Facies (2005)
DOI 10.1007/s10347-005-0039-8

ORIGINAL ARTICLE

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Toarcian–Kimmeridgian depositional cycles of the south-western Morondava Basin along the rifted continental margin of Madagascar

Received: 23 August 2004 / Accepted: 23 November 2005

Abstract After rifting and final breakup of Gondwana along the former East-African-Antarctic Orogen during the Toarcian–Aalenian, passive margins formed around the Proto-Indian Ocean. Sedimentological and stratigraphic studies in the southern Morondava Basin contribute to an improved reconstruction of palaeoenvironmental changes during the syn-rift and post-rift margin formation. Depositional models based on outcrop and literature data in combination with subsurface data sets provide a stratigraphic framework of four transgressive-regressive (T-R) cycles. The incorporation of stratal architecture models derived from seismic images is essential. After a syn-breakup T-R cycle (T-R1), the post-breakup succession commenced with a Bajocian – Early Bathonian carbonate platform (T2). Middle–Late Bathonian sandstones (R2) formed when a global sea-level fall forced the shoreline to move basinward. Incised valleys and palaeokarst known from seismic lines are typical for forced regression cycles. In the Early Callovian again a widespread transgression occurs (T3). During a short regressive phase from the Late Callovian(?) to Early Oxfordian (R3), the siliciclastic shoreface deposits prograded onto the shelf. From the Early Oxfordian onwards a transgressive trend continued (T4). T1–T3 can be explained as the response to the structural development of the breakup rifting but they follow sea-level changes observed in other parts of the world. R3 and T4, in contrast, reflect eustacy.

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Keywords Madagascar · Gondwana · Post-breakup · Jurassic · T-R cycles

Introduction

From the Late Carboniferous onwards, crustal extension in the centre of Gondwana resulted in the formation of three sedimentary basins in western Madagascar: Morondava, Majunga, and Ambilobe (or Diego) basins (Figs. 1 and 2). The basins were aligned with a zone of weakness, which developed along the former Pan-African mobile belt (Montenat et al. 1996; Piqué et al. 1999). Madagascar was situated on the eastern side of this axis of future breakup (Reeves et al. 2002). These basins are characterised by thick successions of Late Palaeozoic, Mesozoic and Cenozoic sediments.

From the Late Carboniferous onwards, a series of localised pull-apart basins and from the Early Permian until the Triassic more extensive intracontinental rifts formed in the centre of Gondwana (Coffin and Rabinowitz 1992; Montenat et al. 1996; Schandelmeier et al. 2004). The Permian–Triassic interval is often considered the Gondwana Breakup strata, which were followed by transitional phase (Early–Middle Jurassic) and the drifting phase from the Callovian onwards (Luger et al. 1994; Montenat et al. 1996; Piqué et al. 1999).

Recent studies of tectono-sedimentary architectures from seismic images have shown that the breakup of Gondwana took place during the late Early Jurassic (Geiger 2004; Geiger et al. 2004). A widespread transgression represents the breakup unconformity and marks the boundary between the Toarcian–Aalenian syn-breakup half-grabens above extensional fault-blocks and the Bajocian–Kimmeridgian/Tithonian post-breakup strata (Geiger et al. 2004). Extensional faulting faded at the breakup unconformity, marking the end of rifting, while Madagascar drifted southwards away from Africa along the Davie Ridge Fracture Zone (Malod et al. 1991).

This paper addresses two issues. Firstly, it aims to review the lithostratigraphy and biostratigraphy for the southern

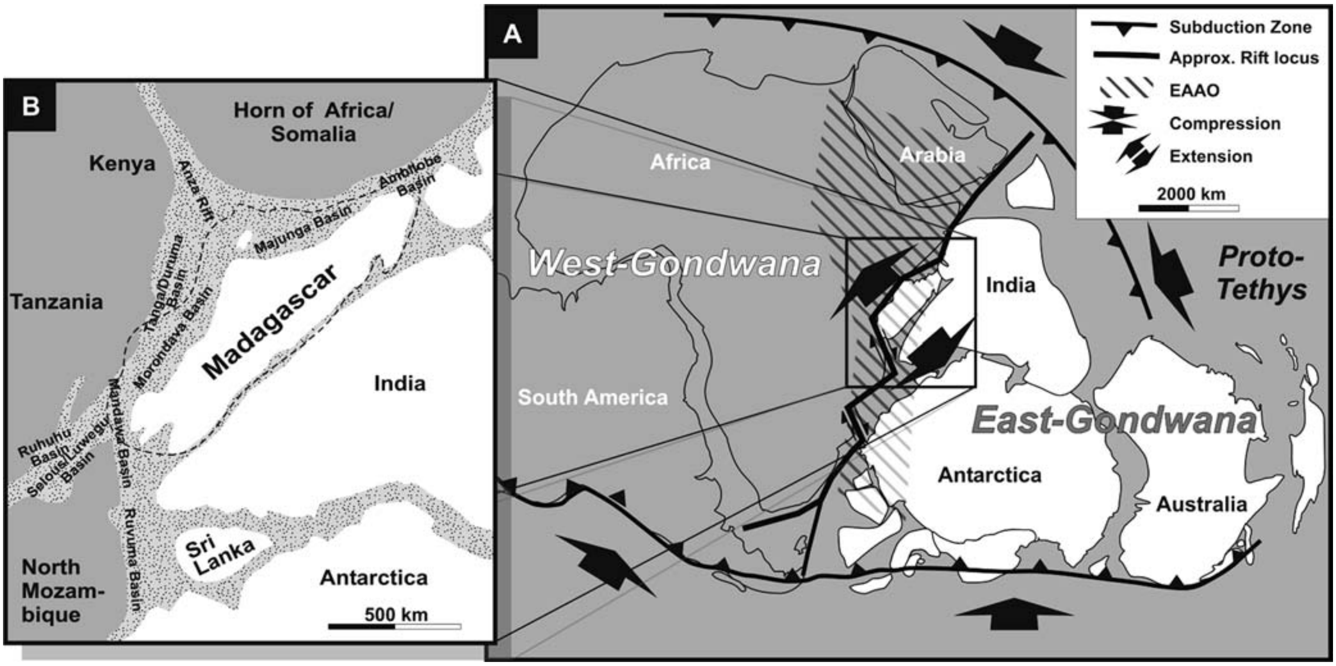


Fig. 1 **A** Gondwana reassembly at 200 Ma with present-day continents. Initial extension, as response to compression at the northern and southern margin of the supercontinent, was localised along the former East African-Antarctic Orogen (EAAO). Compiled from Stollhofen (1999) and Jacobs et al. (1998). **B** Reconstruction of Gondwana fragments by Reeves et al. (2002) based on an interpretation of ocean-floor topography. The outlines of Precambrian crustal fragments are shown in *grey* and *white*. Areas of sedimentary rocks in Karoo-aged basins are shown in *dotted* texture. Present-day Madagascar is shown in *dashed* outline

Morondava Basin to overcome the lack of coherent definitions of lithological units and the poor biostratigraphic data, which in the past have been both a communication challenge and a source of inconsistent stratigraphy. Therefore, 18 sections of Bajocian–Kimmeridgian successions in the southern Morondava Basin have been measured (Fig. 2) to study macroscale- and microscale sedimentary patterns and to characterise depositional environments. Macrofauna-and microfauna were studied for stratigraphic and palaeoenvironmental objectives. Secondly, the revised stratigraphic framework of palaeoenvironmental changes forms the base to identify transgressive-regressive (T-R) cycles in the southern Morondava Basin and to interpret their origin.

Stratigraphy of the Morondava Basin

Studies on the stratigraphy of the Morondava Basin started in the 1950s when SPM (Societé Pétrole de Madagascar) began a hydrocarbon exploration campaign. Most data were never published but sedimentological and stratigraphic concepts were summarized by Besairie and Collignon (1972).

The sedimentary fill of the Morondava Basin can be divided according to their position in the Gondwana Breakup (Geiger et al. 2004). The pre-breakup strata are stratigraphically and lithologically related to the Karoo Mega Sequence of continental Africa (SACS 1980; Kreuser 1995). They comprise the Late Carboniferous – Early Permian Sakoa Group, Middle Permian – Middle Triassic Sakamena Group, and the Late Triassic Isalo Formation (e.g. Hankel 1994; Montenat et al. 1996; Wescott and Diggens 1997, 1998; Geiger et al. 2004). The entire pre-breakup succession varies from 3,000 to 4,000 m in thickness, although it may reach 11,000 m in thickness in the southern Morondava Basin (Boast and Nairn 1982).

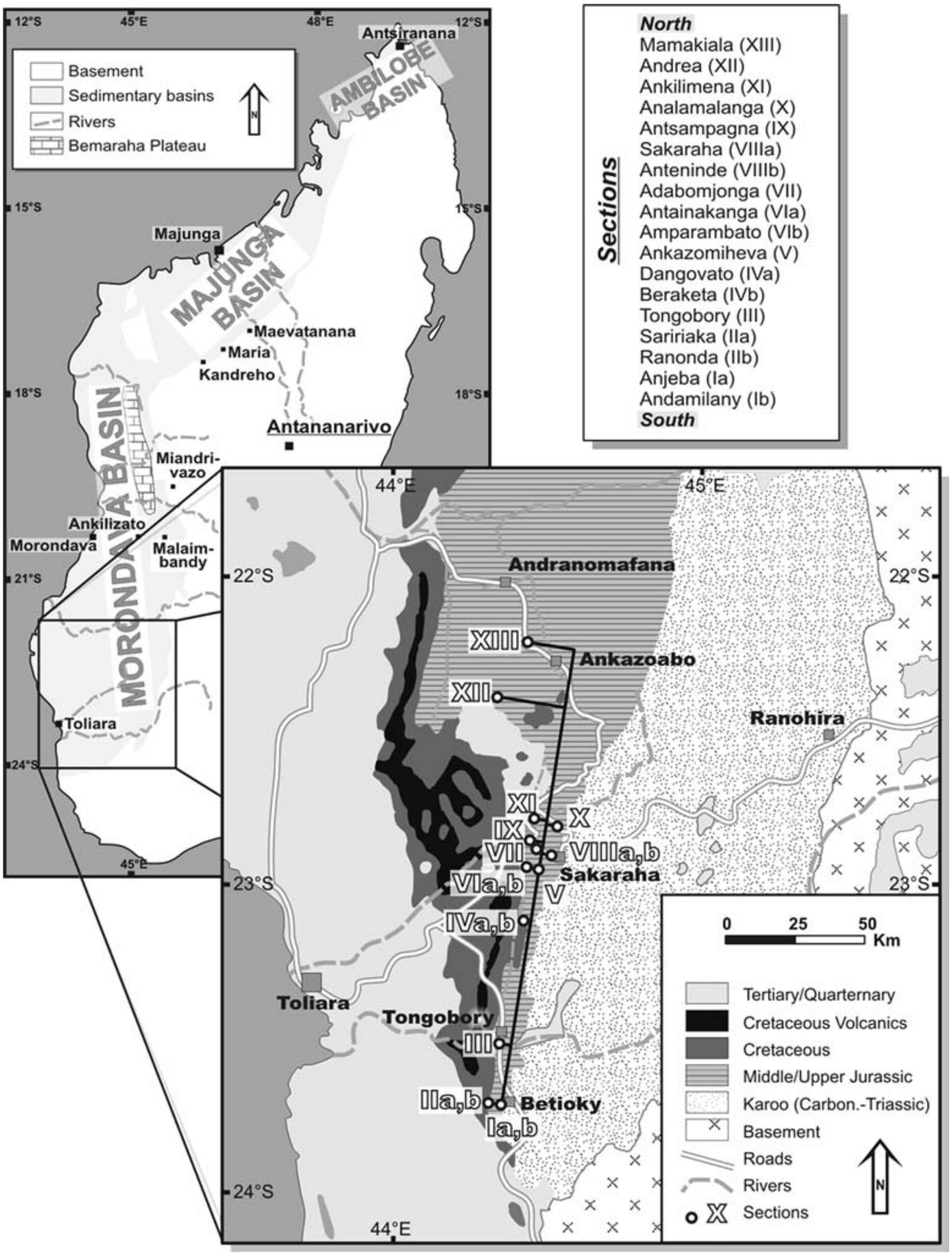
The syn-breakup in particular is documented in the half-graben fillings of mudstones (shales and marls) with a few limestone beds of the Andafia Formation. This formation is exclusively recognized in the seismic record of the former shelf (Geiger et al. 2004), where it, probably together with the Aalenian Sandstone, exceeds some hundred metres in wells (Clark 1996). At outcrop the Andafia Formation can only be found in the Morondava Basin in the Miandrivazo region (Fig. 2), where a few metres are exposed at the type locality at Andafia (Besairie and Collignon 1972; Geiger et al. 2004). In the south-western Majunga Basin yield more than 100 m of the correlative Berono Formation and the overlying Aalenian Sandstone is exposed.

Above the breakup unconformity rests a carbonate platform (Bemaraha and Sakaraha formations), followed by a forced regression sandstones (Ankazoabo and Sakanavaka formations) and deeper shelf deposits (Jurassic Duvalia Marls). The post-breakup succession rarely exceeds 300 m in thickness at outcrop, but is reported to reach more than 2,000 m in the subsurface (Uhmann 1996). The Jurassic record is cut off by two Early Cretaceous transgressive events in the Valaginian/Hauterivian and Aptian (Besairie and Collignon 1972; Luger et al. 1994).

Biostratigraphy of the southern Morondava Basin

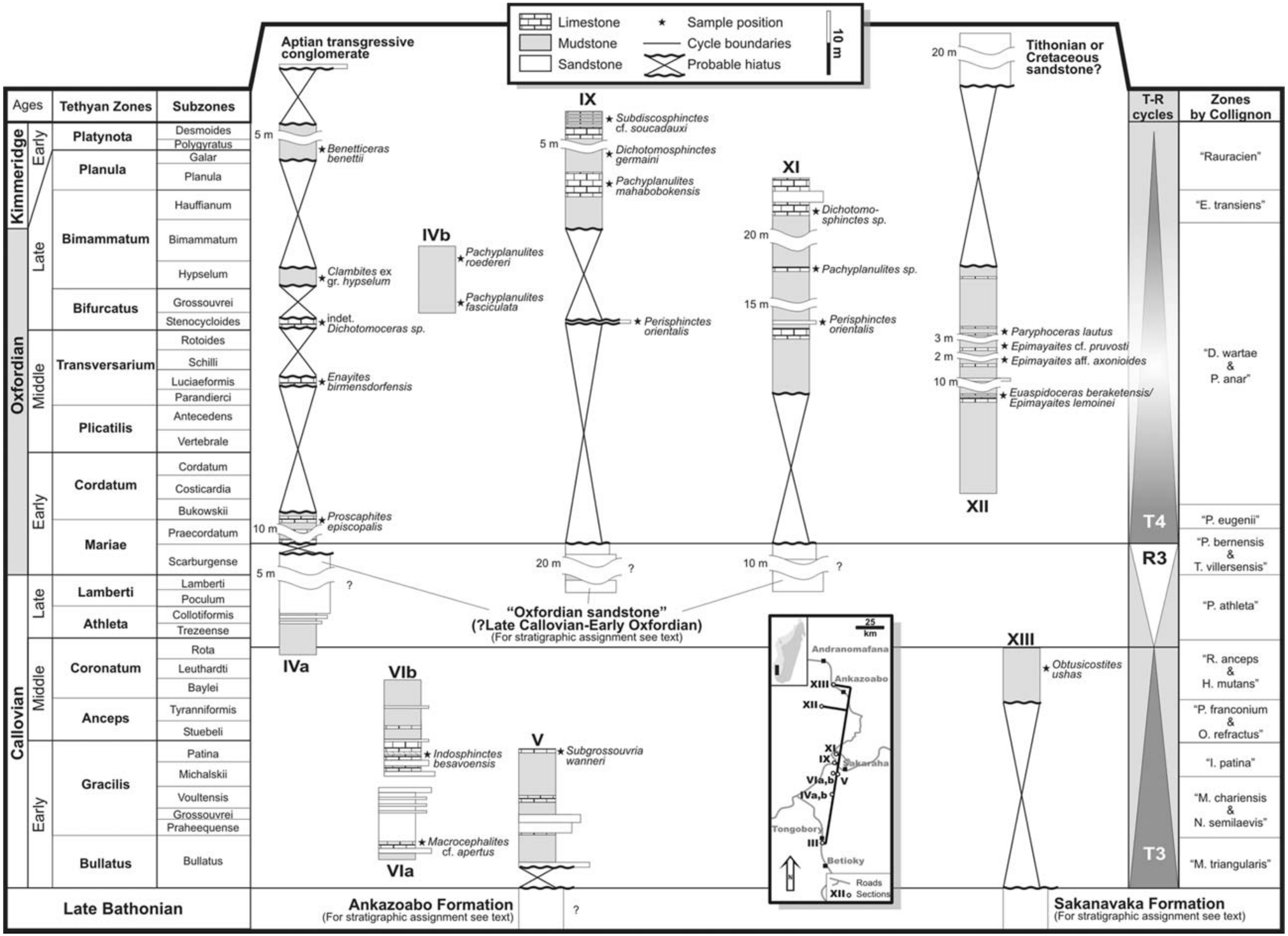
Biostratigraphic concepts of the Jurassic in Madagascar are generally hampered by faunal and floral provincialism due

Fig. 2 Three coastal basins—the Morondava, Majunga and Ambilobe basins—extend along the western coast of Madagascar and are filled with Late Palaeozoic, Mesozoic and Cenozoic sediments. In the southern Morondava Basin 18 sections were measured (zoomed sketch map: Persits et al. 2002)



to palaeogeographical isolation. Madagascar's west coast was part of a narrow embayment that extended southward from the southern Tethyan margin onto the Arabian Peninsula and Somalia. Ammonites are of considerable use for biostratigraphic calibration of Jurassic strata in Madagascar. Numerous species were identified from the measured outcrops and were compared with Tethyan (Cariou and Hantzpergue 1997), western Indian (Spath 1933; Cariou and Krishna 1988) and Madagascan faunas (Collignon 1953, 1960, 1964a,b, 1967; Collignon et al. 1959; Besairie and Collignon 1972; Joly 1976). Apart from the Early Oxfordian age when Boreal zonations are used, the Tethyan zonations of Cariou and Hantzpergue (1997) provide the biochronological scale for this study. Endemism also plays a major role in the composition of microfossil assemblages. Ostracods (Grekoff 1963; Mette 2004; Mette and Geiger 2004a,b,c), and foraminifers (e.g., Espitalié and Sigal 1963a,b; Sigal et al. 1970) have only few East African or Tethyan analogues and usually do not provide stratigraphic information. Palynological analyses of Jurassic strata in the Morondava Basin were performed by Hankel (1994) and Dina (1996), but they did not improve the stratigraphic frame significantly.

Ammonites used for the biostratigraphy of the sections are stored at "Staatliches Museum für Naturkunde Stuttgart, Germany".


Limestone
Mudstone
Sandstone
Sample position
Cycle boundaries
Probable hiatus
10 m
Ages
Tethyan Zones
Subzones
Kimmeridge
Oxfordian
Callovian
Early
Late
Middle
Platynota
Planula
Bimammatum
Bifurcatus
Transversarium
Plicatilis
Cordatum
Mariae
Lamberti
Athleta
Coronatum
Anceps
Gracilis
Bullatus
Desmoides
Polygyratus
Galar
Planula
Hauffianum
Bimammatum
Hypselum
Grossouvrei
Stenocycloides
Rotoides
Schilli
Luciaeformis
Parandierci
Antecedens
Vertebrale
Cordatum
Costicardia
Bukowskii
Praecordatum
Scarburgense
Lamberti
Poculum
Collotiformis
Trezeense
Rota
Leuthardti
Baylei
Tyranniformis
Stuebeli
Patina
Michalskii
Voultensis
Grossouvrei
Praheequense
Bullatus
Late Bathonian
Aptian transgressive conglomerate
Benetticeras benettii
Clambites ex gr. hypselum
indet. Dichotomoceras sp.
Enayites birmensdorfensis
Proscaphites episcopalis
IVa
IVb
Pachyplanulites roedereri
Pachyplanulites fasciculata
IX
Subdiscosphinctes cf. soucadauxi
Dichotomosphinctes germaini
Pachyplanulites mahabobokensis
Perisphinctes orientalis
XI
Dichotomosphinctes sp.
Pachyplanulites sp.
Perisphinctes orientalis
Tithonian or Cretaceous sandstone?
XII
Paryphoceras lautus
Epimayaites cf. pruvosti
Epimayaites aff. axonioides
Euaspidoceras beraketensis/ Epimayaites lemoinei
"Oxfordian sandstone" (?Late Callovian-Early Oxfordian)
(For stratigraphic assignment see text)
VIb
Indosphinctes besavoensis
VIa
Macrocephalites cf. apertus
V
Subgrossouvria wanneri
XIII
Obtusicostites ushas
Ankazoabo Formation
(For stratigraphic assignment see text)
Sakanavaka Formation
(For stratigraphic assignment see text)
Andranomafana
Ankazoabo
Sakaraha
Tongobory
Betioky
25 km
Roads
Sections
T-R cycles
T4
R3
T3
Zones by Collignon
"Rauracien"
"E. transiens"
"D. wartae & P. anar"
"P. eugenii"
"P. bernensis & T. villersensis"
"P. athleta"
"R. anceps & H. mutans"
"P. franconium & O. refractus"
"I. patina"
"M. chariensis & N. semilaevis"
"M. triangularis"

Toarcian–Aalenian

A universal biostratigraphic constraint for the first marine ingression in the East African domain is the appearance of *Bouleiceras* sp. (Luger et al. 1994; Hallam 2001). With the occurrence of *Bouleiceras nitescens* (Collignon) the Beronono Formation (Geiger et al. 2004) in the south-western Majunga Basin is assigned to the homonymous biozone *sensu* Collignon (Fig. 3). The *Bouleiceras nitescens* Zone corresponds to the Early Toarcian *Falciferum* Zone of the Tethyan realm. The subsequent succession lacks biostratigraphic index taxa but intercalating sandstones are commonly attributed as Aalenian Sandstone of suggested Aalenian age (Besairie and Collignon 1972).

Bajocian–Bathonian

The boundary between the syn-breakup and post-breakup successions approximates to the Aalenian/Bajocian boundary. Aalenian–Bathonian sediments are devoid of stratigraphic index fossils.

Bemaraha and Sakaraha formations

In the south-western Majunga Basin Besairie and Collignon (1972) attributed interbedded limestones and mudstones (shales and marls) above the Aalenian Sandstone to the Bajocian carbonate platform (Bemaraha Formation; Geiger et al. 2004). The carbonate platform at the Manambolo and Tsiribihina river gorges, which cut through the Bemaraha plateau (Fig. 2), are likewise assigned to the Early Bajocian (Besairie and Collignon 1972; Piqué et al. 1999). At the eastern margin of the Bemaraha Plateau the carbonate succession directly rests on Isalo sandstones. Farther south at Besabora (along National Road RN35), limestones with the foraminifers *Mesoendothyra croatica* (Gušić) and *Protopeneroplis striata* (Weynschenk) support a Middle Jurassic (Bajocian to Early Bathonian) age (Jekhowsky and Goubin 1964; Piqué et al. 1999). The sandstones and mudstones in the Besabora section are critical for interpretation. While Besairie and Collignon (1972) and Montenat et al. (1996) described debris fan deposits, Clark (1996) and Geiger et al. (2004) argue for a misconception derived from poor mapping and correlate the siliciclastic units with the mixed carbonate-siliciclastic facies of the coastal equivalent to the carbonate platform (Sakaraha Formation).

In the southern Morondava Basin the Sakaraha Formation rests directly on sandstones of the Isalo Formation. Uhmann (1996) gave a maximum Bajocian age at Sakaraha section (VIIIa), and also for Anjeba (Ia) and Andamilany (Ib) sections. The latter two were classified by Besairie and Collignon (1972) as Bathonian. A Bathonian age is also proposed by ostracods (Mette and Geiger 2004a). A Bathonian age for Sakaraha Formation is nowhere else confidently proven. Lathuilière et al. (2002), who studied the coral meadows in this area, identified the rhynchonellids *Burmirhynchia termierae* (Rousselle) and *Baeorhynchia transversa* (Cooper) in a horizon close to the transgressive base, which suggest an Early Bajocian to early Late Bajocian age. Corals also correspond to faunas from the Late Bajocian (Lathuilière et al. 2002).

Ankazoabo and Sakanavaka formations

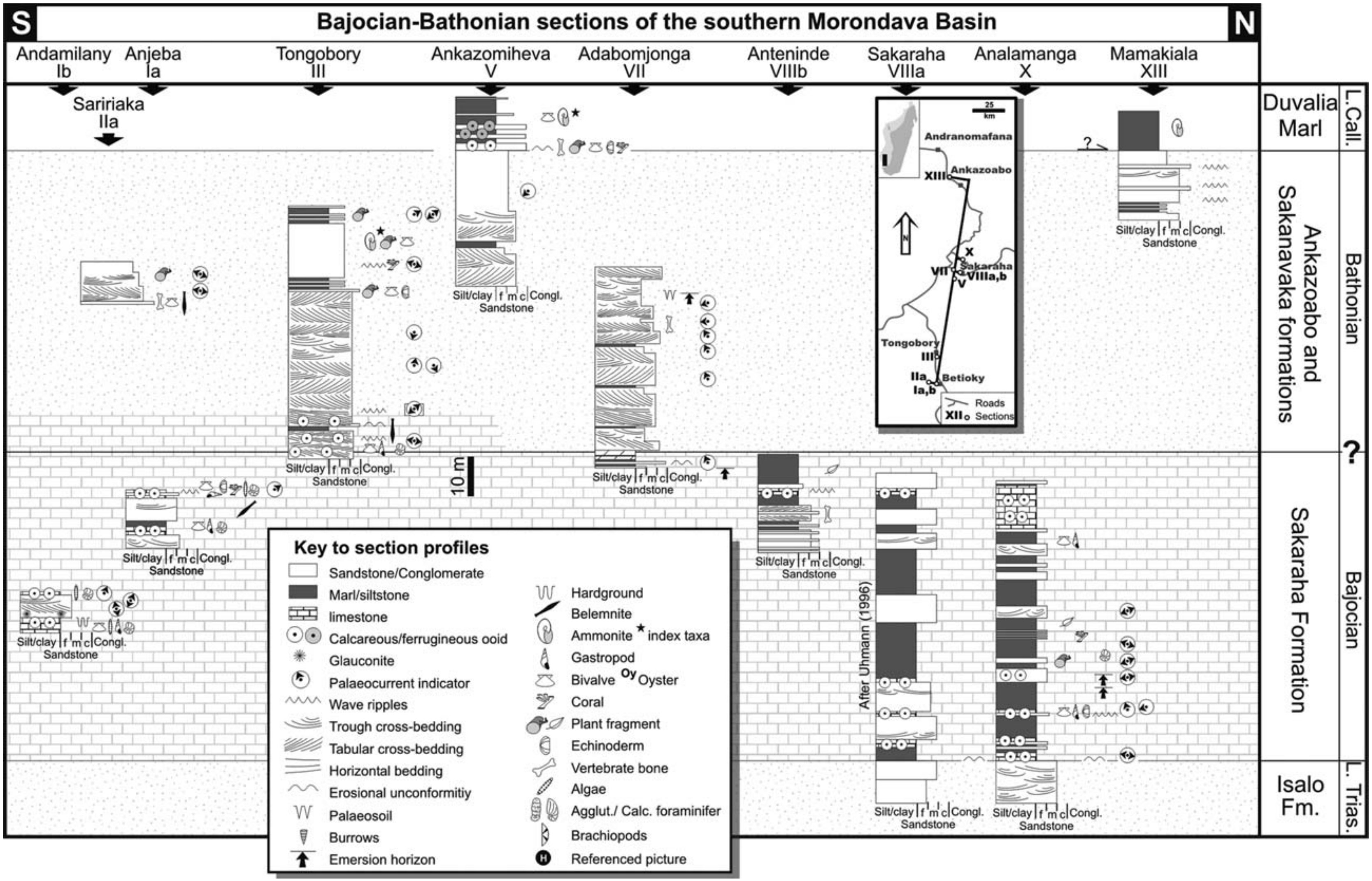
The stratigraphic age of the boundary between the Bemaraha and Sakaraha formations and the overlying sandstones of the Bathonian Ankazoabo and Sakanavaka Formation is unknown. The sandstones appear to be limited to the southern part of the Morondava Basin. Besairie and Collignon (1972) correlate the sandstones lithologically, e.g. at Adabomjonga (VII), but biostratigraphic indications are rare. Luger et al. (1994) correlate the Bathonian sandstones in the south with Late Bathonian units in the central basin based on *Clydoniceras* sp. and *Micromphalites hourcqi* (Collignon) (see Collignon 1964b). At the south end of the Morondava Basin, Tongobory section (III) gives a Bajocian–Bathonian age range due to the occurrence of *Mesoendothyra croatica* (Gušić). Uhmann (1996) classified those strata by indet. *Parachoffatia* sp. as Late Bathonian. Our new specimens could not be determined to the species level, but they agree with a Bathonian age. South of the Onilahy River, there is sandstone with the ammonites *Procerites hians* (Waagen) and *Delecticeras anjohense* (Collignon), which indicate a Middle to Late Bathonian age (Collignon 1964b; Besairie and Collignon 1972).

Another major sandstone succession, the Sakanavaka Formation is described by Besairie and Collignon (1972) from a section along the Sakanavaka River, 20 km NNW of Ankazoabo village, and by Uhmann (1996) and Dina (1996) from the supposed correlative section (XI) along the Mamakiala River (tributary of the Sakanavaka River). None of them gives a biostratigraphically specific fauna. The base of the sandstone is not exposed. Nevertheless, Besairie and Collignon classify this sandstone at Sakanavaka to be younger than the Ankazoabo Formation. They correlate it with the Mandabe Formation north of the Mangoky River, which itself is considered to overlie the Ankazoabo and Besabora formations (cf. Boast and Nairn 1982). However, there is no biostratigraphic support for this lithological correlation.

South of the Onilahy River, in the vicinity of Betioky, the sandstones and bioclastic conglomerates at Saririaka section (IIa) are lithologically correlated to the Ankazoabo sandstone at Tongobory section. Biostratigraphic markers are absent.

The age of the top of the sandstone formations can only be estimated indirectly by ammonites in the overlying mudstones and limestones to predate the Early Callovian (see below).

Fig. 3 Ammonite biostratigraphy of the Callovian – Early Kimmeridgian sections. Taxonomic comparison with Tethyan (Cariou and Hantzpergue 1997), western Indian (Spath 1933; Cariou and Krishna 1988) and Madagascan faunas (Collignon 1953, 1960, 1964a,b, 1967; Collignon et al. 1959; Besairie and Collignon 1972; Joly 1976)



S
Bajocian-Bathonian sections of the southern Morondava Basin
N
Andamilany Ib
Anjeba Ia
Saririaka IIa
Tongobory III
Ankazomiheva V
Adabomjonga VII
Anteninde VIIIb
Sakaraha VIIIa
Analamanga X
Mamakiala XIII
Duvalia Marl
L.Call.
Ankazoabo and Sakanavaka formations
Bathonian
Sakaraha Formation
Bajocian
Isalo Fm.
L. Trias.
Silt/clay f m c Congl.
Sandstone
10 m
After Uhmann (1996)
Andranomafana
Ankazoabo
XIII
Sakaraha
X
VII
VIIIa,b
V
Tongobory
III
IIa
Ia,b
Betioky
25 km
N
Roads
XII Sections
Key to section profiles
Sandstone/Conglomerate
Marl/siltstone
limestone
Calcareous/ferruginenous ooid
Glauconite
Palaeocurrent indicator
Wave ripples
Trough cross-bedding
Tabular cross-bedding
Horizontal bedding
Erosional unconformitiy
Palaeosoil
Burrows
Emersion horizon
Hardground
Belemnite
Ammonite ★ index taxa
Gastropod
Bivalve Oy Oyster
Coral
Plant fragment
Echinoderm
Vertebrate bone
Algae
Agglut./ Calc. foraminifer
Brachiopods
Referenced picture

Callovian – Early Oxfordian

From the Callovian onwards ammonites are useful biostratigraphic tools (Fig. 3). The earliest Callovian is identified with *Macrocephalites* cf. *apertus* (Spath) at the Antainakanga section (VIa). Together with other specimens of the same taxa with wide umbilici, the most likely age ranges between the *Bullatus* and *Gracilis* Zone (Early Callovian). At the Antainakanga section the contact to the underlying Ankazoabo Formation is not exposed.

Amparambato section (VIb), only 0.5 km NE of Antainakanga, belongs to the *Indosphinctes patina* Zone (late Early Callovian) due to the occurrence of *Indosphinctes besavoensis* (Collignon). Thus the Antainakanga section is slightly older than the Amparambato section.The Ankazomiheva section (V) contains the boundary to the underlying Ankazoabo and stratigraphically correspond to the Antainakanga and Amparambato sections. At Ankazomiheva, mudstones can be determined as Early Callovian with *Subgrossouvria wanneri* (Collignon) of the *Indosphinctes patina* Zone (Early Callovian).

The minimum age of the Sakanavaka sandstone was estimated indirectly on the basis of an unspecified fauna from the overlying Callovian mudstones (Besairie and Collignon 1972). At Mamakiala section (XIII) the occurrence of *Obtusicostites ushas* (Spath) confirms a Middle Callovian age ("*Reineckeia anceps* and *Hubertoceras mutans* Zone" *sensu* Collignon = Tethyan *Coronatum* Zone). Since no convincing stratigraphic or lithological concept exists to distinguish the Ankazoabo and Sakanavaka formations, it is more credible that they both belong to the same unit. Thus the Callovian transgression tops both the Ankazoabo and Sakanavaka formations.

The Callovian mudstone succession passes into sandstone, which has been variously reported as Oxfordian (Besairie and Collignon 1972; Luger et al. 1994; Uhmann 1996), but the stratigraphic position remains unclear. It appears that the Oxfordian Sandstone post-dates the *Athleta* Zone (Mamakiala section) and pre-dates the upper *Mariae* Zone. A further but weak stratigraphic constraint for the base of the sandstone comes from Keliambia, 19 km SSE of Ankazoabo, where Besairie and Collignon (1972) describe a Middle–Late Callovian mudstone and iron-oolitic limestone succession underneath the sandstone. Similar Late Callovian mudstones and limestones are also known from the slopes above Mamakiala (Besairie and Collignon 1972). Consequently, the maximum age of the Oxfordian Sandstone ranges from latest Callovian to Early Oxfordian.

The Ranonda section in the very south of the Morondava Basin cannot be unambiguously dated, but the occurrence of a *lytoceratide* suggests a Callovian age.

Fig. 4 Correlation scheme of measured sections of the Sakaraha, Ankazoabo, and Sakanavaka formations in the study area illustrates the stratigraphic difference of the Bajocian platform carbonates and the Bathonian sandstones. See Analamanga (X) and Adabomjonga (VII) sections for further exemplary details

Early Oxfordian – Early Kimmeridgian

Reappearing mudstones from the Early Oxfordian onwards are precisely dated at the Dangovato section (IVa). *Proscaphites episcopalis* (de Loriol) indicates the upper *Mariae* Zone to lower *Cordatum* Zone (Early Oxfordian). Middle Oxfordian is evidenced by *Enayites birmensdorfensis* (Moesch; upper *Transversarium* Zone, *Luciaeformis* Subzone). Late Oxfordian is identified by the occurrence of an unspecified but morphologically typical *Dichotomoceras* sp. which belongs to the lower *Bifurcatus* Zone, and *Clambites* ex gr. *hypselum* (Oppel) which is significant for the lower *Bimammatum* Zone (*Hypselum* Subzone). The occurrence of *Benetticeras benettii* (Checa) dates the top of the succession to the *Planula* Zone (Late Oxfordian) or slightly younger to the *Platynota* Zone (Early Kimmeridgian).

The Beraketa section (IVb) is classified as early Late Oxfordian (*Bifurcatus* Zone) based on a poorly preserved *Pachyplanulites fasciculata* (Collignon). Furthermore, a *Pachyplanulites* cf. *subevolutus* (Waagen) and *Pachyplanulites roedereri* (Collignon) signifies the upper part of the section as late Late Oxfordian (*Bimammatum* Zone).

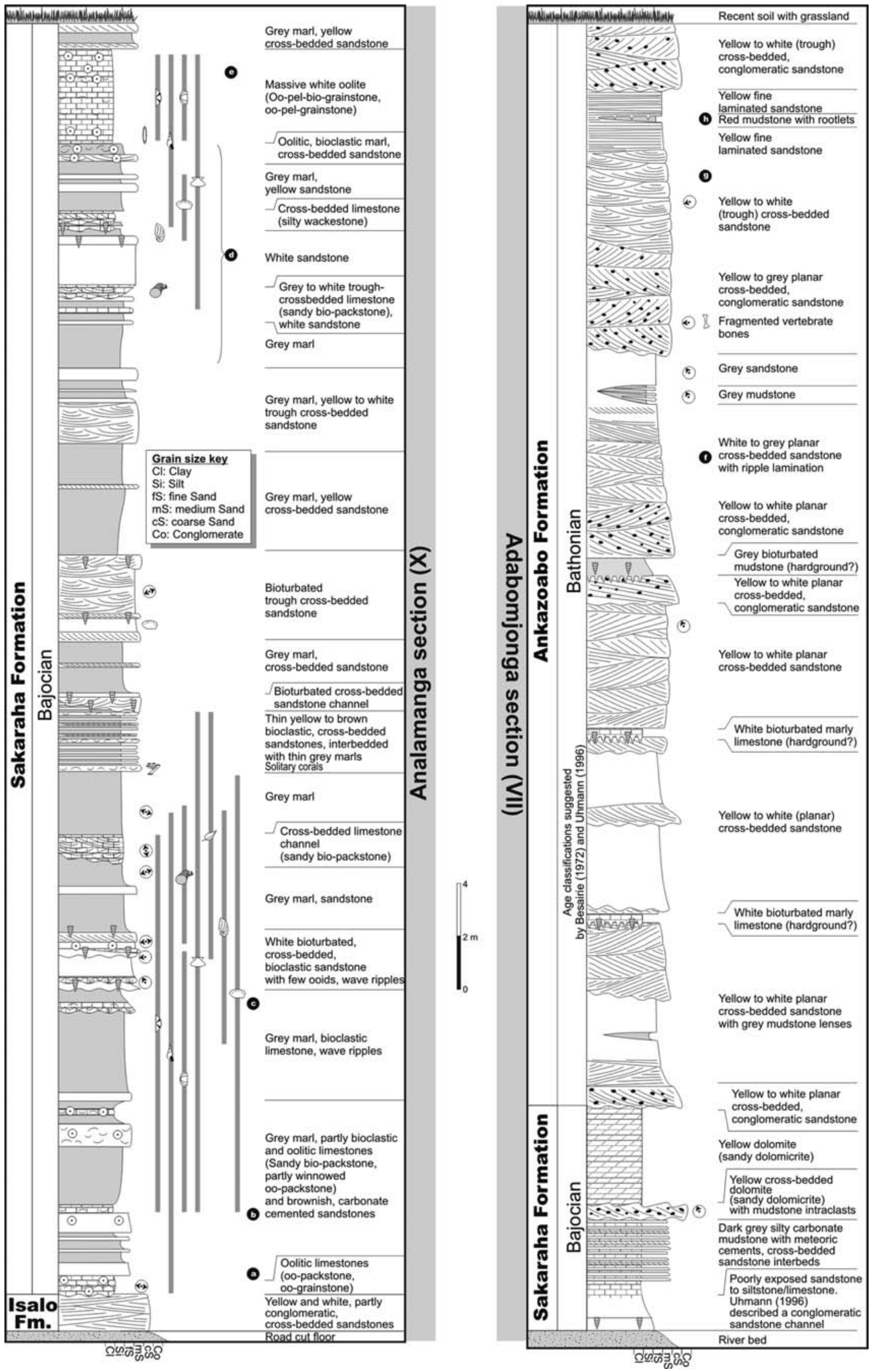
The Antsampagna section (IX) contains *Perisphinctes orientalis* (Siemiradzki) of the *Bifurcatus Zone* (Late Oxfordian). *Pachyplanulites mahabobokensis* (Collignon) and *Dichotomosphinctes germaini* (Collignon) indicates the "Rauracien" *sensu* Collignon (Latest Oxfordian – Early Kimmeridgian). *Subdiscosphinctes* cf. *soucadauxi* (Collignon) marks the uppermost top of the section as being slightly younger than the youngest fauna at Dangovato, possibly correlative to the European *Platynota* Zone (Early Kimmeridge).

At the Ankilimena section (XI) *Perisphinctes orientalis* (Siemiradzki) and a typical *Dichotomoceras* sp. indicate the lower *Bifurcatus* Zone (Late Oxfordian). Typical species of *Pachyplanulites* sp. and *Dichotomosphinctes* sp. are possibly slightly younger.

The Middle Oxfordian is also identified in the Andrea section (XII) based on finds of *Epimayaites lemoinei* (Spath) and *Euaspidoceras beraketensis* (Collignon) of the *D. wartae* and *P. anar*, niveau supérieur Zone of Collignon which correlates with the Tethyan *Transversarium* Zone. Slightly younger but still *Transversarium* Zone is the upper part of the mudstone succession with *Epimayaites* cf. *pruvosti* (Collignon), *Epimayaites* aff. *axonioides* (Spath) and *Paryphoceras lautus* (Spath).

Sedimentary environments

Four major sedimentary environments were recognised in the Jurassic syn-breakup and post-breakup strata of the southern Morondava Basin: (1) a Toarcian–Aalenian dysoxic basin environment, succeeded by marine sandstone, (2) a Bajocian – ?Early Bathonian carbonate platform, (3) Middle–Late Bathonian marine sandstones, and (4) Callovian – Early Kimmeridgian dysoxic basinal



Sakaraha Formation
Bajocian
Isalo Fm.
Grey marl, yellow cross-bedded sandstone
Massive white oolite (Oo-pel-bio-grainstone, oo-pel-grainstone)
Oolitic, bioclastic marl, cross-bedded sandstone
Grey marl, yellow sandstone
Cross-bedded limestone (silty wackestone)
White sandstone
Grey to white trough-crossbedded limestone (sandy bio-packstone), white sandstone
Grey marl
Grey marl, yellow to white trough cross-bedded sandstone
Grain size key
Cl: Clay
Si: Silt
fS: fine Sand
mS: medium Sand
cS: coarse Sand
Co: Conglomerate
Grey marl, yellow cross-bedded sandstone
Bioturbated trough cross-bedded sandstone
Grey marl, cross-bedded sandstone
Bioturbated cross-bedded sandstone channel
Thin yellow to brown bioclastic, cross-bedded sandstones, interbedded with thin grey marls
Solitary corals
Grey marl
Cross-bedded limestone channel (sandy bio-packstone)
Grey marl, sandstone
White bioturbated, cross-bedded, bioclastic sandstone with few ooids, wave ripples
Grey marl, bioclastic limestone, wave ripples
Grey marl, partly bioclastic and oolitic limestones (Sandy bio-packstone, partly winnowed oo-packstone) and brownish, carbonate cemented sandstones
Oolitic limestones (oo-packstone, oo-grainstone)
Yellow and white, partly conglomeratic, cross-bedded sandstones
Road cut floor
Cl Si fS mS cS Co
a
b
c
d
e
Analamanga section (X)
4
2 m
0
Adabomjonga section (VII)
Ankazoabo Formation
Bathonian
Sakaraha Formation
Bajocian
Age classifications suggested by Besairie (1972) and Uhmann (1996)
Recent soil with grassland
Yellow to white (trough) cross-bedded, conglomeratic sandstone
Yellow fine laminated sandstone
Red mudstone with rootlets
Yellow fine laminated sandstone
Yellow to white (trough) cross-bedded sandstone
Yellow to grey planar cross-bedded, conglomeratic sandstone
Fragmented vertebrate bones
Grey sandstone
Grey mudstone
White to grey planar cross-bedded sandstone with ripple lamination
Yellow to white planar cross-bedded, conglomeratic sandstone
Grey bioturbated mudstone (hardground?)
Yellow to white planar cross-bedded, conglomeratic sandstone
Yellow to white planar cross-bedded sandstone
White bioturbated marly limestone (hardground?)
Yellow to white (planar) cross-bedded sandstone
White bioturbated marly limestone (hardground?)
Yellow to white planar cross-bedded sandstone with grey mudstone lenses
Yellow to white planar cross-bedded, conglomeratic sandstone
Yellow dolomite (sandy dolomicrite)
Yellow cross-bedded dolomite (sandy dolomicrite) with mudstone intraclasts
Dark grey silty carbonate mudstone with meteoric cements, cross-bedded sandstone interbeds
Poorly exposed sandstone to siltstone/limestone. Uhmann (1996) described a conglomeratic sandstone channel
River bed
Cl Si fS mS cS Co
f
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h

mudstones (shales and marls) with a thick intercalated? Late Callovian – Early Oxfordian Sandstone.

Toarcian–Aalenian rift basins

At the beginning of the Jurassic marine conditions reached into the East-African- Madagascan domain and partly overstepped onto Karoo rift deposits. Early Toarcian shaly mudstones and siltstones with thin shelly limestones, and more massive limestones in distal oxygen-depleted basins were followed upwards by a prograding shoreface sandstone. Exposures in the Morondava Basin are only localised in the north and their quality and stratigraphic frameworks are poor. Corresponding data, used and discussed in the present paper, are from outcrops in the south-western Majunga Basin (Clark 1996; Geiger et al. 2004).

Bajocian carbonate platform

The Bemaraha Formation of the carbonate platform system is only present at the Bemaraha Plateau (Fig. 2), where the Sakaraha Formation has been entirely eroded. In the south the Bemaraha Formation disappears into the subsurface and only its coastal facies equivalent, the Sakaraha Formation, is exposed (Fig. 4).

Bemaraha Formation

The Bemaraha Formation is made up predominantly of massive limestones (carbonate mudstones, pelletoidal-grainstones or oolitic-grainstones). Typically, they form a lenticular body that runs along the western edge of the Morondava Basin (Clark and Ramanampisoa 2002). The limestones vary in thickness from 30 to 1,000 m and are clearly evident on many seismic lines in the northern Morondava Basin. Clark (1996) and Geiger et al. (2004) provide a comprehensive description and discussion of subsurface data and some additional outcrops. Thicker limestone successions found in the Sakaraha Formation show the interfingering of both lithofacies associations.

Sakaraha Formation

Besairie and Collignon (1972) describe sediments of a mixed carbonate-siliciclastic facies (the so-called Facies Mixte: “mixed facies concept”) in the south-central part of the basin with proposed Bajocian–Bathonian ages. The carbonates are similar to those typically found in the limestones of the Bemaraha Formation, whereas the siliciclastics comprise varying admixtures of mudstone, siltstone, and sandstone. Besairie and Collignon (1972) attributed the regional appearance of the mixed facies to tectonism-induced increase in sediment supply. Siliciclastic successions, such as the Ankazoabo, Sakanavaka, Mandabe and Besabora formations were also included in the mixed facies concept, promoted by stratigraphic uncertainties in all involved formations. This concept strongly influenced stratigraphic nomenclature by SPM in the early 1950s and was widely used by SPM field geologists and for producing geological maps (e.g. Besairie 1969).

Recent studies interpret the Sakaraha Formation as a coastal plain association that accumulated landwards of the barrier/lagoon complex, which belongs to the Bemaraha Formation (Clark 1996; Clark and Ramanampisoa 2002; Geiger et al. 2004). The deposition of the Sakaraha and Bemaraha formations probably terminates during the Early–Middle Bathonian and is followed by the Bathonian sandstones (see below) rather than interfingering with it (Fig. 4).

Sakaraha section (VIIIa) and Anteninde section (VIIIb). The area around Sakaraha is the type locality for the Bajocian Sakaraha Formation (Besairie and Collignon 1972). Uhmann (1996) and Dina (1996) measured a hill section (S22°55.402′/E44°29.983′) along the road RN7 to Toliara (Fig. 2). Above the transgressive contact to the underlying Isalo sandstone, the succession starts with an oolitic-grainstone, followed by a thinly-bedded succession of mudstones, sandstones, and limestones. Calcareous ooids and bioclasts occur in some of the sandstones and limestones. Only some 100 m further west, Anteninde section (S22°55.013′/E44°29.008′) describes the overlying marls in the westward dipping strata.

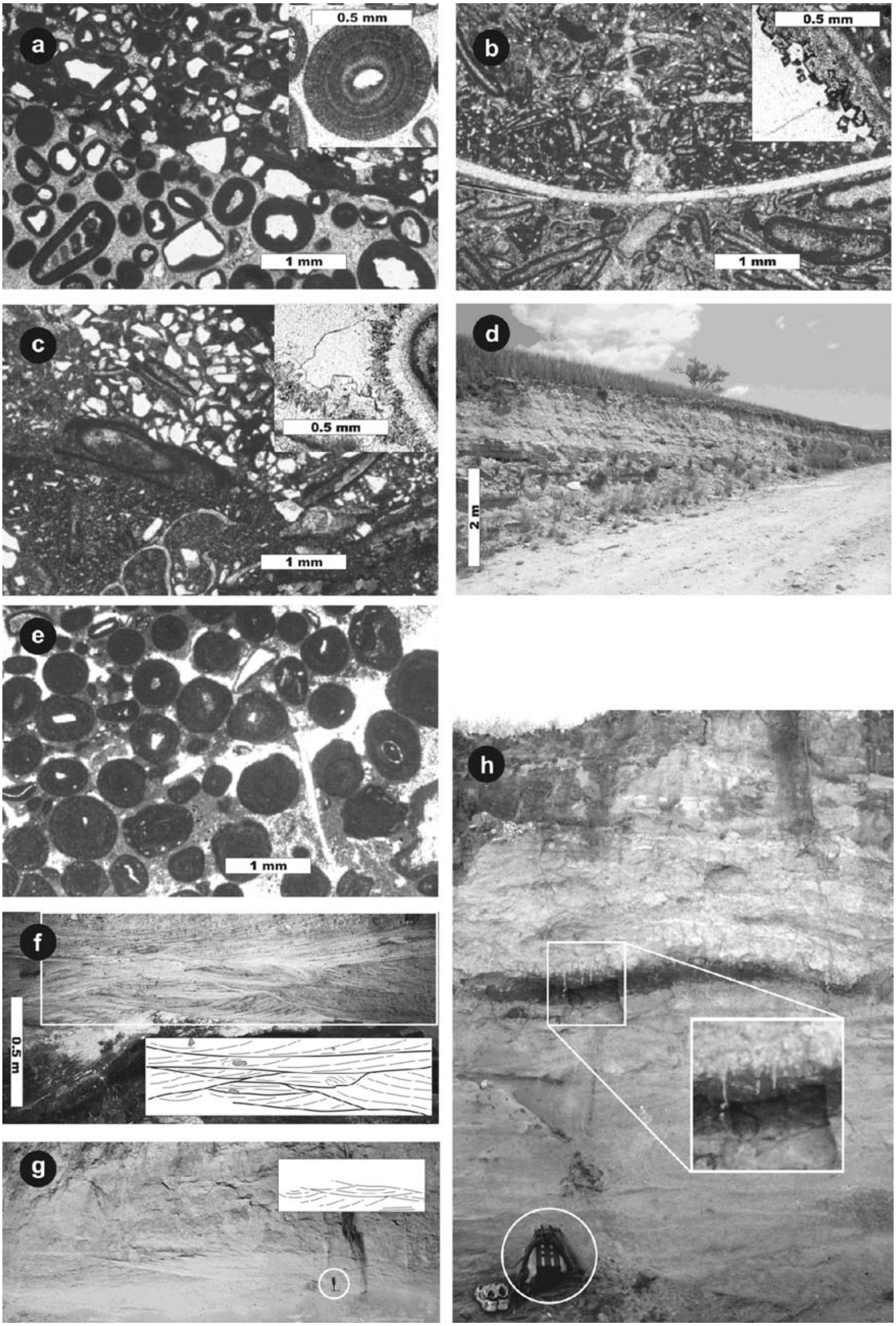
Uhmann (1996) and Dina (1996) interpreted the Sakaraha section as an intertidal to shallow subtidal, open lagoonal environment which is the same as the Anteninde section.

Analamalanga section (X). The Bajocian section at Analamalanga (Fig. 5; S22°51.738′/E44°32.950′) closely resemble those at the type locality at Sakaraha (Geiger et al. 2004). A transgressive oolitic-grainstone unconformity overlies the Isalo sandstone followed by mudstones intercalating with thinner, partly oolitic and bioclastic limestones (packstones and grainstones) and sandstones (Figs. 5 and 6a–d).

Rhynchonellids, bivalves, gastropods, and echinoderms are generally concentrated in the upper and lower parts of the mudstone section. Solitary corals are preserved in one horizon in the central part and possible biohermal structures in the upper part of mudstone unit. Microfossils such as the calcareous algae and lenticuline foraminifers occur in limestones in the upper section. A low diverse ostracod assemblage is also present (Mette and Geiger 2004a).

After the basal transgression, a series of high-energy shallow sand shoals formed during flooding events that reached far landwards. Subtidal, intertidal, supratidal marsh, inner lagoonal, and coastal swamp environments were established (Geiger et al. 2004). The low diversity of the ostracods (Mette and Geiger 2004a) and brackish water bivalve faunas (Geiger et al. 2004) supports the interpretation of intertidal and brackish lagoonal or swamp conditions, whereas the presence of solitary corals and

Fig. 5 Analamanga section (X) illustrates the basal Sakaraha Formation with the contact to underlying Isalo sandstone, whereas Amparambato section (VIb) reveals the sharp boundary to the overlying Ankazoabo Formation. For geographical reference see Fig. 1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bioherms, together with rhynchonellids, is more indicative of subtidal, outer lagoonal conditions. The massive oolite at the top of the section is interpreted as a coastal oolitic barrier complex that marks a major flooding event that shifted the facies belts far landwards.

Anjeba (Ia) and Andamilany (Ib) sections. The Bajocian Anjeba (S23°42.090′/E44°20.607′) and Andamilany (S23°42.265′/E44°20.395′) sections are located west of Betioky (Fig. 2) and were also previously described by Uhmann (1996), where Anjeba represents the upper part of the composed Andamilany section. Geiger et al. (2004) described the Anjeba sandstones, mudstones and partly cross-bedded oolitic and bioclastic limestones. Andamilany contains bioclastic, peloidal and oolitic limestones (mudstones and packstone) and partly conglomeratic, cross-bedded sandstones.

Microfaunas consist of algae, predominantly Dasycladacea, textulariid and lituolid foraminifers (Fig. 7a–f), and ostracods (Mette and Geiger 2004a). Anjeba is topped by a bivalve meadow.

Sedimentological interpretations and faunal assemblages suggest deposition under intertidal to shallow subtidal conditions in a protected lagoonal environment. A hardground, topped with glauconite sandstones, at Andamilany infer a short episode of shallow subtidal conditions.

Bathonian sandstone

In the south-central basin Bathonian sandstone successions rest unconformably on the carbonate platform and appear to be regionally confined. Various formations, e.g. the Ankazoabo and Sakanavaka formations in the south, and Besabora and Mandabe formations in the north, are poorly described and often lack reliable stratigraphic constraints. North of the Tsiribihina River, comparable siliciclastic deposits are unknown. Clark (1996) describes a sandstone lens with off-lapping clinoforms from seismic images that are located almost immediately above the Bemaraha carbonate platform. He attributed them to a set of Upper Jurassic sandstones.

Fig. 6 Photos and photomicrographs from the Analamanga (X) and Adabomjonga (VII) sections. *Analamanga*: **a** Transgressive bioclastic oo-grainstone sometimes with quartz grains as ooid nucleus above Isalo sandstone. **b** Dolomitized bio-grainstone; zoomed inset shows dolomite rhomboids at a rim of a recrystallized shell. **c** Bioclastic fine sandstone and siltstone; zoomed inset depicts marine isopachous cementation, followed by a prismatic meteoric phreatic cementation in various generations. **d** Top of the roadcut section at Analamanga. **e** Bioclastic oo-grainstone sometimes with quartz as ooid nuclei. *Adabomjonga*: **f** Wave-modified bedforms in the cut off walls at the Antenide River, 200 m upstream from the bridge of RN7. Cross-bedding infers turbulent flow as in a tidal environment. **g** Trough cross-bedded sandstone of the Ankazoabo Formation, the *box* outlines prominent bedding planes within the sandstone. River cut parallel south to the RN7, see Hammer for scale. **h** A red clayey palaeosoil with root marks within the *yellow* and *white*, cross-bedded Ankazoabo Formation sandstone, cut off walls at the Antenide River, 500 m upstream from the RN7 bridge. See rucksack for scale

Ankazoabo Formation

At the type locality, in the vicinity of Ankazoabo (Fig. 2), the sandstone succession is interpreted as fluvial and deltaic environment (Besairie and Collignon 1972). A corresponding sandstone facies exists west of Sakaraha at Adabomjonga (VII; Fig. 4), where it directly overlies the Sakaraha Formation. The sandstones lack marine indicators, with the exception of a few calcareous, highly bioturbated horizons. Further south, at Tongobory (III) and Saririaka (IIa), bioclastics of echinoderms indicate an unambiguous marine environment. Stratigraphic distinctive fossils are not known from either the Ankazoabo or the Sakaraha area.

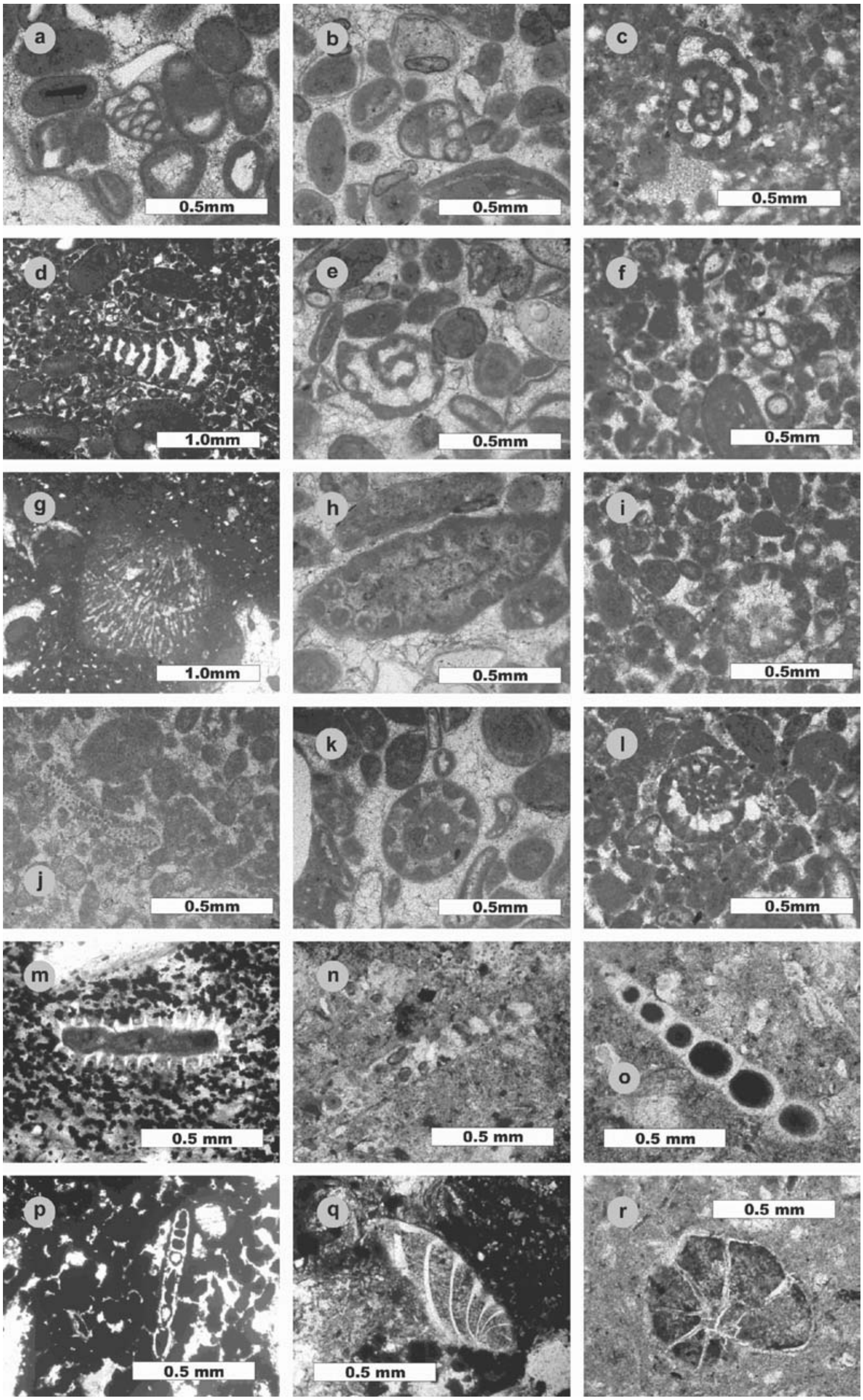
Adabomjonga section (VII). The basal unit of Adabomjonga section (Fig. 6; S22°54.901′/E44°28.618′) is presumably the top of the Sakaraha Formation and consists of thin interbedded mudstones and sandstones, of which the partly ripple laminated sandstones are fining upwards into mudstones. Uhmann (1996) described a metre-thick sandstone channel with conglomerates further down the section but that could not be confirmed. The interbedded sandstones and mudstones are capped by a succession of partly dolomitised conglomeratic sandstones and dolostones. An overlying thick sandstone succession shows several fining upward cycles from conglomerates to trough cross-bedded and finally laminated sandstone (Fig. 6f, g) with claystone interlayers, capped by a horizon of interlacing burrows (*Taenidium* isp.). At the top of the section a red clayey horizon with rhizoids is interpreted as a palaeosoil recording subaerial exposure (Fig. 6h).

Despite the absence of biostratigraphic markers, Besairie and Collignon (1972) and Uhmann (1996), attributed the interbedded mudstones and sandstones, and the dolomites at the base to the Sakaraha Formation with a Bajocian age, while the thick sandstone succession is assigned to the Bathonian Ankazoabo Formation.

Uhmann (1996) interpreted the depositional environment of the basal part (Sakaraha Formation) as supratidal inner lagoonal with intertidal channel. Cross-bedding and ripple lamination indicates wave-influenced subtidal to intertidal sandy shoreface conditions and the mudstones record protected mudflats. Periodic exposure is evidenced by the palaeosoil horizon.

Tongobory section (III). The Tongobory section (Fig. 8; S23°42.265′/E44°20.395′) along both shores of the Onilahy River (Fig. 2) is of Bathonian age (Besairie and Collignon 1972; Uhmann 1996). The section comprises of bioclastic, partly oolitic sandstones with locally planar and trough cross-bedding and bioturbation. Frequently intercalated conglomeratic beds with erosive base contain carbonate mudstone and sandstone lithoclasts, and bioclasts (Fig. 9b). Small and large wave-length wave ripples and sand wave horizons with crest spacings of up to 40 cm are present (Fig. 9a).

Ripples crests and the alignment of wood infer E–W oriented palaeocurrents. Foresets dip predominantly to the east.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The partly oolitic shallow water sandstone at the base was recently interpreted as transition from Sakaraha Formation facies to the Ankazoabo Formation facies, but probably represents more distal facies of the Bathonian Sandstone. Extraformational carbonate mudstone clasts in conglomeratic beds can be reworked debris of the Bajocian carbonate platform. The conglomeratic beds suggest high energy episodes. Coarse, cross-bedded sandstones are interpreted as upper shoreface deposits. Fining upward at the top of the section implies a landward retreating shoreline.

Saririaka (IIa). Saririaka section (Fig. 4; S23°42.125′/E044°18.682′) can be lithologically compared with Tongobory section (III). Local profile descriptions by Besairie and Collignon (1972) suggest a correlation with the Ankazoabo sandstones. The section starts at the base with a thick ferruginous conglomeratic sandstone, containing coarse bioclasts and mudstone and sandstone lithoclasts. The overlying fining-upward, partly bioturbated sandstone is trough cross-bedded.

The basal bioclastic conglomerate contains bone debris and belemnites and bivalves, the latter indicating marine conditions. Fossilised leaf fragments and tree trunks, up to 50 cm in diameter and up to 3 m long, are predominantly WNW–ESE oriented.

The basal conglomerate is interpreted as a transgressive conglomerate with regards to the marine fauna. Trough cross-bedding sandstone succession is interpreted as wave-dominated upper shoreface. The presence of large tree trunks indicates the proximity of the land.

Sakanavaka Formation

The Sakanavaka Formation has its type locality along the Sakanavaka River (20 km NNW of Ankazoabo), where it consists of sandstones with thin mudstone intercalations (Besairie and Collignon 1972). Some horizons are highly fossiliferous with mainly bivalves (*Corbula* sp.) and petrified wood. Ripple marks and cross bedding infer deposition in shallow, agitated water.

Mamakiala section (XIII). The Mamakiala section (Fig. 2; S22°06.924′/E44°26.487′) is the northernmost section at the Mamakiala River. Its biostratigraphic range is widely unknown but overlying mudstone contains ammonites of at least Middle Callovian age. Uhmann (1996) and Dina (1996) described the entire sandstone, siltstone, and mudstone succession, which we found only partly exposed. The outcrop patches yielded partly bioturbated, trough cross-bedded sandstone with mudstone lenses. Wave-ripple lamination was observed in sandstones below the mudstones. Plant debris are common. *Helmintopsis* isp. was found on bedding planes.

Close to Mamakiala village (S22°07.651′/E44°24.634′) a larger outcrop exposes laminated to wavy-bedded sandstone, fine-grained sandstone with ripple lamination in a few places. Conglomeratic beds contain elongated mudstone clasts and cut erosionally into the sandstone succession at several levels. Belemnites, vertebrate bone debris, and bivalve shell debris are found in the conglomerates. Uhmann (1996) also identified the bivalve *Corbula* sp.

The strong interfingering of trough cross-bedded sandstones with mudstones is interpreted as distributary channels framing interdistributary bay deposits. Cross-bedding and coarsening upward cycles reflect a wave dominated shoreface with recurring shallowing. The conglomeratic beds are storm-produced deposits. It is uncertain if the contact to the overlying mudstones is conformable, since it is poorly exposed.

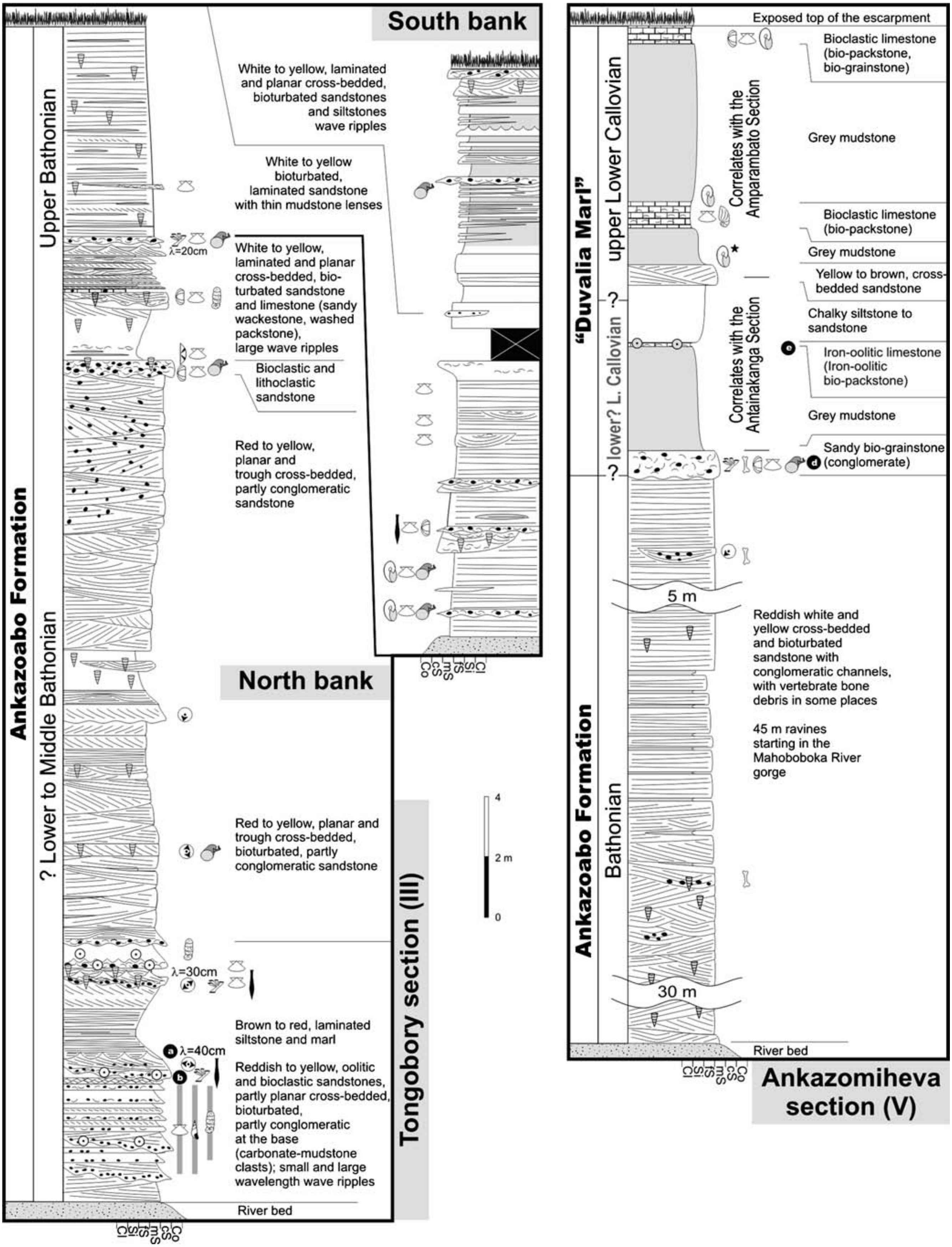
Fig. 7 Thin section microphotographs showing foraminifer and algae assemblage at Anjeba (Ia), Andamilany (Ib) and the Dangovato (IVa) sections. *Bajocian Sakaraha Formation*: Foraminifers: **a** *Textularia* sp., Ib. **b** *Valvulina meentzeni* Klinger, Ia. **c** *Mesoendothyra croatica* Gušic, 1969, Ia. **d** *Haurania* sp., Ia. **e** *Haplophragmoides* sp., Ia. **f** *Valvulina* sp., Ia. Algae: **g** *Rivularia* sp., Ib. **h** *Cylindroporella* sp., Ib. **i** *Terquemella* sp., Ia. **j** *Neomeris* sp., Ia. **k** *Heteroporella* sp., Ia. **l** Indet. *Dasycladacea*, Ia. *Oxfordian Duvalia Marl*, IVa: Ostracod: **m** Cythere. Alga: **n** indet. Dasycladacea. Foraminifers: **o** indet. nodosariid form. **p/q** *Citharina* sp. **r** indet. hyaline rotaliid form. For stratigraphic reference see Fig. 8

Callovian–Kimmeridgian shallow shelf basin

In the Early Callovian, deeper and distal shelf conditions transgressed far landward and formed a thick mudstone succession. Onlapping relationships of corresponding strata have been described from seismic lines (compare Stoakes and Ramanampisoa 1988 and Geiger et al. 2004), where the mudstones can be seen to successively overstep the basin-plain, the slope, and finally the platform carbonates (Clark 1996). In accordance with this, the basal mudstones become progressively younger towards the basin margin. In the deeper part of the basin, the earliest mudstones are possibly even latest Bathonian in age, according to well reports (Clark 1996). The mudstones generally reflect deeper, open marine conditions and are interbedded with sandstones and limestone beds in some places. From the sections at Dangovato (IVa), Antsampagna (IX) and Ankilimena (XI) sandstone of probable Early Oxfordian age is known.

Jurassic Duvalia Marl

Although the name Duvalia Marl originally applied to the Early Cretaceous mudstones (Valanginian – Earliest Aptian?) as introduced by Besairie and Collignon (1972), it became loosely defined thereafter (Clark 1996). Some workers have since included Middle–Late Jurassic and earliest Cretaceous sediments into the Duvalia Marl demonstrating that the depositional history of these sediments is poorly understood. The mudstones are interbedded with thin silt-/sandstones, as well as limestones containing calcareous and iron-oolites and frequently abundant oysters (*Gryphaea* sp.) to form prominent coquinas (“lumachelles”, see Besairie and Collignon 1972).


South bank
White to yellow, laminated and planar cross-bedded, bioturbated sandstones and siltstones wave ripples
White to yellow bioturbated, laminated sandstone with thin mudstone lenses
White to yellow, laminated and planar cross-bedded, bioturbated sandstone and limestone (sandy wackestone, washed packstone), large wave ripples
Bioclastic and lithoclastic sandstone
Red to yellow, planar and trough cross-bedded, partly conglomeratic sandstone
North bank
Red to yellow, planar and trough cross-bedded, bioturbated, partly conglomeratic sandstone
Brown to red, laminated siltstone and marl
Reddish to yellow, oolitic and bioclastic sandstones, partly planar cross-bedded, bioturbated, partly conglomeratic at the base (carbonate-mudstone clasts); small and large wavelength wave ripples
River bed
Upper Bathonian
? Lower to Middle Bathonian
Ankazoabo Formation
λ=20cm
λ=30cm
λ=40cm
Tongobory section (III)
4
2 m
0
Cl Si fS mS cS Co
Exposed top of the escarpment
Bioclastic limestone (bio-packstone, bio-grainstone)
Grey mudstone
Bioclastic limestone (bio-packstone)
Grey mudstone
Yellow to brown, cross-bedded sandstone
Chalky siltstone to sandstone
Iron-oolitic limestone (Iron-oolitic bio-packstone)
Grey mudstone
Sandy bio-grainstone (conglomerate)
Correlates with the Amparambato Section
Correlates with the Antainakanga Section
upper Lower Callovian
lower? L. Callovian
"Duvalia Marl"
5 m
30 m
Reddish white and yellow cross-bedded and bioturbated sandstone with conglomeratic channels, with vertebrate bone debris in some places
45 m ravines starting in the Mahoboboka River gorge
Bathonian
Ankazoabo Formation
River bed
Ankazomiheva section (V)

Ankazomiheva section (V). The Ankazomiheva section (Fig. 8; S22°58.078′/E44°26.941′) consists of Bathonian Sandstone of the Ankazoabo Formation at the bottom and a Lower Callovian mudstone succession at the top. The sandstone is partly bioturbated and trough cross-bedded at the bottom of the section and laminated at the top. Within the laminated sandstone shale lenses and small coarser channels occur. A 1 m thick sandy bioclastic conglomerate (sandy litho-bio-grainstone) transgressively overlies the sandstone and marks the boundary to the overlying Duvalia Marl. The conglomerate is followed by locally fossiliferous mudstones with intercalated cross-bedded sandstones and limestones (partly lithoclast-bearing, oolitic and bioclastic grainstones and packstones).

Vertebrate bone debris were found in the Bathonian Sandstone. The conglomerate contains abundant bioclasts including solitary corals, wood debris, echinoids, and bivalves, but no determinable specimen could be extracted. The mudstones and limestones of the Duvalia Marl contain a few *Lenticulina* sp., macrobenthos and ammonites.

The thick Ankazoabo sandstone at the bottom of the section is interpreted as shoreface deposit. The trough cross-bedding indicates agitated water, as in the upper shoreface. In contrast, the laminated sandstone suggests lower shoreface conditions. The bioclastic conglomerate at the Bajocian–Callovian boundary is interpreted as a transgressive conglomerate and the ammonite fauna in the overlying mudstones and limestones suggests basinal conditions (outer shelf).

Antainakanga section (VIa). The Lower Callovian Antainakanga section (Figs. 2, 10; S22°58.330′/E44°24.247′) starts cross-bedded, well sorted, bioclastic, and carbonate-cemented, bioclastic sandstone which forms a prominent scarp and allows a correlation at outcrop with the Amparambato section (VIb). Above follows bioturbated chalky mudstone succession with abundant bivalve fossils that show an oolitic texture in shell cavities. Several coarsening upward cycles from chalky marlstone to marly siltstone succeed until they are topped by a sandstone bed similar to the one at the base of the section.

Ammonites, bivalves and ostracods are common (Mette and Geiger 2004c). The foraminifer assemblage is highly diverse with a few nodosariid and neoflabelline forms (Geiger 2004).

Ostracods and bivalves reflect normal marine conditions. The bioclastic sandstones at the bottom and at the top of the succession are characteristic for sandbars. Oolites indicate agitated shallow water. Chalk as a product of plankton indicates open marine distal high productive carbonate environments.

Fig. 8 Tongobory section (III) shows a typical siliciclastic lower shoreface setting with shallow marine sandstones of the Ankazoabo Formation. The section is composed of outcrops from the north and the south shore of the Onilahy River. Ankazomiheva (V) section illustrates the boundary to the overlying Callovian mudstones of the basal Duvalia Marl. For geographical reference see Fig. 2

Amparambato section (VIb). At the river cut of Amparambato section (Figs. 2, 10; S23°00.001′/E44°24.929′) the river floor is formed by a highly fossiliferous iron-oolitic limestone and sandstone beds with thin mudstone interlayers. The limestones vary between iron-oolitic bio-wackestone and sandy bio-wackestone (Fig. 11). Mudstone with few thin sandstone and limestone beds overlie the limestone and pass into a thick monotonous mudstone succession.

Within the oolitic limestones at the bottom of the section an abundant ammonite fauna is preserved. The succeeding mudstones contain numerous ostracods (Mette and Geiger 2004c) and foraminifers. Foraminifers are exclusively nodosariid (Geiger 2004).

Ranonda section (IIb). The lithology of the Callovian Ranonda section (Fig. 2; S023°43.438′/E44°18.328′) resembles the one at Antainakanga (VIa). Here, chalky siltstones and mudstones with few interbedded sandstones are topped by an iron-stained limestone (bioclastic oo-packstone).

Chalk points to a distal open marine high productive carbonate environment. The bioclastic oolite at the top of the succession is interpreted as a shoal during a period of shallowing and agitated water conditions.

Beraketa section (IVb). The Middle–Late Oxfordian Beraketa section (Fig. 2; S22°54.901′/E44°28.618′) along the Sakondry River yields dark mudstones (shales) at the base which fade to grey towards the top. Lenses enriched in small, white, thin-shelled bivalve fragments disperse within the dark mudstones. Lenticular calcareous concretions are also present. Red and ochre weathering halos indicate increased iron contents. Several horizons of nodular concretions spread within the middle and upper part of the section.

Concretions contain belemnites and shell and plant debris. Others show bioturbation. Ammonite findings are usually in red fissile clay in the core of a dark nodular concretion. The shells were bleached, white in colour and were highly fragile. In the basal section, a few horizons reveal a highly diverse foraminifer fauna dominated by textulariid and nodosariid forms (Geiger 2004). The middle–upper part of the section contains abundant agglutinating foraminifers which are dominated by lituolid forms.

The dark colour of the mudstones infers a high preservation of organic material which is typical for oxygen-depleted conditions. Ammonites indicate a fully marine environment. In organic-rich, oxygen-depleted, acidic environments calcareous concretions are common. Both interpretations are backed by the foraminifers (see below).

Dangovato section (IVa). The ?Callovian – Early Kimmeridgian Dangovato section (Fig. 2, 12; S23°06.895′/E44°23.254′) along the Dangovato River, a tributary of the Sakondry River, starts at the base with shaly mudstone which contain nodular concretions similar to those of Beraketa section (IVb). Sporadically shelly beds of white thin valve debris occur. The mudstone coarses upwards into a partly cross-bedded and ripple-laminated sandstone, the Oxfordian Sandstone (Fig. 12a). At the top of the sandstone exists a well-developed hardground which is

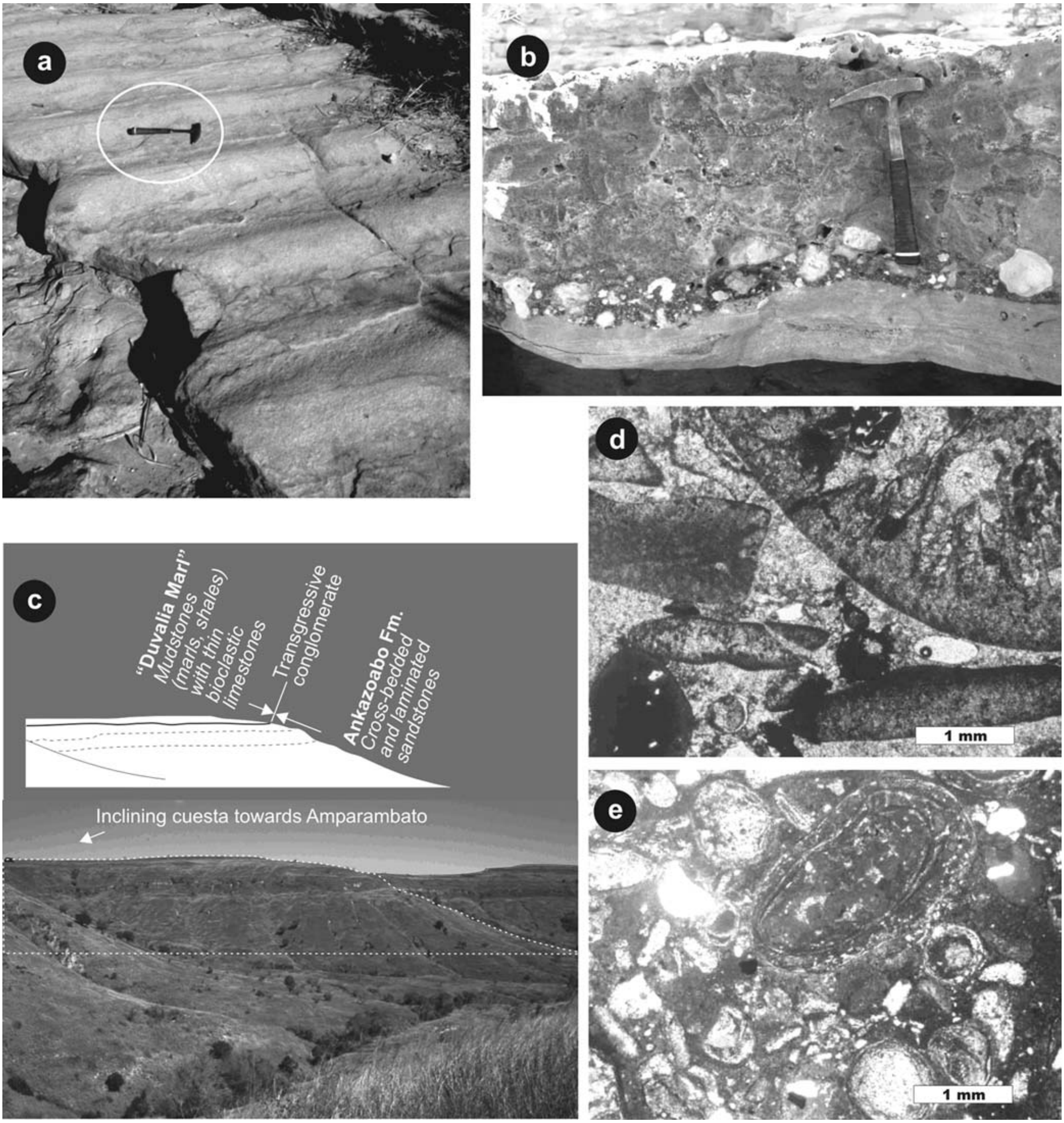


Fig. 9 Photos and photomicrographs from Tongobory (III) and Ankazomiheva (V) sections. *Tongobory*: **a** Sand waves with crest spacing exceeding 40 cm, see hammer for scale. **b** Conglomeratic sandstone with predominantly carbonate mudstones lithoclasts overlies an erosional surface, see hammer for scale. *Ankazomiheva*: **c** Landscape view facing westwards on the Ankazomiheva cuesta. The scarps and ravines provide a section along the slope. **d** Coarse bioclastic debris in a litho-bio-grainstone (bioclastic conglomerate). **e** Sandy iron-oolitic limestone (bio-packstone) with algal oncoid.

overlain by limestone (bio-oo-wackestone). Above succeed interbedded mudstones and limestones. The limestones are often (iron-)oolitic oysters coquinas (Fig. 12b, c). The section terminates with mudstones and a few intercalated thin sandy beds when it is erosionally capped by a major conglomerate (S23°06.066′/E44°22.544′) which is Aptian (Early Cretaceous) in age (Besairie and Collignon 1972; Luger et al. 1994).

Several coalified plant debris, leaf imprints, and belemnites, rhynchonellids, and bivalves were found in the shaly mudstone. The ostracod fauna is highly endemic and stratigraphically and palaeoecologically unspecific (Mette

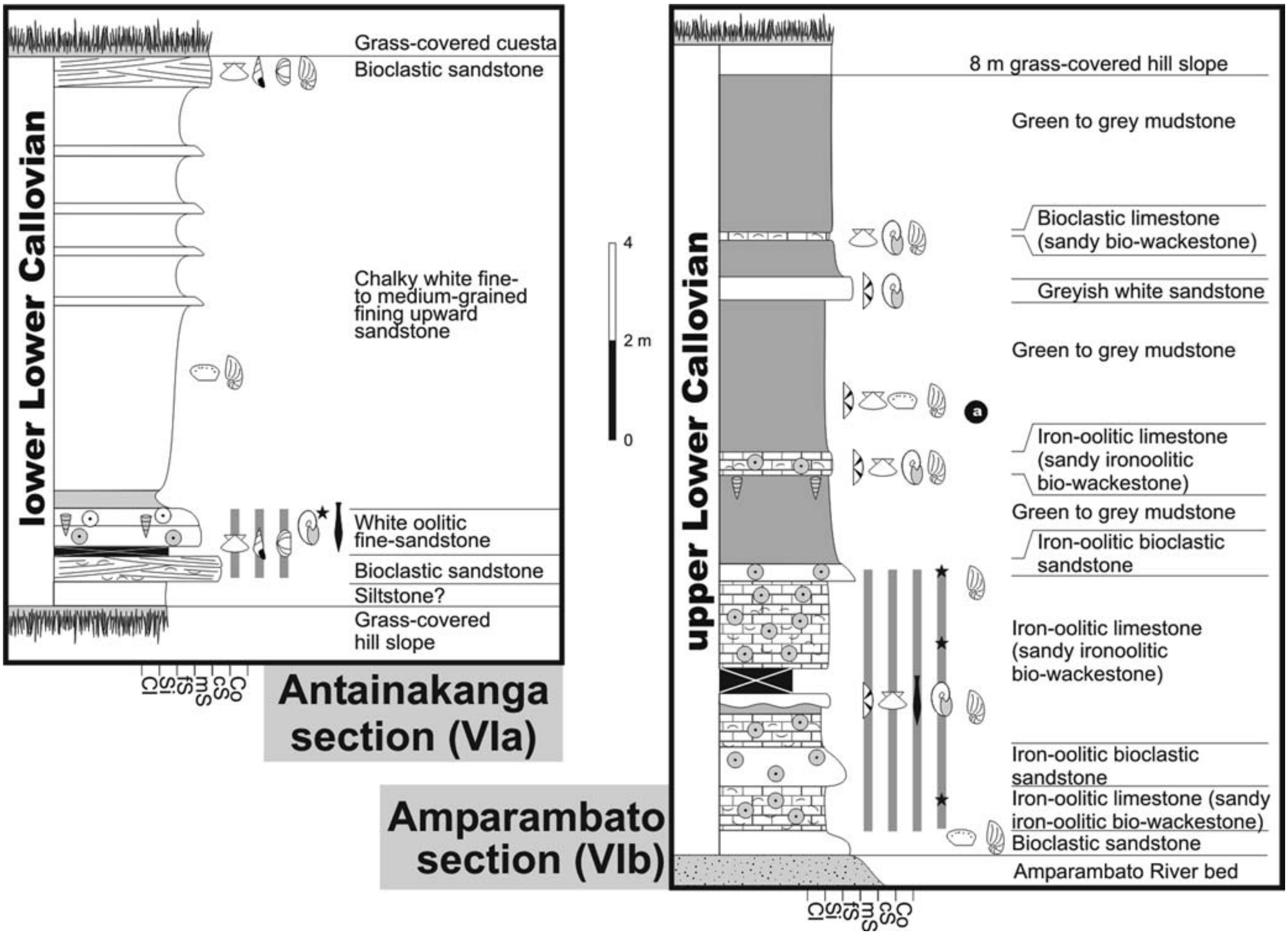


Fig. 10 Lower Callovian Antainakanga (VIa) and Amparambato (VIb) sections illustrate the facies immediately above the Bathonian Sandstone. Compare Ankazomiheva section (V). For legend see Fig. 4

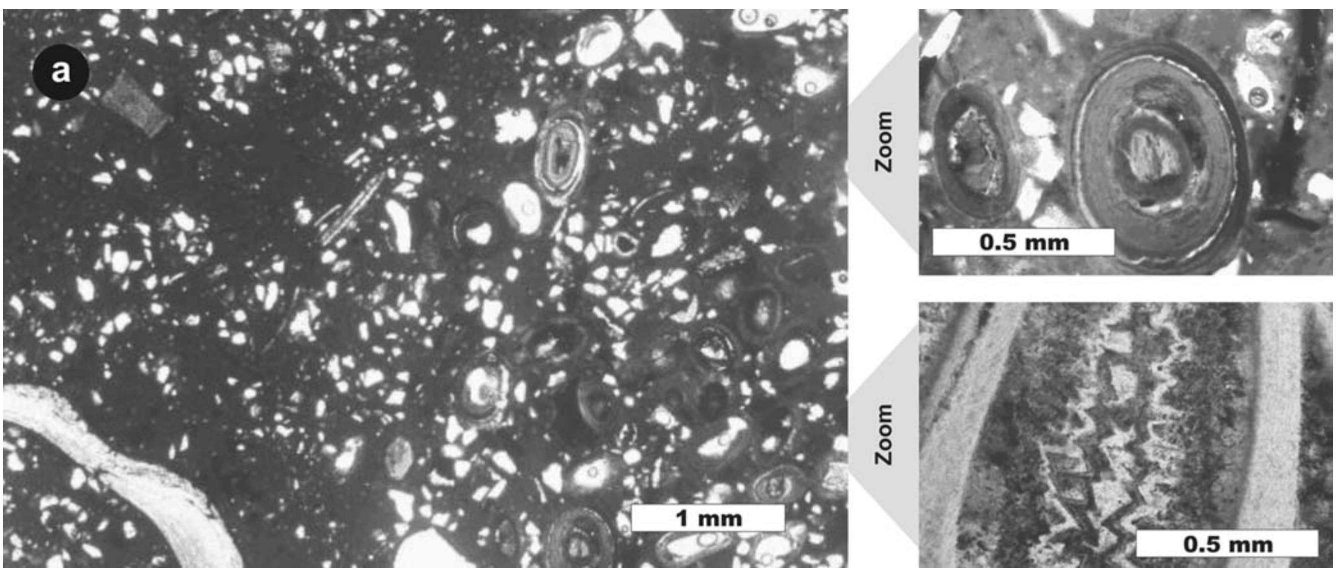


Fig. 11 Silty iron-oolitic bio-wackestone with concentric iron-ooids and bladed calcite rhombus (indicates fresh water mixing). Stratigraphic position as indicated in Amparambato section, Fig. 10

2004; Mette and Geiger 2004b). Shaly mudstones below the Oxfordian Sandstone contain a wide spectrum of agglutinating foraminifers (Geiger 2004). The assemblage is dominated by lituolid forms with accessory textulariid and saccamminid foraminifers. In the interbedding mudstone–limestone succession above the Oxfordian Sandstone, agglutinating forms are successively replaced by calcareous forms (Geiger 2004). In Kimmeridgian samples agglutinating forms reoccur but calcareous forms remain predominant.

Calcareous concretions, microbenthos (see below) and the black colour of the shaly mudstones below the Oxfordian Sandstone reflect dysoxic basinal conditions. The coarsening upward cycle into the Oxfordian Sandstone displays prograding shoreface conditions. The Oxfordian mudstones, and oolitic coquinas and limestones represent distal shallow-water conditions (see below). The comparably thin Oxfordian interval suggests condensation and emphasizes shallow water with erosion and restricted accommodation space.

Antsampagna section (IX). The Antsampagna section (Fig. 2; S22°52.470′/E44°24.950′) is a ground section across the road RN7 from Sakaraha to Toliara, 8 km ENE of Mahaboboka village and 16 km W of Sakaraha village. The section covers the Upper Oxfordian – Lower Kimmeridgian. The section starts 100 m north of the road at the upper bank of the Fiherenana River with cross-bedded red weathering sandstone succession. Clasts are aligned to the bedding planes and bioturbation is present at several levels. This sandstone succession is topped by an erosional surface which is succeeded by a grey mudstone succession of 4 m thickness. At top the mudstone grades into 10 m white sandstone with lamination at the bottom, bioturbated, cross-bedding in the middle part and a massive appearance at the top. About 20 cm of bioclastic coarse-grained sandstone overlies the succession erosionally. This bioclastic sandstone is considered to constitute the inclining surface on which RN7 is built. A few metres south of RN7, 15 cm of pisoidal mudstone is probably overlying the sandstone. The overlying 22 m succession contains mainly mudstones with a few dm-thick iron-oolitic limestones (iron-oolitic bio-packstones). At the top of this succession several cm-thick limestones (pel-packstone) are intercalating with the mudstone.

Petrified wood debris were discovered in the sandstone at the bottom and in the limestone at the top of the section. A few marine bivalves, belemnites and ammonites are present in sandstones and limestones. Several samples revealed foraminifers which are dominated by similar lituolid forms than in the Beraketa (IVb) and Dangovato (IVa) sections (Geiger 2004). *Epistomina* sp. was identified in the

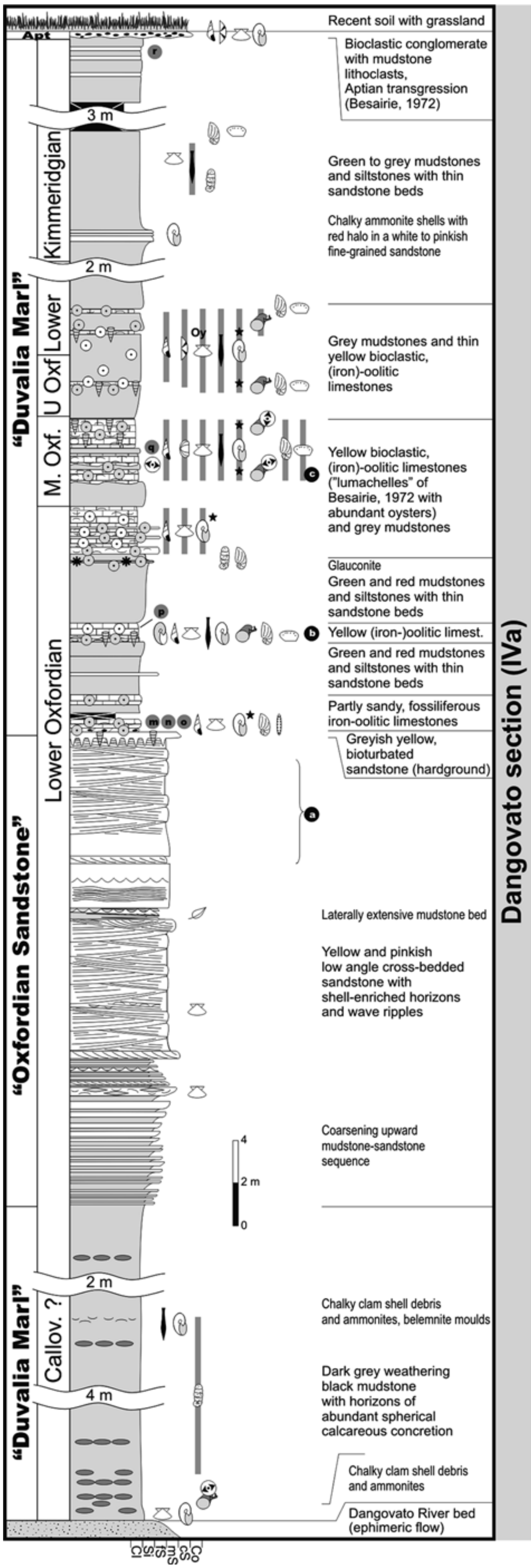


Fig. 12 Dangovato section (IVa) is the stratigraphically most extensive section covering the Lower Oxfordian – Kimmeridgian interval. The top is cut off transgressively by the Aptian conglomerate (Besairie and Collignon 1972). It is the only section to contain the complete succession of the Oxfordian Sandstone. Letters in *black circles* refer to photographs in Fig. 12. Letters in *grey circles* refer to photographs in Fig. 7

mudstone interlayer between the sandstones at the bottom. *Discobotellina* sp. occurs in one horizon at the very bottom of the thick mudstone succession further up. Pisoidal limestone at the top of the sandstone is an emersion horizon which is followed by a transgressional mudstone succession with open marine conditions but episodes of shallowing (oolites).

Ankilimena section (XI). At the Middle–Upper Oxfordian Ankilimena section (Fig. 2; S22°45.417′/E44°25.096′) a cross-bedded sandstone forms the base of the section. The sandstone grades into a mudstone succession which is overlain by a 2 m thick coquina (iron-oolitic bio-packstone) of disarticulated oyster shells and other reworked inner neritic components. Above the coquina mudstone, bioclastic sandstone and recurring iron-oolitic limestone are succeeding. Overlying mudstones contain thin gypsum interlayers. An iron-oolitic limestone (bio-packstone) forms a morphological step at the scarp. The steep slope is apparently of monotonous mudstone which sometimes contains bioclasts. At the top chalky limestone forms the roof of the scarp. Trough and planar cross-bedded sandstone channels with bioclastic horizons bound to bedding planes and wave ripples are spread within the chalk.

Ammonites, rhynchonellids, bivalves, belemnites, echinoderms and wood debris are found in and close to the limestone beds throughout the succession, which elsewhere is devoid of macrofossils. But some lituolid foraminifers were found in the mudstones but nodosariid forms dominate the assemblages associated with the limestones (Geiger 2004), together with ostracod assemblages (Mette and Geiger 2004b).

The sandstone at the base is fining upwards into the mudstone, representing a transition from lower shoreface to deeper shelf deposits. The monotypic character of the macrofossil assemblage within the thick oyster coquina suggests a period of shallowing. This is supported by the oolitic texture of the coquina which infers agitated shallow-water conditions. Similar shallowing events are represented by the iron-oolitic limestones in the succeeding limestones. Chalk production at the top of section infers high productive conditions in distal shallow water. Embedded sandstone channels are of intertidal origin during shallowing.

Andrea section (XII). The Middle Oxfordian mudstone succession at Andrea section (Fig. 2; S22°25.695′/E44°23.459′) is comparable to the Dangovato (IVa), Ankilimena (XI), Antsampagna (IX) and Beraketa (IVb) sections. The Lower–Middle Oxfordian Sandstone known from these correlative sections, is not directly recognised to underlie the section but approximately 2 km NNE at Ampasy, in the Andrea River-bed, 4 m cross-bedded sandstone is exposed. Outcrop correlation to the Andrea section lithological comparison provides an argument for classifying it as underlying Andrea section.

The entire 45 m succession comprises mudstone with only a few iron-oolitic limestone (bio-packstones) and sandstone beds. Spherical concretions align in several horizons. In a few places the mudstone contains common shell fragments. Iron-ooids were also recognised in concretionary horizons in the mudstone. Dina (1996) and Uhmann (1996) measured the same section, but they describe a conglomerate that tops the section at about the central part. This conglomerate was found to represent a Holocene river bed, which cuts into the Jurassic section. However, the section extends further upstream until a thick cross-bedded sandstone truncates the mudstones. Besairie and Collignon (1972) described Oxfordian–Kimmeridgian and Cretaceous sandstones at Manama, which both can be correlative.

The limestones and in places also the mudstones contain a macrofauna of ammonites, bivalves, belemnites and plant debris. Foraminifer assemblages are characterised by a frequent change between lituolid and calcareous assemblages (Geiger 2004) with only a few mixed assemblages. Stratigraphically and palaeoecological insignificant ostracods are described by Mette and Geiger (2004b).

The mudstones represent an outer shelf environment (see above). Episodic shallowing is documented by the oolitic limestones. Besairie and Collignon (1972) described the sandstone at the top of the section as a fluvial sandstone.

Oxfordian sandstone

Several of previously described sections contain a part of or are interbedded with a more or less prominent sandstone succession of probably Early Callovian age (see above). The yellow–pink sandstones are generally well sorted but contain conglomeratic beds. Cross-bedding is a common feature and ripple lamination indicates shallow-water conditions. Common plant debris records a proximal position to the land.

Besairie and Collignon (1972) included several Late Jurassic sandstones in the southern Morondava Basin to what he called "Argovian". Clark (1996) correlated those sandstones with the Sakanavaka and Ankazoabo formations which are considered to have a Bathonian age. With the term "Argovian," as a subdivision of the Lower Oxfordian, Besairie and Collignon (1972) indicated the supposed age, but stratigraphic indicators are missing. The stratigraphy of this study suggests that the age of the sandstones at Dangovato (IVa), Ankilimena (XI) and Antsampagna (IX) ranges in the Late Callovian – Early Oxfordian interval and thus correlate with the "Argovian". They are herein addressed as Oxfordian Sandstone.

A second sandstone unit above the Oxfordian mudstones, such as it is present in the Andrea section, appears as a localised phenomenon. There are three possible explanations: Firstly, the sandstone is not Jurassic, but Cretaceous in age (probably accounts to Andrea section). South of the Onilahy River, Cretaceous sandstones occur, which have a similar lithology and the same order of thickness. Secondly, the Aptian transgression cut off corresponding strata in the very south. The oldest preserved age of Early Kimmeridgian, however, infers a Kimmeridgian age for the sandstone. Thirdly, regional tectonism formed localised facies contrasts.

Iron-ooid formation, ocean floor spreading and sequence stratigraphy

From the Callovian onwards the presence of frequent iron-oolites points to changes in sea water chemistry and thus the depositional environment. Sturesson et al. (2000) propose that iron ooids form in a shallow turbid sea directly next to volcanism. In contrast, Yoshida et al. (1998) suggest rather a non-marine or brackish environment to promote iron-ooid formation. Oxyhydroxides of iron and aluminium, silicon oxide and calcium carbonate are the main components of fossil and modern iron-ooids (Sturesson et al. 2000). Fluid plumes from modern mid-ocean-ridge vents are found to carry increased contents of iron, silicon and various oxides, hydroxides, sulphates and sulphides (Hekinian et al. 1980; Edmond et al. 1982). This argues for a connection of the appearance of iron-ooids and ocean-ridge volcanism from the Callovian onwards and consequently establishes a connection to the onset of sea-floor spreading at minimum Early Callovian. This passive dating of sea-floor spreading, however, enhances the usually cited (e.g., de Wit 2003) oldest oceanic crust measured by magnetic sea-floor anomaly M25 from the Oxfordian/Kimmeridgian boundary (Coffin and Rabinowitz 1992; calibration by Pálfy et al. 2000).

Iron-oolites are often discussed to be of sequence stratigraphic relevance (Jervey 1988; Loutit et al. 1988; Kidwell 1989; van Wagoner et al. 1990). Macquaker et al. (1996) conclude that iron-oolites form in shallow conditions where sediment supply is low and detrital iron is available. This applies to sequence boundaries (e.g. partly Dangovato IVa), major flooding surfaces or maximum flooding surfaces (e.g. Amparambato VIb) and thus limits the unambiguous practicability.

In western India the Oxfordian Dhosa Oolite Member of highly condensed oolites is associated with hardgrounds, reworked concretions, intraformational conglomerates, shell lags, iron crusts and iron oncoids (Fürsich et al. 1992). The top of the member is interpreted as a maximum flooding surface of a relative sea-level highstand. This highstand is a result of the combined Callovian transgression and thermal subsidence after the Gondwana Breakup.

With regards to its sequence stratigraphic relevance, the Oxfordian oolites of Madagascar mark a transgressive event above the Oxfordian Sandstones and the low siliciclastic contamination argues for the deposition in a distal shelf position. The interaction of thermal subsidence and eustacy is unclear. Major regional tectonism has not been proved by thermochronological studies during this time (Emmel et al. 2004).

Transgressive-regressive cycles

The study and intercontinental correlation of sedimentary sequences during the Jurassic is generally hampered by the lack of global biostratigraphic correlation schemes. In Madagascar the poor outcrop conditions and the lack of biostratigraphic completeness require a considerable amount of generalisation. This also means that the destination to pin down events to a biostratigraphic zone must be dismissed. Hallam (2001) outlined that such a generalisation necessarily means that the applicability of sedimentary patterns for interpreting global sea-level changes has to be reduced to second-order cycles though he conceded the influence of eustacy on third–order transgressive-regressive cycles (Embry 1993).

The first transgressive-regressive cycle (T-R1) in the Morondava Basin covers the syn-breakup phase with the transgressive Toarcian shales and marls, followed by the regressive Aalenian Sandstone (Fig. 13). The Bajocian – Early Bathonian marginal carbonate platform sequence which includes the Bemaraha and Sakaraha formations overlies the breakup unconformity and forms T2. It is followed above by the Middle–Late Bathonian siliciclastic sequence (R2) with the Sakanavaka, Ankazoabo, Besabora and Mandabe formations. T3 covers the Early Callovian – Early Oxfordian interval with a deep–shallow shelf sequence of the Jurassic Duvalia Marl. T3 is overlain by the siliciclastic sequence R3 of the Oxfordian Sandstone. Sequence T4 represents an Early Oxfordian – Kimmeridgian/Tithonian deep-shallow shelf environment and is also assigned to the Jurassic Duvalia Marl. The upper part of T4 is diachronically truncated by several Cretaceous transgressions (Luger et al. 1994).

Toarcian–Aalenian cycle (T-R1)

The syn-breakup succession of the Andafia and Beronono formations and the Aalenian Sandstone as described by Geiger et al. (2004) illustrate a transgression and a maximum flooding (Luger et al. 1994) during the Early Toarcian. The successions are characterised by dark shaly mudstones with several thin fossiliferous limestone beds. The shales are rich in organic material and are interpreted to be deposited in oxygen-depleted conditions (Clark 1996; Geiger et al. 2004). Toarcian anoxia has been explained by several global (e.g. Bailey et al. 2003) and regional models (e.g. Röhl et al. 2001). The anoxia coincides with a widespread transgression in the *Falciferum* Zone (Hallam 2001) which is well known from Europe (Hallam 1981; Jacquin et al. 1998), South America (Legarreta and Uliana 1996), Siberia (Zakharov et al. 1998), Middle East (Hallam 1981), and northern Pakistan (Fatmi 1972).

Shallowing and regression in the Aalenian is recorded by the overlying Aalenian Sandstone which are poorly constrained by biostratigraphic markers. The shales grade successively into the 25 m thick cross-bedded sandstone. In the Morondava Basin a transgressive surface indicates a sudden change as a result of basin-wide flooding reaching far onto the rift margins, whereas the transition into the overlying post-rift carbonate platform happens more or less successively at the outcrops in the south-western Majunga Basin (compare Besairie and Collignon 1972; Geiger et al. 2004). Seismic images from the Morondava Basin (Geiger et al. 2004) show the T-R1 strata with syn-rift reflection divergence in half-graben structures (Fig. 14) and infer the coincidental syn-breakup rifting. Such a syn-breakup

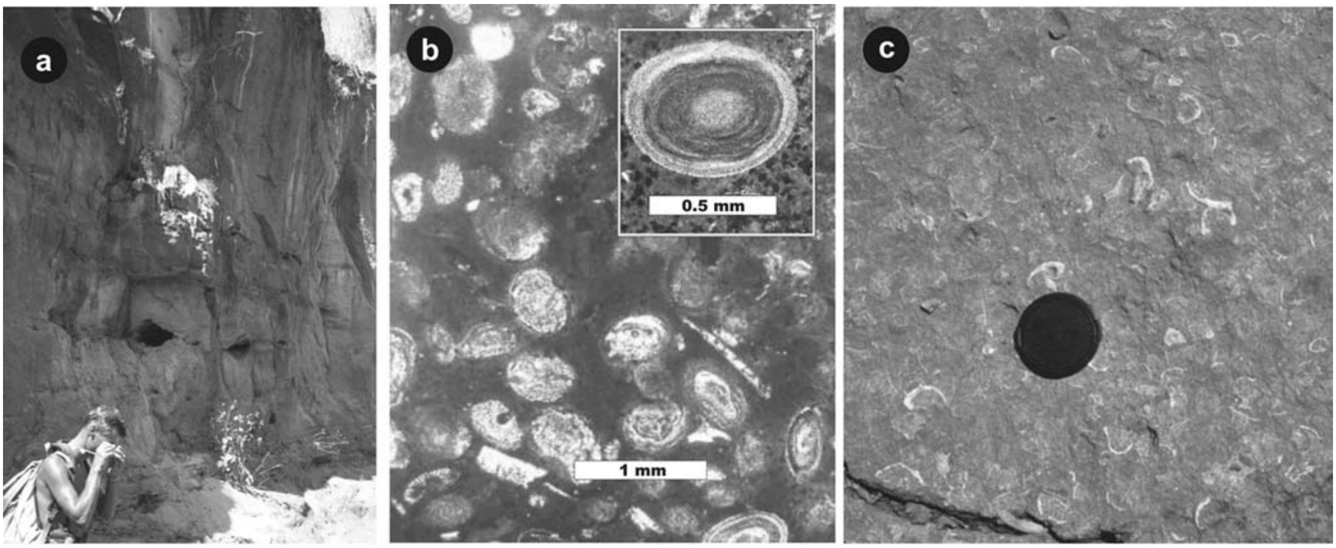


Fig. 13 **a** Dangovato river gorge cuts through the Oxfordian Sandstone. **b** Thin section microphotograph of an iron-oolitic coquina in the Lower Oxfordian; the inset illustrates the concentric character of the ooids. **c** Top surface of an iron-oolitic coquina; the *white* shells are predominantly oysters; see lens cap for scale (7 cm diameter)

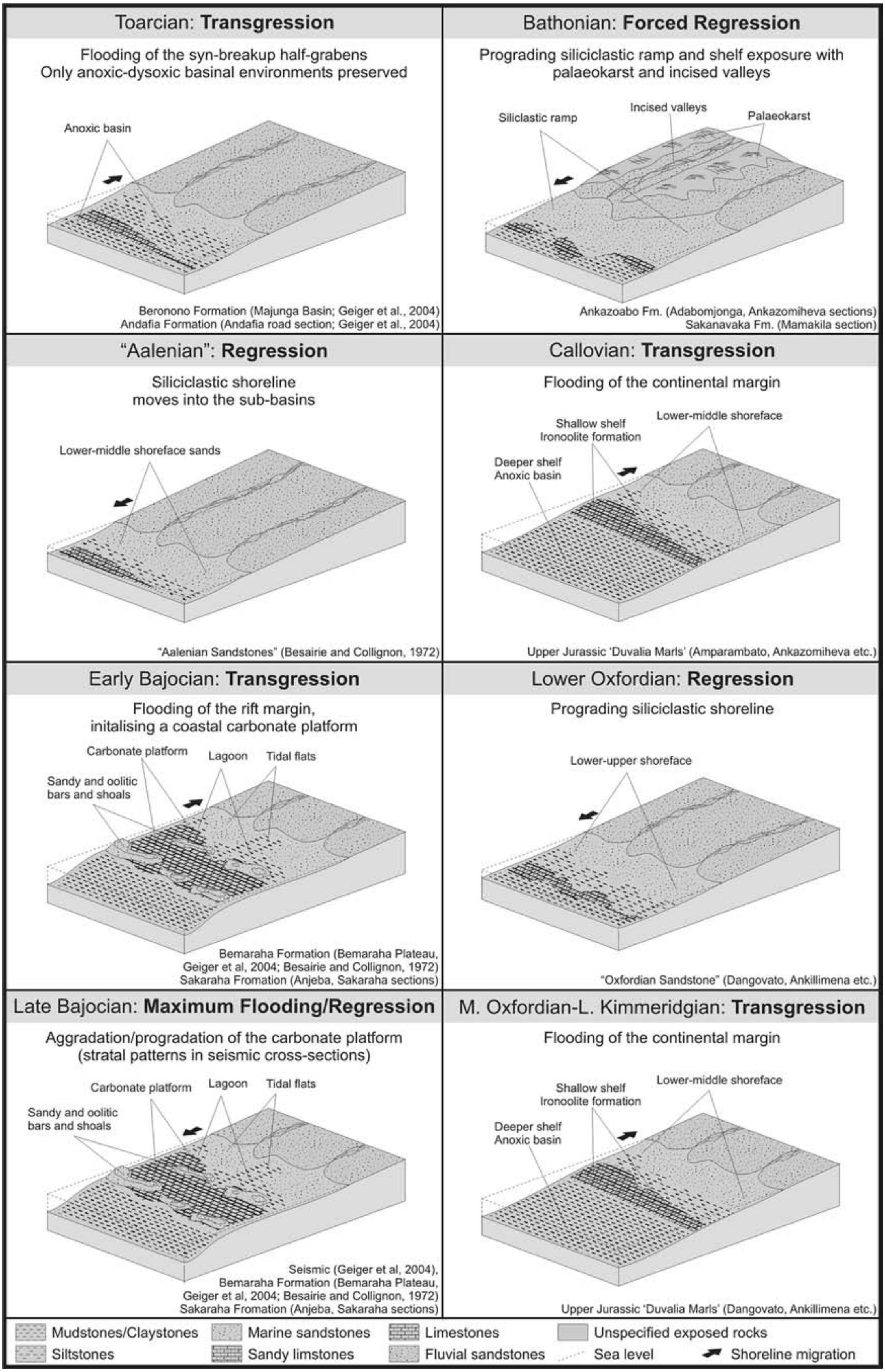
character was inferred by Kreuser (1995) for the possibly? Hettangian–Toarcian prograding deltaic Ngerengere beds in Tanzania. The Ngerengere beds lithologically correlate with the Aalenian Sandstone. In the sequence stratigraphic framework of Somalia the lower Hamanlei Depositional Sequence and the Meregh Fromation (Bosellini 1992) is comparable with T-R1. The presumed more basinal Meregh Formation derived from the first marine ingression is considered to be Pliensbachian (Bosellini 1992). Lithologically, the Meregh Formation resembles the Andafia and Berono mudstones in Madagascar. In the global context the regression with the Aalenian Sandstone deposition is concordant to a regressive event in Europe (Jacquin et al. 1998) while elsewhere in the world (Argentina: Legarreta and Uliana 1996; Himalayas: Li and Grant-Mackie 1993) evidence is rather poor (Hallam 2001).

Early Bajocian – Early Bathonian transgressive cycle (T2)

With a global transgression in the Early Bajocian (Hallam 2001) marine conditions were introduced to the entire basin and the margins were flooded far inland. The transgressive surface unconformably overlies sandstones which belong to the Isalo Formation (Analamanga section X and Sakaraha section VIIIa). Du Toit et al. (1997) described this surface as the MAD 6 seismic marker and put it at the Lower/Middle Jurassic boundary but Geiger et al. (2004) showed that it represents a hiatus ranging almost throughout the Early Jurassic till Toarcian times. The transgressive surface coincides with the breakup unconformity and marks the onset of post-breakup deposition (Geiger et al. 2004). The overlying carbonate succession shows several T-R cycles of higher order at the Bemaraha plateau in the northern part of the basin (Clark 1996) which probably caused the quick facies changes within the Bajocian in the south (Analamanga X).

Similar to Madagascar, the East African margin was widely covered with carbonate-dominated marine deposits of the Lugoba Formation (Kapilima 1984), partly of the Mtumbi and Kidugallo formations of Tanzania (Kreuser 1995), the Baidoa Member of the Baidao Formation of southern Somalia (Kassim et al. 2002) and the Hamanlei of north-eastern Somalia (Bosellini 1992). The transgressive character can often be seen by the overstepping architecture onto basement or older sedimentary strata (Mbede 1991). In Tanzania the major transgression is sometimes dated to have started during the Middle Aalenian with Ruvu Formation (Kapilima 1984) which can be correlated with Kidugallo Formation of Kreuser (1995). A similar basinward younging of the transgressive surface is possible in the south-western Majunga Basin where the transition from Aalenian Sandstone into more distal mudstones and limestones predates the Bajocian (Besairie and Collignon 1972; Geiger et al. 2004).

Comparable transgressive events in the Early Bajocian are known from Europe (Hallam 1988), the Andes (Legarreta and Uliana 1996) and possibly from the Himalayas (Westermann and Callomon 1988; Gradstein et al. 1991). A Late Bajocian transgression which has been suggested from northern Europe (Jacquin and de Graciansky 1998), Greenland (Surlyk 1991) and Argentina (Legarreta and Uliana 1996) has not been clearly recognised in Madagascar. However, changes in the dinoflagellate fraction from the Upper Bajocian is interpreted by Dina (1996) as a regression which is possibly part of the an Upper Bajocian T-R cycle.



Toarcian: Transgression
Flooding of the syn-breakup half-grabens
Only anoxic-dysoxic basinal environments preserved
Anoxic basin
Beronono Formation (Majunga Basin; Geiger et al., 2004)
Andafia Formation (Andafia road section; Geiger et al., 2004)
Bathonian: Forced Regression
Prograding siliciclastic ramp and shelf exposure with palaeokarst and incised valleys
Siliclastic ramp
Incised valleys
Palaeokarst
Ankazoabo Fm. (Adabomjonga, Ankazomiheva sections)
Sakanavaka Fm. (Mamakila section)
"Aalenian": Regression
Siliciclastic shoreline moves into the sub-basins
Lower-middle shoreface sands
"Aalenian Sandstones" (Besairie and Collignon, 1972)
Callovian: Transgression
Flooding of the continental margin
Shallow shelf
Ironoolite formation
Lower-middle shoreface
Deeper shelf
Anoxic basin
Upper Jurassic 'Duvalia Marls' (Amparambato, Ankazomiheva etc.)
Early Bajocian: Transgression
Flooding of the rift margin, initalising a coastal carbonate platform
Carbonate platform
Lagoon
Tidal flats
Sandy and oolitic bars and shoals
Bemaraha Formation (Bemaraha Plateau, Geiger et al, 2004; Besairie and Collignon, 1972)
Sakaraha Fromation (Anjeba, Sakaraha sections)
Lower Oxfordian: Regression
Prograding siliciclastic shoreline
Lower-upper shoreface
"Oxfordian Sandstone" (Dangovato, Ankillimena etc.)
Late Bajocian: Maximum Flooding/Regression
Aggradation/progradation of the carbonate platform (stratal patterns in seismic cross-sections)
Carbonate platform
Lagoon
Tidal flats
Sandy and oolitic bars and shoals
Seismic (Geiger et al, 2004),
Bemaraha Formation (Bemaraha Plateau, Geiger et al, 2004; Besairie and Collignon, 1972)
Sakaraha Fromation (Anjeba, Sakaraha sections)
M. Oxfordian-L. Kimmeridgian: Transgression
Flooding of the continental margin
Shallow shelf
Ironoolite formation
Lower-middle shoreface
Deeper shelf
Anoxic basin
Upper Jurassic 'Duvalia Marls' (Dangovato, Ankillimena etc.)
Mudstones/Claystones
Marine sandstones
Limestones
Unspecified exposed rocks
Siltstones
Sandy limstones
Fluvial sandstones
Sea level
Shoreline migration

Middle–Late Bathonian regressive cycle (R2)

Regarding the accuracy of the Bathonian stratigraphy of the study area, massive sandstones related to a regressional event are presumed to be of an Early/Middle–Late Bathonian age. The sandstones were usually included together with carbonate platform to a Bajocian–Bathonian mixed facies sequence (Besairie and Collignon 1972; Luger et al. 1994; Montenat et al. 1996).

A key locality is at Adabomjonga (VII), where Bathonian sandstones of the Ankazoabo Formation erosionally overly a dolomite succession that is assigned to the Sakaraha Formation of the underlying carbonate platform. Besairie and Collignon (1972) and Uhmann (1996) suggest a Lower to Middle Bathonian age for the unconformity. It sharply separates the Bajocian carbonate platform deposits from the overlying Bathonian sandstones. Such a sharp regressive contact points to a forced regression. In the north of the basin in a loosely defined Bathonian interval Stoakes and Ramanampisoa (1988) described also a gently undulating surface at the top of the carbonate platform sequence with a number of broad valleys cut into it. This relief was formed during a widespread exposure of the platform when karstification and river incision at the top of the Bemaraha limestone took place (Pierce and Yeaman 1986; Stoakes and Ramanampisoa 1988). Stratigraphic correlation ranges from the predating deposition of the Bemaraha and Sakaraha formations to a postdating sandstone succession that is considered to be unspecified Bathonian. The combined occurrence of a sharp contact between a regressive and a transgressive facies, the exposure of wide areas of the shelf, and incised valleys is a strong argument for forced regression (Posamentier and Vail 1988; van Wagoner et al. 1990). The Bathonian sandstones infer a proximal, shallow-water environment (e.g. Tongobory III and Ankazomiheva V). Pisoidal limestones at the top of the sandstone at Antsampagna (IX) are interpreted as calcareous crusts which were produced during soil-forming emersion episodes.

However, the role of the Bathonian sandstones in the southern basin is still not completely understood. Localised descriptions of Bathonian sandstones, e.g. Ankazoabo and Mandabe formations (Besairie and Collignon 1972; Luger et al. 1990) could be further indicators for incised valley fills. However, the extension of the sandstones is still unknown.

The Bathonian regression of Madagascar represents a strong correlation to signals in other parts of the world. In Somali the stratigraphic correlative Goloda Member of the Baidoa Formation is rather indicative of a transgressive phase during the Late Bathonian – Early Callovian (Kassim et al. 2002). Widespread reef formation in Tanzania infers that the transgression died out (Kapilima 1984). For more global comparison, Riccardi (1983) reports a general regression in western Argentina and Chile. Legarreta and Uliana (1996) recognized a widespread discontinuity at the base of a Bathonian regressive succession in central Argentina and identified shelfal exposure and incision. They interpret the features as a forced regression, which exactly reflects the situation in Madagascar. Also the Tethyan signal illustrates a regressive cycle (Jacquin et al. 1998). Nevertheless, Hallam (2001) believes that a Bathonian regression is globally anomalous and clearly marks a regional and not an eustatic event.

Fig. 14 Model of the sedimentary environments during the syn-breakup and post-breakup T-R cycles of the Morondava Basin. Data basis was extended as indicated

Callovian transgressive cycle (T3)

The coarse bioclastic, sandy limestone at the base of the mudstone succession (*Bullatus* Zone: Early Callovian age) marks the beginning of the transgression T3, while the transgressive surface below it represents the sequence boundary (Fig. 14). Nevertheless, Tongobory section (III) shows deepening at the top of the Middle–Late Bathonian sandstone succession prior to the major transgression. *Epsitimona* sp. in mudstones at Antsampagna (IX) are typical for sudden floodings (Gordon 1970). Iron-oolitic coquinas immediate at the base of the Early Callovian mudstones (Amparambato, VIb) point to initial shallow conditions. Proximochorate and for the first-time chorate dinocysts in the succeeding mudstones document deepening of the water (Dina 1996). Chalky mudstones, such as those at Antainakanga (VIa) and Ranonda (IIb), represent a distal, highly marine environment. Abundant ammonites emphasize the open marine character. Further up in the stratigraphically uncertain Late Callovian to early Early Oxfordian at Dangovato (IVa) just below the Oxfordian Sandstone, mudstones with an apparently high content of organic material are classified at outcrop as black shales. Oxygen-depleted deeper water conditions point to an episode of maximum flooding.

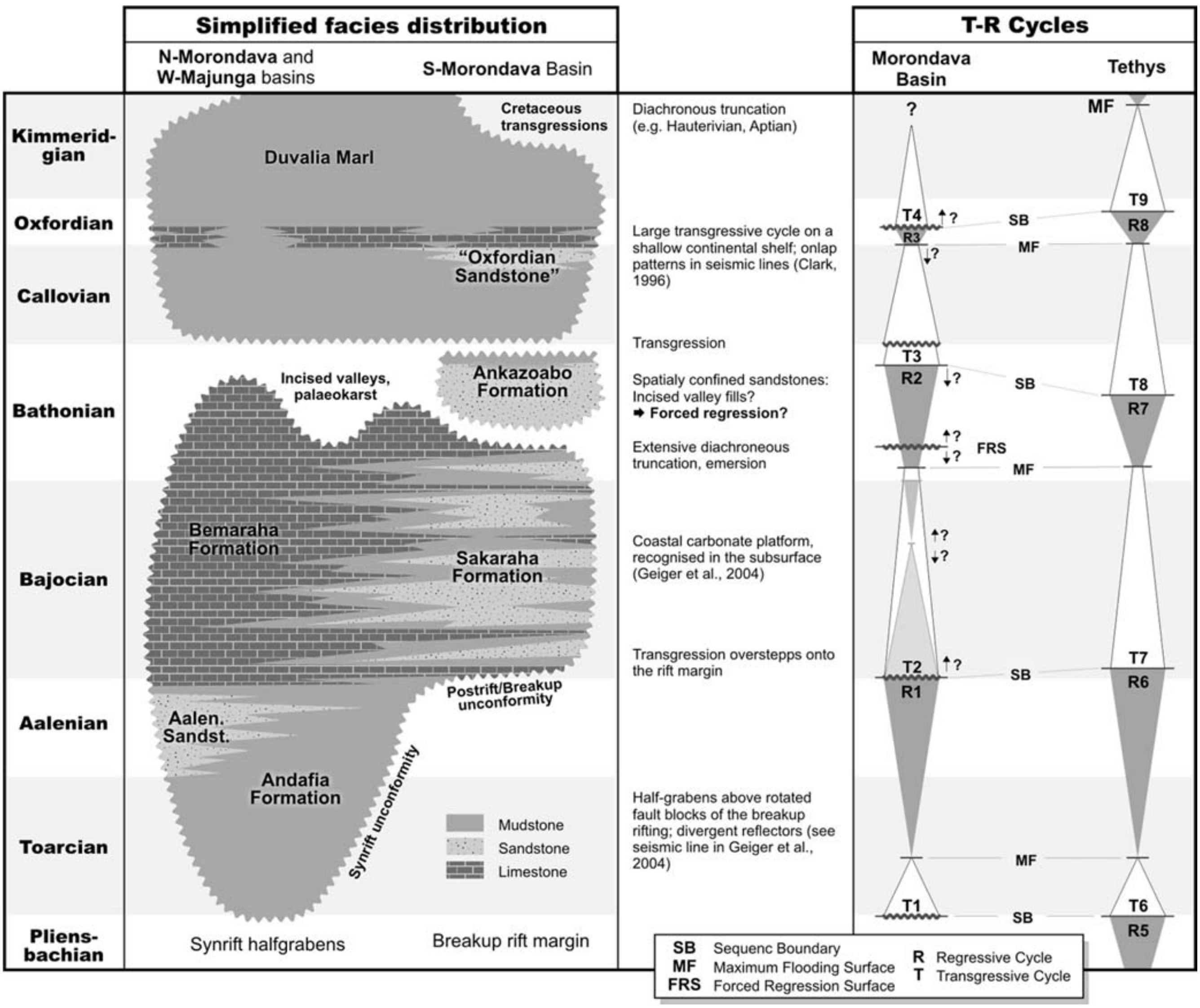
The lithological contrast between the Bathonian sandstones and the overlying Callovian mudstones above is clearly determinable on seismic images and is widely known as sequence boundary. It probably corresponds to seismic marker MAD 7 recognised by du Toit et al. (1997) between the Middle and Upper Jurassic.

In a regional context, major transgressions due to facies changes to more basinal mudstones during the Early Callovian are interpreted from the Anóóle Formation from south-western Somalia (Kassim et al. 2002), from the Pindiro Shales of Tanzania (Kreuser 1995), from the shales of the Malivundo and Magindu formations in Tanzania (Kapilima 1984) and from the Chari Formation in western India (Biswas 1980; Fürsich et al. 1991). Transgression and deepening from the Early Callovian onwards and the deposition of organic-rich shales is a global phenomenon (Hallam 2001). It is locally suggested that this transgressive event already started in the Late Bathonian (Greenland: Surlyk 1991; Europe: Jacquin et al. 1998). A twofold stacking of transgression surfaces in the Upper Bathonian and Lower Callovian and the occurrence of limestones (Amparambato VIb) and sandstones (Antainakanga VIa and Ankazomiheva V) correlate with the southern Morondava Basin (Fig. 14). However, it is in contrast to a sharp regression at the Bathonian/Callovian boundary in the Andes (Legarreta and Uliana 1996) and other parts of the world (Hallam 1988). In Europe (Jacquin et al. 1998) and Greenland (Surlyk 1991) the major sea-level rise was during the Middle Callovian. Iron-oolites at the base of the transgression in Europe (Jacquin et al. 1998), the Himalayas (Li and Grant-Mackie 1993), Pakistan (Fatmi 1972) and India (Gaetani and Garzanti 1991) are in correspondence with the study area and appear to be a world-wide characteristic. The transgressive facies development in Madagascar follows a more or less global trend that is also manifested in the Tethyan cycle pattern (Jacquin et al. 1998).

Late Callovian? – Early Oxfordian regressive cycle (R3)

Nodule-bearing mudstones pass successively upwards into the Oxfordian Sandstone at Dangovato (IVa). This documents the progressive shallowing from deeper shelf to siliciclastic shoreface environment. The age is uncertain but is probably between *Athleta* and *Mariae* Zones (Dangovato and Mamakiala, XIII sections). This sandstone unit probably is also the base at Antsampagna (IX) and Ankilimena (XI). At the top of the sandstone succession at Dangovato a hardground forms the sequence boundary where bioturbation reaches 20 cm into the top surface.

In East Africa a coeval regression has not been clearly described. In the Uegit Formation of Somalia, Kassim et al. (2002) found a regression during Late Oxfordian – Tithonian times but stratigraphic determinations are poor. Further north, Bosellini (1992) combines mudstones with oolitic coquinas of the Uarandab Formation and oolitic and sandy limestones and mudstones of the Gabredarre Formation (Late Oxfordian – Kimmeridgian) to the Uarandab Sequence. At the base of the Uarandab Sequence a transgressive surface is determined as Late Callovian in age (Bosellini 1992). Late Callovian fine-grained siliciclastic sediments in north-western India sug-



gest persistent flooding with hemicycles (Fürsich et al. 1991), although the initiation of the transgressive cycle is uncertain.

A minor regressive event during Early Oxfordian times is known from the Tethyan realm (Jacquin et al. 1998) but has not been recognised on a global scale (Hallam 2001).

Early Oxfordian – Kimmeridgian transgressive cycle (T4)

The Early Oxfordian transgression is stratigraphically well determined at Dangovato section (IVa) with a reliable ammonite age of Early Oxfordian (lower *Cordatum* Zone). Shaly mudstones overlie a hardground at the top of the Oxfordian Sandstone. The same situation has been observed at Ankilimena (XI) and Antsampagna (IX). The succeeding sequence, e.g. at Dangovato (IVa), is rich in iron-oolitic coquinas, marls and limestones and the thin biostratigraphic intervals indicate condensation. Despite unsatisfying stratigraphic control, other sections further north, like Antsampagna (IX) and Andrea (XII), show increasingly thicker mudstone successions with sparsely interbedded thin limestones containing iron-ooids. Iron-oolites argue for shallow agitated-water conditions in the surrounding areas. Thicker stratigraphic intervals could be the result of a basinward increase of accommodation space and a decrease in wave-induced erosion. T4 is concordant with an increased diversity of dinoflagellates and acritarchs in the Oxfordian assemblages indicating marine deep-water environment (Dina 1996). The Kimmeridgian at Dangovato and Antsampagna is devoid of iron-ooids but the mudstones become siltier and indicate increased sediment supply in distal lower shoreface. Calcareous foraminifers, partly with thin tests, indicate normal oxygenation. These are a precursor of a regression. In some places Kimmeridgian and Tithonian palynomorphs argue for a shallowing of the water (Dina 1996). In contrast, localities to the north and towards the basin margin show a reverse trend, partly with exclusively terrestrial floras at the base. In the Upper Jurassic also glauconite, which regenerates in entirely marine shallow shelf conditions (Chamley 1989), becomes more frequent (Uhmann 1996).

The East African biostratigraphy and lithostratigraphy becomes more inaccurate during the Upper Jurassic probably due to poor outcrop quality. In Tanzania mainly shales and evaporites of the Pindiro Shales are found in the coastal basins (Kreuser 1995) and shales and sandstones of the Malivundo Formation (Kapilima 1984) further inland. In Somalia, dark shaly mudstones with oolitic coquinas in parts of the Late Callovian – Late Oxfordian Uarandab Formation and the Oxfordian part of the Gabredarre Formation indicate condensation prior to a maximum flooding during Oxfordian times (Bosellini 1992). Highly condensed iron-oolites of the Dhosa Oolite Member (*Cordatum* Zone) in north-western India (Fürsich et al. 1992) closely resemble those of Madagascar in facies and age. The Late Oxfordian – Tithonian regression of south-western Somalia (Kassim et al. 2002) and a regression within the Kimmeridgian–Tithonian Gabredarre Formation (Bosellini 1992) is a contrast to the prevailing transgressive development in the studied successions.

The Upper Callovian – Lower Oxfordian deposits appear to be intensively condensed in several parts of the world (Norris and Hallam 1995) including the Andes, where the hiatus can be pinched down to the late *Athleta* to early *Cordatum* Zone interval (Legarreta and Uliana 1996). This more or less coincides with a possible age of the Oxfordian Sandstone and the succeeding mudstones with iron-oolitic coquinas. Compared to the Madagascan sea-level change, transgression in the Tethys and other parts of the world are only known from the Late Oxfordian (Jacquin et al. 1998; Hallam 2001).

Fig. 15 Composed lithostratigraphy and interpreted succession of local T-R cycles compared to Tethyan T-R Tethyan T-R cycles (after Jacquin et al. 1998). Facies distribution is mainly based on the present work with supplementary data from Besairie and Collignon (1972), Clark (1996) and Montenat et al. (1996)

Conclusions and discussions

The rifted passive margin along western Madagascar experienced several intensive palaeoenvironmental changes. These changes reflect T-R cycles which all can be related to cycles in other parts of the world. This "global" correspondence hampers interpretations of the influence of regional tectonism on depositional systems and suggests two scenarios: (1) Tectonic control on the depositional environment was too small in comparison to eustacy to have imprinted in the sedimentary patterns, or (2) tectonism coincides with eustacy.

During the syn-breakup event the region was flooded for the first time in the Jurassic and led to the deposition of dark shales. Together with shoreface sandstones of the so-called Aalenian Sandstone this T-R1 cycle coincides with the global sealevel. Possibly slow subsidence and attenuated tectonism produced relatively thin syn-breakup strata with a pronounced marine character. The absence of prominent rift shoulder uplift explains denudation rates beyond thermochronological resolution (Emmel et al. 2004) and suggests a shallow rift system similar to Galicia and ancient Adria (Manatschal and Bernoulli 1999; Wilson et al. 2001).

The T2 reached far onto the rifted margin and formed the carbonate platform. A clear distinction whether this transgression responds to the global Early Bajocian sea-level rise alone or is amplified by expected thermal subsidence of the rift margin is unclear.

The Bathonian R2 has an apparent correspondence with a global eustacy signal. A basinward migration of the siliciclastic shoreline due to increased sediment influx as response to uplift/denudation cannot be determined by thermochronological methods (Emmel et al. 2004).

More critical is the T3. An apparent start of ocean floor production and from the Callovian onwards, could have induced a flexurally downwrapping of the margin by the heavier ocean floor. On the other hand, post-rift thermal subsidence possibly promoted the transgression. However, a strong correlation of T3 with global sea-level changes argues for a major role of eustacy, which at least superimposes an additional mechanism.

The Oxfordian Sandstone follows a regressive event (R3) that is recognizable in other parts of the world. Poor outcrop conditions of the Callovian–Oxfordian boundary interval constrict further models. The following T3 reflects a global event and furthermore demonstrates the relation of the depositional sequences to global eustacy.

Acknowledgements The authors are indebted to A. Dina and T. Razakamanana (University of Toliara) for logistic support and assistance during fieldwork. We would like to thank D. N. Clark (Clark Research Ltd.) for fruitful discussions, F. Fürsich (University of Würzburg) for the identification of the bivalves and their palaeoecological interpretation, and J. Kuss for critically reviewing the manuscript. Research was funded by the German Research Foundation (DFG), project Ku 642/18.

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