Tectonic history of a segment of the Pelagonian zone, northeastern Greece

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Continental metamorphic rocks and ophiolitic bodies within the Pelagonian zone of the Hellenides in the Livadi area, northeastern Greece, show repeated periods of deformation that accompany thermal events of Early Cretaceous and possibly Late Eocene age. Structures associated with the earlier deformation indicate thrusting towards the northeast accompanying regional metamorphism of upper greenschist to lower amphibolite facies. Later structures and a retrogression to lower greenschist facies associated with emplacement of the Livadi ophiolitic rocks into their present position are likewise attributed to northeast-directed thrusting and probably accompanied the allochthonous movement of the Pelagonian basement over the Mesozoic platform carbonates of Mt. Olympos.

Emplacement vectors of northeast polarity are inconsistent with tectonic models of the Hellenides involving large-scale southwestward obduction of Mesozoic ophiolites from a single ocean located northeast of the Pelagonian zone. Tectonic models involving the converging emplacement of Mesozoic ophiolites from two oceans lying northeast and southwest of the Pelagonian zone are more compatible with the observed structures, the latter ocean providing a potential root zone for the deformed ophiolitic rocks at Livadi.

The orientation of minor structures associated with thrusting that postdates the emplacement of the Livadi ophiolitic rocks is consistent with movement from north to south.

Les roches métamorphiques continentales et les appareils ophiolitiques à l'intérieur de la zone pélagonienne des Hellénides dans la région de Livadi au nord-est de la Grèce présentent des déformations répétées à diverses périodes qui accompagnent les poussées thermales d'âge Crétacé inférieur et possiblement Eocène supérieur. Les structures associées avec la première phase de déformation indiquent un charriage vers le nord-est accompagnant le métamorphisme régional du faciès des schistes verts intense au faciès des amphibolites naissant. Des structures subséquentes et une rétromorphose du facies des schistes verts intense associées avec la mise en place en position actuelle des roches ophiolitiques du Livadi, sont vraisemblablement attribuées à un charriage vers le nord-est et probablement accompagnées d'un déplacement allochtone du socle pélagonien au-dessus des carbonates de la plate-forme mésozoïque du Mt. Olympos.

La polarité nord-est des vecteurs lors de la mise en place est incompatible avec les modèles tectoniques des Hellénides impliquant une obduction à grande échelle vers le sud-ouest des ophiolites mésozoïques d'un seul océan situé au nord-est de la région pélagonienne. Les modèles tectoniques s'appuyant sur une mise en place convergente des ophiolites de deux océans s'étalant au nord-est et sud-ouest de la zone pélagonienne sont plus compatibles avec les structures observées, ce dernier océan fournissant l'infrastructure potentielle pour les roches ophiolitiques déformées du Livadi.

L'orientation des structures mineures associées avec le charriage indique une mise en place tardive des roches ophiolitiques de Livadi est conforme au déplacement du nord vers le sud.

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Introduction: the tectonic problem of the Hellenides

The geology of the Hellenides has long been interpreted in terms of a series of north-northwest-trending zones (Fig. 1) that represent broadly distinct facies units displaying differing structural characteristics (Aubouin 1965; Smith and Moores 1974). Simplistically, these "isopic zones" (Aubouin 1959) fall into three groups. Olympos and the Paxos, Gavrovo–Tripolitza, and Parnassos zones are characterized by Mesozoic neritic carbonates and the Ionian, Pindos, Othris, and Vardar zones contain Mesozoic pelagic facies with or without ophiolites. In contrast, the Pelagonian, Serbo-Macedonian, and Rhodope zones comprise largely pre-Mesozoic metamorphic continental basement over which metamorphosed Mesozoic carbonates form a discontinuous cover. In the Pelagonian zone these carbonates are tectonically overlain by ophiolites. The zones are bounded by fault contacts and are considered to represent a succession of superimposed thrust sheets that attained their present geometry during the Tertiary. The sense of emplacement, however, remains controversial.

Aubouin (1965) interpreted those zones lying west of the Pelagonian in terms of a formalized geosynclinal model. Throughout most of the Mesozoic, pelagic eugeosynclinal sediments of the Pindos and Othris zones were separated from the miogeosynclinal pelagics of the Ionian zone by the Gavrovo platform. The Pelagonian

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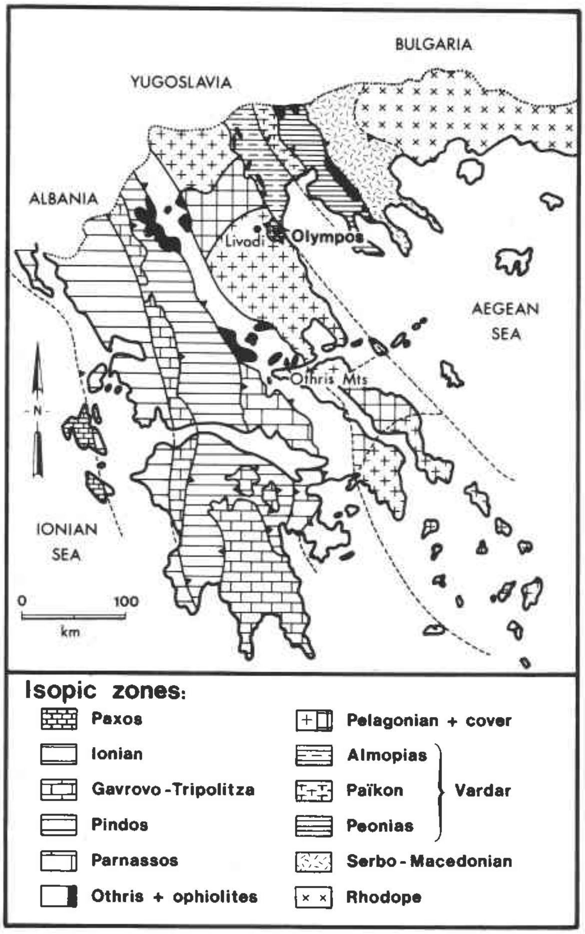

FIG. 1. The isopic zones of the Hellenides after Aubouin (1965) and others. Ophiolites of the Vardar zone after the 1 : 1 000 000 metallogenic map of Greece (Zachos and Maratos 1965).

zone was considered by Aubouin to represent the stable eastern margin of this geosynclinal couple while the neritic carbonates of the Paxos zone constituted the foreland. This simple geometry was subsequently destroyed in late Mesozoic and Tertiary time with the onset of a deformation that progressed westwards. Emplacement of the resulting thrust sheets was from the east.

The subsequent description by Godfriaux (1968) of a major tectonic window (Mt. Olympos) within the Pelagonian zone of northeastern Greece required a modification to Aubouin's model in that the Pelagonian zone was shown to be allochthonous rather than forming a stable geosynclinal margin. On the basis of paleogeography and thrusting in areas to the southwest, Godfriaux concluded that the Pelagonian zone was rooted to the northeast and, on the basis of lithology, equated the neritic carbonates of Olympos with those of the Gavrovo–Tripolitza zone, a correlation later supported by microfauna (Fleury and Godfriaux 1974). Such a correlation implies that the intervening zones have moved southwestward across the Olympos carbonates (Fig. 2*a*), requiring a cumulative displacement on the Olympos thrust of several hundred kilometres (Barton 1975, 1979). The plate tectonic interpretations of Dewey *et al.* (1973), Mercier *et al.* (1975), and Zimmerman and Ross (1976), for example, attribute this southwest-directed movement to the Late Jurassic or Early Cretaceous closure of an ocean now represented by the Vardar zone. Remnants of this ocean are preserved as ophiolites in the Almopias and Peonias subzones of the Vardar, while the ophiolites of the Pindos, Othris, and Pelagonian zones are interpreted as fragments of the same ocean that were transported southwest over Olympos (Fig. 2*a*).

The contrasting tectonic model of Hynes *et al.* (1972) and Smith *et al.* (1975) is based on the tectonic interpretation of the Othris zone in eastern central Greece. In the Othris Mountains (Fig. 1) ophiolites and marine units of successively more continental affinities have been tectonically stacked onto the western margin of the Pelagonian zone. Reconstruction of this deformed Mesozoic continental margin sequence based on gross stratigraphy (Hynes *et al.* 1972; Ferrière 1974, 1976), structural data (Smith and Woodcock 1976*a*; Smith *et al.* 1979), and sedimentary facies relationships (Smith *et al.* 1975; Price 1976, 1977) points to the existence of an ocean that lay *west* of the Pelagonian zone throughout much of the Jurassic. Ophiolite emplacement and tectonic stacking of the Othris continental margin are attributed to the onset of eastward subduction of this Othris ocean during Late Jurassic to Early Cretaceous time (Smith and Woodcock 1976*b*). Final destruction of the ocean is believed to have occurred during the early Tertiary. At this time the Pindos zone, which may preserve sediments from this ocean, moved westwards onto the Gavrovo–Tripolitza zone (Temple 1968; McCaig and Kemp 1979), while the Pelagonian zone moved northeast (Barton 1975) over the Olympos carbonates (Fig. 2*b*). Although the existence of a Vardar ocean and the southwesterly vergence of its destruction is not disputed, it follows that the Vardar could not represent the root zone for the Othris and Pindos sheets (Barton 1975). It further implies that similar carbonate facies can be developed on opposing sides of small ocean basins (Barton 1979).

Support for a two-ocean model may also be derived from the Olympos window itself since the uninterrupted Triassic to Eocene stratigraphy of the Olympos series (Godfriaux 1968) would appear to preclude the possibility of pre-Tertiary tectonic transport across Olympos. Furthermore, a detailed investigation of the structures associated with the Tertiary emplacement of the Pelagonian zone over Olympos has led Barton (1975) to

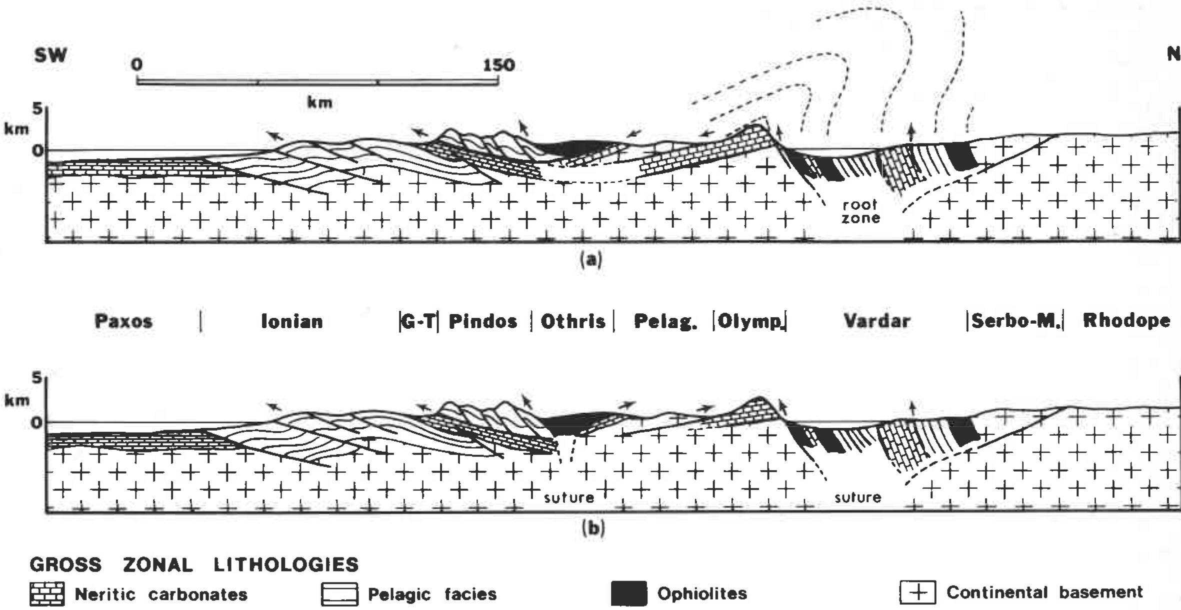

FIG. 2. Schematic sections of northern Greece illustrating the emplacement vectors of isopic zones implied by (*a*) single-ocean and (*b*) two-ocean tectonic models of the Hellenides. Modified after Barton (1975) with additional data from Smith and Moores (1974) and Zimmerman and Ross (1976).

conclude that the Pelagonian allochthon is rooted to the southwest rather than the northeast.

Although the eastward emplacement of the Othris ophiolites has gained some degree of acceptance (Ferrière and Vergely 1976; Cadet *et al*. 1980), the Vourinos ophiolite further north and the Pelagonian zone itself are still widely held to be rooted to the northeast (Vergely 1976; Celet and Ferrière 1978; Zimmerman and Ross 1979). This paper presents structural data from the Livadi area of the Pelagonian zone (Fig. 1), which lies west of Olympos in a region where structural vergence is critical to the problem. The data lend support to Barton's (1975) model of repeated northeast-directed transport.

Geological setting of Livadi

The Olympos region

Godfriaux (1968) first described the Olympos region and recognized three principal lithotectonic units. The continuous Mesozoic platform carbonates and Eocene flysch of Mt. Olympos constitute the autochthonous Olympos series. This series is tectonically overlain to the west by the Paleozoic metamorphics and overlying Mesozoic carbonates and ophiolites of the Pelagonian zone. To the east the series tectonically underlies the Mesozoic pelagic carbonates and ophiolites of the Vardar zone (Fig. 1).

Within the Pelagonian metamorphics Godfriaux (1968) distinguished a stratigraphy in which a lower series of augen schists and associated deformed granites are overlain by metamorphosed mafic–ultramafic lithologies and an albitic mica schist unit. This "basement" sequence is discordantly overlain by probable Triassic to Jurassic marbles and is thus of presumed Paleozoic age. The Pelagonian allochthon overriding Olympos is estimated to be 6 km thick (Barton and England 1979) and records an Early Cretaceous metamorphism (Mercier 1973; Barton 1976; Yarwood and Dixon 1979) that is absent within the Olympos series.

Godfriaux (1968) interpreted the mafic–ultramafic bodies, which outcrop in the vicinity of Livadi, to be a possible Paleozoic ophiolite. Such an interpretation is not consistent with the current tectonic models of the Hellenides since the major ophiolites of this orogenic belt are of Mesozoic age (Smith and Moores 1974). However, if the mafic–ultramafic bodies are Mesozoic then their presence on continental basement far removed from the ophiolite belts of the Hellenides (Fig. 1) necessitates complex late Mesozoic and Tertiary tectonics to allow for their deformation and separation from lithologies typical of the Mesozoic continental margins.

The Livadi area

Structural re-examination of the Livadi area confirms the association of small mafic–ultramafic bodies with the regionally metamorphosed granitic and clastic sedimentary rocks that constitute the base of the Pelagonian zone. However, rather than forming an integral part of the Pelagonian basement, these ophiolitic bodies overthrust and now lie tectonically above the metamorphics

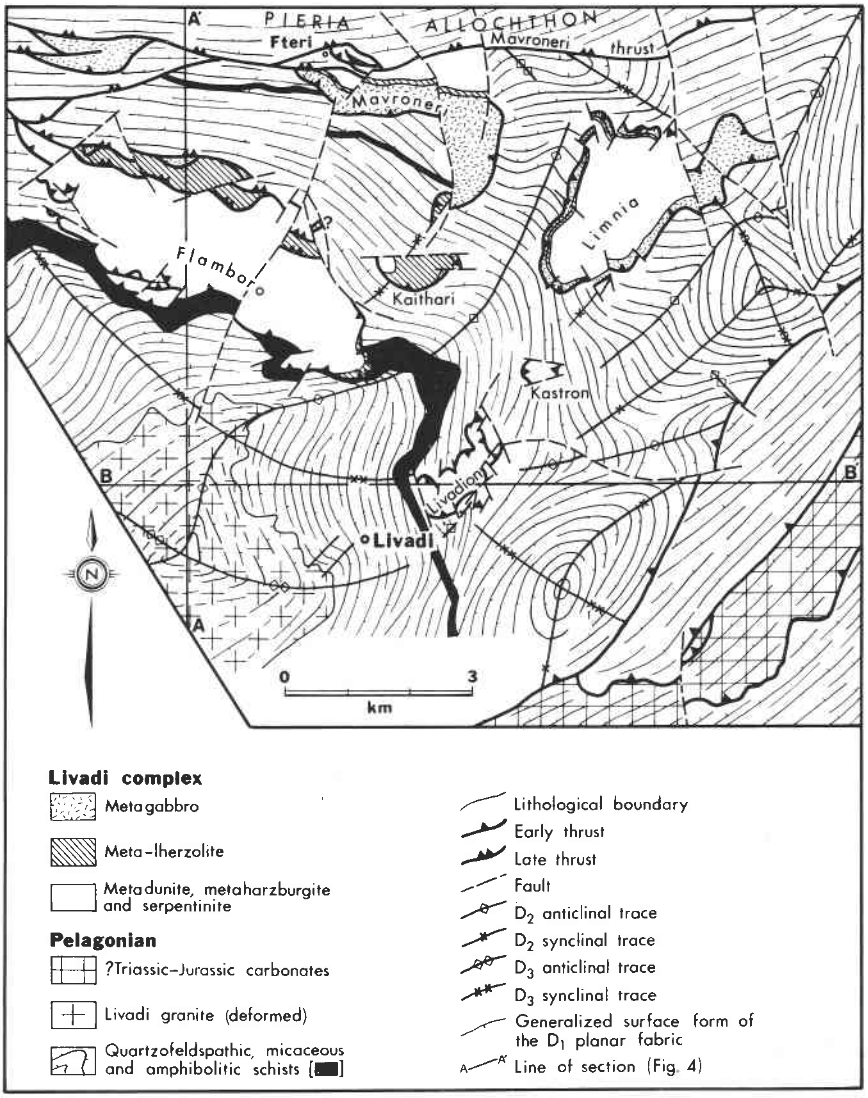

FIG. 3. Simplified original geological map of the Livadi area (see Fig. 1 for location).

(Fig. 3). The thrust klippen collectively constitute the Livadi complex.

The metamorphic rocks of the Pelagonian basement comprise a thick (>4 km) sequence of quartzofeldspathic, micaceous, and amphibolitic schists that represent a succession of probable Paleozoic psammitic and pelitic clastics. Relict primary structures in the form of cross-bedding, grading, channelling, and slump folds are occasionally recognizable and in all cases the compositional banding within the metamorphics corresponds to bedding and the sense of younging is consistently upward. However, portions of the stratigraphy are likely to be missing or repeated since low-angle faulting and shearing are common forms of deformation. In addition to metasediments, granitic and basic (amphibolite) intrusives are widespread. The largest of these, the deformed Livadi granite, occupies the base of the succession (Fig. 4) where its contacts, although sheared, preserve apophyses and are associated with granitic dikes that cut the overlying metasediments.

This gross stratigraphy is tectonically inverted to the north of the map area (Yarwood and Dixon 1979) where similar lithologies occur in a stack of Lower Cretaceous thrust sheets (the Pieria allochthon), which structurally overlie the map area along the younger Mavroneri thrust (Fig. 3). The Kataphygion granite, which forms the structurally highest unit of the Pelagonian basement in the Pieria allochthon, is petrographically similar to the Livadi granite and has been shown (Yarwood and

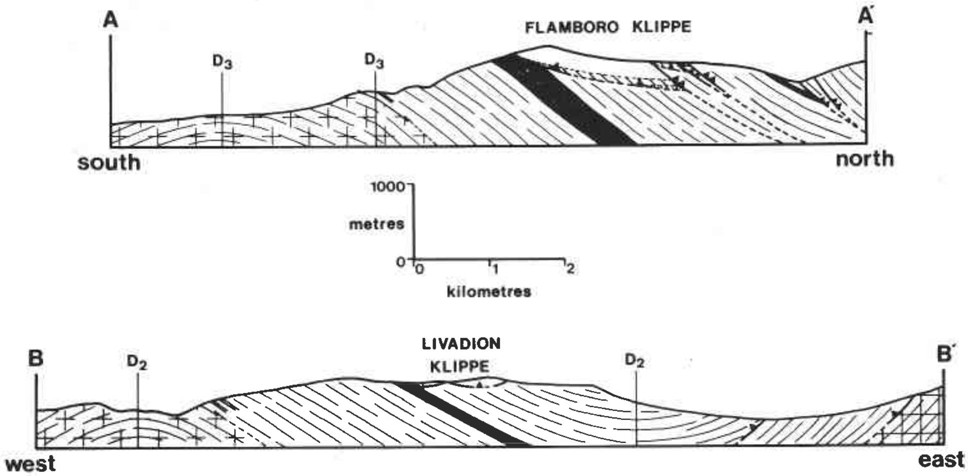

FIG. 4. Schematic cross sections of the Livadi area (see Fig. 3 for location and legend).

TABLE 1. Representative mineral assemblages in the Pelagonian basement and Livadi complex

	Upper greenschist – lower amphibolite facies assemblages	Retrograde lower greenschist facies assemblages	Pre-metamorphic phases
Pelagonian basement			
Quartzofeldspathic schists	Quartz – phengitic muscovite – oligoclase–albite–epidote ± biotite ± actinolitic hornblende	Quartz–phengite–albite–epidote	K-feldspar
Micaceous schists	Quartz – phengitic muscovite – epidote ± biotite ± garnet ± oligoclase ± albite	Quartz–phengite–epidote ± chlorite ± albite	—
Amphibolitic schists	Actinolitic hornblende – phengitic muscovite – oligoclase–albite–epidote–quartz ± biotite	Actinolite–phengite–albite–epidote–quartz	—
Livadi granite	Quartz–oligoclase–albite – phengitic muscovite ± biotite	Quartz–phengite–albite ± epidote	K-feldspar
Meta-igneous amphibolites	Actinolitic hornblende – oligoclase–albite ± phengitic muscovite ± biotite ± epidote	Actinolite–albite–epidote ± phengite ± chlorite	—
Livadi complex			
Metadunite/serpentinite	Metamorphic olivine – antigorite – chromian magnetite	Antigorite–magnetite	Olivine and chromite cores
Metaharzburgite	Metamorphic olivine – talc–antigorite – chromian magnetite	Antigorite–magnetite	Olivine and chromite cores
Meta-lherzolite	High-alumina tremolite – antigorite	Low-alumina tremolite – antigorite–chlorite	Diopside
Metagabbro	High-alumina tremolite – clinozoisite–oligoclase–albite–quartz	Low-alumina tremolite – clinozoisite–chlorite–albite–quartz ± muscovite	—

Aftalion 1976) to have a Late Carboniferous crystallization age.

Mineral assemblages in the basement lithologies reflect two periods of metamorphism. Earlier upper greenschist – lower amphibolite facies phases, characterized by the co-existence of oligoclase (An_{13-23}) and albite (An_{0-5}), predominate but show varying degrees of lower greenschist retrogression. Typical stable assemblages for each event are listed in Table 1. High temperature – low pressure "blueschist" parageneses, reported from the Olympos region by Derycke *et al.* (1974), have not been observed.

The primary stratigraphy of the Livadi complex was one in which dunite gave way in turn to harzburgite, lherzolite, and gabbro to form a broadly layered, cumulate sequence. Varying amounts of the metamorphosed, serpentinized and, in places, structurally inverted equivalents of this stratigraphy are now preserved in the form of klippen, which range in size from 7.5 km by 3 km to less than 0.25 km^2. More or less serpentinized metadunite is the dominant lithology of the more southerly sheets, whereas to the north and northeast occur more complete and locally continuous sequences containing meta-lherzolitic cumulates and metagabbros (Fig. 3). The composite preserved thickness of this stratigraphy is close to 700 m.

The Livadi complex also records two metamorphic events that are compatible in grade and are likely to be of similar age to those of the Pelagonian basement. In the Livadi complex, however, retrograde assemblages are more or less restricted to areas lying close to the basal thrusts of individual klippen. The detailed mineralogy of the complex will be examined in a future paper, but some typical stable assemblages are shown in Table 1.

Structure of the Pelagonian basement

Six phases of deformation have been recognized within the basement, five of which have associated minor structures. On the basis of style, overprinting, and structural truncation, these phases can be placed in a sequential order $D_1$to D_6.

Phase 1 structures

A pervasive, bedding-parallel, foliated, and lineated fabric together with rare isoclinal folds forms the earliest tectonic structures throughout the Olympos region (Barton 1976; Yarwood and Dixon 1979). In the Livadi area this L–S tectonite fabric is defined principally by the dimensional orientation of quartz, mica, and amphibole, and was developed under metamorphic conditions close to the greenschist–amphibolite facies boundary. The foliation wraps around K-feldspar megacrysts in the deformed granites and feldspathic clasts in the metasediments. Feldspars are commonly flattened within the planar fabric and drawn out in the direction of the lineation. Locally, along centimetre-wide shear zones that parallel the foliation, mylonite fabrics are developed. The attitude of the foliation is controlled by later folding and is sharply truncated by the thrust klippen of the Livadi complex (Fig. 3). The mineral lineation, which is broadly parallel or perpendicular to later fold axes, shows a northeast–southwest trend (Fig. 5*a*) and is believed to parallel the slip direction of the D_1 shear zones. Associated mesoscopic, tight to isoclinal folds are most commonly developed in rapidly alternating pelitic and psammitic metasediments. Fold profiles are "similar" in style with thickened hinge zones and limbs that are thinned or sheared out parallel to the foliation. The D_1 fold axes parallel the lineation and the foliation is axial planar such that discordances between foliation and bedding are visible in the fold hinges. Megascopic isoclinal folding is precluded by the consistent sense of younging within the metasediments.

Finite strain estimates from deformed feldspathic clasts yield strain ellipsoids whose X-axes and XY-planes parallel the lineation and foliation, respectively. Ellipsoid symmetries indicate that, while the dominant strain was one of flattening, the greatest amount of strain is located within the shear zones where deformation was constrictional. Since the overall symmetry of the ellipsoids appears to be largely unaffected by later strains, the development of the D_1 fabric requires the association of flattening deformation and shearing. Under such conditions contemporary fold axes, whatever their orientation, would become progressively rotated towards the direction of shear movement (Escher and Watterson 1974) while associated flattening, possibly resulting from tectonic stacking produced during shearing, would modify the folds and give rise to the development of a variable L–S fabric in which isoclinal folds largely parallel the slip direction recorded in the "stretching" lineation.

Phase 2 structures

The most abundant mesoscopic folds are open to tight, often asymmetric structures with a strong buckling component. They represent the second phase of folding and deform the D_1 fabric, which is occasionally crenulated within pelitic layers to form a weak, axial planar cleavage. Although scattered by later folding, axes trend broadly northeast–southwest (Fig. 5*b*) parallel to those of the D_1 folds such that interference produces refolded isoclines of type III (Ramsay 1967) pattern. Axial surfaces generally dip to the northwest or southeast at moderate to steep angles. Second folds of similar orientation occur in the Pieria allochthon (Yarwood and Dixon 1979). On a regional scale (Fig. 3) large structures of this generation can be inferred from variations in the attitude of the D_1 foliation. These megascopic D_2 folds are essentially upright open structures, which plunge gently northeast or southwest and have axial traces in excess of 10 km long.

Phase 3 structures

The third phase of folding is not well represented mesoscopically but is important on a megascopic scale and may correspond to Yarwood and Dixon's (1979) third fold episode in the Pieria allochthon. Mesoscopic folds form large class 1C structures (Ramsay 1967) in competent beds and angular, centimetre-scale, harmonic structures in more pelitic lithologies. An incipient, axial planar crenulation cleavage may be locally developed. Fold axes trend northwest–southeast (Fig. 5*c*) and are broadly perpendicular to those of D_1 and D_2

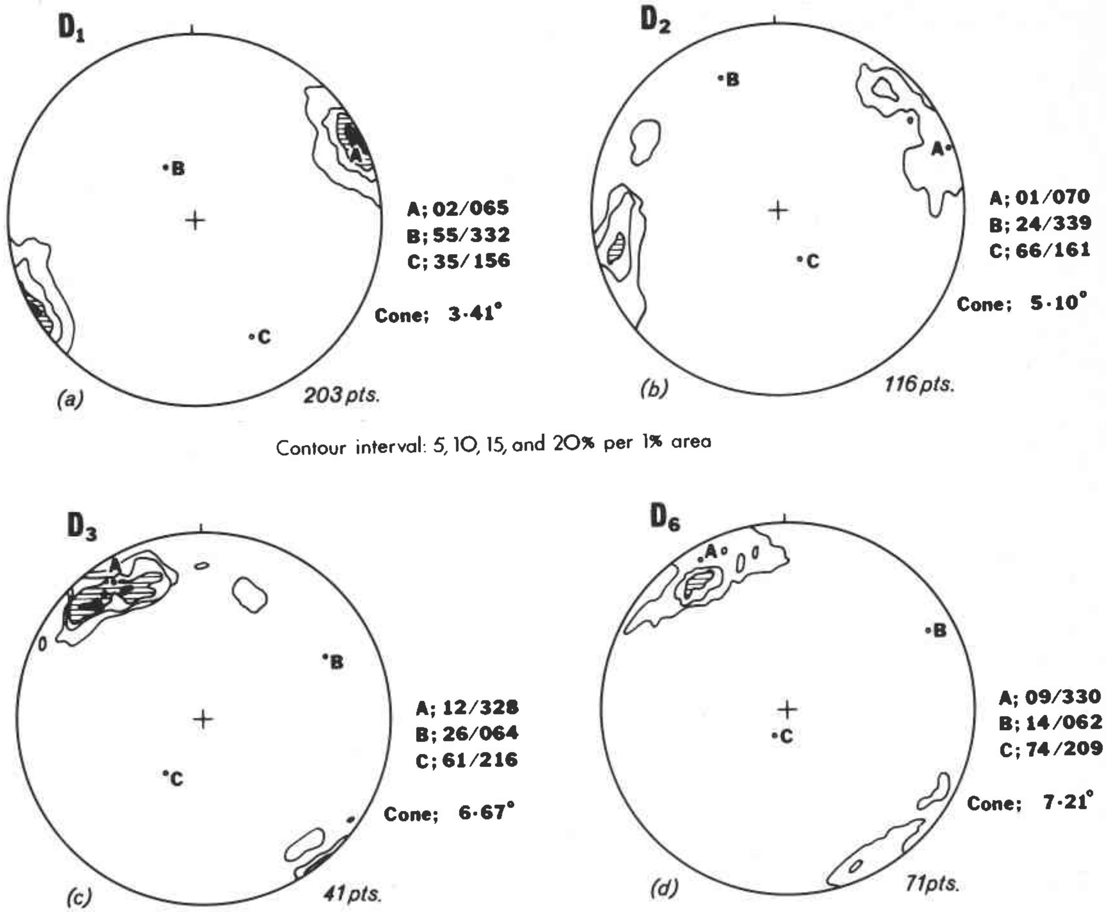

FIG. 5. Equal-area, lower hemisphere projections of minor structures in the Pelagonian basement. (*a*) D_1 mineral lineations. (*b*) D_2 fold axes. (*c*) D_3 fold axes. (*d*) D_6 crenulation axes. *A*, *B*, and *C* are the statistical maximum, intermediate, and minimum axes of the distribution (Watson 1966). Cone of confidence for the mean assumes a spherical (point maxima) distribution (Fisher 1953).

such that interference produces type I and type II (Ramsay 1967) patterns, respectively. Axial planes dip predominantly eastwards at steep angles. The upright, megascopic D_3 structures are similar in size to those of D_2 but plunge gently to northwest and southeast. Interference of these two phases consequently produces the type I domes and basins that characterize the regional structure (Fig. 3). The later nature of the D_3 folds is evidenced in the deflection of the D_2 axial traces. The regional deformation of the mesoscopic D_2 fold axes and, to a lesser extent, the D_1 lineation is such that they trace out complex loci whose form is characteristic of a D_3 fold mechanism involving both buckling and shear components.

Phase 4 structures

The megascopic interference patterns of D_2 and D_3 are truncated to the southeast (Fig. 3) by a series of thrust faults, which farther east cause the Pelagonian metamorphics to override the Eocene flysch of the Olympos series. A retrograde lower greenschist metamorphism is associated with this event while the orientations of linear structures, asymmetric folds, thrust surfaces, and high-angle reverse faults indicate transport towards the northeast (Barton 1975). The emplacement of the Livadi complex (see below) is likely to be contemporaneous with this event since it has a similar vector and associated metamorphism, and likewise truncates the megascopic D_3 folds. Basement structures related to this emplacement, however, have not been observed in either the Livadi area or further north (Yarwood and Dixon 1979).

Phase 5 structures

A fifth period of deformation occurs along the northern margin of the map area where the Livadi complex and its metamorphic basement are tectonically overridden by the Pieria allochthon (Fig. 4). However, structures associated with this deformation are chiefly recorded within the most northerly klippen of the Livadi complex and are described in the following section.

Phase 6 structures

The final deformation of the Pelagonian basement takes the form of kink banding and crenulation of all susceptible lithologies. Crenulation axes trend broadly northwest–southeast (Fig. 5*d*) and are preferentially developed in fine-grained amphibolitic and micaceous schists. Conjugate sets trending north-northwest and west-northwest are present at some locations. Identically oriented crenulations are present in the foliated metagabbros at the base of several klippen of the Livadi complex and in the Pelagonian metamorphics of the Pieria allochthon (Yarwood and Dixon 1979).

Structure of the Livadi complex

Deformational structures within the Livadi complex can be most readily divided into broad groups that predate, are associated with, or postdate the emplacement of the complex into its present position.

Pre-emplacement structures

In addition to exposing progressively more differentiated lithologies to the north and northeast, individual klippen of the Livadi complex display both normal and inverted stratigraphies. Thus in the southernmost Livadion klippe (Fig. 3), metadunite is stratigraphically overlain by metaharzburgite in a normal succession whereas in the more complete sequences exhibited by the Mavroneri, Kaithari, and Limnia klippen, metagabbro is stratigraphically overlain by pyroxene-bearing ultramafics and metadunite in a reversed succession. This stratigraphic inversion implies the presence of a major overturned structure within the complex, which is compatible with the compositional variation of its metamorphic olivine.

Structural control of olivine compositions

Metadunite throughout the complex shows a progressive compositional variation in its constituent olivines. In the Livadion klippe (Fig. 6), the percentage forsterite, which remains homogeneous on a mesoscopic scale, systematically decreases with increasing structural height and, over a vertical thickness of 150 m, exhibits the range Fo_{97} to Fo_{89}. In the Limnia klippe (inset Fig. 6) where the stratigraphy is inverted, this pattern is reversed.

As the Livadi complex metadunites show constant whole-rock Mg/(Mg + Fe) (total iron as Fe^{2+}) ratios (0.91 ± 1), the olivine compositional trend is clearly not primary. However, such ranges are not atypical of olivine produced by the metamorphic dehydration of serpentine (Evans 1977), their more magnesian composition relative to primary ultramafic olivines being attributed to the oxidation of some FeO to magnetite during serpentinization. The composition of such olivines consequently depends on the partitioning of iron between serpentine and magnetite, which for dunites of constant Mg/Fe ratio will be dependent upon the effective oxygen fugacity of serpentinization. In the Livadi complex the metamorphic paragenesis of progressively more magnesian olivine with increasing fO_2 during the preceding or simultaneous serpentinization is consistent with their anomalous Ni contents and the modal distribution of magnetite. Nickel in the metamorphic olivines increases from 0.15 to 0.45% with decreasing forsterite values. This is the reverse of normal magmatic trends (Irvine 1975) but would be expected in rocks that were formerly serpentinized since, in sulfur-poor serpentinites, nickel will follow iron (Bliss and MacLean 1975). Hoffman and Walker (1978) also attribute the production of an olivine compositional trend in the East Dover ultramafic bodies of southern Vermont, which is closely analogous to that of the Livadion klippe, to the presence of a substantial oxygen fugacity gradient.

For the Livadion klippe where the stratigraphy is normal, the olivine compositional trend requires a progressive decrease in the effective fluid fO_2 with increasing structural height. This might be expected for serpentinization and metamorphic recrystallization accompanying tectonic transport where the removal of oxygen through reactions at the base of the sheet would lower the effective fO_2 of the fluid as it migrated upwards. This effect would be enhanced by an increase in temperature, fluid–rock ratio, or fluid movement rate toward the then active thrust surface.

In the stratigraphically inverted Limnia klippe, however, the olivine compositional trend is reversed such that percent forsterite and hence the fluid fO_2 increase with structural height (Fo_{89}–Fo_{95}). This reversal is explained if the klippe represents part of the inverted limb of a major overturned structure (Fig. 7) produced after the effects of the fO_2 gradient had been established, and is therefore compatible with the lithological evidence. Preservation of both limbs at the western margin of the Livadion klippe (Fig. 6) may be indicated by the occurrence of low forsterite values adjacent to the basal thrust.

The attitudes of lithological boundaries and the distribution of olivine compositions suggest a broadly flat-lying, isoclinal form to this structure, while the rapidity with which the inverted limb thins out to the southwest supports its development as the overturned nose of an advancing mafic–ultramafic sheet. A direction of closure towards the north or northeast is indicated by the exposure of progressively more differentiated

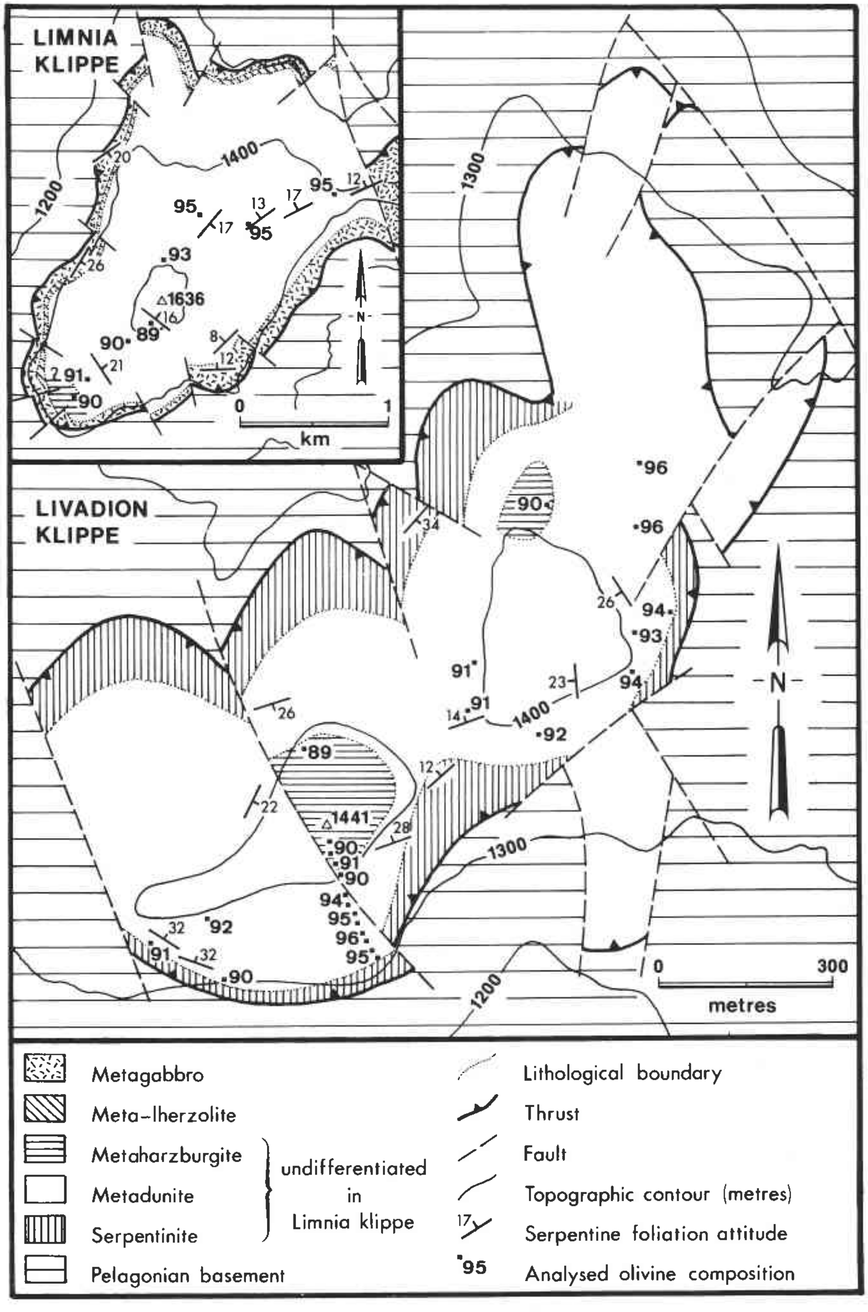

FIG. 6. Distribution of lithologies and olivine compositions in the Livadion and Limnia klippen.

units in this direction and the preferential preservation of parts of the upper limb in the more southwesterly klippen (Fig. 7).

Emplacement structures (D_4)

Subsequent to its regional metamorphism, the Livadi complex was emplaced into its present position along a major tectonic surface now represented by the basal thrusts of the individual klippen. The lower greenschist retrogression that accompanied this event was associated with cataclastic deformation and a second serpentinization. In the metadunites, serpentinization of metamorphic olivine along the thrust surface and shear zones subparallel to it produced antigorite, which

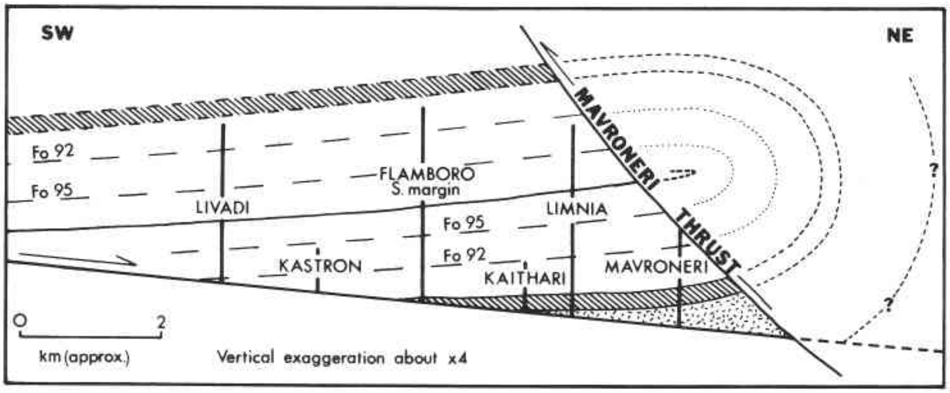

FIG. 7. Highly schematic reconstruction of the overturned structure within the Livadi complex showing approximate sections represented by individual klippen (legend as for Fig. 3).

defines a strong, thrust-parallel, planar fabric. The accommodation of shear stress by this serpentinite would account for the absence of emplacement-related structures in the underlying Pelagonian basement. A thrust-parallel fabric is also developed in the metagabbros where these lie close to the thrust surface. It is defined by the planar orientation of chlorite replacing amphibole. The meta-lherzolites, which preserve relict pyroxene, remain largely undeformed.

Emplacement of the Livadi complex is also associated with the development of linear structures. In the ultramafics, slickensides and oriented antigorite fibres reflect shearing subparallel to the basal thrust surface and lie within the thrust-parallel planar fabric. With the exception of the Mavroneri klippe where overprinting has occurred during the emplacement of the Pieria allochthon, these show a consistent northeast–southwest trend (Fig. 8 *a–c*). In addition, a mineral lineation is contained within the thrust-parallel fabric of the metagabbros. It is defined by the orientation of low-alumina tremolite and chlorite, both of which replace amphibole. This gently plunging "stretching" lineation shows a consistent northeast– southwest trend (Fig. 8*d*) and is considered to parallel the emplacement direction.

A northeast-directed vector for the final emplacement of the Livadi complex can be obtained from regional considerations since the continuous Triassic to Eocene stratigraphy of Olympos (Godfriaux 1968) and the absence of southwest-facing structures in the area studied by Barton (1975) preclude the possibility of emplacement from the northeast.

Postemplacement structures (D_5)

Along the Mavroneri valley, Pelagonian metamorphics of the Pieria allochthon override the rocks of the map area along a series of northerly dipping thrusts (Fig. 4). Tectonic repetition of the Mavroneri klippe indicates that thrusting was later than the emplacement of the Livadi complex, while structures associated with the event are consistent with emplacement from the north. The gently basinal form of the D_4 serpentinite fabric in the Livadion and Limnia klippen (Fig. 6) may also have developed at this time.

Propagation of the Pieria allochthon paralleled the D_1 fabric of the Pelagonian basement and produced a crude, thrust-parallel, D_5 fabric in the Mavroneri serpentinites. Both fabrics are deformed about small-scale asymmetric folds whose sense of rotation, according to the slip vector methods of Hansen (1966) and Scott and Hansen (1968), indicate emplacement from the north parallel to the dip of the thrust surface (Fig. 9*a*). A compatible slip vector derived from equivalent folds developed beneath serpentinites of the Mavroneri klippe at Fteri (Fig. 9*b*) implies oblique motion along its basal thrust surface. This would be expected if emplacement of the Pieria allochthon remobilized a pre-existing tectonic contact that was oriented at an angle to the slip vector. The consistent north–south trend of slickensides and oriented antigorite fibres lying within the thrust-parallel fabric (Fig. 9*c*) is compatible with this slip vector and is considered to reflect an overprinting of the northeast–southwest trend seen elsewhere in the Livadi complex by movement accompanying emplacement of the Pieria allochthon.

Further support for north to south movement can be derived from the northern margin of the Flamboro klippe where the most southerly of a series of subsidiary thrusts associated with the emplacement of the Pieria allochthon has caused a meta-lherzolite sheet to override and imbricate the lithologies of the Flamboro klippe. Further north, the meta-lherzolite sheet is itself overridden by the Pelagonian basement (Fig. 10). The angular relationship between the gently north-dipping meta-

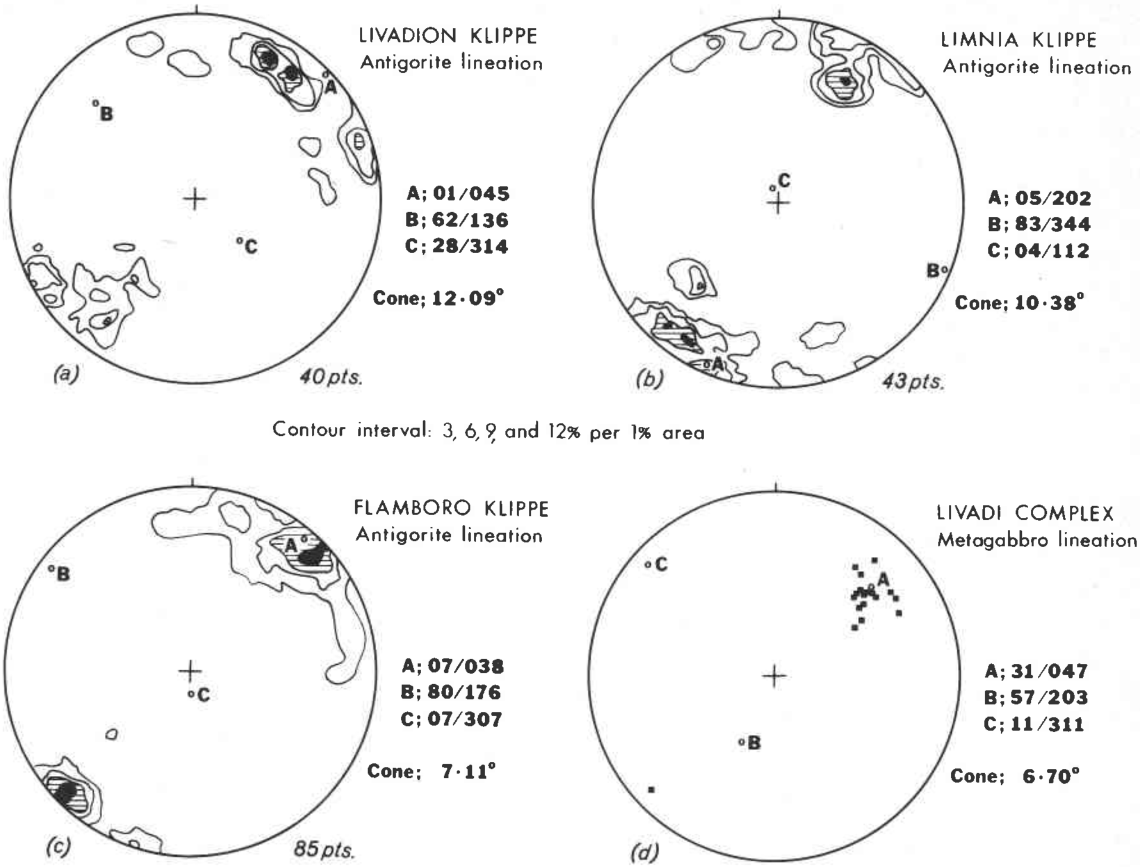

FIG. 8. Stereographic projections of minor structures associated with the final emplacement of the Livadi complex. (*a*), (*b*), and (*c*) Orientation of antigorite fibres in serpentinite. (*d*) Mineral lineations in metagabbro.

lherzolite sheet and the steeply north-dipping imbricate slices developed beneath it is most simply explained by emplacement from the north.

Timing of the deformational phases

The assignment of absolute ages to deformational events within the Livadi area remains tentative as it involves structural correlation across major tectonic contacts. However, a possible interpretation of available data is given in Table 2.

The first deformation (D_1) and associated upper greenschist to lower amphibolite facies metamorphism of the Pelagonian basement represent a period of flattening and northeast–southwest shearing that produced foliation-parallel mylonite zones. That this is an Early Cretaceous event is strongly implied by Rb/Sr determinations on identical mylonites to the east (127 ± 3 Ma, Barton 1976) and on foliated Kataphygion granite to the north (122 ± 3 Ma, Yarwood and Dixon 1979). It is further supported by the presence of the D_1 fabric in marbles of probable Triassic to Jurassic age (Godfriaux 1968) to the southeast and the occurrence of similar mylonites to the north that are discordantly overlain by undeformed Upper Cretaceous limestone (Mercier 1973). Furthermore, the D_1 event invites comparison with the earliest recognizable deformation of the Livadi complex where northeast-directed emplacement was associated with the development of a major overturned structure and regional metamorphism close to the greenschist–amphibolite facies boundary. Such a comparison would imply an Early Cretaceous age for the metamorphism of the complex and would provide a minimum estimate for the age of the overturned structure.

The megascopic D_2 and D_3 structures clearly postdate the D_1 metamorphism, but are truncated by thrust surfaces relating to the emplacement of the Pelagonian allochthon over Olympos. Phyllonites produced during this overthrusting have been dated as Late Eocene (40 ± 1 Ma) on the basis of Rb/Sr determinations (Barton 1976). Consequently the development of both D_2 and D_3 is restricted to the Cretaceous and early Tertiary.

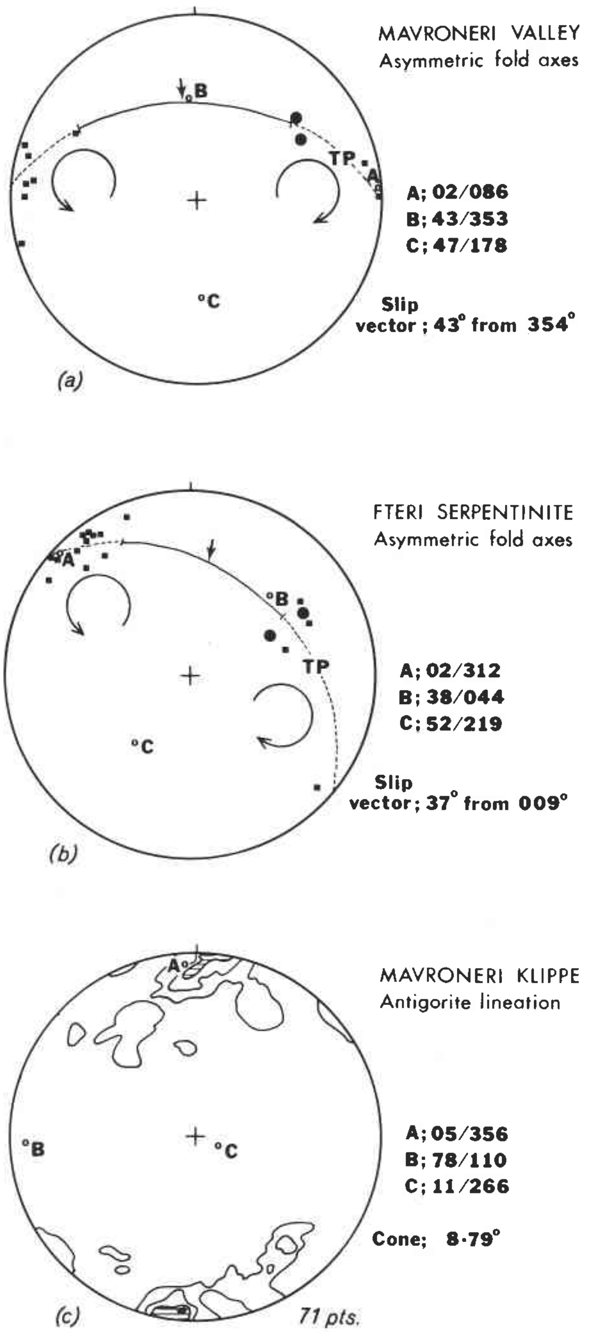

FIG. 9. Stereographic projections of D_5 minor structures associated with the emplacement of the Pieria allochthon. (*a*), (*b*) Hansen plots of asymmetric fold axes and their senses of rotation. The partial great circle defined by the fold axes closely approximates the orientation of the thrust plane (TP) with which they are associated. Vector of movement (small arrow) is from the north. Large symbols are megascopic folds. (*c*) Orientation of antigorite fibres in serpentinites of the Mavroneri klippe.

Yarwood and Dixon (1979) have obtained an Rb/Sr age of 103 ± 2 Ma from an augen schist unit near the southern margin of the Pieria allochthon, which contains a weak second cleavage. Although they are tempted to allocate this age to the emplacement of the Pieria allochthon, they note that it could also correspond to the second fold episode of the allochton, which strongly correlates both in style and orientation with the D_2 event of the Livadi area. This latter explanation seems more plausible in the light of the structural sequence at Livadi where the emplacement of the Pieria allochthon is clearly very late.

Since D_1 to D_3 structures in the underlying basement are truncated by faults along which the Livadi complex was finally transported to its present position, this emplacement event must be of post-Early Cretaceous age. It is most reasonably correlated with the early Tertiary (D_4) emplacement of the Pelagonian allochthon over Olympos (Barton 1976) as both events show lower greenschist retrogression accompanying northeasterly transport. The absence within the complex of megascopic structures analogous to those of D_2 and D_3 presumably places some minimum constraint on the amount of final movement. However, this constraint may be small as such structures have not been observed in the areas studied by Barton and Yarwood.

Based on the above correlation, the later southward movement of the Pieria allochthon would be of post-Late Eocene age. Undeformed Neogene and Quaternary sediments unconformably overlie a possible westward extension of the Mavroneri thrust belt north of Servia (Godfriaux 1968). The amount of displacement represented by this emplacement is unknown although the tectonic stacking of the Mavroneri and Flamboro klippen requires shortening of several kilometres. Although the nature of this event remains unresolved, it may be of regional significance as similar southerly directed thrusting affects the Late Paleocene or younger Dhivri Formation of eastern Othris (Welland 1972). Emplacement of the Pieria allochthon was followed by the development of the D_6 crenulations and kink bands.

Summary and discussion

The principal deformations of both the Livadi complex and its metamorphic basement prior to the emplacement of the Pieria allochthon reflect Early Cretaceous and possibly early Tertiary phases of southwest to northeast movement. Both units are consequently rooted to the southwest and, in the absence of a phase of southwest-directed motion, must always have lain west of Olympos. This evidence is incompatible with the single-ocean tectonic model for the Hellenides (Dewey *et al.* 1973; Mercier *et al.* 1975; Zimmerman and Ross 1976), which involves southwesterly transport

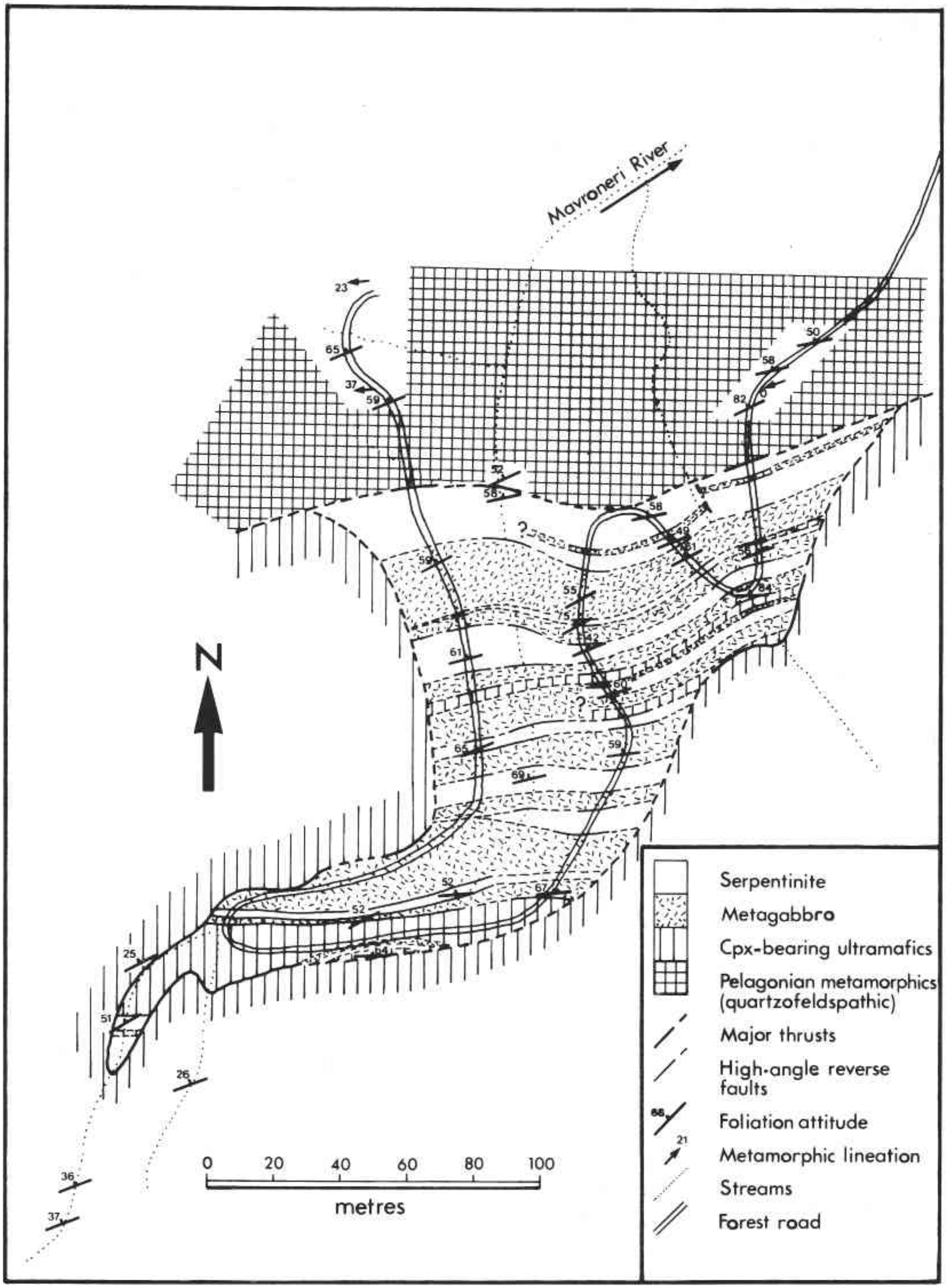

FIG. 10. Original geological map of an imbricated segment of the northern margin of the Flamboro klippe, southwest of Fteri.

TABLE 2. Inferred tectonic correlations in the Olympos region

Event		Vector	Metamorphism	Possible age
Pelagonian basement	Livadi complex			
D_6: NW–SE crenulation and kink banding	NW–SE crenulation and kink banding			
D_5: Emplacement of the Pieria allochthon; asymmetric folding	Emplacement of the Pieria allochthon; asymmetric folding, N–S serpentinite lineation, imbrication	N→S	Serpentinization	?Tertiary
D_4: Final emplacement of the Livadi complex; emplacement of the Pelagonian allochthon	Final emplacement of the Livadi complex; NE–SW serpentinite and gabbro lineations	SW→NE	Lower greenschist + serpentinization	?early Tertiary (40 ± Ma Rb/Sr on Olympos phyllonites; Barton 1976)
D_3: NW–SE folding				
D_2: NE–SW folding				?mid-Cretaceous (103 ± 2 Ma Rb/Sr on Lower Mavroneri augen schists; Yarwood and Dixon 1979)
D_1: Bedding-parallel foliation and NE–SW mineral lineation; NE–SW isoclinal folds; mylonite zones	Initial emplacement of the Livadi complex; development of overturned nose of the advancing thrust sheet	SW→NE	Upper greenschist – lower amphibolite + serpentinization	Early Cretaceous (127 ± 3 Ma Rb/Sr on Olympos mylonites; Barton 1976); (122 ± 3 Ma Rb/Sr on foliated Kataphygion granite; Yarwood and Dixon 1979)
	Crystallization of cumulate sequence			?Triassic–Jurassic
Intrusion of Livadi granite and associated igneous activity				?Late Carboniferous (302 ± 5 Ma U/Pb on Kataphygion granite; Yarwood and Aftalion 1976)
Clastic sedimentation				Paleozoic

across Olympos. Repeated periods of northeast-directed motion, however, are implied by the two-ocean model of Hynes *et al.* (1972) and Smith *et al.* (1975), in which the existance of an Othris ocean lying west of the Pelagonian zone is proposed. Subduction and the tectonic stacking of the eastern margin of this ocean, which are considered to have occurred during Late Jurassic to Early Cretaceous time, correspond to the earlier D_1 thermal event of the Livadi complex and its Pelagonian basement. An early Tertiary age of final closure of the Othris ocean is compatible with what is known of the retrograde thermal event at Livadi, during which the Pelagonian zone moved northeast over the Olympos platform. Thus the Mesozoic Othris ocean might provide a root zone for the Livadi complex since its primary chemistry and mineralogy (discussed in a future paper) are closely analogous to the cumulate sequence of ophiolite bodies, and the Paleozoic age assigned to it on the basis that it formed part of the Pelagonian metamorphic sequence (Godfriaux 1968) can no longer be accepted. However, the majority of the Mesozoic ophiolite bodies that are presently located along the boundary of the Othris and Pelagonian zones and that are considered to have been emplaced from the Othris ocean (Smith and Moores 1974) have not suffered the regional metamorphism exhibited by the Livadi complex. This might be explained if the complex represents an early emplaced fragment of Othris ocean floor, the tectonic overburden of which was increased by stacking of the continental margin and further emplacement of ophiolites. Such a model is supported by the present position of the mineralogically and chemically similar Tranovolto serpentinite 25 km to the southwest. This antigorite serpentinite rests tectonically on Pelagonian basement and is overthrust in turn by continental margin platform carbonates and the Vourinos ophiolite. It may currently occupy a position similar to that of the Livadi complex during its regional metamorphism. Major horizontal movement during the Early Cretaceous and (or) early Tertiary could have resulted in the tectonic separation of the complex from a continental margin sequence, bringing the complex into its present position. The minimum required movement rate is of the order of 5 mm/year. This corresponds to the average rates of tectonic transport of the Othris and Pelagonian zones relative to the Paxos zone during much of the Tertiary (Smith and Moores 1974) and lies well within the established rate of movement (15 cm/year) during the emplacement of the Pelagonian sheet over Olympos (Barton and England 1979).

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AUBOUIN, J. 1959. Contributions à l'étude géologiques de la Grèce septentrionale: les confins de l'Epire et de la Thessalie. Annales Géologiques des Pays Helléniques, **10**, pp. 1–525.

——— 1965. Geosynclines. Developments in Geotectonics 1. Elsevier, New York, NY. 335 p.

BARTON, C. M. 1975. Mount Olympos, Greece: New light on an old window. Journal of the Geological Society, London, **131**, pp. 389–396.

——— 1976. The tectonic vector and emplacement age of an allochthonous basement slice in the Olympos area, N.E. Greece. Bulletin de la Société Géologique de France, Series 7, **18**, pp. 253–258.

——— 1979. Structural evolution of the Vardar root zone, northern Greece: Discussion. Geological Society of America Bulletin, **90**, pp. 125–126.

BARTON, C. M., and ENGLAND, P. C. 1979. Shear heating at the Olympos (Greece) thrust and the deformation properties of carbonates at geological strain rates. Geological Society of America Bulletin, **90**, pp. 483–492.

BLISS, N. W., and MACLEAN, W. H. 1975. The paragenesis of zoned chromite from central Manitoba. Geochimica et Cosmochimica Acta, **39**, pp. 973–990.

CELET, P., and FERRIÈRE, J. 1978. Les Hellénides internes: Le Pélagonien. Eclogae Geologicae Helvetiae, **71**, pp. 467–495.

CADET, J-P., BONNEAU, M., CHARVERT, J., DÜRR, S., ELTER, P., FERRIÈRE, J., SCANDONE, P., and THIÉBAULT, F. 1980. Le Chaînes de la Méditerranée moyenne et orientale. *In* Geology of the Alpine chains born of Tethys. *Co-ordinated by* J. Aubouin, J. Debelmas, and M. Latreille. Memoire du Bureau de Recherches Géologique et Minières, No. 115, pp. 98–118.

DERYCKE, F., GODFRIAUX, I., and ROBASZYNSKI, F. 1974. Sur quelques paragenèses du métamorphisme de haute pression – basse température dans l'Ossa et la pourtour de la fenêtre de l'Olympe (Grèce). Compte Rendu de l'Academie des Sciences (Paris), **279**, pp. 227–230.

DEWEY, J. F., PITMAN, W. C., RYAN, W. B. F., and BONNIN, J. 1973. Plate tectonics and the evolution of the Alpine System. Geological Society of America Bulletin, **84**, pp. 3137–3180.

ESCHER, A., and WATTERSON, J. 1974. Stretching fabrics, folds and crustal shortening. Tectonophysics, **22**, pp. 223–231.

EVANS, B. W. 1977. Metamorphism of alpine peridotite and serpentinite. Annual Review of Earth and Planetary Sciences, **5**, pp. 397–447.

FERRIÈRE, J. 1974. Étude géologiques d'un secteur des zones helléniques internes subpélagonienne et pélagonienne (massif de l'Othrys, Grèce continentale). Importance et signification de la périod orogénique anté-Crétacé supérieur. Bulletin de la Société Géologique de France, Series 7, **16**, pp. 543–562.

——— 1976. Sur le signification des séries du massif de l'Othrys (Grèce continentale-orientale): la zone isopique maliaque. Annales, Société Géologique du Nord, **96**, pp. 121–134.

FERRIÈRE, J., and VERGELY, P. 1976. A propos des structures tectoniques et microtectoniques observées dans les nappes anté-Crétacé supérieur d'Othrys centrale (Grèce continentale): consequences. Compte Rendu de l'Academie des Sciences (Paris), **283**, pp. 1003–1006.

FISHER, R. A. 1953. Dispersion on a sphere. Proceedings of the Royal Society of London, Series A, **217**, pp. 295–305.

FLEURY, J. J., and GODFRIAUX, I. 1974. Arguments pour l'attribution de la série de la fenêtre de l'Olympe (Grèce) à la zone de Gavrovo–Tripolitza: Présence de fossiles du Maastrichtien et de l'Eocene inférieur (et moyen?). Annales, Société Géologique du Nord, **94**, pp. 149–156.

GODFRIAUX, I. 1968. Étude géologique de la région de l'Olympe. Annales Géologiques des Pays Helléniques, **19**, pp. 1–284.

HANSEN, E. 1966. Methods of deducing slip-line orientations from the geometry of folds. Annual Report, Director of the Geophysical Laboratory, Washington, DC, **65**, pp. 387–405.

HOFFMAN, M. A., and WALKER, D. 1978. Textural and chemical variations of olivine and chrome spinel in the East Dover ultramafic bodies, south-central Vermont. Geological Society of America Bulletin, **89**, pp. 699–710.

HYNES, A. J., NISBET, E. G., SMITH, A. G. WELLAND, M. J. P., and REX, D. C. 1972. Spreading and emplacement ages of some ophiolites in the Othris region (eastern central Greece). Zeitschrift der Deutschen Geologischen Gesellschaft, **123**, pp. 455–468.

IRVINE, T. N. 1975. Crystallization sequence in the Muskox intrusion and other layered intrusions—II. Origin of chromitite layers and similar deposits of other magmatic ores. Geochimica et Cosmochimica Acta, **39**, pp. 991–1020.

Mercier, J. 1973. Contributions à l'étude du métamorphism et de l'évolution magmatique des zones internes des Hellénides. Annales Géologiques des Pays Helléniques, **20**, pp. 597–792.

Mercier, J., Vergely, P., and Bebien, J. 1975. Les ophiolites helléniques "obductées" au Jurassique supérieur sont-elles les vestiges d'un océan téthysien ou d'une mer marginale péri-européene? Compte Rendu Sommaire des Seances de la Société Géologique de France, **17**, Fasc. 4, pp. 108–112.

McCaig, A. M., and Kemp, A. E. 1979. Emplacement history of the Pindos ophiolite, Greece. Geological Association of Canada/Mineralogical Association of Canada, Program with Abstracts, **4**, p. 66.

Price, I. 1976. Pre-Upper Cretaceous sedimentary facies of the Othris Mountains, Greece. Bulletin de la Société Géologique de France, Series 7, **18**, pp. 273–279.

——— 1977. Deposition and derivation of clastic carbonates on a Mesozoic continental margin, Othris, Greece. Sedimentology, **24**, pp. 529–546.

Ramsay, J. G. 1967. Folding and fracturing of rocks. McGraw-Hill, New York, NY. 568 p.

Scott, W. H., and Hansen, E. 1968. Movement directions and the axial plane fabrics of flexural folds. Annual Report, Director of the Geophysical Laboratory, Washington, DC, **67**, pp. 254–258.

Smith, A. G., and Moores, E. M. 1974. The Hellenides. *In* Data for orogenic studies: Mesozoic–Cenozoic orogenic belts. *Edited by* A. M. Spencer. Geological Society of London, Special Publication No. 4, pp. 154–185.

Smith, A. G., and Woodcock, N. H. 1976*a*. The earliest Mesozoic structures in the Othris region, eastern central Greece. Bulletin de la Société Géologique de France, Series 7, **18**, pp. 245–251.

——— 1976*b*. Emplacement model for some "Tethyan" ophiolites. Geology, **4**, pp. 653–656.

Smith, A. G., Hynes, A. J., Menzies, M. A., Nisbet, E. G., Price, I., Welland, M. J. P., and Ferrière, J. 1975. The stratigraphy of the Othris Mountains, eastern central Greece: a deformed Mesozoic continental margin sequence. Eclogae Geologicae Helvetiae, **68**, pp. 463–481.

Smith, A. G., Woodcock, N. H., and Naylor, M. A. 1979. The structural evolution of a Mesozoic continental margin, Othris Mountains, Greece. Journal of the Geological Society, London, **136**, pp. 589–603.

Temple, P. G. 1968. Mechanics of large-scale gravity sliding in the Greek Peloponnesos. Geological Society of America Bulletin, **79**, pp. 687–700.

Vergely, P. 1976. Chevauchement vers l'Ouest et rétrocarriage vers l'Est des ophiolites: deux phases tectonique au cours du Jurassique supérieur – Eocrétacé dans le Hellénides internes. Bulletin de la Société Géologique de France, Series 7, **18**, pp. 223–244.

Watson, G. S. 1966. The statistics of orientation data. Journal of Geology, **74**, pp. 786–797.

Welland, M. J. P. 1972. The stratigraphy and structure of part of the eastern Othris Mountains, Greece. Ph.D. thesis, University of Cambridge, Cambridge, England.

Yarwood, G. A., and Aftalion, M. 1976. Field relations and U–Pb geochronology of a granite from the Pelagonian zone of the Hellenides (High Pieria, Greece). Bulletin de la Société Géologique de France, Series 7, **18**, pp. 259–264.

Yarwood, G. A., and Dixon, J. E. 1979. Lower Cretaceous and younger thrusting in the Pelagonian rocks of the High Pieria, Greece. VI[e] Colloquium Géologique de Regions Égéenes (Athens 1977), pp. 269–280.

Zachos, K., and Maratos, G. 1965. Metallogenic map of Greece. Institute for Geology and Subsurface Research, Athens, Greece.

Zimmerman, J., and Ross, J. V. 1976. Structural evolution of the Vardar root zone, northern Greece. Geological Society of America Bulletin, **87**, pp. 1547–1550.

——— 1979. Structural evolution of the Vardar root zone, northern Greece: Reply. Geological Society of America Bulletin, **90**, pp. 126–127.