Basin Research

IAS EAGE

Basin Research (2011) **23**, 678–701, doi: 10.1111/j.1365-2117.2011.00507.x

Formation and fragmentation of a late Miocene supradetachment basin in central Crete: implications for exhumation mechanisms of high-pressure rocks in the Aegean forearc

Willem Jan Zachariasse,* Douwe J.J. van Hinsbergen†,‡ and Anne Rutger Fortuin§

**Stratigraphy and Paleontology group, Faculty of Geosciences, Utrecht University, Utrecht, The Netherlands*
†Physics of Geological Processes, University of Oslo, Oslo, Norway
‡Center for Advanced Study, Norwegian Academy of Science and Letters, Oslo, Norway
§Department of Sedimentology, Faculty of Earth and Life Sciences, Vrije Universiteit, Amsterdam, The Netherlands

ABSTRACT

We present a new lithostratigraphy and chronology for the Miocene on central Crete, in the Aegean forearc. Continuous sedimentation started at ~10.8 Ma in the E–W trending fluvio-lacustrine Viannos Basin, formed on the hangingwall of the Cretan detachment, which separates high-pressure (HP) metamorphic rocks from very low-grade rocks in its hangingwall. Olistostromes including olistoliths deposited shortly before the Viannos Basin submerged into the marine Skinias Basin between 10.4 and 10.3 Ma testifies to significant nearby uplift. Uplift of the Skinias Basin between 9.7 and 9.6 Ma, followed by fragmentation along N–S and E–W striking normal faults, marks the onset of E–W arc-parallel stretching superimposed on N–S regional Aegean extension. This process continued between 9.6 and 7.36 Ma, as manifested by tilting and subsidence of fault blocks with subsidence events centred at 9.6, 8.8, and 8.2 Ma. Wholesale subsidence of Crete occurred from 7.36 Ma until ~5 Ma, followed by Pliocene uplift and emergence. Subsidence of the Viannos Basin from 10.8 to 10.4 Ma was governed by motion along the Cretan detachment. Regional uplift at ~10.4 Ma, followed by the first reworking of HP rocks (10.4–10.3 Ma) is related to the opening and subsequent isostatic uplift of extensional windows exposing HP rocks. Activity of the Cretan detachment ceased sometime between formation of extensional windows around 10.4 Ma, and high-angle normal faulting cross-cutting the detachment at 9.6 Ma. The bulk of exhumation of the Cretan HP-LT metamorphic rocks occurred between 24 and 12 Ma, before basin subsidence, and was associated with extreme thinning of the hangingwall (by factor ~10), in line with earlier inferences that the Cretan detachment can only explain a minor part of total exhumation. Previously proposed models of buoyant rise of the Cretan HP rocks along the subducting African slab provide an explanation for extension without basin subsidence.

INTRODUCTION

Sedimentary basins provide a geological archive of style and timing of tectonic activity, and can be used to determine a minimum age of crustal extension and related subsidence. On the island of Crete in the southern Aegean forearc, the earliest sediments were interpreted to have been deposited in a supradetachment basin, governed by the activity of an extensional detachment that separated high-pressure, low-temperature (HP-LT) metamorphic rocks from very low-grade metamorphic nappes (van Hinsbergen & Meulenkamp, 2006). At an ill-defined time, high-angle faults cross-cut this low-angle detachment, and (half-) grabens developed, followed by Pliocene strike-slip-dominated deformation, leading to a highly complex mosaic of faults and fault-blocks and severe fragmentation of the oldest basin stratigraphy (ten Veen & Postma, 1999a, b; Fassoulas, 2001; ten Veen & Kleinspehn, 2003). These younger deformation phases hamper the reconstruction of the supposed supradetachment basin and makes the dating of its further subsidence and fragmentation due to high-angle faulting notoriously difficult. As a result, the onset and end of the supradetachment basin history has been dated tentatively until now and the evolution – even the very existence – of such a basin is not yet fully clarified. In this study, we document the late

Correspondence: Willem Jan Zachariasse, Stratigraphy and Paleontology group, Faculty of Geosciences, Utrecht University, Budapestlaan 4, 3584 CD Utrecht, The Netherlands. E-mail: jwzach@geo.uu.nl

Miocene (Tortonian) basin evolution in central Crete based on well over 10 years of fieldwork and biostratigraphic analysis. We established a new high-resolution time-stratigraphic frame for the Miocene basin fill and a detailed map of the central, stratigraphically most complex part of the basin. The goals are (1) to portray the temporal evolution of the supradetachment basin in central Crete from inception to the onset of fragmentation through high-angle normal faulting that today dissects the Cretan detachment and (2) to assess the minimum role of upper crustal extension, as recorded by subsidence and basin formation, in the exhumation of HP-metamorphic rocks on Crete. We discuss the implications of our findings for mechanisms of the early to middle Miocene exhumation of the Cretan HP-LT rocks that occurred before basin formation.

SETTING AND RATIONALE

The Aegean region is one of the best-studied continental extensional provinces in the world. The continental crust of the Aegean consists of a stack of upper crustal nappes that accreted during northward dipping, long-lived and still-active subduction that accommodated Africa-Europe convergence since the Cretaceous (Jacobshagen, 1986; Faccenna *et al.*, 2003; van Hinsbergen *et al.*, 2005a). Extension of the Aegean region has been documented from the southern forearc until southern Bulgaria (Fig. 1), and is generally ascribed to southward roll-back of the subducting slab, and associated extensional collapse of the overriding plate orogen (le Pichon *et al.*, 1982; Gautier *et al.*, 1999; Jolivet, 2001).

Among the most prominent manifestations of extension is the widespread exposure of Cenozoic high-pressure, low-temperature (HP-LT) metamorphic rocks, which in some cases experienced a syn-decompressional high-temperature overprint (Gautier *et al.*, 1999; Tirel *et al.*, 2009; Jolivet & Brun, 2010; Ring *et al.*, 2010). Exhumation of these rocks was invariably associated with extensional detachments, which frequently accommodated large-magnitude displacement of metamorphic footwalls with respect to lower-grade or nonmetamorphic hanging walls (Lister *et al.*, 1984; Gautier *et al.*, 1993, 1999; Jolivet, 2001, 2003; Tirel *et al.*, 2009).

The amount and timing of crustal extension is as a result frequently estimated by the study of timing and amount of exhumation of the metamorphic culminations along these extensional detachments (e.g. Jolivet, 2001; Forster & Lister, 2009; Jolivet & Brun, 2010). However, as stressed recently by Ring *et al.* (2010) and Ring & Glodny (2010), the presence of an extensional detachment does not necessarily indicate upper crustal extension. One of the clearest examples of this is the South Tibetan Detachment in the Himalayas, which accommodated at least several tens of kilometres of extrusion of a slice of metamorphic rocks (the Greater Himalayas) in a setting of overall compression and crustal shortening (Burchfiel & Royden, 1985; Searle & Godin, 2003; Long & McQuarrie, 2010). A similar mode of extrusion may explain part of the exhumation of HP-LT rocks in the central Aegean region (Ring *et al.*, 2010).

The island of Crete today forms a horst in the forearc of the Aegean region. It has been affected by N–S and E–W stretching, forming a complex mozaic of fault blocks and associated late Miocene and younger sedimentary basins (Meulenkamp *et al.*, 1988; van Hinsbergen & Meulenkamp, 2006). Moreover, Crete and the Peloponnesos (Fig. 1) expose two HP-LT metasedimentary nappes (the Lower Nappes, LN, comprising the Plattenkalk, and overlying Phyllite-Quartzite units), separated from a sequence of three low-grade to nonmetamorphic nappes (the Upper Nappes, UN, comprising from bottom to top the Tripolitza, Pindos and Uppermost units) by a top-to-the-north extensional detachment (Jolivet *et al.*, 1996, 2010). The Phyllite Quartzite unit of the LN was buried to a depth of $\sim$30 km (8–12 kbar/350 °C)

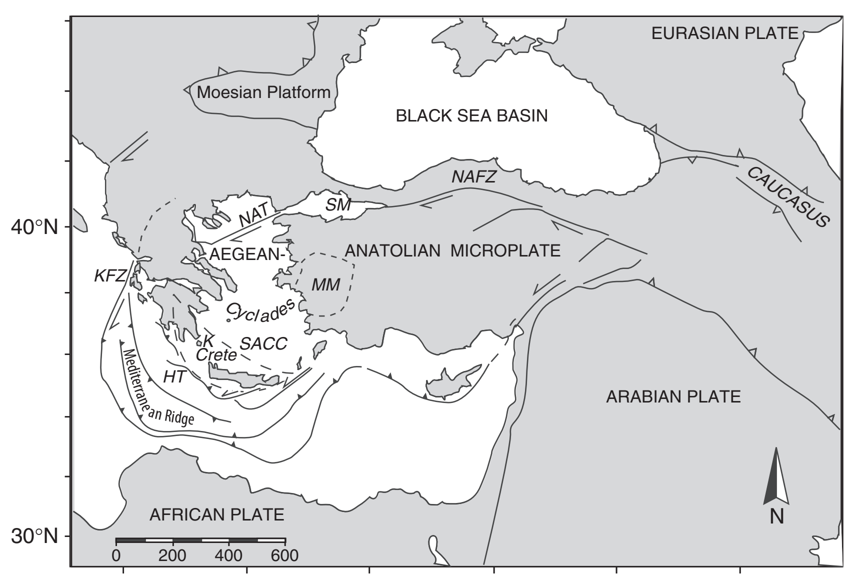


Fig. 1. Map of the Aegean region. HT, Hellenic Trench; NAFZ, North Anatolian Fault Zone; NAT, North Anatolian Trough; KFZ, Kephalonia Fault Zone; SACC, South Aegean Crystalline Complex; K, Kythira; MM, Menderes Massif; SM, Sea of Marmara.

around 24–21 Ma ago, constrained by $^{40}Ar/^{39}Ar$ ages (Seidel *et al.*, 1981; Theye & Seidel, 1991; Theye *et al.*, 1992; Jolivet *et al.*, 1996; Brix *et al.*, 2002), and was exhumed to 2–3 km depth around 15–12 Ma, suggested by apatite fission track ages (Thomson *et al.*, 1998; Marsellos & Kidd, 2008; Marsellos *et al.*, 2010). Assuming that all exhumation of the LN occurred along the Cretan detachment, up to 100 km of horizontal Miocene extension was ascribed to it (Ring *et al.*, 2001; van Hinsbergen *et al.*, 2005c; van Hinsbergen & Meulenkamp, 2006).

However, this interpretation was challenged by Rahl *et al.* (2005), who showed that there is only a small peak metamorphic temperature contrast of $\sim$60 °C across the Cretan detachment, and that the lower parts of the UN experienced low-grade metamorphic conditions. They suggested that the Cretan detachment only accommodated $\sim$6 km of vertical exhumation. Although translating temperature to depth in a forearc is not straightforward, their intepretation seems to be supported by the fact that although the three nappes of the UN are all preserved, their combined thickness on Crete is no more than $\sim$2 km, as opposed to their nonextended $\sim$25 km cumulative thickness in western Greece (Bonneau, 1984; Jacobshagen, 1986; van Hinsbergen *et al.*, 2005a). This suggests that exhumation of the LN was for a large part accommodated by essentially pure-shear thinning of the UN. In addition, a strong top-to-the-north asymmetry of the Cretan detachment suggested by Jolivet *et al.* (1996) is not generally agreed upon, and co-axial deformation has been proposed (Fassoulas *et al.*, 1994; Kilias *et al.*, 1994; Papanikolaou & Vassilakis, 2010).

Dating and assessment of the tectonic style of the sedimentary basins of Crete may shed light on the role of crustal extension and subsidence associated with exhumation of the LN in the southern Aegean forearc. We will discuss the implications of this basin evolution for the syn- and presedimentary exhumation history of these rocks.

GEOLOGY OF CENTRAL CRETE

The Heraklion Basin in central Crete is the island's largest Neogene basin covering some 900 km^2 (Fig. 2a). It is bounded to the W by the Psiloritis Mountains and the Platis River, to the S by the Asteroussia Mountains, and to the E by the Dikti Mountains. An erosional remnant of Neogene sediments is preserved NE of Anogia in the Psiloritis Mts (Fig. 2b). Neogene deposits in the Heraklion Basin and in the Psiloritis overlie basement rocks of the Upper Nappes (UN). These comprise, from bottom to top, the nonmetamorphic (or very low-grade metamorphic: Rahl *et al.*, 2005; Klein *et al.*, 2008) Tripolitza and Pindos nappes and a heterogeneous Uppermost Unit, which consists of ophiolite-related units, some of which are metamorphosed (Seidel *et al.*, 1981; Bonneau, 1984) (Fig. 2a). The UN rocks in the Psiloritis and Dikti Mts are separated from the HP-LT metamorphic rocks of the Lower Nappes (LN) by the Cretan detachment (Fig. 2b). This fault (or faults) is in many places exposed with a sense of shear either to both S and N (Kilias *et al.*, 1994; Fassoulas, 1999; Papanikolaou & Vassilakis, 2010), or dominantly top-to-the-north (Jolivet *et al.*, 1996). The LN comprises the Phyllite-Quartzite and Plattenkalk units (Fig. 2a) (for an overview of the alpine basement rocks in central Crete, see e.g. Bonneau (1984)). The generally lead-grey coloured metamorphic Plattenkalk consist of crystalline carbonates showing mm–cm scale crystals. Chert beds are common. This lithology is easily distinguishable from that of the nonmetamorphic Tripolitza unit with its dominantly bluish grey mud- and wackestones and subordinate intercalations of dark-grey crystalline dolomites. Quartzites and schists occur both in the LN and the Uppermost Unit of the UN. Therefore, determining the erosion products of the UN and LN rocks in Neogene sediments is most straightforward by identification of Plattenkalk debris.

Fault kinematic studies were carried out by Angelier (1975), Meulenkamp *et al.* (1988), ten Veen & Postma (1999a), Fassoulas (2001), and ten Veen & Kleinspehn (2003). These studies distinguished two successive phases of normal faulting along NE–SW, and combined E–W and N–S patterns followed by transpressional faulting along a conjugate system of NNW–SSE and NNE–SSW oblique slip faults. The timing of these phases, however, differs greatly from one author to the other. Foraminifera-based vertical motion studies on central Cretan sections were conducted by Meulenkamp *et al.* (1994), ten Veen & Postma (1999b), van Hinsbergen & Meulenkamp (2006) and Zachariasse *et al.* (2008). Detailed field stratigraphic studies in limited geographic areas of the Heraklion Basin (including detailed geological maps) have been made by Delrieu (1990) and Tsagaris (1991).

MATERIALS AND METHODS

Sections and outcrops

Locations of sections and outcrops in central Crete are given in Fig. 2b and in supporting information (Figure S1). All but two of these sections and outcrops were never previously published. Appendix S1 shows details on location, lithology, biostratigraphy and palaeobathymetry of the studied sections and outcrops along with their lithostratigraphic and biozonal assignment.

Palaeobathymetry

Depositional depth is based on the presence of species of benthic foraminifers, which – in the modern Mediterranean – occupy specific depth intervals (van Hinsbergen *et al.*, 2005b). In some cases, we further documented palaeobathymetry using percentages of planktonic foraminifers and the regression of van der Zwaan *et al.* (1990). All depositional depth estimates are given in Appendix S1. The terms shallow and deep marine as used in text refer to inner neritic ($<$100 m) and outer neritic (100–200 m) to bathyal ($>$200 m), respectively.

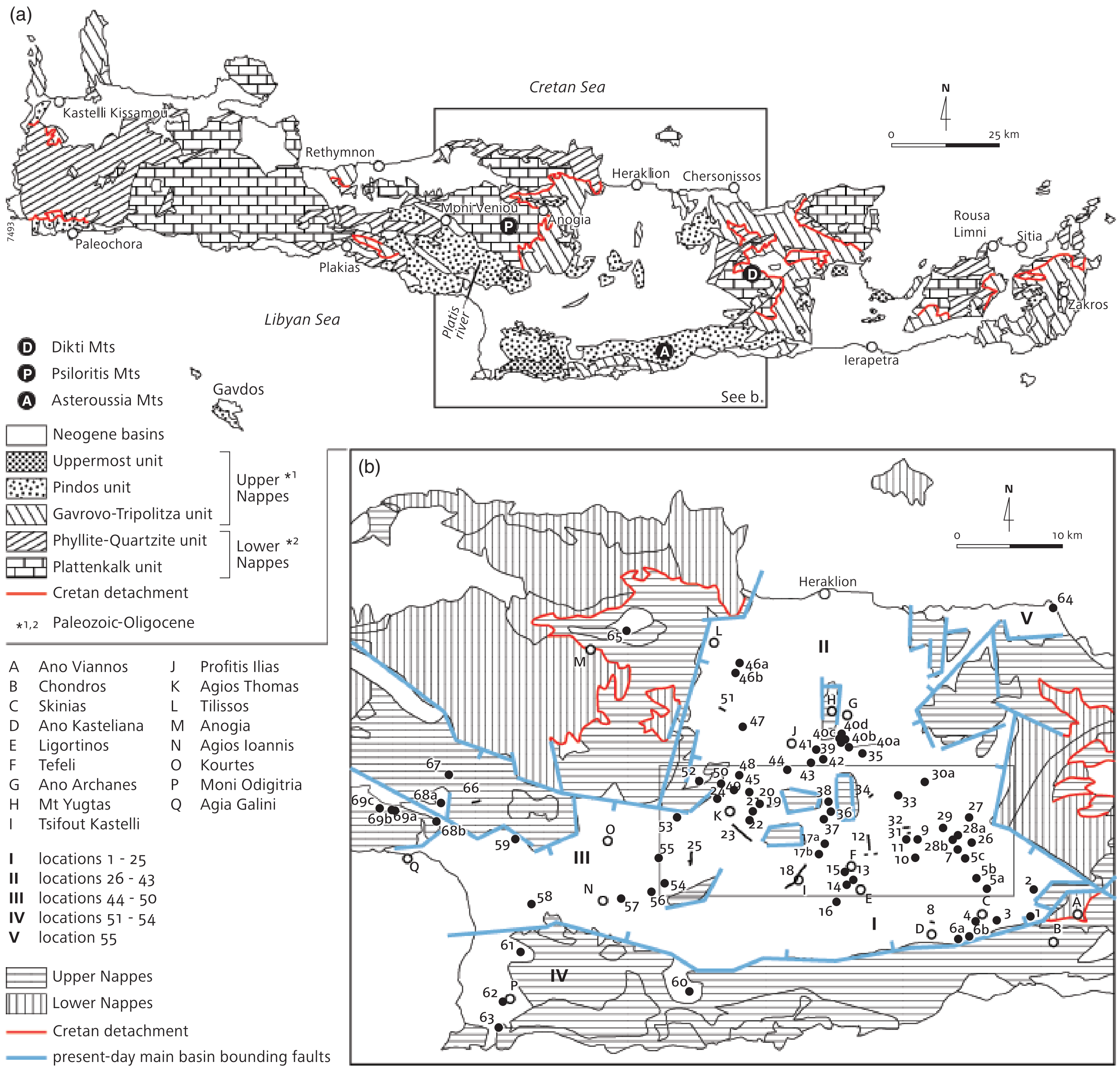


Fig. 2. (a) Geological map of Crete modified from Bornovas & Rontogianni-Tsiabaou (1983) with locations referred to in text. The Lower Nappes (LN) consists of two units showing HP-LT metamorphism: the Plattenkalk unit (below) and the Phyllite-Quartzite unit (above). The Upper Nappes (UN) is without HP-LT metamorphism and made up from base upward by the Gavrovo-Tripolitza unit, Pindos unit and so-called Uppermost Unit. The units of the LN and UN are separated by the Cretan detachment. (b) Location of sections and outcrops in central Crete (see also map in Figure S1). Box refers to geological map in Fig. 7. Information on lithology, age and depositional environment is given in Appendix S1.

Time-stratigraphic frame

Marine sections and outcrops are dated by correlating presence/absence patterns and coiling of planktonic foraminiferal marker species to a middle to late Miocene (~12 - 6.8 Ma) succession of 11 biozones, based on an update of the planktonic foraminiferal biostratigraphy in Krijgsman *et al.* (1995) for sections Gibliscemi on Sicily (see Figure S2) and Metochia on Gavdos (see Figure S3). Definition and description of biozones are given in Appendix S2. Ages for biozonal boundaries in sections Metochia and Gibliscemi are derived from an updated chronology for these sections based on magnetostratigraphy and astronomical tuning of sedimentary cycle patterns (Lourens *et al.*, 2004; Hüsing *et al.*, 2007). Ages for defining bioevents and magnetic chron boundaries are summarized in Table S1.

TORTONIAN-LOWER MESSINIAN STRATIGRAPHY

No formal lithostratigraphic subdivision is available for the Miocene deposits in central Crete. Most authors applied the preliminary formation scheme of Meulenkamp *et al.* (1979), but we found this scheme inadequate to fully portray the geological history of the Heraklion Basin. Therefore, we upgraded and formalized the lithostratigraphy for the Tortonian-lower Messinian deposits in central Crete in terms of six formations, defined below from bottom to top. Each formation is characterized by a unique combination of lithology, depositional environment, and stratigraphic position. Based on the absence and presence of these formations, we have identified four regions (I–IV, Figs 2 and 4 and defined by locations in Appendix S1), each of which is characterized by a unique stratigraphic succession. Figure S4 shows how the new lithostratigraphic subdivision compares with those published earlier.

Viannos Formation

Lithology. This oldest unit is dominated by alternations of clay, silt and cemented sandstone. Sandstone occurs either as 2–6 m thick tabular bodies, or as thinner and finer grained, often lens-shaped interbeds within mudstone-silt intervals. Within the sandstones, sedimentary structures are common, varying from small-scale wave-and current ripple cross-bedding to tabular megabeds, and scour-and-fill structures. Palaeocurrent readings indicate dominantly W-directed drainage (see in Appendix S1). Platy weathering (silty) limestones are common (Photo 3 in Figure S5), whereas lignitic clays occur infrequently within clayey intervals. Limestones and clays may contain fresh water molluscs (*Theodoxus*, *Melanopsis*, *Unio*, and sometimes *Planorbis*), operculae of the freshwater gastropod *Bithynia*, gyrogonites of charaphytes, and smooth ostracods. Conglomerates are relatively rare. They are clast-supported and made up of poorly sorted, (sub) rounded debris from the Uppermost Unit (locations 6b, 27, 37, 67) and only rarely admixed with Pindos limestone pebbles (e.g. along new road immediately south of location 5a), or with pebbles from the Tripolitza unit (locations 36 and 46b). In only one location (30a) conglomerates are dominated by Tripolitza limestone debris (for details, see Appendix S1). Unsorted breccio-conglomerates occur in the NW (W of Kato Kalesa, location 46b) and along the main road from Agia Galini to Spili in the Rethymnon province (locations 69a, c). The (sub) angular debris in both locations is derived exclusively from the Uppermost Unit and alternate with greenish-blue silty and sandy interbeds (locations 69a, c) or coarse to pebbly sandstones (location 46b) (for details, see Appendix S1). The isolated nonmarine deposits at Chersonisos in the NE of the Heraklion Basin are assigned to the Viannos Fm, following IGME (1989). The thick beige-coloured clay beds and some intervening lignites in this area contain freshwater gastropods dated by Willmann (1980) as early Tortonian and/or Serravallian. Intercalated sandstones occasionally show large-scale cross-bedding and wave ripples. Silty clays and silts in the upper part of the sequence in location 64 (Fig. 3) are sometimes laminated and contain abundant calcareous worm tubes and smooth to weakly ornamented ostracods.

Thickness and type section. The maximum exposed thickness of the Viannos Fm is estimated at about 400 m (IGME, 1989, 1994). Unfortunately, the formation can nowhere be traced from bottom to top in a continuous section. The best and most expanded sections, here designated as a composite type section, are found along the main road to Viannos coming from Heraklion (Fig. 3, sections 1.1 and 1.3–1.5; Photos 1–2 in Figure S5).

Underlying unit. Stratigraphic contacts of the Viannos Fm with basement are found in only two places: to the S of location 46b and to the SW of the Heraklion Basin (locations 69a, c). In both locations, locally derived, very coarse angular debris overlies the Uppermost Unit.

Depositional environment. The assembly of sediments of the Viannos Fm covers channel-belt, overbank and lake deposits. The (sub) angular and unsorted character of the components in the breccio-conglomerates at locations 46b and 69a, c probably belong to an alluvial fan with sheetflood deposition. The fine-grained sediments with abundant worm tubes and ostracods in the upper part of the Viannos Fm at location 64 most likely represent deep lake deposits.

Geographic extent. Sediments of the Viannos Fm are widespread in areas I and II. An erosional remnant occurs to the NE of Anogia (location 65). They are entirely absent in areas III and IV. The nonmarine deposits to the N of the road Lagolia – Klima (location 59), assigned to the Viannos Fm on the 1 : 50 000 geological map of Greece (IGME, 1989), belong, in our opinion, to the (younger) Kasteliana Fm (Appendix S1: location 59).

The occurrence of olistoliths. In the western part of area II, the landscape is dotted with exotic blocks of Tripolitza limestone or cemented Tripolitza limestone breccia (Photos 4 and 9 in Figure S5) mapped as Illias Fm on sheet Epano Archanae (IGME, 1994). They are surrounded by less cemented breccio-conglomerates dominated by (sub) angular Tripolitza limestone debris (Photos 6 and 10 in Figure S5), which sometimes incorporate packages of Viannos type of silty clays and sandstones (Photo 11 in Figure S5). Debris-filled fissures and cavities indicate that the blocks were covered by these breccio-conglomerates. This all indicates that these exotic blocks and associated patches of breccio-conglomerates are the erosional remnants of olistostromes of unsorted (sub) angular Tripolitza limestone debris with olistoliths [a nice example is shown on Photo 5 in Figure S5 (location 38)]. A similar olistostrome with several cemented Tripolitza breccia olistoliths in location 44b overlies a debris flow of exclusively (sub) angular fragments of the Uppermost Unit (Photo 12 in Figure S5).

The 20 m thick series of unsorted, predominantly Tripolitza limestone debris with blocks up to 1.5 m, slumped silty clays and folded platy limestones with cemented sandstones below and nonmarine silty clays above (location 42; Photos 7–8 in Figure S5) shows that these olistostromes belong to

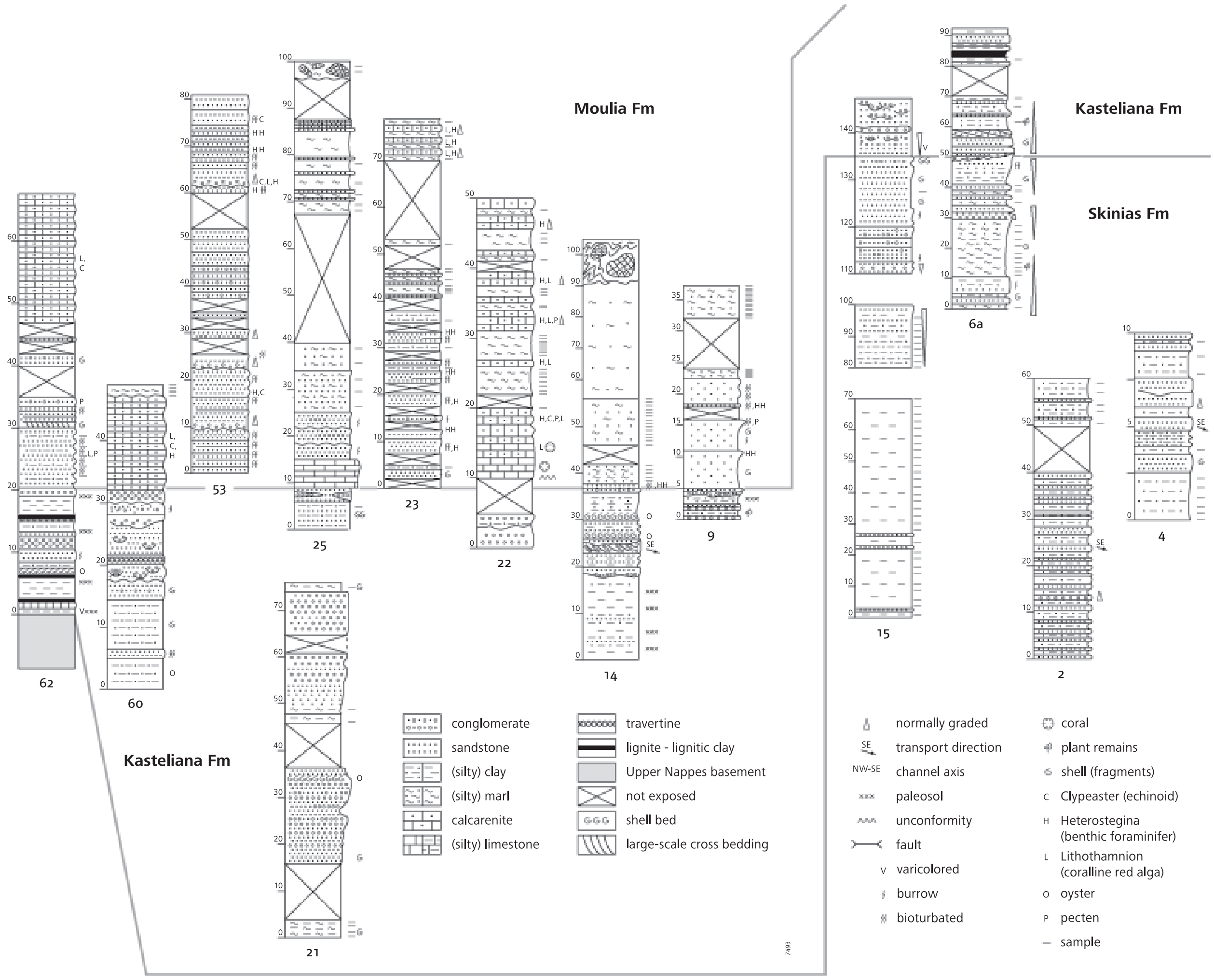


Fig. 3. Schematic lithological columns for selected sections framed in the updated lithostratigraphy (note different scales). Numbers refer to locations given in Fig. 2b and in Figure S1. Details of sections are given in Appendix S1.

the Viannos Fm. Their precise stratigraphic position is difficult to assess because of the lack of vertical trends in the Viannos Fm and the intense faulting of the basin fill. Given the dominance of Uppermost Unit debris in the Viannos Fm, we infer that the Tripolitza limestone dominated olistostromes reflect a later stage of unroofing of the UN in central Crete and are therefore believed to belong to the youngest part of the Viannos Fm. This is in line with the fluvial conglomerate dominated by Tripolitza limestone debris in location 30a which belongs to the uppermost Viannos Fm (see Appendix S1). The exact provenance of the olistoliths is not clear. They are scattered over an area of 14 × 18 km E of the Psiloritis. There is no clear geographic trend in size. Some indication for a nearby western source region comes from the folded limestone bed in location 42 (Photo 8 in Figure S5).

Skinias Formation

Lithology. This formation consists of marine (silty) clays with interbedded sandstones and occasional gravels. Gravity flow deposits, exemplified by slumps and graded pebbly mudstones, are rare except in the SE (locations 2–4 and 5a, see also Fig. 3). Here, (silty) clays are commonly interrupted by sandy turbidites occasionally showing Bouma Ta-c intervals, and by graded pebbly mudstones (Photos 14–15 in Figure S5). The pebbly mudstones to gravel beds are dominated by Tripolitza limestone debris with some flysch and Uppermost Unit fragments. Up to 3 m thick slumped intervals consisting of ripped clay and sandstone beds mixed with gravels or mollusc-rich sands are exposed in location 5a. Palaeocurrent readings indicate transport to the E and SE, i.e. roughly opposite to the dominant W-directed readings in the Viannos Fm. The pebbly sandstone in location 39 (Fig. 3b) representing a mass-flow deposit is dominated by Tripolitza limestone debris.

Thickness and type section. The Skinias Fm is nowhere entirely exposed. A thickness of 150 m has been measured at Vathypetron (location 39 and 40) and of 200–250 m N of Tefeli (location 12). The type

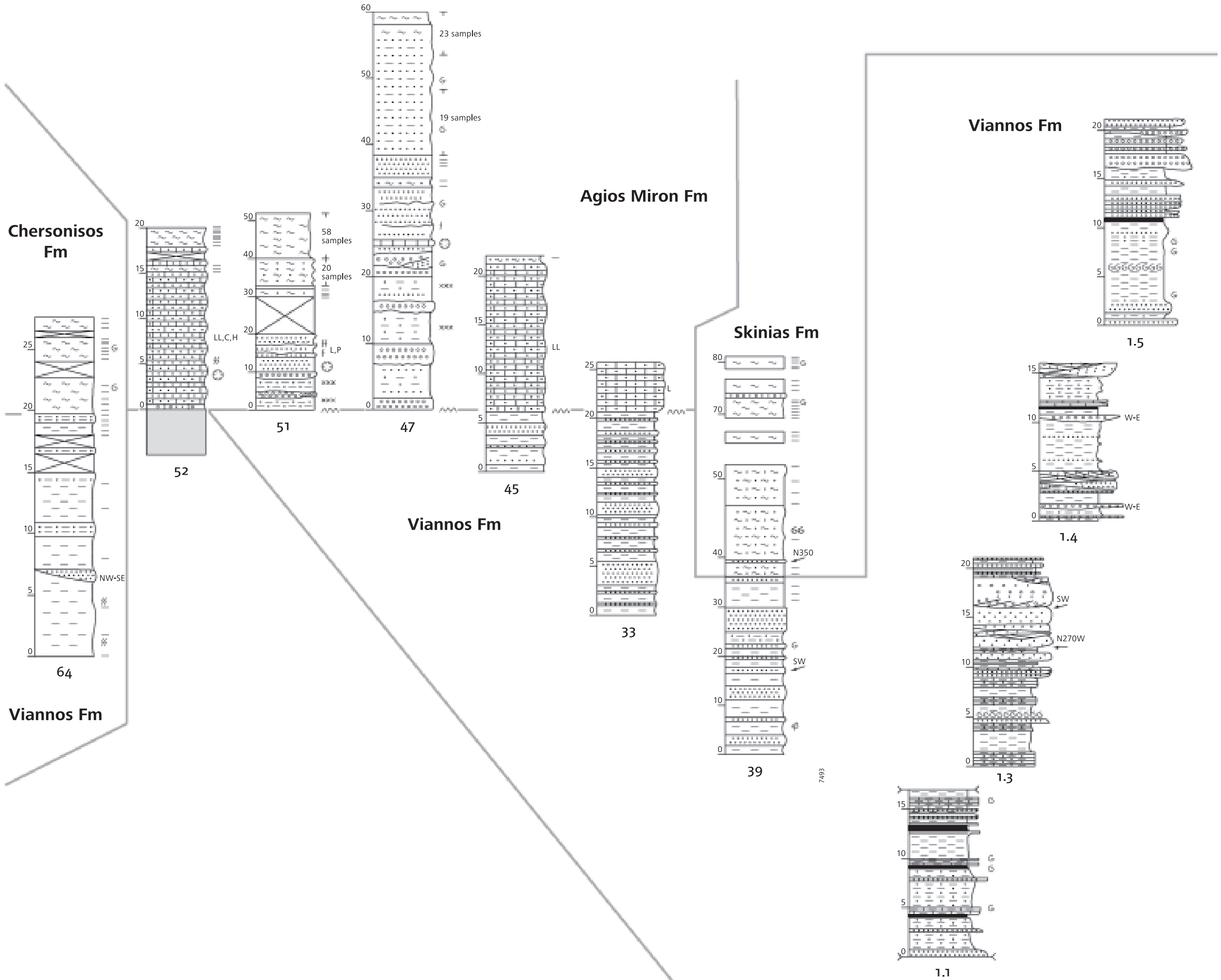


Fig. 3. Continued

section is represented by the 35 m of marine silty clays exposed at location 3.

Underlying unit. The formation conformably overlies the Viannos Fm in locations 13, 34 and 39 (Fig. 3 and Photo 13 in Figure S5).

Depositional environment. The lower part of the formation is characterized by shallow marine silty clays with locally abundant washed *Turritella* (e.g. at Vathypetron, location 39 and NW of Stironas (location 34). Deep marine clays characterize the middle part. The upper part gradually coarsens upward from clay-rich to sand-rich deposits, indicating shoaling, as also indicated by the increasing numbers of malacofossils and the composition of benthic foraminiferal faunas (Fig. 3, section 15).

Geographic extent. The Skinias Fm is widespread in area I and southern part of area II and is absent in areas III, IV and V.

Kasteliana Formation (new name)

Lithology. The Kasteliana Fm comprises a variety of lithologies, varying from silty clays including palaeosols, silts, sands and conglomerates and occasionally travertines, lignites, oyster beds and coral bioherms. The conglomerates are dominantly composed of UN debris but Plattenkalk-pebbles are occasionally present (e.g. in location 16). The coral bioherms are typically 5–10 m high and up to a few 100 m wide consisting of *Tarbellastraea* and *Porites* corals in a (calc) arenitic matrix, surrounded by reef talus.

Thickness and type section. The Kasteliana Fm is approximately 300–350 m thick in area I, as reconstructed from dip data and formation width in the areas of locations 19 and 24. In areas III and IV, however, the formation is significantly thinner [60–80 m at location 56 (area III), 20 m at location 62 (area IV)]. The formation is named after the village of Kasteliana in the SE of the Heraklion Basin, and the Kastellios Hill section of de Bruin *et al*. (1971) is regarded as the type section (Fig. 2b, location 8).

Underlying unit. The formation disconformably overlies the shallow marine top of the Skinias Fm in area I with contacts exposed in locations 6a, 12 and 15 (Photo 16 in Figure S5). In areas III (Photos 17–18 in Figure S5) and IV (location 62), the unit unconformably covers UN basement.

Depositional environment. The Kasteliana Fm consists of an assembly of facies types that accumulated in fluvial-lacustrine, lagoonal and inner neritic environments. The many clast-supported and poorly sorted conglomerates and their associated silty clays and palaeosols, lignites and travertines are interpreted as floodplain deposits dissected by braided channels. Marine transgressions repeatedly drowned the floodplain leading to lagoonal to inner neritic conditions, as indicated by massive oyster beds and silty clays and silts rich in *Limnocardium*, *Terebralia*, *Turritella*, *Murex*, and *Cerithium*. Sands with levels of well-rounded and flattened pebbles arranged parallel to the bedding plane are present throughout the formation and probably represent (reworked) beach deposits (e.g. at location 16). The most open marine conditions are reflected by intraformational bioherms. Corals from one of these bioherms have been used for climate variability studies by Brachert *et al.* (2006) and Mertz-Krauss *et al.* (2008).

Geographic extent. The Kasteliana Fm is widespread in the southern part (areas I–IV) and absent in the northern part (areas II and V).

Moulia Formation (new name)

This formation comprises mixed siliciclastic and calcareous marine sediments and is informally subdivided into a lower and an upper member.

Lithology of the lower member. In the NW of area I (i.e. around Agios Thomas), the lower member is characterized by 10–20 m thick sequences of bioclastic, sandy calcarenites and interbedded calcisiltites (e.g. location 22, Fig. 3). Bedding is poorly developed and expressed by vertically varying concentrations of bioclasts such as *Lithothamnion*, *Heterostegina*, corals, bryozoa, and pelecypods. *Heterostegina* packstones and coral bioherms of *Tarbellastraea* and *Porites* are regularly present, frequently associated with *Clypeaster* and *Pecten latissima*. Elsewhere in areas I, and in area III, the lower member is dominated by siliciclastics. For example, well-sorted, bioturbated sands with irregular hardgrounds and pebble pavements containing scattered pelecypods and *Heterostegina*, are found to the N of Phaestos (location 58) and near Ano Moulia (location 25). These deposits are typically 10–15 m thick. Similar sequences at Vitsilia (location 9) and E of Preveliana (location 23) lack such pebble levels but instead contain levels rich in *Heterostegina* and show indistinct cyclicities of sand and very fine sand-silt couplets (see also Fig. 3). The lower member at these localities measures 20–30 m. An exceptionally thin sequence of 3 m characterizes the lower member S of Tefeli (location 14, Fig. 3). It starts with sands including bedding-parallel pebble strings and passes upward into bioturbated sands rich in *Heterostegina*. Along the new road between Tefeli and Ligortinos at location 13, the lower member is represented by a coral bioherm with mixed clastics of the Kasteliana Fm below and marine silty marls rich in molluscs of the upper member above. Similarly, the bioherm of Kastelli Messara (location 57) may represent the lower member of the Moulia Fm although the contact with overlying marine marls of the upper member is faulted. SW of Moni Odigitria in area IV (Fig. 2b), the lower member comprises 80–90 m of bioturbated fine to pebbly sandstones with hardgrounds and locally abundant *Heterostegina*, passing upward into calcarenites with abundant *Lithothamnion* and fine bioclastic sandstones with molluscs and *Heterostegina* (e.g. at Agiofarangi, location 63). Bioturbated sandstones at location 62 (Fig. 3) contain debris flows of oncoliths and alternate with silty clays. At Sivas (location 61), the base of the lower member includes a *Porites* bioherm.

Lithology of the upper member. The upper member consists of bluish-gray (silty) marls, often rich in molluscs in the basal part. The member contains various types of gravity flow deposits varying in bedding style from thin-bedded siliciclastic turbidites (e.g. Kastelli Messara, location 57) to massive, channeling sandstones with a pebbly base at Mount Kuies (location 24). The overall grainsize varies from graded calcarenites including *Lithothamnion*, bryozoa, *Heterostegina*, and (fragments of) pelecypods at the base to chaotic mixtures of marls and calcarenites along road to Ano Moulia (location 25, Fig. 3) occasionally including outsize clasts (Douli, location 20) or olistoliths (up to 10 m in diameters) of bioclastic calcarenites (location 14, Fig. 3). At Panasos (location 53, Fig. 3) up to 150 m of poorly bedded, sand-rich rudstones and grainstones, and occasionally including pebbly sandstone beds, are exposed. The beds are often strongly bioturbated and characterized by the presence of reworked *Clypeaster* and *Heterostegina* that may be concentrated in distinct horizons. This association reaches its greatest thickness close to the E–W fault bordering the southern Psiloritis Mts and rapidly decreases in thickness and overall grainsize to the S. In contrast, the sequences located in the basin centre contain distinct sapropel-marl couplets with several ash layers in the upper part (Kastelli Messara, location 57 and Agios Ioannis, Fig. 2b), which have been instrumental in the construction of the ATNT2004 timescale (Lourens *et al.*, 2004).

Thickness and type section. Thickness estimates for the Moulia Fm at Ano Moulia (location 25), Preveliana (location 23), S of Tefeli (location 14), and Kastelli Messara (location 57) arrive at 70–80 m. The formation is named after the village of Ano Moulia ~2 km S of Agia Varvara (Figure S1) and the type section is location 25 (Fig. 3).

Underlying unit. The Moulia Fm conformably overlies the Kasteliana Fm, with the exception of the area of Agios Thomas, where an angular unconformity separates these units (Photos 19–21 in Figure S5). Only in area IV and nearby location 53 in area III, do sediments of this formation onlap on UN basement.

Overlying unit. The Moulia Fm is overlain by a cyclically bedded marine series of whitish calcareous marls and sapropels with intervening calcarenitic and siliciclastic turbidites and locally preserved *in situ* evaporites. This series (MUM in Figure S4) belongs to the upper Messinian and is truncated by the terminal Miocene erosional unconformity, covered either by Lago Mare deposits or lower Pliocene mass wasting deposits (Zachariasse *et al*., 2008). Locally, the terminal Miocene erosional event has completely removed the marine upper Messinian. At Kourtes (Fig. 2b), for example, the Moulia Fm is directly overlain by nonmarine Lago Mare deposits (Zachariasse *et al*., 2008).

Depositional environment. Sediments of the lower member are typically deposited in inner neritic environments, in contrast to the deep marine sediments of the upper member. The very thick sequence at Panasos shows evidence for repeated downslope transport and rapid lateral thinning, and is interpreted as a submarine slope apron at the foot of an emergent hinterland.

Geographic extent. The Moulia Fm is restricted to areas I, II and IV.

Agios Miron Formation (new name)

This formation is characterized by a similar deepening marine sequence as the one which characterizes the Moulia Fm but is distinguished from this unit by its underlying units. Also this formation is informally subdivided into a lower and an upper member.

Lithology of the lower member. The lower member consists of bioturbated and fossiliferous sands (*Clypeaster*, *Heterostegina*, pelecypods including *Pecten latissima*). Hardgrounds, pebble levels, and locally *Porites* buildups (e.g. location 47, Fig. 3) are present. At locations 34 and 40a these bioturbated sands pass upward into bioclastic calcarenites. In locations 45 and 52 (Fig. 3) calcarenites prevail showing indistinct cyclicities caused by vertical differences in grain size and in the density of *Lithothamnion*. The member is locally underlain by pockets of up to 25 m thick nonmarine clastics with calcrete palaeosols (locations 47 and 51, Fig. 3) and which we include in the Agios Miron Fm. The nonmarine clastics in the area around the Minoan villa of Vathypetron (Appendix S1: location 40b) are excluded from this denomination because of an intercalated silty clay bed with oysters and unconformable contact with the overlying Agios Miron Fm.

Lithology of the upper member. Silty bluish-gray marls containing molluscs and showing six to seven indistinct sedimentary cycles if complete, are overlain by bluish-gray, homogeneous marls alternating with sapropels and include several ash layers. This sequence is characteristic for the area W of Mount Yugtas (location 51, Fig. 3) but is often incomplete [location 62 at ancient Rhizenia (Fig. 3) and along the new road Heraklion-Agia Varvara (location 50)]. The basal silty marls with six to seven indistinct sedimentary cycles are supposedly time-equivalent to the six to seven indistinct sedimentary cycles present in the calcarenites of ancient Rhizenia (location 52). Until now, sediments of the upper member have been found at only one place in the area E of Mount Yugtas (location 40d, Appendix S1). In contrast, the N-dipping lower member is widely exposed E of Mount Yugtas where it forms a prominent ledge traceable as far E as the village of Avli (location 26).

Thickness and type section. The formation is named after the village of Agios Miron in the NW of the Heraklion Basin (Figure S1). The lower member measures up to *ca*. 50 m (location 40a). The upper member is most complete at location 51, measuring 65 m. This location is chosen as type section of the Agios Miron Fm (Fig. 3).

Underlying unit. The Agios Miron Fm overlies the Viannos or Skinias Fm with an angular unconformity (Photos 22–23 in Figure S5) or onlaps onto UN basement (location 52; Fig. 3 and location 49) or onto Tripolitza limestone olistoliths. One of these olistoliths, N of Profitis Ilias (Fig. 2b), shows an abrasion surface with borings overlain by shallow marine sandstones.

Overlying unit. The Agios Miron Fm passes upward into the upper Messinian described under the Moulia Formation which is truncated at the top by the terminal Miocene erosional unconformity (Zachariasse *et al*., 2008).

Depositional environment. Sediments of the lower member were deposited in an inner neritic environment and those of the upper member in deep water (outer neritic to bathyal).

Geographic extent. The formation is restricted to area II.

Chersonisos Formation (new name)

Lithology. The formation consists of bluish-grey marls with frequent pycnodont oysters and several sapropels.

Geographic extent, thickness and type section. This formation is restricted to Cape Chersonisos and measures minimally 15 m. Type section is location 64 (Fig. 3).

Underlying unit. Viannos Fm. As seen from a distance, the sediments of the Viannos Fm have been slightly tilted before deposition of this unit (Photo 24 in Figure S5). The deep marine marls therefore do not belong to the Skinias Fm, which conformably overlies the Viannos Fm. The presence of many pycnodont oysters and absence of an underlying shallow marine facies, these deep marine marls are not easily assignable to the Agios Miron Fm either. Therefore and despite their limited geographic distribution, these deep marine marls are placed into a separate formation.

Depositional environment. Deep marine (middle bathyal).

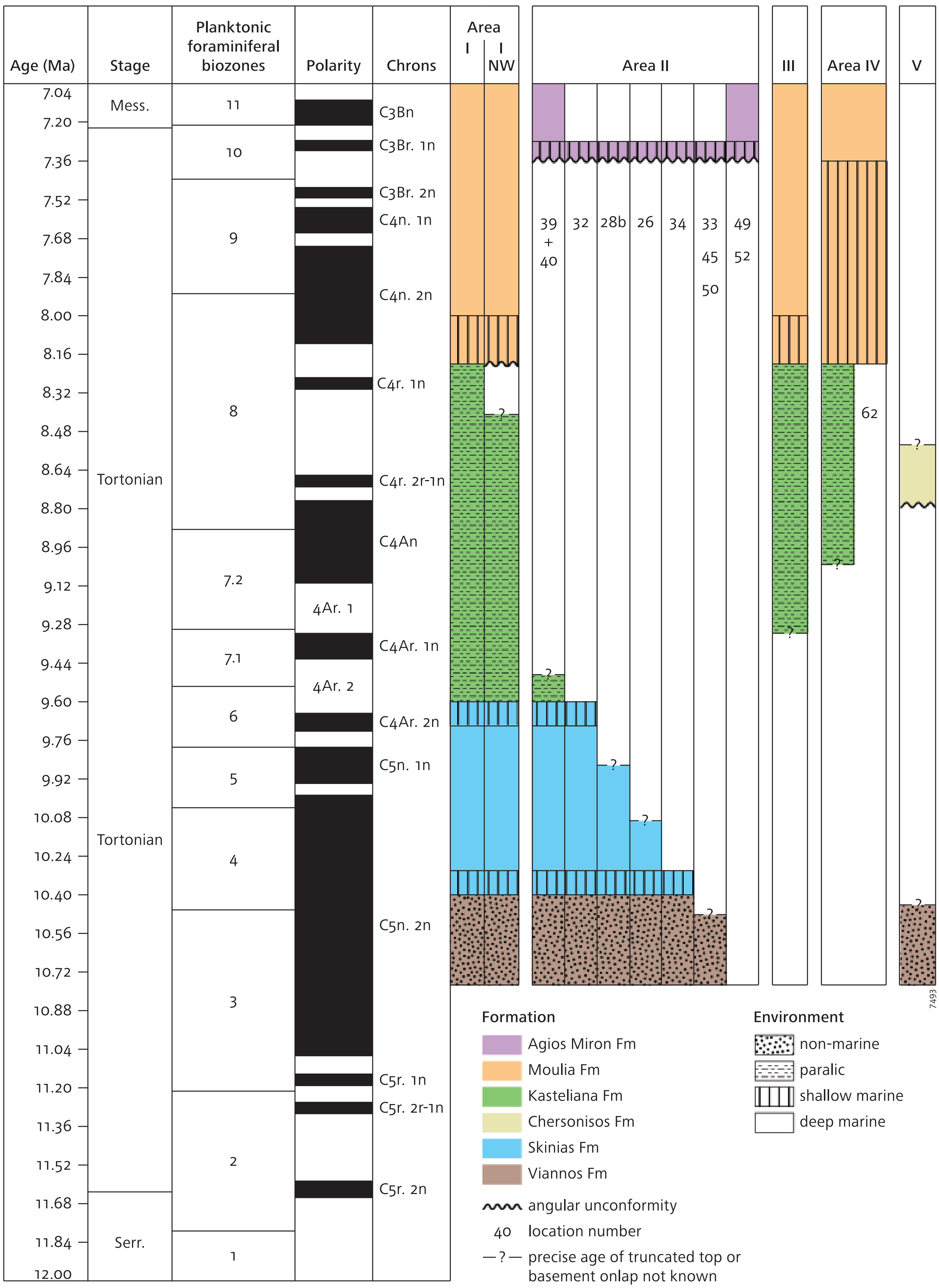
Ages of the formations

The oldest unit that can be directly tied to the updated planktonic foraminiferal biochronology (Figures S2–3; Appendix S2; Table S1) is the Skinias Fm. The shallow marine lower part of this unit (exposed in locations 12, 19 and 39) lacks age diagnostic planktonic foraminifers but the overlying deep marine clays correlate to biozone 4 (Appendix S1). At location 12, deep marine clays of biozone 4 are overlain by marine clays of biozones 5 and 6 (Appendix S1). A biozone 4 and/or 5 age for the Skinias Fm is also found in locations 3 (with a fault separating biozones 4 and 5), 5a, 26, 28, and 32. The upper part of the Skinias Fm is characterized by a shallowing upward sequence, which towards the top suffers from reworked foraminifers and the lack of age diagnostic species. A biozonal 6 assignment for the youngest datable sediments of the Skinias Fm at locations 12, 15 and 19 suggests that the base of the Kasteliana Fm falls in upper biozone 6 (Appendix S1). The normal polarities measured for 12 palaeomagnetic samples in the upper part of location 6a (Fig. 3) may therefore be assignable to Chron C4Ar.2n (see Fig. 4), providing an age of ~9.6 Ma for the top of the Skinias Fm (i.e. slightly younger than age of 9.65 Ma for top C4Ar.2n (Hüsing *et al.*, 2007). The Skinias Fm thus extends from biozone 4 up into upper biozone 6, covering the time span from ~10.4 to ~9.6 Ma, assigning 0.1 Myr for the deposition of both the lower and upper shallow marine Skinias Fm (Fig. 4).

The Viannos Fm, which conformably underlies the Skinias Fm at locations 12 and 39, is thus older than 10.4 Ma. The age of the Viannos Fm can be further specified by biostratigraphic analysis of a ~100 m thick deep marine unit overlying UN basement on the island of Gavdos (Fig. 2a). This unit of bluish-grey silty clays and sandstones with coral limestones at the base is separated by a calcrete palaeosol from the overlying classical section Metochia (Postma *et al.*, 1993a). A 9.8 Ma age for the base of the Metochia section (Hilgen *et al.*, 1995) implies that the marine unit on Gavdos below the calcrete palaeosol is time-equivalent with the fluvio-lacustrine sediments of the Viannos Fm and/or the marine Skinias Fm in central Crete. Neogloboquadrinids in this unit are dominated by small-sized *Neogloboquadrina atlantica* which together with regular occurrences of *Globorotalia partimlabiata* and *Globoquadrina dehiscens* indicate that this older marine unit on Gavdos belongs to biozone 3. The absence of large-sized *Neogloboquadrina atlantica* types suggests an assignment to the upper part of biozone 3 (Figure S2). This is confirmed by the absence of *Helicosphaera walbersdorfensis* and the presence of *Helicosphaera stalis* in a few samples from the middle part of this unit. This calcareous nannoflora is characteristic of lower MNN8 and together with dominant *Neogloboquadrina atlantica* equates the oldest marine unit on Gavdos to the G-42 to G-28 interval in the Gibliscemi section (Hilgen *et al.*, 2000), i.e. upper biozone 3 (Figure S2). Since this unit is truncated at the top by a calcrete palaeosol, deposition may have extended up into lower biozone 4. Hence an age range from 10.78 Ma (i.e. the age of cycle G-42, Hilgen *et al.*, 2000, 2003) to ~10.4 Ma (i.e. an age slightly younger than the 10.467 Ma top of biozone 3 (Table S1) seems an acceptable estimate for the oldest marine unit on Gavdos. Since the Skinias Fm on Crete is estimated to range between 10.4 and 9.6 Ma it is interpreted that the oldest marine unit on Gavdos and the fluvio-lacustrine Viannos Fm of central Crete are time-equivalent units ranging in age from ~10.8 to ~10.4 Ma.

Except for the Agios Thomas region, where the contact between the Kasteliana and Moulia Fms is an angular unconformity, transitions between the Kasteliana Fm with underlying and/or overlying units are everywhere conformable. The age range of the Kasteliana Fm is thus defined by the ages of the youngest marine sediments below, and oldest marine sediments above the Kasteliana Fm. The age of ~9.6 Ma for the top of the Skinias Fm and the base of the Kasteliana Fm is in agreement with an MN 10 assignment of murid rodents from the lower part of the Kastellios Hill section (e.g. de Bruin *et al.*, 1992). Magnetostratigraphy of time-equivalent nonmarine sections in Spain with the GPTS reveals that MN 10 encompasses Chrons C4Ar and C4An (Krijgsman *et al.*, 1996), i.e. from ~9.65 to ~9.11 Ma (Hüsing *et al.*, 2007). Sixteen samples from the lower part of the Kastellios Hill section yielded all reversed polarities (Duermeijer *et al.*, 1998) indicating that the short intervals of normal polarity measured by Sen *et al.* (1986) are either present-day overprints or represent cryptochrons. Hence, the lower part of the Kastellios Hill section correlates to Chron C4Ar.1r or C4Ar.2r. An updated classification of rodents from Plakias on western Crete (de Bruin & Meulenkamp, 1972) and from Kastellios Hill – with *Cricetulodon* in Plakias and two species of *Progonomys* in Kastellios Hill – suggests an older age for the Plakias rodents (Hans de Bruin, 2009, personal communication). The nonmarine deposits of Plakias (Fig. 2a) have been included in the Pandanassa Fm (Meulenkamp, 1969). This unit correlates with the Kasteliana Fm of central Crete and since the base of this unit is dated at 9.6 Ma (this study), a correlation of the lower Kastellios Hill section to C4Ar.1r [Correction added after online publication 18 August 2011 - 'C4Ar.2r' changed to 'C4Ar.1r'] is most likely. This correlation allows the Plakias rodents to be older (9.6–9.3 Ma) than those of Kastellios Hill (9.3–9.1 Ma) (see also discussion under location 8 in Appendix S1).

The regular presence of *Sphaeroidinellopsis seminulina* in the older part of the deep marine member of the Moulia Fm in locations 9, 22, 24, 55, 57, and 58 (see Appendix S1) indicates that the base of the deep marine member of the Moulia Fm is older than 7.91 Ma (Table S1). An age of ~8.0 Ma is calculated by averaging sedimentation rates



for intervals between three bioevents in location 22 and extrapolation back to the base of the section (Appendix S1), and 8.05 Ma in location 58 (Appendix S1). Hence, the Kasteliana Fm is estimated to range from ~9.6 to ~8.0 Ma. Note that 8.0 Ma is a minimum age since the time span of the shallow marine member of the Moulia Fm cannot be directly established. Recently, Köhler *et al.* (2008, 2010) noticed in the Metochia section on Gavdos a sharp decrease in sedimentation rate, coinciding with a sudden decrease in ophiolite-related sediments at ~8.2 Ma. They related these changes to a reduction in the sediment supply from Crete. The change from paralic (Kasteliana Fm) to marine sedimentation (Moulia Fm) in central Crete should have contributed to this sediment starvation event on Gavdos thereby refining the age estimate of ~8 Ma for this facies change to 8.2 Ma. The younger part of the deep marine member of the Moulia Fm is rarely complete and at Kourtes (Fig. 2b) even the entire deep marine member is missing (Zachariasse *et al.*, 2008). The deep member is most complete in sections Kastelli Messara (location 57) and Agios Ioannis (Fig. 2b) (Hilgen *et al.*, 1995, 1997) and extends up into biozone 11 (Messinian).

The oldest datable sediments of the Agios Miron Fm in location 47 contain *Globorotalia menardii 5*, correlating to upper biozone 10 (Appendix S1). The age for the base of this unit is set at 7.36 Ma, i.e. the astronomical age of the First Common Occurrence of *Globorotalia menardii 5* (Lourens *et al.*, 2004). The Agios Miron Fm extends into the Messinian (biozone 11 and younger).

The basal part of the Chersonisos Fm correlates to the bottom part of biozone 8 with an age of ~8.8 Ma (for argumentation, see Appendix S1). Figure 4 shows the chronology of the lithostratigraphic succession for areas I to V (including differences within areas).

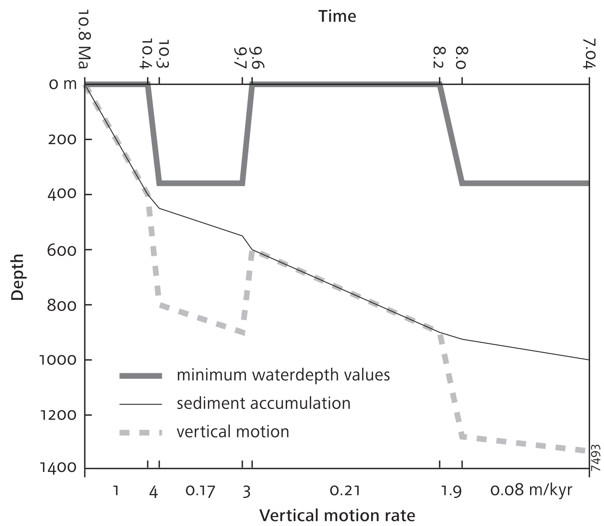


Fig. 5. Vertical motion history of area I for specific time intervals between 10.8 and 7.0 Ma. Vertical motion rates are obtained by adding the amount of accumulated sediment to the average depositional depth estimates between age control points. Depth estimates are minimum values based on the presence/absence of specific benthic foraminifers (see in text under Palaeobathymetry). Age control points for (sub) lithostratigraphic units are from Fig. 4.

TORTONIAN BASIN RECONSTRUCTIONS IN CENTRAL CRETE (10.8–7.36 Ma)

The Viannos and Skinias Basins (10.8–9.6 Ma)

Fluvial conglomerates dominated by whitish Pindos limestone clasts and characteristic for the Males Fm in eastern Crete occur 1 km east of location 1, so forming the westernmost occurrence of such conglomerates and indicating a lateral transition of the Viannos and Males Fms in the SE part of the Heraklion Basin (Fortuin, 1978). From W-directed palaeocurrent readings in both units it follows that the fluvial sediments of the Viannos Fm formed the distal part of a large westward-draining river system originally described by Fortuin (1977) from eastern Crete. Erosional remnants of this so-called Males-Viannos fluvial system have even been found in the SW of Crete (van Hinsbergen *et al.*, 2008). In central Crete, this fluvio-lacustrine system covered an area of at least 25 km wide, from Chrondros (Fig. 2b) in the SE, to Tilissos (Fig. 2b) in the NW, and to Chersonisos in the NE (Fig. 2b). No Viannos sediments are outcropping in area III, where the only sediment-basement contact shows Kasteliana Fm overlying flysch (location 56; Photos 17–18 in Figure S5). The Viannos Fm was thus either not deposited, or, eroded in area III. Because syn-Viannos elevated topography for area III conflicts with the fine-grained fluvial sediments to the E and W of this area, area III is regarded to have formed part of the Viannos Basin. There is also little doubt that the present Psiloritis Mts were part of the Viannos Basin as

Fig. 4. Lithostratigraphic succession plotted per area against a high-resolution planktonic foraminiferal biochronology for the late middle to late Miocene in the eastern Mediterranean as based on the semi-quantitative distribution and coiling of selected planktonic foraminiferal taxa in section Gibliscemi on Sicily (Figure S2) and section Metochia on Gavdos (Figure S3). Biozones are defined and described in Appendix S2. Ages for biozonal boundaries and their defining bioevents are derived from updated astronomical ages for sedimentary cycles in Metochia and Gibliscemi (Lourens *et al.*, 2004) and correlative cycles in Monte dei Corvi (Hüsing *et al.*, 2007). Base Tortonian falls within the lower part of biozone 2 with an age of 11.625 Ma (Hüsing *et al.*, 2007), whereas base Messinian at 7.246 Ma (Lourens *et al.*, 2004) is approximated by the FCO of the *Globorotalia miotumida* group (base biozone 11). The magnetochronology is based on the updated ages for polarity reversals in Metochia (Lourens *et al.*, 2004) and Monte dei Corvi (Hüsing *et al.*, 2007). Position of defining bioevents and polarity reversals in terms of sample number and sedimentary cycle number in sections Gibliscemi and Metochia and their ages are summarized in Table S1.

indicated by occurrences of fine-grained fluvial sediments at location 65.

The Viannos Fm is mostly in faulted contact with UN basement, e.g. in locations 34 and 50 with Tripolitza flysch, and 1.5 km SSE of location 44, and 2.5 km ENE from location 5a with Pindos flysch. Stratigraphic contacts are observable in only two locations (46a and 69) showing an Uppermost Unit basin floor overlain by proximal coarse clastics representing sheetflood deposits on an alluvial fan. These breccio-conglomerates are probably representative of the early infilling stage of the Viannos Basin since they are overlain by finer grained sediments (see Appendix S1). The incipient basin fill in eastern Crete is characterized by similar breccio-conglomerates of Uppermost Unit debris (Mithi Fm, Fortuin, 1977). The westward drainage of the Males-Viannos river system suggests an E-W trending northern and southern margin for the Viannos Basin but their precise location is unknown. The fine-grained sediments in the northernmost locations 46a and 64 suggest a position of the northern margin in the present-day Cretan Sea. A likely candidate for the southern margin is the E–W trending northern margin of the Asteroussia Mts: sediments of the Viannos Fm are absent in the Asteroussia Mts.

The mass-displacement of exotic blocks and breccio-conglomerates during deposition of the youngest part of the Viannos Fm (~10.4 Ma) indicates significant and nearby uplift. The distribution of the exotic blocks across an area E of the Psiloritis and the hint of E-directed transport seems to point to a beginning uplift of the Psiloritis Mts. Olistoliths slid not only to the E but probably also to the N as indicated by the Tripolitza limestone olistolith in breccio-conglomerates overlying fine-grained sediments of the Viannos Fm in location 65 (see Appendix S1). The Tripolitza limestone dominated debris in these olistostromes contrasts with the dominance of Uppermost Unit debris in the remainder of the Viannos Fm and indicates that uplift unroofed a deeper part of the UN in central Crete. The beginning uplift of the Psiloritis Mts and associated gravitional sediment transport precedes the submergence of the fluvial Viannos Basin between 10.4 and 10.3 Ma.

The marine sediments of the Skinias Fm covered a grossly similar area as the fluvial Viannos Fm in central Crete. Contrary to the Viannos Fm, no sediments of the Skinias Fm are present in the Psiloritis Mts. Submergence of the Viannos Basin is accompanied by an increase in subsidence, from ~1 m kyr^{-1} between 10.8 and 10.4 Ma to ~4 m kyr^{-1} between 10.4 and 10.3 Ma (Fig. 5). Subsidence slowed down to ~0.2 m kyr^{-1} during deposition of the main part of the Skinias Fm (10.3–9.7 Ma). Palaeobathymetric data (Appendix S1) indicate that most middle Skinias sediments were deposited in 350–500 m water depth. Definitively deeper conditions, exceeding 600 m, are recorded for the turbidite-rich sediments in the SE of area I. Time-equivalent marine sediments in the Ierapetra area are of comparable depth (Postma *et al.*, 1993b; van Hinsbergen & Meulenkamp, 2006) and rich in sandy-gravelly turbidites (Fortuin, 1977, 1978). Palaeocurrent readings in the SE of area I (locations 2–4) indicate a (north)western source for these turbidites and since this applies also for the Ierapetra equivalents one may argue that both areas were part of one single middle bathyal basin (Ierapetra Basin), shallowing to the (north)west into the upper bathyal Skinias Basin.

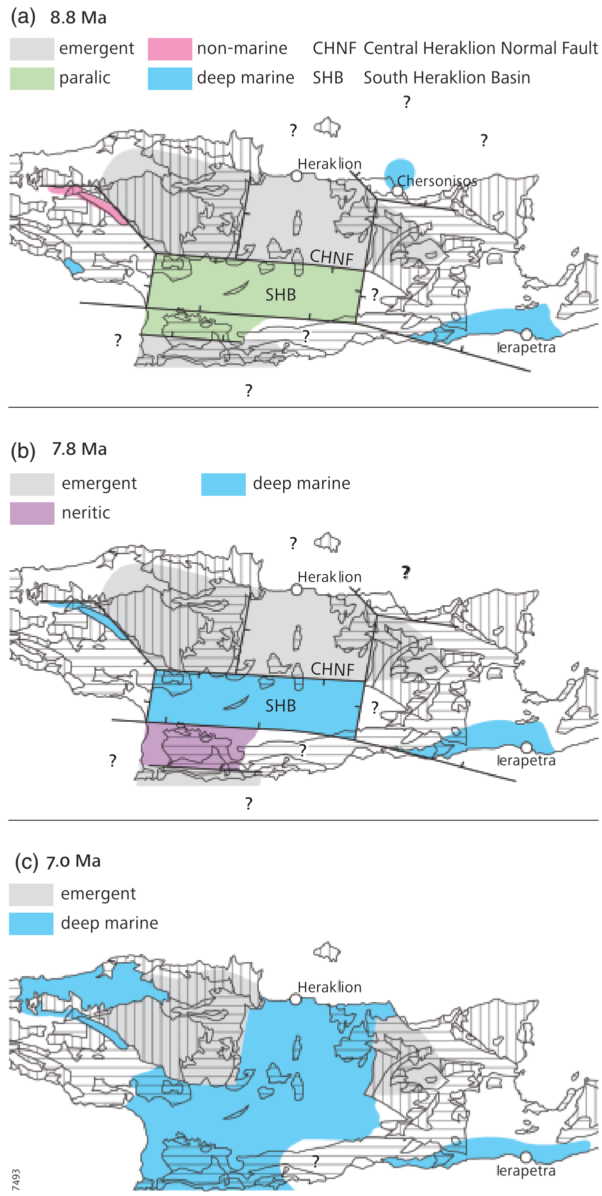


Fig. 6. Schematic palaeogeographic maps for central Crete during three late Miocene time slices postdating the onset of arc-parallel stretching. The distribution of time-equivalent deposits to the west and to the east is based on yet to publish biostratigraphic data and follows present-day contours.

Uplift and fragmentation of the Skinias Basin (9.7–9.6 Ma)

The Skinias Basin was rapidly uplifted between 9.7 and 9.6 Ma (Fig. 5). Only area I subsided shortly after 9.6 Ma. Areas II and V remained emergent and were eroded until ~8.8 Ma for area V, and until 7.36 Ma for area II (Fig. 4). In several places of area II, sediments of the Skinias Fm

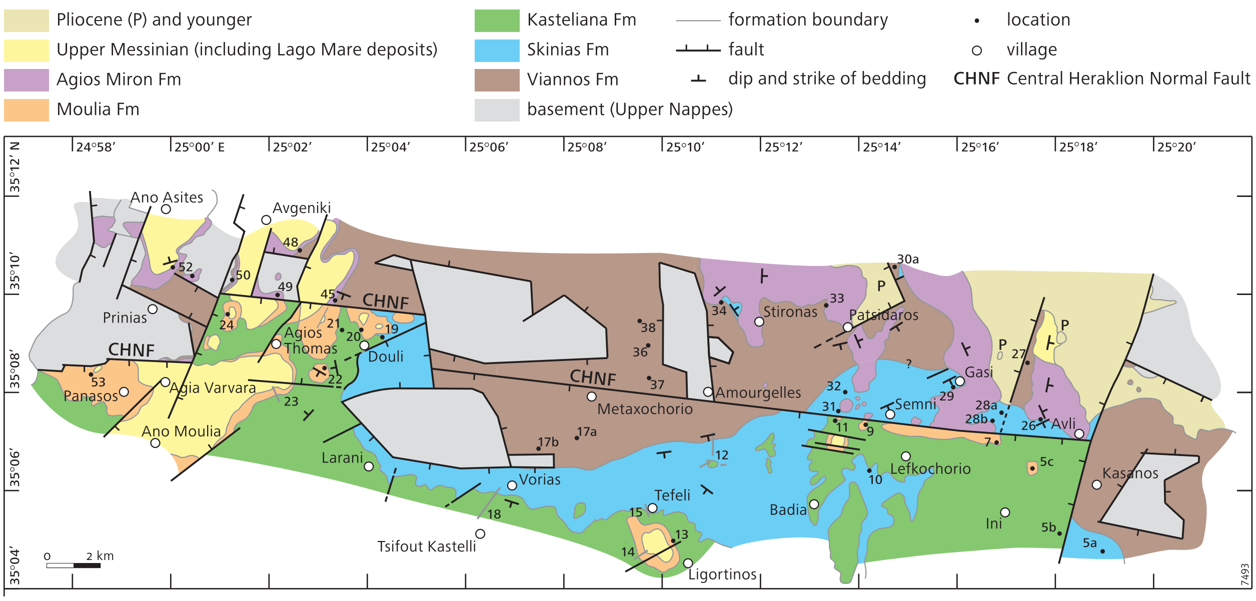


Fig. 7. Geologic map for the central part of the Heraklion Basin showing the outcrop configuration of the lithostratigraphic units defined in this paper. The map is modified after the official geological map of Greece (IGME, 1989, 1994) and the geological map of Delrieu (1990) for the Malevisi area. The map shows the roughly E–W striking Central Heraklion Normal Fault (CHNF) that came into being shortly after the uplift of central Crete between 9.7 and 9.6 Ma and which remained active till 7.36 Ma (see Fig. 6). This fault is responsible for the divergent stratigraphic successions in areas I and II and is offset by NE–SW striking faults.

were preserved, whereas in other places the Skinias Fm and even the Viannos Fm were removed before deposition of the Agios Miron Fm (Fig. 4). This suggests that area II broke up into several smaller and differently elevated fault blocks shortly after 9.6 Ma. One of these even preserves 20–30 m nonmarine to brackish water sediments, which are unconformably overlain by the Agios Miron Fm (locations 39, 40a–c), suggesting deposition in a small faulted basin in an area that was subject to differential uplift. We discussed earlier that area III formed an integral part of the Males-Viannos drainage system. The stratigraphic contact between basement and Kasteliana Fm at location 47 implies that area III lost its Viannos sediments (and most likely also its Skinias sediments) by erosion following uplift between 9.7 and 9.6 Ma (Fig. 4).

The paralic South Heraklion Basin (9.6–8.2 Ma)

Subsidence of area I resumed shortly after 9.6 Ma with rates of $\sim$0.2 m kyr^{-1}, creating accommodation space for the 300–350 m thick paralic Kasteliana Fm (Fig. 5). The timing of subsidence of area III is more difficult to asses because the Kasteliana Fm here lacks micromammal or palaeomagnetic age data. An age of $\sim$9 Ma was suggested for the bioherm at Psalidha (location 54), correlating an ^{87}Sr/^{86}Sr isotope ratio in pristine corals to the time scale of ocean-based variations in strontium isotopes (Mertz-Krauss *et al.*, 2007). The precision of this age is rather low, however: foraminiferal Sr ratios for the astronomically dated samples from the Metochia section on Gavdos (Flecker *et al.* 2002) would assign an age range of 9.6–7.7 Ma for the Psalidha value if we add the Gavdos minimum internal standard error of 24 ppm. The true age of the bioherm at Psalidha, which is in the topmost part of the Kasteliana Fm is probably closer to the 8.2 Ma age for the base of the Moulia Fm (Appendix S1: location 54). Precise age constraints for the base of the Kasteliana Fm are also lacking in area IV. The northern part of area IV accumulated only 20 m of paralic sediments overlying UN basement. The strong differences in thickness of the Kasteliana Fm in areas I, III and IV may result from differences in accomodation space even if subsidence in all these three areas resumed shortly after 9.6 Ma. The uncertainties in age for the base of the Kasteliana Fm in areas III and IV do not change the fact that some time between 9.6 and 8.2 Ma a comprehensive paralic basin [termed South Heraklion Basin (SHB)] extended over much of the southern part of central Crete (Fig. 6a). The geological map of the central part of the study area (Fig. 7) shows an E–W striking fault that is clearly the northern bounding fault of this basin (Fig. 6a). This so-called Central Heraklion Normal Fault (CHNF) is offset by younger NNE–SSW striking faults and continues westward along the southern escarpment of the Psiloritis Mts (Fig. 7). This western segment of the CHNF should have separated the emergent Psiloritis block in the N from the western part of the paralic SHB (i.e. area III) in the S (Fig. 6a). Differences in thickness of the paralic sediments between areas III and IV suggest a southern bounding normal fault along the northern margin of the Asteroussia Mts (Fig. 6a). A subsidiary fault to the S may have controlled the southern limit of onlapping paralic sediments over UN basement (Fig. 6a). The southward continuation of the prominent NNE–SSW fault that presently bounds the Dikti Mts in the W probably bounded the paralic SHB in the E (Fig. 6a). The marine Ierapetra Basin to the E continued to be deep with W-derived sediment transport. Time-equivalent nonmarine sediments are found in the central part of the Rethymnon Province, whereas deep marine conditions prevailed at ~8.8 Ma to the W of Agia Galini (Fig. 6a). The original geographic extent of these erosional remnants is unknown but it suggests the existence of a deep marine basin W of the paralic SHB at about 8.8 Ma. The western bounding fault of the paralic SHB might have been the NNE–SSW striking fault that earlier (during uplift of the Skinias Basin: 9.7–9.6 Ma) confined area III in the W (Fig. 6a). Since the paralic SHB was bound in the N and in the S by elevated areas, marine incursions most likely came from the deep marine basins in the E and – after 8.8 Ma – in the W (Fig. 6a).

Subsidence of the Chersonisos Basin (~8.8 Ma)

The deep marine marls that overlie the Viannos Fm in area V (location 64) can be assigned to the lowermost part of biozone 8 (8.84–8.78 Ma, Appendix S1). The absence of a transgressive sequence between the lacustrine Viannos Fm and overlying 600–750 m deep marine marls suggests the presence of a slump scar, because even in the fastest subsiding basins, a sedimentary record of transgression is normally preserved (e.g. on Gavdos (van Hinsbergen & Meulenkamp, 2006) and Milos (van Hinsbergen *et al.*, 2004)). The original dimension and geometry of the Chersonisos Basin remains unknown and the basin bounding faults in Fig. 6a are speculative.

Tilting (preceding the drowning of the paralic South Heraklion Basin at ~8.2 Ma)

In area II, the Agios Miron Fm unconformably overlies UN basement or tilted sediments of Viannos and Skinias Fm (Photos 22–23 in Figure S5). Tilting in area II thus took place after 9.6 Ma since nonmarine sediments conformably overlying the Viannos and Skinias Fms at Vathypetron (locations 40a–c) are tilted as well. Strongly varying bedding orientations throughout area II suggest the presence of several fault blocks that were tilted in different directions some time between 9.6 and 7.36 Ma (i.e. the base of the Agios Miron Fm). Tilting is not limited to area II. It occurred also in the NW of area I as shown by angular unconformities between the Kasteliana and Moulia Fms around Agios Thomas (Photos 19–21 in Figure S5; see also Appendix S1, locations 19, 21, 22 and 24). Tilting of fault blocks around Agios Thomas occurred before 8.2 Ma (i.e. the base of Moulia Fm) and after deposition of the Kasteliana Fm. The absence of slumping and sliding phenomena in the tilted Kasteliana sediments suggests nondeposition and compaction before tilting. The overlapping time windows for tilting in area II (9.6–7.36 Ma) and in the northwestern part of area I (9.6–8.2 Ma) may indicate that tilting in both areas occurred synchronously shortly before 8.2 Ma.

The marine South Heraklion Basin (8.2–7.36 Ma)

Accelerated subsidence of the SHB to upper bathyal depths between 8.2 and 8.0 Ma, with rates of ~2 m kyr^{-1}, subsequently decreased to ~0.1 m kyr^{-1} during the remaining of pre-Messinian time (Fig. 5). Figure 4 shows that the CHNF remained delimiting the emergent area II from the marine SHB (area I) until 7.36 Ma (see Photo 25 in Figure S5). Between 9.6 and 8.2 Ma, rivers draining the emergent area II extended to the S across the eastern part of the paralic SHB. After 8.2 Ma, when the emergent area II faced the deep marine SHB in the S (Fig. 6b), sediment input was redistributed by submarine gravity flows, as shown e.g. by the massive sandstones with scoured pebbly base intercalated in the deep marine member of the Moulia Fm at location 24. The more distal parts of these mass flows are found farther S where deep marine marls contain thin-bedded, turbiditic sandstones (e.g., at Ano Moulia, location 25). Carbonate shoals fringing the S margin of the emergent area II provided debris found in the calcirudites intercalated in the deep marine Moulia Fm at Douli (location 20) and the calcarenites farther S, at Mikro Oros (location 22) and Preveliana (location 23). The eastern segment of the Agia Varvara Fault Zone (AVFZ) of Delrieu *et al.* (1991, 2004) runs parallel to but S of the CHNF in Fig. 7. In the reconstruction of Delrieu *et al.* (1991, 2004), the deep marine basin S of the AVFZ adjoins an emergent area to the N with onlapping reefal limestones during periods of sea level high stands. The time-equivalency of the deep marine marls S and the reefal limestones N of the AVFZ, however, is incorrect because we found erosional remnants of these deep marine marls overlying the reefal limestones (and calcarenites) at two places N of the AVFZ (see Appendix S1: locations 20 and 22). The eastern segment of the AVFZ is thus a post-sedimentary normal fault, with a dip-slip displacement of ~100 m. The CHNF and AVFZ both bound the Psiloritis Mts in the S. Syn-depositional activity of the CHNF in this area is clearly reflected in the slope apron of mixed siliciclastics and bioclastics that developed shortly after 8.2 Ma S of the Psiloritis Mts at Panasos (location 53). The river that fed this slope apron system must have drained the emergent Psiloritis block. Accelerated subsidence at ~8.2 Ma also affected area IV, but this area remained shallow marine much longer than areas I and III (Fig. 4). The implied differences in subsidence rates between both areas suggest that the basin bounding fault along the northern margin of the Asteroussia Mts remained active (Fig. 6b). The shallow marine deposits in area IV onlap farther southward over UN basement than the underlying paralic sediments of the Kasteliana Fm and covered at least the entire western part of the Asteroussia Mts (Fig. 6b). Time-equivalent deep marine sediments of the Moulia Fm are found in the Ierapetra Basin to the E, and in the central part of the Rethymnon province to the W. The original geographic extent of these marine basins as well as their connection to the marine SHB remains elusive. The overall palaeogeography in Fig. 6b suggests that the $\sim$N–S striking faults bounding the paralic SHB (Fig. 6a) remained active between 8.2 and 7.36 Ma.

The inception of the modern Heraklion Basin (7.36 Ma)

Area II submerged at 7.36 Ma, corresponding to the age of the onlapping shallow marine sediments of the Agios Miron Fm over tilted sediments of the Viannos and Skinias Fm and UN basement (Fig. 4). Further subsidence to upper bathyal depths is evidenced by the deep marine bluish marls overlying these shallow marine sediments. These deep marine sediments are found mainly, but not exclusively, W of Mount Yugtas. The submerged area II was bounded in the W by the Psiloritis Mts and in the E by the Dikti Mts (Fig. 6c). The 350–500 m deep marine marls of upper biozone 10 draped over an irregular and karstified top of shallow marine calcarenites of the Moulia Fm in location 60 (Fig. 3) records rapid deepening of a shoaly area that experienced submarine or even subaerial erosion. The 350–500 m deep marine marls immediately overlying the calcarenites suggest removal of the transgressive sedimentary sequence, possibly by submarine sliding. Nevertheless, we suggest that (part of) area IV subsided synchronously with area II resulting in a marine basin covering the entire modern Heraklion Basin (Fig. 6c). The northern part of the Rethymnon area drowned at about the same time whereas deep marine conditions continued to exist in the Ierapetra region (Fig. 6c). The regional drowning event at 7.36 Ma seems to reflect wholesale subsidence of Crete, rather than subsidence along the basin bounding faults that were active before 7.36 Ma. This is in line with middle bathyal depth estimates in areas I and III whereas upper bathyal conditions prevailed in areas I and IV. Except for a short period at the end of the Miocene when the Mediterranean Basin became isolated from the open ocean, shape and dimension of this marine basin were maintained up into the early Pliocene after which it became uplifted and fragmented (Zachariasse *et al.*, 2008).

DISCUSSION

We will first evaluate whether the Cretan detachment played a role, and if so, during which phase in the basin evolution of central Crete. Then, we will discuss the implications of the reconstructed timing of basin subsidence for the exhumation history of the Cretan HP-LT metamorphic sequence, and the dramatic thinning of the UN.

Arguments for the existence of a supradetachment basin on Crete

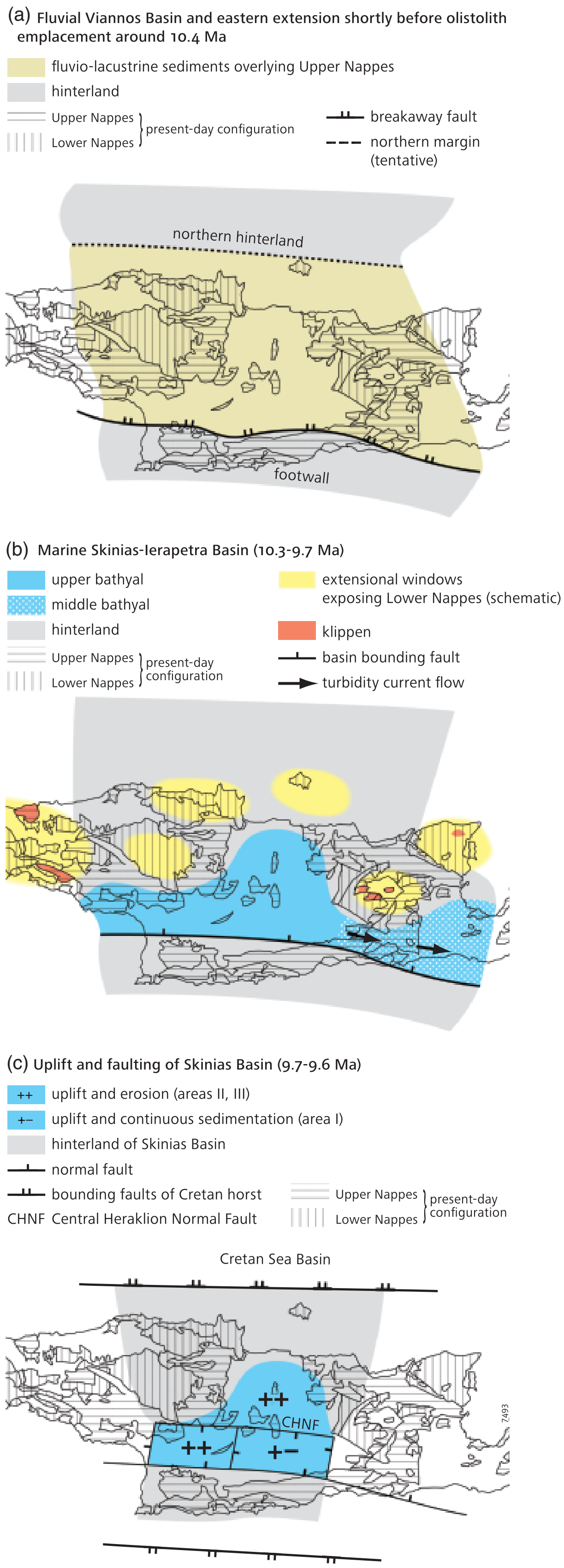
Most supradetachment basins in the Aegean region, or elsewhere, are formed in the hangingwall of an extensional detachment, and their basin fills cover the time-span of exhumation of its metamorphic footwall, and are in faulted contact with that footwall (Friedmann & Burbank, 1995; Sánchez-Gómez *et al.*, 2002; Tirel *et al.*, 2008, 2009; Sen & Seyitoğlu, 2009). On Crete, however, no such direct relationship can be observed: the bulk of exhumation of the LN occurred between 24 and 15–12 Ma by which time the LN was exhumed to apatite fission track cooling conditions ($< \sim 60\,^{\circ}$C), i.e. before the 10.8 Ma onset of continuous sedimentation on Crete.

However, despite the fact that the LN was exhumed to shallow depth, there is no evidence that it was exposed anywhere on Crete at the beginning of subsidence and sedimentation: the Viannos Fm or time-equivalent Males Fm of eastern Crete unconformably overlie the UN and nowhere contain LN debris (Fortuin, 1978; Kopp & Richter, 1983; Fortuin & Peters, 1984; van Hinsbergen *et al.*, 2008). The oldest LN debris is found on eastern Crete in sediments of the Fothia Fm overlying the Males Fm (Fortuin, 1978). The first exposure of LN rocks thus occurred at the end or shortly after deposition of the Viannos and Males Fms.

In several places on Crete, basins of Tortonian age, but postdating the Viannos and Males Fms, have a basin floor of LN rocks with basin margins formed by Tripolitza limestone klippen (van Hinsbergen & Meulenkamp, 2006). Prime examples are found at Zakros and Rousa Limni in eastern Crete (Fig. 2a), and at Moni Veniou and Plakias in western Crete (Fig. 2a). These last-named basins are filled with nonmarine deposits that at Plakias date back to 9.6–9.3 Ma (see discussion under Ages of the formations). The best explanation for the formation of these pre-9.6–9.3 Ma basin configurations is by stretching of the hangingwall of the Cretan detachment with respect to footwall leading to extensional windows wih large UN blocks overlying the Cretan detachment as extensional klippen. This suggests that the Cretan detachment has been active until some time after the deposition of the Viannos Fm but not later than 9.6–9.3 Ma. The early history of Cretan sedimentation is thus governed by activity of the Cretan detachment, and can be considered a supradetachment basin (van Hinsbergen & Meulenkamp, 2006).

Subsidence and fragmentation of the detachment-related Viannos Basin

The E–W trending basin that accommodated the fluvial sediments of the westward-draining Males-Viannos River extended over the entire length of Crete and lasted from 10.8 to 10.4 Ma. The borders of this basin remain vague but we discussed earlier that the southern margin of the Viannos Basin in central Crete could have been located to the N of the Asteroussia Mts. More specifically, these mountains are considered the footwall with the breakaway fault rooting into the Cretan detachment located directly to the N (Fig. 8a). The northern hinterland (termed south


(a) Fluvial Viannos Basin and eastern extension shortly before olistolith emplacement around 10.4 Ma
fluvio-lacustrine sediments overlying Upper Nappes
hinterland
Upper Nappes
Lower Nappes
present-day configuration
breakaway fault
northern margin (tentative)
northern hinterland
footwall
(b) Marine Skinias-Ierapetra Basin (10.3-9.7 Ma)
upper bathyal
middle bathyal
hinterland
Upper Nappes
Lower Nappes
present-day configuration
extensional windows exposing Lower Nappes (schematic)
klippen
basin bounding fault
turbidity current flow
(c) Uplift and faulting of Skinias Basin (9.7-9.6 Ma)
++ uplift and erosion (areas II, III)
+– uplift and continuous sedimentation (area I)
hinterland of Skinias Basin
normal fault
bounding faults of Cretan horst
CHNF Central Heraklion Normal Fault
Upper Nappes
Lower Nappes
present-day configuration
Cretan Sea Basin
++
CHNF
++
+–
7493

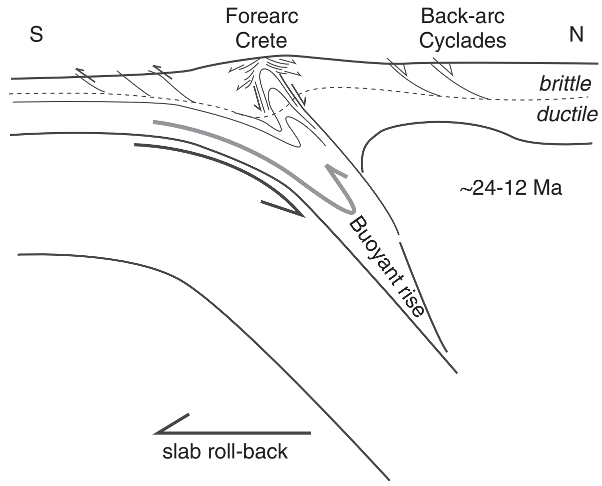


Fig. 9. Schematic cross-section illustrating the regional relationship between buoyancy-driven exhumation between 24–21 and 15–12 Ma in the Aegean forearc (Crete) and back-arc (Cyclades) during early to middle Miocene exhumation of the south Aegean HP-LT metamorphic rocks. Figure modified from Jolivet *et al.* (2003).

Aegean landmass in Drooger & Meulenkamp, 1973) was to be found in the present-day Cretan Sea.

The very uniform character of the Viannos Basin fill in central Crete without LN debris suggests that stretching of the hangingwall above the Cretan detachment was on the scale of observation homogeneous and widely distributed, i.e. essentially thinned by pure shear. The mass wasting deposits of breccio-conglomerates and olistoliths of western provenance deposited during the final stage of the Viannos Basin were interpreted in 'The Viannos and Skinias Basins (10.8–9.6 Ma)' as evidence for significant uplift of the present-day Psiloritis Mts. Similar occurrences in the Ierapetra area are overlain by marine sediments belonging to biozone 4 (our unpublished data) (i.e they correlate with the older part of the Skinias Fm) and thus are time-equivalent to those in central Crete. The olistostromes in the Ierapetra area are S to SE-directed (Fortuin, 1977, 1978; Fortuin & Peters, 1984) suggesting uplift in the N to NW. The uplift of the present-day Psiloritis Mts and the northwestern part of the Ierapetra area most likely reflects the isostatic response to the opening of the extensional windows discussed in 'Arguments for the existence of a supradetachment basin on Crete', probably at places of a weakened hangingwall (Fig. 8a). LN debris in coarse clastics overlying the olistostromes in the northern Ierapetra Basin (Fortuin, 1977, 1978; Fortuin & Peters, 1984) provides an age of 10.4–10.3 Ma for the earliest reworking of LN rocks on Crete which is perfectly in line with the formation and subsequent uplift of the extensional windows at or shortly before $\sim$10.4 Ma. Subsidence of the Viannos Basin hangingwall continued and resulted in the transformation of the fluvial Viannos Basin into the marine Skinias Basin between 10.4 and 10.3 Ma of which the northern margin is formed by the uplifted extensional windows and their adjoining areas (Fig. 8b). The upper bathyal Skinias Basin accommodated 200–250 m of sediments between 10.3 and 9.7 Ma and was connected with the middle bathyal Ierapetra Basin in the E (Fig. 8b). We do not know in this stage whether the Cretan detachment remained active over the period of 10.4–9.7 Ma during which the Skinias Basin subsided $\sim$500 m (Fig. 5). Hangingwall stretching of the floor of the Skinias Basin might have continued since there is no evidence that basin bounding faults were active. With the uplift of the Skinias Basin between 9.7 and 9.6 Ma, followed shortly afterwards by fragmentation along dominant N–S and E–W striking normal faults, subsidence became not longer governed by the Cretan detachment but by block-faulting. Figure 8c portrays two of these fault blocks corresponding with areas I and III with area III experiencing stronger uplift than area I thereby losing its sediment cover by erosion. Subsidence of area III may have resumed at the same time and along the same E–W basin bounding faults as area I (Fig. 8c) but accommodated sediments later resulting in the SHB portrayed in Fig. 6a. Area II remained emergent and subject to erosion for a period of $\sim$1.2 Myr since 9.6 Ma. The Psiloritis Mts continued to be emergent from $\sim$10.4 Ma until today. We propose that the overall uplift of central Crete at 9.7–9.6 Ma and dissection by an E–W and dominant N–S striking fault pattern shortly thereafter is the isostatic response to the formation of the Cretan Sea Basin in the N and a basin S of the Asteroussia Mts under continuous N–S extension (Fig. 8c). The forearc horst which defines Crete today thus started to form 9.7 to 9.6 Ma ago, and became subject to arc-parallel extension since then.

Implications for HP-LT exhumation mechanisms

Our analysis of the earliest Cretan basin history shows that the oldest sediments that appear to be directly governed by activity of the Cretan detachment are $\sim$10.8 Ma old. There are older sediments on Crete (Kopp & Richter, 1983; Peters, 1985; van Hinsbergen & Meulenkamp, 2006; Seidel *et al.*, 2007; van Hinsbergen *et al.*, 2008), but there is no conclusive evidence that these formed in an extensional basin; they could be local pied-

Fig. 8. Schematic palaeogeographic maps for three time slices illustrating the early basin development in central Crete (10.8–9.6 Ma). (a) The geometry of the fluvio-lacustrine Viannos supradetachment Basin (and eastern extension) before olistolith emplacement. (b) The contours of the marine Skinias-Ierapetra Basin. The northern margin is formed by the uplifted extensional windows and adjoining areas. The exhumed LN and extensional klippe at Plakias represent floor and margin of a basin dating back to 9.6–9.3 Ma (see earlier discussion). (c) Uplift and dissection by high-angle faults of the Skinias Basin. Note the role of the CHNF in delimiting the northern margin of the SHB portrayed in Fig. 6a.

mont breccias, broken into the extensional klippen they form today.

Crustal extension, whether accommodated by symmetric (McKenzie, 1978) or asymmetric (e.g. Tirel *et al.*, 2008, 2009) structures, is inevitably associated with basin subsidence. Our analysis shows that in the Cretan segment of the Aegean forearc, 'simple' crustal extension has been active for at least the last 10.8 Myr. However, the vast majority of exhumation of the LN now exposed on Crete, from depths of 30 km or more, to 2–3 km occurred before basin formation, between 24–21 Ma and 15–12 Ma (e.g. Jolivet *et al.*, 1996; Thomson *et al.*, 1998; Marsellos *et al.*, 2010), and was associated with the strong reduction of the combined thickness of the UN, from their combined $\sim$25 km preserved in western Greece (Jacobshagen, 1986; van Hinsbergen *et al.*, 2005a) to $\sim$2 km on Crete (Bonneau, 1984). Early-middle Miocene exhumation of the LN is hence associated with extreme thinning and extension of the UN by a β-factor of 10 or more.

If this was straightforward crustal extension, the present-day crustal thickness of Crete of $\sim$30 km (Tirel *et al.*, 2004) would restore to an original pre-Miocene thickness of 300 km, which is obviously impossible. In addition, such extreme thinning would inevitably lead to strong subsidence, and accumulation of deep marine sediments, of which there is no trace. Indeed, the only relics of pre-Viannos Fm sediments are nonmarine breccio-conglomerates (Kopp & Richter, 1983; Peters, 1985).

Hence, the thinning of the UN, exhuming the LN, must have been compensated by a process that prevented the surface to subside. Previous workers have suggested a mechanism for exhumation of the LN by buoyancy- or extrusion driven upward flow of rock above the subducting African slab (Thomson *et al.*, 1999; Jolivet *et al.*, 2003; Chatzaras *et al.*, 2006; Ring *et al.*, 2010), a process that was numerically modeled by Beaumont *et al.* (2009). This process could be viewed as 'intrusion' of buoyantly rising HP-LT metamorphic rocks into a localized zone of extension (Fig. 9). The fact that the relics of the Uppermost Unit are still present on Crete suggests that erosion had little influence during this process. Rather, the entire UN were tectonically thinned, and the growing evidence for coaxial deformation associated with the exhumation of the LN (e.g. Papanikolaou & Vassilakis, 2010) suggests that this process occurred symmetrically, similar to the modeling results of Beaumont *et al.* (2009).

CONCLUSIONS

(1) The beginning of subsidence and continuous sedimentation on Crete is governed by N–S extension accommodated along the Cretan detachment and lasted from 10.8 to 10.4 Ma, and possibly until 9.7 Ma.

(2) Basin width and fill characteristics of the fluvial supra-detachment basin in central Crete (Viannos Basin) indicate that stretching of the hangingwall above the Cretan detachment was effectively homogeneous, but bounded in the S by an inferred break-away fault along the northern margin of the Asteroussia Mts.

(3) Olistostromes including large exotic blocks deposited during the final stage of the Viannos Basin in central Crete ($\sim$10.4 Ma) and similar and time-equivalent deposits in the Ierapetra region are related to the opening and subsequent isostatic uplift of extensional windows bringing the Cretan HP-LT metamorphic rocks to the surface.

(4) The fluvio-lacustrine Viannos Basin transformed into the marine Skinias Basin between 10.4 and 10.3 Ma which is also the time span in which we find the oldest reworked LN debris on Crete. Subsidence of $\sim$500 m of the marine Skinias Basin between 10.4 and 9.7 Ma might have been governed by continued hangingwall stretching accommodated by the Cretan detachment.

(5) With the uplift of the Skinias Basin at 9.7–9.6 Ma, followed shortly afterwards by block faulting along dominant N–S and E–W striking normal faults, including the here defined Central Heraklion Normal Fault, subsidence became not longer governed by the Cretan detachment but by block-faulting. This period marks the transition on Crete from dominant N–S to combined N–S and E–W stretching. We propose that the uplift of the Skinias Basin was related to the formation of the Cretan Sea Basin to the N and deep marine basins to the S of Crete. The horst structure of present-day Crete may therefore date back to $\sim$9.6 Ma.

(6) The bulk of exhumation of the Cretan HP-LT metamorphic rocks occurred during the early to middle Miocene, before inception of basin subsidence, and was associated with extreme thinning of the hangingwall (by factor $\sim$10) to the Cretan detachment, in line with earlier inferences that the Cretan detachment can only explain a minor part of the total exhumation. This necessitates a process that prevents subsidence despite extension. Previously proposed models of buyoant rise of the Cretan HP-LT and presently still underlying rocks, along the subducting African slab provide such an explanation.

ACKNOWLEDGEMENTS

Wout Krijgsman kindly provided unpublished palaeomagnetic data from section Demati. Elena Turco and Hemmo Abels are thanked for their help in making Figures S2 and S3. Hans de Bruin is thanked for his expert opinion on the age of the mammal faunules of Kastellios Hill and Plakias. This paper further benefited from discussions with Rob Govers, Reinoud Vissers, George Postma, Johan ten Veen and fieldwork reports from Tanja van Kouwenhoven and Joris Steenbrink. Geert Ittmann is acknowledged for washing the many samples. We are further obliged to Margot Stoete, Rien Rabbers and Fred Trappenburg (Geomedia service) and Ton van Hinte (GIS service) for advice and fine draft work. We thank the Institute of Geology and Mineral Exploration (IGME) in Athens for providing working permits and Assi Antonarakou (University of

Athens) for her help in the transport of the many samples. Finally we thank Harms-ic-Verlag in Germany for permission to use a scanned version of their Tourist map 1 : 100 000 of Crete. W.J.Z thanks the people of Tefeli and specifically the Iakovidis and Papadopoulos families for their hospitality and friendship over the years. D.J.J.v.H. acknowledges financial support through an NWO-Veni grant and from StatOil (SPlates project). The thorough and insightful reviews of Laurent Jolivet and an anonymous reviewer, as well as the editorial comments of Brian Horton are gratefully acknowledged.

SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article:

Appendix S1. Data file showing details on location, lithology, biostratigraphy and palaeobathymetry of the studied sections and outcrops in central Crete along with their lithostratigraphic and biozonal assignment. Sections and outcrops in this file are arranged from area I to V and from E to W per area.

Appendix S2. Description and definition of the planktonic foraminiferal biozones defined in Gibliscemi (Sicily) and Metochia (Gavdos). The biozones are considered to be useful for correlating and dating upper middle to lower upper Miocene sediments in the Mediterranean region.

Figure S1. Location of sections and outcrops shown in Figure 2b plotted on a scanned version of the Tourist Map of Crete 1:100,000 (Harms-ic-Verlag, Germany). O = Kourtes; N = Agios Ioannis

Figure S2. (Semi) Quantitative distribution and coiling of selected planktonic foraminiferal taxa in section Gibliscemi (Sicily). This section combines subsections A to F described in Krijgsman *et al.* (1995) and Hilgen *et al.* (2000). The semiquantitative data for Gibliscemi A, B and C (extending from cycle G -3 up to top sapropel of cycle G 85 are based on 125-595 micron fraction counts of all 620 samples (average sample resolution $<$ 5 kyr) including the data-set published in Krijgsman *et al.* (1995) which was based on only 40% of all samples from Gibliscemi A and B. This semi-quantitative record is spliced with the published quantitative pattern in the older part of the Gibliscemi sequence, i.e. Gibliscemi E, D and F (Hilgen *et al.*, 2000; Turco *et al.*, 2001). The spliced record combines *Neogloboquadrina atlantica* (small) and *N. atlantica* (large) into *Neogloboquadrina atlantica* and *N. acostaensis sensu* stricto and N. 4-chambered into *Neogloboquadrina acostaensis*. We further added percentages of *N. atlantica* per total neogloboquadrinids (with–10 values indicating samples where neogloboquadrinids are absent). The distribution and coiling pattern of the selected planktonic foraminiferal taxa allows a subdivision of the spliced record into nine planktonic foraminiferal biozones (numbered 1 to 9) and two subzones (numbered 7.1 and 7.2) up to the level of a significant hiatus at the top of the sapropel of cycle G 85 spanning $\sim$200 kyr (Krijgsman et al., 1995).

Figure S3. Semi-quantitative distribution and coiling of selected planktonic foraminiferal taxa in section Metochia (Gavdos). The data presented here replace the original presence-absence data in Krijgsman *et al.* (1995) and are based on studying the 125-595 micron fraction of all 560 samples providing a resolution of $<$ 5 kyr. Note that specimens referred to as *Catapsydrax parvulus* in Krijgsman *et al.* (1995) are re-labeled *Globorotaloides falconarae* (see also discussion in Hilgen *et al.*, 2000). Lithology, sample position and cycle numbers are from Krijgsman *et al.* (1995) with the mentioning that we extended the original cycle pattern downward with cycles M0 to M-2. Biozones 6 to 9 and subzones 7.1 and 7.2 defined in Gibliscemi are present in Metochia as well and extended upwards by biozones 10 and 11.

Figure S4. Lithostratigraphic units for the Neogene basin fill in central Crete in this and earlier studies. Units are equally boxed using the base of Pliocene as reference line.

Figure S5. Photobook.

Table S1. Ages of planktonic foraminiferal bioevents defining 11 biozones (described in Appendix S2) and their position in terms of sample and sedimentary cycle numbers in sections Gibliscemi (Sicily) and Metochia (Gavdos). Ages for defining bioevents in Metochia are derived from retuning of sedimentary cycles (M-cycles) to the Laskar 2003(1,1) solution by Lourens et al. (2004). Updated ages for bioevents in Gibliscemi A and B are based on retuning G-cycles to La2003(1,1) by Lourens et al. (2004) and for bioevents in Gibliscemi F-C by retuning correlative cycles in the Monte dei Corvi section (northern Italy) to La2004(1,1) by Hüsing et al. (2007). Ages in red are used in this paper. Also shown is position of bioevents in terms of sample number (as registered in the Utrecht collections) and sedimentary cycle number. Table also includes identified magnetic chron boundaries in Metochia (Krijgsman et a., 1995) and Monte dei Corvei (Hüsing et al., 2007) together with their position in terms of M-cycles and correlative G-cycles (from Hilgen et al., 2003). Updated ages for magnetic chron boundaries in Metochia are from Lourens et al. (2004) and in Monte dei Corvi from Hüsing et al. (2007).

[Correction added after online publication 24 August 2011 - in Location 8 in Appendix S1, the sentence 'a correlation of the lower Kastellios Hill section to Chron 4Ar.2r is the most likely' was changed to 'a correlation of the lower Kastellios Hill section to Chron 4Ar.1r is the most likely']

REFERENCES

Aguilar, J.-P., Berggren, W.A., Aubry, M.-P., Kent, D.V., Clauzon, G., Benammi, M. & Michaux, J. (2004) Mid-Neogene Mediterranean marine-continental correlations: an alternative interpretation. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **204**, 165–186.

Angelier, J. (1975) Sur l'analyse des phases superposées de la tectonique cassante: la néotectonique et les failles du massif de l'Ida (Crète, Grèce). *Ann. Soc. Geol. Nord*, **XCV**, 183–200.

Beaumont, C., Jamieson, R.A., Butler, J.P. & Warren, C.J. (2009) Crustal structure: a key constraint on the mechanism of ultra-high-pressure rock exhumation. *Earth Planet. Sci. Lett.*, **287**, 116–129.

Bonneau, M. (1984) Correlation of the Hellenic Nappes in the South-East Aegean and their tectonic reconstruction. In: *The Geological Evolution of the Eastern Mediterranean* (Ed. by J.E. Dixon & A.H.F. Robertson), *Geol. Soc., Lond., Spec. Publ.*, **17**, 517–527.

Bornovas, I. & Rontogianni-Tsiabaou, T. (1983) *Geological Map of Greece*. Institute of Geology and Mineral Exploration, Athens.

Brachert, T.C., Reuter, M., Felis, T., Kroeger, K.F., Lohmann, G., Micheels, A. & Fassoulas, C. (2006) *Porites* Corals from Crete (Greece) open a window into late Miocene (10 Ma) seasonal and interannual climate variability. *Earth Planet. Sci. Lett.*, **245**, 81–94.

Brix, M.R., Stöckhert, M., Seidel, E., Theye, T., Thomson, S.A. & Küster, M. (2002) Thermobarometric data from a fossil zircon partial annealing zone in high pressure-low temperature rocks of eastern and central Crete, Greece. *Tectonophysics*, **349**, 309–326.

Burchfiel, B.C. & Royden, L.H. (1985) North-south extension within the convergent Himalayan region. *Geology*, **13**, 679–682.

Chatzaras, V., Xypolias, P. & Doutsos, T. (2006) Exhumation of high-pressure rocks under continuous compression: a working hypothesis for the southern Hellenides (Central Crete, Greece). *Geol. Mag.*, **143**, 859–876.

de Bruin, H., Daams, R., Daxner-Höck, G., Fahlbusch, V., Ginsburg, L., Mein, P., Morales, J., Heinzmann, E., Mayhew, D.F., Van der Meulen, A.J., Schmidt-Kittler, N. & Telles Antunes, M. (1992) Report of the RCMNS Working Group on fossil mammals, Reisensburg 1990. *Newslett. Stratigr.*, **26**, 65–118.

de Bruin, H. & Meulenkamp, J.E. (1972) Late Miocene rodents from the Pandanassa formation (Prov. Rethymnon), Crete, Greece. *Proc. van de Koninklijke Nederlandse Acad. van Wetenschappen*, **B 75**, 54–60.

de Bruin, H., Sondaar, P.Y. & Zachariasse, W.J. (1971) Mammalia and foraminifera from the Neogene of Kastellios Hill (Crete), a correlation of continental and marine biozones. *Proc. van de Koninklijke Nederlandse Acad. van Wetenschappen*, **B 74**, 1–22.

de Bruin, H. & Zachariasse, W.J. (1979) The correlation of marine and continental biozones of Kastellios Hill reconsidered. *Ann. Geol. Pays Hellenique*, **H.S. 1**, 219–226.

Delrieu, B. (1990) Evolution tectono-sédimentaire du Malevisi et du secteur d'Ano Moulia au Miocene Supérieur (Bassin d'Hérak-lion, Crète Centrale, Grèce). *Mém. Géol. l'IGAL*, **42**, 1–314.

Delrieu, B., Saint-Martin, J.P., Koskeridou, E., Chaix, C., Moissette, P., Tsagaris, S., Perret, A. & Audoult, P.B. (2004). *Late Tortonian eustatic rise in the eastern Mediterranean: recording from an upward backstepping geometry of coral reefs in Central Crete*. 5th International Symposium on Eastern Mediterranean Geology, Thessaloniki, Greece.

Delrieu, B., Saint-Martin, J.P. & Merle, D. (1991) Un modèle d'évolution tectonosédimentaire dans le domaine Sud-Égéen au Miocène Supérieur: L'accident d'Aghia Varvara (Crète Central, Grèce). *C. R. Acad. Sci. Paris*, **313**, 1043–1049.

Drooger, C.W. & Meulenkamp, J.E. (1973) Stratigraphic contributions to geodynamics in the Mediterranean Area: Crete as a case history. *Bull. Geol. Soc. Greece*, **10**, 193–200.

Duermeijer, C.E., Krijgsman, W., Langereis, C.G. & ten Veen, J.H. (1998) Post early Messinian counter-clockwise rotations on Crete: implications for the Late Miocene to Recent kinematics of the southern Hellenic arc. *Tectonophysics*, **298**, 77–89.

Faccenna, C., Jolivet, L., Piromallo, C. & Morelli, A. (2003) Subduction and the depth of convection of the Mediterranean mantle. *J. Geophys. Res.*, **108**, 2099.

Fassoulas, C. (1999) The structural evolution of Central Crete: insight into the tectonic evolution of the South Aegean (Greece). *J. Geodynam.*, **27**, 23–43.

Fassoulas, C. (2001) The tectonic development of a Neogene basin at the leading edge of the active European margin: the Heraklion basin, Crete, Greece. *J. Geodynam.*, **31**, 49–70.

Fassoulas, C., Kilias, A. & Mountrakis, D. (1994) Postnappe stacking extension and exhumation of High-Pressure/Low-Temperature rocks in the island of Crete, Greece. *Tectonics*, **13**, 127–138.

Flecker, R., de Villiers, S. & Ellam, R.M. (2002) Modelling the effect of evaporation on the salinity-$^{87}Sr/^{86}Sr$ relationship in modern and ancient marginal-marine systems: the Mediterranean Salinity Crisis. *Earth Planet. Sci. Lett.*, **203**, 221–233.

Forster, M.A. & Lister, G.S. (2009) Core-complex-related extension of the Aegean lithosphere initiated at the Eocene-Oligocene transition. *J. Geophys. Res.*, **114**, B02401, doi: 10.1029/2007JB005382.

Fortuin, A.R. (1977) Stratigraphy and sedimentary history of the Neogene deposits in the Ierapetra region, Eastern Crete. *GUA Pap. Geol.*, **8**, 1–164.

Fortuin, A.R. (1978) Late Cenozoic history of eastern Crete and implications for the geology and geodynamics of the southern Aegean region. *Geol. Mijnbouw*, **57**, 451–464.

Fortuin, A.R. & Peters, J.M. (1984) The Prina Complex in Eastern Crete and its relationship to possible Miocene strike-slip tectonics. *J. Struct. Geol.*, **6**, 459–476.

Friedmann, S.J. & Burbank, D.W. (1995) Rift basins and supra-detachment basins: intracontinental extensional end-members. *Basin Res.*, **7**, 109–127.

Garces, M., Augusti, J., Cabrera, L. & Pares, J.M. (1996) Magnetostratigraphy of the Vallesian (late Miocene) in the Valles-Penedes Basin (NE Spain). *Earth Planet. Sci. Lett.*, **142**, 381–396.

Gautier, P., Brun, J.-P. & Jolivet, L. (1993) Structure and kinematics of upper Cenozoic extensional detachment on Naxos and Paros. *Tectonics*, **12**, 1180–1194.

Gautier, P., Brun, J.-P., Moriceau, R., Sokoutis, D., Martinod, J. & Jolivet, L. (1999) Timing, kinematics and cause of Aegean extension: a scenario based on a comparison with simple analogue experiments. *Tectonophysics*, **315**, 31–72.

Hageman, J. (1979) Benthic foraminiferal assemblages from Plio-Pleistocene open bay to lagoonal sediments of the western Peloponnesus (Greece). *Utrecht Micropaleontol. Bull.*, **20**, 1–171.

Hilgen, F.J., Abdul Aziz, H., Krijgsman, W., Raffi, I. & Turco, E. (2003) Integrated stratigraphy and astronomical tuning of the Serravallian and lower Tortonian at Monte Dei Corvi (Middle-Upper Miocene, Northern Italy). *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **199**, 229–264.

Hilgen, F.J., Krijgsman, W., Langereis, C.G., Lourens, L.J., Santarelli, A. & Zachariasse, W.J. (1995) Extending the astronomical (polarity) time scale into the Miocene. *Earth Planet. Sci. Lett.*, **136**, 495–510.

Hilgen, F.J., Krijgsman, W., Raffi, I., Turco, E. & Zachariasse, W.J. (2000) Integrated stratigraphy and astronomical calibration of the Serravallian/Tortonian boundary section at Monte Gibliscemi (Sicily, Italy). *Mar. Micropaleontol.*, **38**, 181–211.

Hilgen, F.J., Krijgsman, W. & Wijbrans, J.R. (1997) Direct comparison of astronomical and $^{40}Ar/^{39}Ar$ ages of ash beds: potential implications for the age of mineral dating standards. *Geophys. Res. Lett.*, **24**, 2043–2046.

Hüsing, S.K., Hilgen, F.J., Adbul Aziz, H. & Krijgsman, W. (2007) Completing the Neogene geological timescale between 8.5 and 12.5 Ma. *Earth Planet. Sci. Lett.*, **253**, 340–358.

IGME. (1989) *Geological Map of Greece, Sheet Mochos*. Institute of Geology and Mineral Exploration (IGME), Athens.

IGME. (1994) *Geological Map of Greece, Sheet Epano Archanae*. Institute of Geology and Mineral Exploration (IGME), Athens.

Jacobshagen, V. (1986) *Geologie Von Griechenland*. Borntraeger, Berlin-Stuttgart, 279pp.

Jolivet, L. (2001) A comparison of geodetic and finite strain pattern in the Aegean, geodynamic implications. *Earth Planet. Sci. Lett.*, **187**, 95–104.

Jolivet, L. & Brun, J.P. (2010) Cenozoic geodynamic evolution of the Aegean. *Int. J. Earth Sci.*, **99**, 109–138.

Jolivet, L., Facenna, C., Goffé, B., Burov, E. & Agard, P. (2003) Subduction tectonics and exhumation of high-pressure metamorphic rocks in the Mediterranean orogen. *Am. J. Sci.*, **303**, 353–409.

Jolivet, L., Goffé, B., Monié, P., Truffert-Luxey, C., Patriat, M. & Bonneau, M. (1996) Miocene detachment on Crete and exhumation P-T-t paths of high-pressure metamorphic rocks. *Tectonics*, **15**, 1129–1153.

Jolivet, L., Trotet, F., Monié, P., Vidal, O., Goffé, B., Labrousse, L., Agard, P. & Ghorbal, B. (2010) Along-strike variations of P-T conditions in accretionary wedges and syn-orogenic extension, the HP-LT phyllite-quartzite Nappe in Crete and the Peloponnese. *Tectonophysics*, **480**, 133–148.

Kilias, A., Fassoulas, C. & Mountrakis, D. (1994) Tertiary extension of continental crust and uplift of Psiloritis metamorphic core complex in the central part of the Hellenic Arc (Crete, Greece). *Geol. Rundsch.*, **83**, 417–430.

Klein, T., Reichhardt, H., Klinger, L., Grigull, S., Wostal, G., Kowalczyk, G. & Zulauf, G. (2008) Reverse slip along the contact Phyllite-Quartzite Unit/Tripolitza Unit in eastern Crete: implications for the geodynamic evolution of the External Hellenides. *Z. Deutsc. Geselschaft Geowissenschaften*, **159**, 375–398.

Köhler, C., Heslop, D., Dekkers, M.J., Krijgsman, W., van Hinsbergen, D.J.J. & von Dobeneck, T. (2008) Fuzzy C-means cluster analysis on a late Miocene multiproxy data set of the eastern Mediterranean: constrains the disappearance of Crete as a sediment source. *Geochem. Geophys. Geosyst.*, **9**, Q12018, doi: 12010.11029/12008GC002127.

Köhler, C., Krijgsman, W., van Hinsbergen, D.J.J., Heslop, D. & Dupont-Nivet, G. (2010) Concurrent tectonic and climatic changes recorded in upper Tortonian sediments from the Eastern Mediterranean. *Terra Nova*, **22**, 52–63.

Kopp, K.O. & Richter, D. (1983) Synorogenetische Schuttbildungen und die Eigenstaendigkeit der Phyllit-Gruppe auf Kreta. *Neues Jahrbuch Geol. Palaeontol. Monatsheften*, **165**, 228–253.

Krijgsman, W., Garcés, M., Langereis, C.G., Daams, R., Van Dam, J., van der Meulen, A.J., Agustí, J. & Cabrera, L. (1996) A new chronology for the middle to late Miocene continental record in Spain. *Earth Planet. Sci. Lett.*, **142**, 367–380.

Krijgsman, W., Hilgen, F.J., Langereis, C.G., Santarelli, A. & Zachariasse, W.J. (1995) Late Miocene magnetostratigraphy, biostratigraphy and cyclostratigraphy in the Mediterranean. *Earth Planet. Sci. Lett.*, **136**, 475–494.

Krijgsman, W., Hilgen, F.J., Langereis, C.G. & Zachariasse, W.J. (1994) The age of the Tortonian/Messinian boundary. *Earth Planet. Sci. Lett.*, **121**, 533–547.

Le Pichon, X., Angelier, J. & Sibuet, J.-C. (1982) Plate boundaries and extensional tectonics. *Tectonophysics*, **81**, 239–256.

Lister, G.S., Banga, G. & Feenstra, A. (1984) Metamorphic core complexes of cordilleran type in the Cyclades. Aegean Sea, Greece. *Geology*, **12**, 221–225.

Long, S.P. & McQuarrie, N. (2010) Placing limits on channel flow: insights from the Bhutan Himalaya. *Earth Planet. Sci. Lett.*, **290**, 375–390.

Lourens, L.J., Hilgen, F.J., Laskar, J., Shackleton, N.J. & Wilson, D. (2004) Chapter 21: the Neogene period. In: *A Geologic Time Scale 2004* (Ed. by F.M. Gradstein, J.G. Ogg & A.G. Smith), pp. 409–440. Cambridge University Press, Cambridge.

Marsellos, A.E. & Kidd, W.S.F. (2008) Extension and exhumation of the Hellenic forearc ridge in Kythera. *J. Geol.*, **116**, 640–651.

Marsellos, A.E., Kidd, W.S.F. & Garver, J.I. (2010) Extension and exhumation of the HP-LT rocks in the Hellenic forearc ridge. *Am. J. Sci.*, **310**, 1–36.

McKenzie, D. (1978) Some remarks on the development of sedimentary basins. *Earth Planet. Sci. Lett.*, **40**, 25–32.

Mertz-Krauss, R., Brachert, T.C., Galer, S.J.G., Stoll, B. & Jochum, K. (2007) Seasonal element and Sr isotope ratio variations in late Miocene corals from Crete, eastern Mediterranean. *AGU, Fall Meeting 2007*, PP31D-0657.

Mertz-Krauss, R., Brachert, T.C. & Reuter, M. (2008) *Tarbellastraea* (Scleractinia): a new stable isotope archive for late Miocene paleoenvironments in the Mediterranean. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **257**, 294–307.

Meulenkamp, J.E. (1969) Stratigraphy of Neogene deposits in the Rethymnon province, Crete, with special reference to the phylogeny of uniserial *Uvigerina* from the Mediterranean region. *Utrecht Micropaleontol. Bull.*, **2**, 168.

Meulenkamp, J.E., Dermitzakis, M., Georgiadou-Nikeoulia, E., Jonkers, H.A. & Boger, H. (1979) Field guide to the Neogene of Crete. In: *Publications of the Department of Geology and Paleontology, University of Athens*. (Ed. by N. Symeonidis, D. Papanikolaou & M. Dermitzakis), Series A, **32**, 1–32.

Meulenkamp, J.E., Van der Zwaan, G.J. & Van Wamel, W.A. (1994) On late Miocene to Recent vertical motions in the Cretan segment of the Hellenic arc. *Tectonophysics*, **234**, 53–72.

Meulenkamp, J.E., Wortel, M.J.R., Van Wamel, W.A., Spakman, W. & Hoogerduyn Strating, E. (1988) On the Hellenic subduction zone and the geodynamical evolution of Crete since the late Middle Miocene. *Tectonophysics*, **146**, 203–215.

Papanikolaou, D. & Vassilakis, E. (2010) Thrust faults and extensional detachment faults in Cretan tectono-stratigraphy: implications for Middle Miocene extension. *Tectonophysics*, **488**, 233–247.

Peters, J.M. (1985) Neogene and Quarternary vertical tectonics in the South Hellenic arc and their effect on concurrent sedimentation processes. *GUA Pap. Geol.*, **23**, 1–247.

Postma, G., Hilgen, F.J. & Zachariasse, W.J. (1993a) Precession-punctuated growth of a late Miocene submarine-fan lobe on Gavdos (Greece). *Terra Nova*, **5**, 438–444.

POSTMA, G., FORTUIN, A.R. & VAN WAMEL, W.A. (1993b) Basin-fill patterns controlled by tectonics and climate: the Neogen "fore-arc" basins of eastern Crete as a case history. *Spec. Publs. Int. Ass. Sediment.*, **20**, 326–335.

RAHL, J.M., ANDERSON, K.M., BRANDON, M.T. & FASSOULAS, C. (2005) Raman spectroscopic carbonaceous material thermometry of low-grade metamorphic rocks: calibration and application to tectonic exhumation in Crete, Greece. *Earth Planet. Sci. Lett.*, **240**, 339–354.

RING, U. & GLODNY, J. (2010) No need for lithospheric extension for exhuming (U)HP rocks by normal faulting. *J. Geol. Soc. Lond.*, **167**, 225–228.

RING, U., GLODNY, J., WILL, T. & THOMSON, S.N. (2010) The Hellenic subduction system: high-pressure metamorphism, exhumation, normal faulting, and large-scale extension. *Ann. Rev. Earth Planet. Sci.*, **38**, 45–76.

RING, U., LAYER, P.W. & REISCHMANN, T. (2001) Miocene high-pressure metamorphism in the Cyclades and Crete, Aegean Sea, Greece: evidence for large-magnitude displacement on the Cretan detachment. *Geology*, **29**, 395–398.

SÁNCHEZ-GÓMEZ, M., AVIGAD, D. & HEIMANN, A. (2002) Geochronology of clasts in allochtonous Miocene sedimentary sequences on Mykonos and Paros islands: implications for back-arc extension in the Aegean Sea. *J. Geol. Soc. Lond.*, **159**, 45–60.

SEARLE, M.P. & GODIN, L. (2003) The South Tibetan detachment and the Manaslu Leucogranite: a structural reinterpretation and restoration of the Annapurna-Manaslu Himalaya, Nepal. *J. Geol.*, **111**, 505–523.

SEIDEL, E., OKRUSH, M., KREUZER, H., RASCHKA, H. & HARRE, W. (1981) Eo-Alpine metamorphism in the uppermost unit of the Cretan nappe-pile – petrology and geochronology – Part 1. The Léndas Area (Asteroussia Mountains). *Contrib. Mineral. Petrol.*, **57**, 259–275.

SEIDEL, M., SEIDEL, E. & STÖCKHERT, B. (2007) Tectono-sedimentary evolution of Lower to Middle Miocene half-graben basins related to an extensional detachment fault (Western Crete, Greece). *Terra Nova*, **19**, 39–47.

SEN, S. & SEYITOĞLU, G. (2009) Magnetostratigraphy of Early-Middle Miocene Deposits from E-W trending Alasehir and Büyük Menderes Grabens in Western Turkey, and its tectonic implications. In: *Collision and Collapse at the Africa-Arabia-Eurasia Subduction Zone* (Ed. by D.J.J. van Hinsbergen, M.A. Edwards & R. Govers), *Geol. Soc. Lond. Spec. Publ.*, **311**, 321–342.

SEN, S., VALET, J.-P. & IOAKIM, C. (1986) Magnetostratigraphy and biostratigraphy of the Neogene deposits of Kastellios Hill (Central Crete, Greece). *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **53**, 321–334.

SISSINGH, W. (1974) The Miocene ostracoda from the Hipparion-bearing beds of Kastellios Hill. *Proc. van de Koninklijke Nederlandse Acad. van Wetenschappen*, **77**, 119–128.

TEN VEEN, J.H. & KLEINSPEHN, K.L. (2003) Incipient continental collision and plate-boundary curvature: late Pliocene-Holocene transtensional Hellenic forearc, Crete, Greece. *J. Geol. Soc. Lond.*, **160**, 161–181.

TEN VEEN, J.H. & POSTMA, G. (1999a) Neogene tectonics and basin fill patterns in the Hellenic outer-arc (Crete, Greece). *Basin Res.*, **11**, 243–266.

TEN VEEN, J.H. & POSTMA, G. (1999b) Roll-back controlled vertical movements of outer-arc basins of the Hellenic subduction zone (Crete, Greece). *Basin Res.*, **11**, 223–241.

THEYE, T. & SEIDEL, E. (1991) Petrology of low-grade high-pressure metapelites from the External Hellenides (Crete, Peloponnese). A case study with attention to sodic minerals. *European Journal of Mineralogy*, **3**, 343–366.

THEYE, T., SEIDEL, E. & VIDAL, O. (1992) Carpholite, sudoite and chloritoid in low-grade high-pressure metapelites from Crete and the Peloponnese. *European Journal of Mineralogy*, **4**, 487–507.

THOMSON, S.N., STÖCKHERT, B. & BRIX, M.R. (1998) Thermochronology of the high-pressure metamorphic rocks of Crete, Greece: implications for the speed of tectonic processes. *Geology*, **26**, 259–262.

THOMSON, S.N., STÖCKHERT, B. & BRIX, M.R. (1999) Miocene high-pressure metamorphic rocks of Crete, Greece: rapid exhumation by buoyant escape. In: *Exhumation Processes: Normal Fa ulting, Ductile Fow and Erosion* (Ed. by U. Ring, M.T. Brandon, G.S. Lister & S.D. Willet), *Geol. Soc., Lond., Spec. Publ.*, **154**, 87–107.

TIREL, C., BRUN, J.-P. & BUROV, E. (2008) Dynamics and structural development of metamorphic core complexes. *J. Geophys. Res.*, **113**, B04403, doi: 04410.01029/02005JB003694.

TIREL, C., GAUTIER, P., VAN HINSBERGEN, D.J.J. & WORTEL, M.J.R. (2009) Sequential development of metamorphic core complexes: numerical simulations and comparison to the Cyclades, Greece. In: *Collision and Collapse at the Africa-Arabia-Eurasia Subduction Zone* (Ed. by D.J.J. van Hinsbergen, M.A. Edwards & R. Govers), *Geol. Soc., Lond., Spec. Publ.*, **311**, 257–292.

TIREL, C., GUEYDAN, F., TIBERI, C. & BRUN, J.-P. (2004) Aegean crustal thickness inferred from gravity inversion. Geodynamical implications. *Earth Planet. Sci. Lett.*, **228**, 267–280.

TSAGARIS, S. (1991) La Néogene Supérieur de la Messara Nord-Occidentale (Crète, Grèce)- Etude stratigraphique, sédimentaire et structurale- Essai de reconstruction paleoenvironmentale. *Mém.Géol. l'IGAL*, **47**, 1–262.

TURCO, E., HILGEN, F.J., LOURENS, L.J., SHACKLETON, N.J. & ZACHARIASSE, W.J. (2001) Punctuated evolution of global climate cooling during the late Middle to early Late Miocene: high-resolution planktonic foraminiferal and oxygen isotope records from the Mediterranean. *Paleoceanography*, **16**, 405–423.

VAN DE WEERD, A. (1983) Palynology of some Upper Miocene and Pliocene formations in Greece. *Geol. Jahrbuch, Reihe B*, **48**, 3–63.

VAN DER ZWAAN, G.J., JORISSEN, F.J. & DE STIGTER, H.C. (1990) The depth dependency of planktonic/benthonic foraminiferal ratios: constraints and applications. *Mar. Geol.*, **95**, 1–16.

VAN HINSBERGEN, D.J.J., HAFKENSCHEID, E., SPAKMAN, W., MEULENKAMP, J.E. & WORTEL, M.J.R. (2005a) Nappe stacking resulting from subduction of oceanic and continental lithosphere below Greece. *Geology*, **33**, 325–328.

VAN HINSBERGEN, D.J.J., KOUWENHOVEN, T.J. & VAN DER ZWAAN, G.J. (2005b) Paleobathymetry in the backstripping procedure: correction for oxygenation effects on depth estimates. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **221**, 245–265.

VAN HINSBERGEN, D.J.J. & MEULENKAMP, J.E. (2006) Neogene supra-detachment basin development on Crete (Greece) during exhumation of the South Aegean core complex. *Basin Res.*, **18**, 103–124.

VAN HINSBERGEN, D.J.J., SNEL, E., GARSTMAN, S.A., MARUNTEANU, M., LANGEREIS, C.G., WORTEL, M.J.R. & MEULENKAMP, J.E. (2004) Vertical motions in the Aegean volcanic arc: evidence for rapid subsidence preceding in situ volcanism. *Mar. Geol.*, **209**, 329–345.

VAN HINSBERGEN, D.J.J., ZACHARIASSE, W.J. & FORTUIN, A.R. (2008) Comment on 'Tectono-sedimentary evolution of Lower to Middle Miocene half-graben basins related to an extensional detachment fault (Western Crete, Greece)' by M. Seidel, E. Seidel and B. Stöckhert. *Terra Nova*, **20**, 414–416.

VAN HINSBERGEN, D.J.J., ZACHARIASSE, W.J., WORTEL, M.J.R. & MEULENKAMP, J.E. (2005c) Underthrusting and exhumation: a comparison between the external Hellenides and the "hot" Cycladic and "cold" South Aegean core complexes (Greece). *Tectonics*, **24**, TC2011, doi: 2010.1029/2004TC001692.

WILLMANN, R. (1980) Die Neogenen susswassergastropoden von Chersonisos (Kreta). *Neues Jahrbuch Geol. Palaeontol. Abh.*, **159**, 273–295.

ZACHARIASSE, W.J., VAN HINSBERGEN, D.J.J. & FORTUIN, A.R. (2008) Mass wasting and uplift on Crete and Karpathos (Greece) during the Early Pliocene related to beginning of South Aegean left-lateral, strike slip tectonics. *Geol. Soc. Am. Bull.*, **120**, 976–993.

Manuscript received 18 August 2009; In revised form 8 February 2011; Manuscript accepted 12 March 2011.

