Neogene supradetachment basin development on Crete (Greece) during exhumation of the South Aegean core complex

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ABSTRACT

Tertiary extension in the Aegean region has led to extensional detachment faulting, along which metamorphic core complexes were exhumed, among which is the Early to Middle Miocene South Aegean core complex. This paper focuses on the supradetachment basin developed during the final stages of exhumation of the South Aegean core complex along the Cretan detachment, plus the Late Miocene to Pliocene basin development and palaeogeography associated with the southward motion of Crete during the opening of the Aegean arc. For the latter purpose, the sedimentary and palaeobathymetric evolutions of a large number of Middle Miocene to Late Pliocene sequences exposed on Crete, Gavdos and Koufonisi were studied. The supradetachment basin development of Crete is characterised by a break-up of the hanging wall of the Cretan detachment into extensional klippen and subsequent migration of laterally coexisting sedimentary systems, and finally the deformation of the exhumed core complex by processes related to the opening of the Aegean arc. Hence, three main tectonic phases are recognised: (1) Early to Middle Miocene N–S extension formed during the Cretan detachment, exhumed in the South Aegean core complex. The Cretan detachment remained active until 11–10 Ma, based on the oldest sediments that unconformably overlie the metamorphic rocks. Successions older than 11–10 Ma unconformably overlie only the hanging wall rocks and Middle Miocene sediments form isolated blocks on top of the exhumed metamorphic rocks, which are interpreted as extensional klippen. (2) From approximately 10 Ma onward, southward migration of the area that presently covers Crete was accompanied by E–W extension, and the opening of the Sea of Crete to the north. Contemporaneously, large folds with WNW–ESE striking, NNE dipping axial planes developed, possibly in response to sinistral transpression. (3) During the Pliocene, Crete emerged and tilted to the NNW, probably as a result of left-lateral transpression in the Hellenic fore-arc, induced by the collision with the African promontory.

INTRODUCTION

Rapid, high-strain extension of the continental crust is frequently accommodated along extensional detachments. If displacement along such detachment faults is sufficient, metamorphic rocks can be exhumed to form metamorphic core complexes (e.g. Crittenden et al., 1980; Wernicke, 1981; Davis, 1983; Lister et al., 1984). In the Aegean region, such metamorphic core complexes formed as a result of the southward roll-back of the African slab, extending a late Mesozoic–Cenozoic nappe stack that formed during Africa–Europe convergence (Bonneau, 1984; Jacobshagen, 1986; Jolivet et al., 1994, 2003; Ring & Layer, 2003; van Hinsbergen et al., 2005c; Fig. 1).

Associated with extension and core complex exhumation, subsidence will lead to the formation of supradetachment basins (Dinter & Royden, 1993; Friedmann & Burbank, 1995; Kruger et al., 1995; Janecke et al., 1999; Kyckel et al., 1999; Sánchez-Gómez et al., 2002; Sözbilir, 2002). These basins have frequently been described in 2D cross-section based on seismic analysis (e.g. Kruger et al., 1995) or based on field analysis (e.g. Dorsey & Becker, 1995; Sözbilir, 2002). The reconstruction of the evolution of a supradetachment basin as a function of progressive...
deformation along the extensional detachment is frequently complicated by a lack of three-dimensional information on basin structure and stratigraphy. Moreover, many supradetachment basins are terrestrial, hampering detailed chronostratigraphic control.

The most southern of the Aegean metamorphic core complexes is exposed on the island of Crete, where the South Aegean Core complex was exhumed since the Early Miocene along the top-to-the-north Cretan extensional detachment (Jolivet et al., 1996; Fig. 1). This core complex was rapidly exhumed between approximately 24 and at least 15 Ma (Jolivet et al., 1996; Thomson et al., 1998). Crete exposes a well-described and well-exposed, more or less continuous Middle Miocene to Pliocene sedimentary record. This record is largely marine, and covers large parts of the island, allowing the detailed temporal reconstruction of the three-dimensional Neogene basin evolution (Meulenkamp, 1985; Meulenkamp & Hilgen, 1986; Meulenkamp et al., 1988; Fassoulas, 2001). Moreover, the oldest sediments on Crete are conglomerates, which are uniquely derived from the non-metamorphic hanging wall of the Cretan detachment (Kopp & Richter, 1983; Peters, 1985). This suggests that they predate the exposure of the metamorphic footwall rocks on Crete, suggesting that at least part of the well-described and well-dated sedimentary record of Crete accumulated during the last stages of activity of the extensional detachment. Crete could therefore provide an excellent opportunity to study the three-dimensional temporal development of a supradetachment basin.

In this paper, the three-dimensional evolution is presented through palinspastic reconstruction and analysis of a vertical motion of the Neogene sedimentary basins. The palinspastic reconstruction is based on the integration of the well-described structure, stratigraphy and sedimentology of Crete, together with new observations and age constraints. Additionally, detailed information on

Fig. 1. Geologic map of Greece, modified after Bornovas & Rontogianni-Tsiabou (1983), with outlines of the Aegean metamorphic core complexes and the position of Crete.
the timing of the various tectonic stages recognised in the palinspastic reconstruction is obtained via a novel study concerning vertical motions associated with the foundering, evolution and inversion of the Neogene basins of Crete, based on detailed palaeobathymetry analyses of a large number of sedimentary sequences on Crete, Gavdos and Koufonisi.

OUTLINE OF THE GEOLOGY
Structure and metamorphism of the nappes

The Hellenic nappes consist of sediments that locally unconformably overlie pre-Alpine continental or oceanic basement rocks. The nappes were subdivided based on their lithologies and tectonic positions and include, from bottom to top, the Ionian, Tripolitza and Pindos nappes, which are overlain by an ‘Uppermost Unit’ of mixed Jurassic- and Cretaceous ophiolitic rocks and sediments (Aubouin, 1957; Seidel et al., 1981; Bonneau, 1984; Jacobshagen, 1986). During the Neogene, the Cretan nappe pile was crosscut by the top-to-the-north Cretan extensional detachment, along which the South Aegean metamorphic core complex was exhumed (Ring et al., 2001b; Jolivet et al., 2004; van Hinsbergen et al., 2005b).

The footwall exposed the Lower Sequence (Plattenkalk and Phyllite Quartzite units) in the footwall, and the hanging wall consists of the Upper Sequence formed by the Tripolitza, Pindos and Uppermost units (Fassoulas et al., 1994; Jolivet et al., 1996; Thomson et al., 1999; Ring et al., 2001b; van Hinsbergen et al., 2005b; Figs 2 and 3).

The Plattenkalk unit consists of HP-LT metamorphic clastic rocks (Kastania Phyllites), overlain by Mesozoic clastics, unconformably overlying the pre-Neogene basement or the older conglomerate formations on Crete. Meulenkamp (1969) subdivided the Neogene sedimentary sequence of Crete into six lithostratigraphic groups (Fig. 5), largely based on data published by Freudenthal (1969), Meulenkamp (1969), Gradstein & Van Gelder (1971), Sissinj (1972), Drooger & Meulenkamp (1973), Gradstein (1973, 1974), Zachariasse (1975), Fortuin (1977, 1978) and Meulenkamp et al. (1979).

The Prina Group constitutes the base of the Neogene in some areas – especially in eastern and western Crete – and contains limestone breccias and breccia-conglomerates, in many cases displaying components embedded in an indurated, calcareous matrix. Elsewhere they are found as intraformational mappable units within the Tefelion Group (see below). The Prina Group has nowhere been observed to unconformably overlie rocks of the Lower Sequence. The mapable units between the Tefelion Group formations are formed by olistoliths of the Upper Sequence and are found scattered all over the island. Note that the Prina Group is not the same as the ‘Prina complex’, introduced by Fortuin (1977); the Prina complex forms part of the Prina Group, but the Prina Group also includes the older conglomerate formations on Crete.

The Tefelion Group consists of poorly consolidated marine and fluvi-lacustrine conglomerate, sand, silt and clay unconformably overlying the pre-Neogene basement or the Prina Group and overlain by limestones and marls (calcareous clays) of the Vrysses Group.

The Vrysses Group conformably overlies the Tefelion Group or unconformably overlies the pre-Neogene basement. It consists of marine bioclastic or reeval limestones...
Fig. 2. Geologic map of Crete, modified after Bornovas & Rontogianni-Tsiabou (1983), with the locations of the analysed sections. For profiles, see Fig. 3.
Fig. 3. Cross-sections across Crete. For location of the profiles, see Fig. 2. Note that the vertical scale is exaggerated two times.
or alternating laminated and homogeneous marls, which contain evaporite intercalations in some areas. Its age is late Tortonian to Messinian (Fig. 4; Meulenkamp et al., 1979).

The Hellenikon Group overlies and interdigitates with the Vrysses Group in western Crete and includes red, terrestrial conglomerates and lacustrine, fluvial and lagoonal sands and clays and locally thin evaporite beds. It is overlain by open marine sediments of the Finikia Group. In central and eastern Crete, the Hellenikon Group is generally absent. The Hellenikon Group was deposited during the Messinian salinity crisis (Meulenkamp, 1979).

The Finikia Group consists of open marine marls overlying either successions of the Vrysses Group, deposits of the Hellenikon Group or the pre-Neogene basement. Its basal part frequently contains a breccia that consists of Miocene and earliest Pliocene sediments (marl breccia interbeds). The age of the Finikia Group is Pliocene (Meulenkamp, 1979).

The Aghia Galini Group overlies the Finikia Group and contains terrestrial conglomerates and fluviodeltaic sands and clays (ten Veen & Kleinspehn, 2003). It belongs to the Upper Pliocene and probably also the Pleistocene (Meulenkamp, 1979; ten Veen & Kleinspehn, 2003).

**SAMPLING AND PALAEOBATHYMETRY ANALYSIS**

We analysed 75 sedimentary sections to reconstruct the timing and magnitude of vertical motions associated with Neogene basin evolution on Crete (Fig. 2). The bio-, magneto-, cyclo-, and lithostratigraphic correlations of these sections are given in the Supplementary material.

To estimate the depositional depth of the sediments, we used the general relationship between depth and %P (fraction of planktonic foraminifera among the total foraminiferal population) of van der Zwaan et al. (1990), following sample selection and counting procedures described in van Hinsbergen et al. (2005a). In general, the method is accurate for water depths between approximately 50 and 1100 m. The resolution of the method depends on depth and the amount of averaged values, and will decrease with increasing depth, to give error bars indicated by the 95% confidence limit of about 50 m at shallow levels, to about 100–150 m at deep levels close to 1 km (van der Zwaan et al., 1990; van Hinsbergen et al., 2005a). Moreover, the resolution of the method may be seriously hampered by a number of other phenomena. First of all, %P is also dependent on oxygenation. Van Hinsbergen et al. (2005a) introduced the term %S (fraction of benthic oxygen stress markers), and following their suggestion, we omitted all samples with %S values exceeding 60%. Additionally, downslope transport may lead to winnowing of the sediment and influx of shallower sediments with higher per-
centages of benthic foraminifera. Therefore, samples containing high fractions of rock fragments, or showing evidence for winnowing at deep-marine levels were discarded. The calculated depth was checked by means of the presence or absence of benthic depth markers (see van Hinsbergen et al., 2005a). The results are included in Table 1, and are generally in line with the calculated values. Ideally, the benthic depth markers and the calculated depth should of course coincide. In cases of large discrepancies as a result of reasons listed above, we prefer to follow the (rough) depth indication provided by the deepest depth markers, as these better reflect the biotic conditions during sedimentation.

Some sections show a significant bathymetry change from bottom to top (Fig. 6). From all other sections, the depth estimates were averaged (Table 1). Averaging palaeobathymetry estimates generally decreases the standard deviation, but averaging estimates from different stratigraphic levels throughout a section introduces a new error, because water depth during deposition may have changed because of eustatic sea-level changes, sediment accumulation and tectonics. A section thickness larger than the error bar of the average palaeobathymetry shows deepening during sedimentation (e.g. section Parathiri (Section 63), Table 1), whereas the reverse shows no sea-level changes larger than the error in the palaeobathymetry estimate. Supplementary Table S1 lists the estimates, stratigraphic levels and – if available – numerical age dates of all sections and samples.

Accurate correction for eustatic sea-level changes (in the order of tens of metres) requires a high-resolution, astronomical age control (Lourens & Hilgen, 1997), and could therefore only be carried out for a few sections (Fig. 6). Adding the amount of accumulated sediment to palaeobathymetry provides motion of the base of the section through time and assumed an error for deep-marine (70% confidence limit) of our estimates calculated by averaging multiple samples (> 10) is generally as large as ± 100–150 m (Table 1), and therefore the resolution of a single estimate is much larger than ± 100 m. Moreover, ten Veen & Postma (1999b) and ten Veen & Kleinspehn (2000) did not indicate that they corrected for downslope transport, or that they omitted stress markers from the benthic population, which may strongly decrease the resolution of a palaeobathymetry estimate to an almost meaningless value (van Hinsbergen et al., 2005a). The palaeobathymetry estimates of ten Veen & Postma (1999b) and ten Veen & Kleinspehn (2000) in the shallower marine realm are indeed much more accurate, but variations of only tens of metres of water depth can be invoked by eustatic sea-level changes or tectonics, and it requires a very high-resolution stratigraphic time control to correct for eustatic sea-level changes (Lourens & Hilgen, 1997; van Hinsbergen et al., 2005a), which is not provided by ten Veen & Postma (1999b) and ten Veen & Kleinspehn (2000). Therefore, we do not incorporate their conclusions on the structural history of central and eastern Crete in our reconstructions.

**BASIN RECONSTRUCTIONS**

Here, we combine published information and observations on sedimentary facies, palaeoflow directions and structural geology with the palaeobathymetry reconstructions and age correlations (Table 1 and Figs 2, 4 and 6) to reconstruct the geologic history of the various parts of Crete. The interpreted vertical motion histories for the areas discussed below are summarised in Fig. 7.

**Gavdos**

The basement of Gavdos comprises the Pindos and Uppermost units (Fig. 2; Vinceinte, 1970; Seidel & Okrusch, 1978; IGME, 1993), and is unconformably overlain by upper Serravallian to Messinian sediments (Anastasakis et al., 1990, 1995). Potamos (Section 1) recorded late Serravallian subsidence, containing coral limestone unconformably overlying the pre-Neogene basement, overlain by upper Serravallian mudrocks and sapropels (Anastasakis et al., 1995), deposited at around 500–600 m water depth (Table 1, Fig. 7). The mudrocks and sapropels are overlain by shallow-marine sands, reflecting latest Serravallian to earliest Tortonian uplift. Subsequently, Metochia (Section 2), unconformably overlying both the pre-Neogene basement and the sequence of Potamos (Section 1; Anastasakis et al., 1995), was deposited during a next phase of subsidence from approximately 10 Ma to at least 9.4 Ma (Fig. 6). The 1000–1200 m depth estimate should be considered a minimum, as it corresponds to > 99.9%P (van der Zwaan et al., 1990). Very deep-marine conditions prevailed until at least 6.6 Ma (Fig. 6) Seismic profiles of the shelf Gavdos Rise indicate Messinian evaporite diapirs (Anastasakis, 1987), but Messinian evaporites on Gavdos have probably been eroded. Uplift of the Gavdos rise occurred in the course of the Pliocene (Anastasakis et al., 1984; Anastasakis, 1987).

**Southwestern Crete**

The Neogene sediments of western Crete are exposed to the north and south of the WNW–ESE striking west-Cre-
Fig. 6. Results of the palaeobathymetric analyses and relative motion diagrams obtained from 14 selected sections of the Late Miocene and Pliocene of Crete. Small dots are individual depth calculations; the thin solid curve represents a five to nine point moving average of the depth estimates. Thick solid curves represent geohistory curves, indicating the position of the base of the section with respect to the present-day sea level, and were derived by correcting the palaeobathymetry curve for sedimentation and eustatic sea-level changes, following procedures described in van Hinsbergen et al. (2005a). Vertical grey bars represent the palaeobathymetry indicated by benthic depth markers. The grey area represents the standard deviation of the moving average. Astronomical calibrations of the Pliocene sections are given in the Supplementary material and age calibrations of the Late Miocene sections are taken from Hilgen et al. (1995, 1997) and Krijgsman et al. (1995).
Table 1. Palaeobathymetry estimates including standard deviation, and litho- and chronostratigraphic position of the Cretan sections

<table>
<thead>
<tr>
<th>Number</th>
<th>Section</th>
<th>Sample number</th>
<th>Thickness (m)</th>
<th>Depth (m)</th>
<th>SD (m)</th>
<th>Taxonomic estimate Group</th>
<th>Age range</th>
</tr>
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<td>7 52i</td>
<td>121 500–600</td>
<td>I</td>
<td>u. Serravallian</td>
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<tr>
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<td>Metochia</td>
<td>Gr 4801–5407</td>
<td>102</td>
<td>163  Trend</td>
<td>Trend</td>
<td>I/P</td>
<td>9.8–6.6 Ma</td>
</tr>
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<td>Gr 3878</td>
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<td>1 36</td>
<td>0–50</td>
<td>I</td>
<td>u. Serravallian</td>
</tr>
<tr>
<td>4</td>
<td>Grammenos</td>
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<td>131</td>
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<td>u. Serravallian</td>
</tr>
<tr>
<td>5</td>
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<td>Gr 3899–95</td>
<td>–</td>
<td>1 36</td>
<td>0–50</td>
<td>I</td>
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</tr>
<tr>
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<td>35 0–50</td>
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<tr>
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<tr>
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<td>Lari das</td>
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<td>P</td>
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<tr>
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</tr>
<tr>
<td>39</td>
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<td>12 355</td>
<td>221 500–750</td>
<td>F</td>
<td>Pliocene</td>
</tr>
</tbody>
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The core of the antiform exposes the Lower Sequence locally tectonically overlain by isolated blocks of the Upper Sequence and Prina Group conglomerates (Fig. 2 and profile A–A′ in Fig. 3).

The Prina Group, represented by the Topolia Formation (Fm) (Meulenkamp, 1979; Kopp & Richter, 1983), crops out on both limbs of the west Cretan antiform (Fig. 2) and unconformably overlies and solely consists of fragments of

<table>
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<tr>
<th>Number</th>
<th>Section</th>
<th>Sample number</th>
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<th>Depth (m)</th>
<th>SD (m)</th>
<th>Taxonomic estimate</th>
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<td>355</td>
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<td>P</td>
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</tbody>
</table>

n: number of samples averaged; SD: standard deviation. Key to the groups: I, Ierapetra Group; P, Potamidha Group; V, Vrysses Group; F, Finikia Group; A, Aghia Galini Group. For age calibrations: see Supplementary material. Sample codes refer to the collection held at Utrecht University, Faculty of Geosciences: Gr, Greece-collection; Cp, correlation project collection; TF, thesis collection Tom Freudenthal; JM, thesis collection Johan Meulenkamp. Number refers to section numbers in Figure 2.

Fig. 7. Schematic overview of the lithostratigraphic units and their vertical motion history of the various study areas on Crete, Gavdos and Koufonisi.
the Upper Sequence. Calcareous nanofossils recovered from the base of the Topolia Fm in sections Akros Flomos (Section 11) and Palaeohora Campsite (Section 9) (see Supplementary material) indicate a Middle Miocene age (Table 1, see Supplementary material). Numerous pebble imbrications (n = 121) show paleoflow directions between the southwest and the southeast (bedding-tilt corrected), which is in line with the conclusions of Kopp & Richter (1983). The contact between the Topolia Fm and the Phyllite-Quartzite unit is always formed by the Cretan detachment. Deposition of the Topolia Fm thus precedes both the first exposure of the Phyllite Quartzite on western Crete and the formation of the west-Cretan antiform, and therefore predates the end of activity of the Cretan detachment.

The Topolia Fm is locally conformably overlain by shallow marine sands and marls (Anogia, Section 5; Table 1). The stratigraphic position of the short, deep-marine sections Grammenos (Section 4), Palaeochora (Section 6), Sellino (Section 7) and Voutas (Section 8; Table 1, Fig. 2) with respect to one another is unknown, but most likely post-date the Topolia Fm and, based on the presence of Serravallian foraminifera (W.J. Zachariasse, pers. comm., 2004), these sections are lateral equivalents of Potamos (Section 1) on Gavdos. Sections Voutas (Section 8) and Anogia (Section 5) are overlain by olistoliths of Tripolitza limestone, which were probably emplaced during the early phases of subsidence.

Northwestern Crete

In northwestern Crete, the Tefelion Group unconformably overlies the Upper Sequence, the Lower Sequence and the Prina Group (Topolia Fm; Freudenthal, 1969). Isolated blocks of the Upper Sequence and the Topolia Fm formed a palaeorelief during the deposition of the Tefelion Group, indicating a phase of non-deposition between sedimentation of the Prina and Tefelion Groups. The Tefelion Group exposes shallow marine reefal limestones on the intrabasinal highs and coarse calcarenites along the northern limb of the west-Cretan antcline, which laterally interfinger with deep-marine clays (Freudenthal, 1969). The oldest age date obtained from these clays is 7.6 Ma (Fig. 6, Table 1), indicating (late) Tortonian subsidence. To the east, on top of the horst of Chania, the Tefelion group (Stalos (Section 25) and Mournies (Section 26)) was deposited at significantly shallower levels than east and west of the horst (Table 1) and sedimentary facies of the entire Tefelion and Vrysses Groups on the horst of Chania show very shallow marine deposition (Freudenthal, 1969). East of the horst of Chania, Tortonian lacustrine deposits and overlying uppermost Tortonian and Messinian marine sediments reflect syn-sedimentary subsidence in section Exopolis (Section 29; Benda et al., 1974; Fig. 6). The Messinian sediments of Vrysses (Section 28) are of deep-marine origin (Table 1). Thus, the horst of Chania most likely formed during the late Tortonian to early Messinian, based on the deep-marine origin of the upper Tefelion Group east and west of the horst, and its shallow marine nature on top of the horst, in combination with the syn-sedimentary subsidence in section Exopolis (Section 29). The horst is bounded by N–S striking normal faults (Fig. 2, profile H–H’ in Fig. 3), implying late Tortonian to early Messinian E–W extension. The upper Tortonian marginal calcarenites deposited against the west-Cretan antiform in northwestern Crete reveal thickening towards the north and have a 45° northward dip in the lowermost beds, gradually changing into subhorizontal at the top, showing syn-sedimentary northward tilting of the northern limb of the west-Cretan antiform. This tilting was therefore contemporaneous with subsidence of the basin resulting from E–W extension.

Deep-marine Vrysses Group deposits in places unconformably overlie the pre-Neogene basement (Meulenkamp, 1969), reflecting ongoing subsidence during the Messinian. Hundreds of metres of terrestrial conglomerates of the Hellenikon Group represent the sea-level drop of the Messinian salinity crisis (Meulenkamp et al., 1979), as both the Vrysses and Finikia Groups were deposited at deep-marine levels (Table 1).

Frangokastello

In the Levka Ori Mountains, north of Frangokastello (Fig. 2), we found isolated outcrops of lithified limestones and conglomerates of the Prina Group with shallow marine faunas, separated from the Plattenkalk unit of the Lower Sequence by normal faults.

The Frangokastello area lies south of a large, E–W striking fault scarp (profile B–B’, Fig. 3). Skaloti (Section 32) contains fine-grained, marine sediments and specimens of Clypeaster (Table 1), and was therefore assigned to the Tefelion Group. The Finikia Group sampled near Chora Sfakia (Section 30) is deep marine (800 m: Table 1), but overlies a few metres of coarse conglomerates of unknown age, unconformably overlying the Plattenkalk unit. The absence of a transgressive sedimentary sequence suggests rapid latest Miocene or earliest Pliocene subsidence, as eustatic sea-level changes do not exceed 100 m (Lourens & Hilgen, 1997). The Frangokastello area was uplifted by 500 m after the Late Pliocene, indicated by the palaeobathymetry estimates of Frangokastello (Section 31; Table 1).

Rethymnon area

The south of the Rethymnon area near Plakia (Fig. 2) exposes fluvi–deltaic clays of the Tefelion Group with a late Serravallian to early Tortonian age (De Bruijn & Meulenkamp, 1972; Benda et al., 1974). These form the lateral equivalent of the deep-marine sequences of Potamos (Section 1) and around Palaeohora.

The northern Neogene sequence is younger (Meulenkamp, 1969) and is composed of fluvi–lacustrine successions of the Tefelion Group, exposed in the elongated depressions of Viglotopi (Section 36) and Apostoli (Section 41; Fig. 2). The sequences unconformably overlie the Lower Sequence (Meulenkamp, 1969). In the elongated
basin of Apostoli (Section 41; Fig. 2 and profile C–C' in Fig. 3), the thickness of the marine part of the Tefelion Group increases from zero in the NNE to 200 m in the SSW over a distance of only 2 km (profile C–C' in Fig. 3). This, in combination with 500 m of subsidence recorded in Apostoli (Section 41, Fig. 6), shows syn-sedimentary tilting towards the SSW. This could either have been caused by subsidence in the SSW, along NNE-dipping normal faults, which have not been reported (profile C–C' in Fig. 3), or from uplift in the north. Shallow marine Vrysses Group carbonates overlie the Tefelion Group of Apostoli (Section 41; Drooger et al., 1979). Therefore, before the deposition of the Vrysses Group, the valley of Apostoli must have been uplifted. The onlap of the Vrysses Group on the basement north of Apostoli (Meulenkamp, 1969) shows that in the Messinian, subsidence must have affected the northern part of the Rethymnon area. We propose that the antiform of the Psiloritis Mountains (Fig. 4) continued into the Rethymnon area, but was later submerged along the N–S striking, E dipping normal faults in the west of the Rethymnon area (Figs 2 and 3).

The simultaneous regression in Apostoli (Section 41) and the transgression in the north indicate early Messinian tilting with a northward component, possibly in response to folding of the west-Cretan antiform, which runs south of Apostoli (Fig. 5). The faults presently bounding the Apostoli depression thus post-date the deposition of the Apostoli section.

Messinian sediments contributable to the Messinian salinity crisis have not been identified, but extension and subsidence must have continued throughout the Messinian salinity crisis to explain the much larger palaeobathymetry of the Finikia Group compared with the Vrysses Group (e.g. Gonia, Section 35, and Erisi, Section 40; Table 1). The shallower palaeobathymetry of the Upper Pliocene of Asteri (Section 38) and Stavromenos–Rethymnon (Section 39; Table 1) may be a lateral topographic effect, but most likely reflects Pliocene uplift.

**Central Crete (Heraklion and Messara basins)**

Extensive fluvio-lacustrine deposits (Viannos Fm) of the Tefelion Group form the base of the Neogene on central Crete. Flow directions vary between west and south (Sissingh, 1972; Zachariasse, 1975) and the Viannos Fm underlies shallow marine deposits (Skinias Fm) of lower Tortonian age (Dhemati (Section 61) and Vathiopetro (56; Table 1; Zachariasse, 1975). The Viannos Fm therefore probably formed the lateral equivalent of the deep-marine sediments of Gavdos and southwestern Crete (Postma et al., 1993b). It can be traced to Anoia (Section 51; Fig. 2; Meulenkamp and Hilgen, 1986). The area is characterised by occurrences of large blocks of Tripolitza limestone, which occasionally can be seen to overlie the Viannos Fm. We therefore interpret these blocks as olistoliths. Asymmetric folds associated with olistolith emplacement and the decreasing number and size of these olistoliths from NW to SE suggest a northwesterly provenance. Olistoliths are also found in the Psiloritis mountains near Anoia (Section 51) (Fig. 2), and on the Central Heraklion Ridge; their emplacement therefore predate the uplift of the Central Heraklion Ridge and the uplift of the Psiloritis mountains in the footwall of the Heraklion basin (profile H–H' in Fig. 3). The Central Heraklion Ridge lies in the trend of the fold axis of the large Psiloritis antiform (Fig. 4), and Meulenkamp et al. (1988, 1994) therefore suggested that the Central Heraklion Ridge is a submerging antiform. Alternatively, Angelier (1975) and ten Veen & Postma (1999b) suggested that the Central Heraklion Ridge is normal-fault bounded. The ridge started to form an intrabasinal high in the Late Miocene, during the deposition of the Vrysses Group. The Vrysses Group sediments reveal drag folding along the western border fault of the Heraklion Basin, and Meulenkamp et al. (1994) therefore suggested that the folding of the Central Heraklion Ridge occurred during subsidence and E–W extension. However, the folding of the Central Heraklion Ridge may have started earlier, contemporaneously with the folding of the Psiloritis antiform and the deposition of Apostoli (Section 41).

In the Messara basin (Fig. 2), about 800 m of subsidence occurred during the Late Miocene (Kastelli, Section 46; Fig. 6). Panasos (Section 47) represents the marginal facies of the Kastelli (Section 46) and Aghios Ioannis (Section 45) deposits, indicating that the Psiloritis mountains formed a palaorelief in the late Tortonian. Ten Veen & Postma (1999a) suggested that the present-day fault that bound the Psiloritis mountains in the south formed in the late Miocene. However, the absence of Plattenkalk debris in the marginal basin facies north of Panasos (Section 47) suggests that the present-day contact with the Plattenkalk post-dates Late Miocene deposition. The submergence of Kastelli (Section 46) probably occurred during regional E–W extension, comparable with western Crete. The Lower Pliocene is invariably deep marine (Table 1). Kalithia (Section 57), Aghios Vlassios (Section 55) and Finikia (Section 54) show a detailed correlation between %P (fraction of planktonic foraminifera) and %S (fraction of benthic oxygen stress markers; see above) through time, and the 400 ka cyclic %P and %S variation was interpreted as the response of the benthic life to oxygenation variations forced by Milankovitch cyclicity (van Hinsbergen et al., 2005a; Fig. 6) The palaeobathymetry trend that was constructed from these sections is highly disturbed by the oxygenation effect, but reveals Pliocene uplift (Fig. 6), in line with earlier conclusions of Meulenkamp et al. (1994). Lower Pliocene sediments of Faneromeni–Messara (Section 44) were deposited between approximately 5 and 4.5 Ma and reveal approximately 300 m of uplift (Fig. 6). Also the Upper Pliocene of Atsipadhes (Section 49) indicates syn-sedimentary shallowing. Pliocene uplift of Crete led to the deposition of the fluvio-deltaic Aghia Galini Group (ten Veen & Kleinspehn, 2003). The southern part of the Heraklion area emerged earlier than the northern part, where deep-marine conditions lasted until after 3 Ma, despite the fact that the earliest Pliocene bathymetry was comparable throughout the basin.
Ierapetra area

The lower part of the Tefelion Group in the Ierapetra area consists of three conglomerate formations (Mithi, Males and Fothia Fm) that unconformably overlie the Upper Sequence and (in eastern Crete) conglomerates of the Prina Group (Goudouras Fm). The Males Fm consists of well-sorted conglomerates, which become finer grained and better rounded from base to top and from east to west, in line with the westward palaeoflow directions (Fortuin, 1977). Shallow marine deposits in the top of the formation indicate an earliest Tortonian age (Fortuin, 1977). The Males Fm was interpreted as braided to the meandering river system feeding the lacustrine Viannos Fm of the Héraklion area (Fortuin, 1977) and is therefore the lateral equivalent of deep-marine sediments of Potamos (Section 1) on Gavdos and e.g. Palaeohora (Section 6) on southwestern Crete.

In the Ierapetra depression, the Males Fm is overlain by coarse-grained mass flows, conglomerates and olistoliths of the Prina Group (Fortuin & Peters, 1984). In the south-east, the Males Fm is overlain by the Fothia conglomerates, deposited along a pronounced relief (Fortuin, 1977). The Fothia Fm also onlaps on the Upper Sequence and is overlain by olistoliths that were transported from NNE to SSW over the Fotia Fm, and from NNW–SSE in the Ierapetra basin (Fortuin, 1978). The deposition of the olistoliths is associated with rapid subsidence and SSE-ward tilting during N–S extension (Ring et al., 2001a), evidenced by the deep-marine character of the lower Tortonian Tefelion Group of Kalamavka (Section 64) and Parathiri (Section 63), which interfingers with and overlie the Prina Group. These have constant palaeobathymetries throughout a vertical succession of hundreds of metres of sediment (Postma & Drinia, 1993; Postma et al., 1993a; Table 1), indicating that sedimentation kept pace with subsidence. Subsidence continued throughout the Messinian, evidenced by the onlap of the Vrysses Group on the pre-Neogene basement (Gradstein & Van Gelder, 1971) and the (poorly constrained: 300–900 m) palaeobathymetry of Sikia (Section 69; Table 1).

Isolated outcrops of lowermost Pliocene marls (Aghia Triada, Section 71, Kheratokambos, Section 73 and Kato Zagros, Section 74), deposited at around 1000 m depth, suggest a much wider original distribution of the lower Pliocene than presently observed (Table 1). The Trubi-facies unconformably overlie the Upper Sequence and the Prina Group, suggesting a very rapid latest Miocene subsidence, comparable with the history inferred from Chora Sfakia (Section 30). The palaeobathymetry of Langada (Section 68) shows that uplift of eastern Crete occurred after approximately 4.5–4 Ma (Table 1).

Koufonisi

Koufonisi is part of a largely submerged fault-bounded swell, separated from Crete by an E–W trending graben (Peters, 1985). The island exposes a series of terrestrial conglomerates and sandstones at the base, overlain by late Tortonian marine sandstones. Tortonian subsidence led to deep-marine sedimentation around 700 m in the late Tortonian and early Messinian (Table 1). The Tefelion group is overlain by a finely laminated calcareous sequence of the Vrysses Group, alternating with coarse and unsorted mass–flow deposits and slumps (Peters, 1985). The Lower Pliocene marl breccia interbeds eroded and reworked part of the upper Messinian mass–flow deposits. It is overlain by a sequence of middle Pliocene alternating marls and sapropels (Dermitzakis & Theodoridis, 1978; Theodoridis, 1984). Dermitzakis & Theodoridis (1978) reported the presence of *Uvigerina* spp. from the Pliocene sediments, indicating a depositional depth of more than ~200 m (van Hinsbergen et al., 2005a). The final uplift of Koufonisi must have occurred since the Late Pliocene.

**DISCUSSION**

The geologic histories of the Cretan sub-basins presented above and summarised in Fig. 7 are integrated into a palin-
spastic reconstruction (Fig. 8). The combination of two tectonic processes may explain the Neogene basin development history of Crete: formation of the South Aegean core complex as a result of N–S extension, followed by E–W, arc-parallel extension associated with the opening of the Aegean arc (Jolivet et al., 1996; ten Veen & Kleinspehn, 2003).
Several lines of evidence suggest that the first exposure of the Lower Sequence on Crete occurred during the early stages of Neogene sedimentation. The Topolia Fm and Goudouras Fm of the Prina Group, and the Males Fm and Viannos Fm of the Tefelion Group do not contain fragments derived from the Lower Sequence, but their source areas, indicated by palaeoflow directions, presently widely expose the Lower Sequence. The exposure of the Lower Sequence in these areas must therefore post-date the deposition of the Prina Group and the Males and Viannos Fm.

There is strong evidence that the exhumation of the Lower Sequence during the deposition of the Neogene Cretan sequences was not the result of (only) erosion, but
had a tectonic cause. Firstly, the contact between the Prina Group and the Lower Sequence is always a fault. Secondly the vast areas that at present expose the Lower Sequence in between the scattered occurrences of the Upper Sequence and the Prina Group deposits frequently form the Tortonian basin floors, whereas the basin margins are formed by the Upper Sequence (e.g. in the Toplou and Sitia areas, and on western Crete). Would these basins have been simple half-grabens or synforms, one would have seen the opposite. We therefore propose that the isolated occurrences of the Upper Sequence represent 'extensional klippen' that broke off the hanging wall during the activity of the extensional detachment (Fig. 9). Once extensional klippen have formed, they form passive rock bodies on top of the footwall, and the outcrop of the (top-to-the-north) extensional detachment shifts northwards. This process allows the explanation of the Middle to early Late Miocene sedimentary history and palaeogeography of Crete in terms of a supradetachment basin evolution.

In such a basin, an elongated depression parallel to the outcrop of the detachment will be characterised by longitudinal flow directions and infill from the sides: one sedimentary system flowing from the hanging wall to the basin axis, and one from the exhumed and domed footwall (e.g. Friedmann & Burbank, 1995). The effect the break-up of the hanging wall by the formation of extensional klippen has on Crete has resulted in a northward migration of these co-existing sedimentary systems: During the exhumation of the Lower Sequence between 24 and 15 Ma (Jolivet et al., 1996; Thomson et al., 1998) the Upper Sequence that is presently exposed on Crete still formed a coherent hanging wall. An approximately E–W trending depression south of Crete reflecting the surface trace of the Cretan detachment still existed during the deposition of the oldest Prina Group sediments, as evidenced by the north-to-south palaeoflow directions in both the Topolia Fm and the Goudivas Fm (Figs 8 and 9). It should be noted that the Prina Group sediments preserved on top of the extensional klippen presently have a much wider distribution than during their sedimentation, resulting in a younger N–S extension. As the break-up of the Upper Sequence into extensional klippen continued, the longitudinal sedimentary system in the elongated depocentre parallel to and on top of the surface trace of the extensional detachment shifted northward. As a result, the Prina Group sediments are unconformably overlain by the longitudinal sedimentary system formed by the proximal Male river system, via the lacustrine parts of the Vianno system to the distal deep-marine Potamos-Palaeohora deposits in the west (Figs 8 and 9). The westward deepening perpendicular to the extension direction most likely resulted from the larger extensional strain and probably higher strain rate in the west than in the east (van Hinsbergen et al., 2005b).

This situation ended with rapid subsidence and SSWward tilting, which led to olistolith emplacement on central and eastern Crete around the Serravallian–Tortonian boundary (Fig. 8), and emergence of western Crete and Gavdos (Fig. 8). During this period, the active detachment shifted towards the area north of western Crete, which led to an uplift of western Crete as isostatic response to the removal of the hanging wall. Note that this rebound must have affected entire western Crete; it cannot be explained by the formation of the west-Cretan anticline as suggested by Jolivet et al. (1996), as this folding is much younger (see below). This northward shift as a result of the break-up of the hanging wall into extensional klippen resulted for the first time on present-day Crete in the unconformable deposition of Neogene sediments, on top of the Lower Sequence (Apostoli (Section 41)).

Once the outcrop of the detachment had shifted to the north, Crete entirely consisted of the footwall of the Cre-
Neogene supradetachment basin development on Crete (Greece)

tan detachment or the South Aegean core complex. Therefore, the rest of the Neogene history of Crete cannot be explained in terms of a supradetachment basin, and the exhumed footwall independently deformed from the extensional detachment further to the north. It is possible that the activity of Cretan detachment to the north of Crete continued, but its activity would no longer have affected the rocks exposed on the present-day island.

The oldest indications for the onset of E-W extension deforming the South Aegean core complex are found in the Ierapetra area, where Tortonian palaeoflow directions indicate eastward tilting (Fortuin, 1977; ten Veen & Postma, 1999a; Fig. 8). Although no extension direction was derived, the rapid subsidence on Gavdos around 10 Ma (Figs 6 and 8) following the earliest Tortonian uplift is probably also the result of this change in tectonic regime.

As mentioned above, Jolivet et al. (1996) interpreted the west-Cretan anticline as the result of doming of the footwall of the extensional detachment. Alternatively, Zulauf et al. (2002) suggested that the Orno Oros anticline in eastern Crete resulted from compression. Two facts suggest that the Cretan antiforms are the result of NNE–SSW compression post-dating the activity of the Cretan detachment. Firstly, in the Talea Ori mountains north of the Psiloritis, the Cretan detachment is deformed into a synform with an overturned northern limb (profile D–D’ in Fig. 3), which is unlikely to occur in an extensional setting. Secondly, syn-sedimentary tilting of the northern limb of the west-Cretan antiform and the syn-sedimentary northward tilting in the Apostoli region that we explained above as the result of folding of the Psiloritis antiform (Fig. 4) shows that the formation of the Cretan antiforms post-dates the activity of the Cretan detachment by at least several millions of years (Fig. 8).

We therefore conclude that around 10 Ma, the N–S extension no longer affected Crete, and instead basin formation occurred along high-angle normal faults in response to E-W extension, which occurred contemporaneously with the formation of large-scale antiforms in response to NNE–SSW compression (Fig. 9). These antiforms are not perpendicular to the inferred extension direction. Instead, they are at an angle of approximately 60°–70°. We propose that these antiforms developed as left-lateral transpressional folds in response to the southwestward motion of the Aegean over the northward moving African plate as a logical consequence of the westward extrusion of Anatolia since approximately 13 Ma (Dewey & Sengör 1979; Armiyo et al., 1999; Hubert-Ferrari et al., 2002; Sengör et al., 2003) and the contemporaneous clockwise rotations in western Greece (van Hinsbergen et al., 2005d; in press).

Mammal faunas reported by De Bruijn et al. (1971) and De Bruijn & Meulenkamp (1972) led Drooger & Meulenkamp (1973) to suggest that the connection of Crete to the European mainland was lost in the course of the early Tortonian, as a result of the onset of foundering of the Cretan Basin underlying the Sea of Crete (Fig. 8). We follow this conclusion, as it is in line with the fact that the youngest sediments that provide evidence for a northern sediment source are the deep-marine early Tortonian sediments in the Ierapetra area (Kalavmaka, Section 64; Fortuin, 1978; Fortuin & Peters, 1984).

Subsidence and E-W extension continued throughout the Late Miocene and possibly the earliest Pliocene, as can be concluded from the unconformable contact of deep-marine Lower Pliocene with the pre-Neogene basement in eastern Crete, or with terrestrial conglomerates unconformably overlying pre-Neogene basement in the Frangokastello area. The facies change from dominantly clastic sediments in the Tefelion Group to the carbonate-rich sediments of the Vrysses Group may be associated with ongoing subsidence as a result of E-W extension, leading to submergence of siliciclastic source areas and starvation of clastic sediment supply. Alternatively, this facies change may have a climatic cause.

In the earliest Pliocene, marl breccia interbeds were in many places deposited in the deeper parts of basins (Fig. 8). During (part of) the Messinian salinity crisis preceding the Pliocene, the Mediterranean seawater is believed to have largely evaporated (Krijgsman et al., 1999). We tentatively explain these uniquely Lower Pliocene mass flow deposits as the result of re-equilibration after the Pliocene flooding of instable submarine slopes caused by ongoing subsidence and extension during the late Messinian low stand. This interpretation is supported by the onlap of deep-marine Tufi-facies sediments onto the basement (Kheratokambos, Section 73 and Kato Zagros, Section 74) or separated from the basement by only a few metres of terrestrial sediments (Chora Sfakia, Section 30).

Various sections provide conclusive evidence for uplift of Crete during the Pliocene (Fig. 6, Table 1). Comparing the vertical motion reconstructions of the Heraklion area with palaeobathymetry reconstructions from the Sea of Crete (Wright, 1978) led Meulenkamp et al. (1994) to conclude that the net uplift of Crete was associated with northward tilting. Our data confirm this suggestion; for instance, the terrestrial Aghia Galini group in the southern Heraklion area is time equivalent to the marine sections in the north, despite the fact that the initial Early Pliocene bathymetry for both areas was comparable, or even deeper in the south than in the north (Table 1). Combining this with the palaeobathymetrical and lithological information available from the Upper Pliocene of western and eastern Crete suggests that the tilting occurred along an ENE–WSW striking axis (Fig. 8). The Pliocene uplift of Crete was explained as the result of left-lateral transpression resulting from the oblique collision of the Southern Aegean Arc with the African promontory (Masclè et al., 1999; ten Veen & Kleinspehn, 2003), which is at present still found by GPS analysis (McClusky et al., 2000). This oblique collision most likely is the combined effect of the northward motion of Africa, the westward motion of the Anatolia and the Aegean region, and southward gravitational spreading of the Aegean region over the African plate (Jolivet, 2001).
SUMMARY AND CONCLUSIONS

In this paper, it is shown that the Middle to early Late Miocene basin history of Crete resulted from activity of the Cretan detachment and the late stages of exhumation of the South Aegean core complex, through integration of a new detailed palaeobathymetry-based vertical motion reconstruction and stratigraphy with a synthesis of published structural and sedimentological data.

The oldest deposits in the Cretan supradetachment basin were uniquely derived during the Middle Miocene from the non-metamorphic hanging wall of the Cretan detachment. Subsequent break-up of the hanging wall into extensional klippen resulted in the northward shift of the surface trace of the extensional detachment. This led in the late Serravallian (late Middle Miocene) to an east-west running longitudinal sedimentary system, with the Males river in the east, via the lacustrine part of the Vian-nos system in central Crete, to deep-marine conditions on Gavdos and southwestern Crete. The westward deepening trend may reflect increasing extensional strain along the detachment from east to west.

Until the earliest Tortonian, the hanging wall that covered Crete until the Middle-Late Miocene transition was extended. This led initially to the deposition of large olistoliths in an SSE-direction in central and eastern Crete, whereas western Crete was uplifted, possibly as a flexural response to the removal of the hanging wall. The hanging wall – including the overlying sediments and olistoliths – further extended and formed extensional klippen. Between these, basins developed, with exhumed metamorphic rocks as basin floor until approximately 10 Ma.

Afterwards, E-W extension associated with the outward motion of Crete during the curvature of the Aegean arc formed N-S trending (half-)grabens. Simultaneously, large open folds deformed Crete, striking at a 20°-30° angle to the inferred extension direction. We therefore interpret these as a result of left-lateral transpression, because of the extension direction. We therefore interpret the movement of Crete during the curvature of the Aegean arc.

In the Pliocene, Crete was uplifted and emerged. This uplift was associated with NNW-ward tilting along an ENE-WSW trending axis. This is explained as the result of the collision of Crete with the African promontory.

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Supplementary Material

The following supplementary material is available for this article online:

Appendix SI. Stratigraphic correlations of the analysed sections

Appendix S2. Calculated depth values for all samples of the Cretan sections

Figure S1a. Age correlations, based on litho-bio- and magnetostratigraphy for southwestern Crete, Gavdos and the Frangokastello area.

Figure S1b. Age correlations, based on litho-bio- and magnetostratigraphy for northwestern Crete and the horst of Chania.

Figure S1c. Age correlations, based on litho-bio- and magnetostratigraphy for the Rethymnon area.

Figure S1d. Age correlations, based on litho-bio- and magnetostratigraphy for the Pliocene cyclic sections of the northern part of the Heraklion area.

Figure S1f. Age correlations, based on litho-bio- and magnetostratigraphy for eastern Crete and Koufonisi. Cycles in Appendix Ie are correlated to the precession minima in the target curve of (Laskar et al., 1993).

Italic letters refer to: a = Bianchi et al. (1985); b = Driever (1988); c = Drooger, et al. (1979); d = Freudenthal (1969); e = Hilgen et al. (1995); f = Hilgen, et al. (1997); g = Jonkers (1984); h = Krijgsman, et al. (1994); j = Krijgsman, et al. (1995); k = Langereis (1984); l = Meulenkamp (1969); m = Negri and Villa (2000); n = Postma et al. (1993a); o = Postma and Drinia (1993); p = Spaak (1981); q = Spaak (1983); r = Thomas (1980); s = van der Zwaan (1982); t = Zachariasse (1975); u = Zachariasse (1979); v = Zachariasse and Spaak (1979); w = W.J. Zachariasse, pers. comm. (2004).

Encircled letters correspond to bioevents. Numerical ages are obtained from Perch-Nielsen (1985); Rio, et al. (1990a, b) and Lourens et al. (2004): A = first occurrence (FO) Neogloboquadrina acostaensis (11.78 Ma); B = first common occurrence (FCO) N. acostaensis (10.57 Ma); C = FO dominant left-coiling N. acostaensis (9.44 Ma); D = last occurrence (LO) G. menardii 5 (7.51 Ma); E = FOC G. menardii 5 (7.33 Ma); F = FO G. conomiozea (7.24 Ma); G = FO Tertiary time Irregulariella multiloba (6.42 Ma); H = sinistral to dextral coiling change N. acostaensis (6.35 Ma); J = Sphaeroideinellipses acme (5.30–5.21 Ma); K = FO G. margaritae (5.08 Ma); L = FO Reticulofenestra antarctica (5.04 Ma); M = FO Ammonolithus tricinocalutus (4.52 Ma); N = FOC G. punctulata (4.52 Ma); O = FOC Gephyrocapsa spp. (4.33 Ma); P = FOC Discocysta asymetricus (4.12 Ma); Q = last common occurrence (LCO) G. margaritae (3.98 Ma); R = FO S. pseudoumbilica (3.84 Ma); S = FO G. crassiformis (3.60 Ma); T = FO G. punctulata (3.57 Ma); U = FO G. bononiensis (3.31 Ma); V = FO Sphaeroideinellipses seminula (3.17 Ma); W = FO D. acostaensis (2.54 Ma); X = FO G. bononiensis (2.39 Ma); Y = FO G. inflata (2.09 Ma); Z = FOC common left-coiled N. acostaensis (1.78 Ma).

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