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Tectonic and climatic controls on coastal sedimentation: The Late Pliocene–Middle Pleistocene of northeastern Rhodes, Greece

Jean-Jacques Cornée [a,*], Pierre Moissette [a], Sébastien Joannin [a], Jean-Pierre Suc [a], Frédéric Quillévéré [a], Wout Krijgsman [b], Frits Hilgen [c], Efterpi Koskeridou [d], Philippe Münch [e], Christophe Lécuyer [a], Paula Desvignes [a]

[a] *UMR CNRS 5125 Paléoenvironnements and Paléobiosphère, Université Claude Bernard Lyon I, 27 Bd du 11 Novembre 1918, 69622 Villeurbanne Cedex, France*
[b] *Paleomagnetic Laboratory 'Fort Hoofddijk' Faculty of Earth Sciences, Utrecht University Budapestlaan 17, 3584 CD Utrecht, The Netherlands*
[c] *Institute for Earth Sciences, Budapestlaan 4, POB 80021, 3584 TA Utrecht, The Netherlands*
[d] *Department of Historical Geology–Paleontology, University of Athens, Panepistimiopolis, Zografou, 15784 Athens, Greece*
[e] *FRE 2761 Géologie des Systèmes Carbonatés, Université de Provence, case 67, 3 place Victor Hugo, 13331 Marseille Cedex 3, France*

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Abstract

The Late Pliocene to Middle Pleistocene coastal sediments of northeastern Rhodes (Greece) were deposited in an active tectonic setting. They provide an excellent opportunity to investigate the relative roles played by climate and tectonics in sedimentary processes. The tectono-sedimentary organization of these deposits is revised in the light of an integrated study combining data from field investigations, sedimentology, bio- and magnetostratigraphy, radiometric dating, palaeoecology, and palynology.

Three lithostratigraphic units are recognised: the Rhodes Formation, the newly defined Ladiko–Tsampika Formation, and the Lindos Acropolis Formation. A major erosional surface separates the Rhodes Formation from the Ladiko–Tsampika Formation, which was deposited in deep and narrow palaeovalleys. The Rhodes Formation (Late Pliocene to 1.4–1.3 Ma) comprises three Members: the Kritika, the Lindos Bay clay and the Cape Arkhangelos calcarenite. The shallow-water clastic sediments of the Ladiko–Tsampika Formation (1.3 to 0.3 Ma) are subdivided into two Members: the Ladiko (mostly sandy) and the Tsampika (predominantly clayey). The Lindos Acropolis Formation is Late Pleistocene in age.

Two major transgression–regression cycles occurred prior to the Lindos Acropolis cycle. The deposition of the first cycle (Rhodes Formation) is tectonically controlled, with very high rates of vertical movements (2.6–5.2 mm/year). The second cycle (Ladiko–Tsampika Formation) records sea-level changes controlled by slow vertical motions (around 0.16 mm/year) and glacio-eustatic events with 40-kyr and 100-kyr periods. Finally, the tectonic and sedimentary evolution of Rhodes is integrated into the geodynamic context of the eastern Aegean fore-arc.

Keywords: Tectonics; Climate; Coastal deposits; Pliocene; Pleistocene; Rhodes; Greece; Mediterranean Sea

* Corresponding author.
E-mail addresses: Jean-Jacques.Cornee@univ-lyon1.fr (J.-J. Cornée), pierre.moissette@univ-lyon1.fr (P. Moissette), sebastien.joannin@univ-lyon1.fr (S. Joannin), jean-pierre.suc@univ-lyon1.fr (J.-P. Suc), frederic.quillevere@univ-lyon1.fr (F. Quillévéré), krijgsma@geo.uu.nl (W. Krijgsman), fhilgen@geo.uu.nl (F. Hilgen), ekosker@GEOL.UOA.GR (E. Koskeridou), philippe.munch@up.univ-mrs.fr (P. Münch), christophe.lecuyer@univ-lyon1.fr (C. Lécuyer), paula.desvignes@univ-lyon1.fr (P. Desvignes).

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1. Introduction

Coastal sediments deposited in convergent plate margins record both tectonic and glacio-eustatic sea-level changes. Deciphering these records is critical to understanding the evolution of active tectonic settings where the sedimentary record is often incomplete.

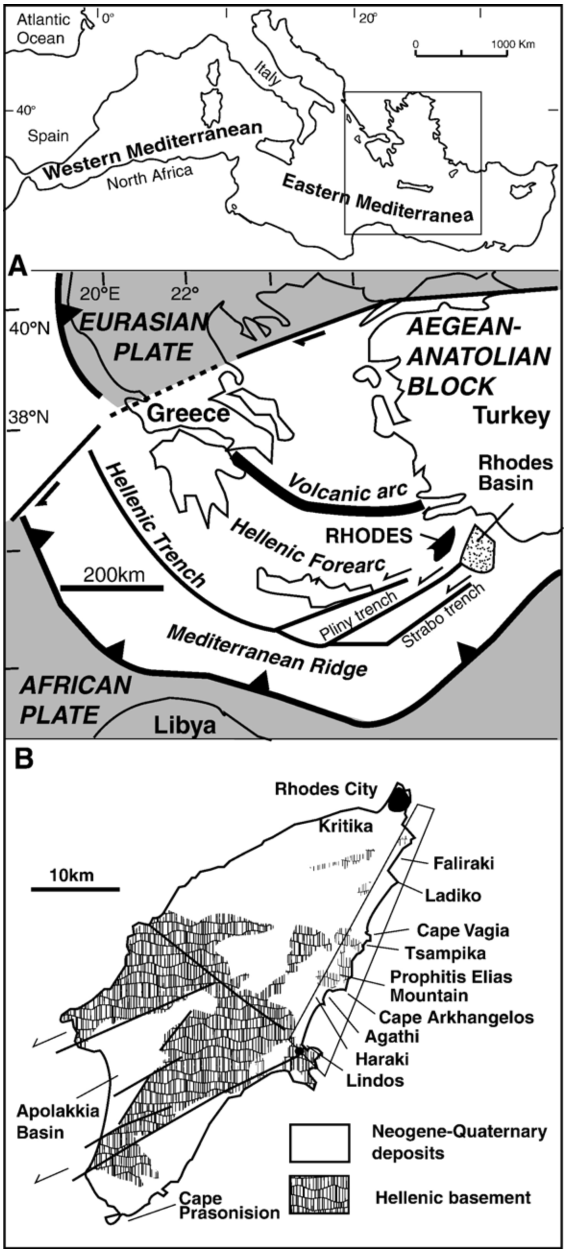

Fig. 1. (A) Location map of the island of Rhodes in the eastern Mediterranean. Modified from ten Veen and Kleinspehn (2002). (B) Location of the investigated sites in Rhodes. Geological data from Mutti et al. (1970) and ten Veen and Kleinspehn (2002), modified.

In northeastern Rhodes, Greece, Pliocene–Pleistocene basinal to littoral fossiliferous deposits are well exposed in steep palaeovalleys that transect the Hellenic Mesozoic basement. The island of Rhodes is located in the easternmost part of the Hellenic fore-arc, which resulted from the Cainozoic subduction of the African Plate under the Aegean–Anatolian Block (Fig. 1). The Aegean Sea collapsed during the Pliocene and Rhodes was exposed to syndepositional transtensional tectonics controlled by N70° trending sinistral strike-slip faults, because of the increasing curvature of the plate boundary. Vertical movements are still active today (e.g. Pirazzoli et al., 1982, 1989; Flemming and Woodworth, 1988).

The northeastern part of Rhodes consists of transgressive Pliocene–Pleistocene sediments resting upon a deformed and deeply eroded, mainly carbonate, Mesozoic basement (Mutti et al., 1970; Lekkas et al., 2000). This basement was repeatedly faulted in various directions, resulting in the formation of steep horsts and grabens that later controlled the nature and distribution of the Pliocene–Pleistocene deposits (Hanken et al., 1996). In such a sedimentary setting, facies changes are common and correlations between separated graben infills are often difficult to establish.

We propose in this paper a revised tectono-stratigraphic scheme for the Late Pliocene–Middle Pleistocene deposits of northeastern Rhodes (Fig. 2), combining data from field investigations, sedimentology, chronostratigraphy, palaeoecology and palynology. This integrated study includes: correlations between formations; geometry of sedimentary deposits; definition of a new formation (Ladiko–Tsampika) in which climatic cycles are recognized; identification of tectonic events (vertical motions and gravity slides); a sequence stratigraphy, and the tectonic and sedimentary evolution of Rhodes within the geodynamic context of the Aegean fore-arc.

2. Geological setting and stratigraphy

Following the general stratigraphic studies of Mutti et al. (1970) and Meulenkamp et al. (1972), an innovative and detailed study was conducted by Hanken et al. (1996), who retained the Damatria and Kritika Formations and defined two new lithostratigraphic units, the Rhodes and Lindos Acropolis Formations, each comprising several different facies group (Fig. 2). The same general lithostratigraphic framework is used throughout this paper. However, the Kritika Formation is now considered as a Member within the Rhodes Formation and the Ladiko sand is part of the newly defined Ladiko–Tsampika Formation.

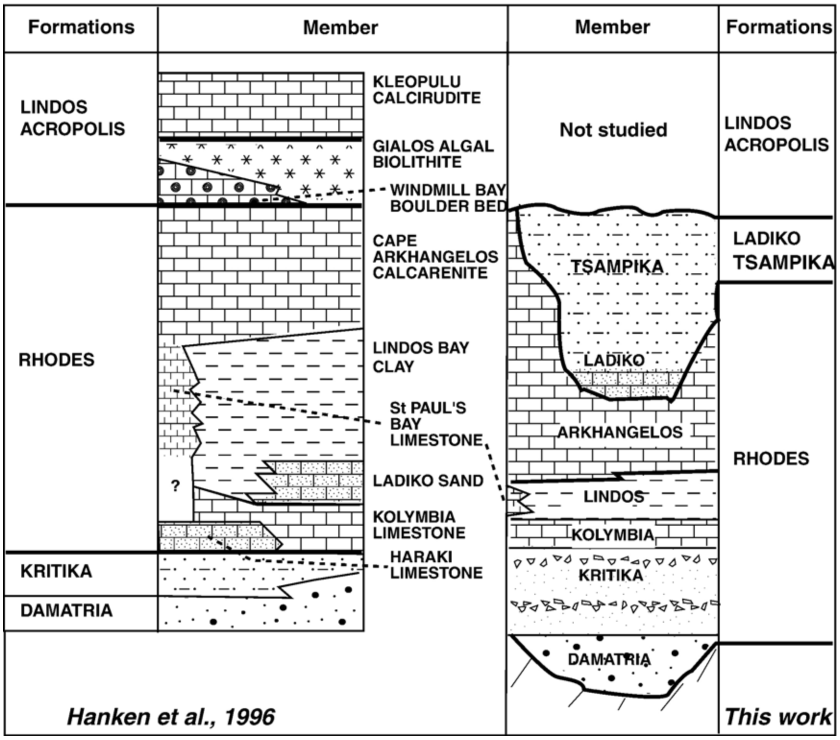

Fig. 2. Pliocene–Pleistocene lithostratigraphy of northeastern Rhodes. Left: after Hanken et al. (1996); right: this study.

2.1. Damatria Formation

This fluvio-lacustrine unit consists of conglomerates, sandstones and siltstones, with local intercalations of clays (Mutti et al., 1970; Meulenkamp et al., 1972; Willmann, 1981). It gradually passes upwards into the mainly marine Kritika Formation (Meulenkamp et al., 1972). The Damatria Formation was referred to the Upper Pliocene by Benda et al. (1977).

2.2. Kritika Formation

The best exposures of this formation occur in the northern and northeastern parts of Rhodes with mostly siliciclastic facies: conglomerates, sandstones, siltstones, clays, and a few lignite beds. The estimated thickness is 180 m (Mutti et al., 1970). Depositional environments were brackish to shallow marine (Keraudren, 1970; Broekman, 1972; Meulenkamp et al., 1972; Hanken et al., 1996; Hajjaji et al., 1998; Benali-Baitich, 2003). The age of the formation is debated. It has been considered as Late Pliocene on the basis of ostracode (Sissingh, 1972) and sporomorph associations (Benda et al., 1977); however, Drivaliari (1993) established that these have no reliable chronological significance. On the basis of calcareous nannofossils, foraminifers and ostracods, Thomsen et al. (2001) established an Early Pleistocene age for its marine strata.

The Kritika Formation (now a Member within the Rhodes Formation) was deposited in palaeovalleys that transect either the continental siliciclastic Damatria Formation, or the Mesozoic basement (Mutti et al., 1970). Some authors have assumed that the top of the Kritika Formation was eroded before deposition of the Rhodes Formation (Hanken et al., 1996; Hansen, 1999); however, this was not documented by Thomsen et al. (2001) in the Faliraki section that we also studied.

2.3. Rhodes Formation

This formation comprises six facies groups (Beckman, 1995; Hanken et al., 1996) from base to top. The first one is placed at the base of the second, four are retained as Members of the Rhodes Formation, and one is now a Member of the newly defined Ladiko–Tsampika Formation.

(1) *Haraki limestone*. A patchily developed 3 m thick subtidal red algal and bioclastic bed, karstified at the top. Its age is probably Late Pliocene. According to Hansen (1999), the Haraki limestone

should be considered as part of the Kolymbia limestone.

(2) *Kolymbia limestone*. A 4 to 20 m thick fossiliferous bioclastic limestone that rests unconformably on the bioeroded Mesozoic basement, on the Kritika Formation, or on the Haraki limestone. The depositional environment varies from shoreface at the base to lower offshore in its uppermost part (Beckman, 1995; Hanken et al., 1996; Spjeldnaes and Moissette, 1997; Steinthorsdottir, 2002). Palaeomagnetic studies by Løvlie et al. (1989) indicate a Late Pliocene age.

(3) *Lindos Bay clay*. About 30 m of blue-grey calcareous to silty clays with foraminifers, deep-sea corals, small bivalves, pteropods, scaphopods, bryozoans, and ostracodes, in which the faunal associations record first a pronounced upward-deepening, with an estimated maximum depth of deposition of about 400–600 m, followed by upwards-shallowing (Moissette and Spjeldnaes, 1995; Hanken et al., 1996). Meulenkamp et al. (1972) mentioned the local occurrence of *Hyalinea balthica* and *Globorotalia truncatulinoides* in the Lindos Bay clay and assigned it to the Pleistocene. Magnetostratigraphic studies by Løvlie et al. (1989) suggest an age of 3 Ma at the base and of around 0.7 Ma at the top of the sequence. These authors placed the Pliocene/Pleistocene boundary 5 m above the base of the Lindos Bay clay. The calibration of the palaeomagnetic interpretation of Løvlie et al. (1989) was based on: (1) the presence in the lowermost part of the clays of a thin reddish bed interpreted as a volcanic ash, which was correlated with the Pliocene volcanism of central Aegean; and (2) the first occurrence of the Pleistocene benthic foraminifer *H. balthica* found at 5.38 m above the ash bed. Thomsen et al. (2001) proposed an Early Pleistocene age (from 1.6 to 1.0 Ma) for the Lindos Bay clay in the Faliraki section.

(4) *St. Paul's Bay limestone*. Deep-sea coral boundstones that encrust the flanks of palaeovalleys, which were filled with the Lindos Bay clay, are regarded as a condensed lateral facies of the clay (Hanken et al., 1996).

(5) *Ladiko sand*. About 30 m of alternating sands, sandy clays and silts, with frequent invertebrate accumulations, the Ladiko sand occurs only in the Ladiko area. Hanken et al. (1996) considered this facies as an atypical near-shore lateral equivalent of the lower part of the Lindos Bay clay. Broekman (1972, 1974) regarded these deposits as part of the Kritika Formation. The rare occurrence of *Globorotalia inflata* suggests an age not older than Late Pliocene (Broekman, 1972, 1974).

(6) *Cape Arkhangelos calcarenite*. Mollusc–bryozoan-rich grainstones and packstones form shelf-down-stepping clinoforms reaching a total thickness of up to 20–30 m (Beckman, 1995; Hanken et al., 1996; Hansen, 1999). The Cape Arkhangelos calcarenite is truncated by erosion in its uppermost part. Its base erodes the Lindos Bay clay or locally interfingers with its uppermost part. This Member was formed during a relative sea-level fall. According to Hanken et al. (1996), the Cape Arkhangelos calcarenite was deposited during the Pleistocene.

2.4. Lindos Acropolis Formation

This formation comprises three facies groups (now Members) deposited in littoral environments (Hanken et al., 1996). At the base is an erosional surface.

(1) *Windmill Bay boulder bed*. Blocks of eroded Cape Arkhangelos calcarenite coated with microbial and bryozoan crusts.

(2) *Gialos algal biolithite*. A 3 m thick red algal boundstone intergrown with microbialite rests on the Windmill Bay boulder bed.

(3) *Kleopulu calcirudite*. A patchily developed 12 m thick algal bioclastic limestone with pebbles derived from both the basement and Pleistocene deposits.

3. Materials and methods

This study is based on field, sedimentological and palaeontological studies of several stratigraphic sections on Rhodes. As no detailed map of Rhodes is available, we used GPS measurements for locations and elevations. More than three hundred rock samples were collected from the outcrops for palaeoecological and palynological analysis. Polished slabs and thin sections of some indurated rocks were made.

Parts of the palaeontological results are from unpublished reports or MSc theses: pollens (Kouli, 1993; Buisine, 2000; Joannin, 2003), bryozoans (Steinthorsdottir, 2002; Maillet, 2003; Lavoyer, 2004), and ostracodes (Benali-Baitich, 2003). Data from an unpublished field guidebook on the geology of Rhodes were also used (Ferry et al., 2001).

Molluscs and bryozoans were studied in the Kritika Member and foraminifers and bryozoans in the Lindos Bay clay. Foraminifers, molluscs, bryozoans, and ostracodes were studied in the Ladiko–Tsampika Formation.

Planktonic foraminifer stratigraphy is based on semi-quantitative analysis of size fractions coarser than 125 μm. About 1500 specimens were picked under a binocular microscope and identified. Taxonomic concepts follow the Atlas of Kennett and Srinivasan (1983).

A total of 116 samples were collected for palynological analysis from sections in the Tsampika area, i.e. about one sample every metre, excepted from the sandy intervals (Fig. 8); 87 samples provided enough pollen grains for analysis. Samples were prepared using a standard chemical technique adapted from Cour (1974). A minimum of 150 pollen grains (apart from those of *Pinus*) were counted and at least 20 taxa were thus identified in each sample. Heusser and Balsam (1977) have shown that the proportion of *Pinus* pollen grains, because of their high buoyancy, is controlled mostly by the distance from river mouths. Their relative amount is consequently low in proximal areas and higher in distal positions (Heusser and Balsam, 1977; Beaudouin, 2003). Percentages of *Pinus* grains are thus used as a proxy for sea-level variations. A simplification of the pollen record was necessary, considering the high floristic diversity. A "pollen index" was consequently developed in order to discriminate between tectonic and climatic controls upon sedimentation. The following ratio was used: [mesothermic + mid-altitude + high-altitude elements] / [herbaceous elements, including steppic ones]. Often over-represented, pollens of *Pinus* and Cupressaceae are excluded. This ratio is used here as an indicator of variations: (1) in moisture (humid episodes, with prevalent *Quercus* and *Cedrus*, versus xeric periods characterized by Asteraceae, including *Artemisia*), and (2), at a lesser level, in temperature (warm- to cool-temperate and humid phases opposed to cooler and more xeric ones).

A total of 41 oriented blocks were sampled from the Vagia section, and 40 hand samples were taken from the Tsampika section, for magnetostratigraphic purposes. Samples from Vagia were drilled in the laboratory and the cores were stepwise thermally demagnetised using standard palaeomagnetic procedures. Samples from the Tsampika section were usually very sandy and drilling in the laboratory was destroying too many samples, so most Tsampika samples were cut into cubes that fit in standard sized plastic holders, which were subsequently stepwise demagnetised using alternating fields. The polarity of the samples was determined by visual inspection of the Zijderveld diagrams. In both sections normal and reversed

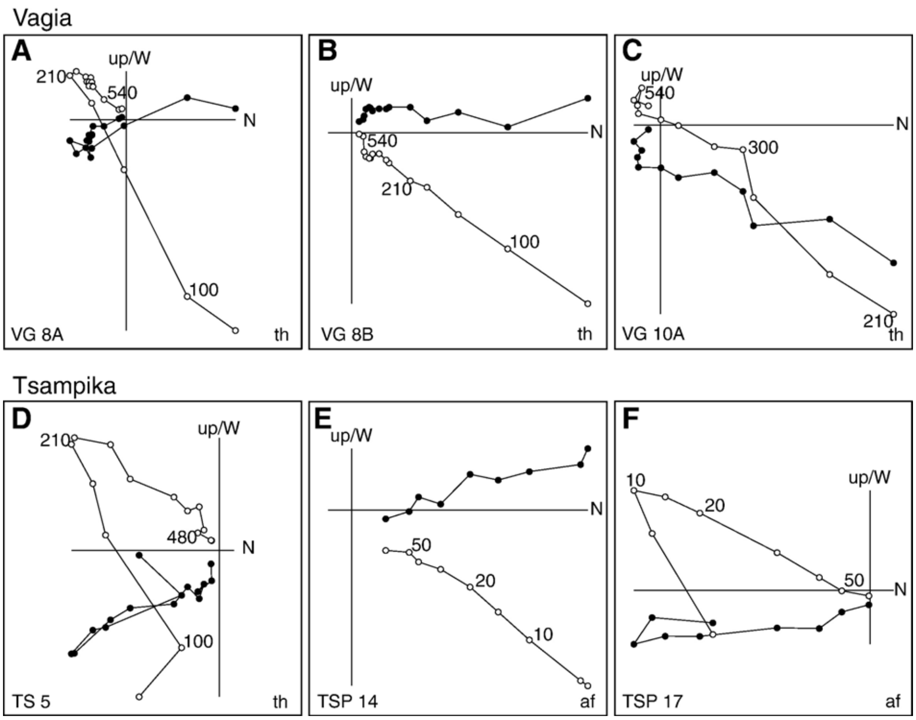

Fig. 3. Orthogonal projections of stepwise demagnetisation diagrams corrected for bedding tilt from the Vagia (A–C) and Tsampika (D–F) sections. In the lower right-hand corner "th" refers to thermal, and "af" to alternating field demagnetisation diagrams. Closed (open) symbols represent the projection of the NRM vector endpoint on the horizontal (vertical) plane. Values denote demagnetisation steps in °C (th) or mT (af).

polarities were observed (Fig. 3), but whether the normal polarities were primary could not in all cases be unambiguously determined, because the sedimentary strata were only slightly tilted.

Amphiboles were separated from a volcanic sand layer found in the Lindos Bay clay of the Haraki section (Fig. 1B). $^{40}Ar/^{39}Ar$ age determinations were performed at the University of Nice-Sophia-Antipolis. The amphibole bulk samples were step-heated in a high frequency furnace following the experimental procedure described by Féraud et al. (1982). The bulk sample analyses were performed with a mass spectrometer composed of a 120° M.A.S.S.E. tube, a Baur-Signer GS 98 source and a Balzers electron multiplier. Plateau ages were defined by at least three successive steps showing consistency within a 2σ error confidence interval and carrying at least 70% of the ^{39}Ar released.

4. Results

4.1. Rhodes Formation

4.1.1. Kritika Member

Apart from in the Kritika area (near Rhodes City) where it was originally defined (Broekman, 1972, 1973), the best-exposed examples of this formation are in the Faliraki area (Fig. 1B) where it was deposited in large embayments open to the east (Mutti et al., 1970). Even though the siliciclastic Kritika Member is poorly fossiliferous, some marine beds contain abundant rhodoliths, foraminifers, solitary corals, molluscs, bryozoans and ostracodes, indicating beach to lower shoreface depositional environments (Hajjaji et al., 1998; Benali-Baitich, 2003; Maillet, 2003). Minor influxes of freshwater organisms are also recorded (Benali-Baitich, 2003; Maillet, 2003). Carbonate facies are scarce and principally consist of communities of skeletal organisms developed on drowned beachrocks: red algae, hermatypic corals (*Cladocora caespitosa*), serpulids, bivalves (oysters, spondylids), and bryozoans (Ferry et al., 2001). Freshwater to brackish characean and dasycladacean algae, benthic foraminifers, molluscs, and ostracodes occur in some beds and are particularly common at the top of the formation. The Kritika Member is organized into transgressive–regressive sedimentary cycles bounded by beachrocks, prograding beach conglomerates, or lower shoreface deposits (Ferry et al., 2001). The duration of the sedimentary cycles cannot be estimated because of a lack of precise dating. Most authors consider that the upper limit of the Kritika deposits is a major erosion surface (see discussion below).

4.1.2. Cape Vagia section

The classical Vagia section was studied by Løvlie et al. (1989), Frydas (1994), Moissette and Spjeldnaes (1995), Hanken et al. (1996), Kovacs and Spjeldnaes (1999), and Steinthorsdottir (2002). Løvlie et al. (1989) placed the Pliocene–Pleistocene boundary at around 5 m above the base of the Lindos Bay clay. This result was based on magnetic polarity stratigraphy and the first occurrence of *H. baltica* above a volcanic ash layer.

Despite a careful scrutiny of the sedimentary succession at Cape Vagia, no volcanic ash bed was found within the clays, but a thin red oxidation horizon in the sediment was revealed by examination of thin sections (Fig. 4A). The Pleistocene foraminifer, *H. balthica*, was not encountered by either Frydas (1994) or us. Throughout the section *Glorotalia inflata* is ubiquitous, indicating an age younger than 2.09 Ma (Lourens et al., 1992; Berggren et al., 1995); as is *Bulimina marginata*, younger than 2.2 Ma. Dextrally coiled assemblages of *Neogloboquadrina pachyderma* are present, but no sample yielded a significant number of sinistrally coiled specimens, which would imply an age younger than 1.8 Ma (Berggren et al., 1995). Our magnetostratigraphic results show predominantly reversed polarities and a clear interval of normal polarity (about 2–3 m thick), located directly above the red layer (Figs. 3, 4A). The normal directions are antipodal to the reversed directions and both indicate a counter-clockwise rotation. This agrees with earlier palaeomagnetic observations on Rhodes (Duermeijer et al., 2000), implying that the magnetic signal is primary. The normal polarity interval could represent the Olduvai (C2n) subchron, since we did not observe *H. baltica* in the sequence. However, if the record of this benthic foraminifer species by Løvlie et al. (1989) is correct, correlations with the Cobb Mountain (C1r.2n) or Jaramillo (C1r.1n) subchron are more likely. We also observed normal polarities in the uppermost part of the Vagia section, but these directions may not be primary as there are no counter-clockwise directions; so they may be due to secondary (weathering) processes. Consequently, we conclude that the Olduvai subchron (1.942–1.785 Ma) is recorded here, in accordance with the latest Pliocene foraminifer association (Lourens et al., 1996) and the occurrence of reversed polarity zones in the overlying Ladiko–Tsampika Formation (see Section 4.2.2.).

4.1.3. Faliraki section

The Faliraki section is exposed along the road adjacent to Faliraki Beach (Ferry et al., 2001;

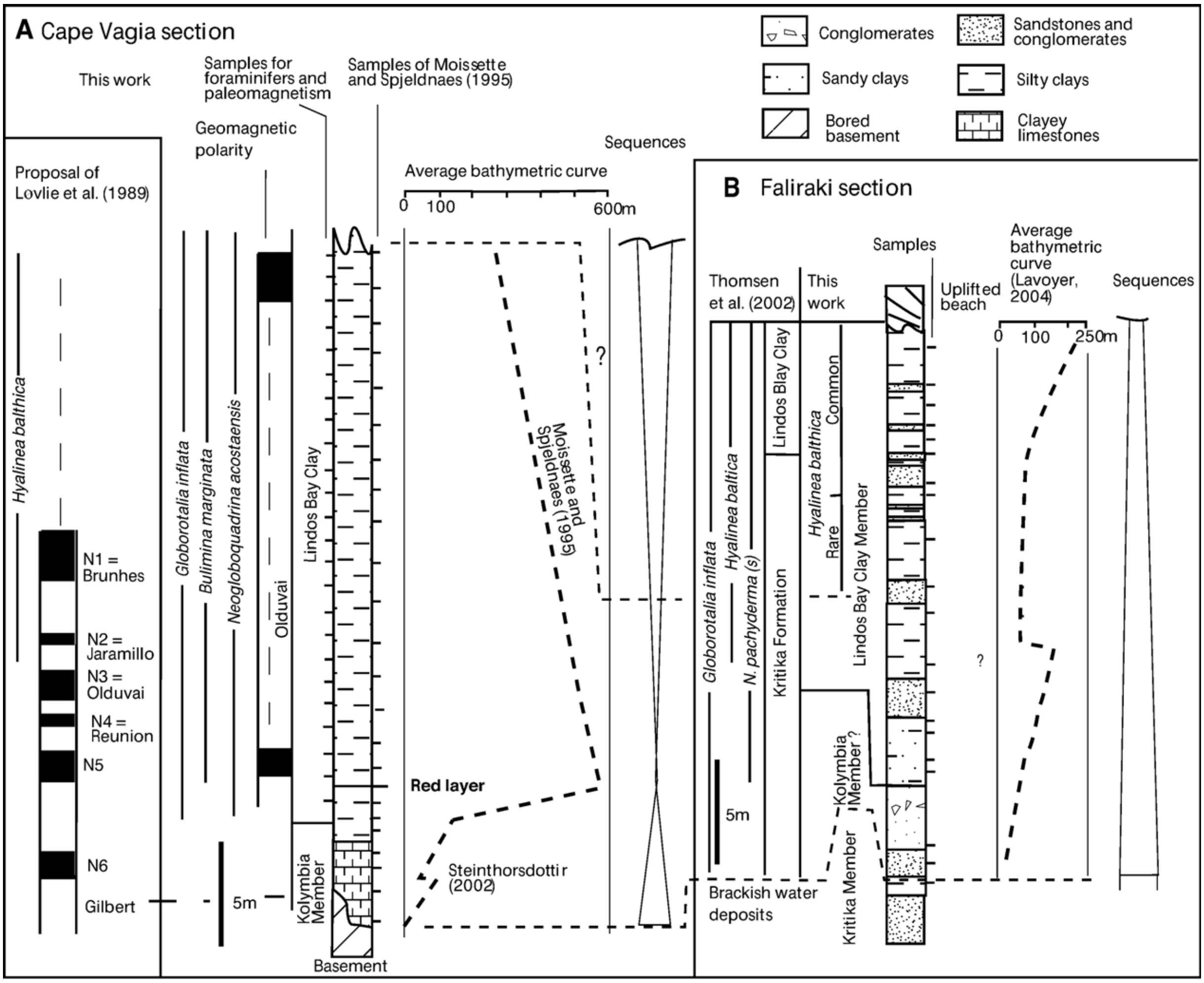

Fig. 4. Lithostratigraphy, biostratigraphy and magnetostratigraphy of Cape Vagia (A) and Faliraki (B) sections. Average bathymetric curves are shown for both sections.

Thomsen et al., 2001; Lavoyer, 2004). At the base are brackish clay and sand beds typical of the Kritika Member (Fig. 4B). They are overlain by 25 m of marine sediments capped by erosive shoreface deposits. The lowermost marine sediments are composed of sandstones and conglomerates with benthic foraminifers, bivalves (pectinids, oysters), bryozoans, and echinoids. Above, clays or silty clays alternate with more sandy and gravely beds containing a rich fauna of benthic and planktonic foraminifers, solitary corals, bivalves (mostly thin-shelled pectinids: *Amussium*, *Chlamys*, *Palliolum*, *Lissopecten*), gastropods (*Gibbula*, *Jujubinus*, Rissoidae), serpulids (especially *Ditrupa arietina*), bryozoans, ostracodes, and echinoids. A number of samples also yielded red algae, scaphopods, brachiopods, and barnacles. Some clayey laminated layers contain a few microfossils, fish remains, and well-preserved continental plant fragments. The uppermost part of the section is composed of the Lindos Bay clay Member, with abundant planktonic foraminifers, pteropods and bryozoans. Based on the analysis of bryozoan colonial morphotypes and the occurrence of various extant species of bryozoans, a palaeobathymetric curve was proposed (Lavoyer, 2004). This curve (Fig. 4B) indicates a progressive upward-deepening from the shoreface to the offshore with a maximum water depth of about 200 m in the uppermost part of the section.

Thomsen et al. (2001) found the Pleistocene foraminifer *Hyalinea balthica* approximately 5 m above the base of the marine part of the section. Our observations show that the first common occurrence of *H. balthica* is located about 14 m above the base of the marine part (Fig. 4B). Underneath, *H. balthica* is extremely rare. Even if the age of the first appearance datum of *H. balthica* is not precisely known, its first common occurrence was estimated at around 1.4–1.5 Ma (Løvlie et al., 1989; Lourens et al., 1998). Since some 8 m of sediments are preserved above the first common occurrence of *H. balthica*, an age of about 1.4 Ma is proposed for the top of the section.

4.1.4. Lindos Bay section

The Lindos Bay section (Beckman, 1995; Moissette and Spjeldnaes, 1995; Hanken et al., 1996) is composed of 28 m of offshore marls. The study of the bryozoan associations (Moissette and Spjeldnaes, 1995) indicates depositional depths ranging from 300 to 600 m in the lower part of the section, to 200–300 m near the top of the section.

Abundant, diversified and well-preserved planktonic foraminifers occur in all thirteen samples from the Lindos section. The assemblages are dominated by species of *Globigerinoides*, associated with *Globigerina bulloides*, *Globigerinella obesa*, and *Neogloboquadrina pachyderma*. *Globorotalia inflata* is common throughout the section suggesting and age younger than 2.09 Ma (Berggren et al., 1995). Although the benthic foraminifer *H. balthica* has not been found in all samples, its common occurrence at the base of the section points to a Pleistocene age.

4.1.5. Haraki section

This small section outcrops along the track that connects the main road to Agathi Beach (Fig. 1B). It is composed of about 5.5 m of Lindos Bay clay containing a volcanic sand layer in its lower middle part.

All samples yielded abundant and diversified planktonic foraminifers. The occurrence of *G. inflata* in the last 1.5 m of the section points to an age younger than 2.09 Ma (Berggren et al., 1995). Typical *Globigerinoides extremus*, found in all samples, indicate an age older than 1.77 Ma (Berggren et al., 1995). *H. balthica* was not discovered in any of the 10 samples. Likewise, none contains a significant number of sinistrally coiled specimens of *N. pachyderma*, which would have indicated an age younger than 1.80 Ma. The $^{40}Ar/^{39}Ar$ analyses performed on amphiboles extracted from the Haraki volcanic sand yielded two identical plateau ages of 2.06 ± 0.17 and 2.06 ± 0.14 Ma, which are in good accordance with the biostratigraphic results. These findings indicate that the section (at least at its top) corresponds to the upper part of planktonic foraminiferal Zone PL6, and represents the interval 2.09–1.80 Ma (Late Pliocene, Olduvai subchron or top of Chron C2r.1r). These results also corroborate the presence of the Olduvai subchron in the Vagia section.

4.1.6. Prophitis Elias area: the Cape Arkhangelos calcarenite

This Member was defined in the Cape Arkhangelos area (Hanken et al., 1996) and is documented also near Kalithea (Hansen, 1999, 2001). This calcarenite occurs at elevations of up to about 70 m above present-day sea level and was interpreted as representing a highstand at the end of a major forced regression (Hanken et al., 1996). Several abrasion platforms were formed on basement limestones during sea-level stillstands. In the Cape Arkhangelos area, the calcarenitic beds rapidly pass into offshore marls containing thin-shelled pectinids (*Amussium*).

We investigated the Prophitis Elias Mountain and surrounding areas in order to get a better understanding of this forced regression. The southern side of the mountain, down to Agathi Beach, shows well-preserved abrasion platforms between sea level and +519 m (Fig. 5A). These platforms often display submarine bioerosional structures (clionid sponges and lithophagid bivalves), or littoral notches infilled with bored basement pebbles. The abrasion platforms transect all structural features of the basement. The tectonic events leading to the development of north–south trending faults took place before deposition of the Rhodes Formation. Remnants of the Cape Arkhangelos calcarenite (with abundant coralline red algae and bryozoans) are found from elevations of about +474 m down to the present-day shoreline (Fig. 5B–C). This implies that numerous calcarenite horizons were in fact formed during the forced regression. Twenty abrasion platforms were thus identified (Fig. 5A and D). They are organized in an irregular pattern with differences in elevation between two superimposed platforms ranging from 12 to 52 m (Fig. 5D). Similar marine terraces have been described in Crete by Peters et al. (1985), who interpreted them as the result of Late Neogene to Quaternary vertical movements. As in Rhodes, the abrasion platforms can now be interpreted as resulting from obliquity-linked glacial cycles. This would mean that the Lindos Bay clay and the Cape Arkhangelos calcarenite are coeval. Such a hypothesis is however highly improbable since field observations indicate that these two Members are superimposed and separated by a major erosion surface (Hanken et al., 1996; Hansen, 2001; and our personal observations).

Stillstands are separated by relative sea-level falls during which the previously deposited calcarenite horizon suffered erosion (Fig. 5E–F). Development

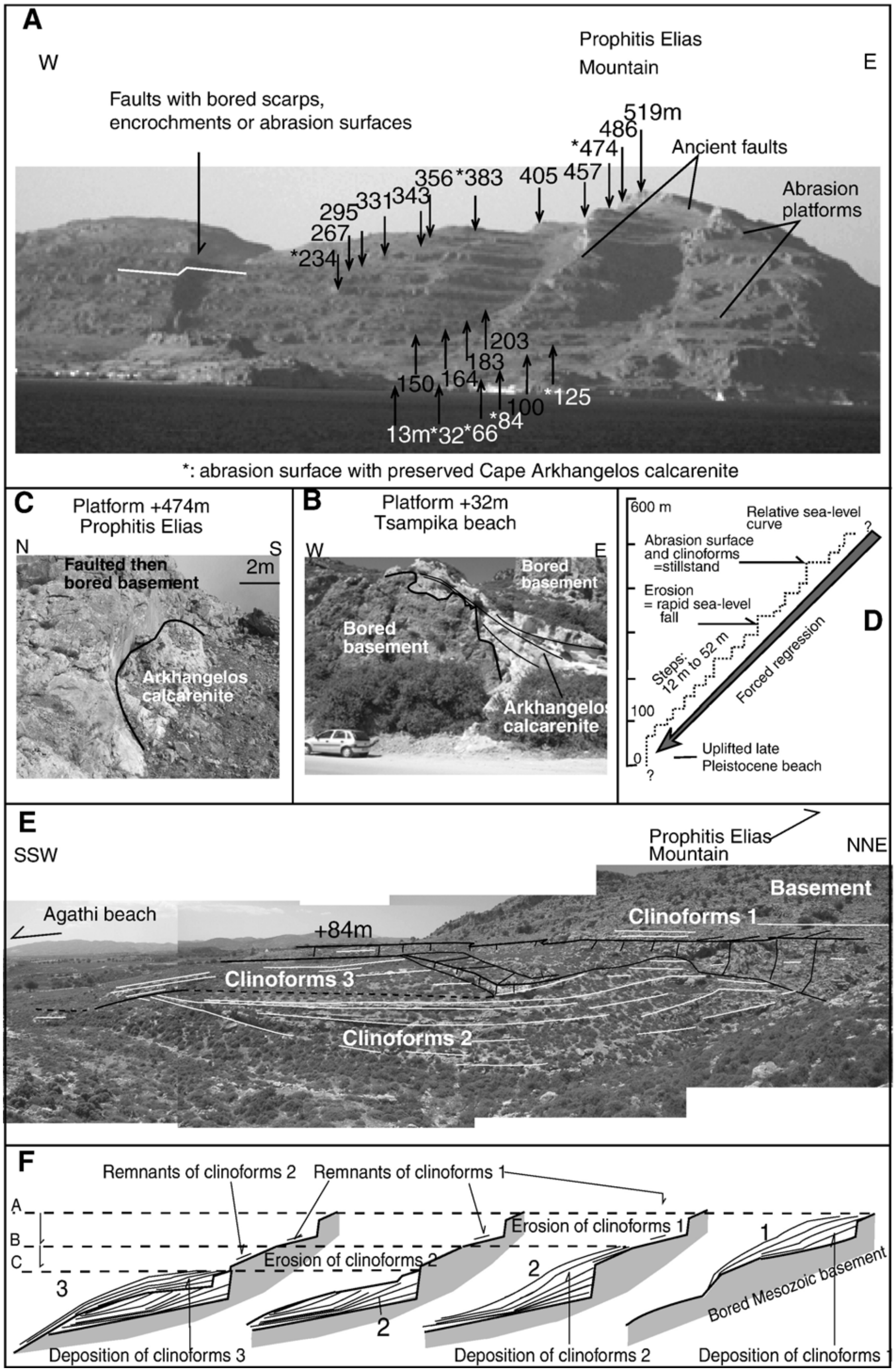

Fig. 5. Prophitis Elias area. (A) Photograph of abrasion platforms on basement limestones. (B) and (C) Field views of the Cape Arkhangelos calcarenite at different elevations. (D) Sketch showing the forced regression during deposition of the Cape Arkhangelos calcarenite. (E) Field view of the eroded clinoforms of the Cape Arkhangelos calcarenite between Agathi Beach and Prophitis Elias Mountain. (F) Sketch showing multiple erosive events linked to the Cape Arkhangelos calcarenite downstepping.

of the calcarenite was principally controlled by morphology of the palaeorelief. Steep flanks are thus characterized by a limited basinward extent of sediment and almost complete later erosion, whereas, where surfaces were sub-horizontal, abrasion platforms developed also on the previously deposited calcarenite

(Fig. 5E). The Cape Arkhangelos calcarenite was deposited after erosion of the underlying Lindos Bay clay, which is now either missing or only patchily preserved (e.g. in the southern part of Agathi Beach). The total sea-level fall cannot be fully estimated, but is at least 519 m, given that the lowermost visible Cape Arkhangelos calcarenite is situated at present-day sea level. This value substantiates the 400 to 600 m depth of deposition proposed for the Lindos Bay clay Member (Moissette and Spjeldnaes, 1995; Hanken et al., 1996).

4.2. Ladiko–Tsampika Formation

Hanken et al. (1996) reported the presence in the Ladiko area of the Lindos Bay clay overlying the Ladiko sand, which consequently was regarded as a lateral equivalent of the lower part of the Lindos Bay clay (Fig. 6). In the Tsampika area, the same authors mentioned the Lindos Bay clay at an elevation of +160 m. Our investigations in the Ladiko and Tsampika areas show that the deposits of the Lindos Bay clay are missing. The greyish to greenish silty clays observed in these areas are different from the typical pelagic sediments of the Lindos Bay clay and have to be considered as a new lithostratigraphic unit, the Ladiko–Tsampika Formation.

4.2.1. Ladiko area

The Ladiko Member is composed of sandy clays, sandstones, and siltstones that were deposited in the vicinity of palaeocliffs (Ferry et al., 2001). These

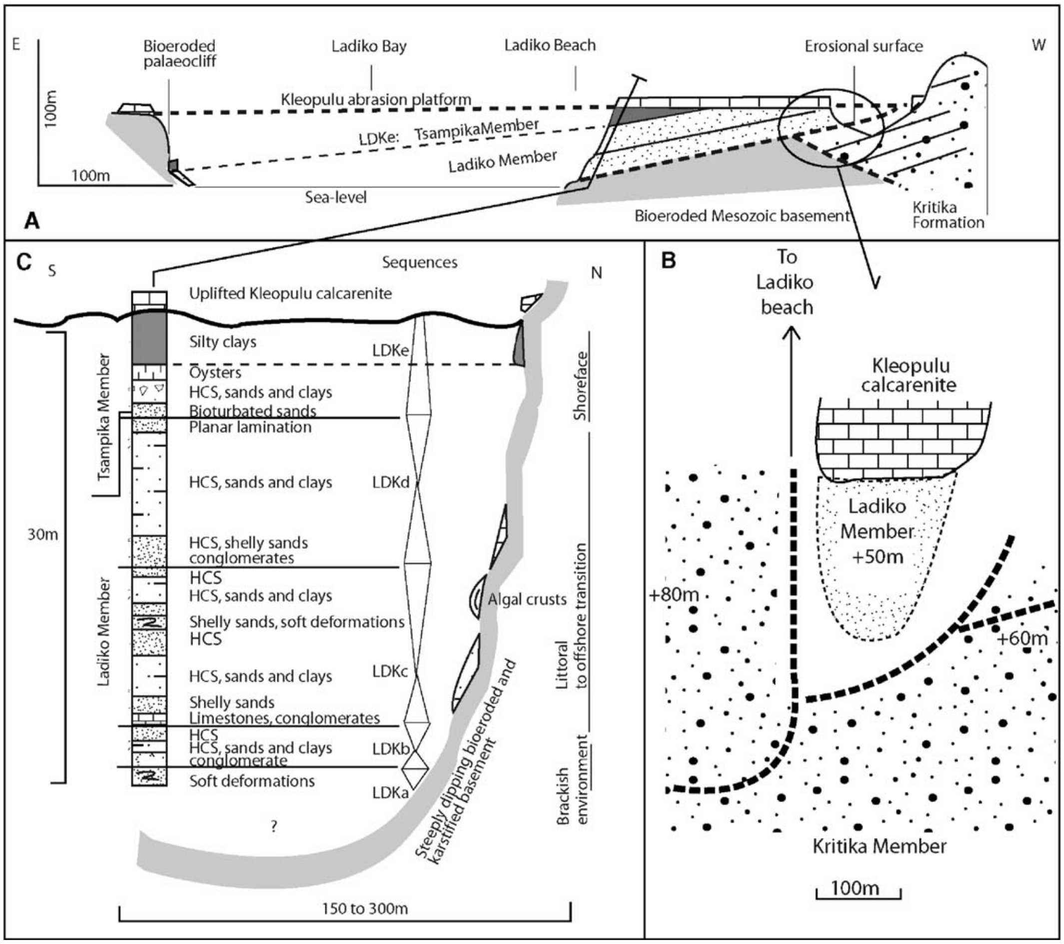

Fig. 6. Ladiko area. (A) E–W section showing the unconformity of the Ladiko sand Member on the Kritika–Rhodes Formation and on the basement. (B) Simplified map of the western part of the Ladiko sand deposits showing that they were deposited in an incised palaeovalley within the Kritika Member. (C) Lithostratigraphic column of the Ladiko sand near its northern palaeocliff. The Ladiko Member is organized in four sequences (Ferry et al., 2001).

sediments are different from those of the typical Kritika Member, which outcrop some hundreds of metres west of the Ladiko section and which are composed of coarse conglomeratic beach sequences. Hidden by recent deposits, the contact between the Ladiko and the Kritika Members cannot be directly observed; however mapping shows that, to the west, the Ladiko sediments rest on the Kritika deposits, which were partially eroded (Fig. 6A and B), and to the east they overlie the bioeroded limestones of the basement. The steep northern flank of the palaeovalley exhibits patchily preserved algal boundstones and calcarenites with hummocky cross-stratifications (Fig. 6C). The clastic deposits of the Ladiko area (Fig. 6C) are organized in at least four retrogradational–progradational beach sequences, as mentioned by Ferry et al. (2001).

The lowermost few beds are poorly fossiliferous and contain only tree leaves and scarce benthic foraminifers and ostracodes. They are provisionally interpreted as indicating fresh-water to brackish environmental settings.

The overlying deposits contain abundant storm-generated concentrations of skeletal marine littoral organisms of diverse origins. Essentially in situ soft-bottom components are represented by benthic foraminifers (mainly *Elphidium*), solitary corals, gastropods (*Bittium*, *Natica*), bivalves (nuculids, pectinids, *Corbula*), serpulids (*Ditrupa arietina*), free-living bryozoans, ostracodes, and irregular echinoids. Allochthonous hard-substrate organisms were transported from nearby palaeocliffs: encrusting and articulated red algae, zooxanthellate corals (*C. caespitosa*), bivalves (*Anadara*, *Chlamys*, *Hiatella*, *Mytilus*, oysters, spondylids), gastropods (*Calliostoma*, *Conus*, *Gibbula*, *Murex*, *Patella*, Rissoidae), erect bryozoans, barnacles, and cidaroid echinoids. Hummocky cross-stratification and ripple-marks are common within the sandstones.

The uppermost part of the Ladiko succession is composed of some 5 m of greenish silty clays with rare macrofauna (oysters mostly). The microfauna is represented by rare to extremely rare benthic foraminifers (*Bolivina*, *Bulimina*, *Elphidium*), ostracodes, and bryozoans. These greenish silty clays are also found against the margins of the palaeovalley with oysters and spondylids. Planktonic foraminifers (globigerinids) are extremely rare in the entire Ladiko section.

The Ladiko Member is interpreted as a littoral and mostly high-energy deposit, formed in narrow embayments where skeletal remains of hard-substrate and soft-bottom organisms accumulated. The age of this unit is Pleistocene as it lies between the Cape Arkhangelos calcarenite and the Tsampika Member (see discussion below).

4.2.2. Tsampika area

The clayey sediments of Tsampika Bay (Figs. 7–11) were previously referred to the Pliocene (Mutti et al., 1970; Hanken et al., 1996). Our study shows that these deposits are preserved in a palaeovalley, above a major erosional surface. The steep sides of the palaeovalley are the result of ancient faults developed in the basement limestones, arranged in two major sets (Fig. 7A): NNW–SSE (eastern side of the palaeovalley) and WSW–ENE (southern side). Some of the NNW–SSE trending faults have been slightly reactivated, as they transect the Ladiko–Tsampika Formation. The Tsampika deposits rest on metre-sized remnants of the Kolymbia limestone (with a typical *Amussium* and bryozoan association), or the red algal–bryozoan Cape Arkhangelos calcarenite, as well as on the basement (Fig. 7A–B). Their relationship to the Cape Arkhangelos calcarenite establishes their Pleistocene age (Ferry et al., 2001). They cannot represent the Lindos Bay clay Member, as this was totally eroded in this area before deposition of the Ladiko–Tsampika Formation.

The Ladiko–Tsampika Formation comprises about 148 m of siliciclastic deposits (Fig. 8). Patchy outcrops were found against the basement at elevations up to +160, indicating that the uppermost part was eroded. Eight depositional sequences have been identified, namely TS1 to TS8 (Fig. 8). A ninth sequence (TS9) probably occurs, but is poorly exposed at the top of the North Tsampika section. These siliciclastic deposits were only identified in the Tsampika area, where the succession is fairly complete, and, as discussed below, at the top of the Ladiko section.

Sequence TS1 (= Ladiko Member) consists of fluvial deposits that grade upward into marine shoreface strata and finally into deltaic/littoral and low-salinity strata, with molluscs such as *Cerastoderma*, *Mactra*, *Martisia*, *Ostrea edulis*, and *Terebralia*. The TS1 sediments infill the deepest part of the palaeovalley and also rest against bioeroded palaeocliffs. The shoreface deposits resemble those of the Ladiko sand, especially in their fauna: coral fragments of *C. caespitosa*, a mollusc association with *Anadara*, *Calliostoma*, *Conus*, *Murex*, *Triphora*, *Truncculariopsis*, and rare planktonic foraminifers. Transported hard-substrate faunas are abundant in the vicinity of palaeocliffs and include *Cladocora*, *Mytilus*, oysters, and spondylids. Sequence TS1 is capped by a fluvial conglomerate composed of centimetre-sized well-rounded basement-derived calcareous pebbles.

Above TS1 are seven retrogradational–progradational sequences of the Tsampika Member, with conglomerates, sandstones and monotonous greenish silty clays. Typically, as at the top of the Ladiko

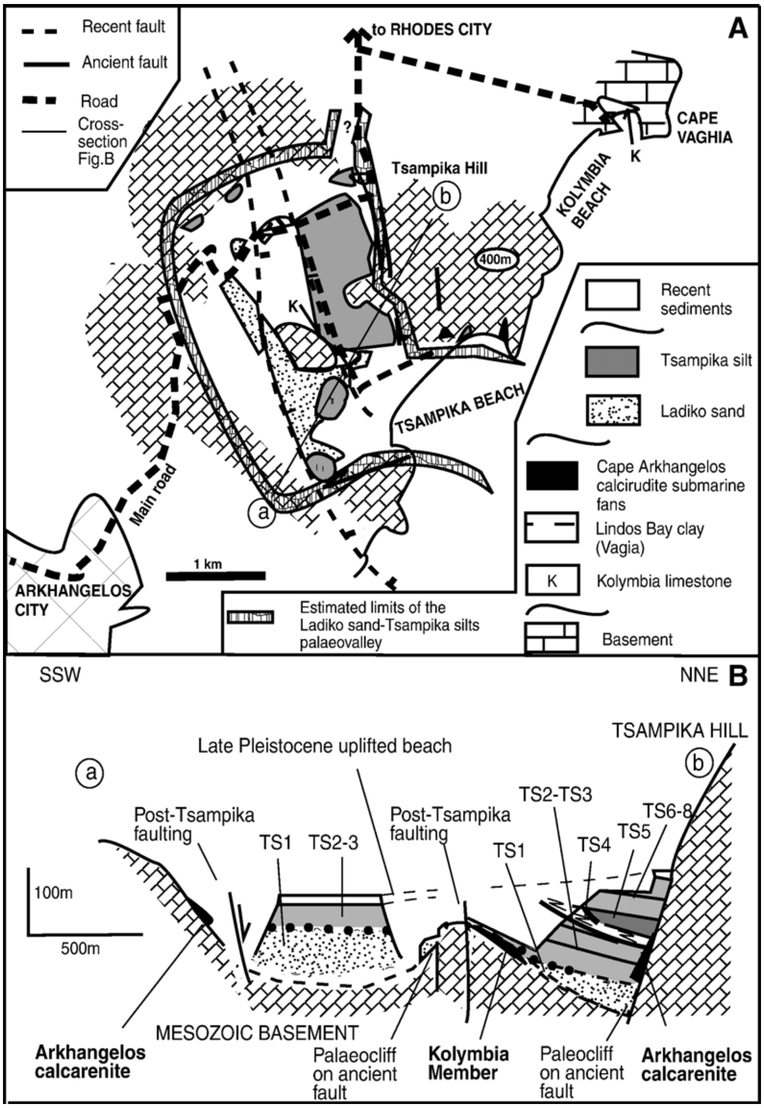

Fig. 7. Tsampika area. (A) Map of the Ladiko–Tsampika Formation (LTF). Note the deposition in a palaeovalley. (B) Cross-section in the Ladiko–Tsampika Formation. Note that the LTF rests on the basement, on remnants of the Kolymbia limestone, or on remnants of the Cape Arkhangelos calcarenite. The Lindos Bay clay is missing because of the pre-LTF erosion.

section, these sequences contain few macrofossils, except in two littoral sandy conglomerates of sequences TS2 and TS6 (Fig. 8). The laminated silty clays contain rare bivalves, especially *Corbula gibba*, and small gastropods (*Epitonium*, *Naticarius*). Sequence TS3 is characterized by mollusc associations indicative of strong turbidity and very stressed dysoxic environments (*C. gibba*, *Dentalium*, *Palliolum incomparabile*, *Parvicardium minimum*, etc.). Marine microfossils occur throughout sequences TS2 to TS8, including ostracodes, benthic (*Elphidium*, *Bulimina*) and planktonic foraminifers (globigerinids). The foliated silty clays represent offshore deposits, whereas the hummocky cross-stratified silty clays are interpreted as representing offshore transition settings. The sandy parts of the sequences often contain hummocky cross-stratification and ripple-marks. They are interpreted as offshore–shoreface transitional deposits. A number of sandy conglomeratic beds contain upper shoreface elements such as infaunal bivalves in life position (*Pinna* and *Panopea*: Hanken et al., 2001), hard-substrate molluscs (*Gibbula*, *Rissoidae*, *O. edulis*, spondylids) and trace fossils (*Ophiomorpha*). Fluvial conglomerates occur sporadically at the base of sequences TS2 and TS6.

Sequence TS6 was deposited over an erosional surface. The basal fluvial conglomerates and beach

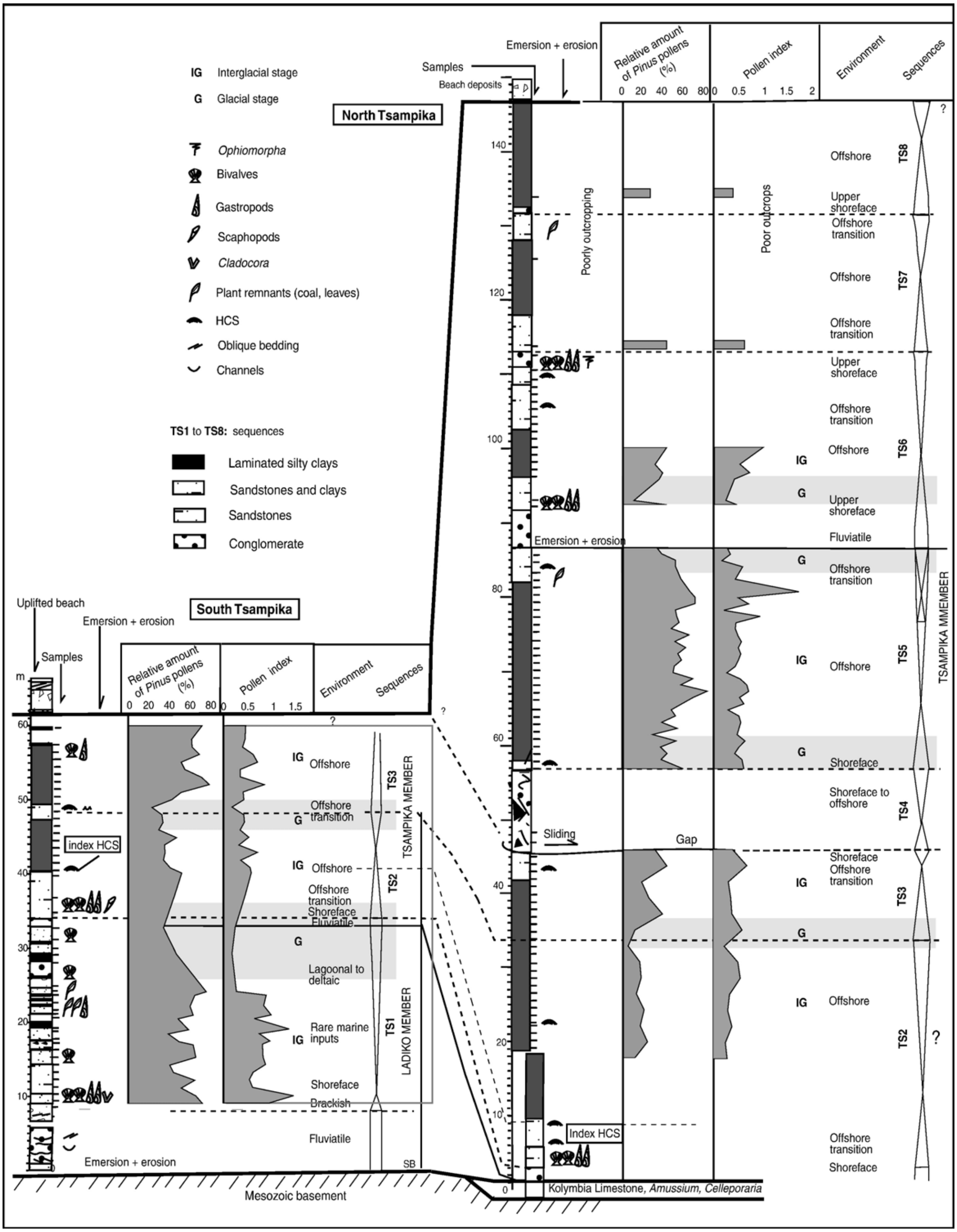

Fig. 8. Lithostratigraphic column of the Ladiko–Tsampika Formation in the Tsampika area with sedimentary sequences and simplified pollen record: (1) changes in *Pinus* amounts, and (2) "pollen index". Littoral sediments (low *Pinus* amounts and relatively high amounts of taxa requiring xeric conditions) are interpreted as deposited during glacial stages. Offshore deposits (high amounts of *Pinus* and taxa requiring warmer conditions) are interpreted as deposited during interglacial stages.

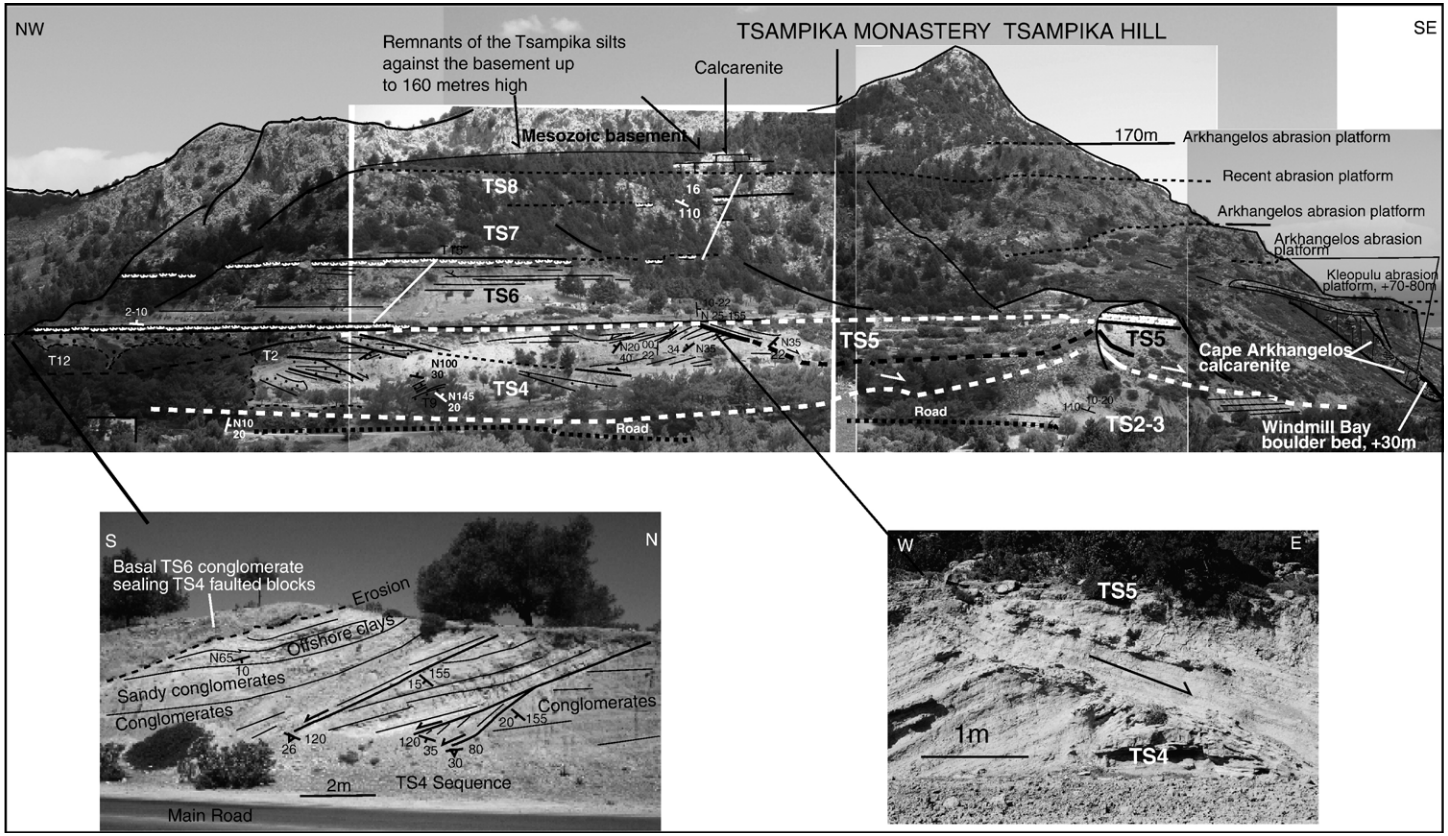

Fig. 9. Field view of the northwestern proximal part of the Ladiko–Tsampika Formation, Tsampika area, with details of the extensional deformations in sequence TS4.

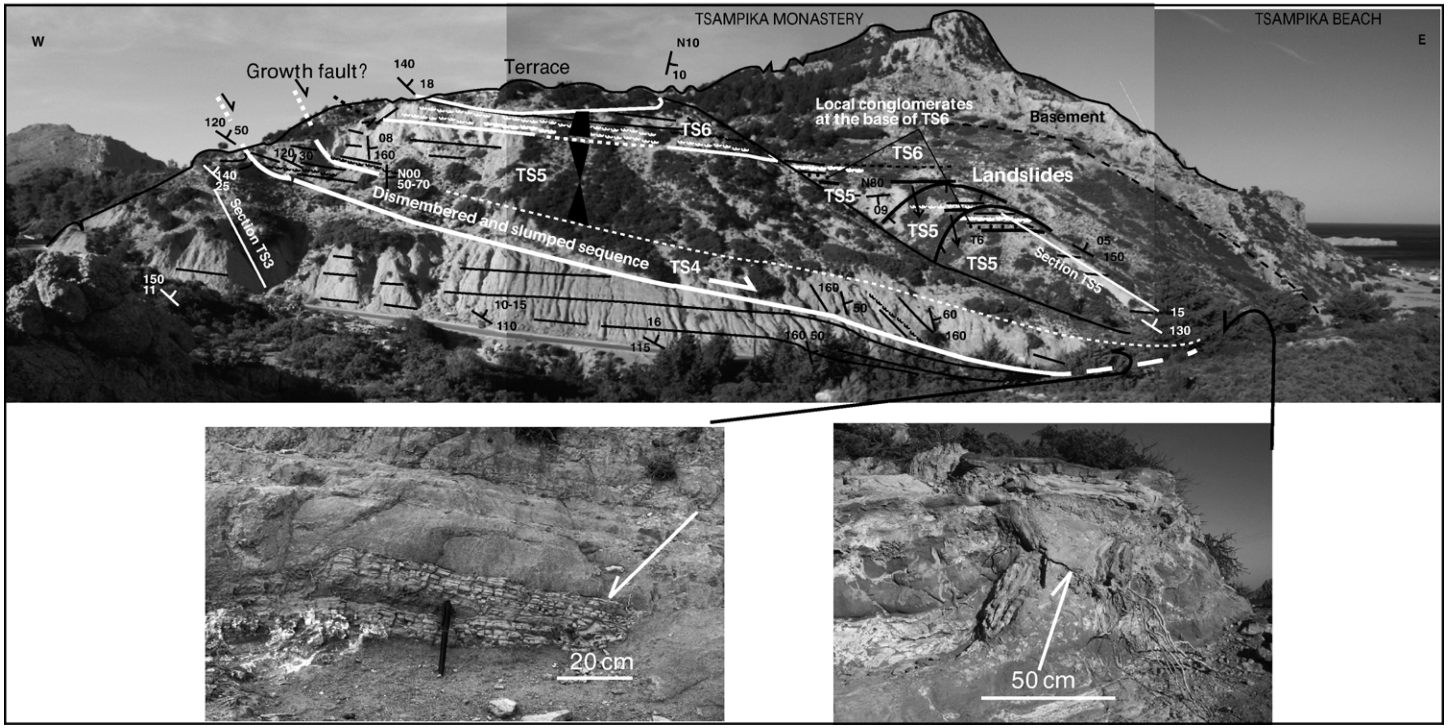

Fig. 10. Field view of the northwestern distal part of the Ladiko–Tsampika Formation, Tsampika area, with details of slumping (left) and diapirism (right) in sequence TS4.

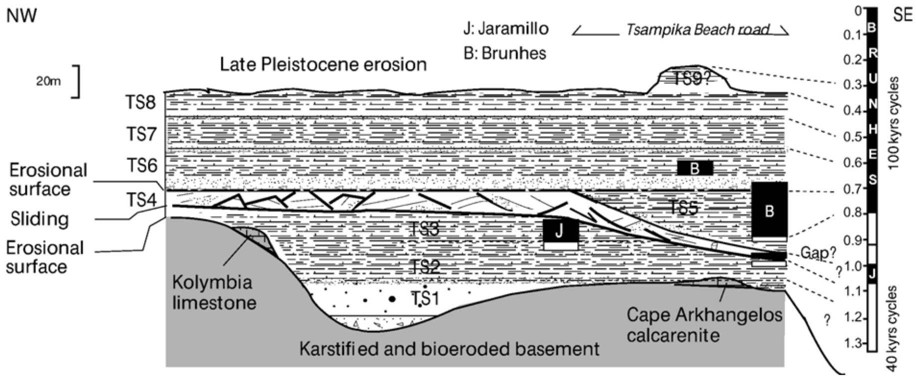

Fig. 11. Synthetic profile of the Ladiko–Tsampika Formation with indication of palaeomagnetic results.

sandstones overlie sequence TS4 in the north (Fig. 9) and sequence TS5 in the south (Fig. 10). Sequences TS7 and TS8 are exposed in isolated patches preserved on basement rocks (Fig. 9). They are also composed of littoral sandstones and conglomerates, or offshore silty clays. The repeated occurrence of beach to offshore sediments within each sequence of the Tsampika Member probably reflects cyclic sea-level changes. We estimate from palaeoecological and sedimentological data that their deposition took place at depths between 0 and 40 m.

Pollen analysis indicates that in each investigated sequence the maximum amount of *Pinus* grains is recorded within the laminated silty clays, reflecting distal conditions of sedimentation (Fig. 8). The minimum percentages of *Pinus* are documented at the base and at the top of each sequence, pointing towards littoral settings. Consequently, the signal registered by *Pinus* fits well with eustatic variations from littoral to shoreface depositional environments. The “pollen index” shows that taxa requiring warm- to cool-temperate and humid conditions (i.e. mostly *Quercus* and *Cedrus*) are prevalent in the offshore deposits of the sequences, whereas taxa indicative of colder and more xeric conditions (i.e. Asteraceae) are better represented in shallow-water sediments. Such variability indicates vertical displacements of vegetation belts along nearby elevations: during warmer climatic episodes, altitudinal vegetation belts spread upwards and conversely (Suc, 1989; Combourieu-Nebout, 1993). Prevalence of taxa living in warm- to cool-temperate and humid conditions is consistent with high *Pinus* content, which itself indicates high sea levels (Suc et al., 1995). In contrast, when taxa reflecting cooler and more xeric conditions are more abundant, *Pinus* content decreases simultaneously, indicating shoreface facies. The “pollen index” is positively correlated with *Pinus* ratio ($r^2 = 0.15$; $p < 0.001$). Variations in percentages of *Pinus* pollen grains are consequently mainly controlled by eustatic changes, indirectly dependent on climatic fluctuations. This pattern is interpreted as recording sea-level changes coeval with glacial–interglacial cycles. This interpretation is further corroborated for sequences TS6–7–8 and probably TS9 by sedimentological and faunal analyses.

An abrupt syn-sedimentary tectonic event occurred during deposition of sequence TS4. In the north, this sequence was dismembered into metric-scale blocks that slid south- to southeastward relative to each other (Fig. 9). Each block is limited by low-angle normal faults and may have rotated, indicating an extensional tectonic control. Normal faults reveal semi-brittle deformation with rare slickensides and folds. This suggests that sequence TS4 was not totally lithified during the extensional tectonic episode. The top of Sequence TS4 was eroded and the fluvial conglomerates at the base of sequence TS6 sealed the tectonic blocks. In the south, the base of sequence TS4 truncates sequences TS2–3 (Fig. 10). Blocks of TS4 slid southeastward and rotated along a sharp, medium-angle fault. Along the road to Tsampika Beach, the southeasternmost outcrops of sequence TS4 are only 3 m thick and display slump folds, diapiric soft deformation, and turbidites. We consider that sequence TS4 suffered a rapid extensional tectonic event that induced temporary emersion and block faulting in the northwest and submarine gravity sliding and re-sedimentation into deeper-water settings in the southeast.

The Ladiko–Tsampika Formation overlies remnants of the partially eroded Cape Arkhangelos calcarenite. It is consequently Pleistocene in age. Palaeomagnetic data (Fig. 11) show a reversed polarity zone at the top of sequence TS2. Sequence TS3 and two samples at the base of TS4 show normal polarity. Sequence TS5 begins with reversed polarity then changes to normal polarity that continues up to sequence TS6. The Ladiko–Tsampika Formation was deposited after the Lindos Bay clay, and therefore is younger than the "Olduvai subchron" proposed for the Cape Vagia section. The normal polarity zones recorded in TS3 and TS4 are therefore correlated with the Jaramillo subchron and the normal polarity in TS5 and TS6 with the Brunhes Chron. Demagnetisation diagrams of the reversed intervals of the Tsampika section indicate counter-clockwise directions (Fig. 3), similar to those in samples taken in the lower part of the sequence (Duermeijer et al., 2000). This suggests that the magnetisation is primary and constrains the vertical axis rotation phase of Rhodes to younger than 1 Ma. Above the top of the Jaramillo subchron (about 1 Ma), the inferred climatic cycles are the Pleistocene glacial–interglacial 100-kyr cycles (Ruddiman et al., 1986; Raymo et al., 1990; Berger et al., 1993; Bassinot et al., 1994; Berger and Jansen, 1994; Mudelsee and Schulz, 1997). As five cycles (TS5–6–7–8–9) are indicated (Fig. 10) above the base of the Brunhes Chron (about 0.8 Ma), the top of the preserved part of the Tsampika Member can be dated at around 0.3 Ma. Before the start of the Jaramillo subchron at 1.07 Ma (Cande and Kent, 1995), climatic cycles are generally related to obliquity (40 kyrs) (Ruddiman, 2003). Three to four climatic cycles have been evidenced within sequences TS1–TS2, so that sequence TS1 may be roughly estimated to begin at around 1.2–1.3 Ma. Consequently, deposition of the Ladiko–Tsampika Formation is considered to have occurred from about 1.3 to 0.3 Ma.

5. Discussion

5.1. Depositional history of the Kritika–Kolymbia–Lindos–Arkhangelos cycle

The erosional surface formerly hypothesized at the top of the Kritika Formation was not found in the Faliraki section. Elsewhere in the investigated areas, the Kritika Formation was mapped as infilling palaeorelief (Mutti et al., 1970) and the Lindos Bay clay is missing above it. There is no direct evidence for an erosional surface between the Kritika and the Rhodes Formations. Our investigations lead us to consider that the brackish to marine littoral Kritika Formation was deposited in low elevation parts of the deformed and eroded basement prior to deposition of the Rhodes Formation sensu Hanken et al. (1996) (Fig. 12A). The Kritika Formation should therefore be regarded as a Member forming a transgressive systems tract at the base of the Rhodes Formation. Some 180 m of lagoonal to littoral sediments accumulated, indicating that the sediment supply balanced the rate of subsidence. Allowing for a Late Pliocene age and a duration of 0.78 Ma (Gradstein and Ogg, 2004) for deposition of the Kritika Member (Fig. 13), the subsidence rate is estimated at a maximum of 0.23 mm/year.

The ages of the Kolymbia limestone and Lindos Bay clay Members are crucial. Based on the presence of *H. balthica* and magnetostratigraphy, Løvlie et al. (1989) assigned a Late Pliocene to Early Pleistocene age, up to 0.7 Ma, for these Members in the Cape Vagia section (Fig. 4A). However, our micropalaeontological study of the Cape Vagia and Haraki sections shows that *H. balthica* is absent, and that a Late Pliocene foraminifer association is present, consistent with the Olduvai subchron proposed at Cape Vagia. However, *H. balthica* was found in the Faliraki and Lindos sections (Fig. 4B), and Orombelli and Montanari (1967) also mentioned its presence in the now obliterated Vasfi section near Rhodes. Furthermore, our radiometric dating of the Haraki volcanic sand yielded an age of about 2.06 Ma. Consequently, deposition of the Lindos Bay clay started during the Late Pliocene (about 2.09 Ma: Vagia and Haraki sections) and continued into the Early Pleistocene (around 1.4 Ma: Faliraki and Lindos sections). A time span of 0.7 Ma is thus suggested for the deposition of the Lindos Bay clay. It cannot be much longer, because of the time needed for sedimentation of the overlying Cape Arkhangelos calcarenite and the subsequent erosion, prior to deposition of the Ladiko–Tsampika Formation, which probably started around 1.3 Ma. This may be a result of either an elevated position of the Faliraki area during the Late Pliocene, or southwestward tilting of northeastern Rhodes during deposition of the Lindos Bay clay.

The Kolymbia limestone onlaps the basement and an accommodation space of some 150 m was thus created (Fig. 12A). As indicated by bryozoan associations and the position of the highest known abrasion platforms in the Prophitis Elias Mountain, deposition of the Lindos Bay clay took place during the Olduvai subchron (Fig. 13), at water depths of more than 520 m in the Vagia and Lindos sections (Fig. 12A). This depth greatly exceeds the magnitude of normal glacio-eustatic sea-level changes (Pillans et al., 1998; Zazo, 1999). It attests to the tectonically induced major drowning of northeastern Rhodes, previously suggested (Meulenkamp et al., 1972; Hanken et al., 1996; Kovacs and Spjeldnaes, 1999), although some low amplitude climate-related sea-level changes probably also occurred (Kovacs and Spjeldnaes, 1999;

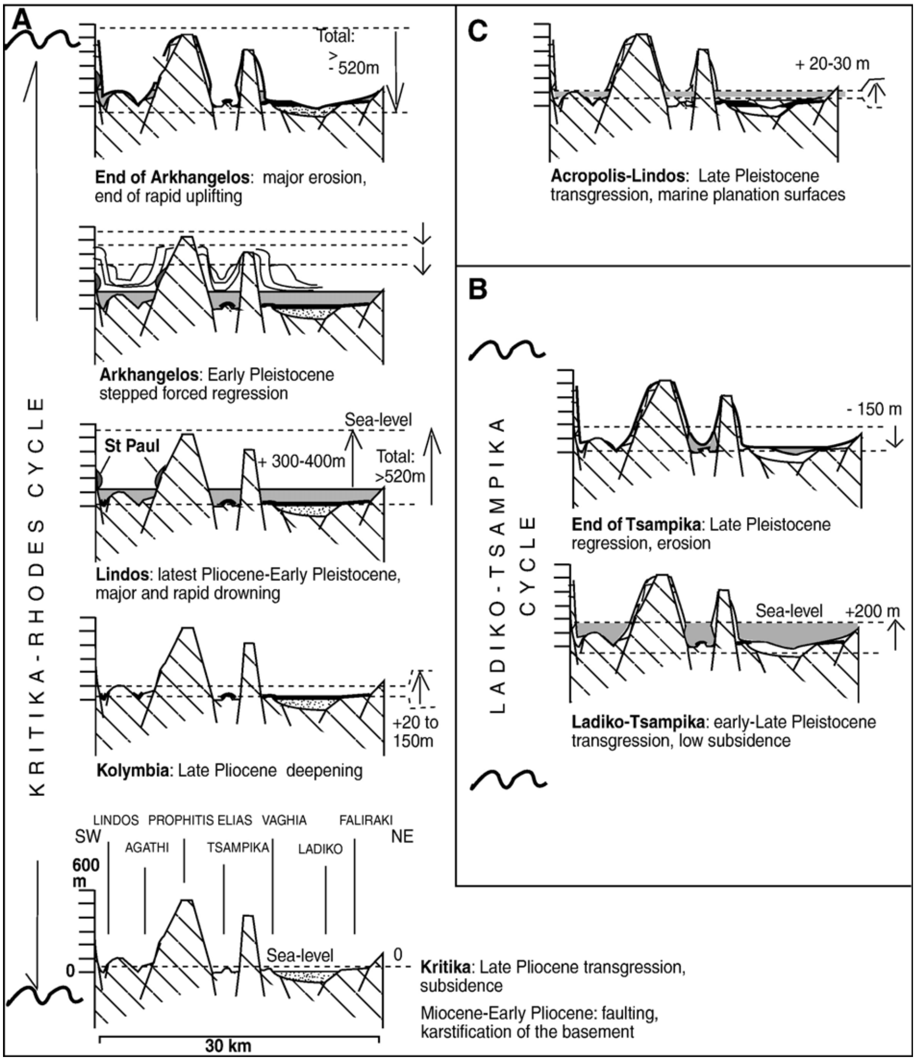

Fig. 12. Diagrams showing the tectono-sedimentary evolution of northeastern Rhodes from Late Pliocene to Late Pleistocene.

Steinthorsdottir, 2002). Allowing for at least 520 m of accommodation space created during the Olduvai subchron (duration 0.18 Ma), the average rate of tectonic drowning is 2.6 mm/year. This is very rapid, even when compared with rates considered to be elevated, as for instance in Milos and Aegina during the Early Pliocene: 1.5 mm/year (van Hinsbergen et al., 2004). As mentioned by Moissette and Spjeldnaes (1995), water depth decreased during the deposition of the uppermost part of the Lindos Bay clay, recording a highstand, which lasted about 0.5 Ma.

Subsequently, the Cape Arkhangelos calcarenite was deposited during a major forced regression between around 1.4 and 1.3 Ma (Fig. 13). This is a stepped regression characterized by an alternation of stillstands and rapid falls in sea level. Stillstands created numerous abrasion platforms due to bioerosion and wave action. The total relative sea-level fall is at least 520 m and most of the previously deposited sediments were eroded. This is interpreted as the consequence of a tectonic inversion accompanying stepped uplift of northeastern Rhodes. Assuming a duration of 100 ka and a sea-level change of 520 m, the average uplift rate was high, around 5.2 mm/year. This rate is much higher than most values measured in present-day volcanic arcs (e.g. 0.8 mm/year of uplift in La Désirade, Lesser Antilles: Feuillet et al., 2004). It is also much greater than the highest known Holocene uplift rates recorded in the Gulf of Corinth (2.4–3.0 mm/year: Stiros, 1998). The stepped pattern of the uplift implies sudden uplift movements separated by tectonically quiet periods. This might be partly related to earthquakes, as suggested for some sudden large Late

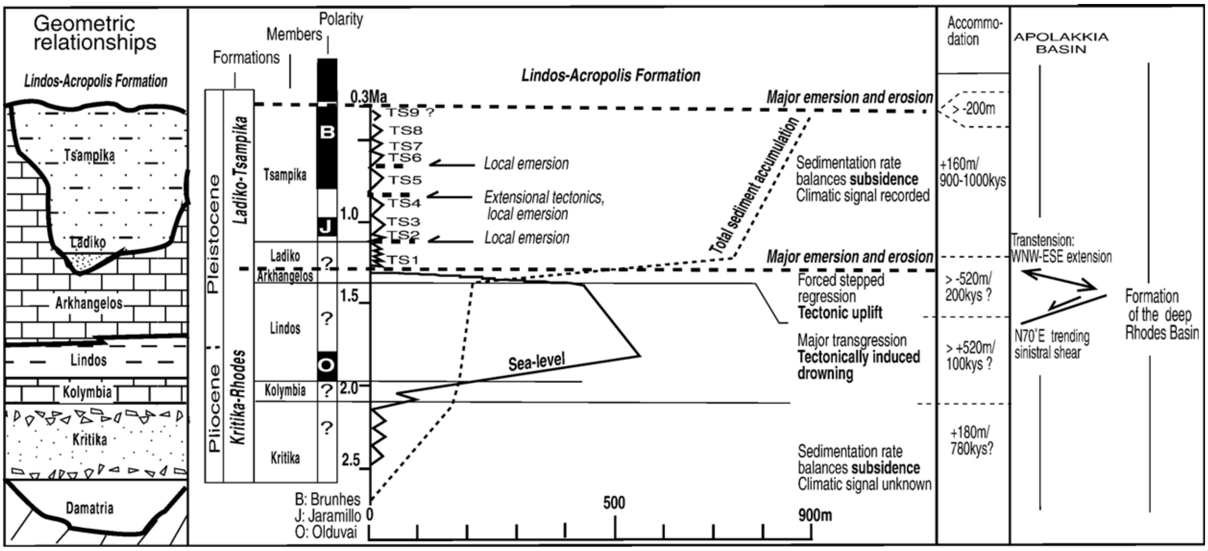

Fig. 13. Main stratigraphic, tectonic, sedimentary and climatic events in northeastern Rhodes.

Holocene vertical movements (Pirazzoli et al., 1989; Stiros, 1998).

Based on data gathered from the four sections mentioned above, we estimate that the deposition of the Lindos Bay clay Member took place between 2.09 and 1.4 Ma. The Cape Arkhangelos calcarenite was deposited from about 1.4 to 1.3 Ma.

5.2. Depositional history of the Ladiko–Tsampika cycle

The Ladiko sand did not yield chronostratigraphical information. However, we propose that this Member is partially coeval with sequence TS1 of the Tsampika area because: (1) beach to offshore sequences occur over a major erosional surface in the Ladiko and Tsampika areas; (2) in both areas deposition begins with continental to brackish sediments; (3) the sandy littoral deposits contain the same and unique faunal association, including transported hard-substrate communities indicating the vicinity of steep palaeohighs; (4) in both areas the lowermost sandy fossiliferous beds are abruptly overlain by fossil-poor greenish silty clays, probably recording a similar sedimentological–ecological event.

The Ladiko–Tsampika Formation is part of a transgressive systems tract comprising some 160 m of mostly marine shallow-water deposits. Based on sedimentological and pollen data, the beach to offshore sequences are correlated with climatic changes and probably record the 40- to 100-kyr climatic cycles evidenced by Bassinot et al. (1994) and Mudelsee and Schulz (1997). Palaeoecological data indicate that sea-level changes were moderate, probably in the order of tens of metres.

These sea-level oscillations are superimposed on minor tectonic movements. The deposition of some 160 m of littoral deposits indicates that subsidence rate was balanced by sediment accumulation. Assuming that deposition of the Ladiko–Tsampika Formation occurred during 1.3–0.3 Ma, the average subsidence rate would have been low, probably around 0.16 mm/year. This value is near that measured for the uplift of western Turkey since the Middle Pleistocene (0.2 mm/year: Westaway et al., 2004). Nevertheless, the abrupt extensional deformation and the gravity sliding identified in sequence TS4 indicate that regional tectonics was still active.

A major erosional episode (Fig. 12B) occurred at the end of the Ladiko–Tsampika Formation deposition, prior to sedimentation of the Lindos Acropolis Formation (Fig. 12C), when most of the Ladiko–Tsampika Formation was eroded down to present-day sea level. The relative sea-level fall was about 150 m. This erosion appears to have occurred since about 0.3 Ma, before deposition of the Lindos Acropolis Formation, but no precise data are presently available. The erosion was probably triggered by one of the Late Pleistocene glacio-eustatic sea-level falls, estimated at about 120 m (e.g. Pillans et al., 1998). It may also be tectonically controlled since Rhodes suffered vertical motions during the Late Pleistocene and the Holocene (Pirazzoli et al., 1989; Hanken et al., 1996). These considerations lead us to propose a latest Pleistocene–Holocene age for the Lindos Acropolis Formation.

5.3. Northeastern Rhodes within the eastern Aegean fore-arc

The Late Pliocene–Middle Pleistocene sediments of northeastern Rhodes record a tectono-sedimentary evolution different from that of surrounding areas. The continental Appolakia Basin of southwestern Rhodes (Fig. 1B) originated as a Late Miocene fault-wedge basin resulting from a regional D1 tectonic phase, with SW–NE trending extension (ten Veen and Kleinspehn, 2002). During the Pliocene and the Pleistocene, subsidence of the basin was controlled by a D2 transtensional phase. Transtension resulted from the combination of WSW–ENE trending extension and 070° sinistral shear (Fig. 1B) (ten Veen and Kleinspehn, 2002). Between the island of Rhodes and Turkey is the deep Rhodes Basin (Fig. 1B), which reaches a depth of some 4400 m (Woodside et al., 2000). This basin has a deformed pre-Late Miocene basement overlain by a weakly deformed and unconsolidated 1 km thick post-Miocene sedimentary cover. Woodside et al. (2000) suggested that the collapse of the basin was related to a ENE–WSW strike-slip motion trending in the same direction as the Pliny Trench, in the submerged part of the fore-arc, during the extension of the Aegean region and western Turkey (Fig. 1B). Sinistral strike-slip motions related to the Strabo and Pliny Trenches are not limited to the Rhodes area. They controlled the deformations in the eastern Aegean arc during the Pliocene and the Pleistocene (e.g. Crete: Duermeijer et al., 1998).

The major collapse of northeastern Rhodes occurred during the latest Pliocene (coeval with the lower part of the Lindos Bay clay). This event is coeval with formation of the Rhodes Basin. However, a unique feature is the major and rapid uplift recorded in northeastern Rhodes during the Early Pleistocene (forced regression of the Cape Arkhangelos calcarenite). The closest tectonic–sedimentary evolution is documented along the southeastern shoreline of Rhodes. In this region, between Lindos and Cape Prasonision (Fig. 1B), the Kritika Member, the Kolymbia limestone and the deep-water Lindos Bay clay have been recorded (Nelson et al., 2001; our preliminary observations). A newly defined formation, the Plimiri Algal limestone unconformably overlies this sequence and is considered to be an equivalent of the Lindos Acropolis Formation (Nelson et al., 2001). The presence of the Ladiko–Tsampika Formation and Cape Arkhangelos calcarenite has not yet been reported and much more detailed work remains to be done in this area.

The easternmost part of Rhodes may be considered as connected to the Rhodes Basin, because both collapsed during the latest Pliocene. But, since around 1.4 Ma, the easternmost part of Rhodes has been rapidly uplifted and thus was disconnected from the subsiding Rhodes Basin. From that time on, the easternmost part of Rhodes has to be treated as a discrete structural block with its own specific tectono-sedimentary evolution.

6. Conclusion

This paper documents the chronostratigraphy, sedimentology, and palaeoecology of the Late Pliocene to Middle Pleistocene coastal deposits of northeastern Rhodes.

Two main transgressive–regressive sedimentary cycles are evidenced: the — the Rhodes Formation cycle (including the Kritika) and the Ladiko–Tsampika Formation cycle, each bounded by major erosional surfaces.

The Rhodes Formation is considered to have been deposited during one tectonically controlled sedimentary cycle. The Kritika Member and the Kolymbia limestone are interpreted as a transgressive systems tract, the Lindos Bay clay as a highstand systems tract and the Cape Arkhangelos calcarenite as a shelf margin wedge to lowstand systems tract. This cycle lasted around 1 Ma, between the Late Pliocene to about 1.4–1.3 Ma. Vertical tectonic movements of the coastal palaeorelief initially controlled the sedimentation. Low-speed subsidence is recorded during the Late Pliocene transgression (Kritika Member). A tectonic drowning later occurred during the latest Pliocene–earliest Pleistocene (Kolymbia limestone and Lindos Bay clay). A major and rapid tectonic inversion, resulting in a rapid uplift and an emersion, developed during 1.4–1.3 Ma (Cape Arkhangelos calcarenite). The major consequence is a stepwise forced regression that formed several abrasion platforms on the basement limestones.

The Ladiko–Tsampika Formation is interpreted as the transgressive systems tract of an incomplete second Pleistocene sedimentary cycle, between approximately 1.3 and 0.3 Ma, abruptly disrupted by a major emersion. Despite an active extensional tectonic setting, the subsidence was very slow and climatic cycles were consequently recorded in the coastal deposits. The sedimentary organization was strongly influenced by 40 and then 100-kyr climatic cycles. The origin of the major erosion at the top of the Ladiko–Tsampika Formation is currently unknown, principally because of the lack of precise dates.

The eastern part of Rhodes provides a good example of the geodynamic evolution in an active fore-arc where vertical movements are probably on a million-year timescale. This region experienced a different tectono-stratigraphic history from that of the neighbouring continental Appolakia Basin and deep-sea Rhodes Basin (Woodside et al., 2000; ten Veen and Kleinspehn, 2002). It is characterized by slow vertical motions (around 0.16 mm/year), disrupted by rapid vertical movements (2.6–5.2 mm/year), including rapid drownings and uplifts. Climatic cycles were recorded in the sedimentation only when subsidence was slow. The easternmost part of Rhodes has to be considered as a separate structural block in the Aegean fore-arc with its own tectono-sedimentary evolution since the Early Pleistocene.

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