Provenance of sediments during subduction of Palaeotethys: Detrital zircon ages and olistolith analysis in Palaeozoic sediments from Chios Island, Greece

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ABSTRACT

Detrital zircon geochronology and analysis of fossiliferous olistoliths from Chios Island, Greece, are used here to constrain terrane accretion processes and the provenance of crustal sources for sediments during the subduction of (a branch of) the Palaeotethys Ocean. U/Pb ages obtained by ion microprobe (SHRIMP-II) analyses of detrital zircons from a Carboniferous greywacke belonging to the tectonostratigraphic Lower Unit of Chios gave major age groups of 2150–1890 Ma, 640–540 Ma, 505–475 Ma and 365–322 Ma. Detrital zircons from a Permian–Triassic sandstone yielded prominent age clusters of 2200–1840 Ma, 1100–910 Ma, 625–560 Ma and 385–370 Ma. The lack of zircon ages between 1.8 and 1.1 Ga in both samples, coupled with the occurrence of ca. 2 Ga-old zircons, imply a northern Gondwana (NW Africa) source. The conodont fauna recovered from an ‘Orthoceras’-bearing limestone from the Carboniferous succession of the Lower Unit indicates a Late Silurian age. The fauna is typical of the Ludfordian Polygnathoides siliculosus conodont zone. The material has a conodont colour alteration index (CAI) of about 1–2, indicating very low-grade thermal alteration of less than 100 °C. The closest localities with similar conodont-bearing limestones are to be found in the Balkan region and in the Istanbul Zone of northern Turkey. The occurrence of ‘Orthoceras’ Limestone can be used as an indicator of palaeosource reconstruction. Our new zircon ages in conjunction with provenance analysis of Silurian to Lower Carboniferous olistoliths strongly suggest that the clastic succession of Chios received its detritus from basement rocks of the Sakarya microcontinent in western Turkey and time and facies equivalents of Palaeozoic units from the Istanbul Zone in northern Turkey and the Balkan region due to subduction of a branch of Palaeotethys close to the southern active margin of Eurasia in Late Palaeozoic times. The multidisciplinary approach of this study underlines the importance of terrane accretion during stepwise closure of Palaeotethys.

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1. Introduction

Palaeogeographical reconstructions show the existence of two major oceanic realms in the eastern Mediterranean area: the Palaeotethys and the Neotethys (e.g., Sengör et al., 1984;Stampfli, 2000 and references therein). Following Stampfli and Borel (2002), the term Palaeotethys is used here to denote a seaway that separated Gondwana from fragments thereof in a period from the Silurian to early Late Triassic, during which these same fragments drifted northward and accreted to Laurussia in a stepwise fashion. Palaeotethys was closed by the northward drift of the Cimmerian terranes in response to the opening of Neotethys in the south. Despite the accumulation of new data over recent years, the Palaeozoic to early Mesozoic history of Palaeotethys is strongly debated (see Robertson et al., 1996; Stampfli, 2000; Stampfli and Borel, 2002; Robertson et al., 2004) because evidence for subduction, in terms of accretionary complexes, blueschists and magmatic arcs, is scarce. Moreover, the complex Mesozoic to Cenozoic structural and metamorphic overprint gave rise to equivocal palaeotectonic models and interpretations. Two of the most controversial topics are (i) the location of peri-Gondwana fragments such as Pelagonia, Sakarya and the Anatolide–Tauride Block (Fig. 1) within the Palaeotethyan frame and (ii) the precise location and polarity of subduction of Palaeotethys under those fragments in late Palaeozoic times; it could have been either northward (e.g., Stampfli et al., 1991, 2003; Eren et al., 2004) or southward (e.g., Şengör et al., 1984; Okay et al., 1996; Romano et al., 2006; Xypolias et al., 2006).

The island of Chios in the eastern Aegean Sea (Figs. 1 and 2) is a key area for understanding the closure of Palaeotethys (e.g., Stampfli et al.,
2003) because it is one of the rare localities where very low-grade to virtually unmetamorphosed fossil-bearing Palaeozoic to Mesozoic sequences are preserved (e.g., Besenecker et al., 1968). A similar succession of late Palaeozoic age can be found in western Turkey opposite Chios in the Karaburun peninsula (Kozur, 1998; Robertson and Pickett, 2000 and references therein), in the Tavas nappe (Lycian nappes) further south (Kozur et al., 1998) and in the Konya region at the northern margin of the Menderes–Tauride Platform (Eren et al., 2004). Several tectonic models have been proposed for the origin of the Chios–Karaburun units, including that of an accretionary complex or a rift setting (see Stampfl et al., 1991, 2003; Robertson and Pickett, 2000; Zanchi et al., 2003).

In the present study, special emphasis has been placed on constraining the provenance of Carboniferous and Permian–Triassic clastic sediments from the Lower Unit of Chios by investigating the age spectra of detrital zircons using the sensitive high-resolution ion microprobe (SHRIMP-II) U–Pb dating method. SHRIMP is a powerful tool in sedimentary provenance analysis to glean information regarding ancient source areas and major magmatic events, crucial for Tethyan plate-tectonic and palaeogeographic reconstructions. The provenance of olistoliths found within the Carboniferous succession is also discussed in order to better constrain the possible sources identified by zircon dating. Our data provide new information about crustal sources and terrane accretion during diachronous closure of Palaeotethys and help clarify the palaeotectonic position of Chios during subduction of this ocean.

2. Geological setting

In the eastern Mediterranean, the Hellenides are an integral part of the Alpine–Himalayan orogenic system and have traditionally been subdivided into the internal (hinterland) and the external (foreland) Hellenides. The internal Hellenides comprise, from SW to NE, the Pelagonian Zone (including the Attic-Cycladic Massif), the Vardar Zone, the Serbo-Macedonian Massif and the Rhodope Massif (e.g., Jacobshagen, 1986, and references therein). All these units consist predominantly of Palaeozoic (but also Neoproterozoic and Mesozoic) basement rocks (e.g., Vavassis et al., 2000; Anders et al., 2005, 2006a,b; Turpaud and Reischmann, 2005; Turpaud, 2006; Himmerkus et al., 2004a,b, 2006a,b, 2007) overlain by or intercalated with sedimentary successions. They experienced quite complex Mesozoic to Cenozoic tectonism, which gave rise to equivocal palinspastic models and interpretations. Chios is generally assigned to the easternmost part of the Pelagonian Zone (e.g., Jacobshagen, 1986; see Meinhold et al., 2007, for discussion). Basically, it comprises two tectonostatigraphic units (Figs. 2 and 3): an ‘autochthonous’ Lower Unit and a tectonically overlying ‘allochthonous’ Upper Unit (Herget and Roth, 1968; Besenecker et al., 1968). The Lower Unit consists of clastic sediments of Late Palaeozoic age containing blocks (of up to 100 m in diameter) of limestones, radiolarites and volcanic rocks. This succession was variably named ‘Chios mélangé’, ‘Chios (wild)flysch’ or ‘Volissos turbidites’ (e.g., Robertson and Pickett, 2000; Groves et al., 2003; Zanchi et al., 2003). Here, we use the non-genetic term Upper Palaeozoic rocks of the Lower Unit (Meinhold et al., 2007). The major rock types of this unit are greywackes, minor sandstones and siltstones as well as intercalated quartz-bearing conglomerates (Fig. 4). The latter mainly contain clasts of quartz, black chert and quartzite embedded in a coarse-grained quartzo-feldspathic matrix. Erosional contacts at the base of greywacke and conglomerate beds and upward reduction in grain size can often be observed. In some outcrops, well-developed sole marks and ripples can be seen at the base of turbidite beds (Fig. 4). The facies of the turbidite–olistostrome succession resulted mainly from turbidity currents, debris flows and submarine slides. Groves et al. (2003) reported Mississippian microfossils from calcareous clasts of a breccia
lying within the turbidite–olistostrome succession and suggested that the Upper Palaeozoic rocks of the Lower Unit are most probably Late Visean or Early Serpukhovian in age.

The Upper Palaeozoic rocks of the Lower Unit were interpreted by Papanikolaou and Sideris (1983) as a post-Middle Carboniferous to pre-Scythian wildflysch bearing Silurian to Carboniferous olistoliths. Stampfli et al. (1991) proposed that the wildflysch represents a predominantly Permian accretionary wedge that can be correlated with the Karakaya Complex of Turkey, which they considered to mark a Palaeotethyan suture. Robertson and Pickett (2000) maintained that the Upper Palaeozoic rocks of the Lower Unit are a Late Carboniferous to ?Early Permian subduction–accretionary complex that developed near the southern margin of the Palaeotethys Ocean. Models invoking a rift setting or a long-lived, deep marine basin receiving olistoliths are also discussed in Robertson and Pickett (2000). Recently, Stampfli et al. (2003) and Zanchi et al. (2003) proposed a tectonic model of an arc-trench system that formed along the Palaeotethyan subduction zone in Carboniferous times. Meinhold et al. (2007) suggested a tectonic setting of a continental volcanic arc, involving upper-continental crustal rocks and/or young differentiated arc material as well as a mixed source of (ultra)basic rocks.

Non-fossiliferous conglomerates and sandstones overlie the Upper Palaeozoic rocks of the Lower Unit (Fig. 3). The conglomerates mainly contain clasts of quartz and black chert embedded in a coarse-grained quartzose matrix. Zanchi et al. (2003) described an angular unconformity at the base of this clastic succession, interpreted as a Variscan unconformity. However, the stratigraphic age of these possibly Lower Triassic conglomerates and sandstones is uncertain because neither biostratigraphic nor geochronological ages exist so far. Following Meinhold et al. (2007) we use the term Permian–Triassic instead of Lower Triassic. The conglomerates and sandstones pass upwards into limestones of late Early to Late Triassic–Early Jurassic age (Besenecker et al., 1968). A Middle Triassic volcano–sedimentary succession is intercalated with the limestones (Besenecker et al., 1968; Gaetani et al., 1992). Non-marine clastic deposits followed by dolomites and limestones mark the uppermost Triassic–lowermost Jurassic.

The Upper Unit of Chios predominantly comprises siliciclastic and carbonaceous rocks of Late Carboniferous and Permian ages (e.g., Besenecker et al., 1968), overlain by transgressive sediments of Early Jurassic age, which are locally overlain themselves by transgressive deposits of Late Cretaceous age. The youngest rocks on Chios are Cenozoic sediments and volcanic rocks cropping out mainly in the southeastern part of the island.

3. Geochronological method

Zircon separation was carried out using standard techniques. The samples were crushed with a hydraulic press and a rotary mill. Zircon enrichment was achieved through hydro-gravimetric (Wilfley table), magnetic (Frantz isodynamic separator) and density (methylene iodide) separation techniques followed by handpicking under a binocular microscope. Zircon grains were mounted in epoxy resin, sectioned and polished together with grains of zircon standard 91500 (ID-TIMS age = 1065.4±0.3 Ma; Wiedenbeck et al., 1995) for calibration.
of the U concentrations and TEMORA 1 (ID-TIMS age = 416.75 ± 0.24 Ma; Black et al., 2003) for calibration of the U–Pb ratios. Prior to the analyses the mounts were gold-coated, photographed under the microscope using both reflected and transmitted light and CL-imaged to reveal the internal structure of the zircons and target specific areas within single crystals, e.g. growth zones, inherited cores, etc. (Fig. 5). Measurements were carried out on the SHRIMP-II mass spectrometer using a secondary electron multiplier operated in a mass-scanning mode at the Centre of Isotopic Research in St. Petersburg, Russia. During U–Pb–Th analysis, the spot size of the O2− primary ion beam was set at about 25 µm. The primary beam intensity was about 4 nA. Data reduction and presentation were processed with SQUID (Ludwig, 2001) and Isoplot/Ex (Ludwig, 2003), using decay constants recommended by IUGS (Steiger and Jäger, 1977). The common lead correction was done by using the measured 204Pb and the model of Stacey and Kramers (1975). Unless otherwise stated 206Pb/238U ages are quoted for analyses younger than 1.2 Ga, whereas 207Pb/206Pb ages are quoted for analyses older than 1.2 Ga. The latter are generally considered as being minimum ages. The analytical data referred to in this paper are available as Supplementary data (see Appendix).

4. Sample description and results

4.1. Carboniferous sample CH52

Sample CH52 was collected from a Carboniferous bluish-grey, medium-grained litharenite (greywacke) cropping out east of Volissos town (Fig. 2). The framework grains are monocrystalline and polycrystalline quartz, and sedimentary and volcanic lithoclasts. Plagioclase grains often display patchy sericitisation and calcite replacement. Single muscovite flakes and minor detrital biotite also occur. The greywacke is poorly sorted by size and dominated by angular to subangular clasts. Accessory minerals include predominantly zircon, rutile, tourmaline and opaque minerals as well as minor chrome spinel and epidote.

Most of the analysed zircons have clear oscillatory growth zoning and appear to be magmatic in origin whereas few are structureless. Inherited cores are rare. The length of single zircon crystals varies between ca. 80 and 290 µm, and their morphology ranges from long prismatic to isometric. Around 65% of the detrital zircon grains have an aspect ratio (length/width) of between 1 and 2. The Palaeoproterozoic to
Neoproterozoic zircons have a well-rounded to rounded shape; the Cambrian–Ordovician to Carboniferous zircons are slightly rounded, subhedral to euhedral. Euhedral zircons are clear. The (110) prism generally dominates over the (100) prism, as does the (221) bipyramid over the (101) bipyramid, corresponding to the S1, S6, S12 and S17 zircon types of Pupin (1980). Such a morphology is typical of an igneous origin and suggests crystallisation temperatures of about 650–700 °C in a magma of crustal origin. The presence of euhedral detrital zircons may
be taken to indicate the last input of magmatic zircons. These zircons are of first-order cycle. Some well-rounded zircons have a pitted surface, which clearly indicates long-distance sedimentary transport.

With the exception of a few grains, the analysed zircons have low to medium U contents (94–661 ppm) and moderate Th/U ratios (0.12–1.19). These values are typical of zircons from acidic igneous rocks (Fig. 6). Three grains seem to be of metamorphic origin with Th/U ratios <0.1. Analysed grains gave four prominent zircon populations that range in age from Early Proterozoic to Phanerozoic. Sample CH52 contains zircon clusters of ca. 2150–1890, 640–540, 505–475 and 365–322 Ma (Figs. 7 and 9). Grain CH52.1.1 has a concordant age of 2542±11 Ma, which indicates the presence of a minor Archaean component. The 206Pb/238U age of 1422±8 Ma of grain CH52.20.1 is strongly discordant and thus may not be geologically meaningful since the time of Pb loss cannot be determined. By contrast, its 207Pb/206Pb age of 1890±18 Ma appears to be geologically significant. Grain CH52.22.1 has a concordant 206Pb/238U age of 322±3 Ma. This zircon crystal is euhedral, clear and has oscillatory growth zoning (Fig. 5) with U and Th values (Table 1) typical of magmatic origin. It is characterized by the dominance of the (100) prism over the (110) prism. Only the (101) pyramid is present. This type of morphology corresponds to the P5 zircon type of Pupin (1980) and indicates crystallisation temperatures of about 800–850 °C in an alkaline magma. Grain CH52.22.1 can therefore be considered as representing the youngest detrital input, thus giving a maximum age of deposition for the Carboniferous greywacke. Four zircon grains (CH52.2.1, CH52.4.1, CH52.10.1, and CH52.21.1) have 206Pb/238U ages of ca. 273 to 206 Ma much younger than the age of the host sediment. These zircons also have very high concentrations of U (909–3453 ppm), Th (370–2016 ppm) and common Pb (3.95–24.99%). Their ages are, therefore, interpreted as the result of radiogenic lead loss probably caused by radiation damage (see Kolodner et al., 2006) and as soiled with common Pb, and were consequently not considered in provenance analysis.

4.2. Permian–Triassic sample CH11

Sample CH11 is a Permian–Triassic, yellowish-grey, medium-grained quartz arenite (sandstone) collected from a quartzose conglomerate.
layer cropping out northwest of Langada village (Fig. 2). Monocrystalline and polycrystalline quartz are the major framework components. The heavy minerals observed are characterised by the stable assemblage zircon + rutile + tourmaline, indicating sediment maturity. Chrome spinel grains were also found with Cr-number (Cr/(Cr+Al)) values between 0.43 and 0.83 and Mg-number (Mg/(Mg+Fe²⁺)) values between 0.24 and 0.68, suggesting a probably mixed (ultra)basic source involving ridge peridotites, fore-arc peridotites and island-arc basalts (Meinhold et al., 2007). The occurrence of iron-hydroxy-oxides indicates secondary alteration processes. Description of zircon structure morphology fits that of sample CH52 above. However, the length of single zircon crystals varies between ca. 80 and 230 µm with only a few long prismatic grains. Around 70% of the detrital zircon grains have an aspect ratio of between 1 and 2. Only a few crystals are euhedral that could be used for determining the type. The (100) prism of the euhedral zircons dominates over the (110) prism, as does the (101) bipyramid over the (221) bipyramid, corresponding to the S19, S20 and S24 zircon types of Pupin (1980). Such a morphology is typical of an igneous origin and suggests crystallisation temperatures of about 800–850 °C in a calc-alkaline to sub-alkaline granitic magma.

In this sample, the analysed zircons have low to medium U contents (35–745 ppm) and, except one grain, moderate Th/U ratios (0.11–1.45). These values are typical of zircons from acidic igneous rocks (Fig. 6). The zircon ages gave populations of ca. 2200–1840, 1100–910, 625–560 and 385–370 Ma (Figs. 8 and 9). Grain CH-11.6 has an inherited core with a magmatic rim of ca. 475 Ma. In a concordia diagram, a regression of the core and the rim yields a discordia with a lower limit for metamorphic zircon (diagram modified after Treuil et al., 2004).

5. Discussion

5.1. Possible source areas according to zircon data

The U/Pb data for detrital zircons from Chios indicate the presence of at least three Proterozoic and a number of Palaeozoic crustal sources. The period between ca. 1.8 and 1.1 Ga is characterised by a total absence of tectonomagmatic activities in the source area of the analysed zircons (Fig. 9). The oscillatory zoning pattern of most of the zircons and the Th/U ratios between 0.1 and 1.2 indicate a source of mainly magmatic origin. To facilitate identification of possible source areas for the Chios samples presented herein we have compiled the distribution of Late Neoproterozoic and Palaeozoic zircon age data from Greece and the surrounding region (Table 3 and Fig. 10). In the ensuing discussion, we make inferences as to possible source areas for the detrital zircon age clusters recorded in the analysed sediments from Chios.

5.1.1. ca. 2.2–1.9 Ga zircon ages

Early Proterozoic (ca. 2 Ga) ages are commonly present among zircons in terranes derived from northern Gondwana (e.g., Zeh et al., 2001; Linnemann et al., 2004 and references therein) as the result of the Icartian–Burmian crust formation event that is known from the West African Craton (e.g., Ennih and Liégeois, 2001; Egal et al., 2002; Walsh et al., 2002), from South America (Keppie et al., 1998 and references therein) and also from Brittany and the Channel Islands in western Europe (Calvez and Vidal, 1978; Samson and D’Lemos, 1998). The lack of late Palaeoproterozoic and Mesoproterozoic zircon ages is often used as a strong argument in favour of a NW African provenance of Precambrian peri-Gondwanan basement terranes (e.g., Zeh et al., 2001; Friedl et al., 2004; Linnemann et al., 2004; Samson et al., 2005). In the case of the Chios sediments then, the lack of detrital zircons with ages between 1.8 and 1.1 Ga, combined with the high abundance of grains aged around 2 Ga can be interpreted as pointing towards a NW African provenance. Moreover, the total absence of zircons with ages of between ca. 1.5 and 1.7 Ga and ca. 2.6–3.7 Ga (e.g., Gorbatchev and Bogdanova, 1993; Bogdanova et al., 2007) indicates that a detrital input from Baltica can safely be excluded.

5.1.2. ca. 1.1–0.9 Ga zircon ages

Ages around 1 Ga are generally restricted to the Grenvillian–Kibaran orogenic belts that resulted from the amalgamation of the Rodinia supercontinent (e.g., Dalziel, 1997). Kemp et al. (2006) have recently shown that remelting of old supracrustal rocks was significant during that period. The Grenvillian orogenic belt can be traced from SW Baltic and eastern Greenland over to western Amazonia and the southern part of North America during the Neoproterozoic (Dalziel, 1992). A branch of this belt also extends along the Africa–Antarctica margin and into eastern India and Western Australia. In the West African Craton and the “Saharan Metacraton” ages around 1.1–0.9 Ga have not been hitherto reported (e.g., Nance and Murphy, 1994; Ennih and Liégeois, 2001; Egal et al., 2002; Abdelsalam et al., 2002). They are, however, known from numerous peri-Gondwanan terranes in Europe (e.g., Moldanubia: Gebauer et al., 1989; Friedl et al., 2004; Mid-German Crystalline Rise: Zeh et al., 2001; Iberia: Fernández-Suárez et al., 2000, 2002). Detrital zircon ages of ca. 1 Ga have been reported from metasedimentary rocks from the Cyclades Islands (Keay and Lister, 2002) and from Crete (Romano et al., 2005). Infracambrian and Carboniferous metasedimentary rocks in the Karachaihar and Sandikli
areas of the Menderes–Taurus block contain ca. 990 Ma zircons (Kröner and Şengör, 1990). Avigad et al. (2003) and Kolodner et al. (2006) described 1.1–0.9 Ga zircons in Cambrian–Ordovician sandstones from Israel and Jordan. Inherited zircons of comparable ages were recently reported for orthogneisses from the Sredna Gora Zone in Bulgaria (Carrigan et al., 2006). The presence of ca. 1 Ga zircons in European peri-Gondwanan crustal fragments and in the northern part of the Arabian–Nubian Shield, thousands of kilometres away from any known source, is still a matter of discussion (see Zeh et al., 2001; Avigad et al., 2003; Gutiérrez-Alonso et al., 2005; Kolodner et al.,

**Fig. 7.** Concordia diagrams showing the results of SHRIMP-II U–Pb zircon analyses for sample CH52. Error ellipses are drawn at the 1-sigma level. The inset shows the enlarged concordia plot for 206U/238Pb ages younger than 800 Ma.

**Fig. 8.** Concordia diagrams showing the results of SHRIMP-II U–Pb zircon analyses for sample CH11. Error ellipses are drawn at the 1-sigma level. The inset shows the enlarged concordia plot for 206U/238Pb ages younger than 1.2 Ga.
2006), Avigad et al. (2003) suggested that they were derived from the Kibaran Belt of central Africa. They proposed that Neoproterozoic glaciers might have transported a large quantity of detritus toward the Gondwana margins where it was later reworked and deposited in the Lower Palaeozoic siliciclastic cover sequences. Keppie et al. (1998) noted that river systems can travel across several cratons, suggesting that the provenance of the zircons might be sought farther afield. Therefore, it might also be possible that the Grenvillian detritus was transported by continental rivers similar to the present-day Amazon or Nile, towards the northern margin of Gondwana, maybe prior to the Cadomian event (Zeh et al., 2001). An alternative scenario was recently suggested by Gutiérrez-Alonso et al. (2005), proposing that Grenvillian zircons in both Iberia and the northern Congo Craton have an ultimate South American (Amazonian and Oaxaquian) source that was widely dispersed after the amalgamation of northern Gondwana. Generally, it seems that Grenvillian–Kibaran zircons have accumulated in post-Cadomian (Ediacaran–Cambrian–Ordovician) cover sequences at the northern margin of Gondwana (Fernández-Suárez et al., 2000, 2002; Avigad et al., 2003; Gutiérrez-Alonso et al., 2005; Kolodner et al., 2006). These sequences may have been reworked during younger orogenic processes, then re-sedimented, thus supplying inherited zircons to younger sediments. In this way, the ca. 1 Ga-old, well-rounded zircons in the Chios sediments studied here probably have a secondary source due to polyphase recycling.

5.1.3. ca. 640–540 Ma zircon ages
Zircons yielding 640–540 Ma ages are probably related to late Pan-African orogenic processes, i.e. to the so-called Avalonian–Cadomian belt (Nance and Murphy, 1994). Magmatic rocks of this event are generally interpreted as proof of a Neoproterozoic to earliest Palaeozoic Andean-type subduction-zone setting at the northern margin of Gondwana (Nance and Murphy, 1994). Cadomian ages are known from orthogneisses in the Menderes Massif of western Turkey (ca. 521–590 Ma, Pb–Pb and U–Pb on zircon: Kröner and Şengör, 1990; Hetzel and Reischmann, 1996; Loos and Reischmann, 1999; Gessner et al., 2004), from units in the Serbo-Macedonian Massif of northern Greece (ca. 540–587 Ma, Pb–Pb on zircon: Himmerkus et al., 2004a, 2006a, 2007) and of SW Bulgaria (ca. 544–569 Ma, U–Pb on zircon: von Quadt et al., 2000; Graf, 2001; Kounov, 2002) and from the Istanbul Zone of NW Turkey (ca. 560–590 Ma, Pb–Pb and U–Pb on zircon: Chen et al., 2002; Ustaömer et al., 2005). Recently, Carrigan et al. (2006) reported concordia ages of 616.9±9.5 Ma and 595±23 Ma for zircons in orthogneisses from the Sredna Gora Zone of Bulgaria. With regard to the latter sample, however, Carrigan et al. (2006) cautioned that its Late Neoproterozoic age of ~600 Ma is not precise because of its widely varying U concentrations and Th/U values. Using all data available at present, it can be concluded that the 640–540 Ma zircons were probably derived from the Avalonian–Cadomian belt and that their source can actually be sought in the terranes mentioned above. An input from the Arabian–Nubian Shield is highly unlikely since this terrane, besides containing 650–550 Ma igneous rocks, is also largely characterised by 900–700 Ma basement rocks (e.g., Kröner et al., 1987, 1990), and zircon ages in the latter range are absent from our Chios data set.

5.1.4. ca. 505–460 Ma zircon ages
Upper Cambrian to Lower Ordovician magmatic rocks are widespread in the pre-Variscan basement of Europe, and have been attributed to rift processes and the opening of the Rheic ocean, separating Avalonia from the northern margin of Gondwana (e.g., Crowley et al., 2000; von Raumer et al., 2002). Some Lower Ordovician calc-alkaline metaluminous granites of arc origin are probably related to a Cambrian–Ordovician active-margin setting (von Raumer et al., 2003). Relatively younger (~450 Ma) magmatic rocks characterise a late- to post-orogenic evolution (von Raumer et al., 2003). Basement rocks with zircon ages of 490–460 Ma are not known so far from the Hellenides of Greece. However, the involvement of an Ordovician crustal source during Carboniferous–Permian magma genesis was recently shown by Anders et al. (2006b) for orthogneisses from the Pelagonian Zone and is also documented in orthogneisses from the Eastern Rhodope of Greece (N. Cornelius, pers. comm. to G. Meinhold). Furthermore, two inherited zircons of 472.4±5 Ma and 481.5±4.8 Ma
were reported for a Middle Jurassic gabbro from the Vourinos ophiolite (Liati et al., 2004). This implies that basement or older sedimentary rocks containing Early–Middle Ordovician zircons were reworked during Neotethyan magmatic activity in Jurassic times. Zircons in basement gneisses from the Sredna Gora Zone in central Bulgaria yielded ages of 480±30 Ma and 485±50 Ma (Arnaudov et al., 1989, cited in Peytcheva and von Quadt, 2004). Furthermore, Carrigan et al. (2005) reported Ordovician ages (~445–467 Ma) from inherited cores of Variscan zircons in metagranites of the Sredna Gora Zone. Zircons in metagranites from the Serbo-Macedonian Massif in SW Bulgaria yielded an upper intercept age of 459.9±7.6 Ma, which was interpreted as the intrusion age (Titorenkova et al., 2003). A Pb/Pb single-zircon evaporation age of around 462 Ma was documented from orthogneisses from the Biga Peninsula in NW Turkey (Özmen and Reischmann, 1999). The Biga Peninsula belongs to the Sakarya microcontinent and constitutes part of the Sakarya Composite Terrane (Gönçüoğlu et al., 1997). It should also be noted here that earlier K–Ar mineral and Rb–Sr whole-rock dating provides evidence that a magmatic or metamorphic event around 460 Ma affected basement rocks in the Pelagonian Zone of Greece (ca. 461 Ma and 465 Ma, K–Ar: Marakis, 1970; ca. 460 Ma, Rb–Sr: Henjes-Kunt and Kreuzer, 1982). However, further work is required to establish the significance of these ages.

5.1.5. ca. 385–370 Ma zircon ages

Devonian ages between 385 and 370 Ma have been reported from several parts of Central and Eastern Europe. Such ages reflect either accretion of Gondwana-derived terranes to Baltica, resulting in HT–MP metamorphism and partial melting during a single-cycle Variscan orogeny (e.g., Góry Sowie, Sudetes: Timmermann et al., 2004) or the timing of Variscan exhumation and associated decompression melting under upper amphibolite-facies to granulite-facies conditions (e.g., Mariánské Lázne Complex: Timmermann et al., 2004). The accretion of Gondwana-derived terranes to Laurussia marks the onset of Variscan orogeny. The spatial distribution and temporal variation of Devonian ages in the eastern Mediterranean realm is not yet accurately known. Crystalline basement rocks of Devonian age have
only been reported from one locality in the Hellenides (Andriesen et al., 1987). These authors documented an upper intercept age of 372 Ma+28/−24 Ma (U–Pb method) on zircons from the core gneisses of Naxos. Nonetheless, incorporation of Devonian basement rocks into Carboniferous–Permian magma genesis cannot be excluded since Keay and Lister (2002) documented inherited zircons of ca. 370–413 Ma from orthogneisses of the Attic-Cycladic Massif. By contrast, Devonian ages are well known from the Biga peninsula in NW Turkey. Zircons from the Çamlık metagranodiorite gave a Pb/Pb single zircon age of 397.5±1.4 Ma (Okay et al., 2006), confirming an earlier Pb/Pb age of 399±13 Ma by Okay et al. (1996). Pb/Pb single-zircon evaporation ages around 372 Ma were obtained from granitic–granodioritic gneisses cropping out in the northern part of the Biga peninsula close to the Sea of Marmara and in the south near Edremit town (Özmen and Reischmann, 1999). Taking into account the data available, we suggest that basement rocks from the Sakarya microcontinent were probably the source rocks of the 370–385 Ma-old detrital zircons of the Chios clastic successions.

5.1.6. ca. 350–320 Ma zircon ages

Lower Carboniferous magmatic rocks are widespread in the Central European Variscides. They reflect ongoing Variscan exhumation and associated decompression melting under upper-amphibolite to granulite-facies conditions at around 340 Ma (e.g., Kröner et al., 2000, and references therein) and regional low pressure–high temperature metamorphism, migmatisation and granitic magmatism between 330 and 320 Ma (e.g., Anthes and Reischmann, 2001, and references therein). By contrast, Lower Carboniferous magmatic rocks are rare in the Hellenides. They have been documented for basement rocks of the Lower Tectonic Unit of the western Rhodope Massif (357±20 Ma, U–Pb on zircon: Wawrzenitz, 1997 cited in Liati and Gebauer, 1999; 345±40 Ma, Pb–Pb on zircon: Kokkinakis, 1978 cited in Vavassis et al., 2000). This, however, has to be revised since Turpaud and Reischmann (2005) and Turpaud (2006) found latest Carboniferous–Early Permian ages in those regions. Recent geochronological studies in the Lower Tectonic Unit of the eastern Rhodope Massif show minor evidence of Lower Carboniferous magmatic rocks (ca. 325 Ma, U–Pb on zircon: N. Cornelius, pers. comm. to G. Meinhold). Early Carboniferous ages, however, are well documented for basement rocks from the Pelagonian Zone (Evia Island, 319±0.7, U–Pb on zircon: Vavassis et al., 2000), the Attic-Cycladic Massif (Paros Island, 325±4 Ma, Pb–Pb on zircon: Engel and Reischmann, 1998; Naxos Island, 316±4 Ma, Pb–Pb on zircon: Reischmann, 1998; 319±1 Ma, 322±2 Ma, U–Pb on zircon: Keay et al., 2001; Sikinos Island, 325±4 Ma, Pb–Pb on zircon: Engel and Reischmann, 1999; Delos Island, 327±4 Ma, Pb–Pb on zircon: Engel and Reischmann, 1999) and recently for orthogneisses from Kithira Island, External Hellenides (324±2 Ma, 323±3 Ma and 320±1.2 Ma, U–Pb on zircon: Xypolias et al., 2006). Carrigan et al. (2006) reported a concordia age of 336.5±5.4 Ma for metamorphic rims of a migmatitic leucosome from the Sredna Gora Zone in Bulgaria that they interpret as the age of zircon recrystallisation during high-grade metamorphism. Recently, Pb/Pb single-zircon evaporation analyses of zircons from a gneiss sample of the Kazdağ Massif on the Biga Peninsula in NW Turkey gave a relatively precise age of 319.2±1.5 Ma (Okay et al., 2006). Thus, the Early Carboniferous zircons found in the clastic sediments of Chios (322±3 Ma from sample CH52 and 326±2 Ma from sample CH11) could have been derived from basement rocks of the Pelagonian Zone/Attic-Cycladic Massif, from Sakarya or the Lower Tectonic Unit of the Rhodope Massif.

![Fig. 11. Compilation of lithology and fossil content for sedimentary olistoliths from the Upper Palaeozoic rocks of the Lower Unit and their interpretation in terms of depositional environment. Data sources: Kauffmann (1965), Besenecker et al. (1968), Herget (1968), Herget and Roth (1968), Fenninger (1983), Robertson and Pickett (2000), Groves et al. (2003), Larghi et al. (2005). Geological time scale according to Gradstein et al. (2004).](image-url)
5.2. Source of zircons

The analysis of magmatic, inherited and detrital zircon ages from different parts of the Eastern Mediterranean helps to identify possible sources for the clastic sediments of Chios. The occurrence of ca. 2 Ga-old zircons, in conjunction with the lack of detrital zircons with ages between 1.8 and 1.1 Ga in our studied sediments from Chios, characterise a northern Gondwana-derived source, especially one from NW Africa. Zircons of Late Neoproterozoic (~540–650 Ma) and Palaeozoic (~280–330 Ma, ~370 Ma, ~430–460 Ma, ~480 Ma) ages are important time-markers for palaeotectonic reconstructions in the Eastern Mediterranean (Fig. 10).

Carboniferous–Permian zircons (ca. 280–320 Ma) were not found in the analysed sediments from Chios, implying that rocks of such an age were either never present or not yet exposed in the source area at the time of sedimentation. The youngest detrital zircon grains in samples CH11 and CH52 are 326±2 Ma and 322±3 Ma respectively. These zircons are slightly younger than 330–335 Ma-old detrital zircons from the Küçükbaşçe Formation of the Karaburun peninsula (Rosselet and Stampfli, 2003). Both Late Carboniferous and earliest Permian zircons (ca. 280–330 Ma) document a major igneous event in the Aegean and the surrounding area (e.g., Reischmann 1998; Engel and Reischmann, 1998, 1999; Vavassis et al., 2000; Reischmann et al., 2001; Anders et al., 2006b). Geochemical analyses indicate a volcanic arc or active continental-margin setting at this time (e.g., Reischmann et al., 2001; Pe-Piper and Piper, 2002; Anders, 2005), due to northward subduction of (a branch of) Palaeeotethys beneath Pelagonia and Sakarya (e.g., Vavassis et al., 2000; Stampfli et al., 2003). Hence, the provenance of the analysed detrital zircons from Chios can be constrained as follows:

Sample CH52 yielded abundant magmatic zircons of Late Neoproterozoic–Cambrian, Ordovician and minor Devonian ages, in addition to very old (~2 Ga) grains. The External Hellenides, the Pelagonian Zone and the Attic-Cycladic Massif consist of Carboniferous–Permian (ca. 280–330 Ma) basement with only rare evidence for older magmatic rocks (see references above), if one excludes the Florina terrane of Anders et al. (2006a). The latter authors suggest that the Florina terrane most probably belonged to East Avalonia and formed the continental basement (ca. 699–713 Ma) onto which the Pelagonian Carboniferous–Permian magmatic arc was founded. The data available suggest that the Pelagonian Zone and the Attic-Cycladic Massif cannot be considered as a possible source of zircons during Carboniferous times. Detrital input from the Florina terrane is also not documented by the Chios clastic sediments. The source should rather be sought in areas lying to the north or northeast of present-day Chios such as NW Turkey or the Balkan region. Detrital zircons of around 540–640 Ma were probably derived from basement rocks of the Avalonian–Cordoman belt. Terrane fragments of these are preserved in northern Greece, Bulgaria and northern Turkey (Fig. 10). Detrital zircons with Ordovician ages could be derived from basement gneisses of southern Bulgaria and of the Sakarya Zone in NW Turkey (Fig. 10).

The Permian–Triassic sample CH11 shows a similar detrital zircon age spectrum to the Carboniferous sample CH52 with two exceptions: (I) Grenvillian ages are present and (II) the amount of Devonian zircons is higher. The source of Grenvillian zircons seems difficult to identify. Grenvillian zircons could be inherited components in magmatic zircons, as already described above for zircons from Cadomian orthogneisses of the Sredna Gora Zone in Bulgaria (Carrigan et al., 2006). The Grenvillian zircons from the Chios sediments, however, are well rounded and homogenous without evidence of magmatic or metamorphic over-growth. Nevertheless, younger magmatic or metamorphic rims on zircons can be removed by abrasion during long-distance river transport, in which case only the inherited components survive.

Grenvillian zircons might have also been derived from Baltica since basement rocks of such age are present there (e.g., Gorbatschev and Bogdanova, 1993; Bogdanova et al., 2007). If so, one could expect also zircons with ages of ca. 1.5–1.7 Ga and ca. 2.6–3.7 Ga (e.g., Gorbatschev and Bogdanova, 1993; Bogdanova et al., 2007), which are, however, totally absent in the Permian–Triassic sample and thus suggest that a detrital input from Baltica is most unlikely. A more plausible scenario is that of recycling of older sedimentary Grenvillian detritus. As already discussed before (see Section 5.1.), zircons aged around 1 Ga have accumulated in post-Cadomian (Ediacaran–Cambrian–Ordovician) cover sequences at the northern margin of Gondwana (Fernández-Suárez et al., 2000, 2002; Avigad et al., 2003; Gutiérrez-Alonso et al., 2005; Kolodner et al., 2006). Outcrops with biostratigraphically dated post-Cadomian clastic cover sequences can be found, for example, in southern Turkey (SE Anatolia and the Taurides: e.g., Gönçüoğlu and Kozlu, 2000) and in northern Turkey (Istanbul Zone: e.g., Göürür et al., 1997; Dean et al., 2000; Yanev et al., 2006).

We propose that the best source candidate for the Chios sediments studied here was the Sakarya microcontinent because of the occurrence there of Ordovician and Devonian basement rocks. Time and facies equivalents of the Lower Palaeozoic sediments from the Istanbul Zone were probably re-worked at the southern margin of Laurussia during the Variscan orogeny (see Section 5.3.) thus supplying
Grenvillian zircon detritus. It becomes then apparent that the ca. 1 Ga-old zircons in the Chios sediments have a secondary source due to polyphase recycling. We infer that the basement of the Sakarya microcontinent was probably exposed at the surface on a regional scale in Permian–Triassic times, thus supplying Devonian detritus.

5.3. Source of olistoliths

The analysis of olistoliths found within the turbidite–olistostome succession of Chios provides additional evidence for possible source areas. Fig. 11 shows a compilation of lithology and fossil content for olistoliths from Chios, including an interpretation for their depositional setting. The Silurian to Carboniferous olistoliths are mainly characterised by a variety of chert, carbonate and fine-grained siliciclastic rocks which were mainly dated by graptolites, conodonts, corals, brachiopods, algae and foraminifers (see Kauffmann, 1965; Besenecker et al., 1968; Herget, 1968; Herget and Roth, 1968; Fenninger, 1983; Groves et al., 2003; Larghi et al., 2005). Furthermore, volcanic rocks of basic, intermediate and acidic composition mentioned by previous researchers were studied in detail by Pe-Piper and Kotoxopoulou (1994) who found clear evidence that at least some of them intruded the deformed Palaeozoic clastic succession and therefore cannot be olistoliths. Based on geochemical data, they suggested that these volcanic rocks probably formed in a back-arc setting above a subducting Palaeotethyan oceanic slab and correlated them with Early Permian granodiorites from the Karakaya Complex in Turkey.

Robertson and Pickett (2000) assumed a Silurian age for some of the volcanic rocks based on the close spatial association of Silurian limestone olistoliths with volcanic rocks north of Spartounta. An in-depth analysis of the provenance of the volcanic rocks, however, is beyond the scope of the present paper. Here, we will only focus on the fossil-bearing sedimentary olistoliths.

The olistoliths essentially define four olistostrome formations with a very clear predominance of younger blocks (Lower Carboniferous) in the lower formation and of older blocks (Silurian) in the upper formation (Papanikolaou and Sideris, 1983; Sideris, 1989). This implies a down-stratigraphy continuous erosion of lithologies in the source area. The limestones and cherts were thermally and tectonically overprinted before they were deposited as olistoliths. Most of the chert and radiolaria-bearing olistoliths found on Chios are strongly fractured and penetrated by veins. At least two generations of veins can be identified in cherty samples. The first generation is characterised by quartz veins and the second by calcite veins. The thermal metamorphic overprint is documented by the conodont colour alteration index (CAI: Epstein et al., 1977; Rejebian et al., 1987). Conodonts from an Upper Visean (?Lower Serpukhovian) limestone breccia cropping out in the Kourounia–Nenitouria area, NW Chios, have a CAI of 5, indicating heating temperatures of 300–480 °C (Groves et al., 2003). In general, a decrease of the CAI from west (CAI of 6–6.5) to east (CAI of 2–3) can be observed in NW Chios (Fig. 2). Detailed investigations by Kozur (1998) on the Karaburun peninsula, opposite Chios, have shown that the Variscan structural and metamorphic overprint cannot be younger than Late Visean because pre-Upper Visean rocks show cleavage and have experienced very low-grade metamorphism (CAI of 4–5), whereas Upper Visean and Triassic rocks lack cleavage and display only very low-grade thermal alteration with a CAI of 2–3 and 1–1.5 respectively. This provides evidence that the Variscan metamorphic overprint affecting the olistoliths is older than ca. 330 Ma.

Recently, the chert olistoliths (ribbon radioliters) were examined biostratigraphically by Larghi et al. (2005), who found age-diagnostic radiolarians and conodonts in two samples thus allowing them to establish Late Silurian (probably Přídolían) and Late Devonian (Famennian) ages, whereas only a more general age range from Devonian to Early(? Carboniferous) could be proposed for the remainder. Based on graptolite fauna, Herget (1968) noted that the Silurian of northern Chios shows similarities to that of Stara Planina and the Kairste area in Bulgaria. The exotic limestone blocks and breccias have been dated as Silurian to Carboniferous (Kauffmann, 1965; Besenecker et al., 1968; Herget, 1968; Herget and Roth, 1968; Fenninger, 1983; Groves et al., 2003). In the course of this study, a small limestone lens, ca. 25 cm in

Fig. 13. Ludford conodont zonation on Sardinia by Corradini and Serpagli (1999), and conodont succession and event stratigraphy on Gotland modified from Jeppsson et al. (2005a). The age of the Ludfordian conodont fauna from the Chios olistolith is indicated by a black quadrangle.
Conodont fauna of an equivalent age (P. siluricus zone) are also known from many other peri-Gondwana areas (e.g., Carnic Alps: Walliser, 1964; Bohemia: Chlupáč et al., 1980; Western Europe and Northern Africa locations: compiled in García-López et al., 1994; Morocco: Sarmiento et al., 1997; Turkey: Kozur, 1998; Poland: Männik and Mållkowskii, 1998; Sardinia: Corradini and Serpagli, 1998, 1999; Australia: Talent and Mason, 1999; Frankenwald/Germany: Blumenstengel et al., 2006), from Laurentia (e.g., Klapper and Murphy, 1975; Uyeno, 1980) and from Baltica (references in Jeppsson et al., 1994, 1996).

Kauffmann (1965) has already reported occurrences of conodonts in cephalopod limestones found in northeastern Chios. The recovered conodont faunas from these limestones have not been illustrated or described, but taxa have been listed in the form sense and assigned by Kauffmann (1965) to the Late Silurian (Ludlow) K. crassa and P. siluricus conodont zones. The faunal lists from locations no. 5 to 7 in the region around Kardamila (Kauffmann, 1965: 653–654; Fig. 2) reflect similar faunal compositions; the material is comparable in age and may tentatively also be placed in the Polygnathoides siluricus zone.

The closest potential source localities of 'Orthoceras Limestone' similar in lithology to the reworked lenses on Chios are located in Turkey and in the Balkan region; the locations are summarised in Fig. 14. Paekelmann and Sievert's (1932) mentioned 'Orthoceras' from Sedef Island (Antirovitha) in the Sea of Marmara and from Kartal and Pendik districts on the mainland SE of Istanbul town. Based on litho- and biostatigraphic correlations, Haas (1968) assumed a Ludlow age for the 'Orthoceras'–bearing limestones from Sedef Island; those from Kartal and Pendik districts were assigned to the 'Obere Soğanlı-Schichten' for which he proposed an Early Emsian age. Recently, Kozlu et al. (2002) described 'Orthoceras Limestone' from the Çamdağ area, dated by conodonts, as being Přídoli in age. 'Orthoceras Limestone' was also reported from the Yukari Yayla Formation of the eastern Taurides and is according to conodont data, of latest Llandovery to earliest Wenlock age (Gönçüoğlu and Kozlu, 2000; Gönçüoğlu et al., 2004). In contrast to the low CAI (1–2) of the new fauna reported in this paper, Silurian conodonts from the Çamdağ area and the eastern Taurides have a CAI value of 5 (Kozlu et al., 2002; Gönçüoğlu et al., 2004). In the northern Central Taurides of the Konya area 'Orthoceras Limestone' from the Ayı Tespe Formation yielded upper Lochkovian–Pragian conodonts (Gönçüoğlu et al., 2004). As far as the Balkan region is concerned, 'Orthoceras Limestone' has been reported from the Suva Planina Mountains in eastern Serbia and from the Bistra Mountains in western F.Y.R.O.M (Gnoli, 2003). Geologically, the Suva Planina Mountains are part of the Kraštite area and the Bistra Mountains belong to the internal zone of the Dinardites. Silurian–Early Devonian cephalopod limestone biofacies is furthermore documented from NW Africa, the Pyrenees, SW Sardinia, the Armorican Massif, the Prague Basin, the Carnic Alps, the Úpomy Mountains in NE Hungary and from the continental platform in Ukraine (Gnoli, 2003). The Early Palaeozoic 'Orthoceras Limestone' may therefore become an important marker for palaeobiogeographic reconstructions of northern Gondwana-derived terranes (e.g., Kozlu et al., 2002; Gnoli, 2003).

With regard to the weathered limestone crust mentioned above, Robertson and Pickett (2000) noted that some limestone olistoliths are locally karstified and cut by laterite-filled fissures, whereas the host sediment is not. The palaeoecologist must be younger than the lithification of the limestones (post-Silurian) but older than their age of deposition as olistoliths (pre-Serpukhovian). The palaeoecologist marks a hiatus in

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*[Plate I](#)*: Representative conodont elements recovered from the cephalopod-bearing limestone lens of Chios. Scale bars equal 100 µm. The figured conodont specimens are deposited in the Senckenberg Museum Frankfurt/Main (SMF numbers in caption to Plate I). (1–5) Elements of the *Ozarkodina excavata* group: (1–5) *Ozarkodina excavata* sp. aff. *excavata*; (1) SMF-78114, Se element, lateral view; (2) SMF-78115, lower-lateral view; (3) SMF-78116, Sa element, anterior view; (4) SMF-78117, Sb element, lateral view; (5) SMF-78118, M element, lateral view; (6) *Oulodus siluricus* (Branson and Mehl), SMF-78119, M element, postero-lateral view; (7–8) *Oulodus cf. siluricus* (Branson and Mehl); (7) SMF-78120, Pb element, postero-lateral view; (8) SMF-78121, Sa element, postero-lateral view; (9) *Polygnathoides siluricus* Branson and Mehl, SMF-78122, Pa element, upper view; (10) *Panderodus gracilis* (Branson and Mehl), SMF-78123; (11, 13, 14) *Panderodus recurvatus* (Rhodes); (11) SMF-78124, lateral view, (13) SMF-78126, lateral view, (14) SMF-78127, lateral view; (12) *Kockelella variabilis ichnusae* Serpagli and Corradini, SMF-78125, Pa element, upper-lateral view.
sedimentation and suggests subaerial exposure somewhere in the Palaeotethyan realm at some point in the Devonian. According to the olistolith data available (Fig. 11), there seem to be two gaps in sedimentation in pre-Carboniferous times, suggesting that karstification could have occurred in the Early Devonian (Lochkovian; and/or in the Middle Devonian (Eifelian and Givetian). Interestingly, in the Çamdağ, the Safranbolu–Araç and the Karadere–Zirze areas of the Zonguldak Terrane, NW Turkey, an unconformity exists between the Silurian and Devonian sequences (Görür et al., 1997; Dean et al., 2000). The stratigraphic gap between the grypochondrium-bearing upper Wenlock and the unconformably overlying Emsian successions, together with an accompanying late Early Devonian thermal event, are characteristic features of the Zonguldak Terrane (Lakova and Göncüoğlu, 2005; Yanev et al., 2006) and are in sharp contrast with the continuous platform-type sedimentation seen in the Istanbul Terrane during the same time period (Yanev et al., 2006). The Early Devonian unconformity and accompanying thermal event were probably due to the docking of the Zonguldak Terrane to Baltica (Yanev et al., 2006). The Istanbul and Balkan Terranes, however, were accreted to the southern margin of Laurussia in Carboniferous times (Yanev et al., 2006). Moreover, a regional unconformity between the Upper Silurian and the Lower Devonian (Middle Lochkovian) also exists in the southern and eastern Taurides and in SE Anatolia (Göncüoğlu and Kozlu, 2000; Göncüoğlu et al., 2004). In the southern Taurides and SE Anatolia this is probably related to the closure of a branch of the Palaeotethys to the north of the peri-Gondwanan terranes (Göncüoğlu and Kozlu, 2000), whereas in the eastern Taurides it could be related to a stepwise detachment of terranes from the northern margin of Gondwana (Göncüoğlu et al., 2004). The sedimentation gap between the uppermost Silurian and Lower Devonian in the Zonguldak Terrane and in the Tauride-Anatolide block most probably represents the time of karstification of the Silurian limestones of Chios.

It is important to note that fossil assemblages of Lower Palaeozoic rocks from the Istanbul, Zonguldak and Balkan Terranes are similar to those from the Avalonian and Armorican Terranes of Central and Western Europe (e.g., Yanev et al., 2006 and references therein). For example, in a palaeogeographic reconstruction for the Ordovician (Dean et al., 2000), trilobite fauna were used to place the Zirze unit of the Zonguldak Terrane in the vicinity of successions from central Europe (Bohemia) and the Anglo-Welsh Basin (Avalonia). The Devonian benthic fauna of this terrane is typical of that of the Rhenohercynian Zone (Yanev et al., 2006). The Zonguldak Terrane may have been located in the eastern continuation of the Avalonian and Moravo-Silesian Terranes (Yanev et al., 2006). Ordovician and Silurian benthic fauna from the Istanbul Terrane are of Avalonian and Podolian affinity, whereas Devonian brachiopods and trilobites are clearly of Bohemian and North African affinity (Yanev et al., 2006). Emsian ostracodes indicate an open marine connection between the Istanbul Terrane, Thuringia and Morocco (Dojen et al., 2004). The Balkan and NW Anatolian Terranes were part of northern Gondwana but not of Baltica (Yanev et al., 2006 and references therein).

To summarise, taken collectively (lithology, fossil content, age and metamorphic overprint together with depositional environment) all data suggest that an input from the Pelagonian Zone or the Attic-Cycladic Massif can be safely excluded because neither Silurian nor Devonian sediments have been reported from these areas so far. We propose that the source area of the olistoliths is located north of present-day Chios, most likely in NW Turkey or southern Bulgaria–Serbia. Alternatively, the source rocks could have been time and facies equivalents of Palaeozoic units in NW Turkey or southern Bulgaria–Serbia, which have not been preserved.

The Visean ‘Tirrachian flysch’ of the Istanbul Zone probably indicates the onset of Variscan orogeny in this part of the Eastern Mediterranean (Görür et al., 1997) and may represent a time and facies equivalent of the Carboniferous turbidite–olistostrome succession of Chios. The time of deposition of the latter was constrained by biostratigraphic data from calcareous clasts of a breccia lying within the turbidite–olistostrome succession and was proposed to be older than Pennsylvanian, probably Late Visean or Early Serpukhovian (Groves et al., 2003). The youngest detrital zircon (grain CH52.22.1) found in this succession is 322 ± 3 Ma-old (this study). According to the Geological Time Scale (GTS) of Gradstein et al. (2004), the Serpukhovian ranges from 326.4 to 318.1 Ma; this range is virtually identical to the Visean range from 326.4 to 319 Ma. We therefore suggest that the time of deposition of the Chios turbidite–olistostrome succession is younger than Visean, probably Serpukhovian, whereas an even younger age (e.g., Bashkirian) cannot be excluded insofar as no further biostratigraphic or geochronological
6. Implications for Palaeotethys

Stampfli and Borel (2002) have presented palaeogeographic maps for the Palaeozoic showing an assemblage of terranes at the northern margin of Gondwana, termed the European Hunic terranes, bordered by the Rheic Ocean to the north and by the Palaeotethys Ocean to the south. According to Stampfli and Borel (2002), the term Palaeotethys is used to denote a seaway that separated Gondwana from Gondwana-derived fragments between the Silurian to early Late Triassic, a time in which these fragments drifted northward and accreted to Laurussia in a stepwise fashion. The Palaeotethys Ocean was closed by the northward drift of the Cimmerian terranes in response to the opening of Neotethys in the south. One of the most controversial topics over the last three decades has been the time of closure of Palaeotethys and the location of its suture zone (e.g., Şengör et al., 1984; Stampfli et al., 1991; Robertson et al., 2004). As already discussed by previous authors (e.g., Stampfli et al., 2003) and shown in the present work, Chios is a key area for understanding the closure of Palaeotethys because it is one of the rare localities where very low-grade to virtually unmetamorphosed fossil-bearing Palaeozoic to Mesozoic sequences are preserved (e.g., Besenecker et al., 1988). Detrital zircon ages and olistolith provenance (this study) imply sediment supply from terranes north of present-day Chios, most probably the Sakarya microcontinent in western Turkey and the Istanbul Zone in northern Turkey, or from time and facies equivalent successions such as those found in southern and central Bulgaria and Serbia. Below we compiled new and pre-existing data regarding Late Palaeozoic magmatic activity, metamorphism, depositional environment and fossil content in clastic successions of Chios and in similar formations in the surrounding region and put forward the following scenario for the geotectonic evolution of the Eastern Mediterranean in Late Palaeozoic to earliest Triassic times:

Variscan orogenic processes were already underway at the southern margin of Laurussia by at least Early Carboniferous times. Carrigan et al. (2006) reported a concordia age of 336.5±5.4 Ma for metamorphic rims of a migmatite leucosome from the Sredna Gora Zone in Bulgaria, which they interpret as the age of zircon recrystallisation during high-grade metamorphism. Subduction of (a northern branch of) Palaeotethys beneath terranes accreted to the southern margin of Laurussia probably ended around 325–320 Ma, testified to by the end of turbidite–olistostrome sedimentation on Chios and Karaburun. The upward transition of the Carboniferous turbiditic fan system to more proximal quartz-rich greywackes and sandstones, followed by carbonate-bearing sandstones with increasing intercalations of limestone beds through to thick limestones marks a shallowing-upward trend above the CCD (calcite compensation depth) and the initiation of a carbonate platform in Late Carboniferous and Permian times (Meinhold et al., 2007). Carboniferous foraminiferal fauna of the Chios–Karaburun units show distinct biogeographic affinities to the southern Laurussian shelf (Kalvoda, 2003). Benthic foraminifera and calcareous algae associations suggest that Chios must have been situated at the southern margin of the Sakarya microcontinent in the late Early Permian (Jenny and Stampfli, 2000). Fig. 15 illustrates the location of Chios in close proximity to the Sakarya microcontinent in a Carboniferous palaeogeographic reconstruction.

Subduction of a more southern or south-western branch of Palaeotethys beneath what was to become the Pelagonian–Cycladic Zone probably started around 330 Ma, producing a voluminous amount of arc-type granitoids between ca. 325 and 300 Ma (e.g., Reischmann, 1998; Vavassis et al., 2000; Reischmann et al., 2001; Anders et al., 2006b; Xypolias et al., 2006). Most of the Carboniferous basement rocks from the External Hellenides, the Attic–Cycladic Massif and the Pelagonian Zone in Greece and the Sakarya microcontinent in NW Turkey have similar geochemical and isotopic signatures (Pe-Piper and Piper, 2002), suggesting their close proximity in Late Palaeozoic times (Xypolias et al., 2006). They are commonly interpreted as being the product of northward subduction of (a branch of) Palaeotethys beneath the above-mentioned areas (e.g., Vavassis et al., 2000; Stampfli et al., 2003). However, southward subduction of Palaeotethys beneath the Cimmerian terranes during the Late Palaeozoic has also been proposed (e.g., Şengör et al., 1984; Robertson and Pickett, 2000; Romano et al., 2006; Xypolias et al., 2006). To solve the problem of subduction polarity of Palaeotethys is beyond the scope of the present paper. Rather, several branches of this ocean were subducted, a process that eventually led to multiple terrane accretion. Furthermore, as was the case in other parts of Europe (e.g., Matte, 1986; von Raumer et al., 2003), orogeny in the Eastern Mediterranean during the Late Palaeozoic was probably accompanied by large-scale transcurrent movements (e.g., Stampfli et al., 2002; Fig. 4), juxtaposing very different oceanic or continental fragments and hence giving rise to equivocal palaeotectonic models and interpretations. Rifting of the Neotethys oceans probably started within older suture zones (e.g., Palaeotethys suture) and later, during the subduction of these oceans in Late Mesozoic and Cenozoic times, a lot of information has been destroyed.

Remnants of one of the Palaeotethyan sutures can be traced from Chios north into the Pontides (‘Thracian flysch’ of the Istanbul Zone: Gürür et al., 1997). Such remnants can also be traced on the Karaburun peninsula in western Turkey (Kozur, 1998; Robertson and Pickett, 2000, and references therein), in the Tavas nappe (Lycian nappes) further southeast (Kozur et al., 1998) and near Konya in central Turkey (Eren et al., 2004). The present-day position of these units, so far south from their possible source is still a matter of discussion. Gessner et al. (2001) noted that the tectonic position of the Lycian Allochthon above the ophiolitic Selçuk mélange implies that the lower units of the Lycian Allochthon were deposited either on a promontory of Adria or Sakarya or on a separate continental fragment, and were not part of Anatolia as has been generally assumed by previous workers. We have shown here that the Upper Palaeozoic–Lower Mesozoic clastic successions of Chios were probably deposited on or close to the Sakarya microcontinent during Permian–Triassic times. If, then, the correlation of units from Chios with similar units from the Karaburun peninsula, the Konya region and the Lycian nappes (e.g., Tavas nappe) is correct, the latter must also have been deposited on or close to Sakarya, which would be in agreement with the model of Gessner et al. (2001).

In order to gain a more regional perspective, we would like to note here that the Upper Palaeozoic Chios olistostrome–turbidite succession may correlate to metamorphic rocks cropping out on Sakarya. Göncüoğlu et al. (2000) described the Tepeköy unit which they interpreted as an accretionary complex formed in an intra-oceanic fore-arc setting due to southward subduction of a branch of the Palaeotethys beneath an arc complex (future Söğüt unit). The Tepeköy and Söğüt units were attached to the northern margin of the Tauride–Anatolide Platform during Middle–Late Carboniferous times. Göncüoğlu et al. (2000) proposed that a back-arc basin developed at the northern margin of the Tauride–Anatolide Platform during the Late Carboniferous, as evidenced by the rocks of the Göktepe Metamorphics. The olistostrome–turbidite unit on the Karaburun peninsula and the Halic Formation at Konya are interpreted by Göncüoğlu et al. (2000) as possible tectonic slices of the Göktepe Metamorphics. If this is the case, the Chios units can also be compared to the latter since they have already been correlated with the Karaburun units (see above).

Which model is correct for the Chios olistostrome–turbidite succession discussed will certainly be the subject of further studies, but it seems clear that Chios was located close to the Sakarya microcontinent in Late Palaeozoic times and received detritus from north Gondwana-derived terranes with similarities to terranes
preserved in northern Turkey and the Balkan region. Furthermore, the virtually unmetamorphosed nature of Chios means that it must have always been in an upper plate position.

7. Conclusions

The presence of ca. 2 Ga-old detrital zircons in the Carboniferous and Permian–Triassic successions of Chios coupled with the lack of ages between 1.8 and 1.1 Ga is a characteristic feature of NW Africa-derived terranes. Zircons aged around 1 Ga almost certainly have ages between 1.8 and 1.1 Ga can be assigned to terranes in western and northern Turkey (e.g., Sakarya microcontinent, Istanbul Zone) and to units in Bulgaria. The analysis of fossiliferous olistoliths (Silurian cephalopod limestones) from the Carboniferous olistolith–turbidite succession indicates a source area similar to that of Palaeozoic sequences exposed nowadays in northern Turkey and the Balkan region.

In summary, the results stemming from the new data presented here establish that NW African-derived terranes were involved in Palaeotethyan subduction–accretion processes in the Aegean region during Late Palaeozoic times. The relationships, however, between these terranes at the southern margin of Laurussia are now masked by younger (Mesozoic to Cenozoic) complex structural and metamorphic events. Our study demonstrates that the combined use of detrital zircon U/Pb ages and provenance analysis of olistoliths can impose tight constraints on terrane accretion processes and the provenance of crustal sources for sediments during the subduction of Tethyan oceans.

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Appendix A. Supplementary data


References


