Geol. Mag. **143** (*3*), *2006, pp. 329–342.* © *2006 Cambridge University Press*
doi:10.1017/S001675680600197X *Printed in the United Kingdom*

On the provenance of mid-Cretaceous turbidites of the Pindos zone (Greece): implications from heavy mineral distribution, detrital zircon ages and chrome spinel chemistry

P. FAUPL*†, A. PAVLOPOULOS‡, U. KLÖTZLI* & K. PETRAKAKIS*

*Department of Geological Sciences, University of Vienna, Althanstrasse 14, A-1090 Vienna, Austria
‡Department of Sciences, Laboratory of Mineralogy-Geology, Agricultural University of Athens, Iera Odos 75, GR-11855 Athens, Greece

(*Received* 24 March 2005; *accepted* 11 October 2005)

Abstract – Two heavy mineral populations characterize the siliciclastic material of the mid-Cretaceous turbidites of the Katafito Formation ('First Flysch') of the Pindos zone: a stable, zircon-rich group and an ophiolite-derived, chrome spinel-rich one. U/Pb and Pb/Pb dating on magmatic zircons from the stable heavy mineral group clearly illustrate the existence of Variscan magmatic complexes in the source terrain, but also provide evidence for magmatism as old as Precambrian. Based on microprobe analyses, the chrome spinel detritus was predominantly supplied from peridotites of mid-ocean ridge as well as suprasubduction zone origin. A small volcanic spinel population was mainly derived from MORB and back-arc basin basalts. The lithological variability of the mid-Cretaceous ophiolite bodies, based on spinel chemistry, is much broader than that of ophiolite complexes presently exposed in the Hellenides. The chrome spinel detritus compares closely with that from the Outer and Inner Dinarides. The source terrain of the ophiolite-derived heavy minerals was situated in a more internal palaeogeographic position than that of the Pindos zone. The zircon-rich heavy mineral group could have had either an external and/or an internal source, but the chrome spinel constantly accompanying the stable mineral detritus seems to be more indicative of an internal source terrain.

Keywords: Hellenides, Greece, flysch, heavy minerals, zircon ages, chrome spinel.

1. Introduction

The Cenomanianto Lower Santonian pelagic succession of the Pindos zone of the Hellenides (Fig. 1) contains considerable amounts of siliciclastic and mixed siliciclastic/carbonate turbidites, termed 'Premier Flysch de Pinde' by Aubouin (1957, 1959). The first thin turbidite interlayers appear within the underlying radiolarites in the Early Cretaceous. In some locations, these turbidites have a high content of ophiolitic detritus. Mineralogical and geochemical investigations on the clay fraction of the Pindos zone by Thiebault *et al*. (1994) demonstrate that this detrital ophiolite influence had already commenced in the Tithonian.

The provenance of the terrigenous material of the mid-Cretaceous turbidites of the Pindos zone is a matter of debate. Palaeocurrent indicators as well as turbidite facies distribution support the model of an external source terrain (Piper & Pe-Piper, 1980; Neumann, 2003), which was active from mid-Cretaceous times until the Palaeogene sedimentation of the Pindos flysch (Piper, 2004). However, the massive occurrence of ophiolite detritus in some places rather points to an internal source. The spatial distribution of heavy minerals from the mid-Cretaceous terrigenous material of the Pindos zone of both the Greek mainland (Faupl, Pavlopoulos & Migiros, 1998) and the Peloponnese seems to support the assumption of an external and an internal source. The minerals of ophiolitic derivation, such as chrome spinel, can be readily supplied from an internal zone, whereas the stable heavy mineral association, comprising zircon, tourmaline, rutile and apatite, defines a second source terrain, which could be situated in either the external or the internal part of the Hellenides. In the case of derivation from a source terrain situated externally to the Pindos zone, a palaeogeographical problem arises, because of the existence of the shallow-water carbonate platform of the Gavrovo–Tripolitza zone external (west at the present time) to the Pindos zone in mid-Cretaceous times. Additionally, the basinal development of the Ionian zone bordered the carbonate platform towards the foreland (Katsikatsos, 1992). Thus, if sediment was supplied from an external source terrain, this must have comprised an exposed crystalline basement high within the realm of the Gavrovo–Tripolitza platform or along the margin of the Gavrovo–Tripolitza zone against the Pindos zone.

The purpose of this contribution is to provide information on the provenance of the mid-Cretaceous terrigenous detritus of the Pindos zone on the basis of spatial distribution of heavy minerals, radiometric ages of detrital zircons and geochemical data from

†Author for correspondence: peter.faupl@univie.ac.at

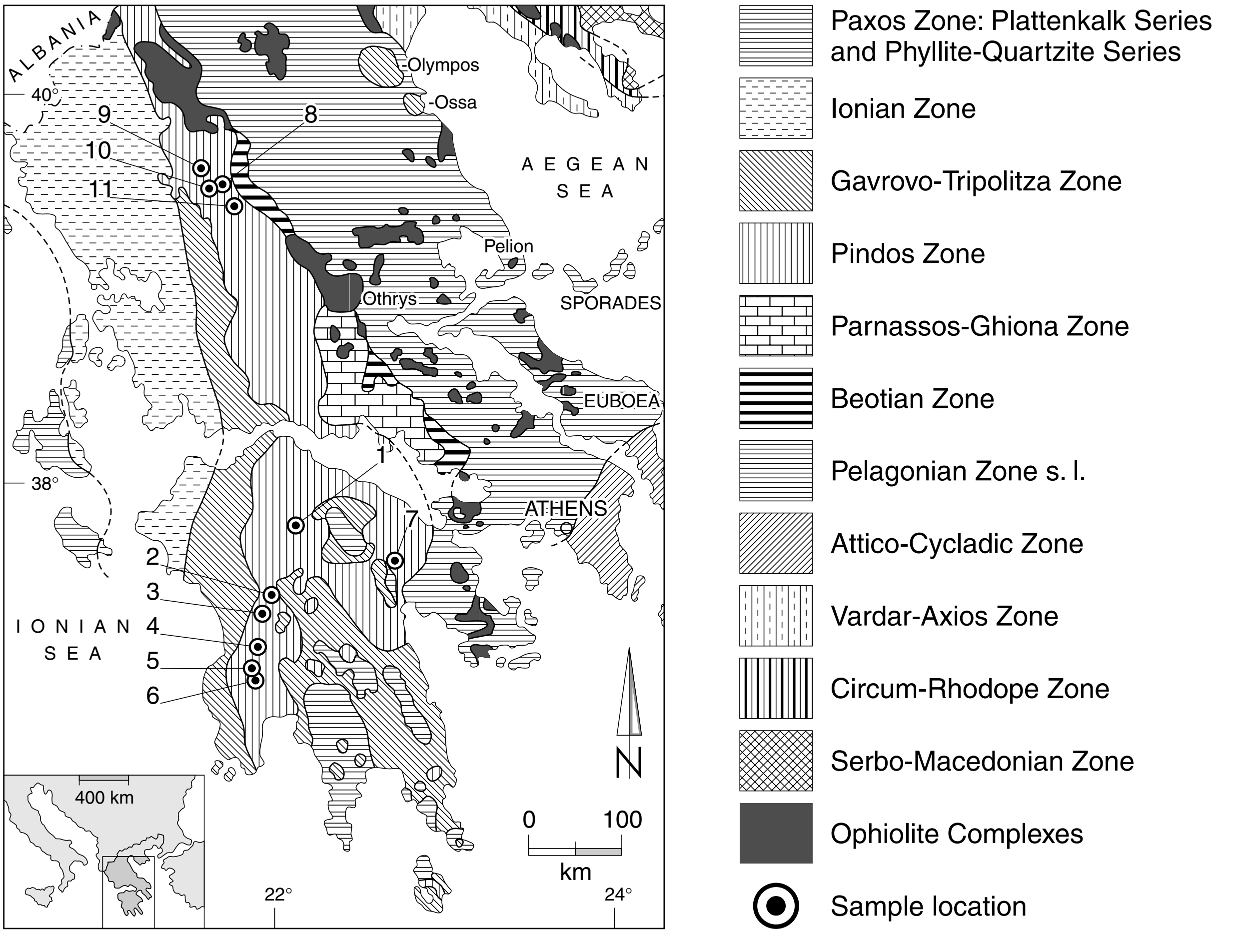

Figure 1. Map of the geotectonic zones of the Hellenides after Jacobshagen (1986). Numbers of sample locations of the mid-Cretaceous turbidites of the Pindos zone refer to Table 1.

detrital chrome spinels. The latter may be compared with spinels from recently exposed ophiolite complexes of the Hellenides (e.g. Pindos, Vourinos, Vermion, Othrys, Euboea: see Gartzos, Migiros & Parcharidis, 1990). The ages of the detrital zircons from the mid-Cretaceous turbidites are considered in relation to ideas of the position of the Palaeotethys suture within the Hellenides, which formed the boundary between the Variscan domain and the Gondwana margin without any Variscan deformation. The position of this suture is under debate (e.g. Robertson & Karamata, 1994; Pe-Piper, 1998; Champod, Stampfli & Kock, 2004).

2. Geological setting

The Pindos zone, which has a width of 160 km as a part of the External Hellenides *sensu* Brunn (1956), represents a characteristic fold-and-thrust belt (Skourlis & Doutsos, 2003) having a long border with the Pelagonian zone *s.l.* (= unmetamorphosed Subpelagonian and metamorphic Pelagonian zone), a tectono-stratigraphic unit of the Internal Hellenides, to the west (Fig. 1). In the southern part of the Greek mainland, the Parnassos–Ghiona zone, representing a carbonate platform, is intercalated between the Pindos zone and Pelagonian zone *s.l.* Palaeogeographically, the Pindos zone, together with the Paxos, Ionian and Gavrovo–Tripolitza zones, forms the margin of the Apulian plate, which was separated by the Pindos Ocean from the internally situated Pelagonian microplate. The existence of oceanic lithosphere within the Pindos zone has been intensively discussed (Smith, Woodcock & Naylor, 1979; Jones & Robertson, 1991; Robertson & Karamata, 1994). Ophiolite complexes, such as the Pindos, the Othrys and Vourinos Ophiolite, are interpreted to be oceanic remnants of this Pindos Ocean domain (Smith, 1993). The Alpine fold-and-thrust belt of the Pindos zone extending from Albania to Crete developed during the Palaeogene orogenic stage because of the collision of the Pelagonian microplate with the margin of the Apulian plate (Mountrakis, 1986; Doutsos *et al.* 1993) and the contemporaneous closure of the Pindos Ocean. In addition to the Pindos zone, the Vardar zone is related to a further oceanic domain which was juxtaposed to the internal margin of the Pelagonian zone (Bernoulli & Laubscher, 1972; Baumgartner, 1985). Both oceanic realms are characterized by a complex history of

https://doi.org/10.1017/S001675680600197X Published online by Cambridge University Press

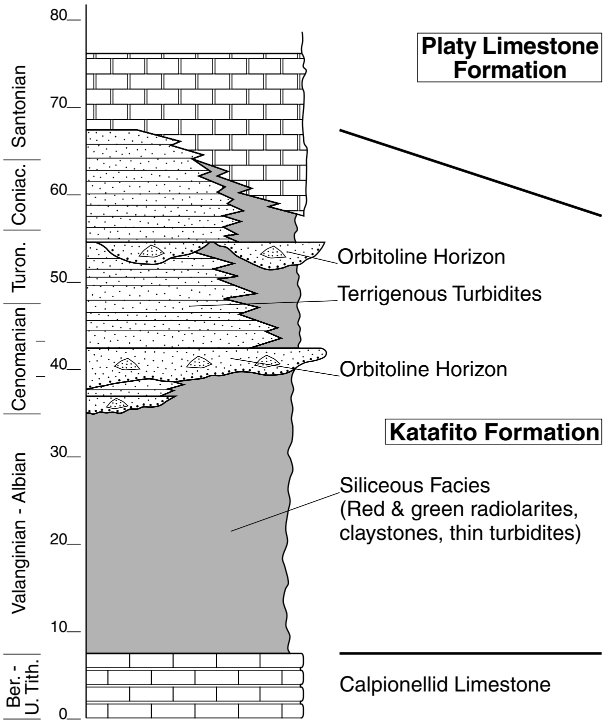

Figure 2. Schematic lithostratigraphic chart of the Katafito Formation from the Pindos zone, based on Neumann (2003).

closure from Jurassic until Palaeogene times. Their possible palaeogeographic position within the Tethys domain is shown by Stampfli *et al.* (2001) and Golonka (2004), among others.

The deep-water succession of the Pindos zone, which ranges from Late Triassic to Eocene/Oligocene, accumulated on the passive margin of the Apulian plate (Robertson *et al.* 1991). The whole sedimentary sequence is believed to be relatively thin (< 1000 m: Degnan & Robertson, 1998) compared with the adjacent carbonate platform of the Gavrovo–Tripolitza zone (> 2000 m: Jacobshagen, 1986). In Jurassic and Early Cretaceous times, radiolarites and cherty limestones accumulated; in the Tithonian and early Neocomian, calpionellid limestones appeared. During Cretaceous times the Pindos zone exhibited a remnant ocean basin with a very low sedimentation rate, especially in the very Early Cretaceous (Neumann & Zacher, 2004). From the Early Cretaceous to the early Santonian, cherty/calcareous sediments were deposited. From the Cenomanian on, these sediments contain rising amounts of siliciclastic and calci-/siliciclastic turbidites, termed ‘Katafito Formation’ by Neumann (2003) (Fig. 2). Due to the intense tectonic deformation, the thickness of the formation is difficult to estimate, but an average of *c.* 60 m seems to be justifiable. The Katafito Formation corresponds roughly to the ‘Premier Flysch de Pinde’ (Aubouin, 1957), the ‘Formation d’Andritsena’ (Maillot, 1979) and the lower part of the ‘Lambia Formation’ (Degnan & Robertson, 1998).

In the mostly reddish chert facies of the Early Cretaceous, there are a few green horizons associated with black shale layers. These are considered to correlate with worldwide anoxic events (Neumann & Zacher, 1996). The maximum siliciclastic supply occurred during the Cenomanian to Coniacian (Wagreich *et al.* 1996; Neumann, 2003). Two characteristic *Orbitolina*-bearing turbidite horizons have been found, one in the upper part of the Lower Cenomanian to the lower part of the Middle Cenomanian and one in the Upper Turonian to Lower Coniacian (Neumann, 2003). In the Santonian the cherty/siliciclastic facies of the Katafito Formation passed into the *Globotruncana*-bearing pelagic carbonates of the ‘Platy Limestone Formation’ (= ‘Calcaires en Plaquette’: Aubouin, 1959). The major part of the ‘Platy Limestone Formation’, the ‘Olonos Limestone Member’ *sensu* Neumann (2003) comprises a calciturbiditic development in the Campanian and Maastrichtian. From the Late Maastrichtian on, thin siliciclastic turbidites are intercalated in the platy pelagic carbonates, which passes into the terrigenous ‘Pindos Flysch Formation’ of the Palaeogene. The transitional facies has been named ‘Couches de Passage’ (Aubouin, 1959) or ‘Kataraktis Passage Member’ of the ‘Lambia Formation’ (Degnan & Robertson, 1998).

On the Greek mainland, the terrigenous turbidite development of the mid-Cretaceous is restricted to the northern Pindos Mountains, whereas towards the south only the cherty facies is present. In the Peloponnese, the turbidite deposits are concentrated in the western parts of the Pindos zone, with a relatively small occurrence in the east, while the cherty facies predominates in between (Neumann & Zacher, 2004, fig. 8A).

3. Heavy mineral distribution of the terrigenous detritus of the Katafito Formation

The sample locations follow the distribution of the siliciclastic facies of the Katafito Formation (Fig. 1, Table 1). Sixty-one samples have been prepared for microscopic investigation using laboratory procedures described by Faupl, Pavlopoulos & Migiros (1998). On both the Greek mainland and the Peloponnese, two major heavy mineral associations have been distinguished. The first association is characterized by the stable heavy minerals zircon, tourmaline, rutile and apatite accompanied by minor amounts of garnet, chrome spinel and traces of staurolite. In the Peloponnese this mineral assemblage is restricted to the western part of the Pindos zone (Fig. 1, locs 1–6; Fig. 3). Garnet-rich associations, which are rather rare, occur in locations 3 and 6 (Fig. 3). In the northern Pindos Mountains the stable mineral assemblage has been found in the western locations 9 and 10 (Fig. 4).

The second association is composed of high amounts of chrome spinel, which is frequently accompanied by green hornblende, whereas zircon, tourmaline, rutile, apatite, garnet, chloritoid and epidote form subordinate

https://doi.org/10.1017/S001675680600197X Published online by Cambridge University Press

Table 1. Locations of the samples

	Latitude	Longitude
Peloponnese		
1 Road Kato Klitoria–Kastelli	N 37°54.51′	E 22°22.03′
	N 37°54.39′	E 22°03.60′
2 Road Dimitsana–Panagia, near Zatouna	N 37°36.08′	E 22°02.20′
	N 37°35.21′	E 22°01.08′
3 Adritsena	N 37°26.97′	E 21°54.98′
	N 37°28.86′	E 21°54.83′
	N 37°30.55′	E 21°54.08′
4 Road Kato Melpia–Sirrizo, NW Diavolitsi	N 37°20.23′	E 21°55.37′
	N 37°21.63′	E 21°54.52′
5 W Meligalas, near village of Mila	N 37°13.78′	E 21°53.51′
6 Mount Ithomi	N 37°10.22′	E 21°55.58′
	N 37°09.74′	E 21°55.56′
	N 37°09.55′	E 21°55.53′
7 Mount Farmakas, NW Argos	N 37°45.69′	E 22°33.06′
	N 37°45.71′	E 22°32.90′
Northern Greek mainland		
8 Palaeokaria valley	N 39°25.80′	E 21°31.50′
	N 39°25.77′	E 21°31.48′
9 W Neraidochori, near Pertouli	N 39°32.50′	E 21°25.00′
10 SW Stournareika	N 39°28.22′	E 21°29.20′
11 Road Oxia–Argithea, W Mouzaki	N 39°21.92′	E 21°33.97′

Number 1–11 refer to Fig. 1.

constituents. In the northern Pindos Mountains this assemblage characterizes locations 8 and 11 (Figs 1, 4) and the isolated location 7 of the Peloponnese. In the western Peloponnese, one spinel-rich sample has been detected (Fig. 3, loc. 3).

The heavy mineral associations reveal two different source terrains, as clearly seen in Figure 5, where apatite/tourmaline and chrome spinel/zircon ratios have high provenance sensitivity (ATi, Czi indices: Morton & Hallsworth, 1994). Detrital blue amphibole, which is very common in the Tertiary Pindos Flysch Formation (Faupl, Pavlopoulos & Migiros, 1998, 2002; Schneider, Bode & Oppermann, 1998), has not been observed in the terrigenous material of the Katafito Formation.

4. U–Pb and Pb–Pb dates of detrital zircons

From locations 1 and 2 of the Peloponnese and 9 and 10 of the Greek mainland, detrital zircons were chosen for U–Pb and Pb–Pb age determination (Fig. 1). These samples represent the stable heavy mineral group (see Figs 3, 4). The primary aim of these investigations was to find the youngest sediment source provenance that contributed to the terrigenous turbidite material of the Katafito Formation.

Zircons were extracted using standard techniques involving crushing, sieving, heavy liquid separation and hand picking. The zircons were classified into different fractions based on their morphology, colour, optical homogeneity, such as presence or absence of inclusions or inherited cores, shape and elongation. The highest quality zircons, that is, idiomorphic non-metamict magmatic grains, inclusion-free and without visible cores, were air abraded to remove the outer surfaces affected by leaching, overgrowth, alteration or cracks in order to produce more concordant ages. These zircons were cleaned in dilute HNO_3 and acetone in an ultrasonic bath, followed by cleaning in suprapure ethanol (for several hours) and warm 3N HNO_3 (for 20 minutes). The clean zircons were dated either by the zircon evaporation or conventional single zircon U–Pb method, respectively. Preparatory handling of the cleaned zircons for evaporation analysis and full details of the technique are summarized in Klötzli (1997). The single zircon U–Pb data were obtained using the methods described in Kebede *et al.* (2005). Preparation and isotope measurements were done in the Geochronological Laboratory of the Department of Geological Sciences of the University of Vienna (MC TIMS Finnigan MAT 262).

The analytical results are given in Table 2. A total of 33 zircons were dated (8 evaporation, 25 conventional U–Pb analysis, respectively). Final probability density distribution plots and age maxima were calculated using AgeDisplay (Sircombe & Stern, 2002). Apparent ages show strongly differing degrees of concordance with $^{207}Pb/^{206}Pb$ ages ranging from 63 to 3937 Ma. Rejecting zircons with a discordance of $> 10\,\%$ and assuming concordance for the evaporated crystals (Klötzli, 1997), 21 crystals remain with apparent $^{207}Pb/^{206}Pb$ ages ranging from 324 to 2050 Ma. In order to combine both evaporation and conventional age data, only the $^{207}Pb/^{206}Pb$ dates will be discussed further. Figure 6 shows a probability density distribution plot of the $^{207}Pb/^{206}Pb$ age data in the range of 200–1000 Ma. The $^{207}Pb/^{206}Pb$ ages are interpreted as minimum magmatic formation ages.

The primary aim of the study was to identify the youngest magmatic event in the detrital zircon

https://doi.org/10.1017/S001675680600197X Published online by Cambridge University Press

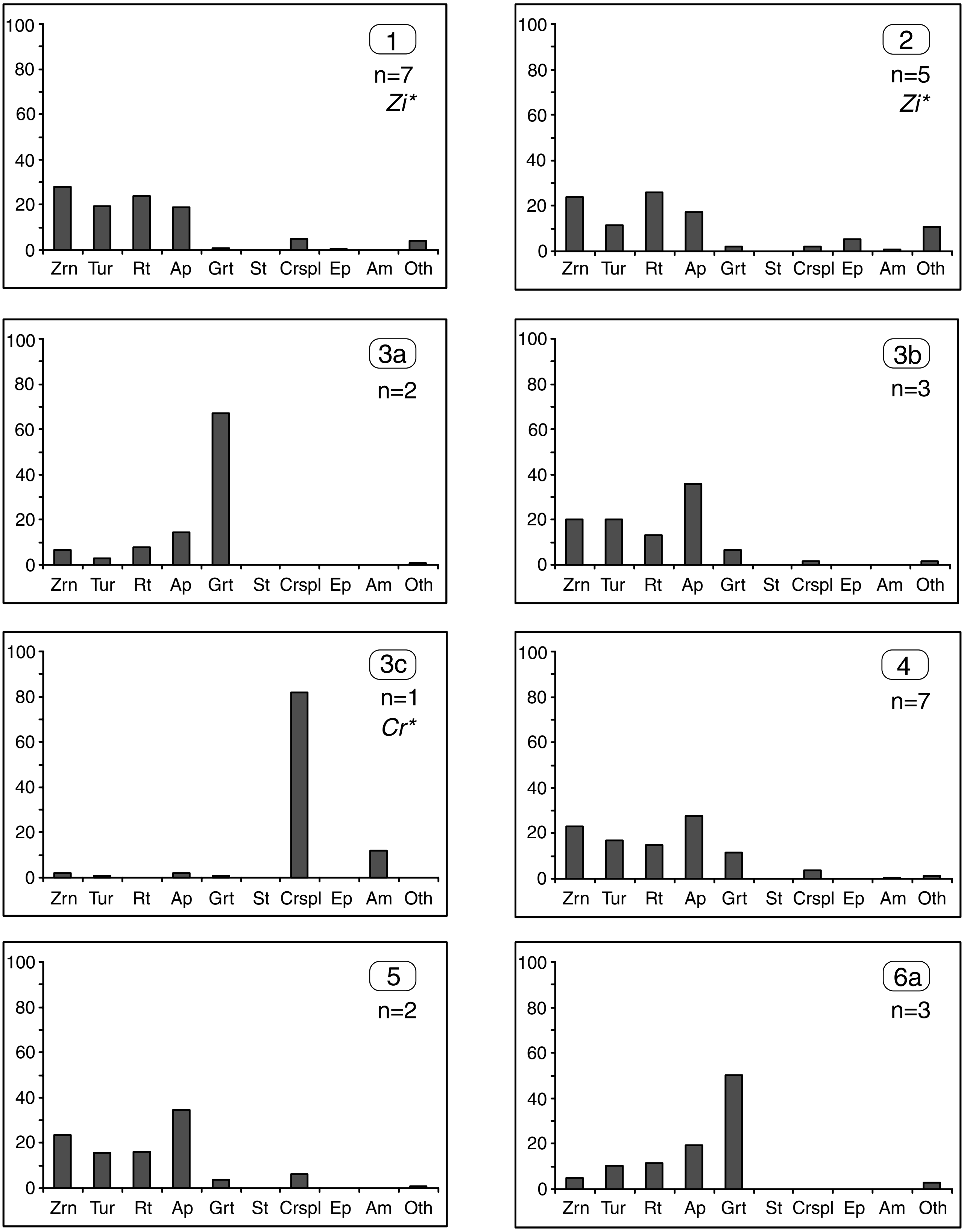

Figure 3. Heavy mineral assemblages of the mid-Cretaceous turbidites of the Pindos zone. For sample locations see Figure 1 and Table 1. Abbreviations: Zrn – zircon, Tur – tourmaline, Rt – rutile, Ap – apatite, Grt – garnet, St – staurolite, Crspl – chrome spinel, Ep – epidote group, Am – green amphiboles, Oth – other heavy minerals; *Zi** – zircon age data; *Cr** – chemical analyses from chrome spinel.

https://doi.org/10.1017/S001675680600197X Published online by Cambridge University Press

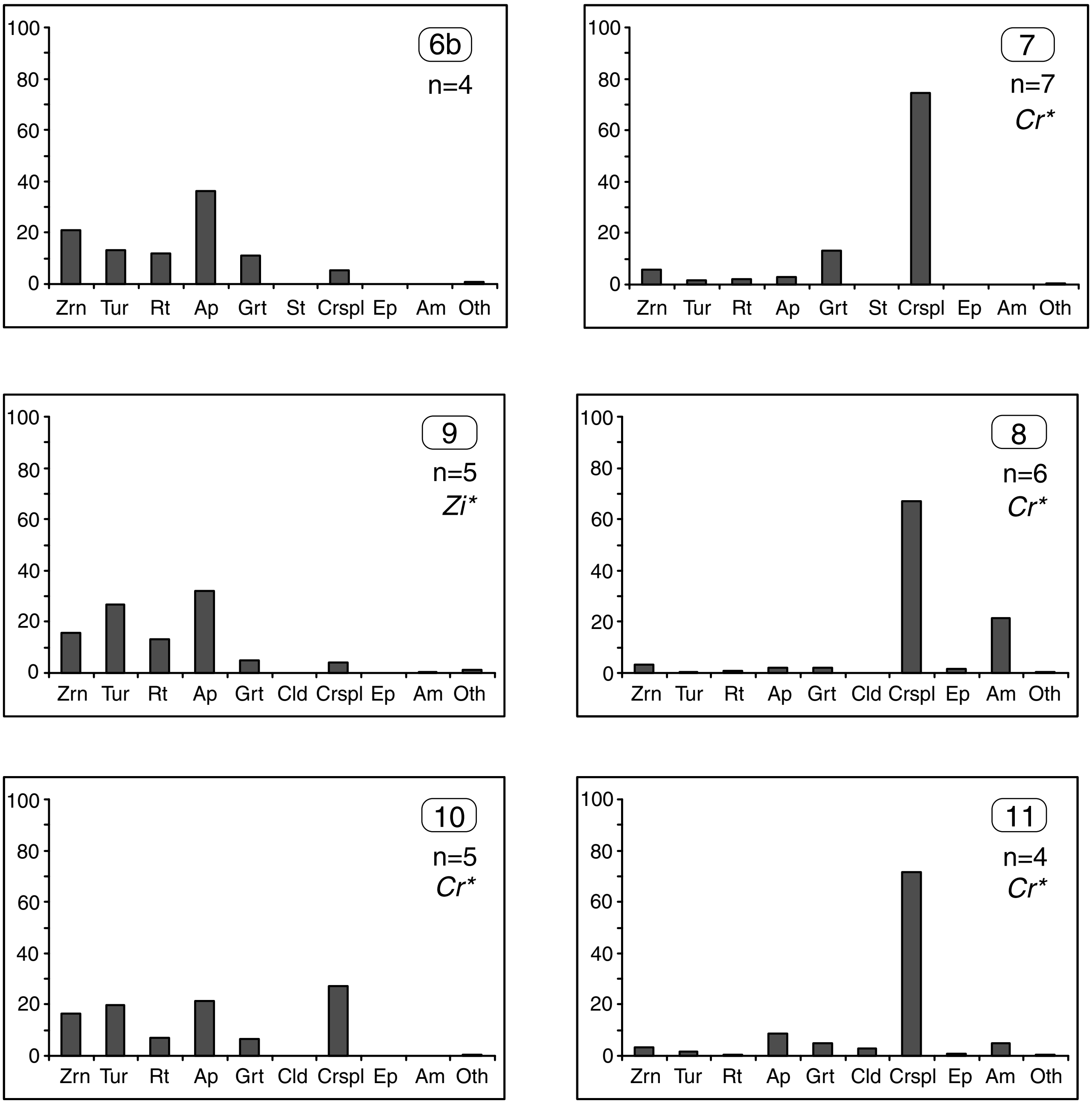

Figure 4. Heavy mineral assemblages of the mid-Cretaceous turbidites of the Pindos zone. For sample locations see Figure 1 and Table 1. For abbreviations see Figure 3. Location 10: the label is mistaken and should read *Zi** not *Cr**.

populations. This should result in defining a minimum age range for the hinterland continental crust. It was not the aim of the investigation to completely reconstruct the age structure of the hinterland. To achieve this on a statistically sound basis, the number of zircons dated would have to be far higher (> 100: Vermeesch, 2004).

Prominent age probability maxima are at 328, 351, 374 and 482 Ma. Additional maxima can be seen at 1440 Ma and 1640 Ma. The reported maxima are interpreted to be geologically meaningful on the basis that they appear in different samples. Maxima derived from only one age are not interpreted as significant. Beside the Cadomian and older Precambrian populations, the Variscan zircon ages are of special interest in the discussion of the source terrain.

5. Chemistry of the detrital chrome spinels

From seven samples of four locations of the northern Greek mainland and the Peloponnese (see Figs 3, 4), the chemistry of 401 detrital chrome spinel grains was studied by microprobe analytical technique (CAMECA SX 100, Department of Geological Sciences of the University of Vienna). Natural minerals have been used as standards. The analytical uncertainty of this instrument is updated periodically by measuring a

https://doi.org/10.1017/S001675680600197X Published online by Cambridge University Press

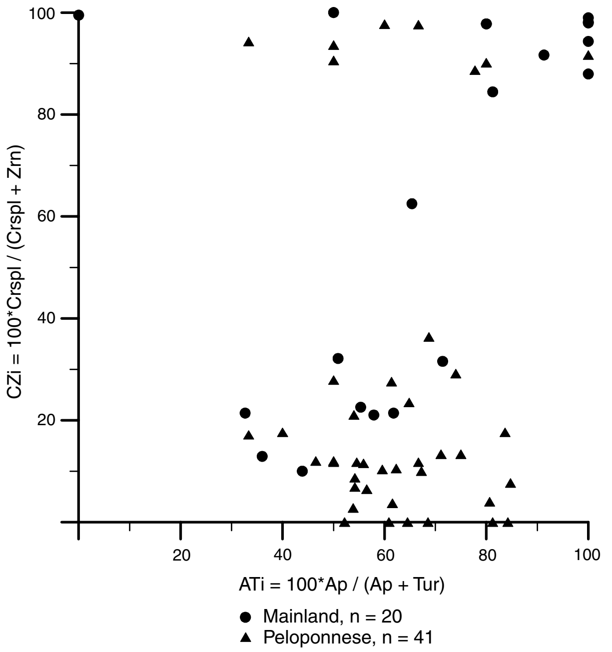

Figure 5. Separation of two heavy mineral populations in the turbiditic sandstones of the mid-Cretaceous Katafito Formation on the basis of the provenance-sensitive ATi and CZi indices introduced by Morton & Hallsworth (1994).

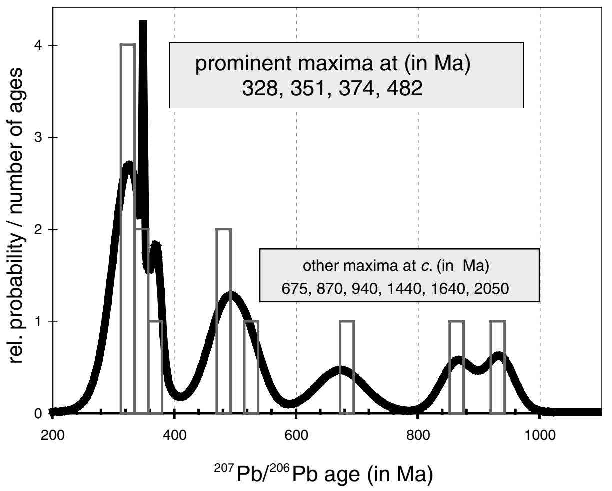

Figure 6. Frequency distribution of $^{207}Pb/^{206}Pb$ ages of detrital zircons from the turbiditic sandstones of the mid-Cretaceous Katafito Formation.

homogeneous subcalcic augite (2.5 % Na_2O, > 200 analyses) and is better than 0.5 % (1σ) for all major and minor oxides. Formula calculations were carried out with the program MINSORT vers. 2 (Petrakakis & Dietrich, 1985).

The variability of the chemical compositions (Table 3) and the Mg- and Cr-numbers (Fig. 7) demonstrates that the chrome spinel detritus from the Greek mainland and the Peloponnese is geochemically in good agreement and shows the typical distribution of spinels from ophiolite complexes (Pober & Faupl, 1988). A regional trend within the Pindos zone between externally and internally situated locations is not observed. The Cr-numbers show a broad spread of 0.2 to 0.9, which has not been observed in the spinel chemistry of the major ophiolite complexes of the Hellenides, where Cr-numbers < 0.4 have not been detected (Gartzos, Migiros & Parcharidis, 1990; Konstantopoulos & Economou-Eliopoulos, 1990; Michailidis, 1993).

Spinels with Cr-numbers < 0.5 were derived from lherzolitic bodies, such as abyssal peridotites of slow-spreading ridges (Lee, 1999), and are also described in bodies from back-arc basins (Cookenboo, Bustin & Wilks, 1997). Peridotite and volcanic spinels have been separated on the basis of the TiO_2 content and the Fe^{2+}/Fe^{3+} ratio (Lenaz *et al.* 2000, fig. 2; Kamenetzky, Crawford & Meffre, 2001). The Al_2O_3 v. TiO_2 diagram (Fig. 8), which uses the data of Kamenetzky, Crawford & Meffre (2001), shows that most of the detrital spinels were derived from peridotites, whereas a small fraction of about 15 % comes from volcanic rocks. The broad spread of the Al_2O_3 values seems to be indicative of two sources: peridotites of MOR-type as well as of suprasubduction zones.

The relatively small population of 'volcanic spinels' (TiO_2 > 0.2 wt %) shows a clear regional trend (Fig. 9), where the majority of the 'volcanic spinels' from the Peloponnese have distinctly lower Al_2O_3 values than those of the Greek mainland. Therefore, it is supposed that most of the 'volcanic spinels' of the Peloponnese were derived from back-arc basin settings, whereas those from the mainland were supplied from MOR-basalts.

6. Palaeogeographic interpretation and discussion

The appearance of siliciclastic material in the mid-Cretaceous 'First Flysch' of the Pindos zone is generally understood as a strong sedimentary signal for early tectonic evolution of the rising Hellenic orogen. The Pelagonian zone was intensively tectonized and metamorphosed during this period, as radiometric data in the range of 130–100 Ma have indicated (e.g. Euboea: Maluski *et al.* 1981; Olympos region: Schermer, Lux & Burchfield, 1990; Ossa region: Lips, White & Wijbrans, 1998). Tectonic activity along the external margin of the Pelagonian zone has frequently been invoked (e.g. Vergely, 1984; Jacobshagen, 1986; Katsikatsos, 1992). Jurassic and Cretaceous ophiolite obduction from both the Vardar and Pindos oceanic domains onto the Pelagonian zone preceded the mid-Cretaceous orogenic events (e.g. Baumgartner, 1985; Robertson & Karamata, 1994).

The palaeogeographic position of the sources of the terrigenous material of the turbidites is a complex problem. On the basis of palaeocurrent indicators, the siliciclastic turbidites of the northern Pindos Mountains show transport directions from the north, parallel to

https://doi.org/10.1017/S001675680600197X Published online by Cambridge University Press

Table 2. Single zircon U–Pb data

		Atomic ratios									Apparent ages (Ma)					
	No.	^{206}Pb/^{204}Pb	^{207}Pb*/^{235}U	± 2 RSD	^{206}Pb*/^{238}U	± 2 RSD	^{207}Pb*/^{206}Pb*	± 2 RSD	^{208}Pb*/^{206}Pb*	± 2 RSD	^{207}Pb*/^{206}Pb	± 2 RSD	^{207}Pb*/^{235}U	± 2 RSD	^{206}Pb*/^{238}U	± 2 RSD
Location 1																
GR 224-2	A	195	2.9479	2.09	0.24033	0.27	0.08896	2.02	0.043	9.83	1403	2.76	1394	1.14	1388	0.25
GR 224-2	B	942	1.8811	1.15	0.07559	0.22	0.18049	0.50	0.464	0.44	2657	0.31	1074	0.71	470	0.21
GR 224-2	C	165	13.096	4.10	0.23301	0.34	0.40763	0.70	0.901	0.76	3937	0.27	2687	1.44	1350	0.31
GR 225-1	A	115	0.3997	0.79	0.05416	0.85	0.05352	0.13	1.898	1.40	351	0.84	341	0.67	340	0.83
GR 225-1	B	972	0.3757	8.54	0.05119	0.53	0.05324	2.70	1.023	0.69	339	18.0	324	7.31	322	0.52
GR 225-1	C	440	0.6302	12.4	0.07914	0.57	0.05775	2.30	0.105	3.01	520	9.69	496	9.85	491	0.55
GR 225-1	D	1236	3.0806	3.61	0.24642	0.99	0.09067	1.30	0.098	2.54	1440	1.72	1428	1.94	1420	0.89
GR 225-1	E	3547	0.3436	7.03	0.04712	0.51	0.05288	2.06	0.099	2.62	324	14.5	300	6.09	297	0.50
GR 225-1	F	1388	0.3689	8.68	0.05057	1.04	0.05291	2.79	0.106	3.31	325	19.5	319	7.45	318	1.01
GR 225-1	G	1755	0.5747	10.3	0.07348	0.58	0.05673	2.16	0.141	2.11	481	9.92	461	8.29	457	0.56
GR 225-4	A	20715	0.3610	1.90	0.04715	0.20	0.05553	1.78	0.091	2.46	434	9.15	313	1.63	297	0.20
GR 225-4	B	266	0.6463	1.71	0.07843	0.18	0.05977	1.30	0.097	1.73	595	4.72	506	1.35	487	0.18
GR 225-4	C	675	2.3650	3.77	0.20546	0.32	0.08348	2.53	0.009	4.76	1281	3.85	1232	2.18	1205	0.29
GR 225-4	D	1604	0.3532	0.33	0.04669	0.27	0.05488	0.33	0.078	0.55	407	1.81	307	0.29	294	0.26
GR 225-4	E	2052	16.015	0.57	0.45907	0.21	0.25302	0.42	0.281	0.67	3204	0.21	2878	0.19	2435	0.17
GR 225-4	F	995	3.1517	0.80	0.24896	0.16	0.09181	0.45	0.122	0.65	1464	0.59	1445	0.43	1433	0.15
Location 2																
GR 213-3	A	129	0.3386	0.71	0.04583	0.45	0.05358	0.69	0.297	2.84	354	4.38	296	0.62	289	0.44
GR 213-3	B	253	1.0354	1.62	0.06505	0.30	0.11544	0.62	0.298	0.58	1887	0.60	722	1.16	406	0.29
GR 213-3	C	124	3.1910	1.08	0.12291	1.83	0.18830	0.43	0.456	0.40	2727	0.26	1455	0.57	747	1.73
GR 213-3	D	827	0.4168	1.22	0.05325	1.74	0.05677	1.05	0.105	1.27	483	4.83	354	1.03	334	1.70
GR 213-3	E	385	3.9425	0.98	0.28362	0.17	0.10082	0.80	0.289	0.62	1639	0.90	1622	0.49	1610	0.15
Location 9																
GR 084-5	A	*					0.06203	3.74			675	11.9				
GR 084-5	B	*					0.07030	2.34			937	5.12				
GR 084-5	C	*					0.06798	2.60			868	6.22				
GR 084-5	D	*					0.10162	2.97			1654	3.33				
GR 084-5	E	*					0.10064	0.64			1636	0.73				
GR 084-6	A	555	0.0483	2.44	0.00715	0.57	0.04893	2.36	0.251	1.33	144	38.3	48	2.38	46	0.57
GR 084-6	B	377	0.3597	9.95	0.04931	0.68	0.05290	3.11	0.123	3.22	325	21.7	312	8.57	310	0.67
GR 084-6	C	1050	0.3799	6.01	0.05200	0.50	0.05299	1.86	0.147	1.65	328	12.8	327	5.14	327	0.49
Location 10																
GR 085-2	A	*					0.05691	3.94			488	17.8				
GR 085-2	B	*					0.05407	0.67			374	4.01				
GR 085-2	C	*					0.12651	8.47			2050	7.32				

*Single zircon evaporation data; for sample locations see Figure 1 and Table 1.

Table 3. Chemistry of the detrital chrome spinels from the Katafito Formation

	Greek mainland						Peloponnese					
	Peridotitic spinels (TiO_2 < 0.2 wt %) n = 149			Volcanic spinels (TiO_2 > 0.2 wt %) n = 34			Peridotitic spinels (TiO_2 < 0.2 wt %) n = 192			Volcanic spinels (TiO_2 > 0.2 wt %) n = 26		
	Mean	Min	Max	Mean	Min	Max	Mean	Min	Max	Mean	Min	Max
SiO_2	0.024	0.000	0.148	0.021	0.000	0.173	0.027	0.000	0.695	0.025	0.000	0.060
TiO_2	0.094	0.003	0.200	0.517	0.202	1.991	0.076	0.000	0.200	0.460	0.203	1.111
Al_2O_3	25.927	4.827	53.543	29.296	13.859	43.283	26.930	4.055	53.169	24.649	6.628	48.586
Cr_2O_3	42.333	14.952	64.719	34.465	23.127	46.072	41.379	13.987	67.240	40.566	17.574	62.244
FeO_{tot}	18.052	12.440	28.290	22.373	15.968	41.581	18.070	12.467	36.065	21.145	15.018	34.392
MnO	0.143	0.052	0.431	0.162	0.068	0.371	0.129	0.028	0.387	0.144	0.061	0.229
ZnO	0.111	0.000	0.480	0.121	0.000	0.388	0.200	0.000	0.561	0.176	0.039	0.347
NiO	0.111	0.000	0.295	0.152	0.059	0.289	0.114	0.000	0.351	0.131	0.037	0.317
MgO	12.701	6.821	19.064	12.246	5.817	16.723	12.858	5.651	18.399	12.435	7.933	18.095
CaO	0.010	0.000	0.083	0.011	0.000	0.039	0.006	0.000	0.208	0.002	0.000	0.029
Cr/(Cr+Al)	0.534	0.158	0.897	0.447	0.264	0.679	0.522	0.150	0.918	0.531	0.195	0.863
$Mg/(Mg+Fe^{2+})$	0.576	0.355	0.759	0.546	0.286	0.696	0.578	0.287	0.747	0.564	0.371	0.756
Fe^{2+}/Fe^{3+}	20.324	2.005	688.833	5.959	1.681	40.287	12.374	2.272	73.000	5.164	2.052	17.512
Fe_2O_3	1.978	0.000	8.964	4.935	0.519	17.236	2.021	0.000	12.249	4.674	1.062	11.531
FeO	16.272	10.535	22.532	17.932	13.007	26.072	16.252	11.122	25.043	16.939	10.391	24.016

the strike of the tectonic structures (Neumann, 2003). On the other hand, based only on a very small set of palaeoflow data, the turbidites of the Pindos zone of the Peloponnese suggest an external source in the west (Piper & Pe-Piper, 1980; Neumann, 2003). The regional distribution of the heavy mineral associations supports the concept of two separate source terrains: the chrome spinel-dominated assemblages seem to be derived from an internal source (see Faupl, Pavlopoulos & Migiros, 1998; Neumann, 2003; Neumann & Zacher, 2004), whereas the stable heavy mineral association characterizes a second source.

Ages of idiomorphic non-metamict magmatic zircons derived from the erosion of Variscan igneous rocks define the most prominent maxima in the age distribution (Fig. 6, 328, 351, 374 Ma maxima). Approximately 35 % of the investigated zircons belong to this age group, which is found in four different samples (Table 2, GR84-6, GR85-2, GR213-3, GR225-1). Whether the maximum at 351 Ma is real or not cannot be judged at the moment. Nevertheless, in a very general way, the three age groups correspond to Devonian (early Variscan), Early Carboniferous and Middle Carboniferous magmatic events found elsewhere in the Variscan orogen of Central and Western Europe. Permian–Triassic (300–240 Ma) or late Variscan (320–300 Ma) zircons have not been detected. As well as the Variscan group, three other discrete times of magmatic activity can be distinguished. An Ordovician (late Cadomian?) event around 480–520 Ma has been detected in three samples (Table 2, GR85-2, GR213-3, GR225-1). Again, this age range is characteristically found in crustal units derived from the northern margin of Gondwana (Romano, Dörr & Zulauf, 2004). Middle Proterozoic age maxima around 1440 Ma (Table 2, GR224-2, GR225-1, GR225-4) and 1640 Ma (Table 2, GR84-5, GR213-3) point to some prominent older crustal sources.

The siliciclastic material characterized by the stable heavy mineral group cannot have been derived from the Apulian foreland *sensu stricto* directly, for the following two reasons: (1) the Ionian basin and the broad Gavrovo–Tripolitza platform were shielding the Pindos zone from the foreland and (2) the Apulian foreland, which forms part of the 'Cimmerian Terranes', was not affected by the Variscan orogeny (e.g. Stampfli, Rosselet & Bagheri, 2004). Therefore, the appearance of Variscan detrital zircons in the turbiditic material (Fig. 6) also excludes the Apulian foreland *s.s.* as a possible source.

Thus, for the zircon-rich heavy mineral population, the only possible external source regions are unknown areas of the Gavrovo–Tripolitza zone or transitional domains between the platform and the Pindos basin with exposed crystalline basement. The transition between the Gavrovo–Tripolitza and Pindos zones may be thought of as an Iberian-type continental margin (Manatschal, 2004). Information about the crystalline

https://doi.org/10.1017/S001675680600197X Published online by Cambridge University Press

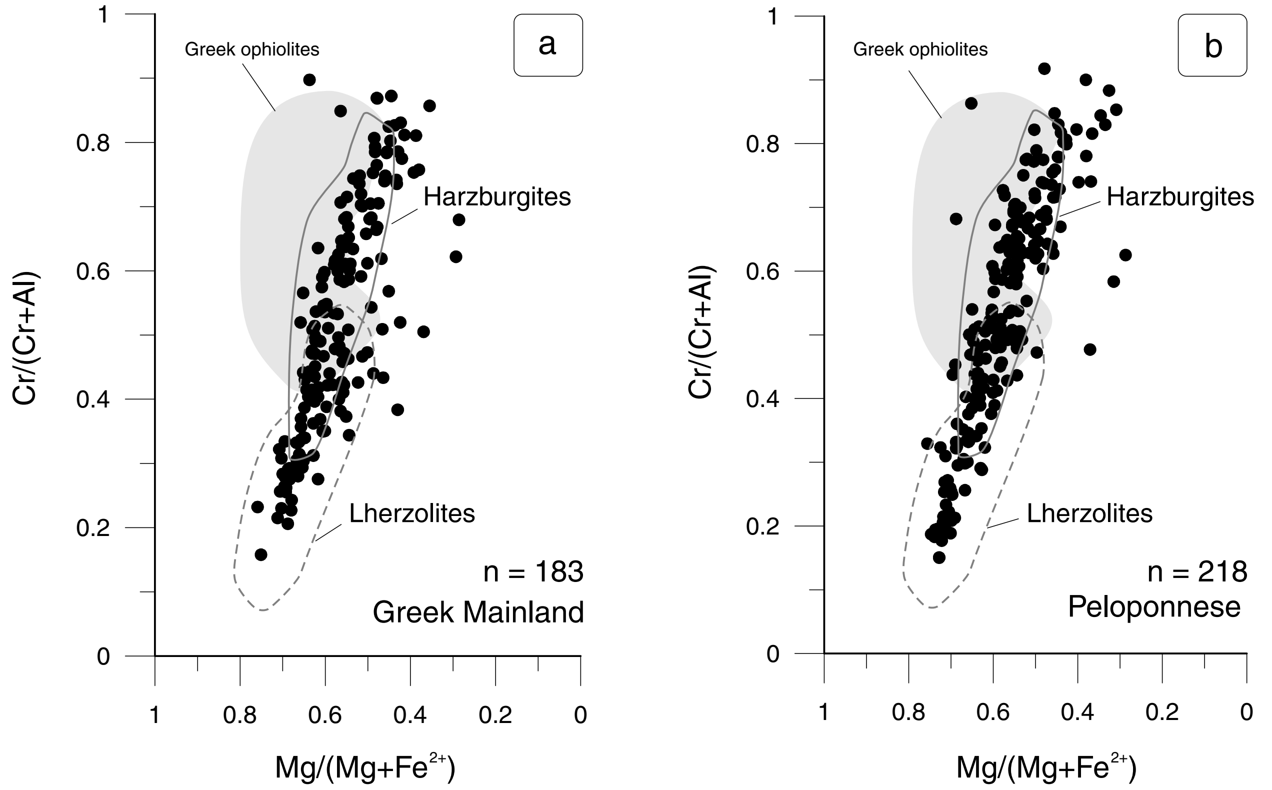

Figure 7. Variation of Mg/(Mg + Fe^{2+}) against Cr/(Cr + Al) of detrital chrome spinels from the turbiditic sandstones of the mid-Cretaceous Katafito Formation. (a) Analyses from locations of the northern Greek mainland; (b) analyses from locations of the Peloponnese. See Figures 3 and 4. Fields of spinels from harzburgites and lherzolites after Pober & Faupl (1988). Field of Greek ophiolites after Gartzos, Migiros & Parcharidis (1990).

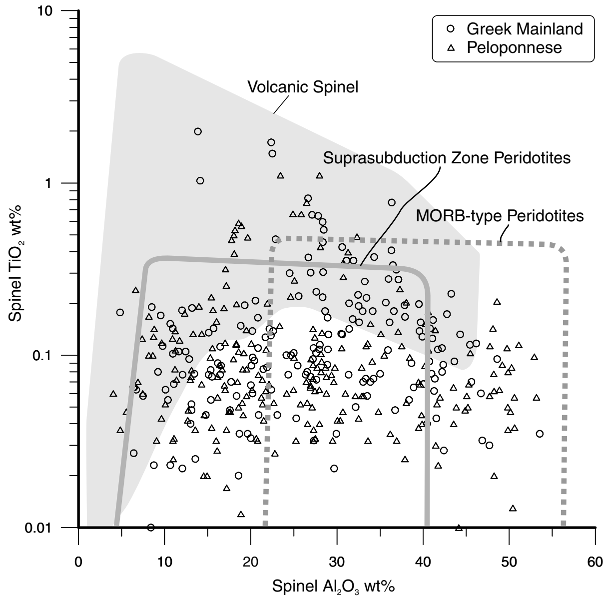

Figure 8. Detrital chrome spinels from the turbiditic sandstones of the mid-Cretaceous Katafito Formation in the discrimination diagram between 'mantle' and 'volcanic' spinels using TiO_2 (Kamenetzky, Crawford & Meffre, 2001).

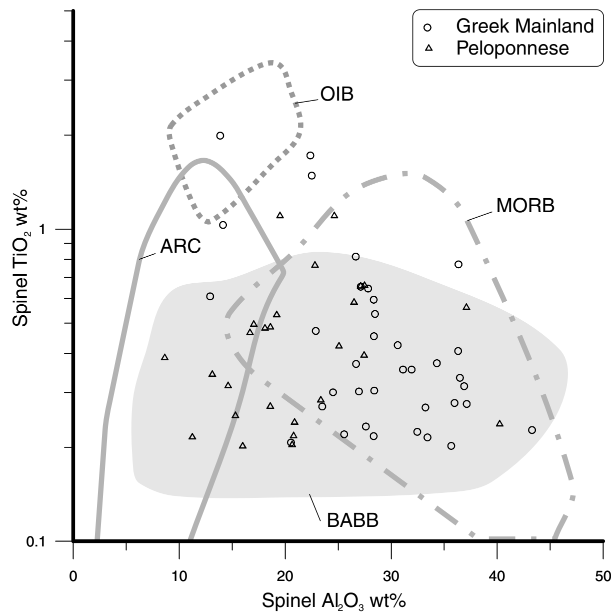

Figure 9. Al_2O_3 v. TiO_2 compositional relations in 'volcanic spinels' (TiO_2 > 0.2 wt %) from turbiditic sandstones of the mid-Cretaceous Katafito Formation. Fields after Lenaz *et al.* 2000; MORB – mid-ocean ridges, BABB – back-arc basins, OIB – ocean islands, ARC – island arcs.

https://doi.org/10.1017/S001675680600197X Published online by Cambridge University Press

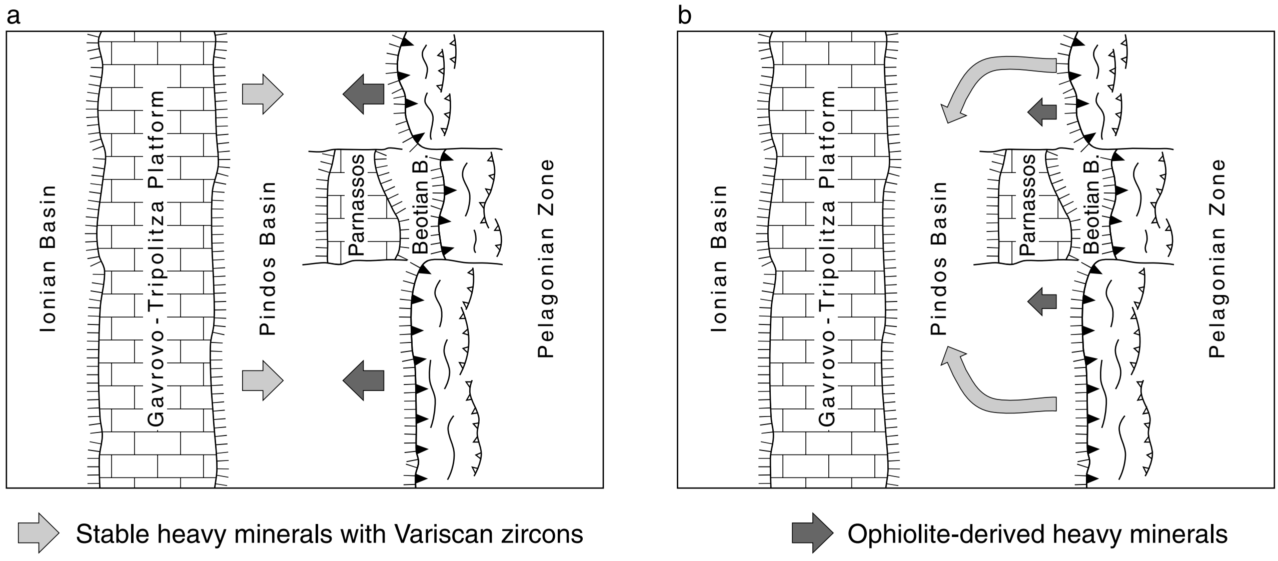

Figure 10. Palaeogeographical sketch maps (not to scale) of the mid-Cretaceous Hellenides showing the two models for the provenance of heavy minerals of the Pindos zone discussed in the text.

basement of the Gavrovo–Tripolitza zone in the area of the Greek mainland and the Peloponnese does not exist, but from eastern Crete, an amphibolite-grade crystalline unit ('Sitia unit') has been described from the base of the Gavrovo–Tripolitza nappe. From this Sitia unit, Variscan radiometric ages have been reported (Papanikolaou, 1988; Finger *et al.* 2002; Champod, Stampfli & Kock, 2004). This implies that the Palaeotethys suture (the boundary between Variscan deformed and Variscan undeformed crust) should be situated between the 'foreland *s.s.*' and the basement of the Gavrovo–Tripolitza zone (Stampfli, Rosselet & Bagheri, 2004). If these findings from the Sitia unit of Crete can be transferred to the Gavrovo–Tripolitza basement of the Peloponnese and the Greek mainland, an external source for the zircon-dominated heavy mineral association of the Katafitio Formation seems to be possible (Fig. 10a).

On the other hand, it is possible that zircon populations with Variscan ages could also be internally derived from the crystalline basement of the Pelagonian zone, where intensive Variscan deformation, metamorphism and magmatism have been documented (e.g. Mountrakis, 1986). If the model of an internal source for the stable heavy mineral group is favoured (Fig. 10b), less significance has to be given to the palaeocurrent indicators reported above. It is a well-known phenomenon in deep-sea basins that the flow directions are strongly controlled by basin floor morphology and current deflections do also occur (e.g. Haughton, 1994). In addition, the geographical land–sea distribution of the Pelagonian zone and the adjacent 'First Flysch' basin could also affect the palaeocurrent flow directions. The concept of an internal source for both heavy mineral assemblages (Fig. 10b) has the advantage that the low amounts of chrome spinel accompanying the stable heavy mineral group are more explicable than in the case of an external source (Fig. 10a).

The chemical investigations of the detrital chrome spinels concentrated exclusively on material extracted from samples of the spinel-dominated heavy mineral association (Figs 3, 4). The Cr/Al values of the detrital spinels, which vary in a conspicuously broad manner (Figs 7, 8), may be explained by erosion of two different ophiolite types, namely, peridotites of mid-ocean ridge setting and peridotites generated in suprasubduction zones, respectively. Such a broad distribution has not been observed in the spinel chemistry of the recently exposed ophiolite complexes of the Hellenides (e.g. Gartzos, Migiros & Parcharidis, 1990, Fig 5). However, Maksimovic & Majer (1981) described Cr-rich spinels with Cr numbers > 0.4 from the ophiolite belt of the Inner Dinarides, which are distinguished from the spinels of the ophiolite belt of the Central Dinarides with Cr-numbers < 0.4. It is therefore supposed that, similar to the scenario in the Dinarides, two different belts of ophiolite complexes were exposed at the same erosional level during the mid-Cretaceous, given that at this time the freshly obducted ophiolites could contain a broad mineralogical–lithological spectrum including types that are now erosionally consumed. A very similar trend in the chemistry of detrital spinels has been observed within the Upper Austroalpine nappe unit of the Eastern Alps, where Early Cretaceous turbidites of the Rossfeld and Schrambach formations are characterized by Cr-rich spinels, whereas the mid-Cretaceous detritus (Tannheim, Losenstein formations) also contains Al-rich spinels (Pober & Faupl, 1988).

The co-occurrence of two different types of peridotite complexes in a mid-Cretaceous reconstruction seems to be only conceivable for the palaeogeographically adjacent internal Hellenic zones, such as

https://doi.org/10.1017/S001675680600197X Published online by Cambridge University Press

the Pelagonian unit (Baltuck, 1982; Richter & Müller, 1993) and the Vardar-Axios zone. Therefore, it is justified to suppose an internal source for the chrome spinel-rich heavy mineral association. The position of the source of the stable heavy mineral group is a more open problem, but for the moment we prefer an internal source terrain (Fig. 10b), because even this heavy mineral assemblage contains detrital spinels, although mostly in trace amounts. To solve this problem more information should be made available, such as data on a possible detrital influence from the Palaeotethys suture, which could also supply detrital chrome spinels.

7. Conclusions

Two heavy mineral assemblages characterize the siliciclastic turbidite material of the mid-Cretaceous Katafito Formation of the Pindos zone: (1) a stable heavy mineral association dominated by zircon and (2) an ophiolite-derived mineral group rich in chrome spinel (Fig. 5).

U–Pb and Pb–Pb dating on detrital magmatic zircons, taken from the stable heavy mineral group, shows clear evidence for the occurrence of Variscan magmatic rocks (around 350 Ma) in the source terrain. In addition to the Variscan ages, three further periods of magmatic activities (480–520, 1440, 1640 Ma) have been distinguished. Palaeogeographically, two possibilities exist to explain the provenance of the zircon-dominated heavy mineral association (Fig. 10). Basement rocks of the Pelagonian zone, where Variscan magmatism and deformation are well documented, would represent an internal source. Alternatively, an external source may be indicated by the regional distribution of these zircon-dominated heavy mineral assemblages and also on the basis of a few palaeocurrent indicators, especially in the Peloponnese. If an external source within the Gavrovo–Tripolitza zone or in the transition between that and the Pindos zone is favoured, the occurrence of detrital chrome spinel accompanying the zircon-dominated heavy mineral assemblages has to be explained, which presents a palaeogeographical problem. Therefore, our preferred palaeogeographical solution involves an internal source terrain for the zircon-rich heavy mineral association.

The chrome spinel-rich heavy mineral association can be derived without any palaeogeographical problems from an internal source terrain, and this is also in accordance with the regional distribution of these minerals. The chemical composition of the detrital chrome spinels documents that the ophiolite complexes which eroded during mid-Cretaceous times had a much broader lithological variability than ophiolite complexes presently exposed in the Hellenides. The chemical data from the detrital peridotite spinels indicate two source types (MORB and suprasubduction zone). The two ophiolite lithologies that existed in the mid-Cretaceous realm of the Hellenides show similarities to the ophiolite belts described in the Outer and Inner Dinarides by Maksimovic & Majer (1981).

Acknowledgements. The authors thank the Hochschuljubiläumsstiftung der Stadt Wien, grant no. H-24/2001, and the authorities of the University of Vienna and the Agricultural University Athens for financial support. A. Besada and R. Faupl were helpful in preparing the heavy mineral samples. W. Füzi produced the polished section for the microprobe. The paper was substantially improved by constructive comments of A. C. Morton and an anonymous reviewer.

References

AUBOUIN, J. 1957. Essai de correlation stratigraphique de la Gréce occidentale. *Bulletin de la Société géologique de France* **7**, 281–304.

AUBOUIN, J. 1959. Contribution à l'étude géologique de la Grèce septentrionale: les confins de l'Epire et de la Thessalie. *Annales géologiques des pays helléniques* **10**, 1–483.

BALTUCK, M. 1982. Provenance and distribution of Tethyan pelagic and hemipelagic siliceous sediments, Pindos Mountains, Greece. *Sedimentary Geology* **31**, 63–88.

BAUMGARTNER, P. O. 1985. Jurassic sedimentary evolution and nappe emplacement in the Argolis Peninsula (Peloponnesus, Greece). *Denkschriften der Schweizerischen Naturforschenden Gesesellschaft* **99**, 1–111.

BERNOULLI, D. & LAUBSCHER, H. 1972. The palinspastic problem of the Hellenides. *Eclogae geologicae Helvetiae* **65**, 107–18.

BRUNN, J. H. 1956. Contribution à l'étude géologique du Pinde septentrional et d'une partie de la Macédoine occidentale. *Annales géologiques des pays helléniques* **7**, 1–358.

CHAMPOD, E. C., STAMPFLI, G. M. & KOCK, S. 2004. Permo-Triassic evolution of the Tethyan margins in the External Hellenides (Crete). *Proceedings, 5th International Symposium on Eastern Mediterranean Geology, Thessaloniki*, 52–6.

COOKENBOO, H. O., BUSTIN, R. M. & WILKS, K. R. 1997. Detrital chromian spinel compositions used to reconstruct the tectonic setting of provenance: implications for orogeny in the Canadian Cordillera. *Journal of Sedimentary Research* **67**, 116–23.

DEGNAN, P. J. & ROBERTSON, A. H. F. 1998. Mesozoic–early Tertiary passive margin evolution of the Pindos ocean (NW Peloponnese, Greece). *Sedimentary Geology* **117**, 33–70.

DOUTSOS, T., PE-PIPER, G., BORONKAY, K. & KOUKOUVELAS, I. 1993. Kinematics of the Central Hellenides. *Tectonics* **12**, 936–53.

FAUPL, P., PAVLOPOULOS, A. & MIGIROS, G. 1998. On the provenance of flysch deposits in the external Hellenides of mainland Greece: results from heavy mineral studies. *Geological Magazine* **135**, 421–42.

FAUPL, P., PAVLOPOULOS, A. & MIGIROS, G. 2002. Provenance of Peloponnese (Greece) flysch based on heavy minerals. *Geological Magazine* **139**, 513–24.

FINGER, F., KRENN, E., RIEGLER, G., ROMANO, S. & ZULAUF, G. 2002. Resolving Cambrian, Carboniferous, Permian and Alpine monazite generations in the polymetamorphic basement of eastern Crete (Greece) by means of electron microprobe. *Terra Nova* **14**, 233–40.

GARTZOS, E., MIGIROS, G. & PARCHARIDIS, I. 1990. Chromites from ultramafic rocks of northern Evia (Greece) and their geotectonic significance. *Schweizer Mineralogische und Petrographische Mitteilungen* **70**, 301–7.

GOLONKA, J. 2004. Plate tectonic evolution of the southern margin of Eurasia in the Mesozoic and Cenozoic. *Tectonophysics* **381**, 235–73.

HAUGHTON, P. D. W. 1994. Deposits of deflected and ponded turbidity currents, Sorbas Basin, Southeast Spain. *Journal of Sedimentary Research* **A64**, 233–46.

JACOBSHAGEN, V. 1986. *Geologie von Griechenland*. Berlin, Stuttgart: Gebrüder Borntraeger, 363 pp.

JONES, G. & ROBERTSON, A. H. F. 1991. Tectono-stratigraphy and evolution of the Mesozoic Pindos ophiolite and related units, northwestern Greece. *Journal of the Geological Society, London* **148**, 267–88.

KAMENETSKY, V. S., CRAWFORD, A. J. & MEFFRE, S. 2001. Factors controlling chemistry of magmatic spinel: an empirical study of associated olivine, Cr-spinel and melt inclusions from primitive rocks. *Journal of Petrology* **42**, 655–71.

KATSIKATSOS, G. X. 1992. *Geology of Greece*. Athens: Katsikatsos, 451 pp. (in Greek).

KEBEDE, T., KLÖTZLI, U. S., KOSLER, J. & SKIÖLD, T. 2005. Understanding pre-Variscan and Variscan basement components of the central Tauern Window, Eastern Alps (Austria): constraints from new single zircon U–Pb geochronology. *International Journal of Earth Sciences* **94**(3), 336–53.

KLÖTZLI, U. S. 1997. Zircon evaporation TIMS: Method and procedures. *The Analyst* **122**, 1239–48.

KONSTANTOPOULOS, G. & ECONOMOU-ELIOPOULOS, M. 1990. Geochemistry of Vourinos chromite ores, Greece. In *Ophiolites – Oceanic Crustal Analogues* (eds J. Malpas, E. M. Moores, A. Panayiotou and C. Xenophontes), pp. 605–13. Nicosia: The Geological Survey Department.

LEE, Y. I. 1999. Geotectonic significance of detrital chromian spinel: a review. *Geoscience Journal* **3**, 23–9.

LENAZ, D., KAMENETSKY, V. S., CRAWFORD, A. J. & PRINCIVALLE, F. 2000. Melt inclusions in detrital spinel from the SE Alps (Italy–Solovenia): a new approach to provenance studies of sedimentary basins. *Contributions to Mineralogy and Petrology* **139**, 748–58.

LIPS, A. L. W., WHITE, S. H. & WIJBRANS, J. R. 1998. $^{40}Ar/^{39}Ar$ laserprobe direct dating of discrete deformational events: a continuous record of early alpine tectonics in the Pelagonian Zone, NW Aegean area, Greece. *Tectonophysics* **298**, 133–53.

MAILLOT, H. 1979. Etude sédimentologique du "Premier Flysch" en Peloponnèse occidental (Grèce): La Formation d'Andritsena. *Bulletin of the Geological Society of Greece* **14**, 94–115.

MAKSIMOVIC, Z. & MAJER, V. 1981. Accessory spinels of two main zones of alpine ultramafic rocks in Yugoslavia. *Bulletin de l'Académie Serbe des Sciences et des Arts, Classe de sciences naturelles et mathématiques* **75**, 47–58.

MALUSKI, H., VERGÉLY, P., BAVAY, D. B. P. & KATSIKATSOS, G. 1981. $^{40}Ar/^{39}Ar$ dating of glaucophanes and phengites in southern Euboa (Greece): geodynamic implications. *Bulletin de la Société géologique de France* **23**, 469–76.

MANATSCHAL, G. 2004. New models for evolution of magma-poor rifted margins based on a review of data and concepts from West Iberia and the Alps. *International Journal of Earth Sciences* **93**, 432–66.

MICHAILIDIS, K. M. 1993. Ferritechromite – chromian chlorite (Kaemmererite) associations in the podiform chromite ores of the Chalkidiki ophiolites in Macedonia, northern Greece. In *Publication in honour of Athanasios G. Panagos, professor at the National Technical University of Athens*, vol. 2, pp. 603–29. Athens (National Technical University).

MORTON, A. C. & HALLSWORTH, C. R. 1994. Identifying provenance-specific features of detrital heavy mineral assemblages in sandstones. *Sedimentary Geology* **90**, 241–56.

MOUNTRAKIS, D. 1986. The Pelagonian zone in Greece: a polyphase-deformed fragment of the Cimmerian continent and its role in the geotectonic evolution of the Eastern Mediterranean. *Journal of Geology* **94**, 335–47.

NEUMANN, P. 2003. Ablagerungsprozesse, Event- und Biostratigraphie kreidezeitlicher Tiefwassersedimente der Tethys in der Olonos-Pindos-Zone Westgriechenlands. *Münchner Geowissenschaftliche Abhandlungen, Reihe A* **40**, 1–156.

NEUMANN, P. & ZACHER, W. 1996. Multistratigraphic investigations and sedimentary cycles in the Upper Cretaceous of the Olonos–Pindos Zone (Greece): paleogeography, facies and basin development. *Göttinger Arbeiten zur Geologie und Paläontologie, Sonderband* **3**, 123–6.

NEUMANN, P. & ZACHER, W. 2004. The Cretaceous sedimentary history of the Pindos Basin (Greece). *International Journal of Earth Sciences* **93**, 119–31.

PAPANIKOLAOU, D. J. 1988. Introduction to the geology of Crete. In *IGCP Project No 276 – Paleozoic geodynamic domains and their alpidic evolution in the Tethys, First Field Meeting, Crete, Greece, October 10–14, 1988* (eds D. J. Papanikolaou, N. Skarpelis, A. Alexopoulos and Th. Markopoulos), pp. 3–16. Chania: Technical University of Crete.

PE-PIPER, G. 1998. The nature of Triassic extension-related magmatism in Greece: evidence from Nd and Pb isotope geochemistry. *Geological Magazine* **135**, 331–48.

PETRAKAKIS, K. & DIETRICH, H. 1985. MINSORT: a program for processing and archivation of microprobe analyses of silicate and oxide minerals. *Neues Jahrbuch für Mineralogie, Monatshefte* **1985**, 379–84.

PIPER, D. J. W. 2004. Sedimentology of the Pindos Flysch of the Peloponnese, Greece. *Proceedings, 5th International Symposium on Eastern Mediterranean Geology, Thessaloniki*, 165–7.

PIPER, D. J. W. & PE-PIPER, G. 1980. Was there a western (external) source of terrigenous sediment for the Pindos zone of the Peloponnese (Greece)? *Neues Jahrbuch für Geologie und Paläontologie, Monatshefte* **1980**, 107–15.

POBER, E. & FAUPL, P. 1988. The chemistry of detrital chromian spinels and its implications for the geodynamic evolution of the Eastern Alps. *Geologische Rundschau* **77**, 641–70.

RICHTER, D. & MÜLLER, C. 1993. Der "Erste Flysch" in der Pindoszone (Griechenland)? *Neues Jahrbuch für Geologie und Paläontologie, Monatshefte* **1993**, 209–26.

ROBERTSON, A. H. F., CLIFT, P. D., DEGNAN, P. & JONES, G. 1991. Palaeogeography and palaeotectonic evolution of the eastern Mediterranean Neotethys. *Palaeogeography, Palaeoclimatology, Palaeoecology* **87**, 289–343.

ROBERTSON, A. H. F. & KARAMATA, S. 1994. The role of subduction–accretion processes in the tectonic evolution of the Mesozoic Tethys in Serbia. *Tectonophysics* **234**, 73–94.

ROMANO, S. S., DÖRR, W. & ZULAUF, G. 2004. Cambrian granitoids in the pre-Alpine basement of Crete (Greece): evidence from U–Pb dating of zircon. *International Journal of Earth Sciences* **93**, 844–59.

SCHERMER, E. R., LUX, D. R. & BURCHFIELD, B. C. 1990. Temperature–time history of subducted continental crust, Mount Olympos region, Greece. *Tectonics* **9**, 1165–95.

SCHNEIDER, W., BODE, S. & OPPERMANN, A. 1998. Zusammenhänge zwischen Deckenvorschub, Suturprogradation, Erosion und Flyschzusammensetzung in den Helleniden. *Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen* **209**, 349–79.

SIRCOMBE, K. N. & STERN, R. A. 2002. An investigation of artificial biasing in detrital zircon U–Pb geochronology due to magnetic separation in sample preparation. *Geochimica et Cosmochimica Acta* **66**, 2379–97.

SKOURLIS, K. & DOUTSOS, T. 2003. The Pindos fold-and-thrust belt (Greece): inversion kinematics of a passive continental margin. *International Journal of Earth Sciences* **92**, 891–903.

SMITH, A. G. 1993. Tectonic significance of the Hellenic–Dinaric ophiolites. In *Magmatic Processes and Plate Tectonics* (eds H. M. Prichard, T. Alabaster, N. B. W. Harris and C. R. Neary), pp. 213–43. Geological Society of London, Special Publication no. 76.

SMITH, A. G., WOODCOCK, N. H. & NAYLOR, M. A. 1979. The structural evolution of a Mesozoic continental margin, Othris Mountains, Greece. *Journal of the Geological Society, London* **146**, 589–603.

STAMPFLI, G. M., BOREL, G., CAVAZZA, W., MOSAR, J. & ZIEGLER, P. A. (eds) 2001. *The palaeotectonic atlas of the Peritethyan Domain*. European Geophysical Society (CD-ROM).

STAMPFLI, G. M., ROSSELET, F. & BAGHERI, S. 2004. Tethyan oceans and sutures. *Proceedings, 5th International Symposion on Eastern Mediterranean Geology, Thessaloniki*, 193–6.

THIEBAULT, F., FLEURY, J., CLEMENT, B. & DEGARDIN, J. 1994. Paleogeographic and paleotectonic implications of clay mineral distribution in Late Jurassic–Early Cretaceous sediments of the Pindos–Olonos and Beotian Basin, Greece. *Palaeogeography, Palaeoclimatology, Palaeoecology* **108**, 23–40.

VERGÉLY, P. 1984. *Tectonique des ophiolites dans les Hellénides internes (déformations, métamorphismes et phénomènes sédimentaires). Conséquences sur l'évolution des régions téthysiennes occidentals*. Published thesis, Universite Paris Sud, 661 pp.

VERMEESCH, P. 2004. How many grains are needed for a provenance study? *Earth and Planetary Science Letters* **224**, 441–51.

WAGREICH, M., PAVLOPOULOS, A., FAUPL, P. & MIGIROS, G. 1996. Age and significance of Upper Cretaceous siliciclastic turbidites in the central Pindos Mountains, Greece. *Geological Magazine* **133**, 325–31.