°: D2 thrusts are now steep and structures face steeply up to the north on the D2 cleavage (figure 10*b*). This northward tilting may be a result of D3 thrusting up a major crustal ramp in the footwall (small-scale section, figure 1) controlled by the geophysically documented northern edge of the Troodos basement slab (Aubert & Baroz 1974), which was also the southern boundary of the postulated earlier strike-slip lineament. It is significant that the large-scale rotation of D2 structures does not seem to occur in the Karpas Peninsula, south of the postulated basement ramp. At occasional localities along the south flank of the Central and Eastern Ranges (for example, 429 877), the D2 structures now face gently southwards, probably in the remnants of a D3 hanging-wall ramp (figure 10*b*).

Structures attributable to the D3 event are less common along the axis of the range. Some of the late fissuring and brecciation of the Trypa Group could be of this age. Along the north flank of the range the Kithrea Group sediments have been tilted northward, but evidence of outcropping thrusts is absent.

In a regional perspective, the Neogene deformation must represent significant N–S shortening. The northward tilting of the range requires that the major N-dipping thrust along its south flank shallows with depth, extending northwards beneath the present range (figure 10*a*) to root below the Cicilia–Ardana Basin. South of this main thrust zone an imbricate stack developed, with thrusts soling out at least as low as the Lapithos Group equivalents beneath the Mesaoria Plain and perhaps even rooting in reactivated fault ramps in the Troodos basement.

(c) *Pleistocene–Quaternary: latest uplift phase*

The Quaternary history of the Kyrenia Range, like the rest of Cyprus, documents rapid vertical uplift which is probably still continuing.

The upper part of the Athalassa Formation in the Mesaoria, ends with a marked erosion surface (Baroz 1979), overlain by continental deposits (fanglomerates) of early Pleistocene (Villafranchian) age, the Old Quaternary of Ducloz (1972) or the Karka Formation of Baroz (1979). Varied deposits include cemented screes, intramontane lacustrine, alluvial, lagoonal and coastal facies (table 6). Extensive travertine precipitates and faunas including pygmy elephants and dwarf hippopotami (Bate 1903, 1904*a*, *b*) point to a warm, humid climate of the glacial pluvial periods. Many old faults are sealed, but some were reactivated by vertical dip-slip at this time (Ducloz 1972). Several raised and tilted coastal terraces (Karka surfaces) exist at altitudes of up to 820 m (de Vaumas 1959*a*, *b*, 1961). Dreghorn (1978) confirms that the southern flank terraces have been folded (pink surface of de Vaumas), the altitude decreasing progressively away from a point of maximum uplift in the Central Range.

Subsequent uplift, superimposed on the effects of fluctuating Quaternary sea levels, generated a series of six terraces at heights ranging from 180 m to locally just below sea level (Ducloz 1968, 1972; Dreghorn 1978; Baroz 1979), corresponding to an estimated uplift rate of 0.3 mm per year (table 5 of Dreghorn (1978)). Each terrace is cyclic, passing upwards from open shallow marine to coastal and continental deposits, followed by renewed transgression.

11. The Kyrenia Range as a crustal lineament

The most obvious features of the range are complexity and longevity. A deep crustal lineament has already been discontinuously active and continues to be so as Africa and Eurasia converge. This complicated history involves episodic passive-margin, active-margin and strike-slip phases. Major crustal lineaments such as this are inherently very complex and any simple solutions should be viewed with scepticism. Large gaps in the record commonly exist (as a result, for example, of strike-slip removal). Key clues are easily destroyed, such as unconformities or early high-level deformation features.

The Kyrenia Range demonstrates similar geological processes at different times. Any attempt to order events on deformation style or volcanic composition would be unsuccessful. Volcanism, probably related to strike-slip faulting, took place in separate Campanian, Maastrichtian, Palaeogene and Miocene phases. Compressional deformation with a variable strike-slip component took place during late Cretaceous, late Eocene and late Miocene–Pliocene phases. Along inferred strike-slip faults major uplift of the order of kilometres must have taken place in the late Cretaceous in geologically negligible time.

Sedimentary rocks emerge as the most sensitive indicators of tectonic setting. For example, strike-slip faulting gave rise to localized scree breccias in the Maastrichtian and Palaeocene, whereas in the mid–late Eocene crustal compression, a transpression produced a frontal flysch basin which rapidly filled with flysch and olistostromes. If there is any moral in this study it is that the overall understanding of major crustal lineaments will only come about through in-depth integrated studies of well preserved individual examples.

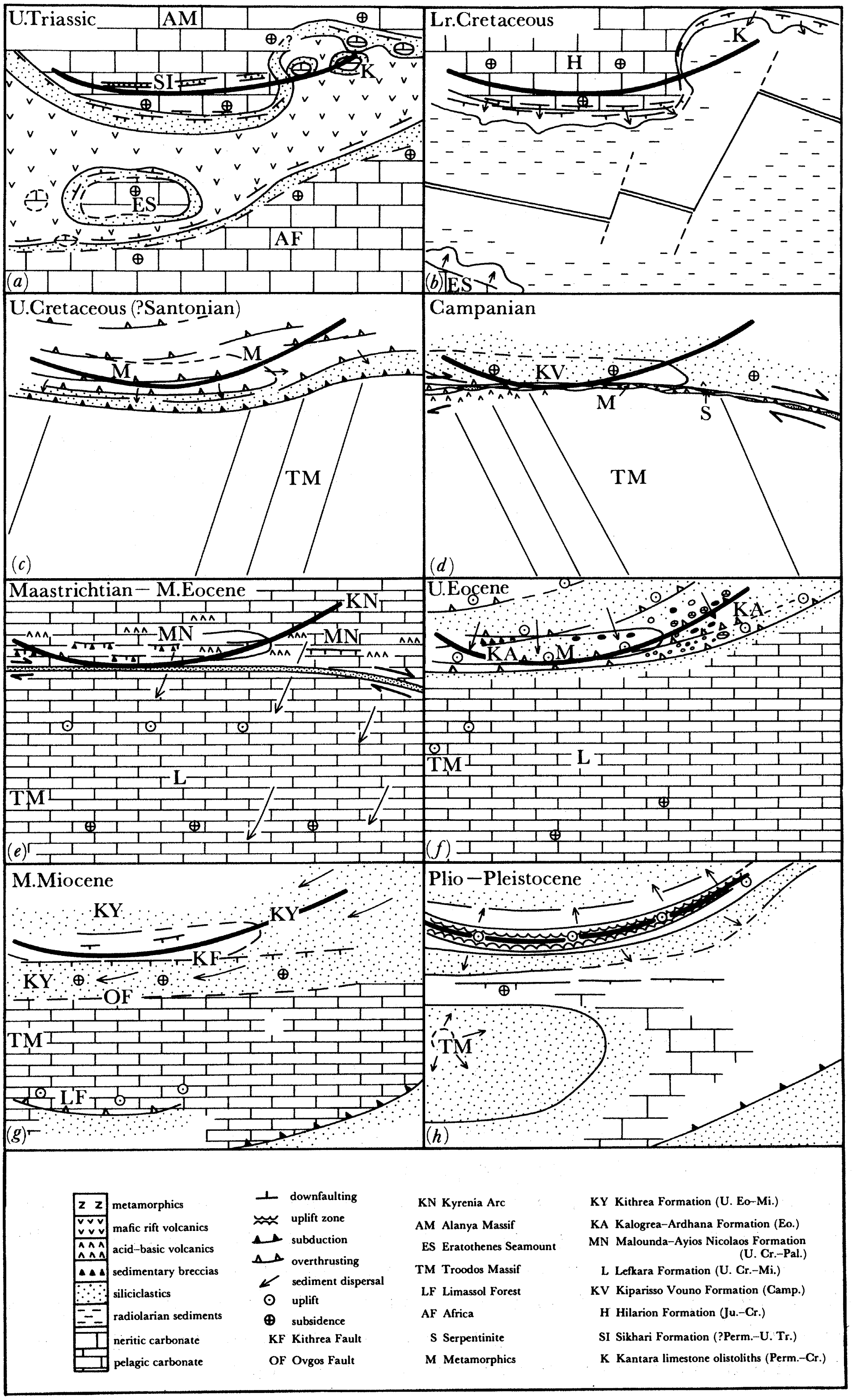

FIGURE 11. For description see facing page.

12. Conclusions

(1) From an integrated study, the Kyrenia Range is shown to exemplify the complexity of a long-lived (*ca.* 200 Ma) deep crustal lineament in the eastern Mediterranean.

(2) The Kyrenia Lineament is located partly along (western and central segments) and partly straddling (eastern segment) the abrupt northward termination of the Troodos ophiolitic crust at depth.

(3) The area rifted from Gondwana in the late Triassic to form a gently subsiding carbonate platform close to the southern edge of a Turkish microcontinent (figure 11).

(4) To the south a small ocean basin offset by fracture zones formed in Cretaceous time (Troodos ocean). From regional evidence, northward subduction began in late Cretaceous time (?Santonian).

(5) During the late Cretaceous (Campanian) the present southern border of the range became an active, probably dextral, strike-slip zone, thought to be associated with the 90° anticlockwise rotation of the Troodos microplate. The Mesozoic passive margin was removed by early Tertiary time, juxtaposing crust similar to the Troodos ophiolite. Mesozoic platform rocks were pervasively sheared and tectonically brecciated (D1). Metamorphic rocks formed, probably by shear heating at depth, and were slivered up sub-vertical fault zones. A very similar strike-slip lineament is preserved in the Mamonia Complex of SW Cyprus and probably also represents part of the Troodos microplate boundary.

(6) In the Maastrichtian and early Tertiary the range area subsided and was blanketed by pelagic carbonate. Within-plate-type bimodal acid–basic volcanics were erupted and scree breccias were shed from active submarine fault scarps in an extensional setting still influenced by strike-slip.

(7) Collisions in the Taurides to the north became more pervasive and by mid Eocene time the Kyrenia Range came under general N–S compression (D2), culminating in major southward thrusting and possible localized metamorphism. A range of syn-tectonic sediments accumulated before (flysch), during (olistostromes) and after (sub-aerial scree breccias) the deformation climax.

(8) Continued convergence of Africa and Eurasia was accommodated by subduction, possibly in a trench located off southern Cyprus, and the range was located broadly in a fore-arc setting during Oligocene and Miocene time. The area subsided rapidly and was enveloped in thick Miocene flysch, fed into the eastern Mediterranean from the Tauride Mountains through a huge submarine fan complex located to the northeast.

Figure 11. Schematic palaeogeography of the Kyrenia Range region during Mesozoic and Cenozoic time for eight time intervals. The present location of the Kyrenia Range central axis is marked as a solid black curved line. (*a*) The Kyrenia Range is rifted from Gondwana in late Triassic time and becomes part of a subsiding carbonate platform located along the southern margin of a Turkish microplate. (*b*) During Cretaceous time a small ocean basin offset by transform faults forms to the south. (*c*) Northward subduction is initiated in late Cretaceous (?Santonian) time, initially deforming the Kyrenia Range. (*d*) During the Campanian, major strike-slip faulting related to the anticlockwise rotation of the Cyprus microplate removes the south-facing passive margin. (*e*) Pelagic carbonate accumulates on the foundered Mesozoic platform, accompanied by extrusion of bimodal volcanics related to local crustal extension during strike-slip. (*f*) Syntectonic sedimentation of mid-late Eocene age culminated in major southward thrusting. (*g*) Northward subduction under Cyprus accompanied strong subsidence of Cyprus in a 'fore-arc' setting, and enveloping by flysch derived from the northeast. (*h*) Plio–Quaternary uplift related to underthrusting of Eurasia by Africa.

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(9) Differential vertical movements south of the range (Kithrea Fault) were coupled with general uplift, isolating the area from the surrounding east Mediterranean sea floor in the late Miocene.

(10) Renewed compressional deformation from end Miocene to mid-Pliocene time (D3) saw large-scale thrusting and northward tilting of the eastern and central range areas and large-scale, more open, folding further east.

(11) Rapid pulsed uplift took place during the Quaternary. If Africa–Eurasia convergence continues, the Kyrenia Range is likely to end up as a far-travelled composite nappe pile underlain by tectonic mélange, comparable with some older collisional counterparts in Tauride areas to the north.

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