SEDIMENTARY AND STRATAL PATTERNS IN JURASSIC SUCCESSIONS OF WESTERN MADAGASCAR

Facies, Stratigraphy, and Architecture of Gondwana Breakup and Drift Sequences

Markus Geiger

Front cover
Karst ridges, pinnacles, and crevices of Jurassic limestones (Bemaraha Plateau) at the Tsingy de Bemaraha Nature Reserve, Morondava Basin (western Madagascar). Insets: Manambolo Gorge through the Bemaraha Plateau; seismic section ML-84-21 (Shell) from the Majunga Basin (north-western Madagascar).

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Eventually, all things merge into one, and a river runs through it. The river was cut by the world's great flood and runs over rocks from the basement of time. On some of the rocks are timeless raindrops. Under the rocks are the words, and some of the words are theirs. I am haunted by waters.

Norman Maclean

Preface

This thesis primarily consists of two publications, which give public access to the results of sedimentological and stratigraphical studies on Jurassic strata in the Morondava Basin, western Madagascar. Each of those publications comprises a chapter in this thesis:

Chapter 2: Geiger, M., Clark, D.N., Mette, W., 2004. Reappraisal of the timing of the break-up of Gondwana based on sedimentological and seismic evidence from the Morondava Basin, SW Madagascar. *Journal of African Earth Sciences*, 38(4): 363-381.

The first publication (Chapter 2) describes the litho- and biostratigraphy and the sedimentary patterns of Karoo-related (Carboniferous-Triassic) and Early-Middle Jurassic (Toarcian-Bajocian) strata in the Morondava Basin. Based on these studies combined with seismic stratigraphy, the paper presents evidence to discriminate the Late Palaeozoic-Late Triassic polyphase Karoo rifting from the Early Jurassic Gondwana breakup rift. The argumentation is based on facies analysis involving lithological and palaeontological interpretations. In combination with subsurface data, a basin-wide depositional concept is established, which relates sedimentary successions to three breakup phases of Gondwana: the pre-, syn-, and post-breakup phases.
Own contribution: field work, facies analyses and interpretation (sedimentary patterns, microfacies, fossils), compilation of a biostratigraphic framework and its discussion, correlations, establishing a tectono-sedimentary model using seismic sections and well data, text and figures.

Chapter 3: Geiger, M., Schweigert, G., submitted. Toarcian-Kimmeridgian depositional cycles of the south-western Morondava Basin along the rifted continental margin of Madagascar. *Facies*.

The second publication (Chapter 3) continues this interpretation and describes the newly formed continental margin after the breakup of Gondwana and during the translational separation, i.e. the drifting. A biostratigraphic framework is established in the study area based on ammonites. Facies interpretation, based on sedimentological and palaeontological indicators, leads to a model of transgressive-regressive (T-R) cycles due to relative sealevel changes. After the syn-breakup T-R cycle in the Early Toarcian-Aalenian, three post-breakup T-R cycles are recognised with transgressions in the Early Bajocian, Early Callovian, and Early Oxfordian. Basin-wide correlations demonstrate the large-scale application of the models. Global correlations of the T-R cycles indicate coherence of regional and global events.
Own contribution: field work, facies analyses and interpretation (sedimentary patterns, microfacies, fossils), palaeoecological interpretations with emphasis on benthic foraminifers and other micro- and macrobenthos, sequence stratigraphic interpretations and discussions, text and figures.

Chapter 4 presents additional basin analytic and geodynamic studies. The sub-chapters present additional geodynamic reviews and considerations affecting the breakup of Gondwana and the subsequent continental drift involving seaway aspects. Moreover, further basin analytic contributions, including isopach models, palaeo-current analyses, and regional considerations on the structure of the breakup rift in comparison with rift basin models enhance the concept of the formation of the Morondava Basin. Two subchapters contain summarized data presented in joint publications:

Chapter 4.5: Three joint publications with Dr. W. Mette, Innsbruck:

Mette, W. and Geiger, M., 2004. Bajocian and Bathonian ostracods and depositional environments in Madagascar (Morondava basin and southern Majunga Basin). *Beringeria*, 34: 37-56.

Mette, W. and Geiger, M., 2004. Taxonomy and palaeoenvironments of Callovian ostracoda from the Morondava Basin (south-west Madagascar). *Beringeria*, 34: 57-87.

Mette, W. and Geiger, M., 2004. Middle Oxfordian to early Kimmeridgian ostracoda and depositional environments of south-west Madagascar. *Beringeria*, 34: 89-115.

These publications deal with the taxonomy, stratigraphic distribution, and environmental implications of ostracod assemblages to consider palaeogeographical distribution patterns. In combination with palaeogeographical information from literature based on sedimentological and macrofaunal studies in the East African domain, the direction of the successive marine invasion and the influence of the continental separation on the palaeogeography are discussed. The results also give suggestions to the opening direction of the seaway between East- and West-Gondwana (Proto-Indian Ocean). I have essentially contributed to the litho- and biostratigraphic framework (latter in cooperation with Dr. G. Schweigert, Stuttgart) as well as to environmental implications based on sedimentological studies at outcrop and palaeontological interpretations based on micro- and macrofaunas.

Chapter 4.9.2: A joint publication with Dr. B. Emmel and Dr. J. Jacobs, Bremen:

Emmel, B., Geiger, M. and Jacobs, J., submitted. Detrital apatite fission-track ages in Middle Jurassic strata at the rifted margin of W-Madagascar - indicator for a protracted resedimentation history. *Sedimentary Geology*.

The publication presents sediment provenance studies on the basis of detrital apatite fission track analysis. Detrital apatite fission-track ages are also used to study the basin-filling history with emphasis on resedimentation processes by comparing detrital apatite fission-track ages with stratigraphic ages. I provided the sedimentological concepts of basin filling, stratigraphic frameworks, palaeo-current data, and perceptions of the basin geometry.

Contents

Summary

Crustal extension and eventual separation of the eastern and western part of the Gondwana supercontinent left imprints in the Jurassic sedimentary record in the Morondava Basin of western Madagascar. The sedimentary successions were studied with an interdisciplinary approach, based on interpretations of sedimentary textures (outcrop) as well as microfacies interpretations (outcrop and thin sections), stratigraphic studies (ammonites), palaeoecological interpretations (macro- and microfossil), interpretations of the stratal architectures (seismic images), provenance studies and resedimentation studies (detrital apatite fission track analyses). Sedimentological and stratigraphical analyses elucidate local and regional palaeo-environmental perturbations and relative sea-level changes. Seismic images allow the nature and temporal order of tectonic events to be interpreted. The combination of sedimentological and seismic concepts results in a tectono-sedimentary model of the breakup of Gondwana.

During the Late Carboniferous-Late Triassic extensional basins formed all across Gondwana. In contrast to earlier concepts, in which Gondwana breakup was seen as a multiphase, long-lasting process during Late Carboniferous-Late Triassic times, the present study infers that the breakup episode can be constrained to the Toarcian-Aalenian interval (late Early Jurassic).

Seismic images from the northern Morondava Basin illustrate that the Late Carboniferous-Late Triassic Karoo succession is involved in breakup faulting with fault block rotation and therefore is by definition a pre-breakup strata. The typical syn-breakup strata are formed by the Andafia Formation (Toarcian) and the Aalenian Sandstones, when they form divergent strata of shales, limestones, and sandstones, deposited in half-grabens above tilted blocks. Observations from outcrop also suggest a basinward repeating, eastward-tilting of the strata and an eastward pinching out of the Andafia Formation. The Andafia Formation in the Morondava Basin correlates with strata of the Beronono Formation in the Majunga Basin. There, the oldest syn-breakup deposits in Madagascar are classified to the *Bouleiceras nitescens Zone* (Early Toarcian). A major unconformity, which is overlain by Early Bajocian sediments, marks the end of this early syn-breakup phase, and is followed by post-breakup deposits of the drift phase. The post-rift phase is characterised by the initial formation of a coastal carbonate platform (Bajocian), which extends along the entire continental margin followed by a siliciclastic sequence of shoreface deposits (Bathonian) and subsequently by basinal shales and intercalating shallow water limestones (Callovian-Kimmeridgian).

During breakup rifting and drifting four major transgressive events, each followed by a regression, were recognised to mainly follow eustatic sea-level changes and to be only minor affected by regional tectonism:

- Early Toarcian
- Early Bajocian
- Early Callovian
- Early Oxfordian

The first transgressive-regressive (T-R) cycle coincided with the Toarcian syn-breakup rifting. Apart from this three T-R cycles are documented in the passive margin succession. An Early Bajocian transgression overstepped the rift flank of the breakup eastwards far inland and covered Toarcian-Aalenian syn-breakup rift strata and extensive parts of the pre-breakup Karoo strata. At the same time a coastal carbonate platform attached to the newly formed continental margin. Previous studies have failed to recognize the basin-wide extent of the carbonate platform (mudstones, limestones, and oolites Bemaraha Formation), because only the coastal equivalents (mudstones, limestones, siltstones, and sandstones of the Sakaraha Formation) are exposed in the southern Morondava Basin.

A major regressive event in the Early-Middle Bathonian resulted in a partial exposure of basinal strata in the northern part of the Morondava Basin (Stoakes and Ramanampisoa, 1988). There palaeokarstification and incised valleys on top of the carbonate platform are known from seismic lines in the north. At the same time shallow water siliciclastics were deposited in the south to form the sandstone-dominated Ankazoabo and Sakanavaka formations. The combination of basin exposure, incised valleys, and localised siliciclastic wedges are interpreted as typical forced regression features.

A widespread transgressive unconformity introduces deeper marine conditions again in the Early Callovian (*Bullatus Zone*) with mudstones, siltstones, limestones, and a few iron-oolites at the base. Deeper basinal conditions prevail, with the exception of prominent shallow water sandstones which post-date the *Athleta Zone* (Late Callovian) and predate the *Plicatilis Zone* (Middle Oxfordian). This “Oxfordian Sandstone” documents a short regression forcing the siliciclastic shoreline to prograde into the basin. The base of the youngest Jurassic transgression of the southern Morondava Basin (*Plicatilis Zone*) is characterised by the occurrence of iron-oolitic limestones. Stratigraphically upwards mudstones and shales are intercalated with bioclastic and iron-oolitic limestones.

The three post-rift T-R cycles are basically comparable with the second-order cycles of the Tethys. Biostratigraphic bias of the Ankazoabo Formation and of the “Oxfordian Sandstone” hampers a precise comparison. However, the Jurassic sea-level curves of both the Morondava Basin and of the Tethys show strong similarities and suggest that the eustatic signal is superimposed on the local sea-level curve.

The depositional concepts derived from studies at outcrop and subsurface are complemented by the compilation of facies maps and isopach maps for the Morondava Basin. Facies maps suggest a basinwide occurrence of the Bajocian carbonate platform, whereas in contrast a pure siliciclastic environment is present during the Oxfordian. Isopach maps of Bajocian-Bathonian and Callovian-Tithonian intervals outline a strong difference in sediment thickness in the southern part of the basin compared to the north. Both time intervals imply two depocentres in the south-central part of the basin, whereas in the north strata are thinner for a large area.

Palaeo-current indicators from the south Morondava Basin suggest a general westward sediment transportation and a N-S trending shoreline during the Middle and Late Jurassic. During the Bajocian-Bathonian a northward directed transport is inferred at the very southern end of the Morondava Basin, where local uplift is probable. During Oxfordian times, at the northern edge of the southern central part of the basin, southward-directed palaeo-currents may indicate an elevated basin margin. Evidence for slightly uplifted marginal areas can also be found at the southern basin margin by the comparison of stratigraphic ages of Bajocian-Callovian sediments with apatite fission track (AFT) ages of detrital apatite grains. The stratigraphic order of detrital AFT ages can only has been formed by a stratigraphic reversal, possibly due to reworking of former Karoo deposits. The erosion of Karoo sediments may refer to uplift of the margin of the breakup basin, which is lined by Karoo deposits.

Considerations of biogeographical vicariance based on new ostracod data indicate high endemism in Madagascar compared to East Africa and other Jurassic findings. Spreading ridge formation and dysoxic basinal conditions in the Proto-Indian Ocean formed a possible migration barrier between Madagascar and the African coast.

Reappraisal of published regional ocean floor ages and consideration of Middle Jurassic volcanism in the region suggests that the maximum age of the Proto-Indian Ocean floor is approximately 180 Ma. This age corresponds to the Toarcian time, which is considered to be the syn-breakup period.

Finally, the new perceptions of basin structure and stratal architecture, as well as their development during the Jurassic can be compared to numerical models of basin formation (Burov and Cloetingh, 1997; Cloetingh et al., 1997; van Balen et al., 1995). Such comparisons indicate that during the Jurassic the Morondava Basin of Madagascar was part of a non-volcanic, fast spreading, narrow, passive rift which rift shoulder experienced no considerable uplift.

Zusammenfassung

Die Extension zwischen den östlichen und westlichen Krustenblöcken des Urkontinents Gondwana und deren Trennung hinterließen Spuren in den jurassischen Sediment-Abfolgen des Morondava Beckens in West-Madagaskar, die mit Hilfe interdisziplinärer Analysen untersucht wurden. Die Untersuchungen stützen sich auf Sedimentstrukturen (Aufschluss), mikrofazielle (Aufschluss und Dünnschliffe), stratigraphische (Ammoniten) und paläo-ökologische (Mikro- und Makrofossilien) Analysen. Desweiteren werden die Schichtgeometrien betrachtet (seismische Schnitte), sowie Provenanz- und Resedimentationsanalysen (detritische Apatitspaltspurenalter) durchgeführt. Sedimentologische und stratigraphische Untersuchungen beleuchten lokale und regionale Änderungen der Paläo-Umweltbedingungen und relative Meeresspiegelschwankungen. Seismische Bilder ermöglichen die Interpretation der Art und der zeitlichen Abfolge tektonischer Vorgänge. Aus der Kombination der sedimentologischen und seismischen Konzepte lässt sich ein tektono-sedimentäres Model für das Auseinanderbrechen Gondwanas ableiten.

Während des späten Karbons bis in die späte Trias entstanden Extensionsbecken in verschiedenen Teilen Gondwanas. Im Gegensatz zu früheren Vorstellungen, das Auseinanderbrechen Gondwanas sei ein mehrphasiger, langandauernder Prozess, zeigen diese Untersuchungen, dass das Auseinanderbrechen auf das Intervall des Toarc bis Aalen (später Unterjura) festgelegt werden kann.

Seismische Schnitte durch das Morondava Becken verdeutlichen, dass die spät-karbone bis spät-triassische Karooabfolge durch tektonische Zerrüttung und Blockrotation gestört wurde. Diese Tektonik steht mit dem Auseinanderbrechen des Kontinents in Verbindung und muss daher vor diesem Ereignis abgelagert worden sein. Die nachfolgend jüngere Andafia Formation (Toarc) und die „Sandsteine des Aalen“ sind in einer typischen Syn-Aufbruchposition, wo sie divergente Schichtmuster, bestehend aus marinen Tonsteinen, Kalksteinen und Sandsteinen, in Halbgräben über rotierten Blöcken bildet. Beobachtungen in Aufschlüssen deuten ebenso auf eine sich beckenwärts wiederholende, nach Osten gerichtete Verkippung der Blöcke und ein allmähliches Auskeilen der Schichten nach Westen. Im Majunga Becken finden sich vergleichbare Ablagerungen, die zur Beronono Formation gezählt werden. Dort sind die ältesten bekannten Syn-Aufbruchablagerungen anhand von Ammoniten aus der *Bouleiceras nitescens Zone* als frühes Toarc identifiziert worden. Eine lateral weitreichende Diskordanz wird von Sedimenten des Unter-Bajoc überlagert, markiert das Ende der Syn-Riftphase (Syn-Aufbruchsphase) und wird gefolgt von den Post-Aufbruchablagerungen der Driftphase. Die Post-Riftphase ist gekennzeichnet durch die initiale Ausbildung einer Küstenkarbonatplattform (Bajoc), die sich den gesamten Kontinentalrand entlang erstreckt, gefolgt von einer siliziklastischen Küstenabfolge (Bathon) und Tonsteinen einer Beckenfazies mit wenigen Einschaltungen von Flachwasserkarbonaten (Callov bis Kimmeridge).

Während der Rift- und Driftphase wurden vier große transgressive Ereignisse erkannt, die jeweils von einer Regression gefolgt sind und deren Verlauf hautsächlich eustatischen Meeresspiegeländerungen folgt und nur wenig durch regionale Tektonik beeinflusst ist:

- Frühes Toarc
- Frühes Bajoc
- Frühes Callov
- Frühes Oxford

Der erste dieser Transgressions-Regressions-Zyklen (T-R Zyklen) fällt mit dem Rift während des Auseinenderbrechens Gondwanas im Toarc zusammen. Daneben sind drei weitere T-R Zyklen in den Ablagerungen des passiven Kontinentalrands überliefert. Eine Transgression im frühen Bajoc überschritt die Riftflanke ostwärts weit landein und bedeckte Syn-Aufbruchsedimente des Toarc-Aalen Intervalls and an vielen Stellen Prä-Aufbruchschichten der Karooabfolge. Zu dieser Zeit bildete sich eine Karbonatplattform am jungen Kontinentalrand. Frühere Untersuchungen übersahen die beckenweite Verbreitung der Karbonatplattform (Mergel, Tonsteine, Kalksteine und Oolithe der Bemaraha Formation), da nur die küstennahen Äquivalente (Mergel, Tonsteine, Siltsteine, Kalksteine und Sandsteine der Sakaraha Formation) im südlichen Morondava Becken aufgeschlossen sind.

Eine bedeutende Regression im frühen bis mittleren Bathon erzeugte eine teilweise Exponierung der Beckensedimente im nördlichen Morondava Becken (Stoakes and Ramanampisoa, 1988). Von seismischen Schnitten des nördlichen Morondava Beckens sind Paläokarst und „incised valleys“ an der Oberfläche der Karbonatplattform bekannt. Die Kombination von Beckenexposition, „incised valleys“ und lokalen siliziklastischen Ablagerungen werden als typische „Forced Regression“ Elemente interpretiert.

Eine weitere ausgedehnte transgressive Diskordanz im frühen Callov (*Bullatus Zone*) leitet über zu tiefer marine Bedingungen mit Ton-, Silt-, Kalksteinen und wenigen Eisenoolithen an der Basis. Diese Bedingungen sind vorherrschend bis zum Auftreten eines Flachwassersandsteins, dessen Höchstalter spätes Callov (*Athleta Zone*) und dessen Mindestalter mittleres Oxford (*Plicatilis Zone*) ist. Dieser sog. „oxfordische Sandstein“ dokumentiert ein kurzes regressives Ereignis, in dessem Verlauf randlich marine Ablagerungsbedingungen in das Becken progradierten. In der *Plicatilis Zone* treten an der Basis der jüngsten Transgression des Oxfords im südlichen Morondava Becken charakteristische eisenoolithische Kalksteine auf. Im stratigraphischen Verlauf wechsellagern Tonsteine mit dünnen bioklastischen und eisenoolithischen Kalksteinen.

Die drei Post-Rift T-R Zyklen sind grundsätzlich vergleichbar mit den T-R Zyklen zweiter Ordnung der Tethys. Die biostratigraphische Unschärfe der Ankazoabo Formation und des „oxfordischen Sandsteins“ lassen keinen genauen Vergleich innerhalb der entsprechenden Zeitscheiben zu. Die großen Übereinstimmungen der Meeresspiegelkurven des Morondava Beckens und der Tethys zeigen, dass das eustatische Signal die regionalen Meeresspiegeländerungen überlagert.

Die Ablagerungskonzepte, die von den Studien an Aufschlüssen, Bohrlochdaten und seismischen Schnitten abgelagert wurden, werden ergänzt durch die Zusammenstellung von Fazies- und Isopachenkarten für das Morondava Becken. Fazieskarten deuten eine beckenweite Präsenz der Karbonatplattform des Bajoc an, während im Oxford ein rein siliziklastisches Ablagerungssystem vorhanden ist. Isopachenkarten der Bajoc-Bathon und Callov-Tithon Intervalle zeigen große Unterschiede in der Sedimentmächtigkeit zwischen dem südlichen und nördlichen Becken. Beide Zeitintervalle zeigen zwei Depozentren im südlichen Zentralbereich, während im Norden die Schichten weitgehend dünner sind.

Paläoströmungsindikatoren aus dem südlichen Morondava Becken implizieren generell einen Sedimenttransport nach Westen und eine Nord-Süd verlaufende Küstenlinie. Im Bajoc-Bathon finden sich Hinweise auf einen nordwärtsgerichteten Sedimenttransport am südlichen Ende des Beckens. Dies deutet auf lokale Hebung hin. Am nördlichen Ende des Feldstudiengebiets im südlichen Beckenteil impliziert eine nach Süden gerichtete Paläoströmung im Oxford ebenfalls regionale Hebung. Die regionale Tektonik im südlichen Beckenteil kann auch durch detritische Apatit-Spaltspurenalter (ASS) in jurassischen Sedimenten nachgewiesen werden. Die stratigraphische Abfolge von detritischen ASS Altern indiziert eine stratigraphische Umkehr der ASS Alter durch Wiederaufarbeitung von wahrscheinlich früheren Karoosedimenten. Deren Erosion deutet auf Hebung des Randes des Riftbeckens, der mit Karoosedimenten bedeckt ist.

Geographische Vikarianzanalysen anhand von Ostrakoden zeigen einen hohen Endemismus im Vergleich zu Ostafrika und anderen jurassischen Vorkommen. Die Bildung eines Mittelozeanischen Rückens und eines dysoxischen Beckenmilieus im Proto-Indischen Ozeans sind mögliche Migrationsbarrieren zwischen Madagaskar und dem afrikanischen Festland.

Die Neubewertung der publizierten regionalen Ozeanbodenalter und die Betrachtung des mitteljurassischen Vulkanismus in der Region deuten darauf hin, dass das maximale Alter des Proto-Indischen Ozeanbodens etwa 180 Mio. Jahre beträgt. Dieses Alter wird dem Toarc zugeordnet (Pálfy et al., 2000), das als Aufbruchsperiode betrachtet wird.

Schließlich können die neuen Vorstellungen über die Beckenstruktur und der Schichtarchitektur, sowie deren Entwicklung während des Juras verglichen werden mit numerischen Modellen der Beckenbildung (Burov and Cloetingh, 1997; Cloetingh et al., 1997; van Balen et al., 1995). Solche Vergleiche zeigen, dass während des Jura das Morondava Becken in Madagaskar Teil eines passiven, nicht-vulkanischen, schnell aufbrechenden, engen Rifts war, dessen Riftschulter kaum Hebung erfuhr.

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Chapter 1 Introduction

From the Late Carboniferous onwards, extension and wide-spread intercontinental rifting across Gondwana prefigured dispersal of the supercontinent. Although extension prevailed throughout the Permo-Triassic, which is known as the Karoo episode, it was only a prelude to the crustal separation that commenced in the Early-Middle Jurassic. During the Jurassic-Cretaceous time the global stress regime favoured crustal separation. The initial divergence during the Early-Middle Jurassic happened along the East African-Antarctic Orogen and split Gondwana approximately in half: East-Gondwana and West-Gondwana. In the Middle-Late Jurassic the western half (mainly Africa and South-America) started to disperse along the Atlantic Rift, and during the Cretaceous, East-Gondwana broke into several pieces (e.g. Madagascar, India, Australia, and Antarctica). The marginal situation along the initial breakup suture makes Madagascar a good candidate to investigate the mechanisms of the early Gondwana Breakup.

1.1 Research Objectives and Methods

This study aims to elucidate the sedimentary record of the Morondava Basin, south-western Madagascar, as a response to the Gondwana separation and the development of the passive margin during the subsequent drift. This basically involves the study of the sedimentary patterns with special emphasis on sedimentology, facies, lithostratigraphy, biostratigraphy, and sequence-stratigraphy. Reliable biostratigraphic control and a consistent framework of the depositional sequences were needed to achieve this goal. Sedimentological studies during the past few decades mostly adopted the stratigraphic concept of Besairie and Collignon (1972). Clark (1996) criticised the unquestioned usage of unrevised biostratigraphic and lithological data of Besairie and Collignon (1972) and outlined that apparently inconsistent mapping on adjacent large-scale map sheets resulted in lithological discrepancy on small-scale maps. These discrepancies were the base for misleading sedimentary models (Chapter 2). Thus one target of this study was to reappraise the stratigraphic relationships of sedimentary sequences (formations) and to establish a tectono-sedimentary model for the Morondava Basin during the Gondwana Breakup.

Eighteen sections were measured at thirteen localities in the southern Morondava Basin and 2 sections were taken in the southern Majunga Basin. Meso-scale facies were analysed by interpreting sedimentary structures at outcrops and by the interpretation of the macrofossil assemblages. Microfacies analysis is based on the interpretation of thin sections and microfossil assemblages.

Biostratigraphy was mainly established with ammonites. With new ammonite data, the biostratigraphic concepts of Besairie and Collignon (1972), Collignon (1964a; 1964b; 1967; 1959), and Joly (1976) was either confirmed, improved or sometimes revised. Due to a lack of ammonite findings in Bajocian and Bathonian strata, stratigraphic classifications from brachiopods and corals (Lathuilière et al., 2002), and from ostracods (Mette and Geiger, 2004a; Uhmann, 1996) were used despite their high uncertainty.

Palynomorph extraction was applied to selected samples but they turned out to be devoid of palynomorphs. Dina (1996) suspected deep weathering to dissolve palynomorphs in outcrop samples in southern Madagascar, because he only able to find dinocysts and spores in well samples.

Subsurface data, such as lithofacies from well logs (Besairie and Collignon, 1972; Dina, 1996; du Toit et al., 1997; Montenat et al., 1996) and seismic facies from seismic cross-sections (Lines 89-KA-05 and TTJ-13, shot by Shell and Amoco, respectively) as well as interpreted data sets from unpublished reports (Pierce and Yeaman, 1986; Stoakes and Ramanampisoa, 1988) were used for basinwide correlation.

1.2 Geological Background

Madagascar, the forth largest island in the world, lies east of East Africa in the Indian Ocean. The Mozambique and Somali channels separate the island from the African mainland. A bipartite topography characterises the island. Low topography and wide coastal plains at the western coast are formed by Late Palaeozoic-Cenozoic sedimentary basins. In contrast, the highlands in the central eastern part are formed by uplifted crystalline basement rocks. At the east coast the coastal plain is narrow and the highland rises steeply.

1.2.1 Gondwana assembly

Madagascar is situated in the centre of the Gondwana supercontinent (Fig. 1). During its assembly in the Neoproterozoic-Early Cambrian crustal fragments collided and amalgamated. Persistent compression gave rise to high orogenic belts (Pan-African Orogen) throughout the supercontinent (Krabbendam and Barr, 2000; Meert, 2002; Stern, 1994; Unrug, 1997). Those orogenic belts served as pathways for extension and translation during the later dispersal of Gondwana (Emmel et al., 2004b; Pili et al., 1997; Piqué et al., 1999). The genetic link between the Pan-African orogeny and the Gondwana Breakup strongly influences reconstructions of both the assembly of the supercontinent and later its breakup. The process of breakup can usually be studied either by forward or backward modelling. On the one hand, the comparison of basement structures and magmatic and metamorphic ages can be used to fit together the crustal segments back together to reconstruct the original supercontinent. From these models the subsequent development is viewed with the aim to finally bring it into agreement with the configuration at the end of the separation. On the other hand, the final configuration can be taken as the starting point, and the breakup process can be retrieved by “backstripping” the continental drift. Although the assembly and breakup of Gondwana have been intensively studied during the last few decades, models of the initial fit of the crustal fragments, in particular, remain controversial.

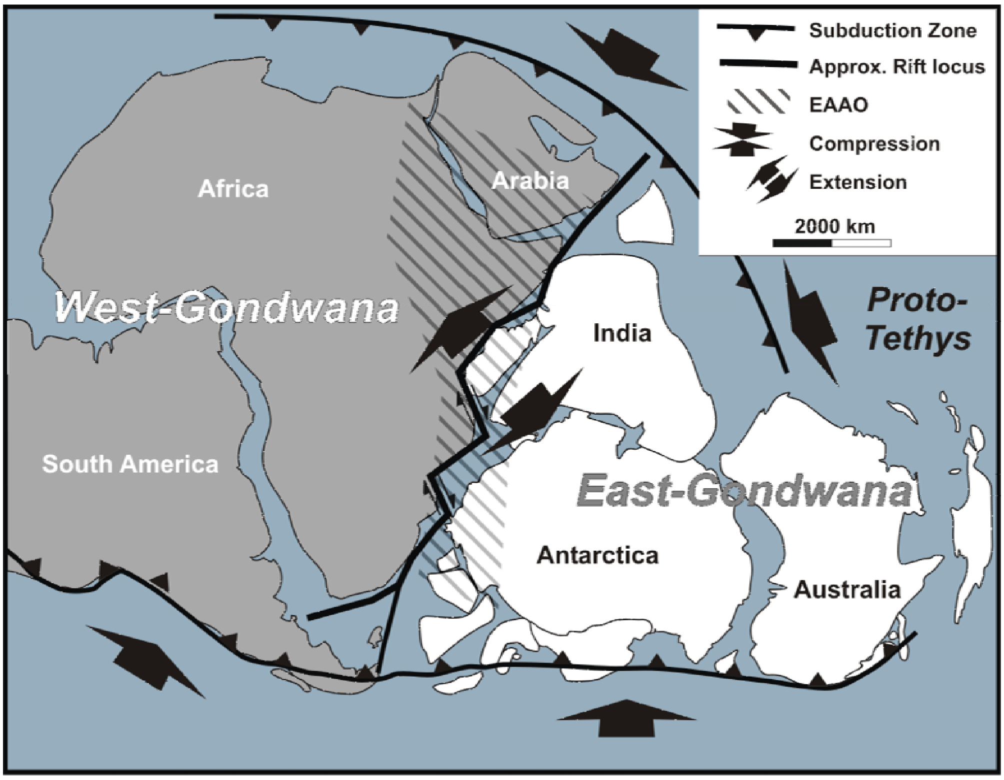


Fig. 1: Gondwana reassembly with the axes of successful separation and opening of the Early Indian Ocean between East- and West-Gondwana. Extension, as a response to compression at the northern and southern margin of the supercontinent, was localised along the former East African-Antarctic Orogen (EAAO). Modified from Stollhofen (1999) and Jacobs et al. (1998).

In previous palaeogeographical reconstructions of the Proto-Indian Ocean, the continental margin morphologies were simply compared. In such a reconstruction of Hobday (1982) Madagascar was relocated adjacent to the Rhodesian craton between the Limpopo and Zambezi rift zones (cf. Daly et al., 1989). In contrast, Segoufin and Patriat (1981), who compared magnetic anomalies in the Mozambique Channel, postulated a palaeo-position for Madagascar more than 3,000 km to the north, alongside north Mozambique, Kenya, and Tanzania. Coffin and Rabinowitz (1987; 1992) enhanced the magnetic sea floor data, in addition, utilized the sedimentary features along the continental margins to find a closer fit of the coastal basins. Based on the correlation of magmatic histories, foliation and isotope ages, the fitting of the exposed basement became more reliable (de Wit, 2003; Kröner et al., 2000; Reeves et al., 2002; Windley et al., 1994). Other authors used gravity data (Lawver et al., 1998) to fit the crustal fragments.

However, the fit of the opposing coasts of East Africa and Madagascar in particular appears to be problematic. Especially for the basement promontory at Cape St. André in NW Madagascar it was difficult to find the complementary gap in East Africa. Reeves et al. (2002) present an enhanced reconstruction for the region which places Cape St. André in the Anza Rift (Bosworth and Morley, 1994) between Kenya and Ethiopia. Moreover, Reeves et al. (2002) conclude that it is not possible to solve the puzzle of the Gondwana reassembly with a simple two-piece model of East- and West-

Gondwana. They suggest that the original concept of two rigid plates (East- and West-Gondwana) breaking apart can be improved by dividing continental plates into several small crustal segments beside the prominent cratons that are loosely connected by shear zones (Reeves et al., 2004).

These shear zones hinge the crustal segments and transfer stress over wide distances, and also allow for limited rotation. Today it is difficult to estimate the mode and amount of movements between those segments. This problem coincides with the question of which basins and basin parts in East Africa and Madagascar were previously connected. The most recent model (de Wit, 2003) place the Morondava Basin, and especially its southern part, adjacent to the coastal basins of Tanzania and northern Mozambique. Modern Gondwana-wide palaeogeographical reconstructions are presented by e.g. Visser and Praekelt (1996), Scotese et al. (1999), and Stampfli and Borel (2002).

1.2.2 Tectonic setting: a Pan-African heritage

When extension within the African sector of Gondwana took place during the Late Palaeozoic-Mesozoic, stress and strain patterns were oriented NW-SE to NNW-SSE. This orientation coincides with Pan-African structures, usually shear zones that were derived during continental collision and the formation of the East African-Antarctic Orogen (EAAO) (Jacobs et al., 1998; Windley et al., 1994).

The NNW-SSE to N-S trending Davie Ridge Fracture Zone (DRFZ, Fig. 2) is almost parallel to the north-western coast of Mozambique and has a similar orientation to the Pan-African shear zones. The DRFZ is a fossil transform fault that guided the southward drift of East-Gondwana (Madagascar, India, Australia, and Antarctica) during the Jurassic-Early Cretaceous (Droz and Mougenot, 1987). As a result, transpressional and transtensional stress were localised at the Davie Ridge and probably applied to the southern margin of the Morondava Basin (Emmel et al., 2004a; Malod et al., 1991). Similar fracture zones (Dhow, VLCC, ARS, Fig. 2) are located along a second possible transform extending from the northern tip of Madagascar to the Somali coast (Bunce and Molnar, 1977; Coffin and Rabinowitz, 1992). However, a clear conjunction to the DRFZ has not yet been demonstrated (Coffin and Rabinowitz, 1987; Coffin and Rabinowitz, 1992). The orientation of the fracture zones follows prominent Pan-African lineations, such as the parallel Ranotsara-Bongolava Shear Zone (RBSZ) more than 500 km to the north (de Wit, 2003; Windley et al., 1994). In contrast to the prominent horizontal displacement at the DRFZ, Mesozoic or Cenozoic reactivation with vertical movements along the RBSZ are small (Seward et al., 1998). According to Emmel et al. (2004b) relative uplift is limited to the region south of the RBSZ. At the northern end of the DRFZ a considerable stretching of the crust appears to have occurred prior to the seafloor spreading (Coffin and Rabinowitz, 1987). The amount of stretching is difficult to estimate since no reliable crustal thickness data are available. Moreover, Cenozoic tectonism strongly obscures possible evidence (Coffin and Rabinowitz, 1984; Mauge et al., 1982).

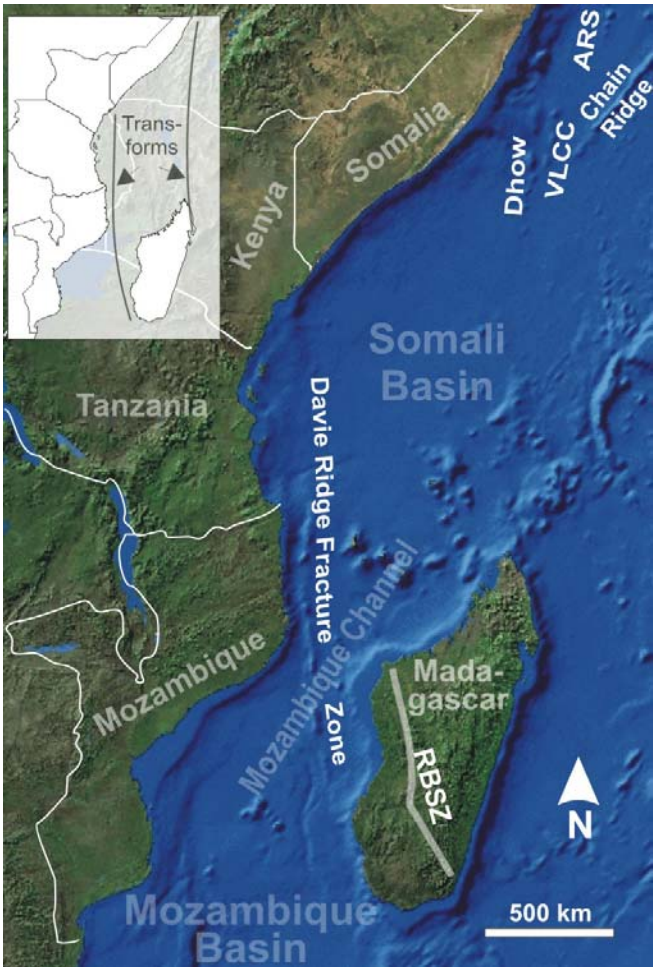
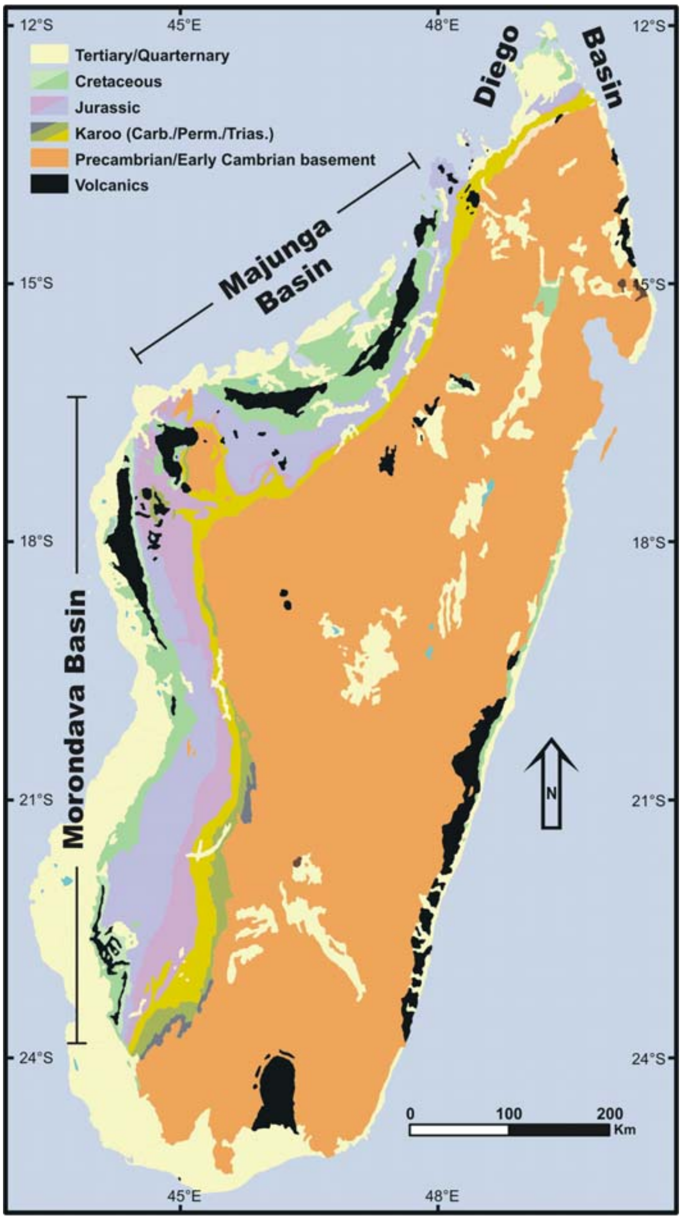


Fig. 2: Present-day sea floor topography in the Mozambique Channel is characterized by the elevated Davie Ridge Fracture Zone which acted as a transform fault guiding the translation of East- and West-Gondwana (Malod et al., 1991). A similar structure is present from the northern tip of Madagascar to the northern Somalia's Indian Ocean coast, documented by the Dhow, Very Large Crude Carrier (VLCC), and ARS (Bunce and Molnar, 1977; Coffin and Rabinowitz, 1987). A subparallel structure, the Ranotsara-Bongolava Shear Zone (RBSZ), crosses Madagascar from SE-NW. The map is modified from the ESRI World Map Background.

1.2.3 Morondava Basin history

The Palaeozoic-Cenozoic sedimentary record of Madagascar is basically restricted to three coastal basins: Diego, Majunga, and Morondava basins (Fig. 3). These basins occupy the coastal plains at the western and north-western margins of the island, extending onto the continental shelf. A narrow coastal strip along the east coast hosts Cretaceous-Tertiary deposits. Within the basement exposures of central Madagascar only sedimentary covers are limited to Cenozoic graben structures, e.g. at the Lake Alaotra and in the Ankay graben, both of probable Tertiary age (Besairie and Collignon, 1972).

The research history of the sedimentary basins of Madagascar is relatively poor. After the voluminous work of Besairie and Collignon (1972) only few studies were published (e.g. Hankel, 1994; Luger et al., 1994; Montenat et al., 1996; Piqué, 1999; Piqué et al., 1999; Rakotosolofo et al., 1999; Razafindrazaka et al., 1999; Wescott, 1988; Wescott and Diggens, 1997; Wescott and Diggens,



1998). Many data derived from oil exploration activities and are held in the archives of OMNIS (Office des Mines Nationales et des Industries Stratégiques) which are closed to public access.

Fig. 3: The geology of Madagascar shows a bipartite distribution. While the central and eastern part is made up of crystalline basement rocks, the western and north-western margins are made up of three coastal basins: The Diego, Majunga, and Morondava basins. Modified from the UNESCO International Geologic Map of Africa at 1:5,000,000, Open File Report 97-470A provided by the USGS.

1.2.3.1 Polyphase Karoo rifting

After a period of stress equality (Cambrian-Middle Carboniferous), a reorganisation of stress patterns led to crustal extension along predefined Pan-African crustal anisotropies (orogens and shear zones). Initially extension was slow and the locus of crustal extension was indistinct. Changing stress orientations (Schandelmeier et al., 2004) formed extensional basins of various types in a broad area. Many authors relate the sedimentary sequences which filled those basins to the comprehensive succession of the Karoo Basin of South Africa (Hankel, 1994; Piqué et al., 1999; Wopfner, 1994). Although the Karoo Basin is a foreland basin of the Cape Fold belt, the term “Karoo” (SACS, 1980) became a synonym for Carboniferous-Early Jurassic sediments of extensional basins all over Gondwana (e.g. Smith et al., 1993; Wopfner, 1999). Generally, the term Karoo applies to Gondwana deposits that predate the breakup (Stollhofen, 1999). In this study the Gondwana breakup event was recognised within the Toarcian-Aalenian period and was discriminated from the polyphase Karoo rifting which failed to 02809880 \h Fig. 25, S23°42.265'/E44°20.395') along both shordeposits of Gondwana formed prior to the time of breakup.

Karoo rift basins chiefly formed along either side of the African continent, where later continental breakup was localised. At the eastern margin of present-day Africa, those basins coincided with the extension of the East-African-Antarctic Orogen.

Sedimentation in Madagascar commenced with Carboniferous-Permian glacial deposits which formed the basal Sakoa Group (Wescott and Diggens, 1997). Later marine, lacustrine, and fluvial deposits built up the succeeding Permian, upper Sakoa and Sakamena groups (Wescott and Diggens, 1997; 1998). A predominantly fluvial environment is recorded by sandstones of the Triassic Isalo Formation (Besairie and Collignon, 1972). After a hiatus during end Triassic-late Early Jurassic time, an Early Toarcian marine incursion is recorded in the subsurface of the Morondava Basin, and at outcrop in the Majunga Basin.

1.2.3.2 Continental breakup and drift

Timing and evolution of the continental separation of East- and West-Gondwana has been the subject of several of studies. Previous considerations of the onset of oceanic spreading were based on palaeogeographical reconstruction (1.2.1). The breakup was assigned to the Oxfordian-Kimmeridgian, according to the first recognized anomaly at chron M25 (Coffin and Rabinowitz, 1992). Early-Middle Jurassic sediments, which post-date the Karoo depositional sequence, were classified as a transitional sequence between the continental Karoo rifting and oceanic spreading (Coffin and Rabinowitz, 1992; Hankel, 1994; Montenat et al., 1996). Montenat et al. (1996) incorporated the tectono-sedimentary patterns and correlated the Middle Jurassic strata of the Morondava Basin to the oceanic spreading phase. They concluded that the underlying Lower Jurassic strata belong to the Karoo rifting.

In the present thesis, the timing of the breakup was interpreted with tectono-sedimentary patterns (Chapter 2). This includes the classification of sedimentary sequences and their position in the pre-, syn-, and post-rift history (sensu Bosence, 1998). In conclusion, after a period of tectonic quiescence during Hettangian-Pliensbachian times (Schandelmeier et al., 2004), the syn-rift phase of the Gondwana Breakup occurred during the Toarcian-Aalenian along an axis with a distinct rift locus.

The Bajocian oceanic spreading and continental drift formed a continental passive margin. The sea transgressed onto the rift shoulder and reached far inland to form a coastal platform along the basin margin. The depositional history was fairly similar in the coastal basins of East Africa (Chapter 3). Interpreting the sedimentary succession in Tanzania, Kreuser (1995) also suggests that the post-rift and drift phase commenced in the Bajocian. The Bathonian of Madagascar is characterized by thick sandstone successions in the southwest (Besairie and Collignon, 1972; Luger et al., 1994). In the north the Bathonian appears to be reduced or missing, since it has not been biostratigraphically recognised. Instead, seismic data suggests intensive erosion at the top of the Bathonian (Chapter 3). From the Callovian onwards a general transgressive event resulted in a deeper basinal environment with a temporary shallowing in the Lower Oxfordian, as evidenced by sandstone units ("Oxfordian Sandstone") that appear in several places in the southern Morondava Basin (Besairie and Collignon, 1972; Luger et al., 1994).

1.2.4 Palaeogeography of the early Indian Ocean

The integrated view on the sedimentary environment of the basins in East Africa and Madagascar, and geodynamic reconstructions of Gondwana (1.2.1), infer episodes of a marine indentation of the southern Tethys into the northern margin of the supercontinent. A contiguous seaway through a corridor of continental extension along the future breakup rift of Gondwana in the Early Jurassic to South-East Africa, prior to the breakup, has been discussed controversially through the last decades (Norton and Sclater, 1979; Prinz et al., 1993; Riccardi, 1991; Salman and Abdula, 1995; Tarling, 1988). In modern reconstructions, the sedimentary basins of present-day Madagascar are shown lying next to the East African coastal basins (1.2.1), framing a broad area of subsidence and deposition. Many authors presume a widespread continental facies to have formed within this basin area. Evidence for this interpretation is found in thick sedimentary successions of the Karoo sequence (Late Carboniferous-Late Triassic) in the marginal basins on either side of the corridor of extension (Coffin and Rabinowitz, 1992; Gordon, 1970; Kreuser, 1995; Luger et al., 1994; Salman and Abdula, 1995; Wopfner, 2002). For all that, the presumed contiguity of extension is purely speculative. In the Early Jurassic, large volumes of volcanics (Duncan et al., 1997; Kamen-Kaye, 1983; Pálfy and Smith, 2000; Salman and Abdula, 1995) covered the early basins south of the extension of the future Davie Ridge Fracture Zone (DRFZ, Fig. 2, Fig. 3). Intercalating continental deposits are known from north-east Mozambique (Jaritz et al., 1977) but the presence of marine environments is not reported. In contrast, marine conditions invaded the corridor of extension southward from the Arabic Peninsula, Ethiopia, and Somalia during the late Early Jurassic (Gordon, 1970; Luger et al., 1994)(Fig. 1). Plume related uplift south of the future DRFZ, as well as large volumes of volcanics, could have been possible morphological barriers for the southward invading sea. Furthermore, the initiating DRFZ was a continental shear zone. These structures often experience local uplift (Lorenzo, 1997) and thus the DRFZ could also have limited the early transgression.

The initial epicontinental embayment during the late Early Jurassic was characterised by considerable terrigenous influx with only minor occasions of evaporate and carbonate formation on either side (e.g. Coffin and Rabinowitz, 1992). Thus the hinterland was highly sediment-supplying. Emmel et al. (2004a; Seward et al., 2004) suggest that provenance areas reached far east across Madagascar, possibly into present-day India.

1.2.5 Jurassic eustacy and geological events

During the Jurassic, a total of eight major episodes of sea-level rise are recognised globally (Hallam, 2001)(Fig. 4). With regards to the Hettangian-Pliensbachian (Early Jurassic) hiatus, Lower Toarcian deposits are the oldest Jurassic rocks in the East African-Madagascan domain (Hallam, 1988). Thus, only five global flooding events have potentially affected the region of the Proto-Indian Ocean (Hallam, 2001): during the Early Toarcian, Early Bajocian, Late Bajocian, Middle Callovian, and Late Oxfordian-Kimmeridgian. Hallam (2001) also recapitulates that during the Aalenian and Bathonian major episodes of sea-level falls are found in different places around the world. Earlier investigations on sea-level changes in the Morondava Basin by Luger et al. (1994) propose three transgressive events: Middle Toarcian, Early Bajocian, and Early Callovian. All three transgressions follow the sea-level curve of Haq et al. (1987) and are thus interpreted as eustatic sea-level rises (Luger et al., 1994).

The present study recognised four transgressive-regressive (T-R) cycles in the Morondava Basin. A major transgression in the Early Toarcian introduced the syn-breakup T-R cycle. The succeeding post-breakup and drift phase (Bajocian-Kimmeridgian) contains three T-R cycles with transgressive events in the Early Bathonian, Early Callovian and Middle Oxfordian (see Chapter 3). Discrepancies exist between regional sea-level curves from different places in the world and for eustatic curves (Fig. 4), and it is difficult to find concordance beyond the resolution of second order sequences.

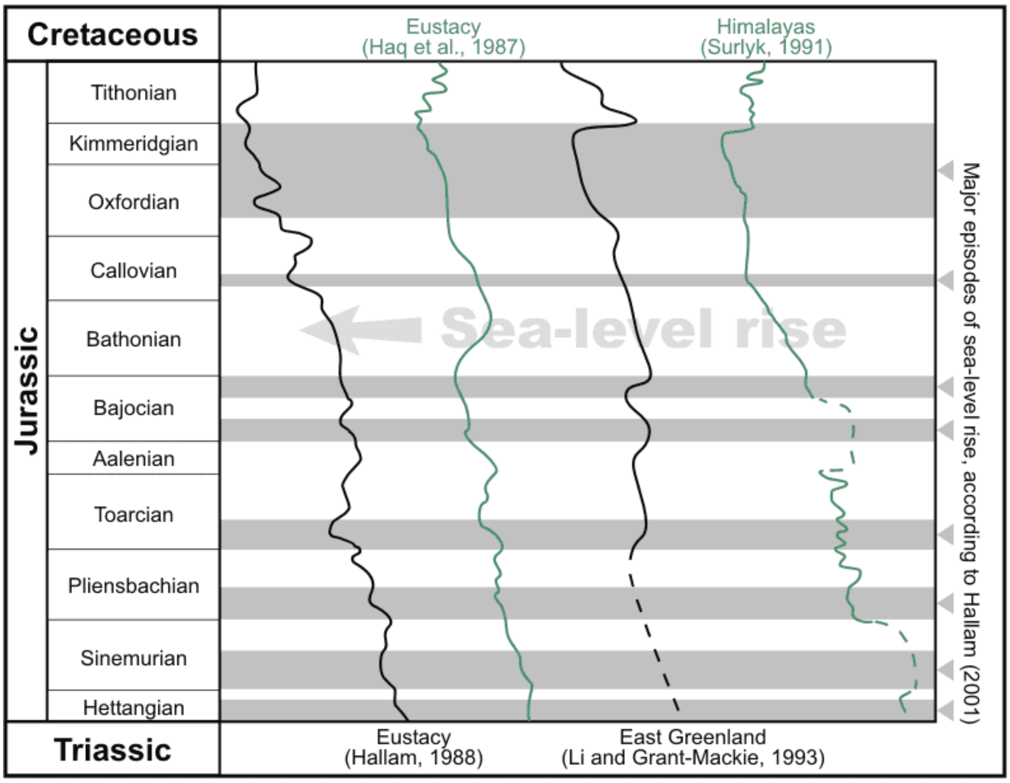


Fig. 4. Proposed Jurassic eustatic sea-level curves (after Hallam, 1988; Haq et al., 1987) compared to regional sea-level curves for the Himalayas (after Li and Grant-Mackie, 1993) and East Greenland (after Surlyk, 1991). The highlighted (light grey) stratigraphic intervals are major episodes of sea-level rise according to Hallam (2001).

A sequence stratigraphic approach for the East African domain is presented by Bosellini (1992). Based on the Somalia record he recognized two flooding surfaces during the Jurassic. The first, on top of the Karoo-age Adigat Sandstone, forms the base of a Bathonian-Callovian carbonate platform; the second flooding surface on top of the carbonate platform represents the base of the Callovian-Kimmeridgian and possibly Tithonian transgression. Moreover, Bosellini (1992) postulated a

Kimmeridgian maximum flooding event, based on a major unconformity that cuts off the Jurassic and is overlain by Late Maastrichian (Late Cretaceous) strata. Other authors (e.g. Kassim et al., 2002; Kreuser, 1995; Mbede and Dualeh, 1997) indicate transgressive/regressive surfaces and facies while describing the lithostratigraphy and palaeo-environments, but do not go into detail of the importance these features to regional cycle architecture.

1.2.6 Jurassic climates

Global climate models for the Jurassic are deduced from palaeofloras and lithological evidence. Rees et al. (2000) suggest that Madagascar was lying at the border of an arid to semiarid (winterwet) climate during the Early and Middle Jurassic. Only in the Late Jurassic did the southern tip of Madagascar feel the influence of a warm temperate climate. Clay mineral analysis of Jurassic strata from the southern Morondava Basin gives a more regional perspective (Uhmann, 1996). During the Bajocian and Bathonian a warm semiarid climate with seasonal humidity and predominantly physical weathering prevailed. During the Callovian-Oxfordian a warm tropical and humid climate predominated. Jarzen (1981) infers also a warm coastal climate from the palynomorph assemblage of the Upper Jurassic at Tendaguru (Tanzanian), which are correlative strata on the conjugated continental margin of E-Africa during the Gondwana dispersal. Dina (1996) presents palaeoclimate interpretations obtained from palynomorph assemblages. He suggests semiarid conditions with humid episodes for the Bajocian-Callovian, whereas Oxfordian assemblages show increased humidity.

In conclusion, climate changes in the study area show minor imprints in the sedimentary record. Temperatures were temperate-warm throughout the Jurassic. Only the humidity changed in a considerable range from semiarid conditions during the Bajocian-Bathonian to more humid climates during the Upper Jurassic. The sedimentary record in the basins of western Madagascar shows no direct correlation to the climate.

Chapter 2 Reappraisal of the timing of the breakup of Gondwana based on sedimentological and seismic evidence from the Morondava Basin, Madagascar

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Abstract

The breakup of Gondwana along the former East African Orogen is widely interpreted to have started in the Late Palaeozoic. The present study indicates that the Permo-Triassic or Karoo phase of rifting was not responsible for the separation of Madagascar from Africa. This rift system failed in the Late Triassic. The separation of Madagascar occurred in the Late Liassic. This event was relatively short-lived, with a distinct episode of rifting followed by separation and continental drift. Previous workers have interpreted the rift and drift phases as a transitional process, starting in the Late Permian and finishing in the Callovian. The pre-rift phase in the Morondava Basin is represented by the fluvial Isalo sandstones which are Late Triassic in age, and the syn-rift phase is recorded by the marine shales of the Andafia Formation which are Toarcian-Aalenian in age. The Early Bajocian unconformity is interpreted as the breakup unconformity. The post-rift or drift phase is represented by the Bajocian-Bathonian carbonates of the Bemaraha Formation, and by the marls, sandstones, and carbonates of the Sakaraha Formation. These two formations are considered to be lateral equivalents of one another, with the Sakaraha Formation representing a coastal plain environment and the Bemaraha Formation a coastal barrier/lagoon complex.

Keywords: Madagascar; Jurassic; Karoo; Gondwana; Breakup; Rift

2.1 Introduction

The breakup of Gondwana is widely interpreted to have occurred in the Middle Jurassic, but the tectonic and stratigraphic history associated with this event is poorly understood. Madagascar was on the eastern side of the suture that propagated through Gondwana during breakup. As a result of the breakup, three sedimentary basins formed in western Madagascar (Morondava, Majunga, and Ambilobe basins, Fig. 5, Fig. 6). These basins contain thick successions of Mesozoic and Cenozoic sediments; the Morondava Basin also contains Late Palaeozoic sediments. This paper examines several stratigraphic units in the Morondava Basin and attempts to correlate them with specific stages of rift evolution, following the rift strata model of Bosence (1998). We recognize seven depositional sequences (Fig. 7) are recognised from the Latest Carboniferous to the end of the Jurassic and link them to a tectonic model of the breakup of Gondwana. A basinwide Bajocian carbonate platform is particularly significant. This platform occurs at the base of the post-Gondwana breakup succession and provides a clear marker for separating the rift and drift phases, rather than invoking the so-called transitional phase of earlier workers (e.g. Hankel, 1994; Montenat et al., 1996). It also allows the breakup event to be dated more precisely as Toarcian.

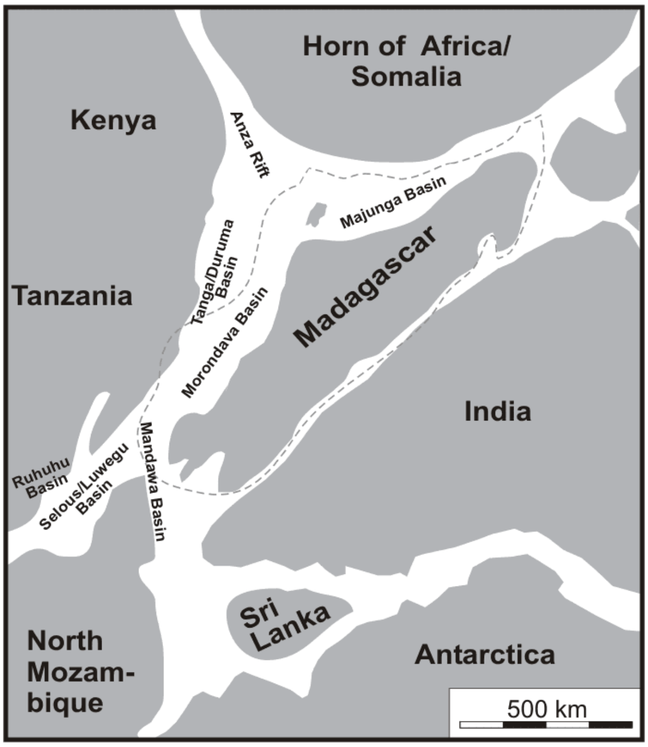


Fig. 5. Reconstruction of Gondwana fragments at 200 Ma by Reeves et al. (2002) based on an interpretation of ocean-floor topography. The outlines of Precambrian crustal fragments are shown in grey. Areas of original continental rocks that have been extended and lost from outcrop by rifting are shown in white. Sedimentary basins in Madagascar and Tanzania are indicated. Present-day Madagascar is shown in outline.

We measured and described five sections, one from outcrops of Toarcian-Bathonian strata near to Kandreho in the southern Majunga Basin (Fig. 8), and four from Bajocian and Bathonian outcrops in the vicinity of Sakaraha and Tongobory in the southern Morondava Basin (Fig. 9, Fig. 11-Fig. 13). We described thin-sections of the limestones and extracted microfossils from more friable marls and siltstones for environmental interpretation. Two East-West seismic cross-sections (Lines 89-KA-05 and TTJ-13, shot by Shell and Amoco, respectively) from the northern Morondava Basin were used to investigate the relationship between structural and depositional history within the basin. The results of the study are discussed in relation to outcrop and subsurface data from elsewhere in the Morondava

Basin (e.g. Besairie and Collignon, 1972; Clark, 1996; Montenat et al., 1996; Piqué et al., 1999 and unpublished reports), and to a more limited extent with information from the Majunga Basin.

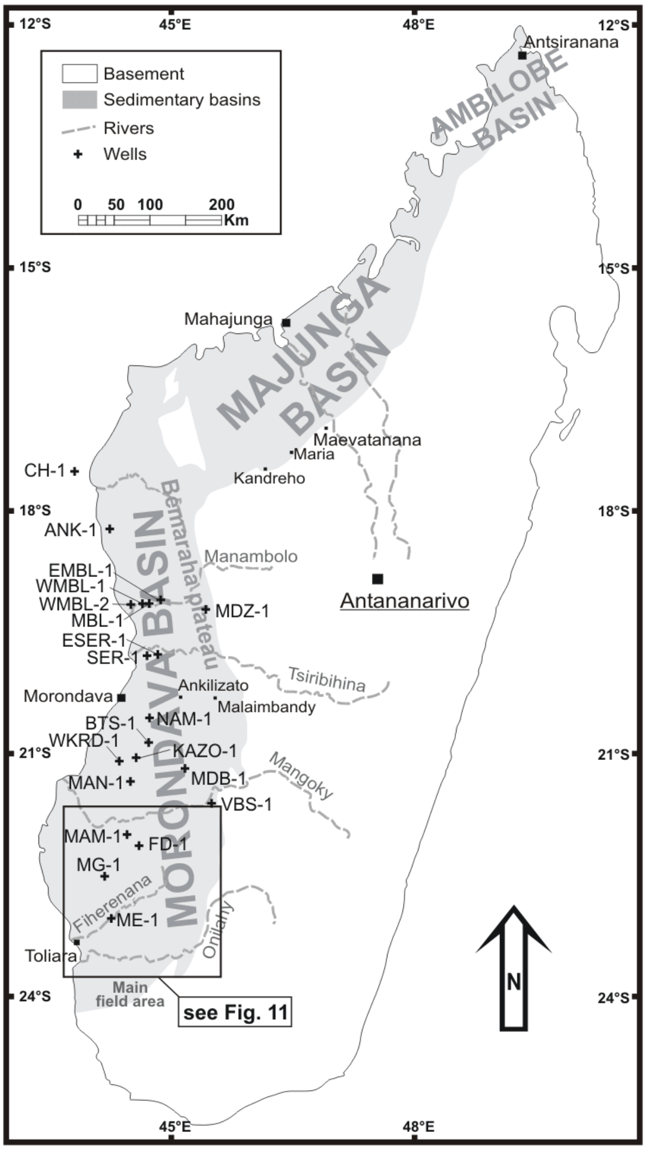


Fig. 6. The three major sedimentary basins of West Madagascar showing important localities mentioned herein. A more detailed map for the outcrop localities is given with Fig. 3. Wells: Chesterfield-1 (CH-1), Ankamotra-1 (ANK-1), West Manambolo-1 (WMBL-1), West Manambolo-2 (WMBL-2), Manambolo (MBL-1), East Manambolo-1 (EMBL-1), Manandaza-1 (MDZ-1), East Serinam-1 (ESER-1), Serinam-1 (SER-1), Namakia-1 (NAM-1), Betsimba-1 (BTS-1), Ankazofotsy-1 (KAZO-1), West Kirindy-1 (WKRD-1), Mandabe-1 (MBD-1), Manja-1 (MAN-1), Vohibasia-1 (VBS-1), Mamakiala-1 (MAM-1), Tandrano-1, also known as Ampandramitsetaka-1 (FD-1), Ambtolahy-1 (MG-1), and Manera (ME-1).

2.2 Tectonic Setting

Madagascar was originally situated in central Gondwana, adjacent to present-day Kenya and Tanzania, before the onset of continental breakup. Recent plate reconstructions by Reeves et al. (2002) have demonstrated a convincing fit between Madagascar and East Africa connecting the basement promontory at Cape St. André to the Anza Rift (Fig. 1). Crustal extension between eastern Gondwana (Madagascar, India, Antarctica, and Australia) and western Gondwana (Africa, Arabia, and South America) commenced at the end of the Carboniferous, with a zone of weakness developing along the former Pan-African mobile belt (Montenat et al., 1996; Piqué et al., 1999). A series of intracontinental rifts and pull-apart basins were formed along the zone of weakness in the Early Permian (Schandelmeier et al., 2004). Coffin and Rabinowitz (1992), Montenat et al. (1996), and Piqué et al. (1999) have suggested that rifting became more widespread in the Triassic and continued uninterrupted until the Early Jurassic. Clark (1996) and Ramanampisoa et al. (1997), however, argue that two distinct phases of rifting can be recognised, one occurring in the Permo-Triassic and the other in the Early Jurassic. They suggest that the initial phase of rifting failed in the Middle Triassic, and that the locus rifting shifted to the west in the Early Jurassic to the present-day continental margin of Madagascar. Similar shifts have been described from the South Atlantic continental margin (Stollhofen, 1999) and the Norwegian margin (van Wijk and Cloetingh, 2002).

The breakup of Gondwana eventually took place along the Early Jurassic rift. Ocean-floor spreading probably started in the Callovian (Coffin and Rabinowitz, 1992; Montenat et al., 1996) and Madagascar drifted southwards away from Africa along the Davie Fracture Zone (Malod et al., 1991). Various authors have argued that there was a relatively long period of transition between the rifting and drifting phases of breakup. Coffin and Rabinowitz (1992) attempted to date this transition using magnetic anomalies and regional seismic surveys from the Indian Ocean. Their study, however, failed to recognise two separate phases of rifting. It was also limited by the occurrence of the so-called Jurassic Magnetic Quiescence before chron M25, at the Oxfordian/Kimmeridgian boundary. Hankel (1994), Luger et al. (1994), and Montenat et al. (1996) dated the transition as Bajocian to Bathonian using stratigraphic and sedimentological data but they also failed to recognise the development of two separate phases of rifting, and the significance of the Early Bajocian breakup unconformity was overlooked. In this study, breakup is seen as a relatively sudden event that occurred in the Toarcian, with drifting occurring thereafter. On seismic, extensional faulting more or less dies out at the breakup unconformity, marking the end of rifting. Thermal subsidence (sag) dominated the tectonic history of the continental margin of Madagascar from the Early Bajocian onwards.

2.3 Stratigraphy of the Morondava Basin

The study of the stratigraphy of the Morondava Basin started in the 1950s when the first geological maps of Madagascar were produced and the Societé Pétrole de Madagascar (SPM) began drilling for hydrocarbons. Most of the borehole data were never published but the sedimentological and stratigraphic concepts based on these data were summarized by Besairie and Collignon (1972). These concepts are still widely used today and the current understanding of the Morondava Basin is still dependent largely on Besairie’s work. A large amount of unpublished data obtained by oil companies are held in the archives of OMNIS (Office des Mines Nationales et des Industries Stratégiques) but very little new information has been made available in the public domain since 1972. Most publications rely heavily on Besairie’s data and only a few recent publications by e.g. Montenat (1996), Luger et al. (1994), Uhmann (1996), and Pique (1999) have presented new sedimentological data.

The pre-breakup succession of the Morondava Basin, ranges from Latest Carboniferous to the end of the Triassic in age and has been divided in three major units, the Sakoa and Sakamena groups, and the Isalo Formation (Fig. 7). The entire succession generally varies from 3,000-4,000 m in thickness, although it may reach 11,000 m in thickness in the southern Morondava Basin (Boast and Nairn, 1982). The overlaying Jurassic sediments, in contrast, are thinner and rarely exceed 200-300 m in thickness at outcrop and several hundred metres in the subsurface. Dina (1996) and Boast and Nairn (1982) report thicknesses exceeding 1500 m in the subsurface, following measurements by Besairie and Collignon (1972). Recent estimates by Clark (1996) suggest the local occurrence of Jurassic strata thicker than 2000 m. Pre-breakup strata are generally correlated with deposits of the Karoo Supergroup of southern Africa (Boast and Nairn, 1982; Kreuser, 1995; Luger et al., 1994; Piqué et al., 1999; Wopfner, 1994).

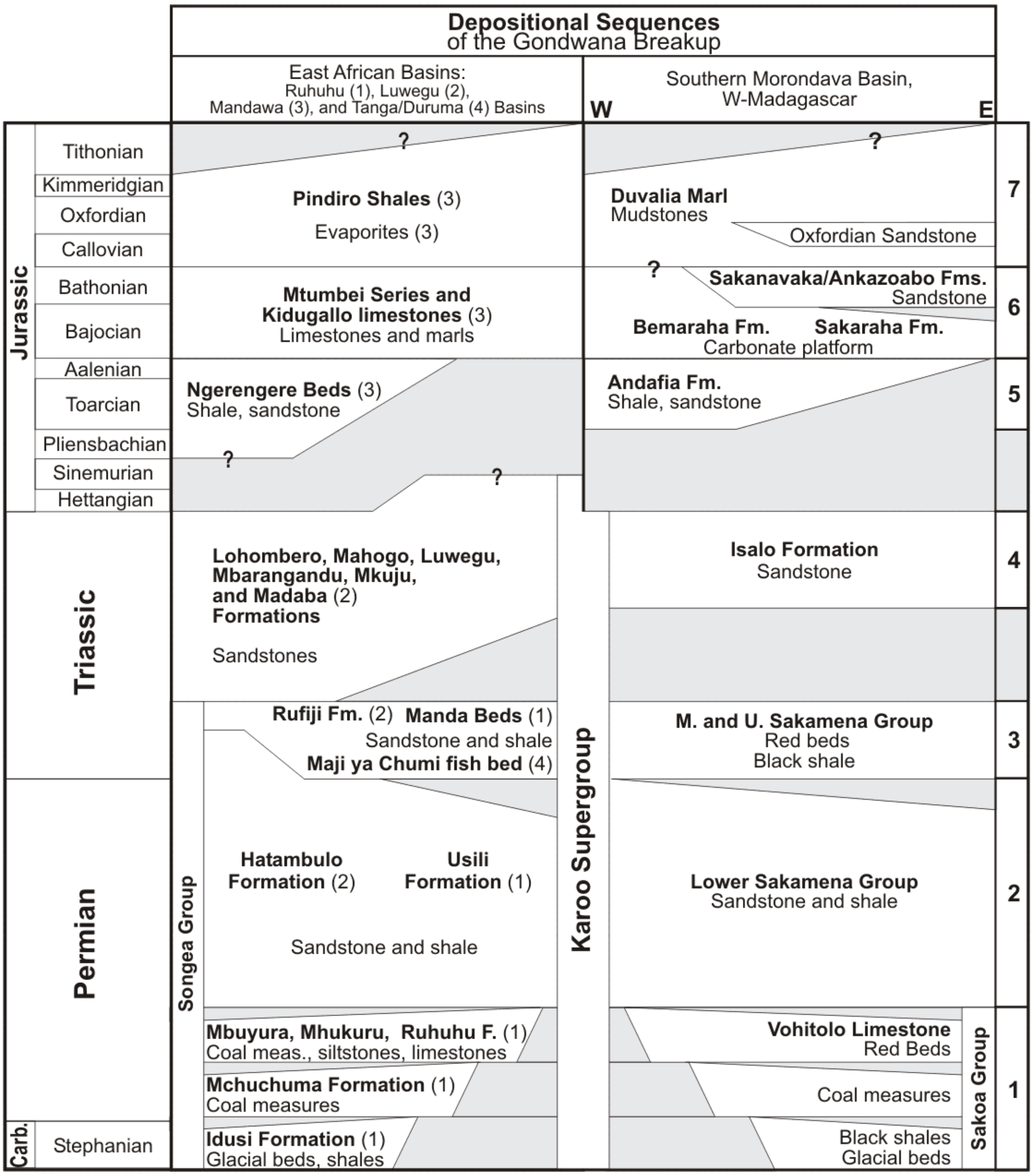


Fig. 7. Depositional sequences recognised in the Morondava Basin from the Latest Carboniferous to the end of the Jurassic. For stratigraphic and lithological information see text. Permian and Triassic are not subdivided into ages because of poor stratigraphic control. The data sources for Karoo sequences are mainly Clark (1996) and Wescott and Diggens (1997, 1998). Jurassic sequences are reappraised and subdivided in the present paper. Correlative sequences of East Africa are compiled from Hankel (1994), Kreuser (1995), and Wopfner (1994, 2002).

2.3.1 Sakoa Group

The Sakoa Group consists of a succession of up to 2,000 m (Wescott and Diggens, 1997) thick succession of tillites, sandstones, shales, limestones, and coals of Late Carboniferous-Early Permian age (Fig. 7). Wescott and Diggens (1997) interpreted these sediments as glacial, lacustrine, swamp, alluvial fan, braided and meandering stream, and marine deposits. They described a palynomorph assemblage of Stephanian (Late Carboniferous) age from the base of the succession, which correlates with the glacial deposits of the basal Dwyka Group of southern Africa (Bangert et al., 2000). The Sakoa can also be correlated with the glacial sediments of the Idusi Formation of East African rift basins (Kreuser, 1995; Weiss and Wopfner, 1997), and with the coals and sandstones of the Mchuchuma Formation of southern Tanzania (Kreuser, 1995; Wopfner, 1994). The top of the Sakoa is marked by the shallow marine Vohitolia Limestone, which contains Early Permian palynomorphs (Wescott and Diggens, 1997). Hankel (1994) correlates this limestone with carbonates of the Ruhuhu Formation in the Ruhuhu Basin, southern Tanzania.

2.3.2 Sakamena Group

The Sakamena Group is made up of three units, the Lower Sakamena Sandstone, the Middle Sakamena Shale, and the Upper Sakamena Sandstone (Fig. 7). According to Wescott and Diggens (1998) the combined thickness of these units is up to 4,000 m. The Lower Sakamena Formation is Middle-Late Permian in age and consists predominantly of sandstones and conglomerates, with subordinate siltstones, mudstones, and shales, together with some stromatolitic, pisolitic, oolitic, and bioclastic limestones in the upper part (the "Vatambe facies" of Wescott and Diggens, 1998). The sandstones and conglomerates were deposited as series of alluvial fans and fan-deltas that spread progressively into developing half grabens (Clark, 1996; Montenat et al., 1996; Wescott, 1988; Wescott and Diggens, 1998). Shallow marine conditions prevailed in the half-grabens, with thick offshore mudstones being deposited in the deeper, more central parts, and deltas prograding in from the margins to form coalescing sheets. The limestones are interpreted as shallow marine sand shoals, intertidal beach deposits and algal reefs that formed around the marine fringes of fan-delta complexes (Clark, 1996). Time equivalent deposits in East Africa belong to the Hatambulo Formation of the Luwegu Basin and probably, in part, to the Usili Formation of the Ruhuhu Basin (Hankel, 1994; Kreuser, 1995).

The Middle Sakamena Formation is Early Triassic in age and comprises dark grey shales and nodular or laminated mudstones, together with local heterolithic and cross-bedded sandstones. The Middle Sakamena Formation rests unconformably on the Lower Sakamena Formation and the basal part progressively onlaps the latter. This unconformity corresponds to the widespread marine transgression in the Early Triassic as reported by Wright and Askin (1987). Wireline log and seismic data suggest that the Middle Sakamena Formation comprises a thick sedimentary blanket across the failed rift complex of the eastern Morondava Basin. It forms a relatively uniform layer on seismic, with strong reflections at the top and bottom. The thickness ranges from about 200 m at outcrop to about 650 m in the subsurface (near the Manandaza-1 borehole, Fig. 6). The shales of the Middle Sakamena Formation at Vohibasia-1, can be correlated with the Maji ya Chumvi Formation fish bed in

Kenya, as described by Hankel (1992), according to unpublished work by John Utting (OMNIS archive). Further correlation within East Africa suggests a correlation with the Sumbadsi Member of the Hatambulo Formation of the Selous Basin in southern Tanzania (Wopfner, 1994). The environment of deposition of the Middle Sakamena Formation has been variously interpreted as marine, lagoonal, and lacustrine, based on different faunal and floral assemblages in the shales. When taken in stratigraphic context, these assemblages suggest that conditions probably varied from marine in lower part of the Middle Sakamena Formation, to fluvio-deltaic in the middle and, finally, lacustrine in the upper part. Wescott and Diggens (1998), however, did not find any marine indicators and they interpret the outcrops of laminated mudstones and sandstones as fluvial and lacustrine deposits.

The Upper Sakamena Formation is Early-Middle Triassic in age and is made up of predominantly sandstones (Besairie and Collignon, 1972; Daly et al., 1989; Wescott, 1988). These sandstones rest conformably on the mudstones and siltstones of the Middle Sakamena Formation. Whereas the Lower Sakamena Formation is restricted to wedge-shaped bodies in half-grabens, the Upper Sakamena Formation comprises a thick and relatively uniform blanket of sediment that stretches across the wider grabens, which became established in the Early Triassic. According to an unpublished field study by Vroon ten Hove (performed on behalf of Shell in 1993 and made available to us by OMNIS), the lowest part of the Upper Sakamena Formation comprises heterolithic sandstones, followed by bioturbated and cross-bedded, coarsening-upward sandstones, and finally succeeded by fining-upward, festoon and trough cross-bedded sandstones. Besairie and Collignon (1972) interpreted the Upper Sakamena Formation as a continental deposit, but more recent studies suggest sedimentation in a range of shallow marine, shoreface, deltaic and fluvial environments. These include distributary channel and mouth bar environments on the delta front and delta plain, inter-distributary bay and crevasse splay environments, and braided and low-sinuosity fluvial channels in an alluvial plain environment (Nichols and Daly, 1989; Vroon ten Hove, 1993; Wescott, 1988; Wescott and Diggens, 1998). Correlative siliciclastic successions of Early Triassic fluvial and alluvial sediments occur in some of the East African rift basins, such as the Kingori Sandstone and the Manda Beds of the Ruhuhu Basin in southern Tanzania, (Kreuser, 1995; Wopfner, 1994; Wopfner, 2002).

2.3.3 Isalo Formation

The Isalo Formation comprises a uniform succession, up to 6,000 m thick, of cross-bedded sandstones and conglomerates with locally developed root horizons and petrified tree trunks (Piqué et al., 1999; Wescott and Diggens, 1998). These sediments are interpreted as fluvial channel deposits and braided stream complexes. Besairie and Collignon (1972) and Luger et al. (1994) assigned an Aalenian age to the uppermost part of the Isalo Formation in the vicinity of Sakaraha (Fig. 3). This age is questionable, however, because cross-bedded sandstones in the Andafia Formation (discussed below) are commonly confused with Isalo sandstones. Superficially, these sandstones bear a strong resemblance to one another but closer examination of outcrops around Sakaraha has revealed immature textures, with angular quartz grains and abundant feldspar and mica grains, more typical of the Isalo Formation. The Andafia sandstones probably represent eroded and reworked Isalo sandstones, and therefore they are much more mature, with rounded quartz grains and much less feldspar and quartz. Besairie and Collignon (1972) also assigned an Aalenian age to the Isalo Formation along the south-eastern rim of the Bemaraha plateau at the Morondava River in the northern Morondava Basin. This age was recently revised by Burmeister et al. (2000). They proposed a Norian (Late Triassic) age for the horizon directly below the Andafia Formation, based on the discovery of a vertebrate fauna in the Isalo Formation about 20 km west of Malaimbandy (Fig. 6). Nevertheless, the age of the Isalo sandstones in the southern Morondava Basin, particularly around Sakaraha, remains unknown. The term “Isalo” as currently applied in Madagascar incorporates sandstones of widely different ages and belonging to several separate depositional sequences. Isalo I and Isalo II are most probable Triassic in age whereas Isalo IIIa, IIIb, and IIIc sandstones are Early and Middle Jurassic. The only common feature that they share is that they are all cross-bedded and yet they are traditionally combined together as the Isalo Formation, or “Isalo Series” (Besairie and Collignon, 1972; Piqué et al., 1999; Wescott and Diggens, 1998). We believe that only the Isalo I and II sandstones are true Isalo sandstones, as defined initially by Besairie and Collignon (1972). These two sandstones belong to the same depositional sequence and can only be distinguished from one another by slight differences in lithology and colour. Consequently, we regard the Isalo sandstone as one sedimentary unit and give it the rank of “Formation”. The Isalo III sandstones are Jurassic in age and should not be included in the Formation as defined herein. Time equivalent strata of the Isalo Formation in East African are mainly fluvial sandstones belonging to the Mahogo, Luwegu, and Madaba formations (Kreuser, 1995).

2.3.4 Andafia Formation

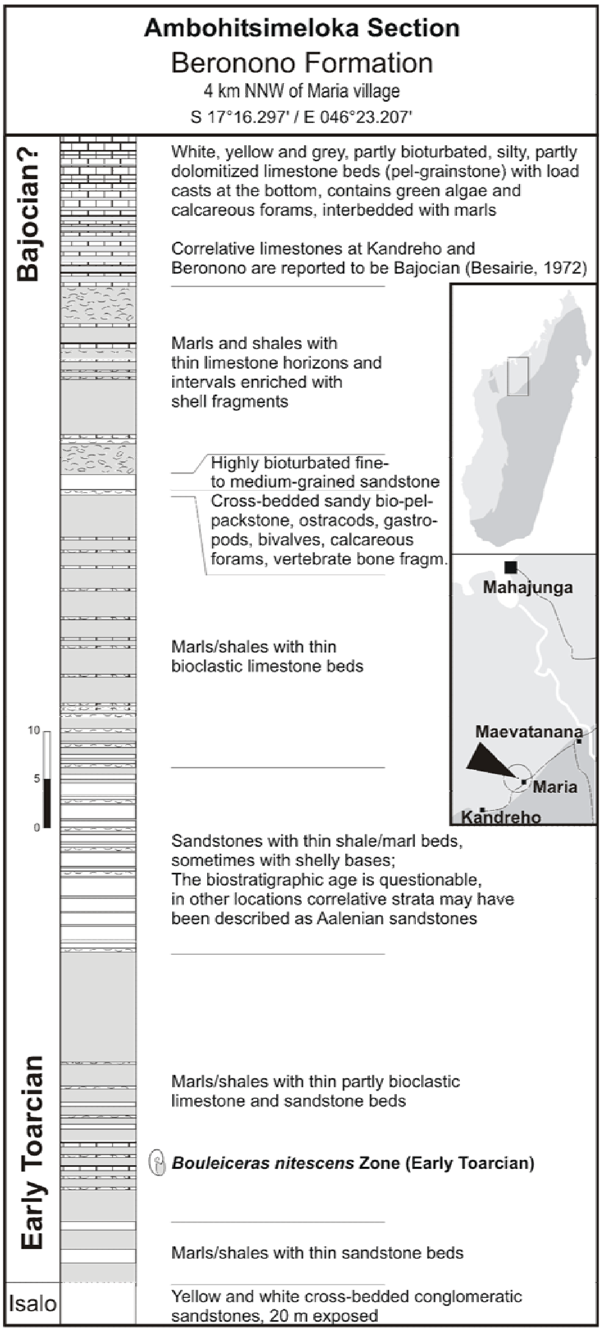
The Andafia Formation rests unconformably on the Isalo Formation. It consists predominantly of marine shales and marls, together with thin limestones and cross-bedded sandstones. Seismic, borehole, and outcrop data from the central and northern Morondava Basin, suggest that these sediments onlap tilted and rotated fault blocks of Isalo sandstone to form wedge-shaped bodies within half-grabens (Clark, 1996). The Andafia Formation is laterally equivalent to the Beronono Formation of the Majunga Basin, and belongs to the same depositional sequence. The Andafia and Beronono formations range from Toarcian to Aalenian in age (Fig. 7) and may even reach into the earliest Bajocian in the northern Morondava and the Majunga basins, according to unpublished biostratigraphic data recorded by Shell (de Jongh, 1990). Besairie and Collignon (1972) assigned a Toarcian date to outcrops of the Andafia Formation, based on the occurrence of typical ammonites. He also reported Toarcian and Aalenian ammonites from the Manera-1 and Ambatolahy-1 boreholes of the southern Morondava Basin. The earliest reported dates for the Beronono Shale in the Majunga Basin are Early Toarcian (*Bouleiceras nitescens* Zone, Besairie and Collignon, 1972).

An understanding of the stratigraphy and composition of the Andafia and Beronono formations is critical to the interpretation of the breakup history of Gondwana. A perusal of the literature, however, suggests that the stratigraphic relationships within the formations are confused and poorly defined. The first attempt to clarify these relationships was made by Boast and Nairn (1982). They produced a comprehensive correlation of the Jurassic based on the published data of Besairie but they also incorporated a number of the inconsistencies inherent in Besairie's work. One such inconsistency concerns the use of the term "Isalo III". This term refers to cross-bedded sandstones of similar character to the Isalo Formation but deposited much later. Some of these sandstones clearly belong to the Andafia Formation (e.g. the Aalenian sandstones of Besairie and Collignon, 1972), but others can be attributed to younger formations such as the Sakaraha, Ankazoabo, and Sakanavaka formations. Thus the stratigraphic range of the top of Andafia Formation is uncertain, which leads to difficulties in dating the ending of the breakup event in Madagascar. Similar inconsistencies have also been noted for some East African basins. Below we argue that the top of the Andafia Formation is Aalenian in age.

2.3.4.1 Ambohitsimeloka Section

To date, only a few incomplete outcrop sections of the Andafia Formation are known in the Morondava Basin. Therefore we measured and described a complete stratigraphic succession of the time-equivalent Beronono Formation in the southern Majunga Basin, where exposures can be found at the escarpment at Ambohitsimeloka, close to Maria (Fig. 6, Fig. 8). These exposures are near the Beronono and Kandreho sections that were described by (Besairie and Collignon, 1972), where, the Beronono Formation rests directly on Isalo sandstone. The lower part of the formation comprises 35 m of shales and marls, interbedded with a few decimetre- or metre-bedded sandstones and some centimetre- or decimetre-bedded shelly limestones. The middle part of the formation consists of about 25 m of sandstones, alternating with decimetre-scale shale beds. The sandstones contain shell fragments at some levels. The thick sandstones are followed by about 20 m of shales and marls, together with few thin centimetre-scale sandstone and limestone beds, and capped by a 2 m thick sandstone bed with oyster shells and bone fragments. The upper part of the formation is made up of approximately 20 m shale with oyster fragments and a few thin bioclastic limestones. The limestones become progressively thicker and more common towards the top of the section and they can be correlated with the basal part of the Bemaraha Formation (Besairie and Collignon, 1972).

An ammonite assemblage comprising *Bouleiceras arabicum*, *B. colcanapi*, *B. nitescens,* and *Protogrammoceras madagascariensis* was recovered from the basal shales. This assemblage belongs to the *Bouleiceras nitescens* Zone, which indicates an Early Toarcian age according to Collignon et al. (1959). The limestone beds at top of the section can be correlated with similar limestones in Kandreho section, 5 km further to the west, where they are dated as Bajocian (Besairie and Collignon, 1972). No typical Aalenian fossils were encountered in this section. Thin sections of the limestones have



revealed a rich fauna of calcareous foraminifera comprising indeterminate lenticuline and dentaline nodosariid forms. Indeterminate valvuline foraminifera and ostracods are also present in minor amounts, together with fragments of thick bivalves, gastropods, echinoderm spiculae and plates, and plant debris. The shales and marls, in contrast, did not yield any microfossils. The limestone beds in the upper part of the section contain algae of the indeterminate genus *Salpingoporella* sp.

Fig. 8. The Ambohitsimeloka section is an example of the Beronono Formation in the Majunga Basin, 4 km north-west of Maria village. This is equivalent to the Andafia Formation of the Morondava Basin.

The shales are interpreted as shallow, open marine deposits, while the bioclastic limestones are considered to slightly shallower, inner shelf deposits influenced from time to time by storms. The sandstones may possibly represent a prograding shoreline but poor outcrops have hampered detailed interpretation. Elsewhere, planar cross-bedded and bioturbated sandstones are developed and these are interpreted as nearshore shelf and shoreface sediments. Local trough cross-bedded sandstones are thought to represent either incised valley fill or channels associated with fluvio-deltaic sedimentation (Clark, 1996). Such sediments are consistent with the overall concept of Ramanampisoa et al. (1997), who interpreted the deposition of the Toarcian sediments as a response to episodic syn-sedimentary block-faulting and tilting.

2.3.4.2 Andafia Road Section

The Andafia Formation is not exposed in the southern part of the Morondava Basin, but the lower 10-20 m of the succession crops out along the road between Malaimbandy and Ankilizato and along the Tsiribihina River (Fig. 6), in the northern part of the basin (Clark, 1996), where the Andafia Formation rests on Isalo sandstones with an erosional unconformity. It is made up mainly of shales and marls, interbedded with subordinate limestones and rare thin siltstones or very fine sandstones.

The basal part consists of decimetre-bedded limestones, interbedded with slightly thicker shales. The shales are dark grey or black and laminated or massive. The limestones are massive or nodular, possibly as a result of bioturbation. They are also fossiliferous in places, with moulds of ammonites and bivalves (particularly large and distinctive trigonids). The upper bedding surfaces of the limestones are irregular or wavy. In some cases, these surfaces might be palaeokarsts, possibly associated with soil profiles. In other cases, the limestones have a domed or columnar morphology, suggestive of stromatolites. According to Clark (1996), the shales and limestones are shallow water, open marine deposits, and the sandstones a prograding shoreline associated with fluvio-deltaic sedimentation (Fig. 10a).

2.3.4.3 Borehole Data

Marine sediments of Toarcian and Aalenian age have been encountered beneath the Bemaraha Formation in the Ambatolahy-1 and Manera-1 boreholes in southern Morondava Basin (Fig. 6)(Besairie and Collignon, 1972; Uhmann, 1996), and also in the Tandrano-1, Ankazofotsy-1, Mandabe-1, and Manja-1 boreholes (Clark, 1996). These sediments are tentatively correlated with those exposed in the Andafia section near Malaimbandy. The Andafia Formation has also been penetrated by a number of boreholes in the central and northern Morondava Basin, including Ankamotra-1, Chesterfield-1, East Serinam-1, Mamakiala-1, Manambolo-1, Namakia-1, Serinam-1, and West Kirindy-1 (Clark, 1996). The thickness of the formation generally varies from 30-600 m, but thicknesses of over 1,900 m appear to be developed at Mamakiala-1 and Chesterfield-1.

2.3.5 Bemaraha Formation

The Bemaraha Formation is made up predominantly of massive limestones, including carbonate mudstones, pelletoidal-grainstones or oolitic-grainstones. These limestones crop out extensively in the northern Morondava Basin. Typically, they form a lenticular body that runs along the eastern edge of the Morondava Basin, adjacent to the failed Permo-Triassic rift (Fig. 16)(Clark and Ramanampisoa, 2002). This concept was originally proposed in several unpublished reports by oil companies working in Madagascar and further developed by Clark (1996) and du Toit et al. (1997) in unpublished reviews complied for OMNIS. The lens varies in thickness from 30-1,000 m and is clearly evident on many of the seismic lines in the northern Morondava Basin. The thickest part of the lens is exposed on the Bemaraha Plateau, in the north-eastern Morondava Basin. To the south, the axis of the lens passes into the subsurface and the thickest successions are encountered in wells such as Betsimba-1 and East Serinam-1 (Fig. 6). The Lower Bemaraha Formation is dated as Bajocian and the upper part is Bathonian (Besairie and Collignon, 1972; Montenat et al., 1996). A carbonate unit of the same age has been described from Tanzania, where marine limestones and oolites of Bajocian age form the Mtumbei Series and Kidugallo limestones (Kreuser, 1995), and from Somalia where it is known as the Hamanlei Fomation (Kreuser, 1995; Luger et al., 1990).

At outcrop, the massive limestones appear to conformably overlie the Andafia Formation. This transition can be observed along the road to Morondava, west of Malaimbandy, where thinly bedded limestones at the top of the Andafia Formation pass upwards into the more massive Bemaraha Limestone (Clark, 1996). Rhodoliths, ooids, and echinoid debris also appear at the base of the Bemaraha Formation and the limestones are thicker and more common than in the upper part of the Andafia Formation. On seismic lines, however, the base of the Bemaraha Formation is clearly unconformable with the Andafia Formation, and in places it can be seen to overstep from the Andafia Formation onto the Isalo Formation or older strata on the crests of some tilted fault blocks (Fig. 16).

Clark and Ramanampisoa (2002) interpreted the Bemaraha Limestone as a coastal barrier/lagoon complex. This complex became established along the western side of Madagascar during the Bajocian and Bathonian and a thick marginal carbonate platform was constructed. The platform now forms the massive limestones of the Bemaraha Plateau, in addition to those found in the subsurface at Betsimba-1 and East Serinam-1. These limestones were deposited in environments ranging from barrier islands with beaches and tidal deltas to lagoons, upper intertidal flats, and supratidal sabkhas. A wedge composed of millimetre- and centimetre-bedded carbonate mudstones occurs basinward of the platform. This wedge is interpreted as an anoxic, submarine slope deposit. A thin unit of millimetre-bedded mudstones is present basinwards of the mud wedge. These sediments probably accumulated in a more offshore, deeper water, anoxic, basin-plain environment.

2.3.5.1 Borehole Data

According to Dina (1996) and Uhmann (1996), three wells have encountered carbonate intervals, that might be equivalent to the Bemaraha Formation, in the southern Morondava Basin (Tandrano-1, Manera-1, and Ambatolahy-1, Fig. 6). These wells were originally drilled in the 1950s by SPM and they have been reviewed and reinterpreted by Dina (1996) and Uhmann (1996). These authors have attempted to date the carbonates using palynomorph assemblages. They also reviewed the sedimentology of the Middle Jurassic in the wells but it is not clear whether their descriptions are based on a re-examination of drill cuttings, or whether they merely reinterpreted the SPM descriptions. Their descriptions also lack any detailed information about the thicknesses of the different lithofacies in the wells.

The carbonate interval in Tandrano-1 is dated as Bajocian and may possibly extend into the Early Bathonian, depending on the reliability of the SPM cuttings descriptions and if there were substantial cavings in the well from overlying sandstone units. According to Dina (1996) and Uhmann (1996), the succession comprises a lower unit of thick oolitic and bioclastic limestones, together with thin marly limestones and possible sandy or silty intercalations, followed by a succession of questionable sandstones (or possibly siltstones and/or sandstone cavings?), marls, limestones, and recrystallized oolites. A similar carbonate interval is seen at Manera-1, although the only the lowermost sediments have been dated as Bajocian. Dina (1996) and Uhmann (1996) noted local plant debris in this well and that dolomitisation appears to have taken place at some levels. They also noted “reef limestones”, based entirely on the interpretation of drill cuttings from the 1950s. It is more likely that these “reefs limestones” reflect the presence of coral meadows, similar to those recently described from outcrops near Betioky (Fig. 6,)(Lathuilière et al., 2002). At Ambatolahy-1, the carbonate interval is dated as Bajocian-Early Bathonian. Here, it is made up of about 600 m of predominantly massive oolitic and dolomitic limestones, together with thin shales and marly limestones, followed by interbedded marls, oolitic limestones and calcareous sandstones. Dina (1996) and Uhmann (1996) interpreted the carbonates from all three wells as representing marginal marine and lagoonal environments, but the significance of the sandstones was not discussed. They also noted that a distinct difference between the carbonate-dominated lithofacies in the boreholes and the siliciclastic-dominated lithofacies as seen at outcrop further to the east.

2.3.6 Sakaraha Formation

The Sakaraha Formation is only found today in the southern Morondava Basin where it occurs at outcrop and in boreholes. Further north, the massive limestones of the Bemaraha Formation are exposed on the Bemaraha Plateau and the Sakaraha Formation has been removed by erosion. Besairie and Collignon (1972) described outcrops of Bajocian and Bathonian sediments in the south as a mixture of siliciclastics and carbonates (the so-called “Facies Mixte”). The carbonates are similar to those typically found in the massive limestones of the Bemaraha Formation, whereas the siliciclastics comprise varying admixtures of shale, marl, siltstone, and sandstone. The Bajocian sediments are generally attributed to the Sakaraha Formation, whereas those of Bathonian age are variously assigned to the Ankazoabo, Sakanavaka, Mandabe or Besabora formations. Besairie and Collignon concluded that these formations are laterally equivalent to the massive carbonates of the Bemaraha Formation, and he developed the “Facies Mixte” concept based on this relationship. This concept was introduced into the stratigraphic nomenclature of Madagascar by SPM in the early 1950s and was widely used by SPM field geologists and by Besairie in the production of geological maps (e.g. Besairie, 1969b). Clark and Ramanampisoa (2002), however, have suggested that only the Sakaraha Formation is laterally equivalent to the Bemaraha Formation. This conclusion was based on borehole, seismic, and outcrop relationships in the northern Morondava and Majunga basins but no dating of the sediments was undertaken to confirm it.

In the present study, four sections of the Sakaraha Formation were examined at outcrop in the southern Morondava Basin to better define the relationship with the Bemaraha Formation. These sections are at Analamanga, Adabomjonga, Tongobory, and Anjeba. The type locality at Sakaraha village was also examined for comparison. Fossil evidence from these sections suggests that the Sakaraha Formation ranges from Bajocian to Early Bathonian, making it time equivalent to the Bemaraha Formation. The remaining formations are almost certainly younger than the Bemaraha Formation and belong to subsequent depositional sequences. The sandstones of the Ankazoabo

Formation, for example, rests on the Sakaraha Formation with an erosional base (as discussed below). It is currently dated as Bathonian but this date is poorly constrained and could be attributed to the presence of reworked Early Bathonian palynomorphs. Thus it is possible that the Ankazoabo Formation is Late Bathonian or possibly Early Callovian in age and that it belongs to a younger depositional sequence than that of the Bemaraha Formation. Clark and Ramanampisoa (2002) interpret the Sakaraha succession as a coastal plain association that accumulated to the east of the barrier/lagoon complex of the Bemaraha Formation (Fig. 14b). A similar association of lithofacies is also found in the Mtumbei clay of the Mtumbei Series of Tanzania (Fig. 7)(Kreuser, 1995).

2.3.6.1 Sakaraha Section

The area around Sakaraha is the type locality for the Sakaraha Formation (Besairie and Collignon, 1972). Uhmann (1996) and Dina (1996) measured a hill section along the main road to Toliara, about 2 km west of Sakaraha. We revisited this section, and although the contact between the Isalo sandstone and the overlying Sakaraha Formation is visible, the remainder of the succession is only poorly exposed. As an alternative, we selected a fresh road cut at Analamanga (see below) for detailed measurement. Here, the succession starts with an oolitic-grainstone that rests uncomfortably on the Isalo Formation. The grainstone is composed almost entirely of superficial ooids with nuclei of quartz grains that were probably reworked from the underlying Isalo sandstones (Fig. 10a). The grainstone is followed by a thinly-bedded succession of marls, sandstones, and limestones. The sandstones are composed of quartz and are generally cemented by carbonate. Calcareous ooids and bioclasts occur in some of the sandstones as well as in some of the limestones.

Besairie and Collignon (1972) dated the exposed succession at Sakaraha as Bajocian, based on the occurrence of the bivalve *Trigonia tenuicostata* Lycett. Uhmann (1996) and Dina (1996) also dated the Sakaraha section as Bajocian using the ostracods *Fastigatocythera malgachica* Grekoff and *Monoceratina striata* Triebel & Bartenst, and a similar age was indicated by the ostracod fauna recently recovered by Mette and Geiger (in press). Uhmann (1996) and Dina (1996) interpreted these sediments as representing an intertidal to shallow subtidal, open lagoonal environment.

2.3.6.2 Analamanga Section

The Analamanga section (Fig. 9) is approximately 10 km NNE of Sakaraha along the road to Ankazoabo. The sediments exposed at this location closely resemble those seen at the type locality at Sakaraha. Here, the Sakaraha Formation starts with a 25 cm transgressive oolitic-grainstone bed (Fig. 10b), which rests unconformably on Isalo sandstones. The basal oolite consists of mostly superficial ooids with quartz nuclei of medium sand grade. It is succeeded by approximately 45 m of silty marl, interbedded with limestones and sandstones, then a 4 m oolitic-grainstone (Fig. 10e), and finally 1 m of thinly interbedded sandstones and marls. Individual beds of marl are up to 5 m thick and contain bivalves, gastropods, wood, crinoids, algae, ostracods, and foraminifera. Orientated wood fragments occur in the marls, indicating E-W currents and flute casts suggest an easterly flow direction. The limestones are 20-40 cm thick and generally consist of bioclastic-oolitic-packstones (Fig. 10d) with echinoderm and bivalve fragments. The sandstones are 0.1-3 m thick and consist mainly of medium sandstones with well-developed carbonate cements (Fig. 10c). Most of the sandstone beds appear to be laterally persistent but a few channelised units occur in the middle of the marl succession. Some of the limestones and sandstones display symmetrical ripples, suggesting wave activity. Sandstones With low-angle trough cross-bedding and some bioturbation are common in the upper part of the marl succession.

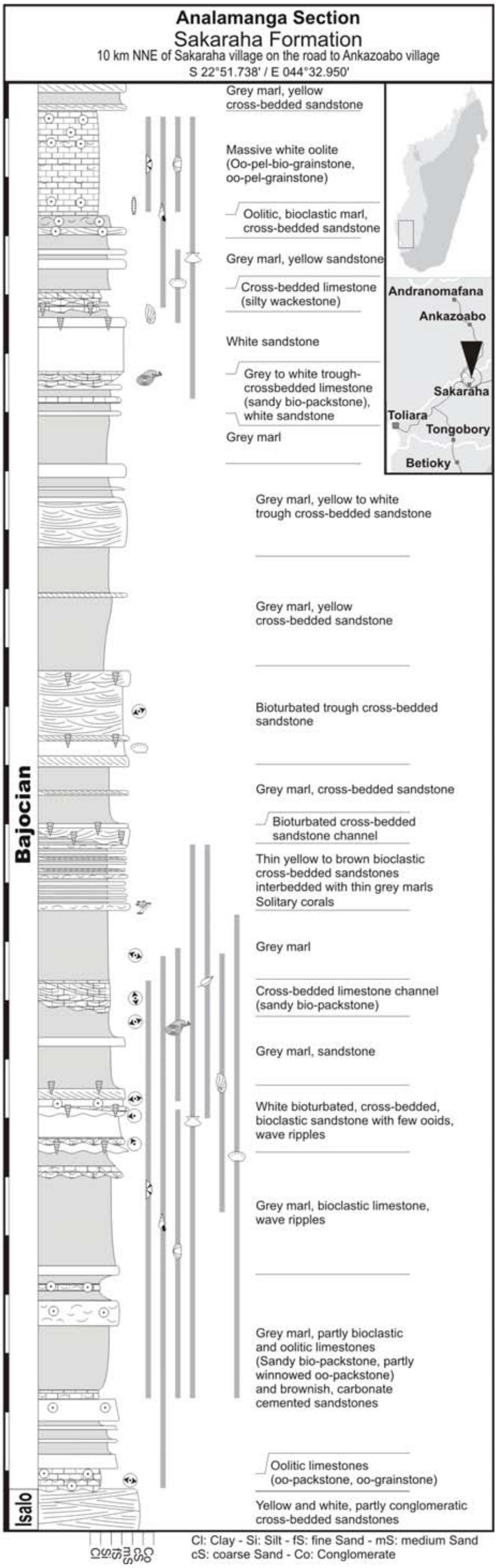


Fig. 9. The Analamanga and section is an example of the Sakaraha Formation in the southern Morondava Basin, 10 km NNE of Sakaraha village. The section is lithologically and stratigraphically correlative to the Sakaraha section (see text). For legend see Fig. 13.

Fossils, such as rhynchonellids, bivalves, gastropods, and echinoderms are generally concentrated in the upper and lower parts of the marl section, while the central part is almost devoid of fossils. Two levels are characterised by the appearance of brackish bivalves, one at the base (*Pronoella* sp.) and the other at top (*Tancredia* sp.). *Protocardia* sp., *Ceratomya* sp., *Frenguelliella* sp., *Modiolus* sp., *Pleuromya uniformis*, *Bakevellia* sp., and indeterminate lucinid forms are also present in the upper and lower parts of the marl section. Other macrofossils include plant debris in the middle of marl unit, together with several tree trunks with diametres up to 30 cm, one level of solitary corals, and possible biohermal structures in the upper part of marl unit. Microfossils such as the alga *Cylindroporella* sp. are present in the massive oolite and indeterminate lenticuline foraminifera and *Palmula* sp. occur in the upper marl section. A low diversity ostracod assemblage dominated by the *Cytheruridae* and *Progonocytheridae* has also been found at this location (Mette and Geiger, in press). This ostracod assemblage strongly resembles the one recovered from the Sakaraha section (see above). It correlates with similar assemblages from Western Australia, South America, Near East, North Africa, and Europe and indicates a Bajocian age (Mette and Geiger, in press).

The appearance of an oolitic-grainstone at the base of the Analamanga section marks the onset of the Early Bajocian marine transgression in the southern Morondava Basin. It also signals a sharp change of environment from the continental sandstones of the Isalo Formation to a marine limestone. Initially, a series of high-energy shallow-marine sand shoals were formed, accompanied by the widespread development of ooids. These shoals appear to have prograded rapidly to the west and inner lagoonal, intertidal, supratidal marsh, and coastal swamp environments became established. As a result, a thick succession of marls and subordinate limestones and sandstones was deposited above the basal oolite. The low diversity of the fauna of ostracods and brackish water bivalves confirms the existence of intertidal and brackish lagoonal or swamp conditions, whereas the presence of solitary corals and bioherms, together with rhynchonellids, is more indicative of subtidal, outer lagoonal conditions.

Repeated flooding events appear to have occurred in the Sakaraha Formation, as evidenced by the presence of interbedded oolitic-grainstones and sandstones. Episodic deepening and the establishment of subtidal conditions is also possibly recorded by the presence of bioclastic-packstones with particularly thin shells (Fig. 10d) in the lower part of the marl unit, although some of these bioclastic beds might represent supratidal beach ridges. With regard to the origin of the oolitic and bioclastic sandstones in the Sakaraha Formation, some undoubtedly formed as shoreface sands. The large, channelised sandstone in the middle of the marl unit can be interpreted as a distributary channel, although it is also possible that it formed by the infill of an incised valley during a lowstand in sea level.

The massive oolite at the top of the section is interpreted as a coastal oolitic barrier complex that formed under agitated shallow marine conditions, similar to those postulated for the basal oolite. Such barriers are thought to have protected the platform from open marine conditions (Clark and Ramanampisoa, 2002). The appearance of this barrier above lagoonal sediments indicates a rapid landward shift of environments as the results of a transgressive event. The sudden absence of quartz in the grainstones also supports the concept of a rapid retreat of the coastline during the landward shift of the facies belts.

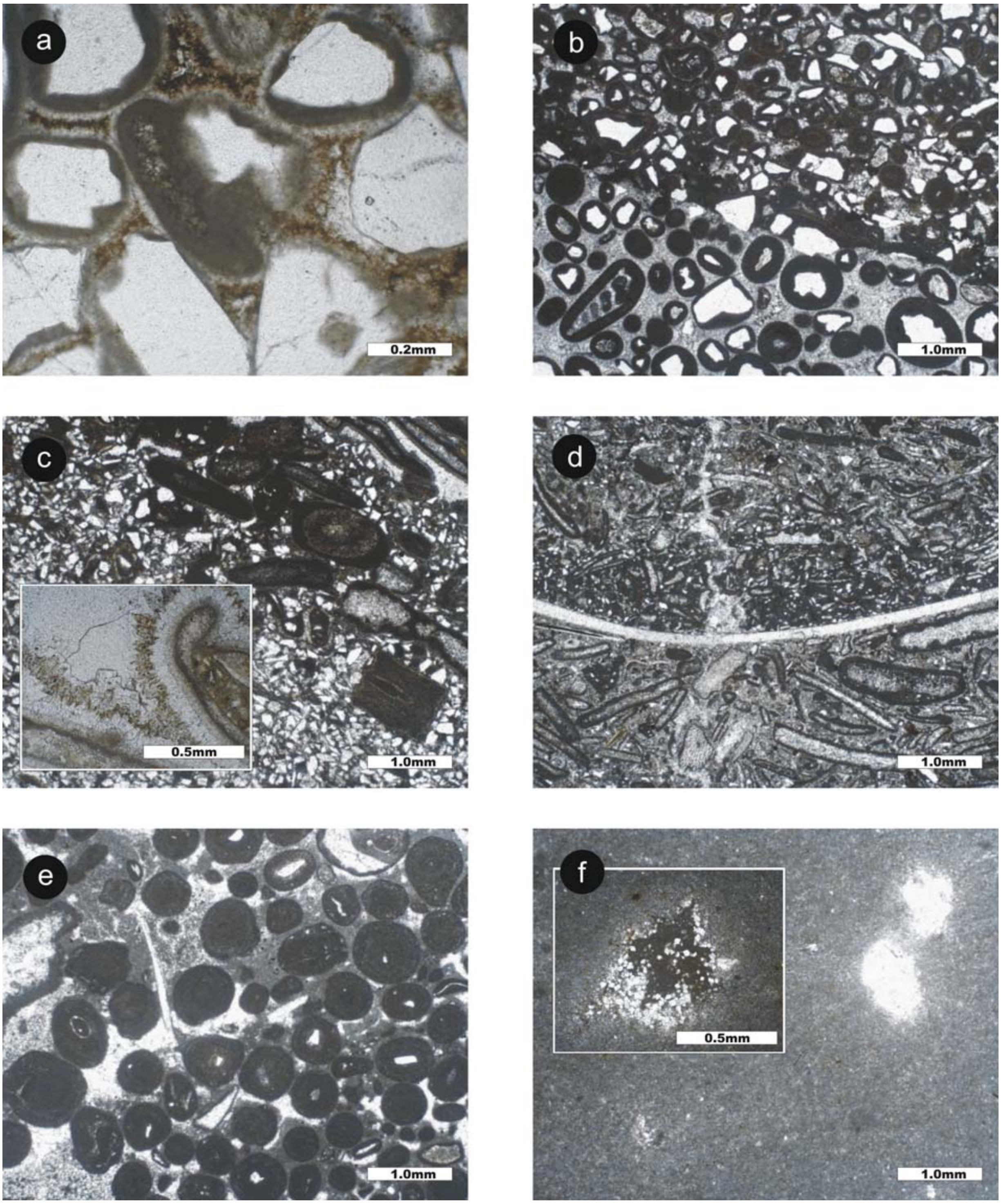
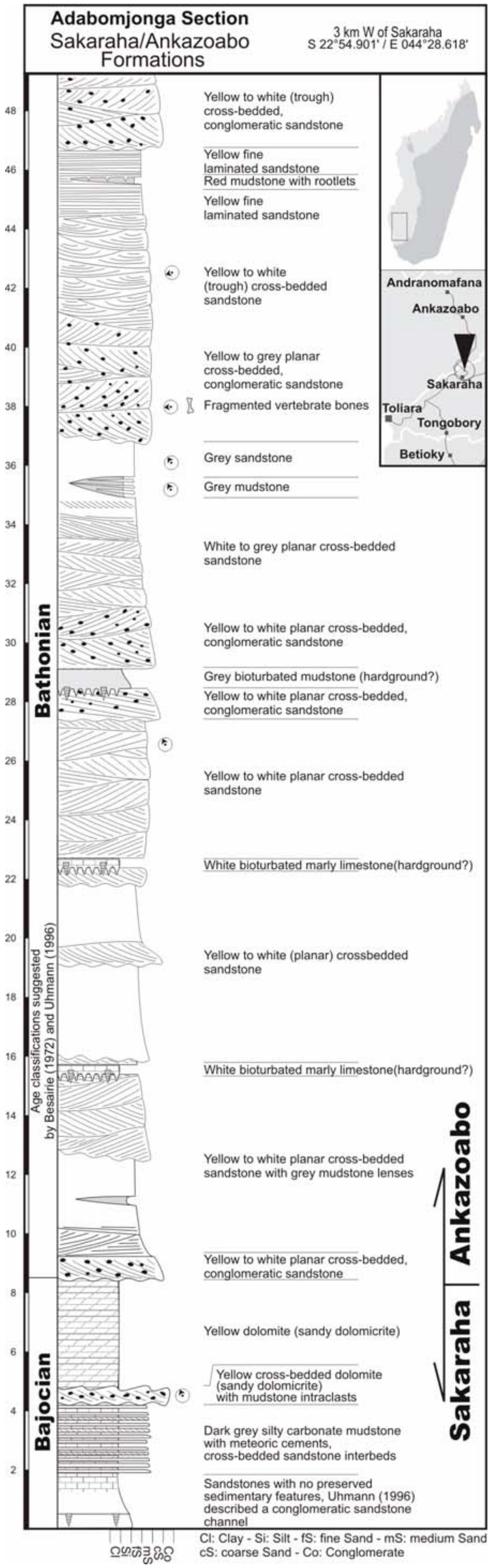


Fig. 10. Thin section microphotographs illustrating the facies associations of the Sakaraha Formation. **a** and **b,** oo-grainstone, superficial ooids with quartz nuclei, high energy shallow water, reworked Isalo Formation sandstone during transgression, Bajocian, (a, Sakaraha, road to Toliara; b, Analamanga); **c,** bioclastic fine sand-siltstone, beach/sandbar deposit, lower shoreface, successive fibrous and prismatic rim cements document particular generations of marine cements due to sealevel fluctuations, Bajocian, Analamanga; **d**, bio-packstone with thin shells, subtidal, Bajocian, Analamanga; **e,** oo-grainstone, high energy shallow water, barrier, Bajocian, Analamanga; **f,** carbonate-mudstone with vugs, meteoric phreatic cementation, emersion, intertidal mudflats, Bajocian, Adabomjonga.



2.3.6.3 Adabomjonga Section

The Adabomjonga section (Fig. 11) is about 1 km west of the Sakaraha section (see above), and 3 km west of Sakaraha village. This section was originally described by Besairie and Collignon (1972), Uhmann (1996) and Dina (1996), and re-measured as part of the current study. The base of the section can be correlated with the top of the Sakaraha Formation such as found at Analamanga (see above). This basal succession consists of 3 m of centimetre- to decimetre-bedded marls and sandstones. Individual sandstone beds are characterised by fining-upward cycles that grade up into marls. They also display small scale cross-lamination. Uhmann (1996) described a 1 m thick channel of coarse sandstones with 3 cm of quartz conglomerates at the very bottom of the section, which we were unable to find. The basal unit is capped by 30 cm of conglomeratic sandstone with an erosional base. This sandstone is trough cross-bedded in places and contains clasts of reworked marl together with quartz pebbles. The conglomerate is followed by about 4 m of homogenous dolomicrite. The top of the dolomicrite is partially eroded by conglomeratic sandstone and this in turn is succeeded by about 40 m of massive sandstone.

Fig. 11. The Adabomjonga section, 3 km W of Sakaraha village, lies stratigraphically above the Analamanga section (Fig. 5). The lower part of the section represents the top of the Sakaraha Formation. The overlying sandstones belong to the Ankazoabo Formation, according to Besairie and Collignon (1972). For legend see Fig. 13.

It has not been possible to date the sediments from the Adabomjonga section because fossils have not been recovered. Following Besairie and Collignon (1972) and Uhmann (1996), the interbedded marls, sandstones, and dolomites are assigned to the Sakaraha Formation, whereas the overlying massive sandstones are placed in the Ankazoabo Formation. A perusal of the literature suggests that the Ankazoabo Formation is of Bathonian age but the stratigraphic evidence is poor and the dating is based mainly on lithological correlations. The overall environment of deposition of the Sakaraha Formation at this location was probably inner lagoonal to supratidal. Periodic exposure of the sediments is also suggested by the occurrence of blocky sparite cements in some carbonate mudstones. These sparites typically line the insides of vugs and are characteristic of early freshwater cementation (Fig. 10f). Uhmann (1996) interpreted the channel at the bottom of the section as an intertidal channel.

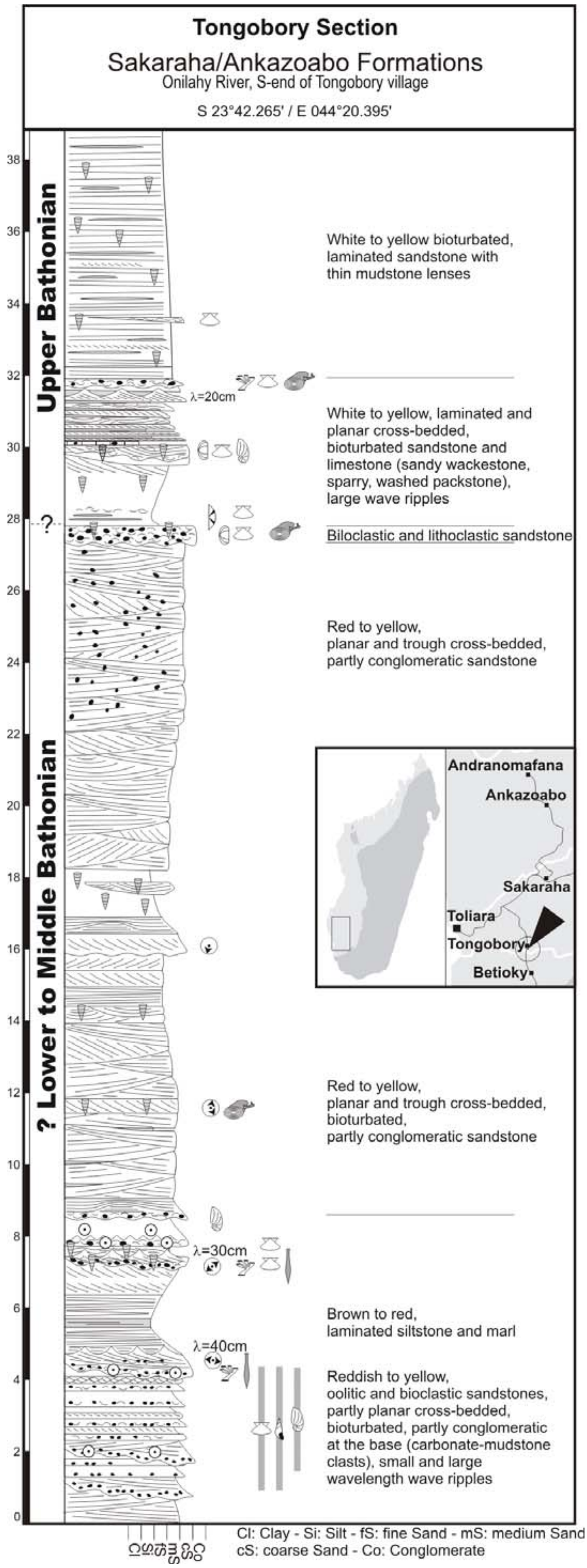
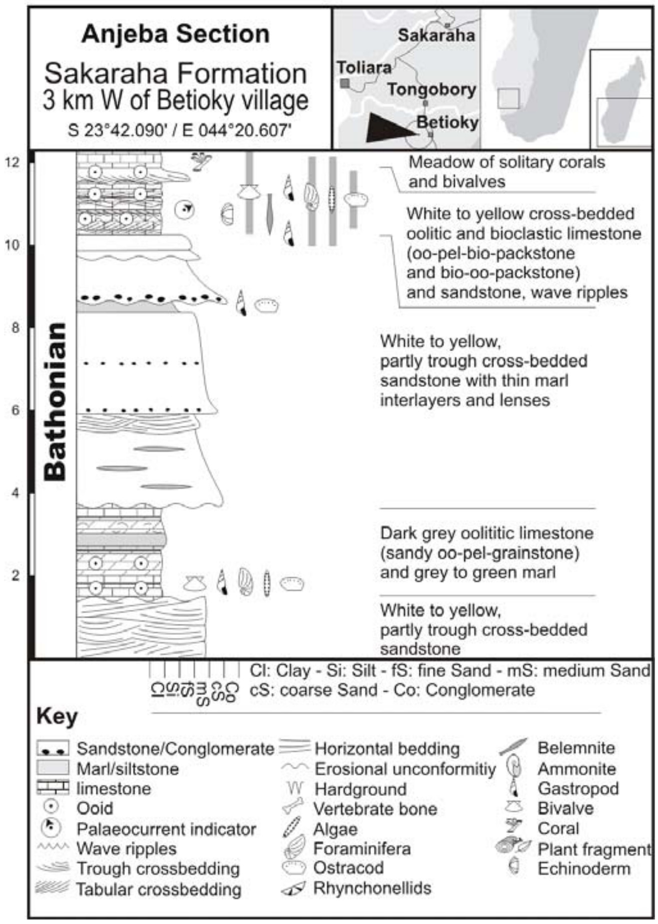


Fig. 12. The Tongobory section at the N-shore of the Onilahy River, S of Tongobory, is an example of carbonate facies in the Bathonian and argues for the extension of the range of Sakaraha into this time. For legend see Fig. 13.

2.3.6.4 Tongobory Section

The Tongobory section (Fig. 12) was described by Uhmann (1996) from the cliffs at Tongobory on the Onilahy River in the southernmost part of the Morondava Basin. The succession consists of about 40 m thick succession of cross-bedded sandstones. We found calcareous ooids in the lowermost sandstones, suggesting that the sediments



at this location might correlate with the upper Sakaraha Formation elsewhere. These oolitic sandstones are topped by several wave ripple horizons, which are overlain by sandstones with the typical sandstone facies of the Ankazoabo Formation. Uhmann dated the sediments as Middle to Late Bathonian based on an unspecified ammonite fauna.

Fig. 13: The Anjeba section, 3km W of Betioky is another example of carbonate facies in the Bathonian. Here coral meadows and oolitic sandstones and limestones suggest strong affinities to the carbonate platform environment of the Sakaraha and Bemaraha formations.

2.3.6.5 Anjeba Section

The Anjeba section (Fig. 13) is 3 km west of Betioky. This section was also described by Uhmann (1996) as the upper part of the bipartite Andamilany section. It consists of about 20 m of sandstones, marls, and oolitic and bioclastic limestones that are generally similar in appearance to the sediments of the Sakaraha Formation as described above. These sediments contain green algae, such as *Cylindroporella* sp., *Heteroporella* sp., and *Halimeda* sp., and an assemblage of ostracods as described by Mette and Geiger (in press). The nature of the sediments and the faunal assemblages that have been recovered suggest deposition under intertidal to shallow subtidal conditions in a protected lagoonal environment. Uhmann dated the sediments as Bajocian. Besairie and Collignon (1972), however, noted that the sediments are Bathonian in age and this was confirmed more recently by the analysis of the ostracod fauna (Mette and Geiger, in press). Interestingly, this is the only occurrence of carbonate sediments of Bathonian age that has been found so far in the southern Morondava Basin. Lathuilière et al. (2002) revisited the area and recovered brachiopods which favour a Late Bajocian age within strata of the similar facies. In conclusion, the Sakaraha Formation here and at Tongobory in the north appears to extend from the Bajocian into the Bathonian.

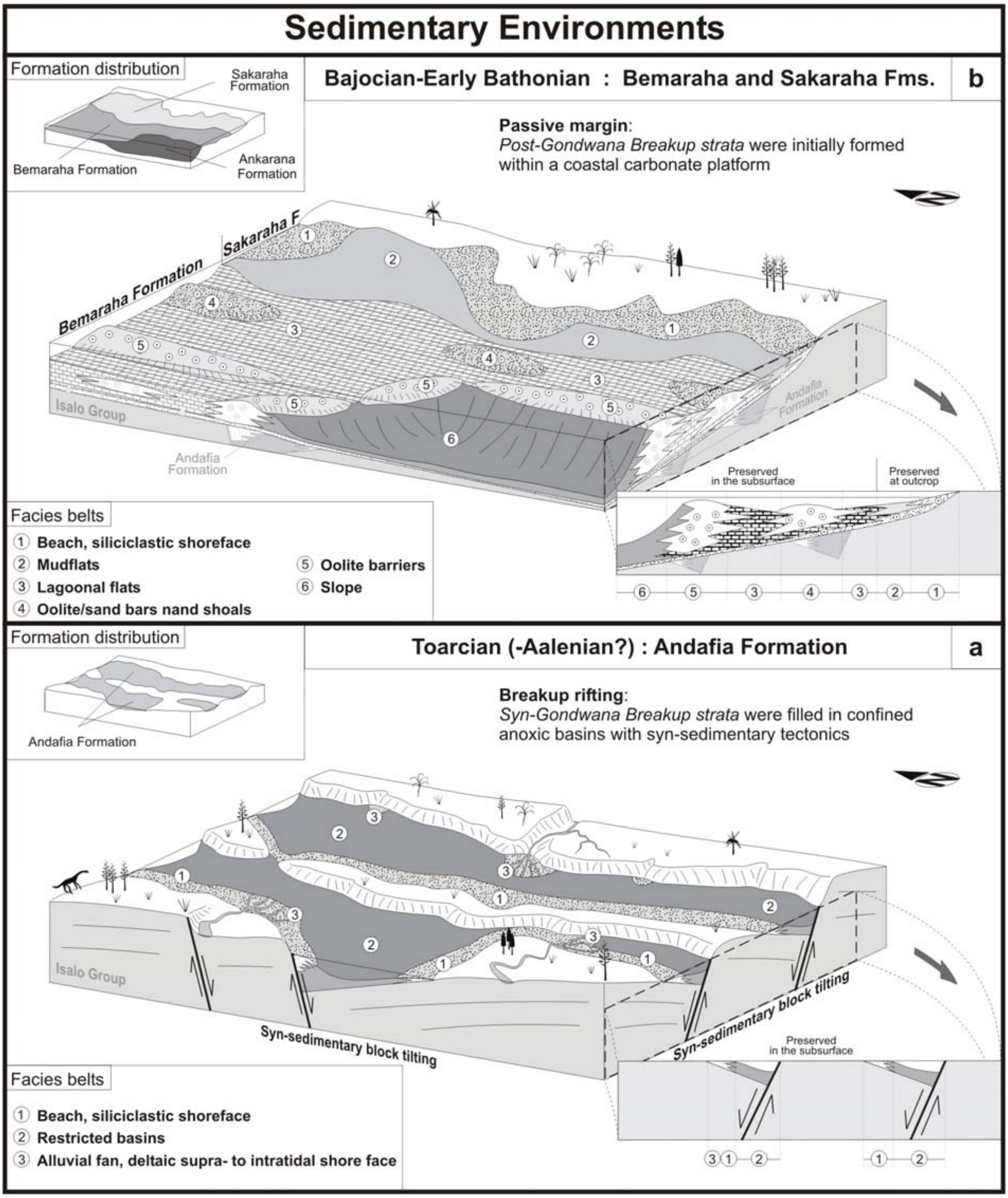


Fig. 14. Facies models of post-Karoo Jurassic strata in southern Morondava Basin based on outcrop and well data. (a) During the Toarcian/Aalenian crustal extension led to widespread faulting and subsidence. Within restricted half-grabens anoxic shallow water deposits of the Andafia Formation accumulated adjacent to continental deposits of the Isalo Formation. (b) After a major erosional event at the beginning of the Bajocian, a stable passive margin favoured the formation of a carbonate platform (Sakaraha/Bemaraha formations).

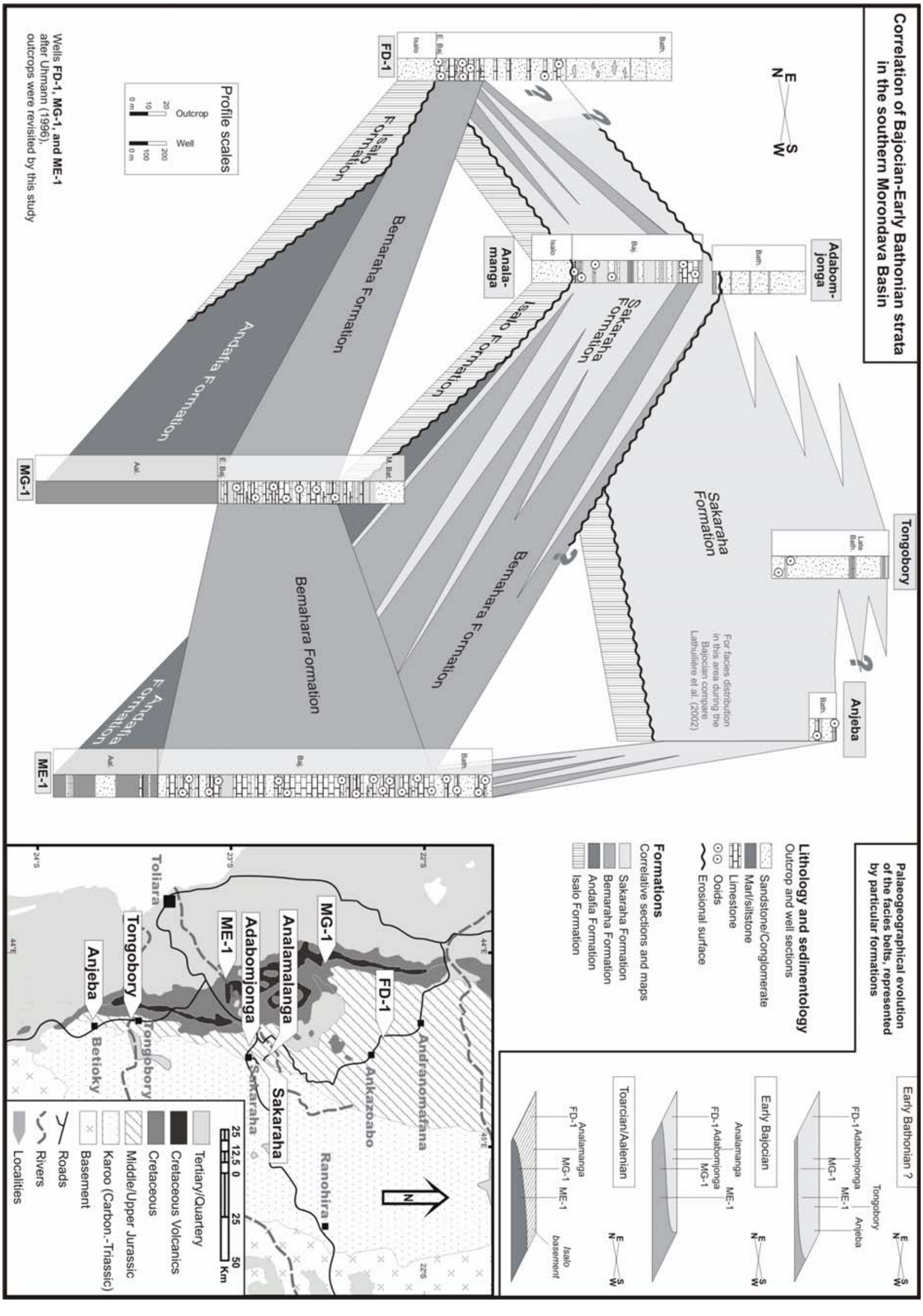


Fig. 15. Correlation scheme of the measured sections in the southern Morondava Basin and their correlation to the selected well logs Tandrano-1, Ambatolahy-1, and Manera-1 after the interpretation of Dina (1996). It shows a thick carbonate succession with massive limestones (Bemaraha Formation) in the wells that is equivalent to a marl-dominated mixed carbonate-siliciclastic succession (Sakaraha Formation) at outcrop.

2.4 Tectonostratigraphy

In order to understand the relationship between the tectonic and depositional history of the Morondava Basin, it is necessary to examine seismic sections in conjunction with borehole and outcrop data (Fig. 15). Previous studies failed to recognise the influence of tectonic activity on sedimentation because good quality seismic data were lacking (e.g. Besairie and Collignon, 1972; Montenat et al., 1996). The seismic cross-section from the Manambolo area (Fig. 16) in the north-central Morondava Basin shows the two distinct rift systems that were described by Clark (1996) and Ramanampisoa et al. (1997). A deep rift basin filled with Karoo-related deposits of Permo-Triassic age can be seen on the eastern seismic line. The western line in contrast, clearly shows a later phase of extensional faulting and rifting, together with a thick succession of post-Triassic strata.

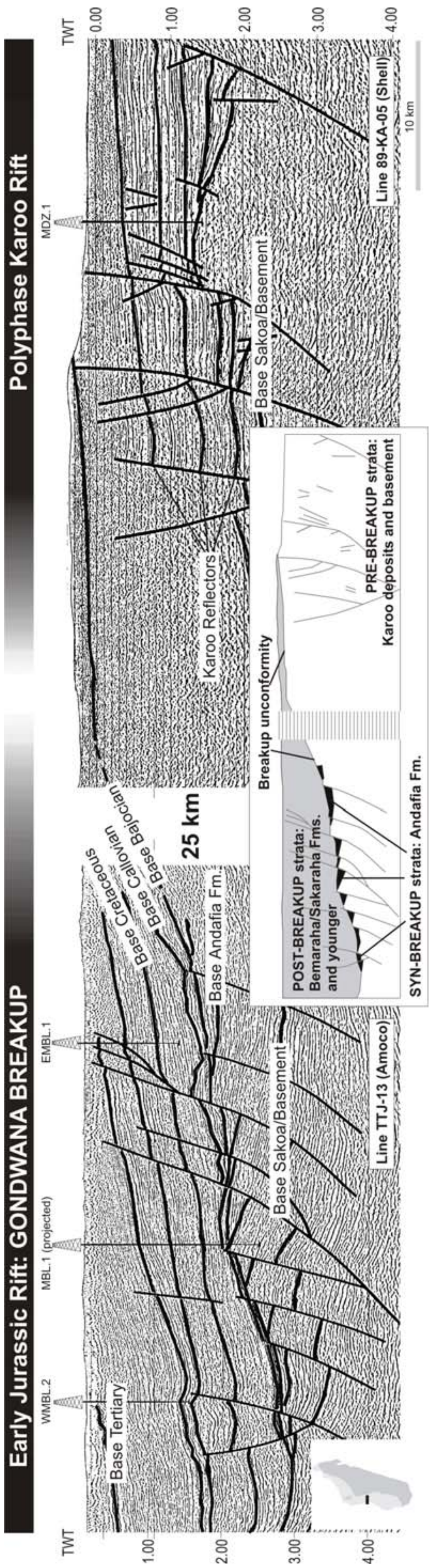


Fig. 16. The compilation of two seismic lines from the northern Morondava Basin shows a subsurface image perpendicular to basin strike. The polyphase Karoo rift (Line 89-KA-05, Shell) at the eastern basin margin is partly overlain by the Bemaraha carbonate platform, which basinward caps the Early Jurassic rift (Line TTJ-13, Amoco). The Early Jurassic rift is the Gondwana breakup rift. For well locations see Fig. 6.

The important observation for this study is that wedge-shaped bodies of sediment occur within late Early Jurassic half-grabens (Fig. 16). These wedges are composed predominantly of shales and are correlated with the Andafia Formation (Chapter 2.3.4)(Clark, 1996). The thickness of the wedges varies in relation to the pattern of block-faulting in the underlying strata (Clark, 1996; Montenat et al., 1996). In most places the shales rest unconformably on Isalo sandstones, but at some localities the Andafia Formation does not completely fill the half-grabens, and the Bemaraha Limestone rests unconformably on the Isalo Formation (Fig. 16). A thin unit of Early Jurassic (Late Liassic) shale may also have extended across the Karoo rift in the Majunga Basin. The Toarcian (Early Jurassic) half-grabens are unconformably overlain by the massive limestones of the Bemaraha Formation. Further to the east, the limestones also partially overstep the Karoo Rift sediments in the vicinity of the Bemaraha Plateau. It is clear from the seismic data that the Late Liassic faults do not generally affect the Bemaraha Formation. Thus heavy extensional faulting appears to have ceased by the Early Bajocian and the Bemaraha Limestone represents the first phase of deposition on a newly formed passive margin.

2.5 Rift Stages of the Gondwana Breakup

According to Bosence (1998), three phases of sedimentation occur during continental breakup: pre-rift, syn-rift, and post-rift. The syn-rift strata are deposited as rifting progresses, and normally show evidence of thickening into hanging wall basins near faults. The post-rift strata in contrast, are deposited after rifting, when extensional faulting has ceased, and record a prolonged period of thermal subsidence on a passive continental margin. The boundary of the pre-rift and the syn-rift stages is marked by the syn-rift unconformity, whereas the post-rift unconformity separates the syn-rift strata and the overlying post-rift strata. We have applied this model to the Morondava Basin, as follows:

(a) Rocks older than the Andafia Formation, comprising the Karoo strata, are considered to represent the pre-rift phase of the Gondwana breakup (pre-Gondwana breakup strata). These sediments were clearly deposited prior to the development of the Gondwana breakup rift because they were not influenced in the Late Liassic (Toarcian) phase of extensional faulting.

(b) The Andafia Formation and its equivalents are interpreted as the syn-rift strata of the Gondwana breakup rift (syn-Gondwana breakup strata). These sediments were clearly deposited in a series of Late Liassic half-grabens as extensional faulting occurred and the fault-blocks became rotated.

(c) The extensive erosional surface that is developed above the wedge-shaped Andafia Formation is considered to be the breakup unconformity (post-rift unconformity).

(d) The deposits that overlay the breakup unconformity (Sakaraha and Bemaraha formations) are post-Gondwana breakup strata.

Two things become clear from the application of this model. First, the Permo-Triassic or Karoo phase of rifting was not responsible for the separation of Madagascar from Africa. Second, a later rift developed to the west of the Karoo rift that eventually led to breakup. This is the “Andafia rift”, which is in Late Liassic age. The Early Bajocian unconformity was recognised by Montenat et al. (1996) but they did not interpret it as a breakup unconformity. Instead, they accepted the concept of a long lasting transitional phase between rifting and drifting through the Bajocian and Bathonian (Hankel, 1994; Wopfner, 1994).

Kreuser (1995) has described a similar stratigraphic relationship in Tanzania where he classified the marine Ngerengere Beds as the marginal continental part of the breakup cycle. He recognised a

Bajocian unconformity at the top of the Karoo, which is followed by shallow marine limestones and oolites of the same age. He also interpreted the unconformity as signifying a breakup event. This suggests that the same pattern of tectono-stratigraphic events seen in the Morondava Basin can be observed on the conjugate margin of East Africa. Thus the rifting history of Gondwana breakup as observed in Madagascar also appears to be valid for the opposing margin in Tanzania and probably for the whole of the East African domain (Fig. 17).

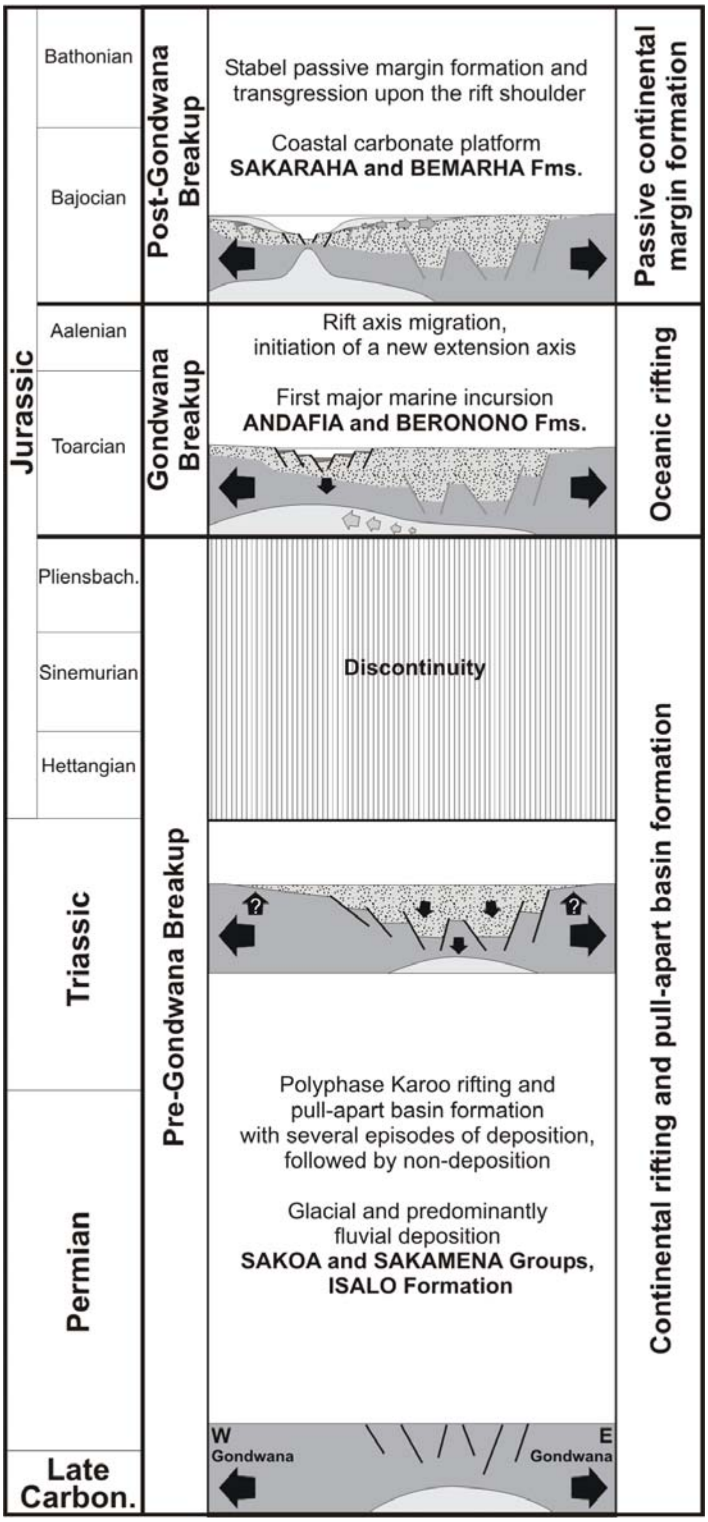


Fig. 17. Gondwana breakup was preceded by several Karoo rifting events, but only the Early Jurassic 'Andafia' rift finally resulted in crustal separation. The Permian and Triassic time intervals are not in temporal scale.

2.6 Conclusions

The confusion in the literature about the stratigraphic relationships in the Jurassic and timing of rifting in Morondava Basin can be attributed to poor outcrops in the region. The tectonic and stratigraphic relationships described herein can only be fully understood with the aid of seismic and borehole data, which were unavailable to most previous researchers.

The present study indicates that the Permo-Triassic or Karoo phase of rifting was not responsible for the separation of Madagascar from Africa. This rift system failed in the Late Triassic. The separation of Madagascar occurred in the Late Liassic (late Early Jurassic) when the "Andafia rift" was formed. This event is thought to have been relatively short-lived, with a distinct episode of rifting followed by continental separation and drift. Previous workers have regarded the rift and drift phases as a transitional process, starting in the Late Permian-early Early Jurassic and finishing in the Callovian. Here we conclude that the pre-rift phase is represented by the crystalline and metamorphic basement and the Late Carboniferous-Late Triassic sedimentary cover. The syn-rift phase is recorded by the Andafia shales of Toarcian-Aalenian age. The Early Bajocian unconformity of the Morondava Basin is interpreted as the breakup unconformity. The early post-rift or drift phase is represented by the Bajocian-Bathonian carbonates of the Bemaraha Formation and by the marls, sandstones and carbonates of the Sakaraha Formation. These two formations are considered to be lateral equivalents of one another, with the Sakaraha Formation representing a coastal plain environment and the Bemaraha Formation a coastal barrier/lagoon complex.

Acknowledgements

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Chapter 3 Toarcian-Kimmeridgian depositional cycles of the south-western Morondava Basin along the rifted continental margin of Madagascar

2005
submitted to *Facies*

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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- 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 to 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.

Keywords: Madagascar; Gondwana; post-Breakup; Jurassic; T-R cycles

3.1 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 (Fig. 18 and Fig. 19). 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).

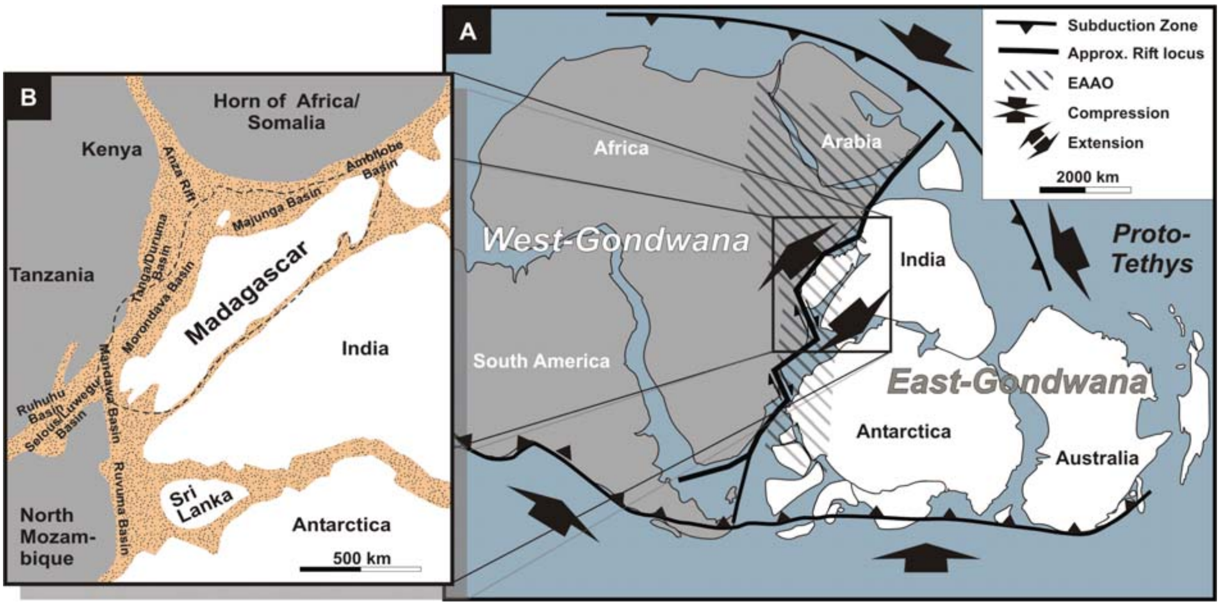

Fig. 18. **A**: Gondwana reassembly at 200 Ma with present-day continents. Initial extension, as response to compression at the northern and southern margin of the supercontinent, was localised along the former East African-Antarctic Orogen (EAAO). Compiled from Stollhofen (1999) and Jacobs et al. (1998). **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.

Recent studies of tectono-sedimentary architectures from seismic images have shown that the breakup of Gondwana took place during the late Early Jurassic (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 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, eighteen sections of Bajocian-Kimmeridgian successions in the southern Morondava Basin have been measured (Fig. 19) to study macro- and microscale sedimentary patterns and to characterise depositional environments. Macro- and microfauna was 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.

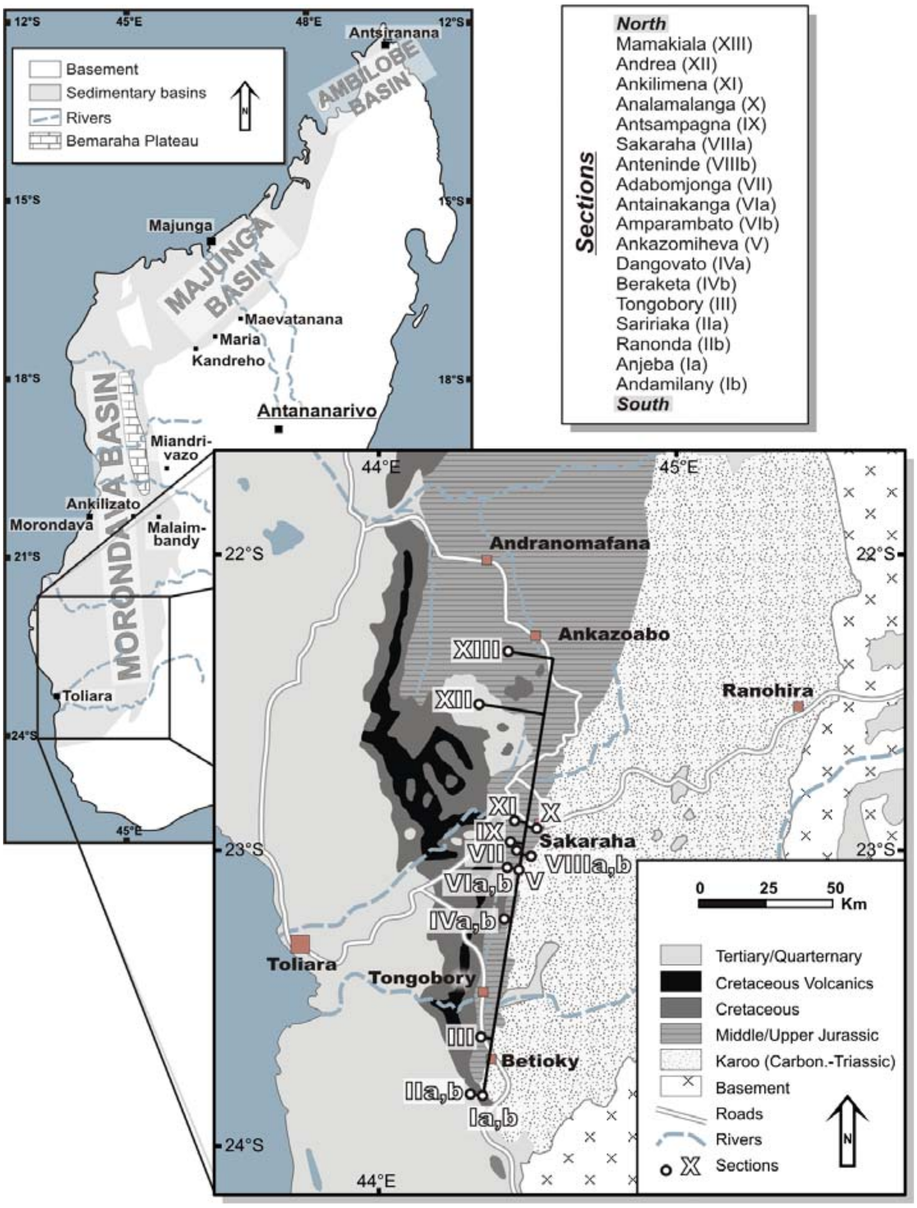

Fig. 19. Three coastal basins – the Morondava, Majunga and Ambilobe basins – extend along the western coast of Madagascar and are filled with Late Palaeozoic, Mesozoic and Cenozoic sediments. In the southern Morondava Basin eighteen sections were measured (zoomed sketch map: Persits et al., 2002).

3.2 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 (Kreuser, 1995; SACS, 1980). They comprise of the Late Carboniferous-Early Permian Sakoa Group, Middle Permian-Middle Triassic Sakamena Group, and the Late Triassic Isalo Formation (e.g. Geiger et al., 2004; Hankel, 1994; Montenat et al., 1996; Wescott and Diggens, 1997; Wescott and Diggens, 1998). 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. 19), 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 Beronono 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).

3.3 Biostratigraphy of the southern Morondava Basin

Biostratigraphic concepts of the Jurassic in Madagascar are generally hampered by faunal and floral provincialism due 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 (Cariou and Krishna, 1988; Spath, 1933) and Madagascan faunas (Besairie and Collignon, 1972; Collignon, 1953; Collignon, 1960; Collignon, 1964a; Collignon, 1964b; Collignon, 1967; Collignon et al., 1959; 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; Mette and Geiger, 2004b; Mette and Geiger, 2004c), and foraminifers (e.g. Espitalié and Sigal, 1963a; Espitalié and Sigal, 1963b; 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”.

3.3.1 Toarcian-Aalenian

A universal biostratigraphic constraint for the first marine ingression in the East African domain is the appearance of *Bouleiceras* sp. (Hallam, 2001; Luger et al., 1994). 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. 20). 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).

3.3.2 Bajocian-Bathonian

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

3.3.2.1 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. 19), 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-early Late Bajocian age. Corals also correspond to faunas from the Late Bajocian (Lathuilière et al., 2002).

3.3.2.2 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)(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 hian*s (Waagen) and *Delecticeras anjohense* (Collignon), which indicate a Middle to Late Bathonian age (Besairie and Collignon, 1972; Collignon, 1964b).

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).

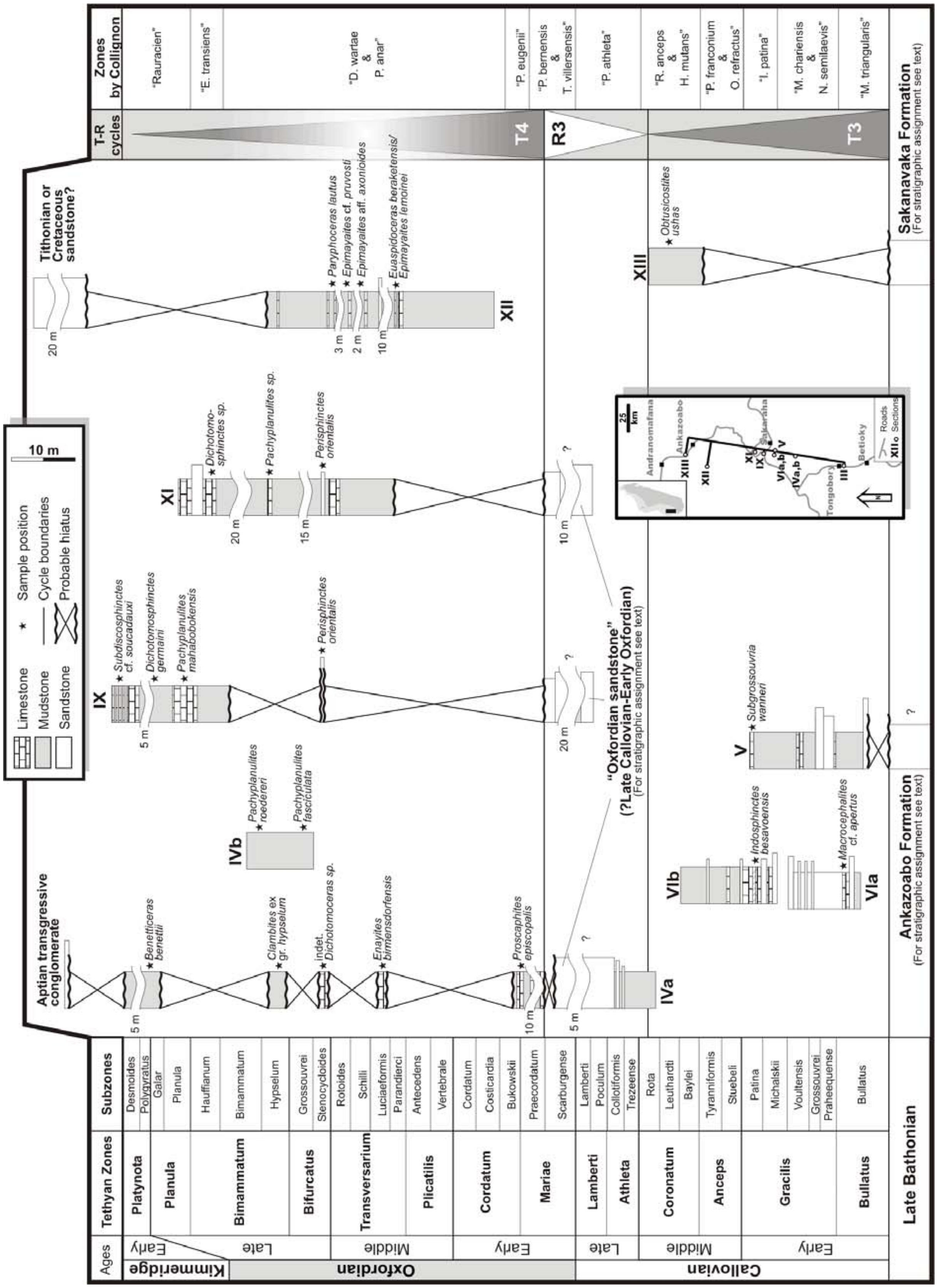

Fig. 20. Ammonite biostratigraphy of the Callovian-Early Kimmeridgian sections. Taxonomic comparison with Tethyan (Cariou and Hantzpergue, 1997), western Indian (Cariou and Krishna, 1988; Spath, 1933), and Madagascan faunas (Besairie and Collignon, 1972; Collignon, 1953; Collignon, 1960; Collignon, 1964a; Collignon, 1964b; Collignon, 1967; Collignon et al., 1959; Joly, 1976).

3.3.3 Callovian-Early Oxfordian

From the Callovian onwards ammonites are useful biostratigraphic tools (Fig. 20). 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 is 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 can not be unambiguously dated, but the occurrence of a *Lytoceratide* suggests a Callovian age.

3.3.4 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. indicates 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 & 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).

3.4 Sedimentary environments

Four major sedimentary environments were recognised in the Jurassic syn- 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 mudstones (shales and marls) with an thick intercalated ?Late Callovian-Early Oxfordian Sandstone (Fig. 32).

3.4.1 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).

3.4.2 Bajocian carbonate platform

The Bemaraha Formation of the carbonate platform system is only present at the Bemaraha Plateau (Fig. 19), 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. 22).

3.4.2.1 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.

3.4.2.2 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 to 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, 1969a).

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. 22).

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. 19). 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 hundred metres 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 Anteninde section.

Analamalanga section (X). The Bajocian section at Analamalanga (Fig. 21, S22°51.738'/E44°32.950', Fig. 19) closely resemble those at the type locality at Sakaraha (Geiger et al., 2004). A transgressive oolitic-grainstone unconformably overlies the Isalo sandstone followed by mudstones intercalating with thinner, partly oolitic and bioclastic limestones (packstones and grainstones) and sandstones (Fig. 21, Fig. 23a-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 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 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.

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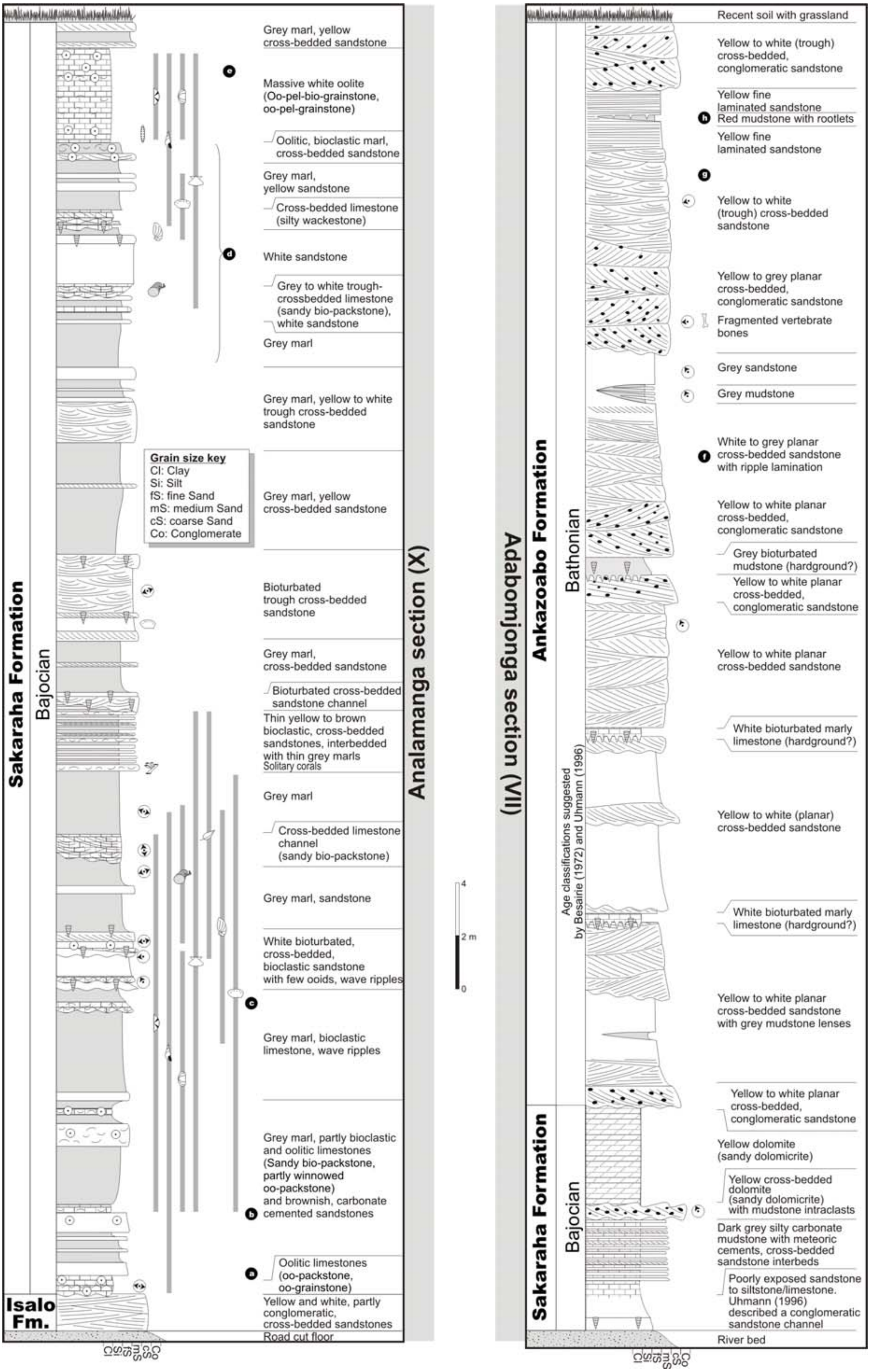
Fig. 21. Analamanga section (X) illustrates the basal Sakaraha Formation with the contact to underlying Isalo sandstone, whereas Amparambato section (VIb) reveals the sharp boundary to the overlying Ankazoabo Formation. For geographical reference see Fig. 19.



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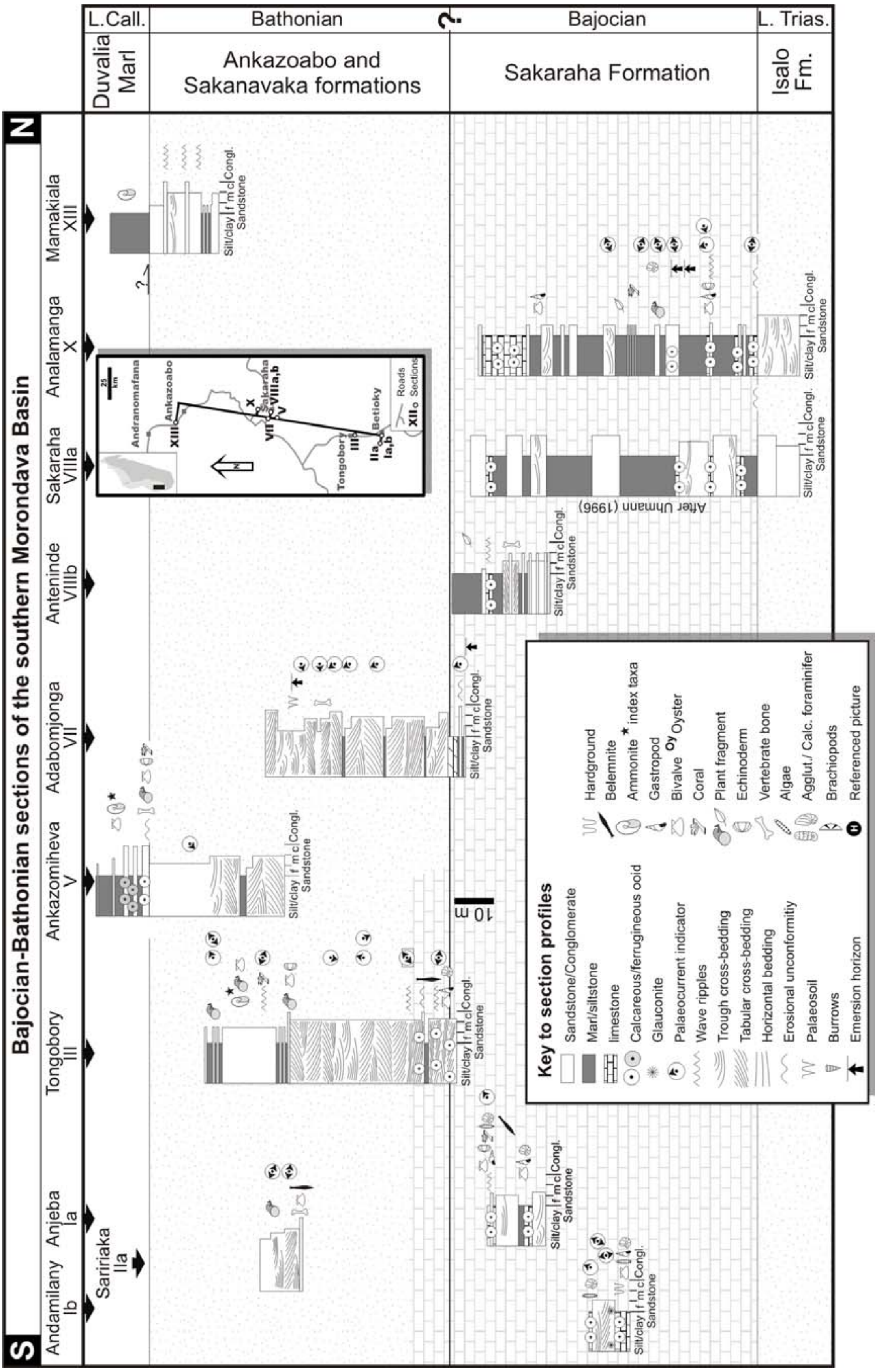

Fig. 22. Correlation scheme of measured sections of the Sakaraha, Ankazoabo, and Sakanavaka formations in the study area illustrates the stratigraphic difference of the Bajocian platform carbonates and the Bathonian sandstones. See Analamanga (X) and Adabomjonga (VII) sections for further exemplary details.

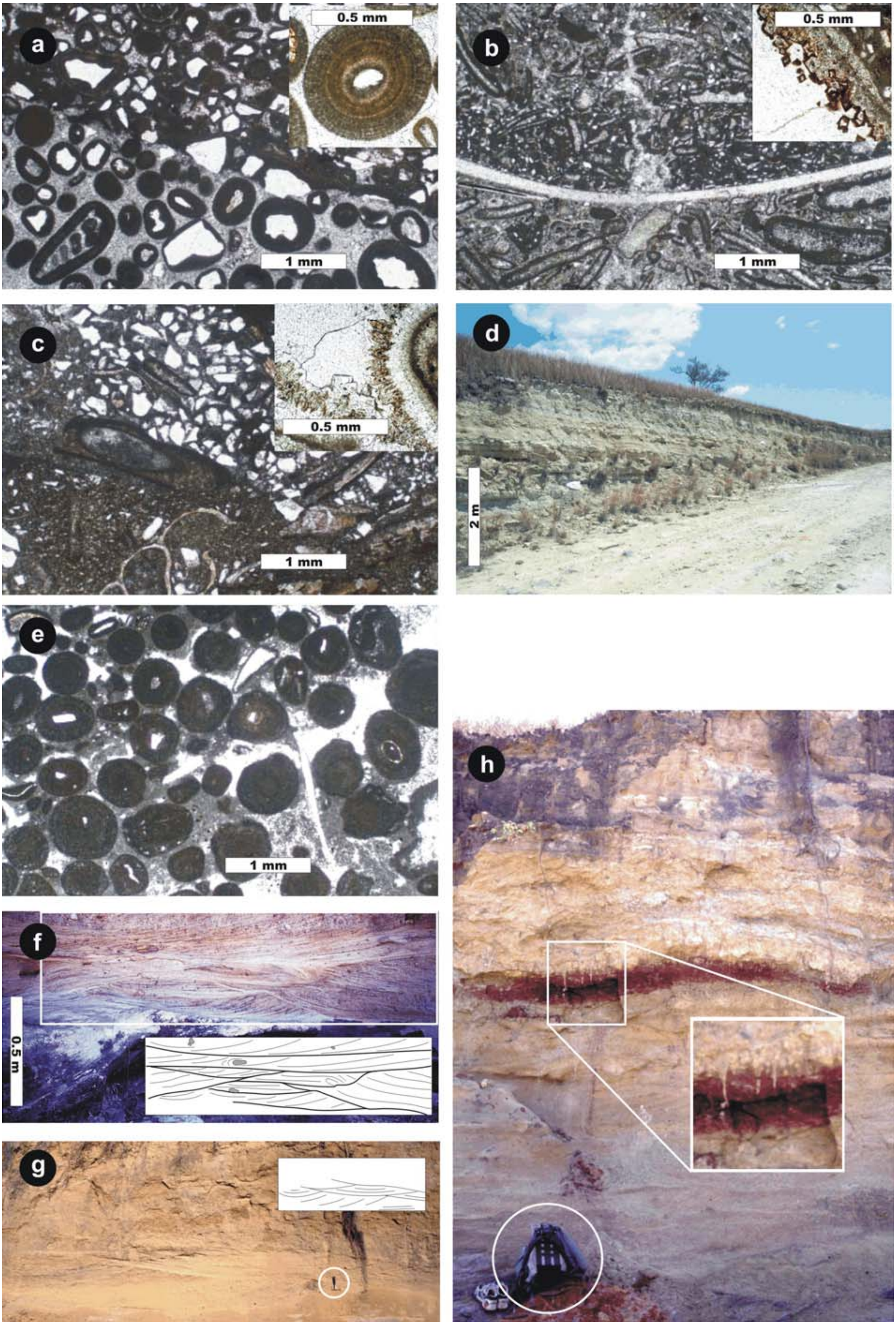
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Fig. 23. Photos and photomicrographs from the Analamanga (X) and Adabomjonga (VII) sections. **Analamanga: a**, Transgressive bioclastic oo-grainstone sometimes with quartz grains as ooid nucleus above Isalo sandstone; **b**, Dolomitized bio-grainstone; 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.

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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. 19) and were also previously described by Uhmann (1996), where Anjeba represents the upper part of the composed at Andamilany section. Geiger et al. (2004) described at 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. 24A-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.

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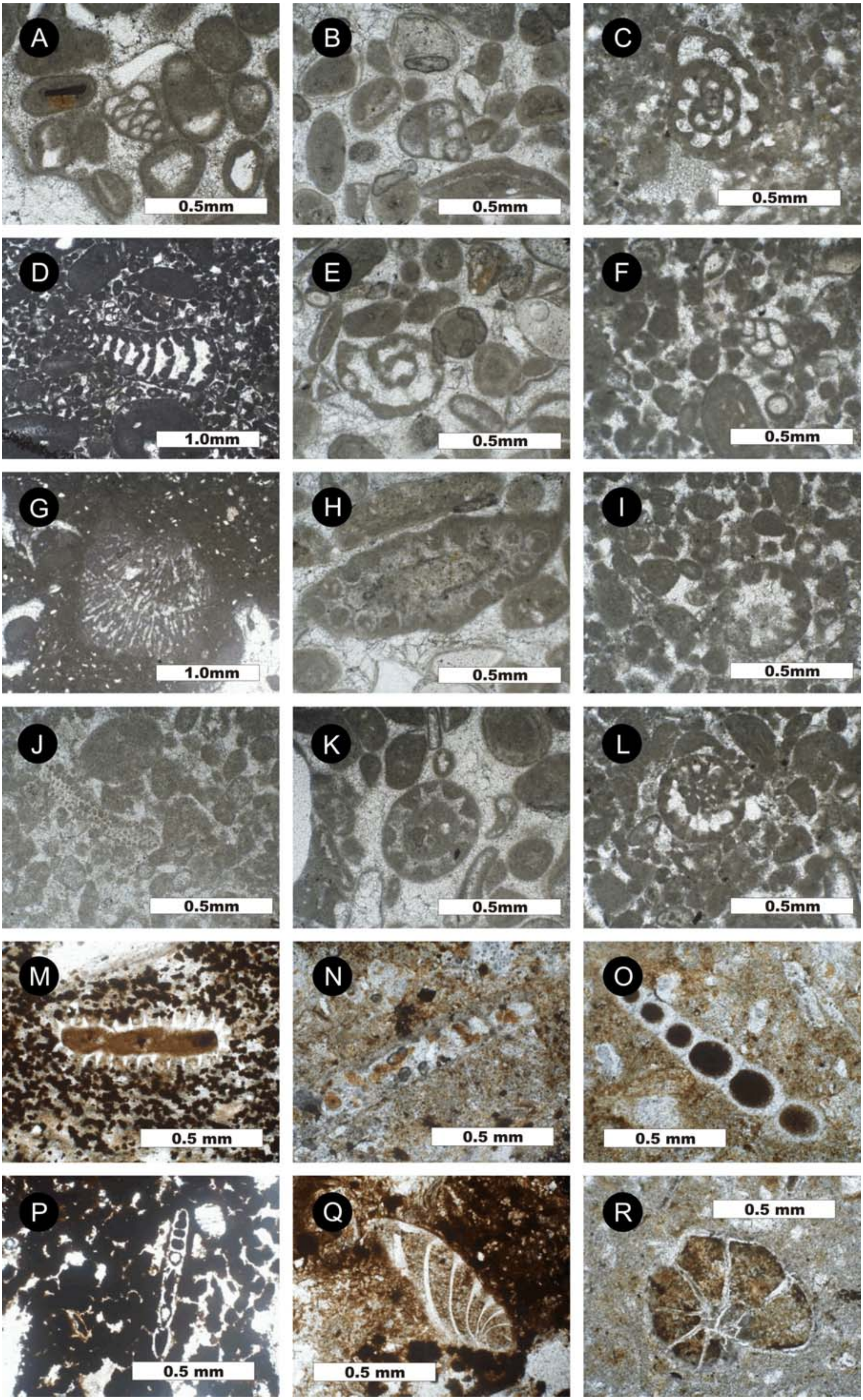
Fig. 24. Thin section microphotographs showing foraminifer and algae assemblage at Anjeba (Ia), Andamilany (Ib) and the Dangovato (IVa) sections. ***Bajocian Sakaraha Formation***: Foraminifers: **A**, *Textularia* sp., Ib; **B**, *Valvulina meentzeni* Klinger, Ia; **C**, *Mesoendothyra croatica* Gušic, 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; **P/Q**, *Citharina* sp.; **R**, indet. hyaline rotaliide.



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3.4.3 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 describes (1996) a sandstone lens with off-lapping clinoforms from seismic images that is located almost immediately above the Bemaraha carbonate platform. He attributed them to a set of Upper Jurassic sandstones.

3.4.3.1 Ankazoabo Formation

At the type locality, in the vicinity of Ankazoabo (Fig. 19), 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. 22), 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 (S22°54.901'/E44°28.618', Fig. 21, Fig. 19) 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 channels 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. 23f, 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. 23h).

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 sub- 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. 25, S23°42.265'/E44°20.395') along both shores of the Onilahy River (Fig. 19) 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. 26b). Small and large wave-length wave ripples and sand wave horizons with crest spacings of up to 40 cm are present (Fig. 26a).

Ripples crests and the alignment of wood infer E-W oriented palaeocurrents. Foresets dip predominantly to the east.

The partly oolitic shallow water sandstone at the base was recently interpreted as transition from Sakaraha Formation facies to the Ankazoabo Formation facies (Geiger et al., 2004), 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. Hummocky cross-stratification and coarse, cross-bedded sandstones are typical for subtidal lower shoreface deposits. Fining upward at the top of the section implies a landward retreating shoreline.

Saririaka (IIa). Saririaka section (S23°42.125'/E044°18.682', Fig. 22, Fig. 19) 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.

3.4.3.2 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 (S22°06.924'/E44°26.487', Fig. 19) 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.

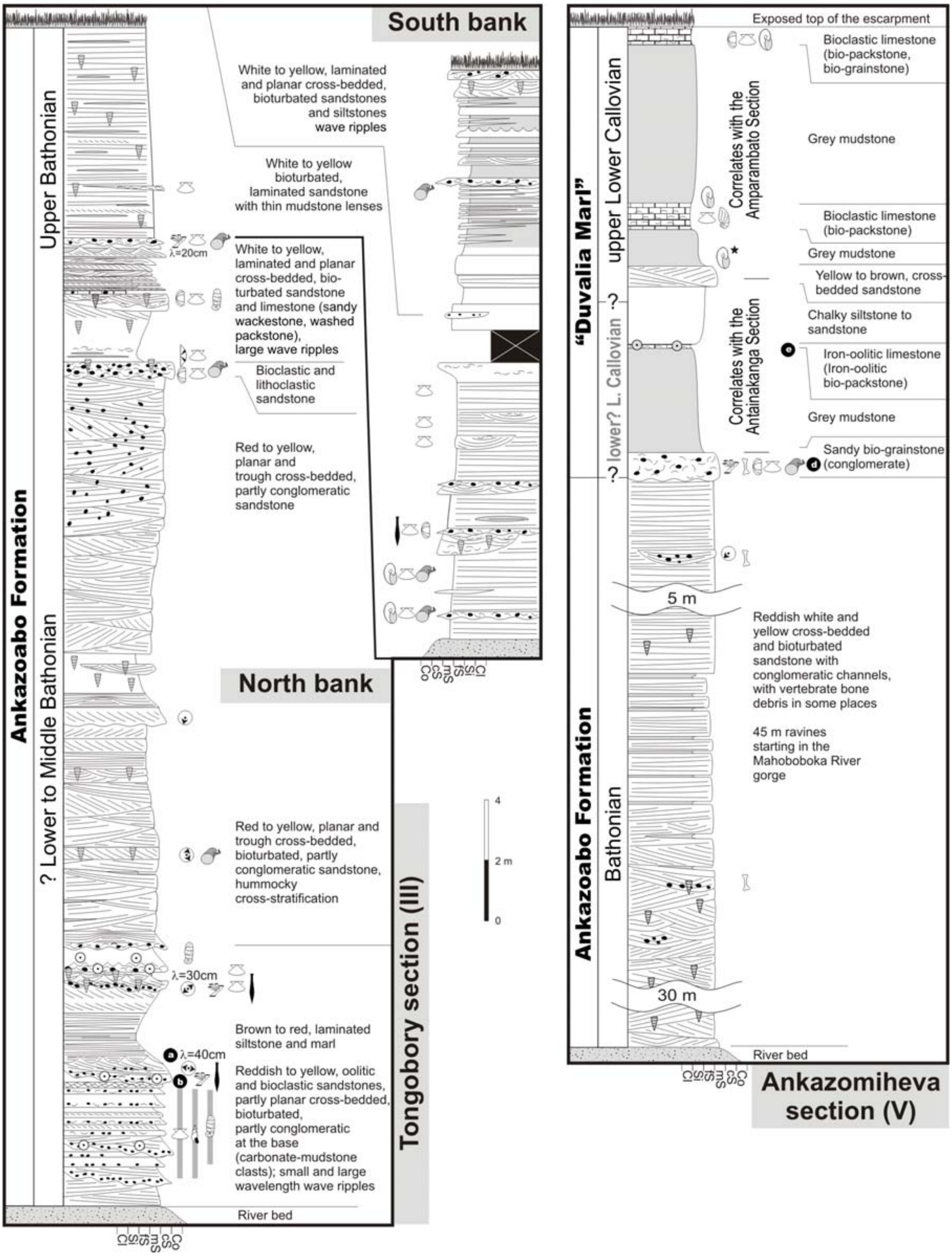

Fig. 25. Tongobory section (III) shows a typical siliciclastic lower shoreface setting with shallow marine sandstones of the Ankazoabo Formation. The section is composed of outcrops from the north and the south shore of the Onilahy River. Ankazomiheva (V) section illustrates the boundary to the overlying Callovian mudstones of the basal Duvalia Marl. For geographical reference see Fig. 19.

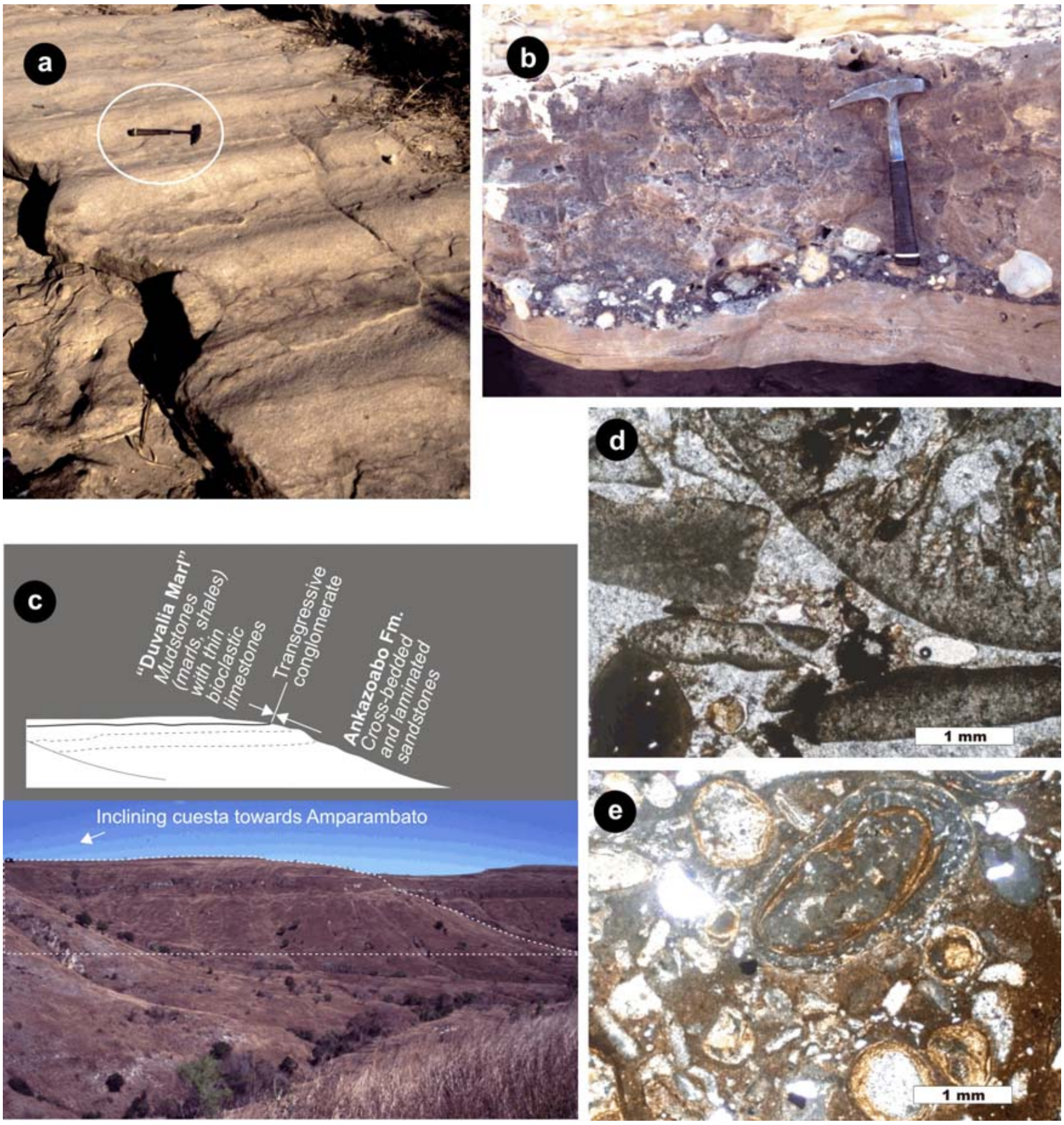

Fig. 26. Photos and photomicrographs from Tongobory (III) and Ankazomiheva (V) sections. **Tongobory: a**, Conglomeratic sandstone with predominantly carbonate mudstones lithoclasts overlies an erosional surface, see hammer for scale; **b**, Sand waves with crest spacing exceeding 40 cm, see hammer for scale. **Ankazomiheva: a**, Sandy iron-oolitic limestone (bio-packstone) with algal oncoid; **b**, Coarse bioclastic debris in a litho-bio-grainstone (bioclastic conglomerate); **c**, Landscape view facing westwards on the Ankazomiheva cuesta. The scarps and ravines provide a section along the slope.

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.

3.4.4 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 Geiger et al., 2004; Stoakes and Ramanampisoa, 1988), 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.

3.4.4.1 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" Besairie and Collignon, 1972).

Ankazomiheva section (V). The Ankazomiheva section (Fig. 25, Fig. 19, 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 (Fig. 19, Fig. 27, 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 toped 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 (Chapter 4.10).

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.

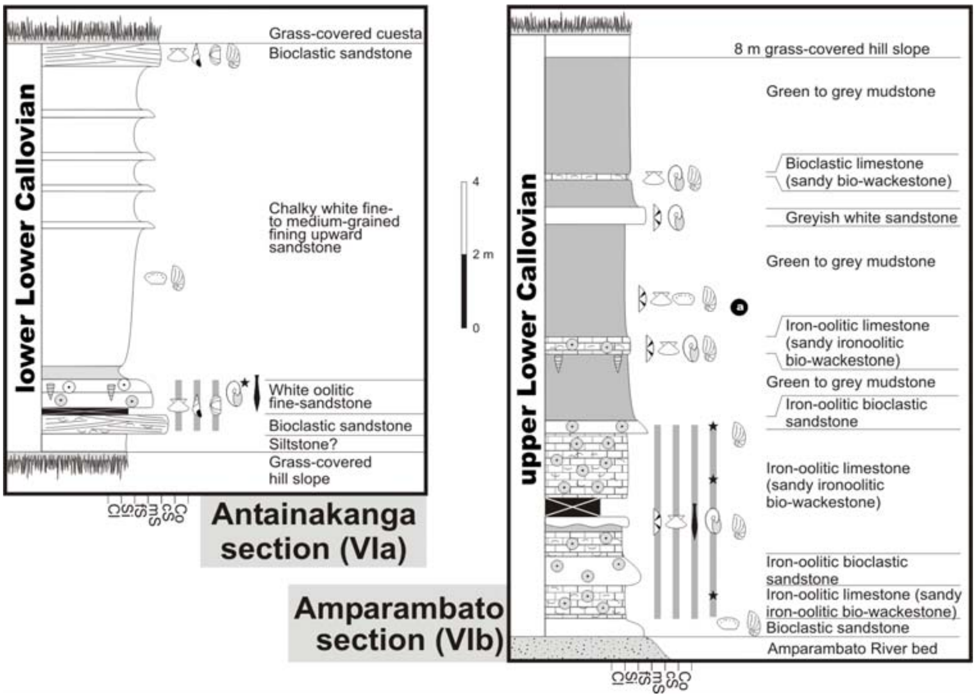

Fig. 27. Lower Callovian Antainakanga (VIa) and Amparambato (VIb) sections illustrate the facies immediately above the Bathonian sandstone. Compare Ankazomiheva section (V). For legend see Fig. 22.

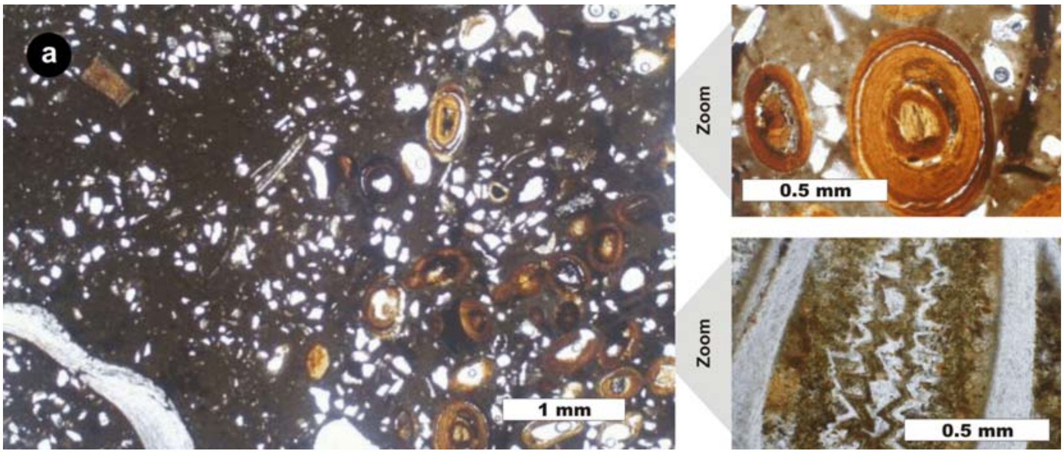

Fig. 28. Silty iron-oolitic bio-wackestone with concentric iron-ooids and bladed calcite rhombus (indicates fresh water mixing). Stratigraphic position as indicated in Amparambato section, Fig. 27.

Amparambato section (VIb). At the river cut of Amparambato section (Fig. 19, Fig. 27, 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. 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 (Chapter 4.10).

Ranonda section (IIb). The lithology of the Callovian Ranonda section (Fig. 19, 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. 19, 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 highly fragile. In the basal section, a few horizons reveal a highly diverse foraminifer fauna dominated by textulariid and nodosariid forms (Chapter 4.10). 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 4.10).

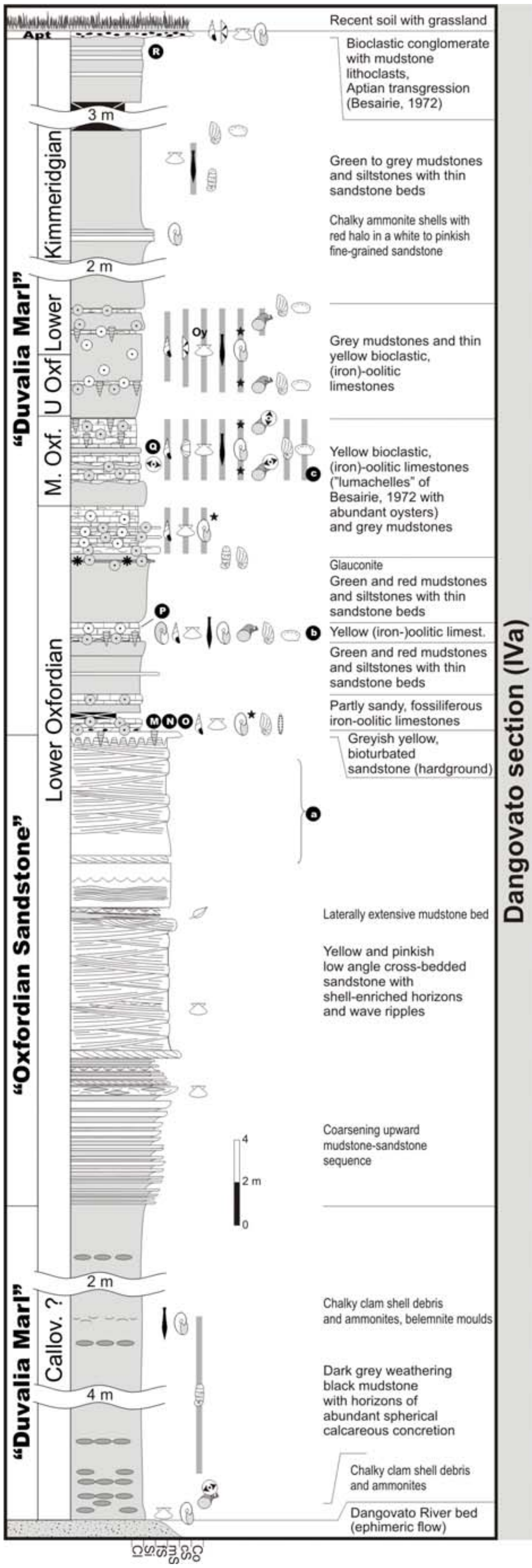

Fig. 29. Dangovato section (IVa) is the stratigraphically most extensive section covering the Lower Oxfordian-Kimmeridgian interval. The top is cut of transgressively by the Aptian conglomerate (Besairie and Collignon, 1972). It is the only section to contain the complete succession of the Oxfordian Sandstone. Small letters in black circles refer to photographs in Fig. 30. Capital letters refer to photographs in Fig. 24.

Dangovato section (IVa). The ?Callovian-Early Kimmeridgian Dangovato section (Fig. 19, Fig. 29, S23°06.895'/ E44°23.254') along the Dangovato River, a tributary of the Sakondry River, starts at the base with shaly mudstone which containing 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. 30a). At the top of the sandstone exists a well developed hardground which is overlain by limestone (bio-oo-wackestone). Above succeed interbedded mudstones and limestones. The limestones are often (iron-)oolitic oysters coquinas (Fig. 30b, 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, 2004; Mette and Geiger, 2004b). Shaly mudstones below the Oxfordian Sandstone contain a wide spectrum of agglutinating foraminifers (Chapter 4.10). 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 (Chapter 4.10). 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. 19, 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 (Chapter 4.10). *Epistomina* sp. was identified in the 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).

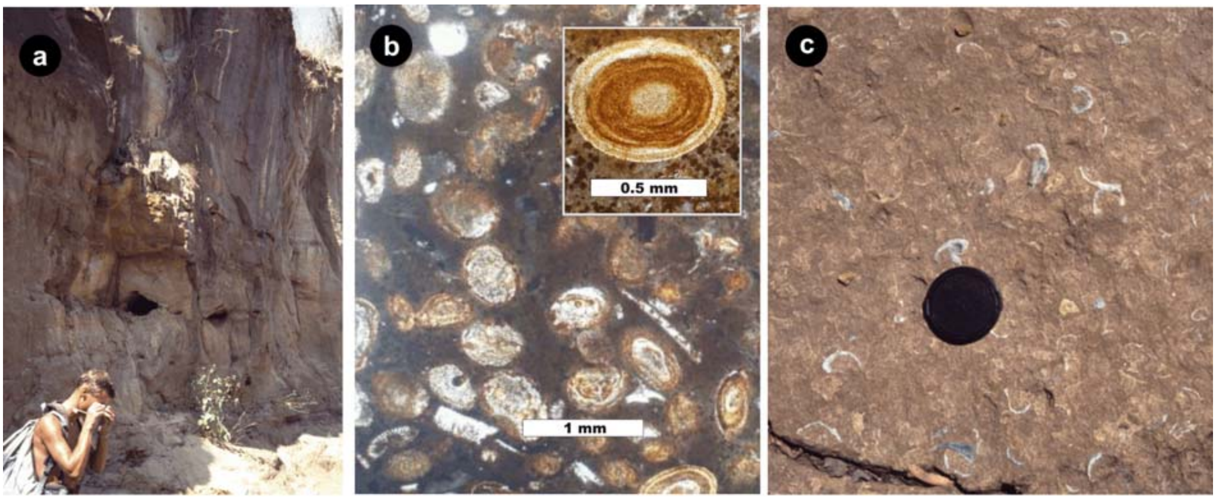

Fig. 30. **a**, Dangovato river gorge exposes the Oxfordian Sandstone. **b**, Thin section microphotograph of an iron-oolitic coquina in the Lower Oxfordian; The inset illustrates the concentric character of the ooids. **c**, Top surface of an iron-oolitic coquina; the white shells are predominantly oysters; see lens cap for scale (7 cm diameter).

Ankilimena section (XI). At the Middle-Upper Oxfordian Ankilimena section (Fig. 19, 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 (Chapter 4.10), 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. 19, 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 (Chapter 4.10) 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.

3.4.4.2 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.

3.4.5 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 (Edmond et al., 1982; Hekinian et al., 1980). 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, 1988b; Kidwell, 1989; Loutit et al., 1988; 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., 2004b).

3.5 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 requires 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 generalisation necessarily means that the applicability of sedimentary patterns for interpreting global sea-level changes has to be reduced to second order cycles but 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. 31). 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).

3.5.1 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 wide-spread 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).

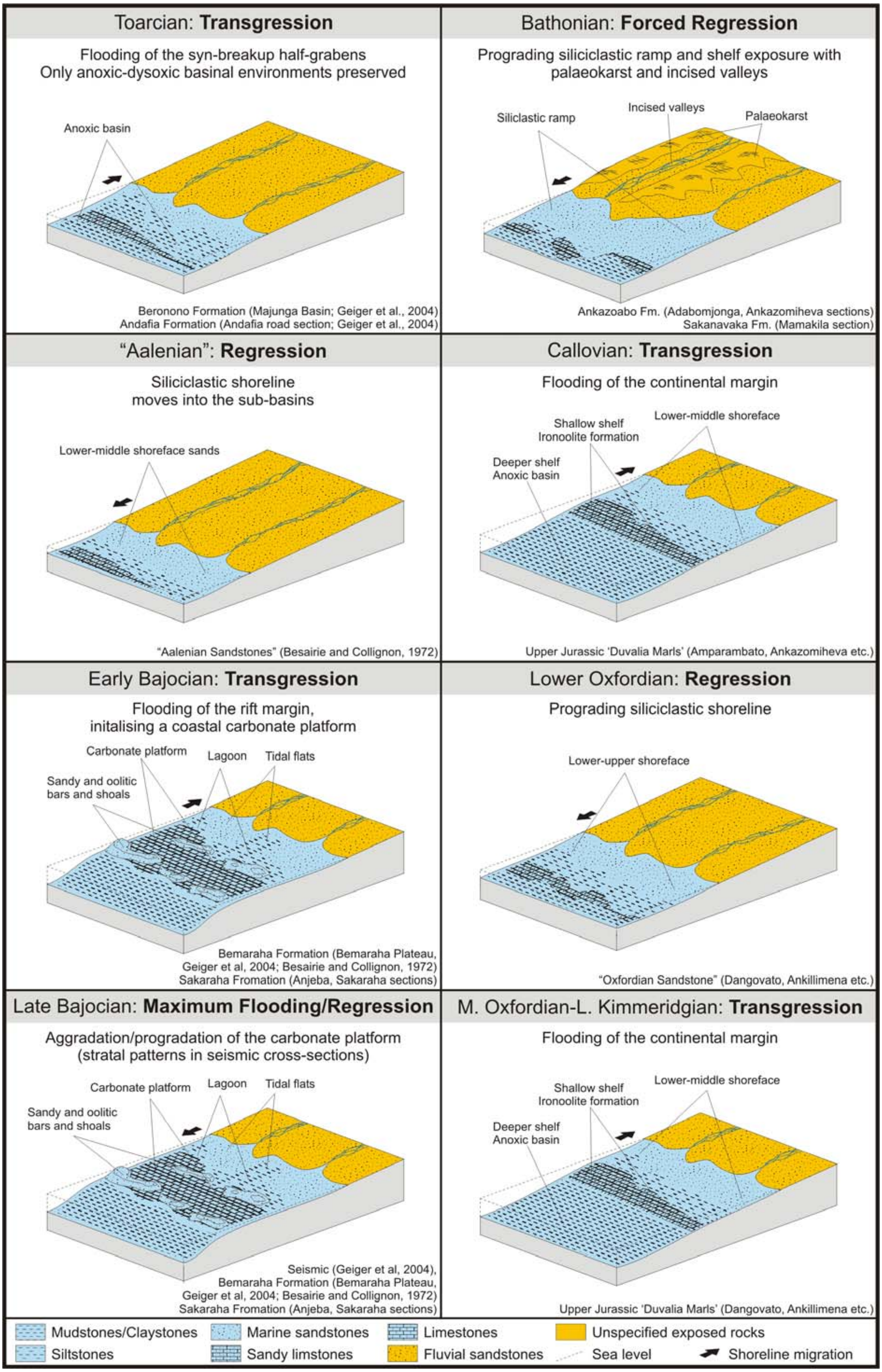
Toarcian: Transgression
Flooding of the syn-breakup half-grabens
Only anoxic-dysoxic basinal environments preserved
Anoxic basin
Beronono Formation (Majunga Basin; Geiger et al., 2004)
Andafia Formation (Andafia road section; Geiger et al., 2004)
Bathonian: Forced Regression
Prograding siliciclastic 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

Fig. 31. Model of the sedimentary environments during the syn- and post-breakup T-R cycles of the Morondava Basin. Data basis was extended as indicated.

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Shallowing and regression in the Aalenian is recorded by the overlying Aalenian Sandstone which is 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. 32) and infer the coincidental syn-breakup rifting. Such a syn-breakup 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).

3.5.2 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 (Gradstein et al., 1991; Westermann and Callomon, 1988). 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.

3.5.3 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.

3.5.4 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. 32). Nevertheless, Tongobory section (III) shows deepening at the top of the Middle-Late Bathonian sandstone succession prior to the major transgression. *Epistomina* sp. (Chapter 4.10) in mudstones at Antsampagna (IX) are typical for sudden flooding (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 of 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-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.

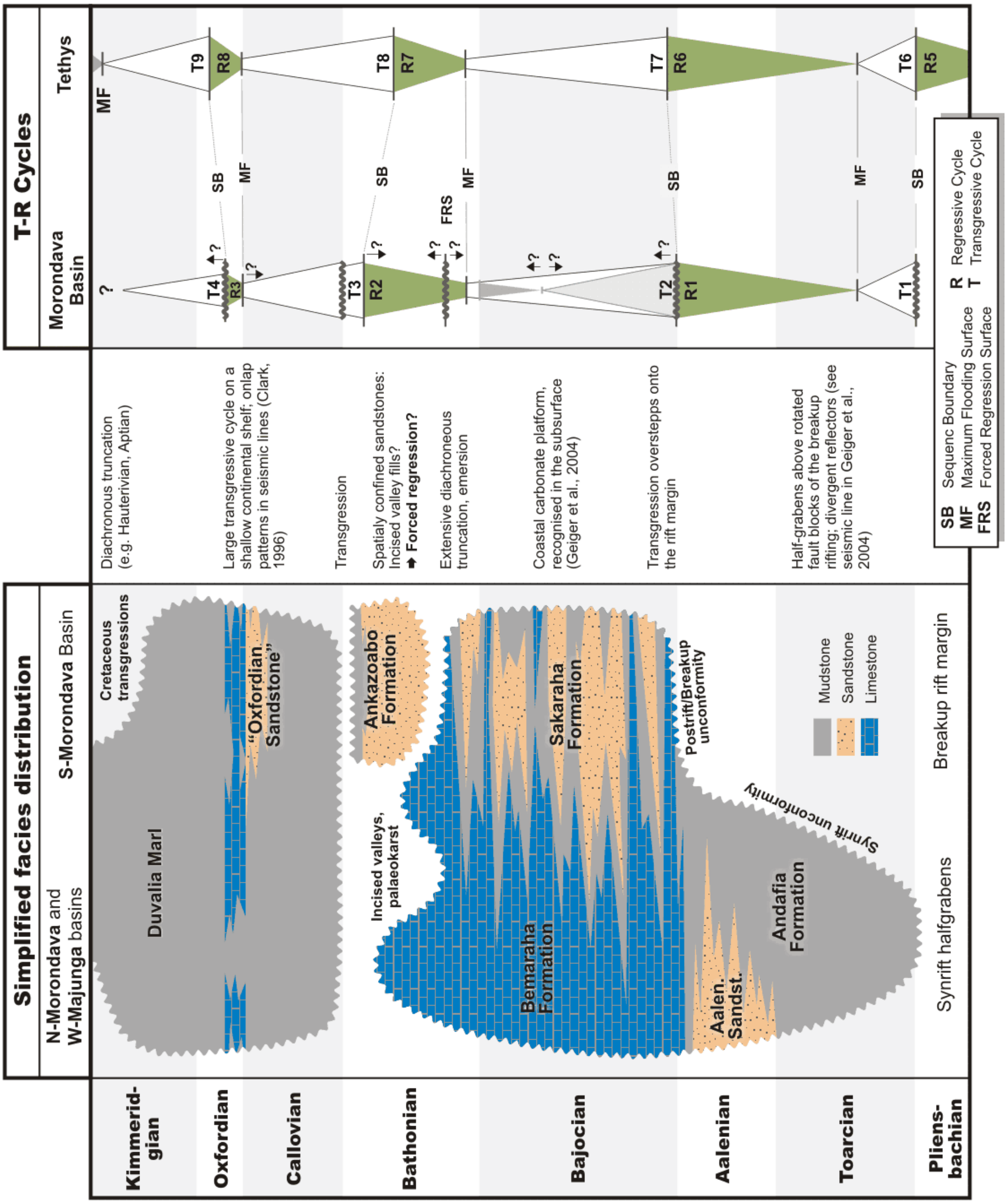

Fig. 32. Composed lithostratigraphy and interpreted succession of local T-R cycles compared to Tethyan T-R Tethyan T-R cycles (after Jacquin et al., 1998). Facies distribution is mainly based on the present work with supplementary data from Besairie and Collignon (1972), Clark (1996) and Montenat et al. (1996).

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 (Europe: Jacquin et al., 1998; Greenland: Surlyk, 1991). 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. 32). 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), India (Gaetani and Garzanti, 1991) are in correspondence with the study area and appears 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).

3.5.5 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 suggests 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).

3.5.6 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 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, such 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 bio- 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 (Hallam, 2001; Jacquin et al., 1998).

3.6 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 sea-level. 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., 2004b) 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 can not be determined by thermochronological methods (Emmel et al., 2004b).

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 on 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.

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Chapter 4 The Morondava Basin: interdisciplinary studies on the eastern margin of the Proto-Indian Ocean

This Chapter presents additional basin analytic perspectives, carried out in cooperation with partners from the joint research project "Rifting Processes and Basin Development: The Early Stages of the Indian Ocean". Apart from summarized presentations of the joint publications, this chapter includes a collection of further investigations and considerations of western Madagascar's shelf basins in connection with the Somali (Ocean) Basin. These basins descended from the Proto-Indian Ocean that formed during the breakup and drift of East- and West-Gondwana. The joint publications are listed below:

Detrital apatite fission-track ages in Middle Jurassic strata at the rifted margin of W-Madagascar - indicator for a protracted resedimentation history

Sedimentary Geology, submitted. (Chapter 4.9.2 and 4.9.3)

Benjamin Emmel [a], Markus Geiger [a], and Joachim Jacobs [a]

Bajocian and Bathonian ostracods and depositional environments in Madagascar (Morondava Basin and southern Majunga Basin)

Beringeria, 34: 37-56. (Chapter 4.5)

Wolfgang Mette [b] and Markus Geiger [a]

Taxonomy and palaeoenvironments of Callovian ostracoda from the Morondava Basin (south-west Madagascar)

Beringeria, 34: 57-87. (Chapter 4.5)

Wolfgang Mette [b] and Markus Geiger [a]

Middle Oxfordian to early Kimmeridgian ostracoda and depositional environments of south-west Madagascar

Beringeria, 34: 89-115. (Chapter 4.5)

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4.1 Introduction

The aim of the joint research project “Rifting Processes and Basin Development: The Early Stages of the Indian Ocean” is to study the processes that led to the formation of the initial marine basins in the Indian Ocean region. To achieve this goal, geochronological, structural, and thermochronological studies were performed on basement units of Madagascar and structural, seismic, sedimentological, palaeontological, and stratigraphical studies of sedimentary units of the Morondava Basin at the western margin of the island. Geochronological and thermochronological studies were carried out by B. Emmel and J. Jacobs (both University of Bremen), structural studies of the pre-breakup strata by F. Bremer and H. Schandelmeier (both TU Berlin), and structural, sedimentological, and palaeontological studies by M. Geiger (Bremen) and W. Mette (Innsbruck, Austria). These cooperations resulted in four joint publications: 1) sedimentological and stratigraphical studies were combined with thermochronological results of detrital apatite; 2-4) sedimentological and palaeoenvironmental studies were integrated into palaeoecological and palaeogeographical interpretations on the basis of micro- and macrofaunal assemblages.

Chapters 2 and 3 have presented the structural and depositional development of the syn-breakup and post-breakup phases. Prior to the breakup, extensive rifting and extension occurred for already 100 Ma (Late Carboniferous?/Early Permian-Late Triassic) and formed accommodation space for a thick succession, the Karoo Supergroup. During this period, various stress and kinematic regimes (Chapter 4.3) formed pull-apart basins, transitional basins, and half-grabens in a generally N-S striking area, parallel to the East-African-Antarctic Orogen (Schandelmeier et al., 2004). After a period of tectonic quiescence during the early Early Jurassic the final breakup occurred and in a narrow, localised area within the wide Karoo-aged rift basin. Chapter 4.4 discusses the underlying processes and mechanisms which led to the breakup rifting. This was the initiation of the opening of the Proto-Indian Ocean which is represented by the Somali and Mozambique basins today. Biogeographical studies infer a rapid separation of the depositional environments (Chapter 4.5). Chapter 4.6 presents alternative indicators for the breakup event on the basis of the sea floor geochronology and datings of the rift-related volcanism from the Mozambique Basin. During the drifting phase, passive margin development was influenced by the continent-continent transfer fault of the future Davie Ridge Fracture Zone (Chapter 4.7, Fig. 33). The total of structural processes affected the structural style of the basins on both sides of the Proto-Indian Ocean (Chapters 4.8 and 4.9). With regards to Madagascar’s energy shortage and the increasing demand for hydrocarbon fuel, Chapter 4.10 concerns hydrocarbon consequences that arise from the new sedimentary, stratigraphic, and structural models, presented in the previous chapters.

4.2 Reconstructions of the Proto-Indian Ocean

The southern Tethys ingressed across Arabia and the eastern part of North Africa southward during the Jurassic and formed a gulf-like embayment that widened with progressive extension. Sea floor spreading started during the Callovian and formed the Somali Basins between East- and West-Gondwana (Fig. 33). A similar continental extension south of the Davie Ridge Transform Zone (DRFZ, compare Chapter 1) formed the Mozambique Basin at about the same time. This was the birth of the Proto-Indian Ocean. Later during the Cretaceous and Tertiary, when East-Gondwana started to separate into the modern continents, the present-day Indian Ocean formed (e.g. Norton and Sclater, 1979).

Reeves et al. (2002) presented a convincing fit of Madagascar with the opposing East African coast. Retracing the transforms of the Indian Ocean and reconnecting shear zones, de Wit (2003) reconstructed the migration pathway of the crustal fragments after the segmentation of Gondwana (Fig. 33). Such reconstructions are mainly derived from retracing ocean floor ages (Müller et al., 1997) and transforms (Coffin and Rabinowitz, 1987), as well as from refitting basement structures (Reeves and de Wit, 2000). The 140 Ma snapshot in Fig. 33 depicts two basins, the Somali and Mozambique basins that simultaneously formed north and south of the DRFZ, separated by the continent-continent transform of the future DRFZ. This continent-continent transform has not been sufficiently investigated yet to give indications for the breakup history.

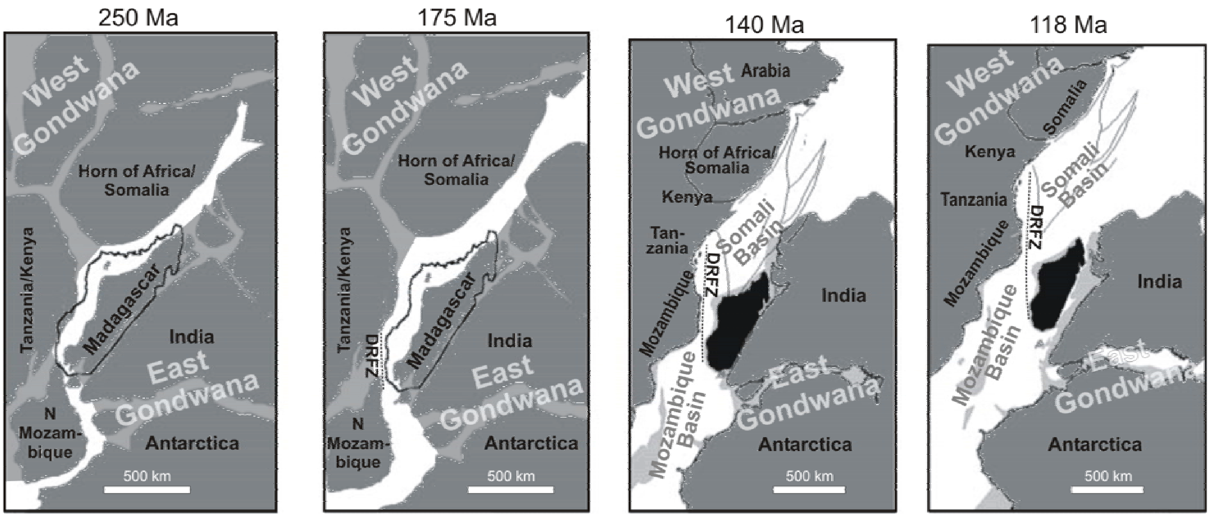

Fig. 33. Palaeogeographical reconstructions of the Proto-Indian Ocean show the palaeoposition of Madagascar at the Permian/Triassic boundary (250 Ma), in the Middle Jurassic (175 Ma), in the Neocomian (140 Ma), and in the late Early Cretaceous. While the Proto-Indian Ocean started opening at the Somali Basin north of Madagascar and the Mozambique Basin in the south of it, the southern tip of the island was connected to the African mainland across the Davie Ridge Transform Zone (DRFZ, compare Chapter 1) for the initial phase of drifting. Pale grey: shallow continental margins and oceanic plateau. White and medium grey zones represent zones of significant strain and limited continental rifting during late Palaeozoic and Early Mesozoic times. After de Wit (2003).

4.3 Gondwana’s stress regime and Madagascar’s mode of rifting

During the persistence of Gondwana from its Late Proterozoic assembly (Stern, 1994) until the Late Palaeozoic when it became part of Pangea (Veevers, 1988), a consistent stress regime acted on the supercontinent that promoted crustal accretion (e.g. Visser and Praekelt, 1996). Throughout the Karoo times (Late Carboniferous-Late Triassic), an overall compressive regime transmitted N-S to NW-SE oriented intraplate compression into the African segment of West-Gondwana (Daly et al., 1989; Visser and Praekelt, 1996)(Fig. 18A). This compression was the response to the orogenies at the Palaeo-Pacific margin in the southwest and the Hercynian and Early Cimmerian plate boundaries at the northern margin of Gondwana, including the India-Tibet region (Milanovsky and Milanovsky, 1999; Ziegler, 1992). Milanovsky and Milanovsky (1999) believe that the Gondwana breakup and the southward migration of East-Gondwana relative to East Laurasia were induced by expansion of the Mediterranean mobile belt (Neo-Tethys) between Laurasia and Gondwana during the Jurassic. Driving forces were the opening of the Tsangpo-Indus and Central Tibet spreading zones and the movements on the north-south-trending West Indian right-lateral pull-apart system (Milanovsky and Milanovsky, 1999).

In contrast to the model of Milanovsky and Milanovsky (1999), which describes an overall passive rifting mode as response to far-field stress, another more widely-accepted model suggests that in the Early Jurassic the Karoo mantle plume below south-east Africa actively governed the breakup process (e.g. Duncan et al., 1997; Stollhofen, 1999; Storey and Vaughan, 2001). Mantle plumes with increased mantle temperatures cause uplift of the mantle and crust, which commonly result in extensional tectonics and finally in the divergence of the crust (e.g. White and McKenzie, 1989). Quantitative modelling can describe the two possible modes of intracontinental rifting (Huismans et al., 2001):

(1) passive extension of the lithosphere due to far-field stress and (2) active rifting due to asthenospheric upwelling. In many rifts a transition from passive extension to active extension, and eventually back to the passive mode is being observed (Huismans et al., 2001).

Probably both modes succeeded in the Madagascan segment of the Gondwana breakup. Far-field stress initialised the crustal extension in a passive mode to form the Karoo rifting phase (Fig. 37). Later during the Jurassic, the Karoo mantle plume forced active rifting at the Mozambique Basin (Storey, 1995)(4.4.6). The lithosphere transferred extensional stress northwards to the East African-Madagascan domain and still passively opened the Somali Basin.

4.4 Rift styles: testing models

The style of rifting, the shape of the forming basins, and the architecture of their fill are strongly controlled by the configuration and temperature and of the lithosphere and asthenosphere (e.g. Huismans et al., 2001; van Balen et al., 1995). Evaluating the features of the Proto-Indian Ocean margin in the East African-Madagascar domain and comparing it to the conjugate margins of Mozambique-Antarctica as well as to the South Atlantic Rift deliver similarities and discrepancies which both characterise the rifting history. Lithospheric and asthenospheric effects, as postulated by numerical models, may significantly deviate from predictions of basin evolution inferred from conventional back-stripping models (Burov and Cloetingh, 1997).

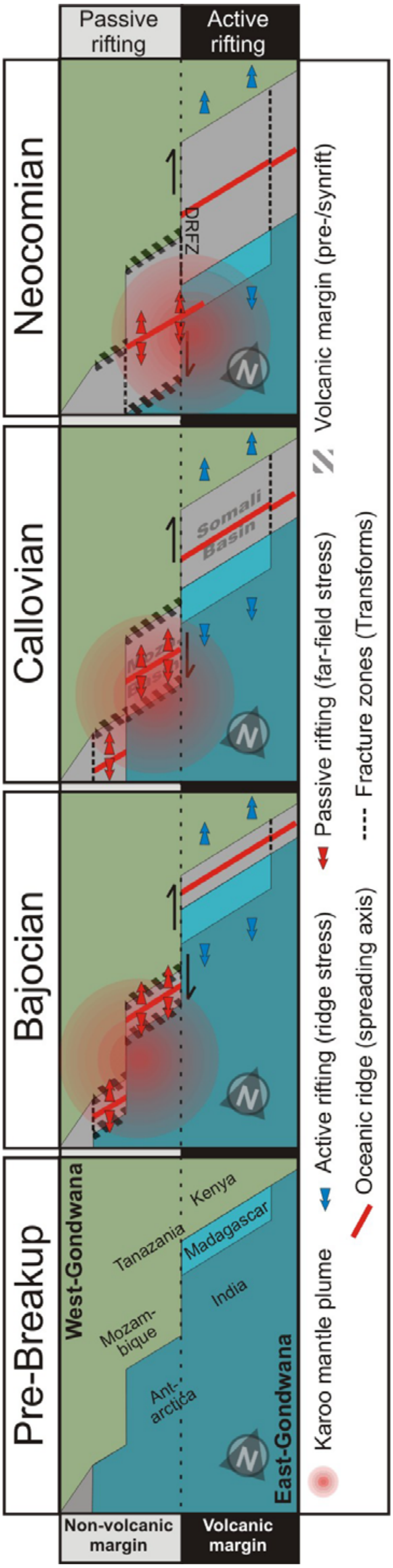

Fig. 34. Schematic snapshots of stages of the Gondwana dispersal. In the pre-breakup configuration Madagascar was in north-eastern centre of the Gondwana. After the Toarcian breakup oceanic spreading probably started in the Bajocian forced by the Karoo mantle plume. In the Somali Basin north-west of the Davie Ridge Fracture Zone (DRFZ) non-volcanic margins formed with passive rifting due to far-field stress. In contrast, plume-related magmatism covered the margins of Mozambique and Antarctica south-east of the DRFZ. During the Callovian crustal decoupling of East- and West-Gondwana and a global sea-level rise initiated a wide-spread transgression onto the continental margins. During the Neocomian the further destruction of Gondwana lasted.

4.4.1 Rifting models

Rifting and crustal extension can be described by different models which differ in the way of rigid and ductile deformation. Two main models have found wide acceptance: (1) the simple shear model of Wernicke and Burchfiel (1982) suggests that extension is accomplished only by brittle deformation along a large-scale, gently dipping shear zone, which traverses the entire lithosphere down to the asthenosphere (Fig. 35A); (2) the pure shear model of McKenzie (1978) with uniform and instantaneous extension above a thermal anomaly assumes the upper crust to fail by brittle fracture, while extension in the lower crust produces ductile deformation and vertical thinning (Fig. 35B). Heterogeneous models like the one of Coward (1986) combine mechanical and thermal subsidence at the rift location (Fig. 35C). In contrast to younger and still active rifts, such as the Red Sea Rift and the Rhine Graben Rift, the Proto-Indian Ocean rift was succeeded by several tectonic and geodynamic events. Structural overprint was produced by the Cretaceous Indian-Ocean rifting during the dispersal of East-Gondwana. The probably coherent Marion Hotspot (87 Ma, Storey et al., 1995) also superimposed thermochronological records. The same was true for Oligocene-Pliocene volcanism at 28-3 Ma (Piqué, 1999). Nevertheless the structure of the rift basin and rift shoulder yields some suggestions of large-scale tectonic and geodynamic relevance (Chapter 4.4.2).

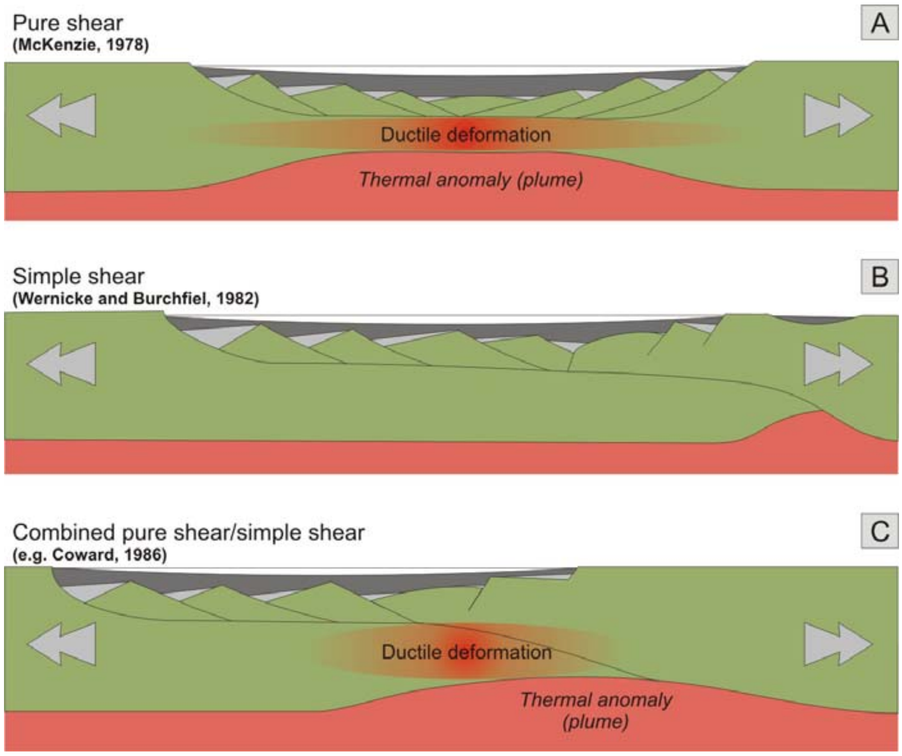

Fig. 35. Tectono-sedimentary models of rift basins: A, pure shear model after McKenzie (1978); B, simple shear model after Wernicke and Burchfiel (1982); D, combined models, e.g. Coward (1986).

4.4.2 Rift shoulder uplift

Rift shoulder uplift is a common feature observed at margins of recent (e.g. Red Sea: Steckler and Omar, 1994) and fossil (e.g. SW Africa: Rust and Summerfield, 1990) rifts, but they have a high variety in the dimension. The promotion of rift shoulder uplift and the formation of the typical rift bounding escarpments are controlled by one or a number of interplaying factors: lateral heat transfer, dynamic effects (spatial strain-rate distribution), flexural forces, small-scale convection or underplating (e.g. Chéry et al., 1992; Cloethingh et al., 1982). Burov and Cloetingh (1997) outline that it is essential to involve surface processes, which modify topography and sedimentary infill, into the considerations and models of rift flank evolution. This is necessary, since geomorphological and sedimentological data suggest coupled surface and tectonic processes.

An applicable method to estimate rift shoulder uplift is the fission track analysis. Fission track analysis measures denudation rates, which can generally be used to interpret vertical uplift (Gallagher et al., 1998; van der Beek et al., 1994). Apatite fission track analyses from the Madagascan basement complex, which borders in the east-central island, date major denudation activity prior to the breakup 180 Ma (Emmel et al., 2004b). This argues for vertical uplift of less than 2000 m and leaves a wide scope for speculation. Only the process of resedimentation of basinal strata, as deduced from detrital apatite fission track analyses (Chapter 4.9.2) suggests minor uplift during the Middle Jurassic. A considerable uplift of the margin would have been an insurmountable obstacle for the rising sea, and the transgression would have not reached as far landward as inferred by seismic and sedimentological studies.

4.4.3 Asymmetry of oceanic rifts and rift shoulder uplift

Oceanic rifts and their spreading ridges show a distinct asymmetry. In a global comparison of ocean floor and continent topographies, Doglioni et al. (2003) discovered that the eastern flank of the ocean ridges and the eastern rift shoulders are generally more elevated than their conjugated counterparts in the west. Depleted, lighter asthenosphere is produced below oceanic ridges and shifts “eastward” relative to the lithosphere, i.e. the asthenosphere shifts relatively “westward”. This results in a density deficit below the eastern flank (Doglioni et al., 2003). The eastward migration of lighter asthenosphere from the ocean ridge below the continent possibly forms an anomalous postrift uplift. Doglioni et al. (2003) assume it to be a global phenomenon.

The application of this model to the Gondwana breakup rift has to be seen critical. Doglioni et al. (2003) assessed only more or less orthogonal rifts, but the breakup rift was associated with a NNE-SSW translation along the DRFZ (Fig. 34). Nevertheless, an E-W extensional component during spreading may have acted on the lithosphere/asthenosphere asymmetry. Nevertheless, comparing the conjugated margins of Madagascar and East Africa, an asymmetry in the rift and the rift shoulder can be seen (Fig. 36). The south-eastern flank of the oceanic rift at the toe of the continental slope of Madagascar lies some hundred meters shallower than its counterpart in the north-western Somali Basin. This relative elevation in sea floor topography may also have other probable causes, e.g. transitional oceanic-continental crust between Madagascar and the DRFZ (Coffin and Rabinowitz, 1987).

By comparing the former adjacent continental margins of Tanzania and Madagascar a similar topography contrast has been recognized. Whereas the Madagascar margin is rising steeply above sea level, the conjugated East African margin at Tanzania is formed by broad belt of coastal low land (Fig. 36). Optionally this coastal flat is caused by sedimentary accretion along the shelf. Although there are various other factors that can explain the asymmetric rift and the discrepancy of continental uplift in Madagascar, the concurrence of both features may indicate the influence of the asthenosphere/lithosphere shifts.

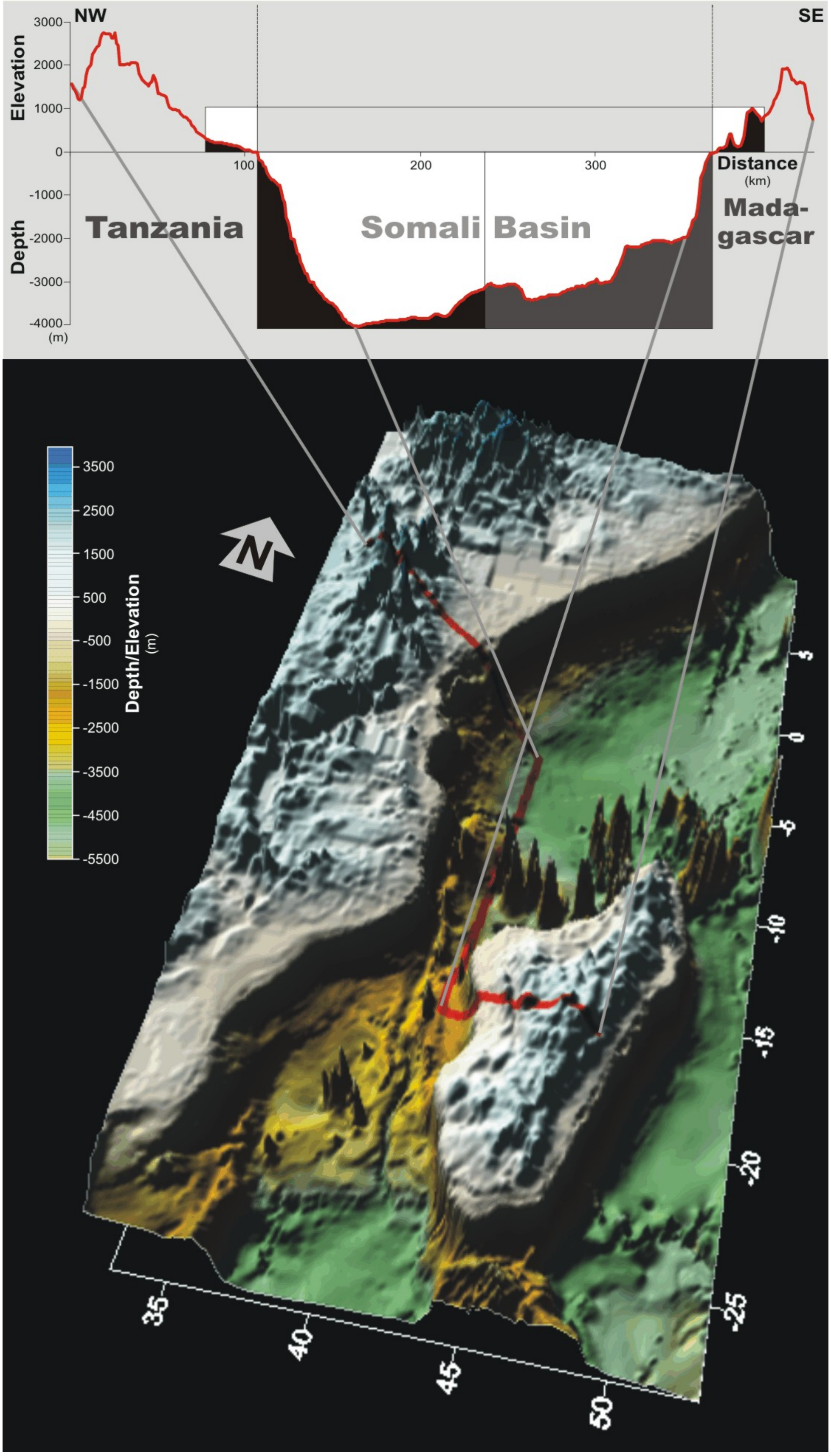

Fig. 36. Bathymetry/topography 3D model and a cross section of the conjugated margins of Madagascar and East Africa (Smith and Sandwell, 1997). Comparing the opposing margins during breakup a relatively elevated ocean floor and continent at the eastern margin (Madagascar) contrasts the western margin (Tanzania). Note that Madagascar rotated during the translation (de Wit, 2003).

4.4.4 Rift episodes: time gaps and dislocations

Continental rifting rarely follows the simplified models of continental extension (Chapter 4.4.1) but rather combines particular elements of different styles which are installed at varying places. Rifting often acts in episodes and the main locus of extension shifts from time to time with more or less the same orientation of the basin axis (van Wijk and Cloetingh, 2002). Thus natural rifts can apparently be divided in multiple rift phases with rift jumps, i.e. migration of the rift locus. Both the migration of the rift locus and the presence of several succeeding, temporally constrained rift phases are often closely related. The migration of rift loci is reported e.g. from the South Atlantic margin (Stollhofen, 1999), the mid-Norwegian margin (van Wijk and Cloetingh, 2002), and the Galicia margin (Manatschal and Bernoulli, 1999). While at the mid-Norwegian margin the locus of extension migrated basinwards to the area of future continental breakup, in Galicia the location of extension shifted from the interior basin to the site of future breakup.

Sonder and England (1989) found that cooling of the continental lithosphere during stretching might increase its strength, so that deformation may migrate to a region of low strain. Every rift shift to a new locus is accounted to form particular rift phases constricted by intermediate periods in which the lithosphere is not under tensile stress. Hereby thinned lithosphere is stronger than the rest of the margin, caused by a decrease of weaker crustal material in proportion to stronger mantle material compared to non-thinned lithosphere (Bertotti et al., 1997). Other authors account the strength variations to the influence of several factors, like thickness and composition of the crust, sediment thickness, geothermal gradient, and the asymmetry of the pre-rift geometry (Sawyer and Harry, 1991; Steckler and ten Brink, 1986). Van Wijk and Cloetingh (2002) modelled the initiation of the rift migration and found that continental breakup eventually occurs when extension velocity exceeds a critical point, while stretching lithosphere with lower velocities does not lead to breakup but forces the locus of maximum extension to migrate. The abandoned rift area eventually is being uplifted.

The model of van Wijk and Cloetingh (2002) helps explaining the complex rifting history of the Morondava Basin, derived from sedimentological and stratigraphical data. As demonstrated in Chapter 2 (Geiger et al., 2004) the Gondwana Breakup rift occurred on a narrow region surrounded by Karoo strata which were deposited during previous rifting episodes but became abandoned. Clark (1996) already assumed a rift jump responsible for this relocation of the locus of the future breakup further off the basin margin based on seismic lines. With regards to van Wijk and Cloetingh (2002), small extension rates formed the multiphase Karoo rift, and high extension rates resulted in the Lower Jurassic Gondwana Breakup. Extension rates for the East African-Madagascan are not known. Nevertheless, in consideration of the palaeo-plate configurations in reconstructions of Gondwana (Lawver et al., 1998; Reeves et al., 2002; Stampfli and Borel, 2002) and the relative position of crustal fragments during the persistence and breakup of the supercontinent, apparently the extension rate of the pre-breakup was small but increased in the syn- and post-breakup. Multiphase extension prior to the breakup lasted from the Late Carboniferous-Late Triassic (100 Ma) but only resulted in horizontal displacement of 50-100 km (Reeves et al., 2002). In contrast the breakup rifting produced a horizontal displacement of more than 2000 km within the late Early Jurassic-Neocomian (<60 Ma). Strain hardening, as a result of cooling induced by the sediment cover (Steckler and ten Brink, 1986) is also a considerable factor, since the pre-rift sequence (Karoo succession) comprises a thick sedimentary succession.

As rift migrations happen in a course of succeeding events, they bring about time gaps intermediating individual rift phases (van Wijk and Cloetingh, 2002). Often they are distinguishable by particular strain patterns in combination with the youngest sediments in which they can be found. Multiphase rifting of the Morondava Basin during Karoo times is described by Schandelmeier et al. (2004). Apart from subsequent rift episodes they found clear changes in the stress pattern for the individual extension phases. The first extensional phase is syn-depositional with the Middle Sakoa Group (Early Permian) and formed NNE-SSW opening pull-apart basins in a strike-slip stress regime. During the Early Triassic (Middle Sakamena Group) NW-SE extension acted in a transtensional sense, followed by normal extension syn-depositional Upper Sakamena Group but pre-dating Upper Isalo Formation (Late Triassic). It is noted here, that the stratigraphy was adjusted to the revised stratigraphy of Chapter 2 (Geiger et al., 2004). The stratigraphical assignments are critical for interpretation, since Schandelmeier et al. (2004) used stratigraphic ages on maps that were adopted to the stratigraphy of Besairie and Collignon (1972)(see the discussion about the reliability of Besairie's stratigraphy in Chapters 2 and 3). A fourth extensional phase in the late Early Jurassic resulted in the breakup of East- and West-Gondwana in the area of the Proto-Indian Ocean. Here, an obvious rift migration and rift focus (Chapter 4.4.5) occurred prior to the breakup.

4.4.5 Wide versus narrow rifting

The mode of lithospheric deformation, i.e. wide extension or narrow necking, as a response to extension is strongly dependent on lithospheric temperature and crustal thickness (Buck, 1991). Numerical models proposed to explain the factors concentrate on distribution of strain rates and strain distribution (e.g. Bassi, 1995; Bown and White, 1995), thermal conditions, like syn-rift magmatism (e.g. Hopper and Buck, 1996), and temperature-dependent rheology (Boutilier and Keen, 1994). Quantitative models support the model that narrow rifting starts from a normal crustal thickness, whereas wide rifting is predicted when the prerift configuration is characterised by a region of thick crust (Huismans et al., 2001). Wide extended zones itself can be shaped by two different mechanisms of strain delocalisation: (1) Diachronous extension with an outward migration of crustal necking at high strain rates (Bassi et al., 1993); (2) near synchronous extension with shifting fault-bound depocentres (multiple boudinage) at constant heat flow (Hopper and Buck, 1996). Van Wijk and Cloetingh (2002) found that the tendency of lithospheric necking, as a focus of strain, is weaker with decreasing extension velocities.

Pre- and syn-rift extensional stress in the Proto-Indian Ocean lasted from the Late Carboniferous until the Early Jurassic breakup rift. Structural style and the resulting basin geometry changed several times (Schandelmeier et al., 2004). Focusing to the Morondava Basin, during the pre-rift stage (Karoo-aged basin, Chapter 2) the basin was extending farther onto the margin (Chapter 4.9.2) and the basin was segmented by swells (Fig. 40), trending along basin-strike (Besairie and Collignon, 1972; Piqué et al., 1999). This infers a multiple boudinage mechanism in the Karoo-aged basin with shifting, fault-bound depocentres. A similar mechanism is known from the Triassic-Liassic of the North Sea (Praeg and Shannon, 2003). In contrast, after a period of non-deposition and probable tectonic quiescence, the Early Jurassic breakup strata (Andafia Formation) is localised close to the present-day continental margin (Chapter 2). Thus a relatively narrow and distinct rift locus resulted eventually in the crustal separation. This sudden localisation of extension at a narrow region of future breakup is, in correspondence with van Wijk and Cloetingh (2002), coherent with a spontaneous increase of the extension rate immediately before breakup rifting (Chapter 4.4.4). Thermal effects on the crust and its rheology, e.g. induced by magmatism, are rather improbable, because lacking syn-breakup volcanism is absent and thermal anomalies were not detected with apatite fission track (Emmel et al., 2004b).

4.4.6 Non-volcanic versus volcanic rifted margin

Rifted margins can be commonly characterised as either volcanic or non-volcanic margins. Although they have already been recognised as opposite endmembers of a wide spectrum of margins, this has remained a useful and robust classification scheme (Whitmarsh and Wallace, 2001). The presence of magmatic activity mainly depends on the presence of a thermal anomaly in the underlying mantle (White and McKenzie, 1989). The igneous rocks form by decompressional melting of asthenospheric mantle as it rises beneath stretched and thinned lithosphere. White and McKenzie (1989) propose in global considerations that this phenomenon is usually bound to mantle plume occurrences (e.g. US Atlantic: Holbrook and Kelemen, 1993; Gondwana margins: Storey, 1995). This indicates the relationship between volcanic rifted margins and active rifting (Chapter 4.3). Volcanic rifted margins are often preceded by large volumes of igneous rocks, the so-called flood basalts or Large Igneous provinces (Menzies et al., 2002).

In some global studies, Madagascar is classified as volcanic rifted margins (e.g. in Menzies et al., 2002), although breakup rift-related volcanics are absent. Exclusively in a few wells, e.g. Tandrano-1 (Uhmann, 1996), volcanics are described that are classified as Middle Jurassic. Rift-related volcanics are instead widely distributed south of the DRFZ around the Mozambique Basin, at the Antarctic margin and the south-east African margins. These volcanics originate from the Karoo mantle plume that produced the Karoo igneous province in southern Africa and the time-equivalent Ferrar igneous province at Antarctica. With regards to the geochronological-stratigraphical correlation scheme of Pálfy et al. (2000), the oldest recognised ages of the Karoo volcanics in South Africa ~183 Ma (Duncan et al., 1997; Encarnación et al., 1996), in north-eastern Mozambique ~177 Ma (Jaritz et al., 1977), and of Ferrar volcanics in Antarctica ~183 Ma (Encarnación et al., 1996; Pálfy et al., 2002) correspond with the postulated initial breakup in the Toarcian-Aalenian interval (Geiger et al., 2004)(Chapter 2).

Two explanations for the presence of a non-volcanic margin in western Madagascar apply:

(1) Numerical models demonstrate that a rheologically homogeneous crust and a moderately rapid extension rate promote the formation of non-volcanic rifted margins (e.g. Bowling and Harry, 2001). Below a critical extension rate asthenospheric ascent to shallow depth is delayed and lithosphere necking can well establish. The minimized production of syn-extensional melt reduces the spatial distribution of syn-rift magmatic rocks. After necking rifting proceeds rapidly to breakup.

(2) Far-field stress transferred from the active rift between Mozambique and Antarctica has been transferred through the crust to the East African domain and caused spreading between Africa and Madagascar.

4.4.7 Basin-fill patterns and shallow rifting

Apart from regional subsidence and uplift as a response of lithospheric extension, erosion and deposition of crustal material is an important factor when investigating passive margins during and after rifting. Numerical modelling of such material transport indicates that the erosion of uplifted rift shoulders has profound effect on offshore stratigraphic patterns. Van Balen et al. (1995) show that flexural uplift, due to isostatic rebound in response to erosion, may extend far into the basin and may cause uplift of the shelf. Contemporaneously sedimentation takes place in an offlap pattern. When the rift shoulder is levelled down, coherent cooling of the lithosphere and subsidence of the shelf promotes onlapping sedimentation.

Both offlapping and onlapping strata are present in seismic lines from the Majunga and Morondava basins (Clark, 1996; Geiger et al., 2004; Geiger and Schweigert, submitted)(Chapters 2 and 3). In the Morondava Basin, an offlapping sedimentary wedge comprises the Bajocian to Early Bathonian carbonate platform and the possibly the Bathonian siliciclastic succession. The latter, which is known from outcrops in the southern Morondava Basin (Geiger et al., 2004)(Chapter 3), has not yet been identified in the present seismic images from the north of the basin (Fig. 40A-C), since relevant intervals have been unroofed. However, forced regression-related features were described from seismic images by Stoakes and Ramanampisoa (1988), when they found incised valleys and palaeokarstification, indicating shelfal exposure. Above the carbonate platform, onlapping reflections illustrate the transgressive character of the succeeding deposits. The “Oxfordian Sandstone”, known form outcrops in the southern Morondava Basin has also not been recognised.

The succession of large-scale sedimentary patterns in the post-rift sequence from offlapping to onlapping structures follows the model of van Balen et al. (1995). These sedimentary patterns are the response to uplift of the rift margins and shelf areas. However, thermochronological studies based on apatite fission track (AFT) analysis do not record considerable denudation and thus implicated uplift along the rift margin during the proposed syn- and immediate post-breakup phase (Emmel et al., 2004b). Sensitivity limits of AFT analysis is limited to denudation of more than 2000 m at a presumed normal geothermal gradient (Lisker, pers. comm.., see also Gallagher et al., 1998). However the absence of an AFT signal does not exclude minor uplift. Rifting without extensive relief and the formation of shallow basins are explained by a shallow level of crustal necking during extension (Kooi et al., 1992). Crustal necking determines the ratio between thinning of the upper crust, where it is replaced by e.g. sediments and the thinning of the lower lithosphere, where crustal material is replaced by dense mantle material (Kooi et al., 1992). Necking as a geometrical concept to describe thinning in a kinematic way can be used for regions with passive rifting (Chapter 4.3)(Fjeldskaar et al., 2004). Thus it is possible to extend the crust and form a basin without uplift at its flanks. Moreover, rift-shoulder subsidence is postulated at certain near-real assumptions (Kooi et al., 1992), where deposition occurs directly on the subsiding rift shoulder. Such a similar situation exists at the West Iberian margin of the North Atlantic Rift, where syn-rift subsidence during the breakup was so small, that the syn-rift strata were resedimented towards the basin (Wilson et al., 2001).

In the Morondava Basin crustal uplift at the rift shoulder and induced flexural uplift of the shelf as proposed by the models of van Balen et al. (1995) was as prominent as to form typical stratal patterns but was not sufficient to produce an outstanding rift shoulder. As a result of this minor uplift the basin margin was subject to marine flooding during a global sea-level rise which was superimposed on the structural development of the basin (Chapter 3).

4.5 Palaeobiogeography and the early seaway

Palaeobiogeographical and palaeoenvironmental considerations involving ostracods are published in:

Mette, W. and Geiger, M., 2004. Bajocian and Bathonian ostracods and depositional environments in Madagascar (Morondava basin and southern Majunga Basin). Beringeria, 34: 37-56.

Mette, W. and Geiger, M., 2004. Taxonomy and palaeoenvironments of Callovian ostracoda from the Morondava Basin (south-west Madagascar). Beringeria, 34: 57-87.

Mette, W. and Geiger, M., 2004. Middle Oxfordian to early Kimmeridgian ostracoda and depositional environments of south-west Madagascar. Beringeria, 34: 89-115.

Depositional environments in the Proto-Indian Ocean are very similar along the entire ocean margins of East Africa, Madagascar, and India (Chapter 3). Eustatic sealevel changes strongly and the symmetrical position of sedimentary basins on either side of the rift influenced the sedimentary patterns forming facies of similar types. On a closer look, this apparent uniform palaeoecology turns out to consist of changes between distinct environments. Such observations were done on benthic foraminifer assemblages in south-western Madagascar (Chapter 3). In a more regional context, palaeoecological changes produced differing faunal assemblages or separated ecological niches to faunal provinces (vicariance biogeography). Marine faunas from the southern Tethys margin are known to have a well developed provincialism (Arkell, 1956; Hallam, 1969; Mette, 2004). Apart from palaeobiogeographical studies concerning the distribution patterns of cephalopods (Cariou, 1973; Enay, 1980; Riccardi, 1991) and benthic fossil groups such as foraminifers (Gordon, 1970), bivalves (Heinze, 1996), and ostracods (Mette, 2004). Especially ostracods are well suited for biogeographical studies, since they have no planktic larval stage and most of the taxa are exclusively benthic organisms. Thus ostracod individuals radiate less than 1000 m during their lifetime.

The total absence of ostracod assemblages from Toarcian black, shaly mudstones of the Ambohitsimeloka section (Beronono Formation, Chapter 2) is contrasting with the low diversity faunas of North Africa, Argentina and Europe (Mette, 2004). Grekoff (1963) also reports a poor fauna that has not been described in detail, probably due to poor preservation. The shaly mudstones of the Ambohitsimeloka section are interpreted to have been deposited in oxygen-depleted basins, unsuitable for benthic faunas. This is supported by the absence of a benthic macrofauna (Chapter 2). Comparisons of ostracod assemblages in other parts of the world suggest an intensive global faunal exchange (Ballent and Whatley, 2000; Whatley and Ballent, 1994).

Bajocian strata in the southern Morondava Basin yield a low diverse ostracod fauna with Cytheruridae and Progonocytheridae. Ostracods were found in the Analamanga, Anjeba, and Anteninde sections (Chapter 3). Facies interpretation on the basis of lithology, sedimentary features, and micro- and macrofossils suggest an intertidal, lagoonal environment with short episodes of brackish influence. This is supported by the low taxonomic diversity of the ostracod assemblages.

During the Early Bajocian Analamanga and Anteninde sections, the presence of globally distributed *Paradoxorhyncha* reflects a faunal exchange between the assemblages of Madagascar, Australia, and South America (Mette, 2004). The occurrence of *Striatojonesia striata* (Triebel and

Bartenstein) which is also known from Morocco (Bassoulet et al., 1989; Boutakiout et al., 1982), Egypt (Rosenfeld et al., 1987), and Jordan (Basha, 1980), shows faunal exchange with North Africa and the Near East. Species similar to the new *Ektyphocythere* sp1. are probably described as *Ektyphocythere mediodepressa* (Boomer, Ainsworth and Exton) from Portugal (Boomer et al., 1998), as *Ektyphocythere bucki* (Bizon) from Egypt (Rosenfeld et al., 1987), and as *E. dierallaensis* from Jordan (Basha, 1980). This close faunal similarities infer that the sea ingressed into the East African domain from the north (Mette, 2004).

The Andamilany section is considered to be slightly younger, probably of Late Bajocian or Early Bathonian age, although it was earlier considered in reference to Uhmann (1996) to be of Middle-Late Bathonian age (Mette and Geiger, 2004a). However, the stratigraphical classification is uncertain and recent considerations and correlations support a more likely Late Bajocian-Early Bathonian age (Chapter 3). This is also supported by the lithofacies which correlates with the Sakaraha Formation, and the Sakaraha Formation is believed to range up to maximum Lower Bathonian. The still low diverse ostracod fauna is dominated by *Oligocythereis* and *Bisulcocypris* which indicate shallow marine conditions with pronounced salinity fluctuations. Fresh water influence and changes in the salinity are also suggested by the macrobenthos and from the microfacies analysis (Chapter 2). Apart from environmental changes, the position higher up in the stratigraphy may have produced the faunal changes within the Sakaraha Formation. Palaeobiogeographical implications are not deduced.

Callovian marine mudstones (Amparambato, Atainakanga, and Behevo sections, Chapter 3), indicating the onset of the “Duvalia Marl” transgression, yield a highly diverse, normal marine fauna of Progonocytheridae, including *Majungaella*, *Pirileberis*, *Fastigatocythere* (*Habocythere*) and *Fastigatocythere* (*Batella*), and a number of endemic taxa. The co-occurrence of sixteen species in India reflects strong biogeographical affinities to there (Mette, 2004) while in contrast, Indian and Madagascan faunas have only four species in common with Tanzania (Mette, 2004). Other species from Tanzania are probably endemic. On the other hand *Fastigatocythere* (*Habocythere*) and *Fastigatocythere* (*Batella*) from Madagascar and India are not known in Tanzania (Mette, 2004). A small faunal exchange with the North Africa and Arabia are documented by six species, but the percentage of endemic taxa within the Indo-East African fauna increases (Mette, 2004). This observation reflects the increasing faunal differentiation that is also reported from benthic macrofossils (Heinze, 1996). These faunal changes suggest favourable environmental conditions in the shallow marine Tethyan indentation into the East African domain. The peripheral situation, warm water inflow from the Tethys (Heinze, 1996; Jansa, 1991; Mette, 2004) and high influx of nutrients (Dina, 1996; Uhmann, 1996) may explain the major radiation along the Proto-Indian Ocean margin of East-Gondwana. As further consequence, specialization as well as regional and ecological diversity increased (sensu Vermeij, 1995) and promoted provincialism (Mette, 2004). Environmental conditions extending contiguously northward (Arkell, 1956; Heinze, 1996) explain intensive faunal exchange with North Africa and Arabia on the one hand, but on the other hand, the faunal differences with nearby East Africa suggest that there must have been a faunal migration barrier. As discussed in

Chapter 3, the onset of sea-floor spreading, associated with possibly oxygen-depleted deep water conditions and submarine volcanism with the release of toxic hydrothermal fluids may have prevented the migration of shallow water benthic organisms (Chapter 3). The rapid increase in depth during the formation of the oceanic basin (Somali Basin) between East Africa and Madagascar is another possible explanation.

Late Jurassic ostracod faunas in Madagascar are constricted to the Middle Oxfordian-Early Kimmeridgian mudstones from Dangovato, Andrea, and Ankilimena sections (Chapter 3). The assemblages are characterised by a high number of endemic taxa of *Majungaella*, *Pirileberis* and a new species which is suggested to belong to a new genus ("*Australophocythere malagachia*" Ms.). Biogeographical reconstructions are limited by the fragmentary fossil record (Mette, 2004). Ostracod faunas in India, East Africa, and Madagascar appear to be highly endemic are different to those in the Near East and Saudi Arabia which rather have affinities to Europe (Mette, 2004).

The global distribution of Middle Jurassic ammonoids and bivalves is considered to be the result of climatic deterioration (Riccardi, 1991). Hallam (2001), however, noted that there are too few localities of Middle Jurassic ammonoids to use them for global considerations. The oldest ammonites from Madagascar are Early Toarcian and have no convincing similarities with ammonites from the Andes. The low diversity and the a strong affinity to the Tethyan taxa (Riccardi, 1991) suggest a migration barrier between Madagascar and the Andes in southern Gondwana. A prominent exception is the pandemic *Bouleiceras* sp. (Chapter 2). The migration barrier between the Proto-Indian Ocean and southern Gondwana prevailed until the Kimmeridgian. This is also supported by the high similarities of genera of Madagascar, Australasia, and the Tethyan realm (Riccardi, 1991). From the Late Tithonian-Berriasian the bivalve *Megacucullaea* that is unknown from other regions outside East Africa and Patagonia, gives the first evidence of the entire opening of a seaway between Africa and the conjugated margins of Madagascar and Droning Maud Land in Antarctica (Riccardi, 1977). Belemnite faunas from Antarctica also suggest a Late Tithonian opening (Mutterlose, 1986). Such a southward progression of the opening of the Proto-Indian Ocean is also supported by the first occurrence of marine sediments in the vicinity of Nacala at the north-eastern margin of Mozambique with a Late Kimmeridgian-Tithonian age according to a bivalve fauna (Prinz et al., 1993).

Conclusively, the faunal and lithofacies distribution patterns of the coastal basins of the Proto-Indian Ocean suggest southward ingressing marine conditions with a migration barrier from the Callovian onwards. This could be a result of favourable ecological conditions or a response to unfavourable conditions resulting from sea-floor spreading.

4.6 Sea floor ages of the Proto-Indian Ocean

The onset of oceanisation which is characterised by the initial formation of oceanic crust appear to have started in the Callovian (Geiger and Schweigert, submitted; Montenat et al., 1996)(Chapter 3), about ~10 Ma after the Toarcian-Aalenian breakup event is recognised in the tectono-stratigraphy of Madagascar (Geiger et al., 2004)(Chapter 2). This time discrepancy between continental rifting and sea floor spreading represents the period when oceanic crust starts to fill the gaps between the partly sediment-covered crustal fragments prior to sea-floor production (Bosence, 1998). Regular spreading later occurs along a defined rift locus.

The oldest directly dated oceanic crust between Madagascar and East Africa correlates with the M25 (~154 Ma, see below) magnetic anomaly (Coffin and Rabinowitz, 1992; Segoufin and Patriat, 1981). However, a clear dating of the oceanisation is problematic, since the Middle Jurassic appears to be devoid of clear changes of the magnetic polarity (Jurassic Quiet Zone, e.g. Vogt and Einwich, 1979). The M25 anomaly belongs to the Mesozoic (M-sequence) block model (Larson and Hilde, 1975). Correlations of the geomagnetic polarity timescale with ammonite zonations correlate the numerical age 154±2 Ma for the M25 with the Oxfordian-Kimmeridgian boundary (Ogg et al., 1984; Pálfy et al., 2000).

Despite of the lack of magnetostratigraphic constraints for the oceanic crust in the Somali Basin older than M25, the considerably extensive oceanic crust between M25 and the continental shelfs at the conjugated margins of the Somali Basin are probably 154-180 Ma in age (Müller et al., 1997)(Fig. 37). An age of 180 Ma correlates with datings of volcanics along the margins of, South Africa, Mozambique, and Droning Maud Land in Antarctica (compare ages in Chapter 4.4.6) and with the breakup model derived from the tectono-stratigraphy (Geiger et al., 2004)(Chapter 2).

However, as long as the age of the oceanic crust at the continent-ocean boundary in western Madagascar is unknown, indirect datings are the best way to meet the deficiency.

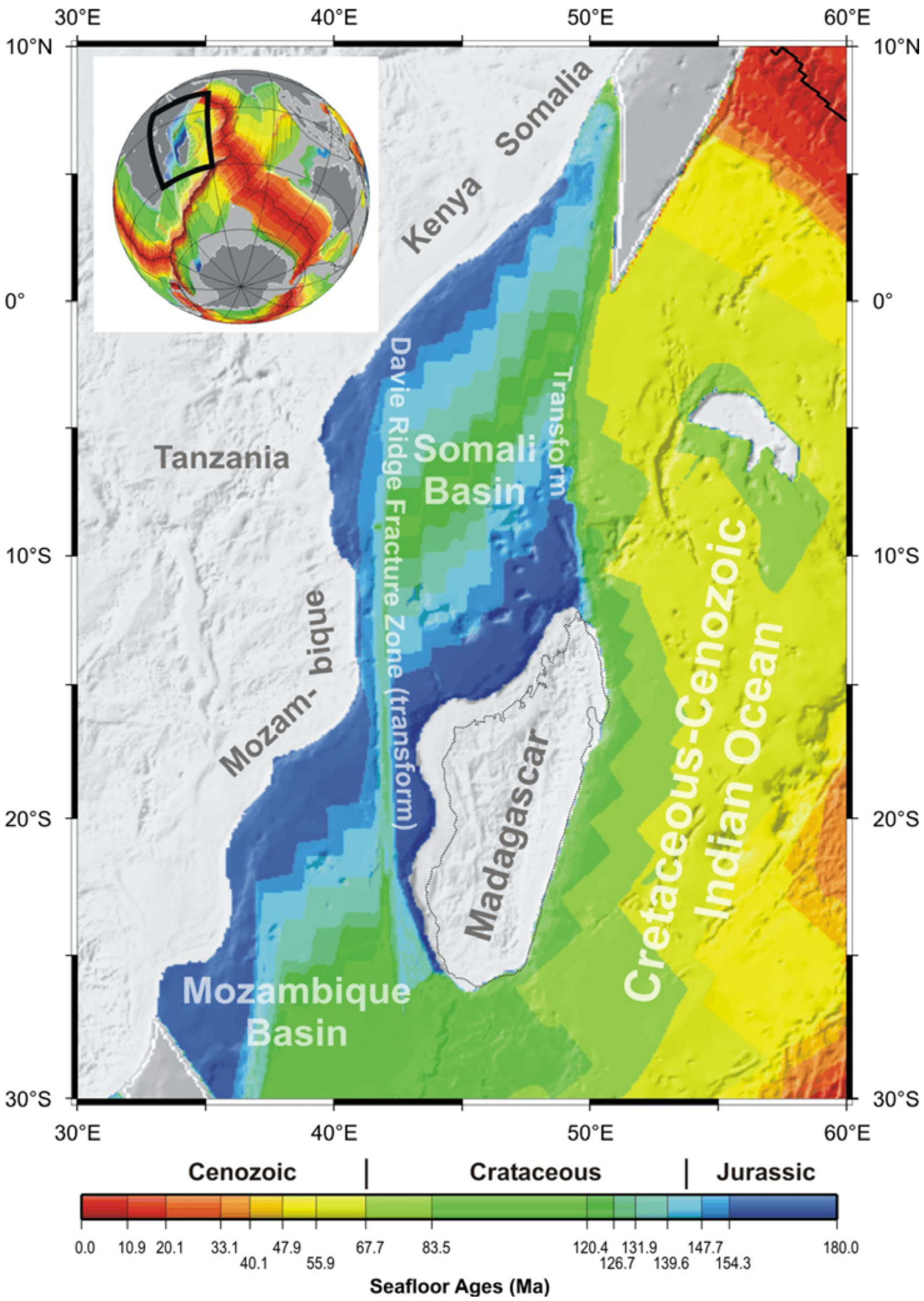

Fig. 37. Sea floor ages in the present-day Indian Ocean. The modern Somali Basin and parts of the Mozambique Basin have been formerly part of the Proto-Indian Ocean. They have the oldest sea floor ages in the present-day western Indian Ocean. Seafloor ages after Müller et al. (1997) draped on the GEneral Bathymetric Chart of the Oceans (GEBCO) data set (Jones, 1997). For a global position see inset.

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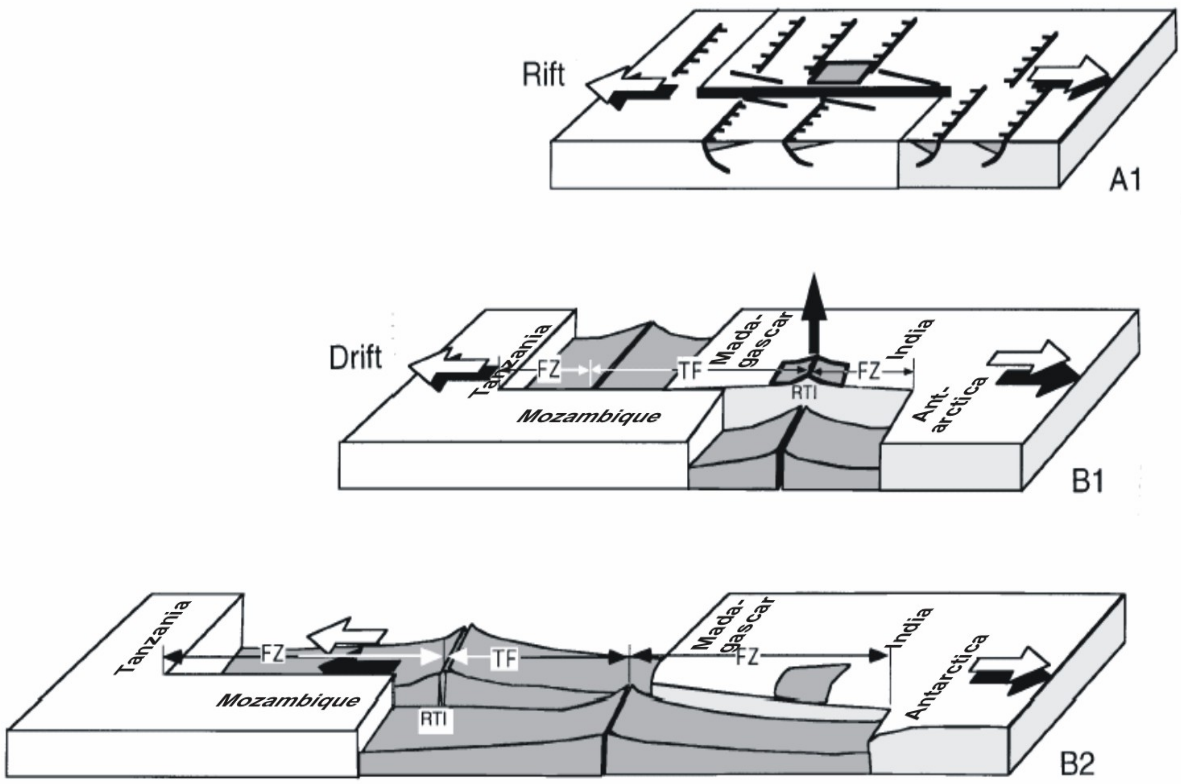
Fig. 38. Schematic three-stage model of shear margin formation during rifting (A) and drifting (B1and B2). During rifting (A1) continent-continent shearing with complex basin formation occurs. During early drifting (B1), the young oceanic block slides against the older continental block possibly inducing thermal uplift. After the passing of a ridge (A2) when the transform becomes inactive, mechanical coupling across the fracture zone is possible. As a result, the faster subsiding oceanic lithosphere may bow down the surface of the continental block. RTI: ridge-transform intersection, FZ: fracture zone, TF: transform fault. Vertical black arrow implies thermally induced uplift on continental block. After Lorenzo (1997). Present-day countries are indicated.

4.7 Davie Ridge Fracture Zone: shear margin characteristics on a fossil transform fault

During the study of the Gondwana Breakup rift with its final separation and subsequent drift, a main structural element comes into focus: the Davie Ridge Fracture Zone (DRFZ)(Fig. 33). The DRFZ is a transform that guided the translation of East- and West-Gondwana during the drift. Strictly speaking, the term “transform (fault)” refers only to active boundaries, but it is often interchangeably used with “fracture zone” (Lorenzo, 1997). Today the DRFZ delineates the continental-oceanic crustal boundary in the Somali Basin (Fig. 33) but initially it formed a continent-continent transform with Madagascar lying north and Mozambique lying south of it. Later the East- and West-Gondwana became detached and proceeding shear formed an ocean-ocean transform. This is in correspondence with the typical stages of shear margin formation of Lorenzo (1997): (1) shearing of continental crust (continent-continent shearing) in a narrow region and complex rifting at its flanks; (2) development of an active transform boundary separating oceanic and continental crust; and (3) passive margin formation along an inactive fracture zone that also separates oceanic and continental crust (Fig. 38). During stage (1) high-standing ridges bound deep sedimentary basins (Fig. 38A1) as indicated by e.g. Guiraud et al. (1997), similar to modern accommodation zones between large half-graben rift systems (Rosendahl, 1987). In stage (2) the younger oceanic block slides along the active transform, heating the continental block on the opposite side of the transform, and possibly inducing thermal uplift (Fig. 38B1). Cooling of oceanic crust in stage (3) may bow down the continental block and initialise passive margin formation (Fig. 38B2). Potential influence of such coupled flexure is a rather modern idea (e.g. Gadd and Scrutton, 1997; Lorenzo and Wessel, 1997).



In contrast to passive margins formed by normal or near normal extension, shear margins differ in three characteristics (Bird, 2001). First, the transition from continental to oceanic crust is relatively abrupt. Second, complex rift basins develop along the continental side of the margin with various structures including normal, wrench, and strike-slip faults due to crustal extension subparallel to the margin. Third, ridges rising over the abyssal floor form along the continental side of the margin. This ridge formation is attributed to heat absorption from the juxtaposed oceanic crust as the ridge transform intersection (RTI, Fig. 38) moves along the plate boundary.

In the Gondwana Breakup context, the DRFZ is a heritage to the modern structural features of the neighbouring ocean basins and the surrounding continental margins. The formation of deep basin alongside the fracture zone is documented by the two depocentres in the south and central part of the Morondava Basin. With distance to the narrow region the transform will eventually rupture the subsidence decreases as recorded by the relatively thinner syn- and pre-rift strata in the north of the Morondava Basin. On the continental block, equivalent basin subsidence refers to e.g. Selous and Luwegu basins (Fig. 18). Thermal uplift of the continental block of Mozambique due to heating from the RTI across the fracture can not been proven as relevant thermochronological studies on the Mozambique-Tanzania margin have not been conducted yet. Nevertheless, abnormal basement uplift and the emersion of the shelf basins of north-eastern Mozambique due to a high-standing basement block in north-eastern Mozambique can be explained by heat transfer from the RTI and the younger oceanic crust across the DRFZ (compare Lorenzo and Wessel, 1997). Flexural bow of the continental margin as oceanic crust cools is a prominent effect at newly formed continental margins, but in the present case extensive regional sea-level rise may overprint this subsidence process (Chapter 3). Thermomechanical coupled flexure of the continental block (north-eastern continental margin of Mozambique, which hosts the Ruvuma Basin) across the fracture zone has not been studied yet (Fig. 18B).

4.8 Structural style of East African basins

Sedimentological studies in the East African domain face a similar shortage of published and accessible data as in Madagascar. The few publications with detailed facies description have been rarely put in a geodynamic context. Moreover, the stratigraphic relation of Mesozoic strata in the many basins in this area have apparently not been completely understood (Kapilima, 1984; Kreuser, 1995; Mbede, 1991; Salman and Abdula, 1995; Wopfner, 2002). Stratigraphic uncertainties and a shortage of lateral correlation due to insufficient wells hamper establishing a convincing concept. Seismic lines, the key to understand the architecture of the sedimentary sequences are often not reaching as deep as the Lower Jurassic breakup strata or are not accessible for public. Instead, schematic cross-sections are available and outline the conspicuous configuration of the East African margin (Mbede and Dualeh, 1997). Middle and Upper Jurassic, Cretaceous, and Cenozoic strata overlay the Gondwana breakup strata at the present-day coast. The Lower Jurassic is bound to symmetric rift basins which are seaward bounded by palaeo-highs, of which the most prominent is Zanzibar Island. The basin that opposes Madagascar's part of the Toarcian breakup rift can be expected further seaward, where data is even sparse. Furthermore, Middle Jurassic salt diapers and pillows have widely disrupted the tectono-stratigraphic architecture in the adjacent strata (Coffin and Rabinowitz, 1992). Especially in Tanzania, several intracontinental basins host Mesozoic marine and continental strata. They commonly are aligned with or coincide with basins of the present-day East African rift system (Bosworth and Morley, 1994; Daly et al., 1989). Although these basins are well studied, the tectono-stratigraphically relation with the breakup rift was not recognized yet. Solely Kreuser (1995) noted that he suspected the late Early Jurassic Ngerengere beds of Tanzania to be syn-breakup deposits. This correlates stratigraphically with the breakup strata of Madagascar (Chapter 2).

The coastal basins of Mozambique (Salman and Abdula, 1995) show completely different characters. In the north-east, the Ruvuma Basin is aligned along the Transform margin of the DRFZ (Fig. 34, Fig. 18B). Jurassic strata are only known form seismic but have not been penetrated by wells and thus the stratigraphic correlation is critical. Nevertheless, seismic lines presented by Salman and Abdula (1995) outline reflection patterns that are very similar to those of the Madagascar with half-graben structures overlain by presumed reef structures, indicating a prominent carbonate facies comparable to the platform. They were interpreted as Bajocian-Bathonian shallow water, lagoonal, and reefal sediments, associated with salt structures, and are overlain by Callovian shales (Salman and Abdula, 1995). The Mozambique Basin further south is a rift margin basin, comparable to the Morondava Basin. Jurassic strata comprise Early Jurassic basalts and rhyolitic tuffs, followed by probably Late Jurassic continental sediments, known as the "Red Beds" (Salman and Abdula, 1995). The absence of Early or Middle Jurassic marine strata suggests that a palaeohigh in the area of the future Mozambique Basin is possibly caused by the thermal anomaly of the Karoo superplume and induced uplift of the region (Chapter 4.4.6). Due to the volcanics, seismic surveys do not image any deeper structures that may define the basin structure.

4.9 Structure style of the Morondava Basin

The structural style of the basins in Madagascar and their relation to the behaviour and configuration of lithosphere have been studied by several authors during the past few years (e.g. Montenat et al., 1996; Piqué et al., 1999; Rakotondraompiana et al., 1999; Razafindrazaka et al., 1999). Nevertheless, those studies are concentrated to gravity data but lack the control by deep seismic and deeper well data to find a perception about the structure of the Mozambique Channel. To meet this handicap, a group of scientists propose to probe Gondwanaland with deep seismic reflection profiling to gain perceptions of the deep crust-upper mantle (Brown et al., 2003). This IGCP-promoted project LEGENDS (Lithosphere Evolution of Gondwana East from iNterdisciplinary Deep Surveys) is aimed to shed light on the deeper crustal and upper mantle structures, which govern the structural evolution of the shelf and oceanic basins. Until the realization of LEGENDS the interpretation of sedimentary patterns and conventional seismic lines already give considerable models of the basin structures and configurations.

4.9.1 Basin geometry and architecture

The geometry of the basins in Madagascar is generally known from seismic surveys which give good outlines of the basin floor topography. Because, only a few wells penetrate the entire successions within the basins, the age and facies of the basal deposits is mostly uncertain. Therefore the Karoo is mostly compiled undifferentiated. The shortage of stratigraphic control in the Mesozoic and Cenozoic, resulted in a generalised stratigraphy in most published seismic images (Anonymous, 1996; e.g. Besairie and Collignon, 1972; du Toit et al., 1997; Geiger et al., 2004; for the Majunga Basin: Hiller, 2000; Montenat et al., 1996). Lithological boundaries between sedimentary sequences as reflectors in the seismic images, however, outline the architecture of the basin fill despite stratigraphic uncertainties. Fig. 40 illustrates the contrast in tectonic activity between the Karoo strata and the overlying Mesozoic deposits. Horst and graben structures are widely known from the basin floor (Besairie and Collignon, 1972; Piqué et al., 1999) and they appear to have lasted far into Karoo times (Fig. 40A, C, and D). Low resolution and the pre-interpreted character of the composed seismic images is probable responsible that the short-lived Toarcian-Aalenian syn-Gondwana Breakup rift (Fig. 39) can not been recognised.

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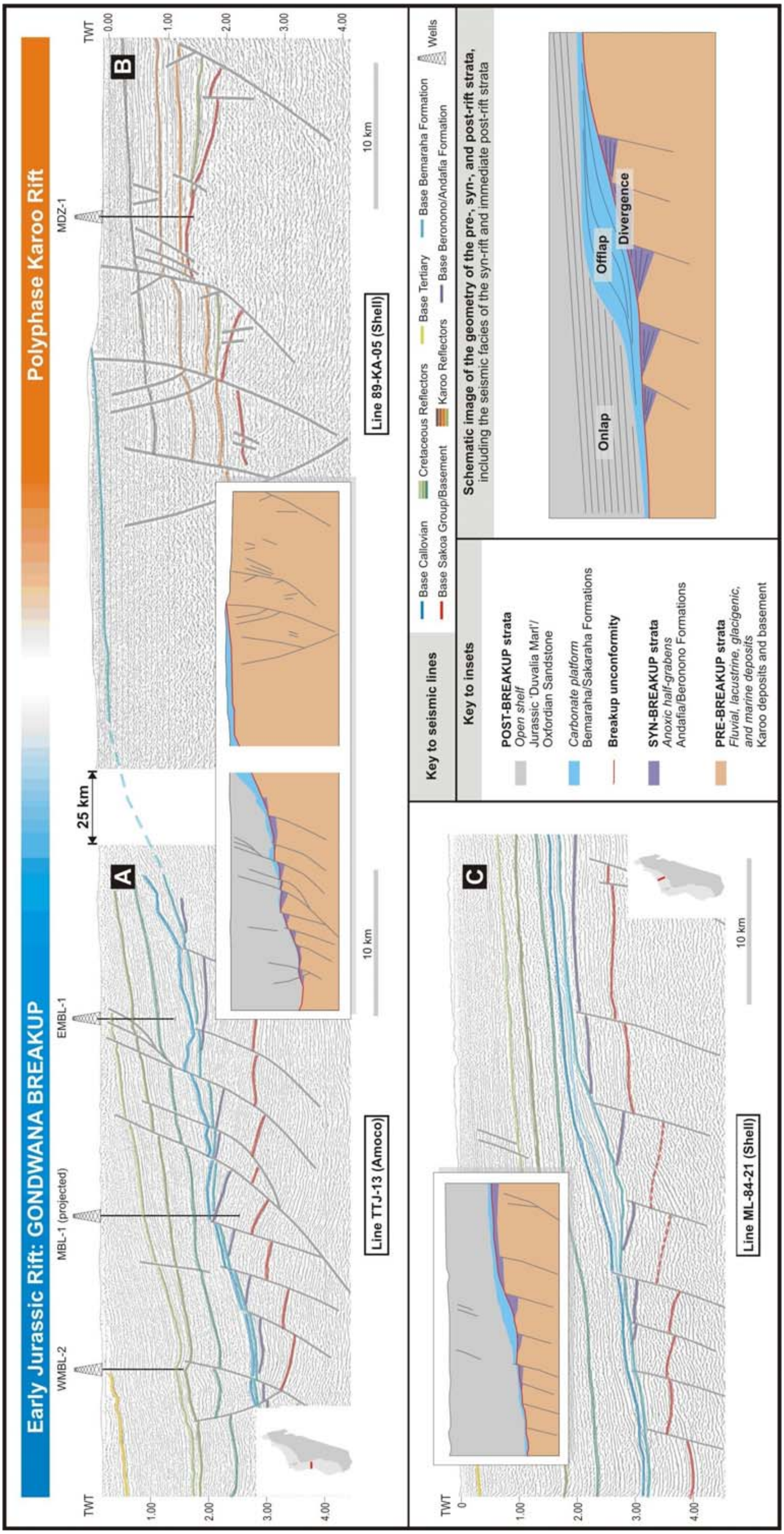
Fig. 39. Seismic images from the northern Morondava and the central Majunga basins (see inset map for location). In the Morondava Basin the overstepping character of the Bemaraha platform onto the Karoo rift system is demonstrated. In contrast the section form the Majunga Basin illustrates the lens shape of the carbonate platform. In both lines the syn-breakup succession is in a typical half-graben structure syn-rift reflection divergence.


Early Jurassic Rift: GONDWANA BREAKUP
Polyphase Karoo Rift
A
B
C
TWT
WMBL-2
MBL-1 (projected)
EMBL-1
MDZ-1
25 km
10 km
Line TTJ-13 (Amoco)
Line 89-KA-05 (Shell)
Line ML-84-21 (Shell)
Key to seismic lines
Base Callovian
Base Sakoa Group/Basement
Cretaceous Reflectors
Karoo Reflectors
Base Tertiary
Base Beronono/Andafia Formation
Base Bemaraha Formation
Wells
Key to insets
POST-BREAKUP strata
Open shelf
Jurassic 'Duvalia Marl'/
Oxfordian Sandstone
Carbonate platform
Bemaraha/Sakaraha Formations
Breakup unconformity
SYN-BREAKUP strata
Anoxic half-grabens
Andafia/Beronono Formations
PRE-BREAKUP strata
Fluvial, lacustrine, glacigenic, and marine deposits
Karoo deposits and basement
Schematic image of the geometry of the pre-, syn-, and post-rift strata,
including the seismic facies of the syn-rift and immediate post-rift strata
Onlap
Offlap
Divergence

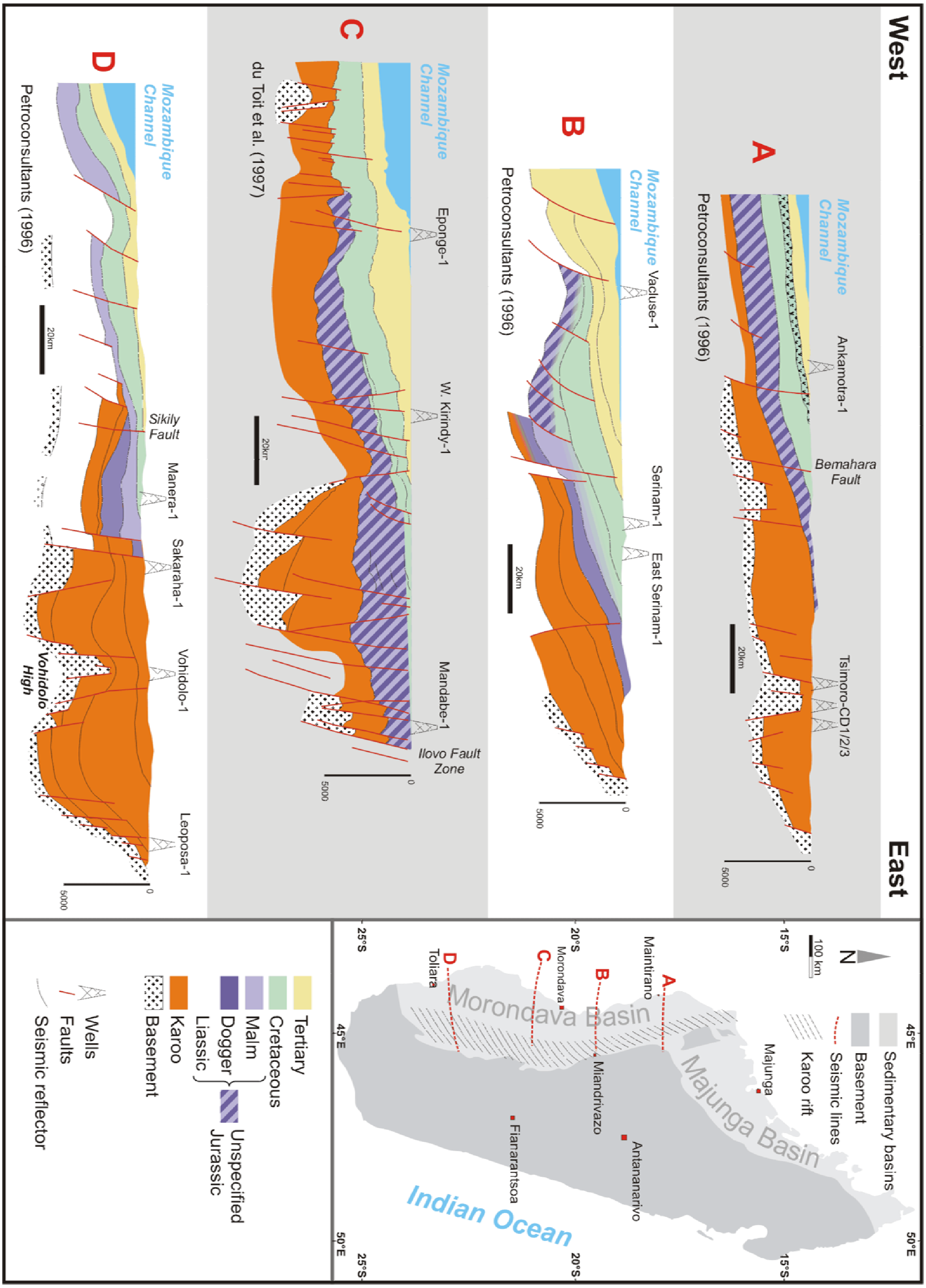


Fig. 40. W-E seismic images (basin to margin transects) from he Morondava Basin to illustrate the N-S variations of E-W. Sections A and B are shallow and less tectonically disrupted than sections C and D further south. Sections A, C, and D show an inter-basinal segmentation by horst and graben structures which trend to the basin strike. Horst and graben tectonism is generally constricted to the Karoo time deposits.

Isopach maps

In addition to the two-dimensional, vertical image of sedimentary sequences from seismic transects, isopach maps of the Morondava Basin illustrate the geographical (horizontal) distribution of the thickness of sequences in particular time slices. Isopach maps therefore indicate places of main deposition, or little erosion respectively. In marginal marine settings erosion and non-deposition are closely related, because deposition is essentially controlled by accommodation space (Jervey, 1988a).

As discussed in Chapters 2 and 3, the previous stratigraphy of the Morondava Basin is strongly based on lithological correlation. As a consequence, stratigraphic boundaries often coincide with lithological boundaries. From this, many of the previous studies used lithological distinctions to delimit sedimentary sequence. Although Chapters 2 and 3 present a more reasonable internal architecture of those particular sequences, the data set derived from previous simplifying stratigraphic concepts (Besairie and Collignon, 1972; du Toit et al., 1997; Montenat et al., 1996; Dina in Uhmann, 1996) has been retained to produce the isopach maps. Stratigraphic and sedimentological deficiencies of the underlying concepts were accepted by using the common stratigraphic groupings:

- Bemaraha and Ankazoabo formations form the so-called Bajocian-Bathonian sequence.
- Callovian-Tithonian sediments are part of one sedimentary sequence.

The younger sequence was limited to the Tithonian upper boundary, since solely the Callovian-Tithonian time-interval comprise a reasonable amount of data observations. Fig. 42 presents the database of well and outcrop sections from literature. Observations from measured field sections have been omitted, because none section contains entirely one of the two selected time intervals.

Linear thickness data was calculated by “Natural Neighbour” interpolation (e.g. Boissonnat and Cazals, 2000) which applies a weighted average of the neighbouring observations (grid nodes), where the weights are proportional to the distance to the "borrowed area". The present-day basin-basement boundary is considered as basin margin, i.e. a line of non-deposition and thus is attributed the value 0 thickness. The basin margin is a line of auxiliary grid nodes that are no observations, wherefore the interpolation between the basin margin and the first observations is less trustful than the area of dense data further into the basin.

In the isopach maps of the Bajocian-Bathonian and the Callovian-Tithonian intervals, two depocentres are outlined, one in the southern and a second in the central part of the basin (Fig. 41). During the Bajocian-Bathonian interval, the one in the south is located at the Manera-1 well, whereas the one in the central part comprises the Manadabe-1 and Betsimba-1 well. Although Betsimba-1 is located further basinward, the succession is less thick. This probably represents the transition from the shelfal carbonate platform and siliciclastic shoreface to a more distal slope position with less sediment supply. The apparent decrease in thickness west of Manera-1 is rather a calculation artefact.

A similar situation is found during the Callovian-Tithonian interval. Both depocentres are still present. The apparent migration of the northern depocentre is caused by the failure of the Mandabe-1 data set. At this location Upper Jurassic strata are eroded and consequently have no value in the interpolation. The situation in the north of the basin is also not representative due to a lack of observations. Small changes of the localities of the observation point due to inconsistent well records also distort the geometry of the northern depocentre. Thus it is not possible to consider these changes as response to external influences.

The depocentres are separated by a "palaeohigh" at Sikily and Sakanavaka outcrop sections during the Bajocian-Bathonian, and during the Callovian-Kimmeridgian at well Tandrano-1 respectively. It represents an area of less deposition or higher erosion. The relatively thin successions at these locations are often set in context with a more sandy facies during the Middle Jurassic as described by Besairie and Collignon (1972), and were interpreted as sedimentary response to a palaeohigh (Besairie and Collignon, 1972; Luger et al., 1994). However, the few known data observations in the south-central part of the basin, a lack of sedimentological reappraisals there, and the questionable reliability of the facies interpretations in the early studies give reason to evaluate those results with a note of caution (compare Chapter 2). Nevertheless, the continuing appearance of thinner successions in the south-central part may be an indication of tectonism-controlled variation in accommodation space. Segmentation with depocentres divided by an E-W trending structural high may has developed due to extensional-compressional stress, induced from the DRFZ transform. Strain interpretation of the structure of the DRFZ by Malod et al. (1991) argues for several changes in the stress field in the region during the post-breakup drift.

The northern part of the basin is of continuously equal thin thickness. Whereas it is supported by several observation points during the Bajocian-Bathonian interval, there is no observation at all during the Callovian-Tithonian. This relatively thin Bajocian-Bathonian succession in the northern part of the basin lies above the sub-basinal extension of the RBSZ (Chapter 1). Although tectonic influence on the accommodation space is rather speculative, thermochronological studies from southern Madagascar infer vertical movements of basement rocks south of the RBSZ relatively to the north at the Lower-Middle Jurassic boundary (Emmel, 2004; Seward et al., 1998). Recently, denudation events of the same age are also dated in the very north of the basin (Seward et al., 2004).

De Wit (2003) interpreted the apparent decrease in total sediment thickness from south to north by rift propagation with the same trend. A longer duration of deposition resulted in thicker sediment successions in the south. De Wit (2003) noted that this is supported by the occurrence of the oldest sedimentary rocks in the south of the basin. However, de Wit's error in reasoning is that the bulk successions with a main proportion of Karoo-aged deposits were not deposited during the initial breakup rift (compare Chapter 2). For this, however, the geometry and high resolution biostratigraphy of the Andafia Formation, "Aalenian Sandstone", and their equivalents is important alone. The southward increasing sediment thickness in Bajocian-Bathonian interval can be interpreted as the formation of a deep sedimentary basin, bounding the continental marginal ridge of the DRFZ during continent-continent shearing (Chapter 4.7).

Facies distribution

Only few palaeogeographical maps of the Morondava Basin with reference to the observed locations exist (e.g. Coffin and Rabinowitz, 1992; Salman and Abdula, 1995). A more detailed facies map was presented by Montenat et al. (1996) who illustrated the distribution of lithologies within the central basin on the basis of well data. Unless exact facies analysis and environmental interpretation of many wells and outcrop descriptions are either missing or are not accessible, the use of simplified lithologies as information for the depositional environment is the best approximation. With a compilation of numerical facies values from new and published data at observation points throughout the basin (Besairie and Collignon, 1972; du Toit et al., 1997; Montenat et al., 1996; Dina, published by Uhmann, 1996), a new data set forms the basis of the facies maps presented in Fig. 42. It shows a compilation of the data sets used. The geographical assignment (coordinates) of localities taken from the literature was obtained from the name file for Madagascar on the GEOnet Names Server (GNS) at the National Geospatial-Intelligence Agency (NGA).

Four lithofacies classes are compiled to describe the major depositional environments of the Lower Bajocian (Tab. 1). During the Upper Kimmeridgian a main carbonate facies is not known, although sporadic observations are recorded (Besairie and Collignon, 1972). Outcrop observations in the southern basin (Chapter 3) infer a basinal setting of those Upper Jurassic limestones. Thus they are included in the k1 distal marine facies of the Upper Kimmeridgian facies codes (Tab. 1).

Lower Bajocian	Upper Kimmeridgian
• **b1** = distal marine mudstones (e.g. Ankarana Formation) • **b2** = limestones and mudstones (e.g. Bemaraha Formation) • **b3** = proximal marine mudstones, limestones, and sandstones (e.g. Sakaraha Formation) (b2–b3: Carbonate platform) • **b4** = continental deposits/fluvial sandstones (not preserved, because eroded!)	• **k1** = distal marine mudstones and limestones (e.g. Duvalia Marl) • **k2** = marine sandstones (e.g. “Oxfordian Sandstone”) • **k3** = continental deposits/fluvial sandstones (?not preserved, because eroded!?)

Tab. 1. Variables (facies codes) with numerical values (1, 2, 3, and 4) for particular lithofacies during two selcted time intervals: the Lower Bajocian (b) and Upper Kimmeridgian (k). In the Upper Kimmeridgian, no limestone facies comparable to the one in the Bajocian (b2) is known. Sporadic limestone beds are included into k2.

With these numerical values and the presumption, that in the entire basin the facies belts are passing continuously into each other in the numerical order (b1-b4, k1-k3), the lithofacies can be plotted in its geographical context. Thus the lateral facies transition behaves like a transition (decrease/increase) of a value at one observation point in relation to the value of the neighbouring observation point. This means that the maps show a simple natural neighbour interpolation between the observation points (see isopach maps above)(Fig. 43C). Similar to the isopach maps, the outline of the present-day basin margin was assumed to represent the basin margin in the maps.

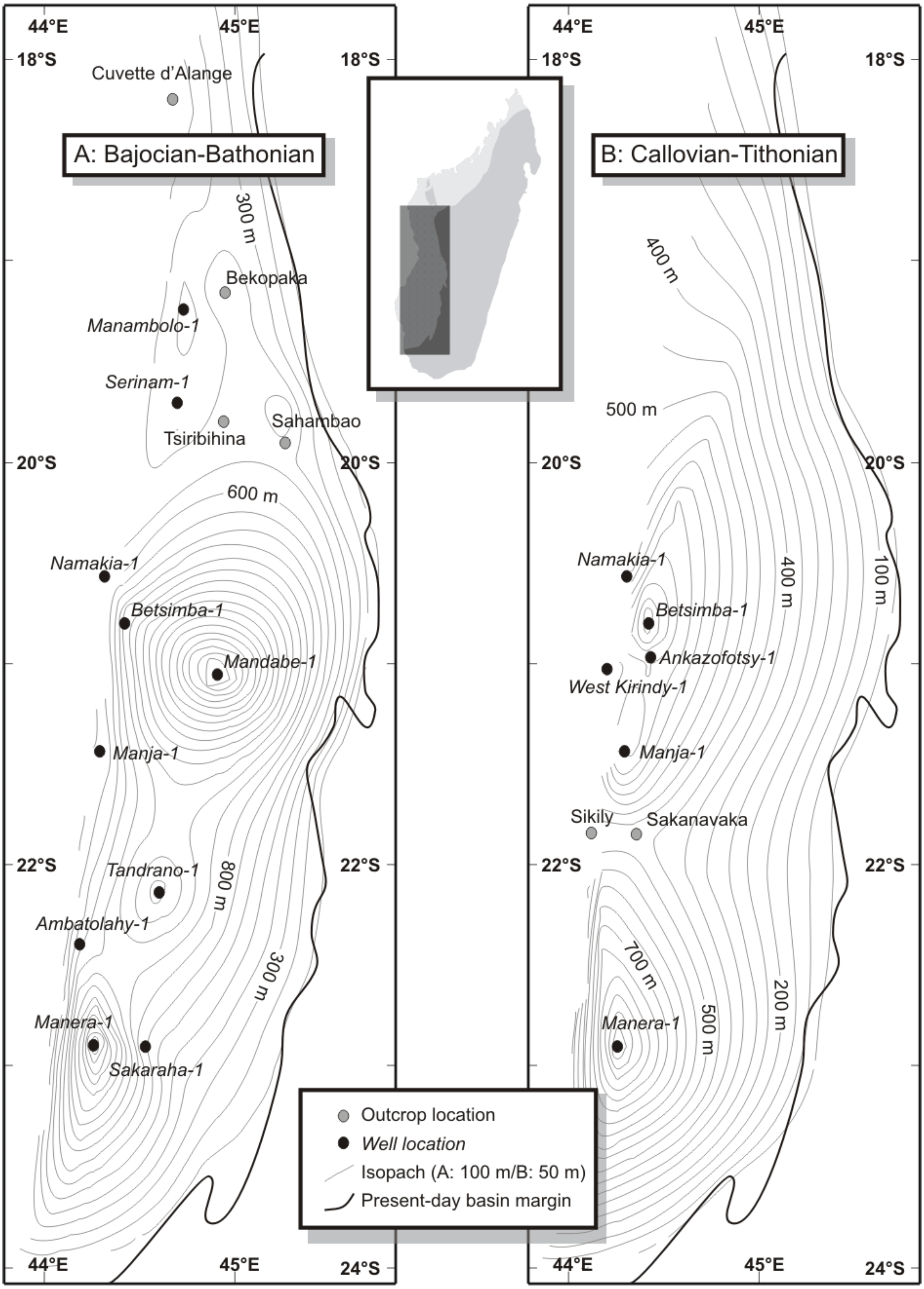


Fig. 41. Isopach maps for the Bajocian-Bathonian (A) and Callovian-Tithonian (B) intervals. The maps are calculated with the natural neighbour gridding method. During the Bajocian-Bathonian two depocentres formed in the central and southern basin. The situation prevailed until the Upper Jurassic, but the isopach gradient is smaller in B than in A. The thick line outlines the present-day basin margin as a clue for the possible former basin margin. The data sets are based on Besairie and Collignon (1972), Dina (1996, in Uhmann, 1996), Montenat et al. (1996), and du Toit et al. (1997). Compare Fig. 42.

Calculations of facies maps were exclusively applied to the Lower Bajocian and the Upper Oxfordian time slices, because other time intervals did not provide a considerable amount of data points (well localities).

The Lower Bajocian facies map (Fig. 43A) illustrates the basinwide extension of the Bemaraha carbonate platform (Chapters 2 and 3). In the very south of the basin, the mixed carbonate-siliciclastic facies of the Sakaraha Formation (Chapters 2 and 3) is present at all outcrops. In the north the entire facies transition from continental facies to the basinal facies (Fig. 43C) appears to crop out at a rather narrow belt on the shelf. In contrast, in the southern and south-central basin, the facies belts appear to be more than double as wide. Because of the wide extend of the environmental uniform depository, facies have probably changed much quicker than in a comparable narrower continental margin, such as in the north. Moreover, the depositional environments are strongly interfingering. This corresponds to observations at outcrops in the study area, where the mixed carbonate-siliciclastic character of the sediments is best developed. A similar wide margin is present in a narrow indentation in the central part of the basin, including the wells Mandabe-1 and Betsima-1 (Fig. 43A). Such a depositional feature can be caused by a palaeohigh situation providing increased sediment supply or by a delta complex. Of neither possibility further indicators are known.

Since Late Bajocian-Middle Bathonian times, the carbonate platform was probably exposed or covered by prograding siliciclastic shore face deposits (Chapter 3). Since Callovian times, deeper basinal environments established in most parts of the Morondava Basin, whereas siliciclastic shoreface deposits formed along the margin. The map (Fig. 43B) of Upper Kimmeridgian successions suggests the presence of basinal lithofacies in areas in the north and in the south of the basin, which approximately coincide with the depocentres, observed in the isopach maps (Fig. 41). These basinal facies areas are separated by a wide siliciclastic-influenced central part of the basin. A clear classification to marine or continental/fluvial environments can not be derived from all literature data. For instance, the sandstones at Andrea section (Chapter 3) lie above Middle Oxfordian mudstones. The Kimmeridgian data set is entirely devoid of prominent carbonate environments, comparable to those of the Bajocian time interval.

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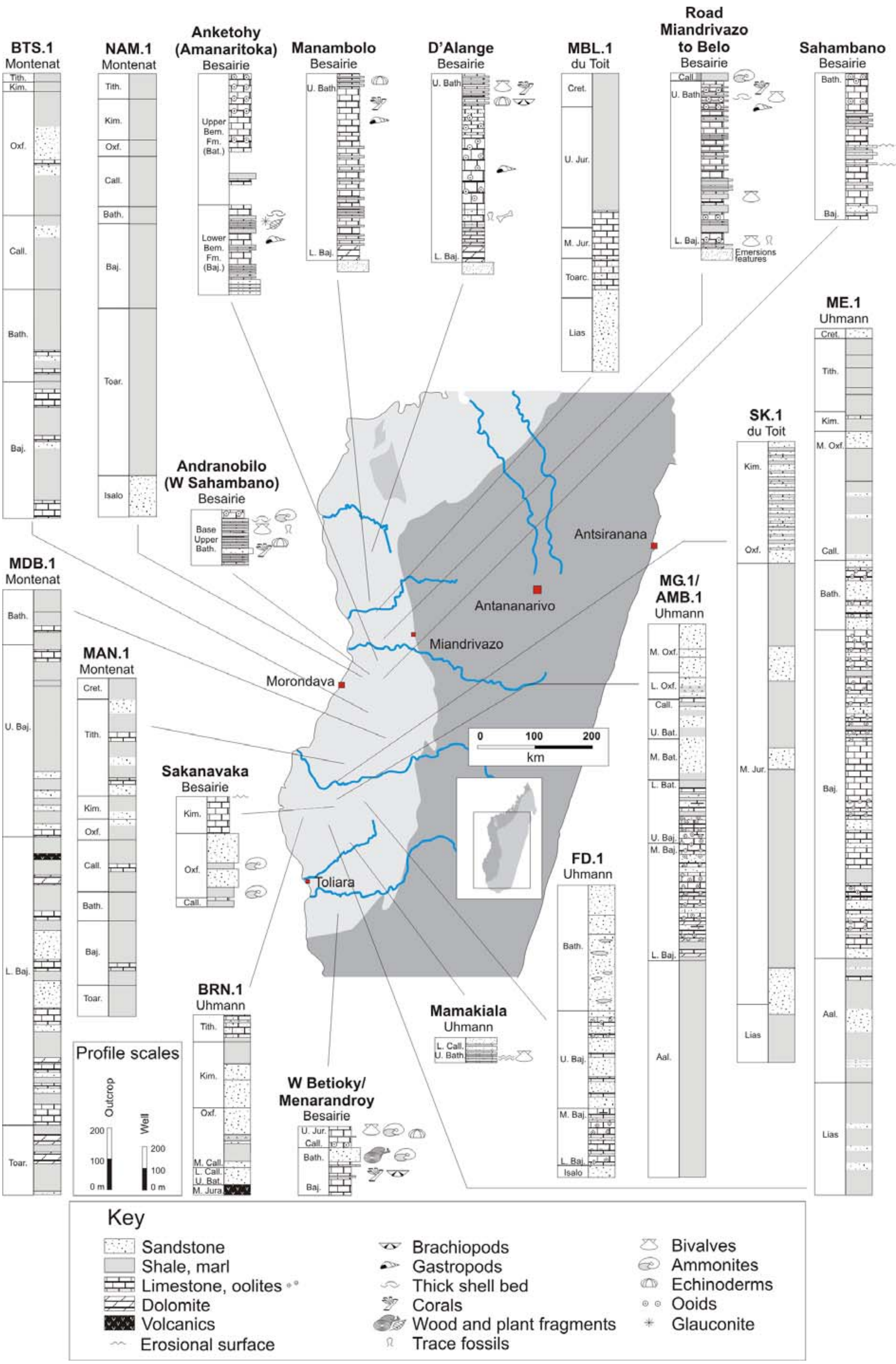
Fig. 42. Compilation of well logs and outcrop sections from Besairie and Collignon (1972), Dina in Uhmann (1996), du Toit et al. (1997), and Montenat et al. (1996). BTS.1: Betsima-1; NAM.1: Namakia-1; MBL.1: Manambolo-1; ME.1: Manera-1; Sk.1: Sikily-1; MDB.1: Mandabe-1; MG.1: Ambatolahy-1; MAN.1: Manera-1; FD.1: Tandrano-1; BRN.1: Beravy-1.



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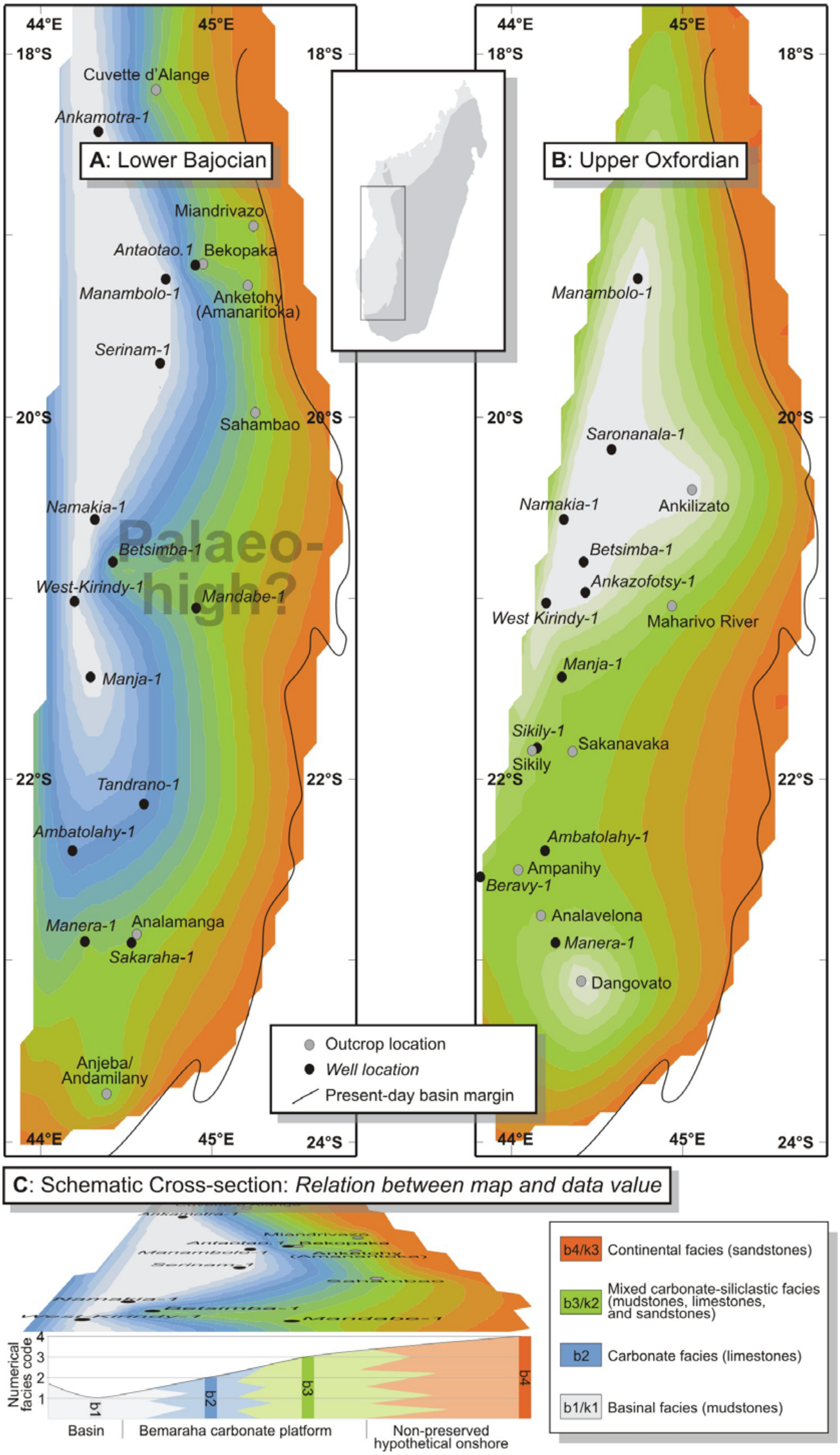

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Fig. 43. Lithofacies maps of the Lower Bajocian (A) and Upper Kimmeridgian (B). Discrete values (1, 2, 3, and 4) correspond to lithofacies codes of b1, b2, b3, b4, k1, k2, and k3 (compare Tab. 1). Shaded colours indicate areas of interpolation with indifferent facies signature. Improbable facies occurrences at the western margins of the maps are calculation artefacts. The perspective map and schematic section (C) illustrates the relation between the interpolated values on the numerical slope between the discrete facies values.

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4.9.2 Apatite fission track: implications to provenance and stratigraphy

Implications from detrital apatite fission track analyses are published in:

Emmel, B., Geiger, M., and Jacobs, J., submitted. Detrital apatite fission-track ages in Middle Jurassic strata at the rifted margin of W-Madagascar - indicator for a protracted resedimentation history: Sedimentary Geology.

Provenance studies of Jurassic strata in the Morondava Basin were applied by Uhmann (1996) on the basis on heavy minerals. Uhmann outlines that heavy mineral suites of sedimentary rocks can only critically be assigned to source areas. Generally, metamorphic, sedimentary, and minor magmatic signatures are observed in Middle Jurassic sediments, while sedimentary source rocks are important during the Upper Jurassic. Correlations with basement areas in Madagascar are not proposed.

New provenance implications for the Morondava Basin are presented by Emmel et al. (2004a) on the basis of thermochronological data from Bajocian-Callovian sediments. The stratigraphic distribution of apatite fission track (AFT) ages shows a partly reversed stratigraphy. AFT provenance analysis compares the detrital AFT ages and modelled T-t paths in the basin with those from possible source rocks in basement areas. AFT ages are cooling ages which are stored to apatite minerals by the statistical distribution and length of radiogenically produced lattice deformations, when the mineral passes through the 110-60° C partial annealing zone (PAZ). AFT is a sensitive low temperature thermochronological technique that is useful to characterise tectonic cycles by estimating the dimension of denudation and burial (Carter, 1999; Gallagher et al., 1998).

Detrital AFT ages in Jurassic strata of the Morondava Basin increase upwards in stratigraphically younger rocks (Fig. 44). This is a result of reworking of the Karoo deposits during the Jurassic. During Karoo times, the detrital cooling ages were reversely deposited in the Karoo basin with oldest cooling ages at the bottom and youngest cooling ages at the top of the sedimentary succession. Later during and immediately after the breakup (Bajocian-Bathonian), the marginal Karoo rift basin was uplifted as part of a rift shoulder and provided erosional material which filled the Jurassic breakup rift basin further to the west. This reworking process inverted the stratigraphy and placed the younger detrital AFT ages from the top of the Karoo rift basin to the bottom of the Jurassic breakup basin (236-381 Ma), whereas older AFT ages were eroded only delayed from the older Karoo strata and represent younger Jurassic strata with older detrital ages (Fig. 44). Thus the oldest detrital AFT ages are found in Callovian strata ranging between 422-429 Ma, whereas younger AFT ages of 233-382 Ma are found in the older Bajocian and Bathonian strata.

The detrital AFT ages in the Callovian sediments even predate the oldest known AFT ages from preserved Karoo sediments (Sakamena Group), which date between 319-399 Ma and are probably derived from sediments of the Late Carboniferous-Early Permian Sakoa Group or even older unknown deposits (Emmel et al., 2004b). Sakoa rocks are rarely exposed which raise the question of the origin of those old AFT ages. A reasonable explanation is that in this time the area of deposition has been extended far eastward onto the present-day basement. Similar implications come from thermochronological study of Seward et al. (2004) from the central basement area. Sakoa Group samples from the basin-basement boundary of the southern Morondava Basin have been exhumed above the PAZ at 180 Ma. The modelled thermochronological path implies a post-depositional burial to approximately 4,000 m during the Permo-Triassic (Emmel et al., 2004a). Such deep burial directly at the present-day basin-basement boundary should coincide with a deposition overstepping this boundary. The sediments with AFT signatures older than 400 Ma could have been delivered from those unpreserved basin parts further east.

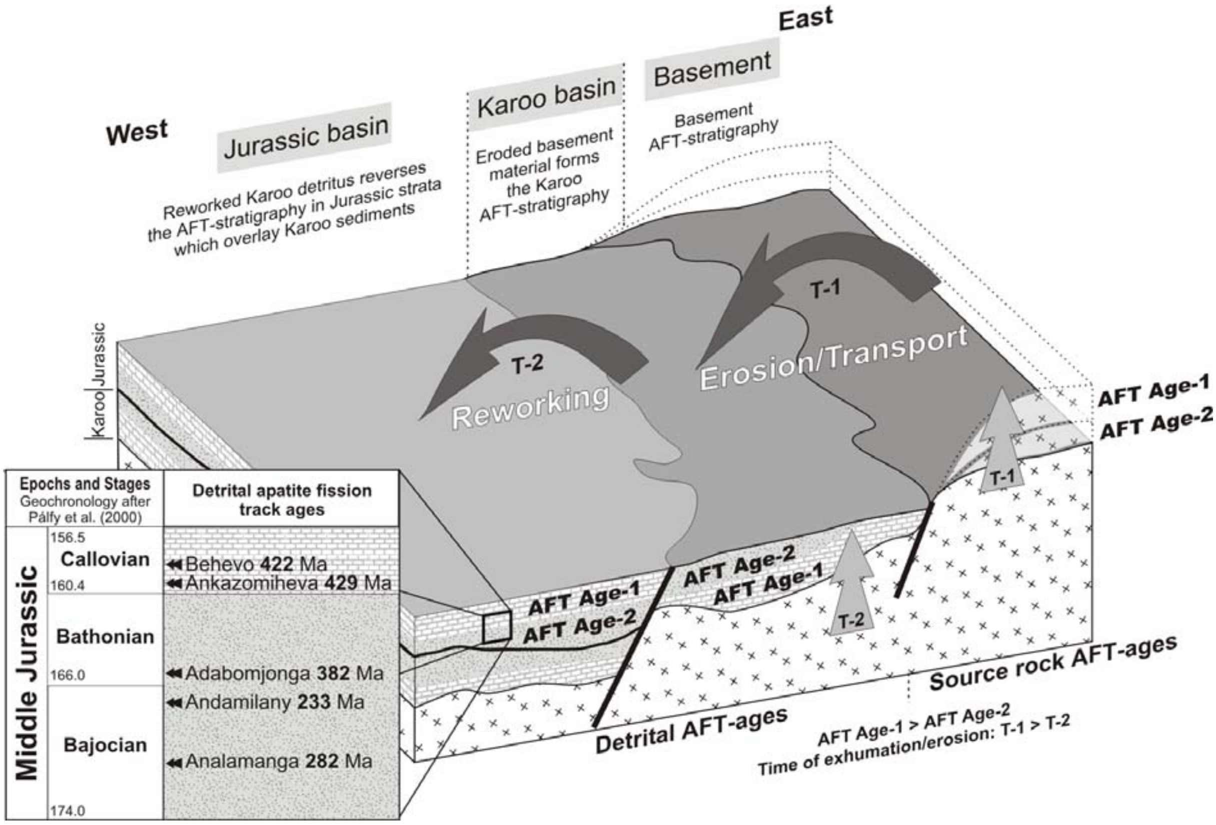

Fig. 44. Schematic model of the stratigraphic successions based on apatite fission track (AFT) ages in the Karoo rift basin and in the Jurassic Gondwana Breakup basin at the western margin of Madagascar. Reworked basement rocks filled the Karoo basin as indicated by inverting the cooling age stratigraphy of the basement strata (T-1). Later the erosion of the Karoo basin provided material for the Jurassic basin at the margin of the Gondwana breakup rift (T-2), and reversed again the detrital AFT age stratigraphy.

4.9.3 Palaeo-current data and basin filling mechanisms

Palaeo-current measurements from sedimentary structures, e.g. ripple marks and cross-stratification, are also used for the interpretation of palaeo-transport directions (Füchtbauer, 1988). Inferences to source areas are mainly based on the interpretation of fluvial transport directions. However, in the studied sections marine strata with coastal sediment transport are dominant, so that tidal and wind-induced wave currents are deduced which rather outline coastline morphologies.

The analysis of palaeo-current patterns within a stratigraphic frame are hampered by discontinuous outcrop conditions and regional occurrences of outcrops of different stratigraphical ages. In conclusion an area-wide comparison of all stratigraphic levels is impossible. Nevertheless, considerable trends in the palaeo-currents are measured by ripple crests, alignments of belemnites and wood debris, channel orientations, and bedding plane orientation (Fig. 45)

In the Bajocian a N-S oriented flow direction (Fig. 45B) implies a physical high in the very south of the basin. This elevated region proposed to have existed in the south-western basin more or less coincides with the Vohibory basement structure which contains AFT ages of ~180 Ma (Emmel, 2004). Bathonian sediments (Fig. 45C) chiefly record E-W directed flows from the elevated basement region in the east. NE-SW directed flow patterns in the northern study area demonstrate a westward indenting shoreline during the Oxfordian (Fig. 45D), which is connected with the siliciclastic unit of the “Oxfordian Sandstone” (Chapter 3). This sandstone unit is proposed by several authors (e.g. Besairie and Collignon, 1972) to occur in the central part of the basin.

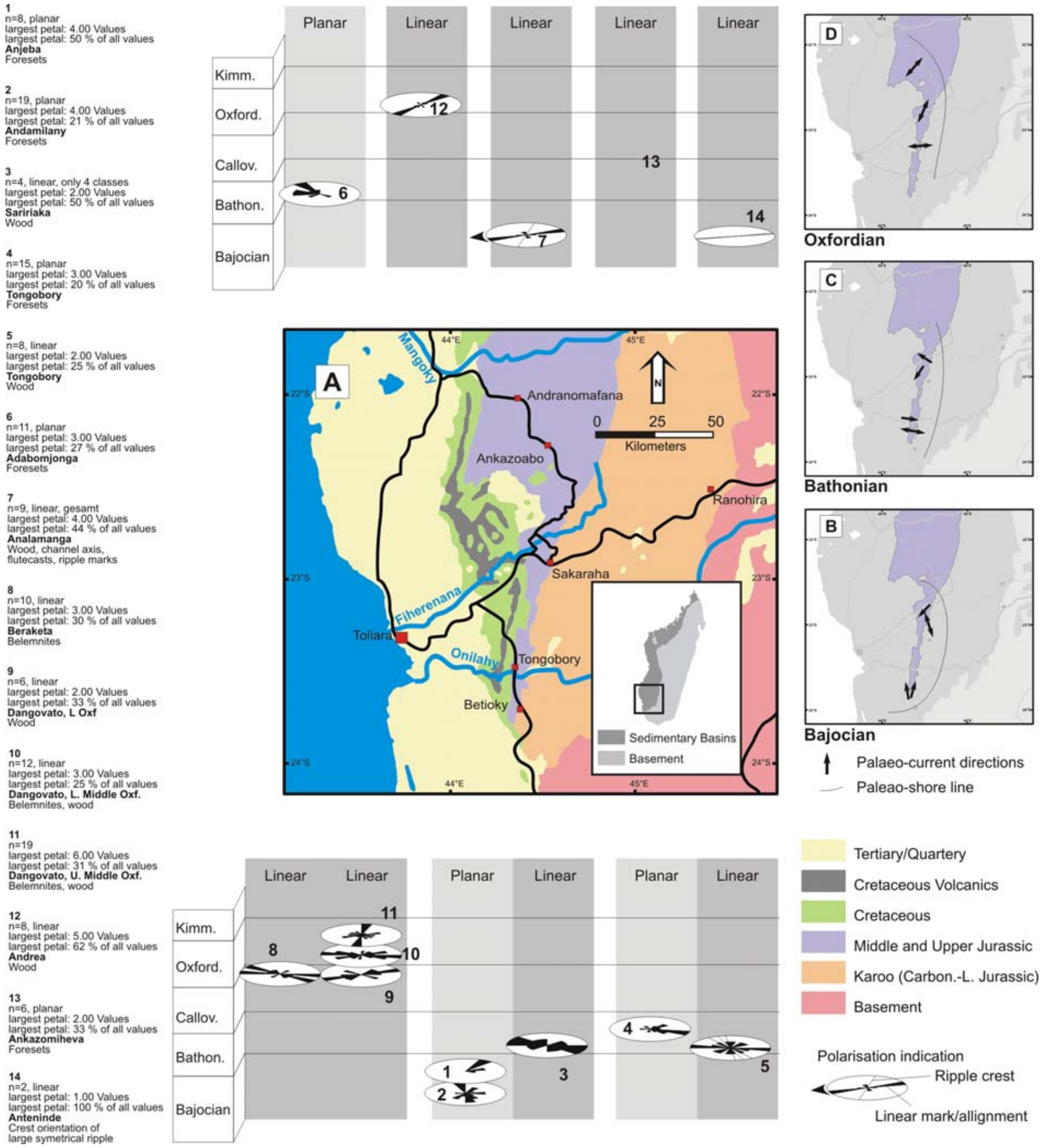

Fig. 45. Palaeo-current diagram with linear data from e.g. aligned wood and belemnites and planar data from bedding planes for various outcrops and time slices. Inset maps B-D illustrate the probable shore line morphology with a general N-S trend. Especially the Bajocian map (B) indicates the E-W bending of the shoreline and a N-S sediment transport. In the Oxfordian (D) a land indentation is suggested by NE-SW directed flow patterns in the south-central part of the basin at the northern margin of the study area.

4.10 Benthic foraminifers and palaeoenvironments

Exclusively in the Callovian-Kimmeridgian successions prominent foraminifer assemblages provide an applicable tool to decipher environmental changes. Generally, the analysis shows bimodal distribution pattern of foraminifer-test composition, agglutinating and calcareous tests (Fig. 46, Fig. 47). The agglutinating forms have a simple inner structure, which is in contrast to the typical Tethyan carbonate platform assemblage of the complex larger agglutinating forms dominated by *Pfenderinidae* (Gordon, 1970). The estimated abundance per standard sample volume of the agglutinating forms is relatively higher compared to those of the calcareous forms, in which a processed sample commonly only reveals a few tens or even less foraminifers. In turn the diversity on genus level is much less in agglutinating assemblages than in calcareous assemblages, often containing only one or few individuals of the same genus.

Monospecific assemblages of calcareous foraminifers are found in the Bajocian Analamanga (X) and Anteninde (VIIIb) as well as in the Early Callovian Ankazomiheva (V), Antainakanga (VIa), and Amparambato (VIb) sections, in the Early Oxfordian-Early Kimmeridgian mudstones intercalating with iron-oolites above the Oxfordian Sandstone at Dangovato section (IVa), and in the Middle Oxfordian Ankilimena section (XI). Apparently, the calcareous foraminifers occur in shallow, agitated water environment, since their occurrence is associated with iron-oolitic coquinas (e.g. Dangovato) and bioclastic cross bedded sandstones (e.g. Ankilimena, V). Calcareous assemblages (Fig. 46, Fig. 47) usually comprise nodosariid forms, especially lenticuline forms dominated by *Lenticulina* sp. and *Astacolus* sp. Accessory forms are often *Dentalina* sp., *Nodosaria* sp., *Citharina* sp., and *Cristellaria* sp. The occurrence of a few *Epistomina* sp. might indicate deeper, more open water conditions for the base and is in the Upper Jurassic of Scotland associated with maximum flooding (Oxford et al., 2002).

Gordon (1970) studied the global distribution of Jurassic foraminifers and explained the occurrence of the test modes with the proximal and distal setting. He considers nodosariid forms to be in middle to inner neritic, normal or nearly normal marine conditions. They typically occur in mixed carbonate-siliciclastic environments on shelf seas with terrigenous input from the nearby lands. Nodosariids are common at the southern Tethyan margin and in more boreal environments (Gordon, 1970). Brassier and Geleta (1993) note that in Jurassic strata of Ethiopia a nodosariid fauna complements the benthos to a more boreal aspect. Brassier and Geleta also see a connection between nodosariid abundances and more open turbid waters. In contrast, Gordon (1970) considers nodosariid forms in middle to inner neritic, normal or nearly normal marine conditions. Simple agglutinating forms dominate cool water or low salinity environments with decreased calcium carbonate secretion (Gordon, 1970).

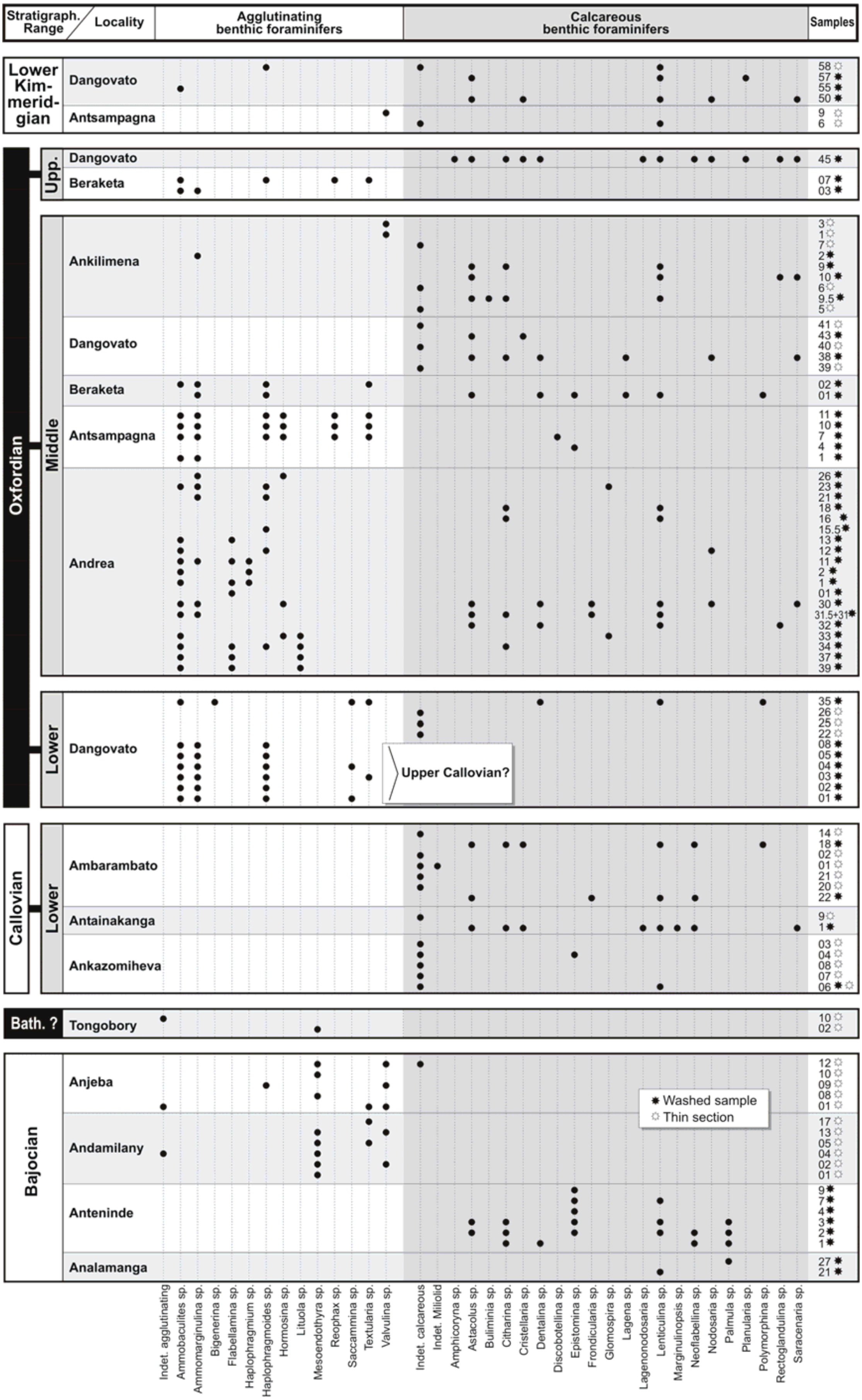

Fig. 46. Stratigraphic distribution of calcareous and agglutinating benthic foraminifers in Bajocian-Kimmeridgian strata of the study area. Sections are listed in stratigraphic order.

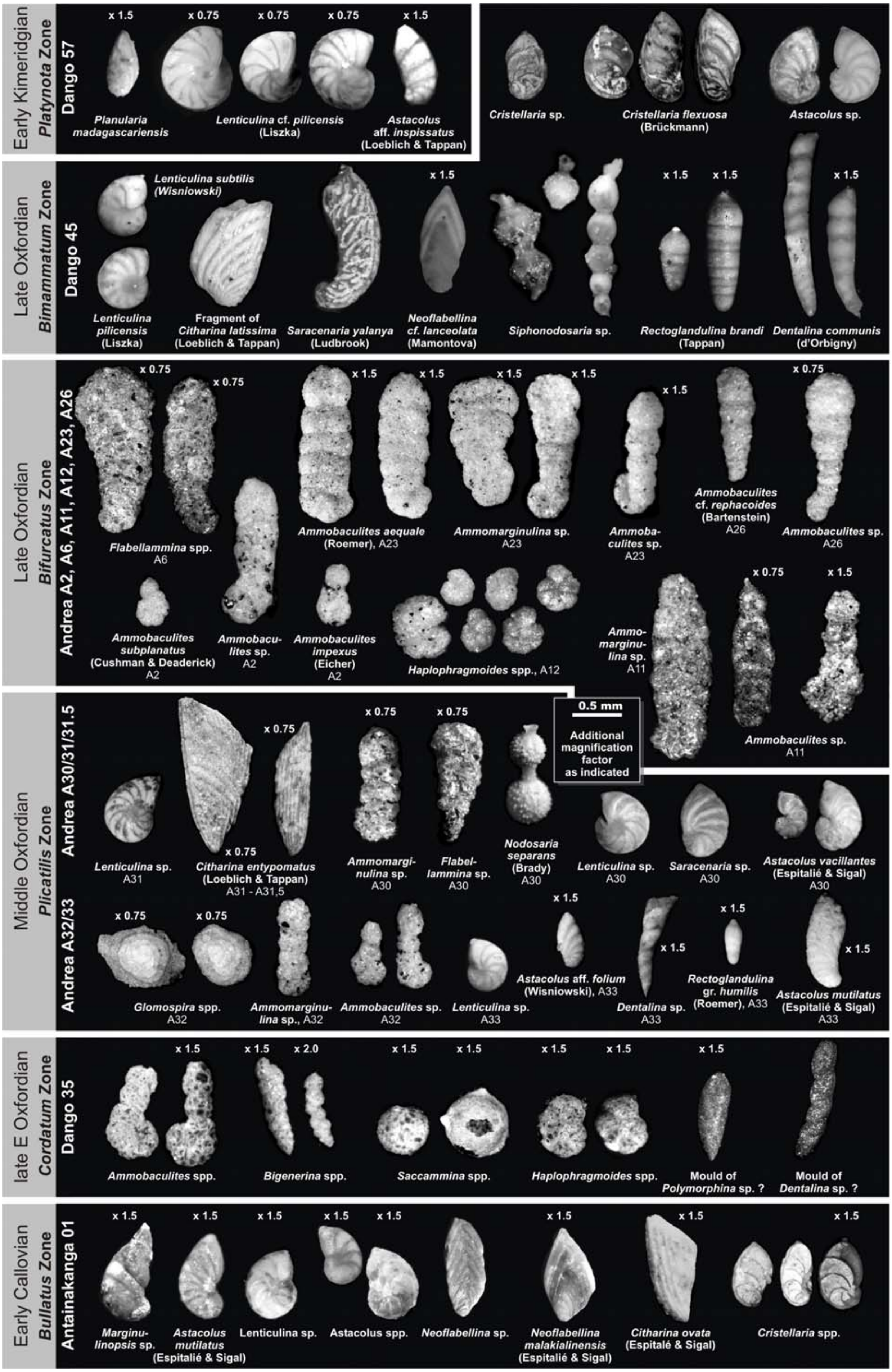
Early Kimeridgian
Platynota Zone
Dango 57
Planularia madagascariensis
Lenticulina cf. pilicensis (Liszka)
Astacolus aff. inspissatus (Loeblich & Tappan)
Cristellaria sp.
Cristellaria flexuosa (Brückmann)
Astacolus sp.
Late Oxfordian
Bimammatum Zone
Dango 45
Lenticulina subtilis (Wisniowski)
Lenticulina pilicensis (Liszka)
Fragment of Citharina latissima (Loeblich & Tappan)
Saracenaria yalanya (Ludbrook)
Neoflabellina cf. lanceolata (Mamontova)
Siphonodosaria sp.
Rectoglandulina brandi (Tappan)
Dentalina communis (d'Orbigny)
Late Oxfordian
Bifurcatus Zone
Andrea A2, A6, A11, A12, A23, A26
Flabellammina spp. A6
Ammobaculites aequale (Roemer), A23
Ammomarginulina sp. A23
Ammobaculites sp. A23
Ammobaculites cf. rephacoides (Bartenstein) A26
Ammobaculites sp. A26
Ammobaculites subplanatus (Cushman & Deaderick) A2
Ammobaculites sp. A2
Ammobaculites impexus (Eicher) A2
Haplophragmoides spp., A12
Ammomarginulina sp. A11
Ammobaculites sp. A11
0.5 mm
Additional magnification factor as indicated
Middle Oxfordian
Plicatilis Zone
Andrea A30/31/31.5
Lenticulina sp. A31
Citharina entypomatus (Loeblich & Tappan) A31 - A31,5
Ammomarginulina sp. A30
Flabellammina sp. A30
Nodosaria separans (Brady) A30
Lenticulina sp. A30
Saracenaria sp. A30
Astacolus vacillantes (Espitalié & Sigal) A30
Andrea A32/33
Glomospira spp. A32
Ammomarginulina sp., A32
Ammobaculites sp. A32
Lenticulina sp. A33
Astacolus aff. folium (Wisniowski), A33
Dentalina sp. A33
Rectoglandulina gr. humilis (Roemer), A33
Astacolus mutilatus (Espitalié & Sigal) A33
late E Oxfordian
Cordatum Zone
Dango 35
Ammobaculites spp.
Bigenerina spp.
Saccammina spp.
Haplophragmoides spp.
Mould of Polymorphina sp. ?
Mould of Dentalina sp. ?
Early Callovian
Bullatus Zone
Antainakanga 01
Marginulinopsis sp.
Astacolus mutilatus (Espitalié & Sigal)
Lenticulina sp.
Astacolus spp.
Neoflabellina sp.
Neoflabellina malakialinensis (Espitalié & Sigal)
Citharina ovata (Espitalé & Sigal)
Cristellaria spp.

Fig. 47. Selected Early Callovian-Kimmeridgian sample faunas of benthic foraminifers and their stratigraphic distribution from Callovian Antainakanga (VIa), and the Oxfordian Andrea (XII) and Dangovato (IVa) sections. See inset for scale.

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Monospecific assemblages of agglutinating foraminifers (Fig. 46, Fig. 47) are found in Bajocian-possibly Bathonian strata at Anjeba (Ia) and Andamilany (Ib) sections, as well as in possibly Early Oxfordian or Callovian mudstones/shales at the base of the Dangovato (IVa) section, and intercalate with agglutinating dominated but calcareous foraminifers-bearing assemblages in the Middle-Late Oxfordian Beraketa section (IVb), in the Middle Oxfordian-Early Kimmeridgian Antsampagna (IX), and in the Middle Oxfordian Andrea (XII) sections. Agglutinating forms are lituolid forms, such as *Haplophragmoides* sp., *Ammomarginulina* sp., and *Ammobaculites* sp.

The occurrence of agglutinating foraminifers is strongly associated with black dark mudstones (shales) that contain calcareous nodules (Beraketa and Andrea section). Nodules are common indicators of oxygen-depleted, non-carbonate but organic carbon-enriched environments, like black shales, with sulphur reduction and hydrogen sulphide and methane production (Ricken and Eder, 1991). Organic matter remineralization in such dysoxic environments can lead to more acidic bottom water and increasing carbonate dissolution (Hilbrecht et al., 1996). Another possible source of acidification is the fluid influx from newly developing oceanic ridge volcanism with commonly lower pH 4-6 in modern analogues (Charlou et al., 2000) compared to pH 8-8.5 of sea water. In conclusion, the preservation of carbonate, even in foraminifer tests, is reduced in dysoxic environments and this environmental stress promotes the replacement of the test material and consequently forms agglutinated tests. Wightman (1990) observed the dissolution of calcareous foraminifers in marsh environments as a result of pH decline due to plant decay. Nagy et al. (1990) found a similar positive correlation of organic-rich shelf sediments and the abundance of coarse agglutinating foraminifers.

Thus the proportional occurrence of agglutinating and calcareous foraminifers is an indicator of the depositional environment (Fig. 48). Calcareous assemblages infer shallow agitated water conditions with a reasonable proximity to land. Agglutinating forms are in organic-rich outer shelf sediments with indications of dysoxic conditions. Similar environmental distinctions on the basis of comparable calcareous and agglutinating foraminifers come from the Cretaceous Walu Shale of Tanzania (Nyagah, 1995). The inferred concurrence of iron-oolitic coquinas and calcareous forms (e.g. at Amparambato and Dangovato sections) emphasizes this distinction, since iron-ooids and coquinas typically form in higher energy environments. Whereas iron-ooid accumulation is presumably promoted by dysoxic conditions but low sulphide activity (Yoshida et al., 1998), iron-oolites form under marginal marine conditions and rather not in the sulphide-enriched basin bottom waters. This can explain why calcareous and agglutinating forms are not found together. Changes of the sea water chemistry within the Early Indian Ocean can be considered as a result of volcanic activity along the ocean spreading ridge (Chapter 3.4.5).

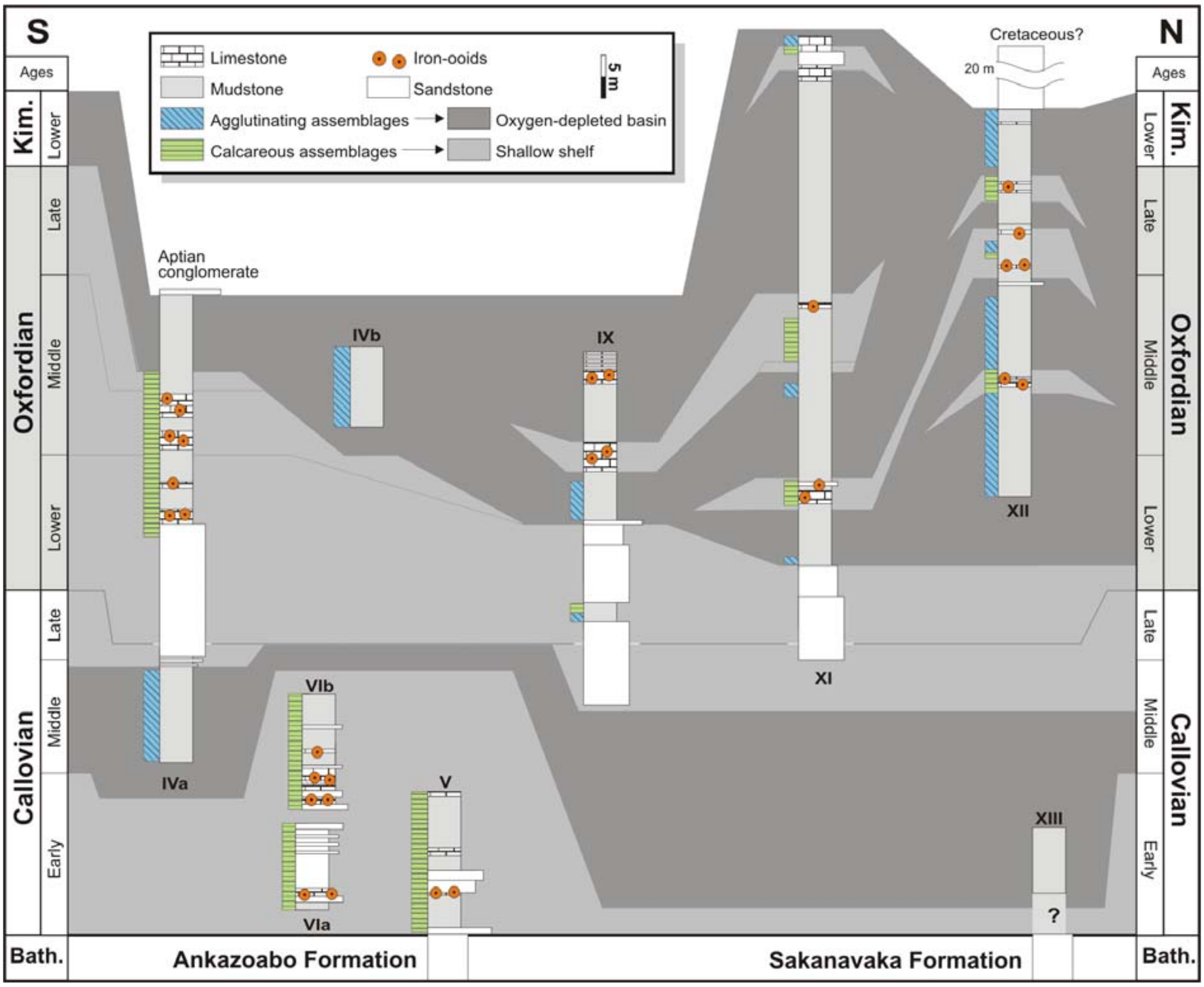


Fig. 48. Callovian-Kimmeridgian lithostratigraphy with palaeoenvironmental correlation on the basis of distribution patterns of iron-ooids and distinct foraminifer tests. Shallow shelf conditions prevail throughout the Early Callovian. In the late Callovian basinal conditions establish and pass gradually into the shallow shelf Oxfordian Sandstone. Despite the Early Oxfordian transgression shallow shelf conditions prevail in the south but in the north basinal conditions establish with only a few episodes of shallowing. See also Fig. 47.

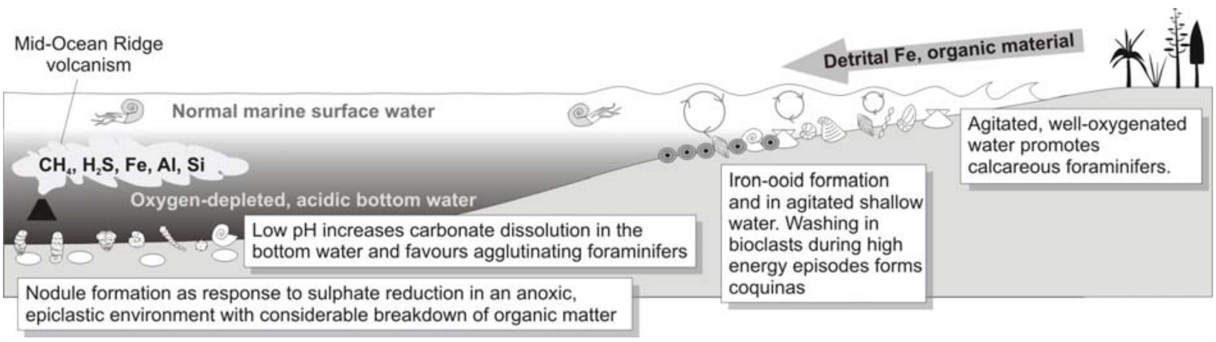


Fig. 49. Callovian-Oxfordian environmental model for iron-oolite production and foraminifer distribution patterns. Agglutinating foraminifers are common in the basinal, dysoxic environment, whereas the calcareous are found in the normal oxygenised, and agitated shallow water. For symbol keys see Fig. 22.

4.11 Hydrocarbon perspectives

Although the hydrocarbon exploration was strongly promoted in the 1950-70s and some oil and gas shows are recognised and a few discoveries are made, they had no commercial success (Hiller, 2000). A few companies still conduct exploration. For the exploration purposes the rough stratigraphical concepts of Besairie and Collignon (1972) are still widely in use, despite the voluminous reappraisal of the exploration potential of Madagascar by Clark (1996). With regards to the new constraints of depositional events and their position in the sedimentary architecture in the basin, it is possible to give a more precise stratigraphical occurrence of source rock-reservoirs relations. Thus in the Jurassic post-breakup succession two potential petroleum systems with oil-prone organic matter are present in the Morondava Basin (Tab. 2). Potential source rocks are (1) carbonate platform limestones of the Bemaraha Formation (predominantly phytoplankton and algae, kerogen type I/II and subordinate land-plant material of kerogen type III (2) and the mudstones of the Jurassic "Duvalia Marl" with kerogen type III and subordinate type II (kerogen types adopted from Clark, 1996; Hiller, 2000). Potential seal for both systems is the Jurassic "Duvalia Marl". Its lower part overlies the Bemaraha and the upper part seals the "Oxfordian Sandstone". Traps are usually in tectonic structures but can also exist in sedimentary structures, such as reef structures in the Bemaraha Formation.

Source rocks	Reservoir rocks	Seals
Jurassic "Duvalia Marl"	"Oxfordian Sandstone", interbedded limestones	Interbedded shales
Bemaraha limestones	Ankazoabo/Sakanavaka sandstones, Bemaraha limestones	Jurassic "Duvalia Marl"
Toarcian shales	"Aalenian Sandstone", Bemaraha limestones	Sakaraha mudstones (marls and claystones), "Duvalia Marl"

Tab. 2. Potential post-breakup petroleum systems of the Jurassic post-breakup succession of the Morondava Basin. Modified from Clark (1996) and Hiller (2000).

The present study shows that the Jurassic successions of the Morondava Basin do not reflect a general N-S facies contrast, as supported by most of the previous authors. Instead we have similar palaeoenvironmental conditions and tectono-stratigraphic patterns along-strike the entire basin. Thus source rock/reservoir rock combinations are also expected to be found basinwide.

4.12 Suggestions for a revision of the Jurassic rock units

Moreover, to account for the newly gained stratigraphic architecture of the Jurassic, I suggest giving some sedimentary units of the southern Morondava Basin the rank of formations (in accordance with NACSN, 1983)(Fig. 50). Based on the characteristics of shallow marine sandstones the covering a large stratigraphic interval the so-called "Aalenian Sandstone" should receive the name of its well exposure at Kandreho in the Majunga Basin: "**Kandreho Formation**".

Based on the characteristics of pelites in shallow (iron-oolites, coquinas) and deeper water conditions (black shales) for a virtual large stratigraphic interval (Callovian-Early Aptian?: Besairie and Collignon, 1972; Luger et al., 1994), I propose to regard the mudstones of the so-called "Duvalia Marl "succession as a group, possibly with the name "**Vineta Group**", according the mountain range SW of Sakaraha in which the reference section of Dangovato lies. The Vineta Group contains the formerly called "Oxfordian Sandstone" which is well mapable and can easily be distinguished by the sandy facies from the under- and overlying mudstones. Its lateral extend is unclear. Therefore it should also obtain the rank of a formation. As the type locality lies at the Dangovato River, I suggest name "**Dangovato Formation**". The "Dangovato Formation" laterally interfingers with the "Vineta Group" and pinches out basinward, where wells only penetrate the latter.

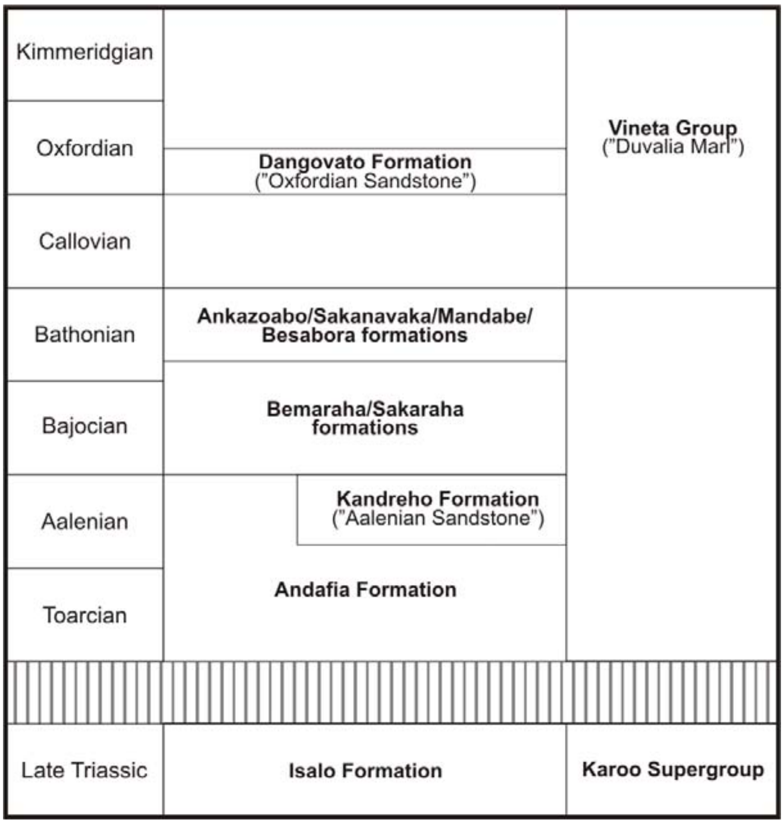


Fig. 50. Revised system of Members and Groups in the Morondava Basin.

Chapter 5 Conclusions and perspectives

In the preceding chapters it has been demonstrated that the structural history of the East African segment of Gondwana was characterised by multiphase extension and a complicated sedimentary response to both tectonism and global sea-level changes.

Initial crustal extension acted on a wide area that earlier formed the Gondwana wide East-African-Antarctic Orogen and pull-apart basins, rift basins, and sag basins formed (Schandelmeier et al., 2004). These basins were mainly filled with continental sediments of the Late Carboniferous-Triassic Karoo succession derived from uplifted rift flanks. The early Early Jurassic was a time of non-deposition. From the Toarcian (late Early Jurassic) onwards, marine conditions ingressed from the north across Arabia and North Africa into the East African domain. The formation of this rift is interpreted to reflect a shift of the rift locus to the west and to have resulted in the separation of East- and West-Gondwana. In half-graben structures above rotated fault blocks the shales of the Andafia Formation (Toarcian) and the "Aalenian Sandstone" were deposited. To compensate for the large horizontal displacement a continent-continent shear, the future Davie Ridge Fracture Zone developed at the southern tip of Madagascar. In contrast to the Karoo rift, the breakup rift did not form prominent rift flanks. Seismic data demonstrate that this relatively shallow margin promoted a transgression of the sea far inland, when the sea-level rose globally during the Early Bajocian. Limited uplift of the margin and the stabilisation of the shelf resulted in the formation of a coastal carbonate platform (Bemaraha and Sakaraha formations) while fission track data from detrital apatite (Bajocian-Callovian strata) show that Karoo sediments were the source for the younger post-breakup sequence. A major regressive event during the Bathonian partly exposed the shelf and formed a thick siliciclastic shoreface in what were probably forced regression wedges (Sakanavaka and Ankazoabo formations). A global sea-level rise during the Callovian flooded the margins of the Proto-Indian Ocean and deposited the "Duvalia Marl" in Madagascar. This transgression favoured the radiation of benthic organisms on the one hand, while on the other hand the coincidental onset of sea-floor spreading with its toxic hydrothermal fluids and the drastic deepening of the ocean basin established a migration barrier to the adjacent East African margin. During the subsequent drift of Madagascar away from Africa, the continental margin of Madagascar experienced a regressive event in the Late Callovian-Early Oxfordian ("Oxfordian Sandstones").

The integrated interpretation of sedimentary patterns at outcrop and stratal patterns in seismic images showed that the rift, which was responsible for the Gondwana breakup in the East African domain, was acting on the present-day western continental margin of Madagascar during the Toarcian-Aalenian. Consequently, the Karoo-aged rift should not to be considered a syn-breakup feature, but a long-lasting failed multiphase rift that predates the onset of Gondwana breakup. This model is in contrast to previous models, which considered the Karoo succession (including Early Jurassic sediments) the syn-breakup strata and the subsequent Bajocian-Bathonian marine succession a transitional phase (Hankel, 1994; Wopfner, 1994).

The study also shows that the breakup rift of Madagascar and Africa was a complex, non-standard rift that did not include many of the expected features. Uplift of the rift margin was of a limited extent such that the marine transgression (Bajocian) that immediately followed breakup could overstep the rift margin far inland. This coincides with the deposition of the Bemaraha carbonate platform. Limited tectonic activity during the breakup is also consistent with the observation that the sedimentary events were generally found to have correlative counterparts in other places in the world. Thus eustacy dominated the basin-filling processes. The sequence stratigraphic model was achieved via an improved litho- and biostratigraphy framework.

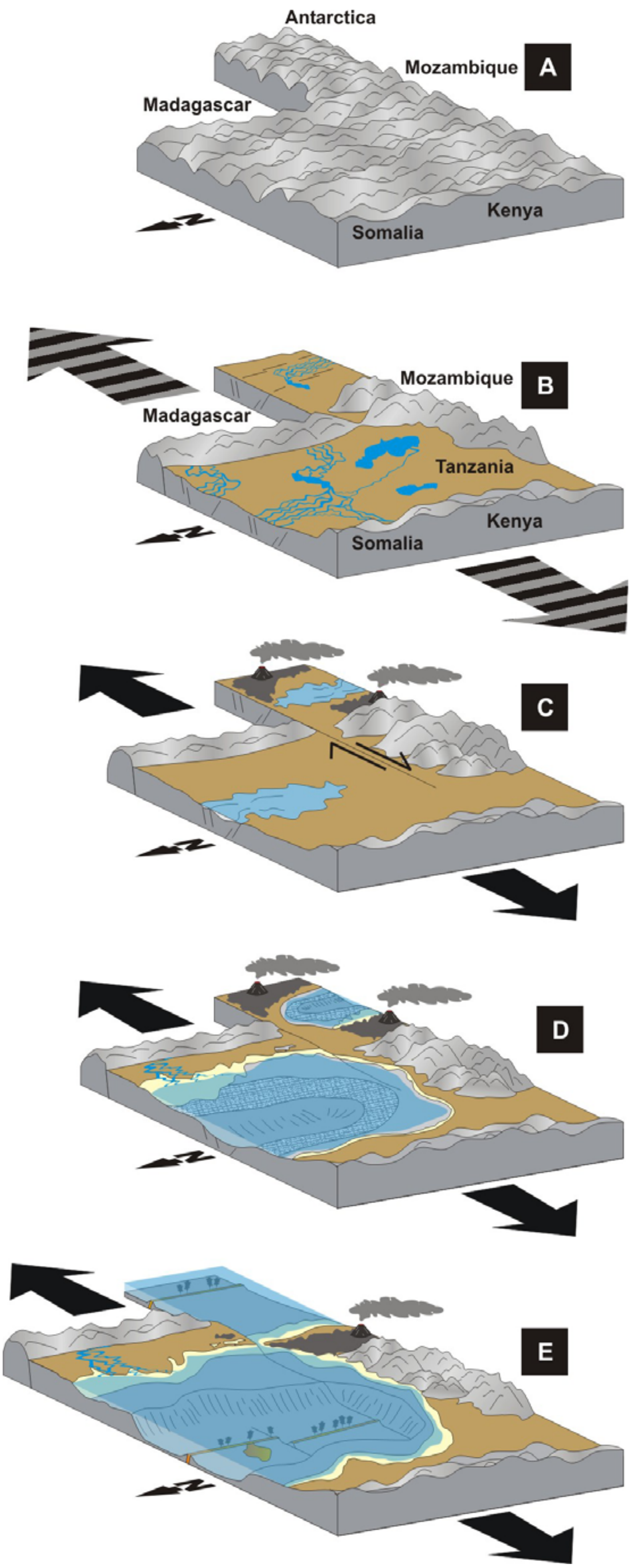
Comparing the new model for basin formation derived from observations at outcrops, well logs, and seismic images with numerical models, the absence of volcanics and rift shoulder uplift suggests a fast and narrow rifting. This avoids the thermal anomalies in the crust/mantle and the production of large amounts of magma. The absence of magma also reflects a typical passive rift situation without magma input from a mantle plume.

The present thesis does not cover all aspects related to the depositional history of the Morondava Basin during the Gondwana Breakup, and several uncertainties remain. For example, the exact stratigraphic age (biozones) of the Isalo, Ankazoabo, and Sakanavaka formations as well as of the "Oxfordian Sandstone" remain unclear. Also, the very restricted availability of precise data from sections across basin strike limits the stratigraphic accuracy of sequence stratigraphic approaches. Furthermore, regional correlation with the East African domain is limited, because sedimentary and stratigraphic concepts, especially for the Jurassic of the coastal basins of Mozambique, Tanzania, and Kenya are poor. With respect to the seaway problem, the location of Jurassic strata in Mozambique mainly in the subsurface hampers palaeobiogeographical studies in the key area. Similarly, tectonic activities along the basin margin during the syn- and post-breakup stages are not entirely understood, as tectonic uplift was too small to be expressed in fission track modelling. Moreover, internal tectonic segmentation parallel to basin strike, as proposed by the palaeoenvironmental studies, remain uncertain. Promising approaches to solve these questions include:

- Improving the biostratigraphic concepts to overcome the present-day lithofacies-oriented stratigraphy on the basis of further ammonite biozonation together with vertebrate and palynomorph biostratigraphy where possible
- Detailed reassessing the lithofacies distribution across basin strike; this basically requires new wells to be drilled
- Modern seismic sections from both the Morondava Basin and the coastal basins of East Africa giving 3D insights on the tectono-sedimentary model to study internal tectonic segmentation of the basin

- Closer estimations of resedimentation processes as response to tectonism along the basin margin can be achieved on the basis of detrital fission track stratigraphy within the syn- and post-breakup successions.

Fig. 51. History of continental extension in the Madagascan segment of Gondwana. **A: Post-East-African-Antarctic Orogeny**: Collapse and erosion of the orogen (Cambrian-Carboniferous); **B: Karoo rifting**: polyphase extension and deposition of thick mainly continental successions (Permian-Triassic); **C: Breakup rifting**: localised, short-lived rift that resulted in the breakup coincided with an initial marine flooding (Toarcian-Aalenian), south of the transform of the future Davie Ridge Fracture Zone volcanism is wide-spread.; **D: Post-rift passive margin formation**: carbonate platform and siliciclastic shoreface deposition (Bajocian-Bathonian); **E: Drifting**: Onset of sea-floor spreading and submarine volcanism (Callovian onwards).



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Erklärung

Hiermit versichere ich, dass ich

1. die Arbeit ohne unerlaubte fremde Hilfe angefertigt habe,

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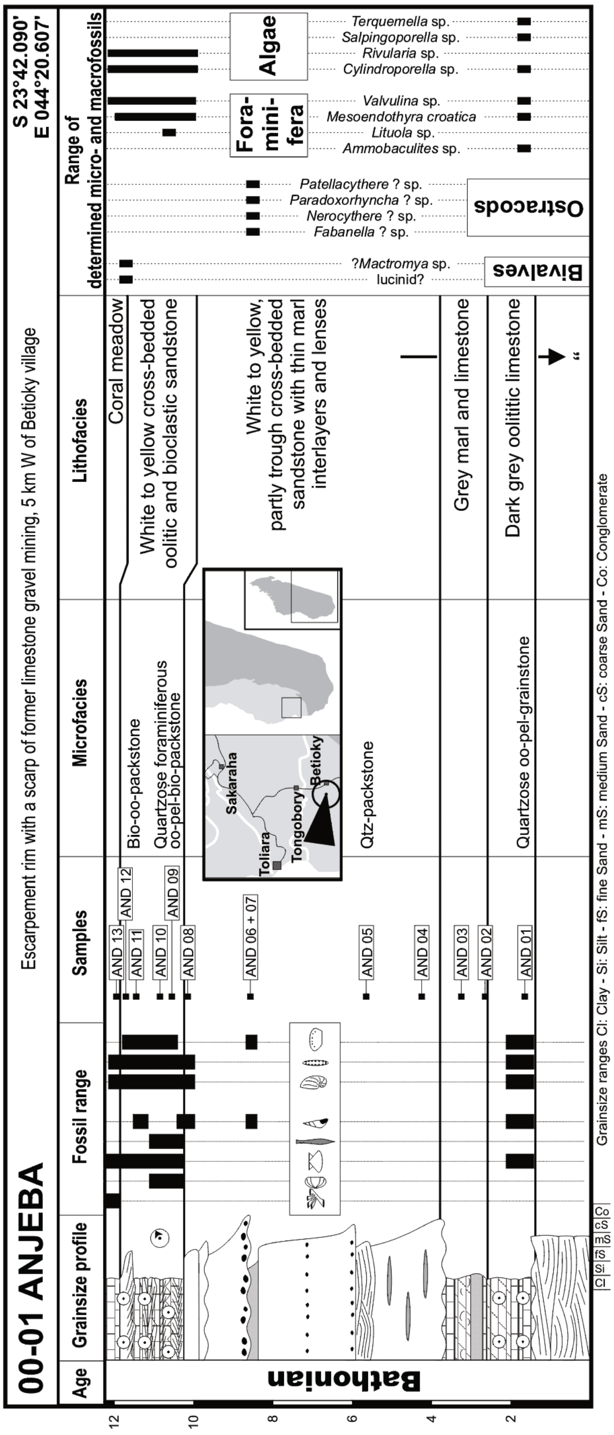
3. die den benutzten Werken wörtlich oder inhaltlich entnommenen Stellen als solche kenntlich gemacht habe.

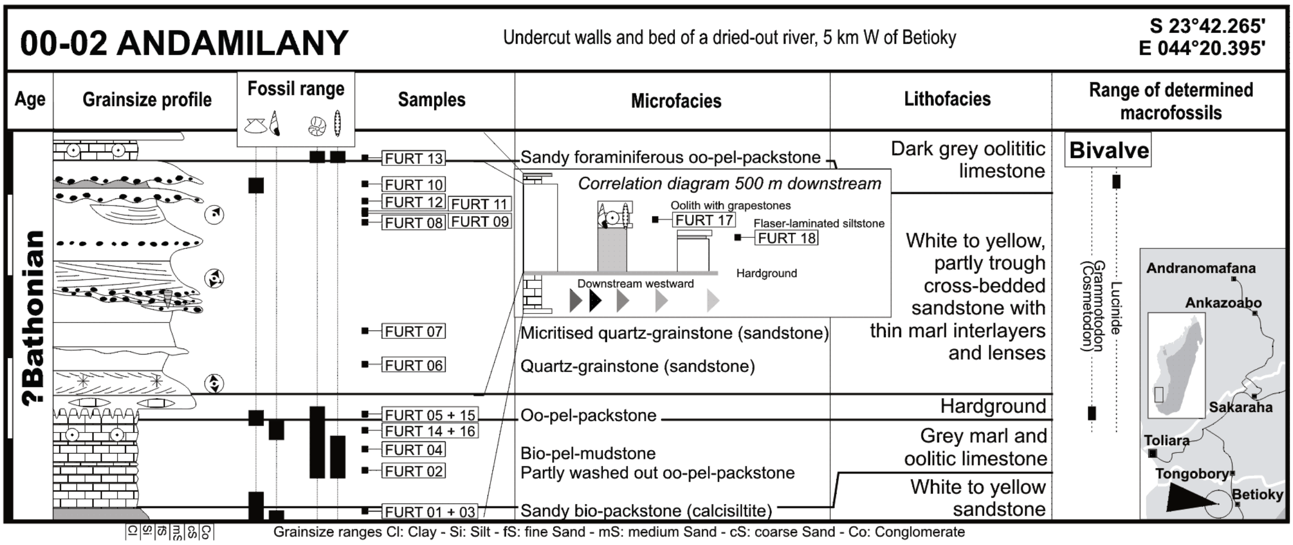
Bremen, den 16. August 2004

Appendix

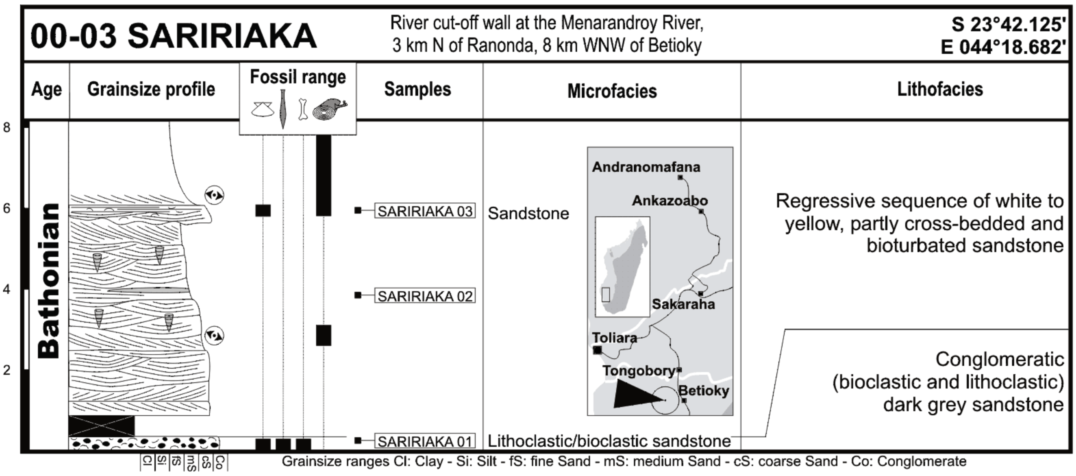
Outcrop sections from the southern Morondava Basin

(For legend see Fig. 13 and Fig. 22)

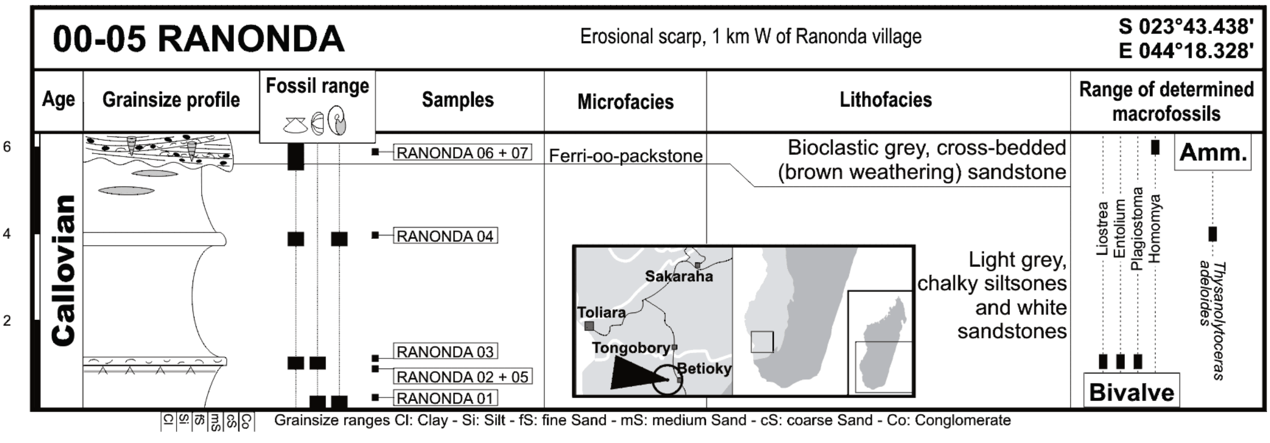




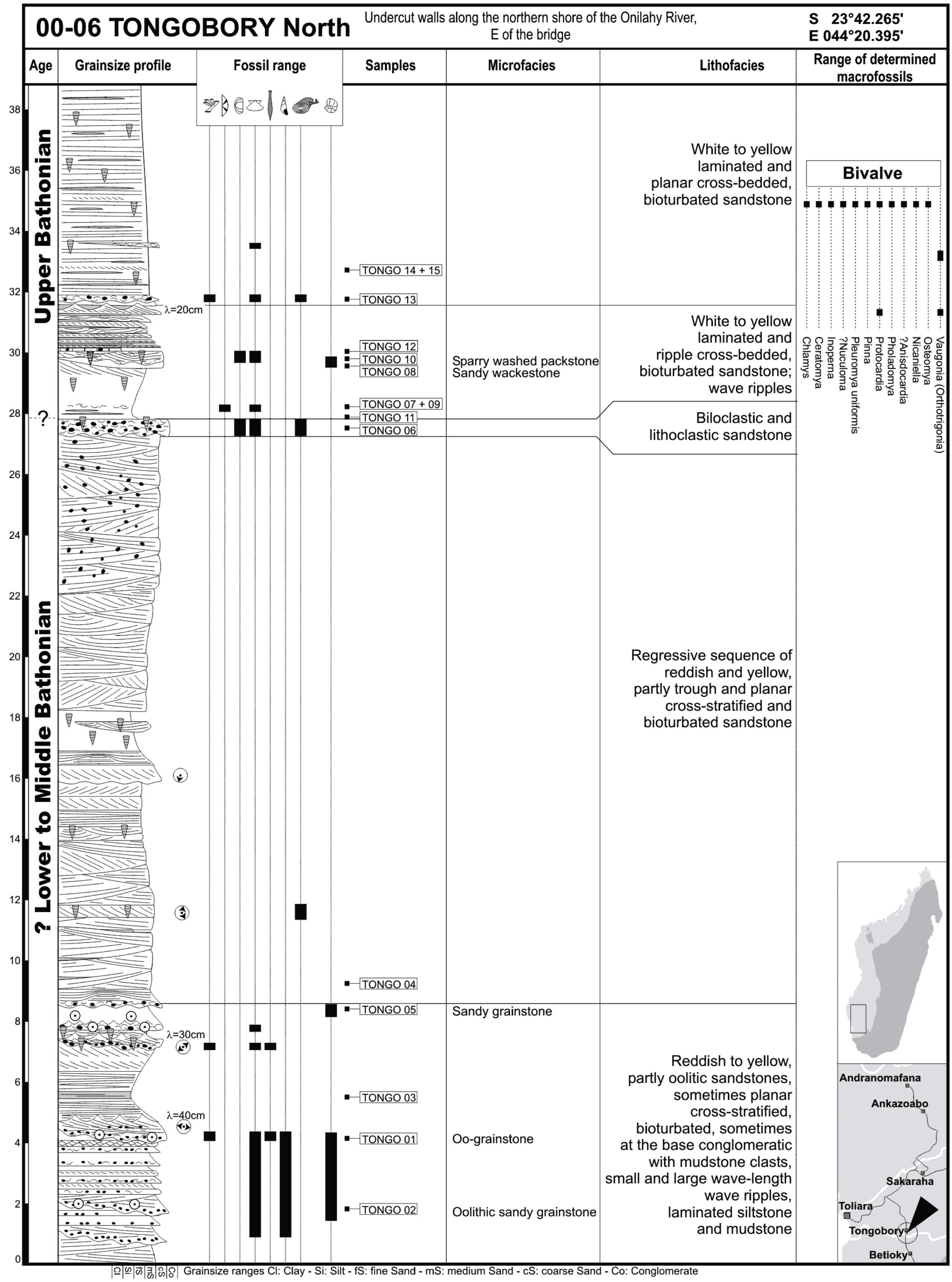
00-02 ANDAMILANY
Undercut walls and bed of a dried-out river, 5 km W of Betioky
S 23°42.265'
E 044°20.395'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined macrofossils
?Bathonian
8
6
4
2
FURT 13
FURT 10
FURT 12
FURT 11
FURT 08
FURT 09
FURT 07
FURT 06
FURT 05 + 15
FURT 14 + 16
FURT 04
FURT 02
FURT 01 + 03
Sandy foraminiferous oo-pel-packstone
Correlation diagram 500 m downstream
Oolith with grapestones
FURT 17
Flaser-laminated siltstone
FURT 18
Hardground
Downstream westward
Micritised quartz-grainstone (sandstone)
Quartz-grainstone (sandstone)
Oo-pel-packstone
Bio-pel-mudstone
Partly washed out oo-pel-packstone
Sandy bio-packstone (calcisiltite)
Dark grey oolititic limestone
White to yellow, partly trough cross-bedded sandstone with thin marl interlayers and lenses
Hardground
Grey marl and oolitic limestone
White to yellow sandstone
Bivalve
Grammatodon (Cosmetodon)
Lucinide
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate



00-03 SARIRIAKA
River cut-off wall at the Menarandroy River,
3 km N of Ranonda, 8 km WNW of Betioky
S 23°42.125'
E 044°18.682'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Bathonian
8
6
4
2
SARIRIAKA 03
SARIRIAKA 02
SARIRIAKA 01
Sandstone
Lithoclastic/bioclastic sandstone
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Regressive sequence of white to yellow, partly cross-bedded and bioturbated sandstone
Conglomeratic (bioclastic and lithoclastic) dark grey sandstone
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

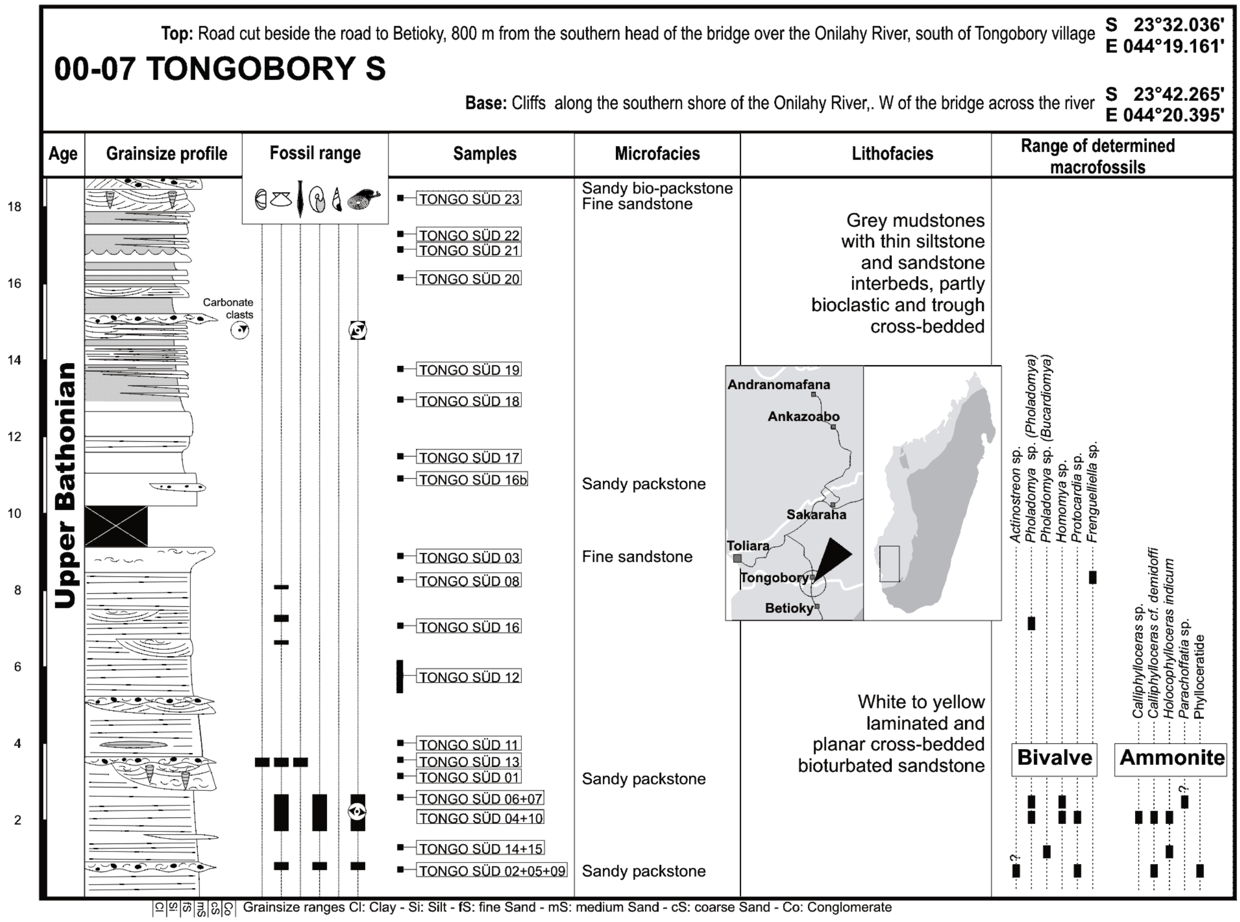


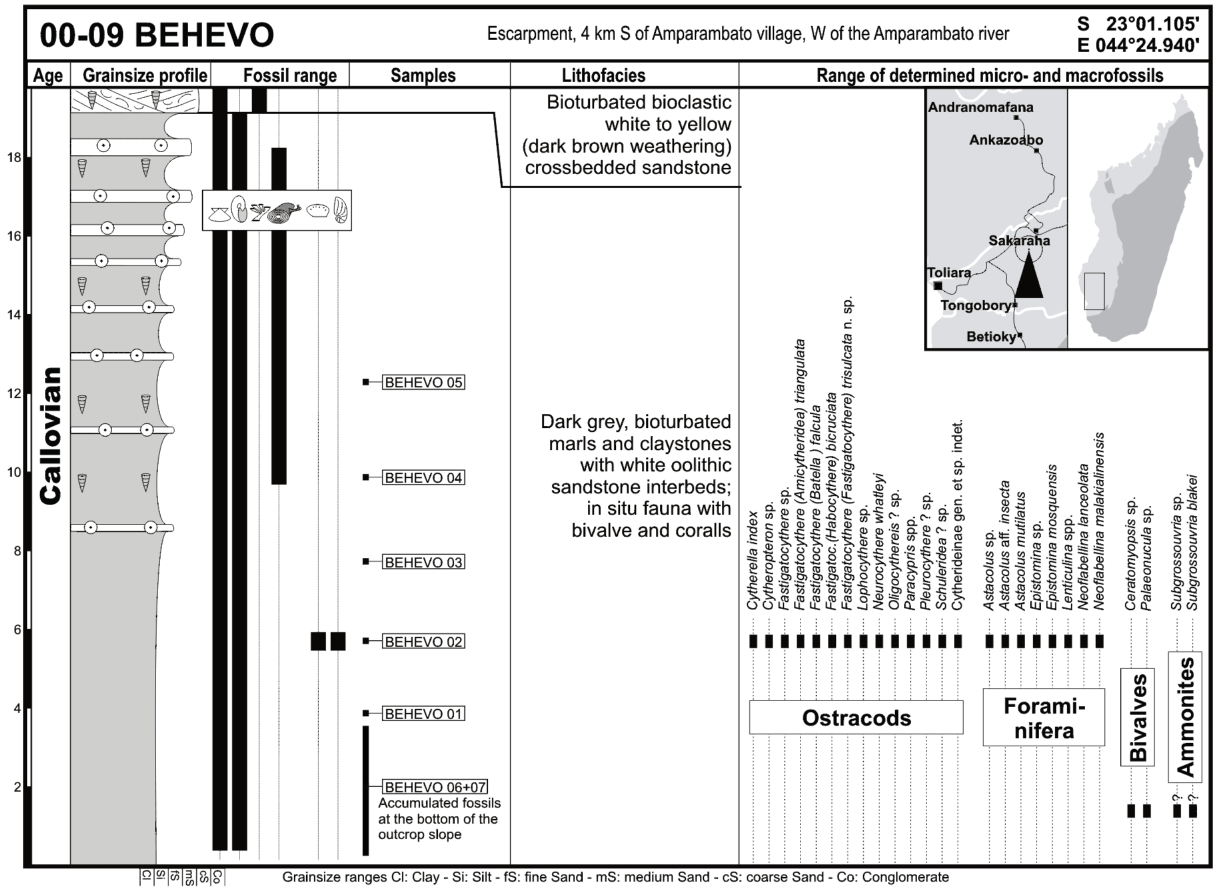
00-05 RANONDA
Erosional scarp, 1 km W of Ranonda village
S 023°43.438'
E 044°18.328'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined macrofossils
Callovian
6
4
2
RANONDA 06 + 07
RANONDA 04
RANONDA 03
RANONDA 02 + 05
RANONDA 01
Ferri-oo-packstone
Bioclastic grey, cross-bedded (brown weathering) sandstone
Sakaraha
Toliara
Tongobory
Betioky
Light grey, chalky siltstones and white sandstones
Liostrea
Entolium
Plagiostoma
Homomya
Amm.
Thysanolytoceras adeloides
Bivalve
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

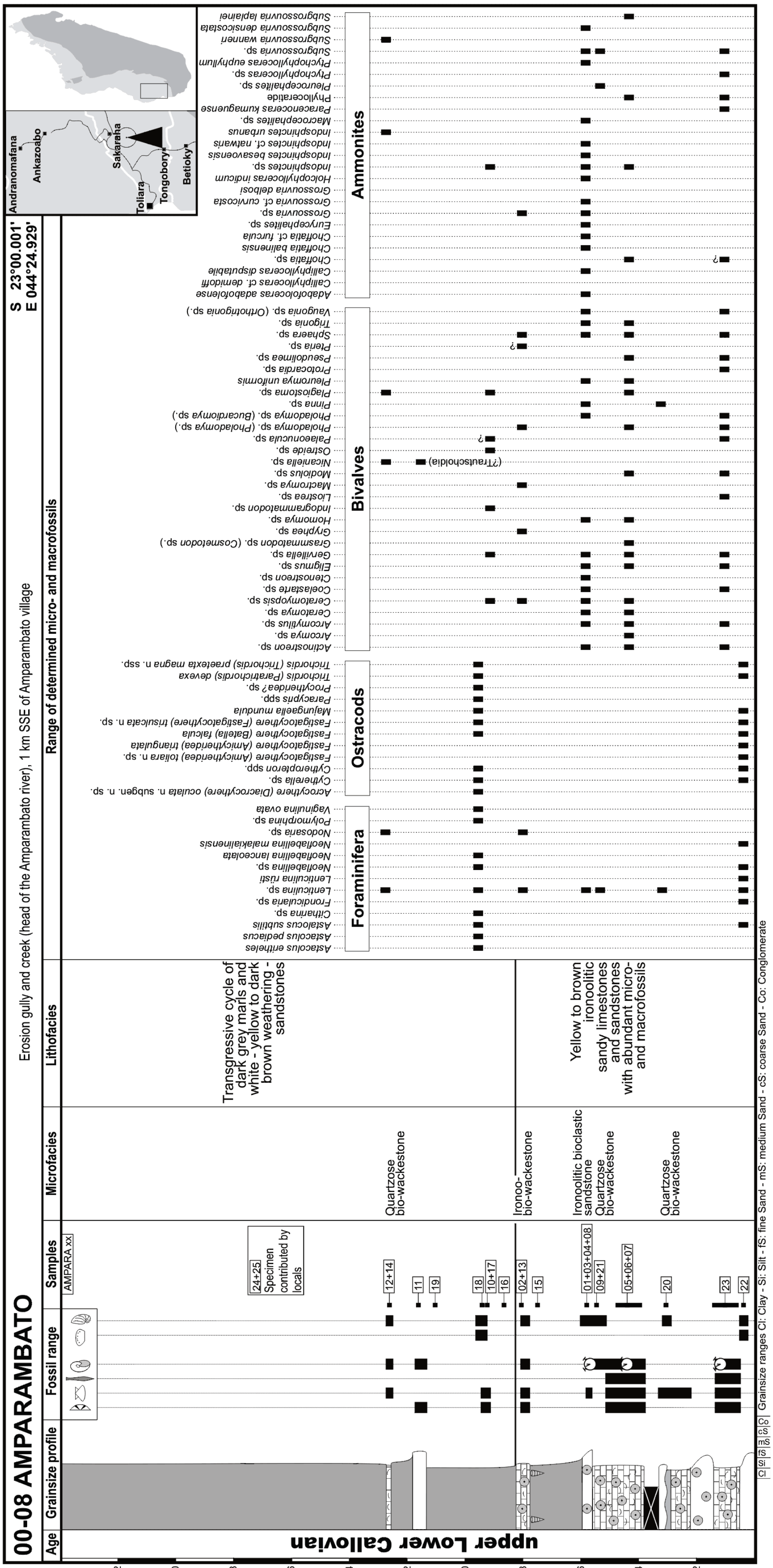


00-06 TONGOBORY North
Undercut walls along the northern shore of the Onilahy River, E of the bridge
S 23°42.265'
E 044°20.395'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined macrofossils
Upper Bathonian
? Lower to Middle Bathonian
λ=20cm
λ=30cm
λ=40cm
TONGO 14 + 15
TONGO 13
TONGO 12
TONGO 10
TONGO 08
TONGO 07 + 09
TONGO 11
TONGO 06
TONGO 04
TONGO 05
TONGO 03
TONGO 01
TONGO 02
Sparry washed packstone
Sandy wackestone
Sandy grainstone
Oo-grainstone
Oolithic sandy grainstone
White to yellow laminated and planar cross-bedded, bioturbated sandstone
White to yellow laminated and ripple cross-bedded, bioturbated sandstone; wave ripples
Bioclastic and lithoclastic sandstone
Regressive sequence of reddish and yellow, partly trough and planar cross-stratified and bioturbated sandstone
Reddish to yellow, partly oolitic sandstones, sometimes planar cross-stratified, bioturbated, sometimes at the base conglomeratic with mudstone clasts, small and large wave-length wave ripples, laminated siltstone and mudstone
Bivalve
Chlamys
Ceratomya
Inoperna
?Nuculoma
Pleuromya uniformis
Pinna
Protocardia
Pholadomya
?Anisocardia
Nicaniella
Osteomya
Vaugonia (Orthotrigonia)
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

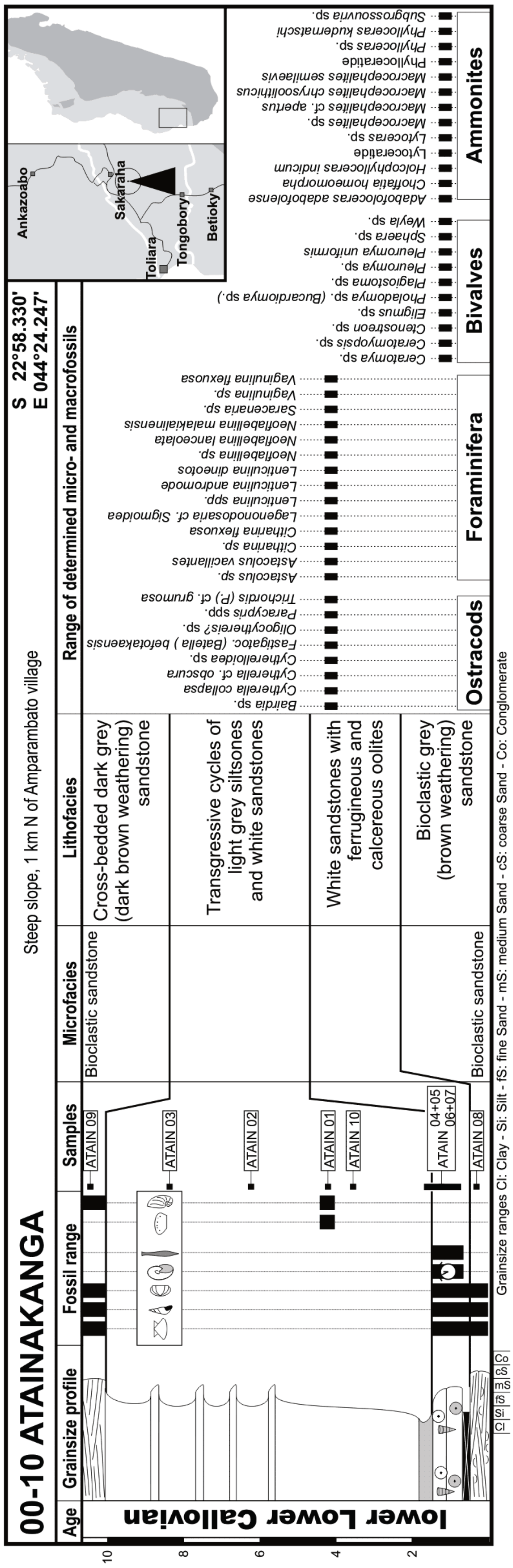
Appendix

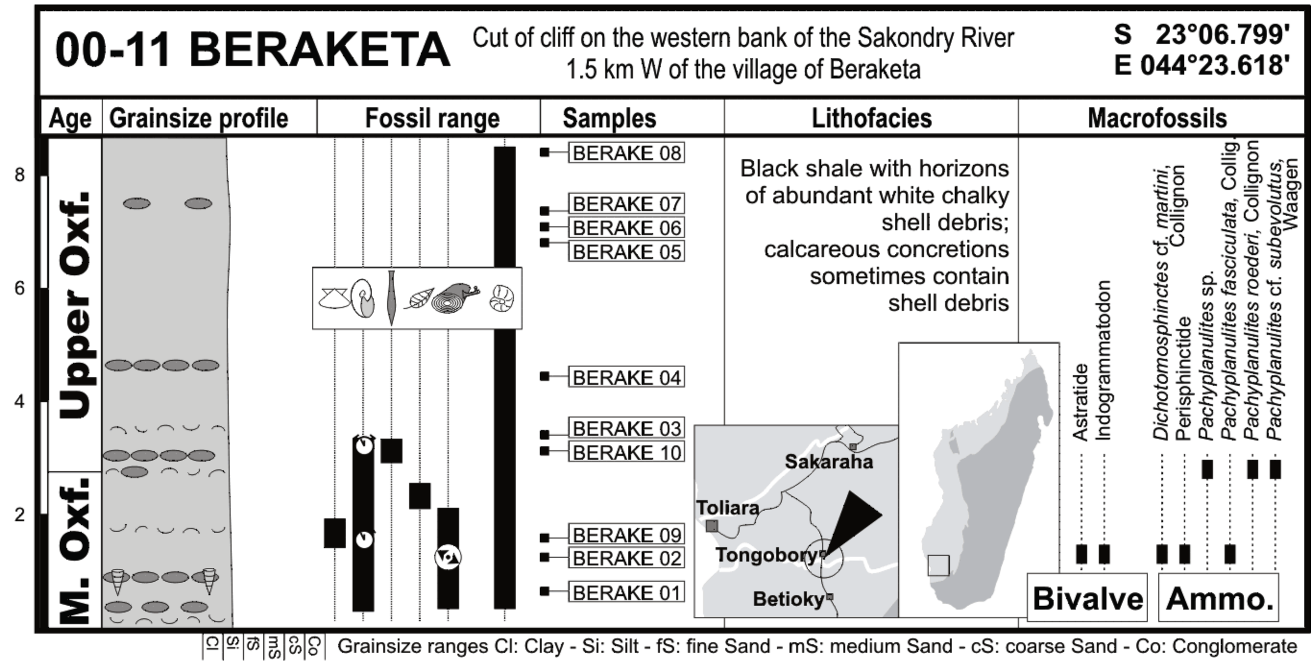

Top: Road cut beside the road to Betioky, 800 m from the southern head of the bridge over the Onilahy River, south of Tongobory village
S 23°32.036' E 044°19.161'
00-07 TONGOBORY S
Base: Cliffs along the southern shore of the Onilahy River,. W of the bridge across the river
S 23°42.265' E 044°20.395'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined macrofossils
Upper Bathonian
Carbonate clasts
TONGO SÜD 23
TONGO SÜD 22
TONGO SÜD 21
TONGO SÜD 20
TONGO SÜD 19
TONGO SÜD 18
TONGO SÜD 17
TONGO SÜD 16b
TONGO SÜD 03
TONGO SÜD 08
TONGO SÜD 16
TONGO SÜD 12
TONGO SÜD 11
TONGO SÜD 13
TONGO SÜD 01
TONGO SÜD 06+07
TONGO SÜD 04+10
TONGO SÜD 14+15
TONGO SÜD 02+05+09
Sandy bio-packstone
Fine sandstone
Sandy packstone
Fine sandstone
Sandy packstone
Sandy packstone
Grey mudstones with thin siltstone and sandstone interbeds, partly bioclastic and trough cross-bedded
White to yellow laminated and planar cross-bedded bioturbated sandstone
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Actinostreon sp.
Pholadomya sp. (Pholadomya)
Pholadomya sp. (Bucardiomya)
Homomya sp.
Protocardia sp.
Frenguelliella sp.
Calliphylloceras sp.
Calliphylloceras cf. demidoffi
Holcophylloceras indicum
Parachoffatia sp.
Phylloceratide
Bivalve
Ammonite
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate


00-09 BEHEVO
Escarpment, 4 km S of Amparambato village, W of the Amparambato river
S 23°01.105' E 044°24.940'
Age
Grainsize profile
Fossil range
Samples
Lithofacies
Range of determined micro- and macrofossils
Callovian
Bioturbated bioclastic white to yellow (dark brown weathering) crossbedded sandstone
Dark grey, bioturbated marls and claystones with white oolithic sandstone interbeds; in situ fauna with bivalve and coralls
BEHEVO 05
BEHEVO 04
BEHEVO 03
BEHEVO 02
BEHEVO 01
BEHEVO 06+07
Accumulated fossils at the bottom of the outcrop slope
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Cytherella index
Cytheropteron sp.
Fastigatocythere sp.
Fastigatocythere (Amicytheridea) triangulata
Fastigatocythere (Batella) falcula
Fastigatoc.(Habocythere) bicruciata
Fastigatocythere (Fastigatocythere) trisulcata n. sp.
Lophocythere sp.
Neurocythere whatleyi
Oligocythereis ? sp.
Paracypris spp.
Pleurocythere ? sp.
Schuleridea ? sp.
Cytherideinae gen. et sp. indet.
Astacolus sp.
Astacolus aff. insecta
Astacolus mutilatus
Epistomina sp.
Epistomina mosquensis
Lenticulina spp.
Neoflabellina lanceolata
Neoflabellina malakialinensis
Ceratomyopsis sp.
Palaeonucula sp.
Subgrossouvria sp.
Subgrossouvria blakei
Ostracods
Forami-nifera
Bivalves
Ammonites
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

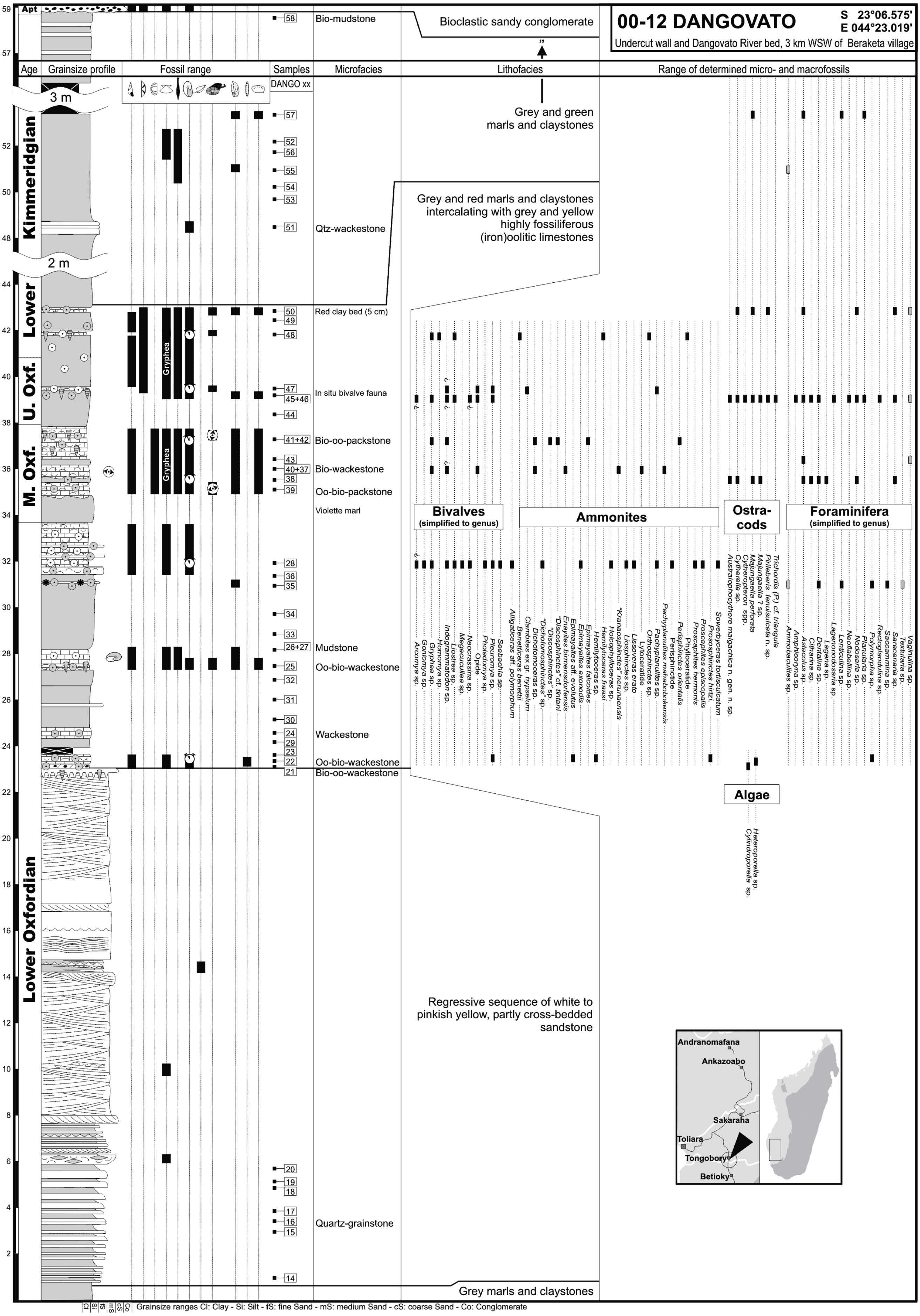


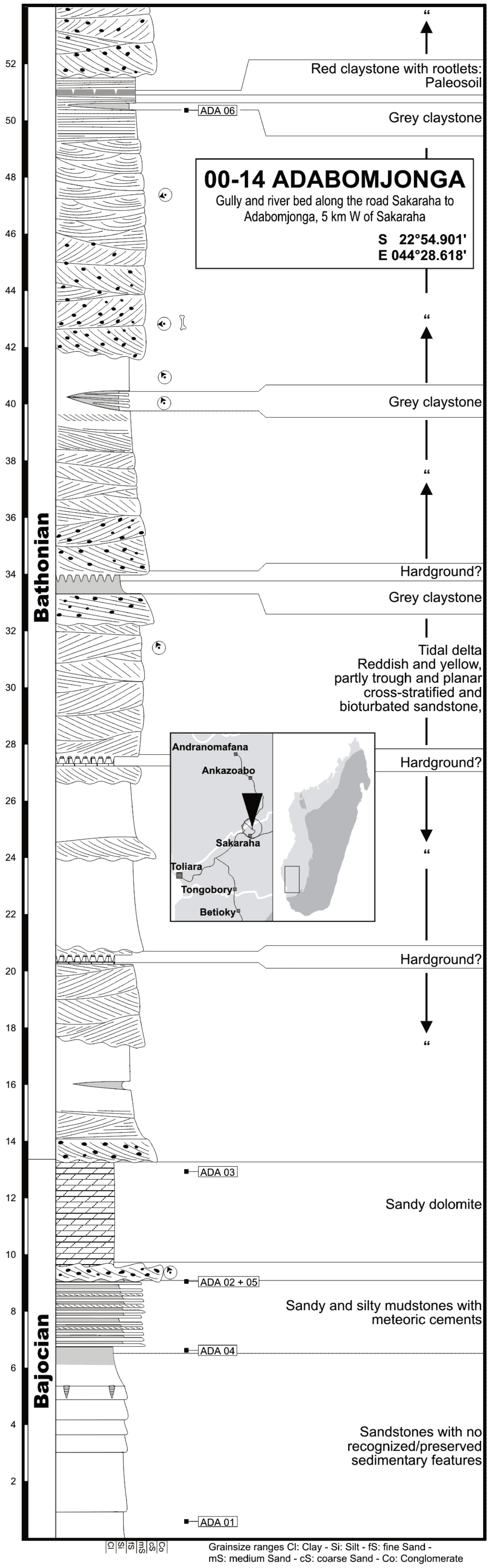
00-08 AMPARAMBATO
Erosion gully and creek (head of the Amparambato river), 1 km SSE of Amparambato village
S 23°00.001'
E 044°24.929'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined micro- and macrofossils
upper Lower Callovian
Foraminifera
Ostracods
Bivalves
Ammonites
Transgressive cycle of dark grey marls and white - yellow to dark brown weathering - sandstones
Yellow to brown ironoolitic sandy limestones and sandstones with abundant micro- and macrofossils
Quartzose bio-wackestone
Ironoolitic bioclastic sandstone
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

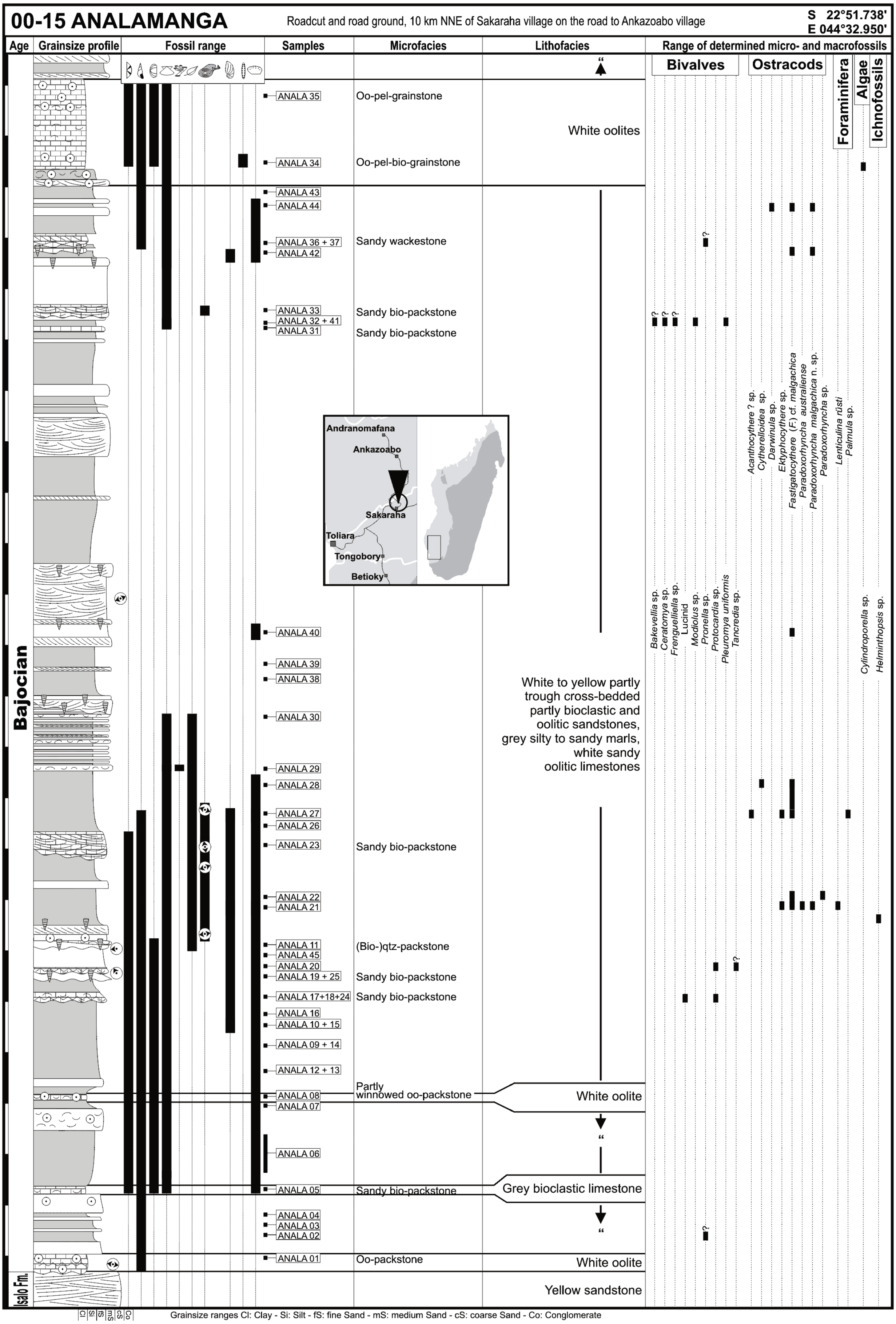

00-10 ATAINAKANGA
Steep slope, 1 km N of Amparambato village
S 22°58.330'
E 044°24.247'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined micro- and macrofossils
lower Lower Callovian
ATAIN 09
ATAIN 03
ATAIN 02
ATAIN 01
ATAIN 10
ATAIN 04+05 06+07
ATAIN 08
Bioclastic sandstone
Bioclastic sandstone
Cross-bedded dark grey (dark brown weathering) sandstone
Transgressive cycles of light grey siltsones and white sandstones
White sandstones with ferrugineous and calcereous oolites
Bioclastic grey (brown weathering) sandstone
Ostracods
Bairdia sp.
Cytherella collapsa
Cytherella cf. obscura
Cytherelloidea sp.
Fastigatoc. (Batella) befotakaensis
Oligocythereis? sp.
Paracypris spp.
Trichordis (P.) cf. grumosa
Foraminifera
Astacolus sp.
Astacolus vacillantes
Citharina sp.
Citharina flexuosa
Lagenonodosaria cf. Sigmoidea
Lenticulina spp.
Lenticulina andromede
Lenticulina dineotos
Neoflabellina sp.
Neoflabellina lanceolata
Neoflabellina malakialinensis
Saracenaria sp.
Vaginulina sp.
Vaginulina flexuosa
Bivalves
Ceratomya sp.
Ceratomyopsis sp.
Ctenosteon sp.
Eligmus sp.
Pholadomya sp. (Bucardiomya sp.)
Plagiostoma sp.
Pleuromya sp.
Pleuromya uniformis
Sphaera sp.
Weyla sp.
Ammonites
Adabofoloceras adabofolense
Choffatia homeomorpha
Holcophylloceras indicum
Lytoceratide
Lytoceras sp.
Macrocephalites sp.
Macrocephalites cf. apertus
Macrocephalites chrysoolithicus
Macrocephalites semilaevis
Phylloceratide
Phylloceras sp.
Phylloceras kudernatschi
Subgrossouvria sp.
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate
10
8
6
4
2

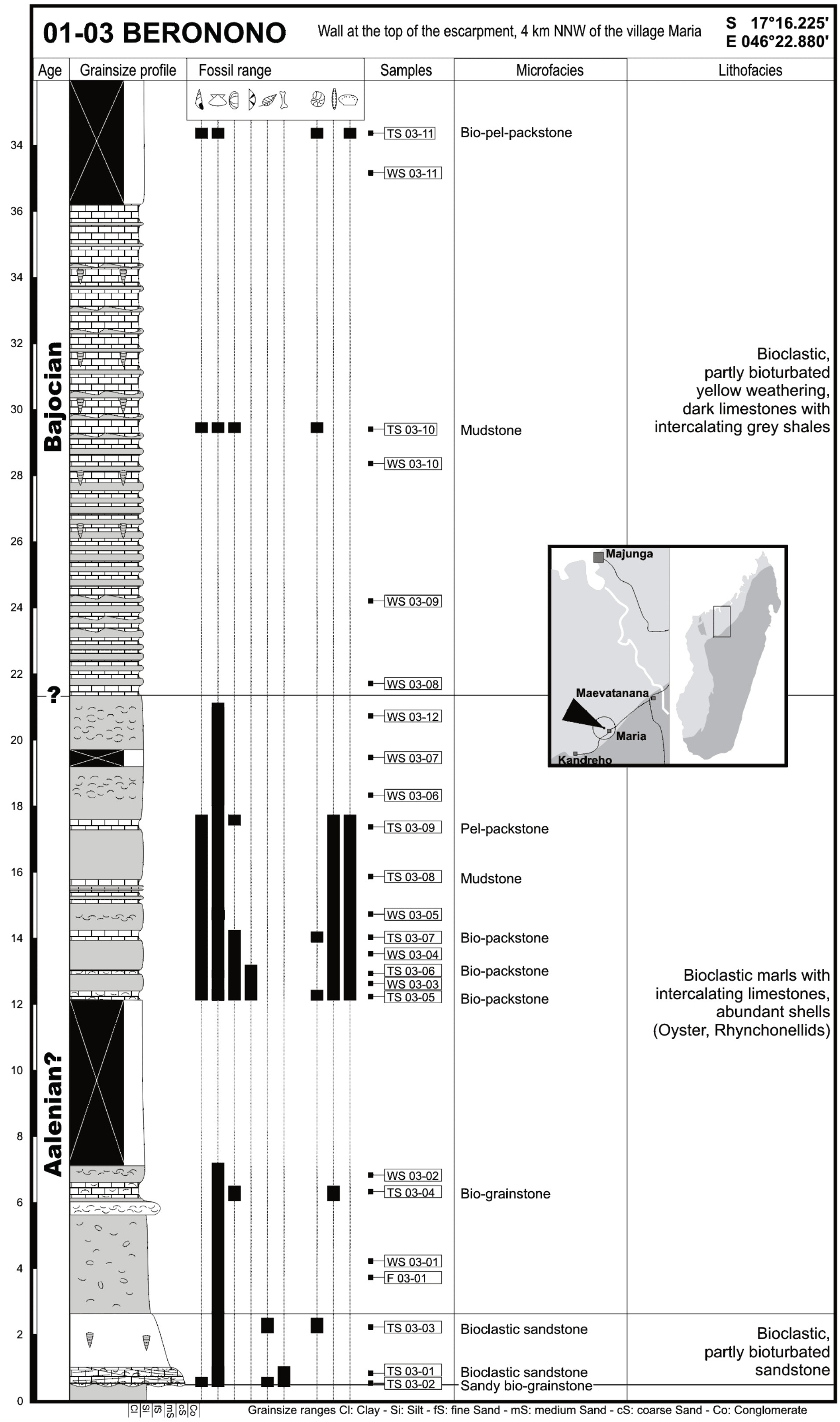


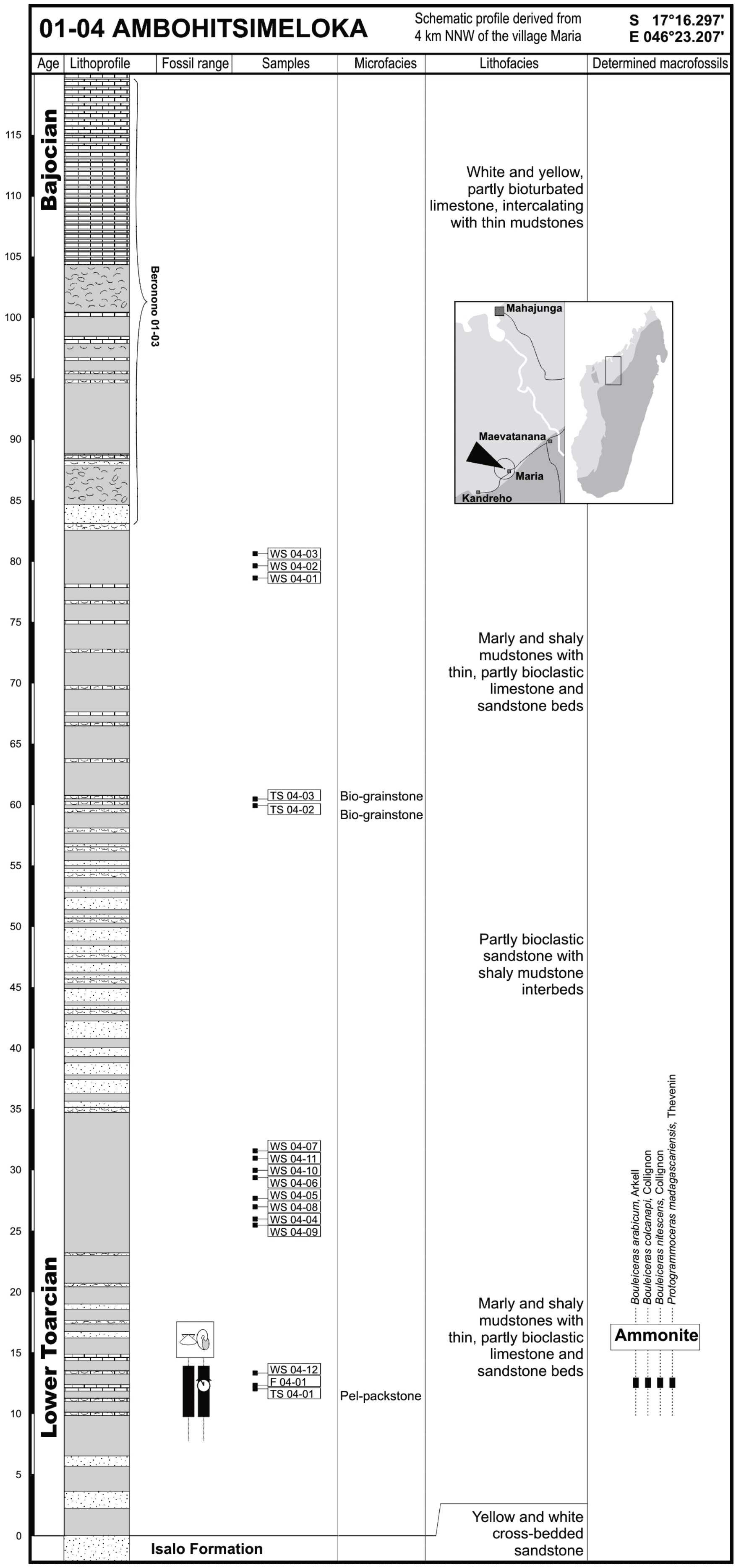
00-11 BERAKETA
Cut of cliff on the western bank of the Sakondry River 1.5 km W of the village of Beraketa
S 23°06.799'
E 044°23.618'
Age
Grainsize profile
Fossil range
Samples
Lithofacies
Macrofossils
Upper Oxf.
M. Oxf.
8
6
4
2
BERAKE 08
BERAKE 07
BERAKE 06
BERAKE 05
BERAKE 04
BERAKE 03
BERAKE 10
BERAKE 09
BERAKE 02
BERAKE 01
Black shale with horizons of abundant white chalky shell debris; calcareous concretions sometimes contain shell debris
Sakaraha
Toliara
Tongobory
Betioky
Astratide
Indogrammatodon
Dichotomosphinctes cf. martini, Collignon
Perisphinctide
Pachyplanulites sp.
Pachyplanulites fasciculata, Collig.
Pachyplanulites roederi, Collignon
Pachyplanulites cf. subevolutus, Waagen
Bivalve
Ammo.
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate


00-12 DANGOVATO
S 23°06.575'
E 044°23.019'
Undercut wall and Dangovato River bed, 3 km WSW of Beraketa village
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined micro- and macrofossils
DANGO xx
Apt
Kimmeridgian
Lower
U. Oxf.
M. Oxf.
Lower Oxfordian
3 m
2 m
Gryphea
Bio-mudstone
Bioclastic sandy conglomerate
Grey and green marls and claystones
Grey and red marls and claystones intercalating with grey and yellow highly fossiliferous (iron)oolitic limestones
Qtz-wackestone
Red clay bed (5 cm)
In situ bivalve fauna
Bio-oo-packstone
Bio-wackestone
Oo-bio-packstone
Violette marl
Mudstone
Oo-bio-wackestone
Wackestone
Oo-bio-wackestone
Bio-oo-wackestone
Regressive sequence of white to pinkish yellow, partly cross-bedded sandstone
Quartz-grainstone
Grey marls and claystones
Bivalves (simplified to genus)
Ammonites
Ostra-cods
Foraminifera (simplified to genus)
Algae
Heteroporella sp.
Cylindroporella sp.
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

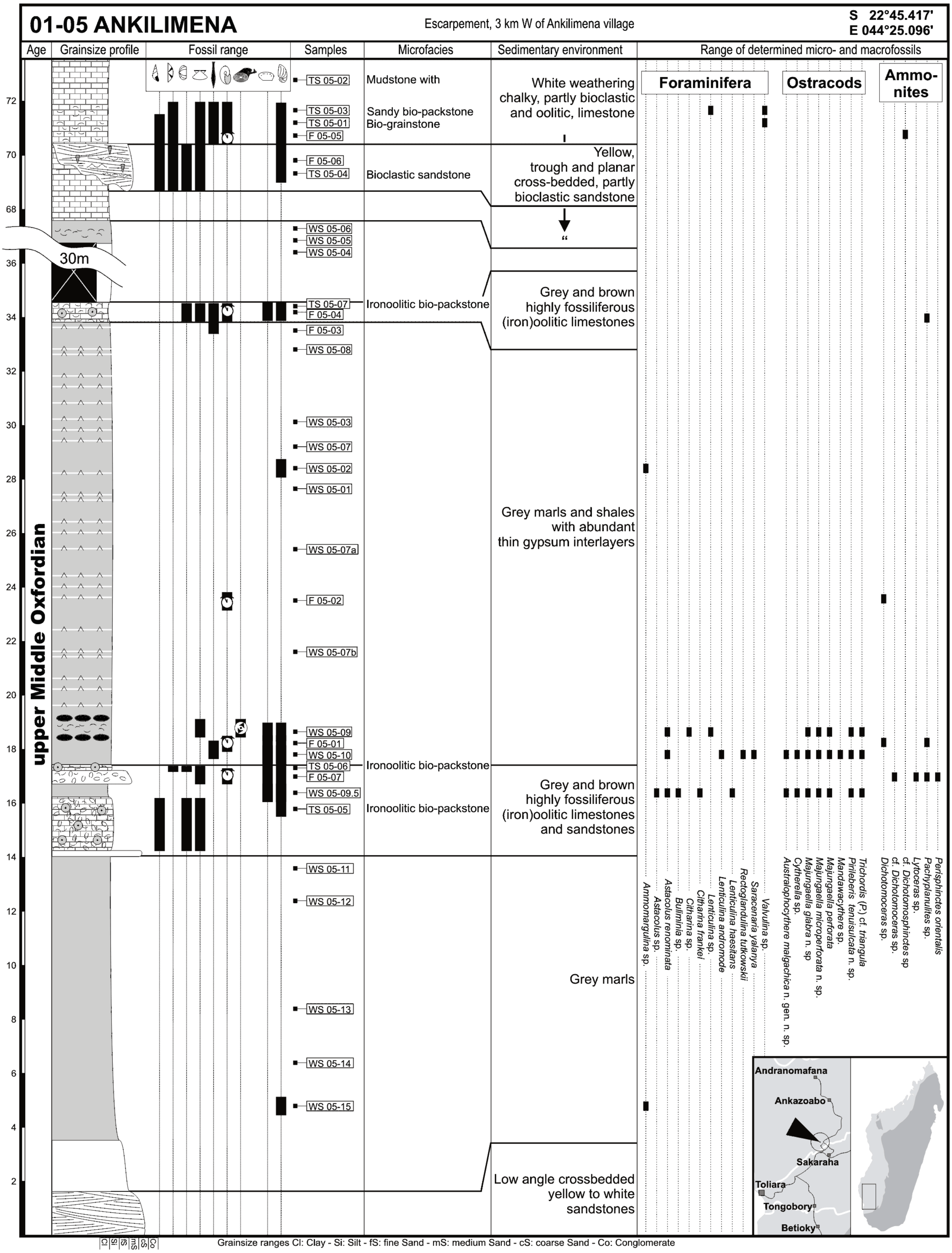

Red claystone with rootlets: Paleosoil
ADA 06
Grey claystone
00-14 ADABOMJONGA
Gully and river bed along the road Sakaraha to Adabomjonga, 5 km W of Sakaraha
S 22°54.901'
E 044°28.618'
Grey claystone
Bathonian
Hardground?
Grey claystone
Tidal delta
Reddish and yellow, partly trough and planar cross-stratified and bioturbated sandstone,
Andranomafana
Ankazoabo
Hardground?
Sakaraha
Toliara
Tongobory
Betioky
Hardground?
ADA 03
Sandy dolomite
ADA 02 + 05
Sandy and silty mudstones with meteoric cements
ADA 04
Bajocian
Sandstones with no recognized/preserved sedimentary features
ADA 01
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

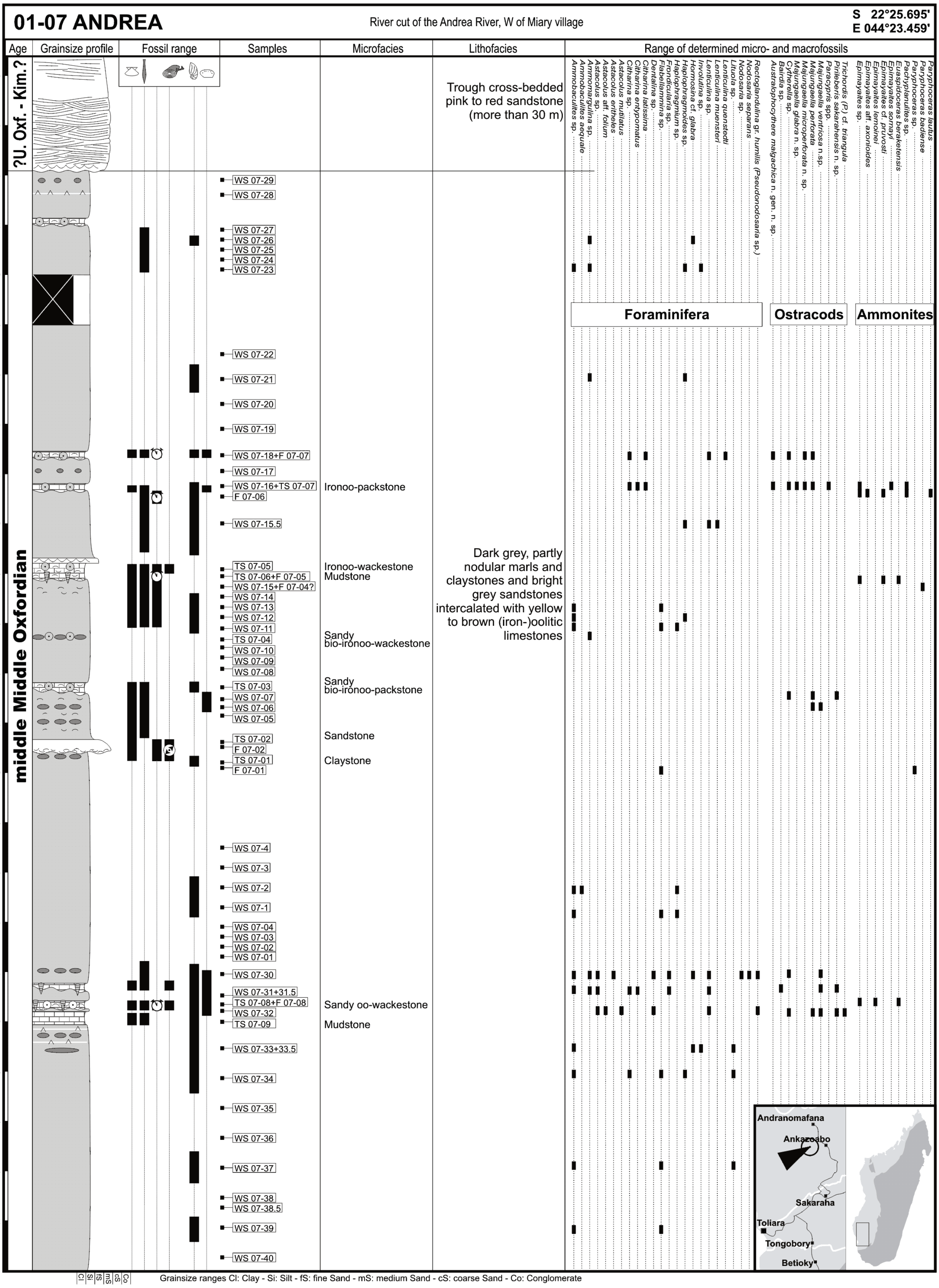

00-15 ANALAMANGA
Roadcut and road ground, 10 km NNE of Sakaraha village on the road to Ankazoabo village
S 22°51.738'
E 044°32.950'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined micro- and macrofossils
Bivalves
Ostracods
Foraminifera
Algae
Ichnofossils
Bajocian
Isalo Fm.
ANALA 35
Oo-pel-grainstone
White oolites
ANALA 34
Oo-pel-bio-grainstone
ANALA 43
ANALA 44
ANALA 36 + 37
Sandy wackestone
ANALA 42
ANALA 33
Sandy bio-packstone
ANALA 32 + 41
ANALA 31
Sandy bio-packstone
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
ANALA 40
ANALA 39
ANALA 38
White to yellow partly trough cross-bedded partly bioclastic and oolitic sandstones, grey silty to sandy marls, white sandy oolitic limestones
ANALA 30
ANALA 29
ANALA 28
ANALA 27
ANALA 26
ANALA 23
Sandy bio-packstone
ANALA 22
ANALA 21
ANALA 11
(Bio-)qtz-packstone
ANALA 45
ANALA 20
ANALA 19 + 25
Sandy bio-packstone
ANALA 17+18+24
Sandy bio-packstone
ANALA 16
ANALA 10 + 15
ANALA 09 + 14
ANALA 12 + 13
Partly winnowed oo-packstone
White oolite
ANALA 08
ANALA 07
ANALA 06
ANALA 05
Sandy bio-packstone
Grey bioclastic limestone
ANALA 04
ANALA 03
ANALA 02
ANALA 01
Oo-packstone
White oolite
Yellow sandstone
Bakevellia sp.
Ceratomya sp.
Frenguelliella sp.
Lucinid
Modiolus sp.
Pronella sp.
Protocardia sp.
Pleuromya uniformis
Tancredia sp.
Acanthocythere ? sp.
Cytherelloidea sp.
Darwinula sp.
Ektyphocythere sp.
Fastigatocythere (F.) cf. malgachica
Paradoxorhyncha australiense
Paradoxorhyncha malgachica n. sp.
Paradoxorhyncha sp.
Lenticulina rüsti
Palmula sp.
Cylindroporella sp.
Helminthopsis sp.
48
46
44
42
40
38
36
34
32
30
28
26
24
22
20
18
16
14
12
10
8
6
4
2
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

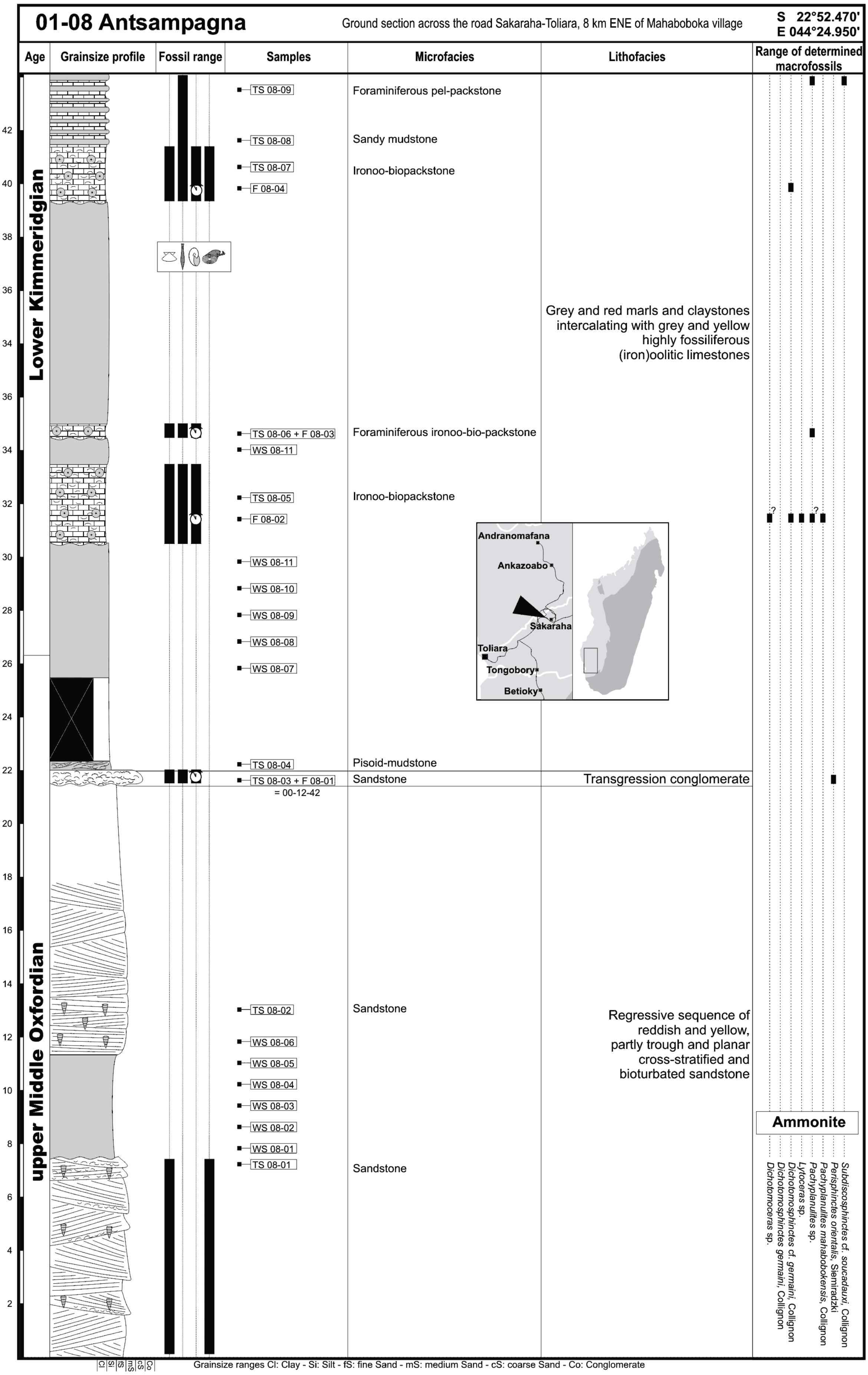

01-03 BERONONO
Wall at the top of the escarpment, 4 km NNW of the village Maria
S 17°16.225'
E 046°22.880'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Bajocian
Aalenian?
TS 03-11
WS 03-11
Bio-pel-packstone
TS 03-10
Mudstone
WS 03-10
WS 03-09
WS 03-08
Bioclastic, partly bioturbated yellow weathering, dark limestones with intercalating grey shales
WS 03-12
WS 03-07
WS 03-06
TS 03-09
Pel-packstone
TS 03-08
Mudstone
WS 03-05
TS 03-07
Bio-packstone
WS 03-04
TS 03-06
Bio-packstone
WS 03-03
TS 03-05
Bio-packstone
Bioclastic marls with intercalating limestones, abundant shells (Oyster, Rhynchonellids)
WS 03-02
TS 03-04
Bio-grainstone
WS 03-01
F 03-01
TS 03-03
Bioclastic sandstone
TS 03-01
Bioclastic sandstone
TS 03-02
Sandy bio-grainstone
Bioclastic, partly bioturbated sandstone
Majunga
Maevatanana
Maria
Kandreho
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

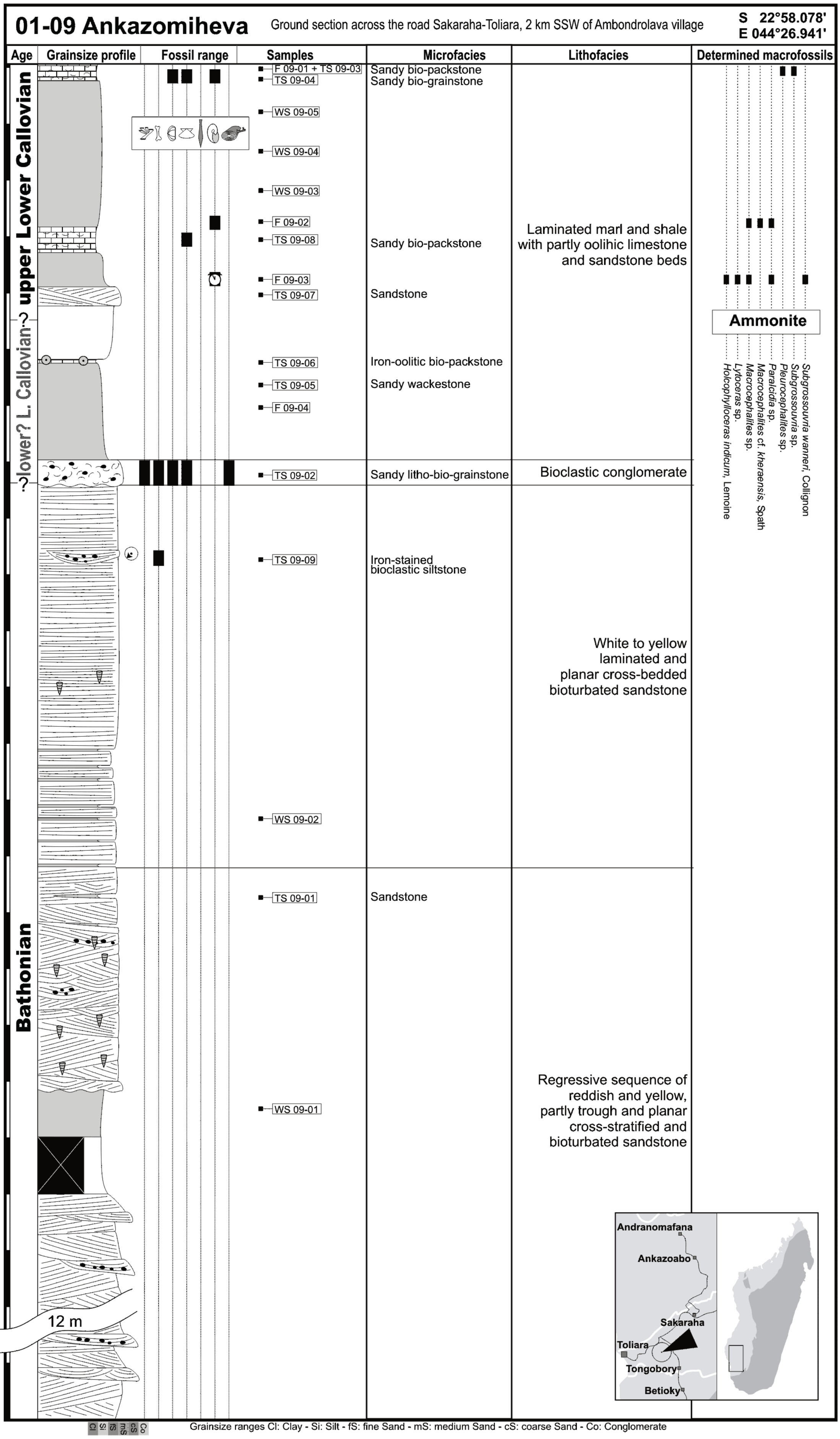


01-04 AMBOHITSIMELOKA
Schematic profile derived from 4 km NNW of the village Maria
S 17°16.297'
E 046°23.207'
Age
Lithoprofile
Fossil range
Samples
Microfacies
Lithofacies
Determined macrofossils
Bajocian
Lower Toarcian
Beronono 01-03
Isalo Formation
WS 04-03
WS 04-02
WS 04-01
TS 04-03
TS 04-02
Bio-grainstone
Bio-grainstone
WS 04-07
WS 04-11
WS 04-10
WS 04-06
WS 04-05
WS 04-08
WS 04-04
WS 04-09
WS 04-12
F 04-01
TS 04-01
Pel-packstone
White and yellow, partly bioturbated limestone, intercalating with thin mudstones
Marly and shaly mudstones with thin, partly bioclastic limestone and sandstone beds
Partly bioclastic sandstone with shaly mudstone interbeds
Marly and shaly mudstones with thin, partly bioclastic limestone and sandstone beds
Yellow and white cross-bedded sandstone
Mahajunga
Maevatanana
Maria
Kandreho
Bouleiceras arabicum, Arkell
Bouleiceras colcanapi, Collignon
Bouleiceras nitescens, Collignon
Protogrammoceras madagascariensis, Thevenin
Ammonite


01-05 ANKILIMENA
Escarpement, 3 km W of Ankilimena village
S 22°45.417'
E 044°25.096'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Sedimentary environment
Range of determined micro- and macrofossils
upper Middle Oxfordian
30m
TS 05-02
TS 05-03
TS 05-01
F 05-05
F 05-06
TS 05-04
WS 05-06
WS 05-05
WS 05-04
TS 05-07
F 05-04
F 05-03
WS 05-08
WS 05-03
WS 05-07
WS 05-02
WS 05-01
WS 05-07a
F 05-02
WS 05-07b
WS 05-09
F 05-01
WS 05-10
TS 05-06
F 05-07
WS 05-09.5
TS 05-05
WS 05-11
WS 05-12
WS 05-13
WS 05-14
WS 05-15
Mudstone with
Sandy bio-packstone
Bio-grainstone
Bioclastic sandstone
Ironoolitic bio-packstone
Ironoolitic bio-packstone
Ironoolitic bio-packstone
White weathering chalky, partly bioclastic and oolitic, limestone
Yellow, trough and planar cross-bedded, partly bioclastic sandstone
Grey and brown highly fossiliferous (iron)oolitic limestones
Grey marls and shales with abundant thin gypsum interlayers
Grey and brown highly fossiliferous (iron)oolitic limestones and sandstones
Grey marls
Low angle crossbedded yellow to white sandstones
Foraminifera
Ostracods
Ammo-nites
Ammomarginulina sp.
Astacolus sp.
Astacolus renominata
Bulimina sp.
Citharina sp.
Citharina frankei
Lenticulina sp.
Lenticulina andromede
Lenticulina haesitans
Rectoglandulina tutkowskii
Saracenaria yalanya
Valvulina sp.
Australoicythere malgachica n. gen. n. sp.
Cytherella sp.
Majungaella glabra n. sp
Majungaella microperforata n. sp.
Majungaella perforata
Mandawacythere sp.
Pirileberis tenuisulcata n. sp.
Trichordis (P.) cf. triangula
Dichotomoceras sp.
cf. Dichotomoceras sp.
cf. Dichotomosphinctes sp
Lytoceras sp.
Pachyplanulites sp.
Perisphinctes orientalis
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

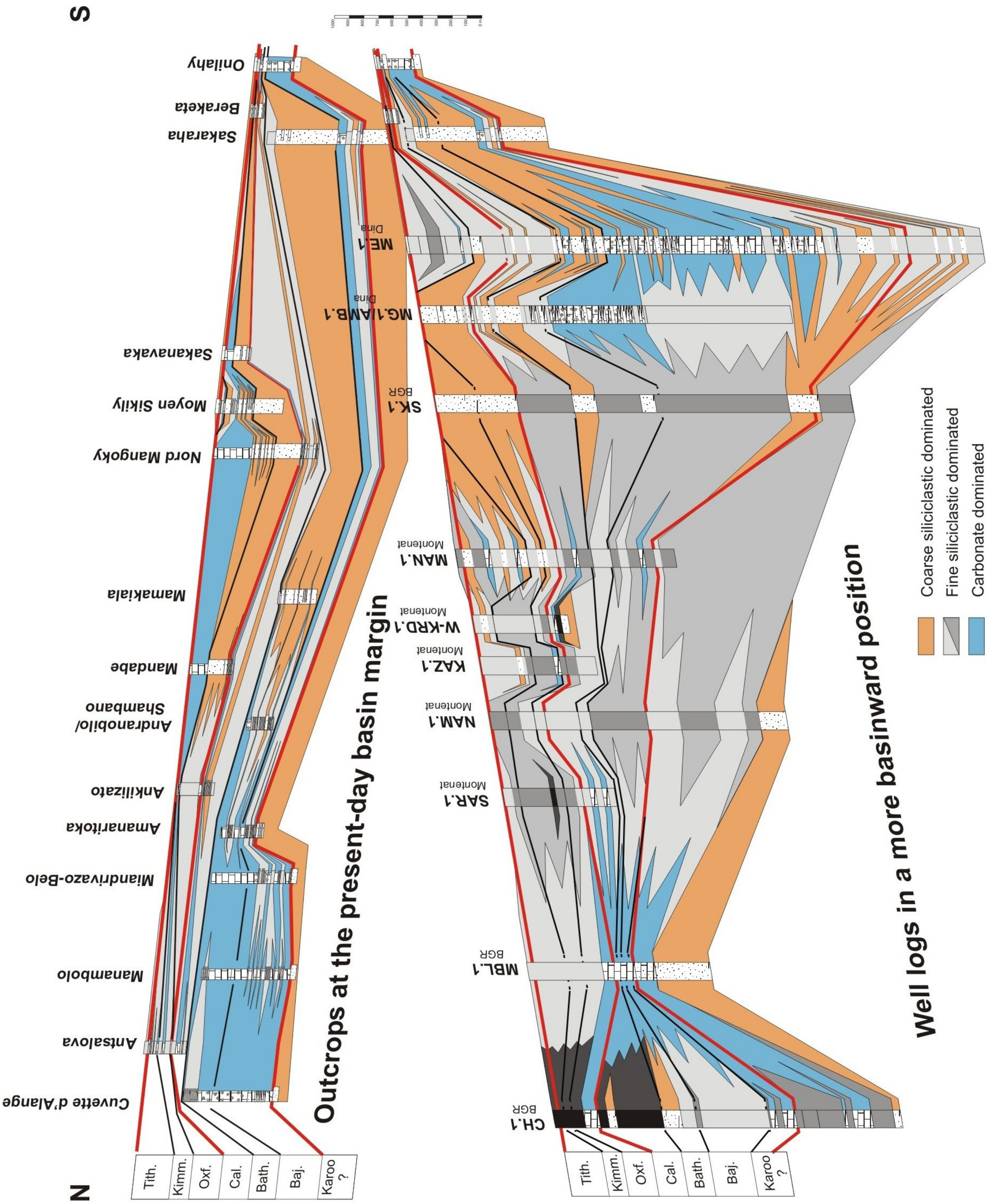

01-07 ANDREA
River cut of the Andrea River, W of Miary village
S 22°25.695'
E 044°23.459'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined micro- and macrofossils
?U. Oxf. - Kim.?
middle Middle Oxfordian
Trough cross-bedded pink to red sandstone (more than 30 m)
Dark grey, partly nodular marls and claystones and bright grey sandstones intercalated with yellow to brown (iron-)oolitic limestones
Ironoo-packstone
Ironoo-wackestone
Mudstone
Sandy bio-ironoo-wackestone
Sandy bio-ironoo-packstone
Sandstone
Claystone
Sandy oo-wackestone
Mudstone
Foraminifera
Ostracods
Ammonites
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate


01-08 Antsampagna
Ground section across the road Sakaraha-Toliara, 8 km ENE of Mahaboboka village
S 22°52.470'
E 044°24.950'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Range of determined macrofossils
Lower Kimmeridgian
upper Middle Oxfordian
TS 08-09
Foraminiferous pel-packstone
TS 08-08
Sandy mudstone
TS 08-07
Ironoo-biopackstone
F 08-04
Grey and red marls and claystones intercalating with grey and yellow highly fossiliferous (iron)oolitic limestones
TS 08-06 + F 08-03
Foraminiferous ironoo-bio-packstone
WS 08-11
TS 08-05
Ironoo-biopackstone
F 08-02
WS 08-11
WS 08-10
WS 08-09
WS 08-08
WS 08-07
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
TS 08-04
Pisoid-mudstone
TS 08-03 + F 08-01
= 00-12-42
Sandstone
Transgression conglomerate
TS 08-02
Sandstone
Regressive sequence of reddish and yellow, partly trough and planar cross-stratified and bioturbated sandstone
WS 08-06
WS 08-05
WS 08-04
WS 08-03
WS 08-02
WS 08-01
TS 08-01
Sandstone
Ammonite
Subdiscosphinctes cf. soucadauxi, Collignon
Perisphinctes orientalis, Siemiradzki
Pachyplanulites mahabobokensis, Collignon
Pachyplanulites sp.
Lytoceras sp.
Dichotomosphinctes cf. germaini, Collignon
Dichotomosphinctes germaini, Collignon
Dichotomoceras sp.
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate


01-09 Ankazomiheva
Ground section across the road Sakaraha-Toliara, 2 km SSW of Ambondrolava village
S 22°58.078'
E 044°26.941'
Age
Grainsize profile
Fossil range
Samples
Microfacies
Lithofacies
Determined macrofossils
upper Lower Callovian
?lower? L. Callovian?
Bathonian
F 09-01 + TS 09-03
TS 09-04
WS 09-05
WS 09-04
WS 09-03
F 09-02
TS 09-08
F 09-03
TS 09-07
TS 09-06
TS 09-05
F 09-04
TS 09-02
TS 09-09
WS 09-02
TS 09-01
WS 09-01
Sandy bio-packstone
Sandy bio-grainstone
Sandy bio-packstone
Sandstone
Iron-oolitic bio-packstone
Sandy wackestone
Sandy litho-bio-grainstone
Iron-stained bioclastic siltstone
Sandstone
Laminated marl and shale with partly oolihic limestone and sandstone beds
Bioclastic conglomerate
White to yellow laminated and planar cross-bedded bioturbated sandstone
Regressive sequence of reddish and yellow, partly trough and planar cross-stratified and bioturbated sandstone
Ammonite
Holcophylloceras indicum, Lemoine
Lytoceras sp.
Macrocephalites sp.
Macrocephalites cf. kheraensis, Spath
Paralcidia sp.
Pleurocephalites sp.
Subgrossouvria sp.
Subgrossouvria wanneri, Collignon
12 m
Andranomafana
Ankazoabo
Sakaraha
Toliara
Tongobory
Betioky
Cl Si fS mS cS Co
Grainsize ranges Cl: Clay - Si: Silt - fS: fine Sand - mS: medium Sand - cS: coarse Sand - Co: Conglomerate

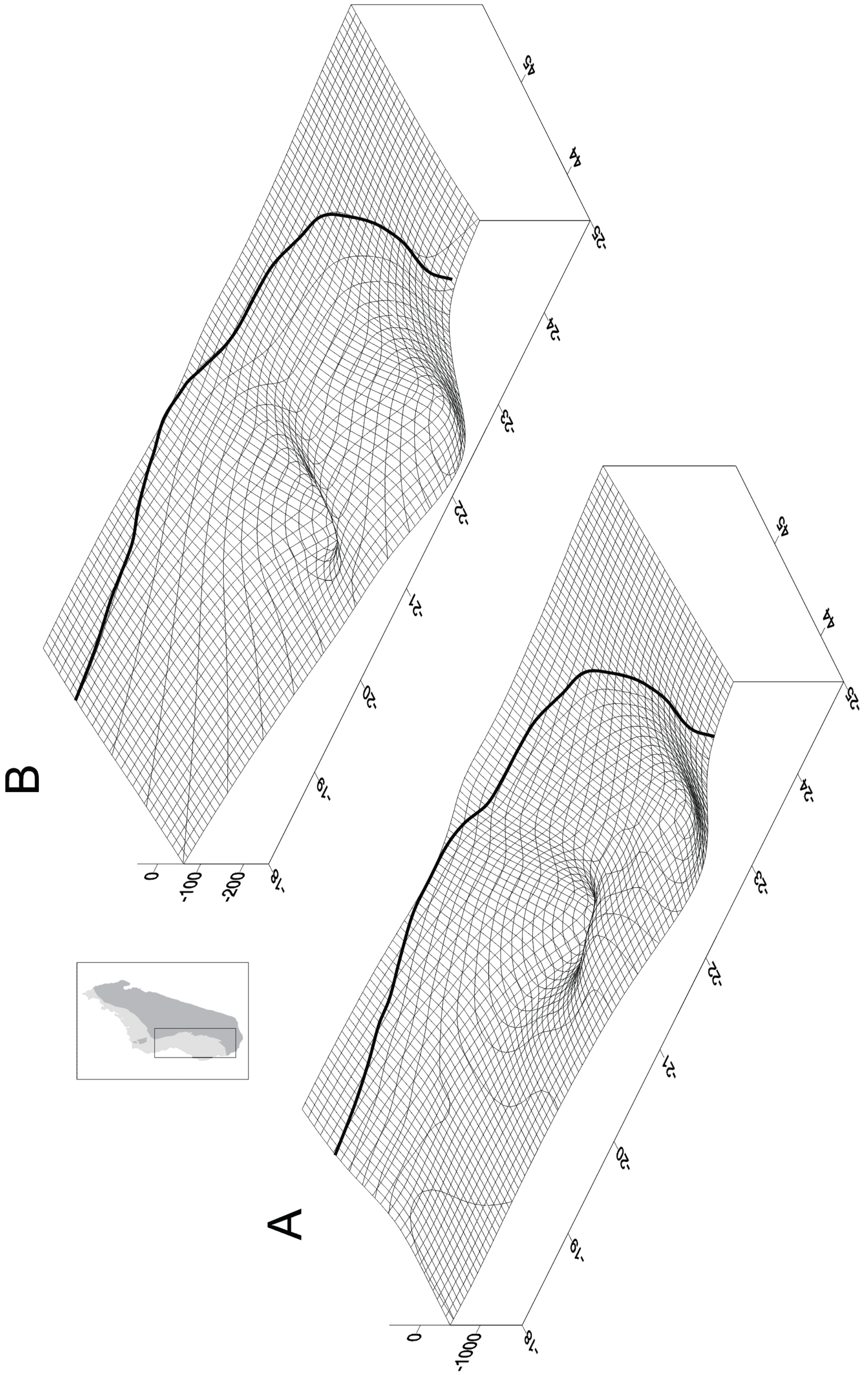
Well and outcrop correlations in the Morondava Basin

N-S Correlation of outcrop and well data of the Morondava Basin illustrate apparent facies contrasts between the northern, central, and southern basin part. This is mainly a result of the lacking facies control across basin strike. The fine siliciclastic successions shown in dark grey are interpreted as a more basinal depositional environment, whereas the light grey depicts rather lagoonal marly shales and marls. **Wells**: BGR: du Toit (1997), Montenat (1996), and Dina (1996) in Uhmann (1996). **Outcrops**: Besairie and Collignon (1972). Well and outcrop locations are presented in Fig. 42:



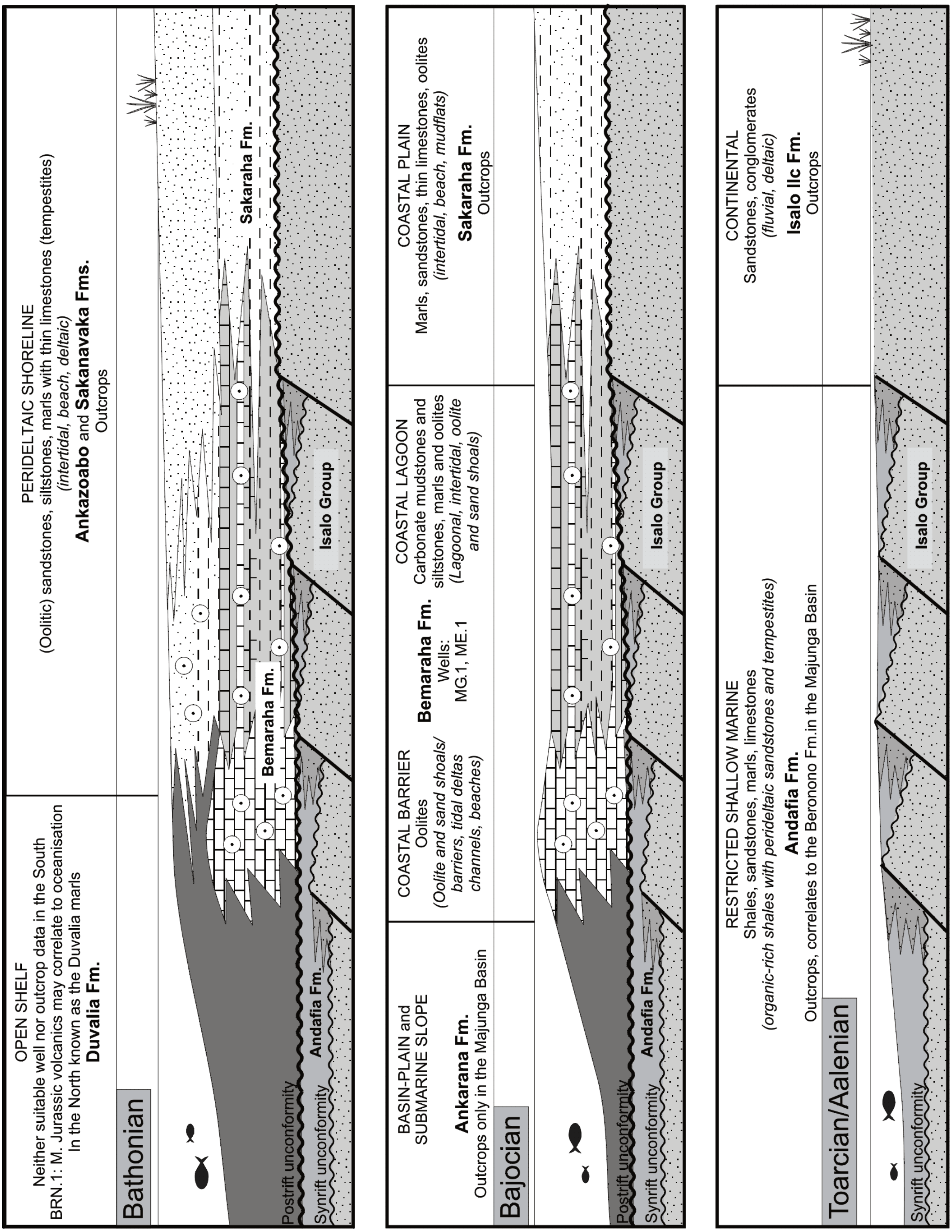
Isopach wireframes of the Morondava Basin

3D visualisation of the isopach data set of the dataset of Fig. 41. A: Bajocian-Bathonian; B: Callovian-Tithonian:



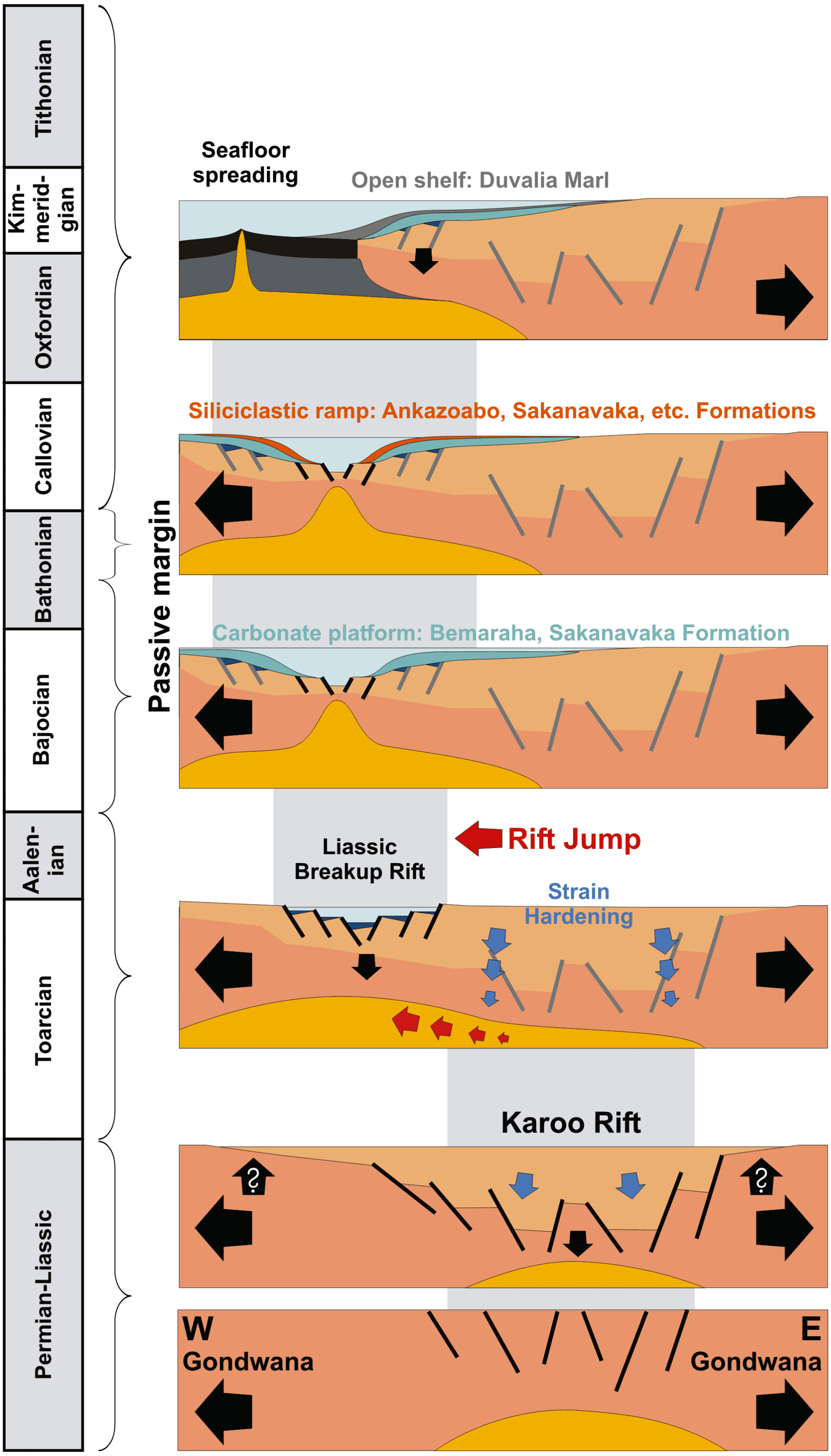
Toarcian-Bathonian facies models

Sedimentary environments of Toarcian-Bathonian strata and their stratigraphic and spatial connection in schematic cross sections across basin strike:



Morondava Basin history and geodynamics

Schematic illustration of the rift stages and the sedimentary response:



Contrasting facies at outcrops in the north and the south

Schematic sections across basin strike depict the facies distribution in comparison to the outcrop appearance at locations in the northern and southern Morondava Basin. A different facies appearance at either part of the basin illustrates the cause of misinterpretation of sedimentological process, when using outcrop data only: